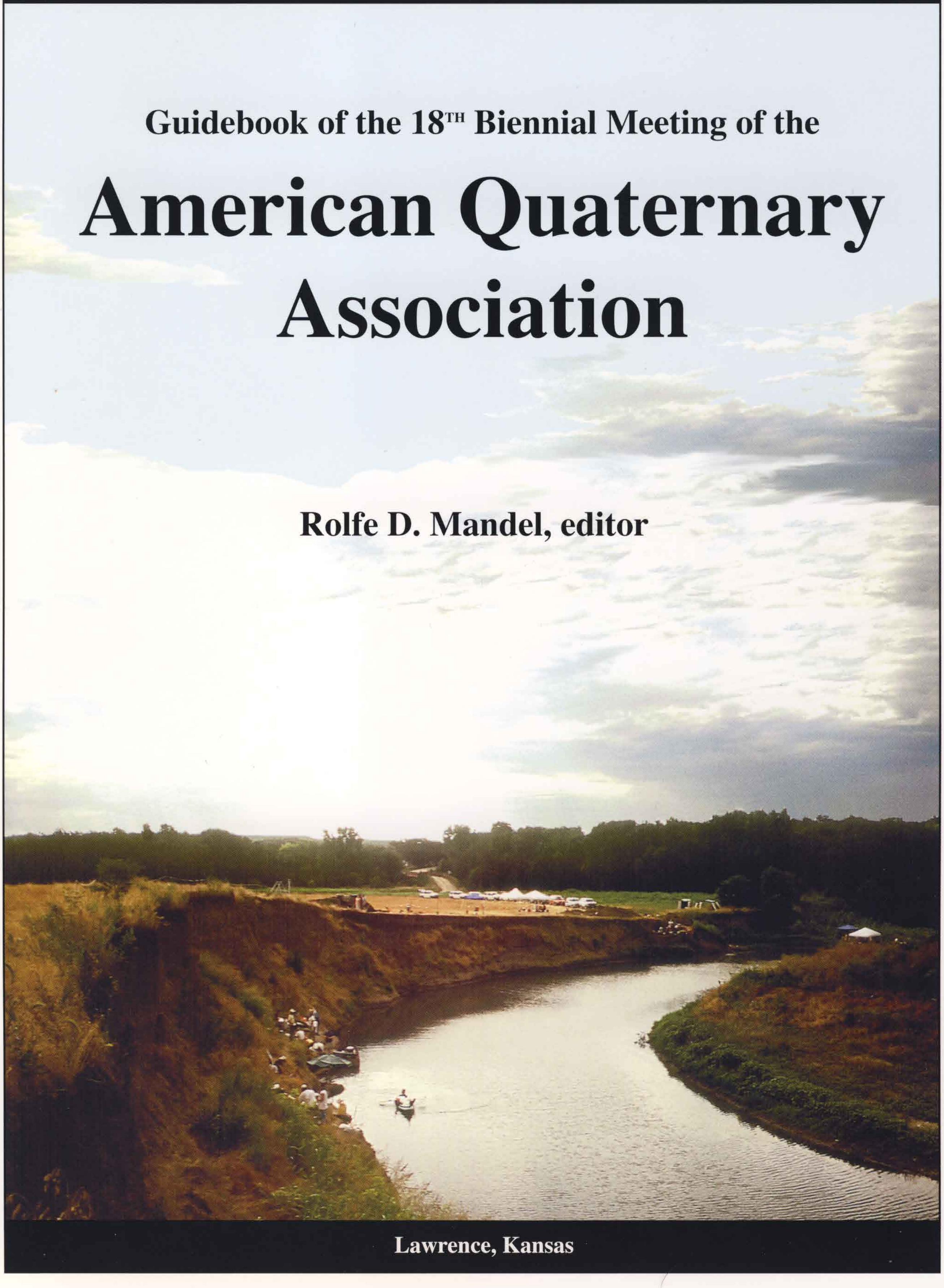
Kansas Geological Survey

Technical Series 21

Guidebook of the 18[TH] Biennial Meeting of the

American Quaternary Association

Rolfe D. Mandel, editor



Lawrence, Kansas

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Rolfe D. Mandel, editor

Lawrence, Kansas 66047
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Preface and Acknowledgments

In June 2004, the University of Kansas hosted the 18th Biennial Meeting of the American Quaternary Association (AMQUA). In addition to a technical session, two pre-meeting and post-meeting field trips were held. Although a guidebook was prepared for distribution at the conference, plans were made to revise it after the trips were completed. Specifically, the authors wanted to include feedback from field-trip participants. Also, by publishing the guidebook in the Kansas Geological Survey's Technical Series, peer review of the manuscripts was ensured, and the volume would gain a broader audience. The resulting guidebook is much more than descriptive narratives for localities along routes; it provides a wealth of new information about the glacial history, loess record, archeology, geoarcheology, and late Quaternary paleoecology of the midcontinent of the United States.

Field Trip 1 focuses on the stratigraphy, paleopedology, lithology, and chronology of Last Glacial loess in eastern Nebraska and western Iowa. Joseph Mason and his colleagues shed new light on the history of loess deposition in areas bordering the Missouri River valley. They use loess thickness and grain-size trends to identify sediment sources, and new data are presented for the Loveland Loess type locality as well as other loess sections.

Field Trip 2 focuses on late Quaternary biogeography and paleoecology along the prairie-forest border in the eastern Plains. The dynamics of prairie vegetation (including the effects of fire and bison) and soils are considered at the Konza Prairie Biological Station in northeastern Kansas. In addition, paleoecological evidence (i.e., pollen, phytolith, and plant macofossil records) indicating late Quaternary bioclimatic change is presented and examined in the South Fork of the Big Nemaha River in southeastern Nebraska.

Field Trip 3 focuses on pre-Illinoian tills and associated sediments and paleosols in northeastern Kansas and central Missouri. Wakefield Dort presents evidence of glaciation and considers the location of the terminal glacial boundary in Kansas. Charles Rovey describes the lithostratigraphy of pre-Illinoian glaciogenic sediments in north-central Missouri and presents the results of recent paleomagnetic studies used to estimate the age of pre-Illinoian tills in north-central Missouri.

Field Trip 4 focuses on late Quaternary alluvial stratigraphy and geoarcheology in the central Great Plains of Kansas and northern Oklahoma. Four early Paleoindian sites—Simshauser, Waugh, Cooper, and Jake Bluff—are included in this trip. Also, the Winger site, a late Paleoindian bison bonebed in southwestern Kansas, and the Claussen site, a stratified Paleoarchaic campsite in northeastern Kansas, are described.

As already noted, all manuscripts contained in this volume were peer-reviewed. A special thanks is extended to these reviewers: Art Bettis, Jeremy Dillon, Jeff Dorale, Glen Fredlund, Andrea Freeman, Tim Kemmis, Lee Nordt, and J. Elmo Rawling. They performed an important duty rarely applied to guidebooks.

I want to thank Rex Buchanan, Associate Director of the Kansas Geological Survey (KGS), for agreeing to publish the 2004 AMQUA field-trip guidebook as a volume in the KGS Technical Series. Other staff members of the KGS also merit recognition. Special thanks go to Marla Adkins–Heljeson for copy-editing and formatting the guidebook, and to Jennifer Sims for preparing many of the illustrations and assisting with other aspects of the guidebook's production. I am also grateful to John Charlton, who prepared many of the photographs.

Thanks also are extended to Amy Macneill and other staff members at KU Continuing Education for assisting with the organization of the field trips. Finally, I am grateful to the landowners, including Verne Claussen, Richard Farwell, Bill Simshauser, Kent and Travis Winger, and Leland Waugh, for allowing us to visit sites on their properties.

Rolfe D. Mandel
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November 2006

AMQUA Pre-Meeting Field Trip 1: Last Glacial Loess Sedimentary System of Eastern Nebraska and Western Iowa

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Introduction

Eolian sediments cover vast areas of the North American midcontinent (fig. 1–1) and contain a record of cratonic sedimentation spanning the Cenozoic Era. An increasing interest in eolian sediments and associated soils as climate proxies has spurred several recent studies of Quaternary loess and eolian sand in the region (Muhs et al., 1999; Arbogast and Muhs, 2000; Mason and Kuzila, 2000; Muhs and Bettis, 2000; Mason, 2001; Forman et al., 1992; Forman and Pierson, 2002; Loope and Swinehart, 2000; Mason, Jacobs, Hanson, et al., 2003). Loess, eolian sand, and volcanic ash deposits provide direct records of atmospheric circulation that can be used to test atmospheric general circulation models (Muhs et al., 1999; Muhs and Bettis, 2000), and buried soils within eolian sequences provide detailed records of environmental change spanning time intervals of centuries to glacial-interglacial cycles. A proliferation of studies over the last two decades have given us an appreciation of the dynamic nature of eolian landscapes, and the complex linkages among geologic, climatic, and biotic systems in the history of the United States' midcontinent's eolian systems.

On this trip, we will visit loess localities in western Iowa and eastern Nebraska (fig. 1–1). The thick loess bordering the Missouri River valley has a long history of investigation dating back to the late 1800's. Recent studies at several localities have yielded new insights on several aspects of the region's loess and eolian record, most notably the chronology and sources of the loess. Some of the data in this guidebook are derived from previous studies reported in Bettis, Mason, et al. (2003), and some are presented here for the first time.

Stratigraphic Framework

A detailed treatment of the trip area's Cenozoic stratigraphy is beyond the scope of this guidebook. Presented here are salient elements of the region's eolian stratigraphy that relate to localities visited on this trip. Extensive wind-deposited tuffaceous silts and sands with interbedded tuff occur in the wedge of Paleogene and Neogene sediment extending eastward from the Rocky Mountain Front Range across the western Great Plains. These sediments represent a range of eolian depositional environments that formed as sediment was shed from the rising Rocky Mountains via volcanic and fluvial processes during the Oligocene, Miocene, and Pliocene. Extensive sand sheets and thin loesses with associated paleosols occur in the Pliocene Fullerton Formation of Nebraska (Reed and Dreeszen, 1965; May et al., 1995) and its correlative, the upper part of the Blanco formation in Kansas (Frye and Leonard, 1952; Holliday, 1988). Yellowstone caldera-source Pearlette Family volcanic ashes provide three timelines for Great Plains stratigraphy: the late Pliocene (2.059 Ma – Huckleberry Ridge ash), early Pleistocene (1.285 Ma – Mesa Falls ash), and the middle Pleistocene (0.639 Ma – Lava Creek B ash) (Boellstorff, 1978; Izett and Wilcox, 1982; Lanphere et al., 2002). As yet poorly understood loess-paleosol sequences occur below the Lava Creek B ash in the Pleistocene Walnut Creek and Sappa formations of Nebraska and Kansas and above the Lava Creek B ash, but below Illinoian-age (marine oxygen isotope stage 6) Loveland Loess in the region (Reed and Dreeszen, 1965; May et al., 1995).

Four middle to late Pleistocene loess units occur in the eastern Plains. From oldest to youngest these are Loveland Loess, Gilman Canyon Formation (equivalent to Iowa's Pisgah Formation), Peoria Loess, and Bignell Loess. Loveland Loess is usually only a few meters thick (except near the Missouri River valley) and sometimes has an eolian sand facies. The last interglacial Sangamon Geosol is developed in the upper part of the Loveland. This paleosol is relatively thick (1.5–2.5 m), exhibits 7.5YR hue colors, and often contains clay coatings in the upper B horizon and carbonate coatings or nodules in the lower part of its profile. The Gilman Canyon Formation is

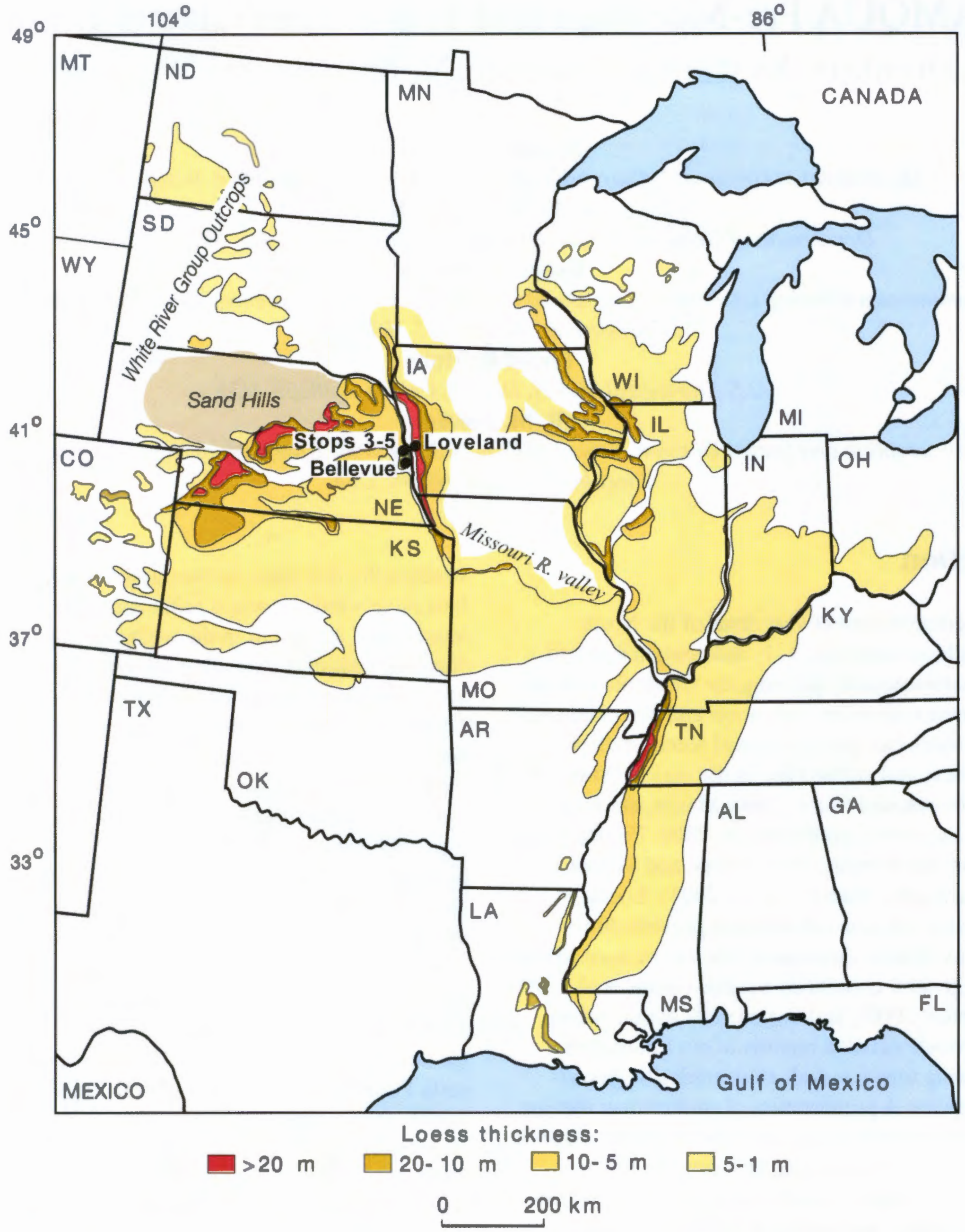

FIGURE 1–1. Map of the midcontinental United States showing thickness of Peoria Loess and field-trip stops. Adapted from Bettis, Muhs, et al. (2003) and sources cited therein.

also thin (usually <2 m) and usually modified throughout by organic accumulation and biological structures (burrows, etc.). The soil developed in the Gilman Canyon Formation (equivalent to the Farmdale Geosol east of the Missouri River valley) is usually welded to the underlying Sangamon Geosol developed in the Loveland Loess. In rare sections where the Gilman Canyon Formation is thick, two or more soils are developed in the unit. Peoria Loess buries the Gilman Canyon Formation. The Peoria is the thickest (up to 50 m) and one of the most aerially extensive last glacial loess deposits in the world (Muhs and Bettis, 2003). At some near-source localities eolian sand is interstratified with the loess and bedding structures are present. On stable upland summits in the western Great Plains of Nebraska and Kansas, the Bignell Loess buries Peoria Loess and the dark-colored Brady Soil is developed in the upper part of the Peoria. The Bignell is patchy in its distribution and appears not to occur as a once-continuous sheet as most of the other loess units in the region do. The Bignell is not present (or at least not discernable) in eastern Nebraska or anywhere east of the Missouri River valley, where the modern soil is developed in Peoria Loess.

Recent applications of luminescence dating to loess in the region are refining the region's loess chronology and have revealed the age structure of individual loess units at several localities. Thermoluminescence (TL) ages from Eustis in south-central Nebraska and TL and infrared stimulated luminescence (IRSL) ages from the Loveland paratype locality in western Iowa (Stop 2 on this trip) suggest that the Loveland Loess accumulated during the penultimate glacial period, equivalent to marine oxygen isotope stage 6 (Forman et al., 1992; Maat and Johnson, 1996; Forman and Pierson, 2002). Luminescence and radiocarbon ages on soil organic matter and charcoal bracket accumulation of the overlying Gilman Canyon Formation and its Iowa equivalent, the Pisgah Formation, to between ~45,000 and 25,000 cal yr B.P. (Forman et al., 1992; Martin, 1993; May and Holen, 1993; Pye et al., 1995; Maat and Johnson, 1996; Muhs et al., 1999; Forman and Pierson, 2002; Bettis, Muhs, et al., 2003). Thus, in the eastern and central Great Plains, the Sangamon paleogeomorphic surface, marked by the Sangamon Geosol, may have developed over a time span that corresponds to all of oxygen isotope stages 5 and 4. A large body of radiocarbon and luminescence ages—TL, IRSL, and optically stimulated luminescence (OSL)—indicates that Peoria Loess began to accumulate near its source areas around 23 ka and continued to accumulate across the region until about 12 ka (Muhs et al., 1999; Forman and Pierson, 2002; Roberts et al., 2003; Bettis, Muhs, et al., 2003). These ages indicate that Peoria Loess accumulated during the late Wisconsin glacial period (oxygen isotope stage 2). Suites of OSL ages through the Peoria at several sections demonstrate that the rate of accumulation varied through time in a systematic fashion across the region (see stops 2–4 on this trip and Roberts et al., 2003).

Maximum limiting ages of the Bignell Loess are based on radiocarbon ages of soil organic matter from the Brady Soil developed in the upper part of the Peoria Loess. These ages in combination with other radiocarbon and luminescence ages within the Bignell Loess indicate that the unit accumulated episodically through the entire Holocene (Martin, 1993; Pye et al., 1995; Maat and Johnson, 1996; Muhs et al., 1999; Mason and Kuzila, 2000; Mason, Jacobs, Hanson, et al., 2003).

Studies of provenance tracers (zircon ages, Pb isotopes in K-feldspar, and loess geochemistry) as well as thickness trends have demonstrated that the Peoria Loess is derived from several sources, some glaciogenic in origin and others nonglaciogenic (Aleinikoff et al., 1999; Muhs and Bettis, 2000; Mason, 2001; Bettis, Muhs, et al., 2003). Eocene and Oligocene volcaniclastic siltstones of the White River Group (and possibly the Arikaree Group) were a significant nonglaciogenic source of the Peoria Loess, while the Platte and Missouri River valleys were the major glaciogenic sources. Although large dune fields, such as the Nebraska Sand Hills, were the area of last entrainment for much of the loess in Nebraska, Pb isotopic composition and the ages of zircons in the loess indicate that most of the silt was ultimately derived from bedrock sources. Nevertheless, eolian sand was an essential component of the regional loess sedimentary system as it served as the tool in ballistic impacts that brought silt and clay particles into suspension in the atmosphere.

Thickness patterns and grain-size and geochemical trends indicate loess- and sand-transporting winds were dominantly from the northwest or west across the region (Smith, 1942; Hutton, 1947; Snowden and Priddy, 1968; Ruhe, 1983; Mason et al., 1994; Muhs et al., 1999; Muhs and Bettis, 2000; Mason, 2001). On this trip we will examine thick deposits of last glacial loess that are located immediately downwind (east) and upwind (west) of the Missouri and Platte River valleys. As we will see, not all of the loess in these sections was derived from these valley sources.

STOP 1. Maas Drive Section, Bellevue, Nebraska

This locality is approximately 7 km north of the Platte River valley and about 5.5 km west of the Missouri River valley (fig. 1–1). Last-glacial loesses, the Sangamon Geosol developed in Loveland Loess, and a sequence of older loesses and paleosols are present here (fig. 1– 2a). Peoria Loess is thicker at localities like Bellevue, along the east side of the Missouri River valley, than to the immediate west (Mason, 2001; Bettis, Muhs, et al., 2003). Mason (2001) estimated that the Missouri and Platte valleys may have contributed from 25% to no more than 50% of the loess deposited at localities in eastern Nebraska (this issue is discussed in more detail at Stop 3). Missouri River valley-source loess west of the valley in Nebraska indicates that although loess-transporting winds were primarily westerlies, easterly or northeasterly winds did at times carry significant quantities of dust (Mason, 2001).

The geochemistry of the loesses and soils in the section provide information pertinent to potential loess sources and post-depositional weathering (fig. 1–2b). Bulk Fe_2O_3 (i.e., ALL Fe) is correlated highly with clay content because of the smectite-dominated clays in loess for this part of the continent (Muhs and Bettis, 2000; Muhs et al., 2001). The higher clay content in the Bt horizon of the modern soil stands out, and the Peoria Loess has greater clay content here than Peoria Loess at Loveland (Stop 2). Because Bellevue is close to the Missouri River, one might expect a clay content that is relatively low, as at Loveland. The higher clay content, reflected in bulk Fe_2O_3, suggests a mix of sources from both the nearby Missouri River and more distal sources to the west (this will be discussed further at stops 2 and 3).

Titanium (Ti) and niobium (Nb) are elements found in resistant, detrital minerals such as ilmenite (basic igneous rocks), rutile (igneous/metamorphic rocks), anatase (metamorphic rocks), and titanomagnetite (igneous/metamorphic rocks). As such, the Ti/Nb ratio changes little or not at all even with considerable chemical weathering and is therefore a useful provenance indicator. Ti/Nb

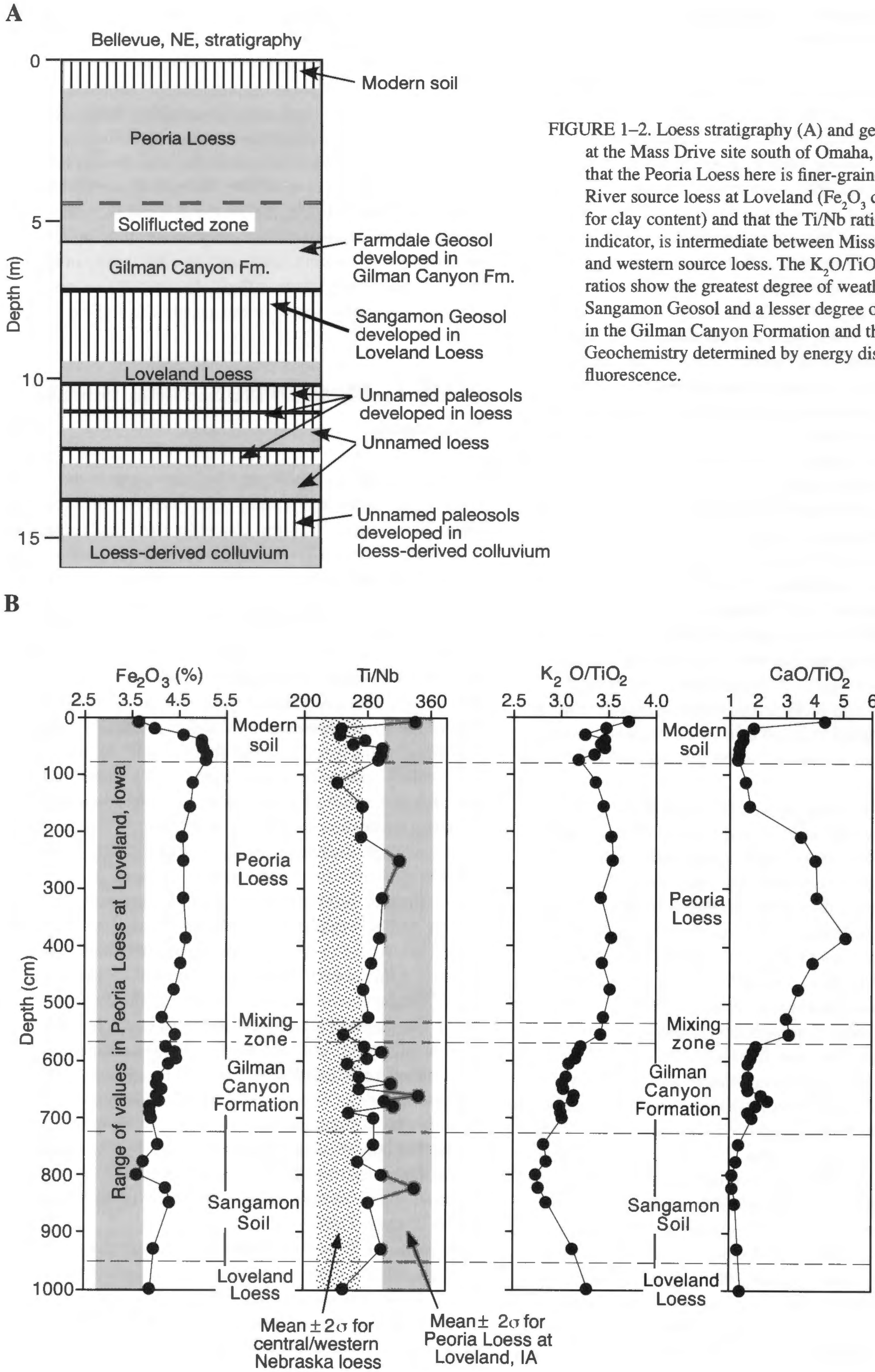

FIGURE 1–2. Loess stratigraphy (A) and geochemistry (B) at the Mass Drive site south of Omaha, Nebraska. Note that the Peoria Loess here is finer-grained than Missouri River source loess at Loveland (Fe_2O_3 content is a proxy for clay content) and that the Ti/Nb ratio, a provenance indicator, is intermediate between Missouri River source and western source loess. The K_2O/TiO_2 and CaO/TiO_2 ratios show the greatest degree of weathering in the Sangamon Geosol and a lesser degree of weathering in the Gilman Canyon Formation and the modern soil. Geochemistry determined by energy dispersive X-ray fluorescence.

values in Peoria Loess at Bellevue show values that are intermediate between the values found in Peoria Loess at Loveland, Iowa (Missouri-dominated source), and Peoria Loess found in central and western Nebraska (western source). Thus, we interpret Peoria Loess at Bellevue to be derived from a combination of an immediate, eastern source (the Missouri River) and the distal facies of loess found in central and western Nebraska. If this interpretation is correct, it explains why Peoria Loess at Bellevue has somewhat higher clay contents (represented by Fe_2O_3) than at Loveland.

Potassium (K) is found mostly in two common loess minerals, K-feldspar (microcline and orthoclase) and mica (biotite, muscovite, and illite). K-feldspar and muscovite are relatively resistant to chemical weathering (although not invulnerable!), so K_2O losses, relative to immobile TiO_2, probably reflect chemical weathering of biotite and illite. Note that zones within the modern soil are depleted in K_2O relative to underlying unweathered Peoria Loess. The Gilman Canyon Formation, which consists mostly of pedogenically altered loess, also has significant K_2O loss. However, the greatest loss is evident in the Sangamon Soil. The greater degree of chemical weathering in the Sangamon Soil compared to either the Gilman Canyon Formation (Farmdale Soil) or the modern soil is consistent with its better-expressed morphology and greater thickness. Is the greater chemical weathering and degree of soil development in the Sangamon Soil (compared to the modern soil) due to a more humid and perhaps warmer paleoclimate or a longer period of pedogenesis (e.g., all of the equivalent of marine isotope stage 5, not just substage 5e) or both? Ruhe (1969) struggled with this question for Sangamon Soils in adjacent Iowa and thought it most likely that both paleoclimate and time were factors.

The ratio CaO/TiO_2 provides another chemical weathering index. This ratio is a proxy for the degree of carbonate mineral (calcite and dolomite) depletion. Note that the ratio is high in the modern soil A horizon (as with K_2O/TiO_2), which is probably a function of biocycling. Deeper in the modern soil profile, however, carbonates are absent. Even in the upper part of Peoria Loess, carbonate content is low, suggesting that loess-deposition rates were low during the final phases of sedimentation. Between 3 and 4 m depth, the CaO/TiO_2 value is relatively high, suggesting rapid loess deposition with little syndepositional leaching of carbonates. The lower part of the Peoria Loess appears to be lower in carbonate content but not as low as during the waning stages of loess deposition. Few or no carbonates are present in the Gilman Canyon Formation, suggesting low rates of loess sedimentation and/or a relatively humid paleoclimate. As with K_2O/TiO_2 values, however, the greatest degree of carbonate depletion is found in the Sangamon Soil. Carbonates are depleted significantly throughout this thick paleosol, supporting the interpretations of a humid last-interglacial paleoclimate, a long period of pedogenesis (perhaps all of marine isotope stage 5), or both.

The lowest 1–1.5 m of the Peoria is distinctly different from the loess above it. This basal zone has more mottles, contains streaks of deoxidized (light-gray) silt and secondary accumulations of manganese oxide, exhibits coarse horizontal parting, and has prominent wavy "streaks" of organic-rich silt (including charcoal) that are sometimes overturned into recumbent folds. AMS radiocarbon ages on humic acids extracted from organic matter in three of the organic-rich zones yielded ages of 22,040 ± 160 ^{14}C yrs B.P. (CAMS–10181; 32 cm above base of Peoria), 20,810 ± 160 ^{14}C yrs B.P. (CAMS–10187; 36 cm above base of Peoria), and 22,210 ± 170 ^{14}C yrs B.P. (CAMS–10188; 38 cm above base of Peoria). These ages are out of stratigraphic sequence and older than would be expected from the base of the Peoria Loess. This can be explained by considering the genesis of the basal part of the Peoria. Several investigators have noted features attributable to periglacial activity in the upper midcontinent (Péwé, 1983; Muhs and Bettis, 2003), and paleoecological studies point to the period from about 21,000 to 16,500 ^{14}C yr B.P. as one that brought arctic fauna and flora into Iowa, Illinois, and eastern Nebraska (Baker et al., 1986; Schwert et al., 1997). The wavy topography and discontinuous nature of the organic-rich beds and the presence of overturned folds and horizontal partings in the basal Peoria sediments suggest that solifluction played a role in the accumulation of this zone. If the basal zone contains sediment deposited by solifluction during the initial phase of Peoria Loess accumulation, then the organic-rich zones were probably derived from the A horizons of the surface soil (Farmdale Geosol) upslope. The radiocarbon ages, therefore, may not date the time of deposition, but rather the age (mean residence time) of carbon in the soil being eroded upslope.

The Peoria abruptly overlies the Farmdale Geosol developed in the Gilman Canyon Formation (fig. 1–2A). Lithologic characteristics of the Gilman Canyon Formation here are very similar to the Pisgah Formation across the river in Iowa. The unit is modified by the Farmdale Geosol which lacks the strong biofabric common to the Gilman Canyon pedocomplex at many sites across Nebraska. Charcoal collected 10 cm below the top of the Gilman Canyon Formation at this section yielded an AMS radiocarbon age of 25,340 ± 260 ^{14}C yrs B.P. (CAMS–10190).

The Gilman Canyon Formation has a gradational boundary with the underlying Sangamon Geosol developed in Loveland Loess. Here the Sangamon is a morphologically well-expressed soil with a 2-m-thick A–AB–Bt profile, thick continuous clay films, strong structure, and strong rubifaction. The base of the Sangamon Geosol is not exposed, but a hand boring below the exposed section indicates that a sequence of older silts (loess) and paleosols are present beneath the Loveland (fig. 1–2A). The ages of these units are presently unknown, but they probably correlate with the "silty clays" discussed at Stop 3.

Between Omaha/Council Bluffs and STOP 2

The motel (Interstate Inn, Council Bluffs, Iowa) is located on a historic Missouri River channel belt (Hallberg et al., 1979). The alluvial fill of the Missouri River valley ranges from 25 to about 40 m in thickness in this area (Miller, 1964; Iowa Geological Survey file data) and consists of a thin veneer of fines (2–5 m) grading downward to medium to coarse sand with a few-meter-thick basal gravel. Near the bluff line the stratigraphy becomes much more complex as wedges of tributary alluvium and slope deposits interfinger with the main valley sediments. The valley floor east of the present river is Holocene in age, but on the west side of the river Wisconsin interstadial floodplain deposits (ca. 25.5–23 ka) are preserved beneath the Peoria Loess-mantled Fort Calhoun Terrace. In the Omaha/Council Bluffs area, the bedrock surface (Pennsylvanian) is deepest on the east side of the Missouri River valley and rises in a series of steps westward. The bedrock surface beneath the Fort Calhoun Terrace is higher in elevation than the base of the modern Missouri River channel. Several large quarries in the area extract Pennsylvanian limestone, primarily for use as aggregate.

This area was last glaciated during the pre-Illinoian, about 500 ka. Multiple glaciations, beginning about 2 Ma, affected the central North American midcontinent and deposited a series of matrix-supported glacial diamictons (Boellstorff, 1978; Roy et al., 2004). It was probably during the last one or two of these glaciations that the Missouri River valley evolved into its present position and major tributary configuration (Wayne, 1985); loess stratigraphy discussed at Stop 3 may ultimately shed more light on this issue.

During the last glacial period, the Missouri River valley drained the western margin of the Laurentide ice sheet and also was fed by rivers draining unglaciated parts of the central and northern Great Plains. The Platte River, which carried outwash from Rocky Mountain Front Range glaciers and also drained a large portion of the unglaciated central Great Plains, joins the Missouri just south of Omaha.

Bluffs along both sides of the valley are mantled with loess. Pre-Illinoian glacial diamicton generally crops out low along the bluff and is overlain by up to 50 m of late Pleistocene loess.

STOP 2: Loess Stratigraphy at Loveland, Iowa

The three most widespread loess units in the North American midcontinent (Peoria, Gilman Canyon/Pisgah, and Loveland) are present at the Loveland section in western Iowa (figs. 1–1 and 1–3). Glacial till deposited during the later part of the Matuyama Chron, ca. 1.2– 0.8 Ma ("A" till of Boellstorff, 1978; R1 till of Roy et al., 2004) underlies the loess sequence here.

This is the reference locality (paratype) for the Loveland Loess, which dates from the penultimate ("Illinoian") glaciation. Thermoluminescence (TL) and infrared stimulated luminescence (IRSL) age estimates from the Loveland here range from 165 to 125 ka, and are in good agreement with TL ages of Loveland Loess from Nebraska (Maat and Johnson, 1996; Forman et al., 1992; Forman and Pierson, 2002). The Sangamon Geosol is developed in the upper part of the Loveland Loess and represents weathering, soil development, and landscape evolution that occurred through the last interglacial and until the overlying Pisgah Formation loess began to accumulate. Charcoal 3 cm above the base of the Pisgah Formation dates 42,150 $\pm$ 2,350 ^{14}C yrs B.P. (CAMS–6359) and 44,790 $\pm$ 3,250 ^{14}C yrs B.P. (CAMS–6638) at this section. Luminescence ages (TL, IRSL, and OSL) are in general agreement and bracket accumulation of the Pisgah Formation between about 46 and 23 ka. Carbonate nodules collected from the Loveland Loess 5.5–6.5 m below the top of the Sangamon Geosol give U-series ages ranging from 57,000 $\pm$ 6,000 to 68,000 $\pm$ 8,000 yrs (table 1–1), suggesting that they formed during the latter part of marine oxygen isotope stage 4 and at the stage 4/3 boundary, rather than during stage 5 (the last interglacial). Carbon isotopic data from speleothems in Crevice Cave, near St. Louis, Missouri, indicates that the interval during which these carbonate nodules formed was one of general warming in the midcontinent that culminated between 55 and 59 ka (Dorale et al., 1998). At Loveland, the Sangamon paleogeomorphic surface, marked by the Sangamon Geosol marks a ca. 80,000-year-long depositional hiatus spanning the last interglacial period and most of the early and middle Wisconsin (oxygen isotope stages 5, 4 and the first half of 3).

The Sangamon Geosol at this locality (solum thickness <2 m, weak to moderate grade soil structure, and no clay coatings) has a much more weakly expressed morphology than the Sangamon we observed at Stop 1. The first impression is that the level of morphologic expression at Loveland is inconsistent with an extended period of weathering covering a full interglacial period. One needs to keep in mind, however, that this is a "bluff line" locality and that local relief on the Sangamon paleogeomorphic surface probably inhibited soil horizonation through runoff, surficial erosion, and reduced infiltration. A short distance east of the bluff line where upland divides are wider, the Sangamon Geosol has thick, well-expressed soil horizons with strong structure and abundant clay coatings, very similar to that observed at Stop 1. Similar patterns in soil development are seen in transects of modern soils extending from the bluff line eastward.

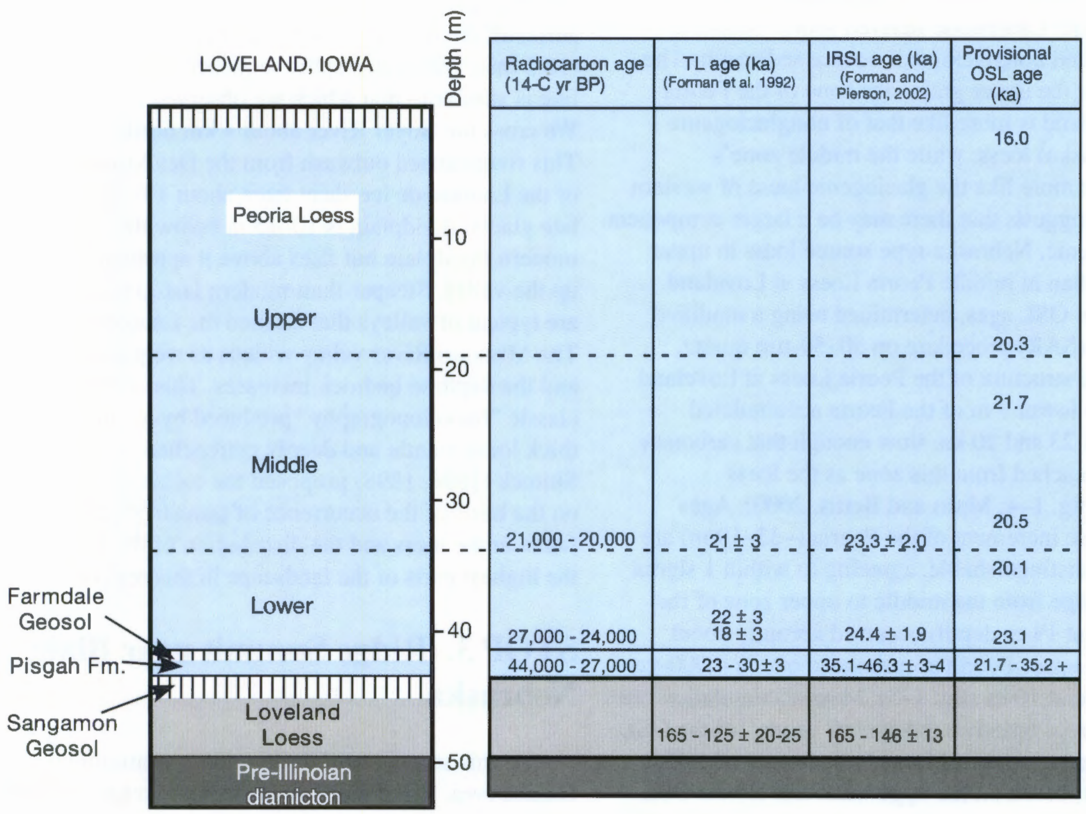

FIGURE 1–3. Stratigraphy and radiocarbon and luminescence chronology of the Loveland locality in western Iowa.

At Loveland, Pisgah Formation loess accumulated between about 46 and 23 ka (fig. 1–3) as the first Wisconsin advance of the Laurentide ice sheet into the midcontinent occurred (Bettis et al., 1996; Curry, 1998; Dorale et al., 1998). The Pisgah (stratigraphically equivalent to the Gilman Canyon Formation in Nebraska) here is exceptionally thick (4 m) and is one of the few examples of relatively unaltered isotope stage 3 loess in the region. Small pieces of *Picea* charcoal and land snails within the unit, as well as rare pollen records, indicate that this loess accumulated on a landscape occupied by vegetation communities that probably shifted from parkland to boreal forest (Baker et al., 1991; Baker et al., 2003). The Farmdale Geosol is developed in the upper 50–60 cm of the Pisgah Formation. This soil marks a significant interstadial period of very slow or no loess accumulation. Abundant pollen and macrofossil evidence indicate that the western Iowa landscape was occupied by boreal forest as this soil formed. As we saw at Stop 1, the Farmdale Geosol is often partially eroded and/or deformed, and at Loveland the Farmdale is a 50–60-cm-thick AC soil profile that is discontinuously present across the section.

During the last decade this section has been the focus of several investigations addressing the chronology, sedimentology, and geochemistry of Missouri Valley-source Peoria Loess. About 41 m of Peoria Loess occurs above the Farmdale Geosol at Loveland. The Peoria is a massive, pale-brown to light-yellowish-brown, calcareous silt loam that extends to the modern land surface. A large number of radiocarbon ages from across the region suggest that Peoria Loess began to accumulate between 23,000 and 18,000 ^{14}C yrs B.P. (Bettis, Muhs, et al., 2003). When the Peoria Loess stopped accumulating is unknown, but estimates based on a few radiocarbon ages of gastropod shells and soil organic matter range from 11,000 to 12,000 ^{14}C yrs B.P. (Wang et al., 2000; Bettis, Muhs, et al., 2003).

Grain-size and geochemical analyses demonstrate that Peoria Loess is not uniform in composition through the section (Muhs and Bettis, 2000). Three grain-size zones are present, an upper finer-grained zone, a coarser middle zone, and a lower zone where grain size is more variable (fig. 1–4). The upper and middle zones also differ geochemically. Muhs and Bettis (2000) showed that Peoria Loess at Loveland has geochemical properties intermediate between nonglaciogenic loess of Nebraska and glaciogenic loess of western Illinois. Eastern Nebraska loess has higher K_2O content (from K-feldspar and micas) and generally higher Na_2O content (plagioclase) than western Illinois, differences that reflect different silicate mineralogy in the source rocks of these loesses (fig. 1–5). Overall, the Peoria at Loveland is intermediate in composition between Nebraska and Illinois loess, indicating that it could contain a mixture of both Laurentide-derived source sediments (glaciogenic) and nonglaciogenic source sediments. The composition of the upper grain-size zone of the Peoria Loess at Loveland is more like that of nonglaciogenic (eastern Nebraska) loess, while the middle zone's composition is more like the glaciogenic loess of western Illinois. This suggests that there may be a larger component of nonglaciogenic, Nebraska-type source loess in upper Peoria Loess than in middle Peoria Loess at Loveland.

A series of OSL ages, determined using a modified single aliquot (SAR) procedure on 30–50-µm quartz, reveals the age structure of the Peoria Loess at Loveland (fig. 1–3). The lower 7 m of the Peoria accumulated between about 23 and 20 ka, slow enough that carbonate was partially leached from this zone as the loess accumulated (fig. 1–4; Muhs and Bettis, 2000). Ages from the middle increment of the Peoria (~33–19 m) are statistically indistinguishable, agreeing to within 1 sigma error. The change from the middle to upper zone of the Peoria Loess (at 19-m depth) occurred abruptly about 20 ka, and the upper Peoria accumulated over a 4,000-yr period until about 16 ka (fig. 1–3). Mass accumulation rates (MAR) calculated based on the central values for the OSL ages and assuming a typical loess bulk density of 1.45 g cm^{-3} show that, based on the uppermost and lowest OSL ages alone, the average mass accumulation rate over the last glacial period is extremely high at ~8,169 g m^{-2} yr^{-1}. This value is one to two orders of magnitude higher than those calculated for last glacial eolian silt elsewhere in the world (Roberts et al., 2003). At Loveland, MAR varies from 3,383 g m^{-2} yr^{-1} in the lower Peoria to 6,070 g m^{-2} yr^{-1} in the upper Peoria. In the middle zone of the Peoria loess, accumulation was more rapid than OSL dating allows us to resolve (>10,000 g m^{-2} yr^{-1}).

TABLE 1–1. Uranium-series isochron-plot ages of carbonate nodules from the Loveland Loess at Loveland, western Iowa. U-series ages were calculated from half-lives of ^{230}Th and ^{234}U of 75,200 and 244,000 years, respectively; errors of ages are reported at 1 standard deviation. Ages were determined by B. Szabo, U.S. Geological Survey (retired).

Sample	Age, ka
LN–2	63±6
LN–5	68±8
LN–6	57±6

Between STOPS 2 and 3

We drive along the base of the eastern bluff of the Missouri River valley for the next approximately 30 km (18.5 mi). The stratigraphic sequence beneath this bluff line is similar to that which we observed at the last stop. We cross the Boyer River about 4 km north of Stop 2. This river carried outwash from the Des Moines Lobe of the Laurentide ice sheet from about 18–16.5 ka. The late glacial floodplain is 10–15 m below the level of the modern floodplain but rises above it approximately 12 km up the valley. Steeper-than-modern last-glacial gradients are typical of valleys that drained the Laurentide ice sheet. The Missouri River valley widens as we proceed north, and the depth to bedrock increases. This is a region of classic "loess topography" produced by a combination of a thick loess mantle and deeply entrenched stream systems. Shimek (1896, 1898) proposed the eolian origin for loess on the basis of the occurrence of terrestrial gastropod fauna in the loess and the distribution of the loess across the highest parts of the landscape in this region.

STOP 3. Ridge Summit near Blair, Nebraska

At this stop we will discuss loess stratigraphy in the Omaha area, based mainly on cores and other subsurface data.

Background

Despite the long history of loess research in the middle Missouri River valley and adjacent regions at the eastern edge of the Great Plains, little of that work has focused on the west side of the valley near Omaha, Nebraska. The east side of the Missouri Valley in southwestern Iowa has attracted more attention in part because of the many well-exposed loess sections there, but also because of the influence of the geomorphic and stratigraphic research programs directed by Robert Ruhe in the 1960's (Ruhe, 1969; Ruhe and Cady, 1967; Ruhe et al., 1971). Furthermore, the middle Missouri Valley is generally interpreted as a classic glaciofluvial loess source; by analogy with other glaciofluvial sources such as the Mississippi River valley, the thickest and most complete loess sections should be on the east side.

Recent research suggests that the stratigraphic record of loess deposition in eastern Nebraska deserves much more attention, because of the information it may provide on loess sources and their climatic and glacial controls during the middle and late Pleistocene. Eastern Nebraska potentially received significant amounts of dust influx from several very different loess systems, during late Pleistocene deposition of Peoria Loess, and possibly during deposition of older units (Mason, 2001;

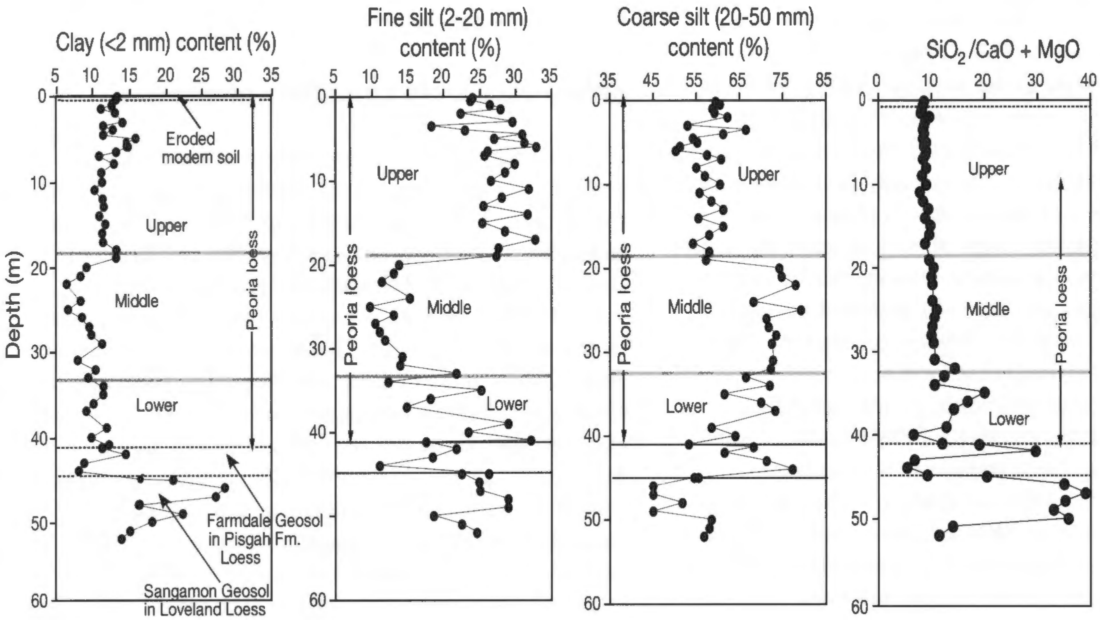

FIGURE 1–4. Grain-size and silica/carbonate depth plots for the loess sequence at the Loveland site. Note the presence of three distinct grain-size zones in the Peoria Loess: upper fine-grained zone, a coarser middle zone, and a lower zone of variable grain size. The silica/carbonate plot indicates that the lower zone is depleted in carbonate relative to the upper zones. Grain-size analysis was done using the pipette method; chemistry was done by wavelength-dispersive X-ray fluorescence. Data from Muhs and Bettis (2000).

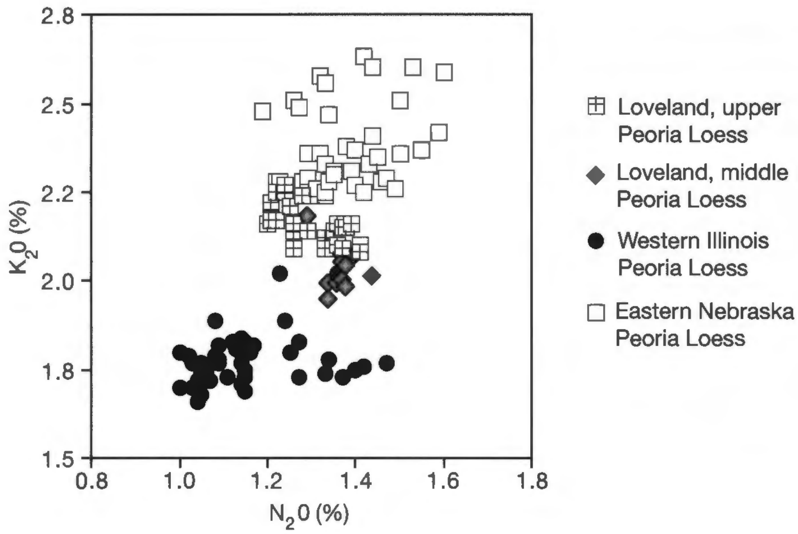

FIGURE 1–5. Concentrations of K_2O and Na_2O in bulk samples of upper and middle Peoria Loess at Loveland, Peoria Loess in eastern Nebraska (Plattsmouth, Elba, and Lincoln sections), and Peoria Loess in western Illinois (Morrison and Rapids City B sections). The plot shows that upper Peoria Loess has geochemical properties similar to those of nonglaciogenic eastern Nebraska loess, while the geochemistry of the middle Peoria Loess at Loveland is more like glaciogenic (Laurentide-source) loess of Illinois. Adapted from Muhs and Bettis (2000).

Muhs and Bettis, 2000). The nearby Missouri Valley is the most obvious loess source near Omaha, and thickness trends (discussed in more detail below) confirm its local importance (Mason, 2001). However, as discussed at earlier stops, there is evidence that Peoria Loess from nonglacial sources on the Great Plains was transported as least as far eastward as the Missouri Valley (Muhs and Bettis, 2000). Finally, there is some evidence that the Platte and Elkhorn river floodplains were secondary Peoria Loess sources in Nebraska, though only detectable within 40 km of the Platte or Elkhorn valleys (Mason, 2001). The lower Platte and Elkhorn rivers share a wide valley within 40 km of many of the new loess sections that have been cored in the Omaha area (fig. 1–6), and thickness trends in those cores support earlier suggestions that the Platte Valley was a locally significant loess source.

The three possible sources of Peoria Loess near Omaha could have had distinctly different controls, and loess transport from these sources may have begun, peaked, and ended at different times. For example, peak dust production from the Missouri River floodplain may have occurred when high glacial sediment input and highly variable meltwater discharge combined to produce abundant wind-erodible sediment on a broad unvegetated floodplain, analogous to glaciofluvial loess sources in Alaska today (e.g., Péwé, 1951). In contrast, Peoria Loess influx to the Omaha area from nonglacial Great Plains sources was almost certainly under the direct control of climatic and vegetation change within the source areas (Mason, 2001; Muhs et al., 1999). Dust production from the Platte River floodplain may have been influenced by glaciers in the Rocky Mountains, but that influence was always remote from the Omaha area. Floodplain conditions favoring dust production may have been more closely related to climatic change in the basins of the Loup and Elkhorn rivers, the Platte's major Great Plains tributaries.

Roberts et al. (2003) recently suggested that loess in the midcontinent is not simply a passive recorder of paleoclimate and/or ice sheet fluctuations but may have actively influenced climate in areas downwind of loess sources. This highlights the importance of sorting out the relative timing of loess influx from various sources to the Omaha area: that information is crucial to understanding the interplay of climate and ice-sheet dynamics in controlling the production of dust that was then carried eastward across the midcontinent. Work at the Loveland section (Stop 2) has already demonstrated distinct temporal changes in Peoria Loess properties consistent with changing provenance (Muhs and Bettis, 2000). A more detailed picture should be available through research on temporal and spatial trends in the abundance of multiple provenance tracers in Peoria Loess across eastern Nebraska, where "dust plumes" from these sources overlapped.

The cores we will see at this stop, and others like them, raise an even more intriguing question: *Were older loess units also transported from the same mixture of glacial and nonglacial sources as Peoria Loess?* If so, then many of the insights gained from studying the climatic and glacial controls on these sources in the case of Peoria Loess may be applicable to older units as well.

Subsurface Data from the Omaha Area

Over the past five years, cores collected during geologic mapping projects have demonstrated that a rich stratigraphic record is preserved in thick loess of eastern Nebraska. These cores, mainly from the area north of Omaha between the Missouri and Elkhorn rivers (fig. 1–6), also illustrate the value of subsurface investigations on broad upland summits. There are many exposures of Peoria Loess in the Omaha area, in gullies, roadcuts, and borrow pits, but few are located on wide ridge tops and many do not expose the entire thickness of Peoria Loess. Many outcrops also expose the Gilman Canyon Formation and the uppermost Loveland Loess, but virtually all of these are on side slopes. During extensive field work in 1997–2002, we located only three exposures of pre-Loveland loess, one a large quarry face cut through a spur ridge descending to the Platte River valley, and the other two small roadcuts poorly exposing the loess section. Given previous work in the Midwest that demonstrated truncation of loess sections on hillslopes (Ruhe, 1967; Ruhe, 1969; Ruhe and Cady, 1967), we turned to subsurface data from broad upland summits to more adequately characterize loess stratigraphy.

Prior to 1997, the major source of information on loess stratigraphy in the Omaha area was a large number of geologic test holes drilled by the Conservation and Survey Division (Nebraska's state geological survey) from the 1930's to the 1980's. Two north-south lines of test holes, 19 km apart, were drilled in the Omaha area between the Missouri and Elkhorn rivers (fig. 1–6A). Test holes were drilled at approximately 5-km intervals along each line, with the exact drilling locations placed on ridgetops where possible. The test holes were drilled using mud-rotary methods, usually to the base of the Cenozoic. Stratigraphy was inferred from drill cuttings carried to the surface in drilling mud, together with changes in drilling action and geophysical wireline logs (examples are shown in fig. 1–7). A pause in drilling was used to clean the mud of old cuttings, at each major change in lithology and every 1.2 m when lithology was uniform; this greatly enhances the accuracy of stratigraphic boundary placement. Geophysical logs typically included single point resistance (SPR) and spontaneous potential (SP) measurements. Higher values of SPR correspond to coarser grain size and/or lower water content. Some fine-grained sediments with strong pedogenic structure also have high SPR values, perhaps because large pores between peds have an effect similar to large intergranular pores in unstructured sediment. The SP log (not shown in fig. 1–7) is generally the inverse of the SPR log.

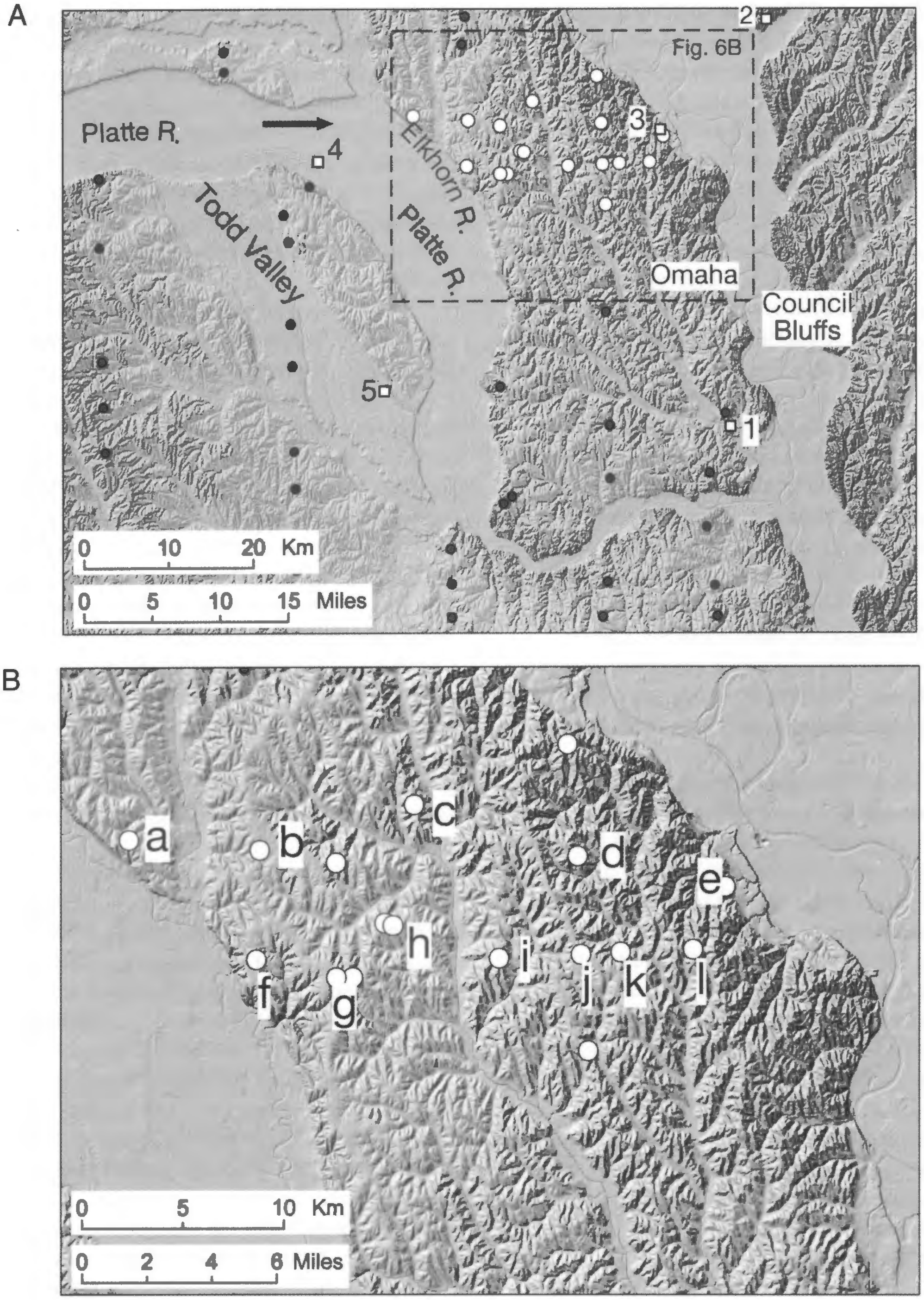

FIGURE 1–6. Shaded relief images of the Omaha area. (A) Regional view, showing field-trip stops, study sites, and the Todd Valley. Arrow indicates new path taken by the Platte when it was captured by an Elkhorn River tributary, causing abandonment of the Todd Valley. Stops on this field trip indicated by numbered squares (exact locations subject to change in some cases). Unfilled circles indicate locations of cores and mud-rotary test holes used to characterize loess stratigraphy north of Omaha. Smaller filled circles are other sites drilled by the Conservation and Survey Division that are in landscape positions suitable for characterizing loess stratigraphy (mainly upland summits). (B) Area north of Omaha where loess stratigraphy is best known because of recent drilling associated with geologic mapping projects. Unfilled circles are loess study sites, as in A. Sites used in constructing the fence diagrams in fig. 8 are labeled with lower-case letters.

Between 1997 and 2001, eight additional mud-rotary test holes were drilled and 15 hollow-stem auger cores were collected in the Omaha area. Most of the new drilling was for the purposes of surficial geologic mapping, with one core collected for an NSF-funded project on loess chronology. The cores were collected mainly on ridge summits, with a few on lower slopes. The new mud-rotary test holes were located on summits, side slopes, or foot slopes, and were drilled by methods similar to those used in earlier years, except that natural gamma logs were collected along with SPR data (fig. 1–7). Natural gamma logs have high values in sediment that contains abundant K, U, and Th, usually because of high clay content. The 5-cm-diameter hollow-stem auger cores were collected continuously to depths of up to 33 m. Core recovery was often near 100% in the loess section, except for two stratigraphic intervals. The middle and lower Peoria Loess often collapsed and jammed the coring tube, leading to poor recovery. The strongly structured upper part of the Sangamon Geosol also jammed and was not fully recovered at some sites. Loess stratigraphy interpreted from cores and test holes in the Omaha area can be compared with abundant subsurface data from elsewhere in eastern Nebraska. This includes many pre-1997 mud-rotary test holes and several auger cores collected in 1998–2002.

Optically stimulated luminescence (OSL) dating is in progress on core samples from one site north of Omaha (Stevens Site, same locality as Core 1-B-98 in fig. 1–8; locality k in fig. 1–6B). From the same core, samples have been analyzed to determine content of meteoric ^{10}Be (^{10}Be formed in the atmosphere and deposited with precipitation), to extend the chronology beyond the current limits of luminescence dating. Another core (3-B-99 in fig. 1–8; locality c in fig. 1-6B) has been used by Greg Balco and John Stone (University of Washington) in testing a new method of dating sediment burial using the content of ^{10}Be and ^{26}Al produced in situ. Minimally disturbed samples from several cores have been used to prepare thin sections, mainly for characterizing the paleopedology of the loess section. Grain-size analysis and X-ray diffraction analysis of clay mineralogy have been carried out on samples from several cores and one outcrop.

Loess Stratigraphy

On the relatively stable parts of the landscape represented by these cores, a regionally coherent record of loess deposition and soil formation is preserved (fig. 1–8). This record is apparently more complete and may represent a longer period of time than the loess sequences previously described at exposures east *or* west of the Missouri River. Cores and test holes on broad upland summits north of Omaha consistently contain the same sequence of depositional units. The oldest sediments in this sequence, informally designated the "silty clays," may correlate in part with poorly understood pre-Loveland loess units that are extensive on the Great Plains (e.g., Feng et al., 1994) and are occasionally observed in the Mississippi Valley (e.g., Jacobs and Knox, 1994; Markewich et al., 1998). The three units above the silty clays can readily be correlated with formal lithostratigraphic units established for loess across the midcontinent: from oldest to youngest, these are the Loveland Loess, Gilman Canyon Formation (stratigraphically equivalent to the Pisgah Formation in Iowa and the Roxana Silt in Illinois and Wisconsin), and Peoria Loess.

Outcrops, test holes, and cores on hillslopes provide an important additional insight on loess stratigraphy in this area. All units observed in the loess sequence on summits are truncated to varying degrees on steep backslopes, and in some cases glacial till is at the surface. However, the entire sequence (including the silty clays) is present on some upper footslopes and low spur ridges, although each unit is thinner than on adjacent summits. The occurrence of all of the units in the sequence both on ridge tops and much lower in the landscape strongly supports the interpretation that all are primarily loess, rather than lacustrine or depression-fill sediment. This distribution also indicates that the present stream-dissected topography was at least partially developed before the end of deposition of the silty clays, and well before the last glaciation. In effect, each loess unit has draped a pre-existing landscape of ridges and valleys but has been preserved to varying degrees depending on the local slope gradient.

The "silty clays" have a total thickness of 6–9 m on wide ridge tops and rest directly on the uppermost glacial diamicton (except in one outcrop described below). Where encountered in older test holes, the silty clays were often labeled "Sappa Formation," but no clear correlation to the type-Sappa section in south-central Nebraska (Reed and Dreeszen, 1965) can be established. The "Yarmouth Clay," described in southwestern Iowa by Ruhe and Cady (1967), was interpreted by those authors as a lacustrine deposit, but could correlate with part or all of the silty clays near Omaha. The stratigraphic position of the silty clays indicates that they post-date the most recent pre-Illinoian glaciation of the Omaha area. Boellstorff (1978) reported no glacial deposits overlying the Lava Creek B ash (640 ka) in eastern Nebraska, although they exist in Iowa and South Dakota. At a quarry south of Omaha, the silty clays rest on fine-grained laminated sediment, probably lacustrine, that overlies Pennsylvanian bedrock. An undated volcanic ash is interbedded with this lacustrine sediment. This ash is most likely the Lava Creek B, which is present at many more localities than any other tephras in eastern Nebraska and Iowa. Thus, the silty clays may have begun accumulating shortly after 640 ka, although much more geochronological work on this unit is clearly needed. *If* this inferred basal age is correct, and the loess sequence does not contain large unrecognized unconformities, the silty clays represent several glacial-interglacial cycles represented by marine isotope stages.

The texture of the silty clays is variable within a single section, typically ranging from clay or silty clay to silty clay loam. High-resolution particle-size analysis by laser diffraction indicates a prominent medium-coarse silt peak (characteristic of loess), with a shoulder or secondary peak in the clay to fine-silt range. Geophysical logs run in mud-rotary test holes indicate that the silty clays have very low resistivity and emit relatively high natural gamma radiation, both indicative of high clay content (fig. 1–7). Gamma logs suggest multiple distinct peaks of clay content (fig. 1–7), but insufficient particle-size data are available to test for such peaks in the cores. Weak to strong pedogenic blocky structure is present throughout much of the thickness of the silty clays, with hints of granular structure in some horizons. However, discrete A–B–C soil profiles are rarely identifiable, except at the base of the unit, where a soil is developed through a gradational contact with underlying diamicton. Root traces and clay coatings on ped faces are often present. Thin sections confirm the presence of illuvial clay coatings in some cases, but also indicate strong stress orientation of clay, probably the result of shrink-swell cycles.

The silty clays are interpreted as a stack of several thin depositional units, primarily loess, but also possibly including some depression-fill or pond sediment toward the base. The lowermost meter of the silty clays often contains scattered coarse sand grains and pebbles; these could have been mixed upward by bio- or cryoturbation, but would also be consistent with a depression-fill origin. This stack of sediment was strongly overprinted by pedogenesis as it accumulated, resulting in such effective welding of overlapping soil profiles that they are no longer recognizable. In effect, the original B horizon in the soil, formed at the basal contact of the silty clays, has grown upward through the rest of the unit because of its slow average rate of deposition.

Loveland Loess north of Omaha is clearly correlative with the same unit at its paratype section immediately across the Missouri in Iowa (Bettis, 1990), based on stratigraphic position and general lithological characteristics. Loveland Loess is 6 to 12 m thick in cores on broad upland summits north of Omaha, exceeding the thickness of this unit at the paratype section and in any other outcrop. Luminescence dating, both at the paratype section and in western Nebraska, places Loveland Loess generally within marine isotope stage 6 (Forman et al., 1992; Forman and Pierson, 2002; Maat and Johnson, 1996). The lower zone of Loveland Loess (described

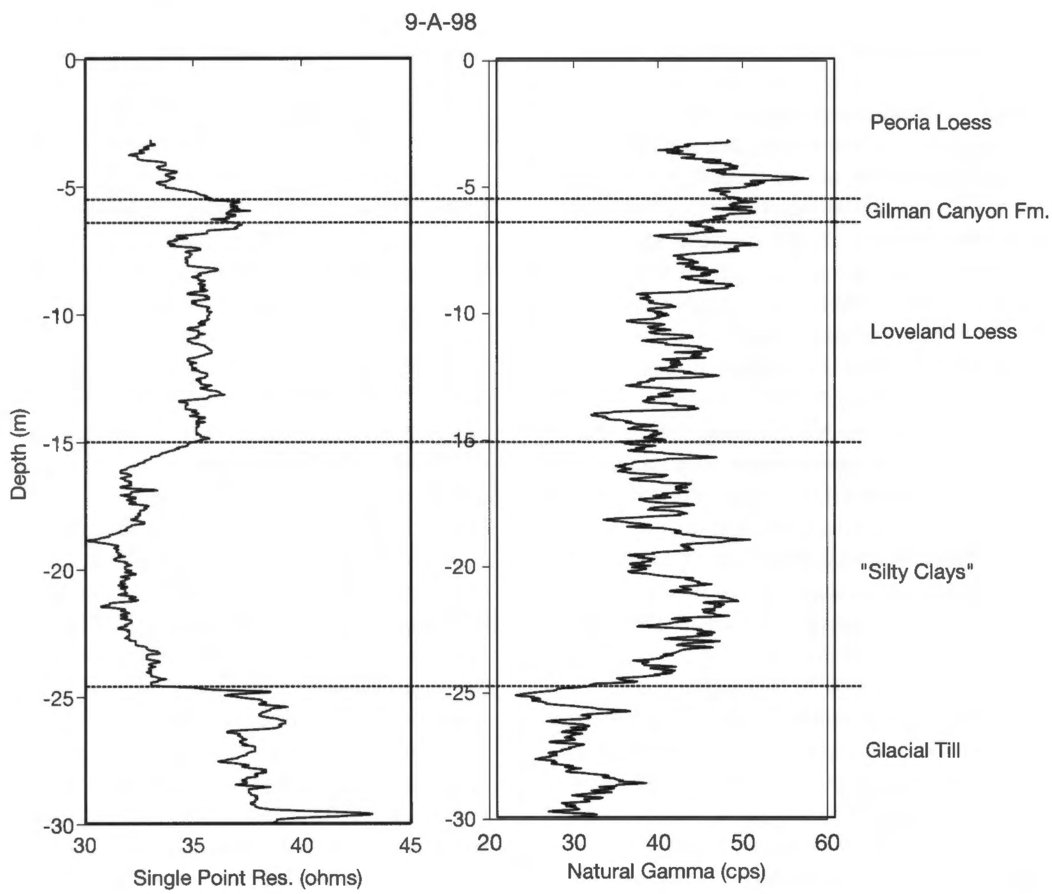

FIGURE 1–7. Example of wireline geophysical logs from a mud rotary testhole in the Omaha area. Basis for log response and loess stratigraphic units are discussed in the text. Test hole 9-A-98 was drilled in a roadcut and the zero point on the depth scale is actually 3 m below the natural land surface. Core 10-B-98 (fig. 1–8) was collected on a nearby summit and has similar loess stratigraphy.

below) is not apparent in descriptions of the paratype section (Bettis, 1990), so that section may not represent the entire duration of Loveland deposition on uplands around Omaha; however, it seems reasonable to assume that Loveland Loess was deposited within the MIS 6 glacial throughout the region.

The Sangamon Geosol is developed in the upper part of the Loveland Loess. In most cores taken near Omaha, the upper solum of this prominent paleosol is marked by horizons with well-preserved granular to very fine subangular blocky structure, which we interpret as A or AB horizons that have lost most organic carbon through post-burial oxidation. The Sangamon Geosol has a strongly expressed Bt horizon in all of the cores, which is generally reddish-brown (7.5 YR to 5 YR hue) silty-clay loam to clay, and has strong subangular blocky structure. In geophysical logs run in mud-rotary test holes, the Sangamon Bt is typically recorded by relatively high natural gamma radiation, and low single point resistance (fig. 1–7). Although the reddish-brown color of the Sangamon Bt may in part reflect iron oxide coatings on ped faces that could have formed after burial (Thompson and Soukup, 1990), pedogenic rubification clearly played a more important role, because ped interiors and clay coatings in the Bt both have redder hues than underlying less-altered Loveland Loess. In one core (2-B-98, fig. 1–8), the Sangamon Geosol Bt, along with the rest of the Loveland Loess, has a gray color interpreted as the result of reducing conditions in an upland swale. The Sangamon Bt horizon has well-developed illuvial clay coatings, clearly identifiable in thin section. Thin sections from two cores suggest physical disruption of the upper Bt of the Sangamon Soil, producing rounded aggregates with embedded clay coatings, along with infillings of uncoated silt that were probably derived from overlying Gilman Canyon Formation loess. Both disruption of pre-existing Bt horizons and silt translocation have been interpreted as evidence of active layer processes in a periglacial environment (Van Vliet–Lanoe, 1985). Based on data from two sites, the predominant clay minerals in the Sangamon Geosol are kaolinite and illite, with small amounts of smectite evident in most diffractograms. There is no evidence of interstratified kaolinite-smectite, which has been identified as a product of in situ clay mineral weathering in the Sangamon Geosol in the lower Mississippi Valley (Markewich et al., 1998).

Below the Sangamon profile, two zones can be distinguished within the Loveland Loess. These zones are identifiable in almost all of the cores in the Omaha area, and in cores or test hole logs across a large part of eastern Nebraska. The two zones were also clearly visible in two large temporary exposures of Loveland Loess north and south of Omaha. The "early Sangamon (?) or Illinoian soil" described by Thorp et al. (1951) at the Yankee Hill Brickyard near Lincoln, Nebraska, may correspond to the lower Loveland Loess in the Omaha area. The upper zone of Loveland Loess is light-colored, generally yellowish-brown silty-clay loam or silt loam. Loess in the upper zone superficially resembles Peoria Loess but often contains sparse clay coatings in larger pores or along fractures, rarely seen in the Peoria. Within the upper zone, there are often one or more thin darker-colored bands that appear to be incipient soils. A common location for one of these dark bands is immediately below the Sangamon Bt or BC horizons. The lower zone of Loveland Loess as a whole is slightly to distinctly darker than the upper zone, and commonly has a slightly redder hue, with a silty-clay loam texture. Within the generally darker-colored lower zone, color value varies significantly, and in outcrops the lower zone often appears to contain a sequence of incipient soils separated by slightly lighter-colored sediment. One obvious interpretation of the lower zone/upper zone contrast is that the lower zone experienced more pedogenic alteration as it accumulated. It is also possible that the color difference (especially the redder hue in the lower zone) in part reflects changing provenance. In both upper and lower zones, thin sections indicate massive to blocky structure, with rare to common illuvial clay coatings in pores. The predominant clay minerals in both zones are illite, kaolinite, and smectite. Smectite may be more abundant below the Sangamon Geosol than within it.

The Gilman Canyon Formation has been dated using radiocarbon and luminescence methods at numerous localities, including Stops 1 and 2 of this field trip. In core 2-B-98 (fig. 1–8), the Gilman Canyon Formation was deposited in a poorly drained setting, in a swale on a broad ridge top. At this site, the unit has very high organic-matter content and contains wood fragments. Wood at 0.6 m below the top of the Gilman Canyon Formation yielded an age of 24,010 ± 90 ^{14}C yrs B.P. (Beta–119107), consistent with ages obtained from the unit elsewhere.

The Gilman Canyon Formation in the Omaha area does not contain two or more distinct, darker-colored buried A horizons, as reported from localities in western Nebraska (Maat and Johnson, 1996). In eastern Nebraska, this unit is usually thin (<1 m), except immediately west of the Missouri. In one core on the western bluff line of the Missouri Valley (3-B-98, fig. 1–8), the Gilman Canyon Formation is almost 6 m thick. The lower boundary of the unit is very difficult to place. In effect, the initial increment of Gilman Canyon Formation loess was incorporated into the Sangamon Geosol A horizon, building that horizon upward over time. Pedogenic alteration characteristically decreases upward within the Gilman Canyon Formation. Macroscopically, there is an upward transition from strong granular or very fine subangular blocky structure to a near-massive condition. In thin section, the lower Gilman Canyon Formation has well-developed granular structure, abundant burrows and excrements of soil fauna, and high porosity, all characteristic of modern grassland soil A horizons. In many geophysical logs from eastern Nebraska, the Gilman Canyon Formation is marked by a distinct peak of single point resistance (fig. 1–7), interpreted as a result of the high porosity and strong aggregation of the

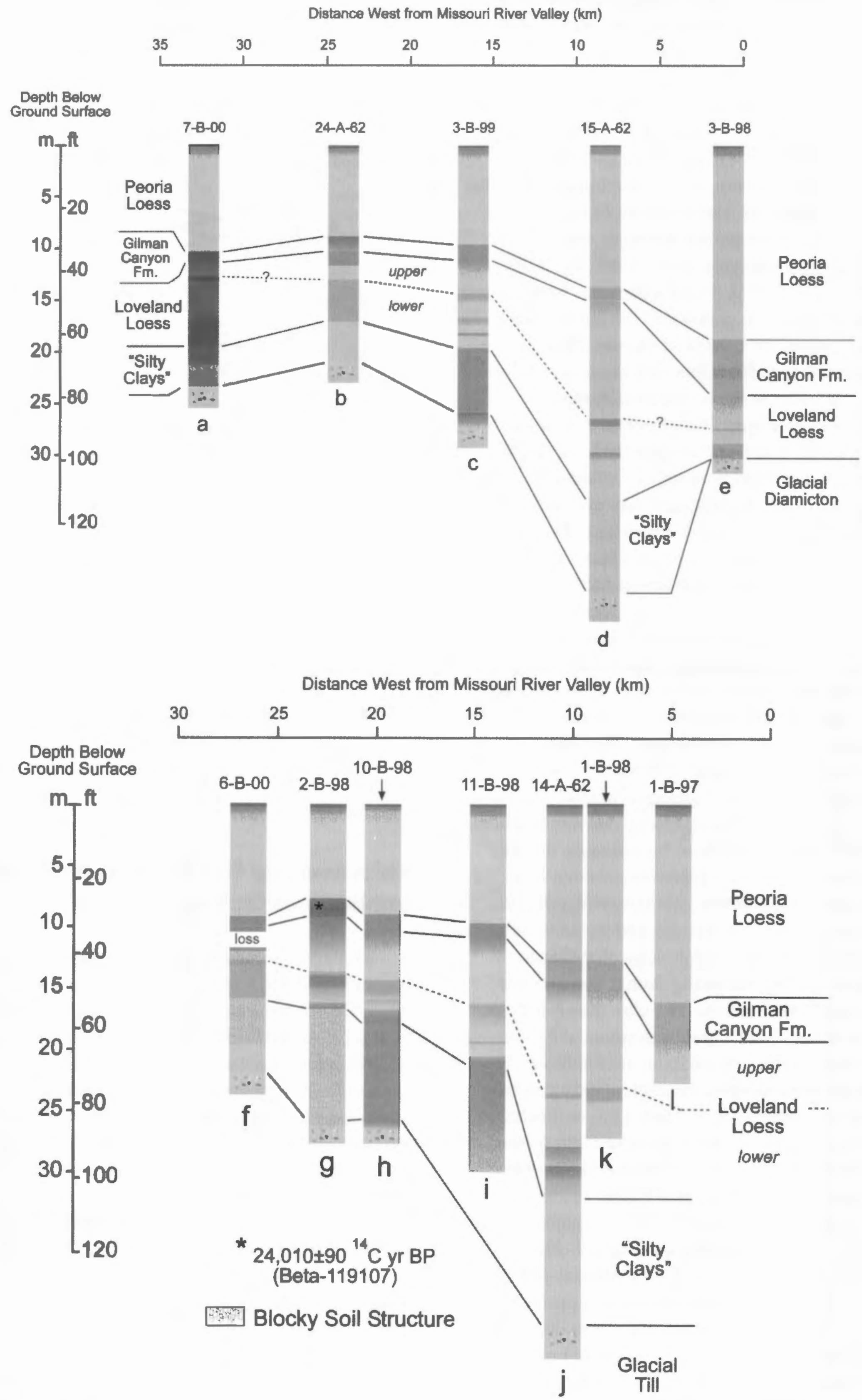

FIGURE 1–8. Loess stratigraphy along two approximately west-east transects north of Omaha (site locations keyed to fig. 1-6B). Gray shades represent lighter and darker colors (i.e., Munsell values), and the occurrence of strongly expressed blocky structure in cores is also indicated.

lower part of the unit. In effect, the aggregation creates a response similar to sand. The clay mineralogy of the Gilman Canyon in cores taken near Omaha is dominated by illite, kaolinite, and smectite.

Peoria Loess (or Peoria Silt) can be correlated across the North American midcontinent. It was deposited generally within the period from 25 ka to 11 ka, but represents different intervals within that range in various parts of the midcontinent. Initial results of optically stimulated luminescence dating of Peoria Loess at the Stevens site (same locality as core 1-B-98, fig. 1–6B and 1–8) are shown in fig. 1–9. These three ages form a nearly linear age-depth trend, in contrast to the large changes in accumulation rate at the Loveland section (Stop 2). Of course, additional ages from this core (in progress) could reveal more variation in the accumulation rate.

Peoria Loess is at least 30 m thick in core 4-A-76 (Boellstorff, 1978), taken at Hummel Park, on the Missouri River bluff line immediately north of Omaha (the contact between Peoria Loess and Gilman Canyon Formation is difficult to identify in that core). Elsewhere, Peoria Loess ranges from 21 to 8 m thick on broad upland summits of the Omaha region. As elsewhere across the Great Plains and Midwest, Peoria Loess in cores collected in eastern Nebraska is light-yellowish-brown silt loam or silty-clay loam with minimal macroscopic evidence of pedogenic alteration below the surface soil profile. Thin sections consistently indicate a distinctive lenticular microstructure in Peoria Loess from the Omaha area. The origin of this microstructure is uncertain, although it resembles structures related to ice lens formation in cold-region soils (Van Vliet–Lanoe, 1985). In any case, this microstructure indicates relatively low levels of bioturbation during loess deposition, suggesting sparse vegetation, low primary productivity, and low abundance of soil fauna. In contrast, Peoria Loess from central and western Nebraska often displays more micromorphological evidence of soil biological activity, including faunal burrows and excrements and root channels. The clay mineralogy of Peoria Loess in the Omaha area is dominated by smectite, with much smaller amounts of illite and kaolinite. There is an abrupt decrease in smectite content between basal Peoria Loess and underlying Gilman Canyon Formation.

A common feature in many loess sections across the midcontinent is a coarsening upward in the upper one-half or one-third of the Peoria Loess, culminating in one or more grain-size peaks. Above these peaks, grain size decreases abruptly (unpublished data, J. Mason). The Loveland section does not fit this pattern, probably because of varying loess input from highly proximal and distal sources. Peoria Loess north of Omaha does fit the pattern, as illustrated in fig. 1–9, despite the probability of multiple sources in this area as well. The grain-size peaks in the upper part of the section may represent strengthening westerly winds as the Laurentide ice sheet retreated.

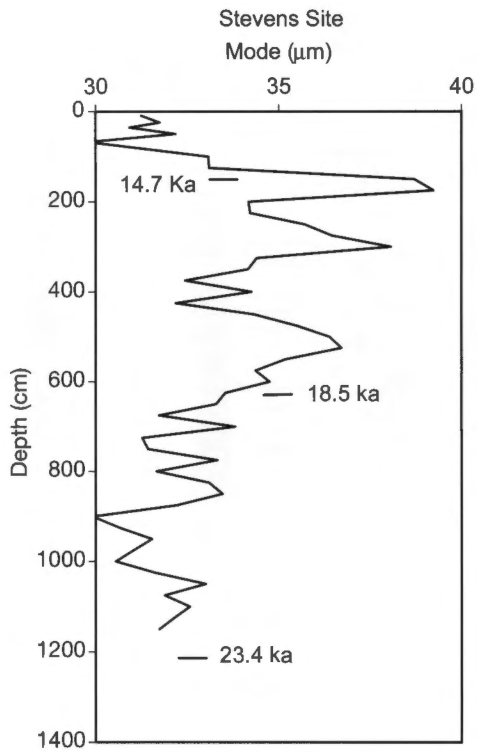

FIGURE 1–9. Optically stimulated luminescence ages and grain-size variation within Peoria Loess at the Stevens site (same locality as Core 1-B-98, figs. 1–6B and 1–8). Modal diameter is calculated from particle-size distributions determined by laser diffraction.

Thickness and Grain-size Trends, and Implications for Loess Sources

Thickness trends of loess units along two east-west transects north of Omaha, between the Missouri and Platte/Elkhorn valleys, are readily apparent in fig. 1–8. Peoria Loess thickness from sites on those transects and several other localities north of Omaha is also plotted against shortest distance to the Missouri in fig. 1–10. Most cores and test holes used to construct these figures are from ridge summits that are among the highest points in the local landscape, so thickness trends in these cores should be a reasonable approximation of trends in loess-deposition rate, relatively unmodified by post-depositional erosion. An important exception is core 3-B-98, collected on a lower spur ridge immediately adjacent to the Missouri Valley in an attempt to capture extremes of loess thickness there. The lower part of the loess section (below Gilman Canyon Formation) appears to be truncated in this core, suggesting that the core as a whole may underestimate loess-unit thicknesses. Figure 1–11 illustrates grain-size data from selected, clearly identifiable stratigraphic positions within the loess sequence. Insufficient data are available from the silty clays to include them here. The modal diameter is probably the single best index of the sorting of individual coarse silt and sand grains during loess transport. Ratios of grain-size fractions, such as 16–31 µm/31–63 µm, respond to sorting in transport but may also vary with changing proportions of fine silt and clay carried in larger sedimentary aggregates (Mason, Jacobs, Green, et al., 2003).

Both thickness and grain-size trends indicate that loess was transported westward from the Missouri Valley into Nebraska, at least as early as MIS 6 (the Illinoian glaciation), and probably earlier in the Middle Pleistocene. Peoria Loess and Loveland Loess clearly thin westward away from the Missouri Valley (excluding the anomalously thin Loveland in core 3-B-98). The Gilman Canyon Formation thins westward at sites near the Missouri, but elsewhere shows no clear trend; furthermore, the uncertainty in placement of the lower boundary is large relative to the total thickness of this unit. The silty clays as a whole also appear to thin westward, although more thickness measurements from this unit are needed. This suggests that the Missouri had already established its present course and was a significant loess source well before MIS 6, perhaps as early as 500–600 ka. Grain-size data clearly show that Peoria Loess, Gilman Canyon Formation, and Loveland Loess all become much finer with distance westward from the Missouri, over a distance of at least 20 km. The trends observed in Loveland Loess in Nebraska, combined with the well-established eastward thinning and fining of Loveland Loess in southwestern Iowa (Ruhe, 1969; Ruhe and Cady, 1967), indicate that the dispersal of glaciogenic loess from the Missouri Valley in MIS 6 followed a pattern similar to dispersal of Peoria Loess in MIS 2.

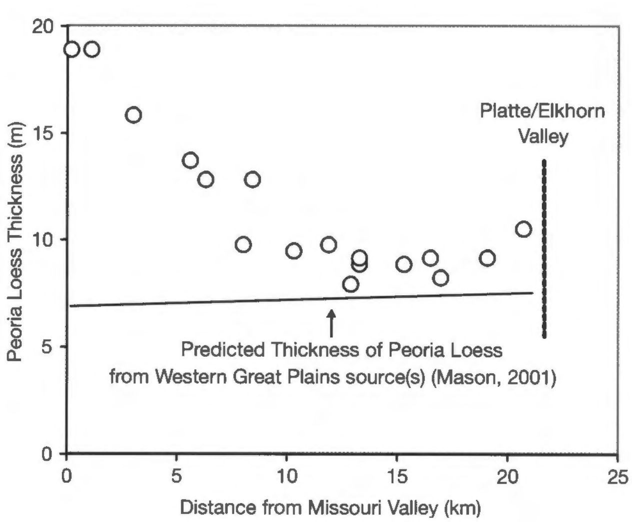

FIGURE 1–10. Peoria Loess thickness as a function of shortest distance to the western bluff line of the Missouri River valley, cores, and test holes north of Omaha (fig. 1–6B). The shortest distance in all cases is in a northeastward direction.

In spite of this strong evidence for persistent loess transport from the Missouri, much of the loess more than about 10 km west of the Missouri Valley may actually have come from major regional sources to the west on the Great Plains. This can be demonstrated for the Peoria Loess and should be considered as a possibility for older units as well. Across much of Nebraska, Peoria Loess thickness decreases eastward, indicating transport from northwestern sources (Mason, 2001). This trend can be extrapolated to the Omaha area, to predict the amount of western-source loess that should be present there, *without any loess transport from the Missouri Valley.* As shown in fig. 1–10, western-source, nonglaciogenic loess probably makes up over 50% of the total Peoria Loess thickness even at sites only 10 km from the Missouri Valley. With more work on available subsurface data and new cores in key locations, it may be possible to make a similar separation of Missouri Valley and western contributions to the Loveland Loess and silty clays.

The possibility of a third loess source in the Platte/Elkhorn Valley is hinted at but not strongly supported by grain size and thickness data from the Omaha area. A Platte/Elkhorn source should be indicated by a steep eastward decrease in thickness and grain size in the area immediately downwind (i.e., at the west end of transects shown in figs. 1–8, 1–10, and 1–11). These source-proximal trends should be distinguishable from very gradual eastward decreases in grain size and thickness of loess transported from the major Great Plains sources (note the very gradual eastward trend of predicted western-source Peoria Loess thickness in fig. 1–10). The modal diameters of the lower Gilman Canyon Formation and upper Loveland Loess do clearly become finer eastward from the Platte/Elkhorn Valley as well as westward from the Missouri (fig. 1–11). Similar trends are not evident for the modal diameters of other stratigraphic horizons, however. The 16–31 µm/31–63 µm ratios are slightly lower (indicating coarser loess) near the Platte/Elkhorn Valley than at sites midway between that valley and the Missouri, for most stratigraphic horizons. In figs. 1–8 and 1–10, Peoria Loess thickness increases slightly at the site farthest west from the Missouri and closest to the Platte/Elkhorn. There is also anomalously thick Peoria Loess at drilling sites immediately east of the Platte/Elkhorn Valley southwest of Omaha, with 1.6 to 2 times the thickness predicted from regional western-source trends (these isolated sites are not included in fig. 1–10).

STOP 4. View of Fremont Bluffs Section

The outcrops across the river from our stop have been studied for many years, as the bluff face has actively retreated (Lueninghoener, 1947; Lugn, 1935; Reed and Dreeszen, 1965; Roy et al., 2004; Wayne, 1987). Most previous work has focused on the pre-Illinoian glacial

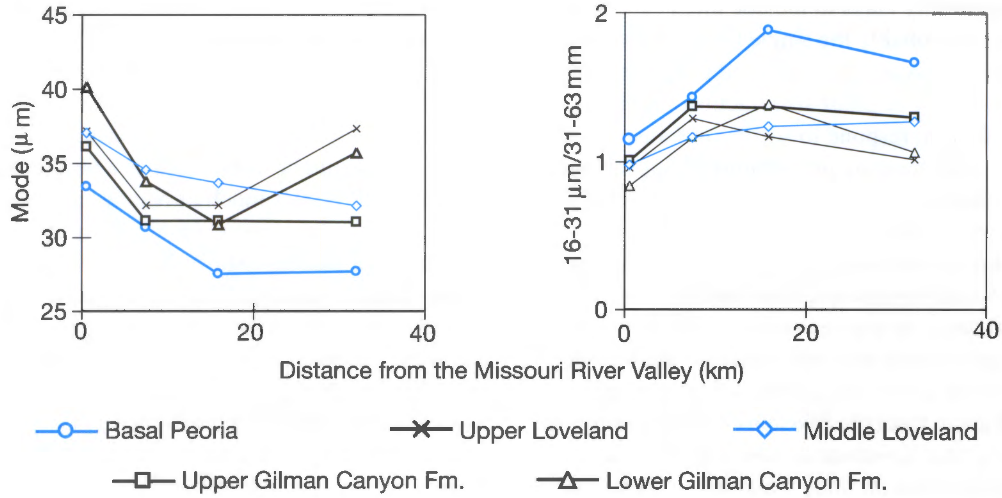

FIGURE 1–11. Grain size of selected samples from easily identified stratigraphic positions within the loess sequence in cores collected north of Omaha, plotted against distance from the Missouri Valley. "Middle Loveland Loess" is simply the midpoint of that unit in a given core, not a separate stratigraphic zone.

tills exposed here. Reed and Dreeszen (1965) described tills of two separate "Kansan" glaciations, separated by the "Fontanelle soil." The latter appears to be more of an organic-rich silt than a well-developed soil and at present is not clearly visible from across the river. Roy et al. (2004) reported two tills, both with normal magnetic polarity; the composition of both units is similar to the younger pre-Illinoian tills at other sites.

The Fremont Bluffs section is also interesting from the standpoint of loess stratigraphy, mainly because of what is *not* there. Peoria Loess at the top of the section overlies the Gilman Canyon Formation, which in turn rests on a well-developed buried soil (a prominent red band crossing the entire exposure). Loveland Loess is thin (<2 m) or absent here, depending on when and where the section has been measured. Because he did not recognize any Loveland Loess here, Wayne (1987) identified the prominent reddish-brown soil as the "Yarmouth–Sangamon Paleosol," implying that soil formation started before the last interglacial (MIS 5). Whatever Loveland Loess is actually present is entirely contained within this soil, and the band of secondary carbonates below the reddish-brown Bt horizon (visible from across the river) is in glacial till. The silty clays appear to be completely absent here.

However, the Fremont Bluffs provide an example of the hazards of relying on outcrops to reconstruct loess stratigraphy, even impressive outcrops such as this one. The Fremont Bluffs exposures are unrepresentative of local loess stratigraphy. Mud-rotary test holes and cores on uplands south and north of the Platte Valley near Fremont (fig. 1–6A) clearly indicate the presence of a significant thickness of Loveland Loess *and* sediments equivalent to the silty clays north of Omaha.

STOP 5. Todd Valley

The Todd Valley is one of the most striking geomorphic features in eastern Nebraska: a broad, well-defined valley containing no major stream, "hanging" above the valley of the Platte River at both northern and southern ends (fig. 1–6A). The Todd Valley was recognized quite early as a former valley of the Platte. Lugn (1935) and Lueninghoener (1947) proposed a straightforward autocapture mechanism to explain this feature, in which a tributary of the Elkhorn River captured the Platte near Fremont (fig. 1–6A). The Platte River has subsequently flowed through the former Elkhorn Valley for a short distance until rejoining its old course near Ashland, Nebraska (where I-80 crosses the Platte). Lueninghoener (1947) concluded that the capture occurred early in a "Peorian cycle" of erosion and subsequent alluviation along the Platte and Elkhorn rivers that was contemporaneous with Peoria Loess deposition on uplands.

Wayne (1987) advocated an alternative sequence of events, in which the Platte initially abandoned the Todd Valley prior to the late Pleistocene. Wayne proposed that the Missouri River was diverted down the present Elkhorn Valley by a late Wisconsin ice advance across the Missouri Valley into northeastern Nebraska. Where the diverted Missouri entered the Platte Valley, near Fremont, a fan was constructed, which in turn caused aggradation of the Platte and led that river to be briefly diverted into the Todd Valley once again. However, there is little stratigraphic or geomorphic evidence for the ice sheet advance into Nebraska and the diversion of the Missouri that Wayne (1987) proposed.

Stratigraphy within the Todd Valley is well known from hundreds of well logs and several geologic test holes and cores. Sand and gravel 30–45 m thick rests directly on Cretaceous bedrock. The gravel lithology appears similar to that of the modern Platte, although detailed provenance studies have not been done. Fine sand of variable thickness overlies the coarser sand and gravel and is in turn capped by Peoria Loess that is 5–10 m thick in most areas. The Gilman Canyon Formation or older loess units have not been identified within the valley, and there is no evidence of significant soil development within the entire sequence between the surface soil and bedrock. Peoria Loess thickness within the valley represents about 75–100% of Peoria thickness on adjacent uplands.

Based on this stratigraphy, fluvial occupation of the valley predates most of the time of Peoria Loess deposition. The absence of Loveland Loess and the Sangamon Geosol within the underlying sediment indicates that the final abandonment of the valley by the Platte occurred after MIS 6 and much of the subsequent MIS 5 interglacial. The fine sand beneath Peoria Loess is interpreted here as eolian sand, for the following reasons. The Todd Valley floor has a distinctive swell and swale topography similar to hummocky sand sheets or fields of low dunes, and some low hills in the valley resemble parabolic dunes. Well logs and a coring transect on University of Nebraska land southeast of Mead both indicate that Peoria Loess thickness is similar across both high and low points of the rolling valley floor and that much of the valley-floor topography is related to varying thickness of the fine sands.

This interpretation leads to the following scenario: During or just after Gilman Canyon Formation deposition, the Platte abandoned the Todd Valley. Extensive eolian sand activity then occurred on the abandoned valley floor, ending early in the period of Peoria Loess deposition. Accumulation of the final increment of Gilman Canyon Formation loess *or* the initial increment of Peoria Loess could have been prevented by active eolian sand movement. Thus, the stratigraphy does not closely bracket valley abandonment within the Gilman Canyon Formation-to-early Peoria Loess interval.

We have recently obtained optical (OSL) ages from the uppermost fluvial sand, and the overlying fine sand interpreted as eolian (fig. 1–12). These ages indicate that the valley was abandoned between about 21 and 24 ka, and probably closer to the latter age. The ages are consistent with the stratigraphic evidence outlined above and would place the abandonment of the valley near the start of Peoria Loess deposition, so in a sense, Loeninghoener's estimated age for the capture was correct. The optical ages also raise the possibility that abandonment occurred around the same time as formation of the Fort Calhoun Terrace on the Missouri River (see discussion of the route between Council Bluffs and Stop 2). A causal connection between incision on the Missouri and abandonment of the Todd Valley is unclear, however. The stratigraphy within the valley does not provide any positive support

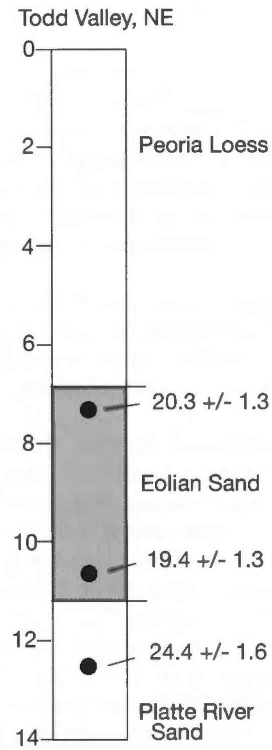

FIGURE 1–12. Optically stimulated luminescence ages from a core collected in the southeastern Todd Valley. Ages determined by Paul Hanson at the University of Nebraska Luminescence Dating Laboratory.

for Wayne's concept of multiple occupations, although such evidence might have been eroded during the final occupation.

Acknowledgments

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AMQUA Pre-Meeting Field Trip 2: Late Quaternary Biogeography and Paleoecology along the Prairie-Forest Border

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Introduction

The morning will be devoted to observing plants and animals at Konza Prairie, one of the largest prairie preserves left in the United States. Gene Towne will discuss some of the current ecological projects at the site. Participants will observe and hear about the dynamics of prairie vegetation, including the effects of fire and bison. Bill Johnson will present results of recent isotopic research at Konza after a tour along the "Bison Loop." He will consider topographic variation in $\delta^{13}C$ of modern soils as an aid to interpreting isotopic signals in buried soils, and biochemical and geochemical markers (e.g., ^{137}C) as indicators of relative surface stability.

In the afternoon, we will drive northeast into Nebraska, where we will examine paleoecologic evidence for what the vegetation was like during the late Wisconsinan and Holocene. We will look at three large cutbanks on the South Fork of the Big Nemaha River in southeastern Nebraska. Detailed studies of the alluvial stratigraphy of the sites by Rolfe Mandel and Art Bettis have provided the stratigraphic and chronologic setting, and the paleoecology is based on studies by Dick Baker and Glen Fredlund. The Farwell locality has exposures of mostly Holocene-age alluvium, plus one mid-Wisconsinan outcrop, and has been analyzed for plant macrofossils, pollen, phytoliths, and stable carbon isotopes. The Old Bridge site has exposures of early Holocene alluvium overlain by late Holocene valley fill and has been analyzed for plant macrofossils, phytoliths, and pollen. The Miles Fan site is late Wisconsinan in age, and pollen and plant macrofossils have been studied. On our way back to Lawrence, we will stop briefly at Muscotah Marsh, studied by Johanna Grüger, 1973, and re-studied by Patrick Moss (unpublished data).

Part 1. Konza Prairie

Fire, Bison, and Vegetation on the Konza Prairie

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The Konza Prairie Biological Station is a 3,487-ha tallgrass prairie preserve located in the Flint Hills of northeastern Kansas (fig. 2–1). Relatively steep slopes and shallow, clayey soils overlying limestone and shale layers make the area more suitable for grazing than cultivation. Consequently, the area has a long history of fire and ungulate grazing. To achieve its primary mission of ecological research, Konza Prairie is divided into 52 watersheds that provide large, replicated experimental units subjected to various fire regimes. In addition, there is a resident bison herd of over 300 animals within a 992-ha enclosure. This tour will focus on changes in vegetation composition and structure resulting from different burn frequencies and different burn seasons, and the interactive effects of fire and large ungulate grazing. We will also talk about the bison herd, its management, and how the bison impact tallgrass prairie compared to cattle.

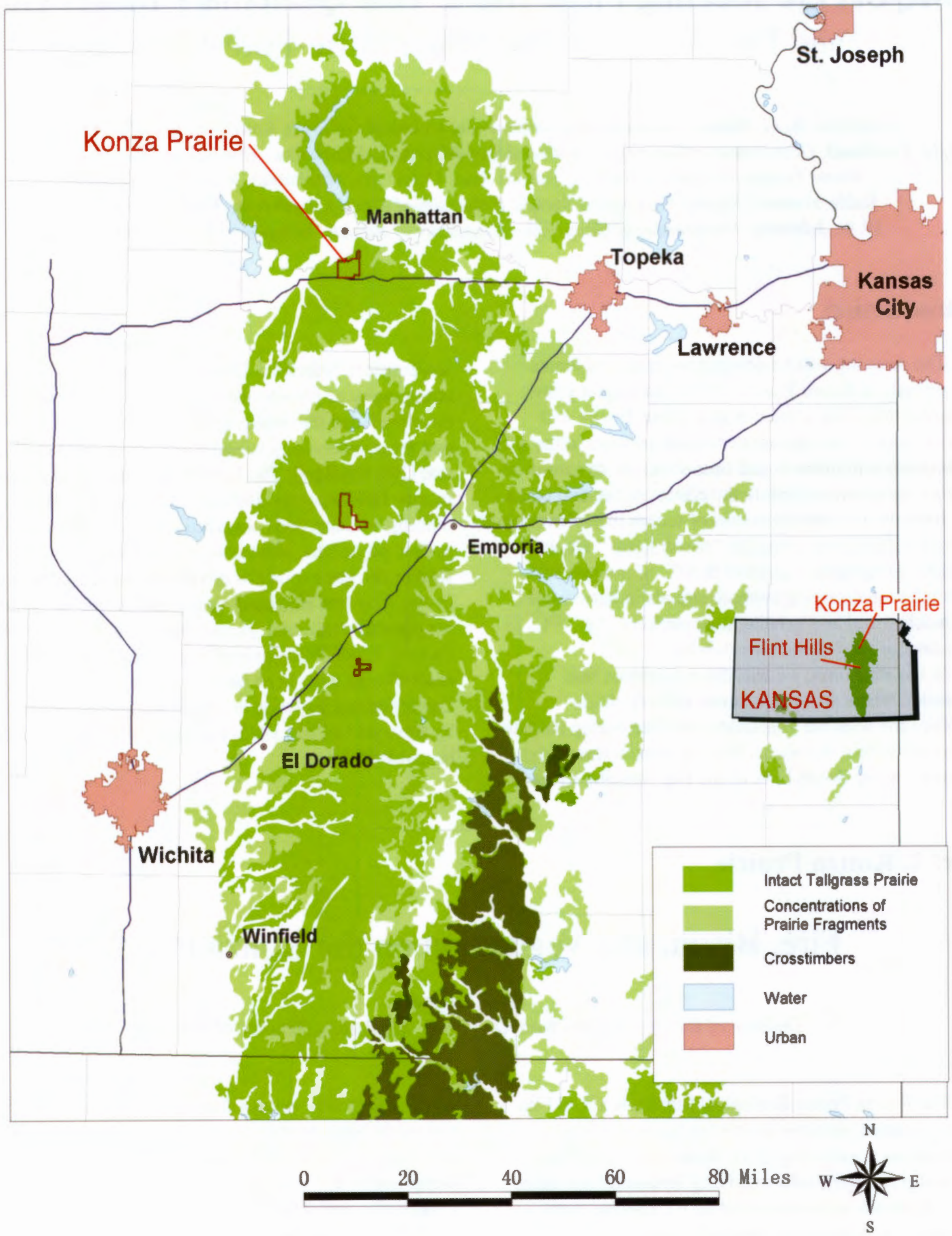

FIGURE 2–1. Map of Kansas showing the location of the Konza Prairie (map courtesy of The Nature Conservancy).

Biogeochemical Signatures within Soils of the Konza Tallgrass Prairie LTER Site

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Purpose

The relatively undisturbed nature of the Konza Tallgrass Prairie Long Term Ecological Research (LTER) site provides a reference landscape for evaluation of biogeochemical signatures within grassland soils. Our first study is evaluating the radioisotope ^{137}Cs as an erosion indicator in this loess-mantled, limestone terrain. Once the Cs inventory and its variability within the natural landscape at Konza have been characterized and modeled, we will apply the model to Fort Riley, an adjacent military reservation, in order to assess the rate of soil loss. While the climate and geology are identical to Konza, the reservation has experienced accelerated erosion caused by mechanized infantry. The second study is examining the vertical (soil profile) and horizontal (landscape) variability in stable C isotopes. Patterns observed are providing modern analog data that can be applied to the interpretation of horizontal and vertical isotopic data sequences from buried soils within the region.

Field Methods

A grid of sample points 50 m apart within a 660 m by 690 m area (fig. 2–2) was defined using a GIS, and specific points for core collection were selected with stratified random sampling. Sample points were then located on the ground using high-resolution GPS. Reference markers were placed, aerial photography flown, and a high-resolution DEM constructed (precision: 30 cm horizontal, 45 cm vertical). For Cs analysis, single cores, 30 cm long, were collected by hand at each of the 188 sample points. To evaluate micro-scale variability, 12 randomly selected points were sampled for a cluster of 12 additional samples each (fig. 2–2), thereby producing a total of 332 core samples. For stable C isotope analyses, soil surface samples (0–2 cm) and associated grass samples were collected at each of the 188 points. Also, representative upland sites were selected for extraction of 1-m to 2-m-deep soil cores to characterize the vertical distribution of stable C.

Laboratory Methods

The 332 30-cm-long cores were dried, ground, homogenized, and sent to J. Kaste and A. Heimsath at Dartmouth College for ^{137}Cs and ^{239}Pu assays. Sub-samples were also analyzed for elemental concentrations, including ^{133}Cs using an ICP-MS. Other analyses of the short cores include clay mineralogy (XRD), particle size (laser diffraction), carbon content (LECO CN analyzer), and rock magnetic analysis. Soil and grass samples were assayed for carbon isotopic content using a system consisting of a CE elemental analyzer, GasBench II, Conflo II interface, CombiPal autosampler, and ThermoFinnigan Delta Plus mass spectrometer.

Radioisotopes

The theoretical basis of using ^{137}Cs as an indicator of soil-surface stability is that H-bomb testing created a ^{137}Cs reservoir in the atmosphere, which is delivered to the soil in precipitation. ^{137}Cs then attaches to soil particles and concentrates within the uppermost part of the soil, with smaller amounts being transported down the soil profile. Generally, the more ^{137}Cs in the upper soil profile, the less erosion that has occurred since bomb testing began in the 1950's. The two research questions focus on defining (1) mechanisms that ^{137}Cs uses to move through soil and (2) scale of variation in ^{137}Cs soil inventories within a natural landscape.

^{137}Cs is thought to be strongly adsorbed onto clay minerals and organic matter in soil, resulting in very low mobility (Ritchie and McHenry, 1990), and the small size of ^{137}Cs plumes emanating from mining and milling waste piles or other low-level radioactive waste discharges have been attributed to nearly irreversible sorption of ^{137}Cs onto the clay fraction in soils (Brady et al., 2002). Sorption onto clays may occur onto high-affinity or low-affinity sites on illite (Poinssot et al., 1999), creating sub-reservoirs of what may be lower-mobility and higher-mobility ^{137}Cs. Presence of other clay minerals with higher ion-exchange coefficients, pH variations, or organic matter could lead to higher mobility (Chappell, 1999), as could changes in the redox state of the soil (Thomson et al., 2002). In addition,

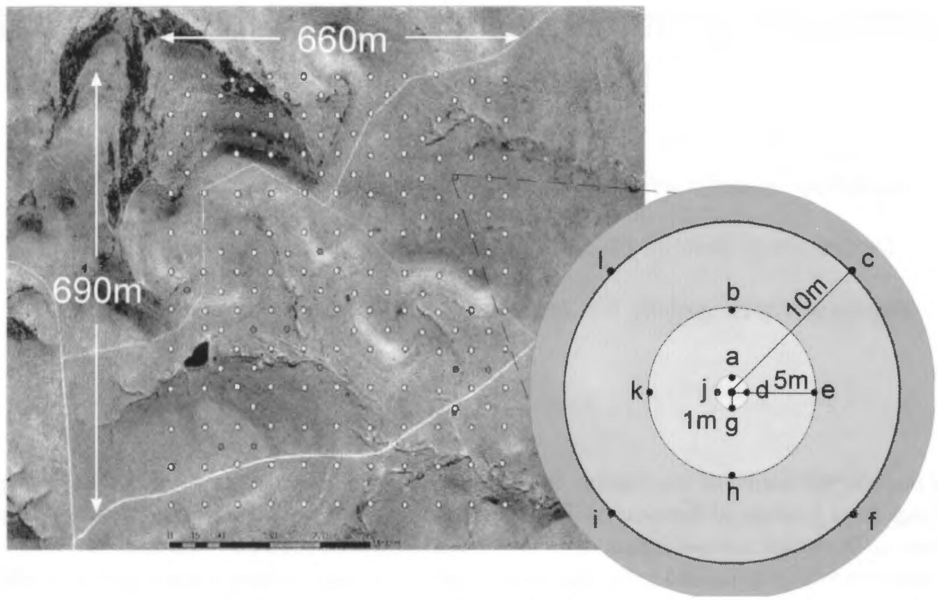

FIGURE 2–2. Landscape grid sampled for this study, including the scheme for micro-scale sampling at the 12 selected points.

because ion exchange is competitive, a change in the dominant cation (especially an increase in the dissolved Na content) has the potential to mobilize sorbed ^{137}Cs (Poinssot et al., 1999). The distribution coefficient for ^{137}Cs is apparently not strongly affected by the amount of humic acid present, suggesting that complexation by dissolved organic carbon is not an important transport mechanism (Langmuir, 1997). Previous work delineating potentially mobile reservoirs of ^{137}Cs have suggested that Chernobyl-sourced ^{137}Cs is probably affiliated minimally with clay minerals in exchangeable sites, and more with organic matter (or possibly fuel particles that liberate ^{137}Cs during the step that liberates material associated with oxidizable phases), and with phases dissolved using 7 M nitric acid at 80°C (Salbu et al., 1994).

At Konza, ^{137}Cs (half life of 30.2 years) and ^{239}Pu (half-life of 24,000 years) radioisotope activities show similar exponential declines with depth (fig. 2–3), which supports the idea that the decrease in the activity of ^{137}Cs with depth is not a function of its short half life. Preliminary findings suggest that Cs at Konza (stable Cs, mass 133, and, by proxy, radioactive Cs, mass 137) has an affinity for SOC. Most (95%) of the stable Cs resides in the weathering-resistant portion of the soil, as isolated using a sequential-extraction technique modified from Tessier et al. (1979). Most of the remaining mobile stable Cs is affiliated with oxidizable phases, probably mostly SOC (fig. 2–4). Enrichment in Cs concentration in SOC with depth occurs, as demonstrated by normalizing Cs concentration to Cs concentration in the oxidizable fraction of the soil and to the shallowest sample (fig. 2–4). Cs content in the other extractable reservoirs (reducible, acid-soluble, and easily exchangeable fractions) shows much less change with depth (fig. 2–4). An exponential decline in ^{137}Cs and SOC with depth contrasts with an exponential increase with depth of stable Cs affiliated with oxidizable phases (SOC).

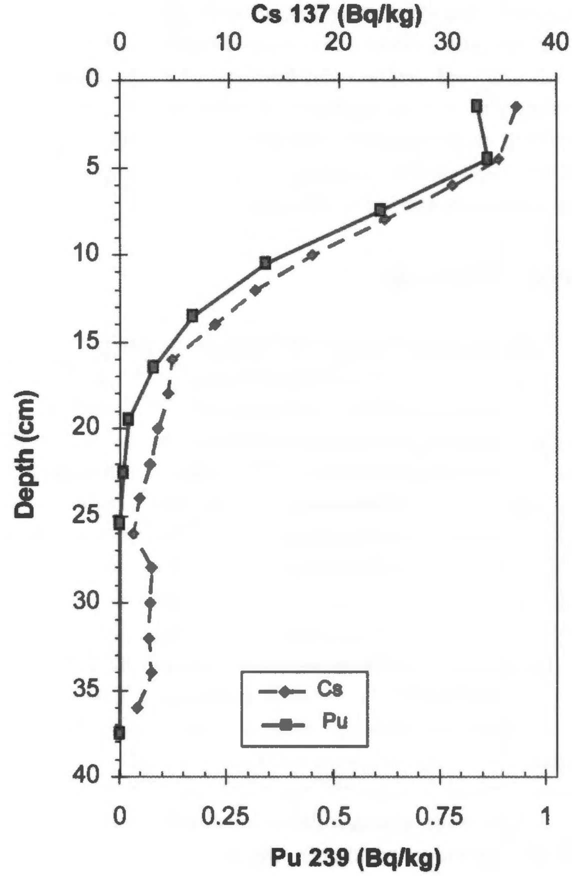

FIGURE 2–3. ^{137}Cs and ^{239}Pu radioisotope distributions in a soil profile from the upland of the landscape grid.

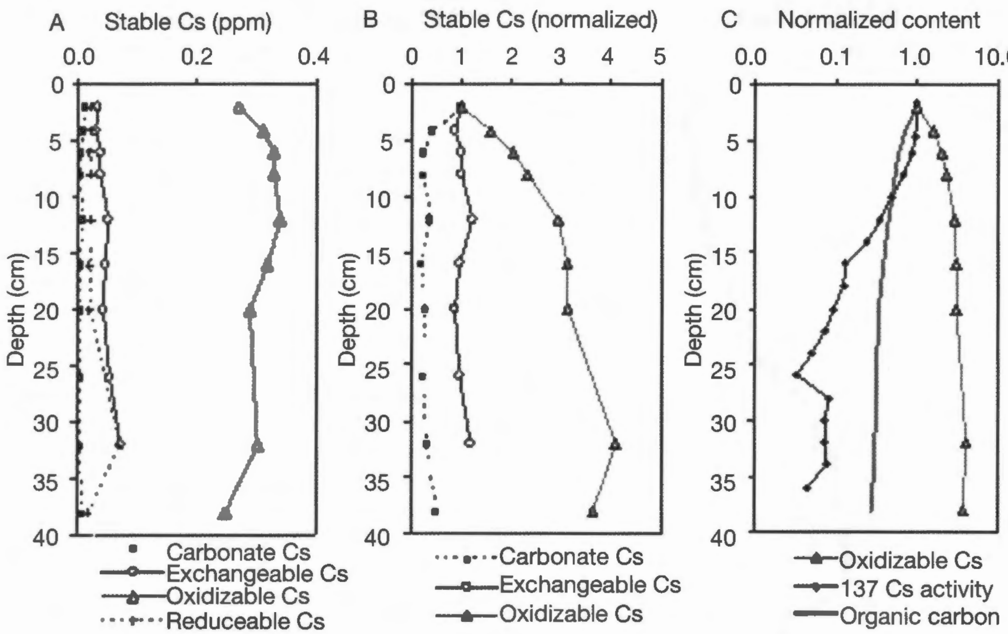

FIGURE 2–4. Cs distribution with depth. A) Cs concentration with depth; B) Cs content normalized to shallowest sample and to the extract reservoir (% SOC for oxidizable soil components; % clay for exchangeable Cs); C) log-normal distributions of ^{137}Cs activity normalized to shallowest sample, SOC normalized to shallowest sample, and stable Cs concentration normalized to shallowest sample and to SOC.

Stable Isotopes

Carbon consists of two naturally occurring stable isotopes, ^{12}C (98.89%) and ^{13}C (1.11%), which are usually expressed in ratio form (^{13}C/^{12}C) or as δ^{13}C (‰). Plants discriminate against ^{13}C during photosynthesis such that the tissue is depleted in ^{13}C relative to the atmosphere (O'Leary, 1981). Grasses fix carbon from atmospheric CO_2 by one of three pathways, the Calvin-Benson (C_3), Hatch-Slack (C_4), and Crassulacean acid metabolism (CAM). Because CAM is employed by desert plants, it is not widespread in the Great Plains. Each of these photosynthetic pathways results in differing levels of discrimination against ^{13}C (Smith and Epstein, 1971). Values of δ^{13}C from C_4 plants produce a range from about -17‰ to -10‰, with an average of -13‰, whereas C_3 plants range from -32‰ to -20‰ and average -27‰ (Cerling et al., 1989; Boutton, 1991). SOC represents an integration of the isotopic inputs from the various sources within a plant community (Troughton et al., 1974; Balesdent and Mariotti, 1996), the value of which approximates the photosynthetic pathway of the dominant species (Stout and Rafter, 1978; Nadelhoffer and Fry, 1988). After removal of carbonates using IN HCl (16 hours @ room temperature), samples were assayed for stable carbon and carbon content using a ThermoFinnigan Delta Plus mass spectrometer and elemental analyzer system.

Konza Tallgrass Prairie is dominated by C_4 grasses, which translate their characteristic isotopic signal to the soil. Soil surface samples from the landscape grid averaged -17.5‰. None of the variance yet appears to be explained, however, by slope and aspect. Senescent grass samples averaged -15.8‰, reflecting a clear C_4 signal. The isotopic profile of an upland soil indicates surface values of about -18‰, but with an increase to about -14‰ at 10–20-cm depth (fig. 2–5). Such ^{13}C depletion in modern surface soils has been widely recognized (e.g., Kelly et al., 1991; de Freitas et al., 2001; Torn et al., 2002) and is due in part to historic, fossil-fuel-induced ^{13}C depletion of the CO_2 in the atmosphere (Boutton, 1996). Consequently, the larger value at 10–20 cm is more representative of the long-term average of the stable grassland vegetation. Likewise, isotopic depletion of soil-surface SOC within the landscape grid likely explains the disparity between the mean of the soil surface and grass sample δ^{13}C. Isotopic values decrease downward in the core to a low of about -24.0‰ (circa 1.6 m), reflecting cool-season C_3 plant domination of the late-Wisconsinan central Great Plains environment. Increase in δ^{13}C values at the core base reflects presence of a middle-Wisconsinan buried soil, the Gilman Canyon Formation soil, one characterized regionally by its C_4 grass-dominated environment (Johnson and Willey, 2000).

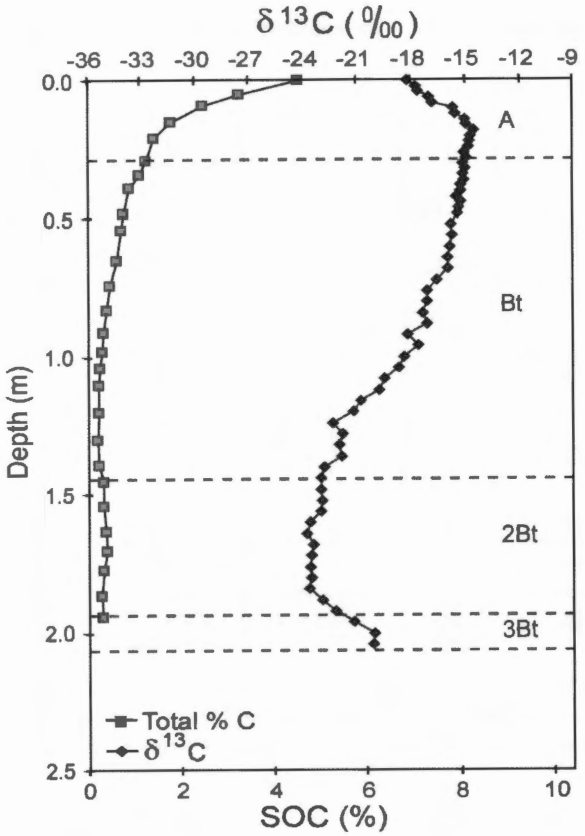

FIGURE 2–5. Distribution of $\delta^{13}C$ and SOC within an upland soil profile.

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Part 2. The South Fork of the Big Nemaha River, Southeastern Nebraska

Late Quaternary Paleoecology of the Eastern Great Plains

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The paleoecological record of much of North America is largely based on studies of fossil pollen from lakes and wetlands. Such depositional basins are widespread, especially in glaciated regions, but pollen sites are much less common on the Great Plains beyond the Wisconsinan glacial margin. Of 328 pollen sites listed for eastern North America (Webb et al., 1993), only one (Swan Lake in the Nebraska Sandhills) is shown for the region from Nebraska to southern Oklahoma. In these regions, other types of paleoecological information have been developed from different depositional environments. Plant macrofossils, where present, can yield species-level identifications and definite local presence of those plants. Phytoliths provide information of useful subgroups of grasses that are indistinguishable from pollen, and $\delta^{13}C$ analyses can be transformed into percentages of C_4 vs. C_3 plants. Both these techniques can be applied to either soils or alluvial sediments, and both can be helpful in distinguishing prairie from forest and distinguishing among various types of prairie. The $\delta^{13}C$ can also be analyzed from cave speleothems, if caves are nearby, giving very high time-resolution. This trip provides the opportunity to examine the potential of stream cutbanks as sources of these paleoecological data and to discuss how these recent paleoecological tools correlate with one another.

The purpose of the afternoon portion of the trip is to show you two sections near Du Bois, Nebraska, that we have studied and to discuss the results. The deposits were sampled from cutbanks along the river. We will examine the lithologic units, look at the fossiliferous deposits, screen them to show you examples of the macrofossils, and discuss their interpretation. At first glance, these deposits would seem to be of questionable use, if not downright useless. They are generally discontinuous spatially, separated by sediments barren of fossils; they are discontinuous snapshots in time; and they would seem particularly prone to reworking. However, by choosing the right sample sites, these problems can largely be overcome, and the benefits can be substantial. First, while it is true that most of the sediments are barren of plant macrofossils, if the creek is permanent and not in a stretch that loses water (in karst terrain, for example), organic-rich deposits near or even below creek level may be present. Second, though the deposits generally do represent a very short time interval, in the streams we have worked on, different ages are exposed in the stream's cutbanks over distances of tens of meters to a few kilometers upstream and downstream. We have put together a series of snapshots representing much of the Holocene. Third, it is true that reworking is possible, but with careful work and site selection, we have determined that there is no evidence of wholesale reworking, and only a few cases of reworking of individual plant remains were recorded. Preservation of very delicate plant remains effectively rules out reworking in many cases. One fairly common exception is spruce needles, which are extremely durable. We occasionally do find needles in middle Holocene deposits, but they are broken and corroded, in the midst of other well-preserved delicate plant macrofossils. Furthermore, we try to minimize reworking by sampling fairly close to the headwaters of the stream; thus, if reworking did occur, it would not involve transport from a distant biome.

There are several advantages to using these organic-rich alluvial deposits. They do contain pollen (Chumbley et al., 1990), and comparisons with the closest lake deposits indicate that the alluvial pollen signal correlates well with the lacustrine record (Baker et al., 1992). A big advantage is that the alluvial deposits contain a much more diverse plant macrofossil record than is found in lacustrine and wetland sediments (Baker et al., 1996; Baker et al., 2002). These fossils represent the whole range of habitats in the drainage basin, including upland, floodplain, weedy disturbed habitats, wetlands that occur on the floodplain, and in the stream (fossils of aquatic plants) (Baker, 2000; Baker et al., 1996; Baker et al., 2000; Baker et al., 2002). In many cases there are well over 200 taxa, and the preservation is often excellent, allowing specific identification. Wood or other upland macrofossils are dated from each section to provide the chronology. Thus, these deposits give a picture of the landscape that has very high ecological resolution. The other big advantage is that the phytolith and isotopic records test and refine the pollen and plant macrofossil interpretations. The climatic record derived from these data is then compared with the stream activity derived from the alluvial stratigraphy and lithology (see Mandel, 1996; Mandel and Bettis, 2001).

On the way back to Lawrence, we will make a short stop at Muscotah Marsh, an important locality cored and analyzed in the early 1970's by Grüger (1973). We will compare and contrast the results from the study sites in the South Fork of the Big Nemaha River with the paleobotanical record at Muscotah.

Methods

We collected ca. 1-liter samples in resealable plastic bags from several organic-rich zones near the base of three large cutbanks along the South Fork of the Big Nemaha River near Du Bois, Nebraska. These were subsampled by Baker for plant macrofossils and by Fredlund for phytoliths, pollen, and stable carbon isotopes. Samples were prepared for phytolith analysis by (1) removal of carbonates with HCl, (2) removal of clays with 0.1 N sodium pyrophosphate, (3) oxidation of the residue with 30% hydrogen peroxide, and (4) isolation of biogenic silicates using heavy-liquid ($ZnBr_2$ at 2.35 specific gravity) as described in Fredlund and Tieszen (1997). Pollen samples were processed using heavy liquid flotation (Johnson and Fredlund, 1985). Pollen and phytoliths were identified using reference materials at the Geography Department, University of Wisconsin–Milwaukee. Plant macrofossils were washed through 0.5- and 0.1-mm sieves, picked by hand, and identified using the reference collection at the Geoscience Department, University of Iowa.

The stable carbon isotope analyses of bulk organic matter were carried out at the Isotope Ratio Mass Spectrometer facility in the Biology Department at Augustana College, Sioux Falls, South Dakota, as described in Fredlund and Tieszen (1997). Pretreated samples were combusted in a Carlo Erba CHN analyzer coupled to an SIRA-10 Isotope Ratio Mass Spectrometer fitted with a special triple trap to isolate cryogenically and purify the CO_2. Lab standard and random replicates were run to ensure precision better than $0.2^{\circ}/_{\circ\circ}$. Isotopic ratios are expressed as per-mil deviations from the PDB standard.

Stop 1: Farwell Site

This site (Stop 1 in fig. 2–6) is an entrenched channel on the South Fork of the Big Nemaha River, about 1.5 km east of the town of Du Bois (ca. 40°01'15" N. Lat., ca. 96°01'15" W. Long.). It is within about 20 km of its source in northeastern Kansas. At this stop, we will park on a bridge and view Holocene and mid-Wisconsinan alluvial sediments exposed in numerous cutbanks along a stretch of the river (Mandel, 1996; Mandel and Bettis, 2001). The Holocene sediments are contained in four members of the DeForest Formation: Gunder (oldest), Roberts Creek, Honey Creek, and Camp Creek (fig. 2-7). The DeForest Formation is a lithostratigrahic unit containing all the fine-grained Holocene alluvium in Iowa (Bettis, 1995). Recent studies have extended the DeForest Formation into eastern Nebraska (Dillon, 1992; Mandel, 1994; Mandel and Bettis, 1995, 2001). At Farwell, the DeForest Formation is laterally inset against the Severance Formation (fig. 2-8). The Severance formation consists of oxidized, sandy alluvium and is Wisconsinan in age (Mandel and Bettis, 2003). The DeForest and Severance Formations are distinguished on the basis of their lithology and geomorphic position, and their chronology has been determined by radiocarbon dating. Plant macrofossils associated with the alluvial fills typically occur at or near water level.

The vegetation of this part of its drainage basin was tallgrass prairie prior to European settlement, with some riparian forest along the stream. Although the prairie has largely been lost to cultivation, its composition was probably typical diverse prairie vegetation somewhat similar to what we saw at Konza Prairie. The riparian forest that remains includes *Acer negundo* (boxelder), *Acer saccharinum* (silver maple), *Fraxinus pennsylvanica* (green ash), *Populus deltoides* (cottonwood), *Salix* spp. (willow), and *Ulmus americana* (American elm).

Pollen sites from central Nebraska (Wright et al., 1985) and the northern Great Plains (Barnosky et al., 1987; Grimm, 2001) indicate that prairie was present throughout the Holocene, but that there were significant fluctuations. In the early Holocene, open deciduous forest on the northeastern Great Plains gave way to prairie, and there is strong evidence of recurring drought in the mid-Holocene (Grimm, 2001; Laird et al., 1998). The only pollen site on the eastern edge of the central Great Plains is Muscotah Marsh (Grüger, 1973), our last stop, where the Holocene was compressed and not well dated.

Our work in the South Fork of the Big Nemaha River valley provides pollen, plant macrofossil, phytolith, and carbon-isotopic data to reconstruct the Holocene vegetation. Because of the coarse time resolution, we cannot resolve short-term climatic changes, but the cutbank sediments give a far better ecological resolution of a snapshot of the past vegetation than marsh or lake sediment samples.

The Farwell site (fig. 2-8) is part of a sequence of sites along the South Fork of the Nemaha River that was analyzed for the plant macrofossil dataset. Uncalibrated radiocarbon ages for these sites are shown in table 2-1. Figure 2-9 shows the pollen percentages from these sites, and figs. 2-10 through 2-12 show macrofossils of upland trees, riparian trees, prairies, and selected weeds. Macrofossils representing aquatic, wetland, other weed, forest floor herb, and shrub and vine habitats are plotted and discussed in Baker (2000). Figure 2-13 presents the

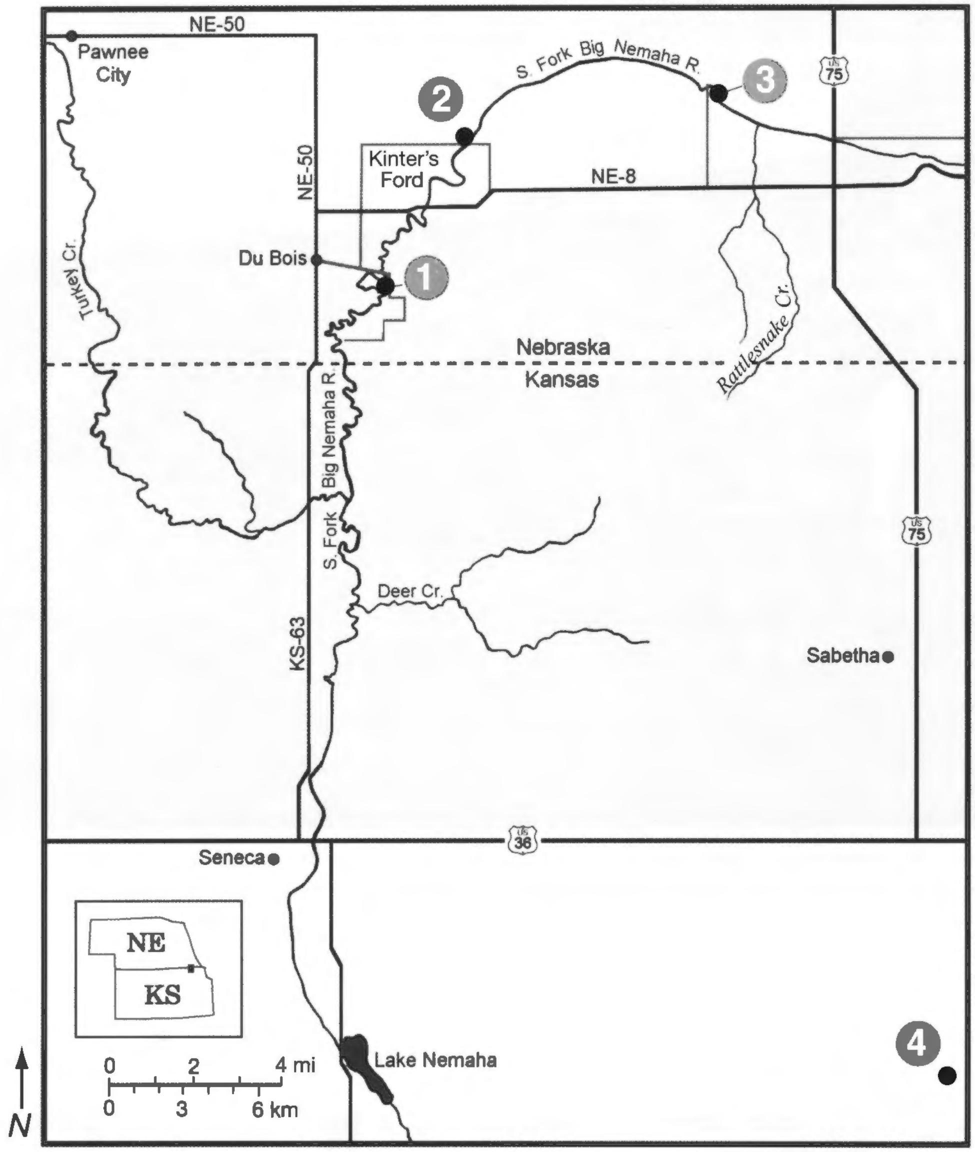

FIGURE 2–6. Map of field-trip stops.

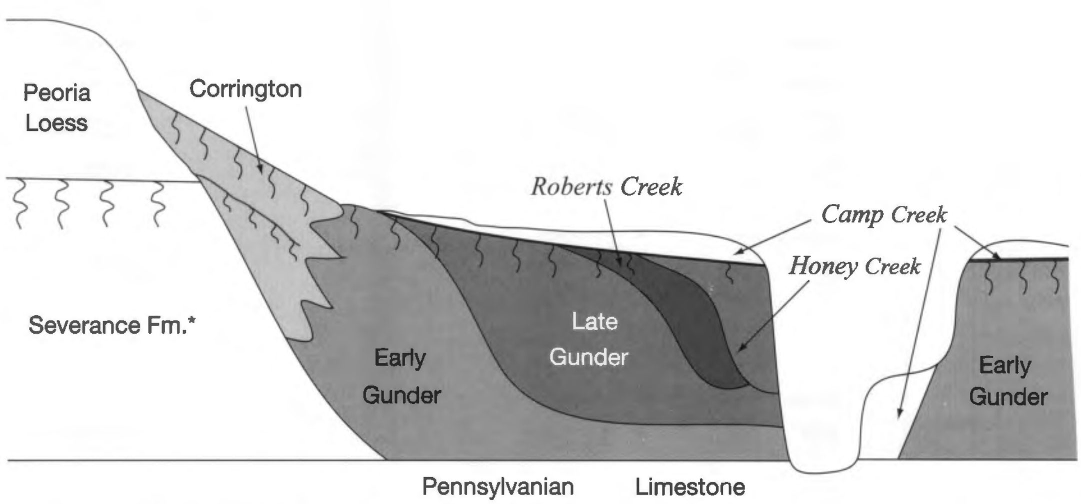

FIGURE 2–7. Stratigraphic and temporal relationships of Wisconsinan formations and units of the Holocene DeForest Formation along the South Fork Big Nemaha River (from Mandel and Bettis, 2001).

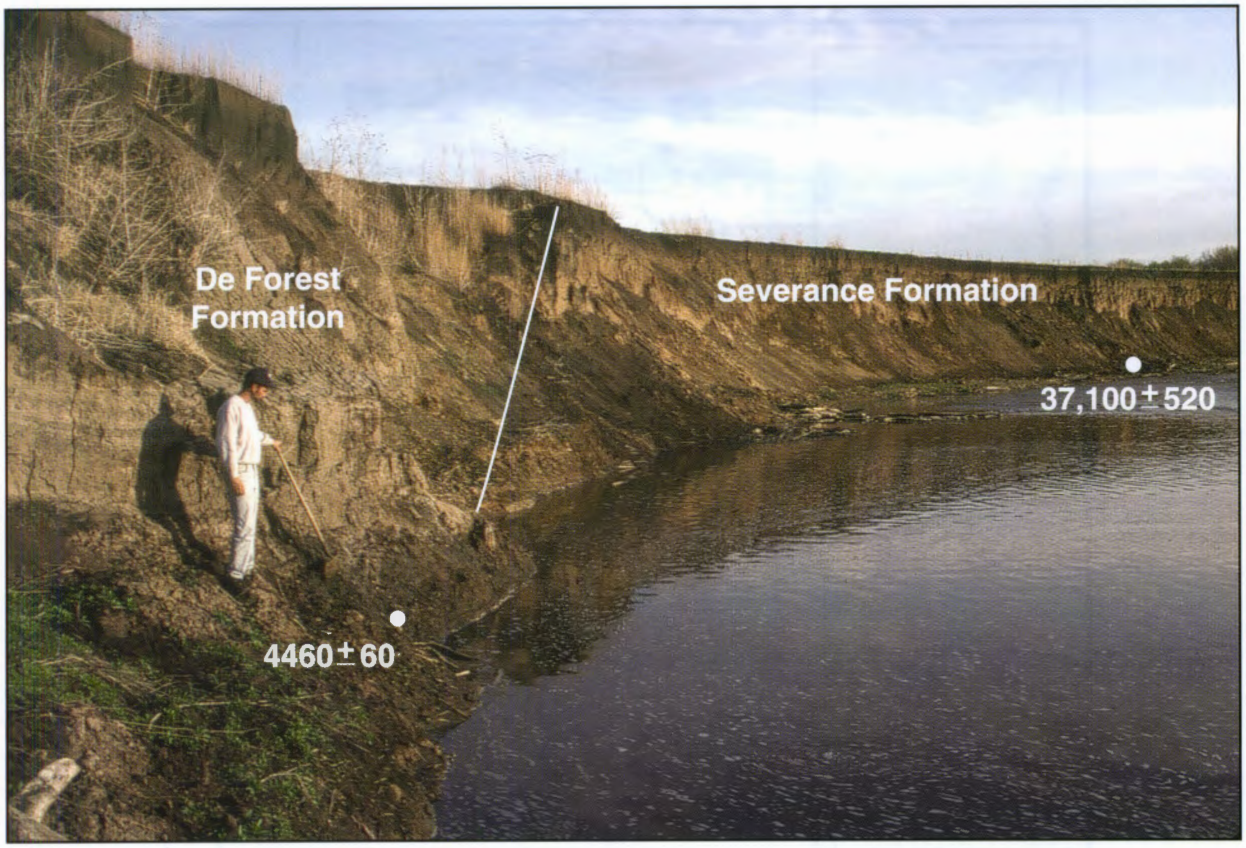

FIGURE 2-8. View of the cutbank at the Farwell site (Stop 1). Radiocarbon ages were determined on plant macrofossils. The white line marks the boundary between the DeForest and Severance Formations.

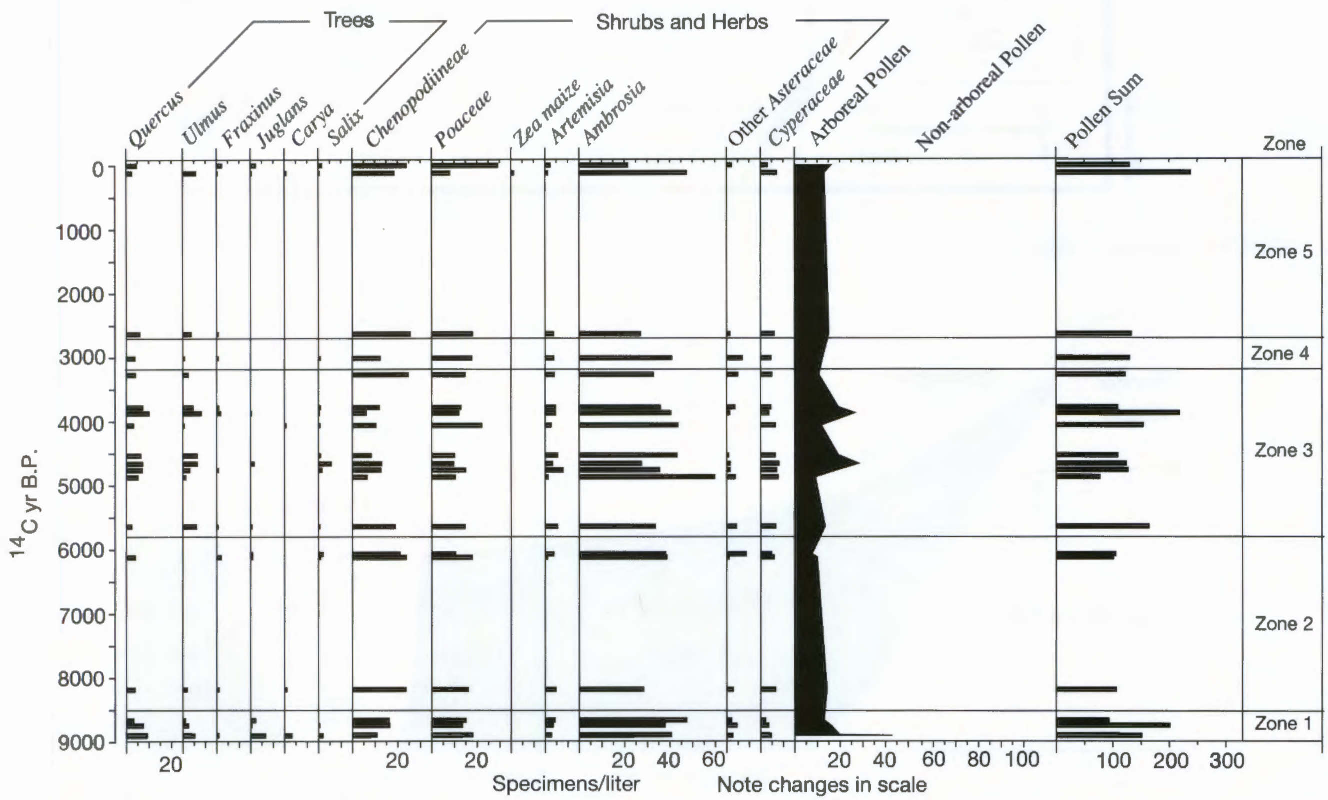

FIGURE 2-9. Pollen spectra for sediment samples collected from cutbanks along the South Fork of the Big Nemaha River near Du Bois, Nebraska. The pollen analysis was conducted by Glen Fredlund.

TABLE 2–1. Radiocarbon ages for study sites in the South Fork Big Nemaha River valley (from Mandel and Bettis, 2001).

Stop No. and Locality	Stratigraphic Unit	Material Assayed	Depth (m)	Uncorrected Age (yrs B.P.)	$\delta^{13}C$	Corrected Age (yrs B.P.)	Lab No.
STOP 1							
Farwell							
25PW63	Late Gunder Mbr.	Wood	6.50–6.70	NR	-28.0	4060 ± 70	ISGS-3416
25PW63	Late Gunder Mbr.	Wood	7.30–7.40	4,460 ± 60	-27.5	4420 ± 60	Tx-9017
25PW63	Late Gunder Mbr.	Charcoal	3.23–3.28	3,150 ± 60	-25.1	3140 ± 60	Tx-9015
25PW63	Severance Fm.	Plant frag.	8.72–8.73	NR	NR	37,100 ± 520	CAMS-41681
25PW63	Severance Fm.	Charcoal	8.81–8.82	NR	NR	53,110 ± 2870^	CAMS-33983
Floodplain~	Camp Creek Mbr.	Wood	3.75–3.90	NR	-28.6	Modern	ISGS-3415
25PW64	Early Gunder Mbr.	Charcoal	1.30–1.70	4,460 ± 70	-26.0	4440 ± 70	Tx-8952
25PW64	Early Gunder Mbr.	Charcoal	1.30–1.70	3,920 ± 70	-25.0	3920 ± 70	Beta-95298
25PW64	Late Gunder Mbr.	Wood	7.69–7.90	NR	-27.9	4660 ± 70	ISGS-3417
25PW64	Late Gunder Mbr.	Wood	7.75–7.98	NR	-27.6	4530 ± 70	ISGS-3418
25PW65	Late Gunder Mbr.	Wood	8.14–8.30	NR	-27.0	4760 ± 70	ISGS-3414
25PW65	Late Gunder Mbr.	Charcoal	1.60–1.75	3,670 ± 180	-26.1	3650 ± 180	Tx-9016
25PW65	Honey Creek Mbr.	Charcoal	4.60–4.75	1,880 ± 40	-26.8	1850 ± 40	Tx-9013
25PW65	Severance Fm.	Soil	2.00–2.10	NR	-23.0	15,110 ± 110	ISGS-4468
STOP 2							
Old Bridge	Early Gunder Mbr.	Wood	12.11–12.31	NR*	-27.7	8890 ± 70	ISGS-3427
Old Bridge	Early Gunder Mbr.	Wood	12.11–12.25	NR	-26.7	8890 ± 70	ISGS-3426
Old Bridge	Early Gunder Mbr.	Wood	12.59–12.70	NR	-25.6	8740 ± 70	ISGS-3430
Old Bridge	Early Gunder Mbr.	Wood	12.05–12.27	NR	-26.9	8630 ± 70	ISGS-3436
Old Bridge	Early Gunder Mbr.	Wood	11.72–11.93	NR	-28.2	8180 ± 70	ISGS-3431
Old Bridge	Honey Creek Mbr.	Wood	9.22–9.33	NR	-28.0	3870 ± 70	ISGS-3429
Old Bridge	Honey Creek Mbr.	Wood	10.78–10.95	NR	-27.0	2630 ± 70	ISGS-3437
Old Bridge	Honey Creek Mbr.	Wood	8.55–8.66	NR	-26.6	1910 ± 70	ISGS-3424
STOP 3							
Miles Fan	Corrington Mbr.	Soil	1.30–1.40	2,920 ± 60	-12.9	3120 ± 50	Tx-8945
Miles Fan	Corrington Mbr.	Soil	3.60–3.70	10,250 ± 120	-12.9	10,450 ± 120	Tx-8944
Miles Fan	- - - - - - - - - -	Peat	9.90–10.0	23,530 ± 310	-27.0	23,490 ± 310	Tx-8946
Miles Fan	- - - - - - - - - -	Peat	10.3–10.4	24,690 ± 410	-27.3	24,640 ± 410	Tx-8947
Miles Fan	- - - - - - - - - -	Peat	10.6–10.7	27,620 ± 550	-27.4	27,580 ± 550	Tx-8948
Miles Fan**	- - - - - - - - - -	Wood	ca. 10.0	NR	-25.1	19,920 ± 240	ISGS-4681
Miles Fan**	- - - - - - - - - -	Wood	ca. 10.0	NR	-24.5	20,290 ± 120	ISGS-4680
OTHER LOCALITIES							
DuBois Quarry	Severance Fm.	Soil	2.36–2.46	18,780 ± 140	-22.4	18,830 ± 140	Tx-9308
DuBois Quarry	Severance Fm.	Soil	4.16–4.26	25,730 ± 360	-28.5	25,670 ± 360	Tx-9309
DuBois Quarry	Severance Fm.	Wood	7.62–7.82	NR	NR	33,257 ± 1096	ISGS A-0020
Kinter's Ford	Early Gunder Mbr.	Soil	2.90–3.00	6,930 ± 70	-16.5	7070 ± 70	Tx-8943
Kinter's Ford	Early Gunder Mbr.	Wood	7.05–7.20	NR	-28.6	6120 ± 70	ISGS-3405
Kinter's Ford	Early Gunder Mbr.	Wood	7.15–7.38	NR	-28.5	6070 ± 70	ISGS-3400
Kinter's Ford	Early Gunder Mbr.	Wood	6.70–6.90	NR	-28.3	5640 ± 70	ISGS-3401
Kinter's Ford	Early Gunder Mbr.	Wood	4.44–4.60	NR	-27.0	4870 ± 70	ISGS-3398
Kinter's Ford	Early Gunder Mbr.	Soil	1.80–1.90	4,630 ± 110	-13.3	4780 ± 110	Tx-8942
Kinter's Ford	Honey Creek Mbr.	Wood	5.22–5.44	NR	-27.3	3780 ± 70	ISGS-3402
Kinter's Ford	Honey Creek Mbr.	Wood	6.48–6.80	NR	-26.4	3300 ± 70	ISGS-3446
Kinter's Ford	Roberts Creek Mbr.	Wood	3.49–3.70	NR	-27.5	3270 ± 70	ISGS-3399
Kinter's Ford	Honey Creek Mbr.	Wood	6.56–6.70	NR	-28.3	3010 ± 70	ISGS-3404
Kinter's Ford	Honey Creek Mbr.	Wood	6.53–6.64	NR	-25.9	2800 ± 70	ISGS-3403
Kinter's Ford	Honey Creek Mbr.	Wood	7.23–7.55	NR	-26.4	2190 ± 70	ISGS-3452
Kinter's Ford	Camp Creek Mbr.	Wood	6.56–6.79	NR	-27.5	370 ± 70	ISGS-3453
25PW83	Late Gunder Mbr.	Wood	8.30–8.50	NR	-26.4	3590 ± 70	ISGS-3450
25PW83	Late Gunder Mbr.	Charcoal	200–205	2,290 ± 70	-26.2	2270 ± 70	Tx-9014

*NR = Not Reported. ^Age rejected. **Sample was collected ca. 100 m downstream from the Miles Fan.
~Sample was collected on the west side of the South Fork of the Big Nemaha River, immediately west of site 25PW63.

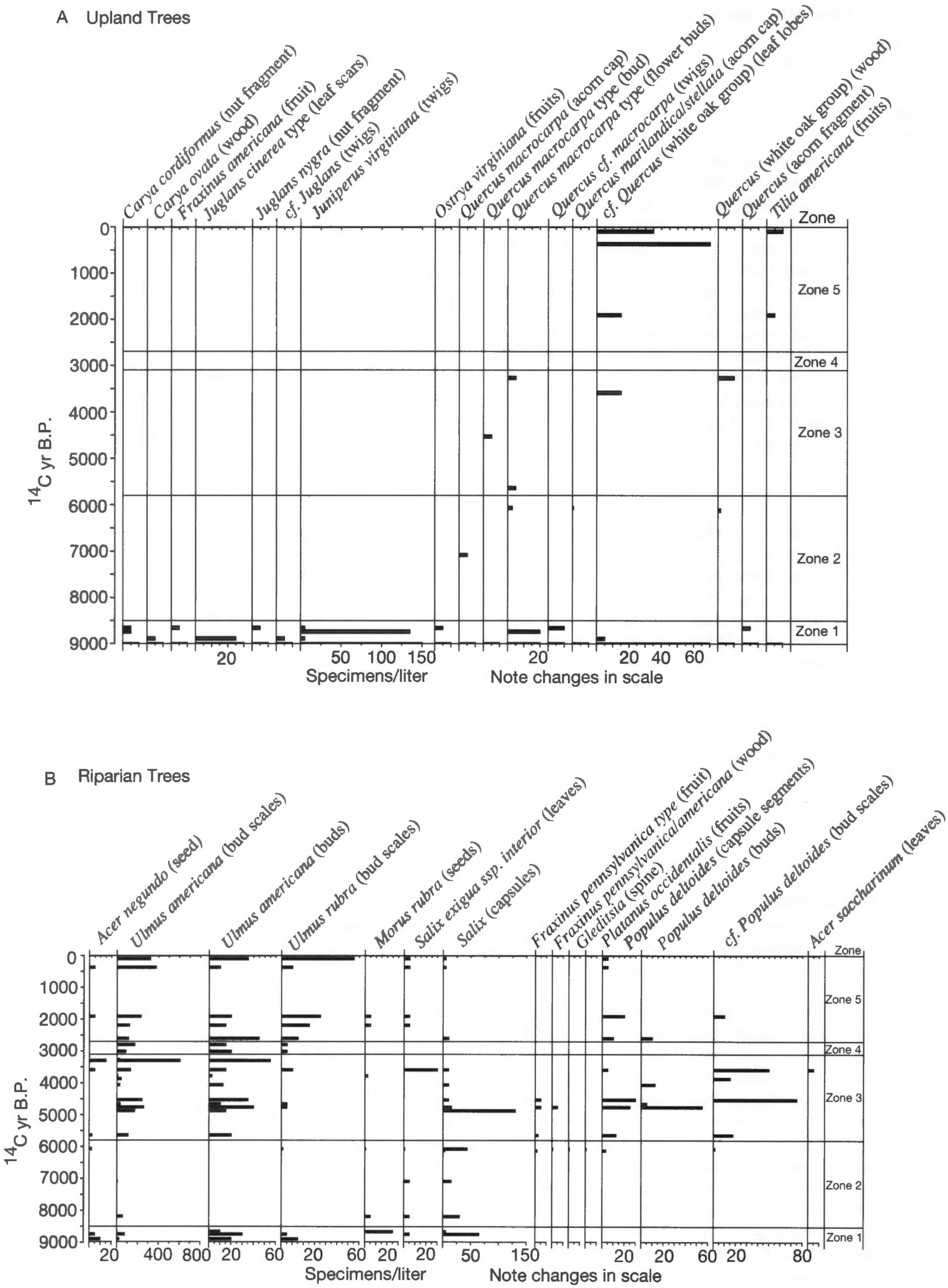

FIGURE 2-10. Summary of plant macrofossils derived from trees. The macrofossils were recovered from Holocene sediments exposed in cutbanks along the South Fork of the Big Nemaha River near Du Bois, Nebraska.

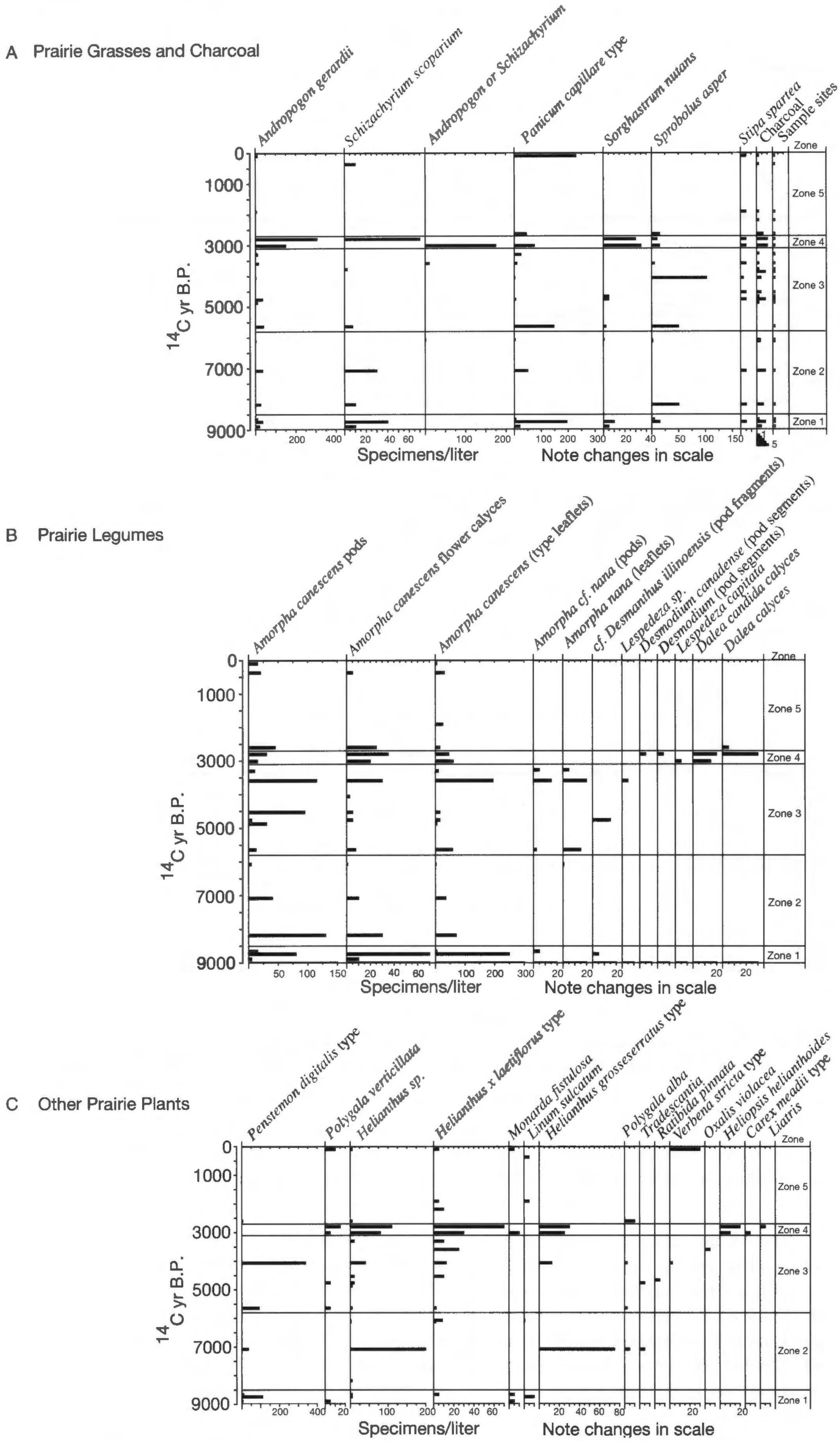

FIGURE 2-11. Summary of plant macrofossils derived from prairie species. The macrofossils were recovered from Holocene sediments exposed in cutbanks along the South Fork of the Big Nemaha River near Du Bois, Nebraska.

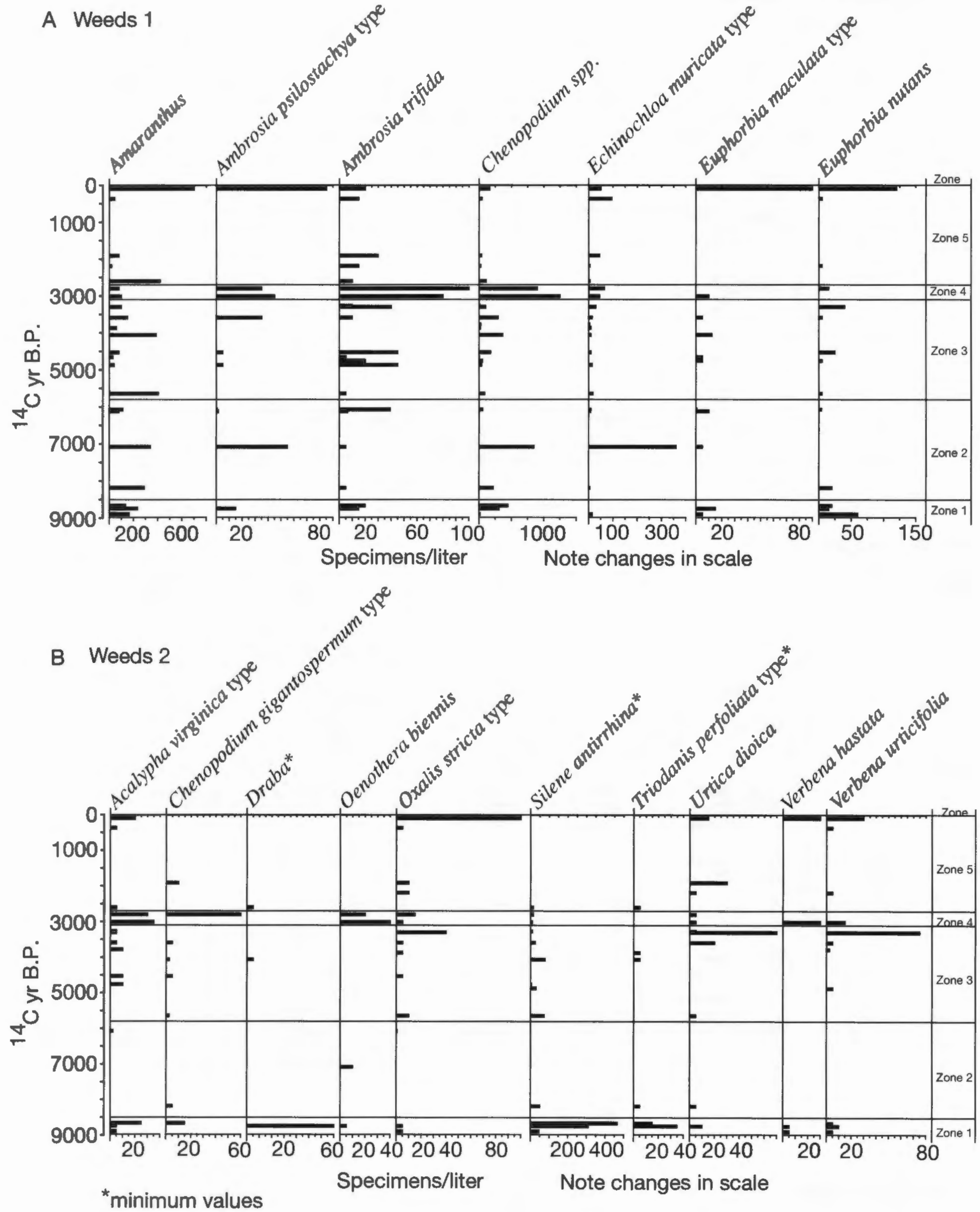

FIGURE 2-12. Summary of plant macrofossils derived from weeds. The macrofossils were recovered from Holocene sediments exposed in cutbanks along the South Fork of the Big Nemaha River near Du Bois, Nebraska.

phytolith and $\delta^{13}C$ data. All are arranged vertically by radiocarbon age of each site, and the macrofossils are grouped by plant habitat and physiognomy. Five zones are recognized, mainly on the basis of macrofossils; all ages are expressed in uncalibrated radiocarbon years. The basal zone (Zone 1, 9000–8500 yrs B.P.) has a diverse mix of macrofossils of deciduous tree taxa including hickories, white ash, butternut type, walnut, ironwood, and bur oak. Riparian trees including boxelder, elms, willows, and mulberry. Despite the diverse mix of forest trees, prairie also is well represented, with the major tall prairie grasses and important legumes, as well as several other taxa. Pollen of deciduous trees is relatively low, and typical prairie pollen types including grass, chenopods, and ragweed are all high. The $\delta^{13}C$ values are highly negative, but they increase upwards, and the highest percentage of C_3 phytoliths occurs in this zone, although both C_3 and C_4 types are present.

The pollen record for Zone 1 suggests that prairie was prominent on flat uplands, but the combined records

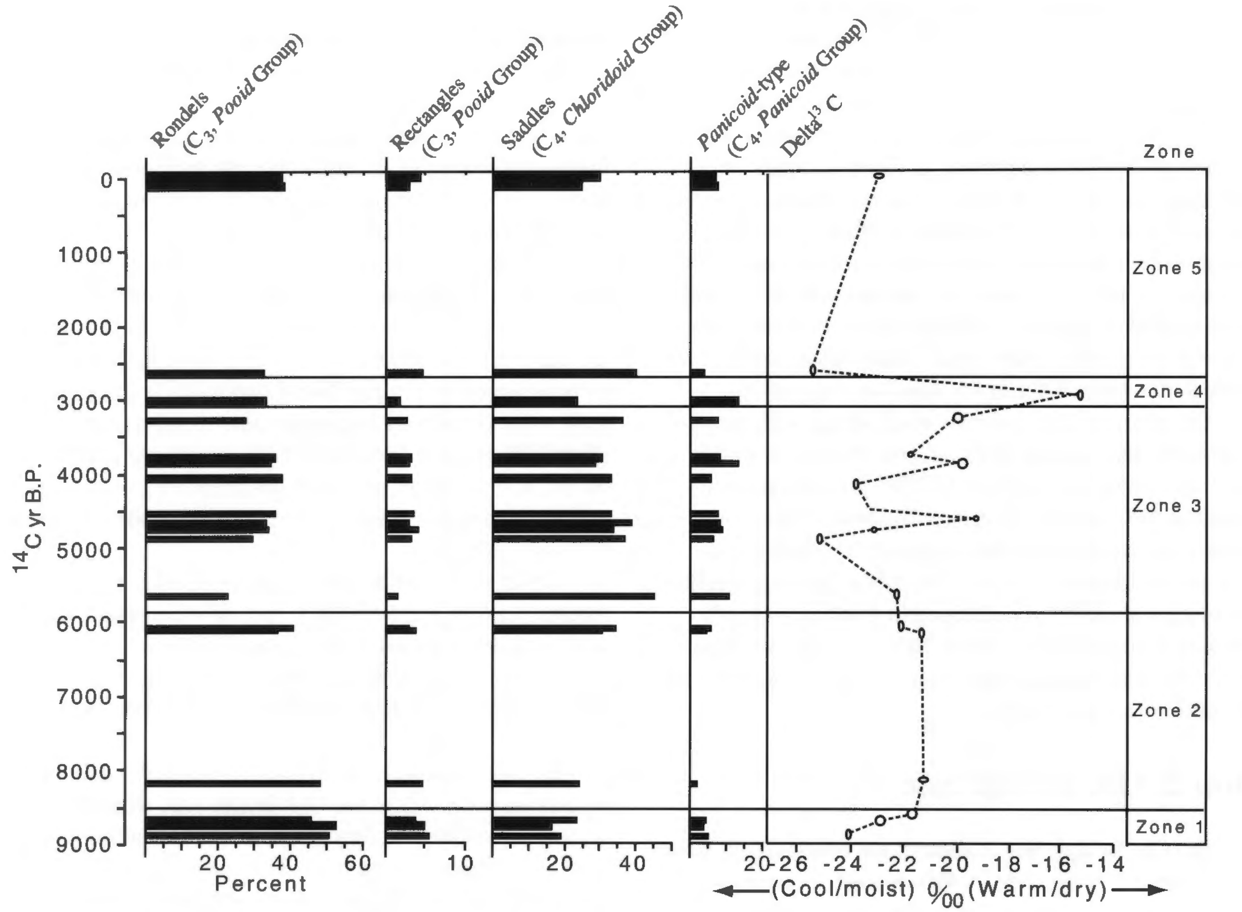

FIGURE 2-13. Phytolith and stable carbon isotope records for sediment samples collected from cutbanks along the South Fork of the Big Nemaha River near Du Bois, Nebraska.

indicate that a mesic forest was still widespread on valley walls, in ravines, and probably on the floodplain. Riparian elements would likely be confined to areas close to the river. The macrofossils indicate that C_4 grasses are abundant in this zone (fig. 2-11), yet the strong presence of C_3 grasses is indicated by phytoliths (fig. 2-13), and the $\delta^{13}C$ values indicate mixed C_3 and C_4 grasses (fig. 2-13).

Only a few sites represent Zone 2; nevertheless, a noteworthy change from Zone 1 is apparent. Macrofossils of most upland and riparian trees have disappeared. Arboreal pollen also declined. Elements of prairie continue and some weeds increase in abundance. The $\delta^{13}C$ values increase about 2 per mil and remain steady. Phytoliths of C_4 grasses increase at the top of this zone.

These changes suggest a distinctly drier climate between 8500 and 5800 yrs B.P. Studies of the 1930's droughts show that the prairie vegetation became much less dense and more patchy, giving room for weeds to become established (Weaver, 1968; Tomenek and Hulett, 1970). Streams were generally undergoing aggradation early in this zone, becoming stable towards the end, and alluvial fans were aggrading throughout Zone 2.

A major change in Zone 3 (5800–3100 yrs B.P.) is that there is a marked increase in the abundance of riparian trees, including boxelder, elm, willow, green ash, and cottonwood. Pollen and plant macrofossils of prairie and weedy plants continue to be present in Zone 3, and only oak was present as a macrofossil. Phytoliths of C_4 plants are slightly more prominent than in Zone 2, and the $\delta^{13}C$ values fluctuate significantly during this interval.

Zone 4 (3100–2700 yrs B.P.) marks another distinct interval. Upland trees remain absent, and all riparian trees disappear except elm. In contrast, most other elements increase strikingly. These include many prairie, weed, wetland, and aquatic species. Charcoal is abundant in this zone, and $\delta^{13}C$ values are at their peak, indicating an abundance of C_4 grasses. Phytoliths of the chloridoid group of C_4 plants (which include grama grasses and buffalo grass of the midgrass-shortgrass prairie) decline, but those of panicoid C_4 plants increase.

The increased abundance of many macrofossil taxa and the increase in charcoal suggests that fires were more common at ca. 3100–2700 yrs B.P., allowing more runoff and increased deposition of plant remains. The disappearance of all riparian elements except elm, which is a late-successional tree, suggests that climatological and hydrological factors prevented the establishment of early succession on sandbars of cottonwood and willow for a few hundred years. One hypothesis is that flow was generally low in spring and early summer, allowing spring establishment on sandbars. Occasional large floods may have occurred in Zone 3, providing fresh sand for willow, cottonwood, and other early successional plants to colonize reworked sandbars. In Zone 4, floods may have been smaller and droughts more frequent, favoring late-successional elms. The absence of new sandbars would make it more difficult for willow and cottonwood to germinate. The loss of most tree species and the increase in diversity and abundance of prairie plants indicate a warmer and/or drier climate at this time. Valley-floor surfaces were stable at this time, but alluvial fans were aggrading.

The return of oak and basswood, along with forest-floor herbs and riparian trees, and the decline of prairie and weed species occurs in Zone 5 (2700 yrs B.P.–present). Values of $\delta^{13}C$ decline sharply at the base of this zone, but no other analyses in this interval were done except for a post-settlement sample. Chloridoid-group phytoliths increased and panicoid decreased in this basal sample. These data suggest that climate had become cool and moist enough to support some trees that provide enough shade for some forest-floor herbs.

Stop 2: Old Bridge Site

The Old Bridge Site is about 8 km downstream from Stop 1 (Stop 2 in fig. 2-6). A 400-m-long cutbank on the east side of the South Fork of the Big Nemaha River exposes a complex sequence of alluvial deposits beneath the T-1 terrace (fig. 2-14). Three members of the DeForest Formation are represented at this locality: Gunder, Honey Creek, and Camp Creek (fig. 2-15).

The Old Bridge Site displays important early Holocene and late Holocene sections (ca. 8000–9000 and ca. 3900–1900 uncalibrated ^{14}C yrs B.P.) that contribute to our overall paleoecological reconstruction along the South Fork of the Nemaha River. These sections have been analyzed for pollen, phytoliths, stable carbon isotopes, and a diverse macrofossil assemblage that reflects what was growing on the landscape. Other sections nearby fill in some of the gaps in time. Table 2-2 shows the variety of plant parts that have been identified just from the Old Bridge site, including buds, bud scales, leaf fragments, leaflets, nut fragments, wood (identified by Laura Strickland), seeds, and fruits. Most samples contain between 200 and 600 macrofossil types picked from just 200 ml of sediment.

Table 2-2 is organized by habitat. Each group represents a specific habitat that must have been present at or near the site. The habitat groups are trees, shrubs, forest-floor herbs, aquatics, wetlands, prairie plants, weeds, and "other" (not sufficiently identifiable to assign to a habitat).

The early Holocene macrofossils, prior to 8500 yrs B.P., had a good representation of mesic deciduous forest trees including oak, hickory, American elm, slippery elm, walnut, ironwood, and red cedar. During this time, prairie elements are also very abundant and diverse, including such indicator species of tallgrass prairies as big bluestem, little bluestem, indian grass, needle and thread, wild bergamot, and lead plant. Phytoliths and $\delta^{13}C$ values indicate a preponderance of C_3 plants, although deciduous-tree pollen is never less than 20 percent. The forest elements in the macrofossil assemblage all but disappear by 8200 yrs B.P. Although sites on the eastern plains are few, data from pollen sites in the Dakotas suggest that deciduous forests were present for between about 11,000–10,000 yrs B.P. on the eastern Great Plains, but died out soon thereafter and were replaced by prairie. These exposures suggest that at least some deciduous trees were able to survive locally until after 8500 yrs B.P. Presumably prairies began to dominate on the drier uplands, and the trees hung on along valley walls and ravines, where cooler, moister conditions were still available.

Other plant communities represented include riparian forests (willow and boxelder), wet meadow/marsh habitats, and disturbed ground/weedy areas. All these habitats are commonly found along floodplains, where erosion and deposition cause disturbance and create oxbows and other wet areas.

By late Holocene (3870 to 1910 yrs B.P.), only riparian forests were present at the Old Bridge site, with cottonwood, willow, boxelder, and elm (a late-successional floodplain as well as an upland forest tree). There was apparently enough shade at 1910 yrs B.P. for some shade-loving forest-floor plants to be present. Macrofossils of prairie species, though less abundant and diverse, are still well represented, and prairie must have been the dominant upland plant cover. Wetland and weedy disturbed habitats were again common on the floodplain.

Stop 3: Miles Fan Site

The Miles alluvial fan is located 8 km downstream from Stop 2 (Stop 3 in fig. 2-6). This fan developed at the mouth of an unnamed intermittent stream that flows into the South Fork of the Big Nemaha River. The channel of the Nemaha is along the north side of the valley wall and is cutting into the midsection of the fan. This has resulted in the development of a steep cutbank exposing a thick section of fan and floodplain deposits (fig 2-16). In addition, a Farmdalian peat is exposed at the bottom of the cutbank.

Based on its lithology, the Miles fan is composed of the Corrington Member of the DeForest Formation (fig. 2-17). The fan deposits are oxidized and consist of multiple upward-fining sequences. Two buried soils are developed in the fan: soil 2 at a depth of 130 cm and soil 3 at a depth of 360 cm (figs. 2-16 and 2-17). Decalcified organic carbon from the upper 10 cm of soils 2 and 3 yielded radiocarbon ages of 3130 $\pm$ 50 yrs B.P. (Tx-8945) and 10,450 $\pm$ 120 yrs B.P. (Tx-8944), respectively.

The alluvium below soil 3 is distinctly stratified and consists of fine-grained, late Wisconsinan floodplain

FIGURE 2-14. View of the cutbank at the Old Bridge site.

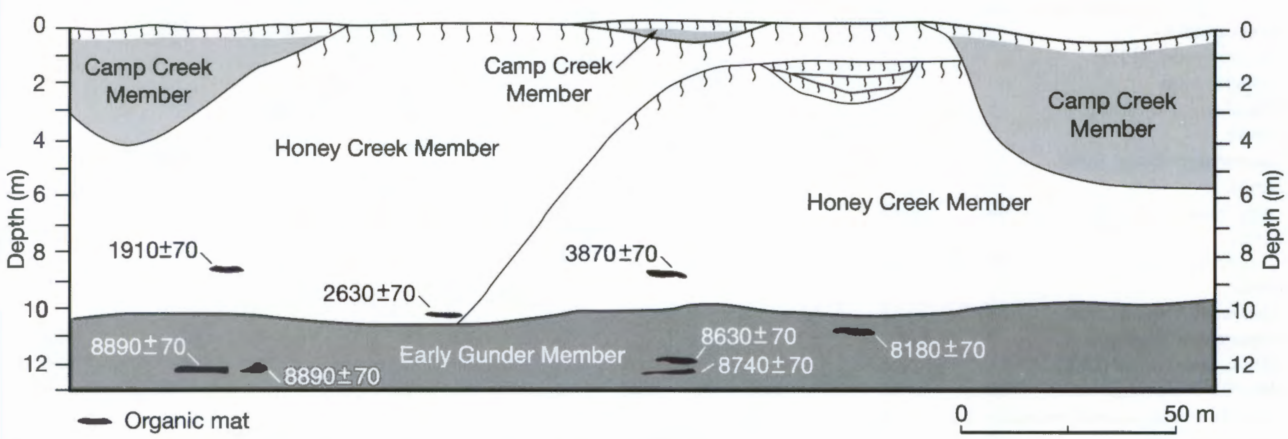

FIGURE 2-15. Outcrop along the east bank of the South Fork of the Big Nemaha River (Stop 2).

deposits (figs. 2-16 and 2-17). A razor-sharp boundary separates these deposits from a 1-m-thick peat that is 9.9–10.7 m below the surface of the alluvial fan. Peat samples collected at depths of 9.9–10.1 m, 10.3–10.4 m, and 10.6–10.7 m yielded radiocarbon ages of 23,490 ± 310 yrs B.P. (Tx-8946), 24,640 ± 410 yrs B.P. (Tx-8947), and 27,580 ± 550 yrs B.P. (Tx-8948), respectively. The peat grades downward into a 75–100-cm-thick unit of gray, organic-rich, pebbly clay loam.

At Stop 3, we will focus on the Farmdalian peat at the base of the fan, and the fossiliferous full-glacial alluvium 100 m downstream from the fan. The pollen record of Fredlund (fig. 2-18) covers only the peat section, and the macrofossil records (fig. 2-19) cover both sections, which are plotted together on diagrams.

We recognize two zones at this locality. The lower zone (ca. 28,000–22,000 yrs B.P.) is characterized by less than 10 percent spruce and very high percentages of non-

TABLE 2-2. Plant macrofossils collected from the Old Bridge site (Stop 2). The data are presented by habitat.

site # / info. date (yrs B.P.)	Profile 1 (upper) 1910 ± 70	Profile 4 2630±70	Profile 2 (upper) 3870±70	Profile 5	Profile 6	Profile 3 8180±70	Profile 2 (middle) 86660±70	Profile 2 (lower) 8740±70	Profile 1 (lower) 8990±70
Trees / fossil type									
Acer negundo / seed	1							1	2
Acer negundo / fruit wing	1 cf.								1 cf.
Carya cordiformus / nut frag.							1	1	
Fraxinus americana / fruit							1		
Juglans nigra / nut frag.							1		
Juniperus virginiana / twigs							1	27	1
Morus rubra / seed	1					1	5		
Ostrya virginiana / fruit							1		
Populus deltoides / capsule segments	4	2							
Populus deltoides / buds		2							
cf. *Populus deltoides* / bud scales	2 cf.		3						
Populus cf. *Populus* / wood (LES)		5				1			
Quercus macrocarpa type / flower bud scales								4	
Quercus cf. *macrocarpa* / twigs							2		
cf. *Quercus* / leaf lobes (white oak group)	3								1
Quercus / proximal tip (under cap)							1		
Salix exigua ssp. / leaf equivalents interior type	1					1		1	
Salix/capsules		2				6	1	13	
Salix sp. / wood (LES)						6			1
cf. *Salix* / buds						1			
Tilia americana / fruits	1				1				
Ulmus americana / seeds / fruits								3	4
Ulmus americana / bud scales	46	22	8	18	8	11	1	15	4
Ulmus americana / buds	4	9		3	1		2	6	4
Ulmus americana / wood (LES)	6	2							2
Ulmus rubra / bud scale	7	3 cf.						1 cf.	3 cf.
Ulmus rubra / wood (LES)		1							1 cf.
other bud scales / bud scales	30	11	14	60	2	28		31	16
Thick-bottomed, pointed bud scales		2	1						1
Other buds / buds	8			6	1		3	2	3
Large-leaf fragments	8	~5		7		3		5	
Forest floor herbs / Fossil type									
Campanula americana / seeds	1							2	
Laportea canadensis / fruits	2								
Salvia lyrata / seeds	1						1		
Shrubs									
Cornus / nutlet fragments		1 cf.					1		
Rubus / fruits	1								
Sambucus canadensis /seeds	1					1	1		2
Sambucus / seeds				1					
Staphylea trifolia / seeds							2		
Vitis / seeds		2				1			

TABLE 2-2 continued.

site # / info. date (yrs B.P.)	Profile 1 (upper) 1910 ± 70	Profile 4 2630±70	Profile 2 (upper) 3870±70	Profile 5	Profile 6	Profile 3 8180±70	Profile 2 (middle) 86660±70	Profile 2 (lower) 8740±70	Profile 1 (lower) 8990±70
Aquatics									
Heteranthera cf. *reniformis* / seeds								1	
Wetlands									
Bidens cernua / fruits							1		9
Bidens comosa / fruits					4				
Bidens frondosa type / fruits	3				2				9
Boehmeria cylindrica / fruits				1	1				3
Carex conoidea / fruits							1		
Cyperus aristatus / fruits			1		1				
Cyperus esculentus / fruits		1							
Cyperus odoratus / fruits					1				1
Echinodorus rostratus / seeds	1								
Eleocharis palustris / fruits				2					
Hemicarpa micrantha / fruits					3				
Juncus / seeds		14 cf.				2			
Leersia oryzoides / florets	4			3					
Lippia lanceolata / seeds					1				
Ludwigia type *alterniflora* / seeds	1								
Lycopus americanus / seeds				2					3
Penthorum sedoides / seeds	1			1			1		5
Pilea pumila / fruits	2								
Polygonum hydropiperoides / fruits				1					
Polygonum lapathifolium / fruits		3			4	12	1	12	
Polygonum pensylvanicum / fruits		2		1		1	1		
Polygonum punctatum / fruits				1			2		
Rorippa palustris / seeds							1		
Rumex altissimus / fruits with perianth							1		2 cf.
Sagittaria brevirostra (= *engelmanniana*) / fruits									1
Scirpus americanus / fruits				1					
Scirpus atrovirens type / fruits									1
Scutellaria / seeds	1						1		
cf. *Spartina pectinata* / partial floret								1	
cf. *Suaeda depressa* / inner seed coat								1	
Prairie plants									
Amorpha canescens / pods		9		9		26	3	16	1
Amorpha canescens / flower calyces		5		7		6		14	2
Amorpha canescens type / leaflets	5	3		50		14	1	50	
Amorpha cf. *nana* / pods							1		
Amorpha cf. *nana* / leaflets			2			2		2	
cf. *Amorpha* / calyx fragments								4	
Andropogon gerardii / florets	1			19		5	2	7	4

TABLE 2-2 continued.

site # / info. date (yrs B.P.)	Profile 1 (upper) 1910 ± 70	Profile 4 2630±70	Profile 2 (upper) 3870±70	Profile 5	Profile 6	Profile 3 8180±70	Profile 2 (middle) 86660±70	Profile 2 (lower) 8740±70	Profile 1 (lower) 8990±70
Andropogon (*Schizachyrium*) *scoparius* / florets				5		2		8	2
Androsace occidentalis / seeds		1					2	22	1
Ceanothus americanus type / seed					1				
Ceanothus americanus type / fruit				12					
Dalea / calyx base		1 cf.			1				
cf. *Dalea* / calyx				1					
cf. *Dalea* / leaflet				1					
Dalea cf. *purpurea* / calyces				3					
Desmanthus illinoensis / pod fragment				1				1 cf.	
Desmanthus illinoensis / leaflet				1				1 cf.	
Helianthus grosseserratus ty / fruits				1					
Helianthus laetiflorus type / fruits	1			1	1		1		
Helianthus sp. / fruits		1		4		1	1	1	
Linum sulcatum / seeds	1							2	
Monarda fistulosa / seeds							1		1
Monarda pectinata / seeds							1		
Oenothera biennis type / seeds								1	
Panicum capillare type / florets		9					1	39	4
Penstemon digitalis type / seeds		1					2	22	1
Polygala alba / seeds		2							
Polygala verticillata / seeds									1
"popped seeds" / seeds/fruits				1					
Ratibida pinnata / fruits									
Schrankia nuttallii / leaflets				1				1 cf.	
Sorghastrum nutans / florets				2				2	1
Sporobolus asper / caryopses (maybe not!)—see note		3				10	1	3	
cf. *Sporobolus vaginiflorus* / caryopses (maybe not)					13				
Stipa spartea / awns (fragments)	1			1		1		1	
Vernonia baldwini type / fruits								2	
cf. *Vernonia* / fruits						1			
Weeds									
Acalypha virginica type (incl. A. rhomboidea) / seeds		1			1		5	1	1
cf. *Acalypha* / seeds					1				
cf. *Amaranthus* / seeds							27		
Amaranthus / seeds	16	84	12	6	14	58		47	33
Ambrosia psilostachya type / fruits				1				3	
Ambrosia trifida / fruits	6	2		1	4	1	4	3	
Asclepias verticillata type / seeds				9					
Chenopodium berlandieri type / fruits	4				2				

TABLE 2-2 continued.

site # / info. date (yrs B.P.)	Profile 1 (upper) 1910 ± 70	Profile 4 2630±70	Profile 2 (upper) 3870±70	Profile 5	Profile 6	Profile 3 8180±70	Profile 2 (middle) 86660±70	Profile 2 (lower) 8740±70	Profile 1 (lower) 8990±70
Chenopodium bushianum / fruits						5			
cf. *Chenopodium capitatum* / fruits									
Chenopodium gigantospermum type / fruits	2				1	1	3		
Chenopodium spp. / fruits	4	22	4	3		45	89	62	3
Descurainia pinnata (combine) / seeds					1				
Draba / seeds		1						14	
Echinochloa muricata type / florets	9	1	2	6	9	1			3
Euphorbia glyptosperma / seeds							3		
Euphorbia maculata type / seeds								3	1
Euphorbia maculata type / perianth parts								4	1
Euphorbia nutans / seeds					3	4	4	3	12
Euphorbia cf. *nutans* / perianth parts									4
Euphorbia serpens / seeds		5	1			1		1	
Euphorbia cf. *serpens* / perianth parts		1							
Euphorbia spathulata / seeds						2			
Euphorbia sp. / locules		1		2					
Lepidium virginicum type / seeds						4		5	
Lepidium / siliques						4		5	
Mollugo verticillata / seeds		1							
Oxalis stricta type / seeds	2		1	1				1	1
Phyla lanceolata / seeds				1					2
Polygonum aviculare type / fruits		1							
Polygonum erectum / fruits						1 type			
Polygonum scandens / fruits								1	
Portulaca oleracea / seeds			1	1	4			1	4
Silene antirrhina / seeds		3	1 cf.			10	93	62	9
Triodanis perfoliata type / seeds		1	1			1	3	7	
Urtica dioica / seeds	6			2		1		2	
Verbena hastata / seeds								1	1
Verbena urticifolia / seeds				1			1	2	1
Verbena cf. *urticifolia* / seeds	3								
Other									
cf. *Ammania robusta*		3						1	4
Amorpha sp. / pod fragment									1
Carex (biconvex) / fruits	2	1	1	4	1	1	7		16
Carex (trigonous) / fruits	2	1			1	2	3		4
Cirsium / fruits	1								

TABLE 2-2 continued.

site # / info. date (yrs B.P.)	Profile 1 (upper) 1910 ± 70	Profile 4 2630±70	Profile 2 (upper) 3870±70	Profile 5	Profile 6	Profile 3 8180±70	Profile 2 (middle) 86660±70	Profile 2 (lower) 8740±70	Profile 1 (lower) 8990±70
Composites / fruits	2	1		1	3	3		17	1
other crucifer / seeds						1		1	
Cyperus / fruits									
Eupatorium purpureum type (incl. *E. maculatum* and *E. fistulosum*)								1	
Eupatorium sp. / fruits	1							1	
medium grass florets								3	
long, narrow grass / caryopses						12			
medium grass / caryopses		1	1		5				
small mottled cf. grass / caropyses		1				5			
misc. other caryopsis-like forms				1					
large grass (prairie) florets (damaged or incomplete)	2	4	2	48		11		35	
other legume calyces	3			3					
large legume seed (charred)									1
small leaflet bases / tips	6		1	14					
medium-sized leaflets / leaves	8								
monocot leaf tips	2								
Panicum / florets					6	5			
Polygonum sp. / fruits	1 cf.	2				1			1
cf. Scrophulariaceae / seeds					38				
Solanaceae / seeds	1			1					
legume pod fragments				1					
misc. small leaves / leaflets						1		14	
tiny "flowers"				3					
umbel / fruit						3			
Viola / seeds	1	2	1	2			1		
Totals	**229**	**240**	**57**	**340**	**140**	**325**	**285**	**594**	**192**
unknowns/indeterminate	20	11	11	44	13	35	20	68	38
unidentifiable wood		3				5		2	
insects	common		uncommon		present	present	uncommon		
bryophytes	rare				1		none	rare	
charcoal	rare (woody when present)						rare	herbaceous charcoal common, woody rare (1 woody chunk saved)	
bones					rare	1 cf fish vertebrae			

arboreal pollen, especially grasses and sedges (fig. 2-18). There is also a rich aquatic and wetland macrofossil flora (figs. 2-18 and 2-19).

This assemblage indicates that the site was a small pond surrounded by sedges and other wetland plants. Aquatic plants are prevalent in the lower part of the zone but decline as wetland plants increase towards the top. This sequence suggests that the early wetland was deeper, supporting more aquatic plants, but gradually filled in or dried up as wetland elements prevailed. The upland vegetation is represented only by the pollen record and clearly is an open, probably prairie, environment. Such ponds characteristically receive very few macrofossils derived from upland habitats. This interpretation correlates well with the old segment from the Farwell site, with nearby Du Bois quarry, and with the Logan quarry site in southwestern Iowa (Baker et al., 2003, and unpublished data), all of which have prairie plant macrofossils from this interval.

Zone 2 is represented by two upper pollen counts from Miles fan, and by macrofossils from the site 100 m downstream. The pollen from the fan shows an abrupt rise in spruce pollen and a decline of non-arboreal pollen. Spruce macrofossils do not appear in the peat section, but they are very abundant at the downstream section, dating from 20,290 to 19,920 yrs B.P. Most aquatic and wetland macrofossils were not present at this site, but sedges are abundant.

Fredlund interpreted this zone with spruce pollen from the fan as representing local vegetation. However, spruce needles are absent there, whereas they are generally abundant where spruce is locally present. Spruce had clearly arrived in the region, however, because both pollen and macrofossils are abundant at about this time at Logan quarry (Baker, unpublished data) and Muscotah Marsh (Grüger, 1973). It apparently was not growing directly onsite here until about 20,000 yrs B.P. Peat deposition ceased when Wisconsinan alluvium was deposited on top of it. At the downstream section, sedimentation continued for a short time, and spruce was clearly growing on site. It was not possible to determine if it was black or white spruce, but the low-diversity assemblage of abundant spruce and sedge would be consistent with a black spruce wetland. This assemblage is also typical and widespread in the late-glacial sediments of the Midwest.

Stop 4: Muscotah Marsh

Muscotah Marsh is located on the floodplain of the Delaware River about 2.5 km south of the village of Muscotah (Stop 4 in fig. 2-6). The marsh is supported by an artesian spring and, as discussed previously, provides the only continuous Holocene pollen record on the eastern edge of the Great Plains (Grüger, 1973). The vegetation of the region prior to European settlement was tallgrass prairie, and the site is located close to a mosaic of deciduous forest and prairie to the east (northeastern Kansas and northwestern Missouri). The Arlington Marsh site, located 5 km south of Muscotah Marsh, also contains pollen-bearing sediments and provides a picture of late Wisconsinan vegetation (Grüger, 1973).

Grüger's (1973) initial analysis of the Muscotah and Arlington marsh sites provides a 25,000-yr record of vegetation change for the region. She identified five zones. The first zone from 25,000 to 23,000 yrs B.P. is composed of relatively open vegetation, with some pine, spruce, and birch, with local stands of alder and willow. It was suggested that the vegetation was similar to what is now found in southern Saskatchewan and southeastern Manitoba. The second zone from 23,000 to 15,000 yrs B.P. is dominated by spruce pollen and represents a spruce forest. There is a hiatus from the second to third zone between 15,000 to 11,000 yrs B.P. The third zone from 11,000 to 9,000 yrs B.P. was composed of a mix of prairie and deciduous oak forest, which was then replaced by prairie around between 9,000 to 5,000 yrs B.P. (fourth zone). The final zone (5,000 yrs B.P. to present) again represented a mix of deciduous oak forest and prairie.

A new core (~10 m in length) was collected from Muscotah Marsh in September 2000 by Baker, Bettis, Mandel, and Moss to examine at higher resolution the Holocene environments of the eastern Great Plains. Moss (unpublished data) has produced a high-resolution (sample interval of 10 cm) pollen record for the first 180 cm of this core (fig. 2-20), which probably covers the late Holocene period. The results of the pollen analysis indicate that the vegetation around Muscotah Marsh was dominated by

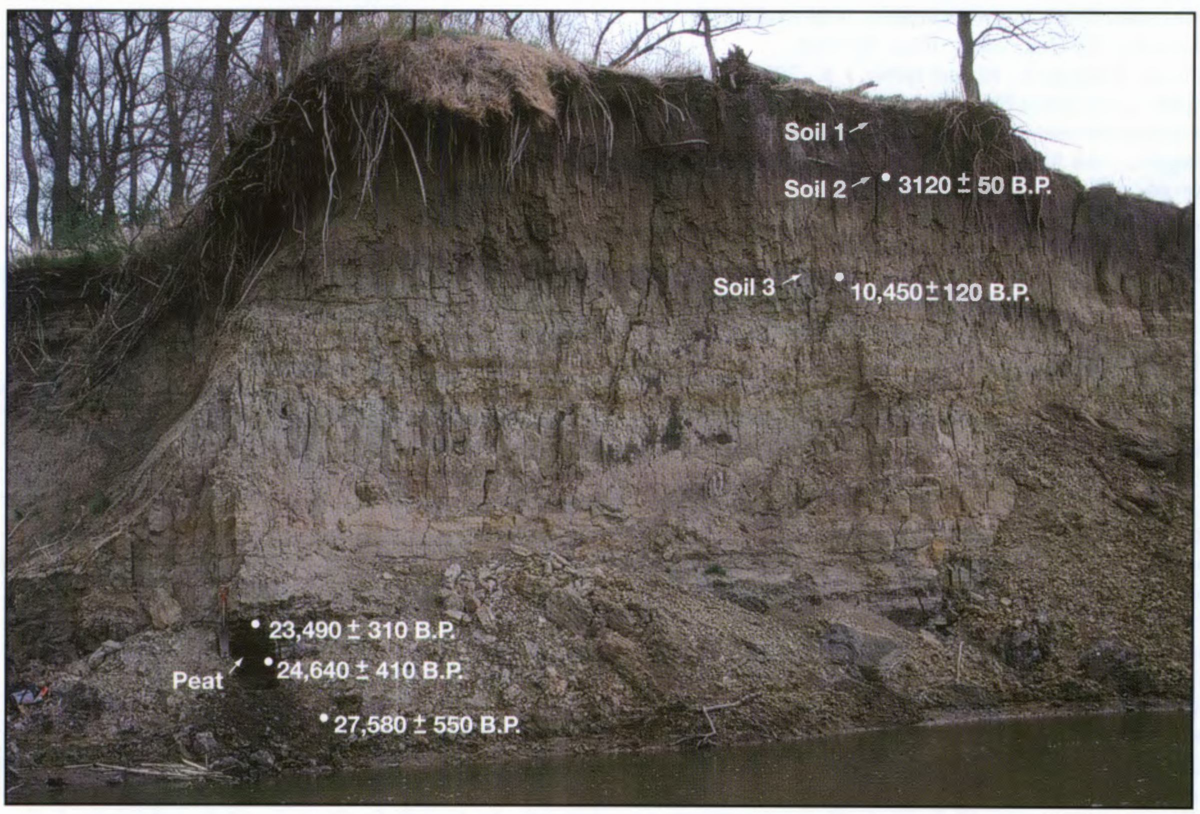

FIGURE 2-16. Photograph of the section at the Miles alluvial fan. Note the dark peat at the bottom of the section.

non-arboreal taxa (relative abundance between 60 to 70%). The dominant taxon from 180 to 30 cm was Asteraceae (*Tubuliflorae*) and probably reflects the presence of tallgrass prairie. Sagebrush (*Artemisia*) was also important from 160 to 100 cm. *Ambrosia* (ragweed) increases from 90 cm and may indicate increased human disturbance (European settlement) and the increased representation of Chenopodiaceae-Amranthaceae from 30 to 0 cm probably reflects intensive agricultural activity in the region. Grasses and sedges are important parts of the non-arboreal vegetation throughout the length of the record. Deciduous forest taxa (particularly oak) maintain consistently low values (5 to 10%) throughout the length of the record and reflect the regional presence of deciduous forest. Conifers also maintain low representation (less than 5%), except at 180 cm (~30%) and 110 cm (~ 40%), where spruce and pine peak; this may reflect contamination of the record or stronger westerly winds that were transporting these pollen types over long distances from the Rocky Mountains. The presence or absence of spruce needles at these depths in the record will hopefully resolve the cause of these conifer peaks. Carbonized particle analysis of the Muscotah Marsh record suggests relatively high fire regimes from 180 to 50 cm and with lower values from 40 to 0 cm, probably reflecting the imposition of fire suppression by Euro-Americans.

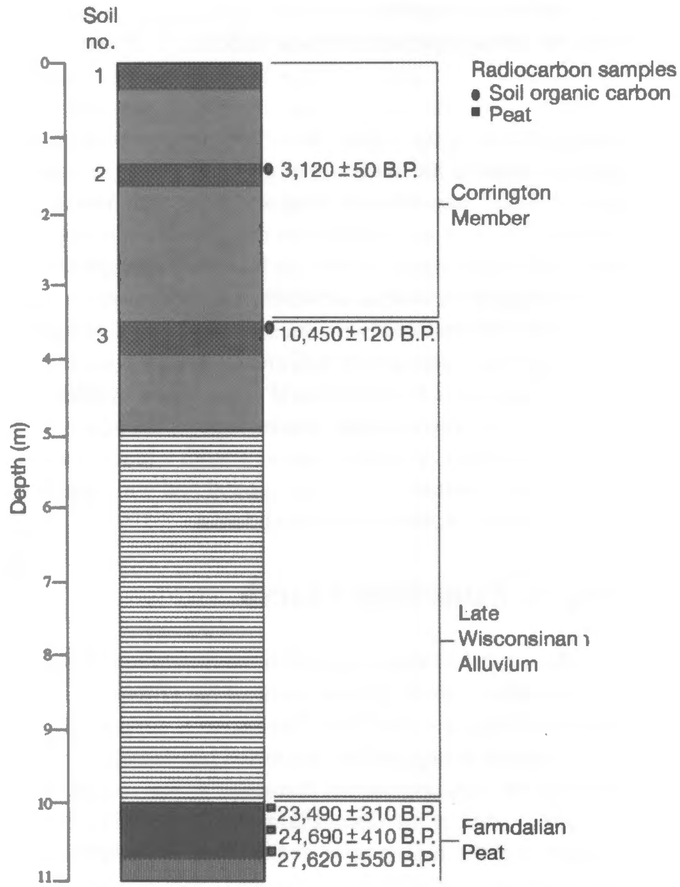

FIGURE 2-17. Stratigraphic section for the Miles alluvial fan.

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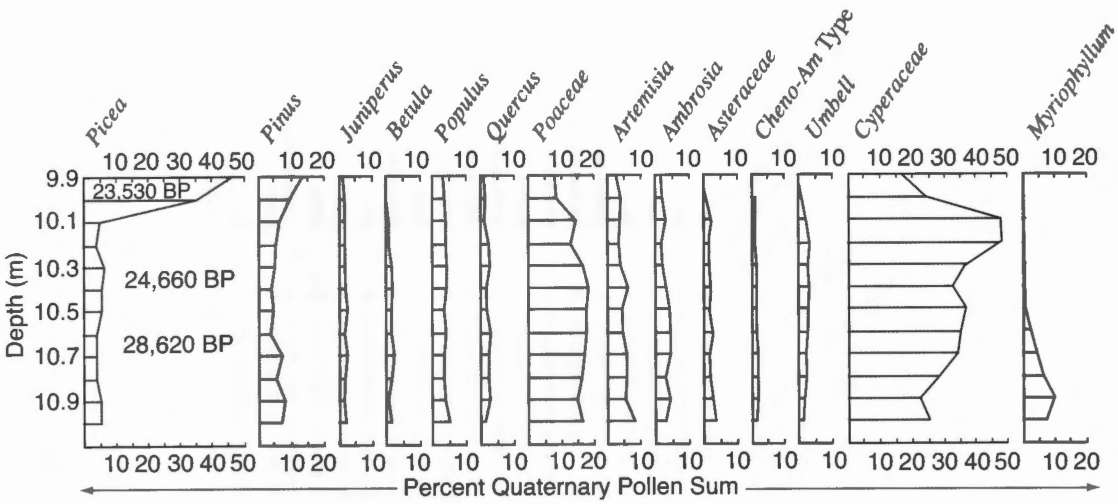

FIGURE 2-18. Miles Fan pollen data.

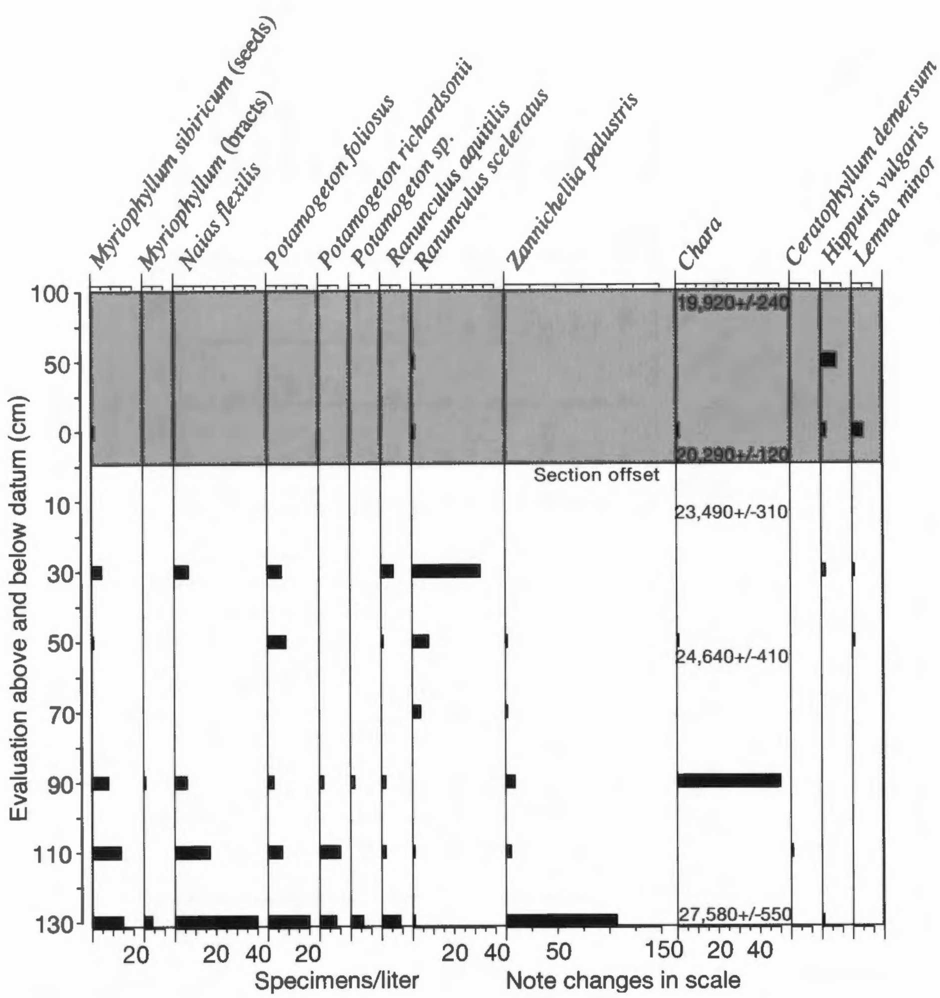

FIGURE 2-19. Miles Fan aquatic-plant data.

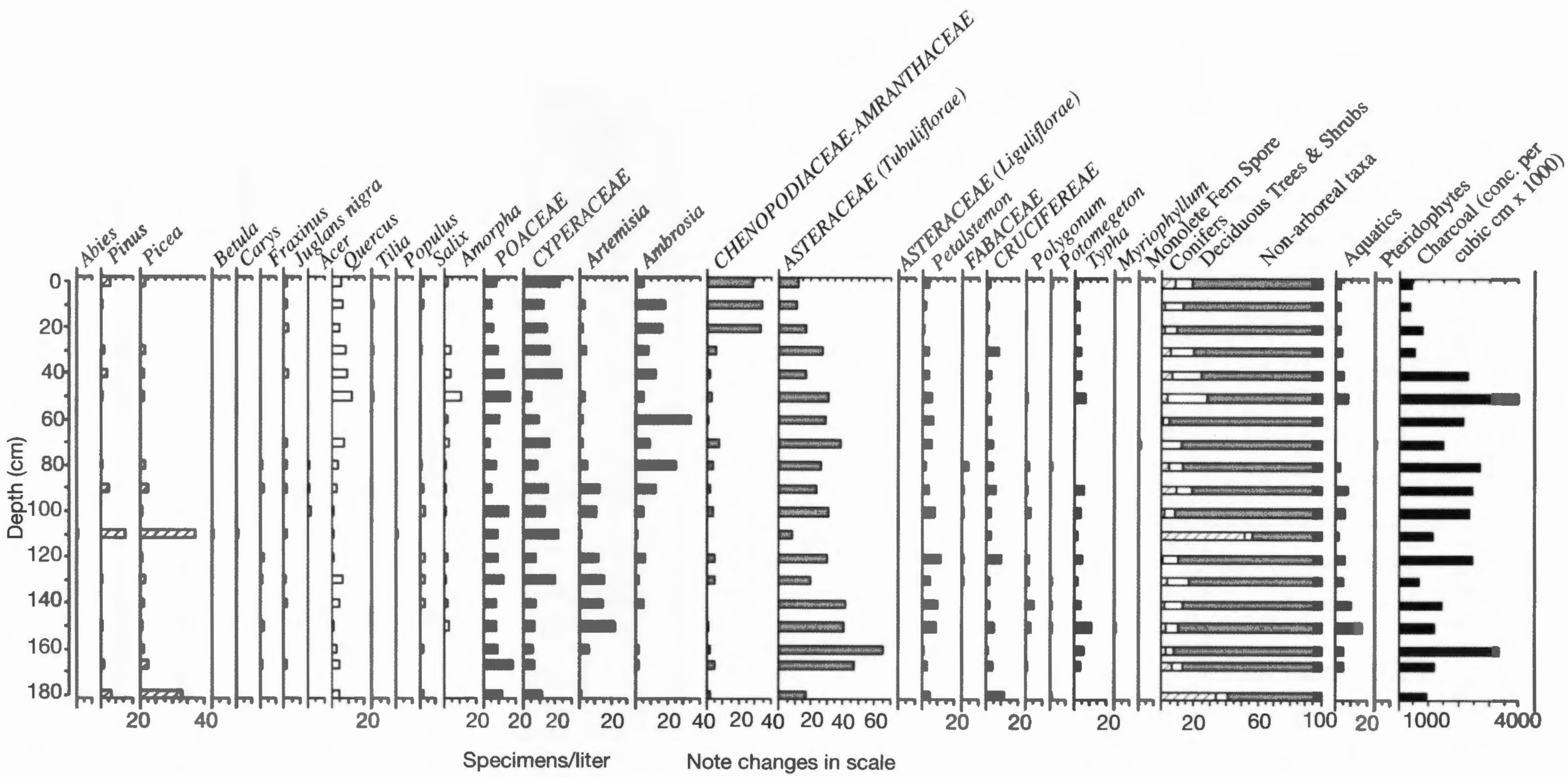

FIGURE 2-20. Pollen record for Muscotah Marsh.

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AMQUA Post-meeting Field Trip 3: Multiple Pre-Illinoian Tills and Associated Sediments and Paleosols, Northeastern Kansas and Central Missouri

Part I: Kansas

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Preface

During a trip across the plains in 1868, Louis Agassiz realized that the presence of red boulders resting on gray limestone bedrock in northeastern Kansas constituted evidence of past glaciation. By 1881, less than 20 years after the American Civil War, Thomas C. Chamberlin was able to follow a glacial boundary from Massachusetts westward into Dakota Territory. He also recognized evidence for what he termed the "attenuated border" of an older glaciation. An 1885 map shows the extent of this earlier ice in northeastern Kansas and adjacent areas.

These earliest investigations developed the general framework of the glaciation of North America. It was established that Early Pleistocene ice sheets covered the northeastern corner of Kansas and most of the northern half of Missouri. Although several studies of local areas were undertaken at various times in the 1900's, no detailed regional syntheses have been forthcoming. There remains an abundance of unanswered fundamental questions.

Kansas terrain has notoriously low relief; Missouri is little better. Consequently, there are very few places where glacigenic sediments—till and outwash—can be viewed in natural exposures. Most information must be obtained from artificial displays that are usually very short-lived. During the 1950's and 1960's, numerous quarries were opened along the western side of the Missouri River valley to provide the Army Corps of Engineers with material for construction of jetties and levees. Every one of those excavations exposed Pleistocene sediments, but all were soon abandoned and have been covered by vegetation and slumped debris.

In the 1960's and 1970's, several borrow pits were opened in and near Atchison to provide material for levees and small dams along and adjacent to White Clay Creek. All are now closed; most have been bulldozed to restore the original landscape. And in the 1970's and 1980's, extensive housing developments in the southwestern part of Topeka were accompanied by opening of excavations for water mains, sewer lines, street grades, and house foundations. These provided excellent, though very short lived, three-dimensional views of glacial deposits in and close to the terminal zone. All were quickly closed completely and essentially nothing of the subsurface relationships can be seen today.

Because of limitations imposed by terrain and modern history, this field trip must be an unusually high-mileage event in order to visit exposures of tills and associated sediments even within Kansas. The route and locations of stops are shown in fig. 3–1. Because of severe time constraints during the initial running of this field trip in June 2004, some of the stops may be omitted. However, they are listed in this guidebook to provide references of possible use during future trips. A much greater distance has been added to provide opportunity to see unusual sections in east-central Missouri, as described by Rovey in Part II: Missouri.

Road Log

0.0 **West Lawrence toll booth entrance to the Kansas Turnpike. Go west.**

18.9 **Turnpike Exit 183. Continue west (straight ahead) on I–70 (fig. 3–1).**

Although the route lies well within the glaciated area, no evidence of this is visible, at least to the casual traveler on the highway.

20.0 Pay toll.

24.2 Passing through downtown Topeka; the State Capitol building can be seen to the left.

Continuing westward, the Kansas River is visible on the right for a short distance. Roadcuts expose only Pennsylvanian bedrock, mostly limestone.

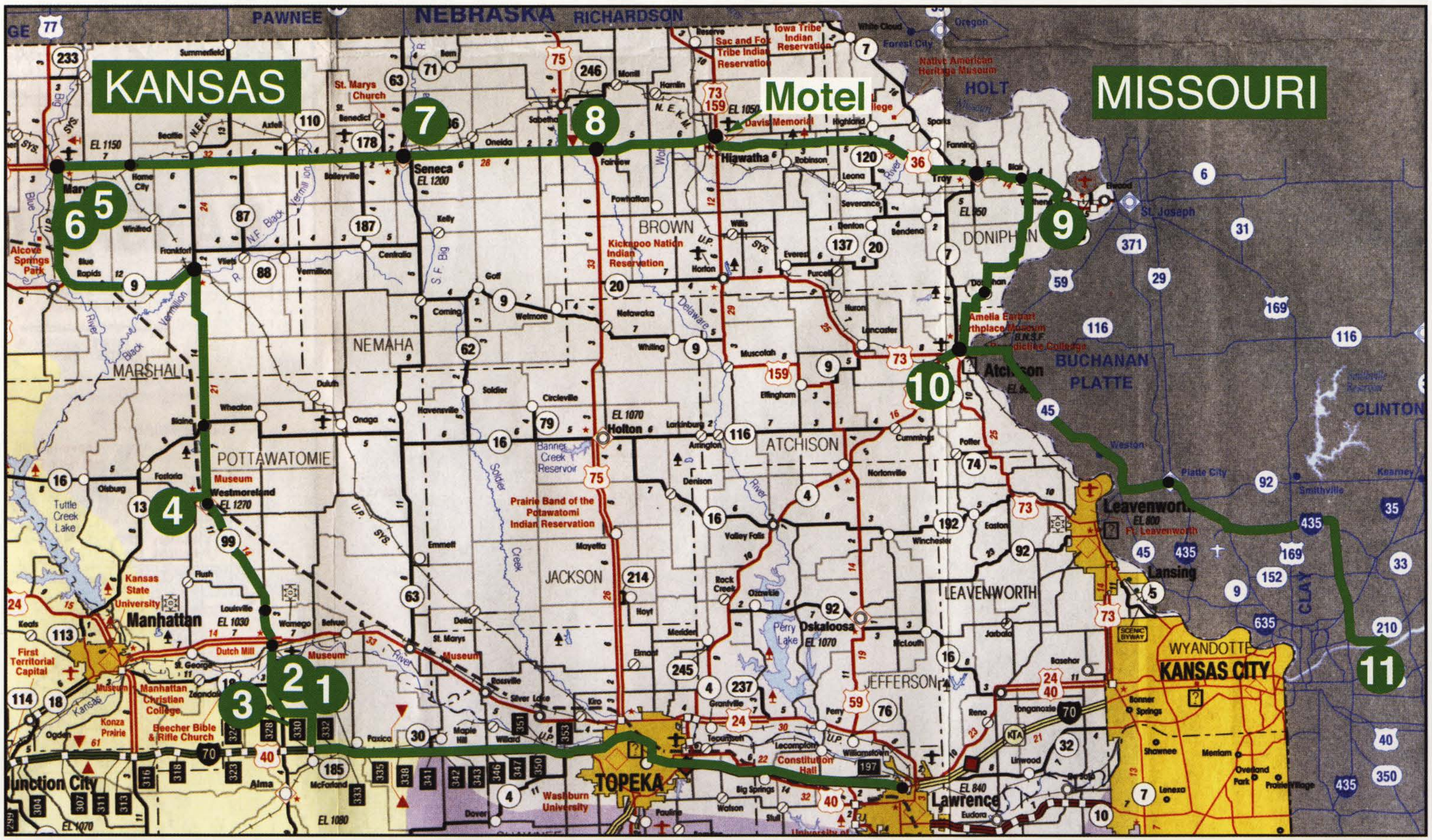

FIGURE 3-1. Route map for the Kansas part of Field Trip 3. Black dashed line marks route of the Oregon Trail.

43.6 The hilltop house south of the highway is situated almost exactly on the terminal line of glaciation in Kansas (as presently recognized). Large erratic boulders of Precambrian Sioux Quartzite are scattered through the pasture west of this house.

The prominent hill directly ahead, known locally as Buffalo Mound, also lies on the glacial boundary. Ice butted against the northern (right-hand) slope, but did not cover the summit. The col south of the summit was briefly occupied by an ice marginal stream that connected proglacial lakes draining eastward (fig. 3–2). The horizontal lines of vegetation crossing the eastern face of the hill express stratigraphic control by limestone and shale units of Permian bedrock; they are not ice-marginal kame terraces. The Carboniferous (Pennsylvanian)–Permian contact is on the hillside on both sides of the road here.

46.9 The glacial boundary angles up from the southeast and here crosses the highway, which is oriented due west.

48.3 Another prominent hill ahead had ice butting against its northern side and an ice-marginal stream flowing through the col to the south.

50.4 **REST STOP**

53.5 A few large boulders of Sioux Quartzite in the field directly north of the highway mark the glacial boundary at this point. Attempts at cosmogenic nuclide (Be-10) dating produced inconclusive results because there was so much spread from sample to sample.

55.5 Valley floor of Mill Creek. This lowland was occupied by a small proglacial lake. The impounding ice front was located a short distance north of the highway, probably with a tongue extending somewhat southward into the lake from time to time.

57.2 **Mile Post 330. Exit from I–70.** Sign says McFarland and Kansas Route 185.
Go north (left) on gravel North McFarland Road (fig. 3–3).

Opposite the first house on the right there are a few pink Sioux Quartzite boulders in the field that slopes up to the west. Their presence indicates that the glacial boundary has been crossed and the trip route has returned to glaciated terrain. The white and light-gray rocks are fragments of local limestone bedrock.

Proceed northward uphill through a dissected landscape. Soil is thin; Permian bedrock lies close to the surface and is exposed in shallow draws. Pink Sioux Quartzite erratics are scattered across the landscape on both sides of the road. White boulders are fragments of the local limestone bedrock.

61.0 **Intersection of North McFarland Road and Homestead Road on Tower Hill.**

STOP 1.

The flat summit of this hill is almost exactly 1,300 ft (396 m) above sea level. The Kansas River in its broad valley lies 6 mi (9.6 km) to the north at an elevation of 940 ft (287 m). Soil on this summit is very thin; Permian limestone bedrock is exposed in several places. Of greater significance are boulders of Sioux Quartzite. The presence of these erratics resting on ancient bedrock on the summit of the highest hill in the area provides proof that glacial ice surmounted this location. Looking toward the east and southeast one can identify large erratic boulders scattered across the landscape, and erratics were visible from the road as this Stop was approached. Although there is nothing visible here that could be called till, the erratics constitute clear evidence that glacial ice crossed the high divide of which Tower Hill is the culmination and continued almost 3 mi (4.8 km) farther south, reaching nearly to the line of highway I-70. However, because the limit line of the erratics bends toward the northwest, the boundary is only about 1 mi (1.6 km) southwest of this Stop (fig. 3–3).

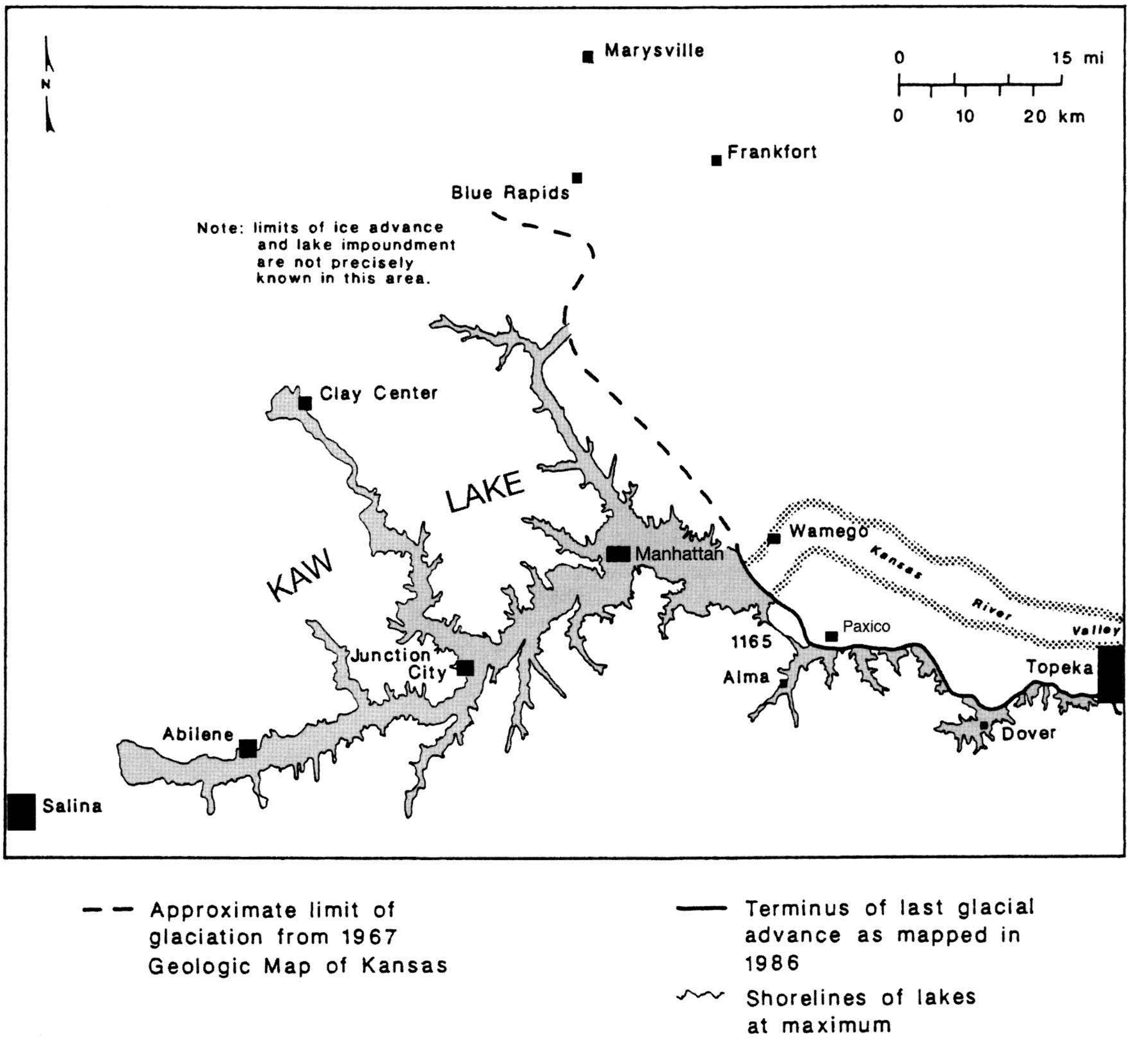

FIGURE 3-2. Proglacial lakes west of Topeka at the time of maximum ice advance. Shorelines depict water levels controlled by spillways against or close to the ice front. Location of the Kaw Lake spillway, elevation 1,165 ft (355 m), is shown. Drainage was eastward. It is believed that there has been no significant warping of the land surface since that time and that the glaciated terrain closely approximated the landscape of today, except for deposition in bayhead locations. This permits plotting on modern topographic maps.

DISCUSSION: Evidence of glaciation and location of the terminal boundary in Kansas

Classical evidence that an area of low relief was formerly invaded by a continental ice sheet includes the presence of both erosional forms, especially smoothed and striated bedrock surfaces, and depositional materials—i.e., stratified outwash, unstratified diamicton (till), and, especially, far-traveled erratics.

In Kansas, glacially scoured bedrock surfaces are extremely rare, usually restricted to chance uncovering in quarries or along highways. Exposures of till are slightly more common, but most of these are created by artificial excavations and are quickly filled or collapse.

The best indicators of the former presence of glacial ice are erratics, up to large boulder size (fig. 3–4), that are prominently scattered across the countryside. In fact, the limits of former glaciation are most conveniently mapped in Kansas by charting the southernmost and westernmost extent of erratics. Most of the solitary erratic clasts are fragments of Sioux Quartzite, a tough Precambrian quartzite that is locally conglomeratic. Its color ranges through various shades of pink, making it easily recognizable at a distance and differentiable from local gray limestones. The Sioux is also present as pebbles and cobbles in most, but not all, till and outwash deposits. All Sioux erratics were derived from a closely delineated outcrop area that is restricted to southern Minnesota and small adjacent areas in southeastern South Dakota and extreme northwestern Iowa (fig. 3–5).

Common in a specific few exposures are clasts of various igneous and metamorphic rocks eroded from sources in far northern United States or in Canada. Included are many varieties of granitic rocks, plus

FIGURE 3-3. Southwesternmost continental glaciation of North America, showing interpretation of younger and older limits, the shoreline of ice-impounded Kaw Lake, and its controlling easterly draining spillway. Locations of individual large Sioux Quartzite erratics are shown by triangles only where close to and defining the younger limit. Area shown is in Wabaunsee County south of Wamego and west of Topeka, Kansas.

metavolcanics and exotics such as Lake Superior agate, Duluth area iron ore, and Keweenaw volcanics and native copper. Present only rarely are small pieces of the soft Pipestone variety of the Sioux Formation. Many exposures display remarkable concentrations of a single clast lithology. Most common are those showing ca. 95% to as high as 100% Sioux Quartzite. In sharp contrast are deposits composed almost entirely of granitic boulders or of various-sized fragments of local limestones.

Field mapping indicates that the southern and southwestern limits of ice advance are very sharply expressed. Throughout most of the distance from Kansas City westward through Topeka and Wamego, and northward toward Nebraska, there is an abrupt edge to the Sioux Quartzite erratics. This permits plotting of a very precise boundary.

It is readily observable that far-traveled erratics, whether contained in till or outwash, or scattered about as solitary boulders, cannot easily be found throughout the entire area of Kansas that is believed to have been covered by continental ice sheets. Instead, in most parts of this area, it is notably difficult to find any evidence whatsoever of the former presence of glacial ice. Indeed, one can drive for miles past roadcut exposures on major highways and observe only bedrock capped by thin soil. This sparseness of ice-deposited debris led Chamberlin (1895) to speak of the "attenuated border" of glaciation, a term that was subsequently adopted by Schoewe (1938). Ongoing discussions have debated whether this situation is a product of original glacial deposition or later modification.

In marked contrast, till, outwash, and erratics are common along the terminal zone of glaciation. Indeed, there is such a close coincidence of the southernmost evidence of the former activity of glacial ice and the presence of concentrations of these glacial deposits that one is tempted to engage in circular reasoning. If numerous erratics are present, the locality must be in the terminal zone—or—if this place is in the terminal zone, till and erratics should be commonplace. Nevertheless, observed facts frequently, though not always, uphold this conclusion.

Following a transect southward from the Nebraska-Kansas border, glacial deposits are present though sparse, abruptly become thick and widespread, then are found no more. It seems indisputable that the ice advanced no farther and that melting along an essentially stable glacier front yielded an accumulation of englacial and supraglacial debris brought from northern sources.

At a very few localities situated well north of the terminal zone, there are limited occurrences of thick, erratic-rich till, generally exposed in small gravel pits or quarries. These deposits have been informally identified by several field geologists as remnants of recessional moraines. In addition, at least three "outliers" of glacial debris, including Sioux Quartzite clasts, have been found 3–6 mi (4.8–9.6 m) south of "the glacial boundary," thus well outside of the supposed limit of ice extent. No clear explanations have yet been determined.

This locality has been the scene of vigorous debates regarding the extent and accomplishments of post-glacial erosion in Kansas. Opposing end-member conclusions are outlined below:

(1) The landscape has been modified considerably by post-glacial erosion. Undoubtedly, thick deposits of till and outwash were left on an ancestral Tower Hill, as well as at other localities, and perhaps were even capped by younger loess and weathering products. The few Sioux Quartzite erratics that we see now are the last remaining vestiges of this accumulation. All the rest was removed by post-glacial erosion. Furthermore, it is unreasonable to suggest that the terrain present as the ice sheets disappeared from Kansas has undergone little to no erosion and change during the last few hundred thousand years.

(2) The landscape that we see now is in many (most?) places almost exactly the same as it was at the time the ice sheets invaded and then disappeared from Kansas. The ice that overran Tower Hill, and other parts of northeastern Kansas, was nearly clean and left behind only those erratics that are now lying on the bedrock surface.

FIGURE 3-4. Sioux Quartzite boulder 2 mi (3.2 km) northwest of Dover, Kansas.

Compilation of field observations indicates that Conclusion 2 is close to the truth. For one thing, a thick section of till deposited on an ancestral Tower Hill would surely have contained numerous rock fragments up to boulder size. A high percentage of these would have been composed of tough Sioux Quartzite, as is seen in existing exposures of till at other localities, some close nearby, such as at Stop 2. As erosion attacked such a deposit, the finer fragments would have been moved down the gentle summit slopes by gravity and rainwash. The larger clasts would have become differentially concentrated more or less in place for a while, but ultimately even these would have moved downslope. Chemically weaker granitic rocks might succumb to weathering, but the resistant quartzite would be little affected.

The expected end result would be that all stream courses, from fingertip rills to larger creeks, would have become choked with a lag concentrate of fragments of Sioux Quartzite. However, such a situation is not found in the field other than at a few limited occurrences where thick till is present today. Most valleys, of all sizes, are entirely free of erratics. This argues for non-deposition of till over much (most?) of the glaciated area, and that, in turn, bears glaciological and climate implications.

Equally strong support for a proposal that post-glacial erosion was ineffective is provided by examination of the relationship between the line connecting the southernmost occurrences of glacial erratics and local topographic features. Detailed mapping demonstrates that there is no consistent link, no topographic control in situations where it would be most expected. Especially noteworthy are places where the drift border crosses a ridge at an angle to the linear crest, but the ridge causes no deviation in the line of most southerly erratics (fig. 3–3). If post-glacial erosion had been effective, downslope movement of glacial debris would have at the very least smeared the terminal line and more likely completely cleaned off the slope.

Proceed west on Homestead Road.

Sioux Quartzite erratics are especially numerous to the right (north).

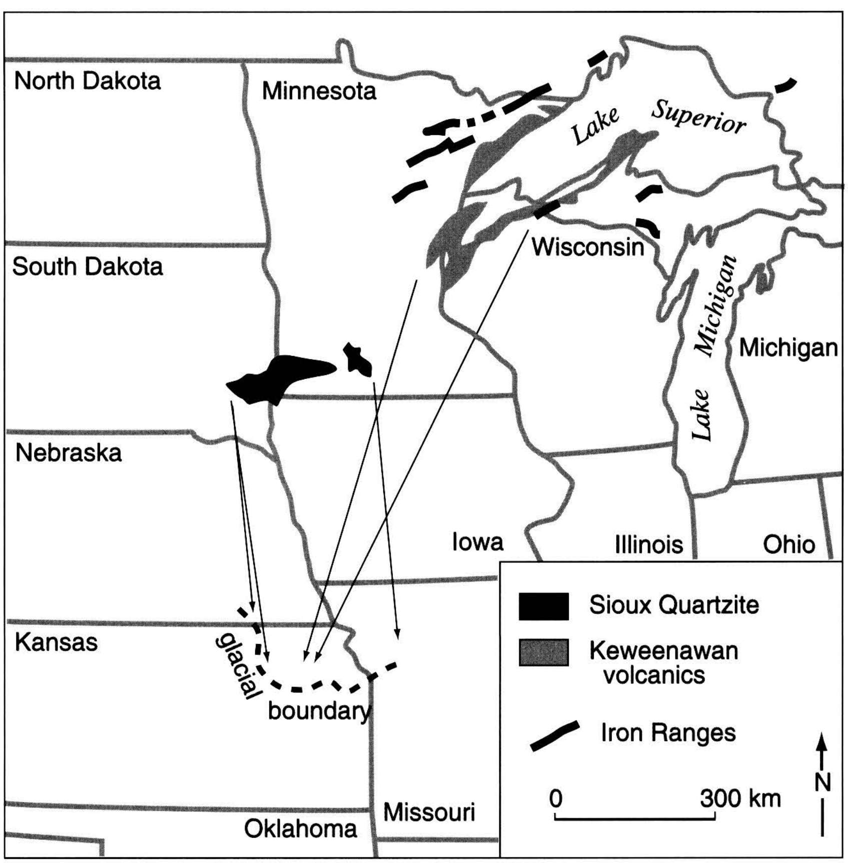

FIGURE 3-5. Sources of identifiable glacial erratics found in northeastern Kansas. Fragments of Sioux Quartzite are common throughout the terminal zone, and eastward into Missouri. Specimens of ore from the Iron Ranges are scarce. Distribution of Lake Superior agates from the Keweenawan volcanics is highly localized, seemingly restricted to an area in and near Topeka.

61.9 STOP 2.

A small fingertip stream system draining the western side of Tower Hill is shallowly incised. Unlike other drainageways in the area, this valley exposes bouldery till resting on limestone bedrock. On the northern side of the road, the thickness is about 15 ft (4.6 m); south of the road, a short distance to the west, a limestone cliff has only a thin cover of till. At both locations in excess of 99% of the large clasts are Sioux Quartzite. Only one granite boulder, 3 ft (0.9 m) long, has been found. Smaller clasts, mainly pebble size, include other lithologies, yet still are about 99% Sioux Quartzite. It is of interest that the non-quartzite clasts exhibit no megascopic evidence of weathering, even though lying unprotected on the surface of the ground. Although it is expected that the resistant Sioux Quartzite might not have been affected by chemical or mechanical weathering, it is puzzling to find granite clasts that appear fresh and untouched. It is especially difficult to explain the preservation in the till of well-striated limestone boulders.

The pattern of associated bedrock exposures indicates that the till here is basically confined to a small pre-glacial valley that has been partially re-excavated by the modern stream. The waterways here are choked with Sioux erratics, providing evidence that till has been removed from the slopes by erosion and transported into the nearest channels. As was discussed above, this is what would be expected everywhere if thick, widespread till had been eroded in the past. Because no such concentrations of erratics are generally found, it is believed that thick till deposits were spatially very limited.

Continue west on Homestead Road.

The frequency of Sioux Quartzite erratics decreases and the average size becomes smaller. A few erratics are visible in roadside ditches and nearby fields for a distance of about 1 mi. Then no more are seen. The interpretation is that the straight road has crossed the terminal limit of glaciation that is here oriented southeast-northwest (see fig. 3–3). That of course would mean that the route has entered the unglaciated area.
However. . . .

63.1 **Turn left (south) on new gravel driveway.**

63.3 STOP 3.

Excavation of the foundation for a new house was undertaken on about April 15, 2004. Preliminary observations indicate that the subsurface material is a weathered till that contains ghosts of rotten granite, a few fragments of Sioux Quartzite, pebbles of various other lithologies, and very numerous small, angular bits of chert. Initial, spur-of-the-moment interpretation is that this is a previously unseen and unsuspected till that is distinctly older than and out in front of the till that contains the numerous large Sioux Quartzite boulders that are seen here scattered across the landscape. It would therefore be a record of a distinctly older advance of the ice sheet.

Three miles (4.8 km) due south of this new excavation there is an old, very shallow cut beside a dirt road (figs. 3–3 and 3–6). This cut exposes a limestone pebble gravel that contains only a few small Sioux Quartzite clasts. This occurrence is enigmatic. The presence of the Sioux pebbles requires some association with glacial activity, but its location well outside of the supposed glacial limit defied ready explanation. Discovery of the old "outside-the-limit" till in the new house excavation suggests that there is some connection; even though one exposure is highly weathered, the other is not. Further investigation, particularly between these two "old" exposures, may produce an explanation.

Through many years of exploring glacial deposits in Kansas, it has always been believed that the most southerly boulder of Sioux Quartzite along any specific line of ice flow marked the limit of southward advance of the continental glacier at that point. Indeed, it became an article of faith that the east-west line connecting adjacent, closely spaced, most-southerly boulders clearly delineated the ice front at the time of maximum advance. This postulation appears to be sound—but only within limits. It omits the possibility that multiple advances of the ice sheet could produce a more complex record, and almost surely would have done so.

It seems certain that a single, specific glacial advance deposited the numerous, large Sioux Quartzite boulders that collectively compose what has herein been called the terminal zone. The southerly margin of this zone has been identified as the limit of glaciation. At the same time, a very few scattered occurrences of debris of apparent glacial origin were known to exist farther south—outside the limit of glaciation. Something was wrong!

The presence of a small exposure of erratic-bearing gravel 3 mi (4.8 km) south of the Tower Hill limit of glaciation was recorded some years ago—and was essentially ignored in a decidedly unscientific fashion. The discovery of a weathered, fine-grained till in the new house excavation at Stop 3 provided the necessary clue. The current interpretation is that this is a separate till body that predates the big-boulder till seen at Stop 2 and extends farther south—beyond the limit of glaciation as defined by the large Sioux Quartzite boulders. Thus, this is taken as evidence of a new, hitherto unrecognized glaciation.

Perusal of a recently published soil report for Wabaunsee County (Graber et al., 1991) finds that the USDA soil scientists identified and mapped areas underlain by the Pawnee Soil, which is defined as having developed on glacial till and outwash. Some of these areas extend from Tower Hill south to I-70 and beyond, all the way to the town of McFarland (fig. 3–6).

On the basis of these discoveries, it appears that in the Tower Hill/McFarland area there are present patches of tills of two distinct ages, tills that were deposited by two separate advances of continental ice. Furthermore, these tills appear to be the products of contrasting glacial erosion and transportation regimes, one leaving numerous large boulders, the other very few. And their episodes of aggradation were separated by a period of time long enough to permit considerable weathering of the first deposit prior to superimposition of the second.

Return to Homestead Road.

63.5 **Continue west on Homestead Road.**

63.9 Another new house excavation on the southern side of the road exposes limestone and shale overlain by a ferruginous soil, probably a paleosol, and a sparse scattering of pebbles of erratic lithologies.
In terms of the regional pattern, the terminal line was oriented approximately east-west subparallel to I-70 westward from Topeka to almost exactly the point where this field trip turned north on North McFarland Road. From that point the glacial boundary extends northwestwards to the Kansas River, thence northwards into Nebraska. That point of change in orientation is the point of farthest southwestward advance of continental ice in North America (fig. 3–7).

64.0 **Turn right (north) on Kansas Highway 99.**

64.7 The highway skirts the western side of a narrow, steep-sided ridge. Limestone bedrock crops out about halfway up the sideslope, but the crest is covered by a felsenmeer of small, rounded Sioux Quartzite boulders (fig. 3–8). Levelling indicates that this cap may be as much as 33 ft (10 m) thick. The nearly flat summit of the ridge may be the original end-of-deposition surface. Although somewhat similar concentrations of Sioux erratics occur in a few other places along the terminal zone, this occurrence is particularly outstanding. The boulders are so numerous as to be in contact with one another. If this situation extends into the subsurface, then the deposit is a clast-supported boulder gravel. The clasts are all Sioux Quartzite, all are subrounded to rounded, and there is a distinctly limited size range; almost all are between 6 and 20 inches (15–50 cm) long.

The high degree of sorting of rounded boulders makes this some sort of outwash deposit rather than a typical till. Indeed, at least one observer has proposed that this is a glacial flood gravel. However, this gravel caps a narrow ridge that lies almost precisely on the line of farthest advance of the ice sheet. The terminal line can be traced from the southwestern flank of Tower Hill to this ridge and then diagonally across the valley and the ridge to the northwest.

This gravel cap is, therefore, a terminus feature on a preglacial high. Furthermore, the apparent high degree of size sorting does not, in this case, require selectivity by running water. Outcrops of Sioux Quartzite in the Minnesota source area locally expose strata that are riven by joints spaced 12 to 18 inches (30–45 cm) apart and intersecting at right angles, providing a ready source for a field of boulders all about the same size. This explanation does, however, require transport en masse from southern Minnesota to northern Kansas, a phenomenon for which there is supporting evidence elsewhere along the terminal zone but which suggests puzzling ice mechanics.

Continue north on Kansas Highway 99.

The route crosses a prominent remnant of a high terrace of the Kansas River, then the broad valley floor and the river itself.

At the time of maximum glacial advance, the valley of the Kansas River was covered by the continental ice sheet all the way east to Kansas City, a straight line distance of 92 mi (147 km) , but 127 (203 km) mi by today's river channel. The western limit of glaciation is approximately 4 mi (6.4 km) upvalley from Wamego. In order for the glacier to flow southward across the river valley and still be able to surmount Tower Hill, 360 ft (110 m) above the present floodplain, there must have been a continued pressure gradient toward the south and the surface of the ice undoubtedly maintained a southward topographic gradient. Without considering other factors, one can crudely estimate that the ice must have been at least 300 ft (91 m) thick on the upland north of the river and perhaps 500 ft (152 m) thick over the ancestral valley of the Kansas River.

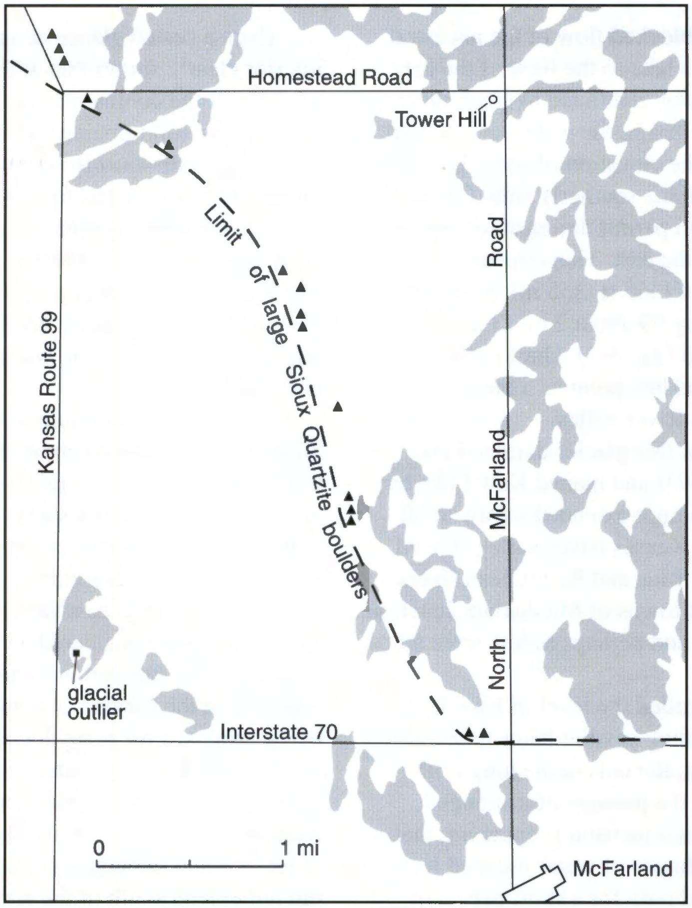

FIGURE 3-6. Distribution of the Pawnee Soil in the Tower Hill–McFarland area as shown in the Wabaunsee County Soil Survey report. By definition, the Pawnee Soil developed by weathering of glacial sediments. Its presence identifies similar substrate but not necessarily constant age. The Pawnee Soil extends significantly outside the limit of prominent Sioux Quartzite boulders, only the farthest southwest of which are shown here (see fig. 3–3).

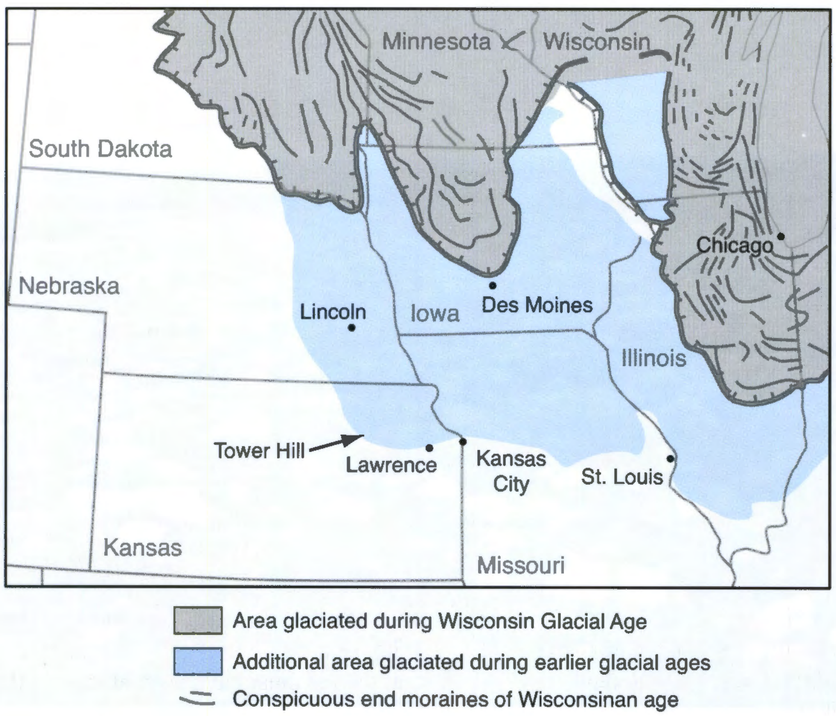

FIGURE 3-7. Extent of pre-Wisconsinan ice sheets in central United States. Note the location of Tower Hill at the southwesternmost limit of glaciation.

Glacial ice effectively blocked flow of the ancestral Kansas River, impounding a lake to the west of the ice front (fig. 3–2). Because water continued to reach this area from the upstream drainage basin, the water level would have risen until it reached the elevation of some low point in the topographic divide on the southerly side. Overflow across a spillway would then permit drainage to continue to lower elevations toward the east. This critical low point, with a threshold elevation of 1,165 ft (355 m), is located just west of Kansas Highway 99 about 2 mi (3.2 km) south of the Tower Hill road (fig. 3–3). Until a few years ago a railroad track used this low point as a means of exit southward from the Kansas River valley.

The former existence of this glacier-dammed lake was recognized by Hay (1893) and named Kaw Lake by Smyth (1898). At its maximum it extended westward at least 70 mi (112 km) in the Kansas River valley as well as up the valleys of the Big Blue and Republican rivers. The locations of the present towns of Manhattan, Junction City, Abilene, Clay Center, and perhaps Salina were all lakefront properties.

The spillway that controlled the level of Kaw Lake, as well as the spillways holding smaller lakes in lower tributary valleys farther east, the cols seen along I-70 Highway, show no signs of the passage of prolonged, high-volume flow. No channel incision is apparent; that suggests that only small volumes of water escaped from Kaw Lake by this route, and only for a short period of time. A consequential implication is that the ice front remained in its terminal position for only a brief interval.

Even a casual glance at a map of the North American interior clearly shows that at the time of maximum advance of the continental glacier into Kansas and adjacent states, there still remained a large part of the drainage basin of the Missouri River that was ice free. Precipitation falling on that area had to flow around the margin of the ice. Considerable volumes of water, at times, must have been augmented by meltwater from a receding glacier. The Kaw Lake spillway and overflow cols to the eastward did not have the capacity necessary for discharges of the requisite magnitude. The water must therefore have gone elsewhere.

The present floodplain of the Kansas River overlies alluvial fill as much as about 190 ft (58 km) thick at Kansas City and several tens of feet thick upstream. Thus, there is a deep, wide bedrock valley and a narrow inner gorge now concealed beneath younger sediment. There is the possibility that high-volume stream discharge followed a sub-glacial tunnel eastward to Missouri. Such an occurrence was considered an impossibility only a few years ago, but streamflow in tunnels beneath glaciers, resulting in the carving of subice tunnel valleys is now being proposed for many localities in North America and Europe. Even so, it does not appear reasonable to suppose that a large stream could maintain tunnel flow for a distance of almost 100 mi (160 km) beneath an active ice sheet. A better explanation might be that the glacier front did not extend south of the Kansas River valley for more than a few years.

FIGURE 3-8. Sea of erratics on ridge crest 5 mi (8 km) south of Wamego. More than 99% of these clasts, displaying a remarkably small range in size, are composed of Sioux Quartzite.

70.3 **Pass through downtown Wamego.**

71.0 **Continue north to U.S. Route 24. Turn right (east) one block.**

REST STOP at McDonald's.

Leaving McDonald's, turn left (west) one block, then right (north).

Continue north on Kansas Highway 99.

74.3 Pass through edge of town of Louisville.

80.0 The route joins the line of the Oregon–California Trail (see fig. 3–1).

85.1 A major campground on the Oregon–California Trail occupied all of the lowland directly ahead. Water was secured from nearby springs.

85.7 Town of Westmoreland.

Turn west (left) on the first street (State Street).

Note Sioux Quartzite boulders in house yards and beside the road.

Continue west through several stop signs, eventually joining a highway.

86.9 **Drive into Highway Department storage yard on the southern side of the highway.**

STOP 4.

The presence of small boulders of Sioux Quartzite in house yards is not proof that the area has been glaciated. This is an attractive ornamental rock that is often transported considerable distances for purely decorative uses. However, visible from this stop are several large Sioux boulders lying in the field to the south. Almost surely these lie where they were left by the melting glacier (except where they have been moved by presently active construction of new houses). Therefore, the terminal boundary must be located farther to the west (10 mi [16 km] according to the *Geologic Map of Kansas* [Kansas Geological Survey, 1991]). The route northward from Wamego was entirely within the area of glaciation, but no till or erratics were visible.

Return eastward through downtown Westmoreland to Kansas Highway 99.

88.2 **Turn left and continue northward on Kansas Highway 99.**

The route passes through a highly dissected terrain cut in Permian limestones. This is, in essence, a northern extension of the Flint Hills. The area has been glaciated, but no evidence of the former presence of ice is visible from the road. It is especially noteworthy that none of the draws are choked with Sioux Quartzite boulders, as would be expected if a thick cover of till had been largely removed by postglacial erosion.

110.1 Town of Frankfort.

Turn left (west) on Kansas Highway 9.

122.6 **Turn right (north) on U.S. Highway 77.**

129.5 **Turn right (east) on gravel Osage Road.**

130.1 **Turn right (south) into borrow pit.**

130.2 **STOP 5.**

This excavation, which has been active for several years, is a source of sand and gravel for the Marshall County road and public works department. Exposed is a jumble of pebble to small boulder gravel plus a small quantity of actual till. Most of the clasts are granitic, some weathered to grus ghosts. Also present are fragments of dark igneous and metamorphic rocks, but only a few of Sioux Quartzite. Interpretation of this site must be combined with that for the closely adjacent next Stop.

Return to U.S. Highway 77.

130.9 **Turn right (north) on U.S. Highway 77.**

131.1 **Turn right (east) on quarry access road.**

131.5 **STOP 6.**

This excavation is a private commercial operation. However, the sediment deposit is a continuation of that seen in the County pit at Stop 5 and the two must be considered as a single entity.

The suite of clasts exposed here contains the same general distribution of lithologies as is seen in the County pit. Sioux Quartzite is scarce. The prevalence of small- to medium-sized boulders provides evidence of a vigorous stream, the northern margin of which has been exposed as an irregular, nearly vertical bedrock cliff that effectively limited quarry development. Foreset bedding indicates flow toward the east, a surprising direction because that is toward the interior of the glacier. Additional spatial information is needed for further interpretation.

Return to U.S. Highway 77. Turn right (north).

135.6 Marysville. **Turn right (east) on U.S. Highway 36**

167.1 Passing the town of Seneca.

168.3 **Turn right (south) on Kansas Highway 63.**

168.9 **Turn right into borrow pit.**

STOP 7.

This small borrow pit exposes a type of till not seen at any of the preceding stops. It has a clay matrix and contains very few clasts larger than pebble size. Casual search finds no Sioux Quartzite, though at least a few fragments are probably present. Oxidation has produced tan iron-staining in the upper part of the section, an effect that has penetrated more deeply along cracks in the original gray portion.

The Midwest Friends of the Pleistocene field trip in June 2000 stopped at this locality. Its guidebook states that when examined in 1999 the excavation face exposed two "matrix-dominated loamy diamictons with inclusions of deformed sand and gravel," heavily jointed, and oxidized in the upper part (Mandel and Bettis, 2000). Till diapirs in deformed stratified sediments suggested ice movement from the northeast. The 2000 guidebook correlates these two diamictons with the upper and lower tills of the Independence Formation as has been proposed by Aber from exposures west of Atchison. In a personal communication in 2004, Rovey suggested that these are the A and B tills of Boellstorf.

Return to U.S. Highway 36.

169.7 **Continue east on U.S. Highway 36.**

Decorative boulders of Sioux Quartzite in house yards may have been brought in from nearby sources.

183.5 **REST STOP.**

Proceed north on U.S. Highway 75.

186.5 **Turn right (east) on dirt road.**

186.7 **Turn right into borrow pit.**

STOP 8.

This small borrow pit has been sporadically active for several years. Subsurface relationships must be complex; details of the exposure have changed frequently as excavation continued. In the past, a generalized description reported the presence of silty colluvium overlying fine-grained till in the upper part of which is a weathering zone believed to be the Sangamon Paleosol. A very few Sioux Quartzite boulders are present. Most noteworthy is a localized concentration of generally elongated, white caliche nodules scattered along fractures in the till. This phenomenon has been observed at only three other localities in northeastern Kansas and northwestern Missouri. Its significance is unknown.

Return to U.S. Highway 75. Turn left (south).

190.2 **Turn left (east) on U.S. Highway 36.**

193.5 Passing through Fairview.

204.6 **Exit to U.S. Highways 73/159 north.**

205.1 **NIGHT STOP. Hiawatha Inn.**

SECOND DAY

0.0 **Depart Hiawatha Inn.**

Proceed south to U.S. Highway 36.

Turn left (east) on U.S. Highway 36.

33.3 **Enter Wathena. Turn right (south) on 4th Street.**

Bend right to cross bridge and railroad.

Bend left and proceed south on Monument Street (also known as the River Road).

Enter the floodplain of the Missouri River and follow the base of the valleyside bluff.

34.1 STOP 9

An eroded, overgrown roadway leads to a long-abandoned borrow pit, source of sand, gravel, and fill-dirt in the late 1960's. To the left (south) there is a large face nearly 140 ft (43 m) high, having north-south orientation, that displays interbedded and deformed till and glaciofluvial sand. To the right (north), and completely separate, there was an excavated east-west face about 38 ft (12 m) high. As of early 2004, this had completely collapsed and was further obscured by trees and brush. An attempt will be made to re-excavate this site before the field trip.

This small pit is known colloquially as the Sudi site because it was the subject of a 1971 Master's thesis by Sudi Einsohn (since deceased). It was more formally referred to as the Wathena South Pit No. 1 (Bayne, 1968) even though, as Einsohn (1971) noted, it was not the first pit south of Wathena, nor was it the first to be excavated. Nevertheless, its uniqueness gives it considerable geological importance.

As far as is known, this is the only North American location where faunal remains have been found stratigraphically beneath a pre-Illinoian till (fig. 3–9). Designated the Wathena Local Fauna by Martin and Schultz (1985), the assemblage includes 14 species of fish, a salamander, two species of turtles, four waterfowl and a towhee, plus 12 species of shrews, ground squirrels, mice, muskrats, lemmings, and voles. Accumulation in a fluctuating, swampy pond environment, perhaps an abandoned river channel, was suggested.

At the time the thesis was being written in 1970, assignment of an age to this fauna was difficult. According to information then current, several of the species present were believed to be of Late Kansan or even post-Kansan (Yarmouthian) age; two were assigned to the Illinoian! But the fauna occurred beneath a till—and conventional wisdom demanded that no till present in Kansas could be younger than the then-accepted Kansan Glaciation.

In 1985 Martin and Schultz included the Wathena Local Fauna in the Sappan, a subprovincial age within the Irvingtonian. They stated that the Sappan faunas in Kansas are post-Blancan, but pre-Kansan. They also reported that the type Sappa Local Fauna in Nebraska occurred beneath a volcanic ash that had been dated at 1.20 my B.P.

About the best that could be said was that the fauna present at the Sudi site lived sometime before some advance of glacial ice into northeastern Kansas and that that advance was most probably part of what has in the past been called the Kansan Glaciation, although the location lies within the area previously believed to have been covered by the preceding Nebraskan Glaciation. (The terms Kansan and Nebraskan are now in disrepute.)

In a 1988 publication (Aber, 1998a), repeated in 1991, Aber presented a measured section attributed to a "gravel pit south of Wathena" and referred to Einsohn's work. He called attention to the presence of the Wathena local fauna at the base of the diagrammed exposure. In support of scientific integrity, attention must be directed to the fact that this diagram appears to be erroneous.

The diagram shows the 140-ft (43-m)-high face of the *southern* excavation. This is *not* the location of the Wathena Local Fauna. The faunal assemblage was present at the smaller Sudi site, which is situated entirely separate from and about 300 ft (92 m) north of the large face. As pointed out in Einsohn's thesis (1971), there was no similarity in the stratigraphy of the two pits and cross correlation was impossible. Einsohn specifically stated that "no vertebrate remains were found in any of these new exposures (the large, high face south of the Sudi site). The Wathena South No. 1 pit is the only Pleistocene exposure in northeastern Kansas in which vertebrate remains have been reported" (Einsohn, 1971, p. 34).

The once spectacular high face of the pit adjacent to and south of the Sudi site is now badly slumped. However, the upper part still reveals some of the deformed interbeds of both sand within till and till within sand. When freshly exposed, the most noteworthy features were sharply bounded, angular blocks of well-stratified sand diversely oriented and encased by till. These were outstanding examples of frigites, blocks of sand that maintained their identity and shape during transportation because the loose sediment was frozen and acted as a solid. It is clear that this assemblage was deposited in a highly dynamic icefront environment. Limestone boulders are prominent in the till, as well as clasts of igneous and metamorphic rocks.

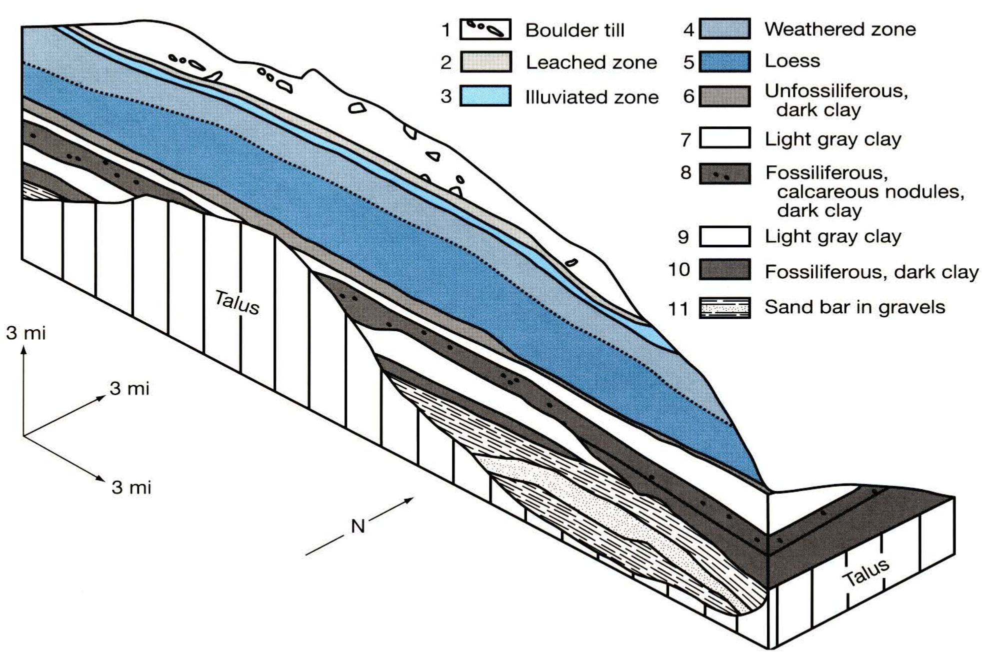

FIGURE 3-9. Diagram of the Sudi site stratigraphy copied from Einsohn's 1971 Master's thesis (Einsohn, 1971).

Return northward to Wathena.

34.9 **Turn left (west) on U.S. Highway 36.**

43.5 **Exit to Kansas Highway 7 and proceed southward.**

Follow Highway 7 into the outskirts of Atchison.

62.9 **Turn right (west) at sign that says "To South 59." Then veer left.**

63.7 **Turn sharp left (south) at sign that says "To South 59."**

63.8 **Junction U.S. Highway 59. Turn right (west).**

At this junction, note the exposure straight ahead.

65.1 **Turn left (south) on Ottawa Road.**

Turn left again into private driveway.

STOP 10.

Although not a large stream, White Clay Creek, as a consequence of its flashy flow regime and displacement against its southern bank by both natural and artificial forces, has accomplished disproportionately great erosion. The result has been creation of the high cutbanks seen here and at the highway junction a mile and a half (2.4 km) east, as well as several smaller faces. These expose spectacular sections of till and glaciofluvial sand and gravel that have attracted the attention of geologists since the early 1900's.

Schoewe (1938) described the presence of two distinct tills here but was uncertain whether or not the lower till was of Nebraskan age. This observation was repeated and extended by Frye (1941), who regarded the lower till as being a product of the Nebraskan Glaciation. This conclusion was repeated, though with less detail and certainty, by Frye and Leonard (1952). It was accepted by Dort as a working hypothesis during the 1960's and 1970's. In 1973 Ward reported the presence of Nebraskan(?) till in the subsurface but not exposed. Dellwig and Baldwin (1965), when analyzing deformation of the two tills and intervening sands, accepted a Nebraskan(?) age for the lower till.

In a paper presented at a meeting in 1982, but not published until later, Aber (1985) discussed the exposure seen at this Stop. (He diagrammed this exposure in this and other papers. See fig. 3–10.) He unequivocally stated that the lower till is of Lower Kansan age and the upper till is of Upper Kansan age, correlating them respectively with the Nickerson Till and the Cedar Bluffs Till of Reed and Dreeszen (1965) in Nebraska. He did not, however, provide clear evidence or reasoning for thus discarding a Nebraskan age for the lower till. In an abstract, Aber (1986) proposed that the Nebraskan should be "demoted" to become "the first substage of the Kansan."

In subsequent publications, Aber lumped all of the strata—lower till, sand and gravel, upper till—into a newly named Independence formation, which then became the type section for all deposits of the Kansan Glaciation. By 1991 he placed on a map the "Early Independence Glacial Limit" and the "Late Independence Glacial Limit" (Aber, 1991). These mimicked the Nebraskan and Kansan limits of previous authors. The result appears to be a grouping of all glacial and glacially related deposits in the state of Kansas into a single unit—the Independence formation, and ignores information from other exposures situated along the Missouri River bluffs north of Atchison (fig. 3-11).

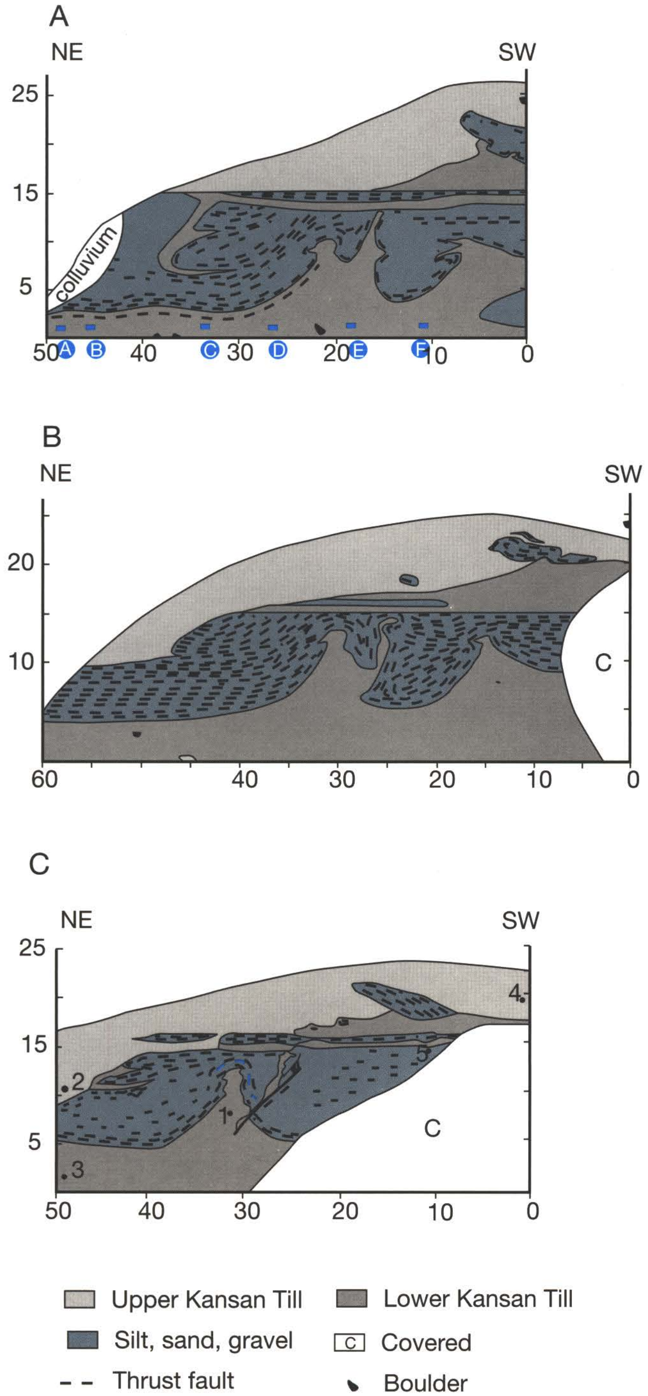

FIGURE 3-10. Measured sections of the exposure at Stop 10 as drawn by Aber in (A) 1988b, (B) 1985, and (C) 1985. Variations may be due to changes in the plane of viewing or in modifications in interpretation. Aber has designated the entire sequence as the type section of his Independence formation, coeval with the Kansas Glaciation.

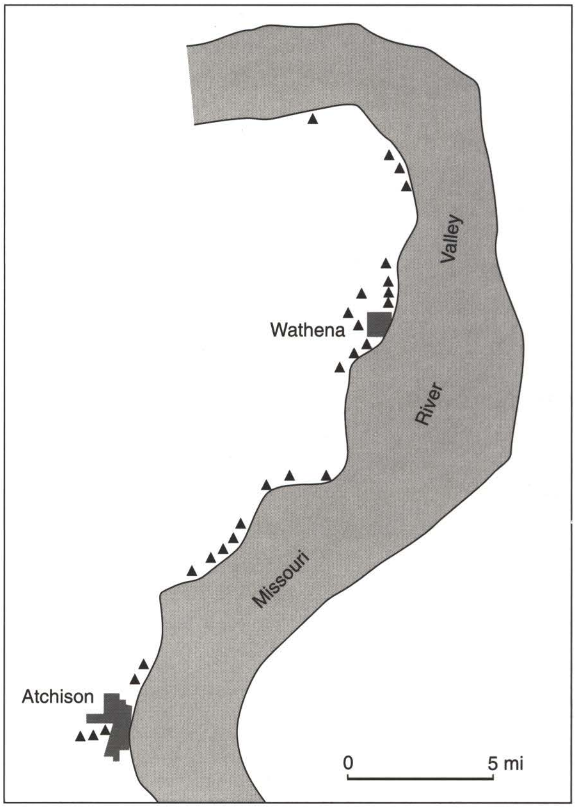

FIGURE 3-11. Some of the quarries and pits that could provide information about Pleistocene deposits along the western side of the Missouri River valley in northeastern Kansas.

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Part II: Missouri

Description of the Harrison and Musgrove Clay Pits

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Harrison Clay Pit

At the intersection of I-70 and State Highway 54 at Kingdom City, Missouri, turn north onto Highway 54 and proceed about 0.4 mi (0.6 km) to the outer or north frontage road. Turn right (east) onto the north frontage road, drive for approximately 3.5 mi (5.6 km), and turn left (north). Follow this gravel road north for slightly more than 0.5 mi (0.8 km) to where the road jogs west for approximately 100 ft (30 m) before turning back to the north. Turn back to the north and continue for another 0.5 mi (0.8 km) to where the road turns to the east. Do not turn east, but continue straight and past the gate into the Harrison Clay Pit.

Discussion

The stratigraphic sequence exposed in the Harrison Clay Pit is summarized in fig. 3–12. This figure is generalized from exposures dating back to the early 1990's; today's exposure is somewhat different. From the top down, the Quaternary sediments include ~2 m of highly weathered, undifferentiated loess overlying a thick Sangamon(?)-Yarmouth weathering profile developed in till of the McCredie Formation. Presently the Columbia member cannot be distinguished as shown in fig. 3–12, although it may simply be obscured by pedogenesis.

In the current exposure the Sangamon(?)-Yarmouth profile is developed mostly, if not exclusively, within the Fulton member of the McCredie Formation. The Fulton member is about 3 m thick and can be distinguished from the younger Columbia member by its lower content of igneous and metamorphic lithologies (fig. 3–12). The basal portion of the Fulton locally includes ~0.5 m of stratified sediments, mostly silts. These sediments bury the underlying Moberly Formation; hence, the weathering profile preserved atop the Moberly was not truncated by subglacial erosion.

The Moberly Formation is distinguished from the younger McCredie Formation here, as at all locations, by a lower content of expandable clay minerals (fig. 3–12). The Moberly also retains a reversed detrital remanent magnetization in contrast to the normal magnetization within the McCredie Formation. The weathering profile within the Moberly is modest here, as is true in most instances. In most cases, the upper Moberly contact is marked by a thin (< 0.5 m) leached zone with or without weak soil structure. In this exposure rare clay skins coat ped surfaces, but this is the only such exposure known; generally the weak soil development is marked primarily by a concentration of redox features with very little (if any) clay enrichment and/or clay mineral alteration.

The oxidized zone within the upper Moberly Formation is generally <2 m thick in this exposure, but locally extends farther downward along polygonal joints. The lower 3 m of the Moberly generally remain unoxidized and preserve abundant wood fragments ranging in size from twigs to logs.

The oldest unlithified sediment at the Harrison Pit is buried by the Moberly Formation. This material is known informally as the "Whippoorwill formation" (see discussion by Rovey et al., this volume, p. 3-B-1). The Whippoorwill has been poorly exposed for most of the past 14 years, but it is a very fine grained diamicton with local accumulations of pebbles and cobbles. These clasts are exclusively of local bedrock lithology and include no igneous or metamorphic varieties; hence, they do not indicate glacial transport. The Whippoorwill apparently originated as some type of mass-flow into local topographic lows. This interpreted origin is consistent with a history of clay mining, because the best clay deposits in this region are often localized within paleokarst networks.

Musgrove Clay Pit

From the Harrison Pit, follow the route backwards to the I-70 and Highway 54 intersection at Kingdom City. Turn east back onto I-70 and follow it for 27 mi (43 km) to the I-70 and Highway 18 intersection at New Florence, Missouri. Exit I-70 south onto Highway 19. From the end of the exit ramp, follow Highway 19 to the south for approximately 2.7 mi (4.3 km) and turn left (east) past a gate and onto the gravel haul road leading into the Musgrove Pit.

Discussion

The stratigraphy at the Musgrove Pit is summarized in fig. 3–13. This figure is based on exposures from the mid- 1990's; the current highwall has a greater thickness

(~12 m) of Moberly Formation. Additionally, the paleosol preserved atop the Atlanta Formation is somewhat thinner now than depicted in fig. 3-13, and the calcic horizon (the 3B2kgb horizon) is much less prominent.

The uppermost sediment exposed at the Musgrove Pit includes several meters of undifferentiated loess. As the highwall has been moved back to higher elevations, the thickness of loess has increased and multiple increments of loess are apparently preserved.

The uppermost till exposed in the Musgrove Pit is that of the Moberly Formation. In today's exposure the basal portion remains unoxidized and preserves abundant wood fragments. The Moberly till at this location contains distinct indicator clasts, including pink quartzite and most importantly Neda Formation, a distinctive oolitic iron ore. The westernmost exposure of the Neda Formation is just south of Dubuque, Iowa; hence, its presence within the Moberly indicates a regional flow direction from the northeast. The Moberly Formation here again preserves a reversed detrital remanent magnetization, indicating that it was deposited prior to the Matuyama/Brunhes reversal at 0.78 Ma.

The Moberly Formation buries a thick, well-preserved paleosol developed atop the underlying Atlanta Formation. This unnamed paleosol locally preserves a B horizon >2 m in thickness, which indicates extreme weathering and prolonged development (see the texture and clay mineral trends in fig. 3-13). The upper portion of this paleosol preserves abundant striae and stress cutans, indicating extensive shearing and shrink-swell deformation, which are consistent with the extremely high (~80%) content of expandable clay minerals.

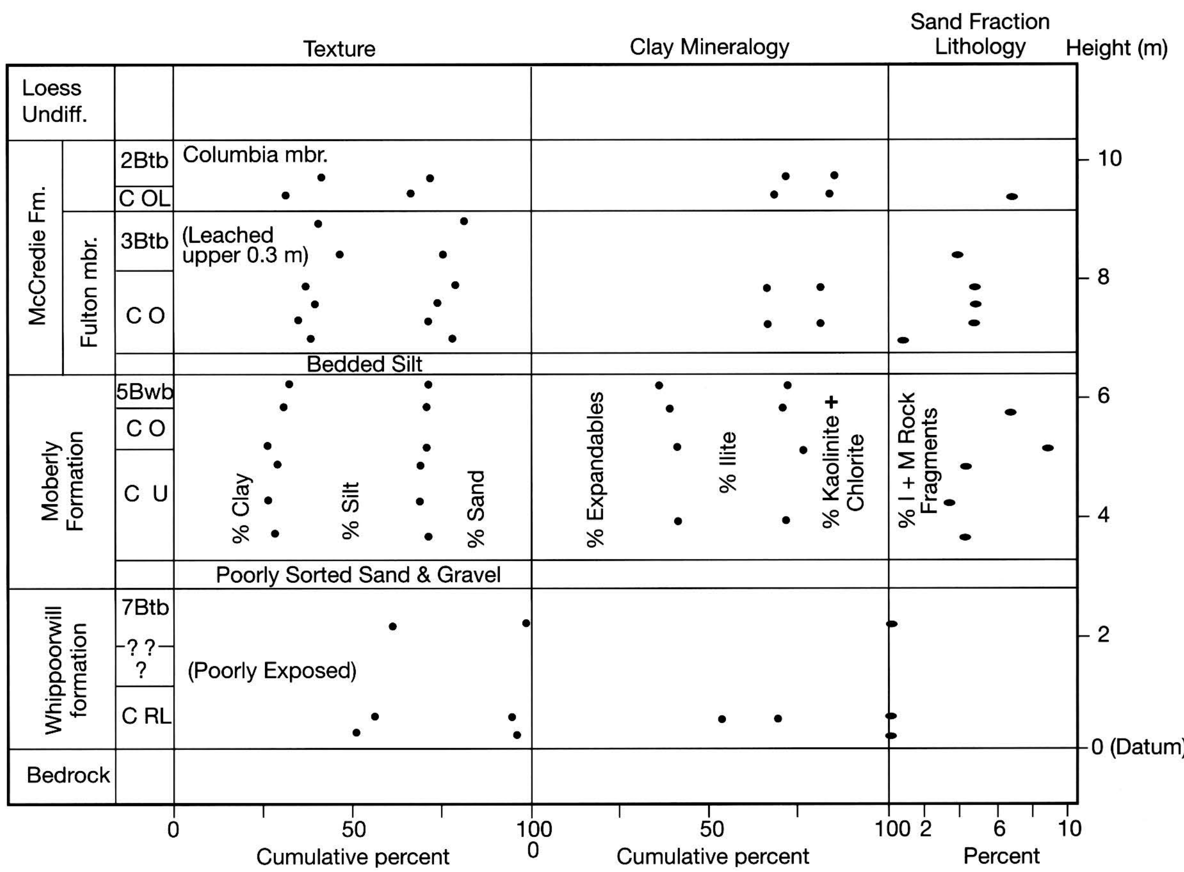

FIGURE 3-12. Lithostratigraphy of glacial and related deposits at the Harrison Pit.

A laminated silt at the base of the Atlanta Formation preserves a reversed detrital remanent magnetization (as does the underlying Whippoorwill formation), indicating that the Atlanta Formation was also deposited during the Matuyama Reversed Chron. The extremely well developed weathering profile preserved atop the Atlanta, however, indicates that the Atlanta Formation was deposited during a much earlier portion of the Matuyama Chron than the overlying Moberly Formation (see also the discussion in Rovey et al., this volume, p. 3-B-1). If so, then the Atlanta Formation is probably Pliocene in age.

The Atlanta Formation buries ~3 m of the Whippoorwill formation, which in turn overlies Pennsylvanian-age bedrock. A weakly developed paleosol atop the Whippoorwill is preserved locally beneath a laminated silt within the base of the overlying Atlanta Formation. Thus, the Whippoorwill was apparently deposited shortly before the area was glaciated.

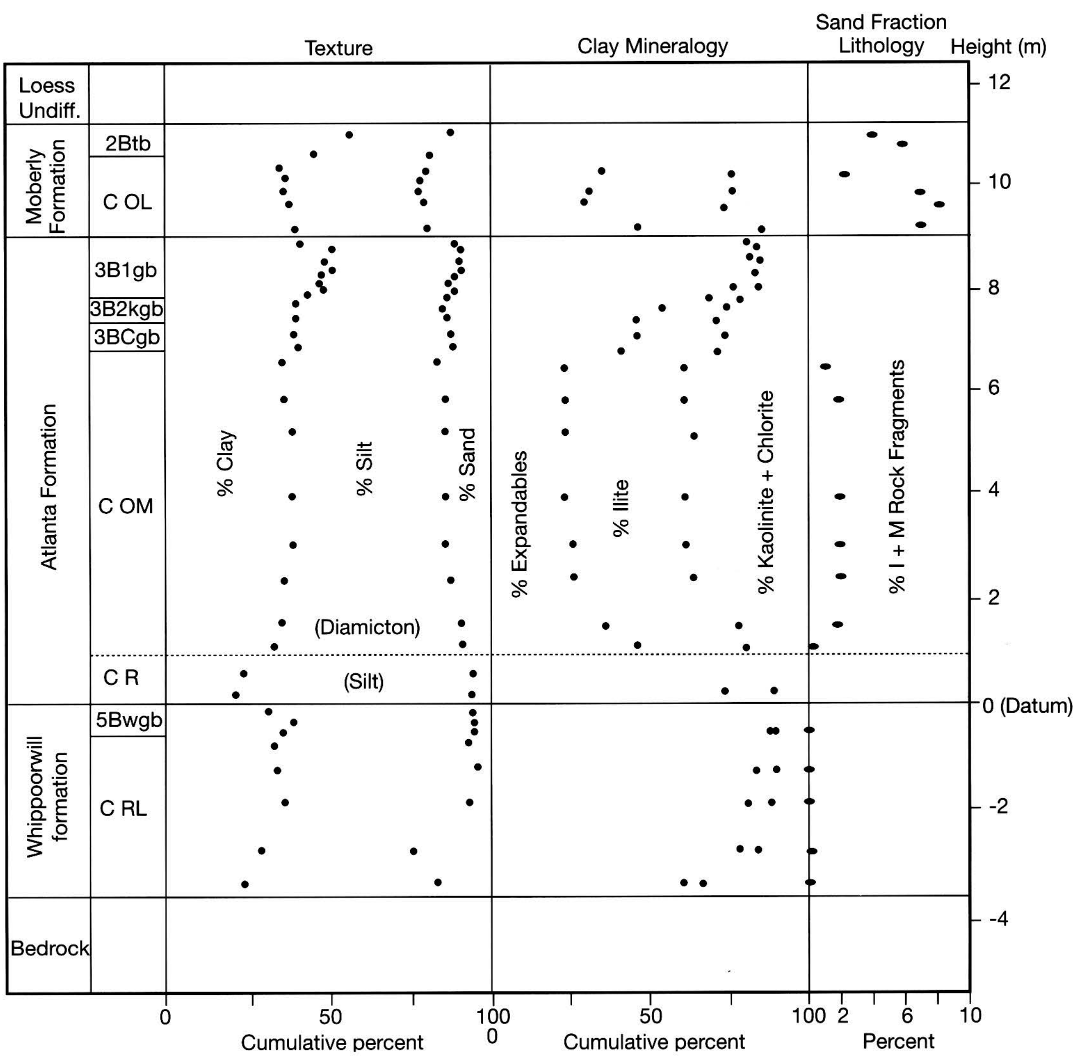

FIGURE 3-13. Lithostratigraphy of glacial and related deposits at the Musgrove Pit.

Lithostratigraphy of Glacigenic Sediments in North-central Missouri

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Abstract

Five glacial tills are present in north-central Missouri along the southernmost boundary of the pre-Illinoian Laurentide ice sheet. These five tills and associated sediment are grouped into three units of formation rank, based upon clast and matrix lithology.

The oldest unit, herein named the Atlanta Formation, is preserved in stable landscape positions and is capped by an extremely well developed weathering profile. The Atlanta Formation is also distinguished by its high clast content, most of which are chert and limestone. The Atlanta preserves a reversed detrital remanent magnetization, indicating that it is older than the Matuyama/Brunhes boundary at 0.78 Ma.

The Moberly Formation locally overlies the Atlanta Formation but is generally the oldest till preserved above bedrock. The Moberly Formation is distinguished primarily by its clay-mineral content, which averages 39% expandable clay minerals. The Moberly Formation also retains a reversed detrital remanent magnetization, indicating that the Moberly was deposited during a later portion of the Matuyama Reversed Chron, before being buried by the normally magnetized McCredie Formation.

The McCredie Formation includes at least three lithologically similar tills, which are separated vertically from each other by mature weathering profiles. The McCredie Formation is distinguished primarily by its high (>50%) content of expandable clay minerals. From oldest to youngest, three informal subdivisions of the McCredie are designated as the Fulton, Columbia, and Macon members, based on stratigraphic position and differences in texture and sand-fraction lithology.

Introduction

In north-central Missouri, five glacial tills are separated vertically from each other by well-preserved soil/weathering profiles, and all are present within a few kilometers of the southern glacial boundary. This boundary marks the southernmost extension of the pre-Illinoian Laurentide ice sheet (fig. 3-A-1) and is virtually coincident with the northern margin of the Ozark Dome. Apparently the largest pre-Illinoian glaciations all reached this same approximate terminus in northern Missouri where a thinning ice margin abutted the dome and was stopped by the reversal in regional topographic gradient. Therefore, presence of a till along the southern glacial boundary in north-central Missouri is a useful criterion for designating and distinguishing major pre-Illinoian continental glaciations. Accordingly, northern Missouri is a key area for understanding the pre-Illinoian glacial sequence throughout North America.

The subglacial conditions near the terminal ice margins apparently favored deposition over erosion; buried paleosol/weathering profles are often preserved nearly intact. Thus, B horizons preserved atop each till may eventually afford the opportunity to obtain burial dates using cosmogenic-isotope techniques, and hence to determine dates of the major pre-Illinoian glaciations. Hopefully, the glacial sequence here is complete enough to establish a general stratigraphic framework for grouping and correlating pre-Illinoian deposits throughout a large portion of North America.

Overview and Background

This paper summarizes recent work on the glacial stratigraphy of north-central Missouri, drawing heavily on Rovey and Kean (1996, 2001), Rovey (1997), Tandarich (1992, 2001), Tandarich et al. (1994), and Guccione and Tandarich (1993). We formally define three Cenozoic glacigenic formations and describe three informal members within the youngest of these formations. For continuity and to avoid confusion, we retain previously used informal names, even when recent exposures warrant a type section some distance from where a unit was first named and described.

The basic stratigraphic succession is shown in fig. 3-A-2; lithologic characteristics of each unit are summarized in table 3-A-2. In compiling table 3-A-2, we excluded samples which were either pedogenically altered or were collected from the "basal entrainment zone,"

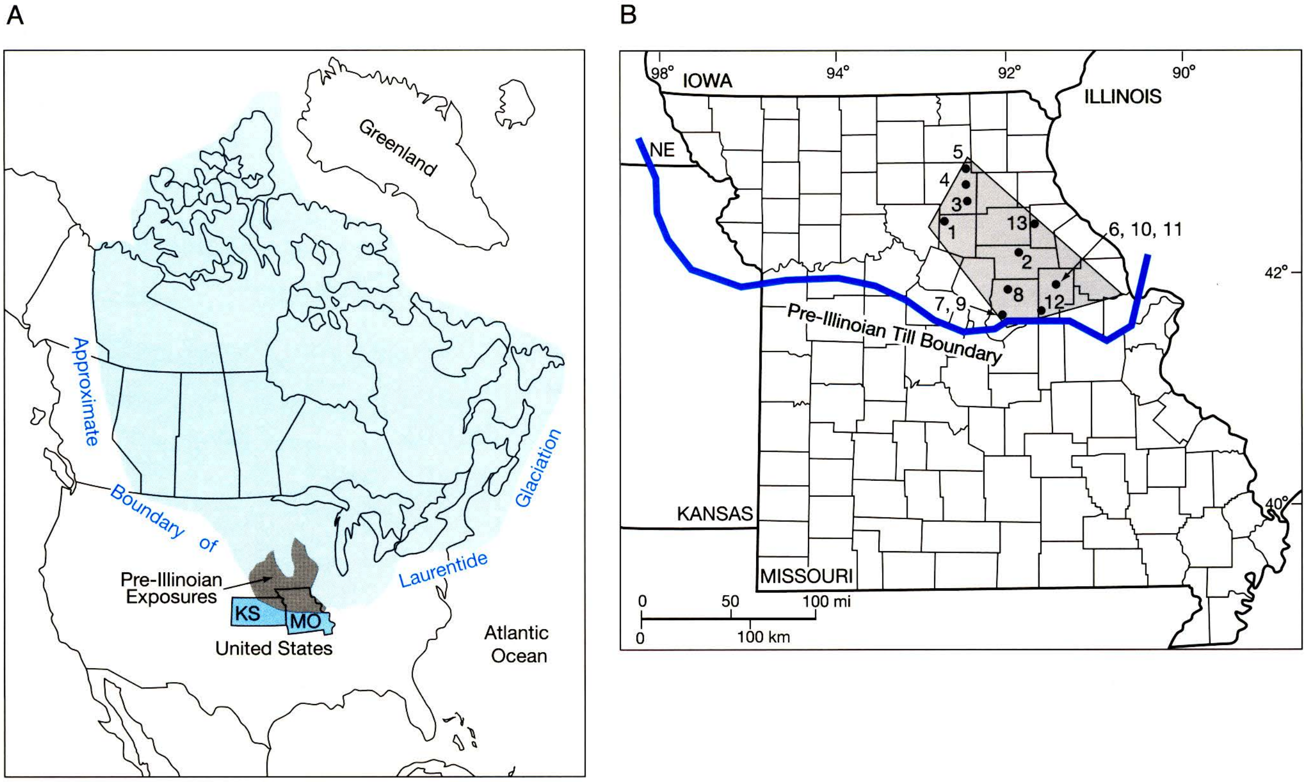

FIGURE 3-A-1. Map showing the present study area and its relationship to the maximum expanse of pre-Illinoian glaciation. Numerals correspond to sites listed in tables 3-A-1 and 3-A-2.

an interval of ~0.5–1 m at the base of a till. Within this lower interval, a till's composition may differ significantly from the more representative overlying material due to a concentration of locally entrained substrate materials.

Atlanta Formation

Source of Name and History: The Atlanta Formation is named for the town of Atlanta, Missouri, in Macon County, near where it was encountered by Rovey and Kean (1996) in corehole SMS-92b (fig. 3-A-1B). The Atlanta Formation was first described and named informally by Allen and Ward (1974), based on exposures within the Deeker Clay Pit. They included it within their Whippoorwill Creek formation, along with an underlying less-pebbly diamicton. Guccione (1983) later restricted the (informal) name Whippoorwill formation to the subjacent (preglacial) diamicton, without designating a name for the overlying cobbly diamicton, which is defined here as the Atlanta Formation.

Although the name "Atlanta" is associated with a major metropolitan area in another state, it is retained for conformity with historical usage—e.g., Rovey and Kean (1996, 2001), Rovey (1997), and Colgan (1999). Moreover, many other geographic names near the designated type area have already been applied formally or informally to different rock or soil stratigraphic units, and few would mistake the metropolitan area of Atlanta, Georgia, as the type area for a glacial deposit.

Type Section/Area: The type section is in the area approximately 5 km south of New Florence, Missouri, in and around the Musgrove, Johnson, and Deeker clay pits (secs. 1, 2, and 11, T. 47 N., R. 5 W., Pinnacle Lake 7.5-min quadrangle).

Description of Unit: The bulk of the Atlanta Formation is a massive, pebbly-cobbly diamicton with a very fine grained matrix (figs. 3-A-3 and 3-A-4). Most of the pebbles and cobbles are whitish chert with subordinate limestone, and these two lithologies dominate the coarse-sand fraction as well. The most common type of crystalline erratics seems to be pink quartzite, but these clasts are rare. Although the diamicton is very cobbly and pebbly, its matrix generally consists of >80% fines. The clay mineral content of the <2-micron fraction is dominated by variable amounts of illite and kaolinite with only small amounts of expandable clay minerals (table 3-A-2, figs. 3-A-2B, 3-A-3).The color of this diamicton is nondiagnostic and is related in all known cases to its weathering history. Most of the clasts were apparently derived from Mississippian-age carbonates, which crop out in numerous areas north of the type area. If this is correct, the Atlanta diamicton may become less cobbly northward.

In the type area, the diamicton described above commonly rests upon a thin (<1 m), unweathered, laminated, clayey silt. This silt feathers against subjacent paleotopographic highs and locally contains scattered pebbles (dropstones) that deform subjacent laminae. The silt also contains pink quartzite in the coarse-sand fraction and is obviously genetically related to the overlying

TABLE 3-A-1. Locations and names of sections shown in fig. 3-A-1 and listed in table 3-A-2.

1. AECI Coal Pit (type area of Moberly Formation). Area around Thomas Hill Reservoir and the large AECI coal mine centered approximately 2 km northwest of Thomas Hill, Missouri. Data in this report were collected from highwalls in the NE sec. 26, T. 55 N., R. 16 W., Prairie Hill 7.5-min quadrangle.

2. Blum Clay Pit. NE sec. 33, T. 52 N., R. 8 W., Mexico East 7.5-min quadrangle.

3. Core SMS-92a. Intersection of Highways 63 and DD, sec. 28, T. 58 N., R. 14 W., Axtell 7.5-min quadrangle.

4. Core SMS-92b (type area of McCredie Formation). Intersection of Highways 63 and M, sec. 21, T. 59 N., R. 14 W., Atlanta 7.5-min quadrangle.

5. Core SMS-92c. Intersection of Highways 63 and D, NW sec. 5, T. 60 N., R. 14 W., LaPlata 7.5-min quadrangle.

6. Deeker Clay Pit (type area of Atlanta Formation). NW sec. 11, T. 47 N., R. 5 W., Pinnacle Lake 7.5-min quadrangle.

7. Fulton Exposure. South of Fulton, Missouri, along Highway 54, NE sec. 2, T. 46 N., R. 10 E., Guthrie 7.5-min quadrangle.

8. Harrison Clay Pit. SW sec. 1, T. 48 N., R. 9 W., Calwood and Kingdom City 7.5-min quadrangles.

9. Highway J Intersection. Intersection of Highways 54 and J, SE sec. 31, T. 46 N., R. 10 W., New Bloomfield 7.5-min quadrangle.

10. Johnson Clay Pit (type area of Atlanta Formation). NW sec. 2, T. 47 N., R. 5 W., Pinnacle Lake 7.5-min quadrangle.

11. Musgrove Clay Pit (type area of Atlanta Formation). Approximately 5 km south of New Florence, Missouri, SE sec.2 and SW sec.1, T. 47 N., R. 5 W., Pinnacle Lake 7.5-min quadrangle.

12. Reedsville Clay Pit. SW sec. 7, T. 46 N., R. 6 W., Reedsville 7.5-min quadrangle.

13. Riedell Coal Pit. SE sec. 20, T. 54 N., R. 7 W., Perry 7.5-min quadrangle.

diamicton. This silt is also present in the basal meter of the overlying diamicton as discrete undeformed inclusions and more rarely along shear planes, which originate in the underlying silt and continue upward into the basal diamicton.

Contacts: The Atlanta Formation is the oldest glacigenic unit recognized in north-central Missouri. As such it overlies various Paleozoic bedrock formations throughout the region. In the type area, however, it generally overlies a preglacial diamicton known informally as the Whippoorwill formation (Guccione, 1983; Rovey et al., this volume, p. 3-B-1). In the clay-pit exposures of the type area, the upper surface is undulatory to hummocky, typically with ~3 m of local relief. The upper surface is marked by an extremely well developed paleosol and weathering profile (fig. 3-A-3). In the Musgrove and Johnson pits, this paleosol preserves a B horizon >2 m thick in the highest paleolandscape positions.

In most cases the upper surface of the Atlanta Formation is buried by the Moberly Formation. In the type area, this contact is typically marked by discontinuous, thin (<0.3 m), laminated, contorted silt and poorly sorted gravelly sand, which in turn is sharply overlain by Moberly diamicton. Locally, however, the Atlanta Formation may be overlain by younger glacial diamictons or undifferentiated loess.

Differentiation: Diamicton of the Atlanta Formation is distinguished visually by its high content of chert and limestone clasts. Laboratory data also distinguish the Atlanta from all other local units (table 3-A-2). The Atlanta has a very low (average 2%) content of igneous and metamorphic rock fragments in the coarse-sand fraction and a low (average 16%) content of expandable clay minerals in the <2 micron fraction. Various mixed-layer clays are also more prominent within the Atlanta Formation, compared to the younger glacial diamictons.

There appears to have been long-standing confusion in distinguishing the present Atlanta Formation from an older subjacent diamicton now known informally as the Whippoorwill formation (Guccione, 1983). However, where both are present, the Atlanta Formation typically rests upon a discrete paleosol that caps the Whippoorwill formation. The unweathered Whippoorwill formation below this paleosol may contain scattered pebbles. Nevertheless, there is little doubt that the Whippoorwill is not a direct glacial deposit (Guccione, 1983; Rovey et al., this volume, p. 3-B-1), although its origin may be related to climate change associated with glaciation. The rare clasts and pebbles in the Whippoorwill include only local sedimentrary lithologies without crystalline erratics, and its lithology varies substantially from one exposure to the next and even from one end of an exposure to the other.

TABLE 3-A-2. Summary of lithologic parameters. Values within each category are normalized percentages. Abbreviations: E, expandable clay minerals; I, Illite; K+C, Kaolinite + Chlorite; Sa, sand; Si, silt; Cl, clay; Q+F, quartz + feldspar; Carb, carbonates + chert; I+M, igneous + metamorphic; n, number of samples. See text and Appendix (p. 3-A-10) for additional explanations. Minor differences between percentages listed here and those in Rovey and Kean (1996) are due to analysis of additional samples at some sites and/or reanalysis of samples at the SMS laboratory.

	Clay Mineralogy				Texture				Sand-Fraction Lithology			
Location	**E**	**I**	**K+C**	**(n)**	**Sa**	**Si**	**Cl**	**(n)**	**Q+F**	**Carb**	**I+M**	**(n)**
Macon member, McCredie Formation												
Core SMS-92a	47	25	28	3*	36	34	30	5	58	28	9	3
Core SMS-92b	48	23	29	3*	37	34	29	6	61	28	11	3
Core SMS-92c	49	21	30	12	38	33	29	13	61	29	10	6
Avg./#Sites	48	23	29	3	37	34	29	3	61	29	10	3
Columbia member, McCredie Formation												
AECI Pit	60	21	19	5	29	38	33	5	61	30	9	7
Blum Pit	55	20	25	4	37	30	33	5	67	23	10	2
Core SMS-92a	52	25	23	4*	35	33	32	2	62	28	10	3
Core SMS-92b	52	25	23	3*	36	30	34	5	55	34	11	5
Core SMS-92c	50	23	27	7*	37	33	30	7	62	28	10	6
Harrison Pit	53	26	21	2	33	35	32	2	51	39	10	6
Fulton Exposure	51	27	22	5	29	40	31	8	49	42	9	6
Riedell Pit	55	19	26	5	37	31	32	13	70	21	9	5
Avg./#Sites	54	23	23	8	34	34	32	8	60	30	10	8
Fulton member, McCredie Formation												
AECI Pit	60	22	18	6	25	36	39	5	60	36	4	4
Core SMS-92a	56	23	21	5*	28	37	35	8	63	31	6	7
Core SMS-92b	50	25	25	5*	28	35	37	5	75	23	2	4
Core SMS-92c	49	22	29	4*	33	34	33	6	62	33	5	5
Harrison Pit	56	27	17	4	23	38	39	6	58	38	4	4
Hwy. J Intersection	49	28	23	6	23	40	37	5	43	52	5	4
Deeker Pit					25	36	37	3	58	38	4	2
Avg./#Sites	53	25	22	6	26	37	37	7	60	36	4	7
Moberly Formation												
AECI Pit	41	35	24	5	29	39	32	16	57	32	11	5
Core SMS-92a	41	27	32	3*	30	38	32	4	60	32	8	4
Core SMS-92b	39	28	33	4*	31	40	29	6	63	30	7	7
Deeker Pit	42	38	20	5	27	43	30	5	43	52	5	6
Fulton Exposure	35	39	26	5	23	43	34	5	49	46	5	5
Harrison Pit	39	36	25	3	30	43	27	4	57	38	5	4
Musgrove Pit	34	39	27	6	24	41	35	8	44	51	5	8
Avg./#Sites	39	34	27	7	28	41	31	7	53	40	7	7
Atlanta Formation												
Core SMS-92b	2	57	41	4	27	40	33	3	27	70	3	4
Johnson Pit	21	46	33	8	17	44	39	8	12	86	2	8
Musgrove Pit	26	36	38	5	13	50	37	6	7	91	2	6
Reedsville Pit	15	56	29	2	14	50	36	2	19	79	2	3
Avg./#Sites	16	49	35	4	18	46	36	4	16	82	2	4

*Analysis by IGS

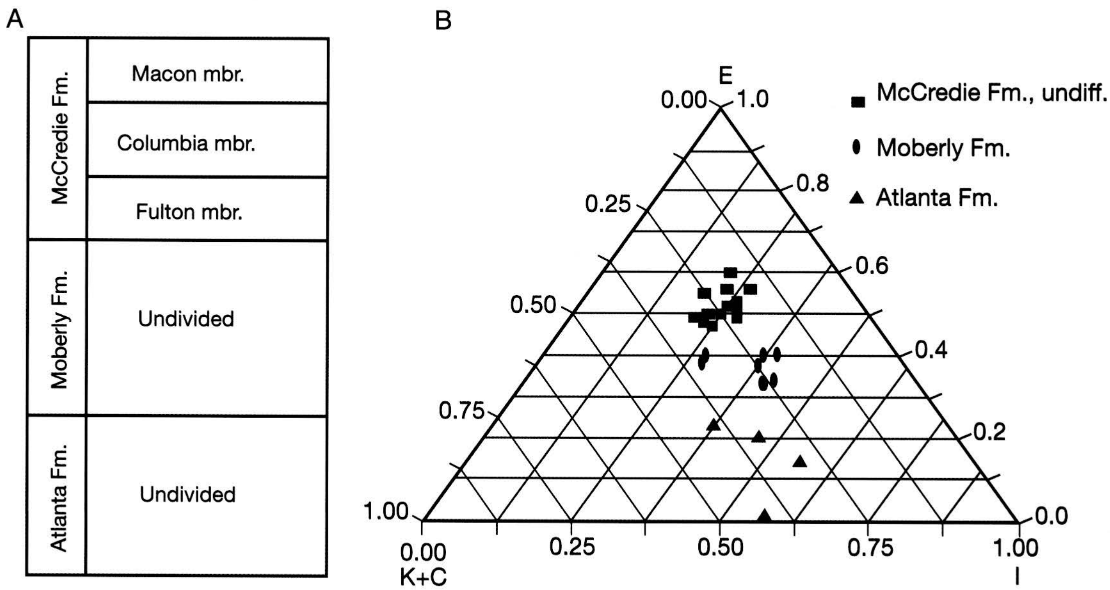

FIGURE 3-A-2. Stratigraphy of pre-Illinoian glacial deposits in north-central Missouri. (A) Stratigraphic column, and (B) Summary of clay mineralogy. Each formation has a unique clay-mineral composition. Data points are the site averages listed in table 3-A-2.

Thickness and Regional Extent: The thickness of the Atlanta Formation is quite variable, due to prolonged erosion prior to burial. In the type area it is typically ~6 m thick with a range between 4 and 7 m.

The Atlanta Formation is generally preserved only locally in very stable landscape positions; elsewhere it is completely eroded. The type area south of New Florence, Missouri, appears to be part of a paleokarst network, which explains the ubiquitous presence of paleotopographic lows ideally suited for protecting the Atlanta Formation from erosion prior to burial by younger deposits. Despite its patchy preservation, the Atlanta is easily recognized in numerous clay-pit exposures south of I-70 extending at least 40 km along a nearly east-west traverse centered around New Florence. Guccione (1973) also described an identical diamicton isolated within bedrock lows and below unweathered, reverse magnetized intertill silts (Moberly Formation) from the Columbia, Missouri, area. North of I-70 preservation of the Atlanta Formation is rare and it has been found at only one locaton to date (corehole SMS-92b). Nevertheless, the widespread occurrence of remnants testifies to a former continuous distribution prior to an extended period of weathering and erosion.

Interpreted Origin: The diamicton that constitutes the bulk of the Atlanta Formation is interpreted as a basal melt-out till, based on its massive unstratified nature, uniform texture, and lack of flow structure and fissility. Supraglacial till is generally quite variable in texture, reflecting partial sorting during mass flowage and resedimentation (e.g., Hallberg, 1980a; Kemmis et al., 1981), and also preserves flow structure and a quasi-stratification. The lack of fissility and "streaked" inclusions near the base of the till are evidence against a lodgment origin. The shear planes and partially deformed and remobilized substrate material in the basal meter may indicate local accumulations of deformation till, but this facies is very minor. The laminated silt, which is present beneath the till in the type area, is interpreted as a proglacial lacustrine silt that accumulated locally in paleotopographic lows.

Age and Correlation: The laminated silt facies at the base of the formation and the underlying Whippoorwill formation both have a reversed detrital remanent magnetization (DRM) (Rovey et al., this volume, p. 3-B-1). A local accumulation of silt near the top of the diamicton at the Johnson Pit (but below the weathering profile) also preserves a reversed DRM. Considering also that the formation is capped by the reverse magnetized Moberly Formation (Rovey and Kean, 2001), the Atlanta was deposited during the Matuyama Reversed Chron (0.78–2.58 Ma). The extremely well developed weathering profile developed atop the Atlanta, along with its general absence in all but the most stable landscape positions, are evidence that it is considerably older than the overlying Moberly Formation, and therefore must date from a much earlier portion of the Matuyama Reversed Chron than the Moberly.

Rovey and Kean (1996) suggested that the Atlanta Formation is closely related to the "C-till" in eastern Nebraska and western Iowa (e.g., Easterbrook and Boellstorff, 1984). The age of the "C-till" is >2.2 Ma, based on a single fission-track data of overlying tephra (Boellstorff, 1973a, b)

Source Area: The overwhelming dominance of sedimentary lithologies within the Atlanta diamicton may be partly related to its stratigraphic position as the oldest glacial deposit; the first glacial advance would have entrained a highly weathered local regolith. Nevertheless, the very low percentage and *variety* of crystalline lithologies indicates a distinctly different source area than

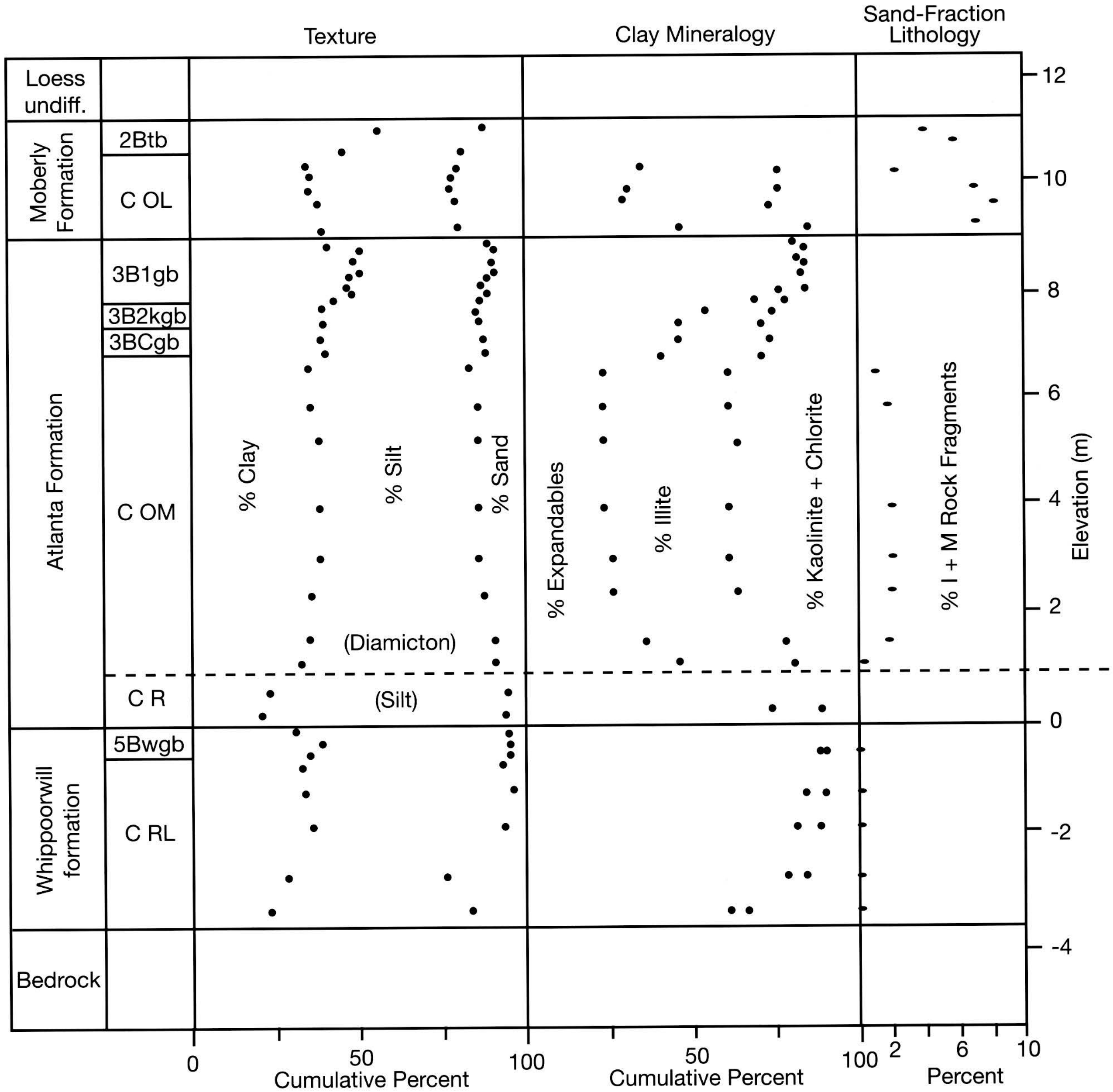

FIGURE 3-A-3. Stratigraphy and lithology of glacial sediments at the Musgrove Clay Pit (type area of Atlanta Formation). Weathering zone (subsolum) terminology: O, oxidized; L, leached; M, mottled; R, reduced; U, unoxidized. Datum is the base of the Atlanta Formation.

the younger glacial deposits. If the rare clasts of pink quartzite had been derived from Baraboo-type quartzite, which is exposed to the northeast, the diamicton should have additional crystalline lithologies derived from the Canadian Shield area. If, however, the quartzite is Sioux Quartzite, which is exposed to the northwest, flowlines would have followed sedimentary lithologies over a much longer distance, accounting for the overwhelming dominance of sedimentary lithologies. Therefore, the ice which deposited the Atlanta Formation probably advanced from a northwesterly direction.

Moberly Formation

Source of Name and History: The Moberly Formation is named for the town of Moberly, Missouri, in Macon County. This unit was originally included within the undifferentiated and informal "McCredie formation" of Guccione (1983). Tandarich (1992, 2001), however, recognized that the basal portion of Guccione's McCredie formation includes a (visually) subtle weathering profile above an interval with a distinct matrix lithology. Tandarich (1992, 2001) therefore proposed the name "Moberly" for this interval, as was reiterated by Rovey and Kean (1996.)

Type Section/Area: The type section is located in an area around Thomas Hill Reservoir and the Associated Electric Co-Op Incorporated (AECI) coal mine, centered approximately 2 km northwest of Thomas Hill, Missouri (T. 55 N., R. 15 and 16 W., Prairie Hill 7.5-min quadrangle), and approximately 20 km northeast of Moberly, Missouri.

Description of Unit: The Moberly Formation is typically a massive diamicton with a fine-grained matrix but locally includes stratified glacigenic silts and sands, particularly at the base. Moberly diamicton includes a wide variety of crystalline erratics, and at least in the exposures along I-70, it commonly contains concentrations of wood fragments near the base. Where the Moberly Formation is buried by one or more younger glacial diamictons, the lower half to two-thirds generally remains unoxidized with a nearly black color due to disseminated organic matter. The uppermost portion is generally oxidized to various brownish hues. Preliminary data obtained from boreholes taken by the Missouri Department of Natural Resources, Geological Survey and Resource Assessment Division, in St. Charles County, Missouri (east of the present study area), indicate that the Moberly may contain multiple tills with virtually identical lithology.

Contacts: Throughout most of north-central Missouri, the Moberly Formation is usually the oldest glacigenic unit preserved atop Paleozoic bedrock. In rare cases, but more commonly south of I-70, it may overlie the Atlanta Formation or deeply weathered remnants of the Whippoorwill formation. The upper formation contact is typically marked by a subtle weathering profile and soil horizon. This soil/weathering profile is generally buried by glacial diamicton and related sediment of the (informal) Fulton member of the McCredie Formation. Locally, however, the Moberly Formation may be buried by younger informal members of the McCredie Formation or by undifferentiated loess.

Differentiation: Moberly diamicton has fewer clasts than the subjacent Atlanta Formation, although the Moberly Formation contains a much higher concentration and variety of igneous lithologies. The Moberly can also be provisionally distinguished from the overlying Fulton member of the McCredie Formation by clast lithology; the Fulton diamicton also contains fewer igneous clasts than the Moberly diamicton.

Laboratory measurements are generally necessary to confirm an assignment to the Moberly Formation. The most distinct and useful parameter is the clay-mineral content (figs. 3-A-2B, 3-A-3, 3-A-5, and 3-A-6; table 3-A-2). The Moberly Formation averages 39% expandable clay minerals, whereas the McCredie Formation typically has >50%. Although some compositional parameters of the Moberly diamicton overlap with those of the overlying Fulton member of the McCredie Formation over a regional scale, the Moberly diamicton has less clay and more igneous and metamorphic rock fragments in the coarse-sand fraction at any given location.

Thickness and Regional Extent: The Moberly Formation is nearly always present where younger glacial diamictons of the McCredie Formation are preserved. Where it is buried by the McCredie Formation, the Moberly Formation is typically 4–5 m thick; however, a much greater thickness (>18 m) is present in the type area.

Interpreted Origin: The bulk of the Moberly Formation is interpreted as basal meltout till, based on the same criteria discussed earlier for the Atlanta Formation.

Age and Correlation: Both the diamicton and unweathered subjacent silts have a reversed DRM (Rovey and Kean, 1996, 2001). Therefore, the Moberly Formation is older than the Matuyama/Brunhes boundary at 0.78 Ma. The weathering profile and soil developed atop the Moberly is relatively modest in most exposures examined to date. Even where the upper paleosol surface was buried by proglacial silt and thus protected from subglacial erosion, the B horizon is relatively immature. This soil horizon is characterized by blocky structure with minor (≤4%) maximum clay enrichment, few to no clay cutans, and virtually no clay mineral alteration. Rovey and Kean (2001) therefore concluded that this paleosol developed during a relatively short interglacial period prior to being buried by the normally magnetized McCredie Formation. If this interpretation is correct, the age of the

FIGURE 3-A-4. Atlanta Formation. Photograph shows the contact between the Atlanta Formation diamicton (above) and the Whippoorwill formation (below) at the Johnson Clay Pit.

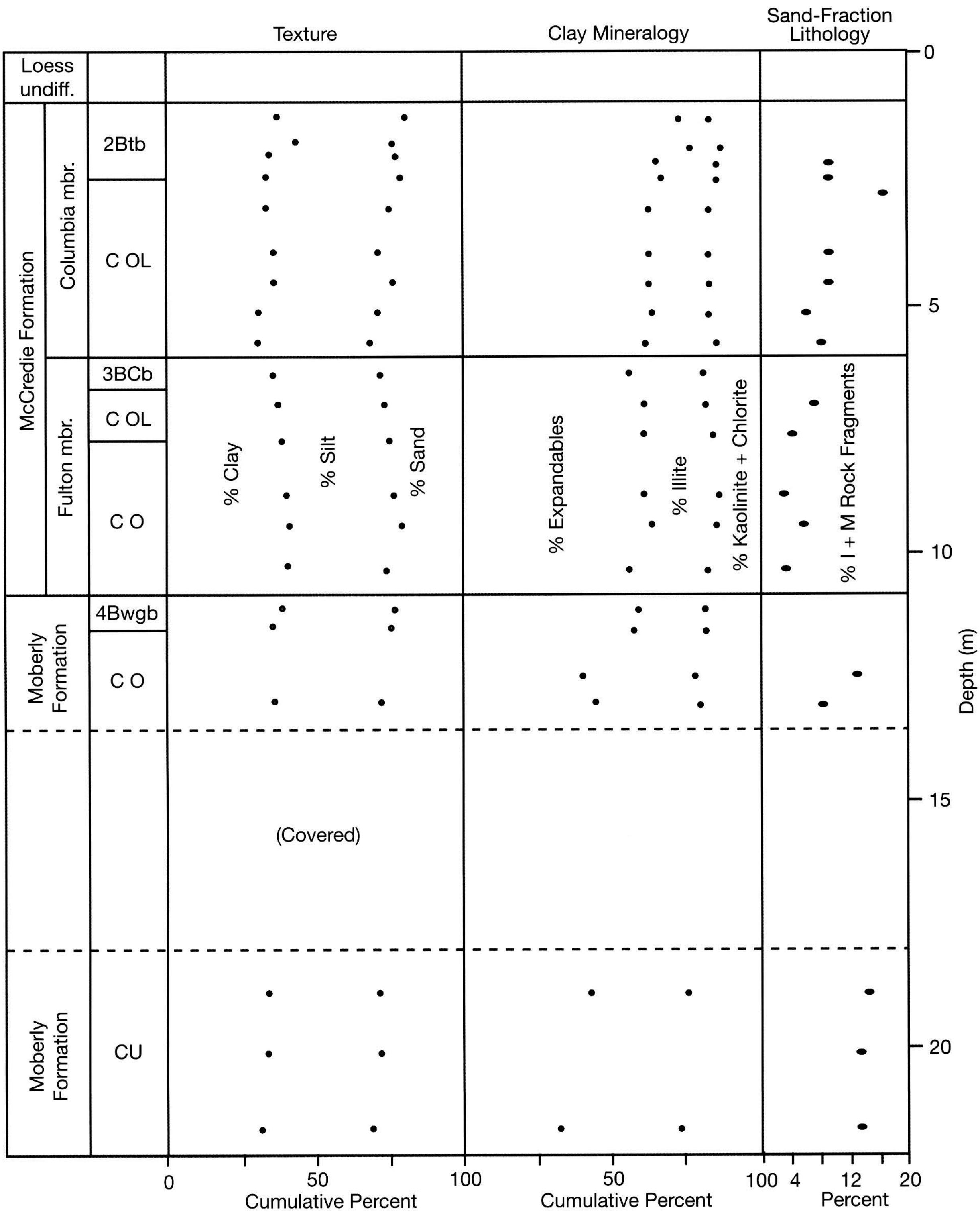

FIGURE 3-A-5. Stratigraphy and lithology of glacial sediments at the AECI Pit (type area of Moberly Formation). See fig. 3-A-3 for weathering-zone terminology. The base of the Moberly Formation overlies bedrock at a depth >24 m, but this contact was inaccessible at the sample location.

Moberly Formation would be slightly older than that of the Matuyama/Brunhes reversal at 0.78 Ma. Recent cores from St. Charles County, however, recovered a very mature B horizon approximately 2 m thick preserved atop the Moberly. Thus, the Moberly Formation may be older than previously believed.

Rovey and Kean (1996, 2001) correlated the Moberly Formation with the Alburnett Formation in eastern Iowa. The two formations are nearly identical in lithology (especially the diagnostic clay mineralogy), both occur within the same position of a nearly identical stratigraphic sequence, and both share a reversed DRM.

Source Area: The Moberly Formation contains distinct indicator clasts that isolate a source area to the northeast. Like the older Atlanta Formation, the Moberly Formation contains clasts of pink quartzite. However, clasts of oolitic iron ore (Neda Formation) recovered at the Musgrove Clay Pit are evidence that these must be Baraboo-type quartzite derived from the northeast. The Neda oolitic iron ore is a very distinct lithology that is present locally at the Ordovician–Silurian boundary throughout the upper midwestern United States. The lithology is soft enough that clasts are unlikely to survive prolonged fluvial transport or recycling through multiple glaciations, and hence they are exceptional indicators of regional provenance. The westernmost exposure of the Neda Formation is approximately 550 km northeast of the Musgrove Pit near Dubuque, Iowa, and all other known outcrops occur farther east of Dubuque. Therefore, the regional direction of ice advance must have been from the northeast.

McCredie Formation

Source of Name and History: The McCredie Formation is named after the town of McCredie, Missouri, in Callaway County. Guccione (1983) introduced "McCredie formation" as an informal name for glacigenic sediments in northern Missouri. Tandarich (1992, 2001) restricted this name to those younger deposits characterized by high contents of expandable clay minerals and that are stratigraphically above the Moberly Formation. This restriction was followed by Guccione and Tandarich (1993) and Rovey and Kean (1996, 2001).

Type Section/Area: The type section is in the area at and around the intersection of Highways 63 and M in Macon Country, Missouri (NW sec. 21, T. 59 N., R. 14 W., Atlanta 7.5-min quadrangle), where a complete sequence of McCredie Formation was obtained in a continuous core (Rovey and Kean, 1996). Tandarich (1992, 2001), followed by Rovey and Kean (1996), suggested the Harrison Clay Pit as the type section for the oldest two (informal) members within the McCredie Formation, but the youngest diamicton of the McCredie Formation is not preserved there. All of the (three) currently recognized diamictons are present in core SMS-92b (Rovey and Kean, 1996). Therefore, the location of that core is designated as the type section of the McCredie Formation.

Description of Unit: The McCredie Formation is dominated by one or more massive, fine-grained diamictons. North of I-70, related glaciofluvial sands and gravels become more common beneath individual diamictons.

The McCredie Formation includes at least three distinct diamictons (fig. 3-A-6). Where these diamictons are preserved in direct superposition, they are separated from each other by paleosol/weathering profiles and/or by bedded sands and gravels. All three diamictons within the McCredie Formation are characterized by a relatively high percentage of expandable clay minerals (figs. 3-A-2B, 3-A-5, and 3-A-6; table 3-A-2). The color of these diamictons is nondiagnostic and is related to the local redox conditions and history.

Contacts: The McCredie Formation generally overlies the Moberly Formation, although locally it may occur directly atop Paleozoic bedrock. The McCredie is generally overlain by (undifferentiated) loess resting upon a highly weathered interval associated with the Yarmouth Geosol. In some landscape positions, the upper surface is marked by a pedocomplex spanning the till-loess contact (Guccione, 1983; Rovey, 1997).

Differentiation: The McCredie Formation is distinguished from older diamictons by its higher content of expandable clay minerals (figs. 3-A-2B, 3-A-5, and 3-A-6; table 3-A-2).

Thickness and Regional Extent: The McCredie Formation is generally present within several meters of the land surface wherever younger loess deposits are preserved. The thickness of the McCredie Formation is highly variable and is related to the number of individual diamictons preserved at any given location. South of I-70 the thickness generally ranges between 0–6 m, and typically only the oldest diamicton unit (Fulton member) is preserved. Northward from I-70 the thickness ranges up to ~20 m, as the younger diamictons and associated glaciofluvial sediment become thicker and better preserved.

Interpreted Origin: The bulk of the McCredie Formation diamicton is interpreted as basal meltout till, based on the same crieria as discussed previously for the Atlanta Formation. The bedded sands and gravels are interpretted as proglacial outwash.

Age and Correlation: Diamictons of the McCredie Formation have a normal DRM (Rovey and Kean, 1996, 2001). Thus the formation is apparently younger than the Matuyama/Brunhes reversal (0.78 Ma). Rovey and Kean (1996, 2001) correlated the McCredie Formation with the Wolf Creek Formation in eastern Iowa, which is also normally magnetized. The Wolf Creek Formation also includes three separate members and shares the same diagnostic clay mineralogy as the McCredie Formation.

Source Area: Diamictons of the McCredie Formation contain clasts of various crystalline lithologies, but no distinct indicator clasts have yet been recognized. The high percentage and variety of igneous clasts, however, probably preclude a far westerly source area and are more consistent with ice advance over relatively nearby outcrops of igneous terrain— i.e., the Canadian Shield area nearly directly north of Missouri.

Subdivisions: The McCredie Formation includes at least three individual diamictons. These are described below as informal members.

Fulton member: The (informal) Fulton member includes the oldest diamicton and related sediment within the McCredie Formation. The Fulton member is distinguished from younger glacial diamictons within the McCredie Formation by its stratigraphic position, a finer texture, and by its lower (~4%) content of igneous and metamorphic lithologies in the coarse-sand fraction (figs. 3-A-5 and 3-A-6; table 3-A-2).

The Fulton member normally overlies the Moberly Formation, although locally it is present directly atop bedrock. Its upper surface is generally marked by a well-developed paleosol/weathering profile, including a Bt horizon of variable thickness. Generally the Fulton member is overlain by a younger diamicton (Columbia member) within the McCredie Formation, but locally may be buried by loess. Where overlain by the younger Columbia member, the Fulton member is typically ~5–6 m thick.

The name of this unit follows the designation of Tandarich (1992, 2001). If this unit is formalized, however, the prior usage of the name Fulton at the formation level in another state may necessitate a name change.

Columbia member: The (informal) Columbia member includes the second oldest diamicton and related sediment within the McCredie Formation. The Columbia member is distinguished from the older Fulton member by stratigraphic position, its coarser texture, and by a higher content of crystalline lithologies in the coarse-sand fraction (figs. 3-A-5 and 3-A-6; table 3-A-2). In many locations the Columbia is the uppermost diamicton preserved beneath undifferentiated loess. In these locations the upper surface is marked by a thick Yarmouth Geosol. Where the younger Macon member is preserved, the Columbia member is typically 5 m thick, and its upper contact is still generally marked by a Bt horizon of variable thickness. At the McCredie type section, however, this paleosol/weathering horizon is truncated to the upper C horizon.

The name of this unit follows the designation of Tandarich (1992, 2001). If this unit is formalized, however, the prior usage of the name Columbia at the formation level in another state may necessitate a name change.

Macon member: The (informal) Macon member includes the youngest glacial diamicton and associated sediment within the McCredie Formation as suggested by Rovey and Kean (1996). It is similar lithologically to the underlying Columbia member, but it can be distinguished by its stratigraphic position above a weathering profile in the underlying Columbia member. Also, at any given location, the Macon diamicton is slightly coarser in texture and contains several percent less expandable clay minerals in the <2 micron fraction (fig. 3-A-6; table 3-A-2). Where preserved, it is generally present directly below undifferentiated loess, and its upper surface is marked by a thick Yarmouth Geosol.

The Macon member is generally preserved in stable upland positions along the axis of interfluves; its preservation becomes more common northward of I-70. Its thickness is accordingly quite variable, but ranges up to 11 m in the northern part of the present study area.

Appendix: Laboratory Methods

All of the lithologic analyses summarized in table 3-A-2 were completed at the Southwest Missouri State University (SMS) Sediment Analysis Laboratory, except for several sets of clay-mineral analyses as noted. Those sets were completed at the Iowa Geological Survey (IGS) laboratory, and their respective values are adjusted in table 3-A-2 (see discussion below) for slight but systematic numerical differences between the two labs.

Clay Mineral Analysis

Clay mineral determinations of the <2 micron fraction were made using the technique known as the "semi-quantitative" or "Glass" method (e.g., Hallberg et al., 1978; Moore and Reynolds, 1997). This technique has been widely used to characterize till lithologies throughout the midwestern U.S., and is a standard procedure at both the Illinois and Iowa Geological Surveys.

The semi-quantitative method is generally robust, but differences in preparation technique can cause slight differences in calculated percentages among different laboratories. Because the data herein are most likely to be compared to those of pre-Illinoian tills in southern Iowa (Hallberg, 1980a, 1980b), the consistency of results between the IGS and SMS laboratories was assessed using split samples of widely varying composition. To briefly summarize the clay-mineral results, the SMS laboratory consistently gives ~3% more illite and ~3% less kaolinite + chlorite than IGS analyses. There is no systematic difference between the percentages of expandable clay minerals, however. The systematic differences in the illite and the kaolinite + chlorite fractions were subsequently confirmed by resampling sections with multiple tills at two different locations: first at Conklin Quarry in Iowa (Bunker and Hallberg, 1984) and then the archived material of Core SMS-92c in Missouri (this report).

Texture

Matrix textures for the <2 mm fraction were determined following the general procedures in Walter et al. (1978). Sand percentages were determined directly by wet sieving, using 4 ϕ (0.0625 mm) as the sand-silt boundary. Clay percentages were determined with the pipette method using 8 ϕ (0.0039 mm) as the silt/clay boundary. For comparison, the IGS laboratory uses 0.05 mm for the sand/silt boundary and 0.002 mm for the silt/clay boundary. Therefore, the textural percentages in table 3-A-2 are not directly comparable to textural analyses determined at the IGS laboratory. Based again on interlab comparisons of samples with a wide range in texture, the larger sand/silt boundary used at the SMS laboratory consistently gives 2% less sand compared to the IGS laboratory. The larger silt/clay boundary used at the SMS laboratory generally gives clay percentages approximately

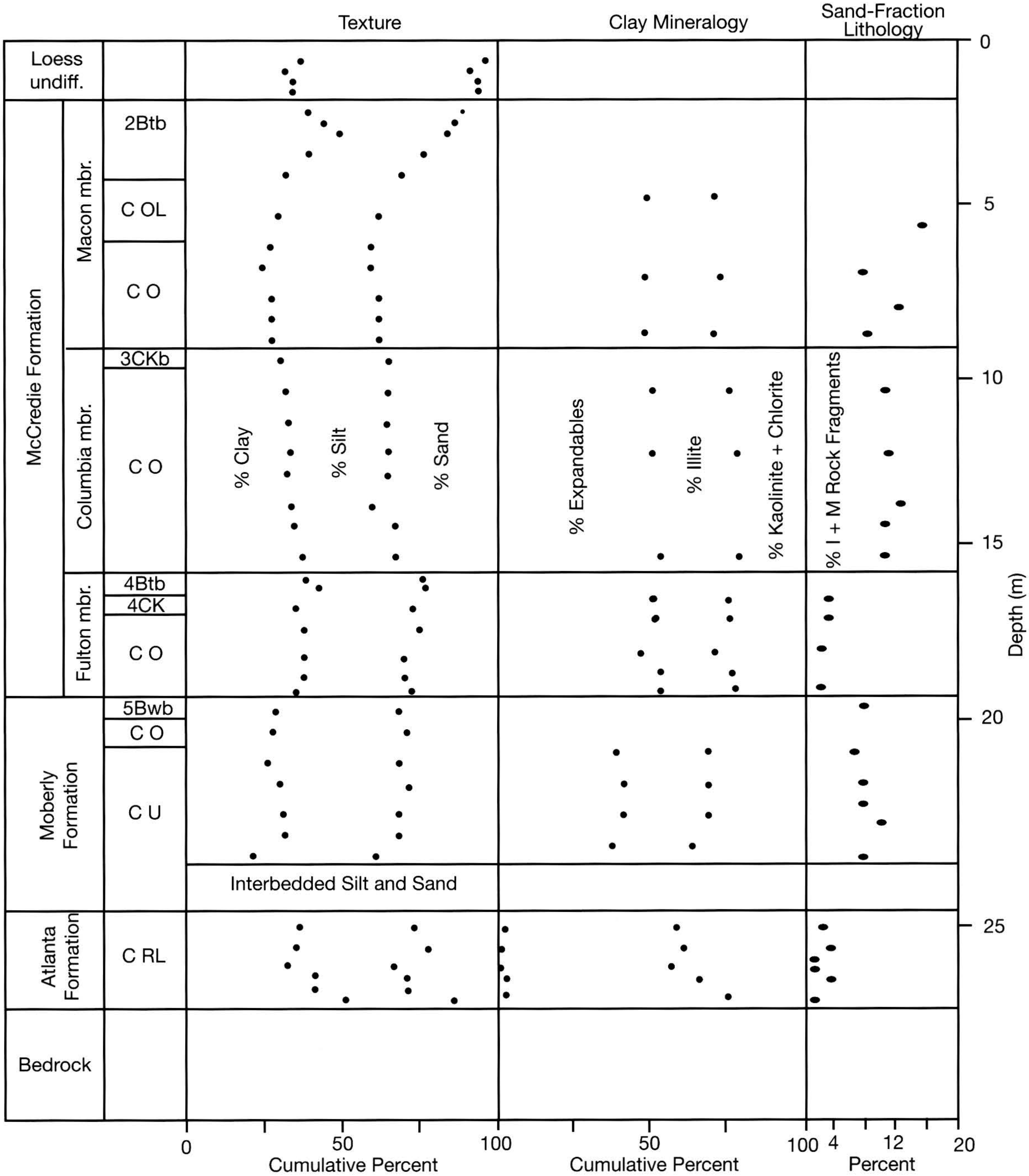

FIGURE 3-A-6.Stratigraphy and lithology of glacial sediments in core SMS-92b (type area of McCredie Formation). See fig. 3-A-3 for weathering-zone terminology.

5% higher than those determined at the IGS laboratory, but ranging from 3% to 6%.

Sand-Fraction Lithology

The coarse-sand fraction (2.0–1.0 mm) was separated from the total sand fraction by wet sieving, and the composition of >100 grains was determined for each sample with a binocular microscope. Based on experience, the best distinction among tills in northern Missouri is achieved by normalizing the various components into three groups: Quartz + Feldspar, Carbonates + Chert, and Igneous + Metamorphic Rock Fragments. Additional lithologies, especially detrital sedimentary grains, are present but these are ignored in determining the normalized percentages in table 3-A-2.

An Igneous Rock Fragment is defined here as a compound grain containing three or more types of common igneous minerals (e.g., quartz, feldspar, micas, mafics) or a compound grain with two different types of igneous minerals and at least three individual subgrains of quartz, feldspar, or mafic minerals. Based on a comparison of SMS-laboratory analyses with published results for Conklin Quarry in Iowa by the IGS laboratory (Bunker and Hallberg, 1984), it appears that the Iowa laboratory uses different criteria to define an Igneous Rock Fragment, and sand-fraction percentages determined at the two different laboratories could not be uniquely related to each other.

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Paleomagnetism of Sediments Associated with the Atlanta Formation, North-central Missouri, USA

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Abstract

Age estimates of the earliest continental glaciation in North America vary by nearly half a million years and span the boundary between the Gauss (Normal) and Matuyama (Reverse) Chrons at 2.58 Ma. Part of this uncertainty is caused by a mixed magnetic polarity within the oldest glacial till throughout the midcontinental United States. Thus, previous paleomagnetic studies have been unable to resolve the till's age relative to the Gauss-Matuyama boundary.

The major pre-Illinoian glaciations all reached the same approximate southern limit in Missouri where the ice was stopped by the regional reversal in topographic gradient along the margin of the Ozark Dome. In northern Missouri the oldest glacial diamicton (Atlanta Formation) is underlain by preglacial and proglacial sediments, which accumulated within paleokarst depressions. These preglacial and proglacial sediments have a well-defined, reversed, detrital remanent magnetization. A localized silt facies within the upper Atlanta Formation also retains a reversed magnetic remanence. Thus, the initial continental glaciation to reach the maximum pre-Illinoian expanse occurred after the Gauss-Matuyama boundary at 2.58 Ma.

Introduction

The timing of the Early and Middle Pleistocene (pre-Illinoian) glaciations in the midcontinental U.S. is constrained by fission-track dates of tephra interbedded between tills and by the magnetic polarity of the tills themselves (fig. 3-B-1). The Matuyama-Brunhes paleomagnetic datum separates three younger, normal-polarity tills from one or more older units with reverse polarity.

One sequence of pre-Illinoian glaciation began during the late portion of the Matuyama Reversed Chron and continued through the early Brunhes Normal Chron. Deposits from the initial advance within this sequence include the reverse magnetized Alburnett and Moberly Formations in Iowa and Missouri, respectively (fig. 3-B-1). In both states the timing of this particular glaciation is estimated to be slightly older than the Matuyama-Brunhes reversal at 0.78 Ma (Richmond and Fullerton, 1986a; Rovey and Kean, 2001). Thus, these deposits would date from the early portion of enhanced glacial activity known informally as the "mid-Pleistocene climate transition" or "MPT" (Raymo et al., 1997). Later advances within this same general sequence deposited the normally magnetized Wolf Creek and McCredie Formations.

A much earlier (Pliocene) glaciation is represented by the "C till" in eastern Nebraska and western Iowa and most likely by the Atlanta Formation in Missouri (fig. 3-B-1). The timing of this older glaciation is less certain, however. A single fission-track date for tephra above the "C till" in southern Iowa (figs. 3-B-1 and 3-B-2) indicates that continental glaciation occurred prior to ~2.2 Ma (Boellstorff, 1978a, 1978b). The date implies that this early glaciation is related to the "North American glacial expansion/intensification" during the late Pliocene, ca. 2.7–2.4 Ma (e.g., Raymo, 1994; Maslin et al., 1998), and which spanned the Gauss/Matuyama boundary (2.58 Ma). Hallberg (1986) and Richmond and Fullerton (1986a) identified this tephra or ash above the "C till" as the "Pearlette B" with an accepted age of 2.01 Ma. The "Pearlette B" and the facies-equivalent Huckleberry Ridge Tuff in the Yellowstone area are associated with a transitional magnetic field (normal to reverse), presumably reflecting the upper boundary of the (upper) Reunion Normal Subchron. Recently Lanphere et al. (2002) suggested that this tuff was actually deposited during a magnetic event slightly younger than the Reunion. Nevertheless, an association with some magnetic event is supported by the fact that the underlying "C till" has mixed (samples with both reversed and normal) remanent polarity (Easterbrook and Boellstorff, 1984), in contrast with the younger pre-Illinoian tills which retain a consistent magnetic remanence of one polarity or the other. Boellstorff (1978a), however, considered this ash to be distinct from the "Pearlette B" and correlated it instead with the "Arcadia Canal" ash dated at approximately 2.3 Ma. Thus, depending on which (if either) interpretation is correct, the underlying "C till" is older than either 2.01 or 2.3 Ma.

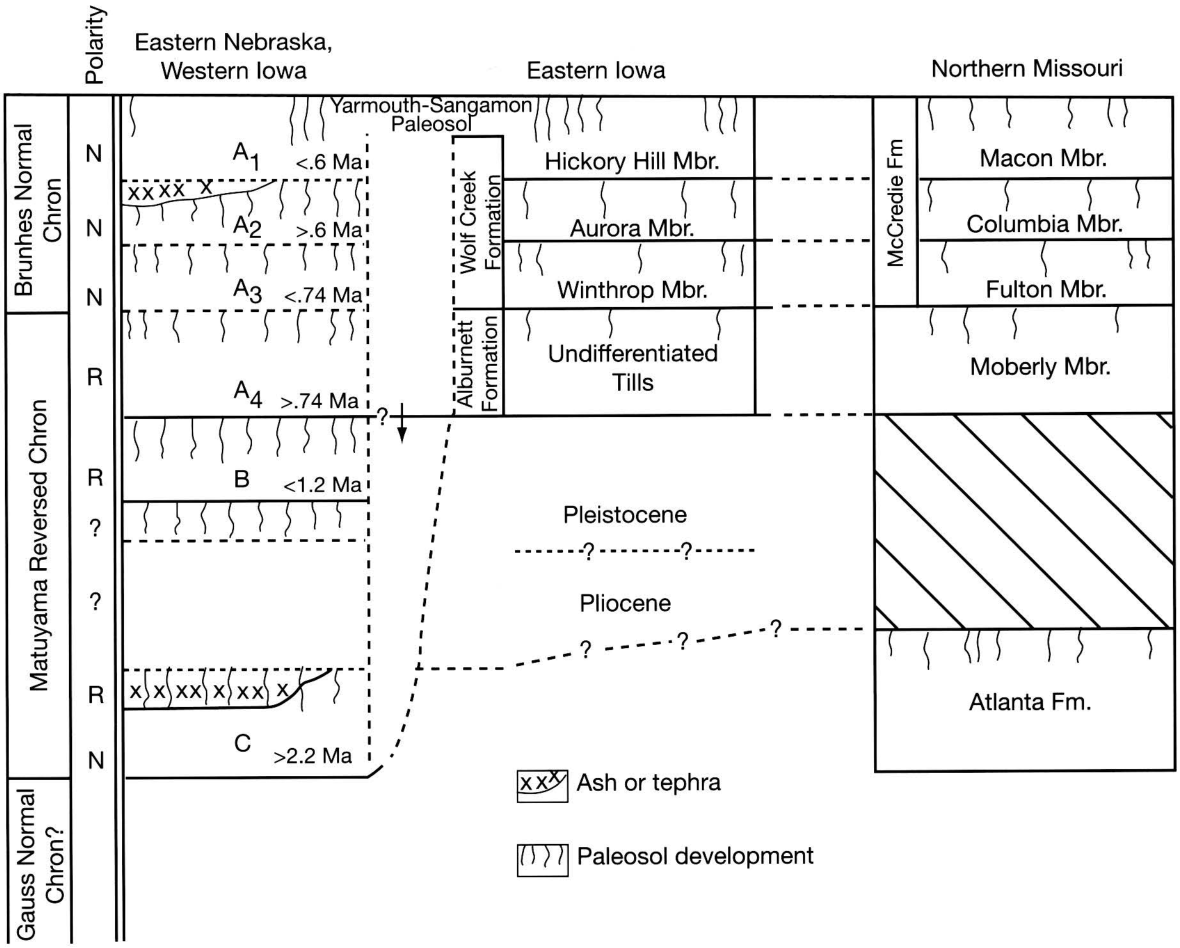

FIGURE 3-B-1. Pre-Illinoian glacial and magnetic sequence in the midcontinental U.S. Modified from Kemmis et al. (1992), after Easterbrook and Boellstorff (1984). Age constraints of the tills in eastern Nebraska and western Iowa are from fission-track dates of interbedded tephra (Boellstorff, 1978a, 1978b; Richmond and Fullerton, 1986a). Ages of paleomagnetic datums as discussed in the text are from Cande and Kent (1995).

Richmond and Fullerton (1986a), who accepted the younger ash date, assigned a tentative age of ~2.14 Ma to the "C till" and hence the earliest continental glaciation. This date would place the till deposition coincident with the transitional magnetic field of the (lower) Reunion Normal Subchron, thereby accounting for the till's mixed polarity. Boellstorff (1978a, 1978b), however, assigned a minimum age of 2.2 Ma and an actual age of ~2.4 Ma, based on paleotemperature trends inferred from oceanic sediment cores. Easterbrook and Boellstorff (1984) speculated that the "C till's" normal remanent polarity may be a primary depositional remanence acquired during the latest portion of the Gauss Normal Chron (prior to 2.58 Ma), whereas the reversed remanence would be a later overprint from the Matuyama Reversed Chron. Recently, however, Roy et al. (2004) reported a reversed magnetic remanence for the "C-till" equivalent in eastern Nebraska and western Iowa. Thus, the age estimates for the earliest North American continental glaciation span nearly half a million years and portions of two polarity chrons.

Objectives

We propose that the timing of the earliest *major* North American continental glaciation can be constrained by examining the glacial and preglacial sediments preserved in Missouri, USA, near the southernmost advance of the pre-Illinoian ice sheet (fig. 3-B-2). Rovey and Kean (1996, 2001) presented initial interpretations of this glacial sequence, along with correlations and magnetostratigrahy, summarized here as fig. 3-B-1.

Rovey and Kean (1996) and Rovey and Tandarich (this volume, p. 3-A-1) documented the presence of five individual pre-Illinoian tills along the southern margin of the former Laurentide ice sheet (figs. 3-B-1 and 3-B-2). The southern margin coincides with a major regional reversal in the topographic gradient along the margin of the Ozark Dome. Hence, the largest of the pre-Illinoian glaciations all reached the same approximate southern terminus. Restated, once a critical threshold was exceeded, the southern limit of the pre-Illinoian advances was no longer constrained by the volume of ice, but by the point of reversal in topographic gradient. Therefore, we believe that presence along the margin of the Ozark Dome

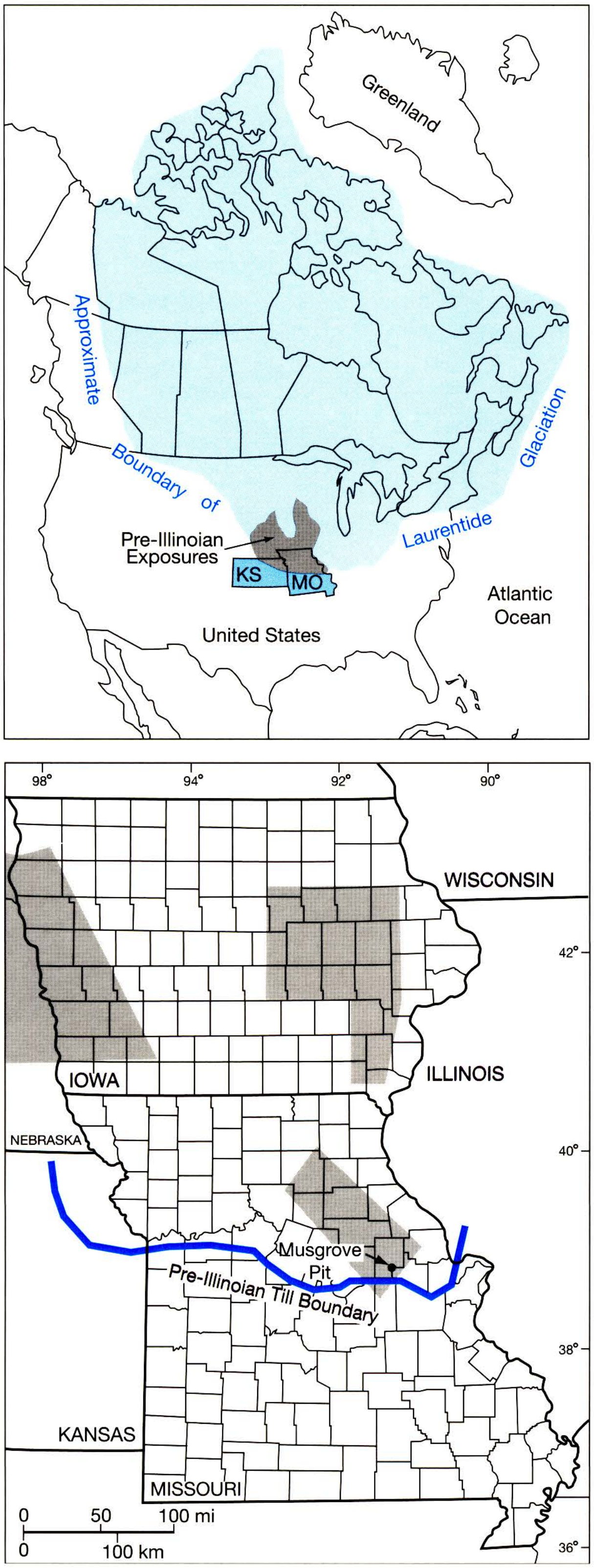

FIGURE 3-B-2. Location of Musgrove and Johnson clay pits. Only the Musgrove Pit is labeled, because their locations cannot be separately resolved at this scale. Shaded areas show locations from which the stratigraphic sequences depicted in fig. 3-B-1 were developed.

distinguishes "major" pre-Illinioan glaciations from less-extensive and hence minor advances.

The oldest of the five tills in Missouri is the Atlanta Formation, which occupies the same position as the "C till" within a nearly identical sequence farther north in Iowa and Nebraska. The two units also share similar lithologies and weathering histories; the most mature weathering profiles within the pre-Illinoian sequence are developed atop these two tills. Moreover, the Atlanta Formation till in Missouri also has a mixed magnetitic polarity, further supporting the conclusion that these units ("C till" and Atlanta Formation) are at least broadly correlative. We recognize that deposits from still earlier glaciations may be present farther north of our study area and that conclusive evidence for correlating the Atlanta Formation and "C till" is lacking. Nevertheless, should the "C till" (or any other older till) date from an entirely different glaciation, that advance would have been considerably less extensive. Therefore, we consider the Atlanta Formation to be a deposit from the first *major* North American (Cenozoic) continental glaciation, regardless of its exact correlation to other units.

Recent activity in Missouri's clay mining district has exposed preglacial and proglacial sediments beneath Atlanta Formation till. Therefore, the primary objective of this project was to measure the magnetic polarity of these subjacent sediments to determine whether the first glaciation known to have reached Missouri occurred before or after the Gauss-Matuyama boundary at 2.58 Ma.

Location and Procedures

The new data presented here were obtained by analyzing samples collected from highwalls of the Musgrove and Johnson clay pits, located in Missouri's Fire Clay District. These pits are approximately 8 km north of the pre-Illinoian glacial boundary (fig. 3-B-2) and are within the type area of the Atlanta Formation (Rovey and Tandarich, this volume, p. 3-A-2). Previous samples of the Atlanta Formation's till facies gave mixed results (Rovey and Kean, 1996, unpublished data this study). Most samples had a reversed magnetic remanence, but some samples were normally magnetized, while others had multiple components of both normal and reverse polarity. Thus, the original detrital remanent magnetization was unclear, based upon analysis of the till facies. Recently, mining in these pits exposed more-localized silty facies within or subjacent to the Atlanta Formation that are more likely to faithfully record the earth's magnetic polarity during deposition.

At the Musgrove Pit, paleomagnetic samples were taken along vertical transects at three different locations. In the following tables and text, we refer to Transect 1 (midlocation of 1996 northern highwall), Transect 2 (west end of the 1999–2000 northern highwall), and Transect 3 (east end of the 1999–2000 highwall). In 2002 a suite of samples was also collected from new exposures in the Johnson Pit.

Samples were taken by pressing a small, oriented plastic box into a leveled surface. Each sample was subjected to stepwise a.f. demagnetization. Demagnetization was generally halted after reaching 100 mT, or after the sample's natural remanent magnetization (NRM) intensity declined to <10% of its original value or to random noise level.

After each demagnetization step, the magnetic remanence was measured in multiple orientations using a Molspin GS-1 magnetometer. The replicate vector components at each step were used to calculate a reproducibility factor, which is a guide to both the stability of remanence and the optimal demagnetization level. After demagnetization, the low-field magnetic susceptibility was measured for each sample. For selected samples the remanence intensity was measured under a series of applied D.C. fields to construct isothermal remanent magnetization (IRM) curves to help assess the mineral carrier of the magnetic remanence.

Site Stratigraphy

Landscape Position

The sedimentary sequence preserved and exposed within the Musgrove and Johnson pits is related to the local bedrock topography. In this district, kaolinitic clay is mined from the Pennsylvanian-age Cheltenham Formation, and the purest facies is restricted to paleokarst features developed in the underlying Mississippian-age limestones. Hence, the mining activity is often localized within a paleokarst network. These paleokarst features remained as isolated bedrock lows prior to glaciation and thus functioned as small, localized depositional basins in which preglacial and proglacial sediments accumulated.

Whippoorwill Formation

The oldest diamicton within the Musgrove and Johnson pits is known informally as the Whippoorwill formation (Guccione, 1983). This deposit is present within the preglacial paleotopographic lows but is absent from higher landscape positions. The Whippoorwill is a leached, mildly gleyed, massive diamicton with rare oxidized mottles and scattered pebbles. In places it resembles a till, but only local, and no erratic, lithologies are present. The matrix lithology varies substantially from one pit to another, suggesting that these sediments were derived from local source areas immediately surrounding each depocenter.

The Whippoorwill formation's geometry is consistent with a mass-flow or sheet-flow origin. Its presence is limited to bedrock depressions, and its thickness is greatest around the perimeter of paleosinks. This ramp-like geometry suggests gravitational transport of saturated soil/residuum from surrounding upland areas into the paleosinks. Thus, the leached condition reflects its depositional state, not subsequent weathering. Likewise, the mildly reduced condition is syndepositional; gleying does not extend below the lower Whippoorwill boundary, nor upward into the overlying till (fig. 3-B-4a). The deposit lacks flow structure, discrete internal weathering surfaces, and stratified interbeds, indicating that the unit may have originated as a solifluction/gelifluction deposit. If this interpreted origin is correct, deposition may be related to climatic cooling associated with the onset of glaciation.

Guccione (1983) described a mature paleosol developed within the upper Whippoorwill formation, but only from locations where it is overlain by much younger materials. In contrast, the Whippoorwill at the Musgrove Pit is capped by a subtle weakly developed paleosol marked only by a blocky structure, slightly more intense gleying, and by carbonized root traces. This horizon has very little clay accumulation or alteration, despite developing under gleying conditions, which are conducive to the formation of expandable clay minerals. In the Johnson Pit, no weathering profile is present atop the Whippoorwill, and its upper contact is gradational to a laminated silt facies with scattered dropstones, the basal part of the overlying Atlanta Formation. Thus, at the Johnson Pit, deposition of the Whippoorwill formation apparently ceased immediately prior to burial by the earliest glacial sediments.

An upward increase in the expandable clay mineral content over the basal meter of the Whippoorwill formation at the Musgrove Pit (fig. 3-B-3) might be taken as evidence for minor clay neoformation, qualifying most of this unit as a BC horizon and indicating a longer duration of weathering. However, a pedogenic increase in expandables would be accompanied by a decrease in illite, which is easily degraded (see for example the clay-mineral trends in the overlying 3B horizon). The nearly constant illite percentage throughout most of this unit indicates a lack of pedogenic alteration except within and just below the $5B_{wg}$ horizon (fig. 3-B-3). Therefore, at the Musgrove Pit, a break occurred between deposition of the Whippoorwill formation and burial by the Atlanta Formation, but this time interval was only sufficient for a weak soil profile to form.

Atlanta Formation

The Atlanta Formation is the oldest unquestioned glacigenic deposit recognized in northern Missouri (Rovey and Kean, 1996; Rovey and Tandarich, this volume, p. 3-A-2). The Atlanta is completely eroded at most locations; its preservation is rare and limited to stable landscape positions such as the paleosinkholes within the present study area. Where locally preserved beneath younger deposits, a very thick (>2 m) B-horizon

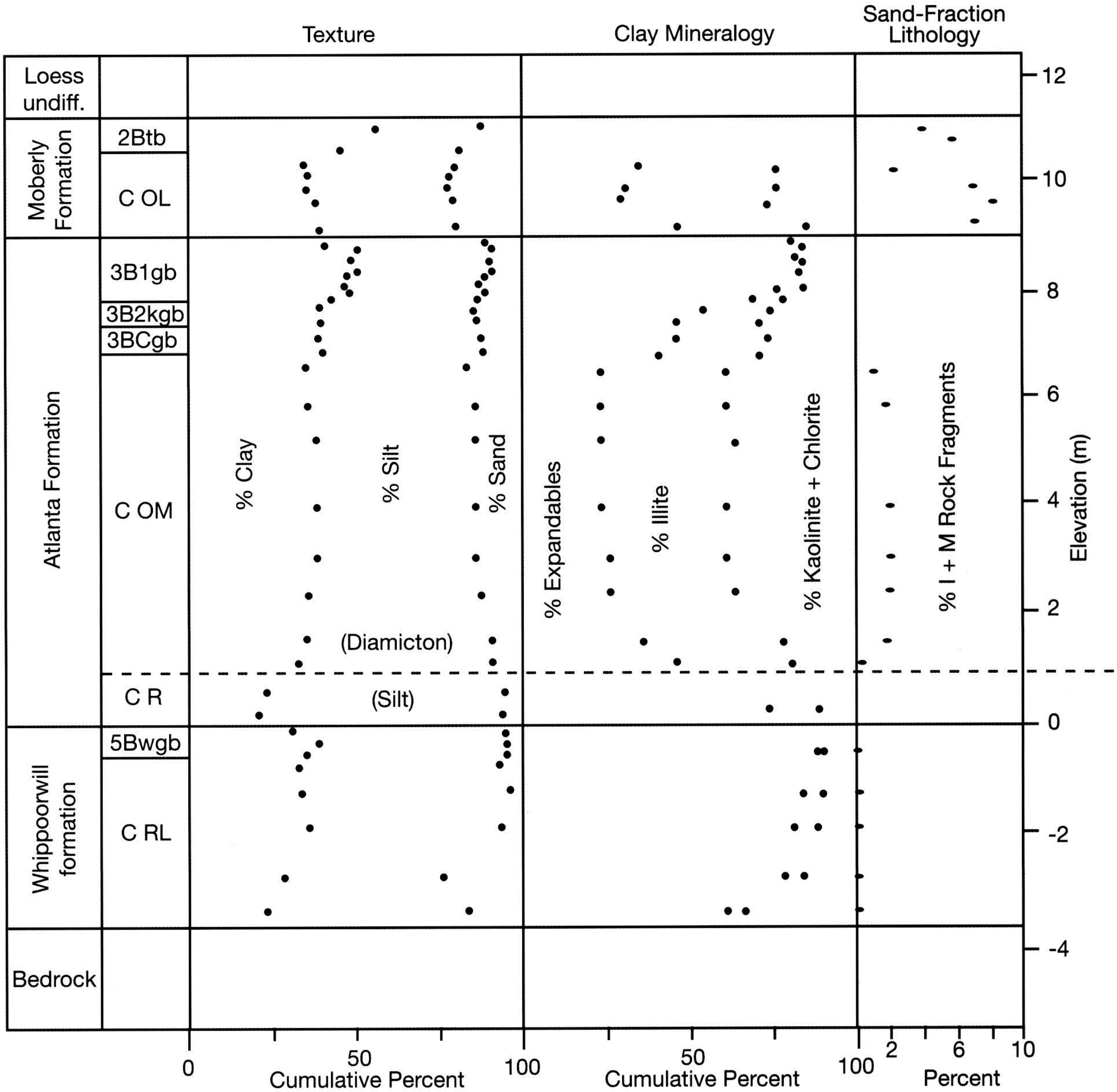

FIGURE 3-B-3. Matrix properties of Quaternary sediments at Musgrove Clay Pit. Results shown are a composite of several individual sampling transects. Subscripts denote the following: O, oxidized; M, mottled; U, unleached; R, reduced; L, leached.

paleosol is developed atop the Atlanta Formation till (see the discussion and figures that follow). This soil horizon, along with the Atlanta Formation's general absence, attest to a long duration of weathering and erosion between its deposition and that of the younger tills, which date from the MPT, ca. 0.8–0.5 Ma (fig. 3-B-1).

The lowest increment of the Atlanta Formation at the Musgrove and Johnson pits is proglacial silt (fig. 3-B-4b). This silt superficially resembles the subjacent Whippoorwill formation, but is slightly calcareous, less-intensely gleyed, and locally rests upon a discrete paleosol. Moreover, the silt is faintly laminated, implying a stratified aqueous origin. The reduced condition does not extend upward into the overlying till, and hence, the gleyed coloration again appears to be syndepositional, possibly caused or enhanced by incorporation of reworked sediment from the underlying Whippoorwill formation.

The silt's composition indicates an approaching glacial source. At the Musgrove Pit, the illite percentage, which is quite low in the underlying bedrock and residuum, increases sharply within the silt. Moreover, the sand fraction includes rare, crystalline rock fragments which are absent from the subjacent materials, and the same facies at the Johnson Pit contains scattered pebbles (dropstones) that deform subjacent laminae.

The silt's upper contact with the overlying till is sharp and erosional. Most of the erosion was apparently subglacial; discrete blocks of the silt are commonly incorporated into the basal few centimeters of the overlying till.

A

FIGURE 3-B-4. Basal contact of Atlanta Formation till. (A) Contact between Atlanta Formation till and underlying Whippoorwill formation, Johnson Pit. (B) Contact (at box) between Atlanta Formation till and underlying proglacial silt, Musgrove Pit. The silt is present locally within swales on the surface of the underlying Whippoorwill formation. Contact between the silt and underlying Whippoorwill formation is at shovel. The upper Atlanta contact with the overlying Moberly Formation is at the abrupt transition to an unoxidized diamicton with fewer clasts.

B

The bulk of the overlying Atlanta Formation is a cobbly diamicton (fig. 3-B-4). Most of this diamicton is interpreted as basal meltout till, based on its uniform matrix composition (fig. 3-B-3), a lack of fissility, and the absence of flow structures. Shear structures are also generally lacking, except locally near the base, and extending downward into the upper few centimeters of the subjacent silts.

Several meters of erosional relief are present along the Atlanta Formation's upper contact. This upper surface preserves a paleosol developed during an exceptionally long period of weathering, relative to other paleosols within the pre-Illinoian sequence. Although clay enrichment within this solum appears modest based on fig. 3-B-3, that appearance is deceptive, because the clay enrichment (approximately 12%) is uniform throughout the entire upper meter. Extreme weathering can also be adduced from trends in clay mineralogy. The expandable clay minerals increase upward from approximately 25% of the clay fraction below the solum to 80% within the 3B1 horizon. Concomitantly, illite decreases upward from approximately 40% to zero and is completely degraded within the upper meter. Also, the solum grades upward from moderately stony (16–35%) at the base to very slightly stony (1–5%) at the top. This gradation must be due to weathering and not an accretionary accumulation of fine-grained material, because it is best expressed in the topographically highest portions of the solum and not within lower swales.

This paleosol retains abundant stress cutans and striated slickensides in the upper 3B1 horizon, which underwent extensive shearing and homogenization (note the uniform texture within the 3B1, fig. 3-B-3). Considering also that the upper portion is approximately 50% clay, of which 80% is expandable, this soil must have been a Vertisol or vertic intergrade. Therefore, the soil's maturity cannot be assessed using standard weathering indices (e.g., Harden, 1982; Harden and Taylor, 1983). Nevertheless, the pedogenic changes within this solum are more extreme than any we have yet encountered within a Quaternary-age paleosol and exceed those of the composite Yarmouth-Sangamon Geosol, even where the latter developed under similar reducing conditions (e.g., Rovey, 1997).

The composite Yarmouth-Sangamon Geosol formed over an approximate 0.5 Ma duration (Richmond and Fullerton, 1986b). Therefore, by comparison, we estimate that the Atlanta Formation till weathered for at least 0.5 Ma prior to burial by the overlying Moberly Formation. Given that the Moberly Formation was deposited during the latest portion of the Matuyama Chron (Rovey and Kean, 2001), the *minimum* age of the Atlanta Formation would be about 1.3 Ma. The actual age, however, is probably much greater.

Paleomagnetic Results

To date we have been unable to isolate a consistent magnetic remanence of just a single polarity within the Atlanta Formation till. Therefore, we focus primarily on results from subjacent materials to determine the state of the magnetic field shortly before glaciation.

Whippoorwill Formation

Samples from the upper Whippoorwill formation were taken along Transect 3 and span the interval below and within the upper paleosol. Each sample had a stable and highly significant reversed characteristic remanence (ChRM) carried by magnetite. These samples lacked normal components of magnetization, except those obviously acquired within the current field and expressed as a weak, viscous remanent magnetization (VRM). The grouped samples have a high Fisher Precision Parameter (κ) up to 38, and a 95% confidence limit (α_{95}) of <10° about the mean orientation (table 3-B-1).

These samples had a consistent reversed declination through the entire demagnetization sequence. Inclinations flipped from shallow normal to reverse by 5 mT demagnetization, as the VRM from the current magnetic field was removed (fig. 3-B-5). Most of the VRM was removed by 5 mT, although slight viscous effects persisted in some samples up to 20–30 mT demagnetization, as indicated by the removal of shallow difference vectors. Above this range, difference vectors stabilized and coincided with the measured remanence directions, implying that demagnetization had isolated the ChRM. Principal Components Analysis (PCA) also isolates a highly significant reversed remanence beginning at around 30 mT, consistent in each case with the optimum demagnetization level inferred from the reproducibility factor (table 3-B-1).

This reversed ChRM is carried by magnetite. Sample coercivities (median destructive a.f. fields) range from approximately 20 to 50 mT, consistent with single-domain magnetite (Lowrie and Fuller, 1971; Heider et al., 1987). Very little remanence remains after 60 mT demagnetization (fig. 3-B-5). Therefore, despite the oxidized mottles, there is no significant amount of hematite or iron hydroxides, which may form authigenically with a harder secondary chemical remanent magnetization (CRM). Thus, the reversed polarity must be a primary detrital remanent magnetization (DRM).

Atlanta Formation Silt

The Atlanta Formation silt, which is subjacent to the till facies, was sampled along all three transects. Three samples from Transect 1, however, were collected at a location where the silt was sheared. These three samples

TABLE 3-B-1. Summary of measurements, Whippoorwill formation. Grouped samples were taken at the same stratigraphic level; numerals in sample labels refer to elevation (m) relative to the Whippoorwill formation's upper contact. "Single point" values are those at the optimum a.f. demagnetization level, based on the reproducibility of measurements in multiple orientations. Demagnetization levels refer to peak a.f. fields in mT. PCA values are principal components isolated from the demagnetization sequence (maximum angular deviation < 10°). κ is the Fisher precision parameter, alpha 95 is the 95% confidence limit (degrees) about the mean direction.

Sample	Single Point Inc.	Single Point Dec.	Demag. Level	PCA Inc.	PCA Dec.	Range
ML-.30a	-41	152	30	-39	153	20-origin
ML-.30b	-52	176	30	-44	178	40-origin
ML-.46a	-42	175	40	-41	180	30-origin
ML-.46b	-53	176	40	-50	171	40-origin
ML-.61a	-37	194	40	-38	193	30-origin
ML-.61b	-48	201	30	-48	201	30-origin
Vector Mean:	-47	179		-44	179	
κ:	35.			38.		
Alpha 95:	9.7			9.3		

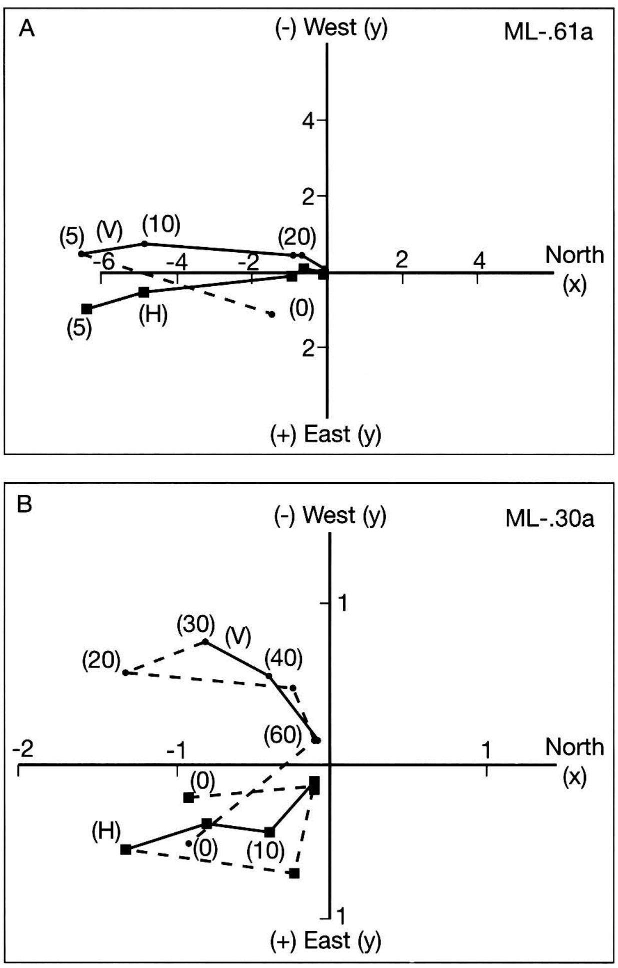

FIGURE 3-B-5. Vector intensity diagrams, Whippoorwill formation. Units are those of intensity (A/m x 10^{-3}).

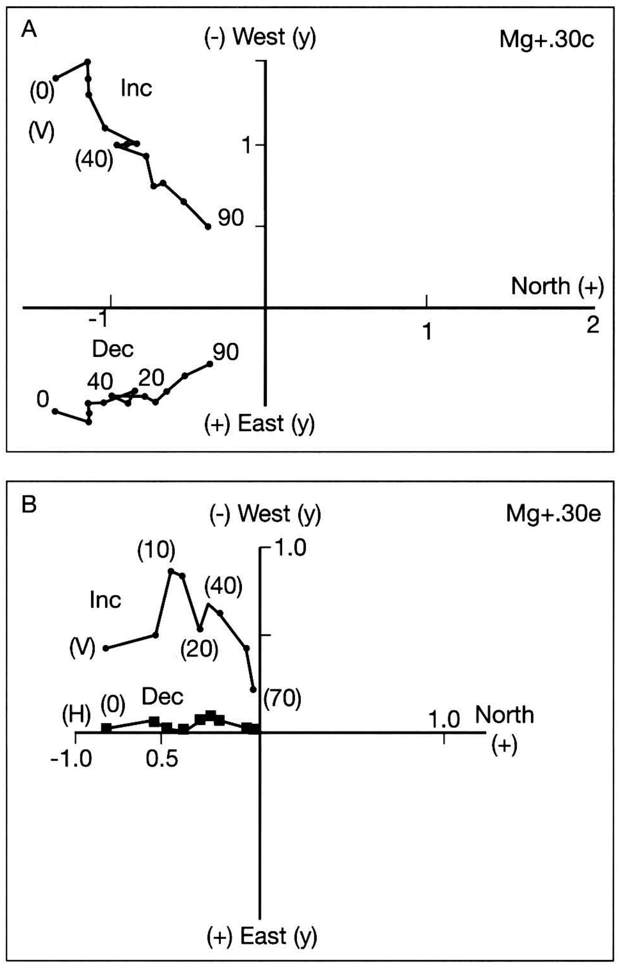

FIGURE 3-B-6. Vector intensity diagrams, proglacial silt, Atlanta Formation. Sample Mg+.30c is unusual in that a.f. demagnetization continues above 60 mT.

TABLE 3-B-2. Summary of measurements, proglacial silt of Atlanta Formation, Musgrove Pit. Grouped samples were taken at the same stratigraphic level; numerals in sample labels refer to elevation (m) relative to the silt's lower contact. See table 3-B-1 for further explanations. Demagnetization of the "ML" series was stopped at 60 mT. The lack of a highly significant vector component (PCA analysis) in some samples is related to the presence of a "hard" remanence above 60 mT. Therefore, some samples stopped demagnetizing and lack a well-defined vector trend toward the origin.

Sample	Single-point Inc.	Single-point Dec.	Demag. Level	PCA Inc.	PCA Dec.	Range
ML+.86a	-40	138	30	-21	182	30-60
ML+.86b	-36	146	20	-48	101	30-60
MG+.61a	-70	99	20	-82	58	60-origin
MG+.61b	-61	130	20	-74	95	40-60
MG+.6ac	-55	126	20	-52	126	20-50
MG+.61d	-61	164	40	None		
MG+.61e	-68	161	40	None		
Group Mean:	-65	137		-71	109	
κ:			41.			20.
alpha 95:			9.8			18.
MG+.30a	-61	149	20	None		
MG+.30b	-72	272	40	None		
MG+.30c	-47	148	20	-53	200	25-60
MG+.30d	-70	162	40	-76	194	15-30
MG+.30e	-50	188	20	-58	197	30-50
Group Mean:	-65	171		-63	187	
κ:			13.			28.
alpha 95:			17.			15.
ML+.06a	-67	212	60	None		
ML+.06b	-43	208	60	None		
ML+.06c	-42	181	60	None		
Group Mean:	-51	198				
κ:			21.			
alpha 95:			18.			
Totals:	-61	162		-66	158	
κ:			13.			7.6
alpha 95:			9.9			18.

apparently had a reversed remanence, but the results are equivocal.

The remaining 15 samples from Transects 2 and 3 span the silt's maximum preserved thickness and have magnetic characteristics very similar to the subjacent Whippoorwill formation. Most importantly, the silt shares a reversed DRM (fig. 3-B-6, table 3-B-2). The silt, however, retains a slightly harder magnetic remanence, implying that minor amounts of hematite are present.

Most of the VRM affecting the silt was removed by 5–10 mT of a.f. demagnetization, although slight effects again persisted to 20–30 mT. In eight samples, PCA isolated a highly significant reversed ChRM between approximately 20 and 60 mT (fig. 3-B-6). Above 60–70 mT of a.f. demagnetization, these samples ceased to demagnetize and the remanence directions stabilized at intensities ranging from 5 to 30% of the peak NRM value. A single thermally demagnitized sample (Transect 1) and IRM curves confirm that this hard remanence is due to small quantities of hematite.

Despite the presence of hematite within the silt, there is no indication that this mineral is authigenic. The hard magnetization isolated above 60 mT is also reversed with nearly identical remanence directions as the softer remanence carried by magnetite and isolated at lower demagnetization levels. Therefore, although the silt has two mineral carriers of magnetization, it has a single reversed DRM. Samples from the same relative elevations have good group statistics (table 3-B-2) with Fisher κ values ranging up to 41.

A very localized silt facies was exposed within the upper Atlanta Formation at the Johnson Pit during the summer of 2002. This silt was intercalated laterally with the diamicton/till facies and in turn was overlain by a separate diamicton with possible flow structure. The geometry and contacts of these different facies closely resembled supraglacial flow deposits that have been extensively documented from the Des Moines Lobe (e.g., Kemmis et al., 1981) and many other modern settings. The paleosol capping the upper till surface continued without dimunition above the silt within the upper part of the overlying diamicton. Therefore, the age of the silt is very close to that of the till facies, even if it was not strictly syndepositional. Like the subjacent silts, this upper silt facies retains a highly significant reversed DRM (table 3-B-3).

Interpretations

The magnetic sequence within the Whippoorwill formation and basal Atlanta Formation silts (fig. 3-B-7) supports the observational conclusion that these units were deposited with just minor interruption. A prominent easterly shift in declination continues across the contact of these two units. These units are not oxidized, and accordingly the reversed magnetic remanence lacks any evidence of a secondary CRM overprint, as might be associated with secondary iron oxides/hydroxides formed

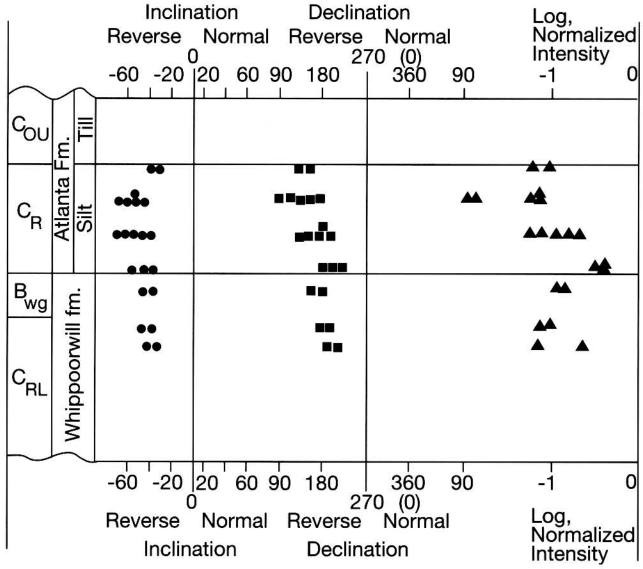

FIGURE 3-B-7. Summary of magnetic remanence directions, Musgrove Pit. Plotted values are the "single point" values from tables 3-B-1 and 3-B-2. Intensities (A/m x 10^{-3}) are measured after 20 mT of peak a.f. demagnetization and normalized by bulk susceptibility (10^{-5} SI).

TABLE 3-B-3. Summary of measurements, upper silt facies in Atlanta Formation, Johnson Pit. Grouped samples were taken at the same stratigraphic level; numerals in sample labels refer to elevation (m) above the formation's lower contact. See table 3-B-1 for further explanations. The two samples labeled as "unstable" had reversed polarity, but had a low magnetic intensity, which decreased to "noise level" before the overprint from the viscous remanent magnetization (VRM) was completely removed.

Sample	Single Point Inc.	Single Point Dec.	Demag. Level	PCA Inc.	PCA Dec.	Range
Js2.3a	-53	201	30	None		
Js2.3b	Unstable			None		
Js2.2a	-27	202	30	-27	177	10-origin
Js2.2b	-27	182	30	-27	158	10-origin
Js1.9a	-50	201	30	-31	212	20-origin
Js1.9b	-34	228	30	-35	223	10-origin
Js1.8a	-25	188	20	-23	188	20-origin
Js1.8b	-84	127	30	None		
Js1.7a	Unstable			None		
Js1.7b	-44	125	30	-44	115	20-origin
Vector Mean:	-46	170		-36	179	
κ:		7.7		3.4		
Alpha 95:		18.		31.		

during oxidation. Therefore, this remanence is a primary DRM. The upper silt facies locally present at the Johnson Pit also retains a reversed DRM. Moreover, the Atlanta Formation is overlain by the reverse magnetized Moberly Formation. Although the upper Moberly Formation is locally overprinted with a secondary (normal) CRM, this later magnetization is easily distinguished from a DRM by its magnetic mineralogy, harder remanence, and erratic demagnetization (Rovey and Kean, 2001).

In summary, sediments below, within, and above the Atlanta Formation consistently have a reversed DRM. Therefore, the Atlanta Formation was deposited during the Matuyama Reversed Chron (2.58–0.78 Ma).

Conclusions

Based on the previous evidence and discussion, we can make several conclusions, albeit with varying degrees of confidence.

- A sequence of preglacial, proglacial, and glacial sediments associated with the Atlanta Formation is reversely magnetized. This formation is also overlain by younger till (Moberly Formation) with a reversed DRM. This constrains the age of the first known glaciation to have reached Missouri (and the southernmost margin of the former Laurentide ice sheet) to within the Matuyama Reversed Chron (2.58–0.78 Ma).

- Based on geomorphic evidence and the advanced weathering profile preserved atop the Atlanta Formation till, the Atlanta is much older than the overlying Moberly Formation. Therefore the Atlanta Formation must date from a much earlier portion of the Matuyama Reversed Chron than the Moberly.

- Based on the paleotemperature record of deep-sea-sediment cores, the most likely time of deposition for the Atlanta Formation is between ~2.5 and ~2.1 Ma. Data from oceanic sediments throughout the northern hemisphere indicate that major continental glaciations culminated near the very beginning (the "North American glacial expansion") and end (the "MPT") of the Matuyama Reversed Chron with little glacial activity between those times (e.g., Beard, 1969; van Donk, 1976; Shackleton et al., 1984; Joyce et al., 1993; Raymo, 1994; Raymo et al., 1997; Prueher and Ray, 1998; Maslin et al., 1998; St. John and Krissek, 2002; Knies et al., 2002). If the Moberly Formation dates from the late portion of the Matuyama Chron (the "MPT"), and the Atlanta Formation is much older, the Atlanta would accordingly date from the early portion of the Matuyama Chron. This conclusion is consistent with the likely correlation of the Atlanta Formation with the "C-till" in southeast Iowa, which has been directly dated at greater than ~2.2 Ma.

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AMQUA Post-Meeting Field Trip No. 4: Late Quaternary Alluvial Stratigraphy and Geoarcheology in the Central Great Plains

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Introduction

This trip focuses on late Quaternary alluvial stratigraphy and geoarcheology in the central Great Plains of Kansas and extreme northern Oklahoma. The first stop is at the Claussen archeological site on the bank of Mill Creek in northeastern Kansas. Mill Creek is a major tributary of the Kansas River and a good example of a Flint Hills stream. An extensive cutbank at the Claussen site provides an opportunity to examine the stratigraphic relationships of early, middle, and late Holocene alluvial fills. Archeological materials are contained in these fills, including Paleoarchaic cultural deposits at depths of 8–10 m below the land surface.

We will cross a steep bioclimatic gradient as we travel west from tallgrass prairie at the Claussen site to the shortgrass prairie of southwestern Kansas. At Stop 2, west of Garden City, Kansas, we will visit the Simshauser archeological site in Mattox Draw. Stratified cultural deposits at Simshauser, including an early Paleoindian component, are contained in an alluvial fan on the margin of the draw. We will examine the soil-stratigraphy of the fan and hear about the results of recent archeological testing at the site. We will also examine thick sections of late Wisconsinan and early Holocene valley fill beneath a broad terrace in Mattox Draw. Buried soils in the valley fill represent former landscapes that were probably occupied by Paleoindians and perhaps by pre-Clovis people.

The third stop is at the Winger archeological site in extreme southwestern Kansas. Winger is a deeply buried late Paleoindian bison bonebed exposed in the bank of Bear Creek, a tributary of the Arkansas River. The stratigraphic sequence at Winger includes early Holocene playa deposits, early Holocene alluvium, and late Holocene eolian sand. In addition to examining this stratigraphy, we will consider site-formation processes that resulted in excellent preservation of the bonebed.

The final three stops will be at the Waugh, Cooper, and Jake Bluff archeological sites in northwestern Oklahoma slightly south of the Kansas border. These three sites are near each other and are very significant archeological localities. The Waugh site includes a Folsom bison bonebed buried in alluvium at the head of Stockholm Canyon. The Cooper site also is a Folsom site and is a rare example of a stratified Paleoindian bison kill. At Cooper, an arroyo was used by Folsom hunters on three separate occasions to trap and kill bison. Jake Bluff, which is located adjacent to Cooper, represents a Clovis bison kill. Archeological excavations will be underway during our visit to Jake Bluff.

A road log is not included with this field trip. Unfortunately, Paleoindian sites are threatened by artifact collectors who often turn to the literature for site locations. The recent destruction of the Burntwood Creek site in northwestern Kansas is a case in point. Hence, we do not want to reveal the precise locations of the sites that will be visited during this trip.

Trip 4, Stop 1: The Claussen Site (14WB322)

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The Claussen site (14WB322) is located near the town of Paxico in the Flint Hills of northeastern Kansas (Stop 1 in fig. 4-1). This site has deeply buried archeological deposits preserved beneath the valley floor of lower Mill Creek, a large stream that joins the Kansas River approximately 20 km northeast of Paxico. The cultural deposits are exposed in a steep cutbank formed by lateral migration of the creek (fig. 4-2). Claussen was discovered by Rolfe Mandel in 1999, and subsequent archeological investigations identified multiple buried cultural deposits, including two Paleoarchaic components in the upper 40 cm of the A horizon of a buried soil approximately 10 m below the surface of the T-2 terrace of Mill Creek. Also, a cultural component of unknown age is about 10 m below the T-2 surface. Charcoal from a hearthlike feature in the upper Paleoarchaic component yielded a ^{14}C age of 8,800 $\pm$ 150 yrs B.P. (ISGS-4684). Charcoal from cultural features 25–35 cm below the upper component yielded AMS ^{14}C ages of 9,225 $\pm$ 30 yrs B.P. (ISGS-AO479) and 9,225 $\pm$ 35 yrs B.P. (ISGS-AO480). Given the paucity of stratified, well-dated early Holocene sites in the Eastern Plains, Claussen is an important "window" into this period. Also, the depth of the buried cultural deposits at Claussen underscores the problem of finding early sites in alluvial settings on the Plains.

Environmental Setting

The Claussen site is in the Flint Hills region of the Central Lowlands Physiographic Province (Fenneman, 1931). This region trends north-south along the western edge of the Osage Cuestas and glaciated region of Kansas. The Flint Hills derived their name from the abundance of chert, or flint, scattered across the uplands. Like the Osage Cuestas to the east, the Flint Hills were formed by erosion of westward-dipping strata. The bedrock geology is characterized by interbedded limestones and shales deposited in shallow seas during the Permian Period about 280 million years ago. Differential erosion of these alternating beds of limestone and shale produce hillsides with stair-step topography. Some of the limestones contain layers or nodules of chert. Because chert is much less soluble than the limestone that encloses it, weathering of the softer rock forms a clay-rich soil containing many chert fragments (Wilson, 1984, p. 19). This gravel-rich soil armors the rocky uplands and reduces erosion rates compared to rates in adjacent areas where the limestone does not contain chert.

From an archeological perspective, the abundance of chert in the limestones is perhaps the most important characteristic of the Flint Hills environment. Due to its superior flaking qualities, Flint Hills chert provided excellent raw material for chipped stone tools and was heavily exploited by prehistoric inhabitants of the region.

The Flint Hills region is one of the largest expanses of tallgrass prairie remaining in North America (Chapman et al., 2001). The dominant grasses are big bluestem (*Andropogon gerardi*), little bluestem (*Andropogon scoparius*), Indian grass (*Sorghastrum nutans*), and switchgrass (*Panicum virgatum*). Trees are mostly restricted to valley floors, rugged ravines, and rocky hillsides. The riparian forests are dominated by cottonwood (*Populus deltoides*), hackberry (*Celtis occidentalis*), willow (*Salix* sp.), and elm (*Ulmus* sp.). Juniper (*Juniperus* sp.) and bur oak (*Quercus macrocarpa*) tend to occur in ravines and on rocky slopes that are insolated from fire, and black walnut (*Juglans nigra*), green ash (*Fraxinus pennsylvanica*), and sycamore (*Platanus occidentalis*) are common on mesic foot slopes and toe slopes.

Geomorphology and Alluvial Stratigraphy

At the Claussen site, the valley floor of Mill Creek is about 3 km wide and consists of a modern floodplain (T-0), a low alluvial terrace (T-1), and a high alluvial terrace (T-2). The T-0 surface is on the west side of the creek and is 1–2 m above mean low water level. As the lowest geomorphic surface in the valley, T-0 is frequently flooded, especially during the spring and early summer.

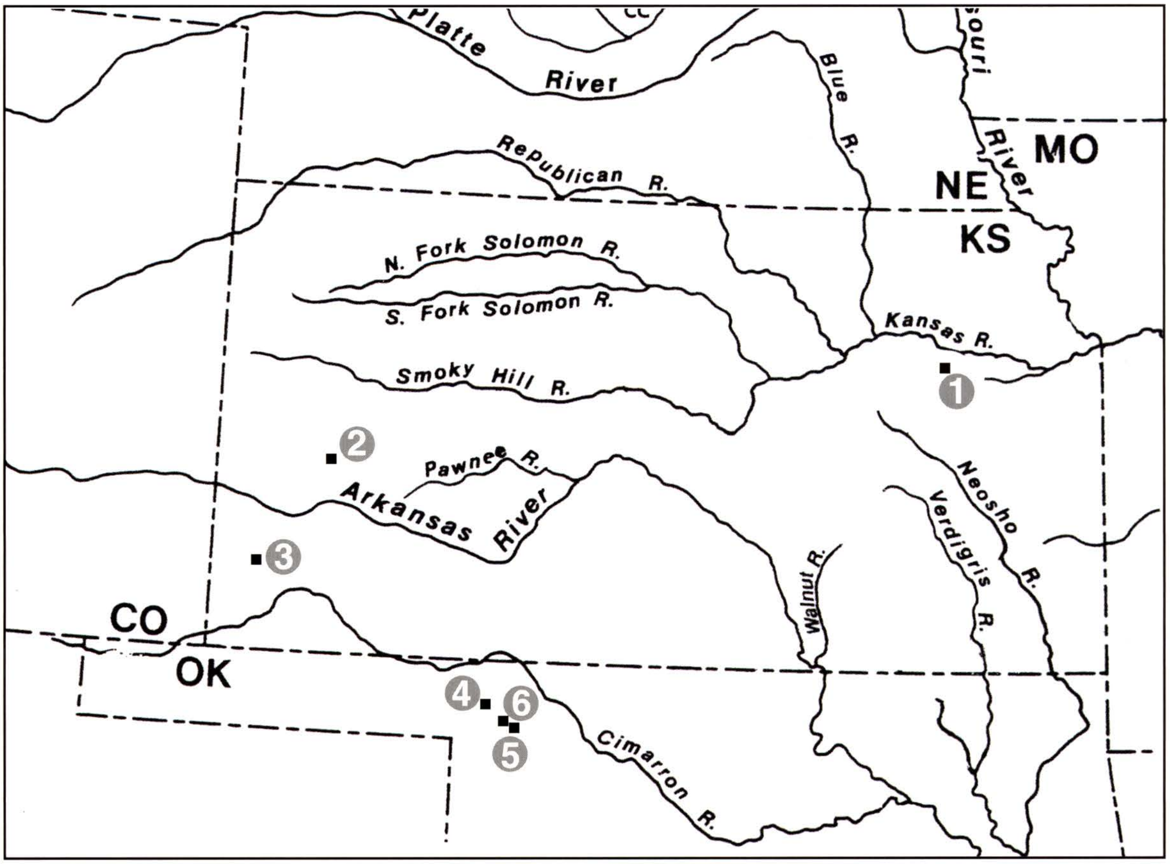

Figure 4-1. Map showing the stops for Trip 4. Stop 1: Claussen site; Stop 2: Simshauser site and Mattox Draw; Stop 3: Winger site; Stop 4: Waugh site; Stop 5: Cooper site; Stop 6: Jake Bluff site.

Figure 4-2. Photograph of the cutbank at the Claussen site. The vehicle is on the T-1 terrace, and the T-2 terrace is in the left foreground. The view is to the southeast. Mill Creek flows to the right in the photograph.

The T-1 surface is 2–3 m above the T-0 surface and is present on the east side of the creek at Claussen. Historic floods are known to overtop T-1, and the remnant of T-1 on the south side of the site is characterized by low-relief flood chutes oriented east-west. At Claussen, a prominent 5-m-high scarp separates the T-1 and T-2 surfaces (figs. 4-2 and 4-3).

The T-2 terrace is the most extensive geomorphic surface within the valley of lower Mill Creek. At the Claussen site, T-2 spans more than three-quarters of the valley floor and is a flat and featureless surface. It gradually merges with colluvial aprons and alluvial fans at the foot of the valley wall north of I-70. However, in some reaches of lower Mill Creek, a 4–5-m-high scarp separates T-2 from narrow remnants of a Pleistocene terrace (T-3).

Three units of the DeForest formation are exposed in the cutbank at Claussen: the Gunder, Roberts Creek, and Honey Creek members (fig. 4-4). The DeForest formation is a lithostratigraphic unit containing fine-grained Holocene alluvium (Bettis, 1995). Although it was originally defined as a sequence of alluvial fills in small valleys of the Loess Hills in western Iowa (Daniels et al., 1963), the DeForest formation has been expanded into eastern Kansas (Mandel et al., 1991; Mandel, 1994; Mandel and Bettis, 2001, 2003).

The Honey Creek member forms most of the valley fill beneath the T-1 terrace at Claussen. This unit consists of at least two laterally inset fills composed of stratified, dark-grayish-brown (10YR 4/2, moist) and dark-brown (10YR 3/3, moist) silt loam grading downward to grayish-brown (10YR 5/2, moist) and brown (10YR 4/3, moist) loam and fine sandy loam.

A buried Middle Ceramic component was observed in the Honey Creek member when the Claussen site was recorded by Mandel. This component is at a depth of 2.1–2.2 m below the T-1 surface and consists of mussel shells, chert flakes, fire-cracked rocks, pottery sherds, and burned-rock features. Charcoal from a burned-rock feature exposed in the cutbank yielded a radiocarbon age of 810 $\pm$ 70 yrs B.P. (ISGS-4831). In 2003, archeological investigations at Claussen included a 22-m^2 excavation block that exposed the Middle Ceramic component in the T-1 fill (fig. 4-5).

Sediments comprising the Roberts Creek member fill a channel (flood chute?) at the foot of the scarp separating the T-1 and T-2 surfaces (figs. 4-4 and 4-6). A shallow buried soil with a thick, black (10YR 2/1, moist), cumulic A horizon is developed in the Roberts Creek member. Five separate cultural horizons were recorded in this buried soil during the 2003 excavations at Claussen. Although the numerical age of the cultural deposits is unknown, Early Ceramic pottery sherds (ca. 2,000-1,000 yrs B.P.) were recorded in the upper cultural horizon.

The Gunder member composes the valley fill beneath the T-2 terrace and mostly consists of moderately oxidized, fine-grained alluvium (fig. 4-7). In the section exposing

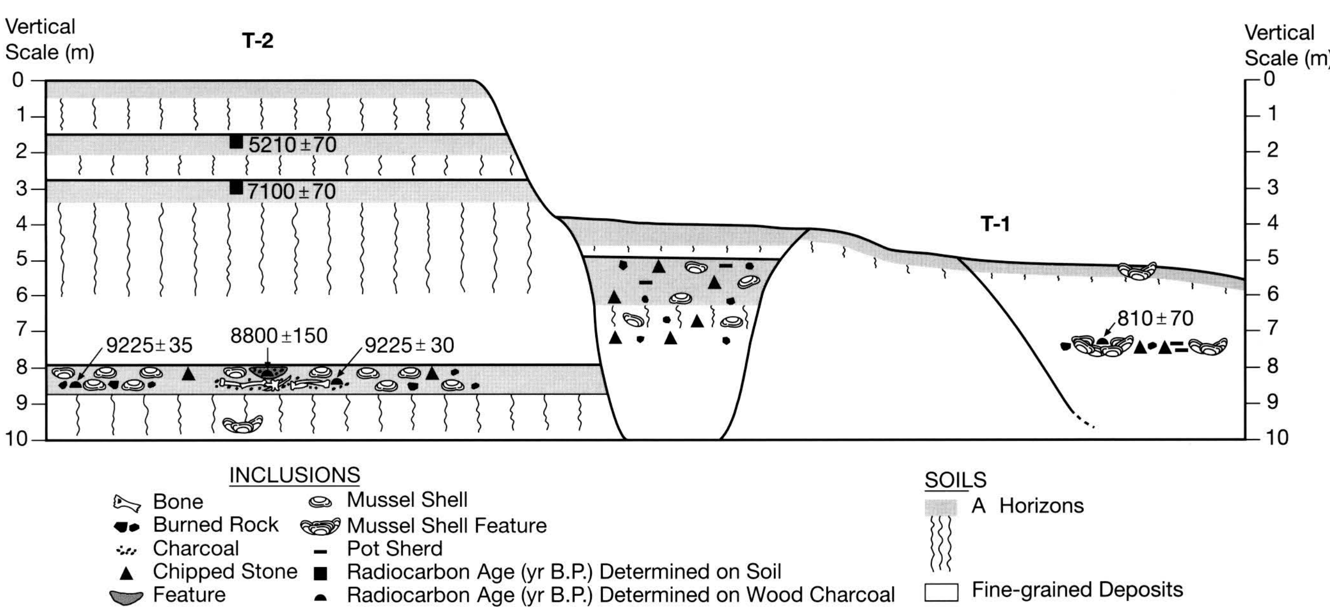

Figure 4-3. Diagram of the cutbank at the Claussen site showing the geomorphology, soil-stratigraphy, archeological components, and radiocarbon ages.

T-2 fill at Claussen, the Gunder member is reduced below a depth of 9.7 m. The surface soil (Soil 1) developed at the top of the Gunder member has a thick, strongly expressed A-BA-Bw-Btk profile (fig. 4-8). This soil is welded to the A horizon of buried soil (Soil 2) at a depth of 1.52–1.87 cm below the T-2 surface. Soil 2 is 133 cm thick and has a strongly expressed A-Btk profile. Decalcified organic carbon from the upper 10 cm of Soil 2 yielded a radiocarbon age of 5,210 ± 70 yrs B.P. (ISGS-4511). Soil 2 is welded to the A horizon of a buried soil (Soil 3) at a depth of 2.85–3.25 m below the T-2 surface. Soil 3 is 310 cm thick and has a strongly expressed A-Btk-Bkss profile. Decalcified organic carbon from the upper 10 cm of Soil 3 yielded a radiocarbon age of 7,100 ± 70 yrs B.P. (ISGS-4514). Laminated silty alluvium separates Soil 3 from the deepest buried soil (Soil 4) in the section. Soil 4 is 2 m thick and has a thick, cumulic A horizon above a Btk horizon. Radiocarbon ages determined on charcoal from cultural features indicate that the rate of sedimentation was very slow around 9,225 yrs B.P., and the early Holocene floodplain was stable by ca. 8,800 yrs B.P.

The soil-stratigraphic record and radiocarbon chronology indicate that Paleoarchaic people repeatedly occupied the early Holocene (ca. 9,200–8,800 yrs B.P.) floodplain of Mill Creek at the Claussen site. Slow aggradation accompanied by pedogenesis formed a stratified record of human occupation within Soil 4. Rapid aggradation soon after ca. 8,800 yrs B.P., as indicated by a radiocarbon age at the Imthurn archeological site about 6 km downstream from Claussen, buried the early Holocene soil and sealed Paleoarchaic and potentially older cultural components beneath an 8-m-thick package of early through middle Holocene alluvium.

Archeology

In 2003, 11.5 1 × 1-m archeological excavation units were placed along a 25-m-long exposure at the base of the cutbank exposing the T-2 valley fill at Claussen. The units were excavated in two areas: an upstream area designated as the west block, and an adjacent downstream area designated as the east block. The excavations yielded 261 bulk soil samples, each weighing approximately 50 pounds. These samples were water-screened on site.

A total station was used to map all cultural materials recorded in the west and east blocks. There are 185 mapped items from the Paleoarchaic component dating to ca. 9,250 yrs B.P., including chipped stone, bone, shell, and charcoal samples. Notable peaks in the vertical distribution of cultural materials and the absence of obvious sorting based on size, shape, or material suggest that the spatial integrity of this component is relatively intact.

No diagnostic artifacts were found within the 8,800-yr-old component or underlying cultural deposits during the 2003 field season. However, inspection of the site in 2005 resulted in the discovery of a Dalton projectile point in a small mass of soil that fell out of the Akb3

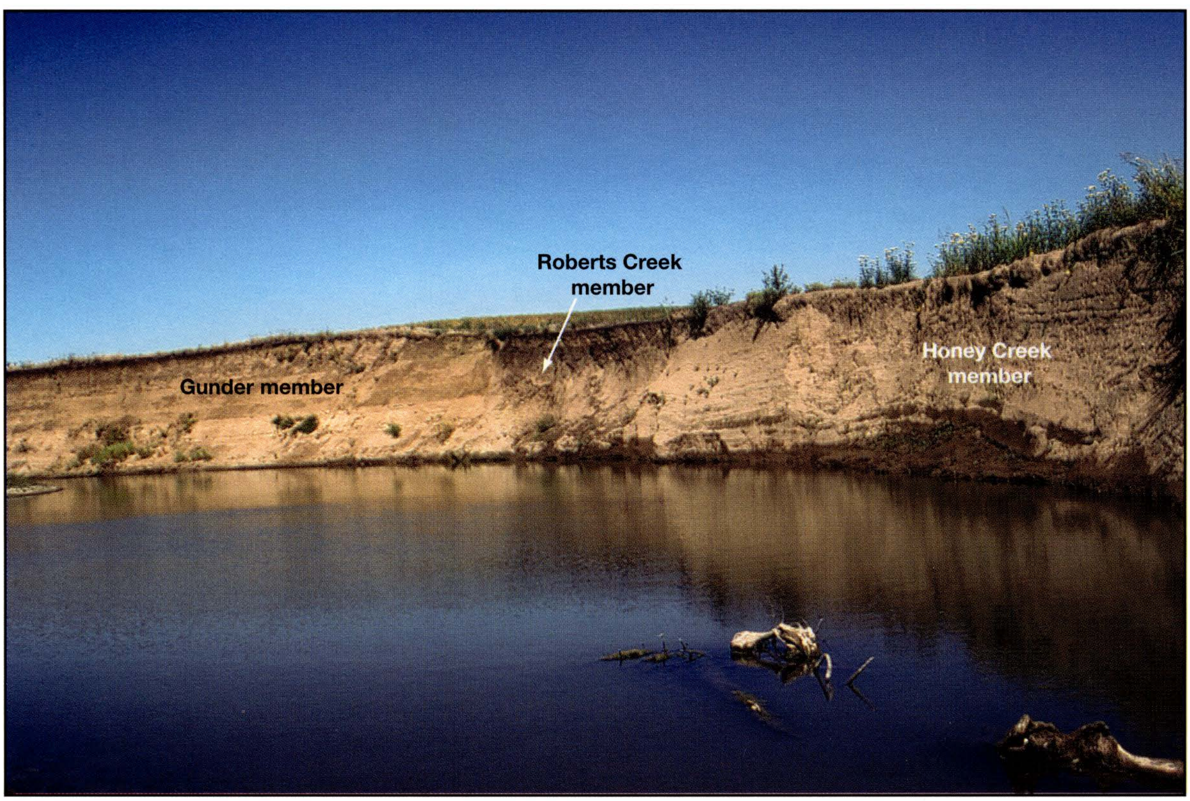

Figure 4-4. Photograph of the cutbank at the Claussen site showing the stratigraphic relationships of the Gunder, Roberts Creek, and Honey Creek members of the DeForest formation. The view is to the east.

Figure 4-5. Photograph of the 2003 excavation block exposing the Middle Ceramic component (floor of the excavation block) in the Honey Creek member at the Claussen site. The view is to the north.

Figure 4-6. Photograph of the channel fill exposed in the cutbank at the Claussen site. Early Ceramic pottery sherds were recorded in the buried soil developed in the upper part of the channel fill. The view is to the east.

horizon and came to rest at the bottom of the section. Also, collectors have found Dalton and lanceolate projectile points on a gravel bar immediately downstream from the site.

Lithic materials identified to date are all derived from local or nearby sources of Permian-age Florence formation cherts. The occurrence of refits, the wide range in sizes and shapes of recovered pieces, and the lack of evidence for edge abrasion or stream rolling suggests that post-depositional alluvial transport has not been significant.

The few tools identified include two modified flakes and three flaked and battered cobbles (core/hammers). All are expedient tools that were made, used, and discarded at the site. One flake tool refits with one of the cobbles, which was apparently collected from the stream gravel in Mill Creek.

The high proportion of flakes that have platforms (75% platform-remnant-bearing based on piece plotted and 1/4-inch screen samples) indicates that post-manufacture breakage due to factors such as trampling has been minimal. This suggests an occupation or activity area usage of relatively short duration.

The rare occurrence of burned lithics (n = 5, 6% of piece plotted and 1/4-inch sample), which does not include the large core/hammers, suggests that burning of lithics was incidental and related to hearths rather than to general surface fires or natural burns. Also, the presence of late-stage thinning and retouch flakes from bifaces indicates that tools were present that have not been recovered. In addition to flake tools, bifacial knives or projectile-point/knives were apparently used and resharpened, or possibly manufactured, at the site.

The archeofaunal assemblage dating to ca. 9,250 yrs B.P. included a variety of vertebrate (22 taxa) and invertebrate (three taxa) species. Medium ungulate taxa (*Odocoileus* or *Odocoileus/Antilocapra*) were the most common economic species. However, avian species, notably *Meleagris gallopavo*, also made up a significant portion of the archeofaunal assemblage. Bison (species indeterminate), *Vulpes velox*, lagomorph, and rodent taxa are all represented by single individuals and low NISPs. Preliminary identification of the mussel fauna indicates the presence of three species: *Amblema plicata*, *Quadrula quadrula*, and *Fusconaia flava*.

Burned bone specimens composed 23% of the mapped faunal sample and 37% of the bone recovered in the screens from 32R-25, the most productive excavation unit. Burning on both the interior and exterior of bone fragments suggests that most heat modification occurred after consumption.

Perhaps the most intriguing (and exciting) aspect of the archeological record at Claussen is the faunal assemblage. There is evidence that Early Archaic and Late Paleoindian occupants of the Western and Central Plains relied heavily on bison. The Winger site in southwestern Kansas, which will be visited during the field trip, is a case in point. However, at Claussen, faunal remains dating to ca. 8,800–9,300 yrs B.P. indicate that occupants of the

Figure 4-7. Photograph of the valley fill (Gunder member) beneath the T-2 terrace at the Claussen site. The person (just to the right of the center of the photograph) is pointing to Paleoarchaic cultural deposits at the top of the lowest buried soil in the section.

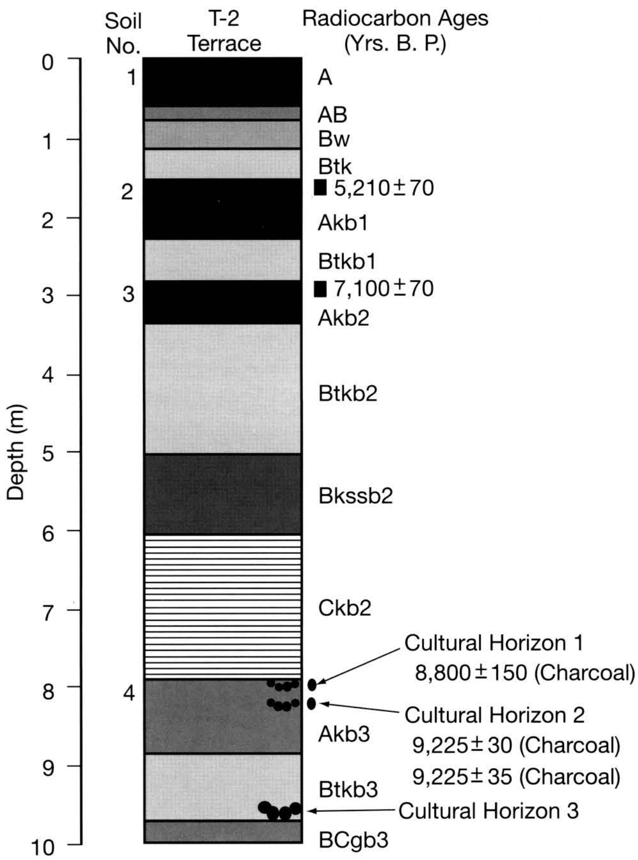

Figure 4-8. Radiocarbon ages of T-2 terrace at the Claussen site.

site relied on a variety of animals for food. A large part of the assemblage is fragmented and burned, indicating a complex range of cultural and noncultural factors were involved in site-formation processes. Nevertheless, the faunal record at Claussen strongly suggests local subsistence behaviors at ca. 9,000 yrs B.P. that were very different from the bison-focused subsistence strategies of Native Americans on the Plains at that time. The diverse fauna and expedient technology at the site indicate an "Archaic" adaptation that may be more comparable to Dalton and later Eastern Woodland cultural complexes. Continued research at Claussen holds considerable potential to address the Paleoindian/Archaic interface.

Acknowledgment

The University of Kansas Odyssey Archeological Research Fund supported our investigations at the Claussen site.

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Trip 4, Stop 2: Simshauser Site (14KY102) and Mattox Draw

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Introduction

At Stop 2 in southwestern Kansas (fig. 4-1), we will visit the Simshauser site (14KY102) and adjacent exposures of late Wisconsinan and early Holocene valley fill in Mattox Draw. Simshauser is a stratified site with at least one Paleoindian component. The site is associated with a small alluvial fan on the margin of the valley floor of Mattox Draw. The channel of Mattox Draw migrated laterally and cut into the fan, opening a "window" into this landform sediment assemblage. We will examine a 4-m-high section of early Holocene fan deposits and discuss the soil stratigraphy and associated cultural deposits.

Following our visit to the Simshauser site, we will walk north several hundred meters and examine late Wisconsinan and early Holocene valley fill exposed in a long, high cutbank along a meander loop of Mattox Draw. The cutbank reveals a complex soil-stratigraphic record and sheds light on why Paleoindian sites are so elusive on the High Plains.

Setting

Stop 2 is located in the semiarid High Plains region of Fenneman's (1931) Great Plains physiographic province. The broad, flat uplands are mantled by 2–3 m of late Wisconsinan-age Peoria Loess. In many areas, a unit of Pleistocene alluvium separates the loess from late Tertiary alluvium of the Ogallala Formation.

Mattox Draw, which drains to the Arkansas River, is deeply entrenched in the Ogallala Formation (fig. 4-9). Like the other dry valleys, or draws, in the region, Mattox is an intermittent stream; it only carries water immediately after heavy rainfalls. Quaternary alluvium is stored beneath two geomorphic surfaces in the draw: a low, narrow floodplain (T-0) and a high, broad terrace (T-1) (fig. 4-9). Small alluvial fans, such as the one at the Simshauser site, are common along the margins of the draw and often merge with valley fill beneath the T-1 terrace.

Geomorphology, Stratigraphy, and Archeology of the Simshauser Site

The Simshauser alluvial fan formed at the mouth of a small, first-order drainage element extending into the adjacent loess-mantled uplands (fig. 4-10). Although the distal end of the fan has been removed by stream erosion, the geometry of the fan is still quite apparent. There are distinct lobes of sediment on both sides of the fan-head trench, and the modern surface of the lobes dips away from the trench. The cutbank at Simshauser provides an excellent opportunity to view the buried paleo-landscape within and adjacent to the fan. The geometry of buried fan surfaces, marked by paleosols, resembles the modern geometry of the fan, i.e., the buried soils dip away from the fan head trench (figs. 4-11 and 4-12).

The alluvium composing the Simshauser fan is predominantly silt loam derived from the loess on the adjacent uplands. However, coarser sediment, including sand and gravel, occurs in the lower 1.5 m of the fan (fig. 4-11). Three buried soils are found in the upper 4 m of the fan (fig. 4-11). The upper buried soil (100–148 cm below surface) has a weakly expressed A–AC profile (fig. 4-11 and table 4-1). The middle buried soil (148–188 cm below surface) is developed at the top of an upward-fining sequence and has a well-expressed Bk–BCk profile (fig. 4-11 and table 4-1); the A horizon was stripped off by erosion before the soil was buried. The lower buried soil (303–358 cm below surface) also is developed at the top of an upward-fining sequence but has a thick, cumulic Ak horizon above an ACk horizon (fig. 4-11 and table 4-1). Decalcified soil carbon from the upper 10 cm of the Akb3 and Bk1b2 horizons yielded radiocarbon ages of 10,170 $\pm$ 70 yrs B.P. (ISGS-4675) and 9,420 $\pm$ 70 yrs B.P. (ISGS-4676), respectively (fig. 4-11).

The initial investigation at Simshauser focused on the cutbank exposing alluvium beneath the west lobe of the fan. Cultural deposits, including flakes, burned rock, and bison bones, were recorded in the upper 15 cm of the A horizon of the surface soil and in the A horizon of the upper buried soil (fig. 4-11). Presently, only one thing that can be said about the ages of these archeological components: they are younger than ca. 9,400 yrs B.P.

In addition to the upper two archeological components, bison bones and a single chert flake were found in the ACkb3 horizon of the lower buried soil (fig. 4-11 and 4-13). Based on the radiocarbon age determined on decalcified soil carbon from the overlying Akb3 horizon, the lowest archeological component in the section is older than ca. 10,100 yrs B.P. Hence, unless the radiocarbon age is problematic, this component represents a Paleoindian camp and/or bison kill site. Archeological testing of the Simshauser site will be conducted by the University of Kansas, June 15–25, 2004, with most of the

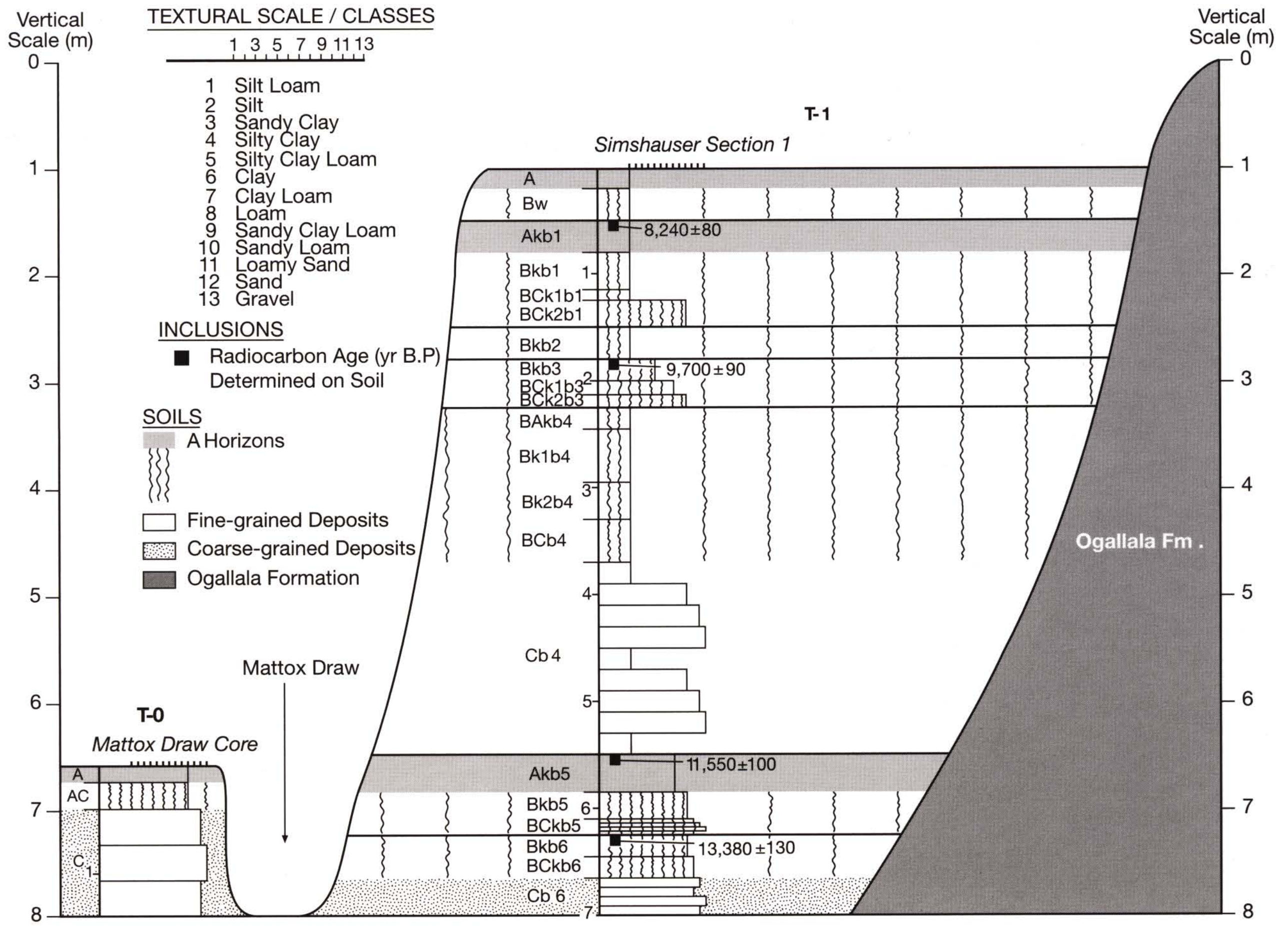

Figure 4-9. Cross section of Mattox Draw showing the landform sediment assemblages, soil stratigraphy, sedimentology, and radiocarbon ages.

Figure 4-10. Photograph of the Simshauser fan (site 14KY102). The view is to the north.

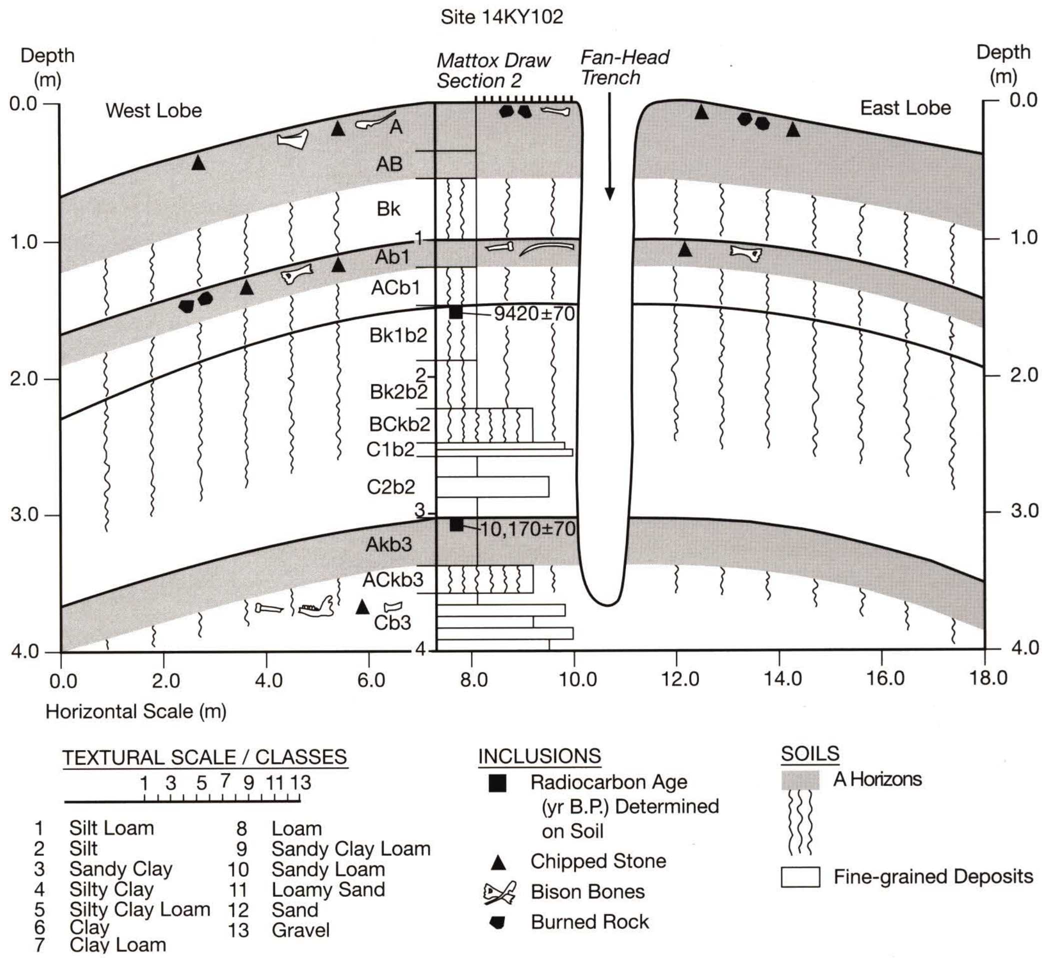

Figure 4-11. Diagram of the cutbank at the Simshauser site showing the soil stratigraphy, sedimentology, archeological components, and radiocarbon ages.

Figure 4-12. Photograph of the west lobe of the Simshauser fan (site 14KY102). The view is to the north. Note the buried soil dipping to the west.

TABLE 4-1. Description of the section exposed beneath the west lobe at the Simshauser site.

Depth (cm)	Soil Horizon	Description
0–5	A1	Brown (10YR 5/3) silt loam, dark-brown (10YR 4/3) moist; weak medium and fine platy structure parting to weak fine granular; friable; many fine and very fine roots; abrupt smooth boundary.
5–35	A2	Brown (10YR 5/3) silt loam, dark-brown (10YR 4/3) moist; weak fine subangular blocky structure parting to weak fine and medium granular; friable; many fine and very fine roots; many worm casts; gradual smooth boundary.
35–55	AB	Brown (10YR 5/3) silt loam, dark-brown (10YR 4/3) moist; weak fine subangular blocky structure parting to weak fine and medium granular; friable; many fine and very fine roots; common worm casts; gradual smooth boundary.
55–100	Bk	Yellowish-brown (10YR 5/4) silt loam, dark-yellowish-brown (10YR 4/4) moist; weak fine and medium prismatic structure parting to moderate fine subangular blocky; friable; common films and threads of calcium carbonate; many fine and very fine roots; common worm casts and open worm burrows; abrupt smooth boundary.
100–120	Akb1	Brown (10YR 5/3) light silty clay loam, dark-brown (10YR 4/3) moist; weak medium prismatic structure parting to moderate coarse granular and very fine subangular blocky; friable; common grayish-brown (10YR 5/2) organic coatings on ped faces; many films and threads of calcium carbonate; many fine and very fine roots; common worm casts and open worm burrows; gradual smooth boundary.
120–148	ACkb1	Brown (10YR 5/3) light silty clay loam, dark-brown (10YR 4/3) moist; weak medium and coarse prismatic structure parting to medium and fine angular blocky; hard; faint bedding; peds part along bedding planes; common grayish-brown (10YR 5/2) organic coatings on ped faces; few laminae of pale-brown (10YR 6/3) silt loam and yellowish-brown (10YR 5/4) loam; many films and threads of calcium carbonate; common fine and very fine roots; few worm casts and open worm burrows; abrupt smooth boundary.
148–188	Bk1b2	Yellowish-brown (10YR 5/4) to brown (10YR 5/3) silt loam, dark-yellowish-brown (10YR 4/4) to dark-brown (10YR 4/3) moist; weak fine prismatic structure parting to weak fine subangular blocky; hard; many films and threads of calcium carbonate; few fine and very fine roots; gradual smooth boundary.
188–223	Bk2b2	Yellowish-brown (10YR 5/4) silt loam, dark-yellowish-brown (10YR 4/4) to dark-brown (10YR 4/3) moist; weak fine prismatic structure parting to weak fine subangular blocky; hard; many very fine and fine threads of calcium carbonate; few fine and very fine roots; gradual smooth boundary.
223–248	BCkb2	Yellowish-brown (10YR 5/4) silt loam, dark-yellowish-brown (10YR 4/4) to dark-brown (10YR 4/3) moist; very weak fine prismatic structure parting to very weak fine subangular blocky; hard; faint bedding; common fine threads of calcium carbonate; gradual smooth boundary.
248–257	C1b2	Stratified yellowish-brown (10YR 5/4) sand interbedded with fine gravel; single grain; loose; abrupt smooth boundary.
257–303	C2b2	Stratified brown (10YR 5/3) and pale-brown (10YR 6/3) silt loam interbedded with very fine sandy loam; few lenses of sand and granules; massive; friable; abrupt smooth boundary.
303–338	Akb3	Dark-grayish-brown (10YR 4/2) silt loam, very dark grayish-brown (10YR 3/2) moist; weak fine prismatic structure parting to moderate medium and coarse granular; friable; common fine and very fine threads of calcium carbonate; gradual smooth boundary.
338–358	ACkb3	Dark-brown (10YR 4/3-3/3) loam, very dark grayish-brown (10YR 3/2) moist; weak fine granular structure; friable; faint bedding; common fine and very fine threads of calcium carbonate; common lenses of granules and pebbles; gradual smooth boundary.
358–400	Cb3	Stratified brown (10YR 5/3) silt loam, pale-brown (10YR 6/3) loam, and pale-brown (10YR 6/3) and yellowish-brown (10YR 5/4) fine sandy loam interbedded with sand and fine gravel; massive; hard.

effort focusing on the lowest cultural component. Results of the testing will be presented during our visit to the site.

Mattox Draw

During our first visit to the Simshauser site in 1998, we also inspected the impressive cutbank spanning the meander loop of Mattox Draw immediately south of the site (fig. 4-14). None of the fills in the draws of western Kansas had been dated prior to our visit. Hence, we could only speculate on the age of the deposits exposed in the cutbank. Previous geomorphological investigations that focused on small streams in western Kansas determined that only late Holocene valley fill is stored in their valleys (see Mandel, 1994, 1995). Therefore, we thought this might also apply to Mattox Draw, even though it is an intermittent drainage element high in the drainage network. Our speculation, however, was proven wrong.

A section of valley fill beneath the T-1 terrace of Mattox Draw was cleaned with hand shovels and described (fig. 4-9 and table 4-2). There are six buried soils in the 7-m-high section. The surface soil and first four buried soils at the top of the section form a pedo-complex spanning the upper 3.7 m of the T-1 fill (fig. 4-9 and table 4-2). We refer to it as a pedocomplex because the bottom of each soil is welded to the top of the soil beneath it. All of these soils have Bk horizons, but the carbonate morphology is weak (stage I-I+ as defined by Birkeland [1999, table A1.5]). Decalcified soil carbon from the upper 10 cm of the Akb1 and Bkb3 horizons yielded radiocarbon ages of 8,240 $\pm$ 80 yrs B.P (ISGS-4645) and 9,700 $\pm$ 90 yrs B.P. (ISGS-4643), respectively. A 1.8-m-thick zone of laminated silt loam interbedded with very fine sandy loam, fine sand, and pebbles (Cb4 horizon in fig. 4-9 and table 4-2) separates the pedocomplex from a pair of buried soils near the bottom of the section (fig. 4-9). The next to the lowest buried soil has a well-expressed A-Bk-BCk profile, and the lowest buried soil is represented by a truncated Bk horizon. Decalcified soil carbon from the upper 10 cm of the Akb5 and Bkb6 horizons yielded radiocarbon ages of 11,550 $\pm$ 100 yrs B.P (ISGS-4673) and 13,380 $\pm$ 130 yrs B.P. (ISGS-4672), respectively.

Based on the suite of radiocarbon ages for the T-1 terrace fill, there is a tremendous volume of late Wisconsinan through early Holocene alluvium stored in Mattox Draw. There is no evidence of middle Holocene alluvium in the draw, and late Holocene alluvium that appears to be less than 1,000 years old is restricted to the modern floodplain (T-0). This pattern of late Wisconsinan through early Holocene sediment storage has been documented in draws throughout western Kansas (Mandel, 2001). Since ca. 8,000 yrs B.P., headward erosion in the draws has not been sufficient to remove the late Wisconsinan and early Holocene alluvium. Bettis (1990) noted a similar pattern of sediment storage high in drainage networks in the Loess Hills of western Iowa.

From an archeological perspective, the presence of multiple buried soils dating between ca. 13,300 and 8,200 yrs B.P. is very significant. These soils represent landscapes that may have been occupied by humans during and immediately after the Pleistocene–Holocene transition. It is important to note that the Clovis-age landscape is about 5.5 m below the T-1 surface. This underscores the problem of finding Paleoindian archeological deposits in a stratified context in the region. Artifact collectors have discovered Early through Late Paleoindian artifacts in the dry channel of Mattox Draw, and we now know the likely source of these materials: the T-1 fill. The task is to find "windows" into the T-1 fill and thereby systematically search for in situ Paleoindian and pre-Clovis cultural deposits. Such an effort is currently underway through the University of Kansas Odyssey Archeological Research Program.

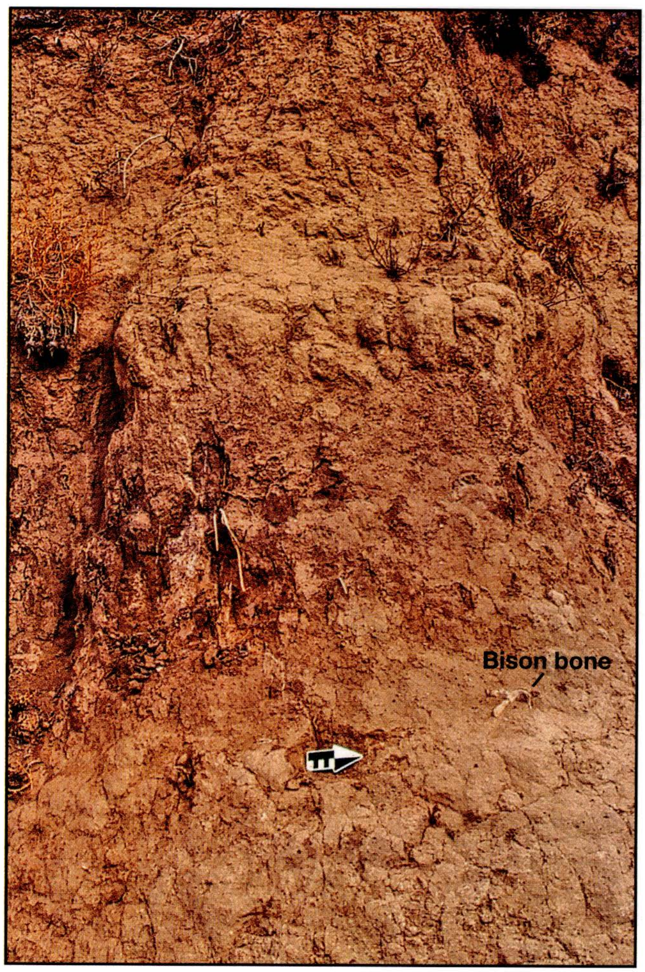

Figure 4-13. Photograph of the thick, buried soil and underlying bison bone in the west lobe of the Simshauser fan (site 14KY102).

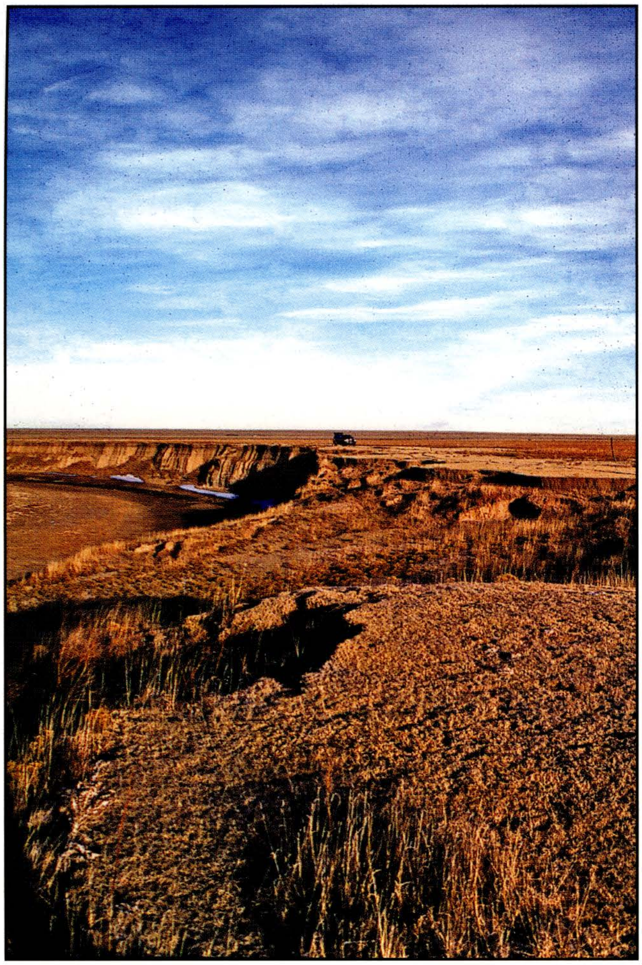

Figure 4-14. Photograph of the cutbank that spans the meander loop of Mattox Draw. The view is to the north.

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TABLE 4-2. Description of the section exposed beneath T-1 terrace of Mattox Draw.

Depth (cm)	Soil Horizon	Description
0–20	A	Brown (10YR 5/3) to dark-brown (10YR 4/3) silt loam, dark-brown (10YR 4/3) moist; weak fine subangular blocky structure parting to weak fine and medium granular; friable; many fine and very fine roots; gradual smooth boundary.
20–50	Bw	Brown (10YR 5/3) silt loam, dark-brown (10YR 4/3) moist; weak fine prismatic structure parting to weak fine subangular blocky; friable; common fine and very fine roots; clear smooth boundary.
50–80	Akb1	Dark-brown (10YR 4/3) silt loam, dark-brown (10YR 3/3) moist; weak fine prismatic structure parting to moderate medium and coarse granular; friable; common films and few threads of calcium carbonate; common fine and very fine roots; gradual smooth boundary.
80–115	Bkb1	Brown (10YR 5/3) silt loam, dark-brown (10YR 4/3) moist; weak fine and medium prismatic structure parting to weak fine and very fine subangular blocky; friable; few fine threads of calcium carbonate; few fine and very fine roots; gradual smooth boundary.
115–150	BCkb1	Brown (10YR 5/3) to yellowish-brown (10YR 5/4) loam grading downward to fine sandy loam, dark-brown (10YR 4/3) to dark-yellowish-brown (10YR 4/4) moist; weak coarse prismatic structure parting to very weak fine and very fine subangular blocky; friable; few fine threads of calcium carbonate; few fine and very fine roots; abrupt smooth boundary.
150–180	Bkb2	Brown (10YR 5/3) to yellowish-brown (10YR 5/4) silt loam, dark-brown (10YR 4/3) to dark-yellowish-brown (10YR 4/4) moist; weak medium prismatic structure parting to moderate medium and fine subangular blocky; friable; common films and threads of calcium carbonate; few fine and very fine roots; abrupt smooth boundary.
180–200	Bkb3	Brown (10YR 5/3) light silty clay loam, dark-brown (10YR 4/3) moist; moderate medium prismatic structure parting to moderate fine and medium subangular blocky; hard; many nearly continuous films of calcium carbonate on ped faces; common fine threads of calcium carbonate; few very fine roots; gradual smooth boundary.
200–225	BCkb3	Brown (10YR 5/3) to yellowish-brown (10YR 5/4) loam grading downward to very fine sandy loam, dark-brown (10YR 4/3) to dark-yellowish-brown (10YR 4/4) moist; weak coarse prismatic structure parting to very weak fine and very fine subangular blocky; friable; very faint bedding; common films and very fine threads of calcium carbonate; few very fine roots; abrupt smooth boundary.
225–245	BAkb4	Brown (10YR 5/3) silt loam, dark-brown (10YR 4/3) moist; weak fine prismatic structure parting to weak fine and to very fine subangular blocky; friable; many fine and very fine dark-grayish-brown (10YR 4/2) organic coatings on ped surfaces; many threads and common films of calcium carbonate; few very fine roots; gradual smooth boundary.
245–295	Bk1b4	Brown (10YR 5/3) to yellowish-brown (10YR 5/4) silt loam, dark-brown (10YR 4/3) to dark-yellowish-brown (10YR 4/4) moist; weak medium and fine prismatic structure parting to weak fine subangular blocky; friable; common very fine threads and few films of calcium carbonate; few very fine roots; gradual smooth boundary.
295–330	Bk2b4	Pale-brown (10YR 6/3) to light-yellowish-brown (10YR 6/4) silt loam, brown (10YR 5/3) to yellowish-brown (10YR 5/4) moist; weak medium and fine prismatic structure parting to weak fine subangular blocky; friable; common very fine threads and few films of calcium carbonate; few very fine roots; gradual smooth boundary.
330–370	BCb4	Pale-brown (10YR 6/3) to light-yellowish-brown (10YR 6/4) silt loam, brown (10YR 5/3) to yellowish-brown (10YR 5/4) moist; weak coarse prismatic structure parting to very weak fine subangular blocky; friable; faint bedding; few very fine roots; gradual smooth boundary.
370–550	Cb4	Laminated pale-brown (10YR 6/3) silt loam interbedded with very fine sandy loam, brown (10YR 5/3) moist; massive; soft, very friable; common lenses of fine sand and small pebbles at 440–445 cm; abrupt wavy boundary.

TABLE 4-2 continued.

Depth (cm)	Soil Horizon	Description
550–585	Akb5	Dark-brown (10YR 4/3) loam, dark-brown (10YR 3/3) moist; very weak medium prismatic structure parting to weak fine granular; friable; common coarse threads and few films of calcium carbonate; few siliceous granules; gradual smooth boundary.
585–610	Bkb5	Brown (10YR 5/3) fine sandy loam, dark-brown (10YR 4/3) moist; weak fine and medium prismatic structure parting to weak fine subangular blocky; friable; common coarse threads and few films of calcium carbonate; few siliceous pebbles and granules; gradual smooth boundary.
610–625	BCkb5	Brown (10YR 5/3) to yellowish-brown (10YR 5/4) loamy fine sand, yellowish-brown (10YR 5/4) moist; very weak coarse prismatic structure parting to very weak fine subangular blocky; friable; common coarse threads and few films of calcium carbonate; common lenses of siliceous granules and fine pebbles; abrupt smooth boundary.
625–645	Bkb6	Brown (10YR 5/3) fine sandy loam, dark-brown (10YR 4/3) moist; weak medium prismatic structure parting to weak fine subangular blocky; friable; common fine threads and few fine films of calcium carbonate; common lenses of siliceous granules and fine pebbles; gradual smooth boundary.
645–665	BCkb6	Pale-brown (10YR 6/3) loamy fine sand, brown (10YR 5/3) moist; very weak coarse prismatic structure parting to very weak fine subangular blocky; friable; common fine threads and few films of calcium carbonate; common lenses of siliceous granules and fine pebbles; gradual smooth boundary.
665–700+	Cb6	Pale-brown (10YR 6/3) very fine and fine sand interbedded with loamy fine sand, brown (10YR 5/3) moist; massive; friable; common lenses of siliceous granules and fine pebbles.

Trip 4, Stop 3. The Winger Site

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Introduction

The Winger site (14ST401), located in Stanton County in southwestern Kansas (see Stop 3 in fig. 4-1), is a deeply buried Late Paleoindian bison bonebed. The bonebed is exposed in the bank of Bear Creek, a tributary of the Arkansas River. At Winger, Bear Creek has cut into the edge of one of the many playa basins that dot the landscape.

In 1966, the Winger site was visited by the late Virginia Buckner, an avocational archeologist who lived in southwestern Kansas. Buckner and several of her associates periodically conducted investigations at the site from 1970 to 1972. During that period, at least 45 linear meters of the bonebed were visible at a depth of 2.5–3.0 m below the surface (Buckner, 1970). They excavated two trenches and four contiguous 5 × 5-ft squares down to the bonebed, leaving the bones in place and covering the exposed bone with plastic before closing the excavation in 1972. From the first trench, Buckner reported remains of at least seven bison, using rib cages and vertebral columns for this estimate. She also noted articulated limbs, phalanges, and tooth rows. Although two "flakes of flint" were found at the bone level in the eastern trench (Buckner, 1973), no diagnostic artifacts were recovered during Buckner's excavations. Buckner (1970) reported that the late Dr. Waldo Wedel, senior archeologist with the Smithsonian Institution, visited the Winger site in August 1970, and conducted "a small exploratory excavation." However, Wedel did not conduct further research at the site, and Winger was abandoned by Buckner in 1972.

The Winger site was relocated in June of 2001 as part of Mandel's statewide geoarcheological research project in Kansas. During our initial visit to the site, we discovered that recent construction of a ranch road leading to the dry bed of Bear Creek exposed part of the bonebed in the area of Buckner's excavation. Also, hundreds of bone fragments were scattered across the streambed because of road construction and sand quarrying. A dense, 35-m-long bison bonebed averaging about 25 cm thick was recorded at a depth of 265–290 cm in the cutbank. At that time, a portion of the cutbank was cleaned from top to bottom just west of what we believe was Buckner's excavation, and a 1.5-m-wide profile was described (Mandel and Hofman, 2003). In addition, bones from the floor of the channel and from a small portion of the bonebed were collected. The collagen fraction from a sample of cleaned rib blade fragments collected from the bonebed yielded a $\delta^{13}C$ corrected ^{14}C age of 9,080 $\pm$ 90 yrs B.P. (ISGS-4934) (Mandel and Hofman, 2003). Given this radiocarbon age, we suspected that the bonebed represented a Paleoindian kill site. Hence, the significance of the Winger site for addressing problems of early Holocene paleoecology and archeology was apparent.

Following preliminary work in May 2002, the combined University of Kansas and University of Missouri archeological field school spent most of June 2002 excavating Winger. Approximately 9 m^2 of the bonebed were excavated. Also, most of the disturbed bonebed sediment was wet screened. Soil and stratigraphic investigations accompanied the archeological studies.

Setting

The Winger site is located in the semiarid High Plains region of Fenneman's (1931) Great Plains physiographic province. Surficial deposits of the region consist of late Pleistocene loess (Frye and Leonard, 1951, 1952). The loess mantles Pleistocene alluvium, which in turn overlies late Tertiary alluvium of the Ogallala Formation.

The High Plains surface is relatively flat and featureless. However, there are thousands of shallow depressions, or playas, scattered across the broad interfluves (Frye, 1950; Mandel, 2000). The playas range from a few meters to several kilometers in diameter, and while most are less than 3 m deep, some of the larger ones are 15–20 m deep.

Winger is at the thalweg of a meander reach of Bear Creek, an intermittent stream, or draw, that rarely carries water. Because of lateral migration, the channel of Bear Creek has cut into the southern margin of a modern playa basin (fig. 4-15). This basin is elliptical and has an area of about 1.5 km^2. The present basin floor is about 5 m below the surrounding surface. There is a loess-mantled rim around the playa basin, except where Bear Creek removed the south rim.

The Winger bonebed is contained in playa fill at the base of the north bank of Bear Creek (figs. 4-16 through 4-19). The stratigraphic relationship of this playa fill to the modern playa basin on the northern fringe of the site is unknown.

Soils and Stratigraphy

The playa fill at Winger is cut into late Wisconsinan Peoria Loess and underlying alluvium (fig. 4-20). The fill is clayey, very dark grayish-brown to olive-brown (moist colors), and noncalcareous. Similar pond deposits have been identified in playas in western Kansas (Mandel, 2000). The Winger bonebed is 13 cm below the top of the playa fill (figs. 4-18 and 4-19).

Two soils are developed in the exposed portion of the playa fill. Only the upper 33 cm of the lower playa soil is exposed in the cutbank at Winger. This portion of the soil has ABk-Bk horizonation (fig. 4-21 and table 4-3) and redoximorphic features (i.e., yellowish-red mottles). The texture is silty clay and the carbonate morphology is relatively weak (stage I+ as defined by Birkeland [1999, table A1.5]).

The upper playa soil has an Ak–Btk–BCk profile and is silty clay loam from top to bottom (fig. 4-21 and table 4-3). Moist colors range from very dark grayish-brown (10YR 3/2) in the 3Akb2 horizon to dark-brown (10YR 3/3) in the 3BCtkb2 horizon, and there are few yellowish-brown (10YR 5/4 and 5/6) mottles in the 3Akb2 horizon. Carbonate morphology is weak (stage I as defined by Birkeland [1999, table A1.5]) compared to the underlying playa soil, but there is evidence of clay illuviation in the form of discontinuous clay films on ped faces and the walls of pores in the 3Btkb2 and 3BCtkb2 horizons. Laminae composed of very pale brown (10YR 7/3) and very light grayish-brown (10YR 7/2) silt loam are common in the 3BCtkb2 horizon. They are laterally discontinuous and probably represent slopewash derived from the adjacent Peoria Loess, although an eolian origin cannot be ruled out. The bonebed is almost entirely within the 3Btkb2 horizon. However, there is some bone in the lower 2 cm and upper 3 cm of the 3Akb2 and 3Btkb2 horizons, respectively.

The presence of the bonebed within the 3Btkb2 horizon indicates that the playa fill was aggrading at the time of the bison kill; it was not a completely stable soil-forming environment. The bonebed was quickly buried in pond deposits, and there was an episode of landscape stability and concomitant soil formation in the playa soon after the bonebed formed (ca. 9,000 yrs B.P.). However, as explained in the following discussion, the episode of pedogenesis that produced the upper playa soil was brief.

The pond deposits are mantled by an 87-cm-thick unit of brown to yellowish-brown silty alluvium (fig. 4-20). A soil with an AB-Bw-Bk profile developed in the alluvium (fig. 4-21 and table 4-3). Decalcified bulk organic carbon from the upper 10 cm of the 2ABb1 horizon yielded a $\delta^{13}C$ corrected ^{14}C age of 8,570 $\pm$ 60 yrs B.P. (ISGS-5192). This age, combined with the age determined on bone from the underlying bonebed (ca. 9,000 yrs B.P. [Mandel and Hofman, 2003]), suggests that approximately 500 years elapsed between the time of the kill and end of soil development in the unit of silty alluvium. Other events during these 500 years include the accumulation of 13 cm of pond sediment above the bonebed, development of

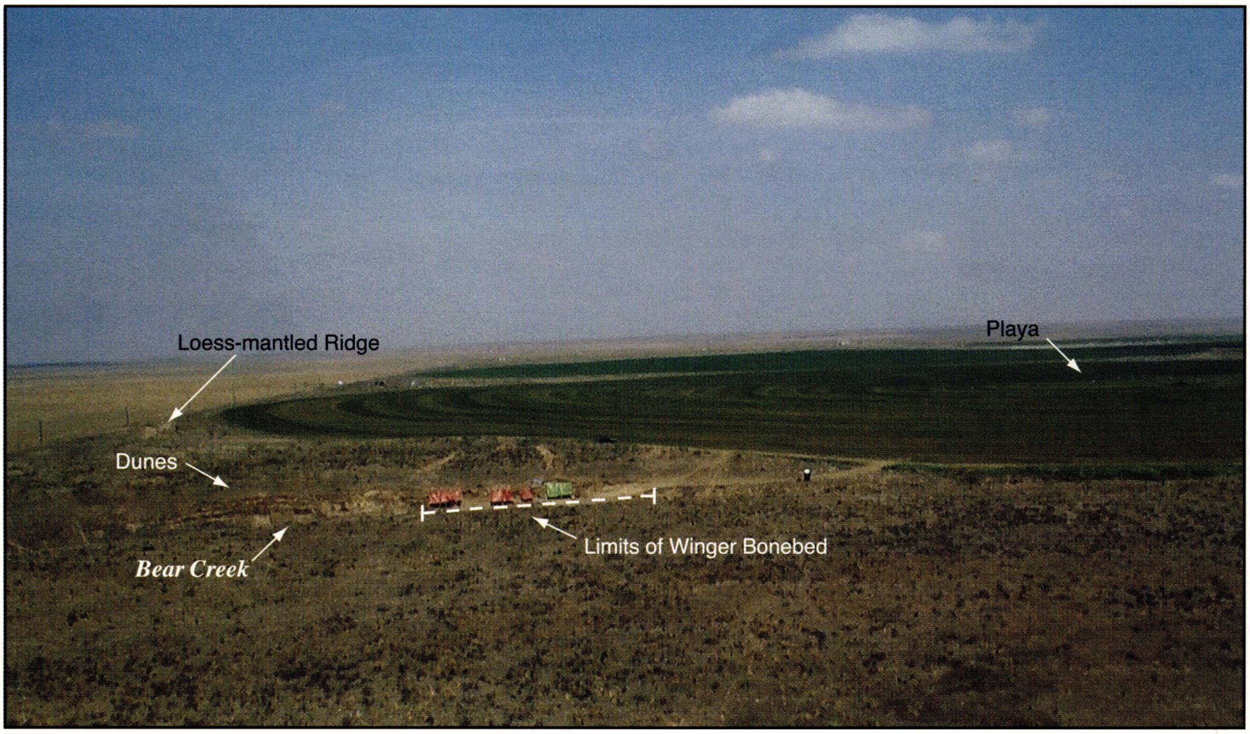

Figure 4-15. Aerial photograph showing the location of the Winger bonebed in relation to Bear Creek and the sand dunes, loess-mantled ridge, and playa. The view is to the north.

Figure 4-16. Photograph of the Winger site. The people are in the dry channel of Bear Creek, and the arrow points to the location of the bonebed in the cutbank. The view is to the west.

Figure 4-17. Photograph of the cutbank at the Winger site prior to excavation. The view is to the west-northwest.

Figure 4-18. Photograph of the section that was described at the Winger site. The stratigraphy is illustrated in fig. 4-20. The photo-scale at the bottom of section is resting on the bonebed. The view is to the north.

Figure 4-19. Photograph of the bonebed at the Winger site. The photo-scale is 20 cm long.

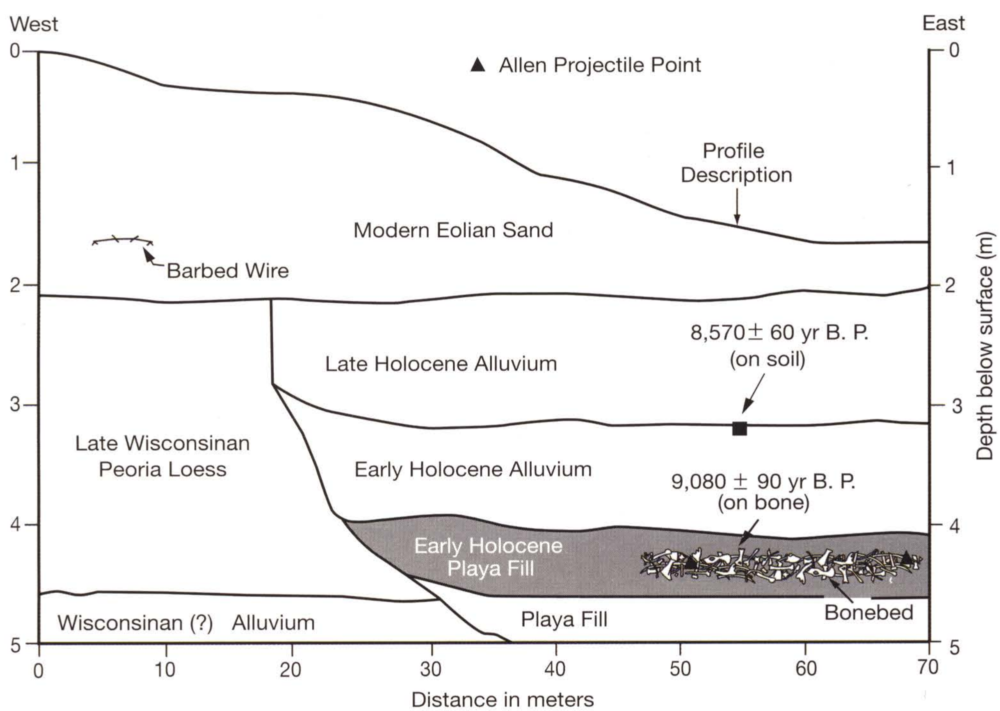

Figure 4-20. Statigraphy exposed along the length of the cutbank at the Winger site.

a soil at the top of the pond deposits, and accumulation of nearly 1 m of silty alluvium above the pond deposits. Hence, it is reasonable to assume that the bonebed was only on the land surface for a short time, and that it was not affected by a long episode of pedogeneis after it was buried. Instead, it was sealed beneath the pond sediments and overlying alluvium soon after ca. 9,000 yrs B.P. This is consistent with the excellent condition of the bone surfaces and preservation of articulated skeletons.

The soil developed in the early Holocene alluvium is overlain by 135 cm of horizontally bedded silty and sandy alluvium interbedded with fine gravel (figs. 4-20 and 4-21 and table 4-3). The numerical age of this alluvium is unknown. However, the absence of soil development in the alluvium suggests that it is late Holocene in age, and it may be modern.

The stratified alluvium is mantled by a recent sand dune. Eolian deposits composing the dune have large-scale crossbedding, with steeply dipping cross laminae attaining angles up to 32°. The thickness of the sand dune ranges from about 2.75 m at the west end of the cutbank to less than 35 cm at the east end (fig. 4-20). Barbed wire 1.7 m below the surface of the dune indicates that the dune was active during modern times.

Archeology

Faunal Remains

Investigations during June 2002 documented numerous articulated or partially articulated bison, including elements with a wide range of sizes, weights, and density. These excavations opened only small windows into the bonebed, but revealed articulated skeletons from the axial units to the distal extremities. In several cases, the limb elements were folded together rather than extended, suggesting that some animals had collapsed while standing, apparently in a playa or pond margin setting. It is possible that an embankment was present to the west and north of the bonebed that could have served as a partial containment or impediment to herd movement out of the playa. The actual method of containing or restraining the bison that enabled many to be killed while standing together remains to be determined. One possibility is that a narrow arroyo or gully at the edge of the playa produced a constriction in herd movement that was exploited by hunters.

The total number of bison represented in the bonebed is unknown, but likely represents at least several dozen animals. Our current estimate of the minimum number of

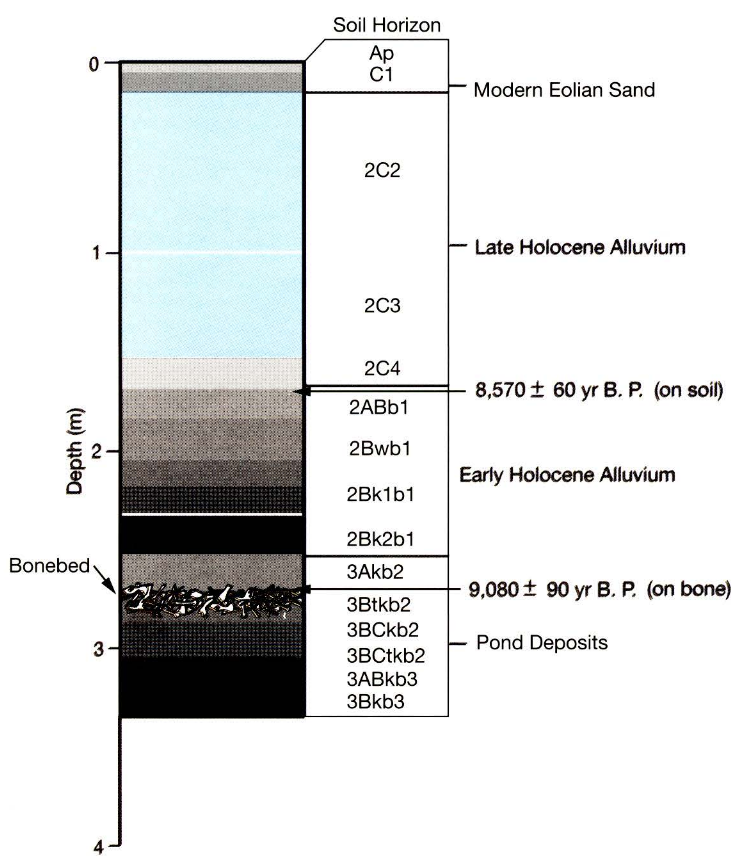

Figure 4-21. Soil-stratigraphic profile described at the Winger site. This section is shown in fig. 4-18.

TABLE 4-3. Description of the soil stratigraphy at the Winger site.

Depth (cm)	Soil Horizon	Description
		EOLIAN SAND
0–15	Ap	Pale-brown (10YR 6/3) fine sandy loam, brown (10YR 5/3) moist; very weak fine granular structure; very friable; abrupt smooth boundary.
15–35	C1	Pale-brown (10YR 6/3) fine sandy loam, brown (10YR 5/3) moist; single grain; loose; horizontal bedding; abrupt smooth boundary.
		ALLUVIUM
35–98	2C2	Stratified light-grayish-brown (10YR 6/2) silt loam, grayish-brown (10YR 5/2) moist; interbedded with brown (10YR 5/3) fine sandy loam, loamy fine sand, and very fine sand, brown (10YR 4/3) moist; massive; very friable; horizontal bedding; few krotovina 6–8 cm in diameter filled with pale-brown (10YR 6/3) fine sandy loam; abrupt wavy boundary.
98–150	2C3	Stratified light-yellowish-brown (10YR 6/4) loamy fine sand and very fine sand, yellowish-brown (10YR 5/4) moist; massive parting to single grain; few krotovina 6–8 cm in diameter filled with light-grayish-brown (10YR 6/2) silt loam; abrupt wavy boundary.
150–165	2C4	Stratified light-grayish-brown (10YR 6/2) silt loam, grayish-brown (10YR 5/2) moist; interbedded with laminae of brown (10YR 4/3) silty clay loam, dark-brown (10YR 3/3) moist; massive; very friable; common lenses of sand, granules, and fine pebbles; horizontal bedding; abrupt wavy boundary.
165–185	2ABb1	Brown (10YR 5/3) to yellowish-brown (10YR 5/4) silt loam, brown (10YR 4/3) to dark-yellowish-brown (10YR 4/4) moist; few fine distinct yellowish-brown (10YR 5/6) mottles; weak fine prismatic structure parting to moderate fine and medium granular; friable; very few fine threads of calcium carbonate; common fine and very fine pores; many worm casts and open worm burrows; gradual smooth boundary.
185–211	2Bwb1	Brown (10YR 5/3) to yellowish-brown (10YR 5/4) silt loam, brown (10YR 4/3) to dark-yellowish-brown (10YR 4/4) moist; few fine distinct yellowish-brown (10YR 5/6) mottles; weak fine prismatic structure parting to weak fine and very fine subangular blocky; hard; friable; common fine and medium pores; common worm casts and open worm burrows; gradual smooth boundary.
211–237	2Bk1b1	Pale-brown (10YR 6/3) silt loam, brown (10YR 5/3) to yellowish-brown (10YR 5/4) moist; weak fine prismatic structure parting to weak fine and very fine subangular blocky; hard; friable; common fine threads of calcium carbonate; common fine and medium pores; few worm casts and open worm burrows; gradual smooth boundary.
237–252	2Bk2b1	Brown (10YR 5/3) silt loam, brown (10YR 4/3) to dark-brown (10YR 3/3) moist; weak fine prismatic structure parting to weak fine and very fine subangular blocky; hard, friable; common fine threads of calcium carbonate; common fine and medium pores; abrupt smooth boundary.
		POND DEPOSITS
252–267	3Akb2	Dark-grayish-brown (10YR 4/2) silty clay loam, very dark grayish-brown (10YR 3/2) moist; common fine faint-yellowish-brown (10YR 5/4) and few fine distinct yellowish-brown (10YR 5/6) mottles; weak fine subangular blocky structure parting to moderate medium and coarse granular; hard, friable; common very fine and fine threads of calcium carbonate; many bison bones and bone fragments in lower 2 cm of horizon; gradual smooth boundary.
267–287	3Btkb2	Dark-brown (10YR 3/3) heavy silty clay loam, very dark grayish-brown (10YR 3/2) moist; weak fine and medium prismatic structure parting to moderate fine subangular blocky; hard, firm; common distinct, discontinuous dark-grayish-brown (10YR 4/2) clay films on ped faces; common very fine and fine threads of calcium carbonate; common fine and medium pores; many bison bones and bone fragments; gradual smooth boundary.
287–307	3BCtkb2	Brown (10YR 4/3) silty clay loam, dark-brown (10YR 3/3) moist; moderate medium and coarse prismatic structure parting to moderate fine angular blocky; hard, firm; common discontinuous laminae of

TABLE 4-3 continued.

Depth (cm)	Soil Horizon	Description
		very pale brown (10YR 7/3) and very light brownish-gray (10YR 7/2) silt loam; common distinct, discontinuous dark-grayish-brown (10YR 4/2) clay films on ped faces and in pores; common fine threads of calcium carbonate; common fine and medium pores; common bison bones and bone fragments in upper 3 cm of horizon; clear smooth boundary.
307–322	3ABkb3	Dark-grayish-brown (10YR 4/2) silty clay, very dark grayish-brown (10YR 3/2) moist; common fine prominent yellowish-red (5YR 5/8) and yellowish-brown (10YR 5/8) and common fine distinct light-olive-brown (2.5Y 5/6) mottles; weak medium prismatic structure parting to moderate coarse granular; hard, firm; many fine threads of calcium carbonate; common shiny pressure faces on peds; gradual smooth boundary.
322–340	3Bkb3	Light-olive-brown (2.5Y 5/4) silty clay, olive-brown (2.5Y 4/4) moist; many fine prominent yellowish-red (5YR 5/8), yellowish-brown (10YR 5/8) and light-olive-brown (2.5Y 5/6) mottles; weak medium prismatic structure parting to moderate fine and very fine subangular blocky; hard, firm; many films and threads to calcium carbonate; common shiny pressure faces on peds.

individuals is not less than six, but this is based on field observation in a very limited portion of the bonebed. Our estimate does not include the area excavated by Buckner, who estimated seven bison in her first trench. For most of the bonebed, skeletons of multiple bison are overlapping. Currently no evidence is known for more than one layer of bone nor for more than one kill.

The faunal assemblage consists of bison of varied age groups from calves to old adults, with most elements of mature animals being relatively small, suggesting females. A few large males are present and numerous immature animals and calves are represented. A cow and calf herd with few adult males is indicated by our available sample, and such herd composition is consistent with a winter through spring kill event(s). Seasonality assessment through dental eruption and wear patterns will be completed in the near future. With the exception of elements that have been exposed along the bank of the creek, condition of the bones is generally good, with little evidence of surface weathering, root etching, or carnivore damage. These factors suggest rapid and fairly deep burial of at least parts of the bonebed. The few crania encountered are highly fragmentary, and this is probably due to their higher weathering profile and the longer period during which they would have been exposed to weathering prior to burial (see Todd and Rapson, 1999). Long bones, flat bones, and vertebrae are typically fractured due to shrink and swell of the surrounding matrix and pressure from overlying sediments. Orientation of elements appears not to have been strongly influenced, if at all, by stream action. Inclination of most elements is roughly horizontal, and the variation is not unexpected for a bonebed in what was a frequently wet substrate.

A concentration of burned bison bone less than 1 m in diameter, including some articulated limb elements, occurred near the center of the exposure. The function of this feature has not been determined, but it probably represents a small hearth area that was used during butchering and processing activities.

Lithic Artifacts

Lithic materials recovered from the Winger site include a few flakes, tools, and projectile points. Three lithic artifacts were found in the disturbed bonebed deposits that had been moved by heavy equipment prior to our investigations. These artifacts include a large bifacial thinning flake of Alibates flint, an ovate biface of locally available Dakota quartzite that may have served as a butchering tool (D in fig. 4-22), and the corner of an Allen point manufactured from Alibates flint (B in fig. 4-22). Within the bonebed we recovered (1) a unifacial flake tool which is triangular in cross section and could have served as a naturally backed knife, (2) a complete Dakota quartzite Allen point (A in fig. 4-22), and (3) the base of an Allen point made from Alibates (C in fig. 4-22). The point base was recovered from the western end of the bonebed at the base of the bone level, and the complete point was found near the eastern end of our excavation. The complete point was recovered in situ with the tip down at an oblique angle, at the bonebed level but just beyond the margin of the dense bone concentration. These artifacts compare well with Allen points recovered from the Norton site in west-central Kansas (Hofman, 1996, p. 73–74), which yielded a radiocarbon age on bone that is essentially identical to that from Winger.

Conclusions

The association of the Winger bonebed with pond deposits underscores the significance of playas as locations for human activities through time. As focal points for water, animal, and plant resources, playas were attractive to human groups in the High Plains environment.

Preliminary results of our investigations at Winger indicate that the site is significant in several respects. First, Winger is the largest intact bison kill/butchery site recorded in Kansas and probably the largest known in the region. The length of the Winger bonebed is similar to the Jones Miller site of Hell Gap age in eastern Colorado near the northwest corner of Kansas (Stanford, 1999), and it is more extensive and less disturbed than the Norton bonebed in west-central Kansas, which also yielded Allen points (Hofman, 1996). Like Winger, the available radiocarbon date for Norton is based on bone and indicates an age of just over 9,000 yrs B.P. for the Allen complex. The Burntwood Creek bison bonebed in the northwest corner of Kansas, originally investigated in 1922 and 1923 by paleontologists from the University of Kansas (see Hill et al., 1992), may originally have been more extensive than Winger, but it is unclear how much of the site remains intact. An Allen point has also been recovered from the

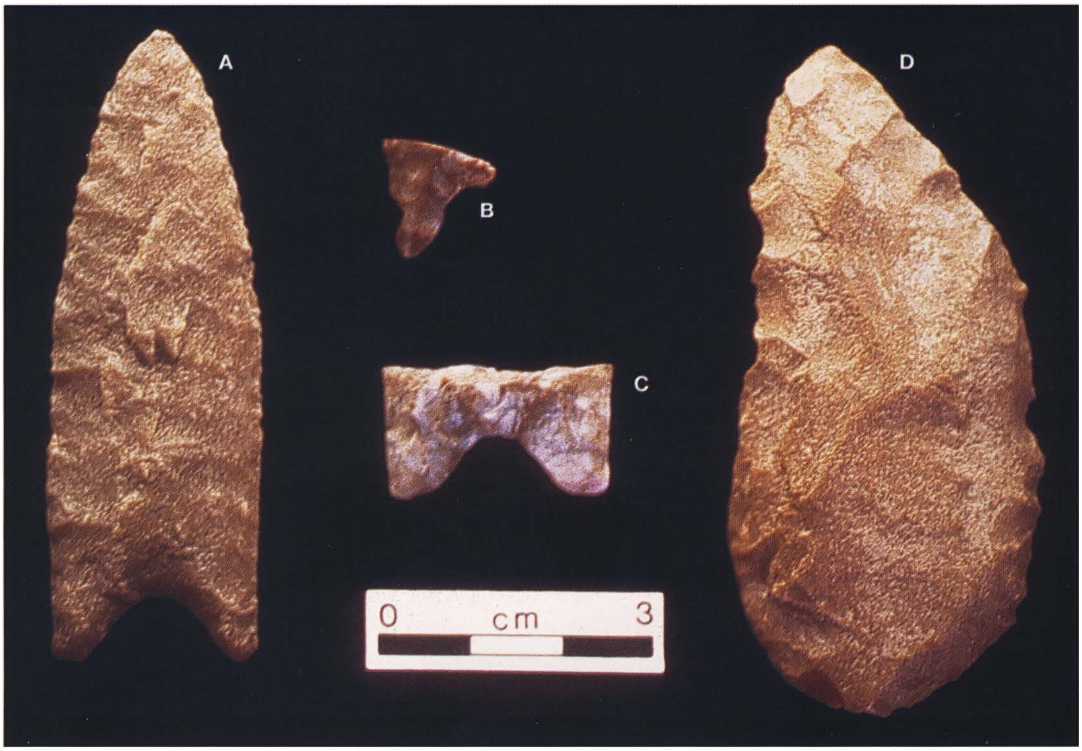

FIGURE 4-22. Artifacts recovered during the June 2002 excavations at the Winger site. **A**, Allen point made from Dakota quartzite; **B**, corner of an Allen point; **C**, base of an Allen point made from Alibates; **D**, ovate biface made from Dakota quartzite.

Burntwood Creek bonebed, and the collagen fraction from a sample of clean rib fragments yielded an AMS ^{14}C age of 9,055 ± 40 yrs B.P. (ISGS-A0728). The somewhat older 12 Mile Creek site in northwestern Kansas was a much smaller bonebed, of Folsom age, that can now be studied only through use of the existing museum collection at the University of Kansas (Hill, 1996). The Late Paleoindian Laird site in northwestern Kansas holds much potential but is much smaller than Winger. Laird yielded an unusual stemmed projectile point (Hofman and Blackmar, 1997), and it has a single bone collagen ^{14}C age of 8,495 ± 40 (CAMS-82397) (Blackmar, 2002).

Second, the bonebed at Winger has undergone minimal disturbance since the time of the bison kill. Primary agents in bone modification have been the shrink-swell action of the playa sediments and the pressure of overlying deposits. Disturbance by carnivores or trampling by hoofed animals appears to be minimal based on surface conditions and orientation of elements. Element density and size is highly variable, with even fragile elements well represented. Bone surfaces are well preserved, except for numerous fractures. Evidence of butchering, dismemberment marks, and fractures can be readily identified and studied. As research progresses, evaluation of the nature and patterns of carcass utilization at the site will be made. The evaluation of butchering and processing behavior, element representation, and seasonality will be investigated in detail. This assemblage will provide a key for comparative studies of bison ecology and evolution for the early Holocene in the Central Plains.

In addition to its size and the quality of bone preservation, the Winger site provides us with an important opportunity to document an apparently unmixed, datable Allen assemblage from deeply buried, stratified deposits. This has not previously been possible, as all known Allen sites in the region have either been (1) shallow, near-surface deposits with mixed artifact assemblages potentially representing multiple technological traditions; (2) deeply buried but disturbed; or (3) without materials for reliable chronometric dating. At Winger, then, we have the potential to add substantially to our information and understanding of the Allen technological complex and the human use of, and decision-making at, an extensive early Holocene bison kill. Information gleaned from the site will also shed light on early Holocene paleoecology and bison ethology and evolution in the Central Plains of North America.

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Trip 4, Stop 4. The Waugh Site (34HP42), Northwestern Oklahoma

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Stop 4 is at the Waugh Folsom site in northwestern Oklahoma (fig. 4-1). The Waugh site is located in Stockholm Canyon, a tributary of Buffalo Creek in Harper County (fig. 4-23). The site is situated in a narrow bottleneck near the upland divide between the Cimarron and Beaver (North Canadian) rivers. This is a common geomorphic setting for bison kill sites in the region.

Stockholm Canyon is cut into the Permian-age sandstone of the Rush Springs Formation and drains to the north. Several exposures of late Pleistocene deposits occur along the margins of the canyon. One of these, designated Area 2, is on the west side of the canyon and dates to >25,000 yrs B.P. Area 2 has mammoth and a varied snail fauna. The Folsom components at Waugh, Areas 1 and 3, are located on the east side of the canyon (fig. 4-24). Area 1 was discovered first and is the northernmost of the areas.

The Waugh site was discovered by Leland Waugh in January 1991. Excavations have been conducted with students from the University of Oklahoma, University of Kansas, Kansas State University, University of Missouri, and other institutions, and members of the Oklahoma Anthropological Society in 1991, 1992, 1993, 1996, and 2000. This research has been supported by the University of Kansas; the Oklahoma Archeological Survey; the Waugh families; the City of Buffalo, Oklahoma; the Buffalo Farmer's Co-op; and citizens of Harper County. Details pertaining to the site are in Hofman (1991a, 1991b, 1995, 1999), Hofman and Carter (1991), Hofman et al. (1992), Hill and Hofman (1997), and Widga and Hofman (2000).

Approximately 50 m^2 have been excavated in Area 1 (fig. 4-24). Area 1 is interpreted as a bison kill and butchery locality with a minimum number of six bison, including three large bulls, two adult cow bison, and one juvenile bison (fig. 4-25). Season of the kill in Area 1 (based on tooth eruption and wear evidence from one animal that is 1.8 to 1.9 yrs old) is late winter to early spring. An unknown amount of the bonebed in the southwest portion of Area 1 was lost to erosion prior to discovery of the site. The densest concentration of bones representing at least four animals occurred in this area.

The bonebed in Area 1 was covered rapidly after the kill event, with some decomposition occurring after burial as indicated by reduced sediment under bones in some of the bonebed. Most bones, except some that are bleached and weathered due to recent exposure, lack root etching

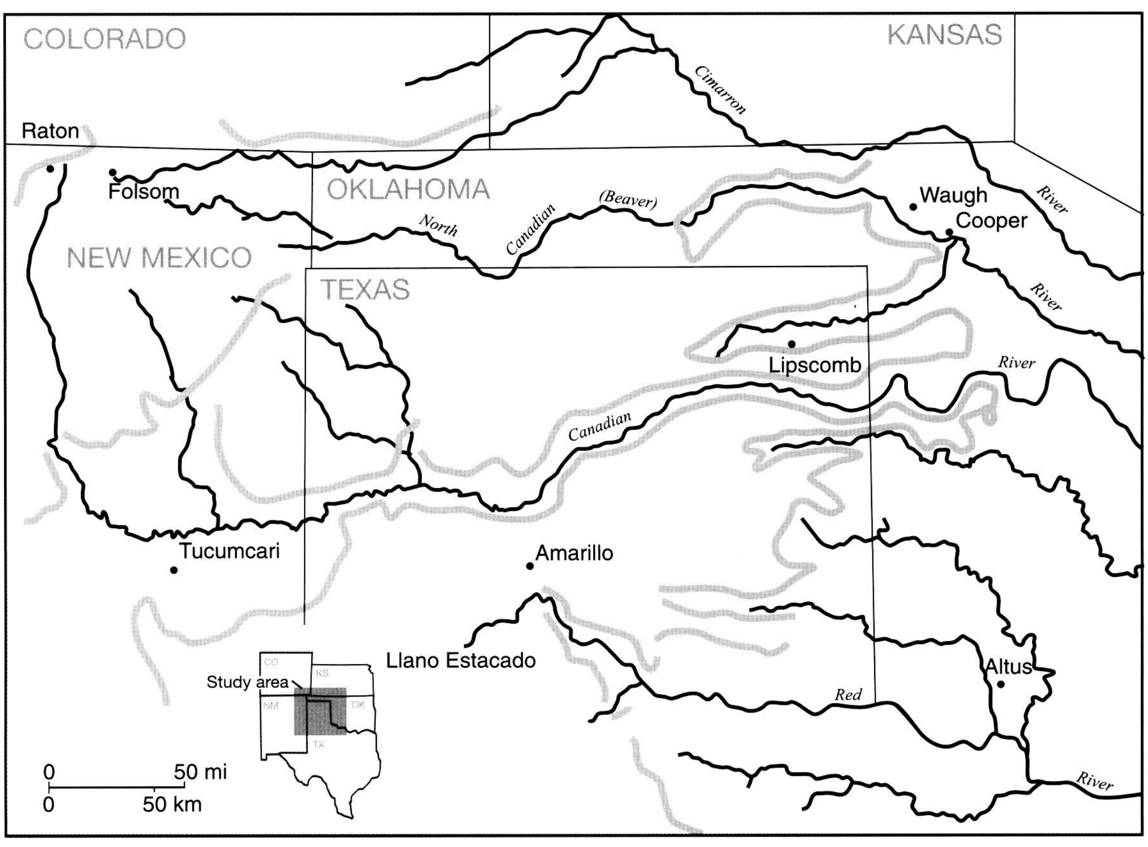

Figure 4-23. Map showing the location of the Waugh site and other Folsom sites on the Southern Plains (from Hofman, 1995).

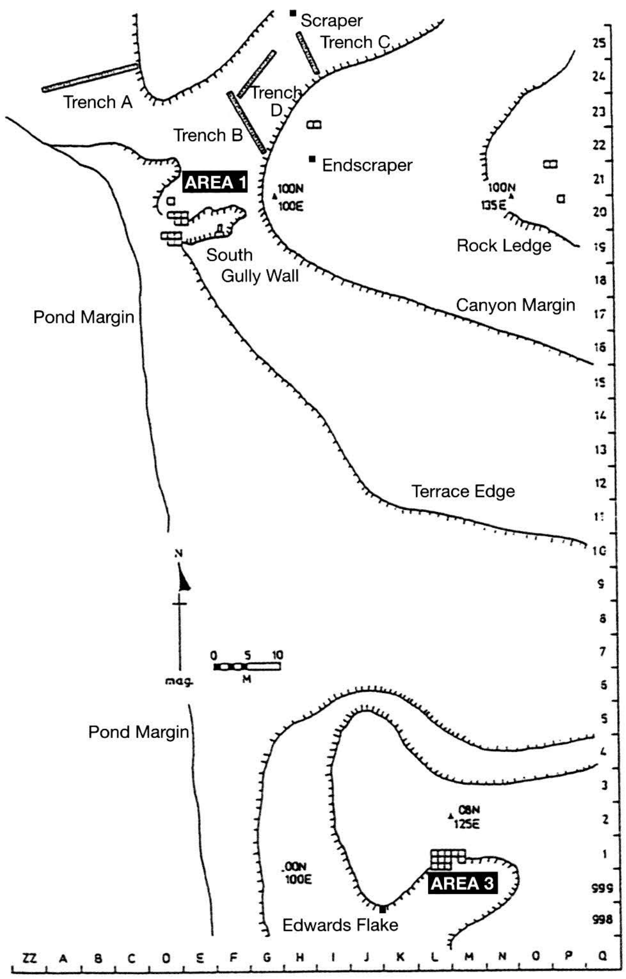

Figure 4-24. Plan view of the 1990–93 excavations at the Waugh site showing Area 1 and Area 3 (from Hofman, 1995).

typical of shallowly buried bones. The eastern portion of the bonebed is more than 3 m below the modern land surface. The western portion of the bonebed is exposed at the surface and resulted in the discovery of the site (fig. 4-26).

There is little evidence for carnivore activity (bone gnawing). Alluvial modification of bone orientation is evidenced in the northwest portion of Area 1, where a series of ribs are oriented north-south parallel to the streamflow. Also, some bones, apparently from the terrace above (east of) Area 1 have washed downslope and are contained in the colluvial fill covering the bonebed. Butchering cut marks occur on several bones. There is no evidence for marrow extraction; the long bones lack impact fractures.

Relatively few artifacts have been found in Area 1. These include one impact-damaged Folsom point of Edwards chert, one Folsom point tip with impact damage made of Alibates, an Edwards chert biface thinning flake, and a small retouch flake (fig. 4-27). The terrace above Area 1 (fig. 4-28) has yielded several lithic artifacts from what is interpreted as a processing and hide-working area. These artifacts include an endscraper of Niobrara jasper, a Folsom/Midland point preform made on a flake of Niobrara jasper, a convergent scraper of Alibates flint, and a wedge/scraper of Alibates flint (fig 4-27). These pieces have all been collected from the surface. Only 5 m^2 have been excavated on the extensive Area 1 Terrace.

Current interpretation for Area 1 is that it represents a small bison kill that occurred in an arroyo head-cut trap. These animals were butchered and processed during the late winter to early spring without evidence for use of the marrow which was probably in poor condition during that season. Carcass processing apparently emphasized meat, hides, and possibly extraction of some bones for technological needs.

Area 2 is located about 200 m south of Area 1 on the western side of the canyon. Investigations in 1991 and 1996 yielded evidence of a single mammoth. The bones are shallowly buried and stratigraphically above a late Pleistocene marsh/pond deposit. No evidence of cultural activity has so far been found in Area 2. A small fragment of mammoth long-bone was, however, found in Area 1 and a potential source for this specimen is the Area 2 mammoth.

Area 3 is located 100 m south of Area 1 and is also on the east side of Stockholm Canyon (fig. 4-24). It is at an elevation about 5.5 m higher than Area 1. This is in concordance with the canyon gradient and the slope of alluvial deposits at the time of occupation (10,400 ^{14}C yrs B.P.). The Area 3 Folsom level is about 3 m below the modern surface and is located to the northeast of a terminal Pleistocene marsh/pond which was drained by the headward erosion of Stockholm Canyon. A total of 36 m^2 have been excavated in Area 2 since it was first investigated in 1992 (fig. 4-29).

Four lithic artifacts and two flakes have been recovered from the Area 3 Folsom level. Three artifacts are made of Alibates and include a spurred end-scraper, a spurred spokeshave or concave scraper, and a large flake knife (fig. 4-27). In addition, a large chopper of Day Creek Dolomite was also found. Each of these artifacts was found near a hearth. The flakes are of Alibates and Edwards chert. A few chipped stone pieces were found on the surface of the west slope of the terrace above the Folsom level. These include a bifacial Clear Fork gouge representing a later brief use of the location.

Bones in Area 3 include remains of at least two bison, one large bull and one juvenile, in addition to small mammal and possible bird bones. The bison bones may be derived from Area 1 bison, but this is inconclusive. Some bone working activity is indicated by a cut-out section from the blade of a bison scapula laying on the Folsom occupation surface in association with the scapula.

Two distinct hearth areas are present in Area 3. These contain ash and abundant charcoal and appear to be very

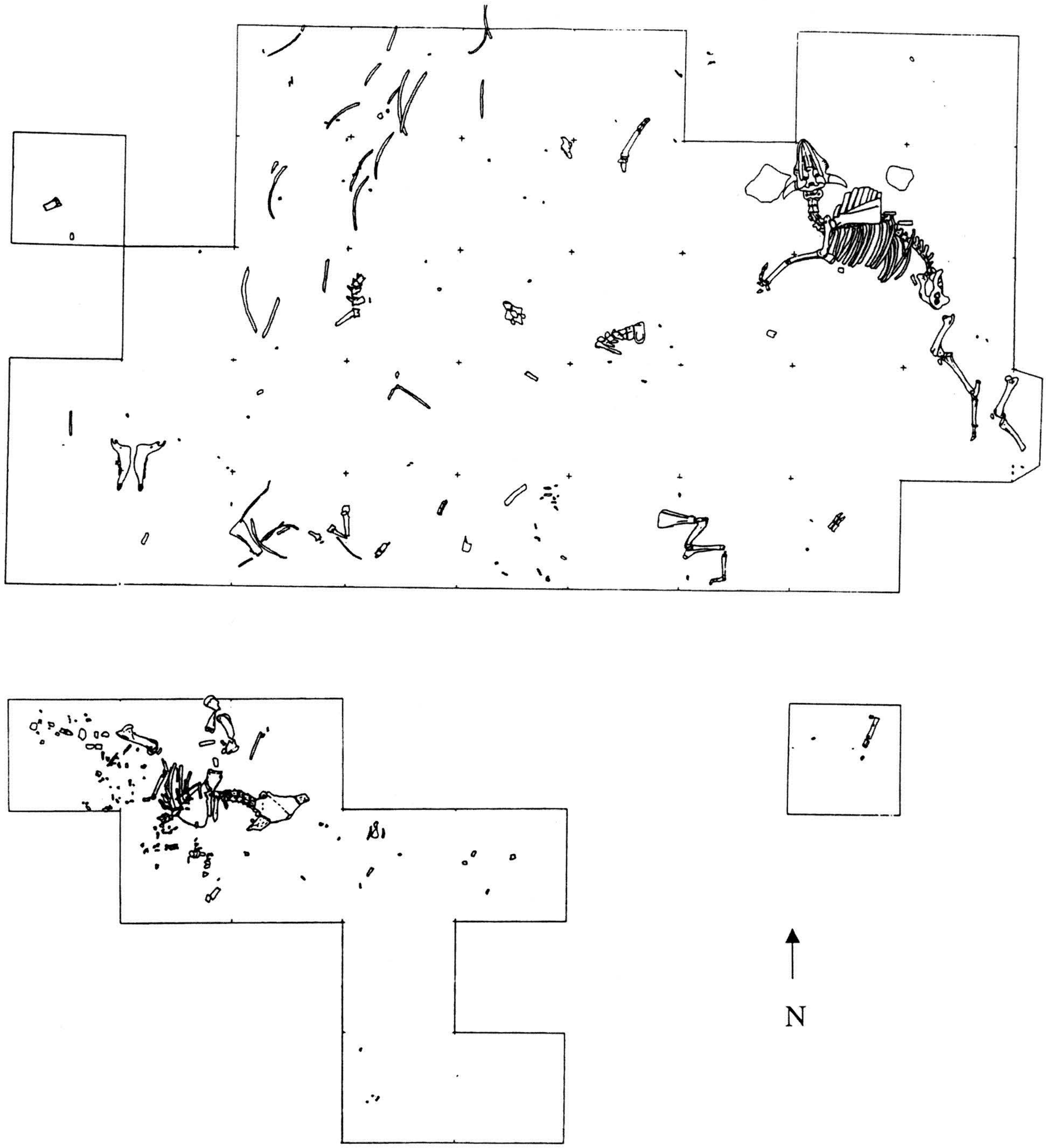

Figure 4-25. Plan view showing the distribution of bone in Area 1 at the Waugh site. Units are 1 × 1 m.

shallow surface hearths (fig. 4-30). They were apparently covered soon after the occupation by alluvial silt and sand, leaving them intact and well preserved. The hearths are about 2.5 m apart. Hearth 1 included a burned bison sacrum with a bison cervical vertebrae series nearby. Also, there were two stone tools near the hearth. Hearth 2 (fig. 4-31) had two stone tools, including the largest artifact in fig. 4-27, and the cut scapula was nearby.

A circular dark stain around Hearth 1 is similar in size and shape to a feature interpreted as a "structure" in the Folsom level at the Agate Basin site in Wyoming. A series of large sandstone rocks north of Hearth 2 may represent part of a Folsom-age structural feature in Area 3.

Two radiocarbon ages on charcoal from Hearth 1 average 10,390 $\pm$ 60 yrs B.P. Dates on the bison vertebrae from nearby are much younger at 9,160 $\pm$ 160 and 9,620 $\pm$ 70 yrs B.P. The ages determined on charcoal are assumed to be the best estimate of the time of Folsom activities in Area 3 at Waugh. At present there is no direct link (e.g., refit) between materials from Areas 1 and 3. Further dating is needed in all areas of the site and for the Pleistocene deposits below Area 2.

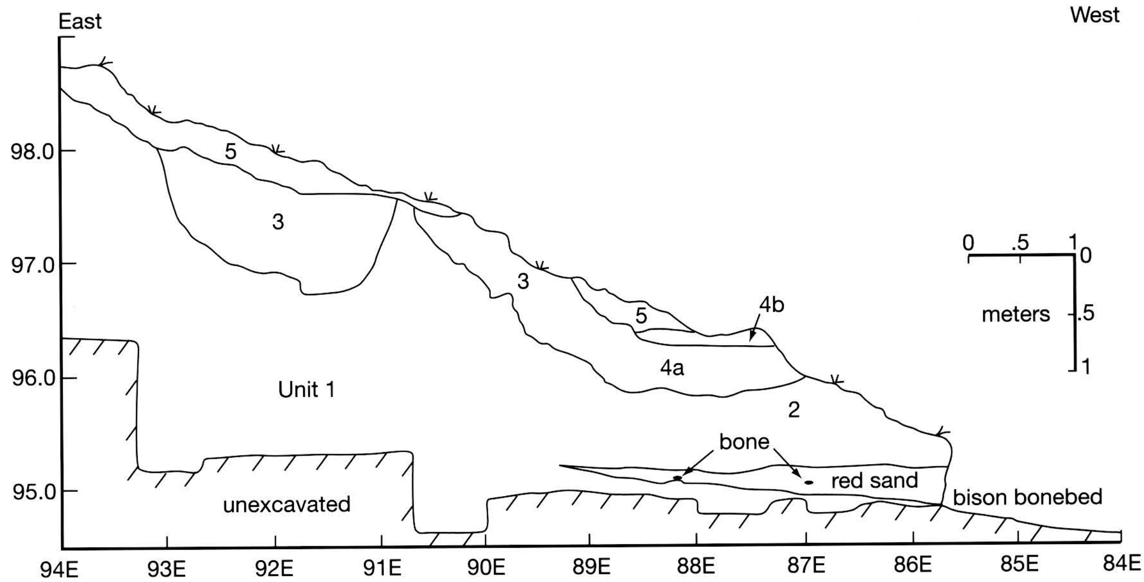

Figure 4-26. Stratigraphy of Area 1 at at the Waugh site (from Hofman and Carter, 1991, p. 32). Unit 1 is early Holocene valley fill. Units 2 through 5 are alluvial fills post-dating Unit 1.

Lithic material sources at Waugh are primarily Alibates flint from the Texas Panhandle, but also represented are Edwards chert from central Texas, and Niobrara jasper from northwestern Kansas. These same materials are represented at the nearby Folsom-age Cooper bison bonebed (see Bement, 1999).

Continued research is needed in the eroded central portion of Area 1 to further determine the number of animals and individual elements, and to search for additional artifacts. Extensive shallow testing of the Area 1 Terrace to the east of the bonebed may provide important clues to secondary processing activities.

Additional investigation of Area 3 is merited to better define the nature and limits of Folsom-age activities there. A much more extensive occupation surface is indicated by preliminary testing and erosional exposures in the area. Also, further investigation of Area 2 is needed to determine whether there is evidence for human involvement with the mammoth remains, and to glean paleoecological information from the marsh/pond deposits.

Geoarcheological, geochronological, and paleoecological studies are continuing in order to understand the site formational history and context of the Folsom activities at the Waugh site. Investigations to date have only provided an initial basis for interpreting the site's complex formation and depositional history.

Figure 4-27 (next page). Dorsal view (A) and ventral view (B) of the artifacts collected at the Waugh site. **A,** fragment of Folsom point from Area 1; **B,** Folsom projectile point from Area 1; **C,** Folsom preform from terrace above Area 1; **D,** convergent scraper from terrace above Area 1; **E,** wedge from terrace above Area 1; **F,** endscraper from terrace above Area 1; **G,** spokeshave from Area 3; **H,** endscraper from Area 3, Hearth 1; **I,** flake tool from Area 3, Hearth 2.

A

B

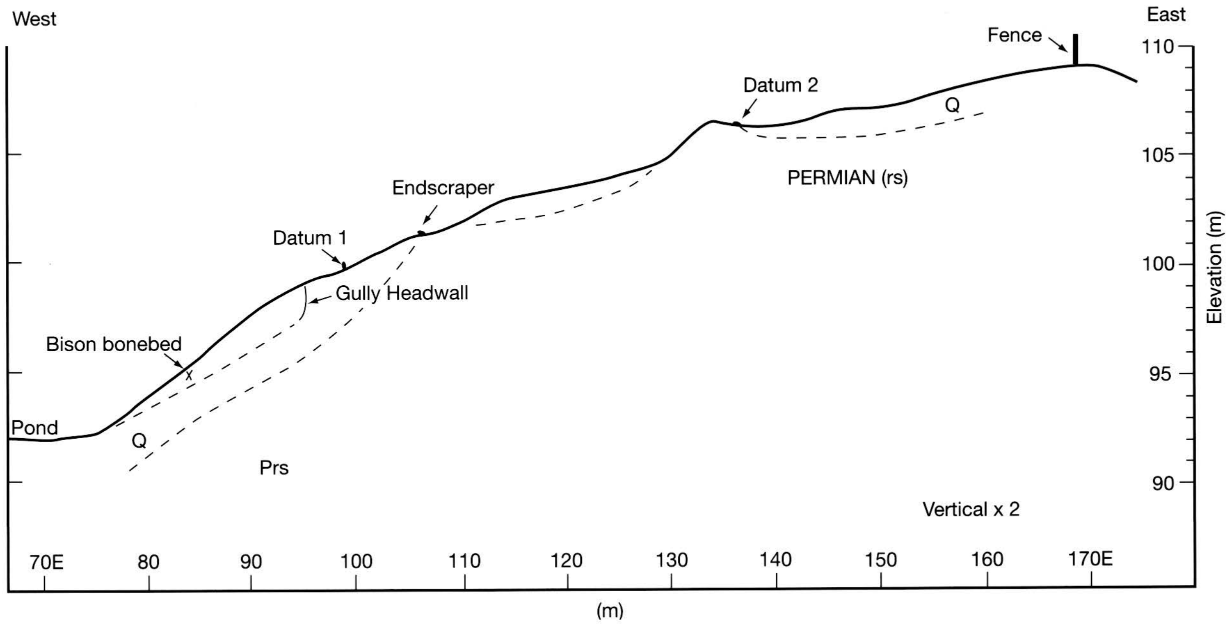

Figure 4-28. Valley cross section in Area 1 at the Waugh site.

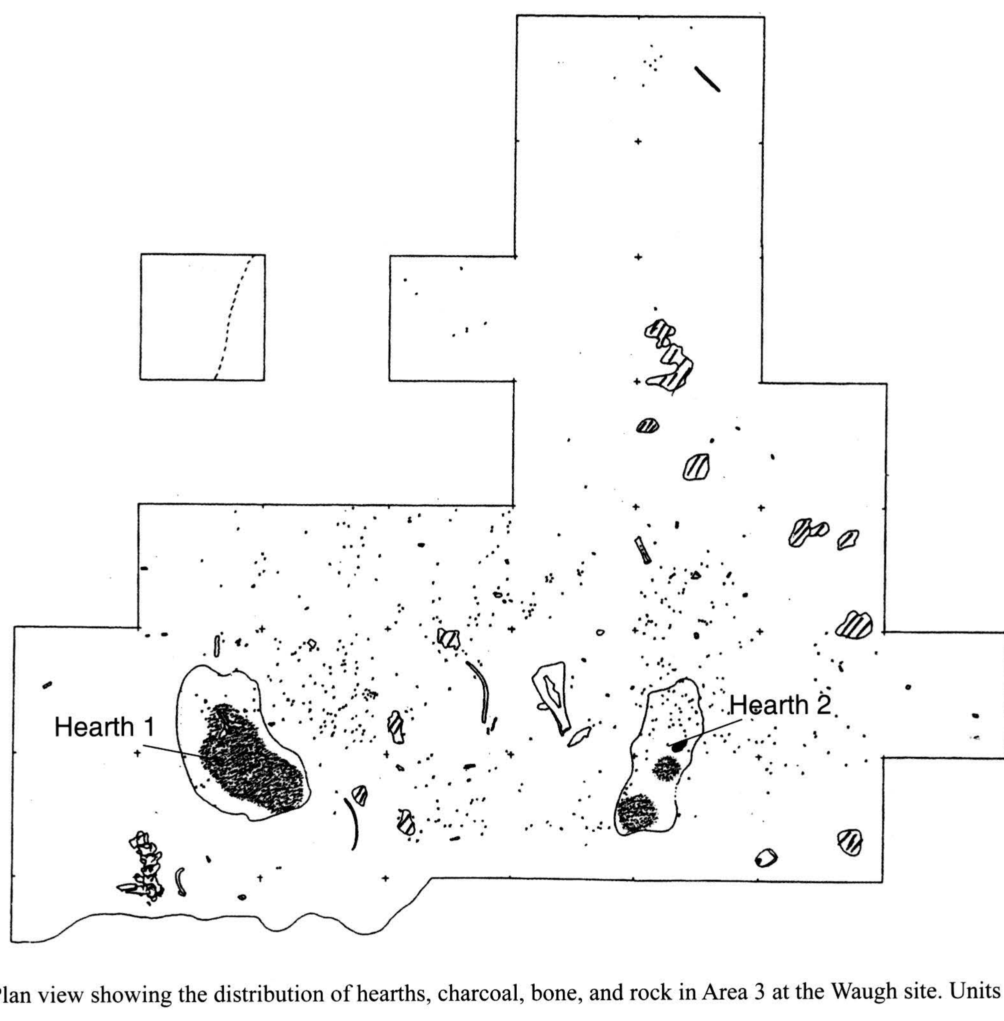

Figure 4-29. Plan view showing the distribution of hearths, charcoal, bone, and rock in Area 3 at the Waugh site. Units are 1 × 1 m.

Figure 4-30. Profile view of Hearth 2, with large flake tool in situ. Arrow is pointing north.

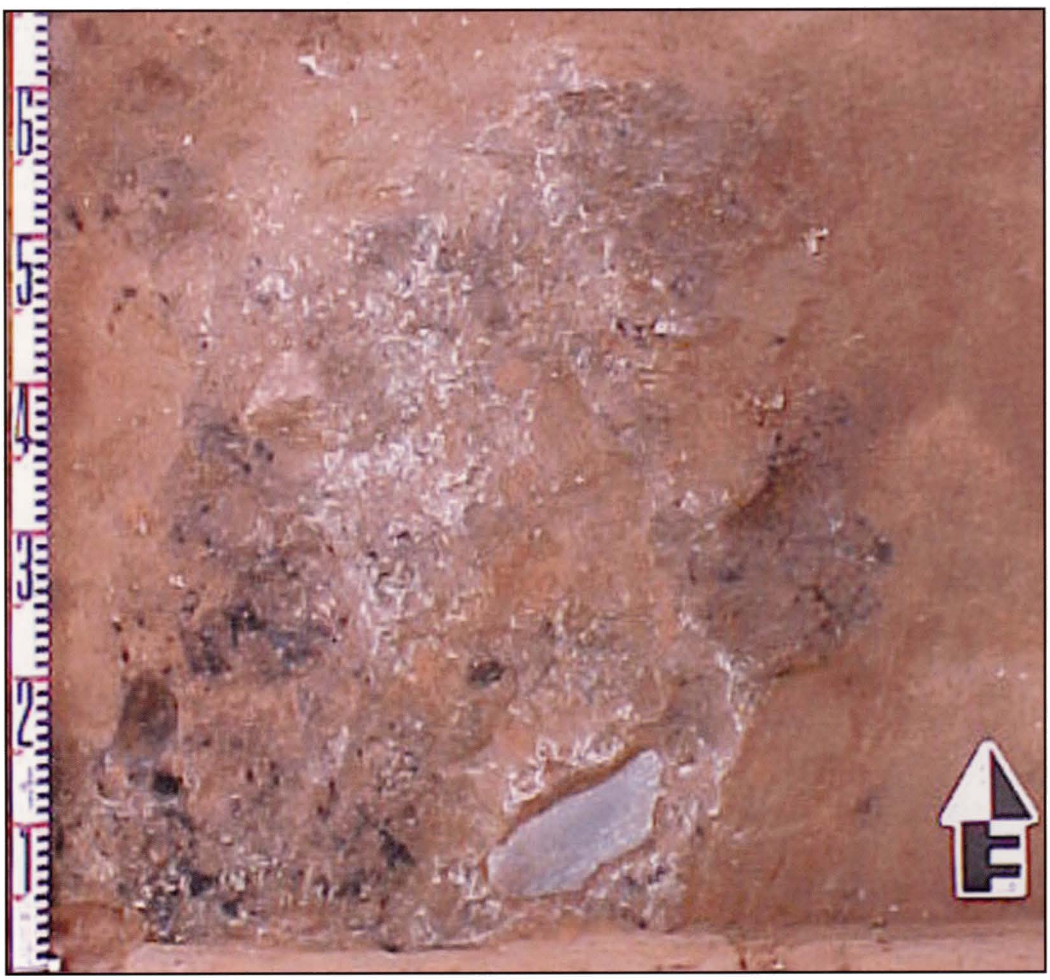

Figure 4-31. Plan view of the northern half of Hearth 2 in Area 3. The large flake tool is in situ. Arrow is pointing north.

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Trip 4, Stop 5: The Cooper Site—Multiple Folsom Bison Kills in Northwest Oklahoma[1]

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Introduction

The Cooper site (34HP45) is a stratified Folsom bison kill site in northwestern Oklahoma (Bement, 1999; Carter and Bement, 1995, 2003). The site is located on the north (left) floodplain margin of the Beaver (North Canadian) River (fig. 4-32). Bones eroding from deposits exposed along a bluff overlooking the Beaver River (fig. 4-33) were monitored for signs of cultural affiliation, and in 1993, the tip of a Folsom point was found. Excavations during the summer of 1993 and again in the summer of 1994 uncovered bison remains and Folsom artifacts within arroyo fill that had been truncated by the meandering of the Beaver River. Only a small portion of each bonebed remained intact (6 × 8 m), although additional bones and artifacts were in a secondary deposit on the floodplain immediately below the bluff edge. The bluff is composed of eolian fine-sandy loam, arroyo fill, and Permian bedrock. During arroyo aggregation, three Folsom bison kill events occurred. Soil formed in this arroyo fill during the late Holocene and was subsequently buried by eolian fine-sandy loam sediments. Detailed description of the site, its setting, excavation, and analyses have been presented elsewhere (Bement, 1999; Carter, 1999; Carter and Bement, 1995). Here, we present a more detailed description of the geoarcheological investigation as it bears on the definition and segregation of the three kill events.

FIGURE 4-32. Aerial photograph of the Cooper site vicinity at the end of the dirt-track road.

[1]*Article is excerpted with permission from Carter, B. J., and Bement, L. C., 2003, Geoarchaeology of the Cooper site, northwest Oklahoma—Evidence for multiple Folsom bison kills: Geoarchaeology—An International Journal, v. 18, no. 1, p. 115–127.*

Sediments and Soils: General Temporal and Spatial Context

The Cooper site contains stratified fine-sandy loam, eolian and sandy loam alluvial sediments, and Permian bedrock. The Folsom bison kill materials are contained in arroyo-fill alluvium that overlies Permian bedrock.

A redoximorphic depletion zone beneath each bonebed indicates that the bison remains decomposed anaerobically. Sublayers were also identified within alluvial sediments and ranged in thickness from several to more than 30 cm (table 4-4). The sublayers indicate that the sediments aggraded by storm-water runoff. The alluvial sediments are buried by fine-sandy loam eolian sediments. Coarse fragments larger than fine gravels (2.0–5.0 mm) are not present (except bison bones and artifacts). The larger clasts within the sediments are limited to the bison bones (articulated skeletons), projectile points and uniface knives, and an occasional fist-sized cobble (11 in the upper kill, one in the middle, and one in the lower) distributed throughout the bonebeds. The distribution of the lithics, including the cobbles among the articulated skeletons, precludes their having been washed into the bonebed.

Two soils are found in the depositional sequence: (1) the modern soil formed in eolian sediments and (2) a buried soil formed in alluvium that dates to 1,100 yrs B.P. Deposition of the alluvium containing the bison bonebeds had ended by the late Holocene as evidenced by this period of soil formation.

Bonebed Features and Interpretation

Each bonebed is separated by an average of 20 cm of finely stratified arroyo fill (fig. 4-34). The bonebeds do intermingle where sedimentation had not completely buried the bones of the earlier event, especially along the east and north bonebed margins between the lower and middle bonebeds and the west bonebed margin between the middle and upper bonebeds. Bonebed diagenesis includes gray redoximorphic zones (2C11b, 2C12b, and 2C13b) immediately below the bonebeds (2C12b, 2C22b, and 2C32b). These gray iron oxide depletion zones formed as organic matter from bison remains was decomposed by soil microorganisms under anaerobic conditions. Bison remains decomposed by micro-organisms represent those left in contact with the soil and not taken by bison hunters. The redoximorphic depletion areas probably formed weeks to months after each kill occurred (see Daniels and Buol, 1992; Obenhuber and Lowrance, 1991; Dobos et al., 1990; Vepraskas and Wilding, 1983; Vepraskas and Bouma, 1976).

Several key observations were made during the excavations that aided in the segregation of skeletons into their respective stratigraphic levels and provide rough estimates of time between kill events. The first and most obvious was the articulated condition of the skeletons such that intermingled extremities from different animals within the same bonebed could be sorted out by following an articulated element toward other body parts. Even when bonebeds were adjacent and overlapped, individual bison

FIGURE 4-33. View of the upper bonebed exposed near the end of the 1993 field season. Buried bones associated with the middle bonebed are visible in the foreground.

remains could be attributed to one of the three distinct bonebeds. If, for example, the leg was from a lower kill episode, then the leg would eventually dip, allowing vertical separation from the upper-level carcasses. This was most visible along the western edge of the site along the steep arroyo wall in the overlap of middle and upper kill skeletons. A second zone of overlap, this time between the lowest and middle kill skeletons, was identified near the north and east edge of the bonebeds.

A second observation in excavating the upper and middle kill overlap was that bones in the intermingled zone that were articulated to animals of the lower level displayed oval to round fractures (Bement, 1999, p. 28, 30–31). The character of the fractures was somewhere midway between green-bone breaks (possible only during or immediately following the death of the animal) and dry-bone breaks (where the fracture follows the cracks that develop years after death). The size and shape of the circular fractures were interpreted as trample breaks caused by the bison of a later kill episode stomping on the exposed bones of a previous episode. A similar series of fractures were found on some of the bones in the lowest kill episode (Bement, 1999, p. 36–37), including the painted bison skull (Bement, 1999, p. 176–178). The trample fractures provide a rough estimate of the interval between kill episodes at the site. The fact that the breaks are not green-bone fractures indicates that a period of several months to a year must have elapsed between kill episodes. Similarly, because the fractures are not dry-bone breaks, the time span is less than 10 yrs in duration (see Bonnichsen and Sorg, 1989; Fiorillo, 1989). Because these figures are based on modern conditions, to extrapolate the rate of drying and cracking back in time, we must consider environmental conditions 10,500 yrs ago. If we assume conditions were cooler and perhaps more moist than today, the rate of bone drying and cracking would be comparatively slower than today and hence, a longer interval between kill episodes would be indicated. Conversely, if conditions were warmer and drier than today, the interval between kills would tend toward a shorter time span between events.

In addition to the trampling evidence, the bones from each layer were rated for extent of weathering, following the procedures of Behrensmeyer (1978). The identification of weathered bone surfaces up to stage three (out of six) in all three bonebeds indicates the bones were exposed on the surface for varying lengths of time before burial (Bement, 1999, p. 131–133), but none was exposed to the point of extensive degradation. The upper surfaces of the uppermost bones in each bonebed exhibited the highest degree of weathering while lower bones in each bonebed displayed less evidence of weathering on both upper and lower surfaces. The end result, then, is a sequence of minimally weathered bones overlain by moderately weathered bones in each kill episode.

Carnivore modification was found on bones from all three kills, with the highest incidence occurring in the lowest kill bonebed followed by the upper kill and then the middle kill (Bement, 1999, p. 133). Although the presence of carnivore gnawing does not provide an estimate of time, it does support the stratigraphic break between kill episodes because such activity can only occur while the carcasses are exposed on the surface. Scavenging activity would cease when the bones were buried by arroyo fill or by the carcasses of a later kill event.

FIGURE 4-34. Photograph of the laminated sediments between the middle and upper bonebeds at Cooper.

TABLE 4-4. Soil-profile description of a south-facing exposure at the Cooper site.[a]

Horizon[b]	Depth (cm)	Color Moist	Structure	Texture	Consistence	Boundary	Reaction	Special Features
A	0–15	10YR4/3	1,m,GR	FSL	vfr	c,s	–	eolian sand; My, f + m roots; Fw, c, krotovinas
C	15–75	10YR6/3	M	FSL	vfr	a,s	ste	eolian sand; My, f + m roots; Fw, c, krotovinas
2Ab	75–81	7.5YR3/4	2,f,SBK	SL	fr	g,d	ste	alluvium (arroyo fill); Fw, f + m roots; My, f worm casts; Fw, c, krotovinas
2Btk1b	81–89	7.5YR4/3	2,m,SBK	SL	fr	c,s	ve	alluvium (arroyo fill); Fw, f + m roots; My, f worm casts; Fw, c, krotovinas
2Btk2b	89–118	5YR4/6	2,c,PR/2,m,	SCL	fr	c,s	ve	alluvium (arroyo fill); Fw, f + m roots; My, m $CaCO_3$ softbodies as threads and ped coatings; Fw, c, krotovinas
2Bkb	118–170	7.5YR4/3	2,c,PR/2,m, SBK	SL	fr		ve	alluvium (arroyo fill); Fw, f + roots; My, m $CaCO_3$ softbodies as threads and ped coatings; Fw, m $CaCO_3$ nodules; Fw, c, krotovinas
Upper bonebed (2C1b) consists of three subhorizons: 2C11b, 2C12b, and 2Cg1b								
2C11b	170–189	5YR4/4	M	SL	vfr	a,s	ste	alluvium (arroyo fill); Fw, f + m roots; Fw, rounded quartzite fine gravels; Fw angular Permian red-bed fine gravels; Fw, c, krotovinas; finely stratified layers of L, SL, S, and LS
2C12b	189–200	2.5Y7/4	M	–	vfr	a,s	ste	articulated bison bones; slight to moderate weathered surfaces; few scavenger marks; projectile points; uniface knives; large cobbles
2Cg1b	200–204	5YR6/2	M	SL	fr	a,s	ste	pinkish-gray, redoximorphic depletion area
Middle bonebed (2C2b) consists of three subhorizons: 2C21b, 2C22b, and 2Cg2b								
2C21b	204–224	5YR4/4	M	SL	vfr	a,s	ste	alluvium (arroyo fill); Fw, f + m roots; Fw rounded quartzite fine gravels; Fw angular Permian red-bed fine gravels; Fw, c, krotovinas; finely stratified layers of L, SL, S, and LS
2C22b	224–234	2.5Y7/4	M	–	vfr	a,s	ste	articulated bison bones; slight to moderate weathered surfaces, few scavenger marks; projectile points; uniface knives; large cobble; trampling damage to bones along western edge; painted bison skull exposed at base of bone bed
2Cg2b	234–239	5YR6/2	M	SL	fr	a,s	ste	pinkish-gray, redoximorphic depletion area
Lower bonebed (2C3b) consists of three subhorizons: 2C31b, 2C32b, and 2Cg3b								
2C31b	239–258	5YR4/4	M	SL	vfr	a,s	ste	alluvium (arroyo fill); Fw, f + m roots; Fw rounded quartzite fine gravels; Fw angular Permian red-bed fine gravels; Fw, c, krotovinas; finely stratified layers of L, SL, S, and LS

TABLE 4-4 continued.

Horizon[b]	Depth (cm)	Color Moist	Structure	Texture	Consistence	Boundary	Reaction	Special Features
2C32b	258–269	2.5Y7/4	M	–	vfr	a,s	ste	articulated bison bones; slight to moderate weathered surfaces; few scavenger marks; projectile points; uniface knives; large cobble; trampling damage to bones in central and eastern bone bed; painted bison skull protrudes into 2C22,b bone bed
2Cg3b	269–274	5YR6/2	M	SL	fr	a,s	ste	pink-gray, redoximorphic depletion area
2C42b	274–332	5YR4/4	M	GSL	vfr	a,s	ste	alluvium (arroyo fill); Fw, f + m roots; Fw, rounded quartzite fine gravels; Fw angular Permian red-bed fine gravels; Fw, c, krotovinas
2C52b	332–379	5YR4/6	M	LVCoS	vfr	a,d	ste	alluvium (arroyo fill); Fw, f + m roots; Fw, rounded quartzite fine gravels; Fw angular Permian red-bed fine gravels; Fw, c, krotovinas
2C62b	379—400	5YR4/4	M	CoSL	vfr	a,w	ve	alluvium (arroyo fill); Fw, f + m roots; Fw, rounded quartzite fine gravels; Fw, c, krotovinas
3Cr2b	400–418	2.5YR4/6	1,m,SBK	SiL	fi	g,s	ste	Fw, f, roots; Fw, black Fe + Mn oxide coatings on joints and fractures of weathered Permian shale and sandstone; Fw, c, krotovinas
3R1b	418–455	2.5YR4/4	M	VFSL	exh	g,s	e	Permian shale and sandstone (weakly consolidated); Fw, f, roots
3R2b	455–530+	2.5YR4/6	M	LVFS	exh	–	ste	Permian shale and sandstone (weakly consolidated); Fw, f, roots

[a]structure: 1 = weak, 2 = moderate, f = fine, m = medium, c = coarse, GR = granular, SBK = subangular blocky, PR = prismatic, / = parting to. Texture: F = fine, Co = coarse, V = very, S = sand, Si = silt, L = loam, C = clay, G = gravelly. Consistence: v = very, fr = friable, fi = firm, exh = extremely hard. Boundary: a = abrupt, g = gradual, c = clear, s = smooth, d = discontinuous, w = wavy. Reaction: st = strongly, v = violently, e = effervescence. Special features: My = many, Fw = few, f = fine, m = medium, c = coarse (Soil Survey Division Staff, 1993).

[b]Horizon designations follow Soil Survey Division Staff (1993) except for subdivision of 2C1b, 2C2b, and 2C3b horizons into subhorizons. These changes were made to assist in the explanation of soil material adjunct to each bonebed and clearly identify buried-soil horizons.

The age distribution of the bison (based on tooth eruption and wear patterns) in all three kills points to a late summer/early fall season of death. This situation suggests that at least one year elapsed between kill episodes. However, several years could elapse between kill events and produce the same pattern as long as each event occurred during the same season of the year.

Cultural modification of the bonebeds was limited to the span of time encompassing and immediately following each kill and consists of damage caused by the projectile points and the butchering tools. The presence of cut marks and various punctures and fractures on bones from each kill indicate animals in all three events were butchered and that similar butchering techniques were employed on all three occasions (Bement, 1999, p. 129–131, 137). The only evidence of pre-kill activity occurred prior to the middle kill with the painting of an exposed lowest kill bison skull.

Conclusion

The geoarcheological investigation of the Cooper site indicates three bison kills were buried by alluvium in the bottom of an aggrading arroyo during Folsom time. The finely stratified sandy loam sediments separating the bonebeds characterized the arroyo fill. The stratified sandy loam alluvium indicates the three bison kills are contained in sediments derived from storm-water runoff. Lack of soil formation within the alluvium, except for redoximorphic depletion areas below each bonebed and evidence from bone characteristics (based on bison tooth eruption and wear pattern) and weathering (based on the degree of weathering and trample-fracture pattern on the bones), indicate that each bison kill occurred at intervals of at least a year apart and probably less than 10 years. Clasts larger than clay, silt, sand, and fine gravel were culturally introduced into the arroyo in the form of Folsom bison kill and butchering materials. Arroyo filling stopped at least by the late Holocene as evidenced by a soil formed at the top of the alluvium and subsequently buried by eolian sediments.

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Trip 4, Stop 6: The Jake Bluff Site—Clovis Bison Hunting Adaptations at the Brink of Mammoth Extinction on the Southern Plains of North America

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Introduction

Recent excavations in gullies along the Beaver River of northwest Oklahoma have unearthed the Jake Bluff site, an apparent Clovis-age bison kill (fig. 4-35). On the southern Plains of North America, the period between 12,000 and 10,000 radiocarbon years before present (yrs B.P.) is generally considered the Pleistocene/Holocene transition and is marked by rapidly changing climatic conditions, flora and faunal associations, and cultures. Whether the extinction of the mammoth and other megafauna species was due to changing environments (Grayson, 1989; Graham and Lundelius, 1984), overkill by human predation (Martin, 1984), or a combination of both (Stanford, 1999), the result was that hunters on the southern Plains of North America had to adapt to hunting the remaining mammal species—principally bison. Our best studied aspects of Clovis hunters are directed toward mammoth hunting with only a secondary, cursory understanding of Clovis bison hunting.

A search of the literature failed to find a model of Clovis bison hunting except for the "opportunistic" taking of large animals near water sources in what is otherwise a generalist subsistence strategy (Meltzer, 1993). Conversely, five models have been proposed for Clovis mammoth hunting (Saunders 1992, p. 133–138). These include the scavenger model (Sellards, 1952), age-selective cull model (Haury et al., 1959), herd stampede model (Haynes, 1966), herd confrontation model (Stevens, 1973; Saunders, 1977, 1980), and opportunistic culling model (Frison, 1978, 1988; Frison and Todd, 1986). Among the Clovis mammoth hunting models are communal hunting scenarios where groups of hunters acted in unison toward the planned trapping and killing of mammoths. If Clovis mammoth hunting included communal hunting techniques, then the development of bison communal hunting techniques is a logical progression, especially following mammoth extinction. Bison, as well as horse and camel remains, occur at Clovis sites including postulated bison kills at Murray Springs (MNI = 6, Hemmings, 1970) Clovis (MNI = 7, Hester, 1972; MNI = minimum number of individuals), and at or near Aubrey (Ferring, 1989). Whether these kills represent opportunistic harvesting of bison near water sources or utilization of some other landform is not fully discussed. However, the presence at Jake Bluff of bison remains in a narrow, deep arroyo suggests the employment of a deliberate, planned (if not communal) utilization of this landform to trap a group of bison.

Results of Previous Excavation at Jake Bluff

During the 1994 excavation of the Folsom-age Cooper site, northwest Oklahoma (Bement, 1999), crew members discovered bison bones eroding from the side of

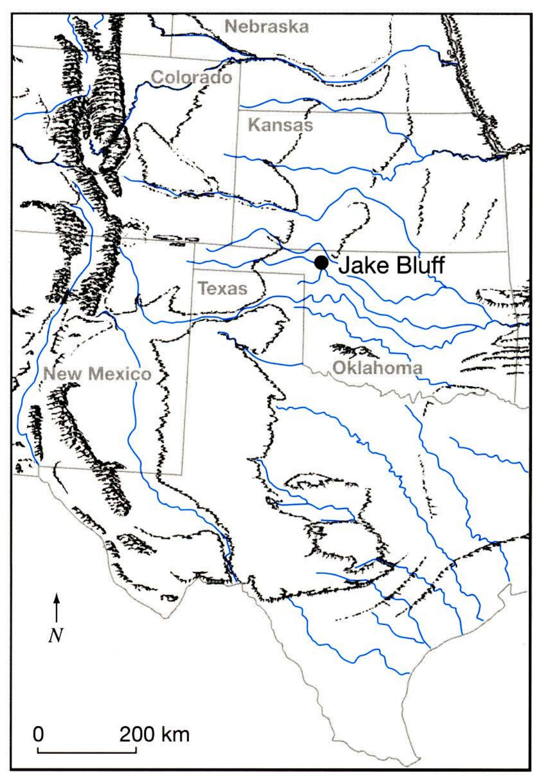

FIGURE 4-35. Location map of the Jake Bluff site in northwest Oklahoma.

Paper presented at the 2003 SAA meetings, Milwaukee, Wisconsin, April 2003.

a modern arroyo only 400 m upstream from the site. Initial assessment found only a handful of bison bones under a drape of late Holocene dune sand. No cultural material was found at that time. In 1998, a 1.5 × 0.5 m section of the bone deposit was uncovered to obtain samples for radiocarbon dating and species identification (Bement and Brosowske, 1999). The bone sample yielded a modern age that was at odds with the species identification of *Bison antiquus*, a late Pleistocene/early Holocene bison species. The stable carbon isotope ratio of -26.1 parts per mil for the dated bone suggested the sample was contaminated by modern plant material. A second sample was submitted. This sample, however, did not contain sufficient carbon for a radiocarbon assay—a condition more in line with the suspected Pleistocene age indicated by the *Bison antiquus* species identification.

The 1998 testing program determined the bones lay on top of bedrock. A chert flake found in the bison bone pile on the bedrock bench provided evidence of human association with the bison remains and the site was recorded as 34HP60, the Jake Bluff site. Subsequent monitoring of the gully exposure yielded two additional flakes and a unifacial flake knife—all of local lithic material. Coring in the area immediately east of this bone pile revealed a 3-m-deep gully. The core removed a piece of bison bone from the depths of this arroyo. Subsequent hand auguring, however, failed to extend the arroyo.

To determine the relationship of the bone deposit and adjacent buried arroyo, three weeks of excavation were conducted during the summer of 2001. This effort included the excavation of a 2 × 1.5-m area of the bone deposit and three 1 × 4-m trenches across the suspected long axis of the arroyo. The block excavation on the bone deposit uncovered the butchered legs of at least three bison. Although dominated by leg elements, the pile also contained a pair of mandibles from a juvenile (3.3-yr-old) bison bull. The tooth eruption and wear pattern identified the season of death as late summer/early fall—identical to that at the nearby Folsom-age kills at the Cooper site. The three trenches confirmed the existence of the arroyo and uncovered bison bone at the bottom, including the mandible from a 4.3-yr-old bison cow. A radiocarbon assay of 10,750 ± 40 (CAMS-79940) on a tooth from the bottom of the central trench suggested an early Folsom age for this site. Based on the radiocarbon date, presence of a buried gully containing bison remains, recovery of flakes and a flake knife, and the late summer/early fall seasonality of the kill, the Jake Bluff site was proposed to be an early Folsom bison kill similar to the three kills at the nearby Cooper site (Bement, 1999).

Summer 2002 Investigation

With funding from the National Geographic Society (Grant #7283-02), excavation was undertaken in 2002 to uncover additional areas within the arroyo, explore the eastern bedrock rim, and obtain additional datable materials and diagnostic projectile points (fig. 4-36). Information from Jake Bluff was intended to address a

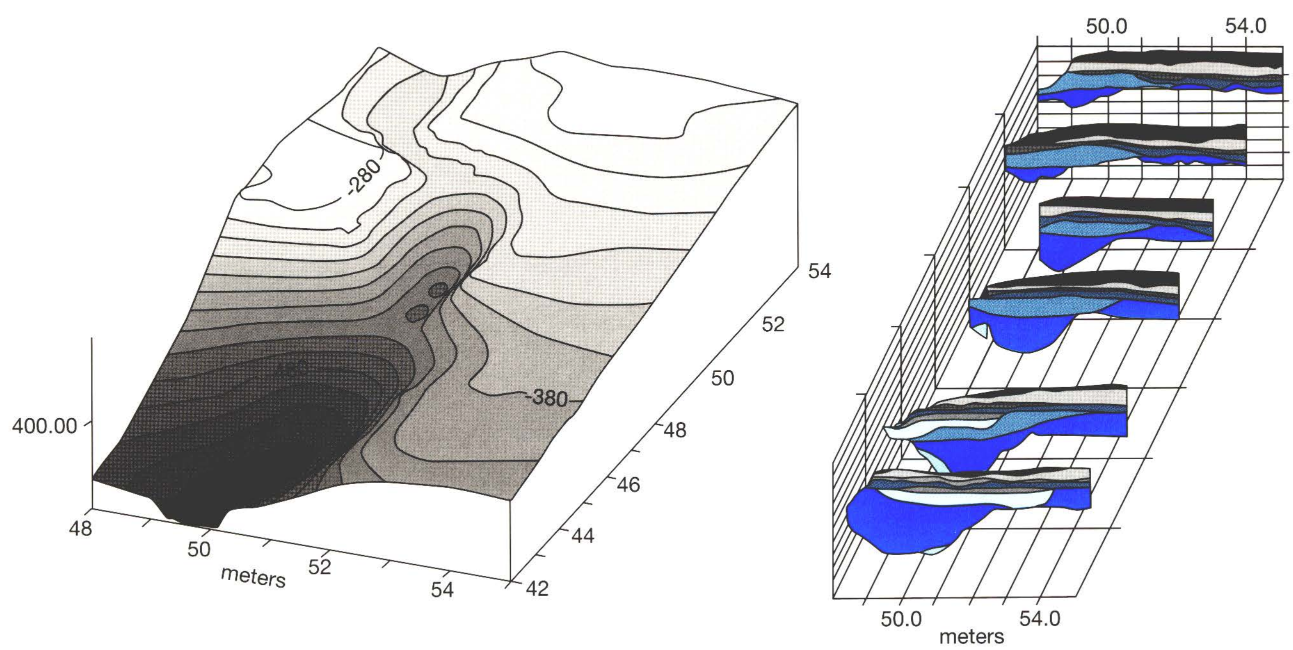

FIGURE 4-36. A 3-D and profile layout of the Jake Bluff gully.

model of Folsom bison-kill organization developed from the excavation of the Cooper site. The presence of three large bison kills in a single gully at the Cooper site raised the possibility for additional kills in other gullies along this stretch of the Beaver River. It was anticipated that the Jake Bluff site was, like the three kills at the Cooper site, a late summer/early fall, large arroyo trap kill along the eastern margin of the southern Plains.

The six-week-long 2002 excavation opened three wide trenches (2 × 6 m) that extended across and onto the eastern bedrock rim of the arroyo. Lateral extensions of these trenches brought the total excavated area to 40 m^2 including a total of 8 m along the arroyo floor.

The gully was found to extend for at least 20 m and was surprisingly narrow (fig. 4-36). The 2-m-wide gully had the appearance of an eroded animal trail incised into the soft sandstone bedrock (fig. 4-37). The excavation across the gully axis identified five zones. Zone I is Permian-age sandstone bedrock. Zone II is the gully fill containing the bison bones and artifacts. Zone III is an erosional cut-and-fill event that removed a portion of Zone II and replaced it with coarse sand. An attempt to date this event failed. The hackberry seeds selected for dating appear to be intrusive into the deposit. Zone IV may be the upward-fining of sediments from the Zone III event or an independent depositional event. Pedogenesis in Zone IV reflects stability on this landform around 500 ^{14}C yrs B.P. Zone V is eolian sand sourced from the Beaver River floodplain alluvium.

The bison remains in the arroyo brought the overall site minimum number of individuals (MNI) to 15 animals (table 4-5). Preliminary analysis suggests the herd consisted of cows, calves, and juveniles. The eastern bedrock rim failed to yield a processing pile such as that seen on the western rim. However, a scatter of broken bones, resharpening flakes, cobble-size hammer stones, and one possible anvil stone indicates processing activities were present.

The bone deposits in the bottom of the arroyo contained resharpening flakes, additional cobble-size hammer stones, and two Clovis projectile points (fig. 4-38). Both projectile points are made of Alibates agatized dolomite and display a single flute on each face (fig. 4-39). One has been reworked from a much larger point (now L = 5.13 cm, W = 2.59, T = 0.74), the other is similar to the Clovis type II defined by Hester (1972) at Blackwater Draw and is rather diminutive (L = 4.25 cm, W = 1.75, T = 0.57).

Workmanship on both specimens is similar to other southern Plains Clovis (Hofman and Wyckoff, 1991). Fine-pressure retouch on the larger point is reminiscent of the workmanship on the Clovis points from the Domebo mammoth kill in central Oklahoma (Leonhardy, 1966) and apparently also resembles specimens from the

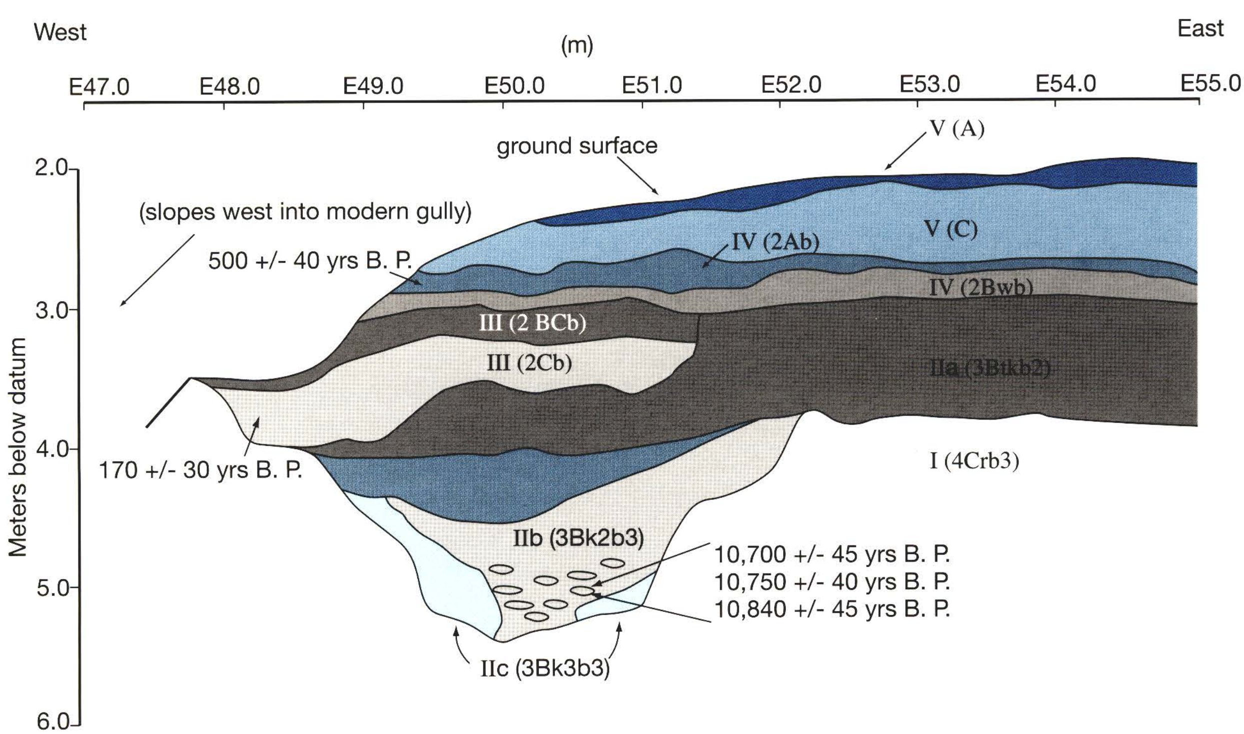

FIGURE 4-37. East-west profile across the gully.

FIGURE 4-38. Two Clovis-style projectile points were discovered in this bone pile on the gully floor.

Mockingbird Gap site in New Mexico—an assemblage believed to fall intermediate between Clovis and Folsom (Stanford, 1999). Basal characteristics place both points in the middle of the range of western Clovis points as discussed by Morrow and Morrow (2002).

The recovery of two Clovis points in direct association with bison remains in the bottom of the arroyo brought new implications to the site. First, it indicated that Jake Bluff was not a Folsom site. Second, there are few Clovis bison kills known (Hemmings, 1970; Hester, 1972). Third, the radiocarbon date of 10,750 ± 40 yrs B.P. suggested the site may post-date mammoth extinction. It was clear that the dating of Jake Bluff was of utmost importance. The evaluation of the initial radiocarbon date and submission of additional samples became the priority of analysis. The first age was obtained from XAD purified collagen from an upper molar recovered from the bottom of the arroyo in 2001. Two additional bone samples were submitted to Stafford Labs for similar processing. This time, a small, dense skull bone known as the petrous was selected from the 2002 excavations. Recent amino acid preservation analysis on a suite of bones from the Cooper site concluded that the petrous contained the least altered amino acids of all bones sampled (Stafford and Bement, n.d.). Because Jake Bluff is only 400 m from Cooper and both sites are in gullies along the same landform remnant of the Beaver River, it was felt that the petrous from Jake Bluff might also be well preserved. Subsequent amino acid analysis of two Jake Bluff petrous bones from the arroyo deposits showed them to be as well preserved as the Cooper petrous. The resultant radiocarbon ages of 10,840 ± 45 and 10,700 ± 45 yrs B.P. (CAMS-90968 and 90969, respectively) provide confirmation of the late age of this site (table 4-6). Combining these two dates with the previous Jake Bluff date yielded an average of 10,754 ± 83 yrs B.P.

Discussion

It is logical that Clovis hunters would turn to bison hunting as mammoths and other megafauna became scarce. In fact, following analysis of the Murray Springs materials, Hemmings (1970) presented the hypothesis that as Clovis hunters turned to pursuit of bison, the Clovis projectile would undergo a concomitant reduction in size (see Bonnichsen et al., 1987, p. 421, for a similar prediction). Hemmings' subsequent analysis, comparing

TABLE 4-5. Field count and minimum number of individuals at Jake Bluff site.

Element	2001 N	2001 MNI	2002 N	2002 MNI	Total N	Total MNI
Humerus	13	7	12	6	25	13
Radius	15	8	8	4	23	12
Ulna	7	4	5	3	12	6
Metacarpal	10	5	10	5	20	10
Femur	16	8	14	7	30	15
Tibia	13	7	12	6	25	13
Astragalus	9	5	7	4	16	8
Metatarsal	8	4	8	4	16	8

TABLE 4-6. Jake Bluff radiocarbon dates.

Lab ID	Material	^{14}C yrs B.P.[a]
Beta-170045	Decalcified Soil Carbon	500±40
Beta-170044	Hackberry Seeds	170±30[b]
CAMS-79940	Tooth-XAD Collagen	10,750±40
CAMS-90968	Petrous-XAD Collagen	10,840±45
CAMS-90969	Petrous-XAD Collagen	10,700±45

[a]Radiocarbon years before present
[b]Intrusive material

point size from mammoth and bison contexts, however, did not support this hypothesis, as some of the smallest points were associated with mammoth and some larger points with bison (Hemmings, 1970). It is not clear, however, that the Clovis bison associations analyzed by Hemmings post-dated mammoth extinction—generally accepted at ca. 10,800 ^{14}C yrs B.P. (Bonnichsen et al., 1987; Mead and Meltzer, 1985). In fact, Haynes (1993, p. 227) considers the mammoth kill at Murray Springs to be the last in the stratigraphic sequence. The presence in the Mockingbird Gap site assemblage of projectile points with characteristics intermediate between Clovis and Folsom types may be evidence of this shift (Stanford, 1999, p. 296). However, without the recovery of datable materials, this site cannot be positively placed in an intermediate time frame between Clovis and Folsom.

Perhaps the post-mammoth adaptation is seen first in the techniques of bison hunting and later in stone use. It is proposed here that it is the method of taking bison that changes from the opportunistic harvesting of one to several animals around water holes to include the deliberate trapping of larger numbers of animals, allowing harvesting of 10+ animals at a time. The arroyo trap technique would be a logical outcome that allows maximum numbers of animals killed with a minimum chance of risk, either in the failure of the overall hunt or risk of bodily harm to the hunters themselves (Meltzer, 1993, p. 301). The arroyo trap would not replace the small kill at pondside but rather would be added as an alternative for situations when large numbers of animals and the required arroyos were available to the appropriate number of hunters. The post-mammoth-extinction Clovis bison kills could include both small kills, possibly around watering areas, and larger kills in arroyos.

A similar dichotomy in bison-kill size has been postulated for Folsom times where small numbers (<6) of animals are taken near water sources while larger kills (>10) are found in arroyo traps (Bement, 1999). Concomitant with the dichotomy in kill size during the Folsom Period is a seasonal dichotomy where the large arroyo traps occur only during the late summer/early fall season while small kills occur throughout the year. Interestingly, the bison (MNI = 15) killed in the Jake Bluff arroyo indicate a late summer/early fall seasonality

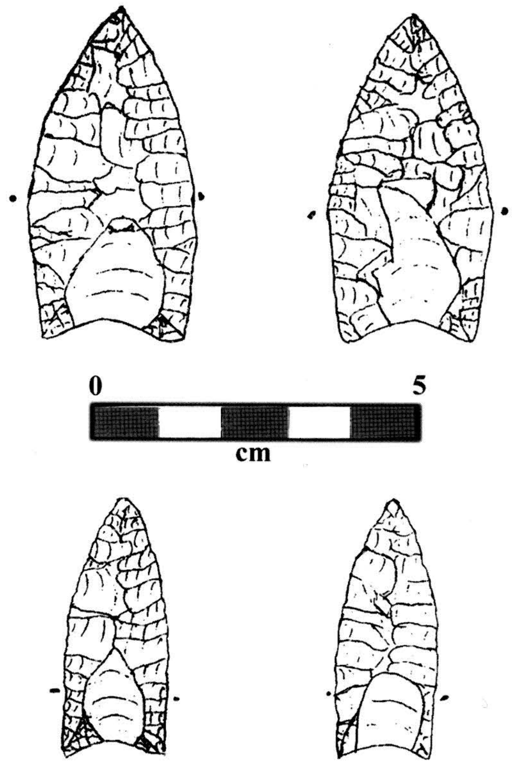

FIGURE 4-39. Drawings of both projectile points from the Jake Bluff site.

of death—the same as that seen at the three Cooper kills only 400 m away and 250 yrs later. Is Jake Bluff a Clovis precursor of the pattern seen during the Folsom Period?

Stemming from this discussion, then, is the hypothesis that the Jake Bluff site is a late-Clovis adaptation to bison hunting. As a consequence, it is predicted that possible shifts in the lithic technology toward that of later Folsom adaptations may exist in the lithic assemblage from Jake Bluff. Finally, is it a coincidence that the Jake Bluff arroyo is among a number of paleo-arroyos including the Cooper arroyo, preserved along this 3-km-long landform segment on the north side of the Beaver River? Perhaps there are additional Paleoindian bison kills awaiting discovery in nearby gullies. Continued work at the Jake Bluff site includes the analysis of lithic technology, butchering technology, landscape evolution, seasonality, and the search for a nearby camp.

In summary, the importance of this research is seen at the level of the site and what it has to say about a specific bison-hunting event and subsequent processing of the kill. Beyond this, the late Clovis age indicated by the radiocarbon dates suggests the site post-dates mammoth extinction. As such, Jake Bluff can offer insights into the development of Clovis hunting techniques and organization pertaining to bison hunting at the scale seen in subsequent Folsom bison kills. This project expands the geomorphological investigation of the Pleistocene/

Holocene transition from the ponded water and draws settings into the short gullies found on valley margins. The expansion of investigation into new landforms follows the expansion of hunting tactics employed first by Clovis and later by Folsom groups on the southern Plains of North America. Whether or not the change in location of bison kills was prompted by the extinction of mammoths or simply the fortuitous archeological investigation of this landform, this project is the first to investigate arroyo-style bison hunting techniques of Clovis hunters.

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