

INQUA
COMMISSION 6

**SECOND INTERNATIONAL
PALEOPEDOLOGY SYMPOSIUM**

FIELD EXCURSION

August 13-20, 1993

An eight-day field trip through the states of
Illinois, Missouri, Kansas, Nebraska, and Iowa
to examine paleosols of various ages and parent materials.

Kansas Geological Survey Open-File Report 93-30

Prepared and compiled by

William C. Johnson

Kansas Geological Survey Open-file Reports

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Introduction To The Excursion

Nature of the Excursion

This field excursion is being held as part of the Second International Paleopedology Symposium, INQUA Commission 6, and is being timed to immediately precede the International Geomorphology Conference being held in Hamilton, Ontario, Canada. In eight days the excursion covers five states and nine or ten physiographic provinces in the central United States. Participants are being exposed to soils ranging in age from Pennsylvanian to Holocene, and developed in a variety of parent materials such as alluvium, till, loess, sand dunes, and marine muds. Also, an interesting diversity of both consolidated and unconsolidated materials is represented.

Organizers and Contributors

Organizers:

Leon R. Follmer, Illinois Geological Survey, Champaign, Illinois 61820

Don L. Johnson, Department of Geography, University of Illinois, Urbana, Illinois 61801

Contributors:

Abbott, Larry R., Department of Geography, University of Illinois, Urbana, Illinois 61801

Arbogast, Alan F., Department of Geography, University of Kansas, Lawrence, Kansas 66045

Balek, Cindy, Department of Geography, University of Illinois, Urbana, Illinois 61801

Bettis, III, E. Arthur, Iowa Department of Natural Resources-Geological Survey Bureau, Iowa City, Iowa 52240

Diekmeyer, Erik C., Department of Geography, University of Kansas, Lawrence, Kansas 66045

Guccione, Margaret J., University of Arkansas, Fayetteville, Arkansas 72701

Hallberg, George R., Iowa Department of Natural Resources-Geological Survey Bureau, Iowa City, Iowa 52242

Joekel, R. Matt, Department of Geology, University of Iowa, Iowa City, Iowa 52242

Johnson, William C., Department of Geography, University of Kansas, Lawrence, Kansas 66045

Kemmis, Timothy J., RUST, Environment and Infrastructure, Sheboygan, Wisconsin 53081

Mandel, Rolfe D., Department of Geography and Geology, University of Nebraska, Omaha, Nebraska 68182

May, David, Department of Geography, University of Northern Iowa, Cedar Falls, Iowa 50614

McCahon, Tom J., Department of Geology, Kansas State University, Manhattan, Kansas 66506

Miller, K.B., Department of Geology, Kansas State University, Manhattan, Kansas 66506

Park, Keyong, Department of Geography, University of Kansas, Lawrence, Kansas 66045

Porter, Donna A., Department of Agronomy, Kansas State University, Manhattan, Kansas 66506

Ransom, Mickey D., Department of Geology, Kansas State University, Manhattan, Kansas 66506

Tandarich, John P., Hey and Associates, Inc., Chicago, Illinois

Woida, Kathy, Department of Geology, University of Iowa, Iowa City, Iowa 52242

Zhaodong, Feng, Lamont-Doherty Geological Observatory, Columbia University, Palisades, New York 10965

Schedule

Assuming that all stays on schedule with bridges being washed out and roads under water or construction, the timing and sequence of the field excursion is to be as indicated on the map and as listed below:

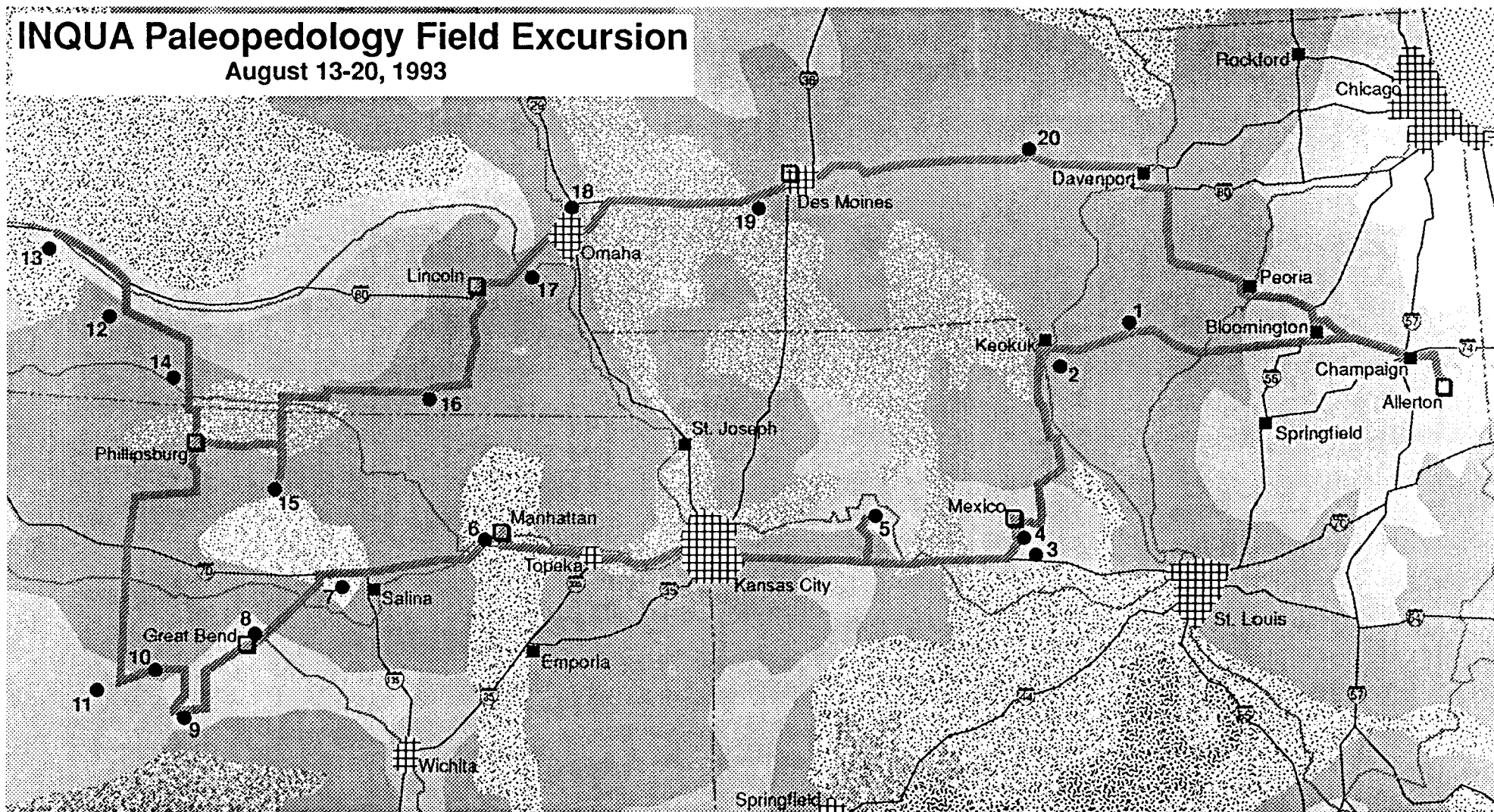
Date	Stop	number and name/location	What to see
13 August/Friday	1	near Macomb, Illinois	Sangamon soil catena
	2	Rocky Run archaeological site overnight in Mexico, Missouri	Holocene soils in loess
14 August/Saturday	3	Winsor Section	welded soils in an alluvial fan
	4	Harrison Clay Pit	Sangamon and Huntsville soils
	5	Franklin Quarry	Farmdale, Sangamon, and Yarmouth soils
		Barbeque on a Flint Hills ranch (evening meal) overnight in Manhattan, Kansas	
15 August/Sunday	6	near Manhattan, Kansas	paleosols in the lower Permian
	7	near Salina, Kansas	hydromorphic Cretaceous paleosol
	8	Barton County Landfill overnight in Great Bend, Kansas	400k yr loess-soil sequence
16 August/Monday	9	Wilson Site	sand-loess paleosol sequence
	10	Pawnee River nr Burdett, KS	Holocene alluvial soils
	11	Pawnee River nr Jetmore, KS overnight in Phillipsburg, Kansas	Holocene alluvial soils
17 August/Tuesday	12	Eustis Ash Pit	volcanic ash-loess paleosols
	13	Bignell Hill	Brady soils and Bignell loess
	14	Schoenenburg Site/Harlan Co. L. overnight in Phillipsburg, Kansas	late Holocene alluvial soils
18 August/Wednesday	15	Lee Site/Lake Waconda	late-Pleistocene alluvial soil
	16	Endicott Site overnight in Lincoln, Nebraska	plinthic Cretaceous paleosols
19 August/Thursday	17	South Bend Site	vertic Pennsylvanian paleosol
	18	Loveland Paratype Section	loess-paleosol sequence
	19	Earlham Quarry overnight in Des Moines	Farmdale, Sangamon, and Yarmouth soils
20 August/Friday	20	Conklin Quarry	Sangamon soil in diamicton

Acknowledgements

Compilation of this guide book would have been extremely difficult or impossible had it not been for the timely submission of field trip stop information by the many contributors. Also, it is often easy to overlook the many landowners who have graciously provided access to their property for pedestrian, backhoe, truck, and other forms of traffic. Erik Diekmeyer and Marla Kerr provided the necessary last-minute efforts to assemble the guide book. Lastly, Beverly Roberts made this publication possible through many long hours of word processing.

INQUA Paleopedology Field Excursion

August 13-20, 1993



University of Kansas Map Associates (George F. McCleary, Jr.), 1993

□ - Overnight Stops

● - Tour Stops

- 1 - Macomb (Sangamon soil catena)
- 2 - Rock Run (Holocene paleosol in loess)
- 3 - Winsor Section (welded soils in an alluvial fan)
- 4 - Harrison Clay Pit (Sangamon and Huntsville soils)
- 5 - Franklin Quarry (Farmdale, Sangamon and Yarmouth soils)
- 6 - Manhattan (paleosols in the lower Permian)
- 7 - Salina (hydromorphic Cretaceous paleosol)
- 8 - Barton County Landfill (400k loess-soil sequence)
- 9 - Wilson (sand-loess paleosol sequence)
- 10 - Pawnee River (Holocene alluvial soils)
- 11 - Pawnee River (Holocene alluvial soils)
- 12 - Eustis Ash Pit (volcanic ash-loess-paleosol sequence)
- 13 - Bignell Hill (Brady geosol and Bignell loess)
- 14 - Harlan County Lake (paleosols in alluvium)
- 15 - Lee Site (Late Pleistocene paleosols)
- 16 - Endicott (plinthic Cretaceous paleosols)
- 17 - South Bend (vertic Pennsylvanian paleosols)
- 18 - Lovelard Paratype (Sangamon soil)
- 19 - Earlham Quarry (Farmdale, Sangamon and Yarmouth soils)
- 20 - Conklin Quarry (Sangamon soil in diamicton)

0 50 100 200 300 Kilometers
0 50 100 200 Miles

Land-Surface Form (after Hammond, 1963)

- Flat Plains (>80 percent gently sloping, 0-100 feet local relief)
- Smooth Plains (>80 percent gently sloping, 100-300 feet local relief)
- Irregular Plains (50-80 percent gently sloping, 100-300 feet local relief)
- Tablelands (50-80 percent gently sloping, 300-500 feet local relief, >50 percent of gentle slope on upland)
- Plains with Hills (50-80 percent gently sloping, 300-500 feet local relief)
- Open Low Hills (20-50 percent gently sloping, 100-300 feet local relief)
- Open Hills (20-50 percent gently sloping, 300-500 feet local relief)
- High Hills (less than 20 percent gently sloping, 500-1000 feet local relief)

Stop 1: U.S. Rt. 67, approximately 4 miles north of Macomb, Illinois: Sangamon Soil in Illinoian till

Location: NW 1/4, Sec. 18, T6N, R2W, McDonough County, Illinois
(USGS Good Hope, IL 7.5 minute quadrangle)

Contributor: Cindy Balek

SANGAMON SOIL BIOHORIZONATION VS GEOGENESIS

The Sangamon Soil is a buried, widely occurring midcontinent paleosol that formed during a significant interval of subaerial weathering and pedogenesis that separates Illinoian from Wisconsinan glaciations (Leverett, 1899; Follmer, 1978). The Sangamon Soil in western Illinois, where developed on better-drained sites, typically exhibits a strongly-developed, texture contrast profile that commonly possesses a stone line marking the boundary between the coarse-textured upper solum and the fine-textured subsoil. Generally, the Sangamon Soil in Illinois was thought to represent a composite soil formed in different parent materials (especially where a stone line was present). According to conventional wisdom, cooling climatic conditions associated with the advent of classic early Wisconsinan glaciation caused the Sangamon landscape to become unstable, resulting in erosion of the Sangamon Soil upper solum and formation of a surface lag marking the top of the truncated subsoil; burial of the surface lag beneath an aeolian-enriched colluvium occurred in conjunction with erosion (Frye et al., 1974) (Fig. 1). Following the classic early Wisconsinan erosional-depositional event, the landscape was said to have stabilized due to climatic amelioration during a purported classic mid-Wisconsinan intraglacial, at which time pedogenic processes produced upper solum characteristics in the aeolian-enriched colluvium and B1 horizon characteristics in the underlying stone line (which overlies the assumed to be truncated 2B2 horizon).

Because of the lateral continuity of Sangamon Soil profile characteristics (i.e., texturally contrasting horizons separated by a stone line) across almost all topographic positions (e.g. flat upland, convex hillcrest, sideslope, and footslope) and because of problems with the inferred timing of climatic-geomorphic events (Baker et al., 1989; Curry, 1989) (it is conceivable that Sangamon Soil development in western Illinois could have spanned the entire post-early Illinoian to late Wisconsinan periods, i.e., ~250ka to 25ka) (Richmond and Fullerton, 1986, Chart 1)), the erosion-deposition model of texture contrast-stone line soil formation for the Sangamon Soil was reevaluated. Based on conceptual models developed by Johnson and Watson-Stegner (1987) and Johnson et al (1987) and on studies of the effects of soil fauna in sorting materials in the tropics and elsewhere (Nye, 1954; Moeyersons, 1978; Lyford, 1963; Darwin, 1881; Mitchell, 1988), it was hypothesized that the texturally contrasting Sangamon Soil horizons, including its stone line, formed due to "normal" pedogenic processes, conditions, and factors in which bioturbation played a significant role, as did "normal" erosion (downwasting) and slopewash accretion (upbuilding).

According to the biopedogeomorphic model developed for this study, Sangamon Soil development in western Illinois presumably began immediately following local deglaciation of the Illinoian ice, at which time the newly-exposed landscape was colonized by plants and animals (Fig. 2a). Gradually, through pedogenesis, which includes the sorting activity of certain mound-building and nonmound-building soil fauna, the originally unsorted parent material (glacial till) began to develop an ordered profile. Fine particles were brought to the surface in the form of burrow mounds resulting in the slow burial of larger

clasts, which eventually would have approached the depth of major biologic activity (Fig. 2b). During periods of rain and high winds, the fines would be removed from exposed mounds; this reworking, together with clay eluviation, would gradually coarsen the texture of the surface horizon relative to the subsoil. Eventually, as it does in some modern soils, this process would produce a subsurface stone line below a sandy silt-textured surface biomantle (i.e., a horizon formed under the dominant influence of biologic activity).

Rainwash and mass wasting processes would cause downslope movement of the biologically-sorted mantle. Translocation of eroded summit materials onto lower sideslopes and footslopes would result in upbuilding at these positions. Because of accretion, the biomantle on the lower sideslopes and footslopes would be thicker than on the crest and upper sideslopes (Fig. 2c). This process of vertical and lateral translocation of Sangamon Soil materials is interpreted to have continued almost uninterruptedly (although the intensity of the rates may have varied with time) until soil development ceased when it became buried by late Wisconsinan (O-18 stage 2) Peoria Loess, about 25ka (McKay, 1979). Cessation of most Peoria Loess deposition, about 13ka (McKay, 1979), led to the formation of the modern soil solum and biomantle (Fig. 2d).

Sangamon Soil Catena with Texture Contrast Horizons and Stone Line

Samples from a Sangamon Soil catena developed on an Illinoian drumlin were collected from an upland boring and back-hoe pits excavated into the east-facing road cut on the west side of U.S. Rt. 67 (Fig. 3). The modern soil (Typic Argiudoll) is developed in the upper 1.5m (5ft) of late Wisconsinan Peoria Loess, which ranges in thickness from 2m (6.5ft) on the drumlin crest to 2.5m (8ft) on the footslope. This downslope thickening of the Peoria Loess on the drumlin is attributed to its being associated with a closed drainage system (where materials are not removed from the system). A distinct break in texture, color, and biofabric separates the C horizon of the Peoria Loess from the underlying 2A horizon of the Sangamon Soil; the Sangamon Soil is draped over the crest and sideslopes of the drumlin. A prominent stone line occurs in the upper part of the Sangamon 2B horizon; the stone line is most distinct on the crest and upper sideslopes and becomes less distinct downslope such that on the north footslope, it was detected only after scraping the back-hoe pit wall with a trowel. The stone line along the north footslope is partly composed of manganese and iron nodules and concretions. The Sangamon 2B horizon is rubified and is developed in Illinoian Glasford Formation till. The 2B horizon is strongly-developed on the sideslopes and crest and more weakly-developed on the footslope. The fact that the Sangamon profile on the footslope is horizonated suggests that vertical pedogenic processes (e.g. eluviation-illuviation) were at least as important as hillslope processes in the formation of the Sangamon Soil; this observation does not conform to the interpretations of other studies in which hillslope processes are thought to be more important than in situ processes for catenas associated with closed drainage systems (Walker and Ruhe, 1968; Kleiss, 1970; Malo et al., 1974).

Morphologically, all Sangamon Soil profiles, regardless of topographic position, generally possess a "complete" sequence of major diagnostic horizons (e.g. 2A, 2E, 2B, 2C) and transitional, or mixed, horizons (e.g. C2A, 2AE, 2EB, 2BE/2B) that display morphologic continuity. Similar observations regarding morphologic continuity of upper solum and subsoil horizons were made by Ruhe (1956b) for the late Sangamon Soil profile in Iowa. Although Ruhe acknowledged that the morphologic continuity of the late Sangamon profile suggested that the horizons were genetically related (as did weathering ratios), the presence of a stone line at the upper solum-subsoil contact was "proof" to him that a plane of truncation separated different-aged materials, and that subsequent pedogenesis produced a welded, or composite, soil. However, those Sangamon Soil catena members developed on sloping, "erosion-prone" surfaces possess upper solum horizons that are, morphologically-speaking, as well-developed as the profiles on the stable flat upland surfaces. Additionally, the extensive mixing of horizons at the upper solum-stone line-subsoil

contact is not consistent with Folster's (1969) contention that allochthonous sediments possess sharp boundaries with the underlying material. However, the mixing of horizons would be consistent for bioturbated soils as soil fauna tend to obliterate horizon boundaries, causing B horizon material to become mixed with A and C horizon materials (Mermut, 1985).

In addition to having similar morphologic characteristics displaying a similar degree of development, the catena members developed on sloping surfaces also possess depth distribution functions of individual sand fractions that are similar to the depth distribution functions found in the stable, flat upland profile (Fig. 4). In all profiles, the very fine and fine sand fractions, especially, maintain relatively consistent profiles from the top of the Sangamon 2A horizon, through the 2E horizon-stone line-subsoil transition, and down into the underlying 2C horizon. If the morphologic continuity and constant sand distribution profiles observed in the Sangamon 2A-2E horizons-stone line-B horizon sequence can be used as evidence against an erosional-depositional mode of texture contrast-stone line soil formation, then it suggests that the Sangamon soils were not subjected to an erosional episode that resulted in stripping of the upper solum and the formation of a surface lag. This would be especially true for those profiles where the sand distribution profiles remain consistent up through the Sangamon profile and into the lower part of the Peoria Loess (which most likely is due to mixing by in situ biologic and/or pedologic processes).

The Sangamon upper solum in all profiles, whether developed on a stable landscape position, erosion-prone sideslope or on a depositional footslope surface, exhibits a similar grading (fining-up sequence) of the medium sand and larger fractions (Fig. 4). The base of the graded sequence is marked by the stone line horizon, where a >2mm-fraction bulge occurs (the >2mm-fraction bulge can be directly correlated to the stone line in those profiles where the stone line was exposed, but can only be inferred to correlate to a stone line in the core-derived flat upland profile). The >2mm-fraction bulge fines up to a very coarse sand bulge in the overlying horizon; the very coarse sand bulge, in turn, fines up to a high coarse sand content in the superjacent horizon. The fact that the flat upland profiles exhibit the same graded sequence as the soils developed on the slopes suggests that the same sorting mechanism, whether it be pedogenic, biologic, and/or geogenic in origin, operated on all landscape positions. Graded soil materials underlain by a stone line are often used as evidence in support of a fluvial erosion-deposition origin for texture contrast-stone line soils (Rohdenburg, 1969; De Dapper, 1989; Embrechts and De Dapper, 1989). While an erosional-depositional origin for the Sangamon Soil texture contrast-stone line profile may be feasible for those soils developed on sloping surfaces, it seems an unlikely process for those soils developed on stable flat upland sites (even Ruhe (1956a, 1956b), who developed the erosion-deposition model of stone line development acknowledged that soils on flat uplands remain unmodified by erosion until the backwasting hillslopes coalesce, at which time the soil would be truncated).

Based on morphologic continuity, depth distribution profiles of individual sand fractions, and the graded sequence in the upper solum, it is concluded that the Sangamon upper solum, on all landscape positions, is developed in the same parent material as the underlying 2B horizon; as such, the stone line and graded sequence formed due to some horizontal but largely vertical, in situ biopedologic processes. Numerous studies have shown that the sorting activity of soil fauna can produce graded soil profiles (Shaler, 1888, 1891; Atkinson, 1957; Lee and Wood, 1971; Akamigbo, 1984; Goudie, 1988; Ponomarenko, 1988; Johnson, 1990; Paton et al., unpublished manuscript). The Sangamon Soil shows much evidence of former biologic activity in the form of a well-developed upper solum granular structure composed mainly of fecal pellets, in clustering of sand grains along channels, chambers, and in the soil matrix (possibly indicating intense earthworm activity (Fitzpatrick, 1984)), and in the presence of numerous discrete fecal pellets of many different shapes and sizes, suggesting that a variety of soil fauna inhabited the Sangamon Soil.

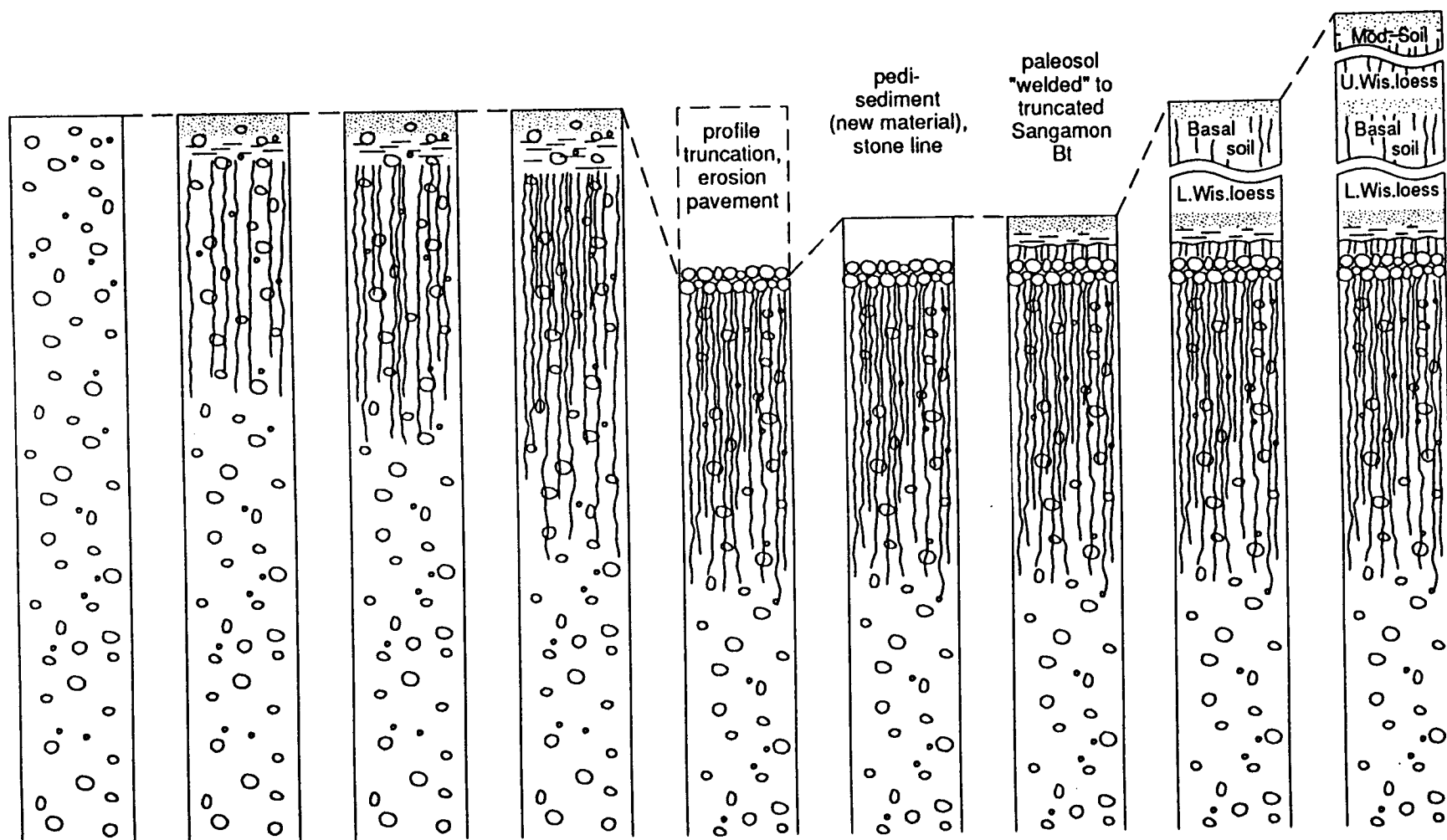


FIGURE 1-1. Diagram showing erosion-deposition origin of texture contrast-stone line soils.

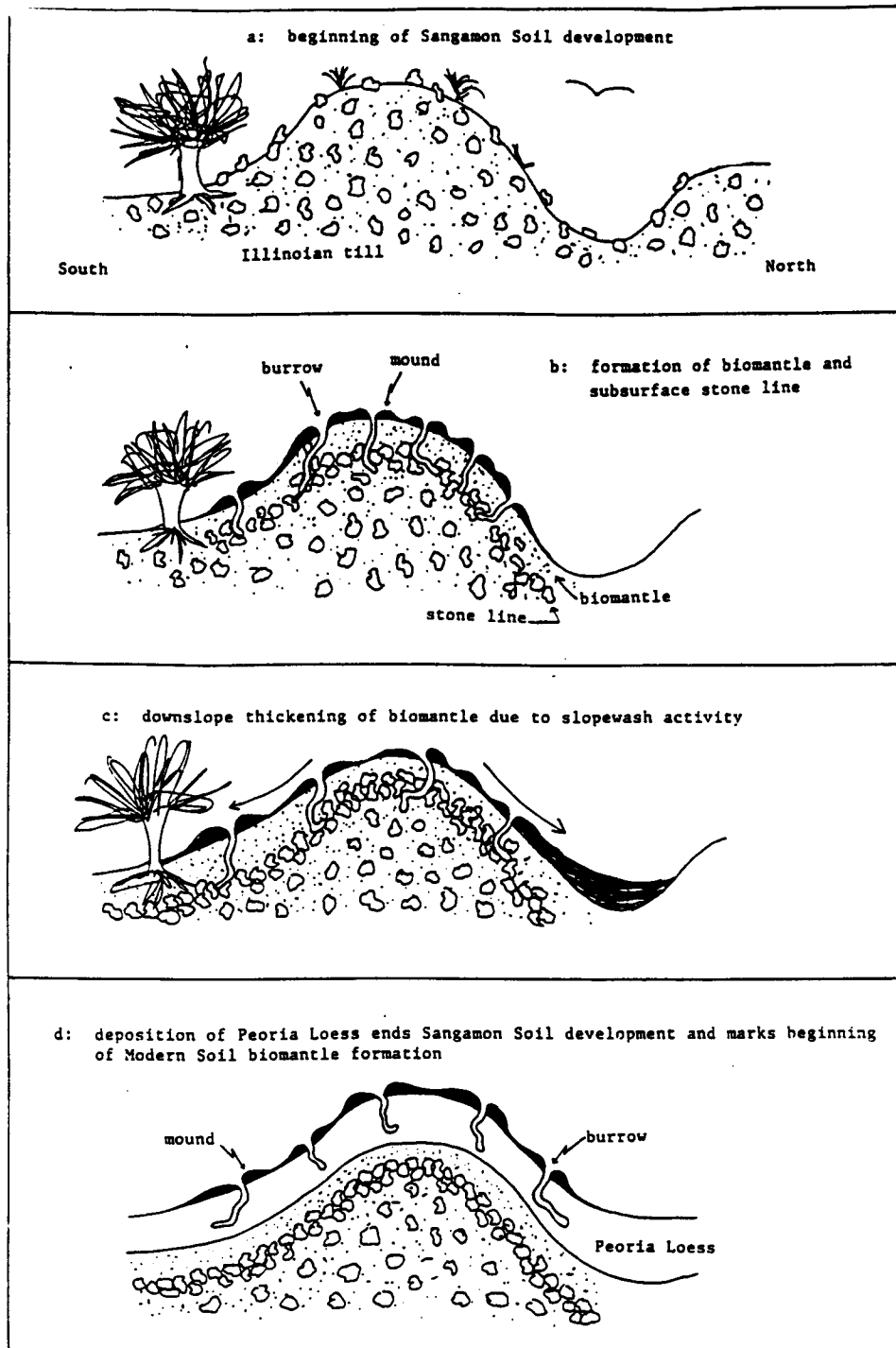
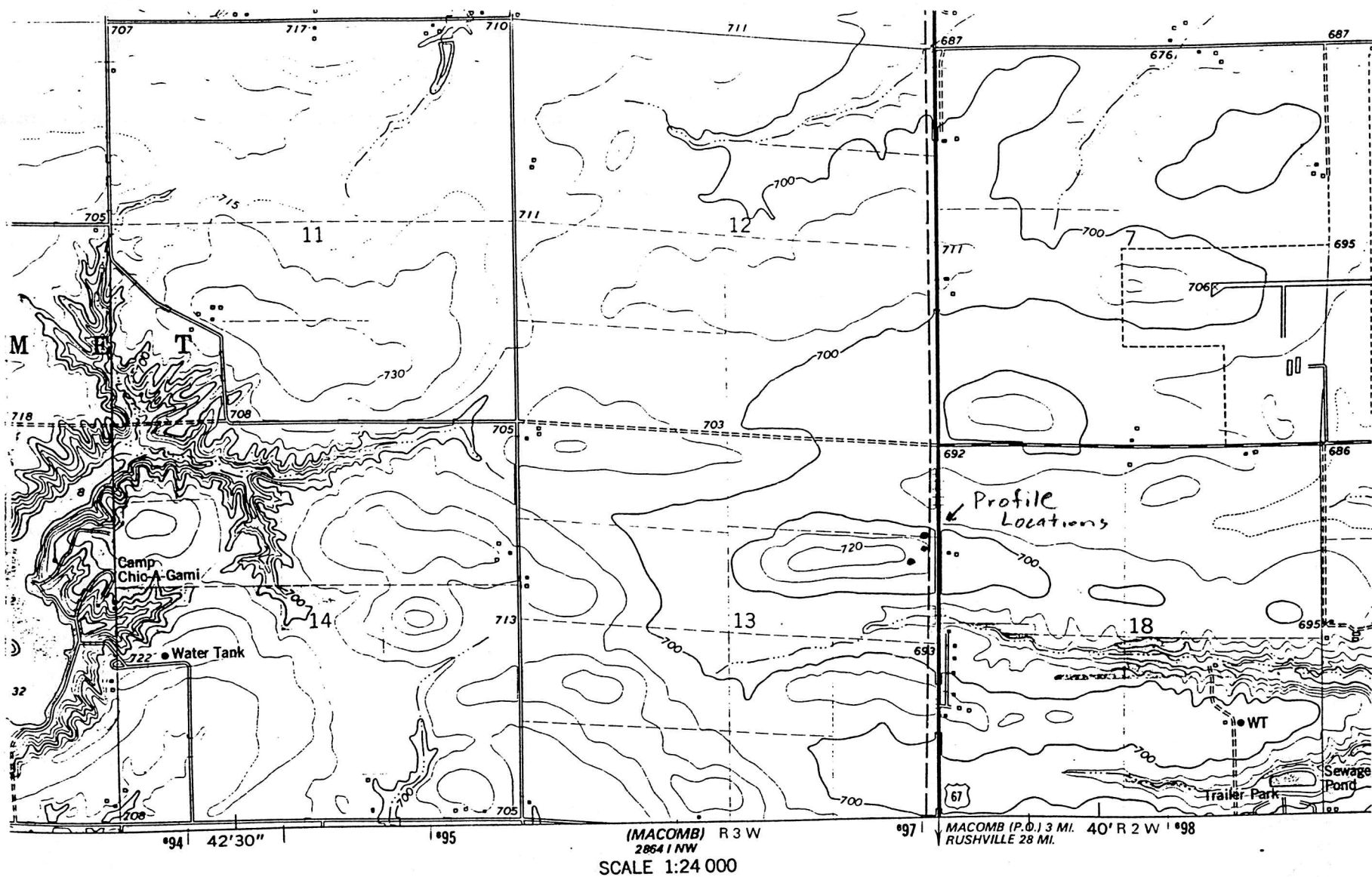
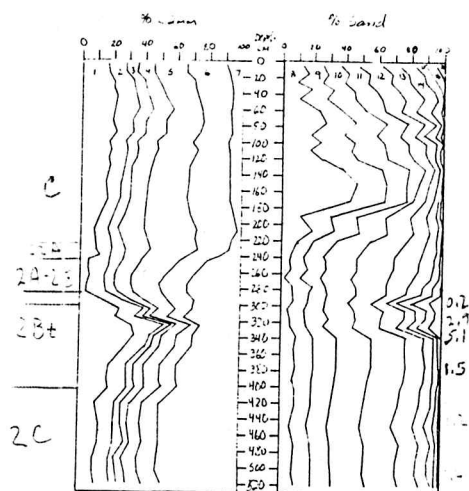
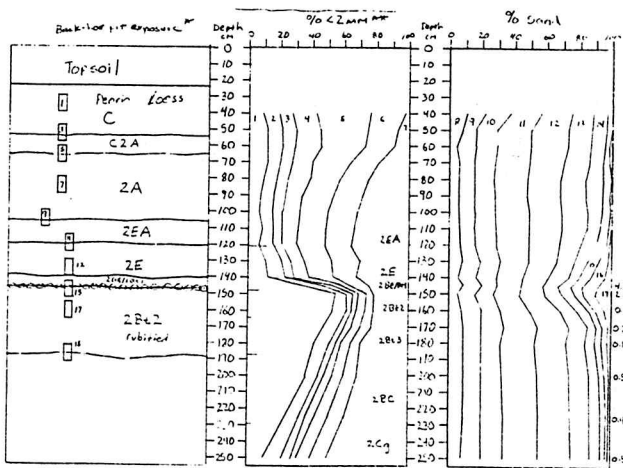


FIGURE 1-2. Diagram showing biopedogeomorphic origin of texture contrast-stone line soils.

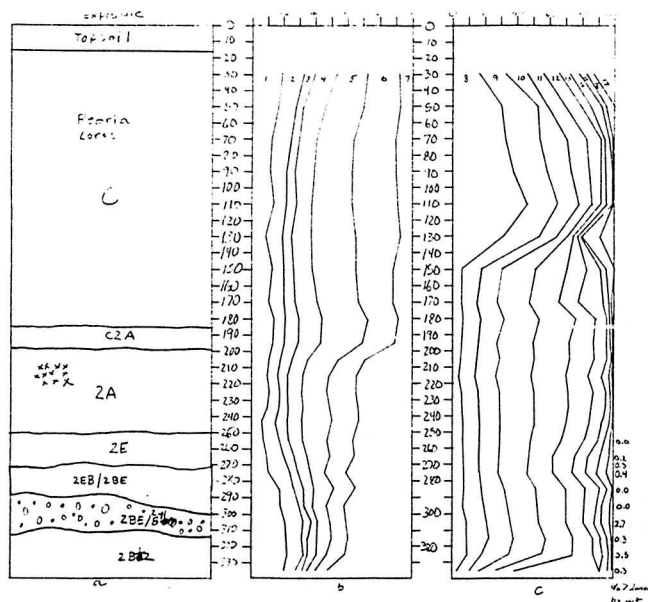




Upland core



Sideslope



Footslope

- 1=fine clay
- 2=coarse clay
- 3=fine silt
- 4=med. silt
- 5=med. silt
- 6=coarse silt
- 7=total sand
- 8&9=very fine sand
- 10&11=fine sand
- 12&13=med. sand
- 14&15=coarse sand
- 16&17=very coarse sand

FIGURE 1-4. Particle size distribution of profiles in Sangamon Soil drumlin catena.

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Stop 2 Rocky Run Archaeological Site, 3/4 mile west of Rocky Run School: Holocene paleosol in loess

Location: N 1/2 of SE 1/4, and S 1/2 of NE 1/4, Sec. 10, R9W, T3N, Hancock County, Illinois
(USGS Warsaw, IL 7.5 minute quadrangle)

Contributor: Larry R. Abbott

HOLOCENE LOESS DEPOSITION AND PEDOGENESIS

Small, usually subtle, dune-like ridges commonly form the bluff edge of major river valleys in the upper Midwest and northern Plains. The genesis of these geomorphic features has been investigated along the upper Missouri Valley but not farther east. In the upper Missouri Valley these features, are called "cliff dunes." Cliff dunes formed during the Holocene by the preferential accumulation of loess at the bluff edge. The loess was derived primarily from the adjacent poorly vegetated bluff face. Loess accumulation is also responsible for the burial of prehistoric occupations on those landforms. Archaeological sites are also buried in the upland soils along and near the bluffs of other major river valleys in the upper Midwest, including the Mississippi. The Rocky Run archaeological site, situated along the Mississippi Valley, has a landscape containing a cliff dune-like landform and has buried archaeological occupations. Such a landform in Illinois suggests that, locally, loess deposition was an important pedogenic vector. By extension the lesser, downwind additions of dust to the soilscape during the Holocene would also have been an important evolutionary vector. Loess in the upper Midwest is typically considered late Wisconsinan in age and glacial in origin. The Rocky Run site was chosen for my thesis research because it offered an opportunity to assess conventional wisdom, as it pertains to loess stratigraphy, and the potential significance of loess deposition during the Holocene.

The Rocky Run archaeological site (11-Ha-662) occupies the interfluvium between Rocky Run and the Mississippi Valley in the N $\frac{1}{2}$ of the SE $\frac{1}{4}$, and the S $\frac{1}{2}$ of the NE $\frac{1}{4}$, of Section 10, R.9W, T.3N, Hancock County, Illinois (Figure 1). The generally flat upland surface is about 585 ft (178 m) amsl. Corps of Engineer bore logs show that gravel were encountered beneath the loess and overlying the bedrock strath. The character of the upland surface and the apparent stratigraphy suggest that the ridge is a remnant of a high, possibly Illinoian-age, terrace. A crescentic, asymmetrical ridge dominates the northern end of the interfluvium overlooking the confluence of the two valleys. The maximum elevation of the crescentic ridge is about 610 ft (186 m) amsl. The bluff-face soil and the bluff-edge soil on this ridge is mapped as Dickinson-Hamburg complex while the soil on the flatter portion of the interfluvium is mapped as Seaton silt loam.

The site area consists of four subareas based on the modern landscape. From north to south these are: (a) bluff edge woods and pasture on the crescentic ridge and part of the flatter interfluvium; (b) bluff edge woods and an eroded grain (oats) field on the relative flat interfluvium; (c) the spillway; and (d) a disturbed and eroded grain (oats) field on the flat interfluvium. The southernmost portion contributed little to the total investigation and is neither shown on the remaining maps nor discussed further. Both oat fields and the pasture were mowed during the summer of 1992. Gullying and piping of the loess is a significant problem along both the eastern and western margins of the interfluvium, particularly north of the spillway.

The Rocky Run site first gained geoarchaeological interest about 1970 during construction of a spillway across the interfluvium from a U.S. Army Corps of Engineers constructed lake on Rocky Run to the Mississippi Valley. In a geoarchaeological investigation conducted at this time Shannon (1972) concluded that a unit of Holocene loess had been deposited over the late Holocene (Woodland) archaeological deposits found along the spillway. No subsequent archaeological or geological investigations were done at the site until the work described here which began in 1992.

Hypothesis

The present investigation at the Rocky Run site was conducted to test the hypothesis that eolian deposition of silt is responsible for burial of the archaeological deposits observed by the previous researchers. The modified loess hypothesis is that local, rapid Holocene loess deposition: (a) was the dominant geopedogenic process responsible for the burial of prehistoric cultural component; (b) contributed to morphological modification of the local upland landscape; and (c) was a significant pedogenic factor of the local upland soilscape. The loess hypothesis is graphically modeled in Figure 2. An alternate hypothesis is that the crescentic ridge has experienced erosion that supplied the sediment that buried the archaeological component.

Procedures

The present field investigation focused on defining the spatial distribution and stratigraphy of the cultural deposits, and the description and character of the geological deposits and associated soils of the interfluvium and the crescentic ridge. This work included collecting soil samples from representative profiles for laboratory analyses.

The site area was hand augered on a Cartesian grid followed by auger transects normal to the crescentic ridge (Figure 3). The auger tests (A.T.'s), designated by a star and respective number on Figures 3, 4, and 5, were excavated in controlled, usually 10 cm, levels to depths ranging from two to five meters depending on the modern landscape position. The extracted soil samples from each level were briefly described and then screened through one-quarter inch hardware cloth. Any residuals, such as artifacts, possible gastroliths, and concretions were collected and bagged. The transects of auger tests normal to the crescentic ridge, besides providing the usual geoarchaeological and pedological information, were excavated for collecting soil profile samples for later laboratory analyses.

Results

Analysis of the temporal and spatial distribution of cultural material is not yet complete, but some tentative conclusions are suggested here. The cultural material recovered during auger testing suggests the presence of at least three major cultural intervals: Oneota, Late Woodland, and Middle Archaic. The Oneota and Late Woodland assemblages of artifacts, or components, are primarily confined to the Ap horizon over much of the site area. Features related to these components are also present. Five Oneota pit features (A.T.'s 6, 7, 15, 16, and 25; Figures 3, 4, and 5) and an Oneota midden were encountered. The pit features are generally greater than one meter in depth and contain considerable quantities of refuse such as chipped stone, ceramics, and bone. About 40-50 cm of midden survives below the Ap around A.T. 34 and 59 (Figure 4). The midden quickly thins eastward to A.T. 32, northward to A.T. 47, and northwestward to A.T. 45 and 50. The midden probably extended southward but does not survive because of the severely eroded character of the soil in that field.

An apparently Late Woodland mound, encountered in A.T. 28 and 48, still survives along the bluff edge at the northwest corner of the site area (Figure 5). The mound has from one to one and one-half meters

of fill and contains burials. Late Woodland ceramics were associated with the remains of two? individuals. Amateur excavation (potting) has left a large crater in the mound center. The Middle Archaic component is still not well defined but was shown by the recovery of a large, straight stem, projectile about 77 cm below the surface in A.T. 51 (Figure 4). Other deeply buried artifacts in surrounding auger tests, particularly those tests to the north and west, may define the spatial component of the Middle Archaic component.

The soils at the Rocky Run site vary most dramatically from the summit of the crescentic ridge at the northwestern edge of the interfluvium toward the south-southeast (Figure 6). Along the bluff edge at the ridge summit the silt loam soils exhibit weakly developed, Ap-C or A-C profiles. Southward about midslope the profiles gain a color B horizon that gradually strengthens, thins, and becomes deeper with increasing distance. At the footslope the profiles have a minimal textural B horizon. From there south to the spillway the Bt horizon gradually becomes better expressed, more enriched in clay, and closer to the surface. A "gray band" was encountered about two and one half to three meters below the surface in several profiles on and near the summit of the crescentic ridge. The field data suggests that the "gray band" merges with the weak Bt horizon near the footslope of the ridge; ongoing laboratory analyses may clarify this suggestion.

Calcium carbonate concretions occur in the Ap horizon and the remainder of the profile of the soils on the summit of the crescentic ridge. The vertical distribution of the recovered calcium carbonate concretions from these profiles shows multiple distinct zones of abundant concretions separated by zones of far fewer to no concretions (Figure 7). The deeper zones with few carbonate concretions usually contain snail shells. This pattern continues downward through the first two to three meters of these profiles. Below about three meters the frequency of calcium carbonate concretions dwindles but snail shells remain common. The species of snails present in the loess have not yet been determined but do appear to change with depth. The tentative interpretation of the pattern of the carbonate concretions in these profiles is that they represent episodes of rapid local accumulation of loess followed by slower rates of loess accumulation and concomitant carbonate (presumably as bicarbonate) translocation and precipitation as concretions. Calcium carbonate concretions occur rapidly increasing depths toward the southern footslope of the crescentic ridge. Concretions were not found within the surface three meters of those profiles along the toeslope of the crescentic ridge nor beyond on the flatter portion of the interfluvium to the spillway. The exceptions were concretions in and below the thickest portion of the midden and below the two pit features closest to the ridge.

Iron-Manganese concretions were only recovered from the profiles on the lowest and flattest portions of the site area next to the spillway. Pebbles, possibly gastroliths, were rarely found in the soils of the site outside an archaeological context. Most potential gastroliths were recovered from the Ap horizon, the pit features, the midden, the burial mound, or in association with a buried cultural component.

Laboratory analyses of the soil profiles are currently underway. The results of these analyses will provide detailed information supporting clearer and stronger interpretations of the genesis of the soils and their relationship to the archaeological components at the Rocky Run site.

Conclusions

The provisional conclusion is that loess accumulation during the Holocene has occurred at the Rocky Run site. The roughly three-meter accumulation of loess dominated the bluff edge pedogenesis and reshaped the local bluff edge morphology into a crescentic, asymmetrical ridge, a "cliff dune." Episodes of regressive pedogenesis appear to have alternated with periods of progressive pedogenesis resulting in the translocation and precipitation of calcium carbonate concretions into zones.

Beyond the immediate bluff edge loess accumulation appears to have been less significant permitting continual progressive pedogenesis. Loess accumulation during the Holocene seems the primary process responsible for burial of the archaeological occupations. Where the three cultural occupations are present in an unplowed context they are found at progressively greater depths as the bluff edge is approached. There is some evidence too for limited accumulations of slopewash along the footslope of the crescentic ridge. The probable slopewash deposits are associated with the Oneota occupation and may have been a byproduct of that occupation.

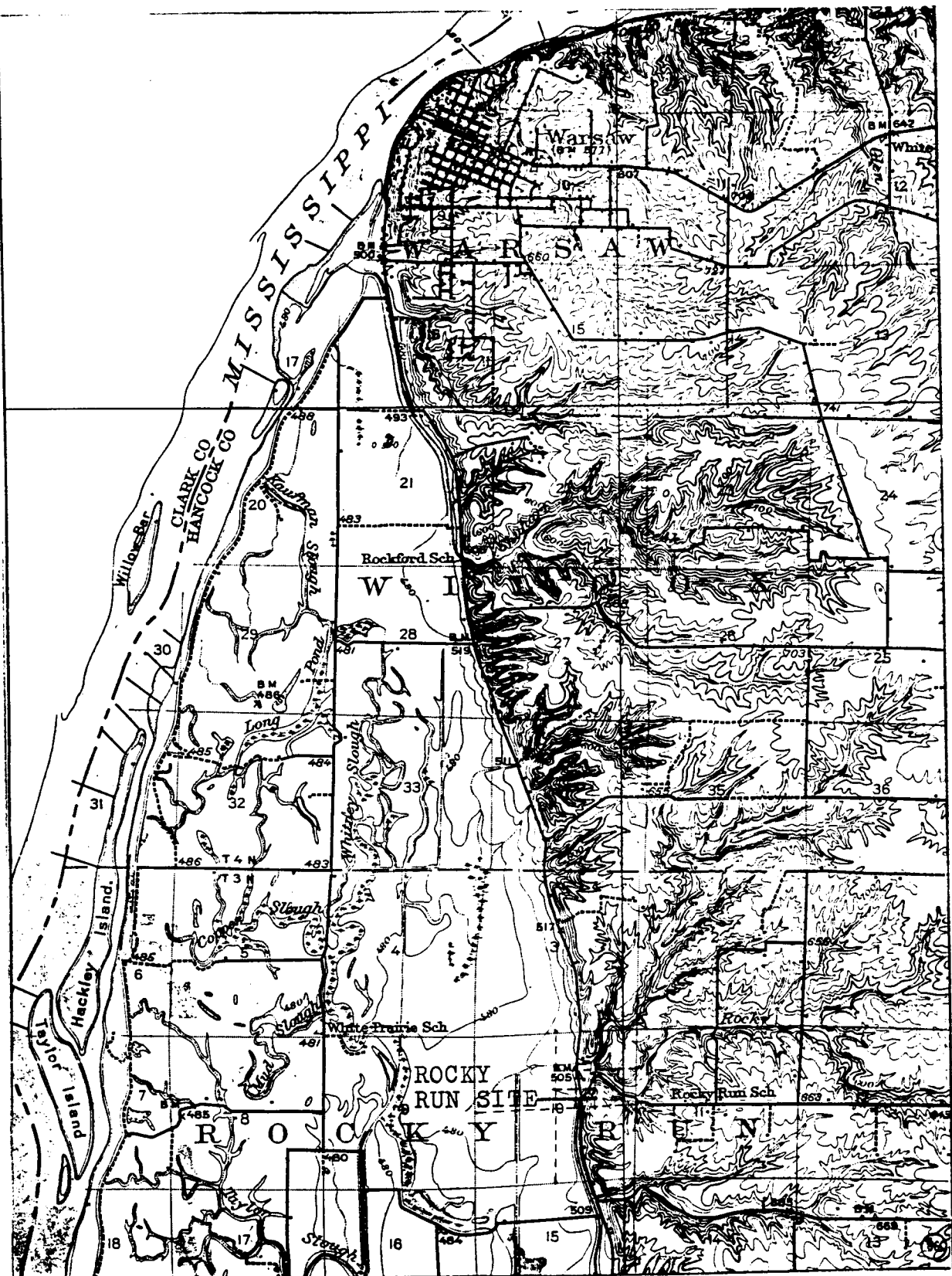


FIGURE 2-1. Locational map of the Rocky Run site.

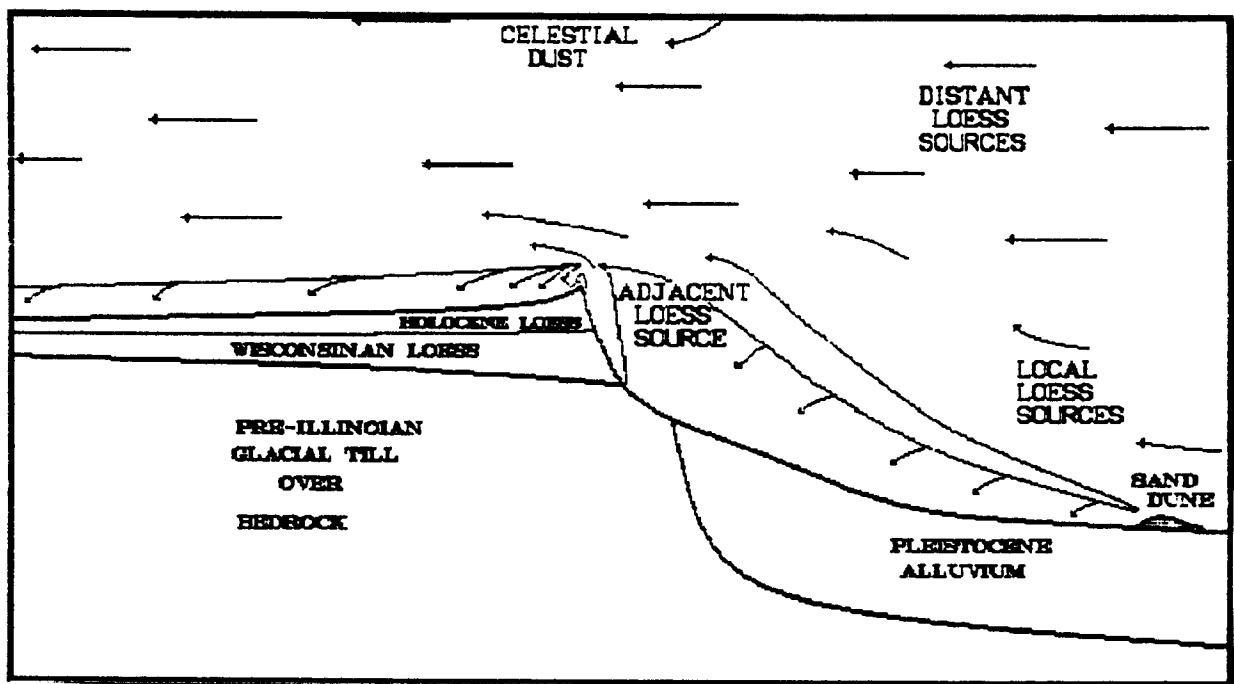


FIGURE 2-2. A model of the modified loess hypothesis showing hypothetical sources and deposition at the Rocky Run site.

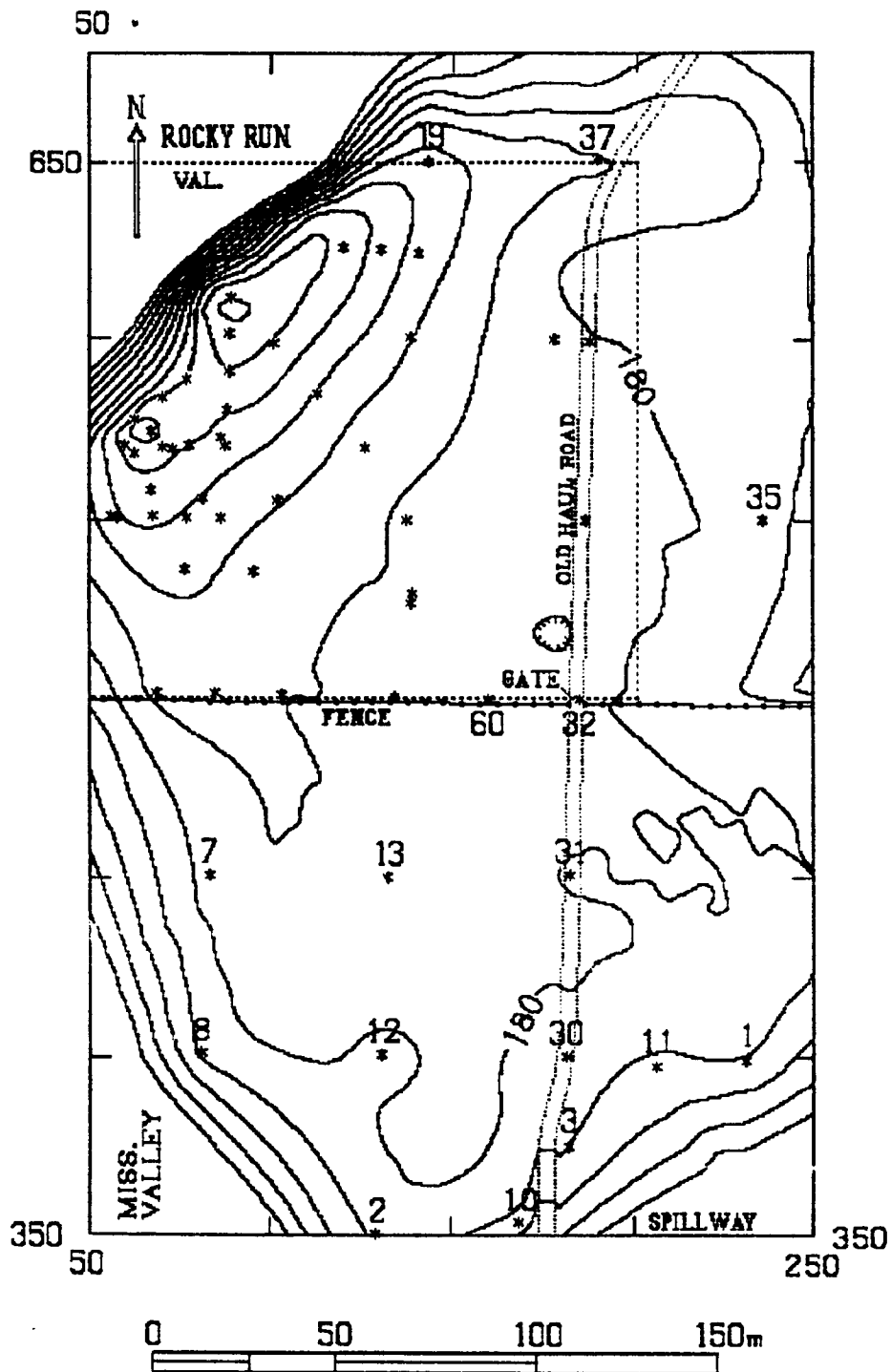


FIGURE 2-3. Rocky Run site. Contour interval: 1 meter.

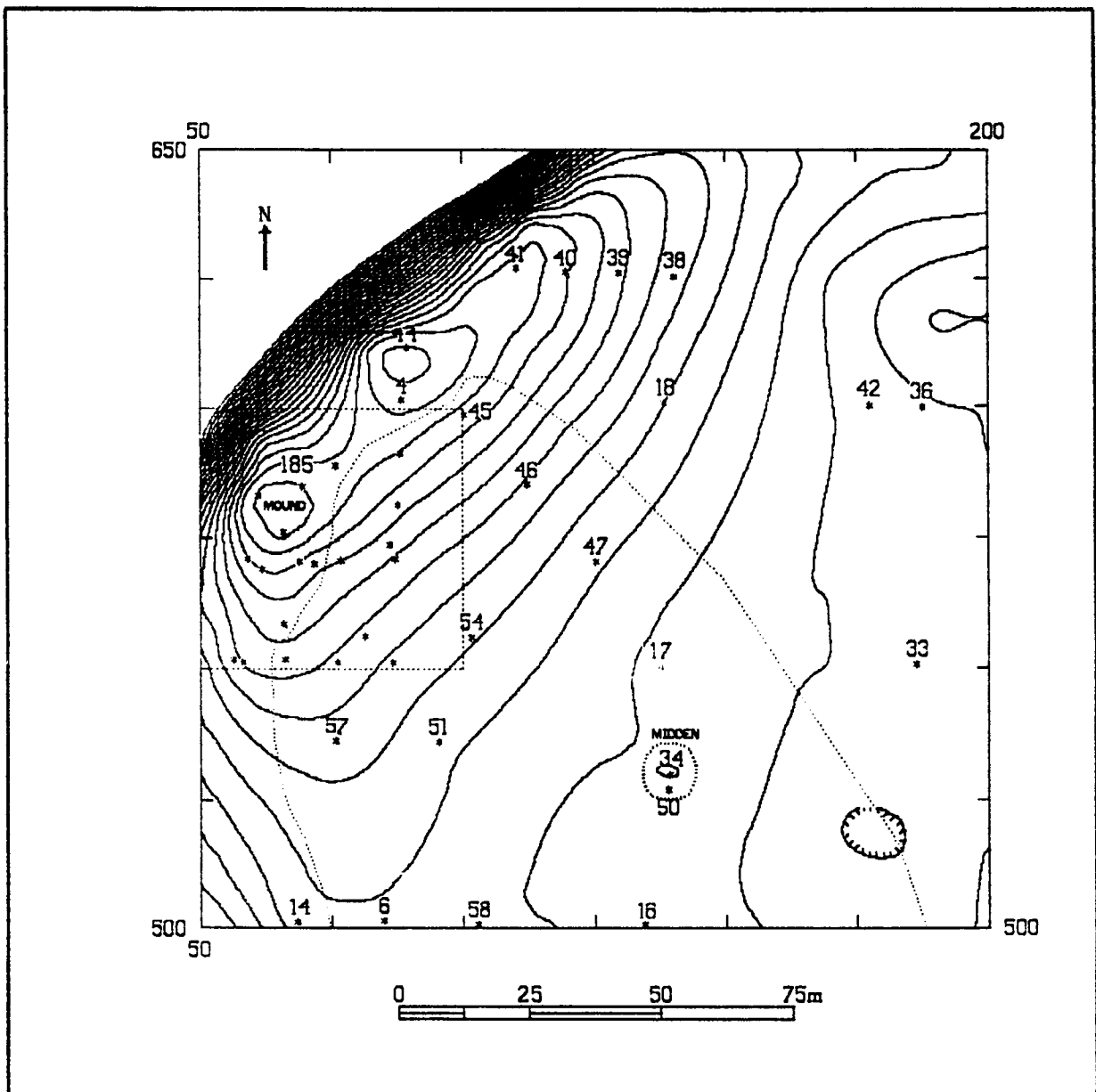


FIGURE 2-4. Northern Rocky Run site. Contour interval: .5 meter.

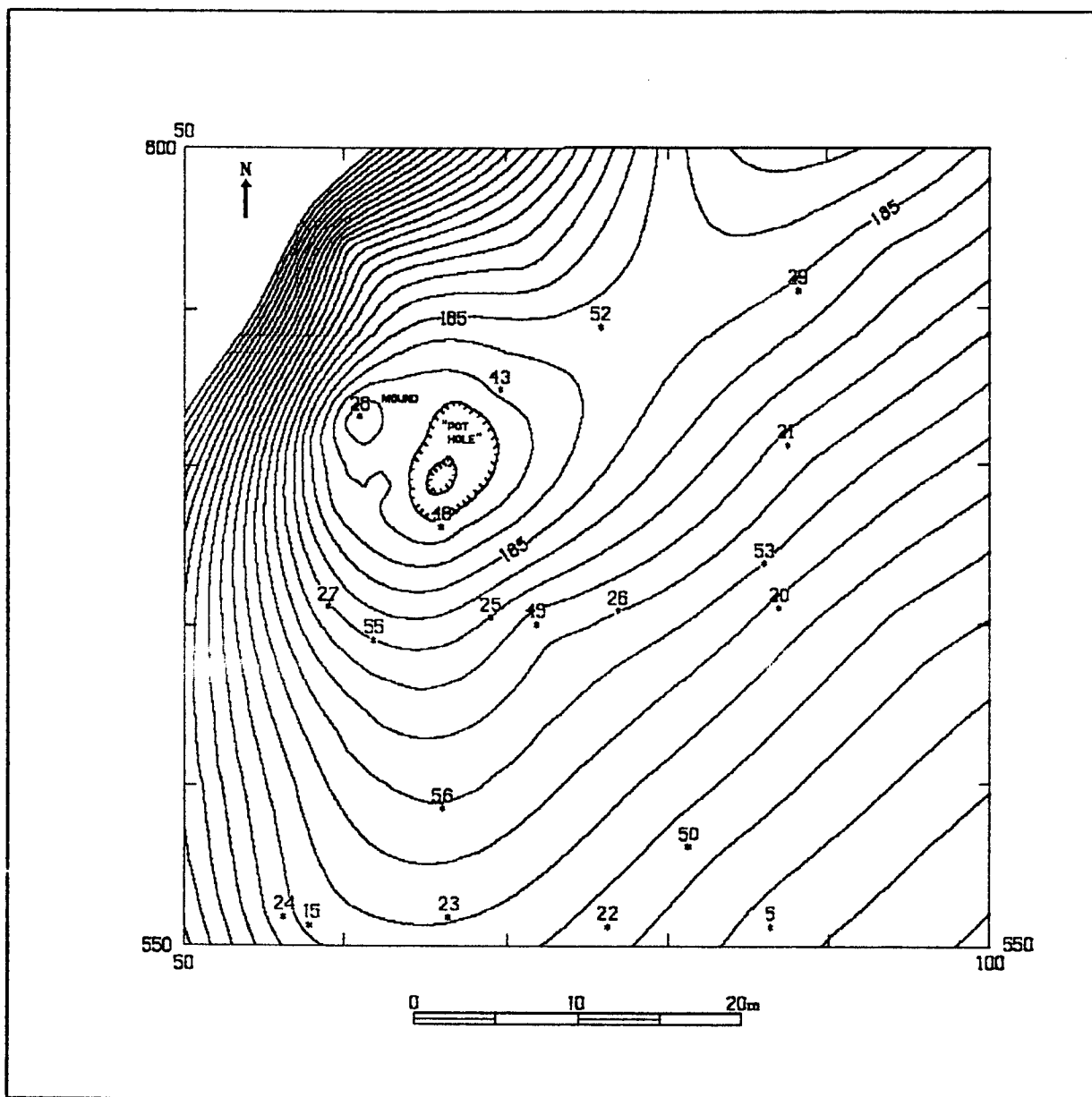


FIGURE 2-5. Rocky Run site mound. Counter interval: .25 meter.

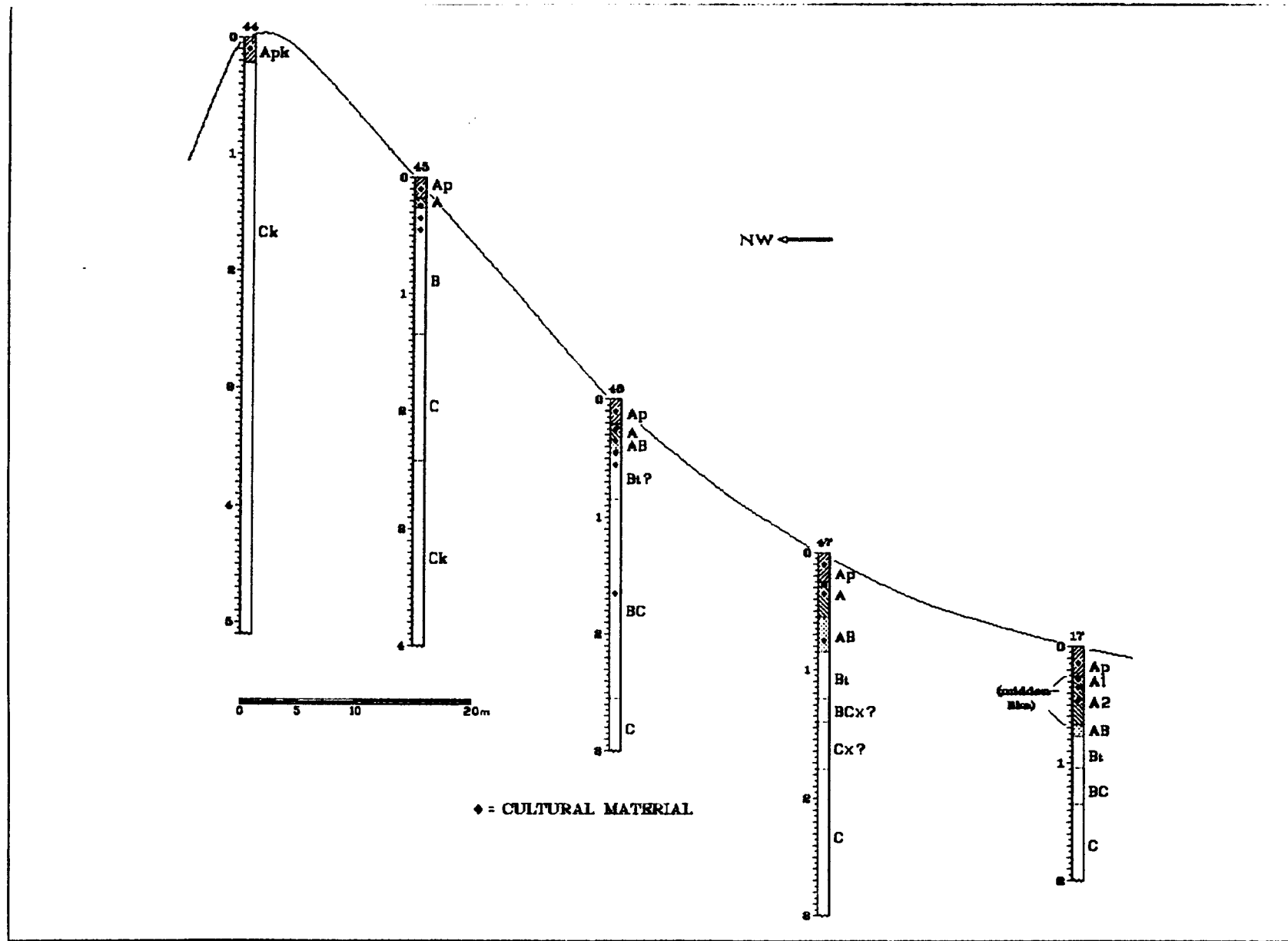


FIGURE 2-6. Rocky Run site soil profile transect.

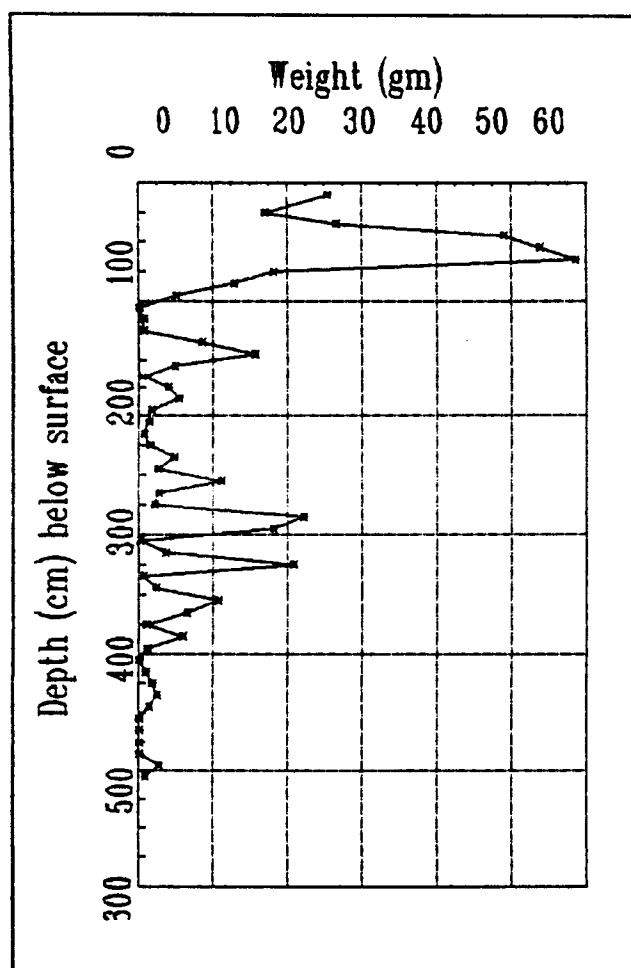


FIGURE 2-7. Vertical distribution of A.T.44 CaCo_3 concretions.

Reference Cited

Shannon, Dennis M., 1972, Possible interpretations of loess deposits in West-Central Illinois: Paper presented to Dr. John J. Alford and Dr. Fred C. Caspall for Paleo-Geography 502I, Fall 1972, copy on file at the Department of Geography Library, Western Illinois University, Macomb.

Stop 3 O.V. Windsor Section (W)

Location: SE 1/4 Sec. 2, T47N, R7W, Callaway County, Missouri
(USGS Readsville and Williamsburg, MO 7.5 minute quadrangles)

Stop 4 Harrison Clay Pit Section (H)

Location: SW 1/4 Sec. 1, T48N, R9W, Callaway County, Missouri
(USGS Calwood and Kingdom City, MO 7.5 minute quadrangles)

Stop 5 Franklin Quarry Section (W)

Location: Sec. 34, T53N, R20W, Saline County, Missouri
(USGS New Frankfort, MO 7.5 minute quadrangle)

Contributors: Margaret J. Guccione and John P. Tandarich

THE QUATERNARY GEOPEDOLOGY OF NORTHCENTRAL MISSOURI

The Quaternary stratigraphy (Table 1) of northcentral Missouri (Fig. 1) includes lower and middle Quaternary alluvium-colluvium and tills and upper Quaternary loess and alluvium-colluvium. Quaternary stratigraphic units for northcentral Missouri proposed are presented in Table 1 and diagrammed by Guccione (1983) in Fig. 2 with some modifications based on results obtained by Tandarich (n.d.). Further Quaternary work in the region is being done by Charles Rovey of Southwest Missouri State University, Springfield, Missouri.

Preglacial

Prior to glaciation, the northcentral Missouri landscape was bedrock-dominated with a veneer of sediments and soils. These bedrock-derived sediments, the Whipporwill Formation, have a limited mineralogy similar to their source materials (shale, sandstone, cherty and chert-free limestone or dolomite) but diverse textures, probably due to a variety of alluvial and colluvial microenvironments. The intensely weathered New Florence soil, which has developed in these materials, indicates that portions of the preglacial landscape were stable for a long period of time preceding glaciation. The unconsolidated materials were eroded in most locations by overriding glaciers but were preserved in isolated sites.

Pre-Illinoian

At least three pre-Illinoian ice sheets advanced from the northwest to northcentral Missouri. The deposits of the oldest of these three ice advances include the proglacial sediments and the loam and silty clay loam diamictos of the Moberly Formation Undifferentiated. An eroded pedo-weathering profile subsolum (C/D) (Tandarich et al., in press), the Huntsville Soil, is present in the uppermost diamicton of the Moberly Formation. Further evidence of erosion of this profile is suggested by the stone line of gravel to boulder-

size materials found at the surface of this profile in some locations, such as the Harrison Clay Pit (Fig. 1). The semi-quantitative clay mineral composition of the Moberly Formation is similar to an unnamed formation seen in southern Iowa by Bettis (personal communication).

The younger two pre-Illinoian ice advances, separated in time by an interglacial, deposited two clay loam diamictons (interpreted to be tills) of the McCredie Formation. The Fulton Till Member is the lower and the Columbia Till Member is the upper of the two till units. Following deglaciation of the area the landscape was modified and the Kingdom City Soil developed in alluvium-colluvium (where present) and in the Fulton till. This soil is moderately developed and occurs in the lower, unstable positions of the landscape, similar to the late Sangamon Soil. Therefore the Kingdom City soil is interpreted as an interglacial soil despite the fact that it appears not to be intensely weathered. The semi-quantitative clay mineral composition of the McCredie Formation tills is similar to the Wolf Creek Formation of southeastern Iowa (H.D. Glass, personal communication).

Yarmouthian

The Yarmouth Soil has developed in alluvium-colluvium and in the upper part of the Columbia till. This soil is preserved only on primary divides where erosion has been minimal (Guccione, 1983). It is on these primary divides where the Sangamon Soil became welded (Ruhe and Olson, 1980) to the Yarmouth Soil, forming the Yarmouth-Sangamon Soil.

Illinoian

Illinoian Loveland Loess, associated with glaciations to the north was derived from the Missouri River valley sediments and deposited over the northcentral Missouri landscape. The Loveland Loess has eroded in dissected areas and is only preserved adjacent to the Missouri River where it is thick (such as in the Franklin Quarry section in Saline County) and on primary drainage divides where erosion has been minimal. On these divides extensive alteration of the thin (<2m thick 35 km north of the Missouri River) deposits during the Sangamonian and the welding of the Sangamon Soil to the underlying Yarmouth Soil have made this deposit difficult, but not impossible to identify.

Sangamonian

The Sangamon Soil developed in the Loveland Loess. Where the Loveland Loess is thick and erosion did not occur, the Sangamon Soil is entirely within the loess. In these sites the soil began to develop during the early Sangamonian and therefore is informally referred to as the early Sangamon soil. However, where the loess is thin, the early Sangamon Soil is welded to the underlying Yarmouth Soil. The Yarmouth-Sangamon Soils have at least a 4.3 m thick leached and iron depleted (redoximorphic terminology is from Vepraskas, 1992) solum. The loess is distinguished from the underlying till by the higher silt content, lower sand content and the presence of less weathered minerals in the loess sand fraction.

Middle Sangamon erosion surfaces, cut into the Loveland Loess and the McCredie Formation, are graded laterally to alluvial depositional surfaces (Fig. 2). These depositional surfaces are 12-30 m above the Holocene floodplains and are underlain by up to 5 m of alluvium. Pedons of the middle Sangamon Soil that developed on these erosion surfaces have a variable morphology (Guccione, 1982). Some of the pedons developed in till on poorly-drained positions are iron-depleted, similar to pedons of the early Sangamon Soil and are without loessial parent material because of erosion. Other pedons of the middle Sangamon Soil developed in till and alluvium-colluvium in somewhat poorly drained situations exhibit red and brown iron masses, and iron depletion zones and are leached of carbonates to a 2.2 m depth. The

argillic horizons, up to 1.3 m thick, have abundant translocated clay and strong subangular blocky structure.

Late Sangamonian erosion surfaces occur topographically below the middle Sangamonian surfaces (Fig. 2). The yellowish brown Fe masses and gray Fe depleted pedons of the late Sangamon Soil developed on these surfaces are less well developed than pedons of the middle Sangamon Soil. The soil structure is weaker, the amount of clay is less, the argillic horizons are not as thick (0.6-0.8 m), and the depth of carbonate leaching is 1.1 m, only half of that found in pedons of the middle Sangamon Soil.

Wisconsinan

The Roxana Silt was deposited during the middle Wisconsinan. At the Franklin Quarry section, a Farmdale Soil is developed in the Roxana. The thick loess that overlies the Farmdale Soil is correlated with the Peoria Loess. The Peoria Loess is recognized over most of the northcentral Missouri landscape.

Holocene

The Holocene Stage is represented by unnamed colluvium and alluvium. Holocene soils have formed in loess, till, colluvium and alluvium and are welded to preexisting soils which have been exposed on erosional surfaces. The soils are predominantly Alfisols (Albaqualfs, Hapludalfs, Paleudalfs and Ochraqualfs), with lesser numbers of Entisols (Fluvaquents, Udifluvents, Udipsamments and Udorthents), Mollisols (Argiudolls, Haplaquolls and Hapludolls), Inceptisols (Dystrochrepts) and Ultisols (Hapludults).

O.V. Windsor Section/Description Stop 3

The O. V. Windsor Section is located within a unique area of eastern Callaway County dominated by an extensive stone line within one-half meter of the surface (Fig. 1). Situated within an alluvial fan, the section appears to exhibit multiple welded soils. The stratigraphic and paleopedologic interpretation of the description shown must be considered tentative. There appear to be at least four lithologic discontinuities beneath the present surface which are distinguishable by morphology and semi-quantitative clay mineral composition (Table 2). More work is needed in the vicinity to understand the geopedologic and paleopedologic relationships.

O.V. Windsor Section: 450 meters west and 510 meters north of the southeast corner, Section 2, Township 47 North, Range 7 West of 5th P.M., Callaway County, Missouri. Surface elevation: 195.1 meters amsl. Original description made on 5 October 1984 by John P. Tandarich and Alice W. Geller.

<u>Samp. No.</u>	<u>Hori- zon</u>	<u>Depth (cm)</u>	<u>Description</u>
		0-18	Road fill/spoil material
<u>Holocene Series</u>			<u>Modern Soil</u> : Mollic Hapludalf (Armster)
<u>Pleistocene Series</u>			<u>Sangamon Soil</u>
<u>Sangamonian Stage</u>			
<u>Unnamed Alluvial Sediments</u> (LS Unit 1)			
W1	A	18-36	10YR 3/2 silt loam (20% clay), weak very fine granular structure, many fine and very fine roots, friable, clear smooth boundary

W2	AB	36-43	10YR 3/2 silty clay loam (28% clay), many 7.5YR 4/4 Fe masses, weak very fine granular parting to weak fine subangular blocky structure, friable, abrupt smooth boundary
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Pre-Illinoian Stages

McCredie Formation

Colluvial & Alluvial Sediments (LS Unit 2)

<u>Samp. No.</u>	<u>Hori- zon</u>	<u>Depth (cm)</u>	<u>Description</u>
W3	2Bt1	43-55	7.5YR 4/4 silty clay loam (39% clay), common 7.5YR 4/6 Fe masses, few fine and few very fine roots, argillans, weak very fine subangular blocky structure, friable, clear smooth boundary
W4	2Bt2	55-66	5YR 4/4 clay loam (45% clay), few fine 7.5YR 5/6 and common fine 2.5YR 4/8 Fe masses, argillans, weak fine subangular blocky structure, <5% angular chert fragments, firm, clear smooth boundary
W5	2Bt3	66-80	5YR 4/4 clay loam (45% clay), few fine 7.5YR 7/6, common medium 2.5YR 4/8 and few fine 7.5YR 5/8 Fe masses, common argillans, Mn masses, strong medium subangular blocky structure, very firm, clear smooth boundary

Colluvial and Alluvial Sediments (LS Unit 3)

			<u>Soil A</u>
W6	3Btb	80-120	5YR 4/3 silty clay loam (36% clay), 5.YR 5/8, 2.5YR 4/8 and 2.5YR 5/8 Fe masses and 5YR 4/2 & 10YR 6/2 Fe depletion zones, 10YR 6/2 argillans, Mn masses, strong medium subangular blocky structure, few very fine roots, <10% angular chert fragments, very firm, abrupt wavy boundary

			<u>Soil B</u>
W7	4Btb1	120-125	2.5YR 4/6 silty clay (46% clay), many coarse 7.5YR 5/2 Fe depletion zones, stone line at top of horizon, strong medium subangular blocky structure, argillans, firm, clear wavy boundary
W8	4Btb2	125-157	5YR 5/6 clay loam (36% clay), many coarse 7.5YR 5/2 Fe depletion zones and many coarse 7.5YR 5/8 & few fine 5YR 4/4 Fe masses, argillans, weak coarse subangular blocky structure, firm, abrupt wavy boundary

			<u>Soil C</u>
W9	5Btb1	157-222	5YR 5/6 clay loam (42% clay), common 10YR 7/1 Fe depletion channels polygonally arranged, and many coarse 7.5YR 6/6 Fe masses, weak fine subangular blocky structure, stone line from 157 to 162 cm, firm, clear smooth boundary

W10 5Btb2 222-245

5YR 5/6 & 5/8 clay loam (42% clay), many 10YR 7/1 Fe depletion channels polygonally arranged, many coarse 7.5YR 6/6 & 6/8 Fe masses, argillans, weak medium subangular blocky structure, firm, clear smooth boundary

Harrison Clay Pit/Description Stop 4

The Harrison Clay Pit (Fig. 1) was described and sampled using the pedo-weathering profile concept of Tandarich et al. (in press). The section contains a mixture of loess and pedisegment which overlies a diamicton interpreted to be the Columbia till member of the McCredie Formation.

Underlying the Columbia till is the Kingdom City Soil. This paleosol may have been developed in alluvium-colluvium as well as the underlying Fulton till. The argillic horizon of the Kingdom City Soil appears to exhibit loading features and is overcompressed. In other locations the soil may have been cryoturbated. The Fulton till rests unconformably on a truncated pedo-weathering profile named the Huntsville Soil and manifested by large lag concentration of pebble to boulder size materials.

The Huntsville soil is developed in diamicton which is part of the Moberly Formation. Morphologically, texturally and mineralogically, the Moberly formation is distinct from the overlying McCredie Formation (Table 3). At the base of the Moberly formation is unweathered sand, probably proglacial in origin. The Moberly Formation overlies the New Florence Soil that had developed in the Pennsylvanian shale and limestone.

Harrison Clay Pit Section: 650 meters east and 150 meters north of the southwest corner, Section 1, Township 48 North, Range 9 West of 5th P.M., Callaway County, Mo. Elevation: 243.9 meters amsl.

Samp. Hori- Depth
No. zon (cm)

Description

Holocene Series

Modern Soil: Typic Hapludalf (Keswick)

Pleistocene Series

Sangamonian Stage

Late Sangamon Soil

Unnamed Sediments (LS Unit 1)

H1	Ap	0-15	10YR 3/2 & 3/3 loam, common 10YR 4/4 Fe masses, weak very fine granular structure, leached, clear smooth boundary
H2	BA	15-38	10YR 4/4 & 3/3 loam, many 10YR 5/8 Fe masses, weak very fine subangular blocky structure, leached, clear smooth boundary
H3	Bt1	38-52	10YR 5/4 clay loam, many 10YR 5/8, common 2.5YR 4/6 & 5/8 Fe masses, few 10YR 4/2 argillans, moderate fine subangular blocky structure, few 10YR 2/2 Mn stains, leached, clear smooth boundary

Pre-Illinoian Stages

McCredie Formation

Columbia Till Member (LS Unit 2)

H4	2Bt2	52-73	10YR 6/6 clay, few 2.5Y 6/2 depletion zones & 10YR 4/6 Fe masses, few 10YR 2/2 Mn stains, few 10YR 4/3 argillans, moderate fine subangular blocky structure, leached, clear smooth boundary
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H5	2Bt3	73-98	10YR 6/6 clay loam, few 10YR 5/8 Fe masses & common 2.5Y 6/2 depletion zones, few 10YR 4/3 argillans, 10YR 2/2 Mn root channel linings, moderate fine subangular blocky structure, leached, clear smooth boundary
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<u>Samp. No.</u>	<u>Hori- zon (cm)</u>	<u>Depth</u>	<u>Description</u>
H6	2BC	98-120	2.5Y 6/6 clay loam, common 2.5Y 7/2 depletion zones, 2.5Y 5/4, & 10YR 6/8 Fe masses, 5Y 3/2 Mn concentrations & stains on ped faces, weak medium subangular blocky structure, leached, clear smooth boundary
H7	2CB	120-133	2.5Y 6/6 clay loam, many 10YR 5/8 Fe masses & few 2.5Y 7/2 depletion zones, 5Y 2.5/1 Mn stains on ped faces, weak medium angular blocky structure, few carbonate concretions, weakly calcareous, clear smooth boundary
H8	2C	133-275	2.5Y 5/4 loam, 2.5Y 6/2 Fe depleted joint linings, 5Y 2.5/1 Mn coatings on ped faces, strong coarse angular blocky structure, calcareous, abrupt wavy boundary

Fulton Till Member (LS Unit 3)

Kingdom City Soil

H9	3Btb1	275-290	Silty clay in laminar horizontal bands of 2.5Y 6/6 & 10YR 6/8 Fe masses, 2.5Y 6/2 & 5YR 6/1 depletion zones, 5Y 2.5/1 root traces, few 5YR 6/1 argillans, moderate fine platy structure, leached, clear smooth boundary
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H10	3Btb2	290-327	Silty clay in laminar horizontal bands of 2.5Y 6/6 & 10YR 6/8 Fe masses, 2.5Y 6/2 & 5YR 6/1 depletion zones grading into interspersed Fe masses and depletion zones, 5Y 2.5/1 Mn root traces, few 5YR 6/1 argillans, moderate fine subangular blocky structure, leached, clear smooth boundary
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H11	3BC	327-402	2.5Y 6/2 clay loam, common 2.5Y 6/6 & 10YR 6/8 Fe masses, 5Y 2.5/1 Mn stains on ped faces, moderate medium subangular blocky structure, leached, gradual smooth boundary
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H12	3CB	402-521	2.5Y 5/6 clay loam, 2.5Y 6/2 depletion zones & 10YR 6/8 Fe masses along joints, strong coarse angular blocky structure, calcareous, abrupt wavy boundary
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Moberly Formation (LS Unit 4)

Huntsville Soil

H14	4C	521-785 531	2.5Y 5/6 loam, 2.5Y 6/2 depletion zones & 10YR 5/4 Fe masses along joints, 5Y 3/1 Mn stains on ped faces, strong coarse & very
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H15	585	coarse angular blocky structure, stone line (including boulders) at top of horizon, calcareous, boundary indeterminate	
H16	635		
H17	685		
H18	735		
H19	785		
	4C/D	785-865	Loam, 2.5 Y 4/4, 4/6 & 5/6 Fe zones and masses (C part) surrounding 5Y 3/1 D cores, strong coarse angular blocky structure calcareous, some sand pods, boundary indeterminate
H20	810		
H21	835		
H22	860		
H23	5D	865-880	10YR 4/1 sandy loam, single grain structure, calcareous, wood & charcoal fragments, abrupt boundary

Franklin Quarry/ Stop 5

The Franklin Quarry is an undescribed section located in an abandoned limestone quarry. The quarry is on the south side of the Missouri River approximately halfway between Columbia and Kansas City, Missouri. It is located on private property.

The deposits in the quarry have been cursorily examined and appear to be as follows:

Holocene Series

Modern Soil: Typic Hapludoll

Pleistocene Series

Wisconsinan Stage

Woodfordian Substage

Peoria Loess

Farmdalian Substage

Farmdale Soil

Altonian Substage

Roxana Silt

Sangamonian Stage

Sangamon Soil

Illinoian Stage

Loveland Loess

Yarmouthian Stage

Yarmouth Soil

Pre-Illinoian Stages

McCredie or Moberly Formation

At the base of the Pre-Illinoian, there are proglacial silts and clays resting on the limestone bedrock.

TABLE 3/4/5-1. Quaternary stratigraphic units of northcentral Missouri.

Chronostratigraphy	Lithostratigraphy	Pedostratigraphy
Stage		
Holocene		Holocene soils
Wisconsinan	Peoria Loess	
		Farmdale Soil
	Roxana Silt	
Sangamonian	colluvium and alluvium	Sangamon Soil late Sangamon middle Sangamon early Sangamon
Illinoian	Loveland Loess	
Yarmouthian		Yarmouth Soil
pre-Illinoian	Columbia Till Member	
	Mc	
	C	Kingdom City Soil
	r	
	e	
	d	
	i	colluvium and alluvium
	e	
	Fm	Fulton Till Member
		Huntsville Soil
	Moberly Formation Undifferentiated	
preglacial		New Florence soil
	Whipporwill formation	

Modified from Guccione (1983).

TABLE 3/4/5-2. O.V. Windsor Section, Clay Mineral Data.

Samp. No.	Geol Unit	Hori- zon	Depth (cm)	Clay Mineral Percentage		
				Exp.	Ill.	K + C
W1	1	A	18-36	22	55	23
W2	1	AB	36-43	23	50	27
W3	2	2Bt1	43-55	31	42	27
W4	2	2Bt2	55-66	31	42	27
W5	2	2Bt3	66-80	29	41	30
W6	3	3Btb	80-120	22	41	37
W7	3	4Btb1	120-125	19	45	36
W8	3	4Btb2	125-157	13	44	43
W9	3	5Btb1	157-222	18	45	37
W10	3	5Btb2	222-245	17	45	38

TABLE 3/4/5-3. Harrison Clay Pit: texture, pH, carbonate and clay mineral composition data.

Samp. No.	LS Unit	Hori- zon	Depth (cm)	Particle Size Distribution Analysis Percentages						pH	Carbonate Percentages		Clay Mineral Percentages		
				Sand	Silt	Silt fractions (u)			Clay				Exp.	Ill.	K + C
						53-32	32-20	20-2							
H1	1	Ap	0-15	34	44	13	10	21	22	6.4	0	0	46	26	28
H2	1	BA	15-38	32	42	11	10	21	26	5.5	0	0	38	32	30
H3	1	Bt1	38-52	26	36	10	7	19	38	4.9	0	0	38	29	33
H4	2	2Bt2	52-73	26	29	17	5	7	45	5.5	0	0	70	13	17
H5	2	2Bt3	73-98	34	28	9	5	14	38	6.3	0	0	77	9	14
H6	2	2BC	98-120	32	34	10	6	18	34	7.2	0	0	76	10	14
H7	2	2CB	120-133	34	34	10	6	18	32	7.6	0	0	71	14	15
H8	2	2C	133-275	37	36	11	6	19	27	8.1	5	0	68	15	17
H9	3	3Btb1	275-290	27	33	9	6	18	40	7.6	0	0	69	15	16
H10	3	3Btb2	290-327	28	31	8	6	17	41	7.5	0	0	66	16	18
H11	3	3BC	327-402	32	31	9	6	16	37	7.6	0	0	66	16	18
H12	3	3CB	402-521	36	34	10	7	17	30	7.6	0	0	76	12	12
	4	4C	521-785												
H14			531	35	41	10	7	24	24	8.1	9	3	44	31	25
H15			585	34	40	9	7	24	26	8.1	9	0	50	27	23
H16			635	34	40	9	7	24	26	8.0	9	1	50	26	24
H17			685	35	40	10	7	23	25	8.1	8	3	45	33	22
H18			735	36	44	10	5	29	20	8.1	12	5	47	29	24
H19			785	11	61	6	9	46	28	8.0	11	5	48	28	24
	4	4C/D	785-865												
H20			810	20	53	7	9	37	27	8.0	10	5	40	31	29
H21			835	52	35	12	7	16	13	8.1	14	7	48	29	23
H22			860	37	40	9	7	24	23	8.0	11	4	39	29	32
H23	4	5D	865-880	55	27	5	5	17	18	8.0	14	3	50	28	22

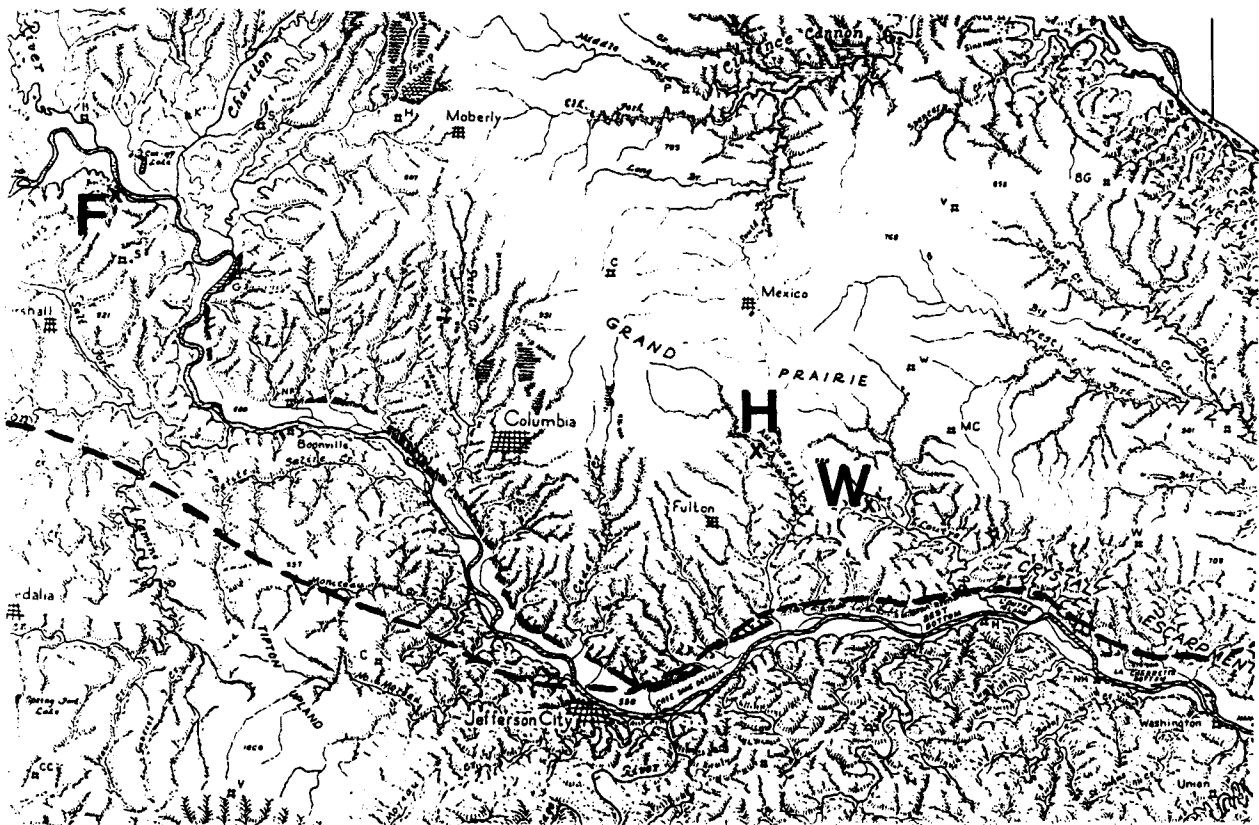


FIGURE 3/4/5-1. Location of northcentral Missouri sections: F - Franklin Quarry, H - Harrison Clay Pit and O.V. Windsor. The dashed line indicates the glacial limit. Map source copyright Terry McBride (1977), publication permission secured.

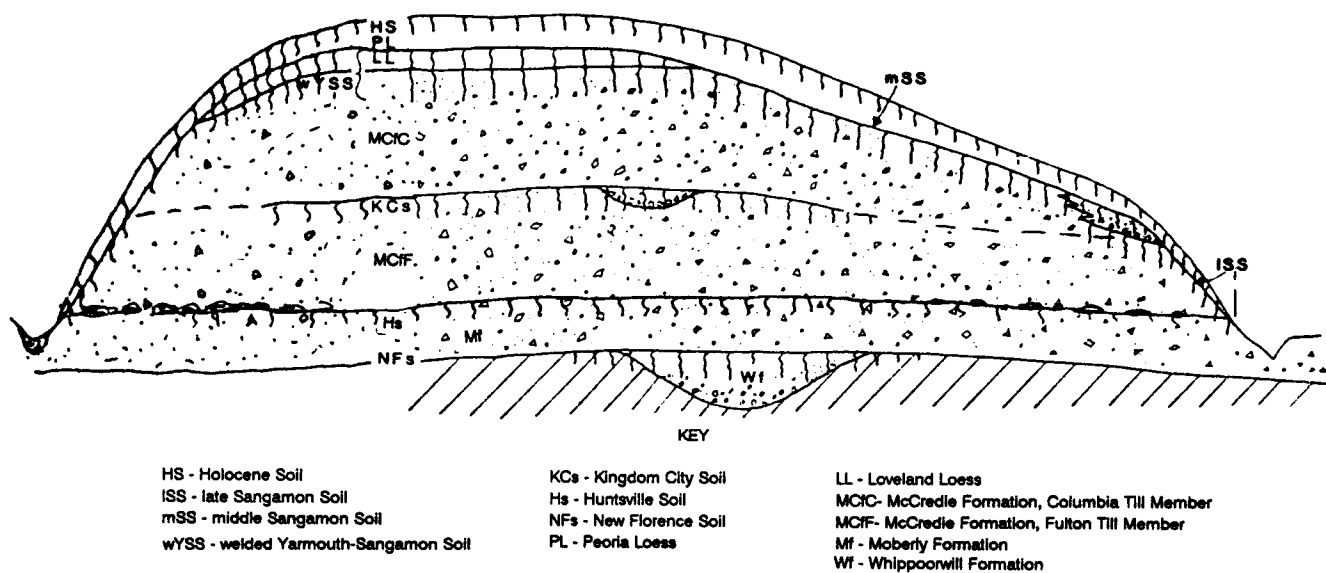


FIGURE 3/4/5-2. Idealized cross section of Quaternary deposits and soils in northcentral Missouri.

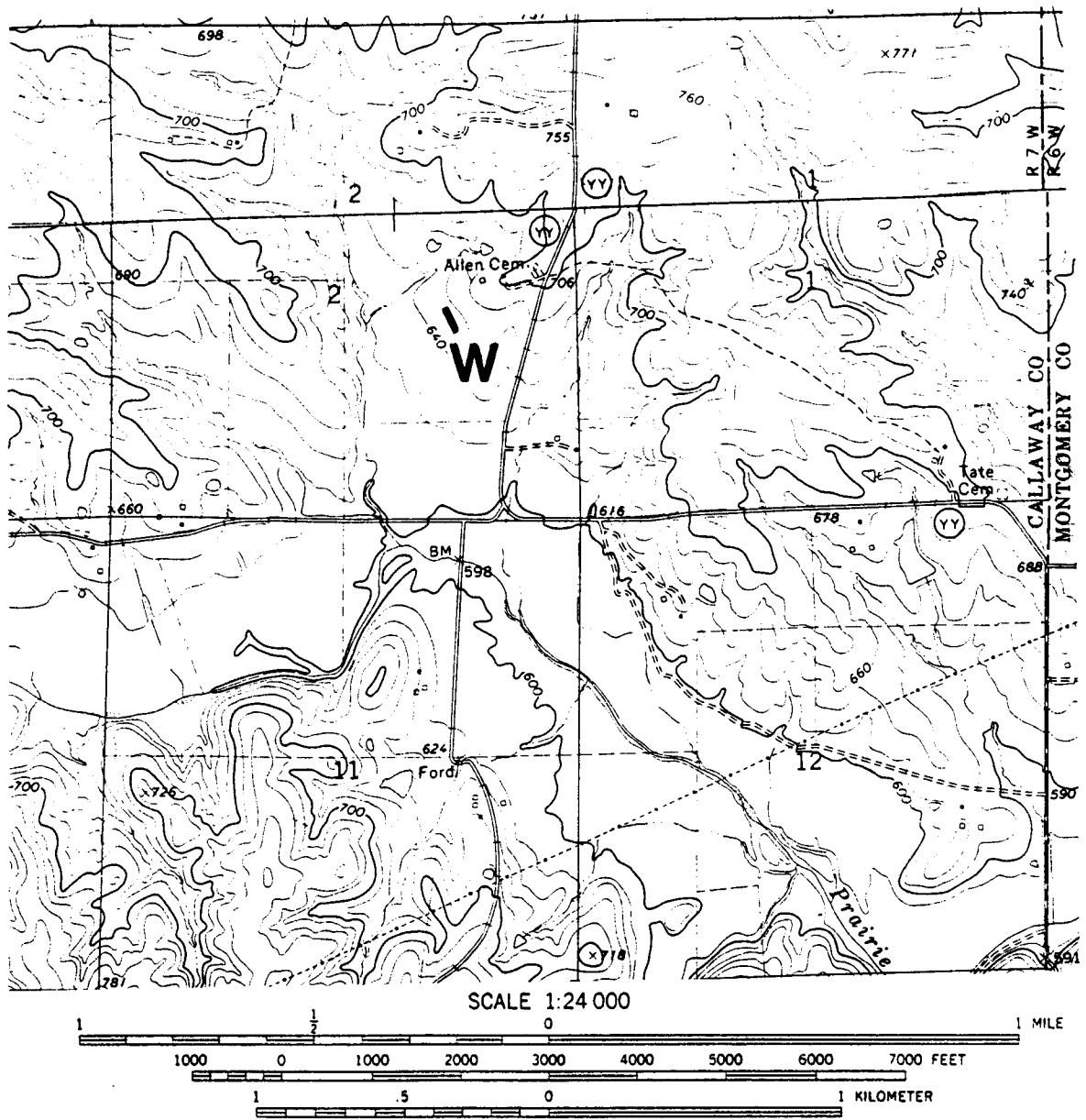


FIGURE 3/4/5-3. O.V. Windsor Section (W), SE1/4 of Section 2, Township 47 North, Range 7 West, Callaway County, Missouri. U.S.G.S. Readsville and Williamsburg 7.5 minute quadrangles, 1973.

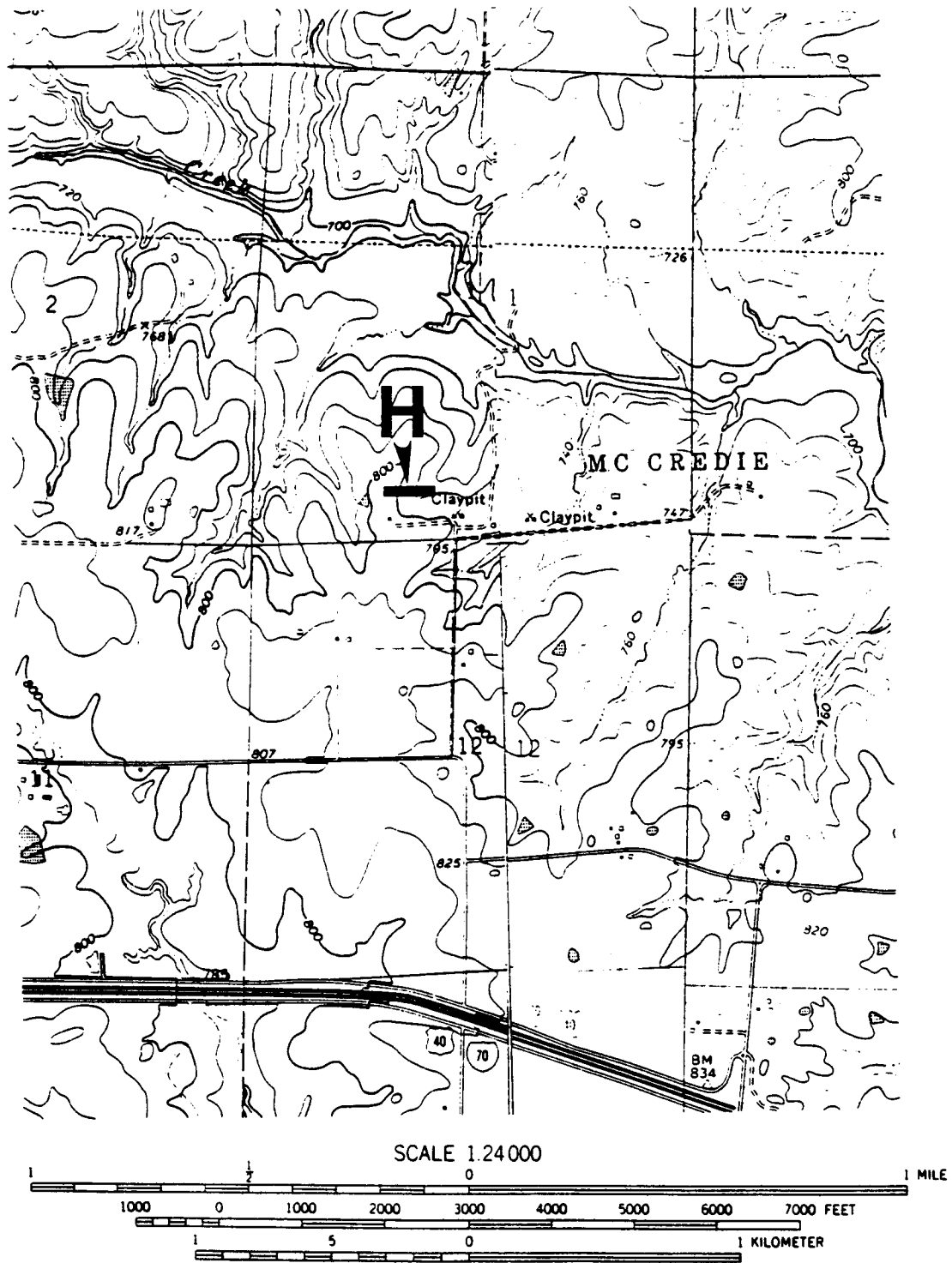


FIGURE 3/4/5-4. Harrison Clay Pit Section (H), SW1/4 of Section 1, Township 48 North, Range 9 West, Callaway County, Missouri. U.S.G.S. Calwood and Kingdom City 7.5 minute quadrangles, 1985.

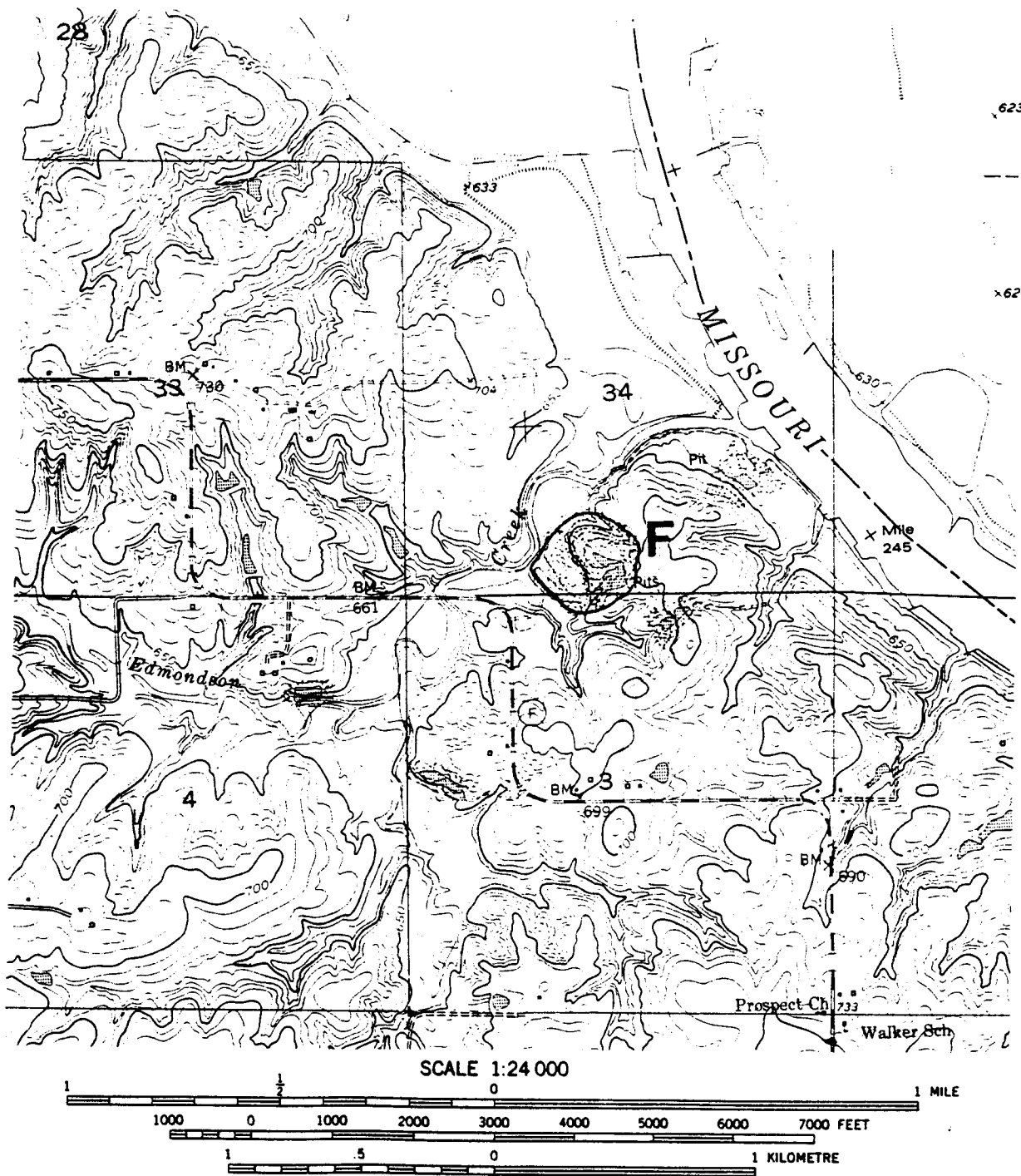


FIGURE 3/4/5-5. Franklin Quarry Section (F), Section 34, Township 53 North, Range 20 West, Saline County, Missouri. U.S.G.S. New Frankfort 7.5 minute quadrangle, 1979.

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Stop 6 West of Manhattan, Kansas, on Scenic Drive, 1.2 miles north of the intersection with K-18: A Permian paleosol

Location: SE 1/4, Sec. 14, T10S, R7E, Riley County, Kansas
(USGS Manhattan, KS 7.5 minute quadrangle)

Contributors: T.J. McCahon and K.B. Miller

PALEOSOLS IN THE BLUE RAPIDS SHALE (LOWER PERMIAN) OF NORTHCENTRAL KANSAS

The Blue Rapids paleosol stop is located west of Manhattan, Kansas, on Scenic Drive 1.2 mi north of the intersection with K-18 (Site 2, Fig.1). This site offers continuous exposure of the lower Permian (Wolfcampian) section from the Bader Limestone of the Council Grove Group through the Wreford Limestone at the base of the Chase Group. Six cyclothemic sequences equivalent to the cycles defined by Elias (1937) occur within this stratigraphic interval.

The general facies pattern of Lower Permian cyclothem is well displayed at this locality. Variegated mudstones with extensive paleosol development are abruptly overlain by thin transgressive marine carbonates typically containing intraclasts, skeletal debris that may be phosphatized, and bone fragments. Above these are gray to yellowish gray fossiliferous shales followed by thick limestone units that may show evidence of subaerial exposure at their tops and are overlain by variegated mudstones. The internal complexity of these formation- to member-scale units can be examined in detail.

Flooding surfaces have proven particularly useful for high-resolution correlation within variegated mudstone intervals, dividing them into meter-scale traceable units composed of single paleosol profiles or series of stacked paleosols (Miller and West, 1993). These paleosols have received little attention, although they have considerable potential as environmental and paleoclimatic indicators. The detailed correlation of paleosol packages, though not necessarily individual paleosols, permits the reconstruction of spatial variability in paleosol development and the assessment of short-term environmental and paleoclimatic change. This is illustrated using detailed measured sections of the Blue Rapids Shale of the upper Council Grove Group in northeastern Kansas. These sections have been correlated at a meter-scale over distances of 25 km. Significant lateral changes in paleosol profiles have been recognized that can be understood as a result of differences in sediment accumulation rates, duration of pedogenesis, and paleotopography. The documentation of such lateral variability over several successive cyclothem provides a means of reconstructing the evolution of the terrestrial and coastal plain environment over time.

A wide variety of pedogenic features can be observed within the stratigraphic interval of the Blue Rapids Shale of north central Kansas. Paleosols in the Blue Rapids are easily identified in the field using soil horization, rooting, structure, and pedogenic clay cutans. Paleosols in the Blue Rapids Shale have been studied at four different sites in a 30 km radius of Manhattan, Kansas. At each site, the Blue Rapids Shale consists of three stacked paleosol profiles of varying morphology separated by discontinuities (Fig.1). Pedogenic morphology varies considerably between the different sites, probably due to the paleogeomorphic position of the paleosol when it was forming.

The variegated mudstone of the Blue Rapids is divisible into four intervals marked by sharp flooding surfaces (Fig.2). Resting on desiccation cracked and brecciated micrite from the Crouse, the lowest interval preserves a nearly complete paleosol profile with well-developed color horizonation, angular blocky ped structure, pedogenic slickensides, and carbonate rhizcretions. This paleosol is sharply overlain by a thin calcareous mudstone containing ostracodes and fish bone. This mudstone in turn lies at the base of another similar, well-preserved paleosol profile, itself abruptly overlain by a laminated mudstone containing ostracodes, fish bone and granule-size intraclasts. The next interval of greenish gray (5GY 6/1) mudstone, though different in appearance, also shows evidence of pedogenesis. The most prominent pedogenic structures in this unit are abundant carbonate-enhanced pseudoanticlines. A horizonally laminated carbonate bed truncates this paleosol and is followed by an interval of dusky yellow (5GY 5/2) shale with thin, wavy carbonate layers. Marking the top of this interval and the base of the next cyclothem is a marine transgressive surface overlain by an intraclastic limestone with abundant marine fossils.

Paleosol Morphology and Interpretation

The bottom paleosol at the fieldtrip stop (Site 2, Fig.1) would classify as a Calcic Argillisol (Mack et al., 1993) that underwent bisequal pedogenesis. A gleyed horizon sits atop the lower paleosol, indicating waterlogging conditions, probably due to the overlying flooding surface that buried the paleosol. There are well-developed clay cutans around the angular blocky peds in the upper and middle profile, indicating clay illuviation. Rhizcretions are confined primarily to the Btk horizon. The Btg1 horizon contains clay cutans and green mottles. Root traces in the Btg1 and Btg2 horizons are not carbonate coated. Subsequent pedogenesis in the upper profile (Btk) and changing environmental conditions are noted by the increased CaCO_3 levels and CaCO_3 rhizcretions. Clay illuviation would have been greatly inhibited in the presence of carbonates, indicating that the carbonate was a pedogenic feature that probably occurred after clay illuviation. Gleyed horizons (Btg) at the base of the profile indicate groundwater gleying in the profile. Water moving down through the profile would have encountered the 2C horizon and flowed laterally, often saturating the lower Bt horizon creating gleyed conditions. The upper contact of the 2C horizon would have provided a natural aquiclude.

The middle paleosol profile shows similarities to the lower profile, and would classify as a Calcic Argillisol. However, carbonate concentration extends to a greater depth in the profile, indicating that groundwater influences were minimal in removing CaCO_3 from the profile. Well-developed clay cutans on the angular blocky peds in the paleosol (Btk) indicate considerable clay illuviation during paleosol development. Carbonate rhizcretions in the paleosol are an indication that CaCO_3 was precipitated: 1) outside the root on a low Ca absorber plant, 2) by microorganisms as a detoxification mechanism (Monger, 1990), or simple microbial respiration (Arnott and Pautard, 1970), or 3) all of the above. Most of the carbonate in the paleosols is located around the rhizcretions, lending support for one of the above hypotheses, as opposed to other suggested sources for pedogenic CaCO_3 (Gile et al., 1966). Rhizcretions in the paleosol extend down to the BC horizon. Accumulation of CaCO_3 is inferred to have postdated clay illuviation as in the lower paleosol. The 2C horizon in the middle paleosol is a calcareous mudstone containing fish bones and ostracodes. Any carbonate dissolution and accumulation in the profile from the 2C parent material would have been minimized in the profile by groundwater leaching. This is apparent in the lowered CaCO_3 content of the Btg and BCg horizons relative to the rest of the profile.

The uppermost paleosol has undergone very little pedogenesis. This paleosol under the classification of Mack et al. (1993) would be a Vertisol. Carbonate-rich pseudoanticlines (1-2 cm thick), along with botryoidal carbonate nodules dominate the mid-profile (Bk horizons). The absence of clearly-developed horizonation in the field is consistent with extensiveurbation associated with seasonal wetting and drying. The top horizon (Btn) of the uppermost paleosol has a weakly developed prismatic structure, and contains small charcoal fragments throughout. This top horizon was probably subjected to sodium saturation (marine

salts) during the flooding event which ended pedogenesis lower in the profile, as similarly identified for Vertic paleosols in the Silurian (Driese et al., 1992). Pseudoanticlines, carbonate concretions, coarse angular to subangular blocky ped structure in the profile middle (Bk), and some root traces, are the only diagnostic pedogenic features evident in this paleosol profile. A heavily weathered laminated mudstone (C) sits atop the middle paleosol and marks the base of this upper paleosol.

There is some indication in the morphology of the paleosols to suggest that environmental conditions were changing within paleosols profiles and between paleosols in the Blue Rapids. While large carbonate rhizocreations are present in the lower and middle paleosols, there are smaller root traces in these paleosols that are carbonate-free. In addition, the occurrence of clay cutans below the horizons of carbonate accumulation suggests increasingly dry conditions during the genesis of these lower profiles. On a larger scale, the uppermost Vertisol appears to indicate that there is a trend toward wetter conditions than were present during the development of the carbonate-dominated lower paleosols. The change from Calcic Argillisols or Calcisols in the lower part of the Blue Rapids Member to Vertisols in the upper part is a trend observed at several other outcrop localities over distances of tens of kilometers (Fig.1). In addition, this same trend in paleosol profiles is recognized within several other lower Permian variegated mudstone intervals. The Speiser Shale Member (Fig.3), for example, has a series of stacked and truncated paleosols in its lower part with reddish brown horizons and well-developed clay cutans. By contrast, the upper paleosol profile is a Vertisol with clearly developed pseudoanticlines and mukkara structure (gilgae), and large concertina root traces. These consistent trends suggest cyclic change from sub-arid or sub-humid conditions to seasonal wet-dry conditions during the course of cyclothem deposition.

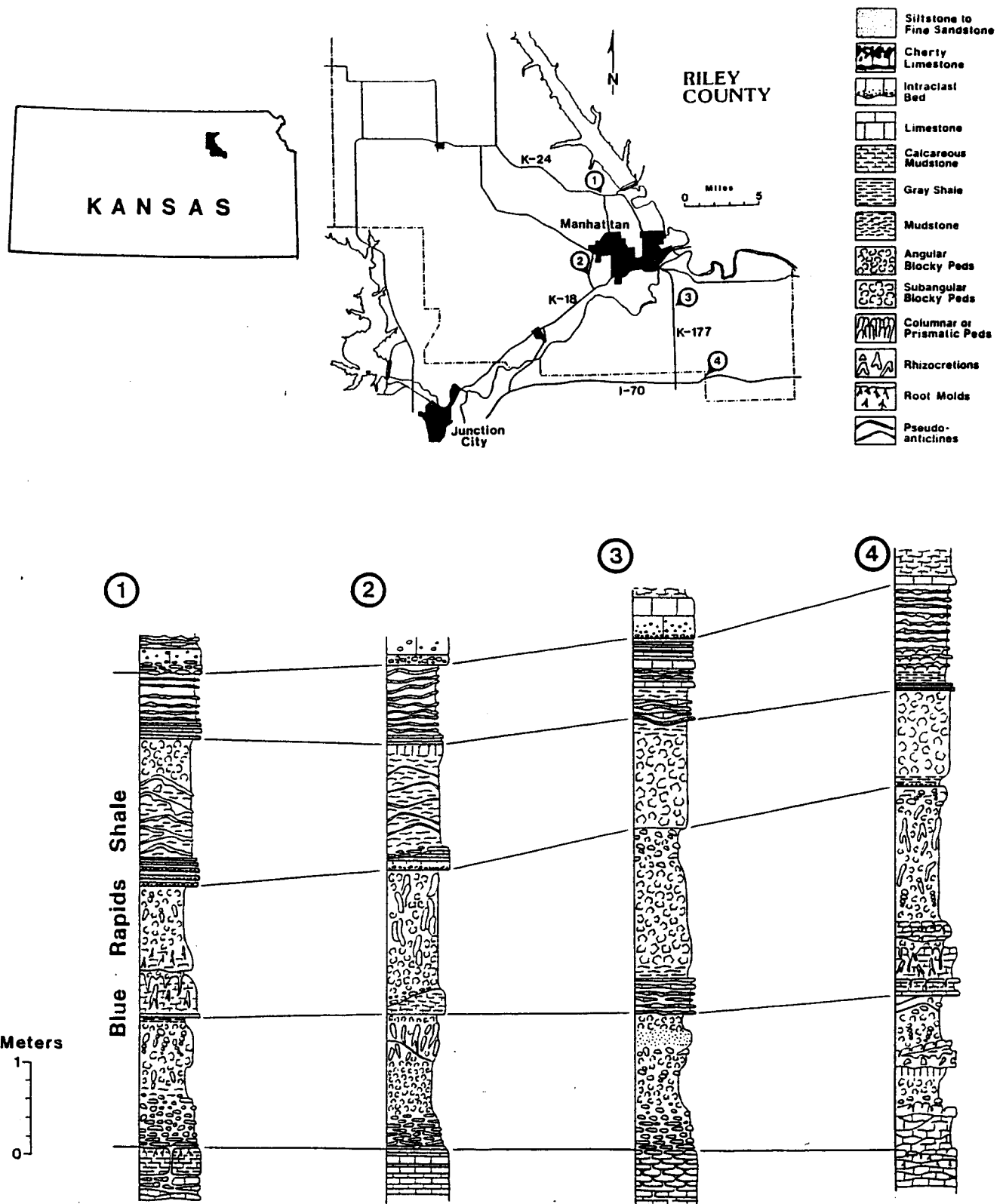


FIGURE 6-1. Blue Rapids (Permian) paleosol Sites 1-4 near Manhattan, Kansas. Stacked paleosol sequences and correlation between sites is shown, along with sedimentological key.

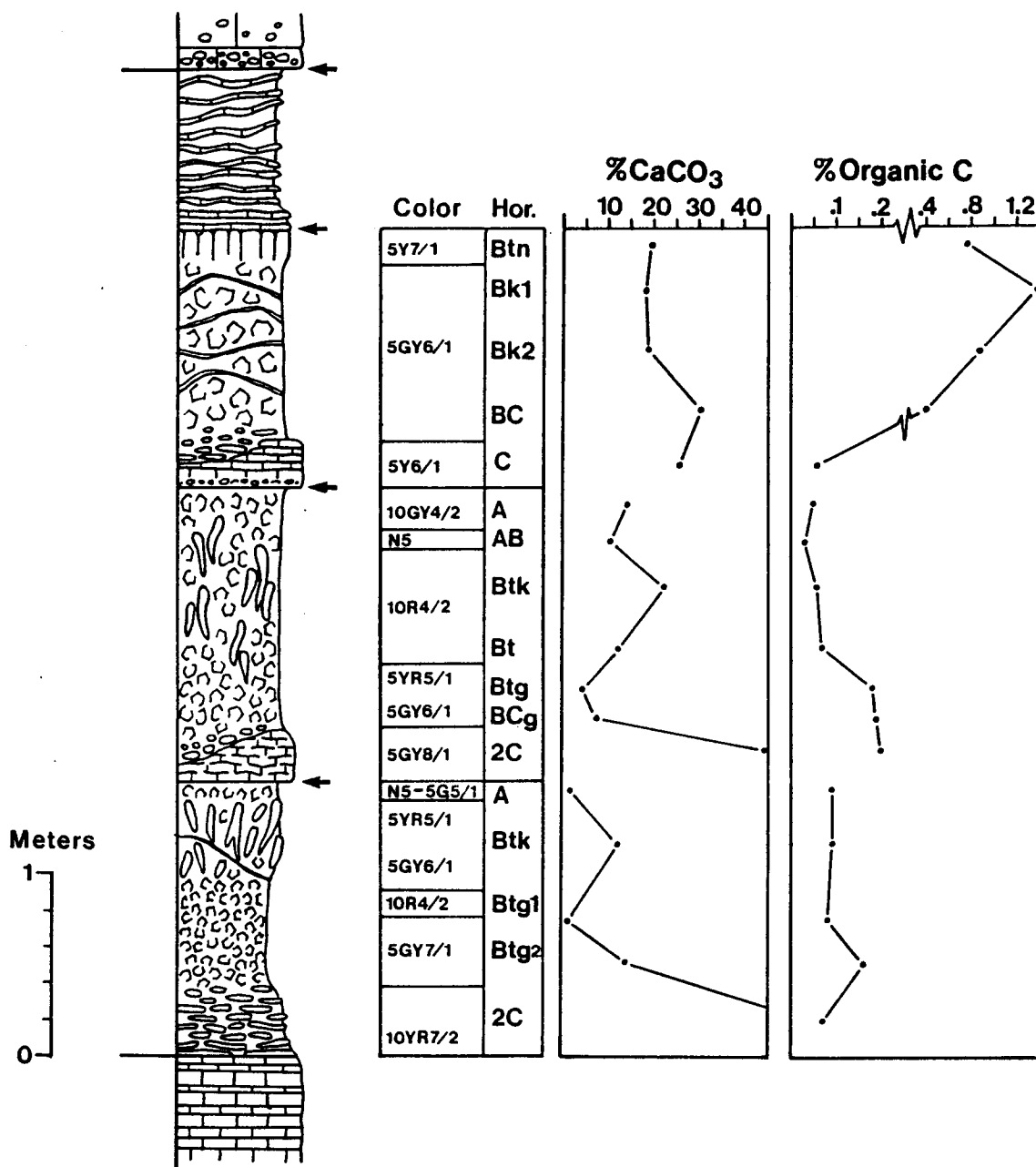


FIGURE 6-2. Blue Rapids Site 2, with three stacked paleosols. Color horizon designation, percent organic carbon and CaCO₃ are given for each paleosol horizon. Heavy arrows mark the flooding surfaces and boundaries between paleosols.

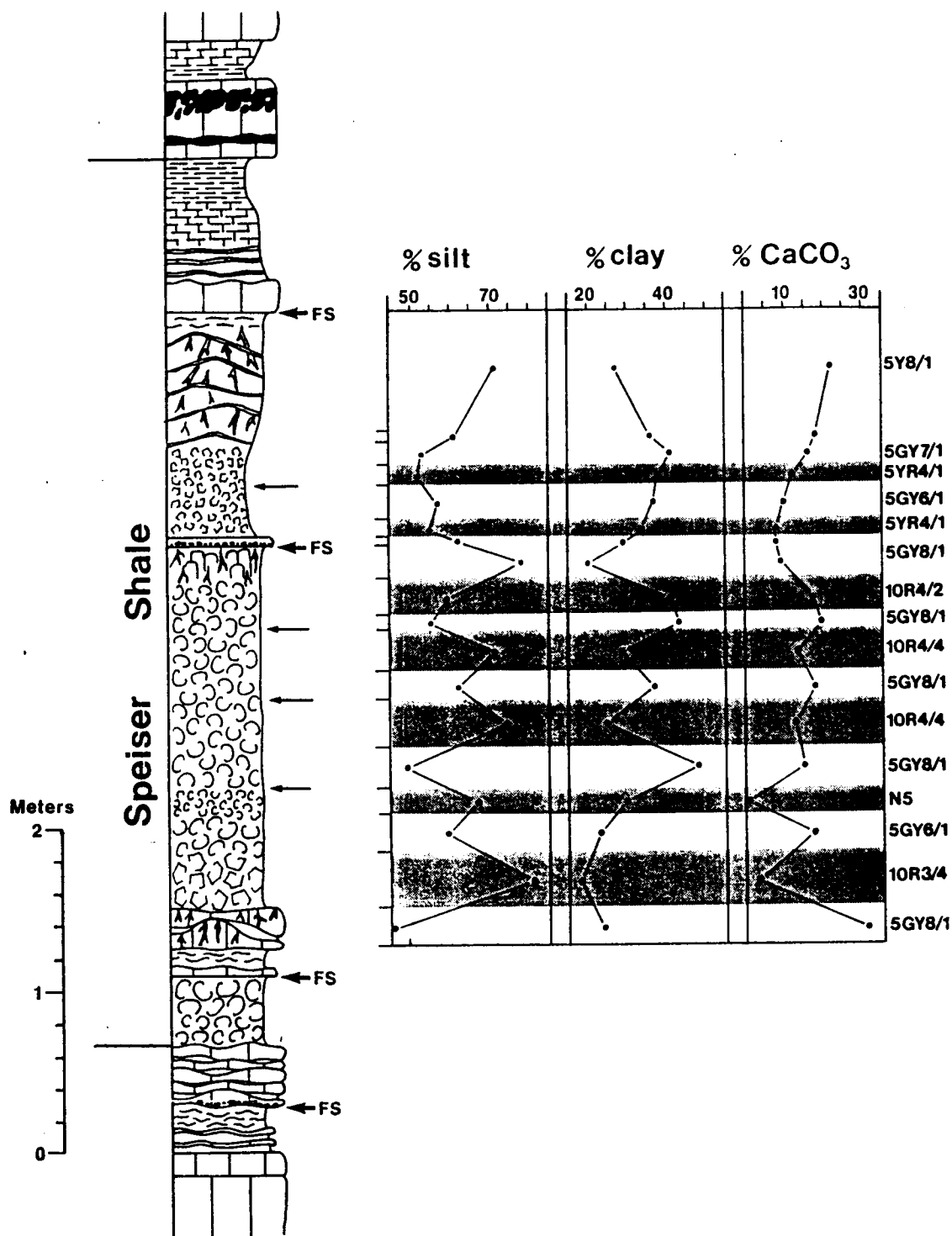


FIGURE 6-3. Measured stratigraphic column for the Speiser Shale at Site 2. The percent silt, clay and CaCO_3 is given for each paleosol horizon. Munsell color designations are shown to the right--shaded horizons are reddish or gray, non-shaded horizons are greenish gray. Heavy arrows mark flooding surfaces (FS), and thin arrows mark sharp boundaries between paleosol horizons indicating truncation of the paleosol profile.

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Stop 7 K-238, 300 m [meters] south of the junction of I-70 north of Brookville, Kansas: A Cretaceous paleosol

Location: SW 1/4, Sec. 34, T13S, R5W, Saline County, Kansas
(USGS Culver, KS 7.5 minute quadrangle)

Contributors: D.A. Porter, T.J. McCahon, M.D. Ransom

A HYDROMORPHIC PALEOSOL IN THE DAKOTA FORMATION (CRETACEOUS), NORTH-CENTRAL KANSAS

The Dakota paleosol stop is located in northeastern Saline County on K-238, 300 m south of the junction of I-70 north of Brookville (Fig. 1). The outcrop exposure containing the paleosol was exposed by county road crews in 1954. The site offers an exposure of the Terra Cotta Clay Member that contains the paleosol, and the Dakota Sandstone caprock that overlies the Terra Cotta. The Dakota Formation in north-central Kansas consists of lenticular sands and clays deposited as overbank sediment within a gradational alluvial plain and deltaic complex (Franks, 1975). These sands and clays were eventually buried by the Graneros Shale during a transgression of the Late Cretaceous Graneros Sea. Palynological evidence from the Terra Cotta Member indicates a sub-tropical vegetation that is native to bogs or wet grassy plains and compatible with acid soils (Snyder, 1963). The Terra Cotta Clay Member is one of two members that make up the Dakota Formation. This lower unit of the Dakota was described and sampled at the Brookville location in north-central Kansas.

The Brookville site consists of variegated mudstone and siltstone intervals (Fig. 2). At the base of the Brookville outcrop is a thick depositional gray shale containing fossil plant compressions. X-ray diffraction of the silt fraction of this gray shale indicates the presence of siderite, kaolinite, and dolomite. Immediately above the fossiliferous gray shale is a depositional carbonaceous unit with charred plant fragments. A gray siltstone sits atop the carbonaceous bed, and can be recognized for the iron-stained pseudoanticlinal features. These fracture cracks in the gray siltstone are filled with a coarser grained sediment than the surrounding matrix. A thin depositional unit of carbonaceous siltstone sits immediately below the paleosol unit. This carbonaceous siltstone also contains charred plant fragments. The upper two meters of the Brookville site have identifiable pedogenic features. However, the upper paleosol profile may have been truncated. Root traces suggest the genesis of this variegated unit is a paleosol with diagenetic overprinting. Interpretation of the Cretaceous paleosol was based on field site study and laboratory analyses of particle size by pipet (Kilmer and Alexander, 1949), organic carbon (Janitzky, 1986), x-ray diffraction (Moore and Reynolds, 1989) of clays and silts, as well as micromorphological examination of thin sections (Brewer, 1976).

Paleosol Morphology and Interpretation

The Brookville paleosol profile is about 2 m thick and massive (Fig. 2). X-ray diffraction data indicates that hematite, goethite, and kaolinite are present throughout the profile in both the silt and clay fraction. Illite is also present in the clay fraction in relatively smaller quantities.

0-40 cm: The upper 40 cm is a very pale brown (10YR 8/2) kaolinitic mudstone. Red mottles (10R 4/3)

in this unit are 5-20 cm in diameter and extend vertically ≥ 1 m, bifurcating downward in the profile. The mottles have developed plinthitic properties from almost 40 years exposure to air. The exposed mottles in the profile have taken on a strong jagged scoria-like appearance. Cutting back into the profile, these red mottles have a soft clayey texture. Observations in 1968 noted that the plinthitic features were faintly observable at the site (P. Twiss, personal communication and photographs). Fresh, crystalline muscovite ≤ 0.5 mm is observed in thin section in this horizon. The muscovite may be diagenetic in origin and therefore neoformed. Redoximorphic features also observed in thin section include common (2-20%) channel neoferrans and common Fe and Mn nodules.

41-189 cm: The second unit is a 70 cm brown (7.5YR 5/6) kaolinitic mudstone that grades into a 110 cm brownish yellow (10YR 6/8, 7/8) siltstone. Plinthitic bifurcating mottles in the upper unit are reddish brown (2.5YR 5/4), and extend into the lower portion of the unit as reddish (2.5YR 4/4) mottles. These plinthitic mottles again range from 5-20 cm diameter. Redoximorphic features increase in this section to include properties of redox depletion as well as concentration. Concentration features include many ($> 20\%$) medium to coarse Fe and Mn nodules (2-2.5 mm diameter), Fe (2.5Y 7/6, 2.5YR 4/4) and Mn (2.5YR 2.5/0) masses in the matrix, common channel neoferrans and ferrans lining root walls. Fe depletion zones in thin section occur around a few root channels and are surrounded by quasiferrans, similar to those observed in modern hydromorphic soils (Bouma et al., 1990). Fe depletion zones (> 2 mm in width) need numerous oxidation-reduction cycles to develop. The death and subsequent decomposition of the root induces reduction of Fe necessary to enlarge the depletion feature (Vepraskas, 1992).

Present in both paleosol units are macro root traces of varying size with multiple bifurcations or branchings. Many of the small root traces are approximately 0.5 to 1 mm in diameter and filled with either hematite or manganese. Stable macropores are developed by the growth of successive roots in one channel. The plinthitic mottles that bifurcate downward into the profile are interpreted as root traces (5-20 cm diameter) of trees or large aquatic plants. According to Cohen (1982), replacement, impregnation, and void-filling of organic matter by mineral matter occurs without total loss of root anatomical structure. The red plinthitic mottles that extend vertically throughout the profile are alone not evidence for pedogenesis since subaquatic plants could develop the same characteristics during diagenesis.

The distinct red, yellow, and black coloration in the second unit conforms to descriptions of zones with fluctuating water tables that produces oxidation and reduction features (Birkeland, 1984). Freyet and Plaziat (1982) indicate that these mottled ferruginous zones are relatively common in hydromorphic paleosols. Iron concretions or nodules usually occur in zones of high pH or high redox potential, and will often survive hydrologic fluctuations such as the lowering of the water table (Gerrard, 1987). Iron and manganese concretions or nodules among red mottles and streaks are an indication that diffusion of oxygen into the paleosol aggregates was rather rapid (Ponnamperuma, 1972). The redoximorphic features observed in this profile such as Fe and Mn nodules and channel ferrans can be explained by both diagenetic and pedogenic processes. The properties that characterize pedogenesis and diagenesis cannot always be differentiated in a paleosol that has been buried and lithified over millions of years. Diagenetic and pedogenic processes have occurred close to 90 myr ago, but differences between them now are small and hard to distinguish.

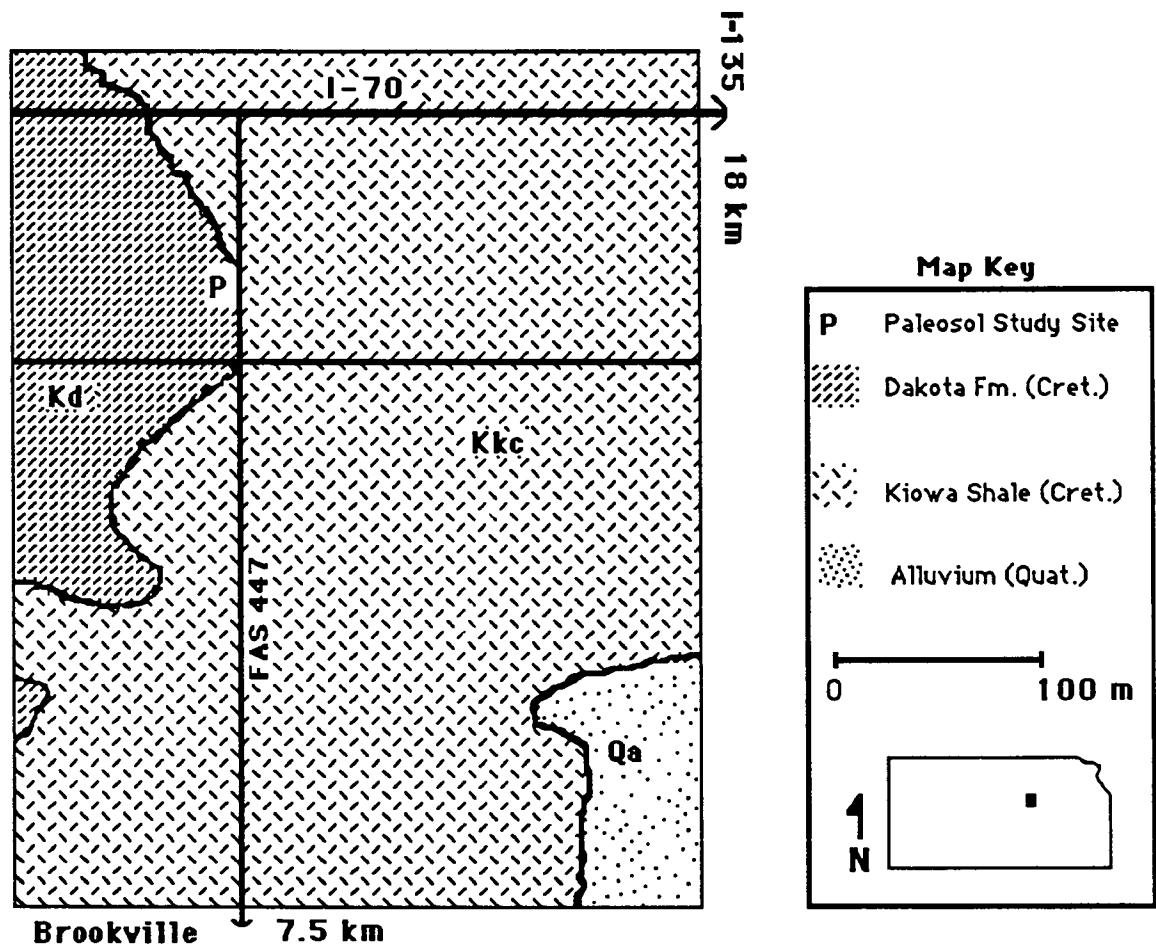


FIGURE 7-1. Dakota (Cretaceous) paleosol site, Saline County, Kansas.

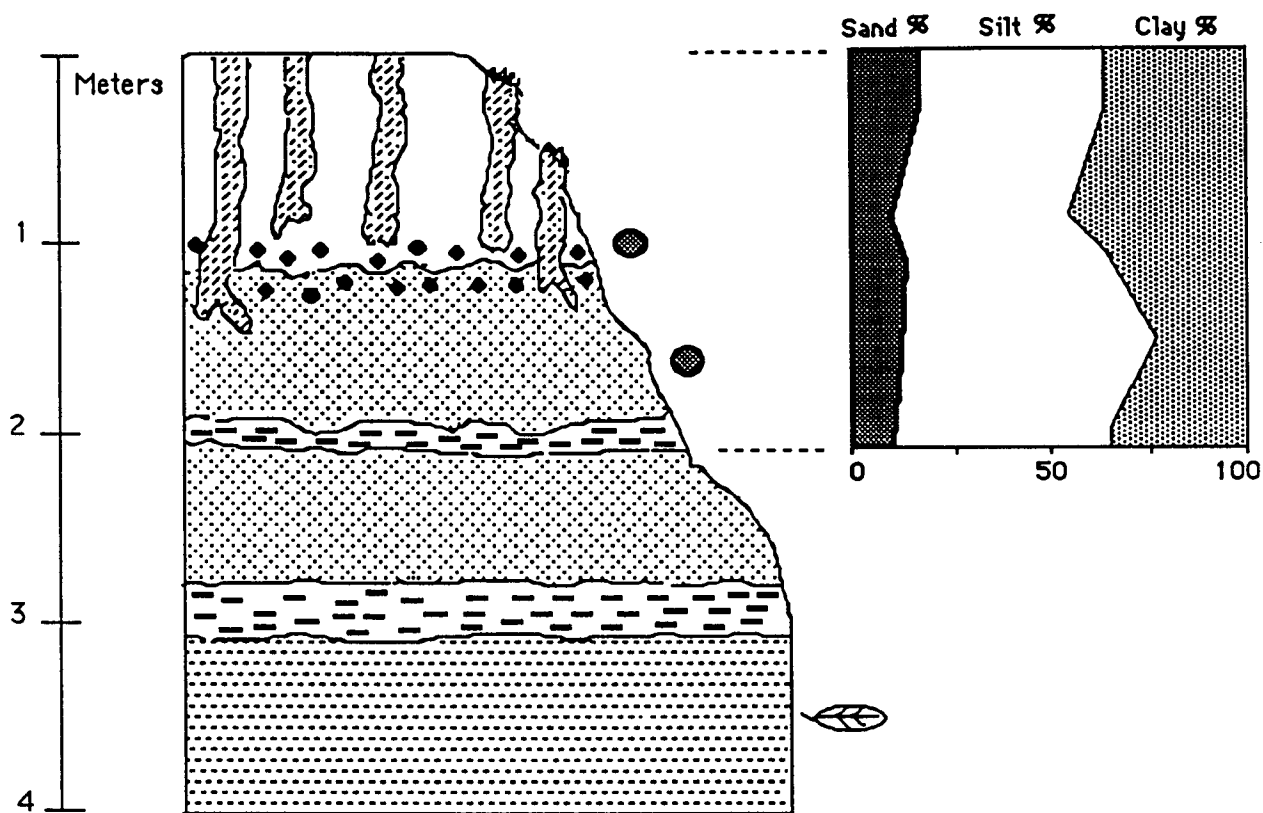


FIGURE 7-2. Lithostratigraphic sequence of Dakota Fm. (Cretaceous); Saline County, Kansas. Interpreted as a hydromorphic paleosol.

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Stop 8 Barton County Sanitary Landfill, 3.5 miles north and 3.5 miles east of Great Bend, Kansas: Soils developed in loess

Location: NW 1/4, SE 1/4, Sec. 12, T19N R13W, Barton County, Kansas
(USGS Ellinwood NW, KS 7.5 minute quadrangle)

Contributor: William C. Johnson and Feng Zhaodong

A 400,000-YR LOESS-PALEOSOL SEQUENCE

Barton County Sanitary Landfill is situated on the loess-mantled divide between the Arkansas River to the south and Cheyenne Bottoms to the north. Approximately 18m of loess overlie a ridge consisting of the Cretaceous Dakota Formation. Quaternary stratigraphy exposed in the landfill consists of the Loveland loess (18-6.4 m), Barton sand (6.4-4.2 m), Sangamon soil (4.2-3.4m), Gilman Canyon Formation (3.4-2.4m), Peoria loess (2.4-1.6m), Brady soil (1.6-0.8m), and Bignell loess with surface soil (0.8-0m). The exposure in the west end of the landfill, which we will examine, lacks the Bignell loess and surface soil, i.e., landfill operations have graded down into the late-Pleistocene/early Holocene Brady soil. Further, we will be able to view only the upper 10-12m due to liquid waste burial at the base of the original exposure.

This site is the southernmost and first of the two we will see on this trip that exhibit the pre-Loveland/pre-Illinoian through the Holocene loess and soil stratigraphy. Major differences and similarities will become apparent during our visit to the Eustis ash pit, the other site with similar stratigraphy.

Regional Stratigraphic Background

Since this is the trip's first stop to examine loess stratigraphy on the Great Plains, an overview of the stratigraphic units exposed in the western face (Loveland, Sangamon, Gilman Canyon Formation, and Peoria) is presented here. The Brady soil and Bignell loess will be discussed at the Bignell Hill stop, as this is their type locality and best exposure on the trip; the Barton sand, recognized only at this locality, is discussed in the site information.

Loveland Formation. The Loveland Formation is the most widespread pre-Wisconsin loess in the Midcontinent. Several investigators (e.g., Reed and Dreeszen, 1965; Ruhe, 1969; Willman and Frye, 1970; Ruhe and Olson, 1980) have described it throughout the Missouri, Mississippi and Ohio River basins. Further, it has been recognized south into Mississippi and Arkansas (McCraw and Autin, 1989). The Loveland, as a loess, has been far less studied (e.g., absolute chronology, geometry, mineralogical composition) than the Wisconsin loesses, particularly the Peoria.

Shimek (1909) first identified the Loveland (as a water-lain deposit) based upon exposures along the east bluff of the Missouri River just northeast of Loveland, Iowa, where it is underlain by "Kansan" till and overlain by Wisconsin loess. Several years later, the Loveland was identified and described in Nebraska by Lugin (1935) who identified both a valley phase (water-lain) and an eolian phase (loess). Condra and others (1950) separated the valley phase, naming it the Crete Formation. Since Leighton and Willman (1950) first designated the Loveland as an Illinoian-age loess, it has been assigned as such in both Kansas (Frye and Leonard, 1952) and Nebraska (Wayne, 1963; Reed and Dreeszen, 1965). Frye and Fent (1947)

identified the Loveland Loess in Kansas on the basis of comparisons made with the type section in Iowa and deposits recognized in Nebraska. More recently, the Loveland of Nebraska has been separated into three distinct units, or members, primarily on the basis of paleosol presence (Reed and Dreeszen, 1965; Schultz, 1968; Schultz and Martin, 1970). The original type section of the Loveland Formation was destroyed by a borrow pit for road construction, and a new type section (paratype) was designated at the end of the borrow pit area (Daniels and Handy, 1959; Bettis, 1990). The Loveland may be described as a yellowish-brown or reddish-brown eolian silt. Red hues increase toward the top of the formation due to development of the Sangamon soil within the uppermost Loveland. The thickest accumulations in Kansas (c. 16m) occur in the northcentral part of the state. A thinning in the loess occurs both southward and westward such that the distribution becomes discontinuous to the southwest. Several paleosols of varying degrees of development are present within the loess, indicating periodic stability or at least dramatically decreased rates of loess fall. The absolute age of the Loveland in the central Great Plains had been uncertain until recent work at Barton and Pratt County Sanitary Landfills in central and southcentral Kansas, although Oviatt and others (1988) reported TL dates of 130 ± 30 and 136 ± 30 ka on the upper part of presumed Loveland loess in northeastern Kansas. TL dating at the Loveland paratype section indicated that the Loveland Loess was deposited approximately $140,000 \pm 20,000$ years B.P. (Forman et al., 1992).

Sangamon Soil. This paleosol is strongly developed and occurs throughout the Midcontinent beneath deposits of the Wisconsin glaciation and within deposits of the Illinoian glaciation or older deposits. States where the Sangamon soil has been recognized include Indiana (Hall, 1973; Ruhe et al., 1974; Ruhe and Olson, 1980), Illinois (Bushue et al., 1974; Follmer, 1979) where the type section is located (Follmer, 1978), Iowa (Simonson, 1941; Ruhe, 1956, 1969), Nebraska (Schultz and Stout, 1945; Thorpe et al., 1951) and Kansas (Frye and Leonard, 1952). In Kansas, the Sangamon soil is very well expressed, occurring throughout the state. Although the soil has received considerable attention in northeastern Kansas (Frye and A.B. Leonard, 1949, 1952; Tien, 1968; Caspall, 1970; Bayne et al., 1971; Schaetzl, 1986), it has been recognized throughout the state (Bayne and O'Connor, 1968) and recently studied here at the Barton County Sanitary Landfill in central Kansas (Feng, 1991). Historically it has been referred to as a "soil in the Sanborn formation" (Hibbard et al., 1944), the Loveland soil (Frye and Fent, 1947), and the Sangamon soil (Frye and Leonard, 1951). The color of the soil ranges from a vivid to pale reddish-brown, with a loss in color occurring westward. Regionally, the soil character varies according to parent material, local drainage and climate which prevailed at the time of pedogenesis. The soil occasionally contains sufficient clay to create a subtle bench on cultivated slopes. Schaetzl (1986) noted that the soil appears to have been a very strongly developed Ultisol or Mollisol.

The Sangamon soil was first used in a time-stratigraphic context to differentiate deposits of the Illionian and Wisconsin glacial stages (Leverett, 1899). An appreciable time span for regional landscape stability and soil formation are indicated by intense oxidation, deep leaching, and high clay accumulation. The main problem associated with the Sangamon soil is its diachronous upper and lower boundaries (Follmer, 1978, 1982, 1983; Johnson, 1986). To further confuse the time element, the lower 1-2 m (3.3-6.6 ft) of the early Wisconsin loess is typically weathered and forms a pedological continuum with the underlying Sangamon soil (Follmer, 1983), and most investigators, until recently, mistakenly included the former in the Sangamon profile. The Sangamon should perhaps be considered a "pedocomplex" rather than a single soil which developed under a unique environmental condition (Schultz and Tanner, 1957; Fredlund et al., 1985; Morrison, 1987). It apparently represents several paleosols welded together to form a complex that reflects significant spatial and temporal variation in environmental conditions and an appreciable time span. Laboratory data from exposures in central Kansas indicate the Sangamon soil was strongly weathered chemically, presumably under a warm, moist climate (Feng, 1991).

Because of apparent time transgressiveness, the age of the Sangamon soil is not precisely known. Follmer (1983) reported a radiocarbon age of $41,700 \pm 1100$ yr B.P. on plant material from the top of the Sangamon

in its type area in Illinois. Forman (1990) reported TL ages of 140 ± 20 and 70 ± 10 ka from loess below the Sangamon soil at two separate sites in Iowa and Illinois, and concluded the Sangamon soil is diachronous and may consist of multiple soils. Feng (1991) and Feng and others (1991) reported a TL age of about 70 ka in the lowermost part of the Sangamon soil exposed here in Barton County and associate it with marine isotope stage 3. Although Richmond and Fullerton (1986) assign Sangamon time to 132-122 ka (isotope substage 5e), they acknowledged reported ages (relative and absolute) ranging from early Illinoian to middle Wisconsin. Basal ages on the overlying Gilman Canyon Formation from Phillips County and elsewhere in Kansas and Nebraska provide a minimum age of about 35 ka for the Sangamon soil (Johnson et al., 1990). Also, Forman and others (1991) obtained TL and radiocarbon ages of 35-23 ka within the loess overlying the Sangamon soil at the Loveland paratype section in Iowa and the Pleasant Grove School section in Illinois.

Gilman Canyon Formation. The Gilman Canyon Formation, first recognized in Nebraska (Reed and Dreeszen, 1965), is an early late Wisconsin (cf. Farmdalian) loess. Equivalents of the formation have been recognized elsewhere: the Loveland Loess is buried by the Roxana silt from Minnesota and Wisconsin to Arkansas and by the Pisgah Formation in western Iowa (Bettis, 1990). The Gilman Canyon of Nebraska and Kansas is typically dark in color, silty, leached of calcium carbonate, and heavily enriched in organic carbon via pedogenesis. It was once considered to be the attenuated A horizon of the Sangamon soil (Thorpe et al., 1951; Reed and Dreeszen, 1965).

Reed and Dreeszen (1965) provide limited textural data and description of the Gilman Canyon Formation at the type section. Their description within the columnar section at the Buzzard's Roost exposures states (p. 62): "Upper 12 inches [31 cm] is medium dark gray, slightly humic, silt; middle 1 foot 1 inch [33 cm] is light brownish-gray silt; basal 3 feet 8 inches [1.12 m] is dark brownish-gray, humic, soil-like silt; entire thickness is noncalcareous . . . 5 feet 9 inches [1.75 m]." Although all of these attributes described at the type section appear representative of the formation as observed in Nebraska and Kansas, the bimodal distribution of humus is curious: this suggests the existence of two periods of relative stability, or low accumulation rates, and an intervening period of accelerated accumulation rates. Consequently, the Gilman Canyon Formation often appears as one or more cumelic A horizons that are developed within a variably to noncalcareous loess, usually no more than 1.2 m (4 ft) thick. In a section revealing an expanded, valley phase of the Gilman Canyon Formation, May and Souders (1988) recognized three distinct organic zones, each of which may represent a separate episode of pedogenesis. Two such zones have been recently observed by the author at the Eustis ash pit in southcentral Nebraska. If two or more distinct periods of soil formation did indeed occur regionally, they are obscured at many localities, likely due to bioturbation. Overall, the formation reflects a sufficiently slow rate of loess fall ($< .08\text{mm/yr}$) such that pedogenesis was operating more or less continuously, but with a decreased intensity at one or more times.

As expressed, the Gilman Canyon Formation is frequently overlain by 0.9-1.5m (3-5 ft) of leached loess which is considered to be basal Peoria Formation. Correlative with the Gilman Canyon and the overlying leached loess zone is the *Citellus* zone (a ground squirrel now recognized as the genus *Spermophilus*) of Nebraska (Condra et al., 1950). The leached zone is transitional between the well developed A horizon(s) in the Gilman Canyon and the calcareous Peoria Loess above, and probably reflects a sufficiently slow accumulation of Peoria Loess such that pedogenesis could keep pace only partially. A.B. Leonard (1951, 1952) supports the contention that the leached, or basal, zone was slowly accumulating, early Peoria Loess experiencing pedogenesis through inference that gastropods were originally present, but subsequently destroyed during weathering of the loess. Above the leached zone, the rate of accumulation of Peoria loess was sufficiently rapid (c. 0.6mm/yr) as to preclude any soil development.

McKay (1979a,b) noted in Illinois that radiocarbon ages on organic materials from within early Wisconsin loess ranged from 40,000 to 31,000 yr B.P., and extrapolated an age of 45,000 years B.P. for the initiation

of loess deposition. Further, McKay (1979a) placed the end of the Farmdalian at about 25,000 years B.P., i.e., when Woodfordian (Peoria) loess fall began. As noted above, Forman (1990) reports TL and radiocarbon ages of 35-23ka for the loess immediately above the Sangamon. Follmer (1983, p.141) indicated that about 5,000 years separates the first deposition of glacially-derived loess and the time of maximum Woodfordian glacial extent in Illinois at about 20,000 years ago. This 5,000-year period may coincide with that for the development of the leached zone overlying the well developed (organic-enriched) part of the Gilman Canyon. Radiocarbon ages obtained from the Roxana in the Upper Mississippi River Valley indicate the loess unit was deposited between 50 and 27 ka (Leigh, 1991; Leigh and Knox, 1993).

Radiocarbon ages from the Gilman Canyon Formation range from approximately 35,000 at the base to 20,000 yr B.P. at the top (May and Souders, 1988; Johnson et al., 1990). Figure 1 and Table 1 present currently known radiocarbon ages for the Gilman Canyon Formation of Nebraska and Kansas, and Figure 2 designates the locations from which the age data were obtained. The basal age of 35 ka agrees well with the time set by Richmond and Fullerton (1986a) for the beginning of the late Wisconsin. While Nebraska has several dated locations forming an arcuate pattern around the eastern and southern sides of the Sand Hills, data come from only three areas in Kansas - Phillips, Barton, and Pratt Counties. The ages in Kansas do show, however, good agreement from the southcentral to the northcentral part of the state and with those from Nebraska. Given the radiocarbon time control and stratigraphic information currently available for the Gilman Canyon Formation within Kansas and Nebraska, it is clear that the associated soil(s) is a *geosol*, i.e. a laterally traceable, mappable, pedostratigraphic unit with a consistent time-stratigraphic position (Morrison, 1965; North American Commission on Stratigraphic Nomenclature, 1983, p. 865). The entire formation may be considered a *geosol*, but because of the possibility for the existence of two or more identifiable cumelic A horizons merged or welded together, it may ultimately be considered a *composite geosol*.

Limited paleoenvironmental data are emerging for the Gilman Canyon Formation. The use of humate-derived $\delta^{13}\text{C}$ and opal phytolith data are showing promise at another stop, the Eustis Ash Pit in southcentral Nebraska (see discussion). Also, interpretation of the fossil pollen assemblage in a core extracted from Cheyenne Bottoms, the large marsh immediately to the north of the landfill, indicates mesic conditions in the marsh and an upland vegetation of grass and sage with scattered trees in the valley and along escarpments during the period from approximately 30 to 25 ka (Fredlund, 1991). The Farmdalian-Woodfordian transition, approximately 25-24 ka, was characterized by increased aridity. The Muscotah Marsh fossil pollen record of northeastern Kansas reflects a mosaic of deciduous forest and prairie (Gruger, 1973; Fredlund and Jaumann, 1987). Regionally, the Farmdalian grasslands were apparently found as far east as Iowa (Baker and Waln, 1985) and north to the Sand Hills region of Nebraska (Fredlund and Jaumann, 1987).

Peoria Formation. Leverett (1899) first proposed the name Peoria for an interglacial period between the Iowan and Wisconsin glacial stages. When Alden and Leighton (1917) demonstrated the Peoria was younger than Iowan, usage shifted to that of a loess, rather than a weathering interval. Within the Midcontinent, several names have been used for post-Farmdalian loess. Ruhe (1983) prefers to use the term "late Wisconsin loess" because of the uncertainties in stratigraphic equivalency from one region to another. The Peoria Formation is typically an eolian, calcareous, massive, light yellowish-tan to buff colored silt that typically overlies the Loveland Formation or an approximate equivalent of the Gilman Canyon Formation.

Ruhe (1983) notes three major features of late Wisconsin (Peoria) loess: it thins downwind from the source area, decreases in particle size systematically away from the source area, and is strongly time-transgressive at its base. The latter feature is unresolved and results in correlation problems. Ruhe (1969) realized a decrease in the age of the soil under the loess from 24,500 years B.P. near the Missouri River to about

19,000 years B.P. eastward across southwestern Iowa. A decrease from 25,000 to 21,000 years B.P. was noted for the base of the loess along a transect in Illinois (Kleiss and Fehrenbacher, 1973). The top of the loess also seems to be time-transgressive, ranging from about 12,500 years B.P. in Illinois (McKay, 1979b) to 14,000 years B.P. in central Iowa (Ruhe, 1969). Despite the amount of attention given the Peoria in Kansas, the source of the silt is not completely certain. Upon a review of the available data, Welch and Hale (1987) conclude that a single source was not likely for all loess deposits in Kansas, and that the loess was derived from a combination of three sources: glacial outwash river floodplains, present sand dune areas, and fluvial and eolian erosion of the Ogallala Formation.

A.B. Leonard (1952) subdivided the Peoria loess of Kansas into three zones on the basis of the molluscan fauna assemblages present. The *basal zone* is equivalent to the leached interval above the Gilman Canyon Formation and is void of molluscan material. The *lower molluscan zone*, or Iowan, produced an assemblage containing 14 species, two of which are diagnostic of the zone. A *transitional zone*, located between the upper and lower faunal zones, contains elements of both assemblages and does not imply any abrupt change in the depositional environment, although the depositional rate may have slowed somewhat. The *upper molluscan zone*, or Tazewellian, contains 26 species, 14 of which do not occur in the lower zone.

Although readily visible stratigraphic breaks such as the Jules soil recognized in Illinois (Frye and Willman, 1973; Frye et al., 1974; Ruhe, 1976; McKay, 1979a,b) and the soil zones in Iowa (Daniels et al., 1960; Ruhe et al., 1971) have not yet been identified in Kansas and adjacent Nebraska, evidence of one or more stable or vegetated surfaces is common. The only indication of soil development recognized is that of a Bt horizon in the Medicine Creek valley (May and Holen, 1993); interestingly, the soil has a probable Paleoindian association (May, 1991). The most common line of evidence for a discontinuity(ies) in Peoria Loess deposition is that of plant remains, usually outcropping as lenses. Many of the age determinations were made from *Picea* remains (see Table 12-3), indicating a cool, moist environment. Although radiocarbon data document the burial of vegetative material throughout the Woodfordian, two temporal clusters or modes of ages appear from the limited data: one 18-17,000 yr B.P. and another 14-13,000 yr B.P. The former time interval represents the last glacial maximum and the latter the time of major deglaciation (Ruddiman, 1987). Interpreting ice core data from Greenland, Paterson and Hammer (1987) record a dramatic decrease in atmospheric dust content from about 13,000 yr B.P.; this period of reduced atmospheric dust may relate to the time of relative surface stability and tree establishment. Regional geomorphic data also support the existence of a hiatus at this time. May (1989), identifies deposition of the Todd Valley Formation in the South Loup River of central Nebraska at about 14,000 yr B.P., which is subsequently buried by loess. Further, Martin (1990) identifies entrenchment in the Republican River of southcentral Nebraska at about 13,000 yr B.P., after which valleys were filled with late Peoria Loess.

Site Data and Discussion

Stratigraphy. The composite stratigraphy of the landfill exposures consists of eight readily distinguishable units:

- Surface soil (0.0-0.2m; truncated by landfill operations): the Crete silt loam (fine, montmorillonitic, mesic Pachic Argiustolls);

- Bignell loess (0.2-0.8): slightly dark, yellowish-brown, calcareous silt with calcium carbonate concretions and manganese specks;

- Brady soil (0.8-1.6): dark, grayish brown silt with calcium carbonate concretions in the B horizon;

- Peoria loess (1.6-2.4): yellowish-tan to gray-tan calcareous silt, with mottles, iron stained grass root channels, and charcoal flecks;

- Gilman Canyon Formation (2.4-3.4): dark grayish brown silt, dominated by an organic-rich soil;

- Sangamon soil (3.4-4.2): reddish silt, noncalcareous, well-developed structure, clay coatings and

oxide staining;

Barton sand (4.2-6.4): slightly reddish, calcareous, well-sorted eolian sand.

Loveland land pre-Loveland loess (6.4-18): yellowish silt, massive structure with four caliche zones at 6.4-8.2, 9.2-11.4, 12.4-14.8, and 16.5-18.0 m.

Age Data. Both ^{14}C and TL ages are available for the site (Table 2). The single ^{14}C age of $24,360 \pm 570$ yr B.P. from the upper Gilman Canyon Formation is consistent with those from other localities (Table 1; see Eustis Ash Pit discussion). Two others, $10,550 \pm 150$ and 9820 ± 110 yr B.P., from the lower and upper 5 cm, respectively, of the Brady soil were obtained from the eastern exposure, now covered with backfill debris. These ages correlate with the Brady ages from other localities (see Bignell Hill discussion) that of $9690 \pm 200/210$ (Dic-3323) obtained by Fredlund (1991) from a Cheyenne Bottoms' core. The apparent base of the Sangamon soil was TL dated at 69 ± 6.4 ka, which is in close agreement with the TL date of 77 ± 8 ka from the Sangamon of Illinois (Forman et al., 1992). Gilman Canyon time may appear to have begun 40-50,000 yr B.P., which leaves about 20,000 years for Sangamon time. The bottom of the Barton sand, or top of the Loveland loess, was dated to 92 ± 7 ka. The TL date of the upper carbonate-depleted loess layer is 193 ± 22 ka. The lowest/oldest TL age of 260 ± 25 ka comes from the boundary between the second (down) carbonate-enriched zone and the second carbonate-depleted zone. Although the base of the loess was not dated here, the stratigraphy and inferred rate of sedimentation agree with a TL date of 416 ± 35 ka obtained from the Pratt County Sanitary Landfill about 85 km to the south (Feng, 1991). With rounding and inferred rates of sedimentation, ages on the four cycles of carbonate enrichment are 415-325, 325-250, 250-195, and 195-95 ka. These periods correspond chronologically to interglacials identified as marine ^{18}O stages 11, 9, 7, and 5, respectively. In an attempt to reinforce the chronology, additional TL samples are pending.

The question remains as to what is Illinoian and what is pre-Illinoian, i.e., where do we draw the time boundary. Richmond and Fullerton (1986) defined the Illinoian glacial stage as marine isotope stages 6-8 based on the terrestrial glacial record, whereas the Illinoian is alternatively defined as isotope stage 6 using the marine isotope surrogate of climate. The beginning and end of isotope stage 6 has been dated using the terrestrial record: Oviatt and others reported TL ages of 130 ± 30 and 136 ± 30 ka on the upper part of presumed Loveland loess in northeastern Kansas; Forman and others (1992) reported TL ages of 125 ± 25 and 165 ± 20 ka on the mid-upper and lower parts, respectively, of the Loveland loess in Illinois; isotopic dating of fossil coral reefs has placed the beginning of the Illinoian at 180-200 ka (Bender et al., 1979), and the end at somewhat less than 125 ka (Ku et al., 1990) or about 120 ka (Edwards et al., 1987a,b; Bard et al., 1990; Chen et al., 1991). Martinson and others (1987) derived an age of 130 ka for the boundary between marine isotope stages 5e and 6 (Sangamon/Loveland boundary) using the Milankovitch orbital tuning technique. An age of 74 ka for the isotope stage boundary 4-5a derived by Martinson and others (1987) should also be considered as a candidate for the Loveland/basal age of the Sangamon (6-5e contact) due to the ages obtained from the terrestrial record, i.e., TL dates of about 70 ka associated with the lower Sangamon/upper Loveland in Kansas (Feng et al., 1993) and Illinois (Forman et al., 1992). Given the limited age data and multiple approaches/definitions for determining temporal boundaries of the Illinoian (Loveland), a lower age of 190 ka and an upper age of 70 ka seem reasonable at this time. Consequently, Sangamon soil development would have occurred from about 70 to 50-45 ka (Gilman Canyon Formation @ 50/45-21 ka), and the upper cycle of carbonate enrichment/pedogenesis (195-95 ka) would represent the lower/basal Loveland/Illinoian and the others would be of pre-Loveland/pre-Illinoian age.

Sediment Character. Selected sediment data is represented in Figure 3. Sand content is anomalously high in the middle of the section (4.5-6.6 m), whereas clay has peaks in the Loveland loess, Peoria loess and Brady soil. The four carbonate zones are identified and enumerated on the figure. The positive correlation between clay and carbonate, especially from 6.4 to 18 m, suggests that the carbonate might have been formed pedogenically. The clay layers are slightly above the corresponding carbonate zones, i.e., clay translocation occurred a bit earlier than carbonate enrichment. Further, carbonate outlines but does not fill

root channels and exhibits a platy habit over nodules within each of the layers.

As a proxy for the degree of loess weathering or soil development, bulk magnetic susceptibility indicates that the weathering intensity was low when the clay and carbonate-enriched zones within the Loveland were formed (Fig. 3). The Barton sand is characterized by relatively high susceptibility, and the reddish Sangamon soil has the highest. Another indicator of weathering, the ratio of kaolinite to the amount of quartz in the clay fraction (Birkeland, 1984; Wang, 1982) indicates that the Brady soil, Gilman Canyon Formation, Sangamon soil and Barton sand were well-weathered. Also, the weakly expressed carbonate-rich zone LS2 and the two loess units LL1 and LL2 are characterized by higher ratio values.

Unstable compounds such as Na_2O , K_2O , MgO , P_2O_5 , and CaO are relatively sensitive to leaching or to effective accumulation by calcisol-forming processes in arid-semiarid environments (Mack et al., 1993). The Bignell loess produces small peaks in Na_2O and P_2O_5 (Fig. 4). The low values in Na_2O and others appear for the Brady soil, whereas Peoria loess is well expressed by high values in Na_2O , K_2O , and MgO . Na_2O , K_2O , MgO , and P_2O_5 are high in the upper and lower parts of the Gilman Canyon Formation where the soil is less well developed. Lowest values occur with the strongly weathered Sangamon soil, and extend downward into the Barton sand. Small peaks in Na_2O , K_2O , and MgO and a large peak in P_2O_5 in the middle of the Barton sand. Variable responses occur in the Loveland loess and associated soils/carbonate zones.

Stable compounds such as SiO_2 , TiO_2 , Fe_2O_3 , and Al_2O_3 (Fig. 5) are controlled primarily by the geologic background of the loess parent material. Excluding CaO , unstable compounds consist generally of < 10 %, and SiO_2 runs high (c. 80 %). SiO_2 is high in the Bignell loess and low in the Brady soil, Peoria loess, and Gilman Canyon Formation. High levels of SiO_2 occur in the Sangamon soil and the upper part of the Barton sand. It is also high in the carbonate-rich zones and low in the carbonate-poor zones. TiO_2 assumes a distribution similar to SiO_2 , whereas Al_2O_3 is negatively correlated since it is second highest in concentration. Fe_2O_3 generally follows the Al_2O_3 curve. From the data, it appears the Brady soil was moderately weathered both chemically and physically, whereas the Peoria loess was not weathered after deposition (high Al_2O_3 values from parent material).

The ratio of $\text{Al}_2\text{O}_3/\text{SiO}_2$ and others permit insight into the intensity of leaching and weathering, as well as geologic background of the parent material (Fig. 6). The ratios of $\text{Fe}_2\text{O}_3/\text{CaO}$ and $\text{Fe}_2\text{O}_3/\text{Al}_2\text{O}_3$ are used to assess the intensity of weathering, as is particularly notable in the Brady and Sangamon soils. Further, pedogenic imprints superimposed on the geologic background can be distinguished by comparing the magnetic susceptibility with Fe_2O_3 and $\text{Fe}_2\text{O}_3/\text{Al}_2\text{O}_3$.

The chemical characteristics indicate that the Brady soil was more strongly weathered than the Gilman Canyon soil, and the Sangamon was the most strongly weathered soil in the sequence. The Bignell loess was relatively poorly weathered, and the Peoria loess has neither traceable leaching or weathering. The Barton sand appears to have been moderately weathered. Within the Loveland and pre-Loveland loess, weathering indices are inversely related to the carbonate content, i.e., the intensity of weathering was low when pedogenic carbonate was accumulating.

Paleoenvironmental Information. The carbonate zones are likely analogous to the soils observed in the Loveland Loess at the Eustis ash pit and elsewhere in the region (Schultz and Martin, 1970; Fredlund et al., 1985; Frye and Leonard, 1951). The zones are temporally equivalent to the nonglacial, or warm marine isotope stages (Feng, 1991). Fossil pollen evidence indicates that the grasslands of the Great Plains expanded during the interglacials and contracted, or perhaps disappeared, during the glacial periods (Kapp, 1965, 1970; Fredlund and Jaumann, 1987). This suggests that the carbonate enrichment was a product of grassland pedogenesis, not unlike that of today. Fredlund and others (1985) extracted the grass opal

phytoliths contained within these multiple soil zones at the Eustis ash pit in southcentral Nebraska and found that the soil-forming periods were warmer and periods of dust accumulation cooler. The Barton sand, dated 95-70ka, represents the early Wisconsin stage. Magnetic susceptibility and the kaolinite/quartz in clay ratio indicates the sand was subjected to strong weathering. This slightly reddish eolian sand, deposited at a relatively high rate (0.22 mm/yr), appears to have been subjected to a relatively warm, dry (and windy?) climate. Data for the Gilman Canyon soil suggest it formed in a grassland environment. The leaching that had apparently occurred could have been a product of the relatively thin nature of the unit and the long duration of pedogenesis, rather than of high precipitation amounts. Peoria loess accumulated at a relatively rapid rate (0.3 mm/yr), with magnetic susceptibility and kaolinite/quartz in clay ratio data indicating a low weathering intensity. Well preserved grass stems in the form of charcoal or as undecayed subaerial remains suggest that the grasses were buried so rapidly that they were not oxidized or decomposed under the cool, moist conditions.

TABLE 8-1. Radiocarbon ages from the Gilman Canyon Formation of Nebraska and Kansas.

Location	Age (sample)
Nebraska	
WI, Winslow Hill ^b	23,000 ± 600 (I-2191)
	31,400 ± 1800, -1500 (I-2192)
YA, Yankee Hill	26,900 ± 1000, -900 (I-2189)
	34,900 ± 2100, -1700 (I-2190)
YO, depression near York (auger holes)	20,940 ± 240 (Beta-12273)
	23,740 ± 220 (Beta-12272)
	28,350 ± 610 (Beta-12274)
CE, Central City	19,640 ± 230, -240 (DIC-3118)
ON, depression near Ong (core holes)	22,590 ± 280 (Beta-24268)
	24,990 ± 430 (Beta-23457)
	26,140 ± 530 (Beta-23456)
SO, UNL South-Central Research Extension Center (core hole)	20,220 ± 330 (Beta-20102)
SP, Spring Ranch	21,140 ± 220 (Beta-20105)
	33,950 ± 290 (Beta-20104)
FU, Funk	24,260 ± 260 (Beta-40551)
BO, Bone Cove	26,330 ± 400 (PITT-0826) ^c
	26,260 ± 680 (Tx-5910)
NO, North Cove	27,850 ± 830 (Tx-7073) ^c
	29,950 ± 900 (Tx-7074) ^c
	30,700 ± 990 (Tx-7072) ^c
CO, Coyote Canyon	21,550 ± 500 (Tx-7294) ^c
NA, Naponee	19,770 ± 590 (Beta-33940)
	29,870 ± 1650 (Beta-33941)
JO, Johnson Lake	24,310 ± 530 (Beta-23343)
	29,450 ± 1250 (Beta-21298)
	29,790 ± 520 (Beta-25769)
	31,210 ± 1420 (Beta-25770)
	34,880 ± 1060 (Beta-23495)
EU, Eustis ash pit	25,090 ± 590 (Tx-6633) ^c
LA, La Sena site (Ft-177)	20,870 ± 1280 (Tx-6707)
	24,830 ± 1340 (Tx-6709)
LI, Lime Creek site (Ft-41)	27,970 ± 1190 (Tx-6774) ^c
BU, Buzzards Roost	21,290 ± 290 (Beta-26826)
	27,900 ± 1100, -1000 (I-2188)
	32,000 ± 2000, -1600 (I-1851)
OX, Oxford	28,710 ± 970 (Tx-6657)
MI, Mirdan Canal	20,660 ± 260 (Beta-28785)
	30,370 ± 750 (Beta-28784)
Kansas	
PH, Phillips County sanitary landfill	24,910 ± 770 (Tx-6746) ^c
KI, Kirwin railroad cut	25,500 ± 820 (Tx-6747) ^c
SI, Siebert pit	24,270 ± 750 (Tx-6624) ^c
	33,590 ± 2260 (Tx-6625) ^c
BA, Barton County landfill	24,360 ± 570 (Tx-5908) ^c
PR, Pratt County landfill	19,640 ± 410 (Tx-6449) ^c
	20,550 ± 590 (Tx-6622) ^c

a. Ages from eolian and alluvial phases.

b. Site designation on fig. 8-1.

c. Reported $\delta^{13}\text{C}$ correction.

Table 8-2. Age Data from Barton County Sanitary Landfill.

Lab. No.	Depth (m)	Age
Radiocarbon		
Tx-7045	1.00-1.05	9820 $\pm$ 110 yr B.P. (corrected)
Tx-7046	1.36-1.41	10,550 $\pm$ 150 (corrected)
Tx-5908	3.00-3.25	24,360 $\pm$ 570 (uncorrected)
Thermoluminescence		
XAL-77	4.15-4.25	69 $\pm$ 6.4 ka
XAL-78	6.55-6.65	92 $\pm$ 7
XAL-76	9.95-10.05	193 $\pm$ 22
XAL-75	12.45-12.55	260 $\pm$ 25

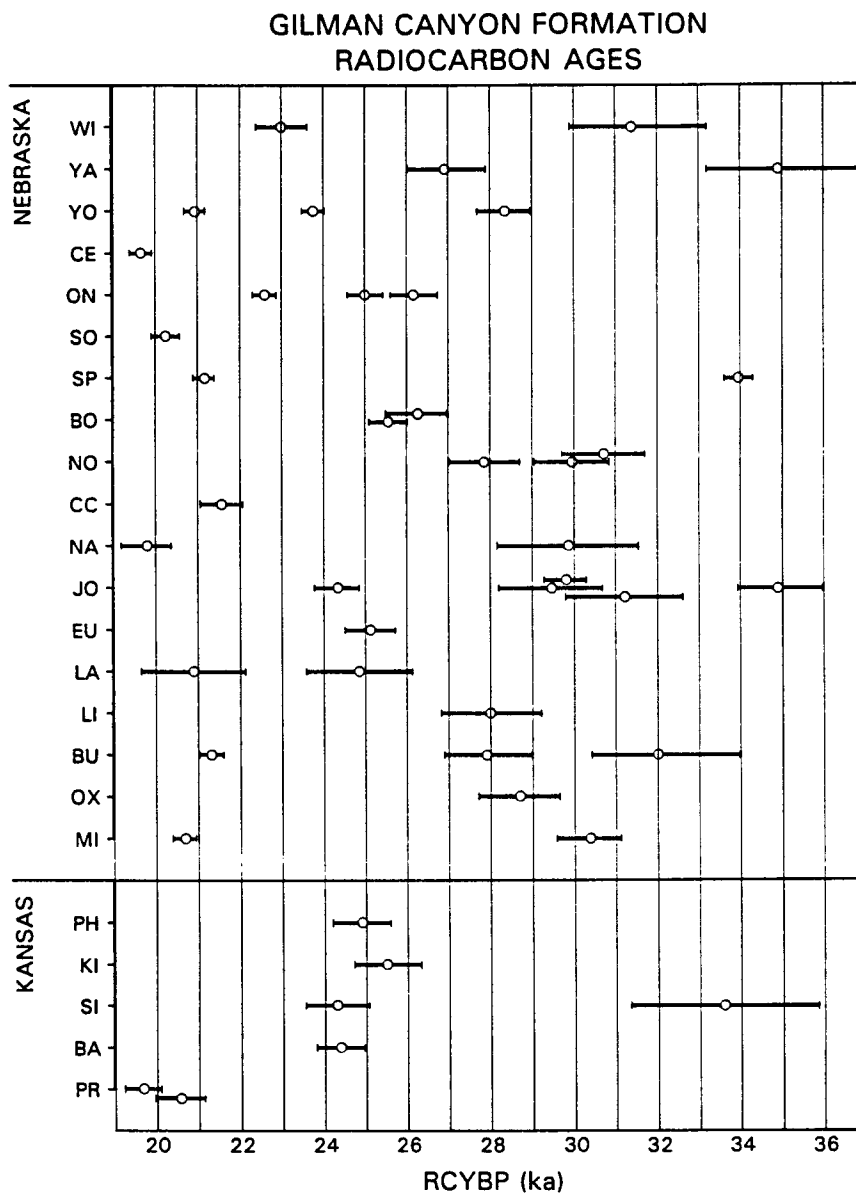


FIGURE 8-1. Temporal distribution of Gilman Canyon Formation radiocarbon ages. See Table 8-1 for site names.

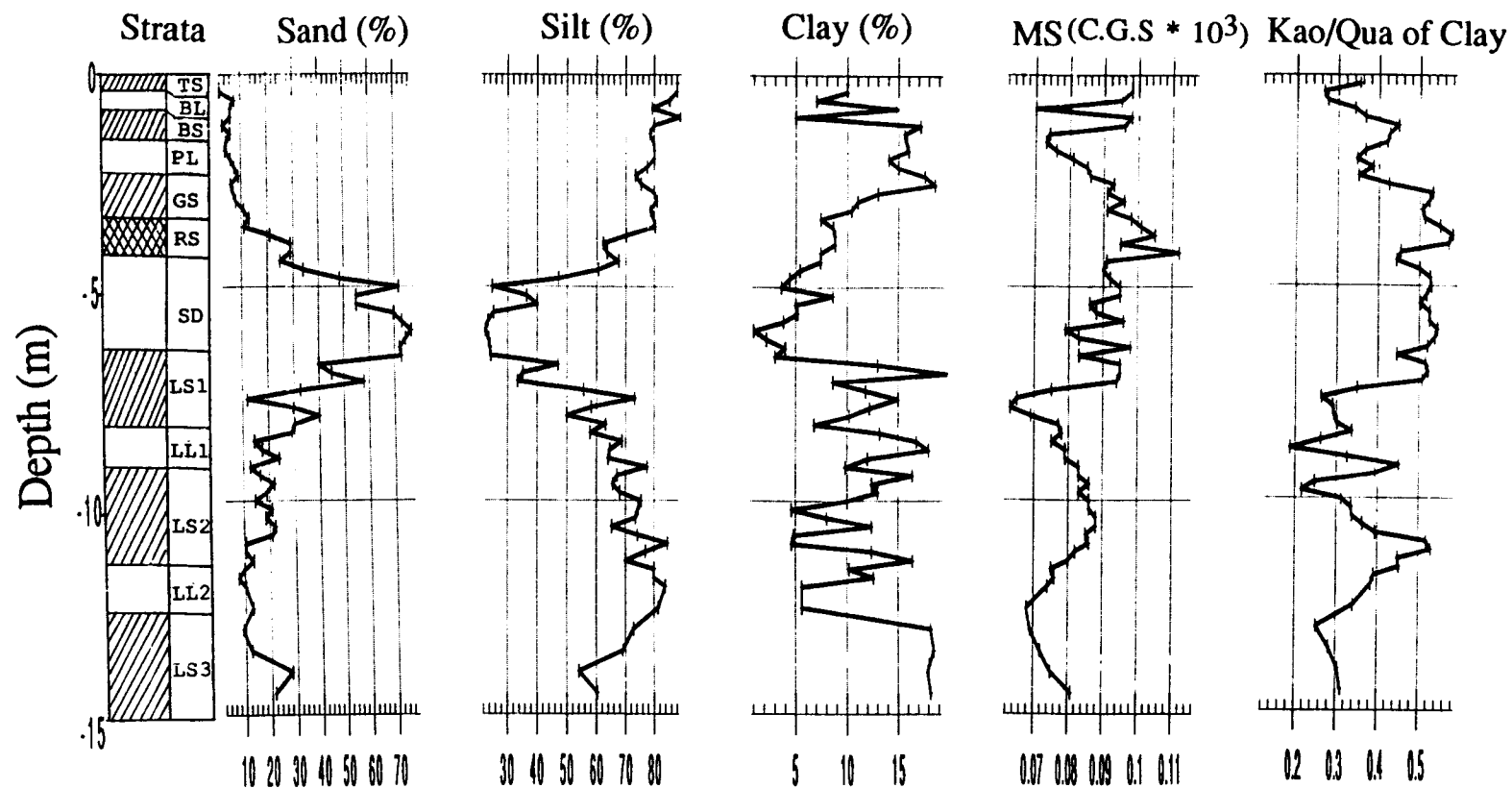


FIGURE 8-3. Stratigraphic distribution of particle sizes, magnetic susceptibility, and the ratio of kaolinite to quartz in the clay fraction. Unit designations are as follows: TS-surface soil, BL-Bignell loess, BS-Brady soil, PL-Peoria loess, GS-Gilman Canyon Formation, RS-Sangamon soil, SD-Barton sand, LS1-First/upper carbonate-rich loessal soil, LL1-upper intervening loess unit, LS2-second carbonate-rich loessal soil, LL2-second intervening loess unit, LS3-third carbonate-rich loessal soil.

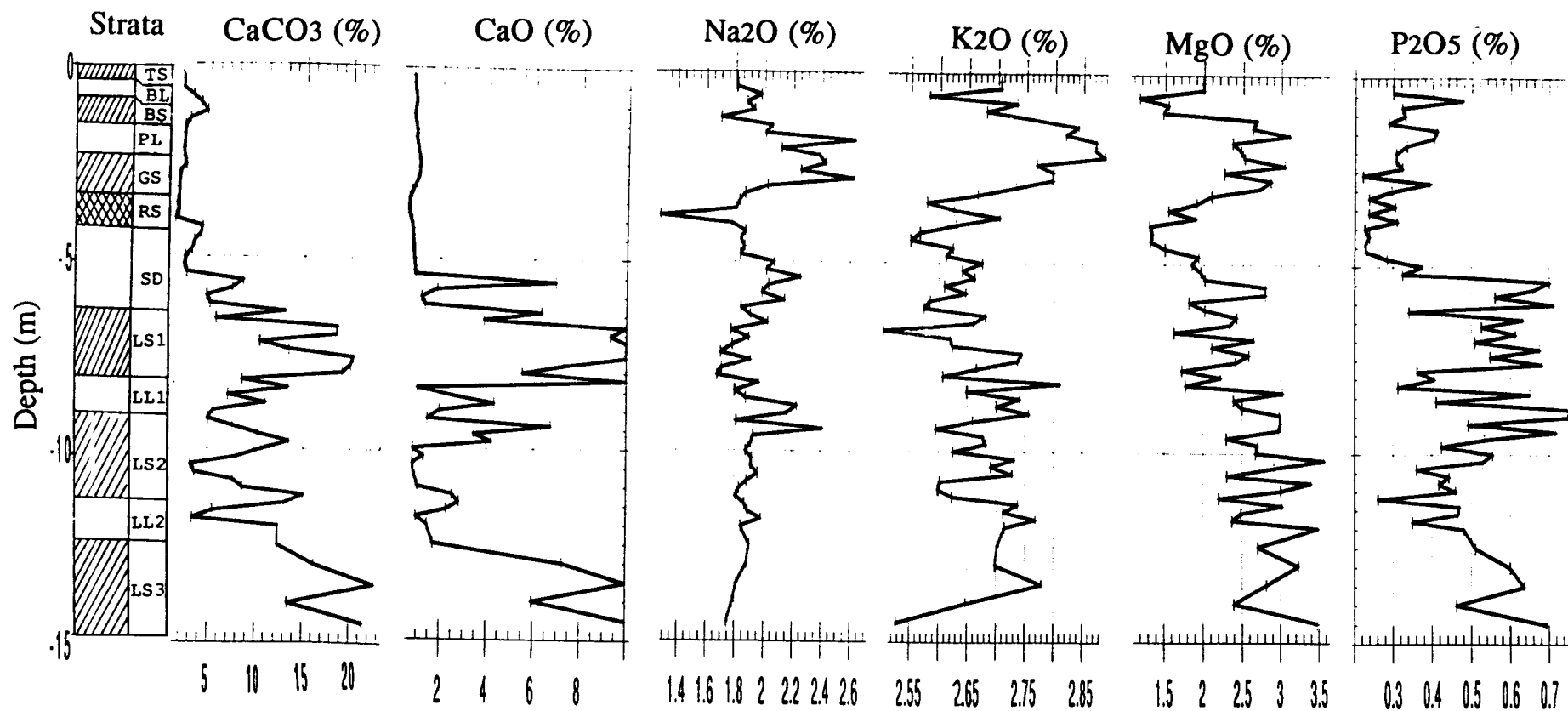


FIGURE 8-4. Stratigraphic distribution of relatively unstable compounds.

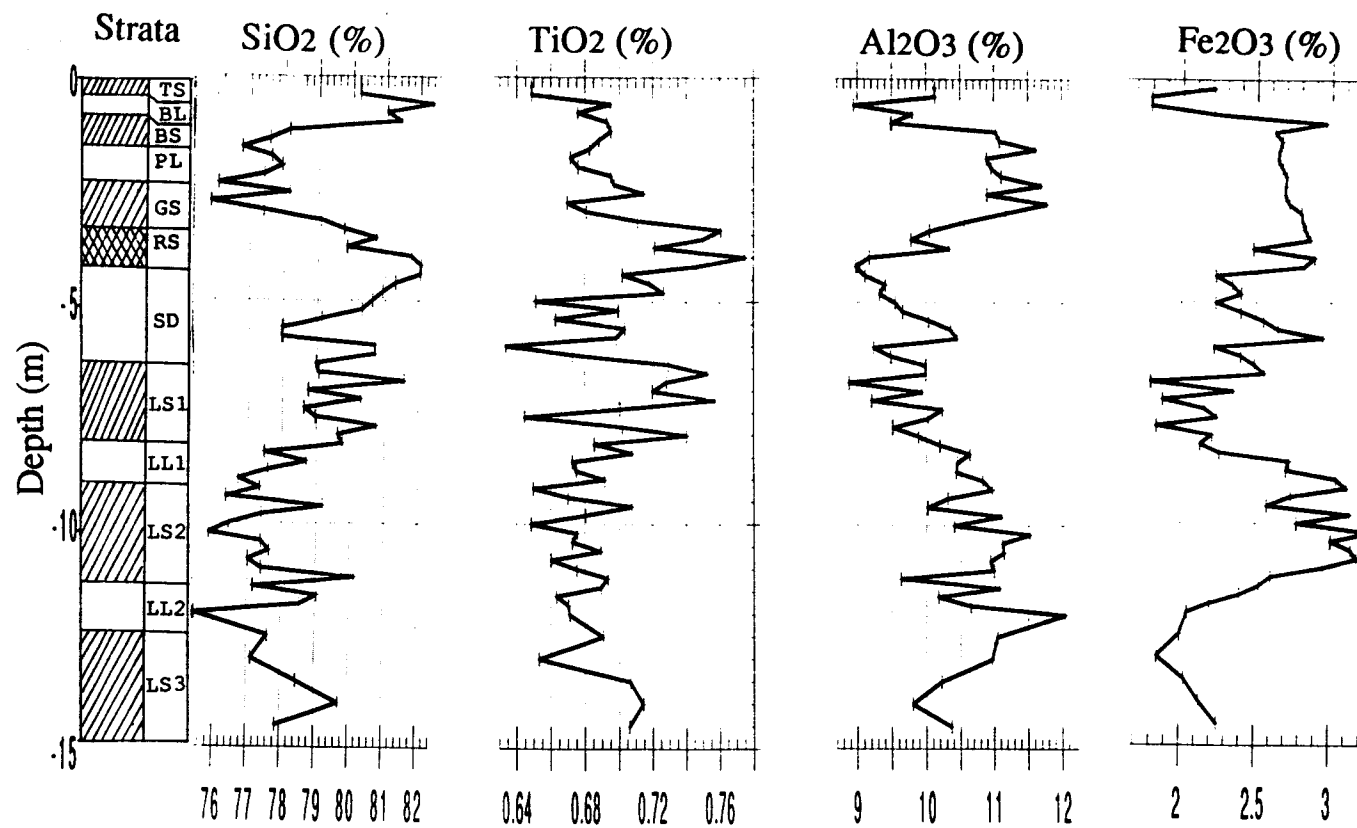


FIGURE 8-5. Stratigraphic distribution of relatively stable compounds.

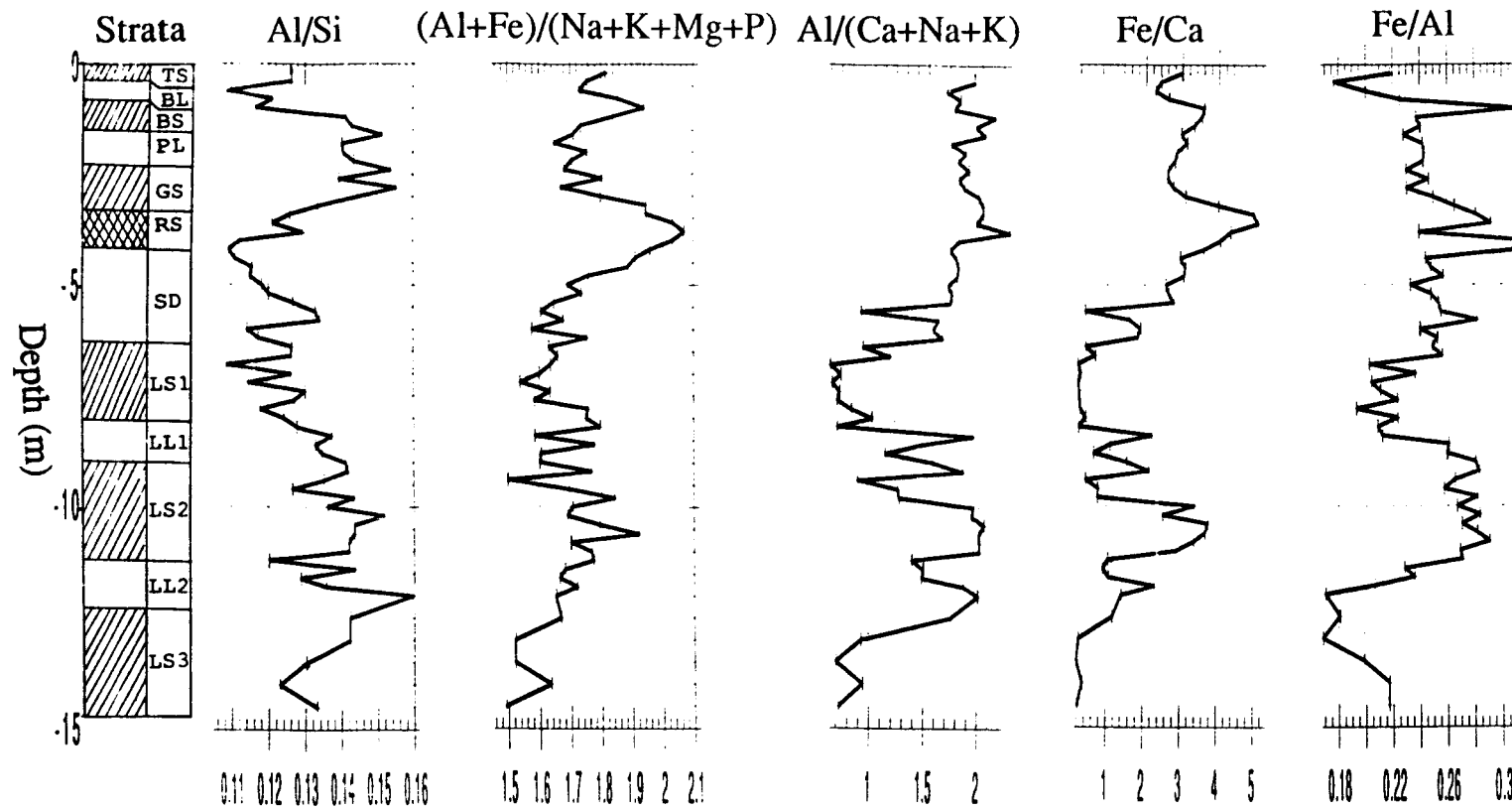


FIGURE 8-6. Stratigraphic distribution of ratios of various compounds. Note that the elemental designations are in oxide form.

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Stop 9 Wilson Ridge, in the vicinity of Lewis, Kansas: Sand-loess paleosol sequence

Location: S 1/2, Sec. 28/29, T25S, R18W, Edwards County, Kansas
(USGS Centerview, KS 7.5 minute quadrangle)

Contributor: Alan F. Arbogast

PALEOENVIRONMENTAL IMPLICATIONS OF BURIED SOILS IN A SOUTHCENTRAL KANSAS LUNETTE

Landscapes of the Great Plains consisting of unconsolidated sand (sand sheets, dunes) are particularly sensitive to climate change. Desertification of these areas was temporally and spatially episodic during the late Quaternary (Muhs, 1985, Ahlbrandt et al., 1983; Holliday, 1989). Inferred paleo-environmental conditions and the chronology of events on sand sheets and dunes can be reconstructed by analyzing the characteristics of buried soils, which indicate a period of paleolandscape stability.

Buried Soils at Wilson Ridge

Wilson Ridge is a lunette located on the southern margin of a small playa lake basin in the western part of the Great Bend Prairie in Kansas. The dune is about 1.5km long and 10m in height (Fig 1). The paleoenvironmental record reconstructed from the sediments and buried soils at the site indicates a complex history of sand and loess deposition and prevailing paleowind directions. Four trenches have been excavated across the dune axis. The surface soil and three buried soils exposed in the trenches indicate that periodic landscape stability has occurred at the site in the last 20,000 years (Fig 2).

The surface soil exposed in the trenches at Wilson Ridge is classified as a Thermic, Pachic Argiustoll that has formed in silt loam sediments. The buried soils are also well developed, consisting of Ab horizons overlying a series of Btb horizons.

Buried soils exposed in trenches 1 and 2 (Fig 2) are morphologically and chronologically equivalent. Each consists of a truncated, light brown (10YR6/4, dry) Ab horizon, formed in sandy sediments and buried by about 2.3 m of sand and silt and overlying a series of silt loam and clay loam Btb horizons. Relatively high concentrations of clay (ca. 15 %) in the Ab horizon of each exposure likely reflects atmospheric fallout of fine sediment during soil formation. Radiocarbon ages (^{13}C corrected) of $17,180 \pm 240$ yr B.P. (Tx-7824) and $16,520 \pm 200$ yr B.P. (Tx-7825) were obtained from the upper .05 m of the buried soil in trenches 1 and 2, respectively (Fig 3).

Trench 3 is located on the south flank of Wilson Ridge (Fig 2). Exposed in this trench is a well developed soil buried by about 2.6 m of sand and silt. The soil consists of a dark brown (10YR2/2, dry) Ab horizon, silty loam texture, that overlies a silt loam BAb horizon and 2 silty clay loam Btkb horizons. Truncation of the Ab and subsequent burial of the soil is evident because stringers of the Ab can be seen in the overlying bedded sands. Radiocarbon age dating of the upper .05 m of the soil exposed in trench 3 yielded an age of $10,360 \pm 100$ yr B.P. (Tx-7826, ^{13}C corrected) (Fig 3).

Trench 1 is located off of the south margin of Wilson Ridge (Fig 2). Exposed in this trench is a very well developed soil buried by about 1.2 m of silt. The soil consists of dark gray (10YR4/3, dry) Ab horizon, silty clay loam texture, overlying 5 Btb horizons totalling 1 m in thickness. A radiocarbon of $5,670 \pm 70$ yr B.P. (Tx-7827, ^{13}C corrected) was obtained on the upper .05 m of the soil (Fig 3).

Discussion

Using the morphology and radiocarbon ages of buried soils, the inferred paleoenvironmental history at Wilson Ridge can be reconstructed. The generalized position and ages of buried soils in the dune is illustrated in Figure 4.

The degree of soil development indicates that pedogenic intervals were of long duration. Periods of instability were probably short but intense, evidenced by thick accumulations of sand and silt burying the soils.

The radiocarbon ages obtained from Wilson Ridge and their laboratory number, trench location, source, and $\delta^{13}\text{C}$ ratios are listed in Table 1. Radiocarbon ages indicate the age of the soil at the time of burial. $\Delta^{13}\text{C}$ ratios can be used to infer vegetation (Krishnamurthy et al., 1982; Delaune, 1986) at the time of soil formation because C4 plants (warm, arid adapted) typically have a mean $\delta^{13}\text{C}$ value of -14‰ while C3 plants (cool, moist season plants) have an average $\delta^{13}\text{C}$ value of -27‰ (Krishnamurthy et al., 1982).

The 17,000 yr B.P. buried soil on the north flank and crest of the dune suggests that Wilson Ridge probably began to form about 20,000 yr B.P., a period of lunette formation recognized on the Southern High Plains (Reeves, 1965; Holliday, 1985). The orientation of the dune indicates that sediment from the adjacent playa was deflated by prevailing northwesterly winds.

During the late-Wisconsin glacial maximum ca 18,000 yr B.P. (Andrews, 1987), the dune temporarily stabilized and a soil formed. A $\delta^{13}\text{C}$ ratio of -11.9‰ suggests that warm season, edaphic plants dominated, a finding contradictory with known late Wisconsin, mesic climatic conditions.

Following landscape stability about 17,000 years ago, the soil was buried by sand, presumably during another period of increased aridity and prevailing northwesterly winds. This interval of instability was likely episodic, but no evidence of latest Wisconsin pedogenesis has been recovered from the site. In all probability, incipient soils formed during brief periods of stability, but were subsequently removed when sand remobilized.

Instability remained episodic until approximately 10,500 years ago, when a well developed soil formed on the south flank of Wilson Ridge. The age of this soil identifies it as the Brady geosol, terminal Pleistocene pedogenesis widely recognized on the Great Plains (Johnson and May, 1992). A $\delta^{13}\text{C}$ ratio of -14.2‰ suggests that edaphic plants continued to dominate.

After the Brady soil formed at Wilson Ridge, another period of landscape instability followed sometime in the Holocene, but probably during the Altithermal (Antevs, 1955), a period of increased aridity and dune formation widely recognized on the Great Plains between 9,000 and 6,000 yr B.P. (e.g., Ahlbrandt et al., 1983; Holliday, 1985; 1989). The stratigraphic evidence from trenches 1 and 3 (Fig 2) indicates that the prevailing winds shifted to the south/southwest during the Holocene since additional sandy deposits accumulated only on the south flank of the dune, burying the Brady soil.

At the end of the Altithermal, a well developed soil formed which is exposed in trench 1 (Fig. 2). A $\delta^{13}\text{C}$ ratio of -13.6‰ indicates that edaphic plants continued to dominate into the middle Holocene. Soil

formation terminated about 5,700 yrs B.P. when deposition of silt increased.

Results obtained elsewhere on the Great Bend Prairie (Arbogast, 1993) suggest that sand was remobilized episodically during the late Holocene at Wilson Ridge. Incipient soils probably formed as the dune stabilized, and were subsequently removed when sand remobilized. Well developed surface soils suggest that the dune has been stable for at least 1,000 years.

Summary

Wilson Ridge is a lunette that formed along the southern margin of a small playa-lake basin on the Great Bend Prairie. Three buried soils are contained in and adjacent to the dune, indicating that deposition of sand and silt was episodic during the late Quaternary. $\delta^{13}\text{C}$ ratios reveal that C4 grasses have dominated throughout the last 20,000 years. Intervals of pedogenesis were of long duration, witnessed by the degree of soil development. Periods of instability were probably short, but intense.

About 20,000 yr B.P. the dune began to form when northwesterly winds prevailed, deflating the adjacent playa. The dune stabilized and a soil developed ca. 18,000 yr B.P. during the glacial maximum. About 17,000 yr B.P., the soil was buried when sand remobilized. General instability, with periods of probable incipient soil development and truncation, persisted until the Pleistocene/Holocene climatic boundary 10,500 yr B.P. when a soil, chronologically equivalent to the Brady geosol, formed. This soil was buried about 10,300 yr B.P. when prevailing winds shifted to the southwest in the early Holocene.

Altithermal instability continued until about 6,000 yr B.P., when a soil formed preserved near the dune. Sand movement in the late Holocene has probably been episodic, with incipient soils forming during short-term, mesic periods and being removed when xeric conditions prevailed. A well developed surface soil indicates that Wilson Ridge has been stable for at least 1,000 years.

TABLE 9-1. Radiocarbon ages obtained from Wilson Ridge.

¹³ C-Corrected age (RCYBP)	Laboratory No.	Trench	Source	$\delta^{13}\text{C}$
5,670 $\pm$ 70	Tx-7827	4	humates	-13.6
10,360 $\pm$ 100	Tx-7826	3	humates	-14.2
16,520 $\pm$ 200	Tx-7825	2	humates	-11.9
17,180 $\pm$ 240	Tx-7824	1	humates	-11.9



contour interval = 3m

FIGURE 9-1. Topographic map of Wilson Ridge. Wilson Ridge is a lunette that formed on the southeast margin of the associated playa as sediments from the lake deflated by strong northwesterly winds during the late Wisconsin.

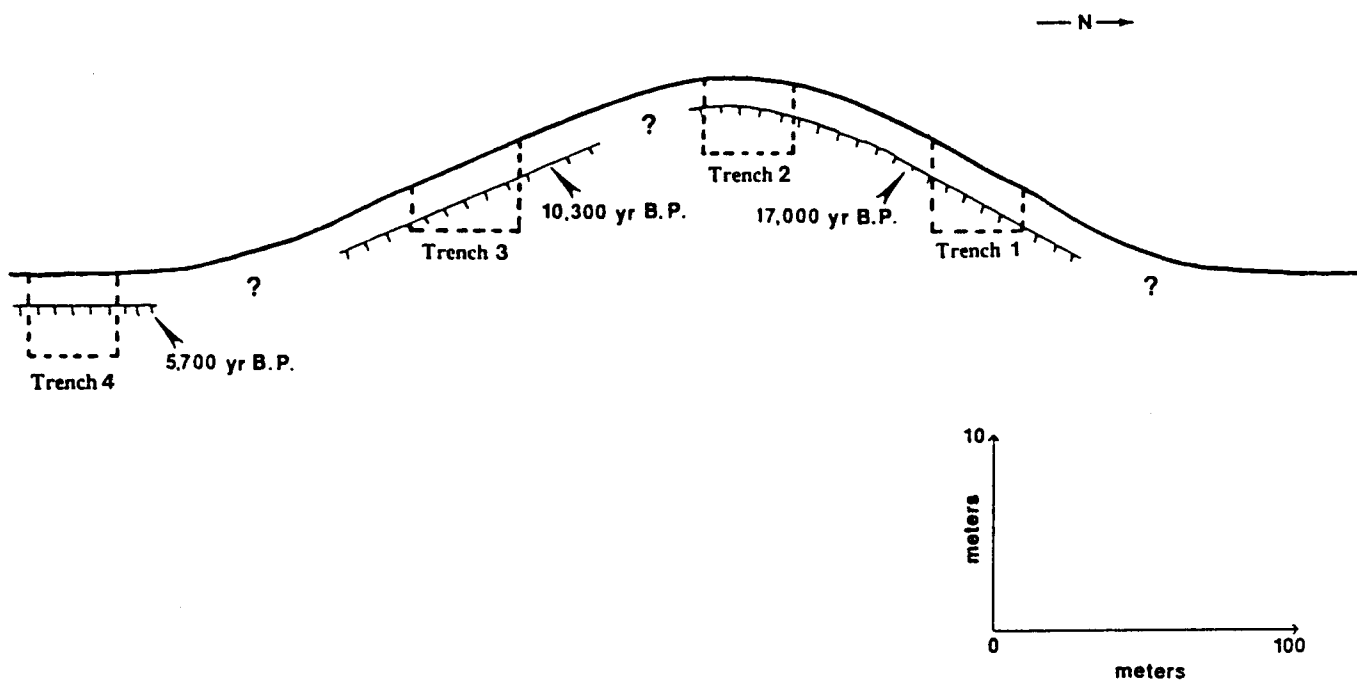


FIGURE 9-2. Generalized cross-section of Wilson Ridge showing the position of backhoe trenches and the location and ages of buried soils. Four trenches were excavated along the dune axis. A soil exposed in trenches 1 and 2 formed during the glacial maximum and was subsequently buried about 17,000 yr B.P. The buried soil in trench 3 formed at the Pleistocene/Holocene climatic boundary and was buried about 10,300 yr B.P. A buried soil exposed in trench 4 is terminal Altithermal in age and was buried about 5,700 yr B.P.

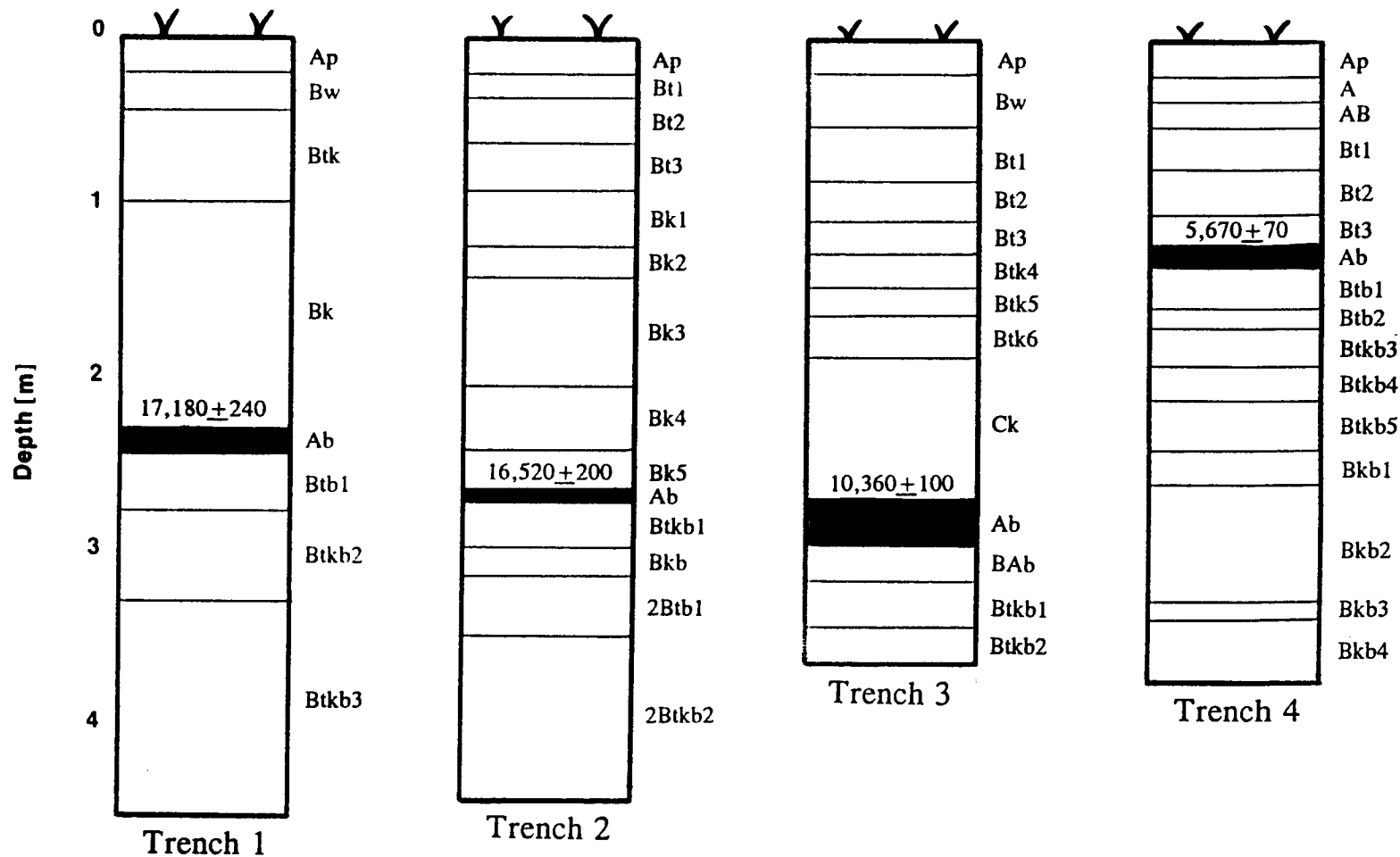


FIGURE 9-3. Trench stratigraphy at Wilson Ridge. Trench number increases from north to south. All soils are well developed, exhibiting a series of Btb horizons. A buried soil found in trenches 1 and 2 are morphologically and temporally equivalent, having formed during the glacial maximum and subsequently buried about 17,000 yr B.P. The buried soil in trench 3 formed at the Pleistocene/Holocene climatic boundary and was buried about 10,300 yr B.P. A buried soil exposed in trench 4 is terminal Altithermal in age and was buried about 5,700 yr B.P.

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Stop 10 Burdett, at the confluence of Buckner Creek and the Pawnee River

Location: SW 1/4, NW 1/4, Sec. 31, R20W, T21S, Pawnee County, Kansas
(USGS Burdett, KS 7.5 minute quadrangle)

Stop 11 Jetmore, approximately 28 miles upstream from that confluence

Location: SW 1/4, NE 1/4, Sec. 16, R25W, T23S, Hodgeman County, Kansas
(USGS Jetmore, KS 7.5 minute quadrangle)

Contributor: Rolfe D. Mandel

HOLOCENE SOILS AND LANDSCAPE EVOLUTION IN THE PAWNEE RIVER BASIN, SOUTHWESTERN KANSAS

A basin-wide study of Holocene terraces, alluvial fills, and buried soils was recently conducted in the Pawnee River valley of southwestern Kansas (Mandel 1990, 1992). The Pawnee River drains nearly 4,000 km² before joining the Arkansas River at Larned, Kansas (Figure 1). This study focused all levels of the drainage hierarchy, evaluating both the similarities among drainage elements composing the basin, and the differences in stratigraphy and chronology that occur in various parts of the basin. The results were used to infer causes of Holocene erosion, alluviation, and landscape stability, and to define their temporal and spatial relationships in the Pawnee River basin.

Two stops will be made in the Pawnee River basin: Stop 10 in the main valley of the Pawnee River, and Stop 11 along Buckner Creek, a major tributary of the Pawnee. Representative sections of Holocene alluvium with buried soils are exposed in cutbanks at both localities. The purpose of these stops is to compare the soil-stratigraphic record in a large valley with one in a small valley.

Description of Stop 10

Stop 10 is at the confluence of Buckner Creek and the Pawnee River (PR-4 in Figure 1). The valley floor is 6.5 km wide and consists of three landform/sediment assemblages: a modern floodplain (T-0), a low terrace (T-1), and a high, loess-mantled terrace (T-3). An intermediate terrace (T-2) identified further downstream was not observed in this portion of the Pawnee Valley. The T-0 surface is about 20 to 50 m wide and is separated from the T-1 surface by a steep 4-m-high scarp. The broad paired surface of the T-1 terrace dominates the valley floor and is separated from the adjacent T-3 terrace by a moderately sloping 3-m-high scarp. T-3 is a paired terrace with a fairly broad, gently sloping surface.

Both Buckner Creek and the Pawnee River have migrated laterally, cutting into the T-1 fill that separates the two streams near their confluence. An 8.5-m-high cutbank located along Buckner Creek about 300 m upstream from its juncture with the Pawnee River provides an opportunity to study a thick section of the T-1 fill. This cutbank exposure, designated as the Rucker Section, is described in Table 1.

Four stratigraphic units and three buried paleosols have been identified in the Rucker Section (Figure 2). The lowest stratigraphic unit (Unit IV) is at least 2.56 m thick and consists of fine-grained alluvium. A

truncated paleosol (Soil 4) is developed at the top of Unit IV. Soil 4 has a 40-cm-thick Btk horizon above a BCK horizon; the A horizon was stripped off by erosion before burial. The 4Btkb horizon is yellowish brown (10YR 5/4, dry) and is a silty clay loam. A few clay flows and many films and threads of calcium carbonate (Stage I+) are present on ped faces in the 4Btkb horizon. Humates from the upper 20 cm of Soil 4 yielded a radiocarbon age of 9320 ± 120 yrs B.P.

Unit III overlies Unit IV except where the former was stripped off by erosion at the west end of the section (Figure 2). Unit III is about 1.2 m thick and consists of silty alluvium. Soil 3 is developed at the top of Unit III and, like Soil 4, is a truncated paleosol. The A horizon of Soil 3 was removed by erosion, leaving a profile with Bk-Btk-BCK horizonation. The thick, strongly-developed paleosolum of Soil 3 stands out in the section. Soil colors range from yellowish brown (10YR 5/4, dry) in the 3Bkb horizon to brown (10YR 5/2, dry) in the 3Btkb horizon. The 3Bkb and 3Btkb horizons are characterized by silty clay loam texture, fine and medium subangular-blocky structure, and Stage II carbonate morphology. Also, distinct clay films were observed on ped faces in the 3Btkb horizon. Humates from the upper 20 cm of Soil 3 yielded a radiocarbon age of 7900 ± 160 yrs B.P.

Unit II mantles Unit III across the eastern two-thirds of the section and overlies Unit IV at the western end (Figure 2). Unit II is about 3.4 m thick and consists of stratified silt loam and fine sandy loam. Soil 2, which is developed at the top of Unit II, has a very dark A horizon above an ACK horizon. The 2Ab horizon is 40 cm thick, dark gray (10YR 4/1, dry), and leached of carbonates. A radiocarbon age of 1660 ± 70 yrs B.P. was determined on humates from the upper 20 cm of this horizon. The 2C horizons are pale brown (10YR 6/3, dry) to very pale brown (10YR 7/3, dry) and show little evidence of secondary carbonate accumulation. The alluvium composing the lower 3 m of Unit II is calcareous based on reaction to HCL. Horizontal bedding is clearly visible throughout the 2C2 horizon, with individual laminae < 5 mm thick collectively making up larger beds 1 - 10 cm thick.

Unit II is mantled by Unit I, except where gullying has removed the latter (Figure 2). Unit I is 1.05 m thick and consists of stratified fine-grained alluvium. The modern surface soil (Soil 1) at the top of Unit I is a Haplustoll (Mollisol) with A-ACK horizonation. The A horizon is 38 cm thick and is a dark grayish brown (10YR 4/2, dry) to grayish brown (10YR 5/2, dry) silt loam. The C horizon displays prominent bedding, with texture of strata ranging from silty clay loam to very fine sandy loam. Preservation of bedding at shallow depths combined with weak surface-soil development suggest that Unit I is very young (Historic?).

Mean residence times determined on organic carbon from buried soils exposed in the section provide a chronologic framework for Holocene aggradation, erosion, and landscape stability within the main valley of the Pawnee River at Locality PR-4. Specifically, Unit IV aggraded sometime before 9320 ± 120 yrs B.P., which is the radiocarbon age of humates from Soil 4 at the top of this unit. Soil 4 was truncated and subsequently mantled by alluvium composing Unit III sometime after ca. 9320 yrs B.P., but before ca. 7900 years B.P. Development of Soil 3 at the top of Unit III was underway by at least 7900 ± 160 yrs B.P. This paleosol was truncated and then mantled by alluvium composing Unit II sometime between ca. 7900 and 1660 years B.P. Mean residence time for organic carbon at the top of Soil 2 indicates that the surface of Unit II was stable by at least 1660 ± 70 yrs B.P. Aggradation of Unit I and concomitant burial of Soil 2 occurred sometime after ca. 1660 yrs B.P.

The episode of stream incision that converted the late-Holocene floodplain into a terrace (T-1) probably took place soon before ca. 1600 years B.P., thereby isolating the T-1 surface from frequent low-magnitude floods. This isolation would account for development of Soil 2 at the top of Unit II. As noted earlier, Unit II appears to be mantled by recent overbank deposits composing Unit I. If the overbank alluvium is Historic in age, it was deposited on the T-1 surface at or before A.D. 1951, which is the last time that

floodwaters covered the T-1 terrace at Locality PR-4 (Don Rucker, personal communication, 1988). According to local landowners, both the Pawnee River and Buckner Creek have been downcutting since about A.D. 1970.

Description of Stop 11

Stop 11 (Locality BC-1) is situated in Buckner Creek Valley approximately 46 km upstream from the confluence of that creek with the Pawnee River (Figure 1). Buckner Creek is a 4th-order stream and has a gradient of 3.3 m/km at this locality. The valley trends east-west and is asymmetric, with steep north-facing slopes and gentle south-facing slopes.

The valley floor at Locality BC-1 is 50 to 200 m wide and consists of two landforms: the modern floodplain (T-0) and a low terrace (T-1). T-1 is a paired terrace throughout most of this locality and is separated from T-0 by a short, steep 3-m-high scarp. In some sections of the valley, however, the T-1 terrace has been removed as a result of lateral migration by Buckner Creek. Hence the side-slopes of the valley descend to the T-1 surface or merge with the modern floodplain.

Buckner Creek has deeply incised its valley fill, creating many steep cutbanks throughout Locality BC-1. A survey of cutbanks resulted in the discovery of site 14HO306. Buckner Creek migrated into a remnant of the T-1 terrace at 14HO306, creating a 7.2-m-high cutbank that reveals multiple buried paleosols and several prehistoric cultural horizons in the terrace fill (Figure 3). The section at 14HO306 is typical of T-1 fill exposed in cutbanks along the middle and upper reaches of Buckner Creek.

Three stratigraphic units and five buried paleosols were identified in the section at site 14HO306 (Figure 3). The lowest unit (Unit III) is 3.18 m thick and consists of gravelly fine sand at its base grading upward to silt loam. A paleosol (Soil 6) at the top of Unit III has a thick, dark Ak horizon above a thin Bk horizon (Table 2). The 6Akb horizon (6Akb1 + 6Akb2) is 77 cm thick and is a dark grayish brown (10YR 4/2, dry) to grayish brown (10YR 5/2, dry) silt loam. Carbonate morphology is weak in the 6Akb horizon (Stage I). However, calcium carbonate content steadily increases with depth below the 6Akb horizon, reaching a maximum of 8.5% in the 6Ck horizon (Table 3). The 6Bkb horizon is 52 cm thick and is a brown (10YR 5/3, dry) silt loam. Films and threads of calcium carbonate are common and there are a few small, hard carbonate concretions (Stage II) in the 6Bkb horizon.

Cultural materials, including chert flakes, fire-cracked rocks, and burned bones are present in the upper 15 cm and lower 10 cm of the 6Akb1 and 6Akb2 horizons, respectively (Figure 3). Although no culturally-diagnostic artifacts were recovered from Soil 6, charcoal from a hearth 5 to 10 cm beneath the surface of the 6Akb1 horizon yielded a radiocarbon age of 2620 ± 220 yrs B.P. In addition, a radiocarbon age of 2820 ± 260 yrs B.P. was determined on charcoal from a zone of fire-cracked rocks and burned bones 70-75 cm below the surface of the 6Akb1 horizon. The great thickness of Soil 6, plus the presence of two Late Archaic cultural levels in the 6Akb horizon suggest that soil formation was accompanied by slow aggradation between ca. 2800 and 2600 B.P. Hence, Soil 6 appears to be a "cumulative soil" (cf. Nikiforoff, 1949).

Unit II is 2.18 m thick and consists of calcareous, loamy alluvium. Soils 5 and 4 are developed in the middle and at the top of Unit II, respectively (Figure 3). Soil 5 has a thin, faint Ak-Ck profile that is difficult to trace across the exposure. Although Soil 4 also lacks a B horizon, it is characterized by a thick, dark A horizon that clearly stands out in the section. The 4Akb horizon (4Ab + 4Akb) is 61 cm thick and has a silt loam texture. Carbonate content steadily increases with depth in Soil 4, ranging from 0.4 % in the 4Ab horizon to 11.1 % in the 4Ckb horizon (Table 3). Carbonate morphology reflects this trend, increasing from Stage I in the 4Akb horizon to Stage II in the 4Ckb horizon.

Several hearths and lithic artifacts and many burned and unburned bones, are present in Soil 4. The artifact assemblage recovered from this cultural component includes a projectile point characteristic of the Keith variant of the Plains Woodland period. Radiocarbon assays from the High Plains place the Keith variant from ca. 1600 to 1200 yrs B.P. (Grange, 1980: 126). A radiocarbon age of 1360 ± 60 yrs B.P. determined on humates from the upper 25 cm of Soil 4 falls within this period.

Unit I, which is at the top of the section, is 1.76 m thick and consists of calcareous, fine-grained alluvium. Two paleosols (Soils 2 and 3) with thin A horizons are developed in Unit I (Figure 3). The Roxbury silt loam, which is developed at the top of Unit I, has A-Bk-BCK horizonation. This surface soil is a Cumulic Haplustoll and is welded into Soil 2. Soil welding helps explain the dark zone within the upper 15 cm of the Roxbury's Bk horizon. The Bk horizon is 40 cm thick and is characterized by weak carbonate morphology (Stage I+).

With the exception of the Roxbury silt loam (Soil 1), soils 4 and 6 are morphologically the best expressed soils in the section at 14HO306. Two dark, overthickened A horizons resembling soils 4 and 6 were observed in cutbank exposures of T-1 fill throughout Buckner Creek Valley. Exposures of T-1 fill in many other small valleys in the Pawnee River basin also display two distinct A horizons among the buried soils. An earlier study (Mandel, 1988) suggested that soils 4 and 6 are pedo-stratigraphic markers and informally named them the Buckner Creek and Hackberry Creek paleosols, respectively. The cutbank exposure at 14HO306 has been designated as the type section for both of these paleosols (Mandel, 1990, 1992).

Two cores were taken from T-1 fill near site 14HO306: Core 1 on the east side of Buckner Creek and Core 2 on the west side (Figure 4). The stratigraphy observed in these cores indicates that depositional units and buried soils exposed in the cutbank extend laterally away from the modern channel of Buckner Creek. The T-1 fill is inset against bedrock along the margin of the valley floor.

Exposures in gullies reveal that the erosion surfaces cut into bedrock immediately upslope from T-1 terraces are mantled by a thin veneer of well-rounded alluvial gravel (Figure 4). Much of the gravel is siliceous material derived from the Rocky Mountains and was transported across the High Plains by eastward-flowing streams during Tertiary and early-Pleistocene times (Frye and Leonard, 1952). The veneer of gravel generally is less than 1.5 m thick at Locality BC-1, and it has been altered by slope processes, especially gully and rill erosion. Although the absolute age of these coarse-grained deposits is unknown, concentrations of *Proboscidean* bones on their surfaces suggest that the gravels accumulated during the Pleistocene.

Radiocarbon assays combined with archaeological evidence indicate that the bulk of the alluvium stored in the valley bottom of Buckner Creek accumulated during the late Holocene. Specifically, most of the T-1 fill aggraded between ca. 2800 and 1360 yrs B.P. Aggradation of T-1 fill was punctuated by episodes of floodplain stability and concomitant soil development. Although some of these episodes of stability were relatively short, as indicated by thin, faint A-C profiles among the buried soils developed in T-1 fill, two episodes were longer. These two episodes are represented by the thick, dark Hackberry Creek and Buckner Creek paleosols. Development of the Hackberry Creek Paleosol was underway by at least ca. 2800 yrs B.P. and continued to at least ca. 2600 years B.P. The Buckner Creek Paleosol was forming by at least ca. 1360 yrs B.P., though bounding dates for development of this soil are not available at Locality BC-1.

Buckner Creek downcut sometime after ca. 1360 B.P., leaving its late-Holocene floodplain as the T-1 terrace. As the T-1 surface became isolated from frequent floods, a Mollisol with A-Bk horizonation developed at the top of the T-1 fill.

A radiocarbon age of 1010 ± 410 yrs B.P. was determined on bison bones recovered from the T-0 fill immediately southwest of 14HO306. Hence the modern floodplain appears to have been aggrading by at least ca. 1000 yrs B.P. Weak soil development at the top of the T-0 fill suggests that the floodplain surface is very young, probably postdating Euroamerican settlement within the region (ca. 1830 A.D.). The absence of prehistoric archaeological sites on T-0 surfaces at Locality BC-1 and other reaches of Buckner Creek Valley supports this hypothesis.

Summary

Only one terrace (T-1) is present in the upper and middle reaches of small streams (< 5th-order), but remnants of a slightly higher terrace (T-2) occur in lower reaches of 4th-order streams near their confluences with the Pawnee River. The bulk of the valley fill underlying the T-1 terrace aggraded between 2800 and 1000 yrs B.P. (Table 4). Most of the alluvium beneath the T-2 terrace accumulated between 10,000 and 5000 yrs B.P.

Three terraces, numbered consecutively upward from T-1, are present in the valley bottoms of large streams (> 4th-order). The modern floodplain (T-0) is the lowest surface; T-2 and T-3 surfaces are loess-mantled Pleistocene terraces, and the T-1 surface is the surface of Holocene valley fill. Radiocarbon assays suggest that the upper 8-9 m of the T-1 fill aggraded between ca. 10,500 and 1600 yrs B.P. Aggradation of the adjacent T-0 fill was underway by 1000-500 yrs B.P.

Radiocarbon ages determined on humates from multiple buried paleosols in valley fill of large streams suggest that the period 10,000 to 5000 yrs B.P. was punctuated by several episodes of floodplain stability and soil development. However, Holocene valley fills in the Pawnee River basin are devoid of any evidence of soil development between 7000 and 5000 yrs B.P. Radiocarbon assays suggest that soil development was underway by at least 5000 yrs B.P. on floodplains in large valleys. Two discrete periods of paleosol development were detected in large valleys: one at 2750-2600 yrs B.P. and another at 2000-1600 yrs B.P. The older of these two episodes partially coincides with the soil-forming period dated to 2800-2000 yrs B.P. in small valleys. However, the most recent episode of paleosol development in large valleys (2000-1600 yrs B.P.) precedes the beginning of the major soil-forming period dated to 1350-1000 yrs B.P. in small valleys. Hence episodes of late-Holocene deposition appear to have been time transgressive throughout the entire extent of the drainage network.

Radiocarbon assays indicate that Holocene erosion and alluviation, as well as periods of net transport and storage of sediment, were diachronous throughout the Pawnee River basin, but were roughly synchronous in similar-sized streams of the drainage network. Early, middle, and late Holocene alluvium is stored in valley fill of large streams, but only late-Holocene deposits are present in valley bottoms of small streams.

Although valley erosion and alluviation may have several causes, major bioclimatic changes explain the pattern of Holocene fluvial activity detected in the stratigraphic record in the Pawnee River basin. Reduced vegetative cover, combined with infrequent but intense rainfalls during the warm Altithermal (8000-5000 yrs B.P.), favored erosion and net transport of sediment in small valleys. As mean annual precipitation increased during the late Holocene, vegetation recovered and erosion rates decreased, promoting sediment storage in small valleys.

The Holocene record of entrenchment, alluviation, and soil formation in the Pawnee River valley generally agrees with alluvial chronologies for valleys elsewhere in the Great Plains and Midwest. Detailed correlation between river basins is yet to be done, however, because of a lack of sufficient detail from adjacent areas.

TABLE 10/11-1. Description of Rucker Section, Locality PR-4, Pawnee River Valley.

Location: Pawnee County, KS; SW 1/4, NW 1/4, Sec. 31, R. 20 W., T. 21 S.

Landscape Position: T-1 terrace

Slope: < 1%

Vegetation: Native grasses

Note: A number in front of a soil horizon indicates the soil number and does not imply a change in lithology.

Depth (cm)	Soil Horizon	Unit	Description
0-27	Ap	I	Dark grayish brown (10YR 4/2) silt loam, very dark grayish brown (10YR 3/2) moist; weak fine granular structure; slightly hard, friable; many fine roots; common fine pores; weak effervescence; clear smooth boundary.
27-38	A	I	Grayish brown (10YR 5/2) silt loam, very dark grayish brown (10YR 3/2) moist; weak fine granular structure; slightly hard, friable; many fine roots; common fine pores; few fine flecks of calcium carbonate; strong effervescence; clear smooth boundary.
38-48	ACk	I	Grayish brown (10YR 5/2) to brown (10YR 5/3) silt loam, dark grayish brown (10YR 4/2) moist; weak very fine granular structure; slightly hard, friable; many fine roots; common fine pores; few fine flecks of calcium carbonate; strong effervescence; gradual smooth boundary.
48-105	C	I	Laminated light grayish brown (10YR 6/2) grayish brown (10YR 5/2), dark grayish brown (10YR 4/2) and very dark grayish brown (10YR 3/2) silty clay loam, silt loam, and very fine sandy loam; many individual laminae are < 2 mm thick; slightly hard, friable; common fine roots; few fine pores; strong effervescence; abrupt smooth boundary.
105-145	2Ab	II	Dark gray (10YR 4/1) silt loam, very dark gray (10YR 3/1) moist; weak fine granular structure; slightly hard, friable; common fine roots; few fine pores; few earthworm casts; noneffervescent; gradual smooth boundary.
145-175	2ACkb	II	Brown (10YR 5/3) silt loam, brown (10YR 4/3) moist; common fine faint grayish brown (10YR 5/2) mottles; weak fine granular structure; slightly hard, friable; few fine roots, few fine pores; few threads of calcium carbonate; violent effervescence; gradual smooth boundary.
175-273	2C1	II	Pale brown (10YR 6/3) silt loam, brown (10YR 5/3 - 4/3) moist; weak fine granular structure; hard, friable; few fine roots; common fine pores; violent effervescence; gradual smooth boundary.
273-441	2C2	II	Very pale brown (10YR 7/3) and pale brown (10YR 6/3) silt loam and very fine sandy loam; stratified; soft, friable; individual strata are <1 to 5 mm thick; few fine roots, few fine pores; violent effervescence; abrupt smooth boundary.

441-482	3Bkb	III	Yellowish brown (10YR 5/4) silty clay loam, dark yellowish brown (10YR 4/4) moist; common fine and medium distinct brown (10YR 5/3) and grayish brown (10YR 5/2) mottles; moderate fine and medium subangular blocky structure; hard, friable; many films and threads of calcium carbonate; few fine hard carbonate concretions; strong effervescence; diffuse boundary.
482-528	3Btkb	III	Brown (10YR 5/2) heavy silty clay loam, dark brown (10YR 3/3) moist; moderate medium subangular blocky structure; hard, firm; few fine pores; common films and threads of calcium carbonate; few fine hard carbonate concretions; common discontinuous clay films on ped faces; weak effervescence; gradual smooth boundary.
528-564	3BCkb	III	Brown (10YR 5/3) silty clay loam, brown (10YR 4/3) moist; moderate fine subangular blocky structure; slightly hard, friable; few fine pores; few films and threads of calcium carbonate; weak effervescence; abrupt smooth boundary.
564-604	4Btkb	IV	Yellowish brown (10YR 5/4) silty clay loam, dark yellowish brown (10YR 4/4) moist; moderate fine subangular blocky structure; hard, firm; few dark brown (10YR 3/3) clay flows on vertical faces; common films and threads of calcium carbonate; few fine pores; strong effervescence; gradual smooth boundary.
604-820+	4BCkb	IV	Yellowish brown (10YR 5/4) light silty clay loam, dark yellowish brown (10YR 4/4) moist; massive; soft, friable; common fine pores; common films and threads of calcium carbonate; strong effervescence.

TABLE 10/11-2. Description of Section at 14HO306, Locality BC-1, Buckner Creek Valley.

Location: Hodgeman County, KS; SW 1/4, NE 1/4, Sec. 16, R. 25 W., T. 23 S.

Landscape Position: T-1 terrace

Slope: <1%

Vegetation: Native grasses

Note: A number in front of a soil horizon indicates the soil number and does not imply a change in lithology.

Depth (cm)	Soil Horizon	Unit	Description
0-18	Ap	I	Brown (10YR 4/3) silt loam, very dark grayish brown (10YR 3/2) moist; weak medium granular structure; slightly hard, friable; common fine roots; common fine pores; noneffervescent; gradual smooth boundary.
18-38	A	I	Grayish brown (10YR 5/2) silt loam, very dark grayish brown (10YR 3/2) moist; moderate medium granular structure; slightly hard, friable; common fine roots; common fine pores; few earthworm casts; moderate effervescence; gradual smooth boundary.
38-46	AB	I	Grayish brown (10YR 5/2) silt loam, brown (10YR 4/3) moist; moderate medium granular structure; slightly hard, friable when moist; few pale brown (10YR 6/3) silt coatings on ped faces; few fine roots; common fine pores; moderate effervescence; clear smooth boundary.
46-61	Bk1 (2Ab)*	I	Dark brown (10YR 3/3) silt loam, dark grayish brown (10YR 3/2) moist; weak medium columnar parting to weak fine subangular blocky structure; hard, friable; common pale brown (10YR 6/2) silt coatings on ped faces; few fine roots; common fine pores; few threads and films of calcium carbonate; strong effervescence.
61-86	Bk2	I	Grayish brown (10YR 5/2) silt loam, brown (10YR 4/3) moist; moderate medium columnar parting to moderate medium granular structure; hard, friable; common pale brown (10YR 6/3) silt coatings on ped faces; few fine roots; common fine pores; common films and threads of calcium carbonate; strong effervescence; gradual smooth boundary.
86-119	BCK	I	Grayish brown (10YR 5/2) silt loam, brown (10YR 4/3) moist; weak medium and fine granular structure; very hard when dry, friable when moist; common fine pores; common films and threads of calcium carbonate; strong effervescence; abrupt smooth boundary.
119-147	3Akb	I	Grayish brown to dark grayish brown (10YR 5/2 - 4/2) silt loam; dark grayish brown (10YR 4/2) moist; weak medium granular structure; hard, friable; common fine pores; common films and threads of calcium carbonate; strong effervescence; gradual smooth boundary.
147-176	3ACKb	I	Grayish brown (10YR 5/2) silty clay loam, dark grayish brown (10YR 4/2) moist; weak fine and medium granular structure; hard, friable;

			common fine pores; common films and threads calcium carbonate; strong efferves-cence; abrupt smooth boundary.
176-210	4Ab	II	Dark gray (10YR 4/1) loam, very dark gray (10YR 3/1) moist; moderate medium granular structure; hard when dry, friable when moist; common fine pores; few flecks of charcoal; few chert flakes and burned rocks; few gastropods; noneffervescent; gradual smooth boundary.
210-241	4Akb	II	Dark grayish brown (10YR 4/2) silt loam, very dark grayish brown (10YR 3/2) moist; moderate medium granular structure; hard, friable; common fine pores; few films and threads of calcium carbonate; few flecks of charcoal; few chert flakes and burned rocks; few gastropods; strong effervescence; gradual smooth boundary.
241-271	4ACkb	II	Brown (10YR 5/3) silt loam; dark grayish brown (10YR 4/2) moist; weak fine granular structure; hard, friable; common fine pores; common grayish brown (10YR 5/2) coatings on ped faces; few films and threads of calcium carbonate; few gastropods; strong effervescence; gradual smooth boundary.
271-317	4Ckb	II	Pale brown (10YR 6/3) silt loam, brown (10YR 5/3) moist; massive with faint lamination; hard, friable; many fine pores; common films and threads of calcium carbonate; few fine hard concretions of calcium carbonate; violent effervescence; abrupt smooth boundary.
317-345	5Akb	II	Grayish brown (10YR 5/2) silt loam, dark brown (10YR 3/3) moist; weak fine granular structure; hard, friable; common fine pores; common films and threads of calcium carbonate; few fine soft calcium carbonate concretions; strong effervescence; gradual smooth boundary.
345-394	5Ck	II	Pale brown (10YR 6/3) silt loam, brown (10YR 5/3) moist; massive with faint lamination; hard, friable; common fine pores; many films and threads of calcium carbonate; strong effervescence; abrupt smooth boundary.
394-432	6Akb1	III	Dark grayish brown (10YR 4/2) silt loam, very dark grayish brown (10YR 3/2) moist; moderate medium granular structure; hard, friable; many fine pores; common films and threads of calcium carbonate; many flecks of charcoal; common bone fragments and burned rocks; few chert flakes; strong effervescence; gradual smooth boundary.
432-471	6Akb2	III	Grayish brown (10YR 5/2) silt loam, dark grayish brown (10YR 4/2) moist; moderate medium granular structure; hard, friable; common fine pores; common films and threads of calcium carbonate; few gastropods; common flecks of charcoal, bone fragments, and burned rocks in lower 10 cm of horizon; strong effervescence; gradual smooth boundary.
471-523	6Bkb	III	Brown (10YR 5/3) silt loam, dark brown (10YR 4/3) moist; weak fine granular structure; hard, friable; common fine pores; common films and threads of calcium carbonate; few fine hard calcium carbonate concretions; violent effervescence; gradual smooth boundary.

523-658	6BCkb	III	Pale brown (10YR 6/3) silt loam; brown (10YR 4/3) moist; massive with faint bedding; hard, friable; common fine pores; common films and threads of calcium carbonate; few fine hard concretions of calcium carbonate; violent effervescence; gradual smooth boundary.
658-712	6Ck	III	Brown (10YR 5/3) gravelly fine sand grading upward to fine sandy loam; brown (10YR 4/3) moist; massive with faint bedding in the upper 25 cm; hard, friable; few fine pores; few films and threads of calcium carbonate; strong effervescence; irregular boundary.
712+	7R	—	Limestone

*Welded Soil

TABLE 10/11-3. Particle size distribution and calcium carbonate content of sediment samples from section at Site 14HO306 (Locality BC-1).

Depth (cm)	Soil Horizon	Particle Size Distribution (mm)										Texture* Class	CaCO3
		Sand-----						Silt-----		Total Clay (<0.002)			
		VC	C	M	F	VF	Total	Fine	Total				
		(0.25- 1.0)	(0.10- 0.5)	(2.0- 0.25)	(0.02 - 0.10)	(0.05- 0.05)	(2.0- 0.05)	(0.02- (0.002)	(0.05- 0.002)				
		-----%						-----%			-----%		
0-18	Ap	0.2	1.9	3.5	7.5	9.3	22.4	23.1	66.2	11.4	SiL	0.4	
18-38	A	0.2	2.0	4.0	6.6	8.4	21.2	20.3	64.1	14.7	SiL	0.6	
38-46	AB	0.2	0.7	2.1	8.4	11.1	22.5	16.5	58.7	18.8	SiL	7.2	
46-61	Bk1(2Ab)~	0.0	0.2	1.0	2.3	6.7	10.2	27.2	69.1	20.7	SiL	5.5	
61-86	Bk2	0.0	0.7	1.1	3.4	6.1	11.3	25.7	66.3	22.4	SiL	6.2	
86-119	BCk	0.2	0.2	0.3	2.0	5.7	8.4	16.9	71.6	20.0	SiL	4.7	
119-147	3Akb	0.1	0.2	0.5	1.8	4.0	6.6	33.0	74.2	19.2	SiL	4.5	
147-176	3ACkb	0.0	0.1	0.5	2.3	5.9	8.8	32.2	63.9	27.3	SiCL	6.4	
176-210	4Ab	0.2	2.6	7.2	11.9	7.7	29.6	17.8	49.6	20.8	L	0.4	
210-241	4Akb	0.4	2.3	7.7	16.1	10.8	37.3	12.4	45.9	16.8	SiL	5.4	
241-271	4ACkb	0.3	1.9	7.4	17.1	11.6	38.3	10.3	45.6	16.1	SiL	9.5	
271-317	4Ckb	0.1	0.8	3.6	10.3	11.7	26.5	16.7	51.6	21.9	SiL	11.1	
317-345	5Akb	0.3	0.5	0.9	11.0	13.1	25.8	20.2	54.3	19.9	SiL	8.0	
345-394	5Ckb	0.4	1.2	1.4	1.3	5.0	9.3	27.3	66.1	24.6	SiL	8.6	
394-432	6Akb1	0.4	0.7	1.2	2.7	3.5	8.5	30.9	64.6	26.9	SiL	3.8	
432-471	6Akb2	0.4	1.0	2.1	3.4	4.0	10.9	28.9	65.3	23.8	SiL	4.7	
471-523	6Bkb	0.2	0.9	3.1	7.3	6.4	17.9	20.1	56.9	25.2	SiL	7.4	
523-658	6BCkb	0.2	1.7	4.5	8.4	5.9	20.7	20.7	56.9	22.4	SiL	8.5	
658-685	6Ck1	0.3	1.9	3.2	22.3	20.4	48.1	17.9	45.7	6.2	SL	3.7	
685-712	6Ck2	2.1	2.9	5.1	37.1	36.2	83.4	15.1	13.6	3.0	LS	2.3	

*Texture Class: L= Loam; SiL= Silt Loam; SiCL= Silty Clay Loam; SL=Sandy Loam; LS= Loamy Sand

~ Welded soil

TABLE 10/11-4. Radiocarbon ages from the study area.

STREAM VALLEY	LANDFORM	MATERIAL DATED	DEPTH (m)	AGE (Y.B.P.)	LAB NO.
Large Valleys					
Pawnee R.	T-1	Soil	6.20-6.40	10,240±120	TX-6391
Pawnee R.	T-1	Soil	3.45-3.70	10,100±130	TX-6374
Pawnee R.	T-1	Soil	5.63-5.83	9320±120	TX-6396
Pawnee R.	T-1	Soil	5.28-5.48	7900±160	TX-6392
Pawnee R.	T-1	Soil	3.30-3.35	7720±110	TX-6476
Pawnee R.	T-1	Soil	3.50-3.70	7150±110	TX-6398
Pawnee R.	T-1	Soil	1.25-1.55	4350±50	TX-6389
Pawnee R.	T-1	Soil	1.42-1.62	2740±80	TX-6375
Pawnee R.	T-1	Soil	0.45-0.65	2600±80	TX-6394
Pawnee R.	T-1	Soil	0.60-0.95	2050±80	TX-6390
Pawnee R.	T-1	Soil	1.80-2.00	1940±60	TX-6393
Pawnee R.	T-1	Soil	2.75-2.95	1850±80	TX-6397
Pawnee R.	T-1	Soil	1.20-1.40	1710±80	TX-6395
Hackberry Cr.	T-2	Soil	6.50-6.70	9820±110	TX-6480
Hackberry Cr.	T-2	Soil	4.51-4.71	7170±120	TX-4732
Hackberry Cr.	T-2	Soil	2.45-2.65	4970±100	TX-5731
Small Valleys					
Pawnee Cr.	T-1	Soil	1.95-2.15	2430±90	TX-6474
Pawnee Cr.	T-1	Soil	1.20-1.35	1500±80	TX-6473
Hackberry Cr.	T-1	Charcoal	5.50-5.15	2680±80	TX-5640
Hackberry Cr.	T-1	Charcoal	4.20-4.30	2450±90	TX-5641
Hackberry Cr.	T-1	Charcoal	4.10-4.15	2340±400	TX-5639
Hackberry Cr.	T-1	Soil	2.90-3.30	2170±50	TX-5645
Hackberry Cr.	T-1	Soil	2.10-2.50	1950±50	TX-5645
Hackberry Cr.	T-1	Bone*	1.55-1.60	1310±130	TX-5643
Hackberry Cr.	T-1	Bone**	1.55-1.60	1190±370	TX-5643
Hackberry Cr.	T-1	Soil	1.07-1.37	1090±70	TX-5646
Buckner Cr.	T-1	Charcoal	4.64-4.69	2820±260	TX-5730
Buckner Cr.	T-1	Charcoal	3.94-4.04	2620±220	TX-5729
Buckner Cr.	T-1	Charcoal	5.28-5.33	1610±100	TX-5642
Buckner Cr.	T-1	Soil	4.98-5.28	1430±50	TX-5644
Buckner Cr.	T-1	Soil	1.76-2.00	1360±60	TX-5642
Buckner Cr.	T-0	Bone**	1.00-1.05	1010±410	TX-5651
Sawlog Cr.	T-1	Soil	2.00-2.30	2260±80	TX-5647

* Appetite

** Collagen

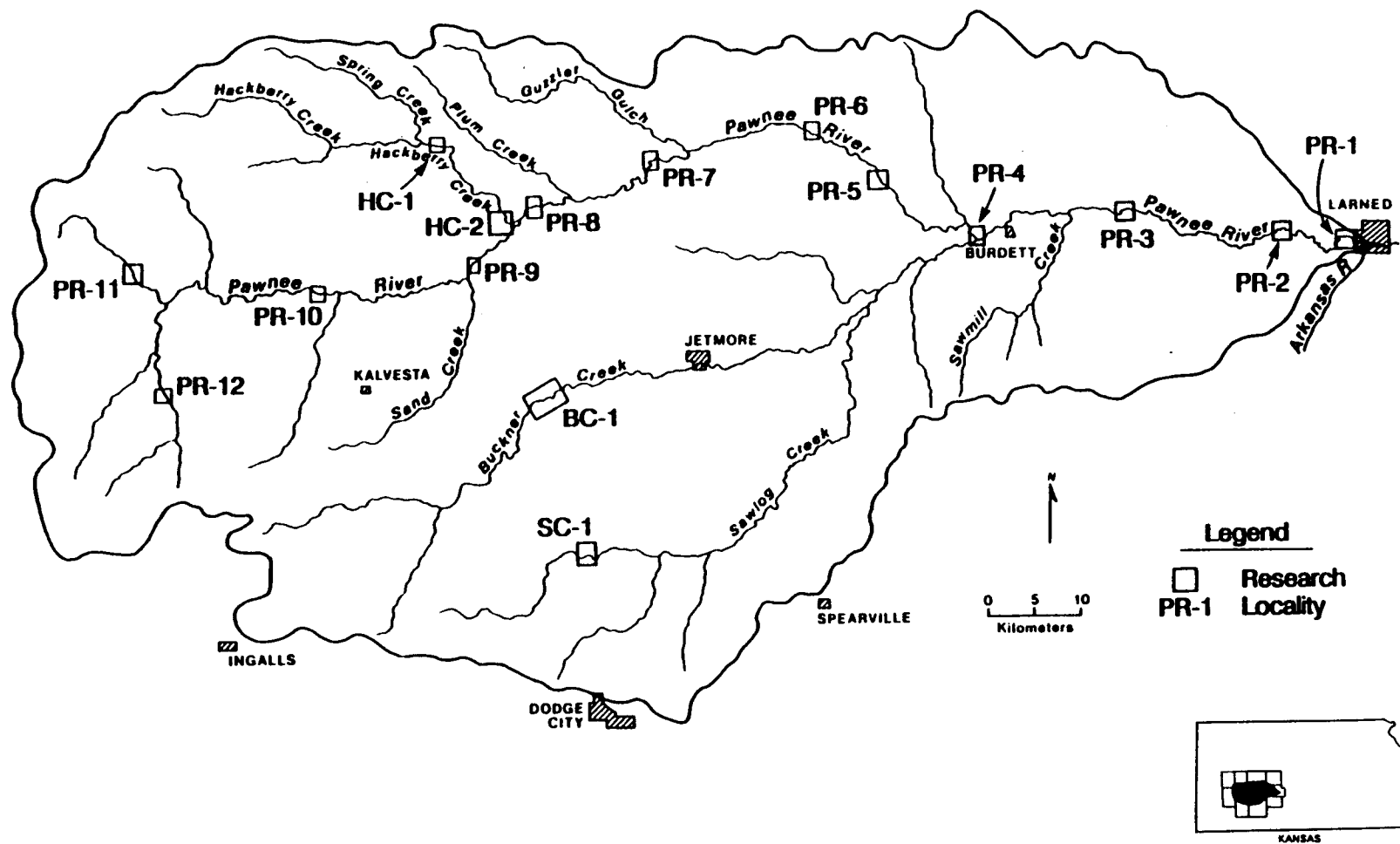


FIGURE 10/11-1. Map of the Pawnee River basin showing localities where detailed soil-stratigraphic investigations were conducted.

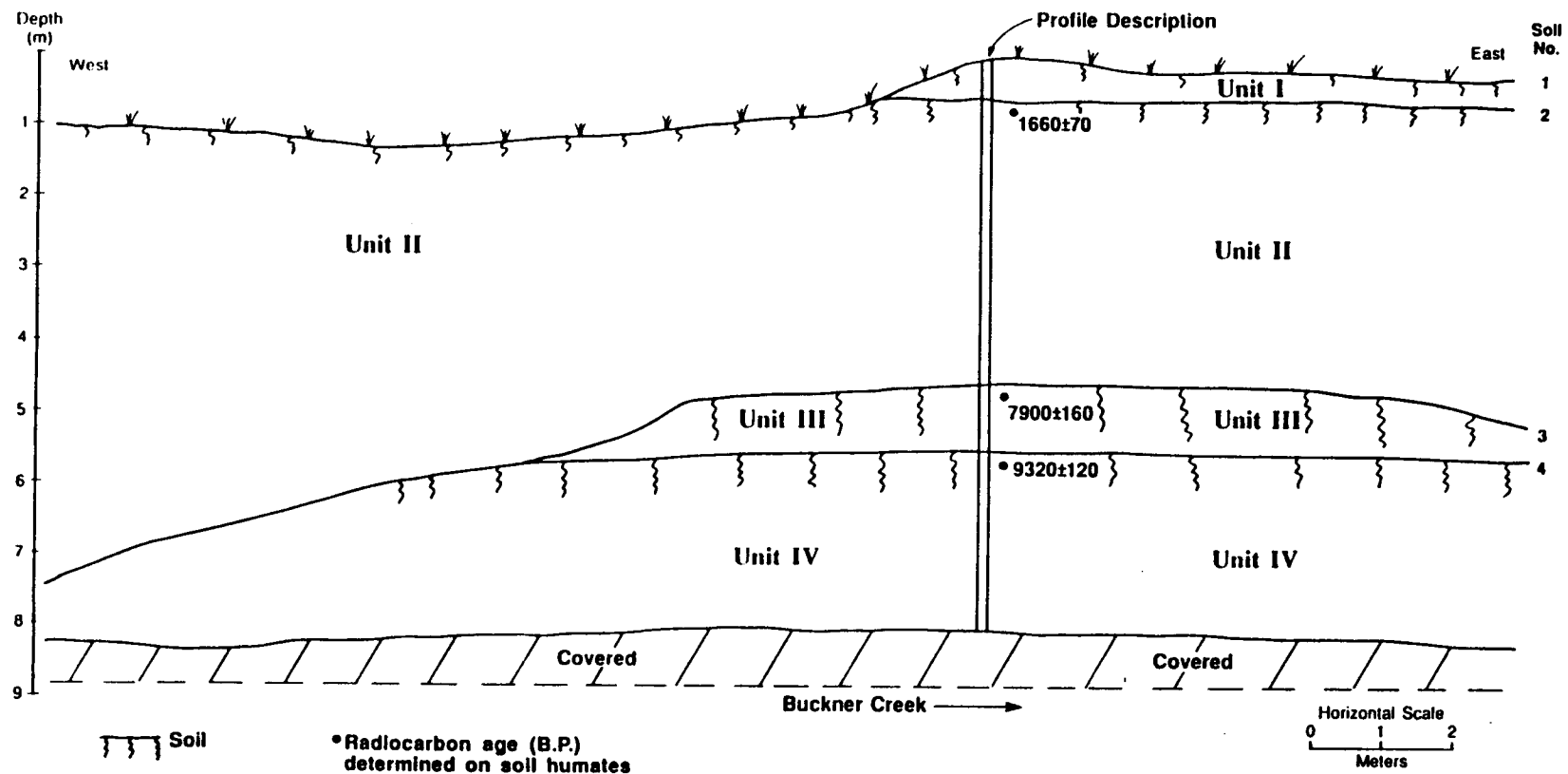


FIGURE 10/11-2. Diagram of the Rucker Section showing stratigraphic units, soils, and radiocarbon ages.

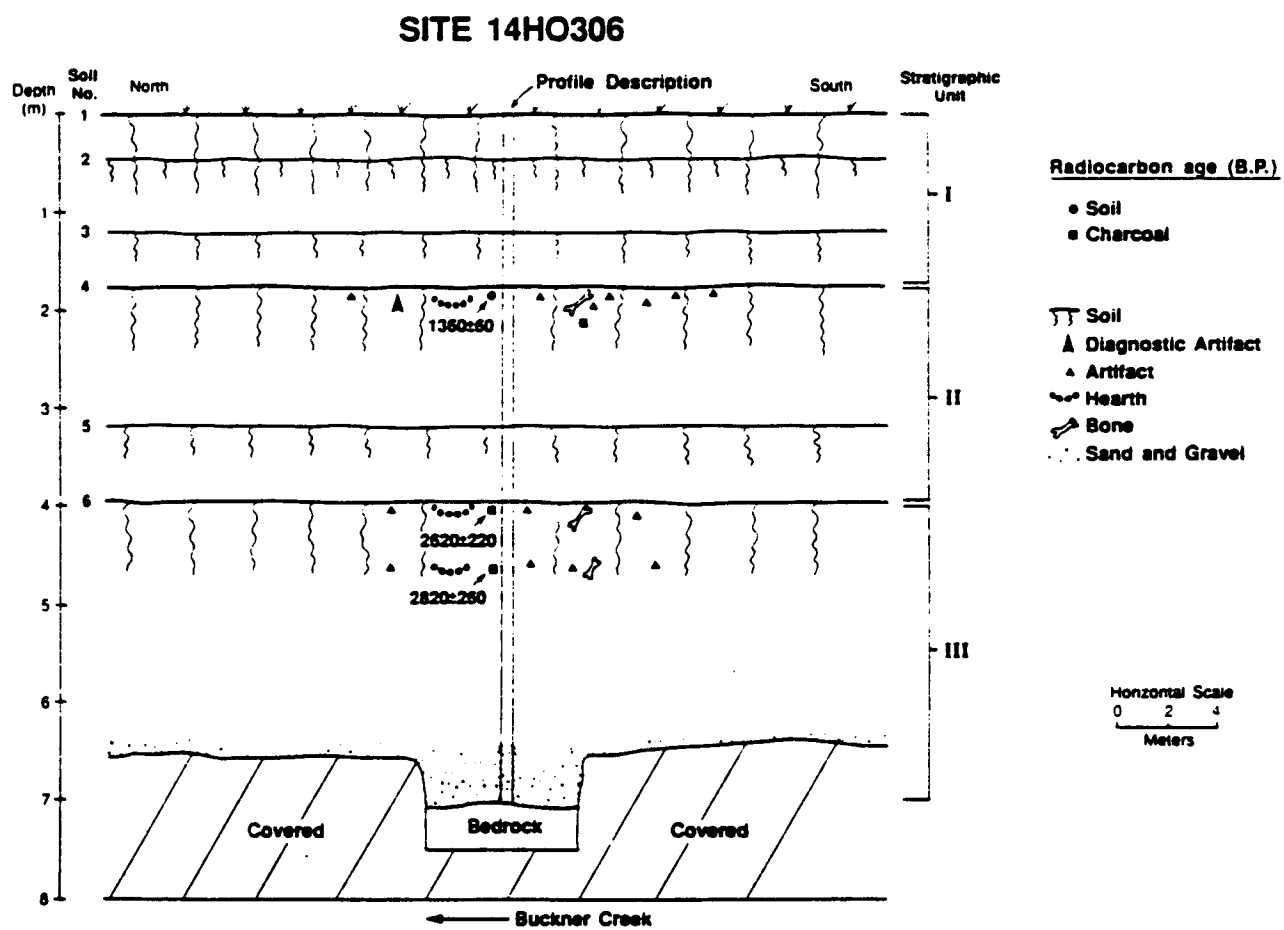


FIGURE 10/11-3. Diagram of section at Site 14HO306 showing stratigraphic units, soils, and radiocarbon ages.

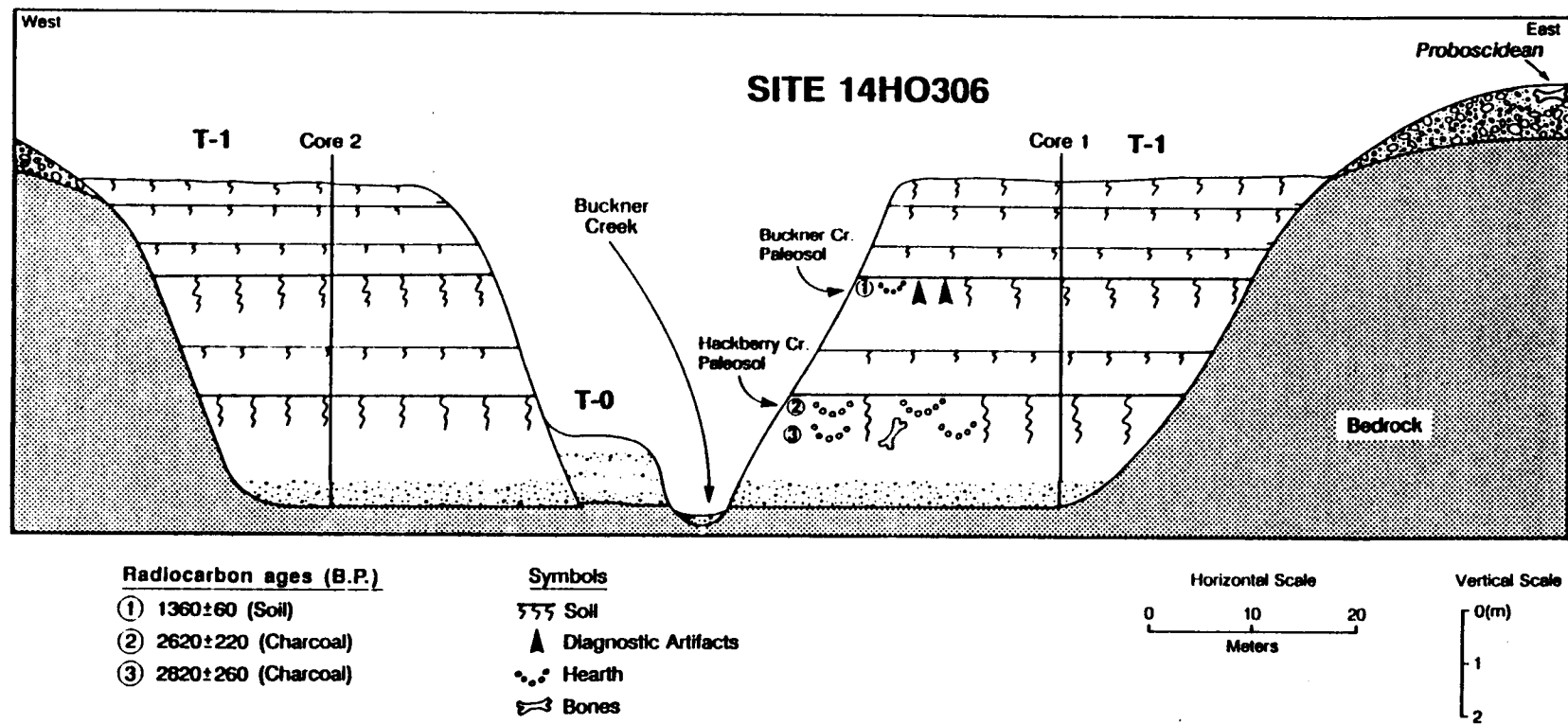


FIGURE 10/11-4. Cross-sectional diagram of Buckner Creek valley at Locality BC-1.

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Stop 12 Eustis Ash Pit, 2 miles southwest of Eustis, Nebraska: Loess-soil sequence over volcanic ash

Location: NE 1/4, SW 1/4 Sec. 20, T8N, R24W, Frontier County, Nebraska
(USGS Eustis, NE 7.5 minute quadrangle)

Contributors: William Johnson, David May, Erik Diekmeyer, M.R. Farr and Kyeong Park,
with contributions of data from S. Valastro (University of Texas Radiocarbon
Laboratory), M. Kerr, and S. Bozarth (University of Kansas Department of Geography)

A 600K (?) RECORD OF LOESS DEPOSITION AND SOIL DEVELOPMENT

Late-Pleistocene paleosols and loess are exposed in an abandoned volcanic ash mine 3.2 km southwest of Eustis, Frontier County, Nebraska. The main exposure, about 35m high, has weathered for nearly 50 years but soil and loess stratigraphy is still visible. Late-Pleistocene stratigraphy of the site includes Loveland (Illinoian) loess with soils, Sangamon soil, Gilman Canyon Formation (late Wisconsin) loess and geosol, and Peoria (late Wisconsin) loess. The late-Pleistocene/early-Holocene Brady geosol and Bignell loess are absent or poorly developed in the profile. The upper 16 m of the exposure is Peoria loess, while the underlying Gilman Canyon Formation is only about 1.5 m thick. Five Loveland complexes, with the Sangamon being uppermost, are expressed in the lower half of the exposure. Although the soils and ash layer (mined as an abrasive) are best viewed in the former working face of the mine, excavation for stratigraphic work has been conducted along an exposure adjacent to the haul road (profiles 1-3) and a smaller face to the southwest of the main face (profiles 4 & 5).

Environmental setting

The Eustis ash pit is close enough to the Rocky Mountains to be in their rainshadow; the present climate is dry, with precipitation averaging 533.4 mm per year (Hoppe and Huber 1978). Annual temperatures reflect the midcontinental location of the county; summers are warm, winters are quite cold. Much of Frontier County is now rangeland and dry or irrigated agriculture, although potential natural vegetation consists of *Andropogon-Bouteloua* prairie (Hoppe and Huber 1978). Most of Frontier county is covered by a thick blanket of Pleistocene-age materials, although the Pliocene-age Ogallala Formation outcrops in valleys sides. Major streams are entrenched 30 to 60 m below the general elevation of the uplands. Modern soils are predominantly Mollisols formed in Peoria loess (Hoppe and Huber 1978).

Sangamon and Pre-Sangamon Soils

The volcanic ash at the base of the section was fluvially redeposited and occurs locally in pockets up to 15 m thick (Schultz and Stout, 1945). Izett and Wilcox (1982) identified the ash as the Lava Creek B (Pearlette Type O), fission-track dated at 620ka. Recent fission-track dating of the ash has, however, produced a significantly older age of 980 ± 300 ka (Feng et al., 1993), but the standard deviation is large. The multiple paleosols above the ash represent stable episodes of variable duration that interrupted loess accumulation. The age of the ash suggests that the lower part of the loess could be around 600ka, which is pre-Illinoian according to most chronologies (see Barton County Sanitary Landfill discussion). Until TL dates or other information come to light, the age of the lower soils will be unknown and the Illinoian/pre-Illinoian boundary undetermined.

The Loveland, considered by some to be a group rather than formation, has been divided into three loess units, or members, punctuated by paleosols (Reed and Dreeszen, 1965; Schultz, 1968; Schultz and Martin, 1970). Given the problems of correlating loessal soils across an area of variable loess thickness and soils expression, any designations of soils should be tentative. Five soil complexes appear in the Eustis section. The lowest of these is poorly expressed; at Buzzards Roost and elsewhere it is either unrecognized or convergent with the adjacent soil complex. The second soil complex up is perhaps the equivalent of the "Yarmouth soil development" (Reed and Dreeszen, 1970) and the "Ingham Paleosol" (Schultz and Martin, 1970). The third and fourth soil complexes up are probably the equivalent of the "Buzzard's Roost Paleosol Complex" (Schultz and Martin, 1970). The paleosols clustering together in the middle of the section are the Sangamon and overlying Gilman Canyon Formation soil.

A study of opal phytolith content and morphology in the lower part of the section (12-33m) characterized the vegetational environment and hence prevailing climate of the soils (Fredlund et al., 1985). Of the three environmentally sensitive classes (Chloridoid, Festucoid, and Panicoid), the Festucoid are the most abundant (Fig. 1) because of the tendency for most grasses to produce them (Brown, 1984). The C4 (warm, dry adapted) grass phytoliths (Chloridoid and Panicoid) exhibit an increase in most of the soils, but particularly the Gilman Canyon (discussed below). Based on these increases, it appears that most of the soil-forming periods must have been warmer and drier than period of loess accumulation, i.e., cooler climates during loess accumulation (glaciation).

The Gilman Canyon Formation Loess and Composite Geosol

The Gilman Canyon Formation is well expressed at the Eustis pit and has been the focus of research here. This is the same formation discussed and examined at the Barton County Sanitary Landfill 300 km to the south/southeast in central Kansas, adjacent to the Arkansas River. The formation was first recognized as the *Citellus* zone by Schultz and Stout (1945) because of its fossiliferous nature and abundant krotovina. *Citellus* is a genus of ground squirrel now recognized as *Spermophilus*. Reed and Dreeszen renamed the zone the Gilman Canyon Formation in 1965. The formation underlies much of the central Great Plains and is a widely used late-Pleistocene pedostratigraphic marker. The Gilman Canyon Formation is considered a geosol because it is a buried, laterally traceable, isochronous lithologic unit that consists of one or more differentiated pedologic horizons (North American Stratigraphic Code 1983). Further, the possible existence of two or more welded A horizons may allow the Gilman Canyon Formation to be classified as a composite geosol. The formation is typically rich in organic matter, leached of calcium carbonate, and melanized via pedogenesis (Kerr 1992). The Gilman Canyon Formation is correlatable with the Pisgah formation of western Iowa (Bettis 1990) and the Roxanna silt extending from Wisconsin to Arkansas (Johnson and Follmer, 1989; Leigh and Knox, 1993).

Eleven radiocarbon ages provide chronological control for the Gilman Canyon Formation at Eustis (Table 1), and bracket the geosol between 36,000 and 20,000 yr B.P. Ages from the Eustis section are in agreement with those from other sections in Nebraska and Kansas (Johnson 1993). Plotting depths of the radiocarbon samples against their ages (Fig. 2) reveals slow sedimentation rates (0.08 mm/yr) during Gilman Canyon time compared to rates in the overlying Peoria loess. Peoria sedimentation rates were highest (0.6 mm/yr) immediately following Gilman Canyon time, approximately 22,000 yr B.P.

Sedimentological data are available for the Gilman Canyon Formation at Eustis, as exposed along the haul road. Silt content is relatively high (75%), while sand and clay contents are low (Fig. 3). Texture is more variable in the Peoria loess, and sand increases upward at the expense of silt. The increased level of sand beneath the Gilman Canyon geosol is also notable but not yet adequately explained. Organic matter (Fig. 4) is concentrated in the modern soil and in the Gilman Canyon Formation (Kerr 1992). Calcium carbonate (Fig. 5) is highest near the modern soil, low in Peoria loess, and shows a sharp increase at the contact

between the Peoria loess and Gilman Canyon Formation. Reed and Dreeszen (1965) found two humus peaks at the Gilman Canyon type section, while May and Souders (1988) observed three organic zones in an expanded phase of the Gilman Canyon Formation.

Paleoenvironmental data are emerging for the Gilman Canyon Formation. $\Delta^{13}\text{C}$ values are a potential source of proxy data for vegetation type and hence climate (Krishnamurthy et al., 1982). When determinations are derived from the organic fractions of the soil, they reflect inputs by the plants, particularly grasses, growing on those surfaces. C3 (cool-season) species have an average $\delta^{13}\text{C}$ composition of -27‰ and C4 (warm-season/arid) species -13‰, relative to the PDB standard (Deines, 1980). $\Delta^{13}\text{C}$ values acquired in association with ^{14}C ages for the Gilman Canyon at Eustis (Fig. 6) indicate that C4 plants were dominant for most of the time of pedogenesis (Johnson 1993). This indicates that vegetation and thus climate during Gilman Canyon time was similar to the present warm, semiarid conditions at the Eustis site. Early and late Gilman Canyon time was apparently characterized by a C3 plant environment (cool, moist adaptation). The former indicates the cooling associated with the glacial maximum occurring in Peoria time at about 18 ka.

Ongoing opal phytolith analysis of the Gilman Canyon and Peoria loesses is producing a climatic signal consistent with that of the isotope data (Fig. 7). C4 types (Panicoidae, Chloridaceae) increase and the C3 type (Pooideae) decrease within the Gilman Canyon pedogenic zone (c. 17.75-16.25 mbs), indicating increased aridity and/or warmth. Continued Panicoidae presence throughout the Peoria likely reflects the more moist C4 types and the presence of those panicoids which are C3 types. An arboreal climate is apparent by the occurrence of woodland phytolith types (spiny spheres, opaque bodies, blocky polyhedra) in the lower Gilman Canyon (pre-pedogenesis) and the Peoria loess. These data are corroborated by opal phytolith morphology data from Kansas (Fredlund et al. 1985, Fredlund and Jaumann 1987), which indicates the existence of a panicoid-dominated (C4) grassland during Gilman Canyon time.

Possible landscape stability and pedogenesis during Peoria time

Stratigraphic breaks in Peoria loess are recognized in Illinois (Frye et al. 1974) and Iowa (Daniels et al. 1960). However, no paleosols are recognized in Peoria loess in Nebraska or adjacent Kansas. Frye and Leonard (1949) observed a partially leached zone lacking molluscan faunal remains in Peoria loess in northeastern Kansas. Lenses of plant remains have been observed in Peoria loess in Kansas (Johnson 1993) and have been interpreted as discontinuities and possible periods of landscape stability during Peoria time. Plant remains, indicative of at least limited surface stability, have been ^{14}C dated at a number of localities in Kansas and Nebraska (Table 2).

The relatively rapid rate of accumulation of Peoria loess at the Eustis pit is evidenced by laminations, differentiated by particle size change and carbonate content. The laminations are best developed in the upper 8 m of the loess, where sedimentation rates, as defined by the ^{14}C ages, are high (Fig. 2). The zone also exhibits low magnetic susceptibility which is indicative of rapid accumulation (see below), slight gleying, high concentrations of different mollusc species (esp. *Vertigo modesto* and *Columella alticola*, cool open woodland species), and occasional mineralized plant detritus and fossil root channels. Similar lamination of the upper Peoria loess is observable at the Bignell Hill locality to the northwest.

Mineral Magnetism

A great deal of attention has been given to paleomagnetic measurements of loess-paleosol sequences, particularly in China (Heller and Liu, 1982; Kukla, 1987; Kukla et al., 1988) and in Europe (Fink and Kukla 1977). Although much research has focused on paleomagnetic events, magnetic data have been used to successfully reconstruct climatic sequences. The magnetic susceptibility record of Heller and Liu (1982)

and later Kukla et al. (1988) exhibited a close correlation to the $\delta^{18}\text{O}$ record of deep-sea sediments, with high magnetic susceptibilities associated with the interglacial intervals and low magnetic susceptibilities associated with glacial intervals. Recent work has explained this magnetic-global climate relationship in terms of increased pedogenic production of ferrimagnetic grains during the warmer interglacial intervals (e.g. Maher and Thompson, 1991; Heller et al., 1991). We have measured a number of rock magnetic properties in samples from the Gilman Canyon formation, Peoria loess, and modern/Brady soil complex at the Eustis ash pit and have noted a similar correspondence to soil development and global climate (Fig. 8).

Magnetic susceptibility reflects the magnetization temporarily induced in a rock by an artificially applied low-amplitude magnetic field. The strength of the susceptibility signal depends on the concentration and, to some extent, grain size of the magnetic minerals. Magnetic susceptibilities in the Gilman Canyon Formation are nearly twice that of the unweathered Peoria loess and comparable to the modern/Brady(?) composite soil (Fig. 8, Table 3).

Frequency dependency (FD) and anhysteretic remanent magnetization (ARM) are magnetic parameters which are useful in indicating the abundance of very fine ferrimagnetic particles. FD measures the presence of grains near the stable single domain/superparamagnetic size boundary (c. 0.03 μm dia.). ARM is a type of artificial remanence given to samples by applying a direct field to a sample during alternating-field demagnetization. ARM is particularly sensitive to grain size with finer magnetic particles acquiring ARM more efficiently than coarser particles. Both parameters are typically enriched within modern soils and ultrafine ferrimagnetic particles from such soils have been identified with transmission electron microscopy and attributed to either inorganic or bacterial production in the soils (Maher, 1986; Fassbinder et al., 1990). Both FD and ARM correlate well with susceptibility at Eustis, although they are more enriched in the Gilman Canyon soil and less enriched in the modern/Brady(?) soil complex than susceptibility (Fig. 8). Saturation isothermal remanent magnetization (SIRM) is a parameter which most accurately reflects the total amount of ferrimagnetic mineral grains. It more closely correlates with magnetic susceptibility than either FD or ARM at Eustis (Fig. 8).

Together, these parameters indicate that the Gilman Canyon soil complex is greatly enriched in very fine to ultrafine-grained ferrimagnetic particles relative to the Peoria loess, and it is likely that this enrichment is due to the pedogenic production of these magnetic particles during warm, moist climatic conditions. Chemical parameters which reflect soil development, such as CaCO_3 and organic content, correlate well with susceptibility (Fig. 9) further suggesting a link between soil processes and magnetic properties.

TABLE 12-1. Radiocarbon ages.

depth b.s.	uncorr. age	$\delta^{13}\text{C}$	corr. age	lab no.
<i>PEORIA FORMATION</i>				
2.10-2.30	n.a.	-24.6	16,570 $\pm$ 80	CAMS-4295
7.00-7.10	n.a.	-28.9	16,780 $\pm$ 120	ETH-9806
7.30-7.40	17,490 $\pm$ 400	-24.7	17,500 $\pm$ 400	Tx-7712
9.20-9.25	n.a.	-26.5	21,200 $\pm$ 170	CAMS-3720
9.90-10.00	21,630 $\pm$ 370	-16.2	21,780 $\pm$ 380	Tx-7816
<i>GILMAN CANYON FORMATION</i>				
16.30-16.35	22,290 $\pm$ 340	-23.6	22,320 $\pm$ 340	Tx-7347
16.40-16.45	23,480 $\pm$ 500	-23.4	23,510 $\pm$ 500	Tx-7348
16.55-16.60	23,950 $\pm$ 330	-23.96	23,970 $\pm$ 330	Tx-7349
16.65-16.70	24,550 $\pm$ 360	-21.43	24,600 $\pm$ 360	Tx-7350
16.75-16.80	25,850 $\pm$ 510	-18.5	25,950 $\pm$ 510	Tx-7351
16.90-16.95	29,370 $\pm$ 640	-18.5	29,470 $\pm$ 650	Tx-7352
17.00-17.05	30,110 $\pm$ 700	-18.0	30,210 $\pm$ 710	Tx-7353
17.10-17.15	30,420 $\pm$ 720	-17.8	30,530 $\pm$ 730	Tx-7354
17.25-17.30	30,950 $\pm$ 840	-17.5	31,080 $\pm$ 860	Tx-7355
17.40-17.45	33,020 $\pm$ 1020	-18.2	33,120 $\pm$ 1020	Tx-7356
17.60-17.65	35,870 $\pm$ 2190	-17.4	36,000 $\pm$ 2230	Tx-7357
18.48-18.58	38,580 $\pm$ 1240	-22.0	38,630 $\pm$ 240	Tx-7817

TABLE 12-2. Wood, charcoal, and humate ages, Woodfordian/Peoria-age deposits in Kansas and Nebraska.

Location	Material	Sample	Age	Source
Kansas				
Superior Sand Co., Sedgewick County	peat	Dic-3101	19,340±200/21 0	Jaumann, 1991
Mt. Hope Sand Co.,	<i>Picea</i> litter	Beta-17900	16,420±190	Jaumann, 1991
Reno County	soil humates	Tx-6742	13,670±290	Johnson, 1991a
GMD5 site 1, Edwards County	<i>Picea</i> charcoal	Tx-6479	17,970±330	Johnson, 1991a
GMD5 site 9, Reno County	<i>Picea</i> charcoal	GX-9355G	17,930±550	Wells and Stewart, 1987
Coon Creek Graham County	<i>Picea</i> charcoal	Beta-9320	14,450±140	Wells and Stewart, 1987
Courtland Canal, Jewell County				
Nebraska				
Coyote Canyon	charcoal	Tx-7295		Martin, unpub. data, 1992
Harlan County	charcoal	Tx-7293	19,730±300	
North Cove, 25HN164	<i>Picea glauca</i>	Beta-12286	21,250±530	Wells and Stewart, 1987
Harlan County	charcoal	Beta-33579	17,880±170	Souders and Kuzila, 1990
Naponee, Franklin County	<i>Picea</i> or <i>Larix</i>	Beta-42015	18,830±180	May and Holen, 1993
Bloomington, Franklin County	charcoal	Tx-7006	18,860±360	May and Holen, 1993
La Sena site, 25FT177,	humates	Tx-6708	16,730±490	
Frontier County	<i>Abies balsamea</i>	Tx-6128		May, 1989
South Loup River, Buffalo and Grant counties	fragments	Beta-27758	14,080±190 13,160±450	Swinehart, 1990

Ages indicate surface stability or discontinuities in Peoria loess deposition.

TABLE 12-3. High frequency magnetic susceptibility and frequency dependency data, with standard deviations, for major stratigraphic units.

Strat. units	HF SUS.	S.D.(SUS)	FD	S.D.(FD)
Modern/Brady(?) soil	70.11	10.85	3.01	1.19
Peoria loess	48.28	5.41	1.93	0.87
Gilman Canyon Fm. (paleosol)	74.82	5.37	3.32	0.76
Gilman Canyon Fm. (loess)	70.35	5.77	2.12	0.62

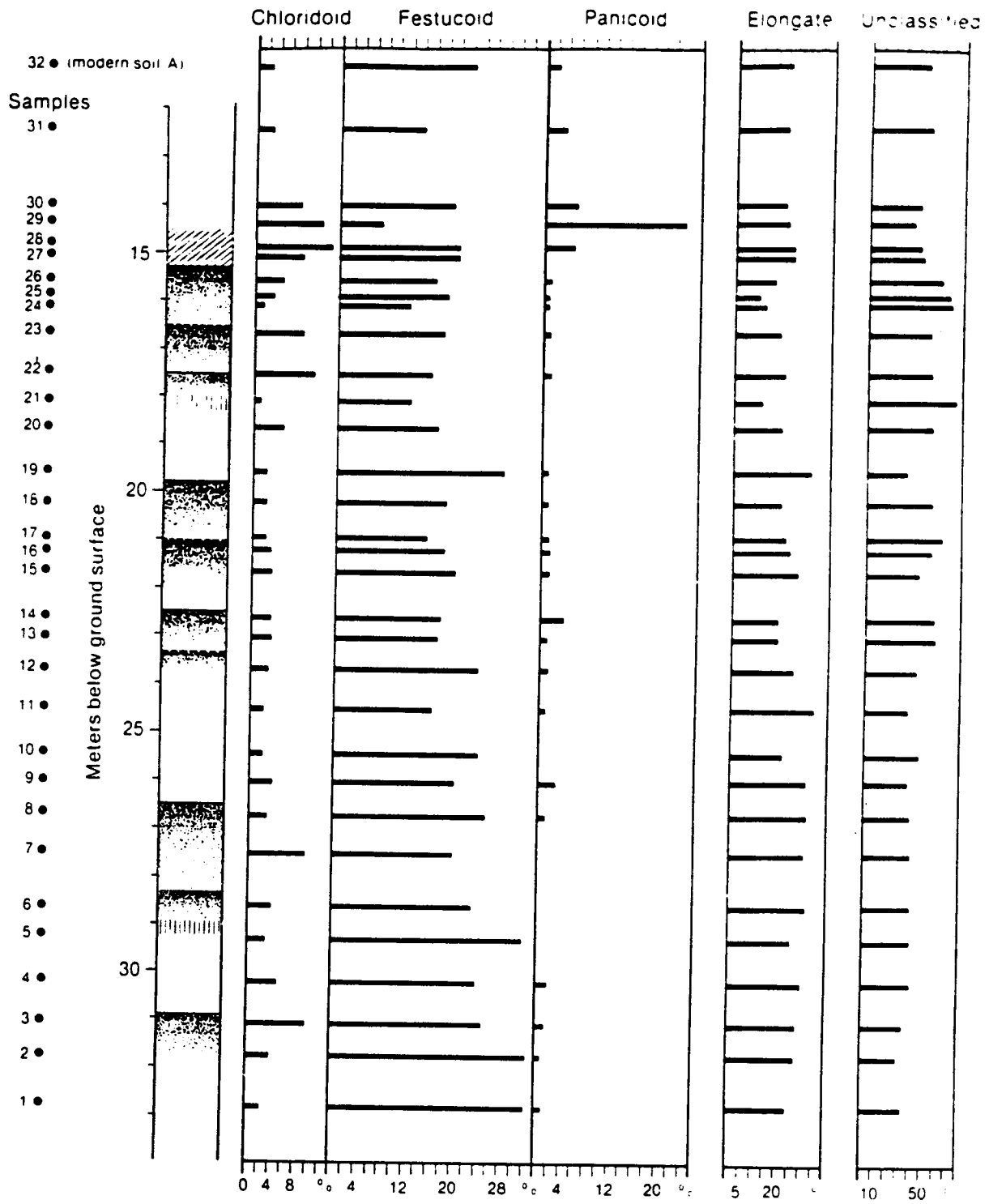


FIGURE 12-1. Relative abundances of phytolith classes from 12-33 m below surface from the main face. From Fredlund *et al.*, 1985.

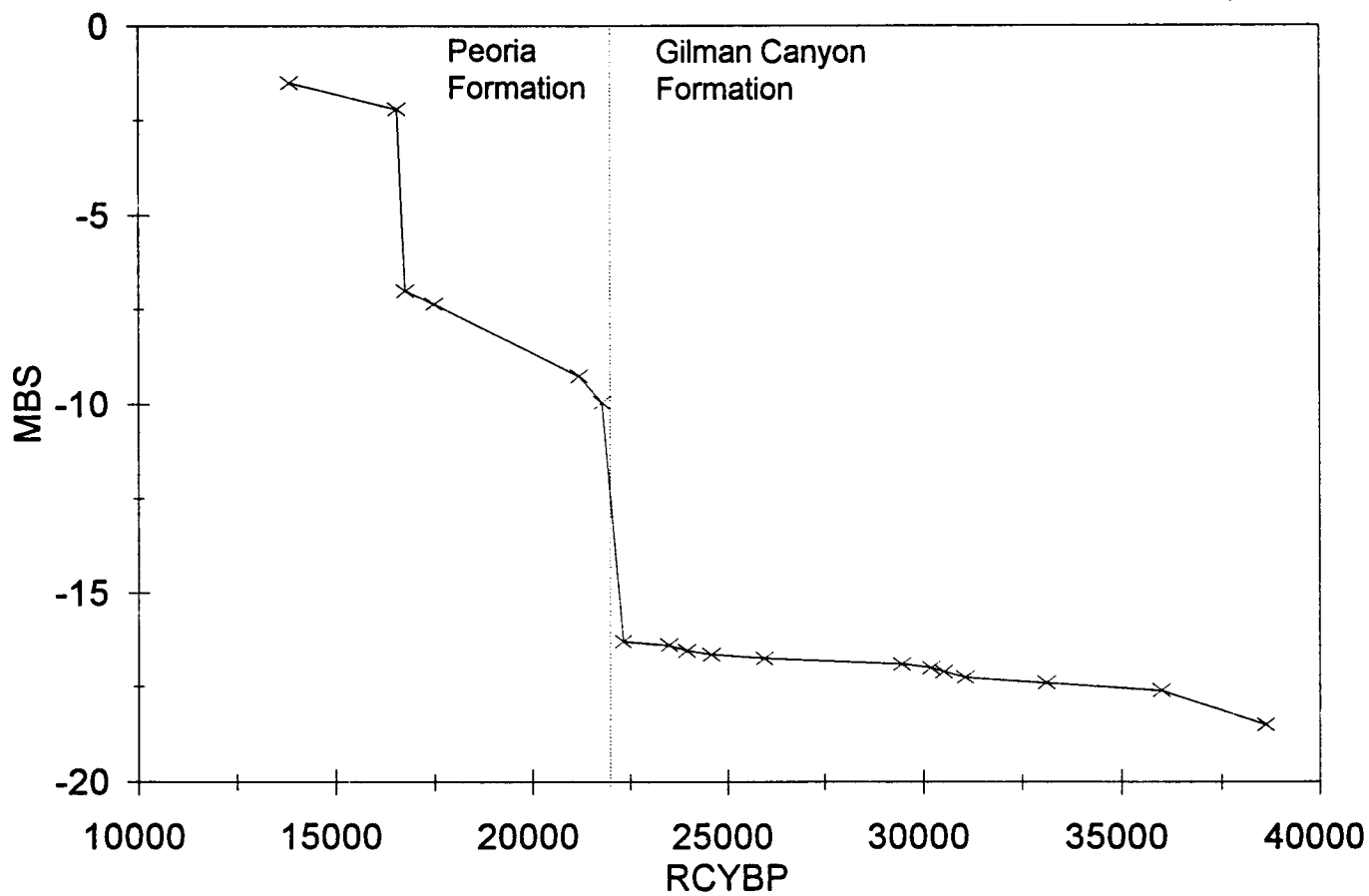


FIGURE 12-2. Sedimentation rate for the Gilman Canyon Formation and Peoria loess, profiles 1-3.

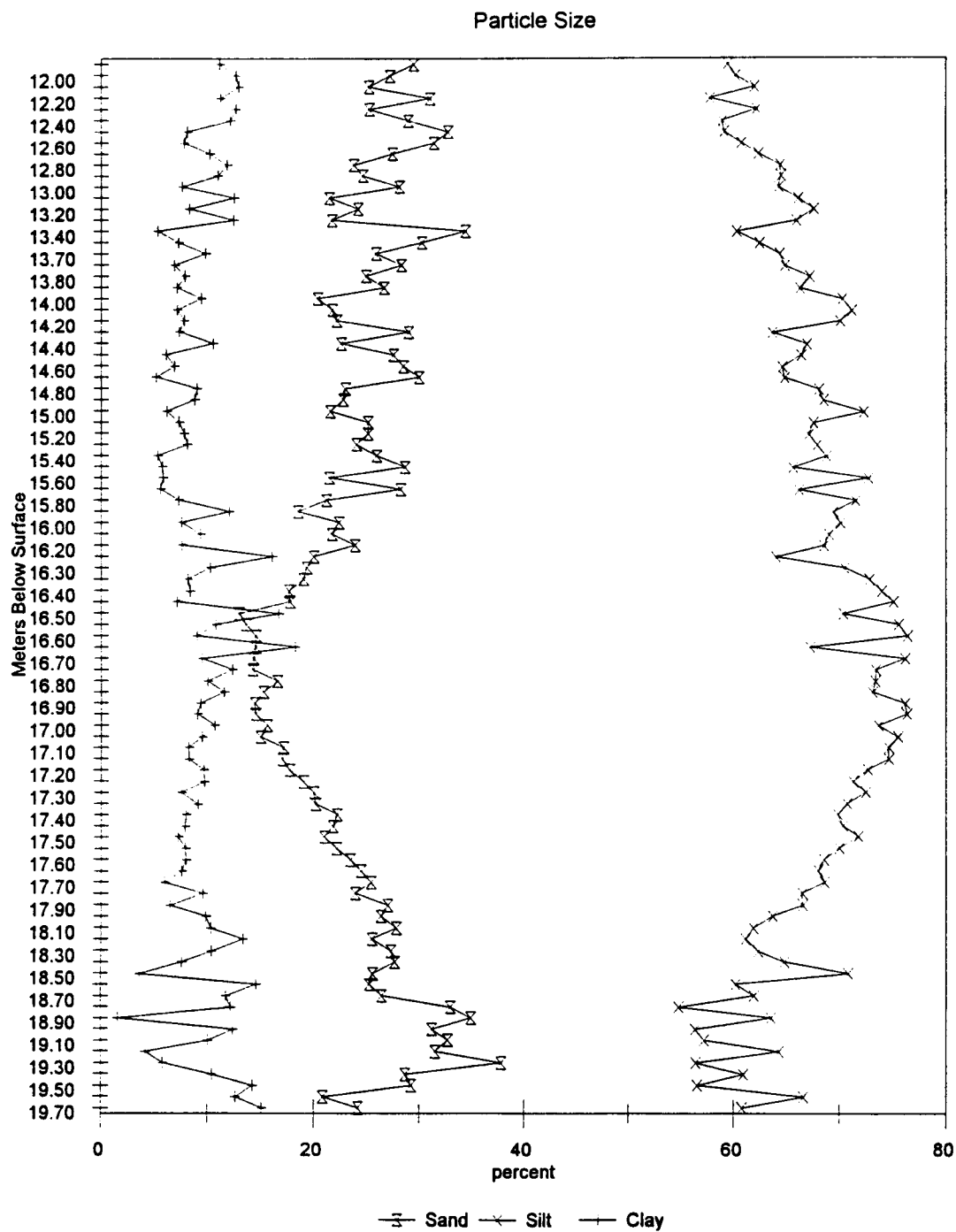


FIGURE 12-3. Particle size distribution for the lowermost Peoria loess and Gilman Canyon Formation, profiles 1-3.

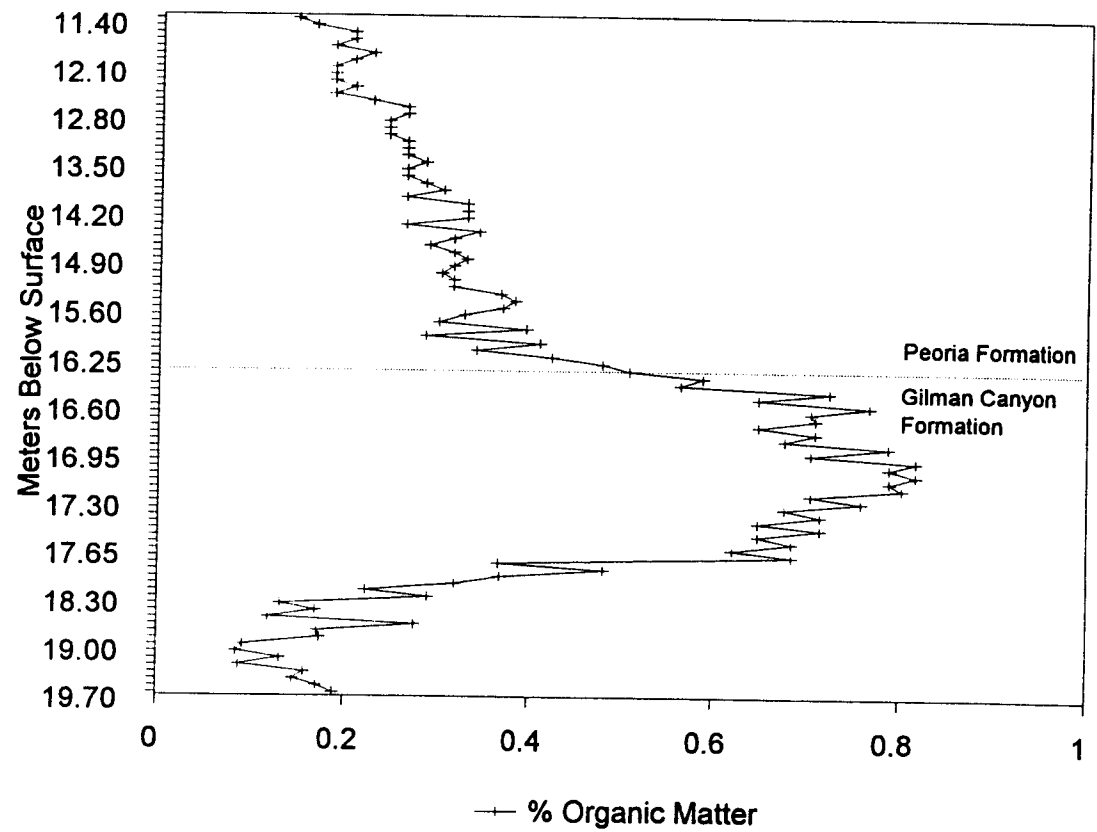
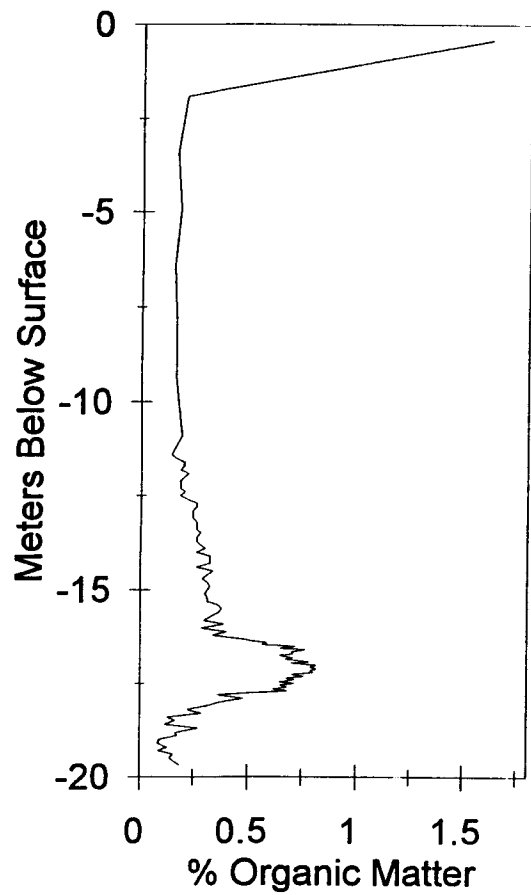


FIGURE 12-4. Organic matter content for the upper 20m, with fine resolution of the Gilman Canyon Formation.

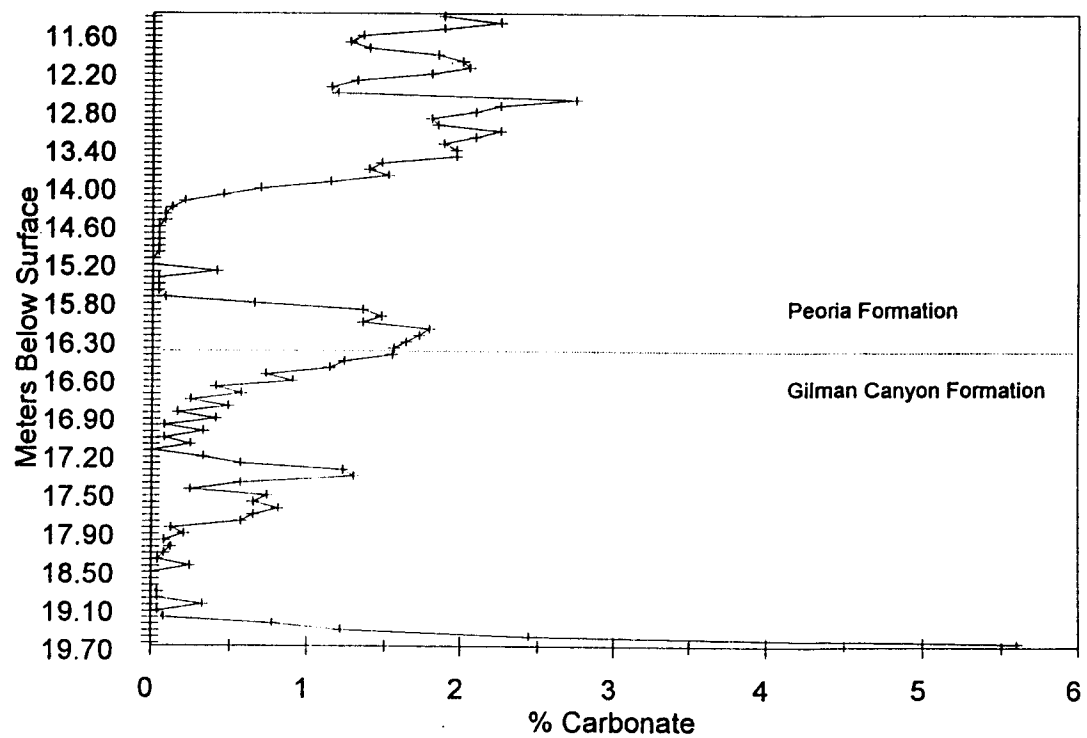
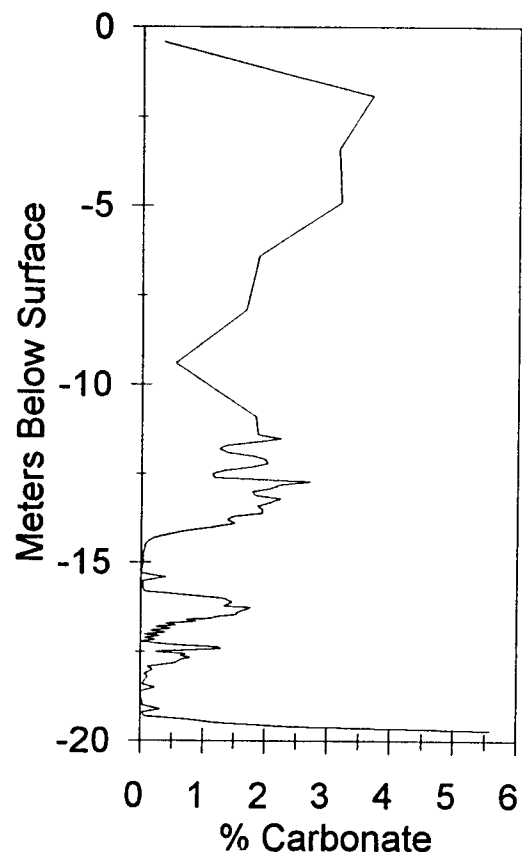


FIGURE 12-5. Carbonate content for the upper 20 m, with fine resolution of the Gilman Canyon Formation.

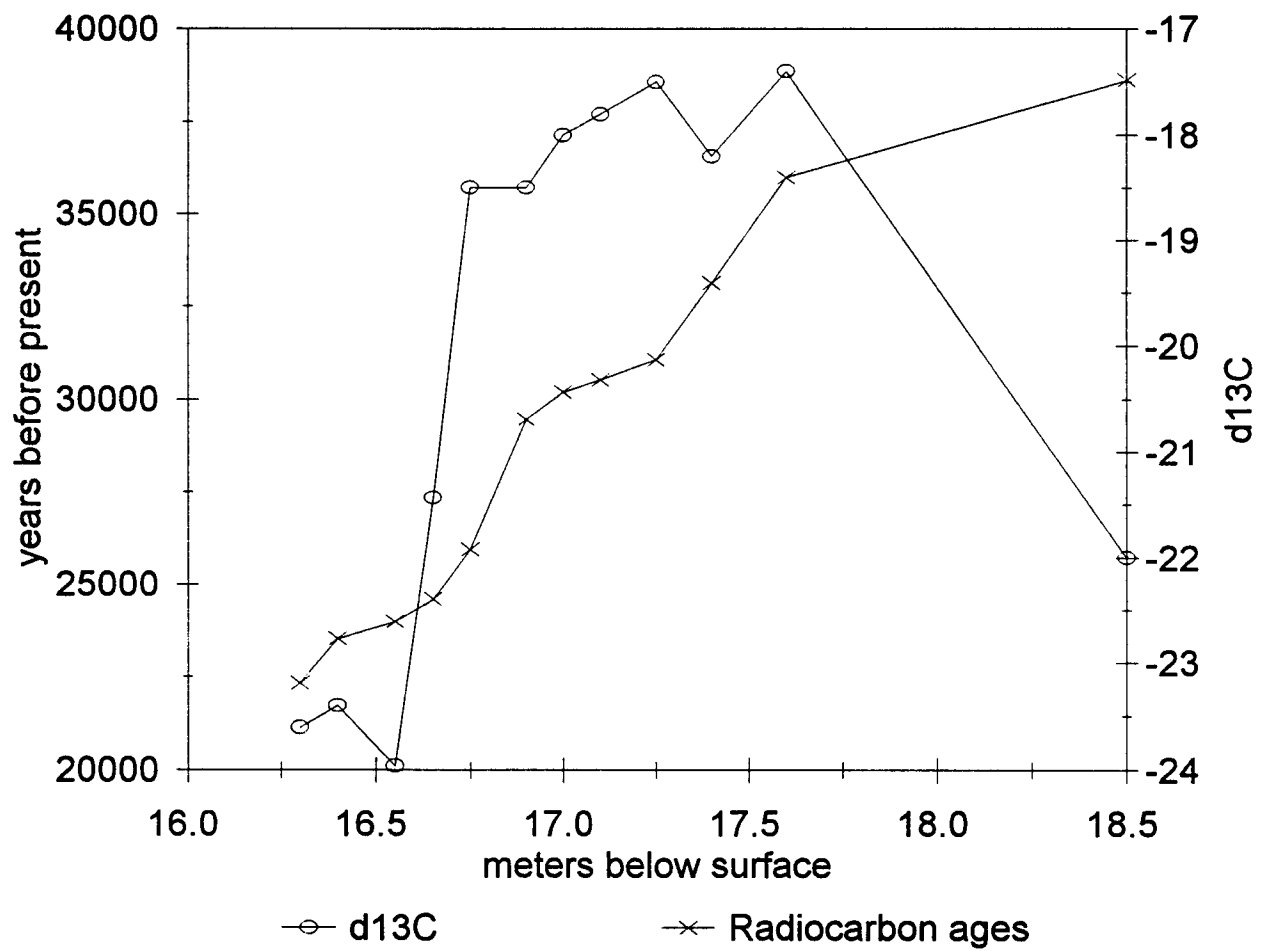


FIGURE 12-6. Radiocarbon age versus $\delta^{13}\text{C}$ value for the Gilman Canyon Formation.

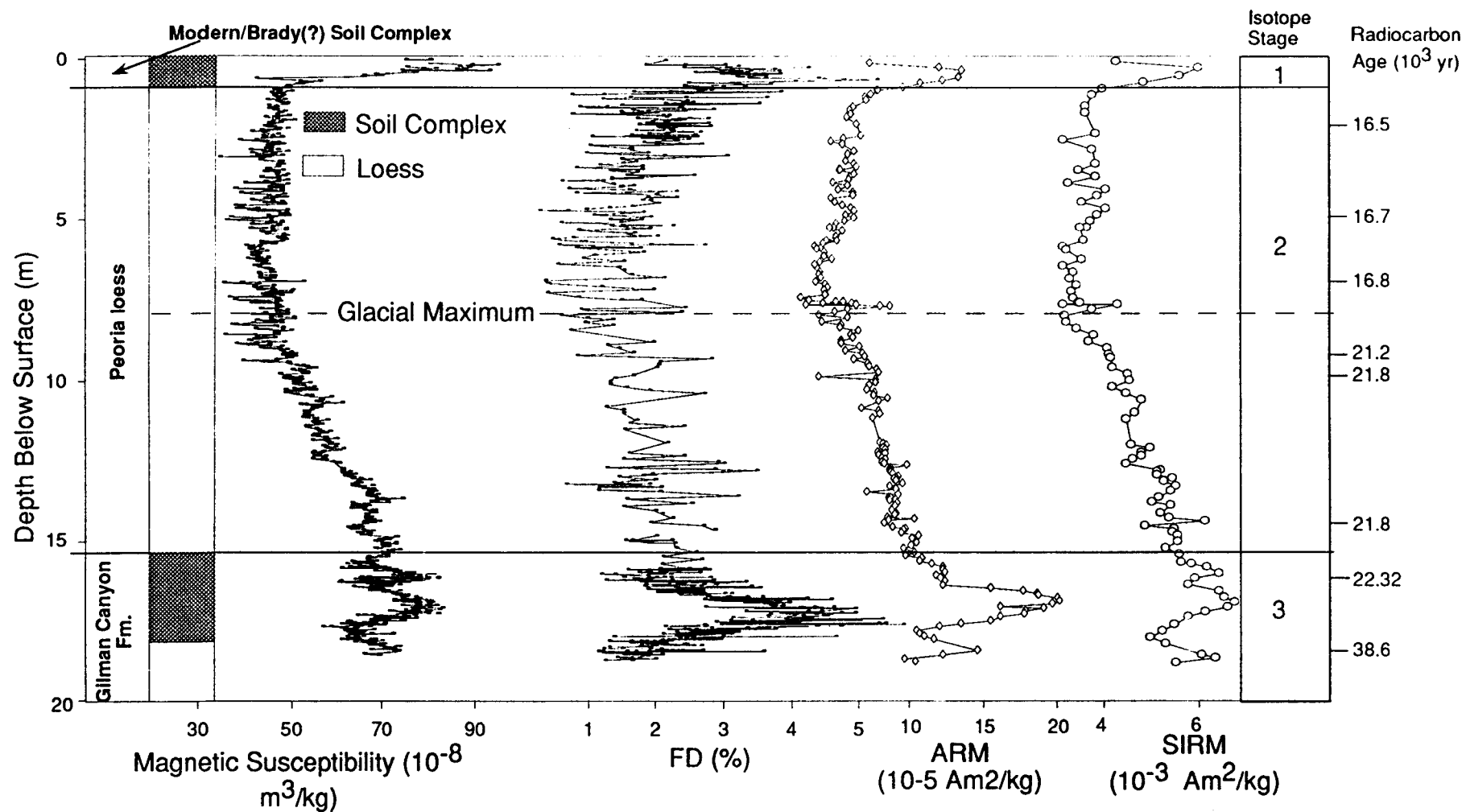


FIGURE 12-8. Stratigraphic variation of magnetic susceptibility, frequency dependence, anhysteretic remanent magnetization (ARM), and saturation isothermal remanent magnetization (SIRM) and soil stratigraphy of the Peoria loess and Gilman Canyon Formation at the Eustis ash pit (profiles 1-3).

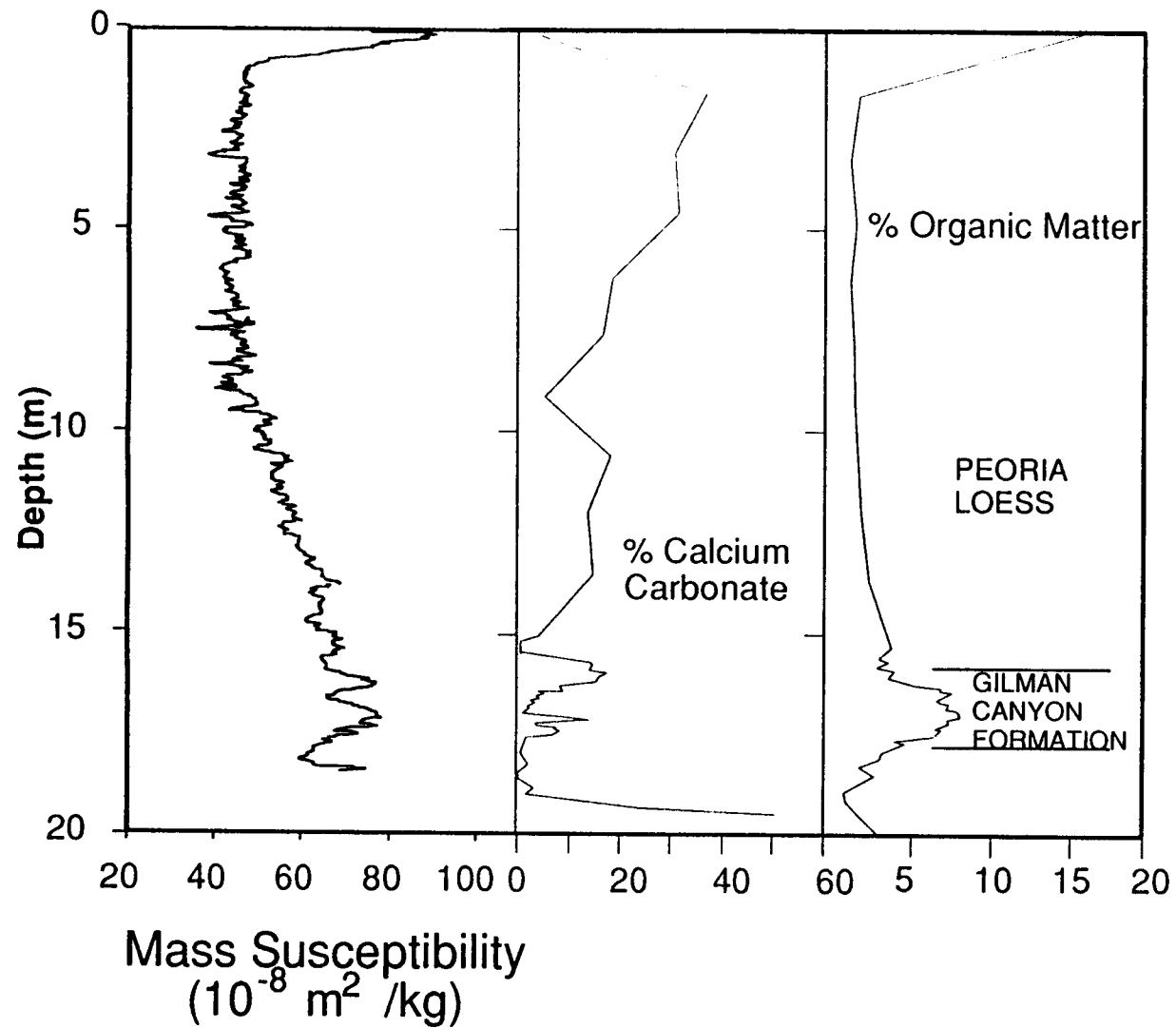


FIGURE 12-9. Magnetic (mass) susceptibility and two chemical parameters (CaCO_3 and OM) from the Peoria loess and Gilman Canyon Formation (profiles 1-3).

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Stop 13 Bignell Hill Roadcut, 1.7 miles due south of Bignell, Nebraska: Loess-soil sequence

Location: E1/2 E1/2 sec. 3, T12N R29W, Lincoln County, Nebraska
(USGS Maxwell, NE 7.5 min. quadrangle)

Contributors: William Johnson, Kyeong Park, and Randy Farr

TYPE LOCALITY OF THE LATE-PLEISTOCENE/EARLY HOLOCENE BRADY SOIL AND HOLOCENE BIGNELL LOESS

Introduction

Late-Quaternary paleosols and loess are exposed in a roadcut on the south valley wall of the Platte River in southwestern Nebraska, southeast of North Platte in Lincoln County. The view from the top of the roadcut provides a nice vista of the Platte River valley and the southern margin of the Sand Hill, which form the north valley wall. The section is the type locality for the late-Pleistocene/early Holocene Brady soil and overlying Holocene Bignell loess. Schultz and Stout (1945) originally described the Bignell Section as follows:

"Loveland fine sand is overlain by the *Citellus* Zone soil, very thick "Peorian" loess, soil X [cf. Brady soil; Schultz and Stout, 1948], Bignell loess, and a complex top soil. This is a representative section for the entire loess-canyon area. The *Citellus* Zone, which includes not only the soils at the top of the Loveland but also the transitional few feet of the Peorian loess is quite fossiliferous."

The roadcut exposure, covering a vertical distance of about 80 m from the top of the hill, or valley wall, to the road intersection at the base of the valley wall, consists of the Sangamon soil (?), Gilman Canyon Formation alluvial phase with soil, Gilman Canyon Formation loess and soil, Peoria loess, Brady soil, Bignell loess, and modern soil. Six ^{14}C ages (+ 2 pending) provide chronological control at Bignell Hill (Table 1).

Peoria Loess

The dominant feature at this section is the immense thickness of Peoria loess, about 50m (cf. Eustis ash pit with c. 16m). The proximity of the locality to the Platte River and the Sand Hills provided a tremendous source of silt and fine sand for transport southward by the north and northwest winds. This valley wall was, of course, the destination for some of that sediment. The overall accumulation rate for the Peoria loess in this section is quite high, i.e., $50\text{m}/10\text{k yr} \approx 0.5\text{mm/yr}$. The rate of accumulation was, however, not likely uniform. Scant stratigraphic information from the middle and lower Peoria suggest dramatically slowed accumulation and perhaps incipient soil development. Stratigraphic breaks in Peoria loess are recognized in Illinois (Frye *et al.*, 1974) and Iowa (Daniels *et al.*, 1960). However, no paleosols are recognized in Peoria loess in Nebraska or Kansas. Frye and Leonard (1949) observed a partially leached zone lacking molluscan faunal remains in Peoria loess in northeastern Kansas. Lenses of plant remains have been observed in Peoria loess in Kansas (Johnson 1993) and have been interpreted as discontinuities and possible periods of landscape stability during Peoria time. The magnetic susceptibility records show that

there are some periods of strong intensities, which may represent zones of incipient soil development/pre-burial weathering associated with pedogenesis or landscape stability (Fig. 1). Magnetic susceptibility data (discussed below) from the Eustis ash pit and Bignell Hill exhibit a high degree of correlation (Fig. 2).

The upper part of the Peoria loess (c. 17-7m) indicates a high rate of accumulation via laminations similar to those observed at the Eustis ash pit. The laminations can be differentiated by particle size and carbonate content as at Eustis. Laminations, 0.5-1.0 cm in thickness, appear to be annual given the current age control and the concentration of *Picea glauca* (white spruce) needles and twigs on the upper faces of the individual laminar units. An ongoing opal phytolith analysis of loess will soon provide a much better picture of vegetation (and climate) prevailing during Peoria time.

Brady soil

The Brady soil was first named and described by Schultz and Stout (1948) at this site. The soil is developed within the Peoria Loess and is overlain by the Bignell Loess. The name was subsequently adopted by researchers in Kansas (Frye and Fent, 1947; Frye and Leonard, 1949, 1951; Frye et al., 1949). It is regionally extensive only in the northwestern and west central parts of Kansas, and even there it occurs discontinuously on the landscape. Frye and Leonard (1951) and Caspall (1970, 1972) recognized Brady development in northeastern and other parts of Kansas. Without the overlying Bignell Loess, the Brady soil does not appear to exist; the modern surface soil may have incorporated post-Bradyan loess fall into its profile. The Brady soil is typically dark gray to gray-brown and better developed than the overlying surface soil within the Bignell. Strong textural B horizon development and carbonate accumulation in the C horizon are typical, although it occasionally displays evidence of having formed under poorer drainage conditions than have associated surface soils (Frye and Leonard, 1951). Feng (1991) noted that the Brady soil, as expressed at the Barton County Sanitary Landfill, is strongly weathered both physically and chemically.

The age of the Brady soil had, until recently, been uncertain, even here at the type locality. Dreeszen (1970, p.19) reported an age of 9160 ± 250 (W-234) obtained in 1954 and another in 1965 of 9750 ± 300 (W-1676), both from the type section but very likely contaminated by modern plant roots. Subsequently, Lutenecker (1985) reported an age of 8080 ± 180 yr B.P. but provided few specifics other than that the source was the A horizon of the Brady soil at the type section. Better age control for the type section was recently secured as part of the ongoing study: ages of $9,240 \pm 110$ (Tx-7425) and $10,670 \pm 130$ (Tx-7358) yr B.P. were obtained on the upper and lower 5 cm (2 in), respectively, of the Brady A horizon (Table 1).

The Brady soil has also been recently dated at other localities in Nebraska and Kansas. Souders and Kuzila (1990) obtained a ^{14}C age of $10,130 \pm 140$ years B.P. on the Brady soil occurring within the Republican River valley of southcentral Nebraska. Sites along Harlan County Lake have yielded a number of ages, ranging from $10,550 \pm 160$ to $9,020 \pm 95$ years B.P., on exposures of the Brady soil (Cornwell, 1987; Johnson, 1989; Martin, 1990; Martin and Johnson, unpub. data). Two ^{14}C ages of 9820 ± 110 (TX-7045) and $10,550 \pm 150$ (TX-7046) years B.P. have been derived from the upper and lower 5 cm (2 in), respectively, of the Brady A horizon exposed at the Barton County site (Feng, 1991). The two radiocarbon ages from the Speed roadcut in Phillips County agree reasonably well with those from the type section in adjacent Nebraska and central Kansas (Table 2; Fig. 3).

Although it appears Brady pedogenesis occurred from about 10,500 to as recently as 8,500 years B.P. (Fig. 4), greater refinement of the Brady soil chronology is necessary, but present data clearly indicate it was a product of a major period of landscape stability at a time when widespread climatic shifts were occurring at the end of the Wisconsin. This was the first significant period of soil development since Gilman Canyon

time, and represents the climate of the early Holocene. There is an isochronous alluvial soil found throughout the region which is particularly well expressed within the Kansas River basin (Johnson and Martin, 1987; Johnson and Logan, 1990). The two ages of 8274 ± 500 (C-108a) and 9880 ± 670 (C-471) years B.P. determined from alluvial fill (Fill 2A) at archaeological sites Ft-50 and Ft-41 on Harry Strunk Lake in southwestern Nebraska (Schultz et al., 1951; Libby, 1955) were the first radiocarbon determinations on the Brady soil. The soil, occurring in both eolian and alluvial contexts, qualifies, based upon present ^{14}C data, as a *geosol*.

Development of the Brady soil correlates well with indicators of regional climatic change. The fossil pollen record at Muscotah Marsh of northeastern Kansas indicates that spruce had essentially disappeared from the region by about 10,500 yr B.P. As this decline occurred, deciduous tree species increased until about 9,000 yr B.P., the time at which grassland expansion began (Gruger, 1973). On a hemispheric scale, the abrupt decrease in atmospheric dust noted in the Greenland ice core at 10,750 yr B.P. (Paterson and Hammer, 1987) reflects decreased loess deposition and possibly Brady-age pedogenesis associated with relative terrestrial stability. Further, ^{18}O levels within the same core suggest rapid warming about 10,750 yr B.P., with the characteristic Holocene temperature regime being established about 9,000 yr B.P. At least the early and perhaps all of Brady soil forming interval coincide with the Younger Dryas cold interval of the North Atlantic region (Table 3). Here at the Bignell Hill type section, the beginning of Brady soil forming interval matches well with the oxygen isotope data from Greenland Icecore (Dansgaard et al., 1985). Also, there is a minor, but notable kink in the susceptibility data just below the Brady soil that may indicate climatic degradation comparable to Younger Dryas cold spell (Fig. 5). A rapid rate of dust accumulation, approaching 1 m/1000 yr in some profiles on the Loess Plateau in China has led to the preservation of a high-resolution record of Late Quaternary environmental changes (An et al., 1993). They found a minor fluctuation which is believed to represent the Younger Dryas cold spell.

Stable carbon isotope data ($\delta^{13}\text{C}$) from the total organic fraction reflect inputs by the plants, particularly grasses growing on those surfaces. C3 (cool-season/humid) species have an average $\delta^{13}\text{C}$ composition of -27‰ and C4 species (warm-season/arid) -13‰, relative to the PDB standard (Deines 1980). $\Delta^{13}\text{C}$ values acquired in association with the ^{14}C ages for the Brady soil forming period indicate gradual changes from a C3 to C4 environment, with minor fluctuations (Fig. 6). This indicates that vegetation and thus climate of Wisconsin age was characterized by a cool, moist C3 plant environment, and that of the latter Brady period by conditions similar to those of the present, warm and sub-humid.

Bignell Loess

The Bignell loess, first described at this site, is typically a gray or yellow-tan, massive silt, calcareous and seldom more than 1.5 m thick. Although it is often somewhat less compact and more friable than the underlying Peoria loess, no certain identification can be made without the presence of the Brady soil. The Bignell loess does not form a continuous mantle on the Peoria; instead, it occurs as discontinuous deposits which are most prevalent and thickest adjacent to modern-day valleys, particularly the south side, and often within depressions on the Peoria surface, such as here at Bignell Hill. Feng (1991) speculates that the Bignell loess of central Kansas is relatively well weathered because it was derived from a preweathered source, the Brady soil surface, perhaps eolian and alluvial phases alike. This is consistent with the earlier interpretation derived in Nebraska that Bignell loess is at least partially comprised of re-worked Peoria loess (Condra et al., 1947, p. 33).

It appears from the ^{14}C -age determinations obtained here at the type section in Nebraska and the Speed roadcut that the Bignell loess can be no older than about 8,000 yr B.P. Snails collected by A.B. Leonard from the lower part of the Bignell in Doniphan County, northeastern Kansas, produced ages of $12,500 \pm 400$ (W-231) and $12,700 \pm 300$ (W-233) yr B.P. (Frye and Leonard, 1965). Because the shell material had

absorbed an indeterminate amount of dead carbonate, Frye and others (1968) proposed an averaged age of approximately 11,000 years. Based upon the age data available for the Brady, the soil humate-derived ages are probably closest to reality.

A pronounced feature of the Holocene climate of the Plains was an extended warm, dry period (Wright, 1970; Benedict and Olson, 1978; Barry, 1983), identified as the Altithermal (Antevs, 1955) or, less commonly, as the Hypsithermal (Deevey and Flint, 1957). This dictates that the Bignell was a warm-climate loess, unlike the cold-climate loess of the Woodfordian. Reconstruction of the general circulation patterns for North America indicates that from the last glacial maximum about 18 to 15 ka there was no detectable change in atmospheric circulation: the westerly jet was split by the Laurentide ice sheet into a north and south flow around a strong glacial anticyclone (Kutzbach, 1985, 1987; COHMAP Members, 1988). By 9 ka, the ice had wasted appreciably, the jet was no longer split, orbital parameters were favoring increased temperatures, and zonal flow was dominating (Kutzbach, 1981, 1985, 1987). Model results produced mean summer temperatures 2° to 4° C higher (COHMAP Members, 1988) and annual precipitation up to 25 % less than at present in the region (Bartlein et al., 1984; Kutzbach, 1987).

Because of the increasing zonal flow and aridity of the Altithermal, species of the tall grass community migrated eastward to the present areas of mixed deciduous-prairie vegetation, i.e., the prairie-forest ecotone shifted eastward (Van Zant, 1979; Semken, 1983; Webb et al., 1983). The fossil pollen record from Muscotah Marsh provides a disrupted but interpretable Holocene signal, indicating a middle Holocene prairie expansion (Gruger, 1973). Fossil pollen data from Cheyenne Bottoms suggest consistently lower water levels in the marsh during the middle Holocene (Fredlund, 1991). Molluscan fauna from the Bignell loess of Kansas suggest that climate was somewhat drier than during Peoria time (Frye and Leonard, 1951). After a period of soil formation near the end of the Pleistocene, pedogenesis is not recognized until about 5,800 years B.P. in the sand of the Great Bend Prairie, central Kansas (Johnson, 1991a). Therefore, based upon various climatic proxies and a limited number of ¹⁴C ages, it appears the Bignell Loess was deposited, for the most part, from the end of Brady pedogenesis at about 8,500 to about 5,500 yr B.P.

Magnetic Mineralogy

A great deal of attention has been given to paleomagnetic measurements of loess-paleosol sequences, particularly in China and in Europe. Although much research has focused on paleomagnetic events such as excursions, magnetic data have been used to successfully reconstruct the climatic sequences. Recent work has explained this magnetic-global climate relationship in terms of increased pedogenic production of ferrimagnetic grains during the warmer interglacial intervals (e.g. Maher and Thompson, 1991; Heller et al., 1991).

Contiguous samples were collected (40-50/m) from the upper 18 m of the Bignell Hill section using demagnetized, plastic cubes. Bulk magnetic characteristics were measured using a Bartington MS-2 type dual frequency susceptibility sensor, pulse magnetizer, and magnetometer. Magnetic susceptibility measures magnetization temporarily induced in a rock by an artificially applied, low-amplitude magnetic field. The strength of the susceptibility signal depends on the concentration and grain size of the magnetic minerals. Magnetic susceptibilities in the Brady soil at Bignell are nearly twice that of the unweathered Peoria loess and higher than in the modern soil (Fig. 1). Susceptibility plots from the Eustis ash pit and Bignell Hill correlate well (Fig. 2) and show similar enrichment in the paleosols and depletion in the relative unweathered loess intervals. The ultimate control on these correlations is likely to be climate.

Frequency dependency (FD) measures the presence of grains near the stable single domain/superparamagnetic size boundary (c. 0.03 μ m dia.). FD can effectively identify weathering and soil profile development, which gives rise to increased concentrations of extremely fine grained ferrimagnetic

crystals. Magnetic susceptibility data correlate well with frequency dependency (Fig. 7). Magnetic susceptibility values in the Bignell loess are intermediate between the Peoria loess and the Brady soil. The Brady soil can be differentiated from the Gilman Canyon Formation soil by examining the FD and high-frequency susceptibility (Table 4). Susceptibility intensities of the Brady soil are considerably higher than the Gilman Canyon soil complex, but FD is consistently lower.

Weathering and pedogenesis are best expressed in the NRM (natural remanent magnetization) data (Fig. 8). NRM measures the residual magnetization possessed by rocks, sediments and soils. NRM intensities in Brady soil are 3-4 times larger than in the unweathered loess. The intensities of NRM and low field susceptibility vary by 2-3 orders of magnitude throughout the section, being closely correlated to one another. Higher intensities characterize the Brady soil, whereas low values are typical of the unweathered Peoria loess. Also, NRM data differentiate between the Brady soil and Bignell loess (Fig. 8). These features are ascribed to the pedogenic processes, which promote magnetic mineral enrichment in the soil due to alteration in a relatively warm climate (Heller and Liu, 1982; Liu *et al.*, 1992).

Saturation isothermal remanent magnetization (SIRM) of the Brady soil is about twice that of Peoria loess samples (Fig. 9). SIRM acquisition and thermal demagnetization diagrams indicate that both loess and paleosol samples contain low-coercivity ferrimagnetic phases (Fig. 9 and 10). The maximum unblocking temperatures between 550° and 600°C indicate that much of this ferrimagnetic material is magnetite. However, the abrupt drop in intensity below 400°C in many samples (Fig. 10) probably reflects the conversion of maghemite to hematite. Thus it is likely that both magnetite and maghemite are present in loess and paleosol samples. Recent work has suggested that both magnetite and maghemite may be formed during either inorganic reduction/oxidation cycles in soils (Maher and Taylor, 1988; Singer and Fine, 1989). Magnetite can also form in soils from the activity of magnetotactic bacteria (Fassbinder *et al.*, 1990).

TABLE 1. Radiocarbon ages from the Bignell Hill section.

Lab No.	Depth	Uncorr. Age	^{13}C	Corrected age	Unit
Tx-7425	1.80-1.85	$9,110 \pm 110$	-17.4	$9,240 \pm 110$	upper Brady soil
Tx-7358	2.15-2.20	$10,580 \pm 130$	-19.3	$10,670 \pm 130$	lower Brady soil
Beta-60,508	6.50	$11,880 \pm 90$	n.a.	n.a.	Peoria loess
Tx-7921	8.45-8.50	Pending			Peoria loess
Tx-7922	17.45-17.50	Pending			Peoria loess
Tx-7706	49.7	$28,130 \pm 610$	-17.6	$28,250 \pm 620$	GCF (loessal soil-upper)
Tx-7707	50.9-51	$30,970 \pm 780$	-24.3	$30,980 \pm 790$	GCF (loessal soil-lower)
Tx-7708	70.2-70.3	$33,750 \pm 1110$	-24.8	$33,760 \pm 1110$	GCF (alluvial phase)

All ages were determined from total humates, except for Beta-60,508 (charcoal; AMS)

TABLE 2. Radiocarbon ages from the Brady soil of Nebraska and Kansas.

Nebraska		
Bignell Hill		
	8,080 ± 180	Lutenegger, 1985
	9,160 ± 250	Dreeszen, 1970
	9,750 ± 300	Dreeszen, 1970
	9,240 ± 110	Johnson, 1993
	10,670 ± 130	Johnson, 1993
North Cove		
West		
	10,550 ± 160	Johnson, 1989
	10,220 ± 140	Johnson, 1989
	10,270 ± 160	Johnson, 1989
East		
	11,530 ± 150	Johnson, 1989
	11,025 ± 90	Martin & Johnson, unpub.
Prairie Dog Bay		
	10,140 ± 110	Cornwell, 1987
	10,360 ± 130	Martin, 1990
	9,020 ± 95	Martin & Johnson, unpub.
Naponee		
	10,130 ± 140	Souders and Kuzila, 1990
Kansas		
Speed		
	8,850 ± 140	Johnson, 1990
	10,050 ± 160	Johnson, 1990
Barton County		
	9,820 ± 110	Feng, 1991
	10,550 ± 150	Feng, 1991

Ages represent the eolian phase only. An alluvial phase has been well documented throughout the Kansas River basin (Johnson and Martin, 1987; Johnson Logan, 1990) and adjacent river systems, such as the Loup River of central Nebraska (Brice, 1964; May, 1990) and the Pawnee River (Mandel, 1991) and Walnut River (Mandel, 1991) of the Arkansas River system.

TABLE 3. Chronology of the Younger Dryas.

Begin	End	Reference	Region
	10,720±150	Dansgarrrd et al., 1989	Greenland Icecore
11,000	10,000	Kudrass et al., 1991	Sulu Sea, SE Asia
11,000	10,300	Shane, 1987	Ohio
11,000	10,000	Broecker et al., 1988	EN32-PC4, Orca Basin
10,800	10,000	Engstrom et al., 1990	Alaska
11,290	10,170	Mathewes et al., 1993	British Columbia
11,000	10,000	Mott et al., 1986	Atlantic Canada
11,200	10,500	Lehman and Keigwin, 1992	North Atlantic
	10,580-10,950	An et al., 1993	China

TABLE 4. Composite high frequency susceptibility and frequency dependency mineral magnetic data.

Strat. Unit	High Frequency Susceptibility	HF stand. dev.	Frequency Dependency	FD stand. dev.
Modern Soil	70.11	10.85	3.01	1.19
Bignell Loess	104.00	4.11	1.58	0.63
Brady Soil	112.75	5.90	3.14	0.75
Peoria Loess	48.28	5.41	1.93	0.87
GCF	74.82	5.37	3.32	0.76
Unaltered GCF	70.35	5.77	2.12	0.62

Values for the modern soil, Bignell loess and Brady soil are from the Bignell Hill section, and Peoria loess and Gilman Canyon formation (GCF) from the Eustis ash pit.

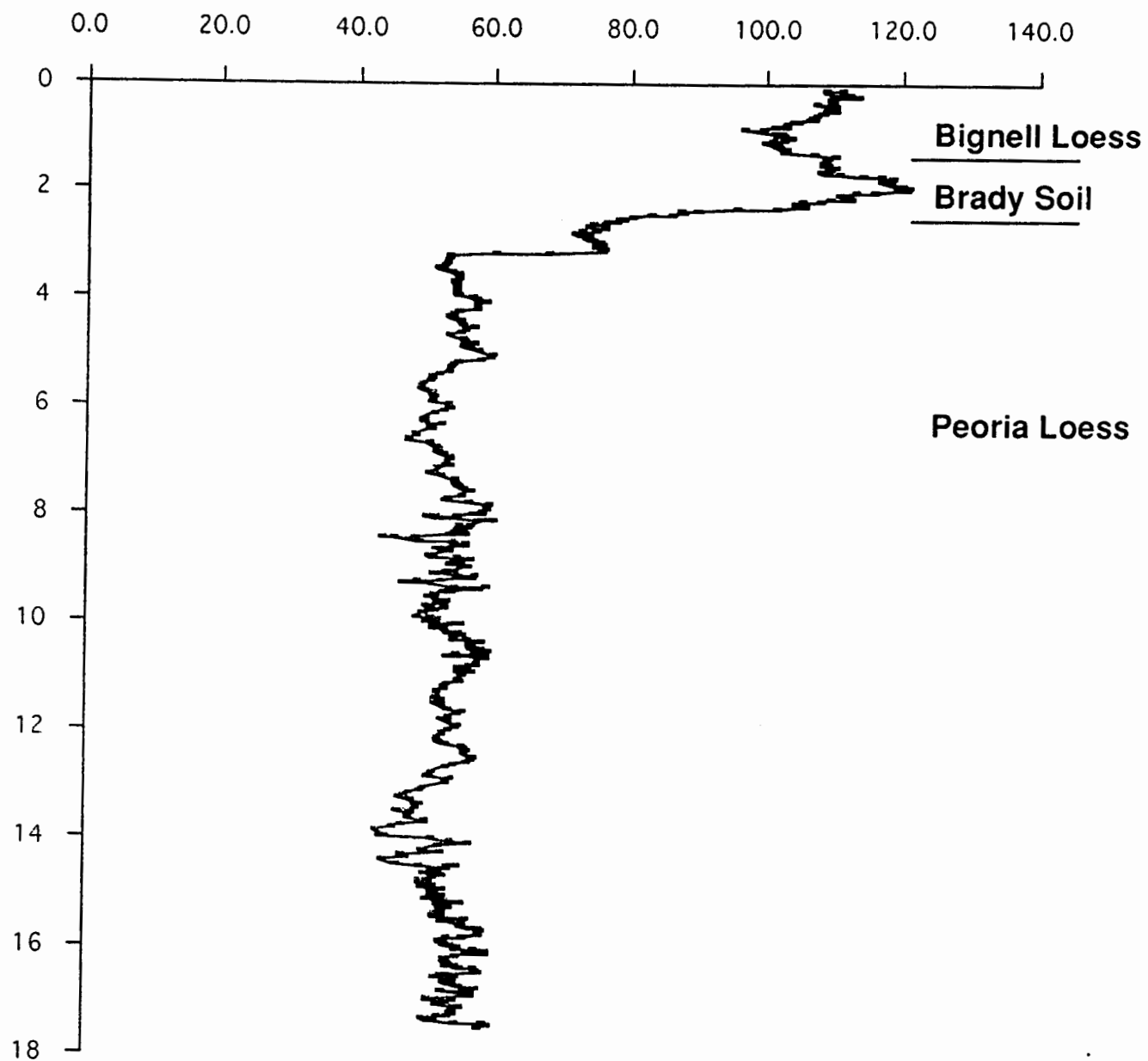


FIGURE 13-1. Mass susceptibility data for the upper 18m of Bignell Hill. A 3-point moving average has been used for the plot.

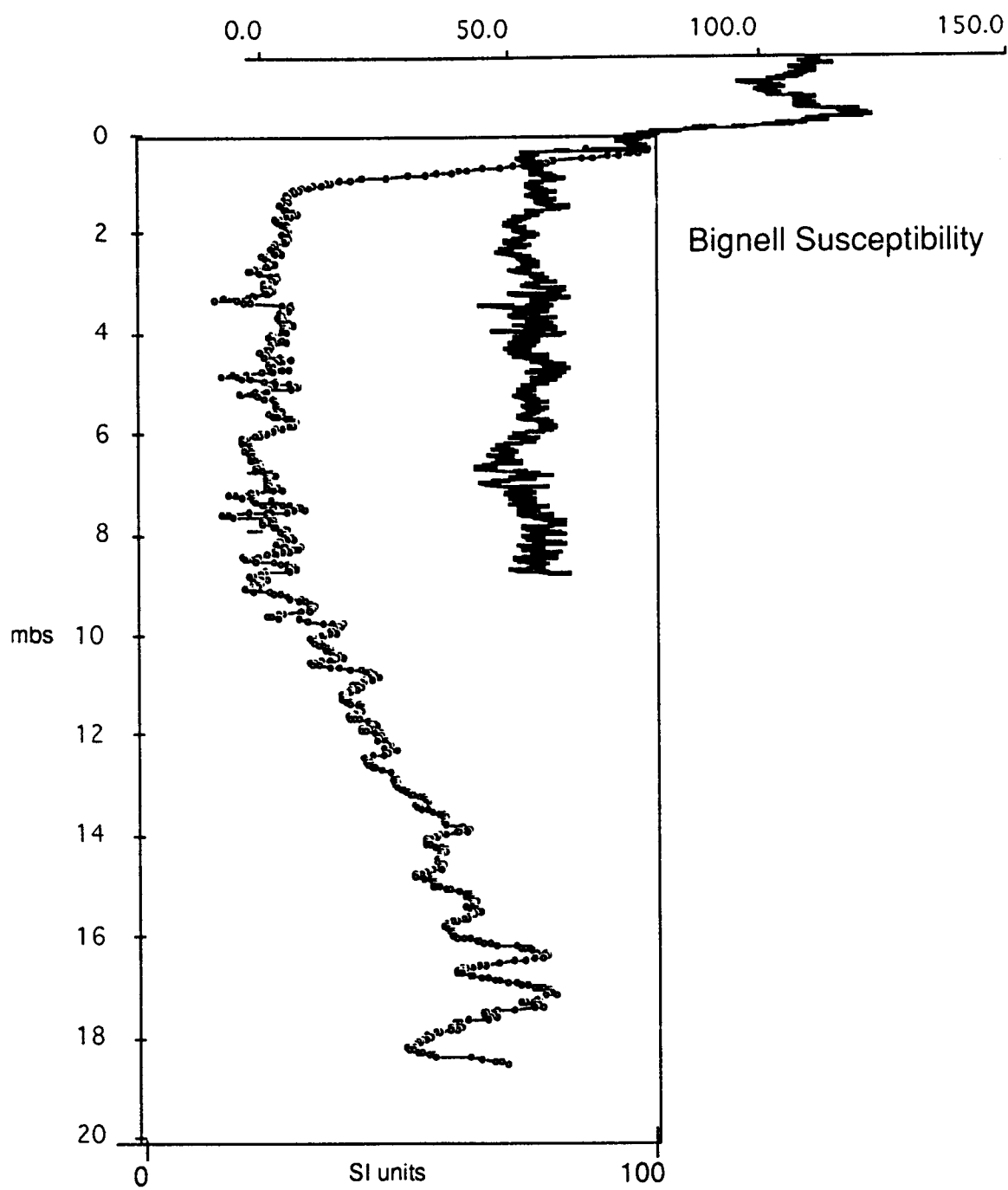


FIGURE 13-2. Mass susceptibility data for the Eustis ash pit and Bignell Hill. Note that the SI scales are offset for greater clarity.

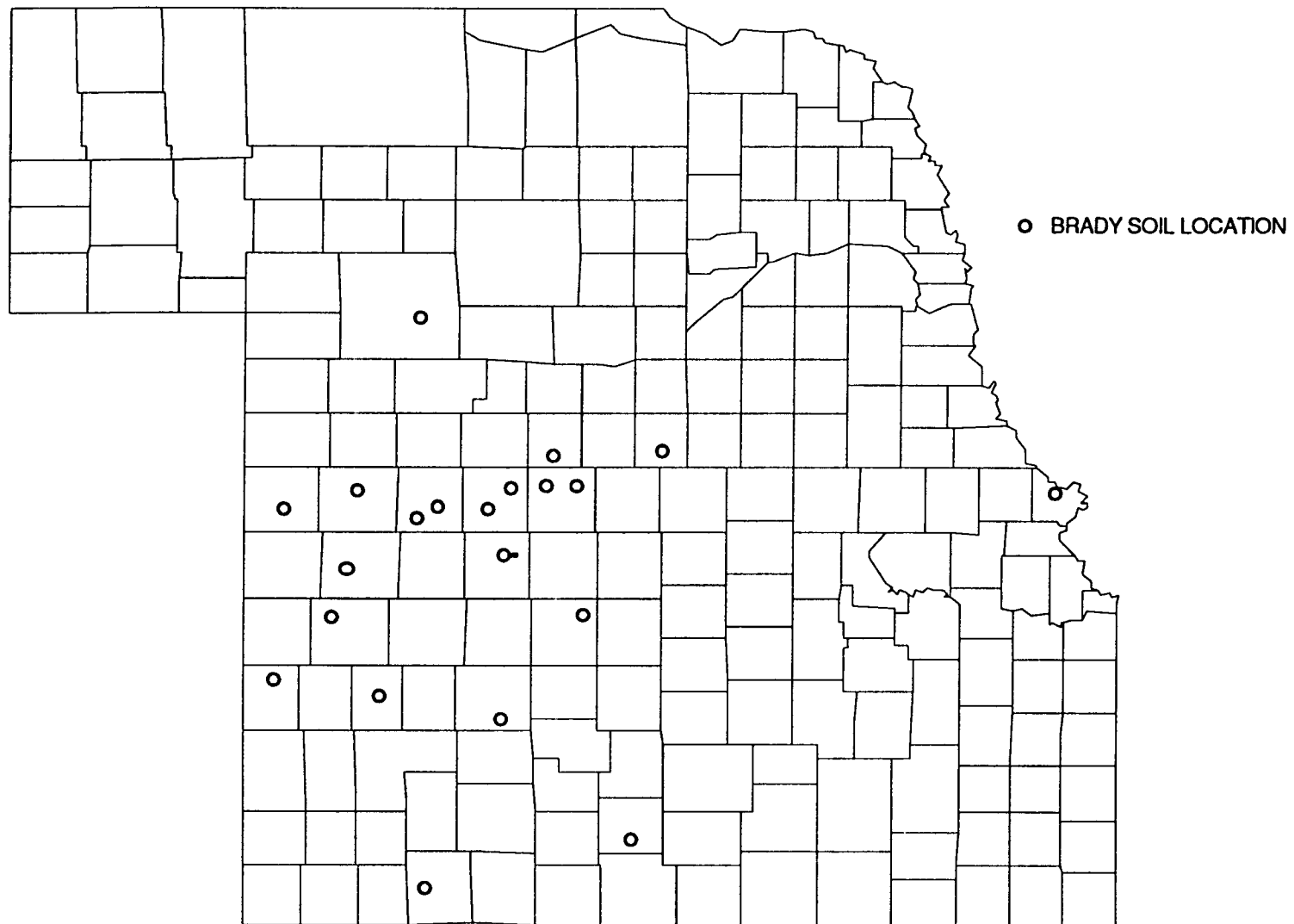


FIGURE 13-3. Reported locations for exposures of the Brady soil. The northwesternmost locality is Bignell Hill.

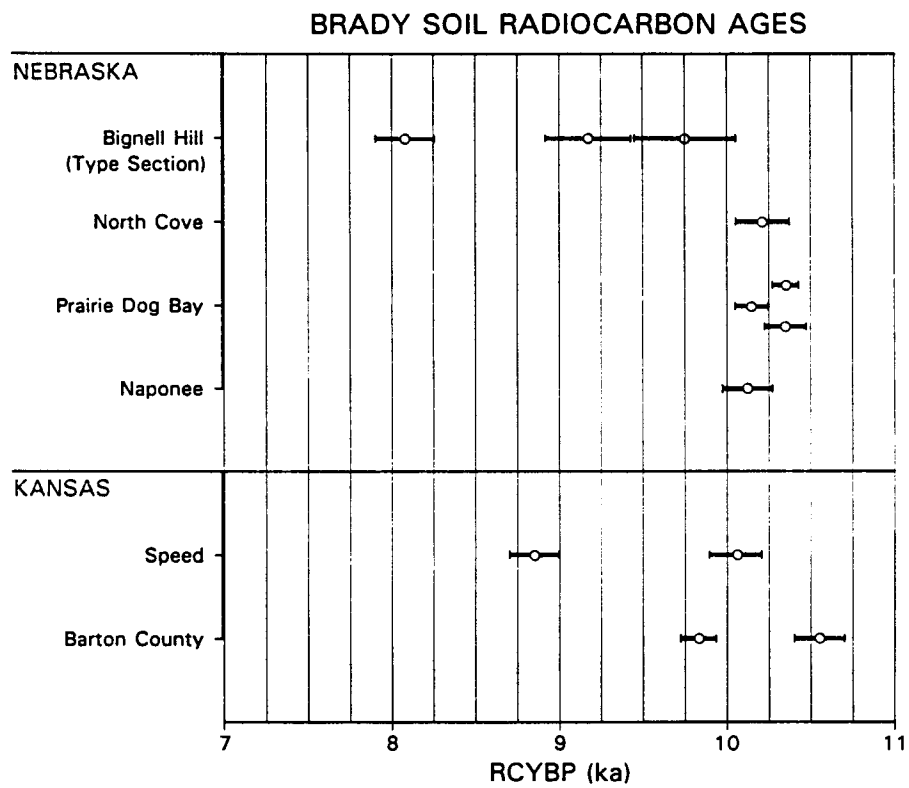


FIGURE 13-4. Graphic representation of the reported ^{14}C ages for the Brady soil.

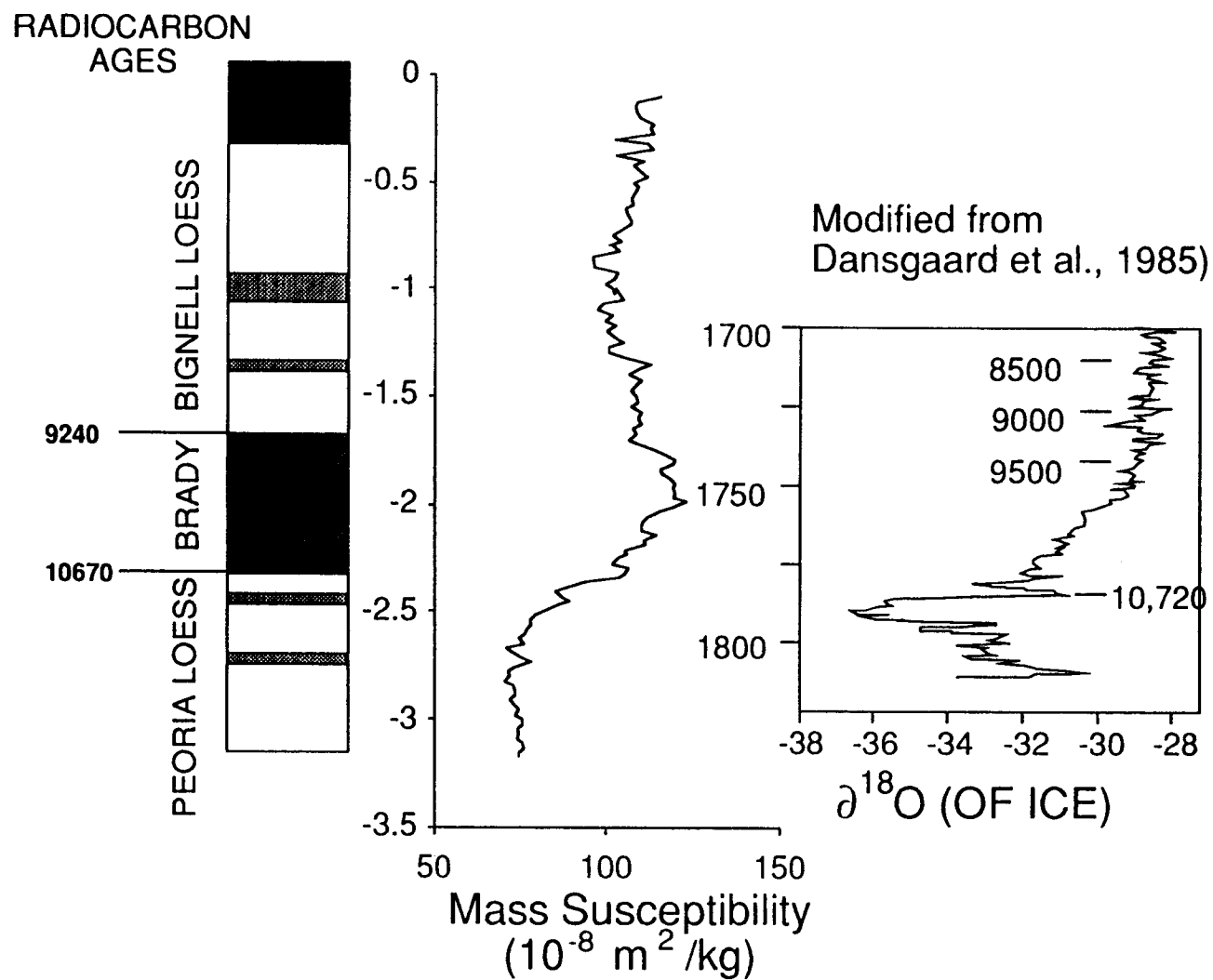


FIGURE 13-5. Mass susceptibility of the uppermost Peoria loess, Brady soil, and Bignell loess at Bignell Hill and the $\delta^{18}\text{O}$ values from a Greenland ice core.

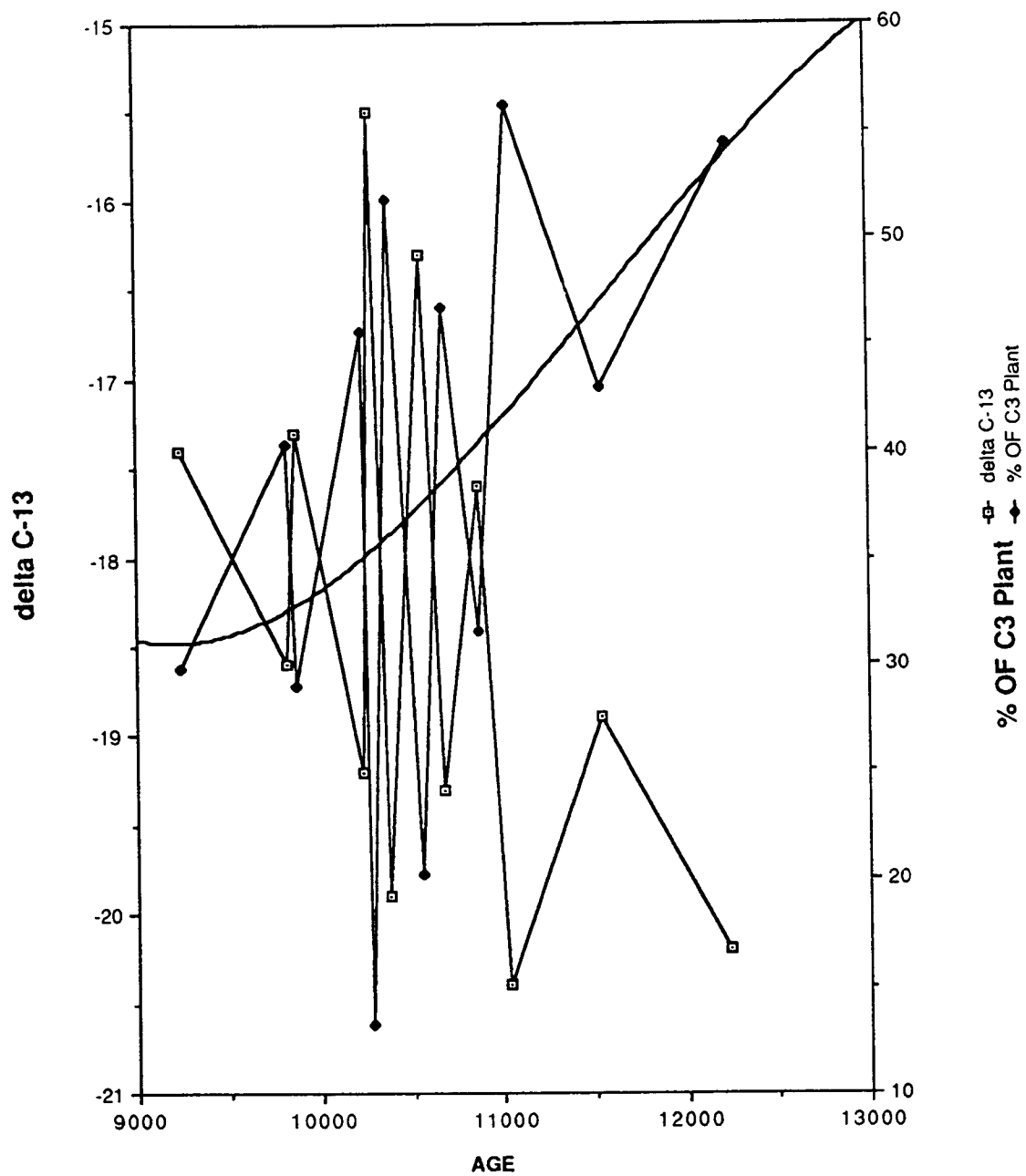


FIGURE 13-6. Percent C3 (cool, moist adapted) plants as derived from the $\delta^{13}\text{C}$ values of ^{14}C -dated Brady soil samples. The curve was fitted from both data sets.

RADIOCARBON AGES

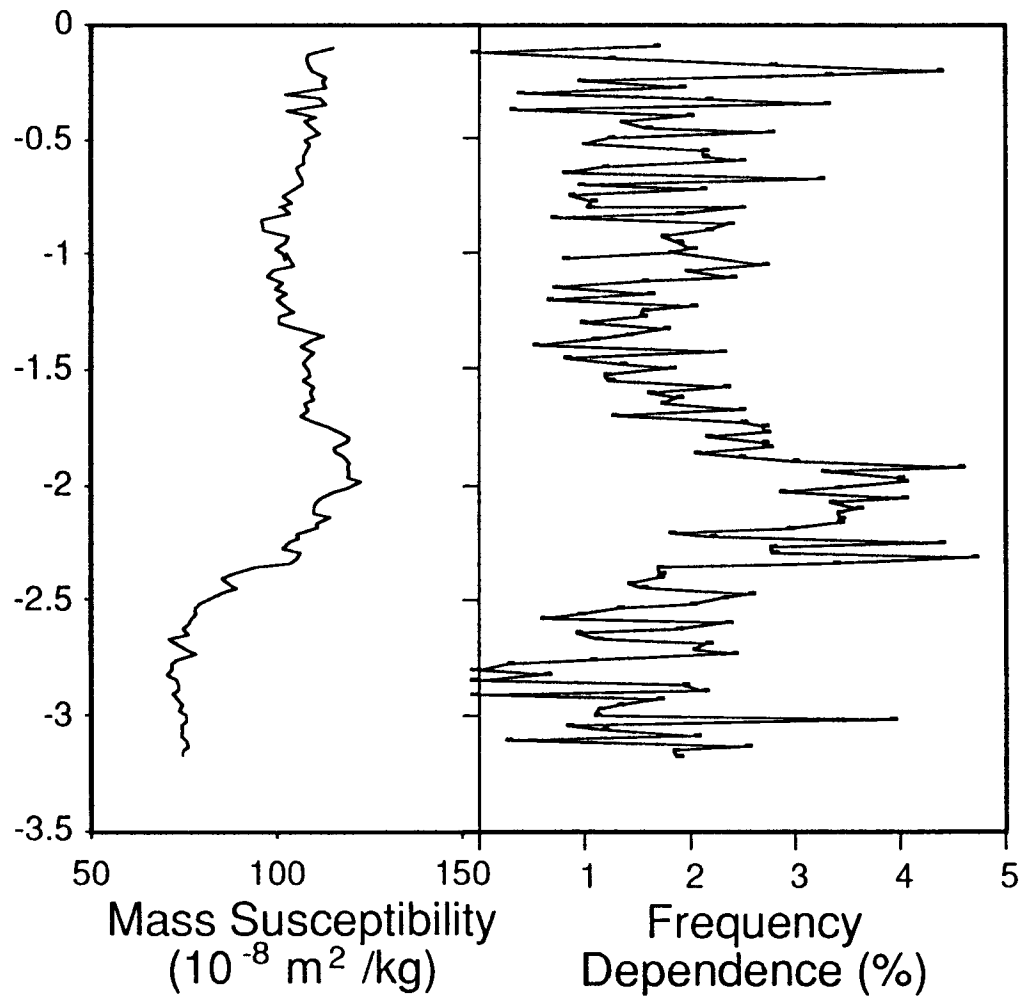
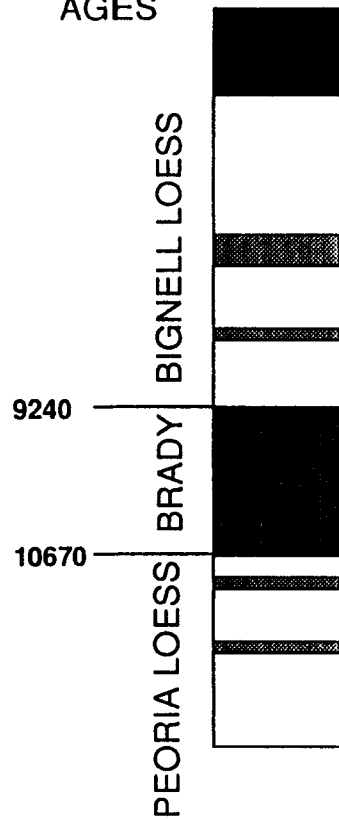


FIGURE 13-7. Mass susceptibility and frequency dependency for the uppermost Peoria, Brady soil, and Bignell loess, Bignell Hill.

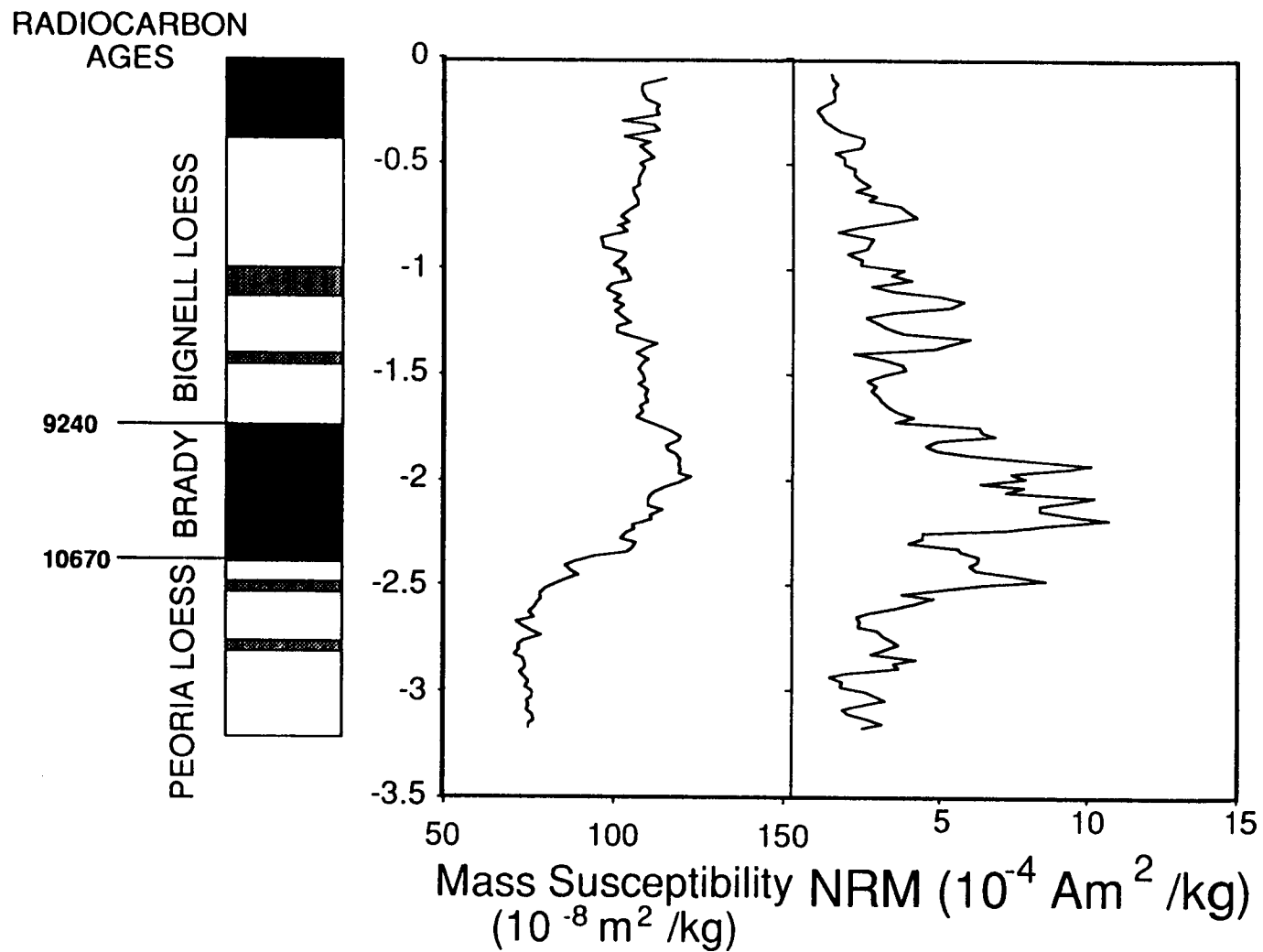


FIGURE 13-8. Mass susceptibility and natural remanence magnetization (NRM) for the uppermost Peoria loess, Brady soil, and Bignell loess, Bignell Hill.

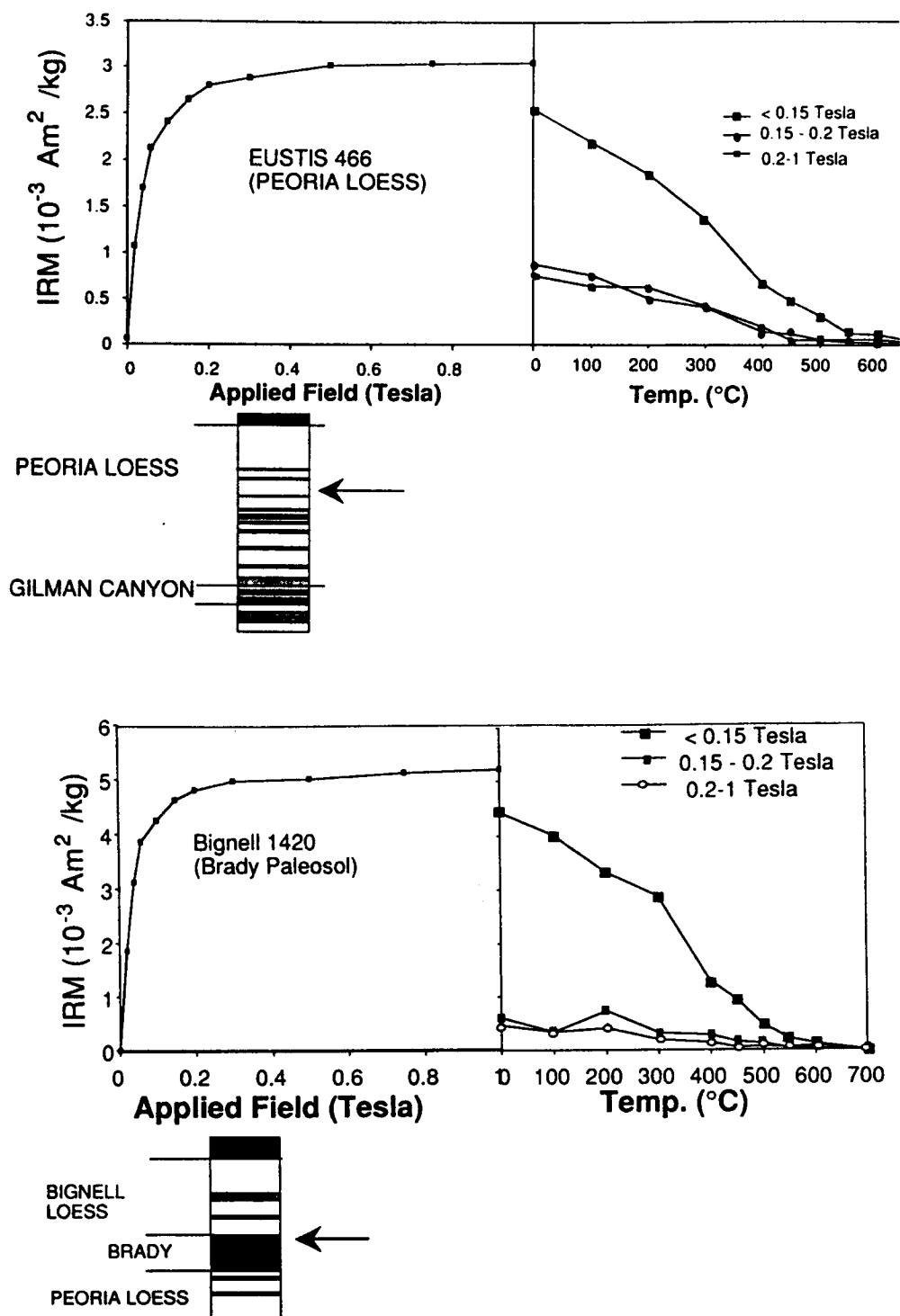


FIGURE 13-9. Saturation magnetism remanence (SIRM) for representative samples of the Peoria loess (Eustis ash pit), Brady soil (Bignell Hill) and Bignell loess (Bignell Hill). Note the difference in IRM values for the plateau among the three curves.

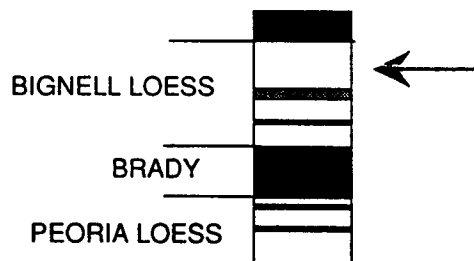
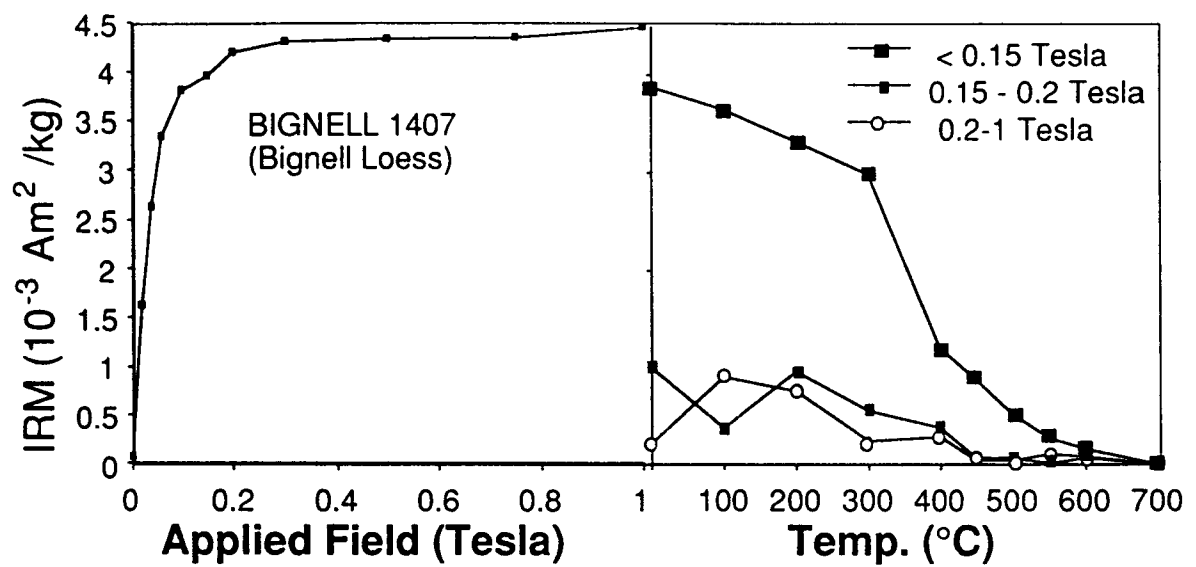


FIGURE 13-9 continued...

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Stop 14 Schoenenburg site, at the upper (west) end of Harlan County Lake, 3 miles south of Orleans, Nebraska: Late-Holocene soils in alluvium

Location: SE1/4 SE1/4 Sec. 34, T2N R19W, Harlan County, Nebraska
(USGS Alma SW, NE 7.5 min. quadrangle)

Contributor: Charles Martin

LATE-HOLOCENE ALLUVIAL SOIL DEVELOPMENT AND CLIMATE CHANGE ON THE REPUBLICAN RIVER

The Schoenenburg site is located on the Republican River immediately above the conservation pool level of Harlan County Lake. It consists of a 7m-high cutbank exposure on the south side of the channel (right bank) featuring three paleosols which can be traced the length of the exposure, about 185m. The focus of the study here was to ascertain the timing of pedogenic initiation and termination, as it relates to the climatically-induced stabilization and destabilization of the Republican River floodplain (Martin, 1990, 1992).

Site stratigraphy

The buried soils are developed in and separated by deposits of silt, silty clay, and medium-to-fine sand. For a distance of 270m back (south) from the channel's edge, the surface of the fill is nearly level, but rises sharply where valley fill intersects upland loess. Along the base of the exposure, there is a layer of cross-bedded, well-sorted medium fluvial sand (Fig. 1). These coarse sands are capped by 0.20 m-thick bed of fine sand that, in turn, is overlain by coarse sand. At a depth of 6m, there is an abrupt change to silty fill, and at 5.8m the deposit has a silty clay texture. These changes are reflected by the high 84th percentile phi values at 6.0m. The silty clay unit is gray (10YR6/1) and highly calcareous (Fig. 2). A well developed paleosol featuring a black (10YR3/1) organic-rich (1.9%), cumulic Ab horizon and dark brown (10YR4/2) Bwb or Btb horizon is formed in the fill at 4.87-5.57m. The Ab is clearly marked by high organic matter and leaching of carbonates between 4.87 and 5.30m (Fig. 2).

The lowermost paleosol is buried by 2m of overbank deposits. Grayish-brown (2.5Y5/2) clay overlies the paleosol, and is in turn overlain by laminated clayey silt that is interbedded with clay drapes. At a depth of 4.20m, there is an abrupt change from clayey silt to fining upwards sequences of interbedded fine and medium sand, a change that may represent an increase in streamflow energy. It is in these overbank deposits that the middle paleosol is developed, at a depth of 2.94-3.20m. The middle paleosol consists of only a dark brown (10YR3/3), cumulic Ab horizon that, in contrast to the lowermost soil, is not enriched in clays; the organic matter content of the middle soil is also somewhat lower (1.7%). It appears that the episode of surface stability during which the middle paleosol formed was of shorter duration than that of the lower soil. The massive coarse silt burying the middle paleosol displays little of the bedding and few of the laminations that characterize most other deposits in the section, suggesting that it may have accumulated in an environment of lower stream energy than that in which the other units were deposited. The upper paleosol, which consists of a black (10YR2/1) Ab and dark brown (10YR3/3) clayey Bwb horizons, is developed in the coarse silt at a depth of 1.63-2.50m. The uppermost paleosol has the highest organic matter content (2.3%) of the three buried soils, a condition likely due, in part, to the relative youth

of the soil. The Ab horizon is leached of carbonates, while the Bwb is relatively rich in carbonates and clays that were probably translocated from above. Profile development of the uppermost paleosol, as measured by degree of horizonation and clay content, is greater than that of the middle soils, and roughly equal to that of the lowermost soil, which in turn relates to duration of surface stability/low overbank sedimentation rates. This third episode of floodplain stability and pedogenesis ended with truncation and deformation of the Ab to produce a wavy upper soil boundary. This was followed by deposition of fining upwards sequences of medium and fine sand. Thirteen discrete sandy laminations, some only 0.02m thick, were noted in the upper 1.60m of the exposure. These laminations likely resulted from individual overbank flows. Following deposition of these fining upwards sequences, floodplain stability returned and the surface soil formed. Subsequent incision of 7m has exposed the late-Holocene deposits.

To trace the lateral extent of fills and paleosols to the south of the exposure, two cores were extracted from terrace fill (Fig. 3: cores 8 & 14) and one from the valley side slope (core 16). Core 8, collected 28.5m from the bank edge, contains three paleosols at depth similar to those in the exposure. Core 14, which was extracted about 92m from the channel, contains two paleosols, which appear to be the upper two exposed in the cutbank. The character of the sediments in cores 8 and 14 differs in that the fining upwards sequences so prevalent in the exposure are not present (Figs. 4,5), and the coefficient of variation increases with distance from the cutbank. These changes relate to the mechanics of overbank flow, i.e., coarser sediments settle out of suspension closer to the active channel, leaving only homogenous sediment suspended in the flow that reached the more distal portions of the floodplain. Another difference between the cutbank and the two cores is that the depth to the paleosols and the thickness of the fills decreases with distance from the exposure (Fig. 3). This follows in that the highest rates of deposition, and therefore thickest fills, occur near the active channel (Allen, 1970). Core 16, extracted from fill at the edge of the uplands about 160m from the bank, contains no visible paleosols and features poorly sorted fill (Fig. 6) comprising interbedded sandy clay and sandy silt, with pebbles scattered throughout. The absence of a paleosol(s) demonstrates that the valley-side fill is sufficiently distant from the zone of active channel movement to avoid overbank sedimentation. Also, the poor sorting suggests that it may not be *in situ* Peoria loess, and is likely colluvium. It is truncated by the valley fill, indicating that it accumulated prior to deposition of the basal fill of the exposure. Further, the colluvium overlies an oxidized and weathered yellowish-brown (2.5Y6/4) clay layer which may represent an eroded Bt horizon dating to the late Pleistocene or earlier.

Paleosol chronology

Bulk samples (6kg) were collected from each of the three buried soils for ^{14}C dating. Initially, ages were obtained on total humates through the University of Texas Radiocarbon Laboratory, but were subsequently resampled for fractionated dating in a laboratory comparison context. The samples were each split and sent to the radiocarbon laboratories at the University of Pittsburgh (PITT) and Southern Methodist University (SMU). The resulting ^{14}C ages exhibit a great deal of intra-fraction and inter-laboratory variation (Table 1; Martin and Johnson, 1993), thereby providing both insight into the source (fraction) of the assay and the differences among laboratories. These are particularly important questions or considerations for those dating soils humates, such as here in the Great Plains, an environment producing little charcoal and peat for possible preservation.

The data indicate appreciable variation in the ages from the different fractions, but the residue (base-insoluble) fraction dates oldest for each laboratory, and, except in two instances, the humic acid (base-soluble) fraction dates youngest for each laboratory. When one is afforded the luxury of dating more than one fraction, the one with the oldest age is generally considered the best estimate of the true age of the soil (e.g., Scharpenseel and Schiffman, 1997; Matthews, 1980; Head et al., 1989; Hammond et al., 1991), but the literature is not in complete accord. The ages clearly provide a chronological stratigraphic order for

the section's buried soils. The lowermost soil appears to have begun developing (upon reduction of overbank deposition) about 3700 yr B.P., and was buried shortly after about 3000 yr B.P. A terminal age of about 2600 yr B.P. seems reasonable for the middle soil, and about 1700 yr B.P. for the upper buried soil.

The climatic factor

The alluvial record from the Schoenenburg site on the Republican River suggests that episodes of entrenchment occurred sometime between 4500 and 3700 yr B.P. and after 1700-1100 yr B.P. This chronology is generally synchronous with that observed in other stream systems of the central Great Plains (e.g., May and Holen, 1985; May, 1989, 1992; Johnson and Martin, 1987; Johnson and Logan, 1990; Mandel, 1992), and is consistent with the notion that climate is the major causal variable in a tectonically stable region (e.g., Brakenridge, 1980; Knox, 1983). Based upon limited regional paleoclimatic information, the early incision appears to correlate with a shift to more mesic climatic conditions, whereas that of the more recent incision seems to have been contemporaneous with increased aridity.

TABLE 14-1. Radiocarbon ages ($\delta^{13}\text{C}$ -corrected) from the Schoenenburg site.

Soil	Lab No.	Humic acid	Fraction dated Residue	Total
Upper paleosol				
middle 6cm	Tx-5911			2020 $\pm$ 60
upper 5cm	PITT-827A	1660 $\pm$ 40		
	-827		1730 $\pm$ 40	
	-827B			1225 $\pm$ 55
	SMU-2579	1100 $\pm$ 50		
	-2711		1320 $\pm$ 40	
	-2721			1100 $\pm$ 55
Middle paleosol				
upper 5cm	Tx-5978			2780 $\pm$ 80
	PITT-828A	2105 $\pm$ 40		
	-828		2670 $\pm$ 40	
	-828B			2155 $\pm$ 90
	SMU-2582	2000 $\pm$ 60		
	-2714		2250 $\pm$ 60	
	-2722			2130 $\pm$ 30
Lower paleosol				
upper 5cm	Tx-5912			3050 $\pm$ 60
	PITT-829A	3135 $\pm$ 50		
	-829		3280 $\pm$ 50	
	-829B			2970 $\pm$ 80
	SMU-2900	3040 $\pm$ 40		
	-2715		3925 $\pm$ 105	
	-2723			3080 $\pm$ 55
lower 5cm	Tx-5977			3720 $\pm$ 90

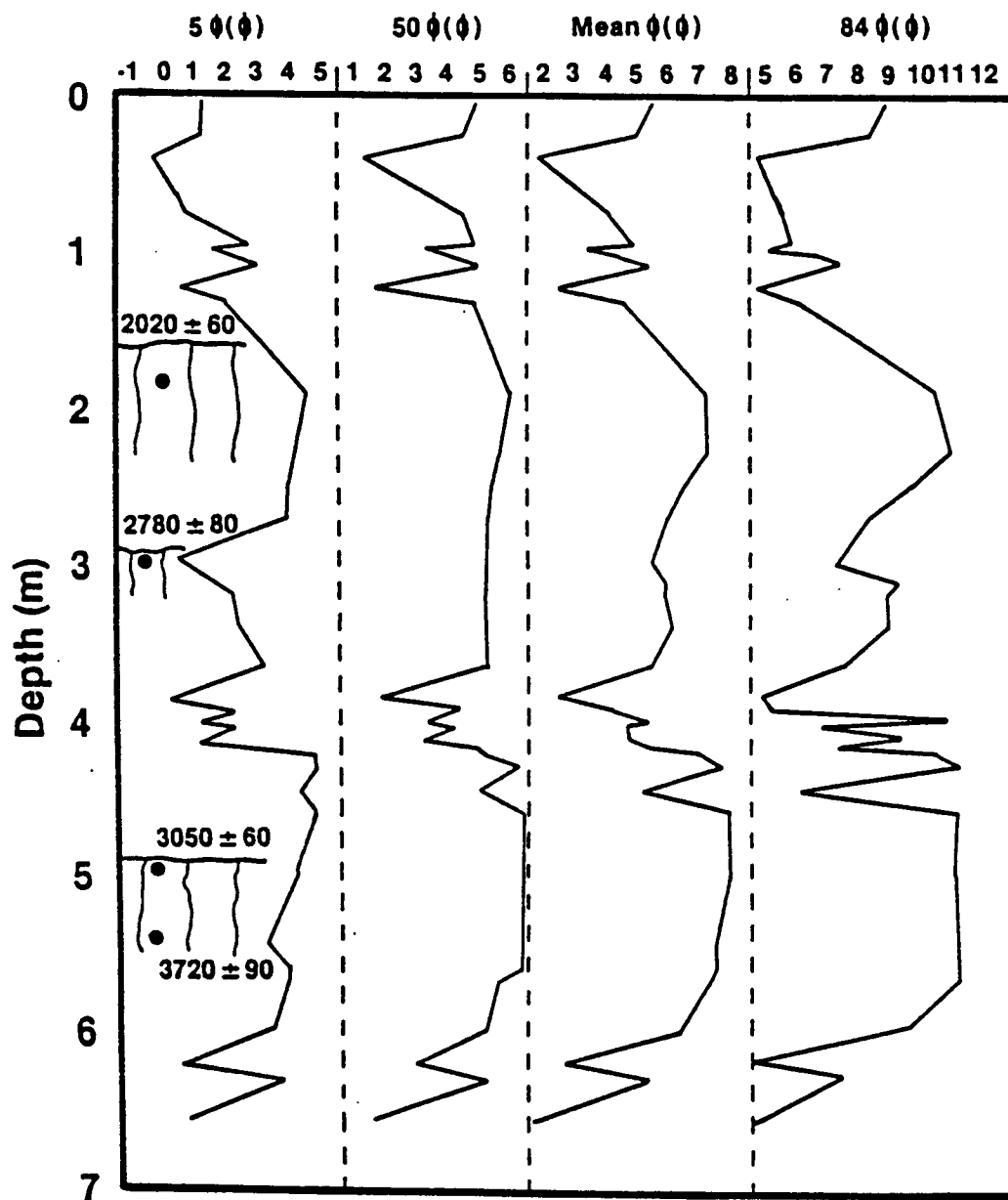


FIGURE 14-1. Particle size trends and soil stratigraphy for the Schoenenburg cutbank exposure. Depth is measured from banktop (603m amsl). Radiocarbon ages are those from the University of Texas.

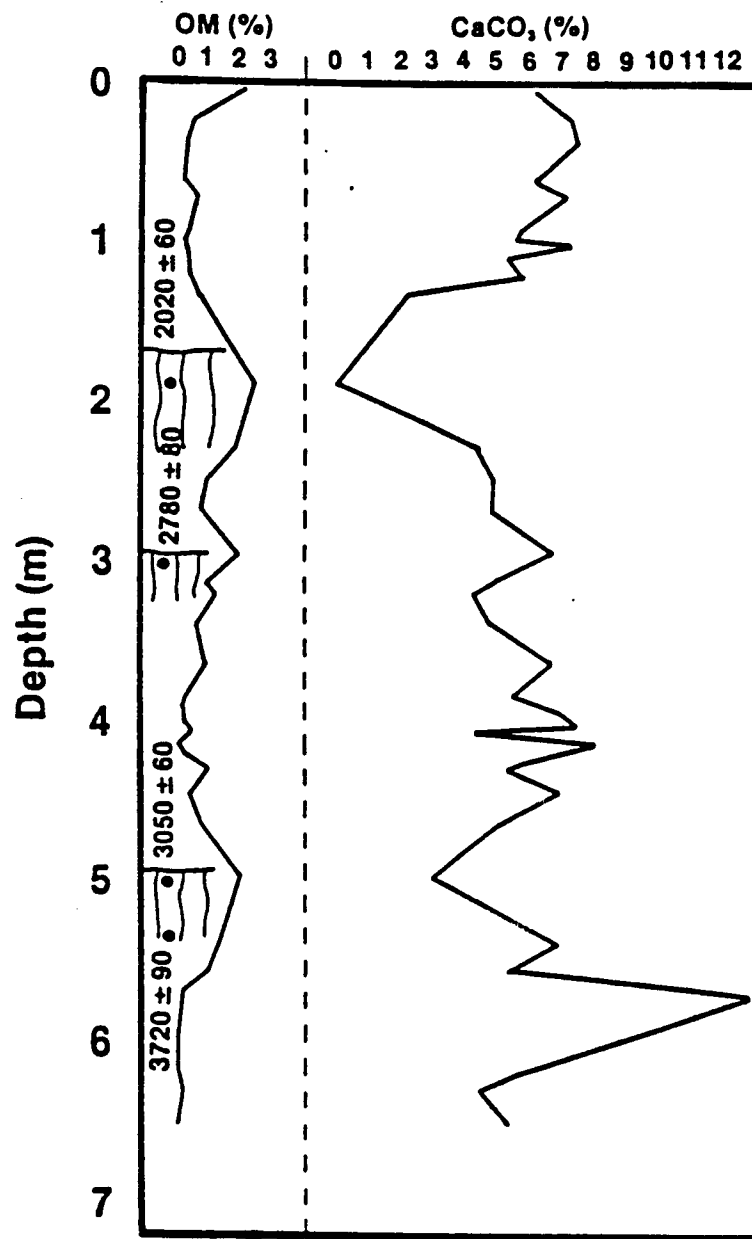


FIGURE 14-2. Organic matter and calcium carbonate trends for the Schoenenburg cutbank. Note high organic matter and low carbonate contents in the three buried soils.

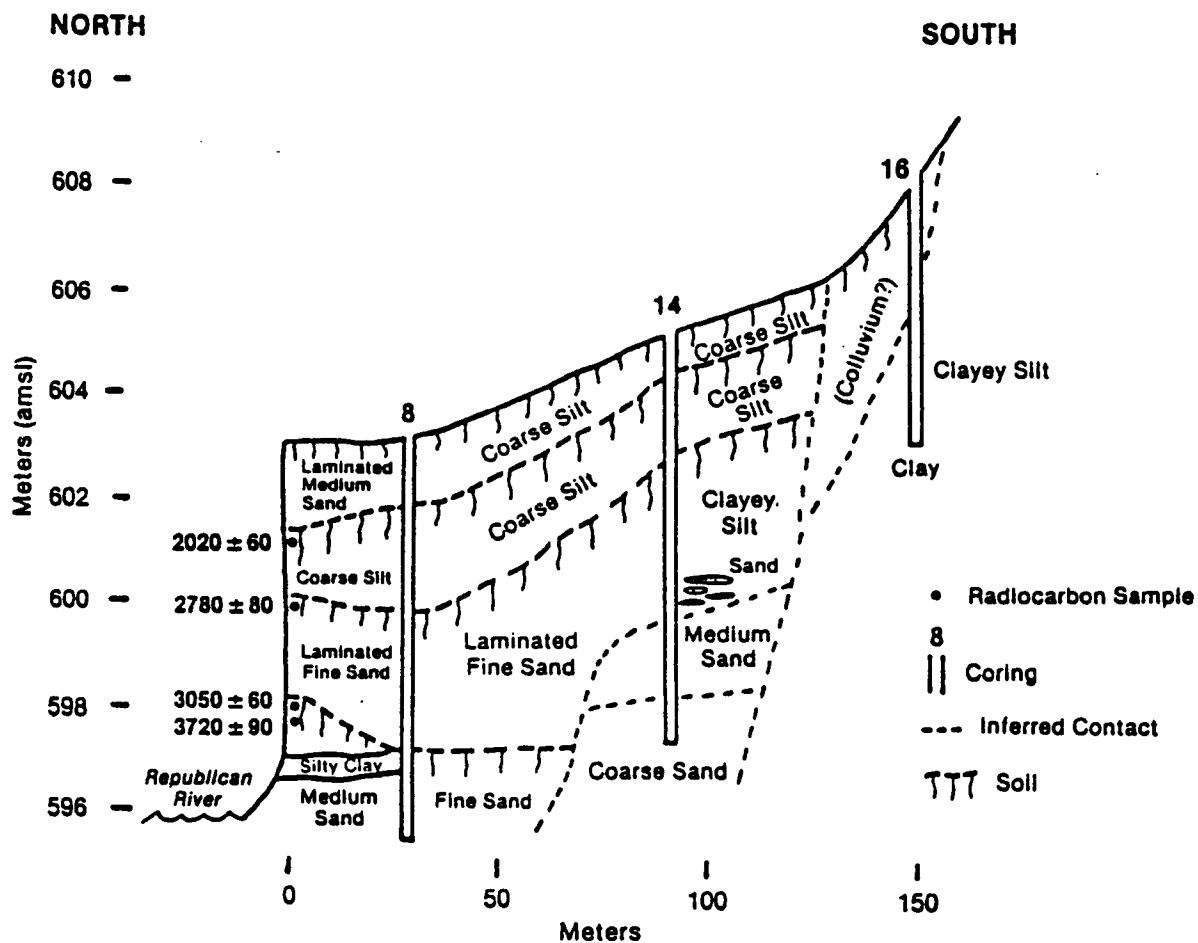


FIGURE 14-3. Cross-section of the Schoenenburg site and the location of core holes in the valley.

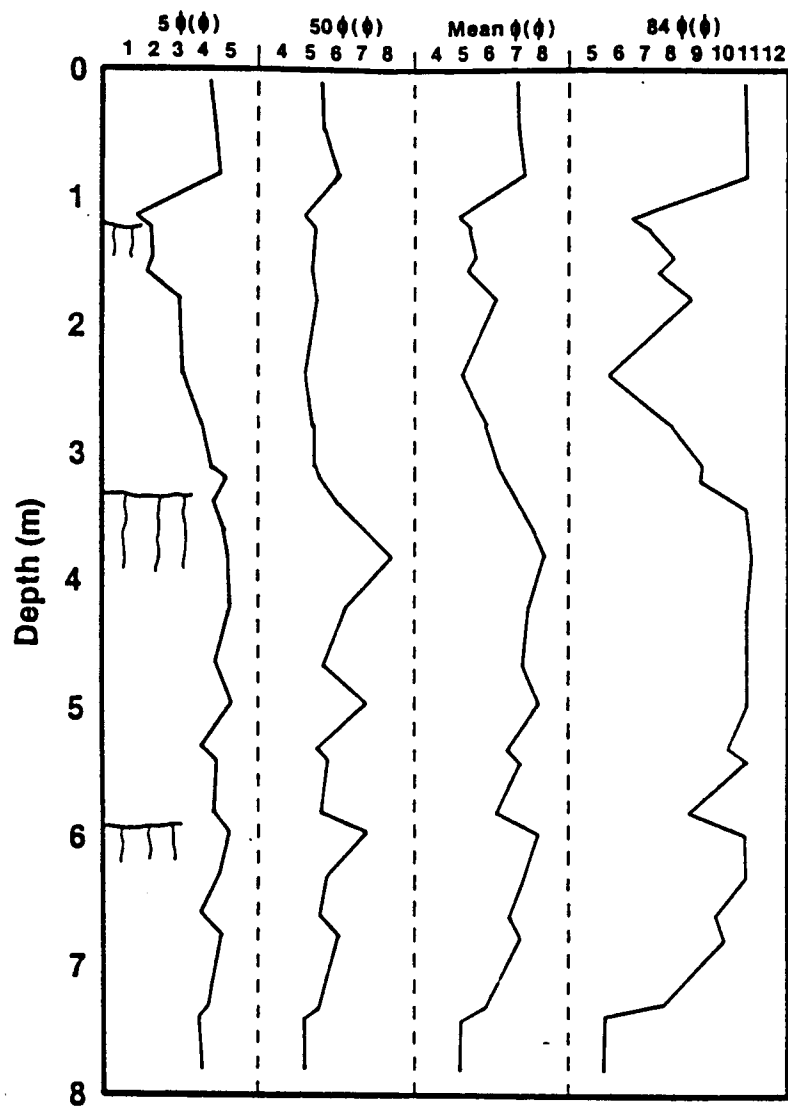


FIGURE 14-4. Particle size trends for core 8, Schoenenburg site.

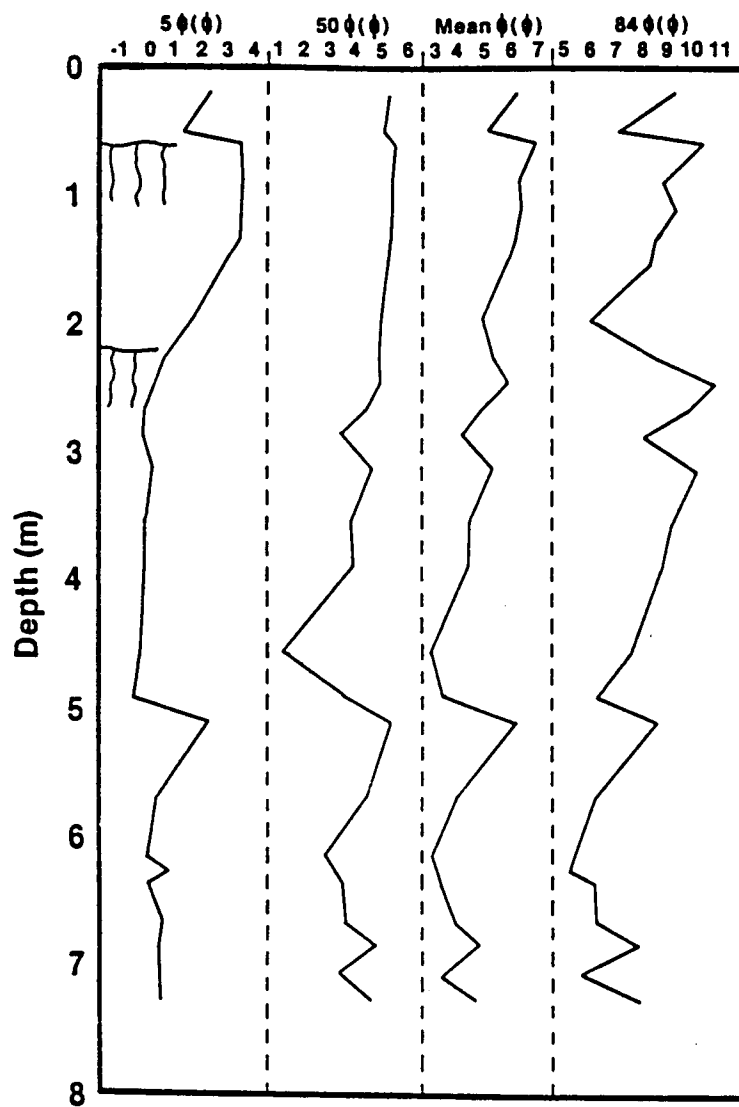


FIGURE 14-5. Particle size trends for core 14, Schoenenburg site.

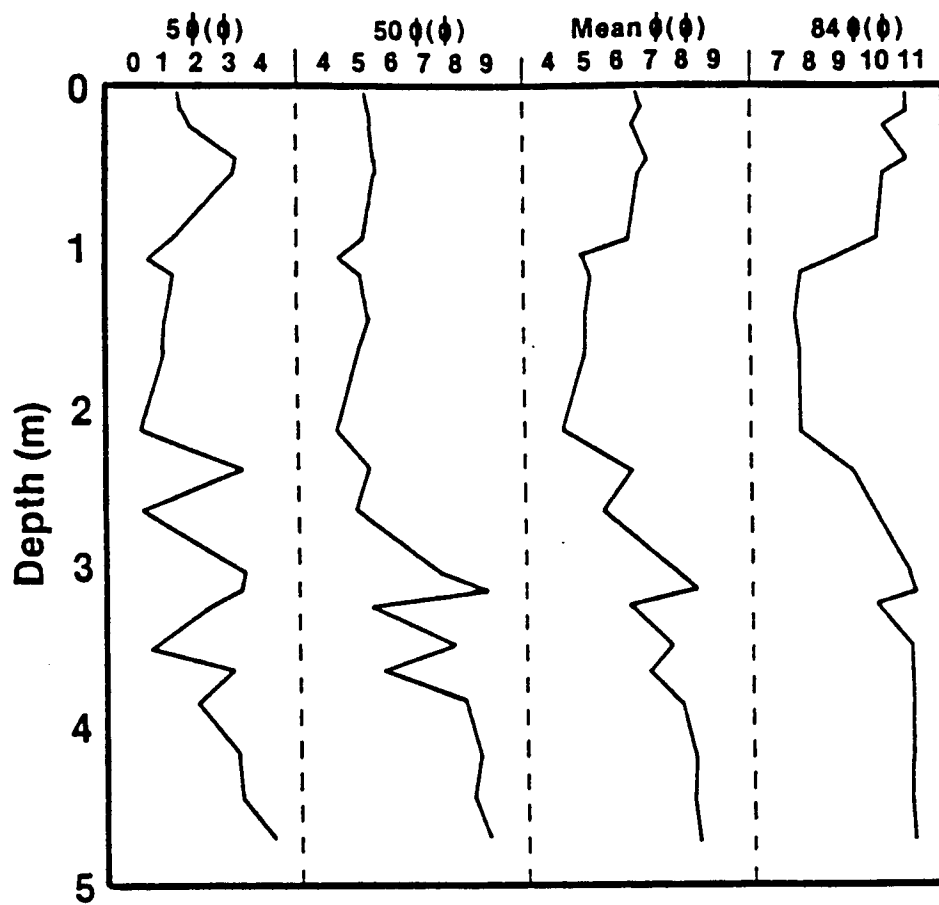


FIGURE 14-6. Particle size trends for core 16, Schoenenburg site.

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Stop 15 Kenneth Lee Site, 7 miles south then 1.3 miles east of Downs, Kansas on the south side of the South Solomon River: Alluvial soils

Location: SE 1/4 of SE 1/4 of NW 1/4, Sec. 26, T7S, R11W, Osborne County, Kansas
(USGS Downs South, KS 7.5 minute quadrangle)

Contributor: David May

SOILS IN MIDDLE AND LATE HOLOCENE ALLUVIUM

The Kenneth Lee site is a cutbank exposure along the south side of the South Solomon River just upstream from Lake Waconda. The site is about 7 miles south of Downs, Kansas on state highway 181, 1.0 miles east on gravel, and about another 0.3 miles east through farm fields. The cutbank exposure is very near the southern edge of the South Solomon River valley close to where the Atchison, Topeka, and Santa Fe Railroad turns south. The legal description of Lee cutbank site is the southeast 1/4 of the southeast 1/4 of the northwest 1/4 of Section 26 in Township 7 south, Range 11 west. The cutbank locality is on the Downs South 7.5-minute topographic map.

The purpose of this fieldtrip stop is to examine both the evolution of the modern soil downward through several Late Holocene alluvial strata, and to examine the properties and degree of development of buried soils in Middle and Late Holocene alluvium. Although many of the issues about Holocene soil development in alluvium will have been addressed in the Pawnee River Valley at two stops, the stop at the Lee cutbank will hopefully provide an opportunity to examine what evidence exists for the formation of argillic horizons in fine-grained Holocene alluvium. This site is located at the edge of a former floodplain, where backswamp (flood basin) deposits should be more common in the valley fill than along the axis of the valley. These deposits should contain more clay than much of the alluvial fill, and thus the alluvial strata, even when unmodified by pedogenesis, exhibit angular blocky structure. Thus, the presence of argillans, the character of lower boundaries of alluvial strata, and the degree of bioturbation tend to be important morphological indices of pedogenesis in the Holocene alluvium at this locality. Below I have first provided a regional context for the alluvial fill and terrace at the Lee site. Next, I present the data that has so far been obtained from this site. Finally, I interpret the morphological, physical, chemical, and chronological data from the perspective of flood-driven valley aggradation and possible pedogenesis in the Holocene alluvium.

Regional Context of Lee Site

The prominent fluvial terrace in the Solomon, North Fork Solomon, and South Fork Solomon River valleys is the Kirwin Terrace (Leonard, 1952). The town of Downs is built on this terrace. Locally the terrace stands 12 meters (40 ft) above low stage of the South Solomon River (Figure 1). Based on its height, spatial extent in the valley, and most importantly, its age, the Kirwin Terrace is correlated with terraces in other drainage basins in the central Great Plains. The Kirwin Terrace is correlative with the Newman Terrace in the Kansas River Basin (Davis and Carlson, 1952; Johnson and Logan, 1990), the Linwood Terrace in the lower Platte River Valley (Lueninghoener, 1947), the Elba Terrace in the Loup River Basin of central Nebraska (Brice, 1964; May, 1989, 1992), Martin's (1992) High Terrace in the middle reaches of the Republican River in southern Nebraska, and Terrace 2 (Schultz and others, 1948) and the Stockville

Terrace (Brice, 1966) in the Medicine Creek Basin in southwestern Nebraska. Many Holocene alluvial stratigraphic sections exposed below these surfaces have been described in these basins. The Lee site is an important site because it is one of the most studied single sections of late-Pleistocene to late Holocene alluvium in the central Plains.

Data

A total of 23 alluvial stratigraphic units were exposed in the alluvial fill below the Kirwin Terrace and above the level of Lake Waconda in early August 1989. These units are numbered from the base of the exposed alluvium to the surface. Physical and chemical data are presented in Table 1 for stratigraphic units 7 to 23 (the upper 6.4 meters), and radiocarbon ages determined thus far are presented in Table 2. Figure 2 is a graphic representation of the upper portion of the Lee cutbank section. Figure 3 is a plot of the physical and chemical data.

Interpretation of the Physical and Chemical Properties of Alluvial Strata (and Soil Horizons)

Introduction. Probably the most significant of the particle-size data are the percentages of sand and clay in each sample. The percentages of sand and clay are in a general way inversely related. High amounts of sand and low amounts of clay probably represent episodes of flooding and rapid floodplain aggradation. Conversely, high amounts of clay in a sample probably represent very slow deposition, or deposition far from the active channel of the South Solomon River (Backswamp or flood basin environment).

The chemical data provide several clues about the rates of alluviation on the floodplain, and diagenesis of the sediments once they are deposited. The percentage of organic matter in a sample of alluvium is a function of either the amount of decomposed organic matter incorporated into sediments in place (soil formation), or the amount of organic matter eroded from elsewhere and redeposited. I attribute most of the variability in the organic matter content of samples at the Lee cutbank site below the modern soil to variations in the amount of detrital organic matter.

Carbonate is present in various quantities in the samples when they were deposited, because carbonate rocks are common in the Solomon River Basin. The amounts that remain in the alluvium after rainwater percolates downward is largely a function of the amount, frequency, and depth of infiltrating water. The texture of the alluvium, in addition to the amount and frequency of infiltrating water, is also a determinant of the depth at which carbonate is precipitated. With the exception of the high amount of carbonate in stratigraphic unit 20, the percentages of carbonate in the Lee cutbank samples generally are low in sandy units and high in the finer-textured clay-rich units.

The percentages of soluble salts in the alluvium in the Solomon River Valleys are probably controlled by three environmental variables: (1) the amount of dissolved salts in the Solomon River when the alluvium was deposited; (2) the rate of valley alluviation (rapid alluviation would reduce the exposure of the sediments to leaching, while slow alluviation and longer exposure of the sediments at the surface would facilitate more complete leaching of salts); and (3) the salt content of the groundwater in the Solomon River Valleys. If the amounts of dissolved salts in the groundwater and surface runoff in the Solomon River Basin are assumed to have varied little during the Holocene, then the soluble salt content of the sediment samples from the Lee cutbank site can be used to infer relative rates of valley alluviation.

Interpretation of units 8-13

The properties of the sediments are discussed and interpreted from bottom up as they were deposited. The high content of sand and soluble salts, and low content of carbonate in samples 8 and 9, which were

deposited after 10,500 before 7250 yr B.P., indicate rapid deposition of these stratigraphic units during a large flood in the South Solomon River Valley. The thick stratigraphic unit 10 contains the highest amounts of soluble salts of any of the samples from the cutbank that have been analyzed. This chemical evidence, in conjunction with the presence of horizontal laminations in this sedimentary unit, indicate rapid deposition. However, sand content in the sample of unit 10 was very low. Perhaps very high sediment concentrations caused by accelerated soil erosion on hillslopes in the basin were responsible for the interval of alluviation immediately before 7250 yr B.P. Also, the Lee site is on the edge of the valley, and probably was within the backswamp (flood basin) of the floodplain, and thus did not receive large inputs of sand during flooding.

Radiocarbon ages of total humates indicate that stratigraphic unit 11 was deposited between 7620 and 6720 yr B.P. However, the 7600-year-old date is apparently too young. I have recollected samples from the base of unit 11 and from the top of unit 9 in an attempt to resolve the dating problem. The recollected samples will be fractionated to determine which is contaminated. The high amounts of soluble salts and clay, and low amounts of sand in unit 11 are indicative of slow sedimentation. The overlying unit, 12, has properties very similar to those of unit 11, except that it contains a lower percentage of soluble salts. The lower salt content of unit 12 could be a consequence of post-depositional leaching. Thus, unit 12 probably was also deposited slowly.

Stratigraphic units 12 and 13 were deposited after 6720 yr B.P. and before 4960 yr B.P. This is a long span of time that must be accounted for by either slow deposition and soil formation, or erosion of some alluvium from the floodplain. The high clay content of unit 12 suggests it was deposited very slowly, was deposited on the floodplain far from the then-active channel, or was both deposited slowly and in a distant flood basin. Conversely, unit 13 was deposited more rapidly, nearer the channel, or both rapidly and near the then-active channel. However, a slow depositional rate for unit 12 is not supported by the organic matter content of the unit, which averages only 0.4%, the lowest of any of the sedimentary units. If the floodplain received only small amounts of vertical accretion sediments, then organic matter should have been incorporated in the evolving soil on the floodplain, and at least a thick cumulic A horizon should have been present and would probably have been preserved.

Both the physical and chemical data indicate that unit 13 has been modified by soil-forming processes. Some of the most convincing evidence of soil formation in unit 13 is the morphology of the unit. Root tubules are common in unit 13, but are not present in the older deposits below. Also, the lower boundary of unit 13 is gradual, rather than abrupt or clear, because of extensive bioturbation. The clay content of unit 13 is 12%, which is low relative to most of the other stratigraphic units, and more importantly, is much less than that of the underlying unit 12. While the variability in clay content from one stratigraphic unit to another could be attributed to variability in flooding and proximity to the active channel, the high amount of silt in unit 13 is probably an artifact of the loss of clay from the unit. The percentage of organic matter does not indicate that the unit is a former A horizon, but the percentage of carbonate, the lowest in any unit below the zone of leaching in the modern soil, suggests that unit 13 was exposed at the surface for a long time. Thus, during much of the interval between 6720 and 4960 yr B.P. the rate of sedimentation on the floodplain of the Solomon River was very low and a soil formed; If this interpretation is correct, then unit 13 was probably the A horizon of the soil, and unit 12 was probably the B horizon of the soil. However, the data for units 13 and 12 could be interpreted differently; erosion of some Middle Holocene alluvium may have occurred at this site. Arbogast (1991) has evidence for considerable Middle Holocene erosion from the Wolf Creek drainage south of this site in the Saline River Basin.

Interpretation of units 14-17

The percentages of clay in unit 14 is the highest of any stratigraphic unit exposed at the Lee site. The carbonate content of this unit is the highest (13.5%), presumably because this unit is the least permeable. The very high amount of clay in unit 14 may indicate very slow vertical accretion in a backswamp (flood basin) environment on the floodplain, or illuviation. The high clay content is at least partially the result of translocation of clay from above, because argillans are present on the very-fine, strongly-expressed, angular-blocky peds. Thus, this stratigraphic unit is a Bt horizon. Unit 15, which immediately overlies unit 14, has many of the same properties as unit 14. Thus, it too, is a Bt horizon.

Units 16 and 17 may be the source for the illuvial clay in units 14 and 15. Unit 16, which contains only 11% clay, has a redder hue (7.5 YR) than any other of the 23 units. Unit 16 may have been exposed at the surface longer than underlying and overlying strata, it may have been exposed when effective precipitation was greater, or it may have been much better drained than other units. The low amount of carbonate (3%) in unit 16 also attests to greater precipitation, better drainage, or both greater precipitation and better drainage. However, the sand content of samples increases sharply up the profile at the contact between units 15 and 16. Such a change signals a lithologic discontinuity within the alluvium. The boundary between these units is clear. Thus, a discontinuity may be present, and soil formation (translocation of clay) may have occurred across the boundary between units 15 and 16.

Unit 17 has many properties in common with unit 16. The properties of unit 17 are transitional between those of an A horizon (dark brown color (10YR 3/3) and root tubules) and those of a B horizon (argillans and organs?, and subangular to angular blocky structure). Thus unit 17 may have been modified into an AB, BA, or Bt horizon by pedogenesis.

Interpretation of units 18-23

A discontinuity is clearly evident between unit 17 and the overlying unit 18. The lower boundary of unit 18 is abrupt and very slightly wavy. The lower portion of unit 18 is horizontally laminated. Ripples are evident near the base of unit 18 in a few places along the cutbank. Thus, truncation, burial, or both truncation and burial of a soil occurred with the deposition of unit 18 sometime after 4600 yr B.P.

Unit 19 is morphologically similar to unit 18, and thus was probably deposited in a manner similar to that of unit 18. However, another lithologic discontinuity occurs between units 19 and 20. Unit 20 is a flood deposit that has been modified slightly by pedogenesis. Unit 21 is similarly a flood deposit that has been modified by Late Holocene pedogenesis. Other discontinuities occur between units 21 and 22, and 22 and 23.

Pedogenesis

An obvious issue to raise when identifying buried soils that have developed in alluvium, and that have been buried by alluvium, is, what role has erosion and alluviation (cumulization) played in the formation of soils that formed on a former floodplain. Soils may have formed, then A horizons may have been removed and B horizons partially truncated. Both B and A horizons may have formed as vertical accretion occurred (cumulization), and this may partially account for B horizons that contain relatively high amounts of organic matter. Still another scenario, is that episodic flooding caused episodic vertical accretion of varying magnitude for several thousand years. The resulting alluvial-soil stratigraphic section exposed today is a combination of alluvial units with various origins and with various degrees of soil horizon development, perhaps welded together. A single section reflects various inputs of sediment. Floods on local tributaries supply eroded soils, residuum, and alluvium, while large floods on the main stem (South Solomon River)

would deliver a mixed load of sediment derived from both sides of the valley from a considerable distance upstream. Furthermore, the position of a single locality on a floodplain and within a valley dictates amounts of sediment delivered, velocities of floodwater, and thus properties of deposited sediment.

My interpretation of the Lee Cutbank site may be biased toward fluvial deposition of clay, rather than translocation of clay and pedogenesis. Both processes appear to be responsible for the morphology of the strata at the Lee site. The pedogenesis that I recognize in the upper part of the section is briefly summarized.

The modern soil has developed in units 18 to 23, or in the upper 1.4 meters of alluvium beneath the Kirwin Terrace. Unit 23 is an Ap horizon of the modern soil, and unit 22 is the A horizon. Unit 21 is a transitional 2AB or 2BA horizon. Units 20 and 19 are 2Bt1 and 3Bt2 horizons, respectively, and unit 18 is a transitional 3BC horizon.

Unit 17 appears to be a truncated 4Ab or 4ABb horizon that has been somewhat modified by later (modern?) pedogenesis. Unit 16 is probably a 4Bb or 4Btb horizon. Although it contains less total clay than the overlying unit (4Ab or 4ABb horizon), it is formed in a sandier alluvial unit and contains argillans. Unit 15 is tentatively labeled a 4BCb horizon; argillans are discontinuous, and the subangular blocky structure is weak.

TABLE 15-1. Physical and chemical properties of stratigraphic units 7-23 exposed at the Lee Cutbank site.

Unit	Depth (cm)	% Sand	% Silt	% Clay	% Organic Matter	% Calcium Carbonate	Soluble Salts (mmhos\cm)
23	0-12	23	71	6	2.6	0.0	0.07
22	12-66	21	67	12	2.0	0.0	0.00
21	66-76	38	46	16	0.7	4.5	0.11
20	76-91	36	45	19	2.1	12.5	0.13
19	91-125	7	70	23	0.7	8.0	0.12
18	125-148	6	69	25	0.9	12.5	0.12
17	148-161	12	73	15	0.6	8.5	0.09
16	161-185	19	70	11	0.9	3.0	0.15
15	185-199	3	69	28	0.7	10.0	0.16
14	199-225	1	53	46	1.0	13.5	0.12
13	225-255	5	83	12	0.9	0.5	0.26
12	255-334	1	61	38	0.4	10.0	0.17
11	334-396	1	64	35	0.7	6.5	0.57
10	396-542	2	72	26	0.7	5.5	1.05
9	542-560	6	87	7	0.6	0.5	0.99
8	560-574	14	79	7	1.5	1.5	0.96
7	574-642	6	81	13	0.5	7.5	0.46

TABLE 15-2. Radiocarbon ages (total humates) of alluvial units described at the Lee Cutbank site.

Strat. Unit	Depth (cm)	Lab no.	C-14 age (yr BP)	$\delta^{13}\text{C}$	Corrected C-14 age (yr BP)
1	1185-1195	Tx-6864	11,940 $\pm$ 170	-18.6	12,050 $\pm$ 170
7	627-237	Tx-7832	10,480 $\pm$ 120	-19.0	10,580 $\pm$ 120
9	542-552	Tx-7135	7,160 $\pm$ 120	-18.4	7,250 $\pm$ 120
		Tx-7463	7,090 $\pm$ 70	-18.3	7,190 $\pm$ 70
11	386-396	Tx-7040	7,550 $\pm$ 160	-20.6	7,620 $\pm$ 160
11	334-344	Tx-6863	6,660 $\pm$ 110	-20.3	6,720 $\pm$ 110
13	225-235	Tx-7041	4,820 $\pm$ 70	-16.5	4,960 $\pm$ 70
17	148-158	Tx-6865	4,530 $\pm$ 70	-19.2	4,630 $\pm$ 70

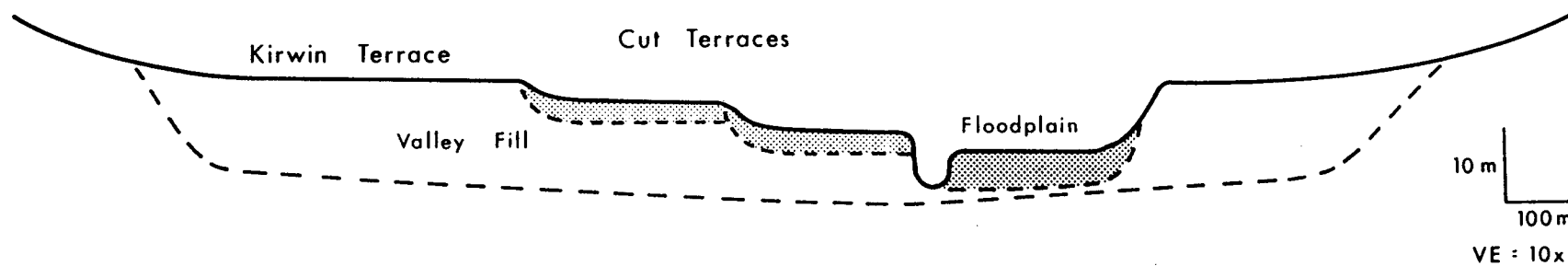


FIGURE 15-1. Generalized cross section of the North Solomon, South Solomon, and Solomon River Valleys in the vicinity of Lake Waconda, Kansas.

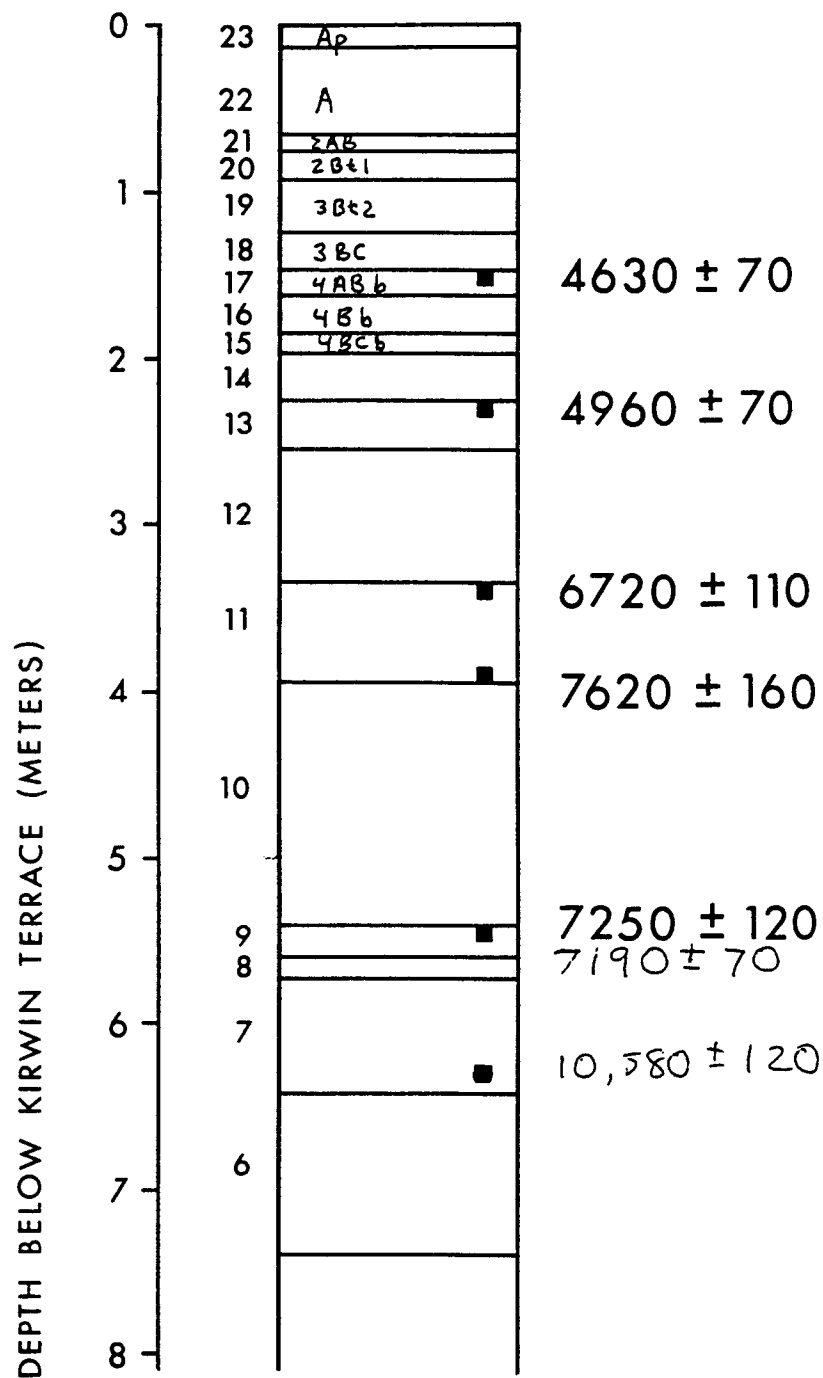


FIGURE 15-2. Stratigraphic section of the Lee Cutbank site showing stratigraphic units, inferred soil horizons, and corrected radiocarbon ages.

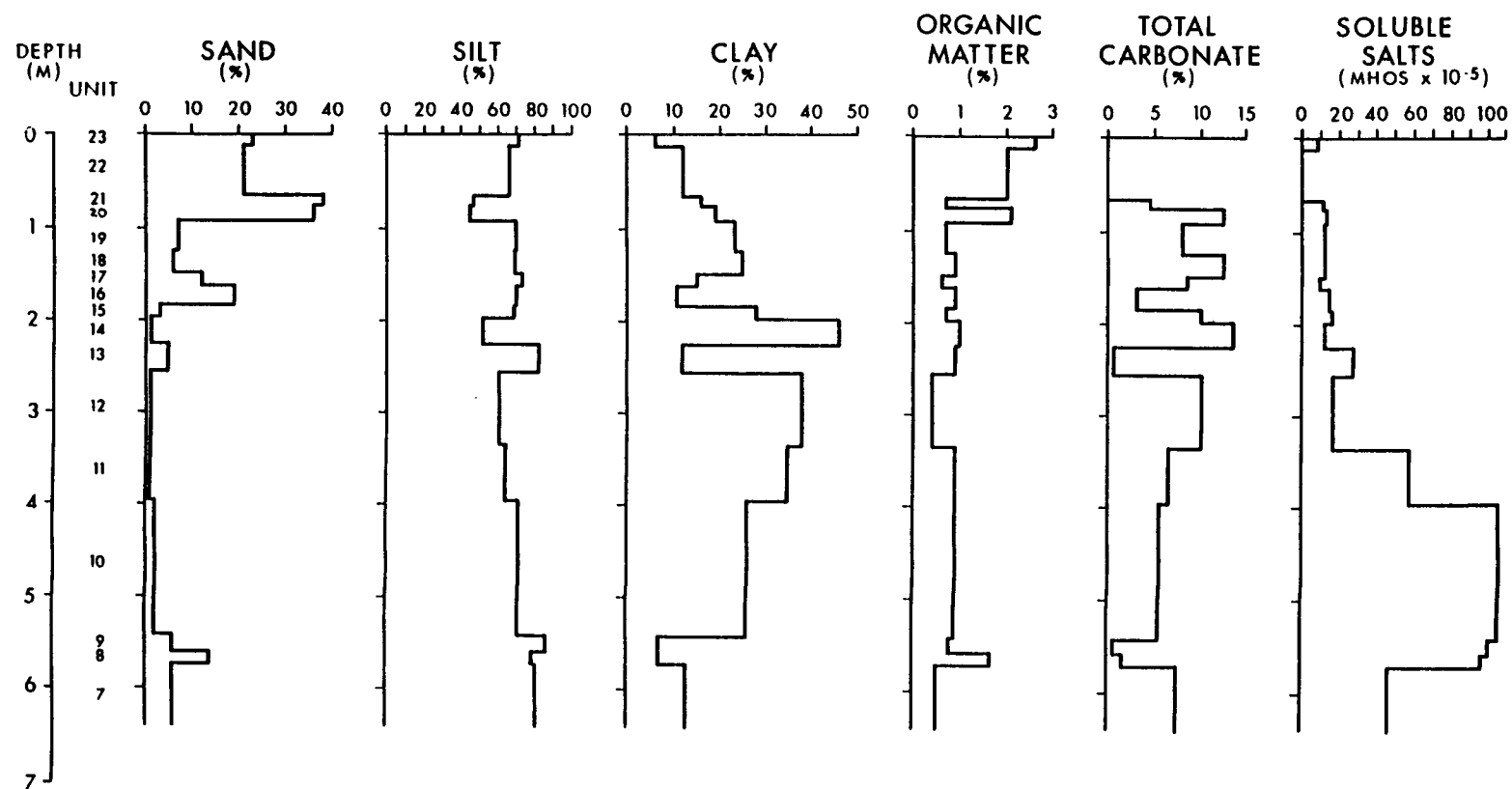


FIGURE 15-3. Particle-size and chemical data for samples from stratigraphic units 7-23 at the Lee Cutbank site.

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Stop 16 Endicott Clay Products Company pit, approximately 1.5 miles southwest of Endicott, Nebraska: Cretaceous paleosols

Location: SE 1/4, Sec. 7, and W 1/2, SW 1/4, Sec. 8, T1N, R3E, Jefferson County, Nebraska
(USGS Fairbury Southwest and Endicott, NE-KS 7.5 minute quadrangles)

Stop 17 Abandoned Quarry, approximately 2 miles WNW of South Bend, Nebraska: Cretaceous paleosols

Location: W 1/2, NW 1/4, SW 1/4, Sec. 15, T12N, R10E, Cass County, Nebraska
(USGS Ashland East, NE 7.5 minute quadrangle)

Contributor: R.M. Joeckel

PLINITHIC PALEOSOLS IN LOWER DAKOTA FORMATION (CRETACEOUS, ALBIAN-CENOMANIAN) AT ENDICOTT CLAY PIT

Age: ~100 Ma

Comments. Massive, clayey to sandy kaolinitic mudstones, with intense reddish mottling and closely-associated hematitic ironstone bodies set in a whitish or grayish matrix, characterize the fine-grained alluvial sediments of the lower and middle Dakota Formation (Albian-Cenomanian) in southeastern Nebraska (Fig. 1). These mottled mudstones are interpreted as paleosols ranging in depth from about 2.5 m to more than 10 m. Paleosols can be traced laterally for up to a few hundred meters in clay pits, and along the Little Blue River between Steele City and Fairbury, Nebraska, a single geosol (exposed in the Endicott pit) appears to be mappable over at least 5 km. In vertical section, the paleosols are often truncated by sandy or muddy fluvial channel fills, very thin sheets of sand, or sandy meanderbelt deposits with prominent lateral accretion surfaces. In many cases, however, mottling is developed within thinner channel fill sediments and sand sheets as well, often extending uninterrupted downward from overlying massive alluvial mudstones. The frequent continuation of mottling across intraformational unconformities demonstrates that pedogenic processes were pervasive and ongoing. Common intraformational erosion surfaces truncating mottled horizons and the frequent occurrence of re-worked paleosol fragments (identical in color and structure to in-situ paleosol material) as intraformational clasts demonstrate without doubt that Dakota paleosols formed contemporaneously with deposition.

The most striking aspect of the Dakota alluvial paleosols is the persistent appearance of large, vertically-elongate, essentially columnar, zones of concentrated reddish mottling, infrequent hematitic nodules, and weak cementation by iron oxides. In horizontal section (parallel to paleo-land surface) these vertical mottled zones (hereafter abbreviated VMZs) are joined to form a coarsely reticulate pattern, thereby often surrounding essentially cylindrical or tongue-like bodies of lighter-colored, very sparsely mottled mudstone. VMZs vary in diameter from 2-30 cm, but are commonly 12-20 cm in diameter, and have horizontal spacings ranging from 2-18 cm. Individual VMZs extend downward through the two or more meters depth of any given paleosol profile; simple upward- or downward-branching bifurcations of VMZs are common.

Mottling within VMZs comprises masses of many coalescing, distinct, fine to medium, vermicular mottles associated both with planar and channel-like voids. Mottling associated with planar voids often consists of weakly-cemented, vein-like accumulations of iron oxide (XRD shows hematite). The most common mottle colors are weak red (10R 4/2, 4/4), red (10R 4/6, 4/8), dark red (10R 3/6), or dusky red (10R 3/3, 3/4) on a background of reddish-gray (10R 6/1), dark reddish gray (10R 4/1), pale red (10R 6/2), or weak red (10R 5/2). Few to common, fine, yellowish (e.g., 7.5YR 6/8 reddish yellow) vermicular mottles are often present along with the reddish mottles. Masses of reddish mottles vertically or laterally within about several cm-wide patches of continuous weak red, red, or dusky red coloration. Frequently, these zones of continuous reddish coloration and preferential hematite enrichment are more resistant than adjacent material, and may actually be developed as friable to moderately indurated masses of weak red (10R 4/2), red (10R 4/8, 5/8), dusky red (10R 3/3, 3/4), and dark red (10R 3/6) ironstone 5-20 cm in diameter. Reworked or weathered-out ironstone fragments often form a continuous pavement several cm thick on post-Cretaceous erosion surfaces cut into Dakota paleosol-bearing mudstones. XRD shows that Dakota paleosol ironstones consist of quartz (mostly as sand and silt), hematite, and whatever clay minerals are common in adjacent paleosol material (dominantly kaolinite, with minor illite).

Original soil structure in VMZs is only poorly expressed, if at all. There appears to be weakly preserved pedality (evidenced by a blocky or hackly fracture) within at least parts of VMZs in a few instances. Vein-like accumulations of hematite developed in planar voids (mentioned above) surround and subdivide blocky units that probably represent fine to medium subangular blocky peds; these interpretive peds are coated with illuvial material (probably both iron oxides and clay). Where such structures are best developed, the mudstone fractures around the interpretive peds in a hackly or blocky fashion.

VMZs are commonly surrounded by a light-colored (e.g., 5Y8/1) halo about 5 mm wide, which gives way to the surrounding, lower-value (e.g., 5Y 7/1) matrix material. The margins of VMZs, their light-colored halos, and bodies of the surrounding matrix are always sharp--zones of gradation between these materials are usually less than 2 mm wide.

The matrix of paleosol material surrounding VMZs is conspicuously lighter-colored: white (5YR 8/1, 5Y 8/1), light gray (5YR 7/1, 5Y 7/1), or reddish gray (10R 6/1) with few, fine to coarse, reddish or yellowish vermicular mottles. Large subvertical and subhorizontal slickensides appear sparsely in the mudstone surrounding VMZs. There is little or no indication of preserved pedality in the low-chroma matrix, although in old outcrops it weathers in a coarse subangular blocky to prismatic fashion.

Aside from lithological discontinuities between mudstone and sandstone units within single profiles, horizonation in Dakota alluvial paleosols is not prominent--VMZ horizons are fairly uniform from top to bottom. If found in a modern profile they would be designated Btv, but would probably also be subdivided into Btv1, Btv2, Btv3, etc. based on relatively subtle differences in color, structure, and expression of plinthite bodies. No pedogenic horizons were found that appeared to be organic-rich A horizons above paleo-plinthite horizons, although carbonaceous sandy mudstones (sometimes stratified) are sometimes preserved in close vertical association with the paleosols. In some profiles, 7-120 cm horizons of white (10YR 8/1, N8/0) or light gray (10YR 7/1) massive, unmottled or relatively sparsely mottled mudstone with clear to gradual lower boundaries overlie prominent VMZ horizons. These unmottled to very sparsely mottled mudstones are interpreted as E or EB horizons by virtue of their leached appearance and their superior position in vertical sequence. Yellow (10YR 7/6, 2.5Y 7/6), brownish yellow (10YR 6/8), and very pale brown (10YR 8/3) horizons 5-25 cm thick with abrupt to clear boundaries (not depositional contacts) also appear near the tops of several VMZ profiles. Such horizons often have reddish mottling similar to horizons below them, but may also be unmottled. Such horizons are probably best designated Btv (if mottled) and, perhaps, E or Bw (if unmottled). At some localities, VMZs lose their reddish hues moving downward 2 or more m in a profile and become yellowish brown (10YR 5/6) throughout, without

discrete mottles. In such cases, the transition from intensely mottled VMZs into vertical, yellowish brown features without internal mottling or other structure is interpreted as a Btv to BCg horizon transition.

The clay fraction ($< 2 \mu\text{m}$) of Dakota alluvial paleosols dominated by kaolinite, although subequal illite appears in profiles near the base of the formation. Quartz and hematite are minor constituents in the clay fraction (hematite is segregated in VMZs). Yellowish, fine- or very fine-grained sandstone stringers that sometimes appear within paleosol profiles have goethite cements.

Mineralogically and morphologically, the most appropriate modern analogs to Dakota alluvial paleosols with VMZs appear to be plinthite-bearing Ultisols, particularly Plinthaquults.. Modern Plinthaquults often develop in geomorphic environments similar to those interpreted for the Dakota paleosols (well-weathered alluvial sediments on low-relief surfaces).

In southeastern Nebraska, Dakota paleosols have been mined for well over a century, primarily for brick clay. The Endicott operation is the largest, but there are active pits in the Lincoln and Omaha areas as well. The Endicott pit exposes a profile over 15 m thick consisting of at least three separate plinthic horizons or groups of horizons (Fig. 2). There are prominent planar to slightly concave-up erosion surfaces atop this profile (Fig. 3). These may relate to changes in fluvial systems prompted by the onset of the Greenhorn marine transgression, which eventually deposited several tens of meters of marine sediments over the Dakota Formation.

VERTIC PALEOSOLS IN UPPER PART OF LAWRENCE FORMATION (UPPER PENNSYLVANIAN, VIRGILIAN, DOUGLAS GROUP) AND IN SNYDERVILLE SHALE MEMBER (UPPER PENNSYLVANIAN, VIRGILIAN, SHAWNEE GROUP, OREAD FORMATION) AT ABANDONED QUARRY

Age: ~290 Ma

Comments. Contrasting paleosols in the upper Lawrence Formation and in the Snyderville Shale Member (Virgilian) are traceable over at least 14,000 km² in Nebraska, Iowa, Missouri, and Kansas. (Figs. 4, 5; see also Table 1) These paleosols appear within classic Midcontinent cyclothems (supposedly cyclic packages of carbonate and clastic sediments) and should be viewed as critical indicators of near basin-wide emergence (peak regression of the Midcontinent seaway) and geomorphic stability, as well as useful stratigraphic markers. Each was drowned and slightly eroded by a subsequent marine transgression. Both paleosols appear to have been Vertisols, as they show have nested, "synclinal-anticlinal" sets of very large, intersecting slickensides. Also, this profile of the upper Lawrence paleosol also has filled cracks extending to depths of 100 cm or more. Significant general contrasts between the paleosols are: high-chroma color, strong preserved soil structure, and small iron oxide nodules (upper Lawrence) versus low-chroma color, weak to very weak preserved soil structure, and pyrite segregations (Snyderville).

Brecciated mudstone fabric (generally common in upper Pennsylvanian and Permian paleosols of the Midcontinent) is characteristic of the thickest and best-developed sola in the upper Lawrence. It appears as a mosaic of small (commonly 1-10 mm, but as large as 25 mm), loosely to closely fitted, subangular blocky (rarely angular blocky or rounded subangular blocky) and small ($< 1-1.5 \text{ mm}$) granular aggregates partially to completely surrounded by "veins" (infillings) of darker mudstone. Larger ($> 3 \text{ mm}$) aggregates consist of a few to several, smaller aggregates partially separated by very fine ($< 1 \text{ mm}$) infillings. Aggregates are commonly pinkish red (5YR 6/2 or 8/2), very pale brown (10YR 7/3 or 7/3-7/4), or light gray (10YR 7/1) in a reddish brown (5YR 3/3 or 4/3) matrix or reddish brown (5YR 4/3)

in a dark reddish brown (5YR 3/4) matrix. The most likely explanation for these aggregates is that they are preserved soil structure (peds) because they appear to have formed *in-situ* and before burial.

Brecciation often appears in zones 1-2 cm wide on either side of very large slickensides if the surrounding material is mostly massive. This phenomenon (and the close association of large infillings with brecciation) implies a very close link between in-situ soil mass movement (driven by wetting and drying/shrink-swell cycles) and the periodic creation of soil structure in the deeper part of the subsoil.

"Veins" or infillings between aggregates are usually partially accommodated to unaccommodated (i.e., their margins or walls are partially fitted to unfitted), although they are sometimes fully accommodated when very narrow. Infillings are impregnated with iron oxide and are usually darker (e.g., 5YR 3/4) than the adjacent aggregates. At locality 3, infillings become gray to light gray (5Y 5/1-7/1) deep in the profile (below 200 cm). In a few cases, the infilling consists of well-sorted medium to coarse silt. More commonly, infillings contain a few to many, ≤ 2 mm, carbonate nodules, iron oxide nodules, and granular to subangular blocky aggregates in a weakly to strongly iron oxide-impregnated matrix of clay and silt. The nodules and aggregates found in veins are identical to features in the surrounding matrix, indicating that they were re-worked via pedoturbation.

Clay minerals provide little insight into weathering processes during late Lawrence and Snyderville times. Clay mineral assemblages in both paleosols are dominated by illite, with subordinate chlorite and minor kaolinite. There is only equivocal evidence for weak pedogenic weathering of illite during the in the upper ~ 100 cm of both soils. This is not surprising, as Lawrence and Snyderville paleosols were originally base-rich soils (pedogenic carbonate is pervasive) without strong chemical weathering, as is generally the case in modern Vertisols.

The upper Lawrence paleosol is thickest and best-developed shelfward (northward, as at this locality) and exhibits topohydrosequence variation at two basinward localities in northeastern Kansas. While the upper Lawrence paleosol is only a few tens of cm below the overlying Toronto Limestone Member in Nebraska and Iowa, from Atchison to Lawrence, Kansas up to 6 m of tidal muds and sands lie between the paleosol and the basal Toronto. Apparently, there was an appreciable time differential in the south to north migration of the Toronto transgression. The Snyderville paleosol shows vertical rather than horizontal variability at the South Bend locality, where two paleosols (a lower, high-chroma paleosol with strong soil structure and an upper, low-chroma paleosol without visible structure). Topohydrosequence variation is not visible at any of the Snyderville localities that have yet been studied. The Snyderville paleosol does not show a north-south stratigraphic-geographic trend equivalent to the upper Lawrence, although there is local evidence for lowstand incision of streams and small lows that underwent little or no subaerial exposure. Contrasting characteristics in the two paleosols are likely to be related to: 1) an inherent change in geomorphic state (poorer drainage in Snyderville times), 2) climate change (wetter conditions during Snyderville times), and 3) differing rates and extents of subsequent transgressions.

TABLE 16/17-1. Comparison of upper Lawrence and Snyderville paleosols.

<u>Upper Lawrence Formation Paleosols</u>	<u>Snyderville Shale Member Paleosols</u>
Similarities	
sets of very large, intersecting slickensides, forming synclinal-anticlinal sets, in subsoil; these are prominent	sets of very large, intersecting slickensides, forming synclinal-anticlinal sets, in subsoil; these are distinctive to prominent
crystallitic b-fabric (frequent in thin sections) overprints other fabrics and features, suggesting later-stage emplacement of pervasive, fine carbonate	crystallitic b-fabric (frequent in thin sections) overprints other fabrics and features, suggesting later-stage emplacement of pervasive, fine carbonate
calcitic carbonate nodules present in upper part of solum	calcitic carbonate nodules present in upper part of solum
large, disc-shaped dolomitic carbonate nodules or discontinuous beds of dolomitic carbonate present in lowest part of paleosol	dolomitic carbonate nodules present in lower part of paleosol; upper part of underlying limestone frequently contains dolomite
some micromorphological features suggest clay illuviation	some micromorphological features suggest clay illuviation
fragments of well-oriented clay, with unit or sweeping extinction, appear in thin section; these are relict fragments of laminated sediment and/or fragments of pedogenic clay coatings	fragments of well-oriented clay, with unit or sweeping extinction, appear in thin section; these are relict fragments of laminated sediment and/or fragments of pedogenic clay coatings
Differences	
high-chroma (reddish to reddish brown: 5YR 3/4, etc.), with an upper, low chroma horizon	low chroma (grayish to greenish gray: 5Y6/1, etc.)
paleosol developed in mudstone over shale with probable large vertical fractures	paleosol developed in mudstone or siltstone over thick limestone with microkarst
mudstone/shale unit atop paleosol grades markedly upward into an overlying marine limestone	mudstone/shale unit atop paleosol has relatively abrupt contact with overlying marine limestone
strong soil structure preserved in several profiles	soil structure very weak or absent

pyrite absent

pyrite present throughout profile

iron oxide nodules present

iron oxide nodules rare or absent

greenish gray or yellowish streaks on slickensides in subsoil

dark gray or very dark gray streaks (containing organic matter?) present in subsoil; dark gray or very dark gray streaks on slickenside surfaces in subsoil; faint, dark-colored "bowls" appear in one outcrop

significant, non-cyclic lateral variability represented in two, long outcrops

lateral variability absent or cyclic ("bowls") at same outcrops; vertical variability represented at one outcrop as two, different, welded profiles

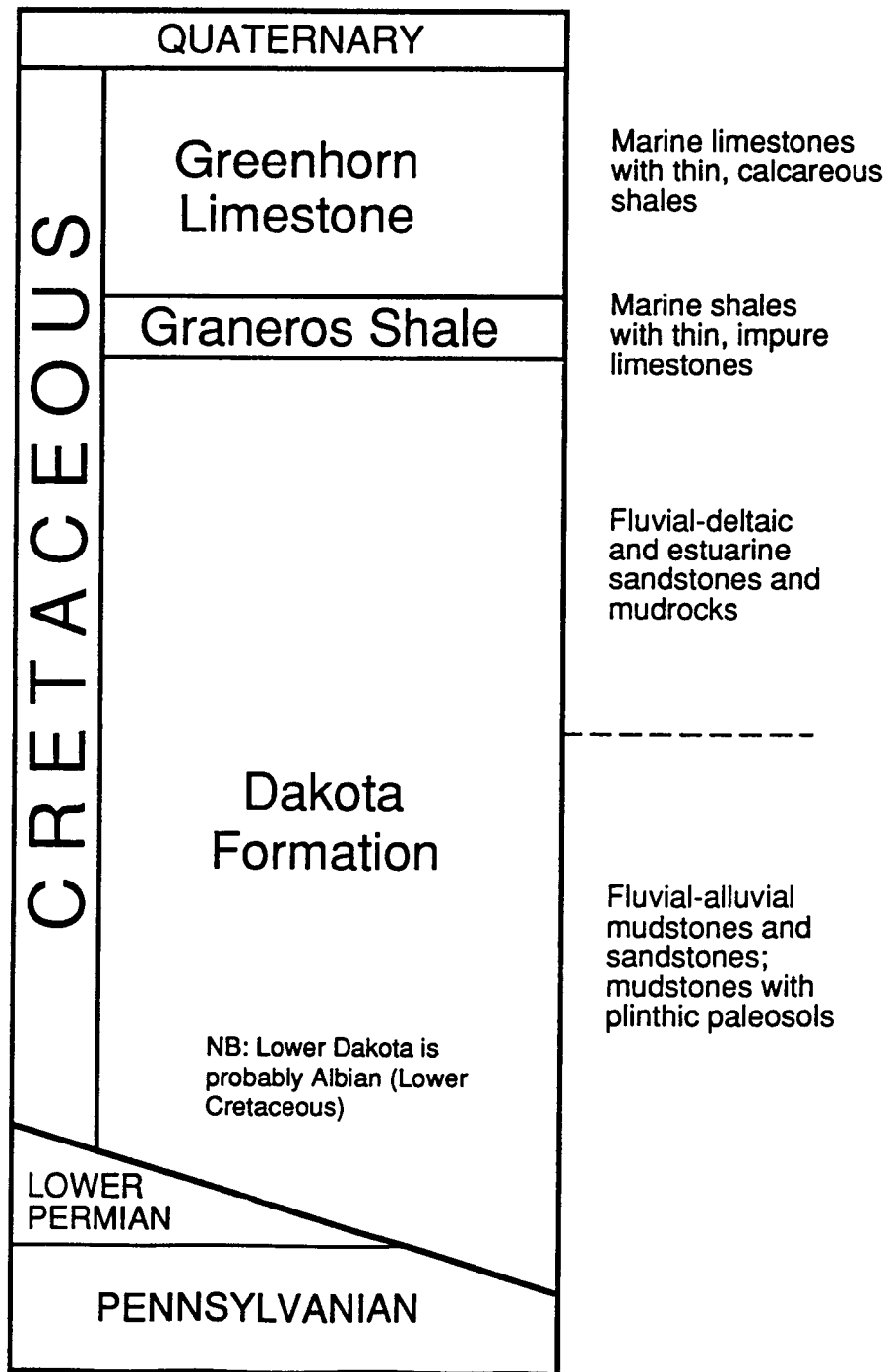


FIGURE 16/17-1. Generalized section for southeastern Nebraska.

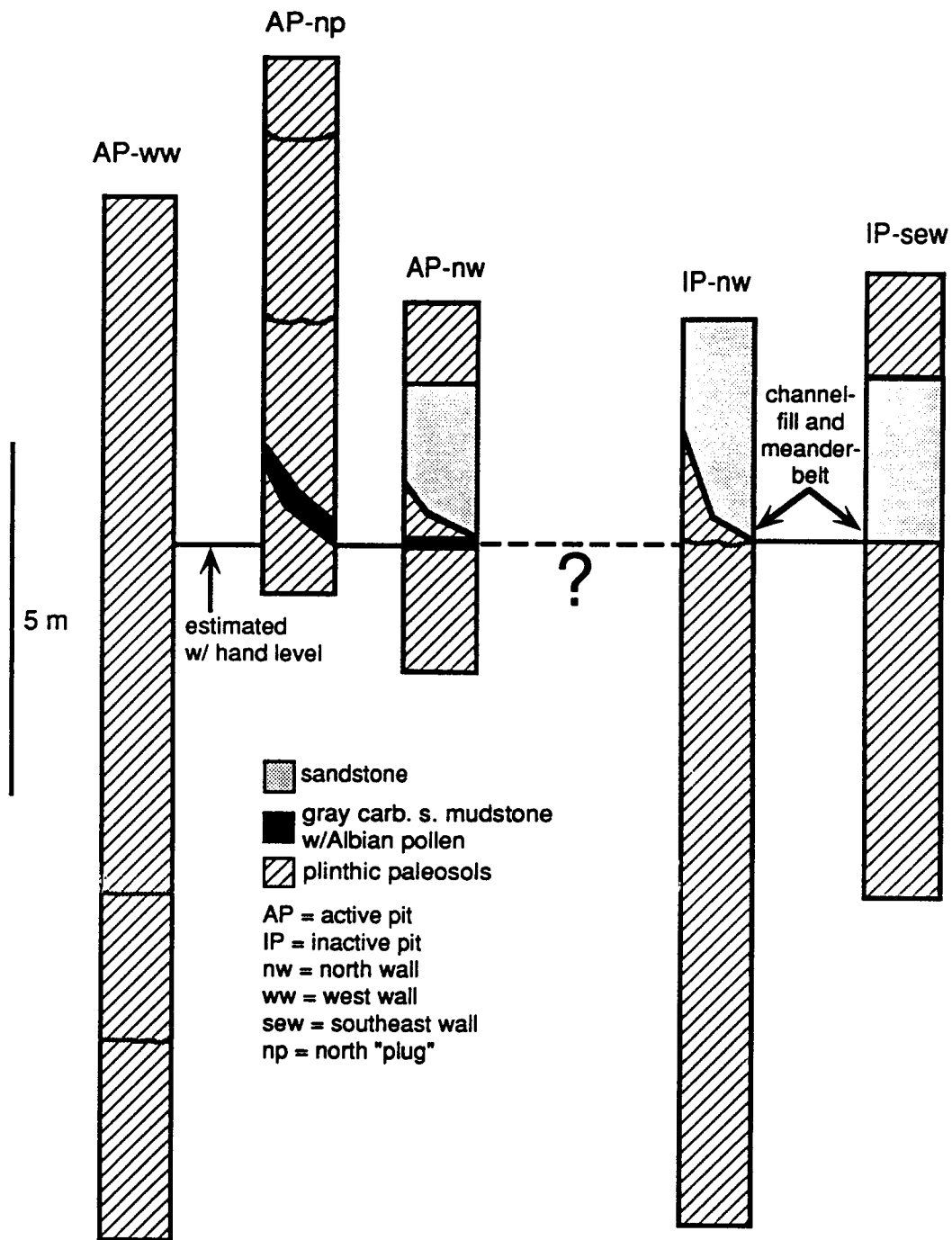


FIGURE 16/17-2. Stratigraphic relationships--Endicott pit.

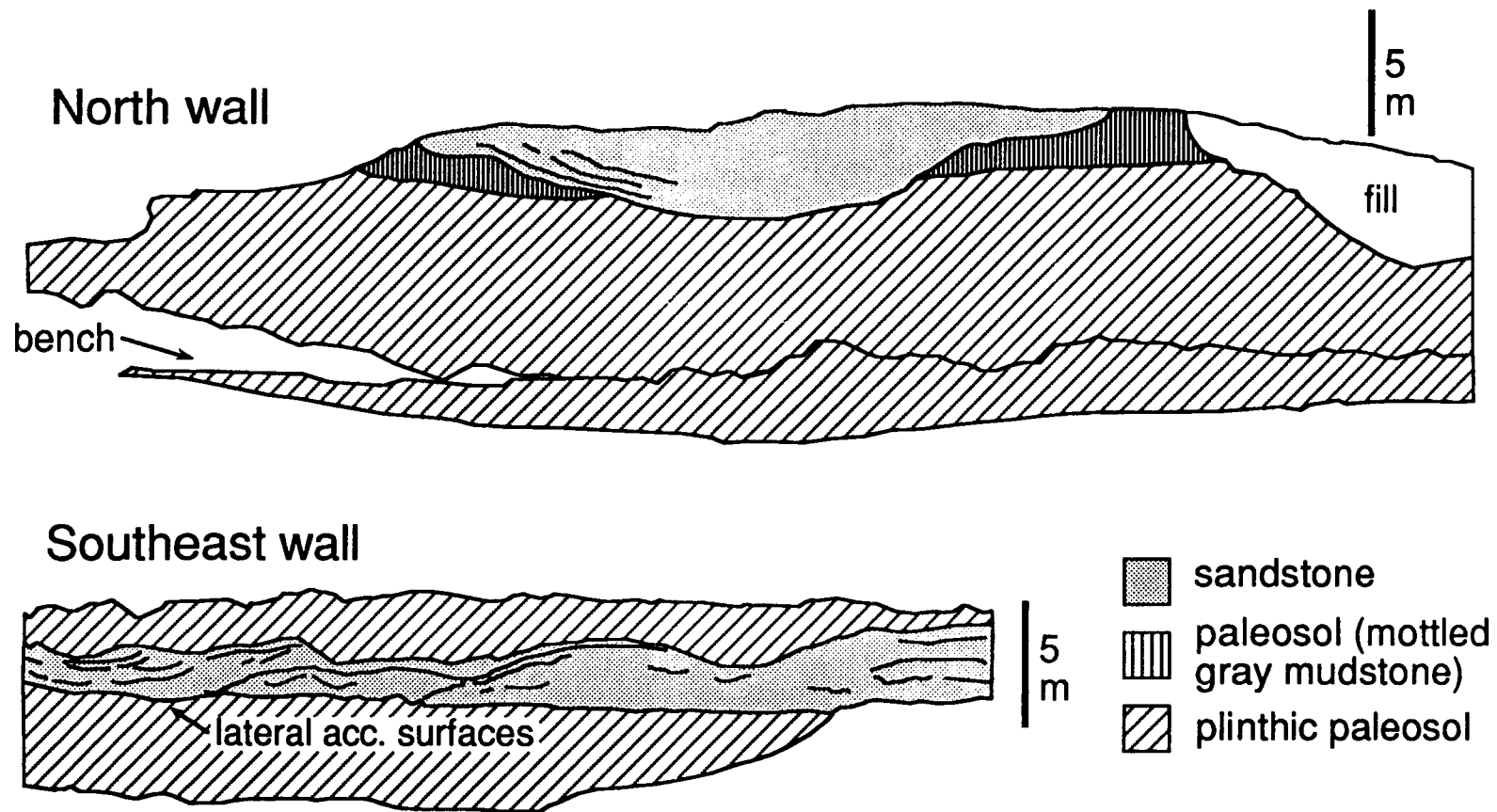


FIGURE 16/17-3. Field relationships of lower Dakota Formation paleosols in inactive part of Endicott pit, Jefferson County, Nebraska.

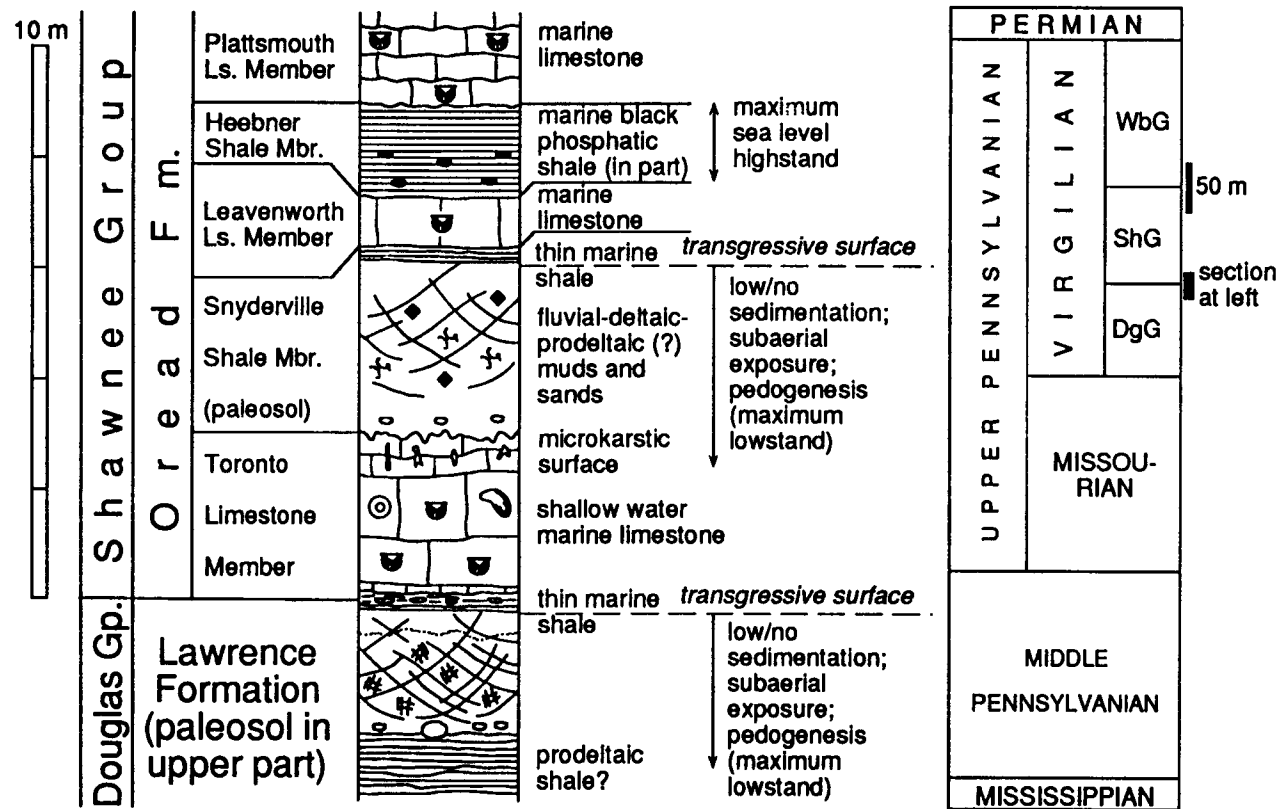


FIGURE 16/17-4. Stratigraphic section of Lawrence Formation, Snyder Member and associated units in northern Midcontinent, U.S.A.

Stop 18 Borrow Pit north of Loveland and west of Hwy 75-29: Loess-soil sequence

Location: N 1/2, Sec. 3, T77N, R44W, Pottawattamie County, Iowa
(USGS Loveland and Missouri Valley, IA 7.5 minute quadrangles)

Contributor: E. Arthur Bettis III

SANGAMON SOIL AT THE TYPE SITE OF THE LOVELAND LOESS

The Loveland Loess (or Loveland Silt) is the most widespread pre-Wisconsinan loess in the Midcontinent. It has been described throughout the Upper Midwest in the Missouri, Mississippi, and Ohio River basins (Willman and Frye, 1970; Reed and Dreeszen, 1965; Ruhe and Olson, 1978; Ruhe, 1969), and identified at several localities adjacent to the Mississippi Valley in Arkansas, and Mississippi (e.g., McCraw and Autin, 1989, and references therein). Unlike Wisconsinan loesses of the Midcontinent, the Loveland Loess has not been extensively studied, either for mineralogical composition, geometry (and thinning relationships), or chronology (other than the infamous "count down" method). Regional stratigraphic relationships suggest that the Loveland Loess is Illinoian in age and that the Sangamon Soil developed in the upper part of the unit is buried by Wisconsinan deposits. At complete sections in the Upper Midwest, the Loveland Loess is buried by the Roxana Silt in Illinois, Wisconsin, eastern Minnesota, Missouri, and Arkansas, or its correlatives in other states (the Pisgah Formation in Iowa and the Gilman Canyon Formation in Nebraska). Recently Miller and others (1985) suggested greater complexity to the loess record in the Mississippi drainage. They propose that an early Wisconsinan loess (Sicily Island Loess) occupies a stratigraphic position between the Loveland Loess and Roxana Silt along the Mississippi Valley south of Thebes Gap, and that the Roxana Silt pinches out somewhere north of Vicksburg, Mississippi (see Figure 4 in McCraw and Autin, 1989). The potential complexity of this loess record, coupled with recently determined "young" thermoluminescence (TL) age estimates of 80-90 ka for the Loveland Loess in Iowa and eastern Nebraska (Canfield, 1985; Norton and Bradford, 1985), raised questions about the Illinoian age of the Loveland Loess and prompted Forman et al. (1992) to obtain TL age estimates from the Type Section.

History of the "Loveland" in Iowa

The definition of the Loveland as well as the types of deposits included in it have changed over the past 90 years. The first descriptions of the unit were made back in the time when the debate still raged between an eolian or aqueous origin for loess. The name "Loveland" was proposed by Shimek (1909, p. 405) for reddish to yellowish silt and clay that overlay "Aftonian beds" at Logan in western Harrison County, and "Kansan" till at the Type Section in the bluff above Loveland in Pottawattamie County. Shimek noted that the Loveland was buried by calcareous loess now known to be Wisconsinan and included associated sand and gravel deposits in the Loveland. He believed that the unit was for the most part glaciofluvial in origin (quite ironic since Shimek was the major proponent of an eolian origin for the overlying Wisconsinan loess). Kay and Graham (1943) discussed the development of the term "Loveland" and applied it to Shimek's silt, which they referred to as the "loess phase", as well as underlying sand, gravel, and stratified silt and clay, all of which they grouped into the Loveland Fm. Further, Kay and Graham (1943) included the volcanic ash and its enclosing beds at the County Line ash site in Harrison County, Iowa (Pearlette

Type B; see Bettis et al, 1990 and Izette and Wilcox, 1982) in the Loveland Fm. Mickelson (1949) restudied the Loveland Fm. of western Iowa and concluded that the sand and gravels were below and could be stratigraphically separated from the Loveland Loess. He suggested the term "Loveland" be restricted to the loess, and that the deposits enclosing the volcanic ash at the County Line site be part of the Sappa of Reed (1948). Condra and others (1950) recognized three phases of the Loveland Fm., a valley phase consisting of stratified silts and clays with laminae of sand, an upland phase consisting of massive loess, and a colluvial or slope phase separating the other two phases. In 1965 Reed and Dreeszen differentiated a lower sand and gravel (Crete) member, and an upper silt member (Loveland loess) of the Loveland Fm. in Nebraska. In 1957 the Type Section of the Loveland Fm. was destroyed during excavation of a borrow pit for road construction. Daniels and Handy (1959) described a new Type Section (the present Type Section) on the north end of the borrow pit where the following sequence was exposed (present stratigraphic units and soil nomenclature indicated in parentheses):

0-78 ft.-----oxidized and unleached Wisconsin (Peoria) loess
 78-83 ft.-----oxidized and leached Wisconsin (Peoria) loess
 83-91 ft.-----oxidized and unleached Wisconsin (Peoria) loess
 91-102 ft.-----oxidized and leached Wisconsin (Peoria) loess
 102-102.25 ft.--A1b soil horizon in Farmdale increment (Pisgah Fm.)
 102.25-104 ft.--C1b soil horizon in Farmdale increment (Pisgah Fm.)
 104-114 ft.----oxidized and unleached Farmdale increment (Pisgah Fm.)
 114-118 ft.----Sangamon Soil (inceptisol) developed in Loveland Loess
 118-123 ft.----oxidized and leached Loveland Loess
 123-130 ft.----oxidized and unleached Loveland Loess
 130-134 ft.----oxidized and unleached Kansan (Pre-Illinoian) till

The present section is a little north of the area described by Daniels and Handy. It differs in that the Loveland Loess is thicker (24 ft. here vs. 16 ft.), and the Sangamon Soil is also a little thicker, but is still weakly expressed morphologically. The contacts between units seen in this section are similar to those described by Daniels and Handy (1959).

Chronology of the loesses at the Loveland Type Section

The Loveland Loess is beyond the range of radiocarbon dating and is not directly associated with other datable materials such as bone or volcanic ash. Loess is an ideal material for TL dating. Until 1989 no one had attempted to apply TL dating techniques to the Loveland Loess at the Type Section. Canfield obtained a TL age estimate of 82.7 ka (regeneration method) from the Loveland Loess near Plattsmouth

Nebraska, about 64 km (40 mi) south of here (Canfield, 1985), while Norton and Bradford (1985), also using the regeneration method, obtained a similar TL age estimate (89.2 ka) from Loveland Loess near Treynor, Iowa. Both of these age estimates suggested that the Loveland was younger in age than Illinoian, or that TL dating was providing only minimum age estimates.

A series of TL age estimates from the Type were obtained in 1989 using several methods, and the results suggest that the Loveland Loess was deposited around 140 ± 20 ka (Figure 1; Forman et al., 1992). Forman et al. (1992) attribute the disparity between these age estimates and those obtained by both Canfield (1985) and Norton and Bradford (1985) to different TL dating methods. TL age estimates obtained at the Loveland Type Section from the overlying Pisgah Fm. loess and the Peoria Loess agree well with radiocarbon ages determined on disseminated organic matter and an AMS-determined radiocarbon age on gastropod shell. These ages indicate that the Pisgah Fm. loess was deposited between about 35 and 25 ka, while the Peoria Loess began to accumulate shortly after 25 ka. Amino-acid ratios determined on gastropod shells from the Pisgah Fm. and Peoria Loess in this section support the TL age estimates and radiocarbon ages, and suggest a relatively short period of time with little or no warming between the end of Pisgah Fm. loess deposition and the onset of Peoria Loess accumulation (Oches et al., 1990). Another series of TL, OSL, and radiocarbon ages are presently being determined for the loess units at the Type Section as well as at Plattsmouth and several other localities in this area. These should help with correlations of loess units in the area and provide a sound chronologic framework for Missouri-Valley source loesses.

The Sangamon Soil and Weathering Profile in the Loveland Loess

Its hard (no pun intended) to get a good look at the morphology of the Sangamon Soil here because of the brick-like nature of the desiccated soil. You can see, however, that the Sangamon Soil has a thick solum (2.0 meters) with weak soil structure, diffuse horizon boundaries, minor evidence of clay translocation, and reddening within the solum. The upper fifty centimeters of the profile are calcareous because of post-burial recharge from overlying calcareous loesses, but the Loveland Loess exhibits a 5.5 meter thick leached zone associated with the weathering profile of the Sangamon Soil. Hard nodules of pedogenic carbonate associated with the Sangamon Soil weathering profile are present in the Loveland Loess beginning about 4.4 meters below the base of the solum. These characteristics are typical for the Sangamon Soil in bluffline exposures, and contrast with those described from sites only a few kilometers distant from the bluffline where the Sangamon Soil is much more well expressed morphologically, has abundant illuvial clay in the Bt horizon, and a much thicker leached zone below the solum.

Why does the morphology of the Sangamon Soil vary so dramatically in proximity to the Missouri River bluffline? An analogous morphologic sequence is present in modern soils developed in the Peoria Loess. Adjacent to the bluffline the modern soil on ridge crests is typically an Entisol (A-C soil profile), while only a few kilometers back from the bluffline Mollisols with A-Bw-C profiles are found on ridge crests. Morphologic properties of the modern bluffline soils are strongly influenced by excessive drainage and surface erosion on the narrow divides and steep slopes. The landscape on which the Sangamon Soil developed had less local relief, broader divides, and probably was not as excessively drained as the modern bluffline area. The morphology of the Sangamon Soil at the Type locality is not consistent with a landscape that was relatively stable for a long period of time. I suggest that this profile developed during the early or middle Wisconsinan following a period of erosion that removed soils developed on the bluffline landscape. This hypothesis could be easily tested using a technique such as ^{10}Be dating, where an estimate of the exposure time of the soil surface could be obtained and compared to values obtained on the Sangamon Soil at greater distances from the bluffline.

Pisgah Formation

Differences and disagreements in stratigraphic nomenclature have tended to overshadow the similarity of the loess sequence across the Upper Midwest. The Pisgah Formation (formerly referred to as basal Wisconsin sediment or basal Wisconsin loess in Iowa) occupies the stratigraphic position of the Roxana Silt of Illinois and the Gilman Canyon Fm. of Nebraska. This unit includes loess, colluvium, slope deposits, and mixing zone materials. It overlies the Sangamon Soil developed in Loveland Loess at the Loveland Type, and has the Farmdale Soil developed in its upper part. The Pisgah Fm. is buried by Peoria Loess. Where it is loess, the Pisgah Fm. differs mineralogically from the Roxana Silt and does not contain the buried soils characteristic of the Gilman Canyon Fm. The Farmdale Soil is developed in the upper part of the Pisgah Fm. (it was formerly referred to as the basal Wisconsin soil or the basal loess paleosol in Iowa).

There are a few features of the Pisgah Fm. loess to observe at this stop. The Farmdale Soil developed in the upper part of the Pisgah Fm. can be examined on the face above the first step. Several discontinuous dark brown burned zones are evident at and around the Pisgah Fm./Peoria Loess contact. These zones contain charcoal flecks, thin orange (burned) zones at their base, and a few pockets of wood ash. Similar burned zones have been observed at this stratigraphic position in other exposures in western Iowa and eastern Nebraska.

Peoria Loess

The Peoria Loess is quite thick at this locality and has two leached zones in its lower part. In some areas it exhibits primary eolian bedding. A good example of the eolian bedding can be observed on the western side of the face above the second step. Here the loess exhibits different bedding structures, including planar stratification, cross-bedding, and ripples (amplitude 3-4 mm) that are inversely graded. The presence of this primary bedding indicates that loess deposition in this interval was very rapid, vegetation was locally absent, and soil mixing processes were overwhelmed by deposition. Note that the bedding disappears laterally as well as vertically in the section.

Pre-Illinoian Glacial Sequence

Last, but not least, is the pre-Illinoian diamicton sequence at the base of the section, and in another section west of the loess exposure. The pre-Illinoian glacial sequence at the site has not been studied in detail, but preliminary observations suggest that it consists of a thick sequence of resedimented diamictons and associated meltwater deposits. This interpretation is based on deposit types (discontinuous diamicton and stratified sand and gravel beds), deposit geometry (discontinuous, dipping beds), contacts between beds (which show the absence of pervasive shear, but evidence for soft-sediment deformation), and the presence of deformation structures suggesting collapse related to ice melt rather than active-ice shearing processes. The diamicton is noticeably more affected by secondary weathering than Wisconsin glacial deposits, and exhibits numerous secondary features such as oxidation, extensive jointing, and accumulation of secondary iron and carbonate along joint faces and some bedding planes. The clay mineralogy of this diamicton (average 60:26:14, EX:ILL:K+C) suggests that this is one of the "A" tills of Boellstorff (1978; data on file at IDNR-GSB), part of the youngest sequence of pre-Illinoian glacial deposits in this part of the Midwest.

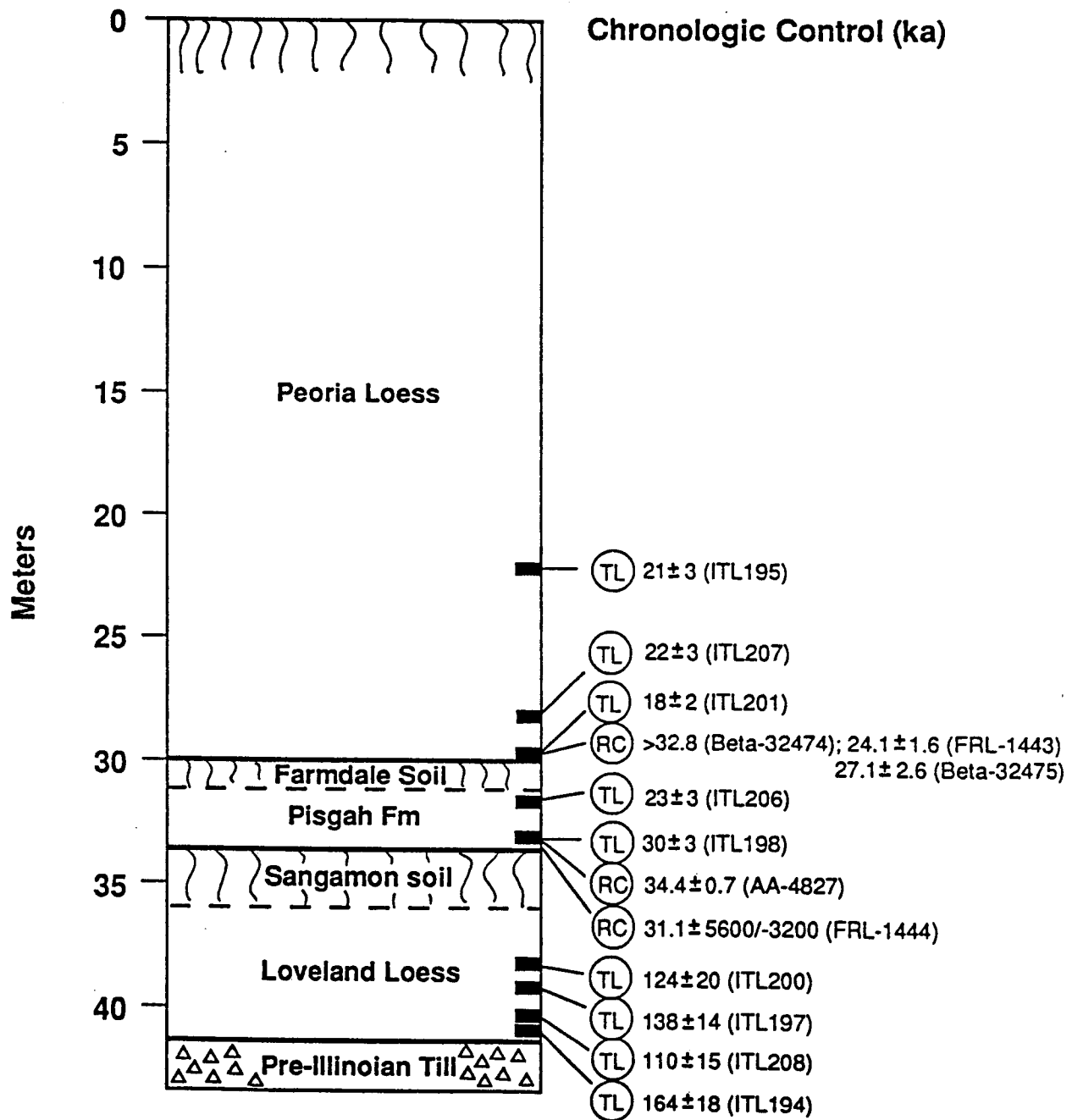
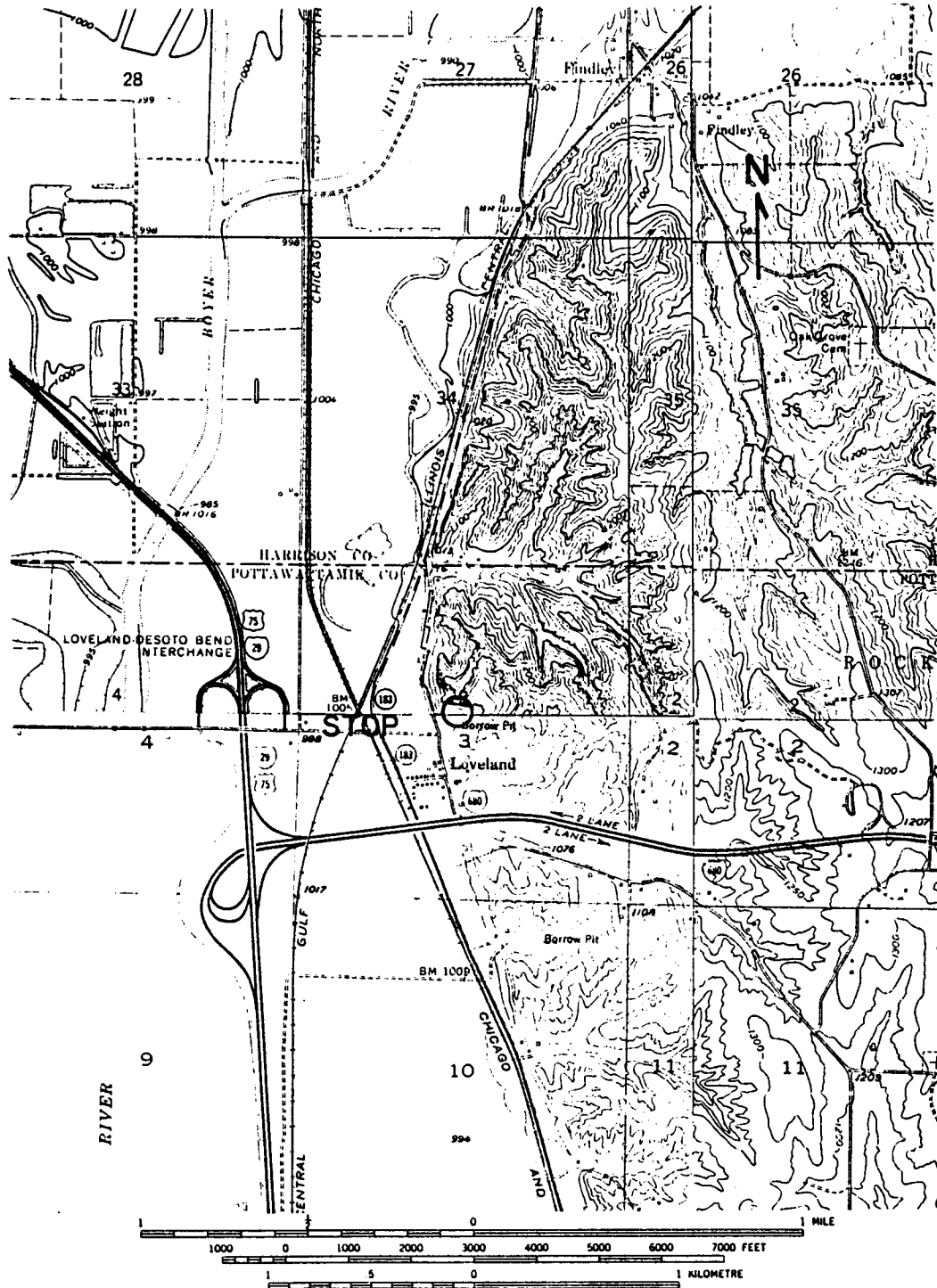


FIGURE 18-1. Stratigraphy of the Loveland Type Section showing radiocarbon and TL age estimates obtained on the loess units. See Forman et al., (1992) for details on the TL age estimates.



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Stop 19 Earlham Quarry, 1 mile southeast of Earlham: Yarmouth soil in "gumbo"

Location: E 1/2, NW 1/4 , Sec. 8, T77N, R28W, Madison County, Iowa
(USGS Earlham East, IA 7.5 minute quadrangle)

Contributor: Kathy Woida

PARTIAL TOPOSEQUENCE OF YARMOUTH-SANGAMON AND FARMDALE PALEOSOLS

Description

Earlham Quarry is situated on a broad, tabular upland divide about 30 km northeast of Ruhe's (1956, 1967) Greenfield quadrangle study area and about 50 km north of the Aftonian type area of Kay and Apfel (1929). The stratigraphy of this area was formerly interpreted as a classic Nebraskan, Aftonian, Kansan sequence (mantled by Wisconsinan loess), but only an undifferentiated Pre-Illinoian stage is currently recognized in Iowa (Hallberg, 1986). The name "Yarmouth" was retained in the Earlham Quarry study to refer to the interglacial stage preceding the Illinoian ice advance.

The general stratigraphic sequence at Earlham Quarry (Fig. 1) consists of Upper Pennsylvanian (Missourian) limestone overlain in turn by Pre-Illinoian till, undifferentiated Yarmouth-Sangamon materials, a basal Wisconsinan loess (Pisgah Formation), and the Wisconsinan Peoria loess. A shallow drainageway or swale on the paleo-landsurface is exposed along the west wall of the quarry. Insofar as local relief on this paleo-landsurface is only about 2.5 m and well-drained or moderately well-drained pedons are absent, the toposequence is incomplete.

Comments

The Yarmouth-Sangamon Paleosol (YSP), the thickest Pleistocene buried soil in the U.S. midcontinent, occurs in southern Iowa, eastern Nebraska, northeastern Kansas, northern Missouri, and portions of other surrounding states. It developed over the longest time span of any Quaternary soil known in this region and records information about the paleoenvironments of at least one glacial and two interglacial stages. However, past attempts to understand the genesis of the YSP have been largely ineffective. The paleosol appears remarkably uniform with depth, lacking any clear horizon differentiation, and has been depleted by weathering of many of the chemical constituents used in soil-development studies. The fine-grained texture ("gumbo") and lack of sedimentary structures have hindered identification of the parent material as well.

Field description and micromorphological, textural, and mineralogical analyses of five soil profiles (Fig. 1) were used to identify the major processes and conditions of Pleistocene soil evolution at Earlham Quarry (Fig. 1; Tables 1 and 2). The results suggest that the YSP is a complex polygenetic soil that developed in multiple parent materials (Fig. 2): (i) a Pre-Illinoian glacial till (DU-5) comprising the base of the paleosol; (ii) a middle unit (DU-4) of colluvium and loess, possibly Pre-Illinoian in age; and (iii) an upper unit (DU-3) of loess and reworked loess, possibly Illinoian in age. A stone zone at the top of the till indicates a period of erosion before burial by the fine-textured sediments. Depositional units 3 and 4 were

differentiated on the basis of medium:coarse silt ratios (Fig. 3). Silt ratios increase gradually upward through DU-4 and the unit is relatively rich in sand. Because the stone zone at the base of the unit suggests little sedimentologic mixing with the underlying till, the sand is likely colluvial in origin, added perhaps during beveling of the Pre-Illinoian landscape. The abrupt change in clay mineralogy at the base of the unit, however, suggests the input of some externally derived material such as loess. A loess rich in medium silt, similar to the overlying Peoria Loess, could also account for the fining-upward trend of DU-4. Slow aeolian deposition would allow new silt to become incorporated into the coarser materials by mixing and, through time, the surface would migrate upward with progressively less mixing until the surface was composed completely of loess.

A few coarse sand grains and granules occur at the bottom of DU-3 and gradually decrease in abundance upward. In the past, such particles were viewed as partial evidence for a parent material of weathered till, but it seems much more plausible that coarse grains from DU-4 worked their way upward in the profile by pedologic mixing processes such as freezing and thawing. A colluvial origin for DU-3 is also unlikely in that DU-3 is laterally continuous across the toposequence both texturally and mineralogically. The study site is situated at the highest level of the modern local landscape and there is no evidence for the existence of a former topographic high that could have supplied 2 m of colluvium to the area.

The similarity in medium-coarse silt ratios between DU-3 and DU-1 (Peoria Loess) suggests a loessial origin for DU-3, which may be equivalent in age to Loveland Loess deposits farther west in Iowa. If Ruhe's (1969) equation is used to calculate the expected thickness of Loveland Loess based on distance from the Missouri River, the Earlham area should have received about 2 m of loess, which is close to the actual thickness of DU-3.

Abundant krotovinas and the presence in thin sections of granular soil structure (probably earthworm pellets) and abundant channel pores (Tables 1 and 2), all of which imply bioturbation close to the landsurface, support the notion of a depositional discontinuity between units 3 and 4 and imply that the YSP is a composite, welded profile consisting of at least two sola (Fig. 4). Several other micromorphological features of the welded profile point to churning of the soil by shrink/swell as a significant soil-forming process that occurred after burial of the lower solum (Solum 2), and possibly before burial of that unit as well. These include slickensides, striated birefringence fabrics, and embedded clay coatings. Slickensides occur in both sola, in a zone 0.9 to 2.9 meters below the top of the YSP, and suggest at least one climatic interval of very dry seasons during the paleosol's history.

Barite nodules and coatings in Solum 2, which are not covered by clay or other colloids, probably formed relatively late in the pedogenic history of the paleosol, after the last episode of clay illuviation that affected Solum 2. Because nearly all the soil barite found today occurs in soils with a pH of about 5 or lower, the barite probably formed before burial and base resaturation of the paleosol by Wisconsinan loess, during a time of relatively acidic conditions (compatible with forest vegetation?). Poorly drained conditions were also likely, since most documented occurrences in modern soils are in hydromorphic settings.

The gradual boundary between the YSP and the overlying Farmdale Soil indicates upward growth of the solum as Wisconsinan silts were assimilated into the profile by pedologic and biologic mixing processes, including earthworm activity. Krotovinas and the worm-reworked A horizon of Solum 1 are compatible with an interpretation of grassland vegetation during this time (about 31,500 radiocarbon years B.P.), which is agreement with paleoenvironmental data from a nearby site (Baker et al., 1991) that suggested open, prairie-like upland environments at about 33,000 years B.P. The YSP's upper boundary is further obscured by numerous involutions, probably evidence of active cryoturbation of the landsurface under periglacial conditions during the last glacial maximum, between 21,500 and 16,500 years B.P.

Theoretically, paleocatenas might be expected to exhibit lateral variations among their members analogous to the differing soil types found in catenary association on the modern landsurface. Results of the Earham Quarry study show less variability than expected, even considering the absence of well-drained members in the toposequence. The most obvious difference is in the degree of mottling and gleization between the YSP in the depression and the better-drained profiles, clearly visible in the field and confirmed by micromorphology and extractable iron-oxide data. Members of the toposequence exhibit relatively few differences in other chemical parameters or in clay mineralogy. This may be largely due to the homogenizing effects of diagenesis, which included base enrichment, oxidation of organic carbon, depletion of phosphorus, and possible destruction of smectite interlayers. Pre-burial pedoturbation and bioturbation may also be partially responsible for the absence of standard depth curves for chemical constituents and the uniform clay mineral composition of the two Yarmouth-Sangamon sola.

Johnson and Watson-Stegner's (1987) "evolution" model of pedogenesis can be applied to the YSP. The composite paleosol underwent several "progressive" processes that promoted horizon differentiation, soil upbuilding, and soil thickening--e.g., slow accretion of aeolian silts, clay illuviation, and formation of soil structure. Other "regressive" processes, including bioturbation, argillipedoturbation, and cryoturbation, destroyed or obscured horizon boundaries and promoted a simplified profile. As a result of these two intersecting pathways of soil evolution, as well as post-burial diagenesis, the YSP's isotropic profile belies a genetically complex history.

TABLE 19-1. Macromorphological features of buried soils and weathering zones, profiles 3 and 5.

Unit/ Horiz. (or Zone*)	Depth (cm)	Texture †	Main Colors	Mottles/ Accumulations ‡	Structure **	Coatings/ Fillings ††	Slickensides ‡	Other Features	Lower Boundary §§
PROFILE 3									
Peona Loess MDL	191-275	sil	2.5Y 5/2	m3p 7.5YR 4/6 cip 5YR 2.5/2	0	f 10YR 4/2-cl			a wavy
Farmdale Soil 2Ab	275-287	sic	10YR 2/1 10YR 3/2 10YR 5/2	m1f 10YR 5/4	1m blk (nonpedogenic)	f 10YR 4/1 f 10YR 3/1		Matrix colors occur as bands. Few charcoal flecks	c
Y-S Solum 1 3Ab	287-310	sic	10YR 4/1 10YR 4/3	m3p 10YR 5/8	1v1&f blk	f 2.5Y 4/4-cl m 5YR 4/6-fe		Common krotovinas	g
3Ab	310-348	sic	10YR 4/1 10YR 4/3	f1p 7.5YR 2/0	2v1 blk	m 5YR 4/6-fe			c
3ABb	348-384	sic	10YR 5/1	f1d 7.5YR 3/2	2-3v1 blk	f-c 5Y 5/1-cl f 5YR 4/6-fe		Few krotovinas	g
3Bb1	384-452	sic	10YR 5/1	f1f 2.5Y 5/4	3 f&m blk	c-m 2.5Y 4/0-cl f 2.5Y 4/4-cl c 5YR 5/8-fe c 10YR 4/6-fe	c		a
3Bgb2	452-549	sic	5Y 5/1	c1-2f 2.5Y 5/4 c1-2p 7.5YR 4/6	2 f-m pr 2f blk	c 2.5Y 4/0-cl f 5Y 4/1-cl f 5YR 2.5/2-fe/mm	c	Few krotovinas	g
3Bgb3	549-574	sic	5Y 5/1 5Y 6/2	v2-3p 7.5YR 4/6	1-2v1&f blk	f 5Y 4/1-cl f 5YR 2.5/2-fe/mm	c	Few coarse sand grains	c
3BCgb	574-617	sic	5Y 5/1 5Y 6/2	c3p 7.5YR 4/6	2m pr 1m blk	c 5Y 4/1-cl c 2.5YR 3/2-fe/mm	a	Common coarse sand grains	c
Y-S Solum 2 4BCgb	617-693	sic	5Y 5/1 5Y 6/1 10YR 5/6	c3p 7.5YR 4/6 c1 f 5Y 5/4 f1p 2.5YR 4/8	1m blk to 0	7.5YR 2/0-mm c 2.5YR 3/2-fe/mm f 10YR 5/8-fe	c	Abundant coarse sand grains. Few krotovinas	c
Pre-Illinoian till 5Cg	693-716	cl	5Y 6/1 10YR 5/6	f1p 2.5YR 4/8	0	m 2.5YR 3/2-fe/mm		Common pebbles	g
MOL	716-820	l	7.5YR 5/8	c2p 5Y 7/2	0	c 10YR 3/1 f 5Y 7/2		Common pebbles	a
PROFILE 5									
Peona Loess MDL	361-376	sic	5Y 5/2	m2d 10YR 5/3 cip 10YR 5/6	0	c 5YR 5/8-fe			a
Farmdale Soil 2Ab1	376-389	sic	10YR 4/1 10YR 4/4	c1f 10YR 3/2 m1d 10YR 5/6	0	c 7.5YR 2.5/0		Few charcoal flecks	a
2Ab2	389-404	sic	10YR 4/1	c2 f 10YR 5/4 f1p 10YR 5/6	0	f 7.5YR 4/6-fe		Few charcoal flecks	a
2Bwb	404-422	sic	10YR 5/2 7.5YR 5/6	m1f 7.5YR 5/8	1v1 blk	c 7.5YR 5/8-fe f 10YR 7/2-si			g
Y-S Solum 1 3AB	422-465	sic	10YR 5/2 7.5YR 5/6	m1-3 f 7.5YR 5/8	1v1 blk- 2-3f blk	f 5YR 5/8-fe c 10YR 7/2-si		Common charcoal flecks	g
3Bb	465-490	sic	10YR 5/2		2-3f blk	f 10YR 4/1-cl c 10YR 7/2-si			g
3Bb1	490-544	c	2.5Y 5/2		2-3f blk	f 10YR 4/2-cl c 5YR 2.5/1-mm			g
3Bb2	544-580	sic	2.5Y 5/2	c2-3p 7.5YR 4/6	2m blk- 2f blk	f 2.5YR 4/6-fe c 10YR 5.5/1-cl c 5YR 2.5/1-mm	f		g
3Bgb	580-610	sic	2.5Y 5/2 5Y 5/1	c2-3p 7.5YR 4/6	2m blk	f 2.5YR 4/6-fe c 10YR 5.5/1-cl c 5YR 2.5/1-mm	f	Few small pebbles	g
Y-S Solum 2 4Bgb	610-670	sic	5Y 5/1 7.5YR 5/6		2m blk	c 7.5YR 5/2-cl c 7.5YR 2/0-mm	c	Common pebbles. Few barite coatings & nodules.	g
4BCgb	670-737	sic	5Y 5/1 7.5YR 5/6		1m blk		c	Common pebbles. Few barite coatings & nodules	g
Pre-Illinoian till 5C	737-864	cl	10YR 5/6 10YR 5.5/2		0	c 7.5YR 2/0-mm	c	Common large pebbles	a
* Weathering zones: D=deacidized; L=leached; M=mottled; O=oxidized. † c=clay; l=loam; si=silt. ‡ f=few; c=common; m=many; a=abundant; 1=fine; 2=medium; 3=coarse; f=finer; d=distinct; p=prominent. ** 0=structureless; 1=weak; 2=moderate; 3=strong; v1=very fine; f=fine; m=medium; c=coarse; pr=prismatic; gr=granular; blk=angular blocky; sblk=subangular blocky. †† cl=clay coating; fe=Fe-oxide coating; mn=Mn-oxide coating; fe/mn=Fe/Mn-oxide coating; si=silt coating. §§ a=abrupt; c=clear; g=gradual; d=diffuse.									

TABLE 19-2. Micromorphological features of buried soils and weathering zones, profiles 3 and 5.

Unit/ Horizon (or Zone*)	Depth (cm)	Micro- structure†	Porosity (%)	Void Type §				Sort- ing	RDP**	B-fabric ††	Textural Pedofeatures §§				Amorphous Pedofeatures §§				Fecal pellets ‡‡	
				pl	ch	pk	vg				clay cont.	emb. clay	dark brown	silt cont.	Fe nod.	Fe mottle	Fe coat.	Mn nod.		Mn coat.
PROFILE 3																				
Peoria Loess MDL	223-229	massive	5-10	50	50			well	sap	stipple sp						o	c	c		
Farmdale Soil 2Ab	275-280	crack	10-15	50	50			well	dap	stipple sp.	o				o	c		r		
Y-S Solum 1 3Ab	289-296	spongy	20-25	25	25	50		mod	dap	circular str.	o	c	o		o	r		r	va	
3A1b	317-323	sbk (w) spongy	15-25	35	35	30		poor	dap	random str. circular str.	c	c	c		r		m		m	
3ABtb	357-363	sbk (m) spongy	25-30	20	15	65		poor	dap	random str. circular str.	o	m	o		r			r	c	
3Btb1	392-399	sbk (s)	15-20	65	10	25		poor	dap	random str. porostriated	a	a	c		o	o	o	r		
3Btb1	423-430	sbk (s)	20-25	65	35			poor	dap	random str. porostriated	a	m	o		r					
3Btgb2	488-495	abk (m)	15-20	65	15	20		poor	dap	random str. porostriated	m	m	c		r		r		r	
3Btgb3	559-566	sbk (m)	15-20	70	25	5		poor	dap	porostriated random str.	c	o			c	a	r	r		
3BCgb	580-587	crack	15-20	65	20	15		poor	dap	porostriated monostriated	o	o			r	c	c	r		
Y-S Solum 2 4BCgb	633-640	abk (m)	15-20	70	10	20		poor	dap	mozaic sp. granostriated	o	r			r	a	c	o		
Pre-Illinoian till 5Cg	697-704	massive	8-10	10	20	70		unsort	dap	mozaic sp. granostriated	r					va		o	o	
PROFILE 5																				
Peoria Loess MDL	371-378	crack	5-10	60	40			well	dap	stipple sp.	r					o	c	r	r	
Farmdale Soil 2Ab1	384-391	platy (wk)	5-10	50	50			mod	dap	stipple sp.	r				o	o	c	r	r	
2Ab2	391-396	platy (wk)	5	40	40	10	10	mod	dap	stipple sp.					o	r	o	r	r	
2Bwb	401-406	channel	5	15	65	10	10	mod	dap	stipple sp.	o		r		o		o	o	o	
2Bwb	414-419	spongy	10-20	40	10	50		mod	dap	circular str.	r		o	o	m	r	o		va	
Y-S Solum 1 3Ab	432-437	spongy	20-30		10	90		mod	dap	circular str.	r	r	o	o	m		r		va	
3Ab	462-467	granular sbk (s)	30-40	5	10	80	5	mod	dap	circular str.	o		r	o	m	r	c	r	va	
3Btb1	493-498	abk (m)	10-15	80	20			mod	dap	random str.	r	a		r	m	o			r	
3Btb2	559-566	abk (m)	15-20	70	10	20		poor	dap	random str. porostriated	c	m	r		o	c	c		m	
3Bgb	604-610	abk (m)	15-20	90	5	5		poor	dap	porostriated monostriated	o	o			c	c	m	o	c	
Y-S Solum 2 4Btgb	610-616	abk (m)	15-20	40	40	20		poor	op	porostriated monostriated	m	c			o	c	c		a	
4Btgb	645-650	abk (m)	15-20	85	15			poor	sap	porostriated granostriated	m	c			r	a			c	
4BCtgb	670-676	sbk (w)	15-20	80	15	5		poor	sap	porostriated granostriated	m	c			r	a			o	
Pre-Illinoian till 5C	808-815	massive	5-10	70	5	10	15	unsort	sap	mozaic sp. granostriated	r				o	a		c	o	

* Weathering zones: D=deoxidized; L=leached; M=mottled.

† abk=angular blocky; sbk=subangular blocky; w=weak; m=moderate; s=strong.

§ pl=planar void; ch=channel; pk=compound packing void; vg=vugh.

** Related distribution pattern (based on a coarse/fine limit of 20 µm): sap=single spaced porphyric; dsp=double spaced porphyric; op=open porphyric.

†† Birefringence fabric: stipple speckled, mozaic speckled; circular striated, random striated.

§§ Estimated abundances: r=rare (<1%); o=occasional (1-3%); c=common (3-5%); m=many (5-10%); a=abundant (10-20%); va=very abundant (>20%).

(Textural pedofeatures: emb. clay=embedded clay coating; dark brown=dark brown clayey patch).

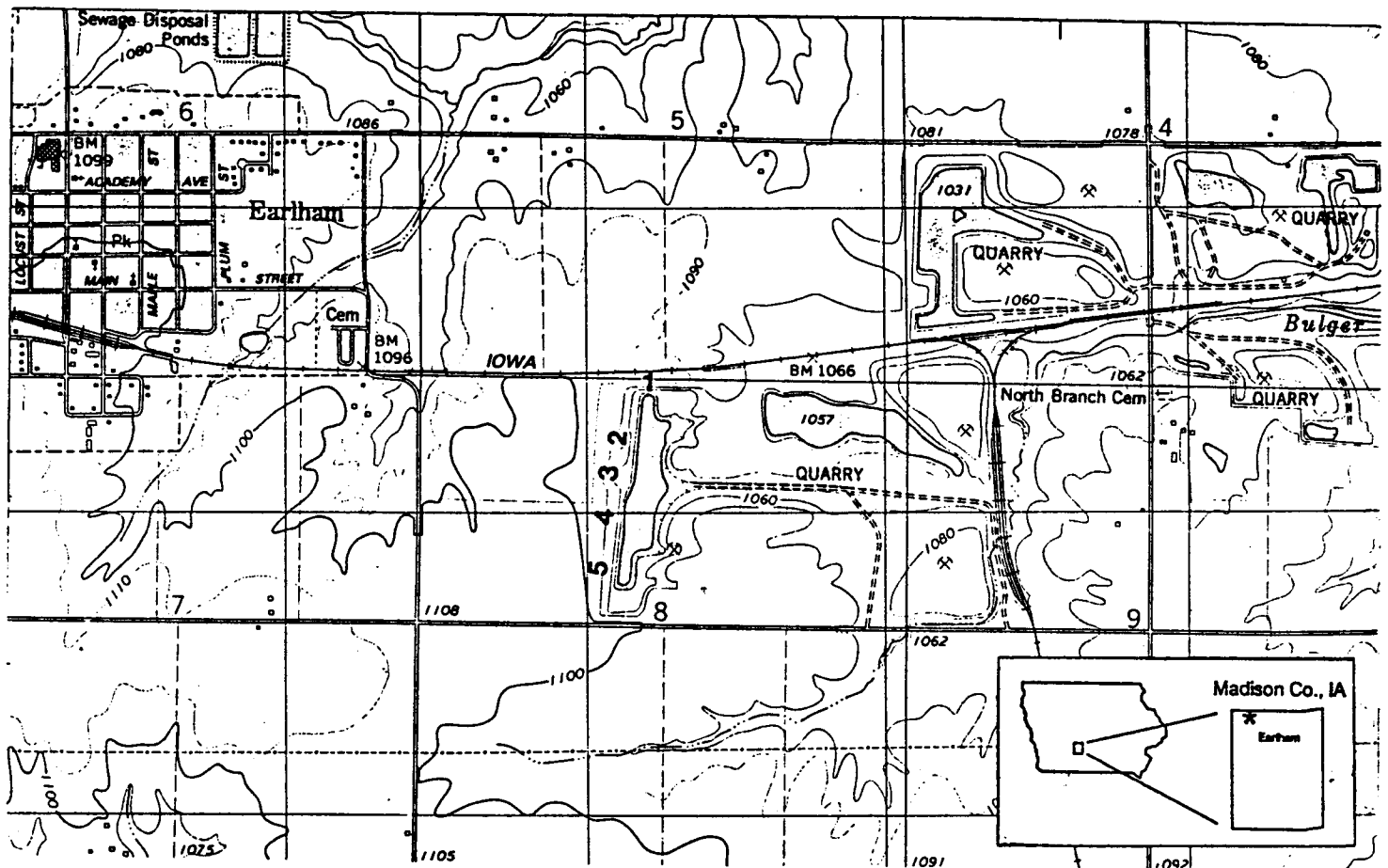


FIGURE 19-1. Portion of the Earlham East 7 1/2 minute quadrangle map, showing the Earlham Quarry study site (E 1/2, NW 1/4, Sec. 8) and the topography of the surrounding area. Inset map shows location of Earlham in Madison County, Iowa. Numbers (1-5) along quarry walls indicate approximate position of soil profiles described in the study.

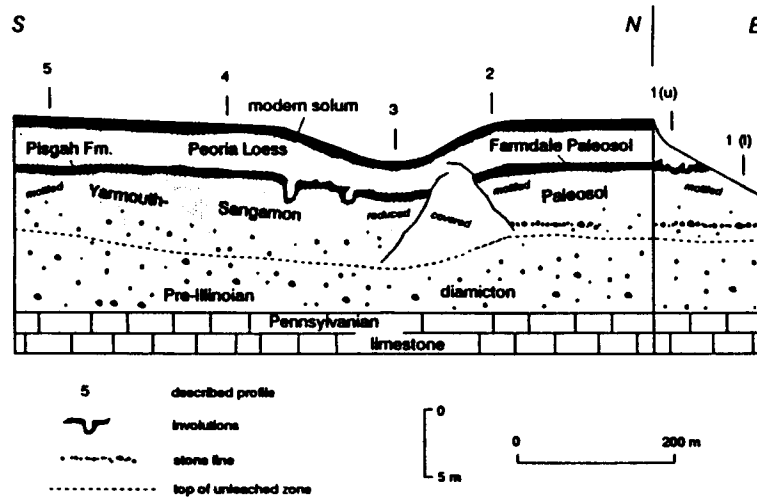


FIGURE 19-2. Site stratigraphy.

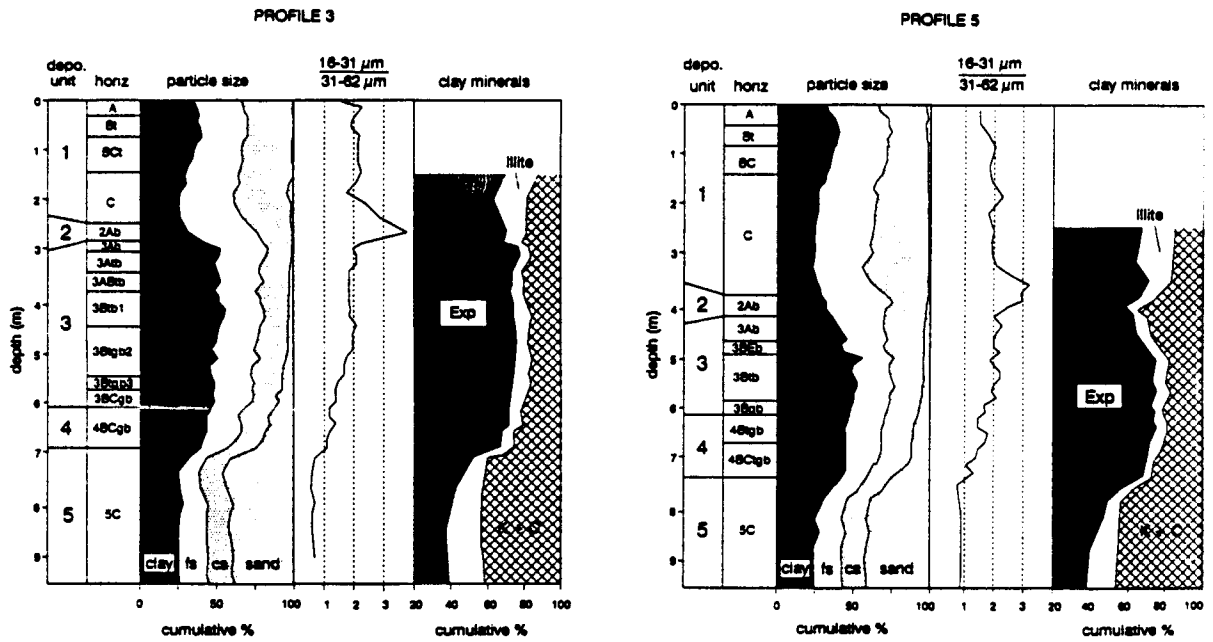


FIGURE 19-3. Profile 3 and 5 data.

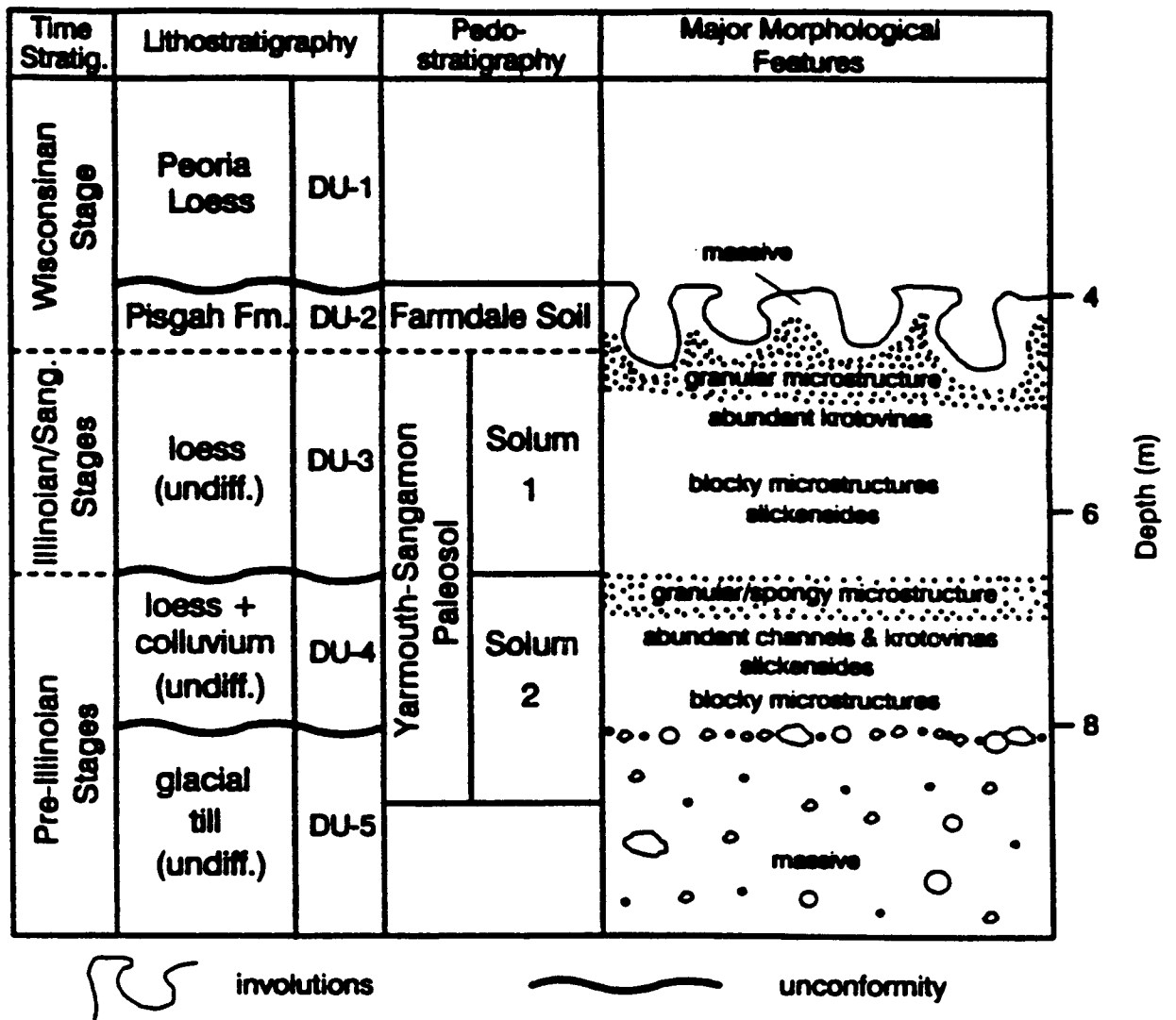


FIGURE 19-4. Stratigraphic sequence.

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Stop 20 Conklin Quarry, north of I-80, Iowa City, Iowa: pre-Illinoian glacial sequence

Location: E 1/2, Sec. 32, T80N, R6W, Johnson County, Iowa
(USGS Iowa City West, IA 7.5 minute quadrangle)

Contributors: E. Arthur Bettis III, T. J. Kemmis and G.R. Hallberg

QUATERNARY PALEOSOLS AND THE LATE SANGAMON PALEOGEOMORPHIC SURFACE

At this stop we will examine a portion of the Late Sangamon paleogeomorphic surface and several facies of the Sangamon Soil developed on glacial diamicton (till). We will also examine the deep weathering profile associated with the Sangamon Soil and discuss the merits of using weathering profiles, rather than soil profiles, to judge the relative rank (interglacial vs. interstadial) of paleosols in glacial sequences.

Conklin Quarry contains the most complete exposure of the pre-Illinoian glacial sequence in the Upper Midwest. The Quaternary sequence in the quarry has been extensively studied during the last 20 years and a detailed picture of the stratigraphic succession and associated buried paleogeomorphic surfaces has been developed. At this time four buried paleogeomorphic surfaces are available for examination. We will examine three of these developed on glacial diamicton and associated alluvial deposits. In order to better place the paleolandscapes and associated soils in context a summary of the pre-Illinoian stratigraphic framework for eastern Iowa follows.

The Quaternary Stratigraphic Succession

The pre-Illinoian glacial deposits of Iowa are monotonously similar: where unweathered, the diamictons (glacial tills) are all gray and have similar loam textures. As a consequence, it has been necessary to make stratigraphic differentiation using laboratory data rather than field properties. Further, the stratigraphic units are differentially preserved beneath different geomorphic surfaces that have developed as the drainage system evolved. A definitive stratigraphic sequence was constructed only after an exhaustive regional drilling program that included transects across the different surfaces and across bedrock valleys where the thickest and often most complete stratigraphic sequences were preserved (Hallberg, 1980a, b).

Pre-Illinoian deposits have been lithostratigraphically classified into the Alburnett Formation and younger Wolf Creek Formation (Hallberg, 1980a). Both formations consist predominantly of glacial diamictons (largely basal tills), but include other deposits as well. The formations and their members are differentiated by various physical and mineralogical characteristics (Hallberg, 1980a, b), but clay mineralogy (Table 1) is the main property used to distinguish between deposits at the formation level. Whereas some lithostratigraphic units, such as the Wedron and Glasford formations, also correspond to chronostratigraphic intervals, the Wolf Creek and Alburnett formations do not; each formation consists of multiple depositional units and paleosols, and includes deposits of several glacial and interglacial episodes. The Wolf Creek and Alburnett formations, therefore, are not correlative in any way to the former concepts of 'Kansan' and 'Nebraskan' glaciations, erroneous concepts based on inadequate stratigraphic data (e.g., Hallberg, 1986; Boellstorff, 1978). However, because of the fragmentary record for the pre-Illinoian, formal chronostratigraphic classification has not yet been developed. In western Iowa, where thicker sequences

are preserved, stratigraphic units correlative to the Wolf Creek and Alburnett formations overlie still older glacial deposits (Fig. 1; Hallberg, 1986). In the absence of radiocarbon dating, absolute ages for the different stages are unknown, although fission-track ages on volcanic ashes buried within the western Iowa sequence provide some time control (e.g., Boellstorff, 1978; Hallberg, 1986). The oldest pre-Illinoian glacial deposit underlies a volcanic ash with a fission-track age of 2.2 million years B.P. ("B" on Fig. 1), and the youngest pre-Illinoian glacial deposit overlies a volcanic ash dated at about 610,000 years B.P. ("O" on Fig. 1), and is estimated to be about 500,000 years old (Hallberg, 1986). Paleomagnetic study of the detrital remnant magnetism (DRM) of sediments at Conklin Quarry indicates that deposits of the Alburnett Formation have reversed polarity, and are thus interpreted to be older than 790,000 years ago (the Bruhnes reversed/Matuyama normal boundary; Johnson, 1982), whereas deposits of the Wolf Creek Formation have normal polarity, and are younger than 790,000 years old (Baker, 1985; Baker and Stewart, 1984).

The Wolf Creek Formation is dominated by massive, texturally and compositionally uniform basal tills, and is subdivided into members based on lithologic differences between these tills. Diamictic units within the Alburnett Formation, however, have virtually identical properties; as yet there is no way to differentiate or correlate Alburnett Formation deposits on a regional scale.

During the Wisconsinan two loess sheets were deposited across eastern Iowa. The oldest is a thin sheet that has largely been reworked by colluvial (and periglacial?) processes and pedogenically altered. This unit is referred to as the Pisgah Formation. Its stratigraphic position is the same as the Roxana Silt of Illinois and the Gilman Canyon Formation of Nebraska, but its lithologic properties differ. The youngest Wisconsinan loess is the Peoria Loess. These loesses mantle Wisconsinan and older geomorphic surfaces.

The Quaternary succession exposed at Conklin Quarry has varied through time with quarry operations, and has provided an evolving picture of stratigraphic and geomorphic relationships at the site. The most complete stratigraphic successions were exposed in 1975-76 and 1983-84. Figure 2 presents particle-size data from the most complete 1983-84 section. The north end of the present site is approximately 150 m south of the 1983-84 site. Additional descriptions and data for exposures at the site are given in Hallberg et al. (1984) and Kemmis et al. (1992).

Regional Setting

Conklin Quarry is located in the Southern Iowa Drift Plain near its northern border with the Iowan (erosion) Surface (Prior, 1991). The Southern Iowa Drift Plain was glaciated several times prior to the Illinoian Episode during the interval from before 2.2 million to 500 thousand years before present (Hallberg, 1986). Quaternary deposits in this area discontinuously cover an irregular bedrock surface. Since the end of pre-Illinoian glaciations, an integrated drainage network has developed from episodic erosion of the landscape. Two Wisconsinan loess sheets mantle the landscape; these are differentially preserved as a result of late Wisconsinan and Holocene erosion.

The Quaternary record of the Southern Iowa Drift Plain is fragmentary because of multiple periods of both glacial and subaerial erosion. No primary glacial landforms remain, although scattered loess-mantled tabular divides underlain by 'Yarmouth-Sangamon' paleosols are considered to be possible remnants of the youngest pre-Illinoian drift plain (e.g., Kay and Apfel, 1929; Ruhe, 1969). The present landscape is thus dominated by stream dissection and erosional landforms. Across the Southern Iowa Drift Plain, at least four different *sets* of surfaces can be identified: the Yarmouth-Sangamon, Late-Sangamon, Iowan and Holocene surfaces.

In eastern Iowa, the loess-mantled Yarmouth-Sangamon surface is preserved only on narrow, nearly flat

upland divides (Fig. 3). This surface is perhaps the remnant of the youngest pre-Illinoian drift plain which was subject to weathering and local modification until burial by Wisconsinan loesses. Generally, a thick gray (poorly drained) soil (now buried) is developed on the Yarmouth-Sangamon surface. This soil was named the Yarmouth-Sangamon Paleosol (Ruhe et al., 1967) because it was presumed to transgress Yarmouth and Sangamon time. Unfortunately, like many other paleosols, it was named in the context of time terms rather than independently, as recommended by the present stratigraphic code (NACSN, 1983). In strict terms, the 'Yarmouth-Sangamon Paleosol' is a facies of the Sangamon Soil (Geosol) because the units which bury and stratigraphically define them are equivalent.

The Late-Sangamon surface is a set of erosion surfaces cut into and inset below the Yarmouth-Sangamon surface of the primary divides. The break between the Yarmouth-Sangamon and Late-Sangamon surfaces is marked by topographic, geomorphic, and pedologic discontinuities (Ruhe et al., 1967; Hallberg et al., 1978b). In eastern Iowa, the Late-Sangamon surfaces generally consist of gently sloping loess-mantled pediments that are now only partially preserved as steps (levels) at about the same elevation on interfluvies. In localized areas, other parts of the Late-Sangamon landscape, from the pediments to the valley-slope fans to the floodplain, are preserved. After, and in part during, the later stages of cutting of the Late-Sangamon erosion surfaces, soil formation occurred, indicating a change to relatively more stable hillslope conditions. This soil developed until burial by Wisconsinan loesses, and has been called the Late-Sangamon Paleosol in Iowa (e.g., Ruhe et al., 1967; Hallberg et al., 1978b). Compared to the Yarmouth-Sangamon paleosols, Late-Sangamon paleosols are generally less weathered, have thinner sola, and are better drained with generally red to reddish-brown colors (Ruhe, 1969; Hallberg et al., 1978b, 1980b). Late Sangamon paleosols are typically developed in multiple parent materials: 1) an upper unit of "pedisegment," sediment derived from upslope erosion of the pediment; 2) a stone line or gravel lag which marks the pediment/erosion surface; and 3) underlying glacial deposits (Canfield et al., 1984; Ruhe et al., 1967; Ruhe, 1969). Pedogenesis commonly continued during slow burial by Wisconsinan-age Pisgah Formation deposits, resulting in upbuilding of the Late-Sangamon Paleosol. In strict terms, the Late-Sangamon Paleosol is also a facies of the Sangamon Soil because the stratigraphic units which bury and define it are the same as those in the type area of the Sangamon Soil.

Another set of erosion surfaces, called the Iowan Surface, are inset below the Late-Sangamon surfaces in the Southern Iowa Drift Plain. This set of erosion surfaces have been referred to by various terms in previous literature: the "Iowan surface" (Ruhe et al., 1968; Prior, 1991), the "Iowan Erosion Surface" (Ruhe, 1969; Hallberg et al., 1978b), and the "Early Wisconsinan pediment" (Ruhe et al., 1967). The break between the Late-Sangamon and Iowan surfaces is again marked by topographic, geomorphic, and pedologic discontinuities. In eastern Iowa, the Iowan surface consists of gently sloping, loess-mantled pediments which are generally shorter in length than the Late-Sangamon pediments. Again, only remnants of these pediments are preserved, and they occur as steps or levels down interfluvies below Yarmouth-Sangamon and Late-Sangamon surfaces and above Holocene surfaces. The Iowan surface represents a renewed period of relatively rapid downcutting, and it is again marked by a stone line or gravel lag that is commonly overlain by a thin increment of colluvium formed as the erosion surface developed upslope. The Iowan erosion surfaces were cut just prior to and during the Wisconsinan glacial maximum (c.a. 21,000-16,500 B.P.; Bettis and Kemmis, 1992). Regional deposition of the Peoria Loess occurred during development of the surfaces: Peoria Loess is thinner on Iowan surfaces than on Yarmouth-Sangamon and Late-Sangamon surfaces upslope, and basal radiocarbon ages are younger (Hallberg et al., 1978b; Ruhe et al., 1968). Early increments of Peoria Loess accumulated on the relatively "stable" Yarmouth-Sangamon and Late-Sangamon surfaces, but did not accumulate where the Iowan erosion surface was actively being developed. During the later phases of Peoria Loess deposition, the Iowan surfaces stabilized and the last increments of loess were deposited on them. No buried soil is present on this surface because it was immediately buried by Peoria Loess.

A final period of downcutting and headward extension of the drainage system has occurred during the Holocene. Holocene erosion surfaces are present on portions of upland slopes and in the upper part of the drainage network as a result of headward stream extension. The Holocene alluvial valleys consist of a series of multiple, often subtle terrace surfaces. These terraces are underlain by deposits of the DeForest Formation (Bettis, 1990; Bettis et al., 1992).

In different parts of the Southern Iowa Drift Plain various surfaces dominate the landscape because of differential preservation (Hallberg et al., 1980; 1978b; Ruhe, 1969). In much of east-central and southern Iowa, the loess-mantled Late-Sangamon surface is best preserved. Conklin Quarry has exposed several cross-sections of a Late-Sangamon interfluvium. Because such exposures are uncommon, the Conklin Quarry exposures present an excellent opportunity to study this ancient landscape.

The Late Sangamon Paleogeomorphic Surface and the Sangamon Soil

The pre-Illinoian glacial sequence at Conklin Quarry is truncated by the Late Sangamon erosion surface (Ruhe et al., 1967; Ruhe, 1969; Hallberg et al., 1984), an erosion surface that developed as the present regional drainage network evolved. The present exposure cuts across a loess-mantled interfluvium of the Late Sangamon erosion surface (Fig. 4). The Sangamon Soil, and the underlying weathering zones associated with it, developed downward from the erosion surface (or any thin overlying colluvium associated with it). [Note: previous studies, following Ruhe, 1956, referred to this soil-stratigraphic unit in relative-time terms as the Late Sangamon Paleosol; present pedomorphostratigraphic practice is independent of time, and soil-stratigraphic units are defined on the basis of the unit which buries them. This soil-stratigraphic unit corresponds to the Sangamon Soil because it is buried by the Pisgah Formation, a unit stratigraphically (but not lithologically) equivalent to the Roxana Silt that buries the Sangamon Soil in the type area (e.g., Follmer, 1978; Willman and Frye, 1970)].

From a distance, the Sangamon Soil appears as a dark reddish band arcing across the interfluvium, but the Sangamon Soil facies here vary with position across the slope (these paleosol 'facies' are analogous to soil 'series' of modern soils). Table 2 gives a description of the Sangamon Soil at the crest of the Late Sangamon erosion surface. At this position, the Sangamon Soil is morphologically well expressed with a thick E-Bt-C horizon sequence preserved. The soil's gradational upper boundary with overlying colluvial and eolian sediments of the Pisgah Formation reflects slow burial and up-building as Pisgah Formation deposits began to bury the surface of the Sangamon Soil.

No A horizon is macroscopically preserved; organic matter is not persistent after burial in well drained positions (Yaalon, 1971) because it is rapidly destroyed by oxidation, and organic-matter content was probably low anyway because soils associated with conifer forests, like that existing at the time of burial, have low organic-matter content. The E horizon is thin, and exhibits weak to moderate platy soil structure, breaking to weak to moderate very fine subangular blocky structure (Table 2). The Bt horizon is over 1.5 m thick - note the clay "bulge" of the Bt horizon in the particle-size profile on Figure 2. The Bt horizon has well expressed strong subangular to angular blocky soil structure that grades downward in size from very fine to coarse. In the main part of the Bt horizon, peds are continuously coated with illuvial clay, and in the lower part of the horizon, larger, joint-like structural units are coated with very dark gray secondary manganese-oxide coatings.

Color has often mistakenly been used to relate paleosols to climate and/or duration of development (Ruhe, 1965). The Sangamon Soil here has not been studied in detail, but micromorphological study by Thompson and Soukup (1990) of a Sangamon Soil in western Iowa reveals that the reddish color is primarily related to secondary iron oxides that coat, and therefore are younger than, the secondary clay coatings on soil peds. This led Thompson and Soukup to suggest that the soil's reddish color post-dates burial, and results

from translocation of iron weathered from the overlying Wisconsin loesses. While the micromorphology of the Sangamon Soil at Conklin Quarry has not been studied, the suggestion that the reddish color results from post-burial iron translocation is a reasonable possibility that needs further investigation.

The weathering profile of the Sangamon Soil is a predictable succession of weathering zones with depth: MOJL, MOJL2, MOJU2, MOJU, MRJU, UJU, and UU (see the appendix for a summary of weathering zone terminology). The Aurora Till Member of the Wolf Creek Formation (the unit in which the Sangamon Soil and its weathering profile are developed along the crest of the interfluvium) is deeply oxidized, with oxidation locally extending to depths over 9 m (30 ft). Leaching extends 0.5 to 1 m below the base of Sangamon Soil solum. Large, hard secondary carbonate nodules occur at the base of this leached zone and extend for about 1.5 m into the calcareous diamict. Zones of secondary carbonate up to 5 cm wide discontinuously cement joints and some of the joint intersections in the diamict. An horizon of sand lenses in the Aurora Till Member occurs at or near this zone of secondary carbonates, and contacts between the till and the lenses are commonly cemented with secondary carbonate several centimeters thick. All of this secondary carbonate was derived from the dissolution of primary carbonate minerals in the Sangamon Soil and reprecipitation with depth.

Vertical joints (the J designation in MOJL, MOJL2, MOJU2, MOJU, MRJU, and UJU) extend from the Sangamon Soil into the top of the unoxidized, unleached diamict of the Aurora Till Member. Locally, where the member is thin, these joints extend all the way through thin, unoxidized, unleached diamict; a few of these joints extend downward into the underlying stratigraphic unit, but most end at the base of the till member. The vertical joints form an irregular polygonal network. Joint frequency varies spatially with depth from the Late Sangamon paleogeomorphic surface. Eight meters (25 feet) below the crest of the Late Sangamon interfluvium joint spacings range from 2-75 mm with a mean of 26 mm in the MOJU weathering zone. Near the top of the underlying unoxidized (UJU) zone at a depth of about 10.5 m (35 ft), spacings range from 370-1090 mm with a mean of 688 mm. But spacings vary with respect to the paleohillslope, so that to the south on a lower step of the Late Sangamon surface, where the Aurora Till Member (and the preserved unoxidized diamict) is thinner, the spacings range between 100 and 690 mm with a mean of 377 mm.

Joints are opening fractures (Pollard and Aydin, 1988), and their occurrence as a polygonal network indicates formation by shrinkage. The association with oxidized (unsaturated) weathering zones (Hallberg et al., 1978a) suggests the joints formed by desiccation, and this is further supported by the relationship of joint frequency to the landsurface, with decreasing frequency downward and away from the paleolandsurface.

Through time, quarry operations have exposed different portions of the Late Sangamon paleolandscape. Locally, the Late Sangamon erosion surface descends to either shallow swales, former drainageways or sidevalleys on the Late Sangamon surface (see also Ruhe et al., 1967). In such positions, the Sangamon Soil thickens into swales and no longer exhibits dark reddish brown colors, but is gleyed, marked by mottled gray and greenish colors indicative of poor soil drainage in these landscape positions. During 1975-76 a large and deep Late Sangamon sidevalley was exposed. The valley was nearly 6 m deep and about 3 m wide. A thick cobble lag at the base of the valley fill was overlain by poorly sorted and crudely stratified deposits grading upward from sandy loam to silty clay loam, clay loam, and clay. Occasional pebbles and cobbles occurred throughout the deposit. A gleyed, poorly drained, but still strongly expressed Sangamon Soil was present in the top of this alluvium, continuous with the better drained Sangamon Soil on the Late Sangamon erosion surface. There was evidence of multiple soils developed within the alluvial fill, each with A and/or cambic B (Agb/Bgb) horizons. The soil and stratigraphic sequence of this buried sidevalley are similar to those of low-order drainages on the modern land surface, and illustrate one of the facies that the Sangamon Soil may have.

Franklin Soil Developed in Alluvium Buried by the Aurora Till Member

A weakly expressed example of the Franklin Soil (Hallberg, 1980b) is developed in the fluvial deposits locally incised into the Winthrop Till Member of the Wolf Creek Formation (Fig. 4). At Conklin Quarry, horizon sequences of this paleosol have varied from O-A-C to A-C to A-Bwg-C. Only incipient soil development has taken place. Even in the most morphologically well expressed profile (A-Bwg-C), the cambic B horizon was gleyed and had only weak to moderate grade soil structure. Locally, a compressed peat or muck (O horizon) capped the paleosolum. Little well preserved pollen was present, but a few spruce, chenopod, and grass pollen grains were identified. The Franklin Soil would appear to be a first- or lower-order interstadial soil in the terms defined by Follmer (1983).

Although the equivalent soil on higher landscape positions in the Winthrop diamicton has been truncated, a full set of associated weathering zones has been preserved. Successively from the truncated upper surface, these include the MOJU, MRJU, UJU, and UU weathering zones (Figs. 4 and 2). As much as the upper 5 m of the Winthrop Till Member have been oxidized, and jointing extends down into the upper part of the unoxidized weathering zone.

Evaluating the relative importance of stratigraphic breaks represented by soils in the pre-Illinoian sequence is not an easy task. Most soil profiles buried by pre-Illinoian glacial diamicton are developed in alluvium in paleovalley settings. Direct comparison of their horizon sequence with 'Quaternary standards' such as the Sangamon or Modern soil, developed on relatively stable upland landforms, is probably not a valid approach for determining relative rank (interglacial, first-order interstadial, etc.). Alluvial environments contain a host of time-independent facies that strongly affect the morphology of soils developed in them. In addition, weathering zones, especially the oxidized and leached zones, are usually developed to much shallower depths in valley environments. These considerations cast doubt on relative rankings based on direct comparisons of soil profiles and weathering zones developed in upland and valley positions.

The Franklin Soil in Conklin Quarry exemplifies some of these complexities. Based solely on the preserved soil profile (developed in alluvium) a first- or lower-order interstadial rank is suggested. On the other hand, a 5 m thick truncated weathering profile (associated with a valley wall or upland) located lateral to the Franklin Soil suggests significant landscape development and weathering of the Winthrop Till Member prior to the glaciation during which the Aurora Till Member was deposited.

Westburg Soil Developed in Alluvium Buried by Winthrop Till Member Diamicton

The southern margin of a buried paleovalley cut into Alburnett Formation diamicton is exposed near the base of the section (Fig. 4). The Westburg Soil is a cumulic, poorly drained, morphologically well expressed soil developed in the upper, fine-grained portion of the alluvial fill (Fig. 2). The soil exhibits catenary relationships from the footslope/toeslope position on the southern part of the exposure to the floodplain position farther to the north, and has Bt and Btg horizons with continuous clay coatings and strong soil structure. The strong structure may be related, in part, to the fining-upward sedimentology and to later compaction by overriding glaciers. The contact with diamicton of the overlying Winthrop Till Member of the Wolf Creek Formation is abrupt. Locally the Westburg Soil has been truncated, and in some places it has been deformed into massive sheared and slickensided blocks. Small block inclusions of the paleosol are locally found in the basal part of the overlying diamicton.

The lateral equivalent of this soil on higher landscape positions, developed in Alburnett Formation diamicton, was truncated by glacial erosion prior to deposition of Wolf Creek Formation tills. Even so, the deep weathering profile (MOL, MOJU2, MOJU, UJU, and UU) preserved beneath the truncated valley wall on the south end of the exposure, and the relatively thick, well-expressed O-A-Bt-C horizon sequence

of the preserved paleosol in alluvium indicate the Westburg Soil is of interglacial rank.

In previous exposures, a thin, compressed O horizon (peat and mucky silty clay loam) locally filled small channels cut into the Westburg Soil. Little pollen was preserved in the peat, but fragments of spruce wood were present. The spruce probably grew as climate became colder prior to the glacial advance associated with the overlying Winthrop Till Member of the Wolf Creek Formation.

Peaty O horizons are commonly associated with soils developed in alluvium and buried by glacial tills in the Upper Midwest. Invariably, pollen spectra from the peats are dominated by taxa indicating boreal conditions. These O horizons probably developed in proglacial settings during advance of the glacier that deposited the overlying diamicton. If this is the case, the O horizons and associated paleoecologic record do not necessarily reflect the range of environmental conditions under which the soil profile beneath the O horizon developed.

Discussion

Paleosols in the Quaternary sequence at Conklin Quarry are typical of those encountered in till sequences throughout the Upper Midwest. Traditional interpretations of the conditions accompanying development of these soils are fraught with problems produced by burial diagenesis, pedogenic (and ecologic) overprinting, and glacial erosion. We suggest a more trustworthy approach to interpretation of the climatic conditions accompanying development of soils in these sequences, and the duration of stratigraphic breaks marked by paleosols is to compare weathering zone sequences and properties.

Weathering zones, or parts of weathering zones are usually preserved in till sequences, they have a logical depth progression of properties related to a former landsurface, and vary relative to the drainage conditions of the associated landsurface (see Hallberg et al., 1978a). Oxidized weathering zones extend downward as the local and regional drainage network evolves. Since a regional drainage network progressively develops as time passes following a glaciation, the thickness and character of the oxidized portion of the weathering profile beneath paleo-upland positions should provide us with a reasonable way to compare the relative rank (interglacial vs. interstadial) of breaks in the glacial record.

TABLE 20-1. Summary of clay mineralogy data for the Wolf Creek and Alburnett formations of eastern Iowa.

Clay Mineralogy - % ¹			
	Ex. mean $\pm$ s.d.	Ill. mean $\pm$ s.d.	K + C mean $\pm$ s.d.
WOLF CREEK FORMATION			
N = 476	61 $\pm$ 4	18 $\pm$ 3	22 $\pm$ 3
ALBURNETT FORMATION			
N = 207	43 $\pm$ 6	25 $\pm$ 4	32 $\pm$ 4

¹Ex. — expandable clays (smectite group); Ill. — illite; K + C — kaolinite and chlorite

TABLE 20-2. Description of buried Sangamon Soil on crest of paleo-interfluvial at Conklin Quarry.
Description by G.R. Hallberg, S. Esling, and T. Bicki.

Depth meters	Horizon or Zone	Description
PEORIA LOESS		
0 - 0.25	MOU	Mottled yellowish brown (10YR5/4) silt loam; common fine mottles; massive; clear, smooth lower boundary; few snail shells.
PISGAH FORMATION Farmdale Soil		
0.25 - 0.37	Eb	Brown (10YR5-4/3) silt loam (more sand than above); common faint yellowish brown (10YR5/4-6) mottles, few fine light gray (10YR7/1) grainy silt coats (silans); weak thin platy, breaking to weak very fine subangular blocky; common fine root tubules; friable; few fine charcoal flecks; clear, smooth lower boundary.
WOLF CREEK FORMATION Late Sangamon sediment Sangamon Soil		
0.37 - 0.50	2Eb	Mixed reddish yellow (7.5YR6-5/6) to yellowish brown (10YR6-5/6) loam; moderate thin platy breaking to moderate very fine subangular blocky; common moderate silt coats (as above), common thin strong brown (7.5YR4-5/6) coatings on plates and tubules; few thin discontinuous clay films; very friable; few fine charcoal flecks; clear, smooth lower boundary.
0.50 - 0.59	2B11b	Strong brown (7.5YR5/6) clay loam; common fine dark brown (7.5YR4-4/6) mottles and coatings; moderate fine subangular blocky; thin to medium discontinuous clay films, few fine silt coats; firm; few fine charcoal flecks; gradual lower boundary.
0.59 - 0.70	2B12b	Dark brown (7.5YR4/4) clay; common fine strong brown (7.5YR5/6), yellowish red (5YR5/6), and yellowish brown (10YR5/4-6) mottles; strong fine subangular blocky; thin nearly continuous clay films, continuous moderate films on vertical tubules; few light gray (10YR7/1) and yellowish brown (10YR5/6) silt coats; firm; few charcoal flecks; gradual boundary.
0.70 - 0.76	2B13b	Stone line (gravelly clay loam) at contact between units.
Aurora (or Hickory Hills) Till Member Sangamon Soil (continued)		
0.76 - 0.89	3B14b	Mottled dark brown (7.5YR4/4), strong brown (7.5YR5/6), yellowish red (5YR4/6), dark yellowish brown (10YR4/4) clay with some pebbles; few fine red (2.5YR4/6 and 3/6) mottles; very strong, fine subangular to angular blocky structure; continuous thick clay films; very firm; gradual lower boundary.
0.89 - 1.11	3B15b	As above; with few, medium, dark gray and gray mottles and coatings; slickensided pressure faces on thick clay coatings along vertical cleavage planes; abrupt lower boundary.
1.11 - 1.18	3B16b	Dark yellowish brown (10YR4-5/4) and light olive brown (2.5Y5-4/4) heavy clay loam with some pebbles; common fine and medium strong brown (7.5YR4/6 and 5/6-8) and few medium yellowish red (5YR4/6) mottles; strong medium and fine subangular to angular blocky structure; nearly continuous moderate clay films; some peds with bare interior; firm; common pressure faces on vertical cleavage planes; gradual lower boundary.
1.18 - 1.40	3B17b	Light olive brown (2.5Y5/4) clay loam with some pebbles; common brown and red mottles as above; strong medium subangular blocky, breaking to moderate fine subangular blocky; nearly continuous thin clay films on medium peds, common thin coatings on fine peds; firm; clear, wavy lower boundary.
1.40 - 1.69	3B18b	Yellowish brown (10YR5/6) clay loam with some pebbles; common strong brown (7.5YR5/6) mottles and coatings; common, coarse, thick, very dark gray (5YR3/1) manganese-oxide and clay coatings on larger structural units; strong medium, breaking to weak fine subangular blocky structure; common thin clay films; clear, wavy lower boundary.
1.69 - 2.04	3B19b	Yellowish brown (10YR5/6) heavy loam with pebbles; common manganese-oxide coatings as above; moderate medium to coarse subangular blocky; common thin clay films; diffuse, irregular boundary.
2.04 - 2.50	3C-MOJL	Dark yellowish brown (10YR4/4) loam with pebbles; common yellowish brown (10YR5/6-8), brown (10YR5/3), strong brown (7.5YR4-5/6), and pale olive (5Y6/4) mottles; moderate coarse angular blocky to massive; firm; few manganese-oxide and clay coatings; brown mottles along vertical joints.

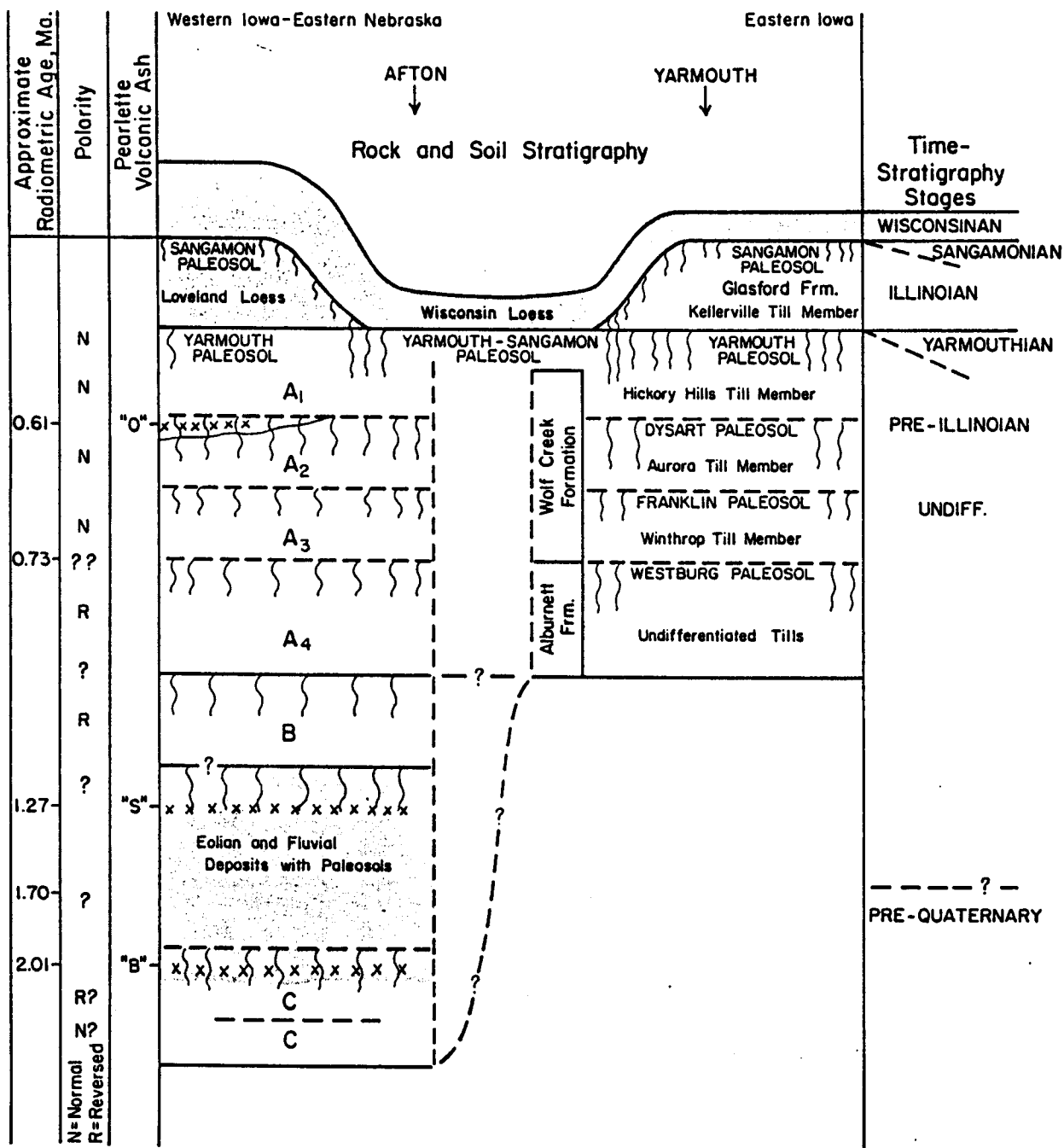


FIGURE 20-1. Correlations between eastern and western Iowa Quaternary deposits showing approximate radiometric age, magnetic polarity, and position of volcanic ashes in the stratigraphic sequence.

STRATIGRAPHY

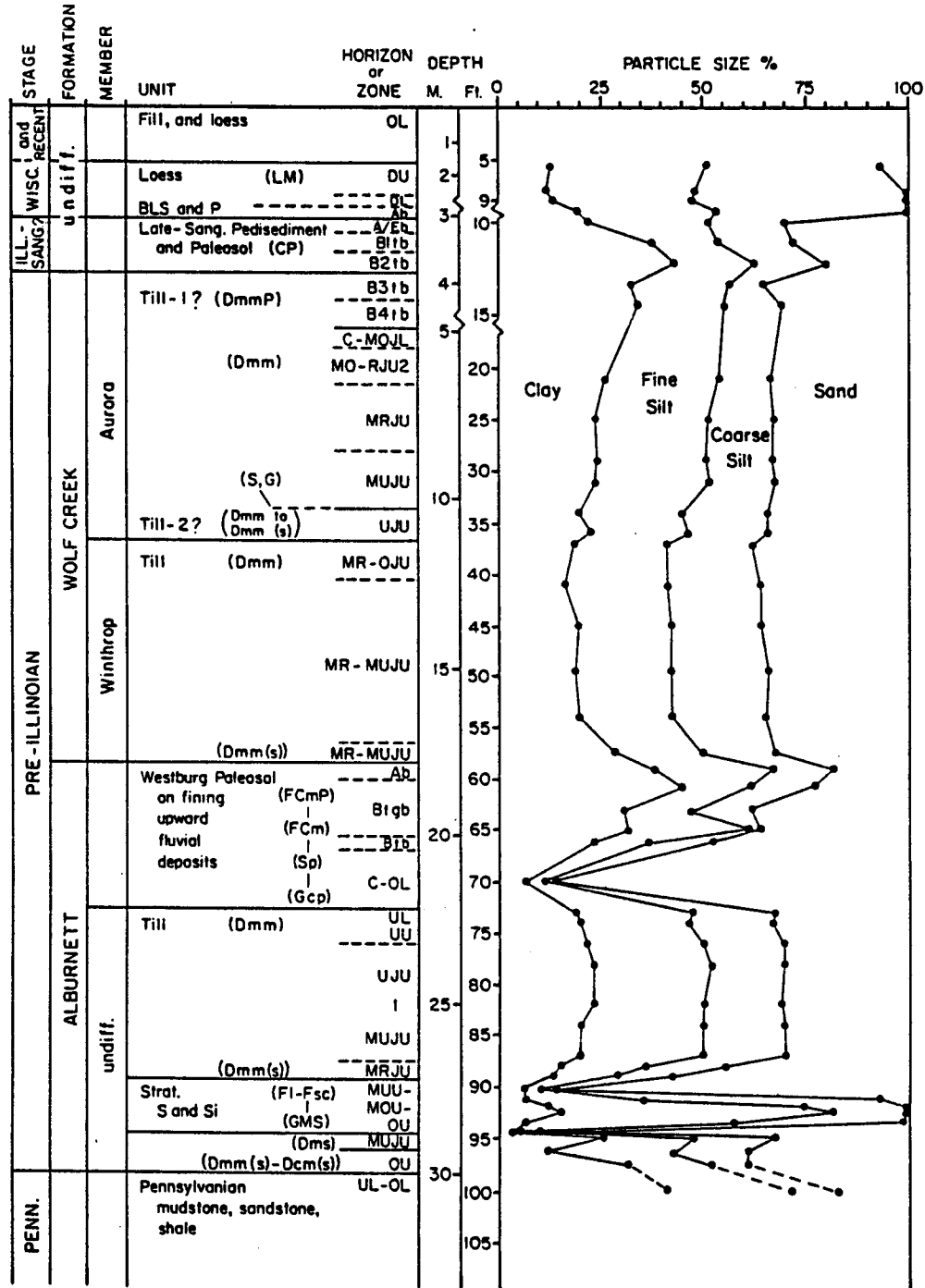
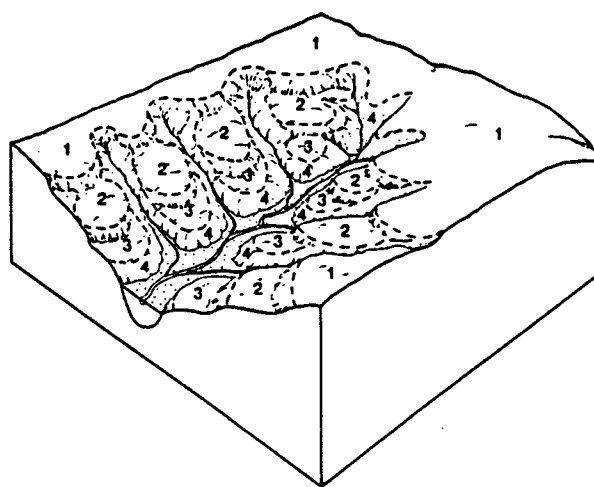


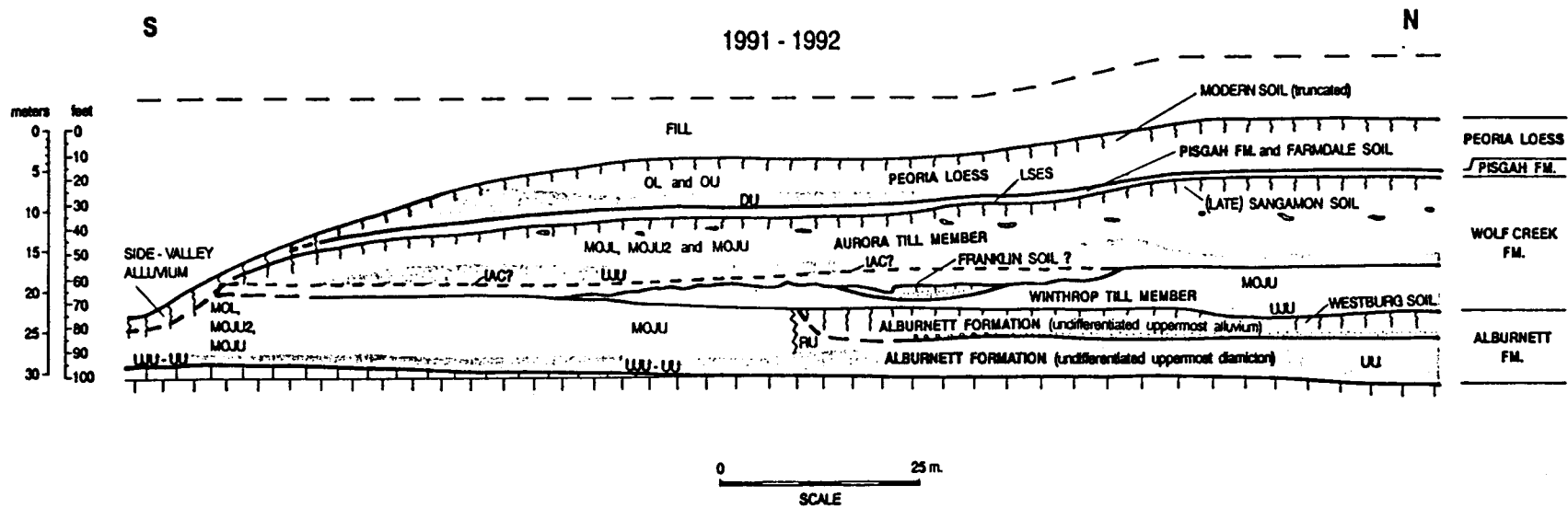
FIGURE 20-2. Stratigraphy and particle-size data for the 1983-84 exposure at Conklin Quarry.



LEGEND

- | | |
|--|--|
| 1 | Loess-mantled Yarmouth-Sangamon surface and paleosol |
| 2 | Loess-mantled Late Sangamon erosion surface and paleosol |
| 3 | Loess-mantled Wisconsinan age ("lowan") erosion surface |
| 4 | Location of major Holocene age erosion |
| <div style="display: inline-block; width: 10px; height: 10px; background: repeating-linear-gradient(45deg, transparent, transparent 2px, black 2px, black 4px); border: 1px solid black;"></div> | Location of major Holocene age deposition |

FIGURE 20-3. Schematic diagram of geomorphic surfaces in the Southern Iowa Drift Plain of eastern Iowa.



APPENDIX: WEATHERING ZONE TERMINOLOGY

Standard abbreviations and symbols are used to describe weathering zones below the soil solum (O, A, E, and B horizons). These are defined by Munsell color (related to the state and distribution of free-iron oxides and organic matter) and the presence or absence of matrix-carbonate minerals (see Hallberg et al., 1978a for a complete discussion). In brief:

First symbol: color reference

O	oxidized
D	deoxidized
R	reduced
U	unoxidized

Second symbol: leached or unleached state

U	unleached
L	leached; no carbonates detectable (with dilute HCl)
L2	leached; primary carbonates absent, secondary carbonates present
U2	matrix carbonates unleached, secondary carbonates present

Modifier symbols:

M	mottled; zones containing 20-50% contrasting mottles; when used with unoxidized zone designation it infers 20% or less mottles of reduced colors; precedes first symbol when used.
J	jointed; describes the presence of well defined vertical joints; joints often show oxidized and reduced colors and often have coatings or rinds of secondary iron-oxides and occasionally other secondary minerals such as calcite or gypsum; second symbol when used.

Examples of descriptions using weathering zone terminology:

OL	oxidized, leached; yellowish-brown (10YR 5/6) or strong brown matrix color; leached.
MDU	mottled, deoxidized, and unleached; grayish brown (2.5Y 5/2) matrix color with strong brown (7.5YR 5/8) mottles and pipestems, unleached.
RJU	reduced, jointed, and unleached; mixed olive (5Y 4/4 and 5Y 4/3) and very dark grayish brown (2.5Y 3/2) matrix color with common gray (5Y 5/1) and light olive brown (2.5Y 5/4) mottles; prominent vertical joints with 1-cm thick strong brown (7.5YR 5/8) segregations along joints; unleached.
UJU	unoxidized, jointed, unleached; uniform dark greenish gray (5GY 4/1) matrix color with few thin vertical joints which have mottled light olive brown (2.5Y 5/6) and olive gray (5Y 5/2) faces and a 3-cm thick rind of greenish gray (5GY 5/1); unleached.

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