PALEOENVIRONMENTAL RECONSTRUCTION VIA OPAL PHYTOLITH AND CARBON ISOTOPE ANALYSIS OF LATE-WISCONSINAN LOESS:

GEOARCHAEOLOGICAL INVESTIGATIONS ON FORT RILEY, RILEY AND GEARY COUNTIES, KANSAS

by

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Kansas Geological Survey
Open-file Report 94-38

Kansas Geological Survey
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94-38

A Technical Report
for

U.S. Army Construction Engineering Research Laboratory (CERL)
P.O. Box 9005
Champaign, Illinois 61826-9005

in fulfillment of Contract No. DACA88-93-M-1036

by

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May, 1994

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Introduction

Geomorphological and geoarchaeological research on Fort Riley, Kansas for the U.S. Army Construction Engineering Research Laboratory (CERL) is being conducted to assist in the development of a dynamic paleoenvironmental model of Holocene landscape evolution. Research began with an overview study by D.L. Johnson (1992). On-site work by D. Johnson, an assistant and the contractor in June, 1993 involved further reconnaissance, subsurface exploration with a vehicular-mounted coring rig, and documentations and sampling of natural exposures. Eighteen localities, numbered 21-38, were studied, with 16 yielding sediment and soil cores. These sites represented various landscape positions within Fort Riley. Of the sites which were cored, all except 21, 26, and 33 extended to bedrock or to the residual soil developed on the bedrock. Laboratory analyses were conducted on samples from cores 21, 24, and 25 by D.L. Johnson and subcontractors (D. Johnson, 1994). Analyses included particle size determination (pipet, hydrometer, sieve, and elutriation), various wet chemical analyses (pH, cation exchange, organic matter, phosphorous, elemental ppm, base saturation), and ^{14}C dating.

After a laboratory examination of the cores collected, core 21 was selected for a pilot study to evaluate the potential for recovering a paleoenvironmental record from the carbon isotope and opal phytolith content within the sediments, which was the scope of this project. This particular core was taken from a bluff above Camp Funston and considered to be representative of the upland environment, past and present. At the University of Illinois, the 10.80-m long core was carefully described and sampled for sediment analysis by D.L. Johnson and for carbon isotope and opal phytolith analyses by W.C. Johnson and E. Diekmeyer. The latter samples were transported back to the University of Kansas, Department of Geography, where they were subsequently analyzed in the palynology laboratory. Description and sedimentary data for core 21 are provided in Appendix A.

The two parameters selected for the development of a time series of climate change have been successfully applied to paleoenvironmental reconstructions in the loess deposits of the region. W.C. Johnson, S. Bozarth, and E. Diekmeyer, as well as others associated with the University of Kansas, have studied the past environments of loess deposits ranging from the Platte River valley of western Nebraska to southcentral Kansas (e.g., Johnson, 1993a,b).

Regional Late-Quaternary Loess Stratigraphy

Classically, the Pleistocene has consisted of four glacial advances, or stages: the Nebraskan, Kansan, Illinoian, and Wisconsinan. In recent years this traditional Pleistocene chronology has been largely abandoned because of the complexity of the stratigraphic record and faulty correlations made by early workers. Within the United States, ten pre-Illinoian glaciations have been recognized, seven of which occur in the Pleistocene (three in the Pliocene). In Iowa and Nebraska alone, pre-Illinoian time is known to contain deposits from one Pliocene and five Pleistocene glaciations (Richmond and Fullerton, 1986a), and during the last 900,000 years there have been 11 episodes of glaciation in the U.S.A. (Fullerton and Richmond, 1986).

Because the glacial stages "Nebraskan" and "Kansan" and interglacial stages "Aftonian" and "Yarmouthian" appear to be no longer appropriate stratigraphic indicators in their type regions, many investigators have called for abandonment of the classical pre-Illinoian stratigrahic framework (e.g., Hallberg et al., 1980; Dort, 1987; Hallberg, 1986; Richmond and Fullerton, 1986a). The classical stage names have been used in chronostratigraphic, lithostratigraphic, pedostratigraphic, and event-stratigraphic contexts, and continued adherence to the nomenclature will likely propagate the error and confusion (Hallberg, 1986, p. 12). Boellstorff (1978a) was among the first to call for the development of a new regional stratigraphic framework for the Central Plains.

Richmond and Fullerton (1986a) and others have presented an informal scheme of geologic divisions and associated time scale for the Quaternary of the United States (Fig. 1). The Pliocene-Pleistocene boundary was set at 1.65 ma, in accordance with the estimate of 1.64 ma derived by Aguirre and Pasini (1985) on the basis of radiometric ages from the Virca section in Italy; this move has apparently resolved the problem of identifying an appropriate stratotype (Rio et al., 1991). Until recently, one age commonly proposed for the boundary was 2.4 ma (Bowen, 1978; Nilsson, 1983; Pecsi, 1985). The boundary is strictly geochronometric, i.e, without stratigraphic basis, in the United States, primarily because many have attempted to define it climatically (e.g., Beard, 1969; Lamb and Beard, 1972; Boellstorff, 1978c; Beard et al., 1982). Specifically, it can not be located stratigraphically and has no significance to the stratigraphic and chronologic sequence of glaciation, to the history of vertebrate fauna of the North American land mammal ages, or to the record of climatic or environmental change (Richmond and Fullerton, 1986b). For the central Great Plains, Boellstorff (1976) has argued, however, for a change to 1.8 or 2.0 ma on the basis of land mammal ages and a major ash marker bed, respectively. The Pleistocene was subdivided by Richmond and Fullerton (1986) into early, middle, and late, with the middle being further broken down into early middle, middle middle, and late middle Pleistocene. In most instances, numerical ages of the boundaries were arbitrarily set within the period of radiometric dating and the marine ^{18}O record. Exceptions were stratigraphic markers at the early-middle Pleistocene boundary at 788 ka (Matuyama-Brunhes polarity reversal) and the early middle-middle middle Pleistocene boundary at 610 ka (Pearlette O ash).

A very good empirical relationship exists between cold stages in the marine oxygen isotope record and documented times of glaciation (ice volume) in the United States (Richmond and Fullerton, 1986). Given the close correspondence between the glacial record and marine isotope record and the fragmentary nature of the former, it follows that the nearly continuous loessal record should be an excellent terrestrial cognate of the marine isotope record. Therefore, the climatic and chronological record assembled for the marine sequence should match the loessal record well. The generally accepted model relating climate to the loessal record indicates that periods of stability and pedogenesis are usually associated with warm interglacials, and periods of significant loess accumulation coincide with the colder glacial times (Kukla 1977, 1987). Morphologic and isotopic analyses of plant opal phytoliths from loess exposed at the Eustis ash pit in southwestern Nebraska support the model and the relationship with the marine isotope record for the Illinoian, Sangamon and early middle Wisconsin stages (Fredlund et al., 1985; Fredlund, 1993).

Pre-Illinoian Stages

Little is known of the pre-Illinoian loesses because far fewer exposures exist than of the Loveland (Illinoian) loess and certainly the Peoria (late-Wisconsinan) loess. Pedogenesis has been recognized in these early loesses, however. Zones of carbonate enrichment, occurring at about 410-360, 330-290, and 250-200 ka, within the Barton County landfill exposure were interpreted to be pedogenic in origin (Feng et al., 1994). These carbonate zones are likely analogous to the soils observed in the pre-Loveland loesses at the Eustis ash pit, Nebraska (Fredlund et al., 1985) and elsewhere in the region (Schultz and Martin, 1970; Frye and Leonard, 1951). The zones are temporally equivalent to the nonglacial, or warm marine isotope stages (Feng et al., 1994).

Fossil pollen evidence indicates that the grasslands of the Great Plains expanded during the interglacials and contracted, or perhaps disappeared, during the glacial periods (Kapp, 1965, 1970; Fredlund and Jaumann, 1987). This suggests that the carbonate enrichment was a product of grassland pedogenesis, not unlike that of today. Fredlund and others (1985) extracted the grass opal phytoliths contained within these multiple soil zones at the Eustis ash pit in southcentral Nebraska and found that the soil-forming periods were warmer and periods of dust accumulation cooler.

Illinoian Stage

All Pleistocene deposits, Illinoian and younger, were originally included in the Sanborn Formation of Kansas. Consequently, the formation was different from all others in that it included two glacial stages and substages, as well as two major unconformities, namely the Sangamon and Brady soils. The Sanborn Formation was originally considered a convenient term for designating, primarily for mapping purposes, the unconsolidated Pleistocene deposits. The Sanborn name was given to unconsolidated deposits observed by Elias (1931) in northwestern Cheyenne County, Kansas. Subsequently, Frye and Fent (1947) designated a Sanborn type section and subdivided the formation into three members, the Loveland silt, Peoria silt, and Bignell silt. Frye and Leonard (1949) recognized a fourth member underlying the Loveland silt and the Crete sand and gravel. Because of subdivision into discernable, mappable units, use of the Sanborn Formation designation has become inappropriate. Also, confusion results from the name Sanborn Group being used for a thick siltstone exposed in the Borchers Badlands, Meade County, Kansas fission-track dated to about 1.08 ma (Carter and Ward, 1991). In 1959 the Kansas Geological Survey abandoned the name Sanborn Formation, elevated its members to the rank of formation (Jewett, 1959), and later defined phases within formations (Bayne and O'Connor, 1968).

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

Shimek (1909) first identified the Loveland (as a water-lain deposit) based upon exposures along the east bluff of the Missouri River just northeast of Loveland, Iowa, where it is underlain by "Kansan" till and overlain by Wisconsinan loess. Several years later, the Loveland was identified and described in Nebraska by Lugn (1935) who identified both a valley phase (water-lain) and an eolian phase (loess). Condra and others (1950) separated the valley phase, naming it the Crete Formation. Since Leighton and Willman (1950) first designated the Loveland as an Illinoian-age loess, it has been assigned as such in both Kansas (Frye and Leonard, 1952) and Nebraska (Wayne, 1963; Reed and Dreeszen, 1965). Frye and Fent (1947) identified the Loveland Loess in Kansas on the basis of comparisons made with the type section in Iowa and deposits recognized in Nebraska. More recently, the Loveland of Nebraska has been separated into three distinct units, or members, primarily on the basis of paleosol presence (Reed and Dreeszen, 1965; Schultz, 1968; Schultz and Martin, 1970). The original type section of the Loveland was destroyed by a borrow pit for road construction, and a new type section (paratype) was designated at the end of the borrow pit area (Daniels and Handy, 1959; Bettis, 1990).

The Loveland may be described as a yellowish-brown or reddish-brown eolian silt. Red hues increase toward the top of the formation due to development of the Sangamon soil within the uppermost Loveland. The thickest accumulations occur in the northcentral part of the state: recorded thicknesses approach 15 m. A thinning in the loess occurs both southward and westward such that the distribution becomes discontinuous to the southwest. In Kansas, the Loveland is typically less than 10 m thick, but produces a very distinctive mark on the landscape via its variation in stratigraphy. It occurs on uplands and valley side slopes. As a result, the Loveland and its capping Sangamon soil are well expressed in natural exposures, particularly freshly cultivated fields.

The absolute age of Loveland loess in Kansas is largely uncertain, but recent work at sections exposed in a Geary County quarry in northeastern Kansas, the Barton and Pratt County sanitary landfills of central Kansas, and the Eustis ash pit of southwestern Nebraska provided the first absolute-age information on the Loveland beyond that carried out at the paratype section. Oviatt et al. (1988) reported TL ages of 136 ka and 130 ka for the upper part of the presumed Loveland loess exposed in an abandoned quarry near Milford in northeastern Kansas. TL age data from the Kansas sites indicates that the Loveland loess began accumulating sometime before 130 ka (Feng et al., 1994). Recently, Maat and Johnson (in press) have derived a TL age of approximately 160 ka immediately below the Sangamon soil in Loveland loess at the Eustis ash pit. Dating at the Loveland paratype section was the first attempt to employ the TL technique on loess in the Midcontinent, and results were consistent with the data obtained from loess at other localities in the North American Continent, including those in central Kansas. Four TL ages derived from the Loveland loess indicate the sediment was deposited approximately 140,000±20,000 years B.P. (Forman, 1990b).

Sangamon soil. This paleosol is strongly developed and occurs throughout the Midcontinent beneath deposits of the Wisconsinan glaciation and within deposits of the Illinoian glaciation or older deposits. The Sangamon soil has been recognized in Indiana (Hall, 1973; Ruhe et al., 1974; Ruhe and Olson, 1980), Illinois (Bushue et al., 1974; Follmer, 1979) where the type section is located (Follmer, 1978), Iowa (Simonson, 1941; Ruhe, 1956, 1969), Nebraska (Schultz and Stout, 1945; Thorpe et al., 1951) and Kansas (Frye and Leonard, 1952). In Kansas, the Sangamon soil is very well expressed, occurring throughout the state. Although the soil has received considerable attention in northeastern Kansas (Frye and A.B. Leonard, 1949, 1952; Tien, 1968; Caspall, 1970; Bayne et al., 1971; Schaetzl, 1986), it has been recognized at many localities in the state (Bayne and O'Connor, 1968) and recently studied in central Kansas (Feng et al., 1994). Historically it has been referred to as a "soil in the Sanborn formation" (Hibbard et al., 1944), the Loveland soil (Frye and Fent, 1947), and the Sangamon soil (Frye and Leonard, 1951). The color of the soil ranges from a vivid to pale reddish-brown, with a loss in color occurring westward. Regionally, the soil character varies according to parent material, local drainage and climate which prevailed at the time of pedogenesis. The soil occasionally contains sufficient clay to create a subtle bench on cultivated slopes. Schaetzl (1986) noted that the soil appears to have been a very strongly developed Ultisol or Mollisol.

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

Because of apparent time transgressiveness, the age of the Sangamon soil is not precisely known. Follmer (1983) reported a radiocarbon age of 41,700±1100 yr B.P. on plant material from the top of the Sangamon in its type area in Illinois. Forman (1990a) reported TL ages of 140±20 and 70±10 ka from loess below the Sangamon soil at two separate sites in Iowa and Illinois, and concluded the Sangamon soil is diachronous and may consist of multiple soils. Feng (1991) and Feng and others (1994) reported a TL age of about 70 ka in the lowermost part of the Sangamon soil exposed in central Kansas and associate it with marine isotope stage 3. Although Richmond and Fullerton (1986b) assigned Sangamon time to 132-122 ka (isotope substage 5e), they acknowledged reported ages (relative and absolute) ranging from early Illinoian to middle Wisconsin. Basal ages on the overlying Gilman Canyon Formation from numerous locations in Kansas and Nebraska (Johnson, 1993) provide a minimum age of about 40 ka for the Sangamon soil. Also, Forman (1990b) obtained TL and radiocarbon ages of 35-30 ka within the loess overlying the Sangamon soil at the Loveland paratype section in Iowa and the Pleasant Grove School section in Illinois.

Post-Sangamon time was one of extensive landscape instability including upland erosion, as evidenced by the partial or complete removal of the Sangamon soil. In a quarry near Woodruff in Phillips County, Kansas, the Loveland and Sangamon have been removed and the top of the Ogallala eroded. Similarly, the same units may have been stripped and channels cut into the Smoky Hill Chalk prior to deposition of the Gilman Canyon Formation. Consequently, erosional truncation of the soil may be in part responsible for the apparent diachronous character of the soil. Deposition occurred at some locations: for example, a sandy zone overlying the Sangamon soil in the Phillips County sanitary landfill suggests a dry and windy transition to the Gilman Canyon Formation above. A similar unit has been observed by the author at the Eustis ash pit in southwestern Nebraska.

Wisconsinan Stage

As the most recent glacial episode, the Wisconsin has the greatest resolution, and has been traditionally defined with five substages. It has been generally accepted that the stage began approximately 79,000 to 70,000 radiocarbon years ago. The substages, as defined in Illinois, include the Altonian, (70,000 to 28,000 B.P.), the Farmdalian, (28,000 to 22,000 B.P.), the Woodfordian, (22,000 to 12,500 B.P.), the Twocreekan (12,500 to 11,000 B.P.), and the Valderan, (11,000 to 5,000 B.P.) (Willman and Frye, 1970; Frye and Willman, 1973). Frye and Leonard (1965) refered to the Altonian, Farmdalian, and Woodfordian Substages collectively as the pre-Bradyan and to the Valderan as the post-Bradyan, the name Brady coming from the Brady soil. In recent years, time divisions within the Wisconsin have been rescaled and renamed: the stage is defined as extending from about 122 ka to 10 ka, or isotope stages and substages 5d through 2.

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

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

As expressed, the Gilman Canyon Formation is frequently overlain by .9-1.5 m of leached loess which is considered to be basal Peoria Formation. Correlative with the Gilman Canyon and overlying leached loess zone is the *Citellus* zone (a ground squirrel now recognized as the genus

Spermophilus) of Nebraska (Condra et al., 1950). The leached zone is transitional between the well developed A horizon(s) in the Gilman Canyon and the calcareous Peoria loess above, and probably reflects a sufficiently slow accumulation of Peoria loess such that pedogenesis could keep pace only partially. A.B. Leonard (1951, 1952) supports the contention that the leached, or basal zone was slowly accumulating, early Peoria loess experiencing pedogenesis through inference that gastropods were originally present, but subsequently destroyed during weathering of the loess. Above the leached zone, the rate of accumulation of Peoria loess was sufficiently rapid (c. .6mm/yr) as to preclude any soil development.

McKay (1979a,b) noted in Illinois that radiocarbon ages on organic materials from within early Wisconsinan loess ranged from 40,000 to 31,000 years B.P., and extrapolated an age of 45,000 years B.P. for the initiation of loess deposition. Further, McKay (1979a) placed the end of the Farmdalian at about 25,000 years B.P., i.e., when Woodfordian (Peoria) loess fall began. As noted above, Forman (1990a) reports TL and radiocarbon ages of 35-30 ka for the loess immediately above the Sangamon. Follmer (1983, p.141) indicated that about 5,000 years separates the first deposition of glacially-derived loess and the time of maximum Woodfordian glacial extent in Illinois at about 20,000 years ago. This 5,000-year period may coincide with that for the development of the leached zone overlying the well developed (organic-enriched) part of the Gilman Canyon. Radiocarbon ages obtained from the Roxana in the Upper Mississippi River Valley indicate the loess unit was deposited between 50 and 27 ka (Leigh and Knox, 1993).

Radiocarbon ages from the Gilman Canyon Formation range from approximately 35 ka at the base to 20 ka at the top (May and Souders,1988; Johnson et al., 1990). The basal age of 35 ka agrees well with the time set by Richmond and Fullerton (1986a) for the beginning of the late Wisconsin. While Nebraska has several dated locations forming an arcuate pattern around the eastern and southern sides of the Sand Hills, data come from only three areas in Kansas - Phillips, Barton, and Pratt counties. The ages in Kansas do show, however, good agreement from the southcentral to the northcentral part of the state and with those from Nebraska.

Given the radiocarbon time control and stratigraphic information currently available for the Gilman Canyon Formation within Kansas and Nebraska, it is clear that the associated soil(s) is a *geosol*, i.e. a laterally traceable, mappable, pedostratigraphic unit with a consistent time-stratigraphic position (Morrison, 1965; North American Commission on Stratigraphic Nomenclature, 1983, p. 865). The entire formation may be considered a geosol, but because of the possibility for the existence of two or more identifiable cumulic A horizons merged or welded together, it may ultimately be considered a *composite geosol*.

Limited paleoenvironmental data are emerging for the Gilman Canyon Formation. $\delta^{13}C$ values are a potential source of proxy data for vegetation type and hence climate (Krishnamurthy et al., 1982). When determinations are derived from the organic fractions of the soil, they reflect inputs by the plants, particularly grasses, growing on those surfaces. C_3 (cool-season) species have an average $\delta^{13}C$ composition of -27 o/oo and C_4 (warm-season/arid) species -13 o/oo, relative to the PDB standard (Deines, 1980). Terrestrial plant ecology of the Gilman Canyon Formation appears to have been characterized by primarily C_4-type grasses, or a relatively warm, possibly dry climate. From plant opal phytolith morphology (Fredlund et al., 1985; Fredlund and Jaumann, 1987) and isotope data (Fredlund, 1993), it is evident that there existed a panicoid-dominated grassland, i.e., one of moist, temperate-adapted tall grasses. These data are not inconsistent with the $\delta^{13}C$ values, since panicoid grasses are C_4 types. Further, some of the C_3-level values derived, specifically those from La Sena and Lime Creek sites, are reflecting former peaty or otherwise local, wet valley bottom environments (May, pers. comm.) which are characterized by C_3 plants meso- or hygrophytic in habit.

Interpretation of a fossil pollen assemblage from a core extracted from Cheyenne Bottoms, a large marsh in central Kansas, indicates mesic conditions in the marsh and an upland vegetation of grass and sage with scattered trees in the valley and along escarpments during the period from approximately 30 to 25 ka (Fredlund, 1991). The Farmdalian-Woodfordian transition, approximately 25-24 ka, was characterized by increased aridity. The Muscotah Marsh fossil pollen record of northeastern Kansas reflects a mosaic of deciduous forest and prairie (Gruger, 1973; Fredlund and Jaumann, 1987). Regionally, the Farmdalian grasslands were apparently found as far east as Iowa (Baker and Waln, 1985) and north to the Sand Hills region of Nebraska (Fredlund and Jaumann, 1987).

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

Ruhe (1983) notes three major features of late-Wisconsinan (Peoria) loess: it thins downwind from the source area, decreases in particle size systematically away from the source area, and is strongly time-transgressive at its base. The latter feature is unresolved and results in correlation problems. Ruhe (1969) realized a decrease in the age of the soil under the loess from 24,500 years B.P. near the Missouri River to about 19,000 years B.P. eastward across southwestern Iowa. A decrease from 25,000 to 21,000 years B.P. was noted for the base of the loess along a transect in Illinois (Kleiss and Fehrenbacher, 1973). The top of the loess also seems to be time-transgressive, ranging from about 12,500 years B.P. in Illinois (McKay, 1979b) to 14,000 years B.P. in central Iowa

(Ruhe, 1969).

In Kansas, the Peoria is a reddish, yellowish, or tan buff color, homogeneous, massive, locally fossiliferous, variably calcareous, and ranges from coarse silt and very fine sand to medium to fine silt and clay (Frye and Leonard, 1952). Thicknesses vary from in excess of 30 m adjacent to the Missouri River valley to .6 m in discontinuous patches. Any accumulation less than .6 m is presumed unrecognizable in the field because it has become incorporated into the existing surface soil. The Peoria loess typically rests conformably upon the Gilman Canyon Formation. Thickness of the Peoria is highly variable but tends to be greatest on uplands in the northwestern part of the county with a slight decrease southeastward. Exposures are common in roadcuts and may reveal as much as 9 m of the Peoria.

Despite the amount of attention given Peoria loess in Kansas, the source of the silt is not completely certain. Upon a review of the available data, Welch and Hale (1987) conclude that a single source was not likely for all loess deposits in Kansas, and that the loess was derived from a combination of three sources: glacial outwash river floodplains, present sand dune areas, and fluvial and eolian erosion of the Ogallala Formation. Research on trace element concentrations in loess currently underway by Diekmeyer (unpublished data) indicates, however, that the Platte River valley was the primary source, with secondary inputs from the major river valleys to the south (e.g., Republican, Smoky Hill, Solomon, Arkansas).

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

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

Holocene Series

Until recently, the Kansas Geological Survey recognized the Recent Stage of the Pleistocene Series (Bayne and O'Connor, 1968). The Recent is defined, as accepted by the Kansas Geological Survey, to be the last 5000 years, or the time since the end of the Valderan Substage of the Wisconsinan Stage. This nomenclature, awkward and regional in nature, has now been replaced by the use of the term "Holocene" (series status) (Baars, 1994), as was done by the U.S. Geological Survey (Cohee, 1968).

The beginning of the Holocene, about 10 ka (Hopkins, 1975), is a time of dramatic environmental change and attendant stratigraphic discontinuities. This boundary is generally considered only geochronometric, i.e., without specific stratigraphic reference, although a stratotype in Sweden has been proposed for the boundary (Morner, 1976) and has a reported age of 10,000±250 years B.P. (Fairbridge, 1983). Watson and Wright (1980) contended that major climatic and environmental change at 10 ka may be documented only on a local scale, i.e., all changes recorded in the stratigraphic record are diachronous. This notion now seems to be faulty on the regional and subcontinental scale in that research of the last decade has documented major pedogenesis at 10 ka in both alluvial and eolian/upland settings. This is the first major geosol to occur in the stratigraphic record of the region since the Gilman Canyon geosol 10,000 years earlier.

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

A typical exposure of the Brady soil would be that in Phillips County, located in west-facing roadcuts in the SW4, SW4, Sec.24, T4S, R19W. The locality was recognized by A.R. Leonard (1952, p.42-3) and revisited by Johnson (1993). It is the east face of a road cut about .8 km north of Speed, Kansas in which the Peoria loess, Brady soil, and Bignell Loess are visible. In the late 1940s and early 1950s the Loveland loess, Sangamon soil, and Gilman Canyon Formation were also exposed in the roadcut; they can yet be distinguished in a poor quality exposure around on the north face at the end of the roadcut. A profile within the road cut was excavated and sampled for radiocarbon dating in the uppermost and lowermost 5 cm of the A horizon: ages of 8,850±140 (Tx-6626) and 10,050±160 (Tx-6627) years B.P. were obtained, respectively (Johnson, 1993). The results indicate a soil forming interval lasting a minimum of 1200 radiocarbon years; our research elsewhere in the central Great Plains suggests an interval of 1500-2000 years.

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

The Brady soil has been recently dated at localities in Nebraska and Kansas. Souders and Kuzila (1990) obtained a radiocarbon age of 10,130±140 years B.P. on the Brady soil occurring within the Republican River valley of southcentral Nebraska. Sites along Harlan County Lake upstream from Naponee have yielded a number of ages, ranging from 10,550±160 to 9,020±95 years B.P., on exposures of the Brady soil (Cornwell, 1987; Johnson, 1989; Martin, 1990; Martin and Johnson, in review). Two radiocarbon ages of 9820±110 (TX-7045) and 10,550±150 (TX-7046) years B.P. have been derived from the upper and lower 5 cm, respectively, of the Brady A horizon exposed in Barton County, central Kansas (Feng, 1991).

Although it appears Brady pedogenesis occurred from about 10,500 to as recently as 8,500 years B.P., greater refinement of the Brady soil chronology is necessary, but present data clearly indicate it was a product of a major period of landscape stability at a time when widespread climatic shifts were occurring at the end of the Wisconsin. This was the first significant period of soil development since Gilman Canyon time, and represents the climate of the early Holocene. There is an isochronous alluvial soil found throughout the region which is particularly well expressed within the Kansas River basin (Johnson and Martin, 1987; Johnson and Logan, 1990). The two ages of 8274±500 (C-108a) and 9880±670 (C-471) years B.P. determined from alluvial fill (Fill 2A) at archaeological sites Ft-50 and Ft-41 on Harry Strunk Lake in southwestern Nebraska (Schultz et al., 1951; Libby, 1955) were the first radiocarbon determinations on the Brady soil. The soil, occurring in both eolian and alluvial contexts, qualifies, based upon present radiocarbon data, as a *geosol*, like the Gilman Canyon Formation soil.

Development of the Brady soil correlates well with indicators of regional climatic change. The fossil pollen record at Muscotah Marsh of northeastern Kansas indicates that spruce had essentially dissappeared from the region by about 10,500 years B.P. As this decline occurred, deciduous tree species increased until about 9,000 years B.P., the time at which grassland expansion began (Gruger, 1973). On a hemispheric scale, the abrupt decrease in atmospheric dust noted in the Greenland ice core at 10,750 years B.P. (Paterson and Hammer, 1987) reflects decreased loess deposition and possibly Brady-age pedogenesis associated with relative terrestrial stability. Further, ^{18}O levels within the same core suggest rapid warming about 10,750 years B.P., with the characteristic Holocene temperature regime being established about 9,000 years B.P.

Bignell loess. The Bignell loess was first described and named at the type locality in a bluff exposure on the south side of the Platte River valley southeast of North Platte, Nebraska (Schultz and Stout, 1945). It is typically a gray or yellow-tan, massive silt, calcareous and seldom more than 1.5 m (5 ft) thick. Although it is often somewhat less compact and more friable than the underlying Peoria, no certain identification can be made without the presence of the Brady soil. The Bignell loess does not form a continuous mantle on the Peoria; instead, it occurs as discontinuous deposits which are most prevalent and thickest adjacent to modern-day valleys, particularly the south side, and often within depressions on the Peoria surface. Feng (1991) speculates that the Bignell loess of central Kansas is relatively well weathered because it was derived from a preweathered source, the Brady soil surface, perhaps eolian and alluvial phases alike. This is consistent with the earlier interpretation derived in Nebraska that Bignell loess is at least partially comprised of re-worked Peoria loess (Condra et al., 1947, p. 33).

It appears from the radiocarbon-age determinations obtained at the type section in Nebraska and the Speed roadcut that the Bignell loess can be no older than about 8,000 years B.P. Snails collected by A.B. Leonard from the lower part of the Bignell in Doniphan County, northeastern

Kansas, produced ages of 12,500±400 (W-231) and 12,700±300 (W-233) years B.P. (Frye and Leonard, 1965). Because the shell material had absorbed an indeterminate amount of dead carbonate, Frye and others (1968) proposed an averaged age of approximately 11,000 years. Based upon the age data available for the Brady, the soil humate-derived ages are probably closest to reality.

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

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

Parameters and Methodology

Carbon Isotopes

Fractionation. Carbon isotope fractionation occurs during photosynthesis (Smith and Epstein, 1971), and fixation of carbon by plants proceeds along one of three pathways C_3 (Calvin-Benson), C_4 (Hatch-Slack), and CAM (Crassulacean). The latter is not relevant, as it is a desert adaptation which uses both photosynthetic pathways. The carbon isotopic composition ($^{13}C/^{12}C$) of the plant material is highly correlated with the type of photosynthetic pathway followed by the plant (Deines, 1980). Further, vascular plants segregate into two groups on the basis of their isotopic composition, or $\delta^{13}C$ value. C_4 plants (warm, dry-adapted plants) have an average $\delta^{13}C$ value of -14‰, while C_3 plants (cool, moist season plants) average -27 ‰ (Krishnamurthy et al., 1982).

Carbon isotope fractionation analyses have proven to be of use in determining past vegetation and associated climatic conditions. Analyses have been performed on pedogenic carbonate (Cerling, 1984; Cerling and Hays, 1986; Cerling et al., 1989; Gu et al., 1991; Cole and Monger, 1994; Humphrey and Ferring, 1994), lacustrine carbonate (Humphrey and Ferring, 1994), alluvium, loess and other sediments (Jasper and Gagosian, 1989; Lin et al., 1991; Aucour et al., 1994; Nordt et al., 1994), soil organic matter (Krishnamurthy et al., 1982; DeLaune, 1986; Schwartz et al., 1986; Guillet et al., 1988; Schwartz, 1988; Ambrose and Sikes, 1991; Cole and Monger, 1994), and opal phytoliths (Kelly et al., 1991; Fredlund, 1993).

Methods. The procedure utilized is identical to that used by our laboratory for the preparation of soil and sediment samples for ^{14}C humate dating, which renders the results compatible with those obtained in the course of age correction for the effects of isotopic fractionation. Fifty-four 250 to 300-gram samples collected from 10-cm intervals of core 21 were prepared for $\delta^{13}C$ analysis. Samples were first disaggregated in 20-quart aluminum pots filled with distilled water. They were then skimmed with a 60-mesh screen to remove floating organic debris. Next, the samples were washed through a 230-mesh screen with distilled water into a second pot in order to remove the sand and coarse silt fractions; the fine fraction remaining is assumed to contain the adhering organic carbon. The samples were then treated with concentrated HCl in order to remove the inorganic carbon contained within the carbonate. This step is particularly important because of the prevalence of limestone bedrock in the study area and the significant amounts of carbonate transported in with the loess. Following washing and oven-drying (100°C) in 4-liter beakers, the samples were pulverized and packaged. They were then sent to Geochron Laboratories for stable isotope ratio analysis.

Opal Phytoliths

Grass opal phytoliths are the best studied and can be separated into morphologic categories related to the plant photosynthetic pathways and the major subfamilies of grasses. C_4 grasses are represented by the Panicoideae and Chlorideae, whereas the C_3 grasses relate to the Pooideae. Loess deposits of the central Great Plains have been found to contain large amounts of grass opal phytoliths, which produce an interpretable climatic record (Fredlund et al., 1985; Bozarth, 1992b; Fredlund, 1993; Johnson, 1993; Johnson et al., 1993). Opal phytolith data from the Gilman Canyon Formation at the Eustis ash pit indicate the dominance of C_4 grasses throughout most of the formation, and appear to correlate nicely with the $\delta^{13}C$ values (Fig. 2). Opal phytoliths, unlike $\delta^{13}C$ values, can indicate the presence of an arboreal component (e.g., Rovner, 1971; Geis, 1973; Wilding et al., 1977) and differentiate to some extent between deciduous and coniferous trees (Bozarth, 1992a, 1993).

Opal phytoliths are generally well preserved in most sediment and can be isolated from buried sediment samples and analyzed to reconstruct the paleoenvironment for a particular area. This has been successful on a number of sediment types, including loessal sites in China (Lu et al., 1991), Nebraska (Fredlund et al., 1985; Bozarth, 1991c, 1992b; Johnson et al., 1993) and the Southern High Plains (Bozarth, 1993), as well as alluvium in Kansas (Bozarth, 1986) and the Southern High Plains (Bozarth, 1993), and swamp and upland sediment in Panama (Piperno, 1988).

Formation and stability. Growing plants absorb water containing dissolved silica through their roots. Microscopic amorphous silica bodies are subsequently produced by the precipitation of hydrated silicon dioxide ($SiO_2 \cdot nH_2O$) within the plant's cells, cell walls, and intercellular spaces. Silica bodies with characteristic shapes are called opal phytoliths. Phytolith is derived from the Greek words *phyton*, meaning plant, and *lithos*, meaning stone. Opal is the common name for hydrated silicon dioxide. Opaline bodies formed in plants without specific shapes are simply plant opal. Phytoliths form in most plants and are produced in many shapes and sizes. Many phytolith types are specific to particular groups of plants. Phytoliths are largely a "decay in place" fossil (Rovner, 1975) and represent the vegetation of a site at the time of deposition (Piperno, 1988).

The dissolution and stability of phytoliths in soil is not fully understood. Laboratory experiments demonstrate, however, that the solubility of silica is a function of temperature, particle size, pH, and the presence of a disrupted surface layer. Studies show that the solubility of amorphous silica increases linearly with temperature from 0°C. Particle size is another factor affecting stability as opal dissolution is greater with a decrease in size (Wilding et al., 1977, 1979). Pease (1967) experimentally determined that there appears to be a slight increase in phytolith solubility in the range of 5.0 to 8.5, an added increase between pH 8.5 and pH 9.0, and a large increase beginning at pH 9.0. Opal stability is also a function of the presence of certain metallic ions and sesquioxides. The adsorption of Al and Fe ions onto the surface of opal will decrease silica dissolution due to the formation of relatively insoluble silicate coatings. The presence of sesquioxides may increase dissolution of phytoliths due to the adsorption of monosilicic acid (Wilding et al., 1977).

Morphology and taxonomy. Monocotyledons, particularly the Poaceae (grasses), produce a wide variety of morphologically distinctive phytolith forms. The most taxonomically useful types of grass phytoliths are silicified short cells. Several types of trapezoidal circular, rectangular, and elliptical short cells are diagnostic of the Pooideae (Brown, 1984; Twiss, 1987; Bozarth, 1991b), a C_3 grass subfamily adapted to cool temperatures (Twiss, 1987). Saddle-shaped bodies occur most commonly in the Chloridoideae, a C_4 grass subfamily (Brown, 1984; Twiss, 1987; Mulholland and Rapp, 1992) that flourishes in areas with warm temperatures and low available soil moisture. Saddle-shaped phytoliths are similar in appearance to double-edged battle axes formed by two opposite convex edges and two opposite concave edges. However, a few saddle-shaped phytoliths have only one concave side (Brown, 1984).

Bilobate and cross-shaped phytoliths are formed in the Panicoideae, another C_4 grass subfamily (Brown, 1984; Twiss, 1987; Mulholland and Rapp, 1992) that thrives in warm temperatures and high available soil moisture (Twiss, 1987). Bilobates with indented, concave, or pointed lobes are formed only in grasses in the Panicoid subfamily. Bilobates with raised lobe edges and round or flat ends which are symmetrical in side view are also formed only in Panicoids (Bozarth, 1991c, 1992b).

Bilobate phytoliths with raised lobe edges and round ends are also formed in *Aristida* (needlegrass, wiregrass), a genus in the Chloridoid subfamily (Gould and Shaw, 1968). However, bilobates formed in *Aristida* differ from Panicoid bilobates in that the raised edges on the top (the longer part) slope down at the ends. In addition, they are asymmetrical in side view as the top is more concave than the bottom (Bozarth, 1992b). *Stipa*, a genus in the Pooid subfamily (Gould and Shaw, 1968), also produces bilobates (Bozarth, 1992b). These bilobates differ from those produced in Panicoids and *Aristida* by not having raised lobe edges. Many have a small lobe on one side in the middle. Unlike most Pooids, *Stipa* species grow in dry areas (Pohl, 1968).

There are several other types of phytoliths produced in grass in addition to short cells. Long cells are relatively large, elongate bodies with smooth or wavy edges. Bulliform cells are large keystone shaped-cells. Dendriforms are cylindrical rods of varying length that have protrusions or spines radiating from a central core. Asteriforms are roughly spherical spiky phytoliths. Trichomes are silicified prickly-hairs composed of two parts, an outer sheath and an inner core. The outer sheath dissolves soon after being deposited on the soil, while the inner core remains well preserved. The silicified stomata are taxonomically useful at various levels but are typically not well preserved. The other types are not specific to any particular subfamily but are preserved in most sediment. Piperno (1988) reported that dendriforms and asteriforms are apparently formed only in grass floral bracts.

Non-grass monocots also produce numerous taxonomically valuable phytoliths. *Cyperus* (sedge) produce distinctive phytoliths in the form of cone shaped-bodies with round wavy margins. These phytoliths are occur both singly and in multiples. Truncated cones with multiple peaks and round wavy bases are formed in *Scripus pallidus* (bulrush). Both of these phytolith types appear to be diagnostic of the genera that produce them (Bozarth, 1993).

Several types of phytoliths are produced in woody dicotyledons (deciduous shrubs and trees) and herbaceous dicotyledons (forbs and weeds). The two most common types of diagnostic dicot phytoliths are flat polyhedrons with 5-8 sides and anticlinal cells (Rovner, 1971; Wilding and Drees, 1971; Geis, 1973; Wilding et al., 1977; Bozarth, 1992a). Anticlinal cells have wavy, undulating walls with the appearance of jigsaw-puzzle pieces. Most of these polyhedral and anticlinal phytoliths consist only of silicified cell walls and are not well preserved in sediment (Wilding and Drees, 1974; Bozarth, 1992a). Other phytolith types formed only in dicots include branched elements with spiral thickening and honeycomb-shaped assemblages (Geis, 1973; Wilding and Drees, 1973, 1974; Bozarth, 1992a).

Several species of arboreal dicots produce opal spheres that range in size from 1 to 50 micrometers (Wilding and Drees, 1973, 1974). Opal spheres are also produced in conifers (Klein and Geis, 1978), but are much smaller (3 to 8 micrometers). Opaque opal spheres have been extracted from the A horizon of several forested soils in Ohio demonstrating that they are well preserved (Wilding and Drees, 1973, 1974).

Spiny spheres are formed in neotropical palms (Piperno, 1988) but have not been reported in temperate vegetation. However, the association of spiny spheres with deciduous tree phytoliths in a loessal site in Nebraska (Bozarth, 1992b) suggests that they are also formed in this, or an associated, group of plants.

Wilding and Drees (1973) reported opaque bladed forms (which appear to be opaque platelets), in white oak (*Quercus alba*). Similar particles were observed in isolates from a soil formed under deciduous forest.

Several families and genera of dicots produce phytoliths unique to those taxa. Opaque platelets with systematic perforations and certain types of segmented hairs are diagnostic of Asteraceae (the sunflower family). Platelets with irregular edges and echinate (spiny) sculpturing on one side are formed in the fruit of hackberry (*Celtis occidentalis*) and appear to be unique to that genus. These types are well preserved in sediment. Flat polyhedrons with 5-8 sides that are filled with coarse verrucae (bumps) appear to be unique to Ulmaceae (the elm family) (Bozarth, 1985, 1987b, 1992a).

Certain types of stalked verrucate phytoliths are specific to hackberry, mulberry (*Morus*), false nettle (*Boehmaria*), or nettle (*Urtica*). Elongate verrucate phytoliths with one or both ends tapering to a point are unique to *Pilea* (Bozarth, 1992a). Phytoliths with deeply scalloped surfaces of contiguous concavities are unique to *Cucurbita* (Bozarth, 1987a).

Several types of phytoliths are produced in the Pinaceae (pine family). Silicified, irregularly-shaped, polyhedral cells are the most common taxonomically useful Pinaceae phytolith. This type of phytolith is produced in *Picea rubens* (red spruce), *P. mariana* (black spruce), *P. glauca* (white spruce), *P. engelmannii* (Engleman spruce), and *Pinus banksiana* (jack pine) (Norgren, 1973; Klein and Geis, 1978; Bozarth, 1988, 1991b). Blockly polyhedra with smooth surfaces and at least eight non-parallel sides are characteristic but not diagnostic of Pinaceae as they are also produced, although relatively infrequently, in grasses (Bozarth, 1991b).

In contrast to smooth polyhedrons, polyhedrons with bordered pit impressions on the surface are unique to the Pinaceae. This type of phytolith is abundant in *Pinus* (pine), *Picea* (spruce), Douglas-fir (*Pseudotsuga*), and less commonly in *Larix* (larch), *Tsuga* (hemlock), and *Abies* (fir) (Klein and Geis, 1978). *Pseudotsuga menziesii* (Douglas-fir) needles produce distinctive, branched, silicified particles (Brydon et al., 1963). This same type of phytolith was also reported in Douglas-fir by Garber (1966) as irregular shapes with spiny processes and by Norgren (1973) as amoeboid bodies with tapering, conical protrusions. Thin plates with wavy margins on all four sides are formed in needles of *Picea glauca* (white spruce) and appear to be unique to that species. Phytoliths with spiny irregular bodies are commonly formed in needles of *Pinus banksiana* (jack pine) and appear to be diagnostic of that species (Bozarth, 1991b).

Methods. The same 54 sediment samples from Core 21 subsampled for $\delta^{13}C$ analysis were also subsampled for opal phytolith analysis. Phytoliths were isolated from the 5-gram subsamples using a procedure based on heavy-liquid (zinc bromide) flotation and centrifugation (Bozarth, 1991a). This procedure consists of five basic steps: 1) removal of carbonates with dilute hydrochloric acid; 2) the removal of colloidal organics, clays, and very fine silts by deflocculation with sodium pyrophosphate, centrifugation, and decantation through a 7-micron filter; 3) oxidation of sample to remove organics; 4) heavy-liquid flotation of phytoliths from the heavier clastic mineral fraction using zinc bromide concentrated to a specific gravity of 2.3; 5) washing and dehydration of phytoliths with butanol; and 6) dry storage in 1-dram glass vials.

A representative portion of each phytolith isolate was mounted on a microscope slide in immersion oil under a 22x40 mm cover glass and sealed with clear nail lacquer. Each isolate was then studied at 400x with a research-grade Zeiss microscope. At least 200 short cells, in addition to other phytoliths, were counted in all of the samples with adequate preservation. A complete slide was scanned and all phytoliths classified in those samples with poor preservation to a depth of 320-340 cm. Below this depth, a complete slide was scanned on every other sample as poor preservation precluded significant results. All other samples were checked for preservation.

Estimates of phytolith concentration were made using an indirect method reported by Piperno (1988). A known number of exotic spores (in this case *Lycopodium*) were added to each sample after the oxidation stage. The concentration of phytoliths (per gram) was computed as follows:

Phytolith conc. = no. of phytoliths counted x (total no. exotics added / no. exotics counted) / 5

Concentration permits an evaluation of the phytolith production, preservation, and sedimentation rate for a given sample interval.

Phytoliths were classified according to a convention that has been developed and used by other reports and publications. An extensive reference collection of plants native to the Great Plains has been developed in the palynology laboratory through field collection, research plots, solicited samples, and specimens supplied by the University of Kansas Herbarium. The phytolith reference collection consists of phytoliths extracted from complete or representative aerial portions of the following: 1) 25 species of 20 genera of 11 tribes of 6 subfamilies of the Poaceae (grass); 2) 11 species of 4 genera of 4 non-grass monocot families; 3) 65 species of 62 genera of 11 families of herbaceous dicots; 4) 20 species of 18 genera of 13 families of woody (mostly arboreal) dicots; 5) 14 species of 7 genera of 5 families of gymnosperms; and 6) 2 species of Equisetum. These reference materials include all the dominant species in the study area as reported by Kuchler (1974).

An unknown group consists primarily of phytoliths too poorly preserved to be classified any other way. There were a few other phytoliths included under this heading that were stuck under the cover glass and could not be rotated for three-dimensional viewing, thereby precluding positive taxonomic classification.

Results

The core exhibits a number of visual diagnostic changes through its length, such as color and soil horizons (Fig. 3a,b). As many as six soils may be present, including that of the surface. For example, major textural and color changes occur at 358 cm, the 2B1b/3Ab contact. Texture changes from silt above to silt loam below, and color shifts from yellowish brown (10YR 4.5/3 moist) to brown or strong brown (7.5YR 5/4 moist). Regionally, these colors are typically associated with the Peoria loess and Loveland loess, respectively. Unfortunately, no absolute time control exists for this core; two radiocarbon ages of about 19 ka come from the upper part of a buried, dark soil horizon at two different localities in the study area (see the D.L. Johnson companion report on the geomorphology). The soil zone is consistent in appearance and age with the loessal geosol of the Gilman Canyon Formation. The core was selected, however, because of the relative length and presumed high-resolution possible for this study. Based upon these observations, the initial working hypothesis was that the color break reflected the Peoria-Loveland contact.

Carbon Isotopes

The $\delta^{13}C$ values derived from the core samples are presented in Table 1 and displayed in Figure 4. Values on the abscissa are in parts per mil (‰) and decrease to the left. Therefore, the C_3 component in the vegetation increases to the left and the C_4 component to the right, with the middle (most of the field) representing a proportional mix of the grass types.

Significant variation occurs in the data, establishing some trends and different regimes with unique means. The five surface samples reflect the nature of the late-Holocene surface soil's vegetation, one dominated by the C_4 grama and buffalo grasses. The uppermost sample cuts back to the left, a common phenomenon in these data; this likely represents a historical, anthrogenic change in vegetation (increase in tree cover and/or change in grass species). There is a transition between values of the surface soil and a group of lower values of about -17‰ persisting from about 200 to 300 cm. The apparent stratigraphic and pedologic break which occurs at 358 cm is also exhibited in the carbon isotope data, i.e., values are consistently lower (more C_3) below the break than above and exhibit increased variance. From 4 to 5 meters there was an apparent period of C_3-C_4 mix, with the latter dominating; this is zone also exhibits a probable A horizon (4Ab:463-493 cm). A period of C_3 grass types seems likely at about 540 cm since it is defined by 2 to 4 assays. The return to more C_4 vegetation around 6 m is firmly established but does not associate with a buried A horizon, but rather a Bt horizon (also an indication of stability). Low values (C_3) associate with a paleosol between 7 and 8 meters (5Ab); this is a distinct isotopic period with well developed trends on both the upper and lower ends. The lowermost portion of the core reflects a mix in grass types consistent with the other two zones above at 4 to 5 meters and about 6 meters, and is markedly more C_3 than the surface meter. Overall, the groupings or zones of similar values tend to associate with paleosols, reflecting a time of stability, i.e., a lack of major change or shift in the vegetative cover.

$\delta^{13}C$ values have been determined for several samples from the Eustis ash pit and other loess exposures in the central Great Plains, mainly as a consequence of ^{14}C age correction of humate samples (Fig. 5; Johnson, 1993; Johnson et al., 1993). The time of Gilman Canyon Formation geosol development was dominated by C_4-type vegetation, comparable to the Pleistocene/Holocene Brady soil and Holocene to modern vegetation. Transitions from the Gilman Canyon geosol to periods of increased loess accumulation above and below it were dominated by C_3-type plants. During the glacial maximum (c.18 ka, Peoria time) C_3-type vegetation dominated, followed by a gradual shift to C_4-type plants, culminating in the Brady soil. Mid-Holocene (post-Altithermal?) soils reflect C_4-type grasses as well. Carbon isotope data from the Beisel-Steinle site, Russell County, are the closest available data to the study area (Fig. 6). The curve is much smoother than the composite curve and displays a pattern comparable to that of the regional composite. With the exception of the surface zone (disturbed by deep plowing and planting of a cultural C_3 grass/grain), C_4 values of the Holocene are evident, as is the Peoria loess, and Gilman Canyon Formation.

Opal Phytoliths

Results from the extraction and analysis were unanticipated, based upon extensive experience in the extraction and analysis of phytoliths from loess deposits elsewhere in the region. Both concentration and preservation of phytoliths were poor, except for the upper part of the core. From 0 to 40 cm phytolith concentration and preservation were good, and from 40 to 60 cm concentrations and preservation were adequate for interpretation. Below 60 cm, phytolith preservation and concentration were inadequate for analysis. Extraction procedures were repeatedly verified and samples rerun in order to check the concentrations. The effectiveness of the extraction procedure was attested to by good *recovery of Lycopodium* spike grains. Poor preservation typically associates with low concentrations. Further, phytoliths recovered frequently exhibited deep solutional pitting.

The degree of phytolith preservation correlates well with the pH stratigraphy of the core (Appendix A). Within the upper part of the core where preservation is good, pH values are 5.94 to about 7, but below that pH rises to a precarious level for the preservation of phytoliths (Pease, 1967). Other locations where phytolith analyses have been conducted on loess, pH, where measured, has been generally between 7 and 8. Although high pH levels may not be exclusively responsible for the poor preservation, they certainly seem to be a contributing factor.

All 54 isolates were studied microscopically, but, because of concentration and preservation, every sample was quantitatively analyzed in the upper 350 cm and every other sample below that level. Results are presented in two formats: relative frequencies of all phytolith types encountered (Fig. 7) and relative frequencies of grass short cells alone (Fig. 8). The latter presentation was produced so that the presence of the other phytolith types would not obscure the interpretation of the phytoliths diagnostic of the various grass subfamilies.

The percentage diagrams (Figs. 7,8) have been subdivided into three zones: A,B,C. Zonation is a common tool in botanical microfossil analysis, particularly palynology (fossil pollen analysis), and is rapidly being adopted in opal phytolith studies. Zonation is generally used to identify periods which are similar with regard to the assemblages of the various microfossil types. Here zonation has been used to differentiate periods of concentration and preservation as well. Zone A represents the upper part of the core which is dominated by the surface soil development and presumed late Holocene sedimentation. As both figures indicate, C_4-type grasses dominate: panicoid and chloridoid subfamilies are high relative to the C_3 pooids. These data are in agreement with the $\delta^{13}C$ values derived from the same sample. Zone B exhibits a chloridoid peak centered on about 200 cm, which agrees with the $\delta^{13}C$ data. At about 275-350 cm an increase in C_4 panicoid types occurs, showing a change in the grass composition not recognizable in the $\delta^{13}C$ data, an advantage of the phytolith analysis over the isotopic approach. Zone C represents the bulk of the core in which phytolith data are not interpretable. The break between Zones B and C coincides with the sedimentary, pedogenic, and isotopic break at about 358 cm. With such small short cell sums, the percentage bars are inflated and largely meaningless. The carbon isotope record must be relied upon exclusively for this part of the core.

Discussion

The lack of absolute time control on core 21 prevents certain correlation of the isotope and opal phytolith data to the regional loess stratigraphy. Time control is available, however, for two localities in this study (D.L. Johnson, 1994). A well developed buried soil at Site 19, the Bala Cemetary, yielded an uncorrected ^{14}C age of 19,070±280 yrs B.P. (ISGS-2622). Further, Site 20, located east of the Manhattan airport, exposed a soil developed in a Pleistocene cutoff meander buried in an alluvial fan-apron complex by reworked loess from the uplands. This site produced an uncorrected age of 19,990±450 yrs B.P. (ISGS-2623). The similarity of the ages and of the two soils indicates that they are the same soil. D.L. Johnson (1994) designated this buried soil the Riley paleosol, and correlated it to the Gilman Canyon Formation geosol. The ages are consistent with those derived elsewhere in the central Great Plains from the upper part of the Gilman Canyon Formation (Johnson, 1993).

Without absolute time control, the interpretation of the core is not clear. The core does, however, exhibit a number of buried soils (Appendix A). Buried A horizons have been identified at 252-280 (2Ab), 358-410 (3Ab), 463-493 (4Ab), 755-807 (5Ab), and 867-902cm (6Ab). Descriptions and laboratory data provided by D.L. Johnson (1994; Appendix A), regional field observations by W.C. Johnson, and sediment characteristics noted in conjunction with the isotopic and opal phytolith analyses were useful in developing a working temporal model for the stratigraphy of core 21.

At least three possible chronostratigraphic interpretations exist for the core. First, the late-Wisconsinan Peoria loess may be resting upon the Illinoian Loveland loess. This interpretation is based on the respective colors characteristic of the two loesses in the region. The Gilman Canyon Formation loess is readily apparent due to the lack of the characteristic enriched in organic matter as at sites 19 and 20. Also, the $\delta^{13}C$ values are not consistent with those derived from the Peoria elsewhere in the region. An insufficient number of assays have been made on the Loveland loess to provide a regional isotope model. The second possible interpretation is that the Peoria is overlying the Gilman Canyon Formation, but again the color is anomalous for the region, and the $\delta^{13}C$ signature is reversed from the pattern observed at all Gilman Canyon Formation localities measured to date. The third, based upon the isotopic signature and other information and observations, is that the upper 358 cm is the Holocene Bignell loess, the lower loess is Peoria exhibiting the glacial maximum from 7 to 8 m, and the Gilman Canyon Formation extends down from 7 or 8 m.

On the basis of an interpretation of data from this study and that reported by D.L. Johnson (1994), we have adopted the third possible interpretation above, i.e., the sequence represented in core 21 is, bottom to top, Gilman Canyon Formation (loess), Peoria loess, Bignell loess, and intercalated soils (Fig. 4). The Gilman Canyon Formation loess and associated geosol are represented by soil 6Ab (867-902 cm) and perhaps by soil 5Ab (755-807 cm). Based upon regional observations by W.C. Johnson (Johnson, 1993, unpublished data) and others (Reed and Dreeszen, 1965; May and Souders, 1988), the Gilman Canyon geosol often exhibits two or more episodes of pedogenesis, with the uppermost pedogenic event being least developed or expressed. Therefore, soil 5Ab may represent the late geosol development. An alternate interpretation for 5Ab is that it formed during the glacial maximum (*c.* 18ka), i.e., post-Gilman Canyon time. Recent ^{14}C dating of buried soils in the region by W.C. Johnson and students has revealed presence of a glacial maximum soil in a number of landscape settings. This alternative interpretation is supported by occurrence of the lowest $\delta^{13}C$ values in association with 5Ab; this reflects the dominance of C_3 grasses and perhaps a limited distribution of trees at this time of a cooler, more moist climate. The $\delta^{13}C$ values were elevated during 6Ab pedogenesis, reflecting a drier, warmer C_4 vegetation, which is indicative of most of Gilman Canyon time (Fig. 5). Given the probabilities, interception of a krotovina in 6Ab is consistent with the ubiquitous occurrence of ground squirrel burrows in the Gilman Canyon Formation, the so-called *Citellous zone*.

Peoria loess overlies the Gilman Canyon Formation and extends upward to soil 3Ab (358 cm). The $\delta^{13}C$ curve was used to designate this core segment as Peoria, i.e., the isotopic values increase upward as in the regional model (Fig. 5). Climate was gradually ameliorating during the late Pleistocene as the Wisconsinan ice sheets wained. Peoria loess is capped by development of the Brady soil, which like the Gilman Canyon geosol, appears to have developed under more than one period of pedogenesis. Recent paleomagnetic investigations of loess by W.C. Johnson, K. Park and M.R. Farr (e.g., Johnson et al., 1993) have strongly suggested two episodes of weathering, or pedogenesis, which is interpreted to be a function of a short-lived episode of renewed loess deposition associated with the Younger Dryas climatic event. Therefore, soil 4Ab may represent early Brady pedogenesis and 5Ab, late Brady pedogenesis. This interpretation is also consistent with the regional observation that notably greater structural development appears in the lower part of the Brady soil sequence.

The sediment above soil 3Ab (358 cm) is believed to be Bignell loess. Although Bignell loess is distributed discontinuously, the emerging model would indicate a thick accumulation on this bluff location: it is on the upland immediately adjacent to the north side of a relatively wide valley in the central Great Plains. Our magnetic data (K. Park, M.R. Farr and W.C. Johnson, unpublished data) suggest that Bignell loess is more weathered than Peoria loess, indicating a local source. An increase in sand typical of the Bignell loess occurs in the core data. Soil 2Ab would, in this interpretation, be an early Holocene soil, a common occurrence where the Bignell loess is thick. The increase in sand above this soil probably represents the Altithermal, the extended period of drought conditions widespread during the middle Holocene. The C_4 isotopic signature above 358 cm is typical of the Holocene.

Another possible interpretation (second one above) is favored by D.L. Johnson (1994), i.e., Peoria and Bignell loess extend down to the 358 cm which is the buried Riley paleosol (Gilman Canyon Formation geosol). This interpretation is based upon the laboratory and descriptive analysis of the core, particularly color of the Ab horizons. Localities where the Riley paleosol was clearly indicated, moist colors exhibited hues of 7.5YR and 10YR with chromas greater than 3. In core 21 the soil at 358-410 cm is 7.5YR with a chroma greater than 3, like the documented Riley paleosols.

Conclusions

Through a combination of two relatively new investigative techniques, carbon isotope and opal phytolith analyses, past vegetation and associated climate can be reconstructed with a relatively high degree of accuracy. The application of these two techniques to loess samples collected from core 21 produced an interpretable carbon isotope curve and partially interpretable phytolith data. The following conclusions can be made based upon these analyses:

(1) The core contains a recognizable and interpretable $\delta^{13}C$ signal. In the absence of absolute time control on the core, we can only speculate on the paleoenvironment represented in the data. An educated guess based upon a regional composite $\delta^{13}C$ curve, vertical distribution and degree of development of buried soils, and particle size distribution is that the upper 358 cm represents the Holocene (Bignell loess), 358-755(867) cm the Peoria loess with the capping Brady soil, and below 755(867) cm the Gilman Canyon Formation. Deposits predating the Gilman Canyon Formation may have been eroded earlier or were never deposited. The former is most likely due to the bluff location of the drill core site.

(2) Poor opal phytolith recovery was unanticipated. The data extracted and interpreted were, however, consistent with the isotopic data. The reason for low concentration and poor preservation of phytoliths is uncertain, but, given our successes elsewhere in the region, must relate to site-specific conditions. Specifically, samples from some loess exposures have contained up to 60% biogenic opal, most of which was phytoliths. The site condition(s) responsible for the dearth of phytoliths most likely related to the limestone bedrock of the bluff site: contamination of the incoming loess by locally-derived limestone fragments or upward migration of carbonate-rich water would have elevated the pH above levels that can be tolerated by phytoliths. The solutional pitting attests to the deterioration, i.e., phytoliths were produced in large quantities but did not survive due to dissolution, except within the uppermost part of the core. If the Gilman Canyon Formation had been eroded from the site prior to Peoria loess deposition, limestone would likely have been exposed as a source of contamination. Mechanical degradation of the phytoliths would have occurred if the lower loess unit had been reworked, e.g., colluvium, alluvium. No evidence for such mechanical loss exists; no fragments of crushed or fractured phytoliths appeared in any of the isolates. Also, the carbon isotope data would have formed a chaotic, uninterpretable curve.

Recommendations

We recommend that further studies be undertaken at Fort Riley in order to develop a dynamic paleoenvironmental model of Late Quaternary landscape evolution and to provide the prehistoric cultural resource information base required for development of a meaningful temporal and spatial model of cultural history in the study area. The field and laboratory strategies should be modified to insure the creation of a chronologically controlled paleoenvironmental data base/time series. This should include further study to determine if opal phytolith analysis has potential in the study area.

Specifically, it is recommended that:

(1) W.C. Johnson and D.L. Johnson together conduct an on-site evaluation of drill core site 21 and use the observations to develop criteria for the selection of potentially viable phytolith and isotope study sites. These sites should then be cored to reveal the stratigraphy and bedrock situation. After narrowing site selection on the basis of exploratory drilling, a sample-quality core should be extracted or, if the loess thickness is less than 5 meters, the site should be backhoe trenched to increase sampling ease and stratigraphic control. As much of the late-Quaternary stratigraphjic column should be present as possible, at least from the Gilman Canyon Formation up; this would vastly reduce the likelihood of problems such as chronological control and limestone contamination.

(2) Adequate time control is necessary for the core or exposure being analyzed for phytoliths and carbon isotopes. Backhoe trenches and freshened natural exposures would provide the opportunity to collect the volume of loess required for humate ^{14}C dating and increase the probability of discovering charcoal for dating. Samples for dating should be collected from the Riley paleosol (Gilman Canyon Formation geosol) and the overlying loess.

(3) Select and sample a number of sites and spot analyze the phytolith content in order to insure selection of the best possible locality(ies) for development of the high-resolution paleoenviromental data base.

(4) If it is realized through spot analyses that core 21 was anomolous and phytolith analysis will produce interpretable results, two to three sites should be sampled at close interval and analyzed for the phytolith content.

(5) Undertake a close interval $\delta^{13}C$ analysis of two or three sites (coincident with any phytolith analysis).

The information gathered herein is useful in of itself and as a guide for future research in the study area. We have learned more about the potential variability in opal phytolith content of loess and how to deal with it. The carbon isotope record extracted from the core appears viable and consistent with the regional record, if the temporal assumptions are correct.

Table 1. Δ[13]C values from core 21, Fort Riley, Kansas.

Sample number	cm below surface	$\delta^{13}C$
1	1070	-21.6
2	1050	-20.7
3	1030	-20.0
4	1010	-18.5
5	990	-20.7
6	970	-18.5
7	950	-20.2
8	930	-21.0
9	910	-20.2
10	890	-20.0
11	870	-21.0
12	850	-20.3
13	830	-21.6
14	810	-23.0
15	790	-23.4
16	770	-25.5
17	750	-23.3
18	730	-24.6
19	710	-24.0
20	690	-22.5
21	670	-21.8
22	650	-21.3
23	630	-20.3
24	610	-19.1
25	590	-20.8
26	570	-19.6
27	550	-20.8
28	530	-24.1
29	510	-22.2
30	490	-20.7
31	470	-19.0
32	450	-20.3
33	430	-20.1
34	410	-21.3
35	390	-18.4
36	370	-18.2
37	350	-16.8
38	330	-17.0
39	310	-17.1
40	290	-17.8
41	270	-18.1
42	250	-18.0
43	230	-17.9
44	210	-17.7
45	190	-18.3
46	170	-17.8
47	150	-17.1
48	130	-16.2
49	110	-16.7
50	90	-14.6
51	70	-13.9
52	50	-13.3
53	30	-12.8
54	10	-14.6

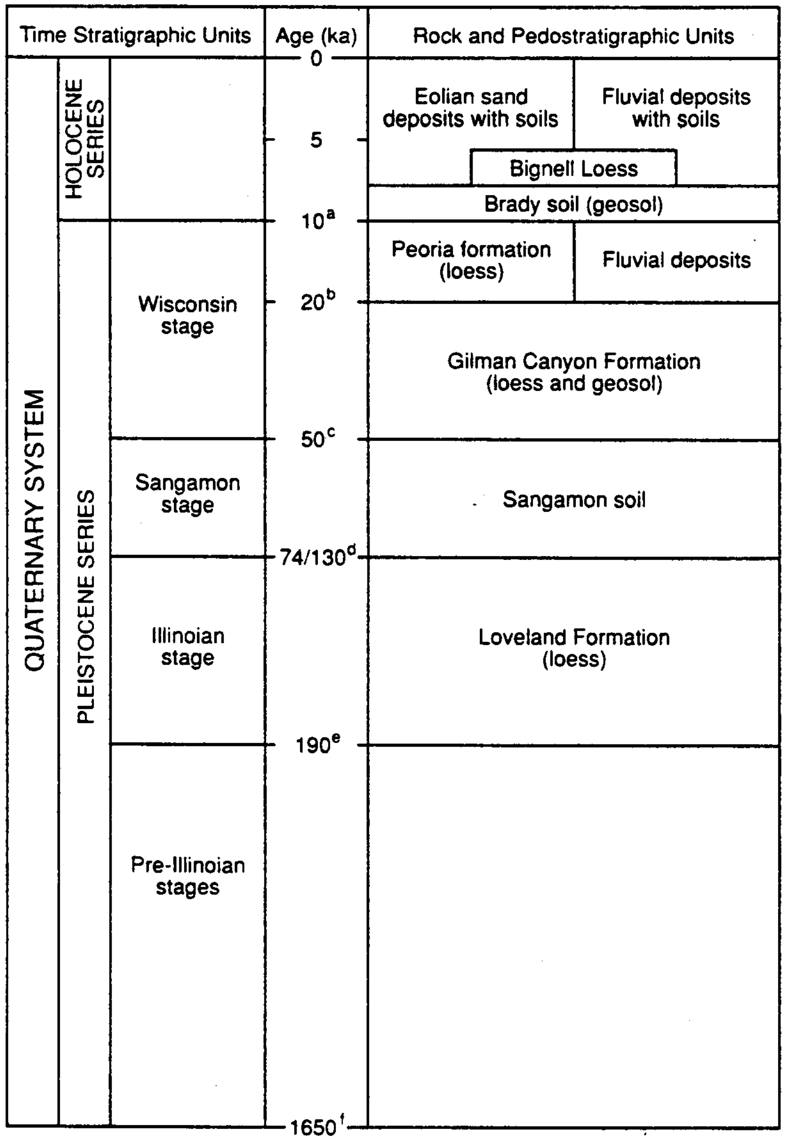


Figure 1. Late Quaternary stratigraphic succession for the nonglaciated part of Kansas. Notes: (a) Pleistocene-Holocene boundary after Hopkins, 1975. (b) The uppermost Gilman Canyon Formation ^{14}C dates at about 20 ka (Johnson, 1993). (c) The age of 50 ka is an estimate of the beginning of Gilman Canyon time, not the ending of the Sangamon; the boundary is an unconformity. (d) The age of 74 ka is used because of TL ages obtained from below the Sangamon by Forman et al.(1992) in Illinois and by Feng (1991) in cnetral Kansas; the 130 ka age is probably a better choice since it is consistent with the sea-level record and the age of marine isotope boundary 5e-6, presumed end of the Illinoian glaciation. (e) Age of marine isotope boundaru 6-7 as determined by Martinson et al.(1987). (f) Age of the Pliocene-Pleistocene boundary at the Virca, Italy, section (Aguirre and Pasini, 1985).

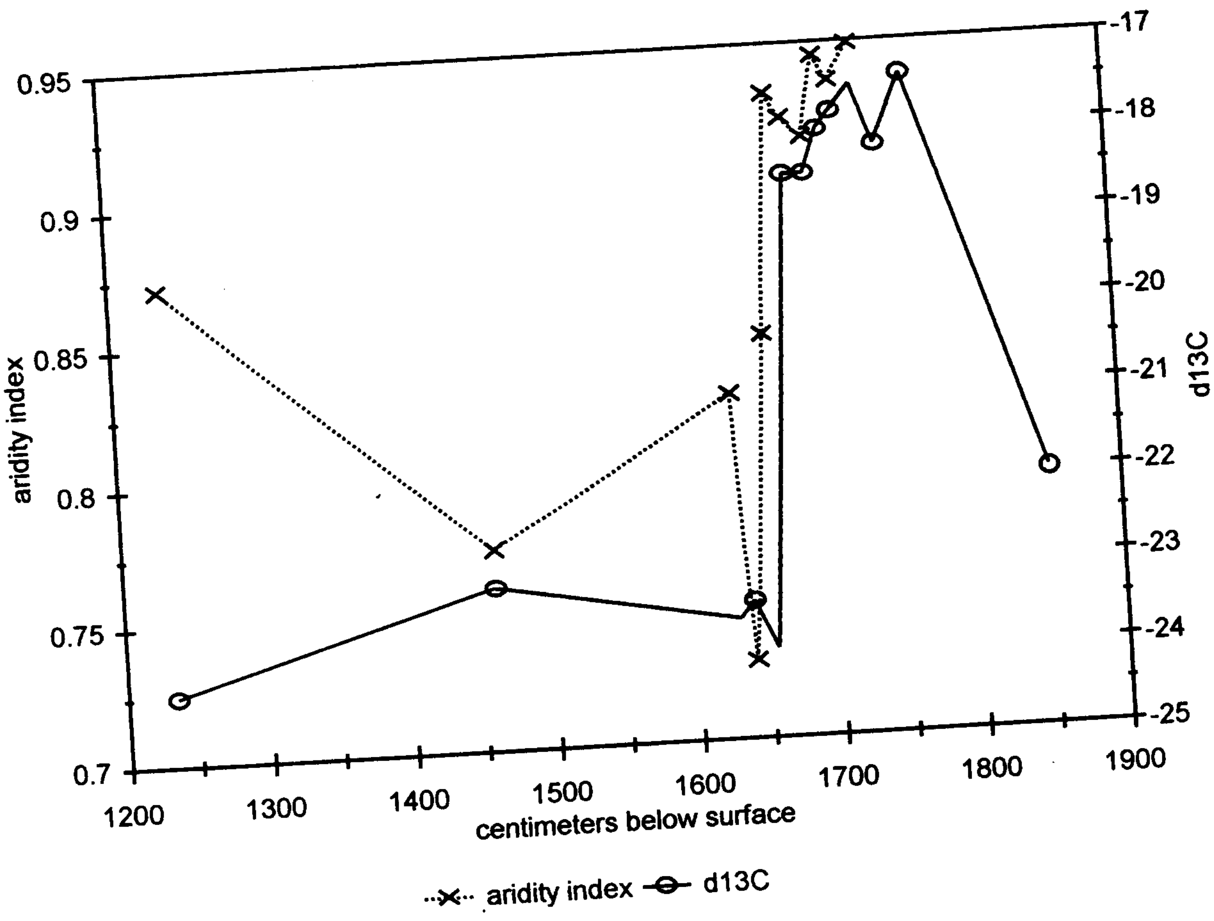
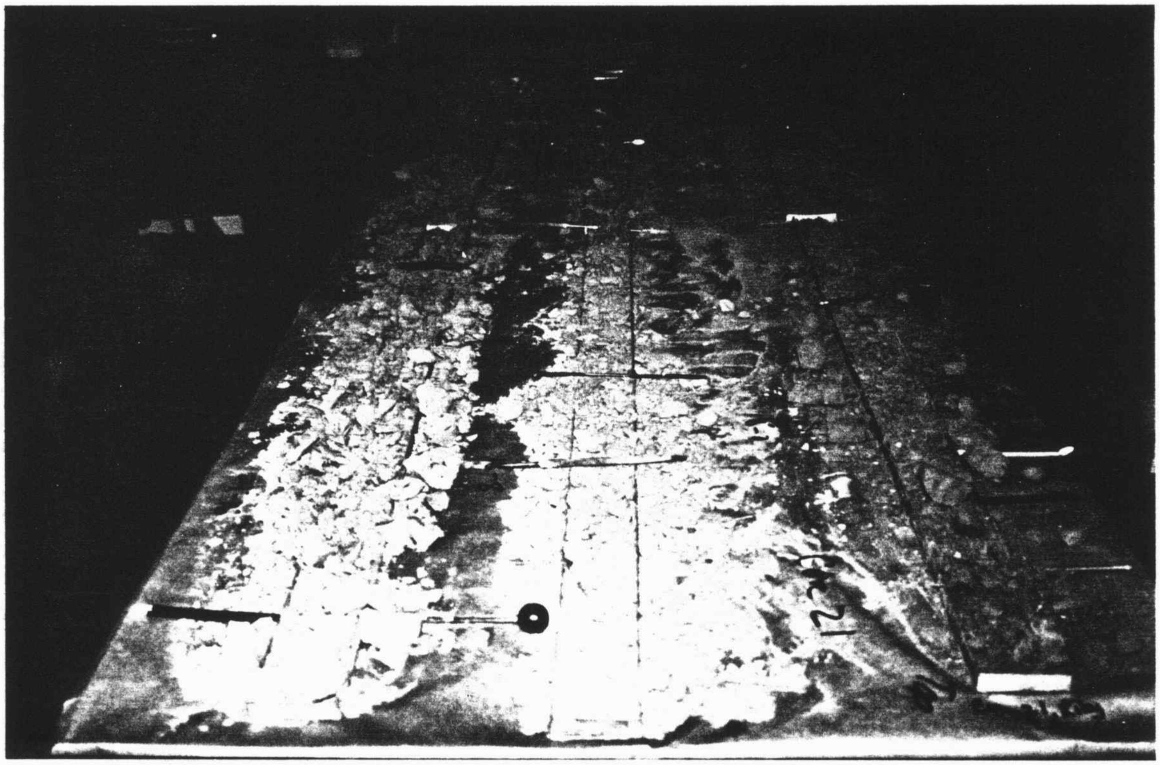


Figure 2. Comparison of the $\delta^{13}C$ content and the opal phytolith-derived aridity index (Chloridoideae/Chloridoideae + Panicoideae) from the Gilman Canyon loess at the Eustis ash pit, Nebraska (W.C. Johnson, D.W. May, S. Bozarth, E. Diekmeyer, unpublished data). Larger values indicate increasing aridity.



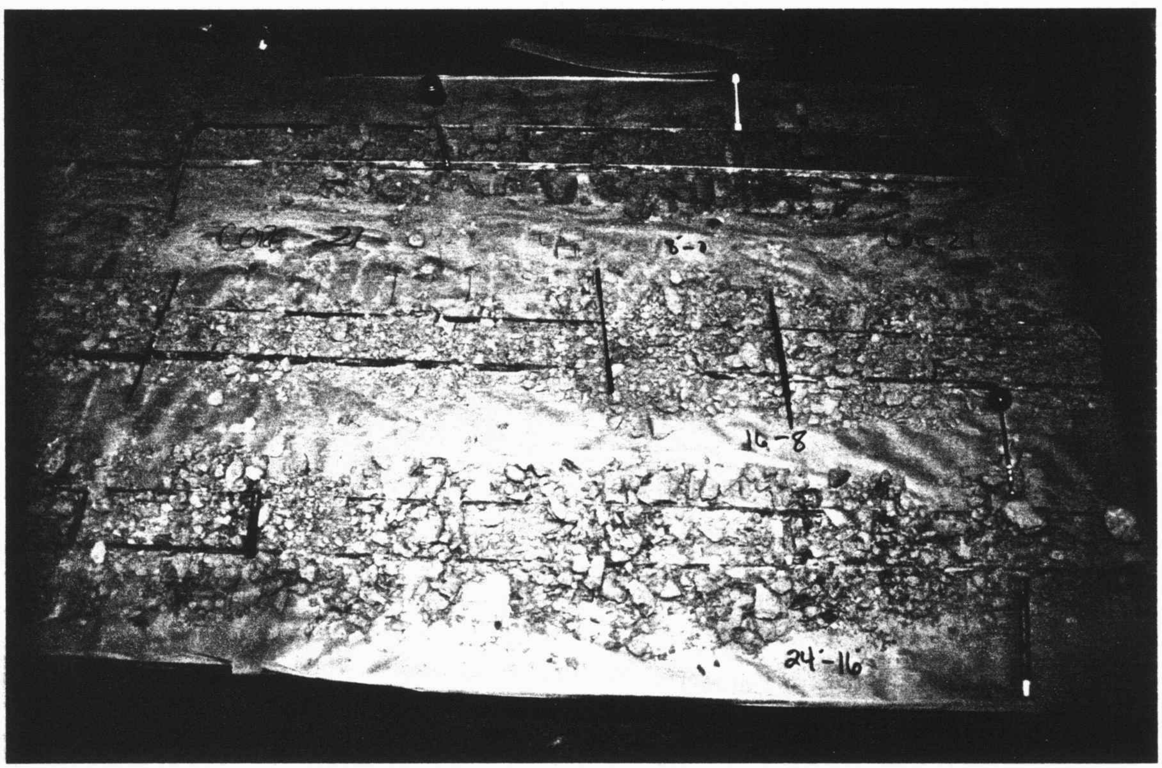


Figure 3a (top). Core No. 21 after sampling and description. Top (modern soil) is in the right foreground, and the bottom of the core (10.80 m) is in the left rear.

Figure 3b (bottom). Same as above, but a lateral view with the top in the right rear. The 358-cm stratigraphic break is located at the middle marker in the middle core segment.

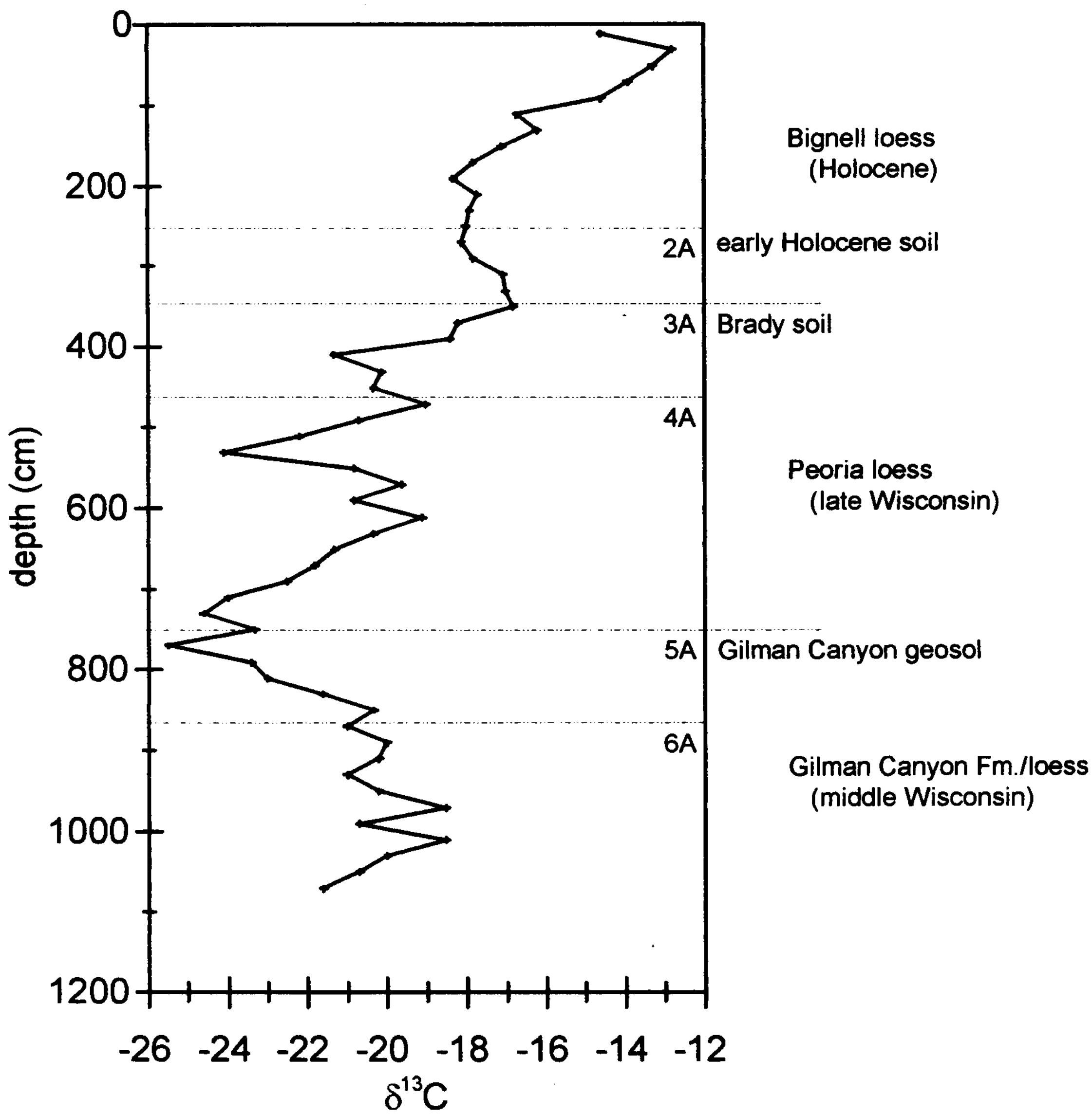


Figure 4. Change in $\delta^{13}C$ values with depth and presumed lithostratigraphic and pedostratigraphic units for core 21, Fort Riley.

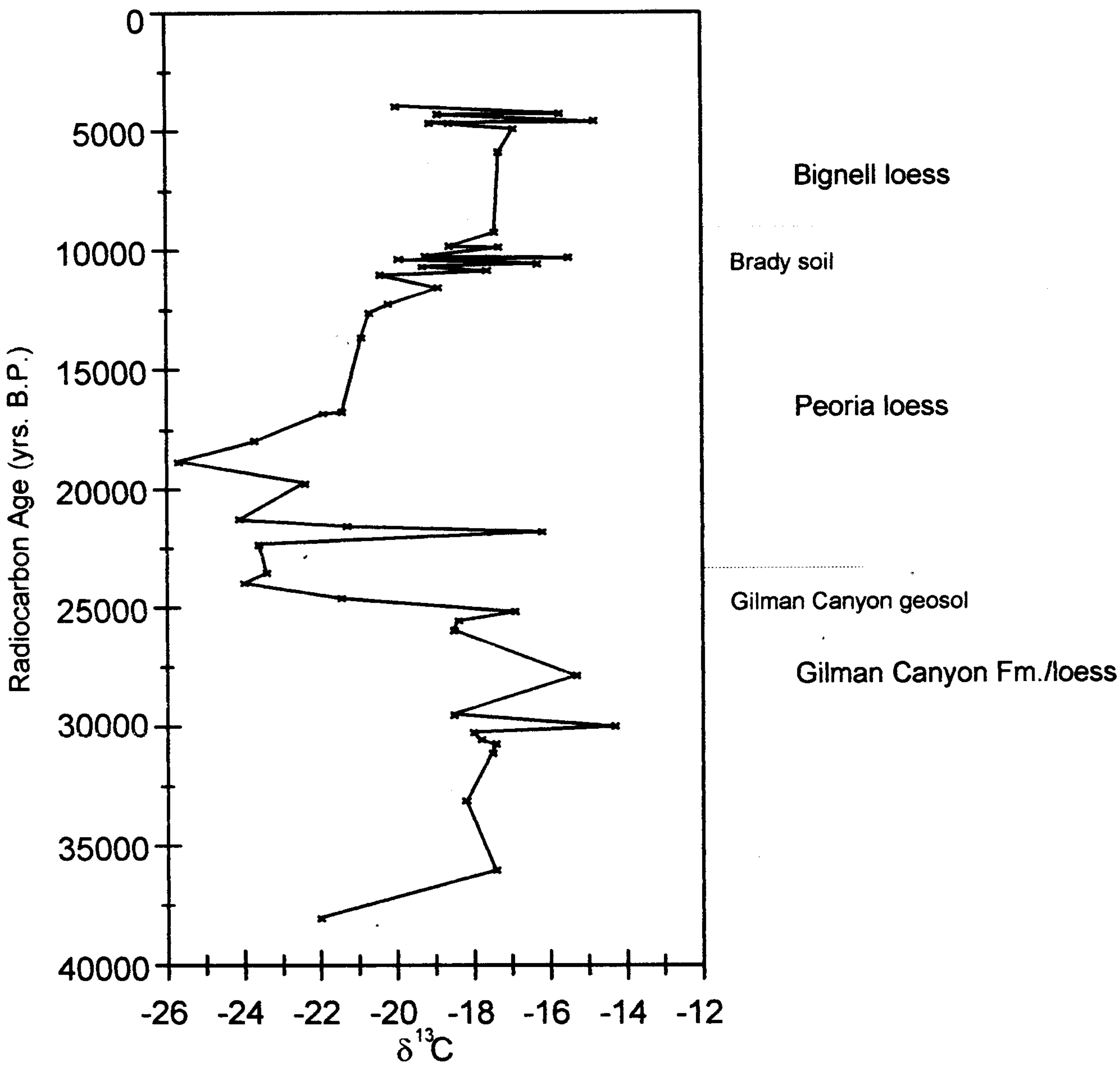


Figure 5. Variation in $\delta^{13}C$ values with climate, as a composite of sites from the central Great Plains. (W.C. Johnson, unpublished data, and other sources)

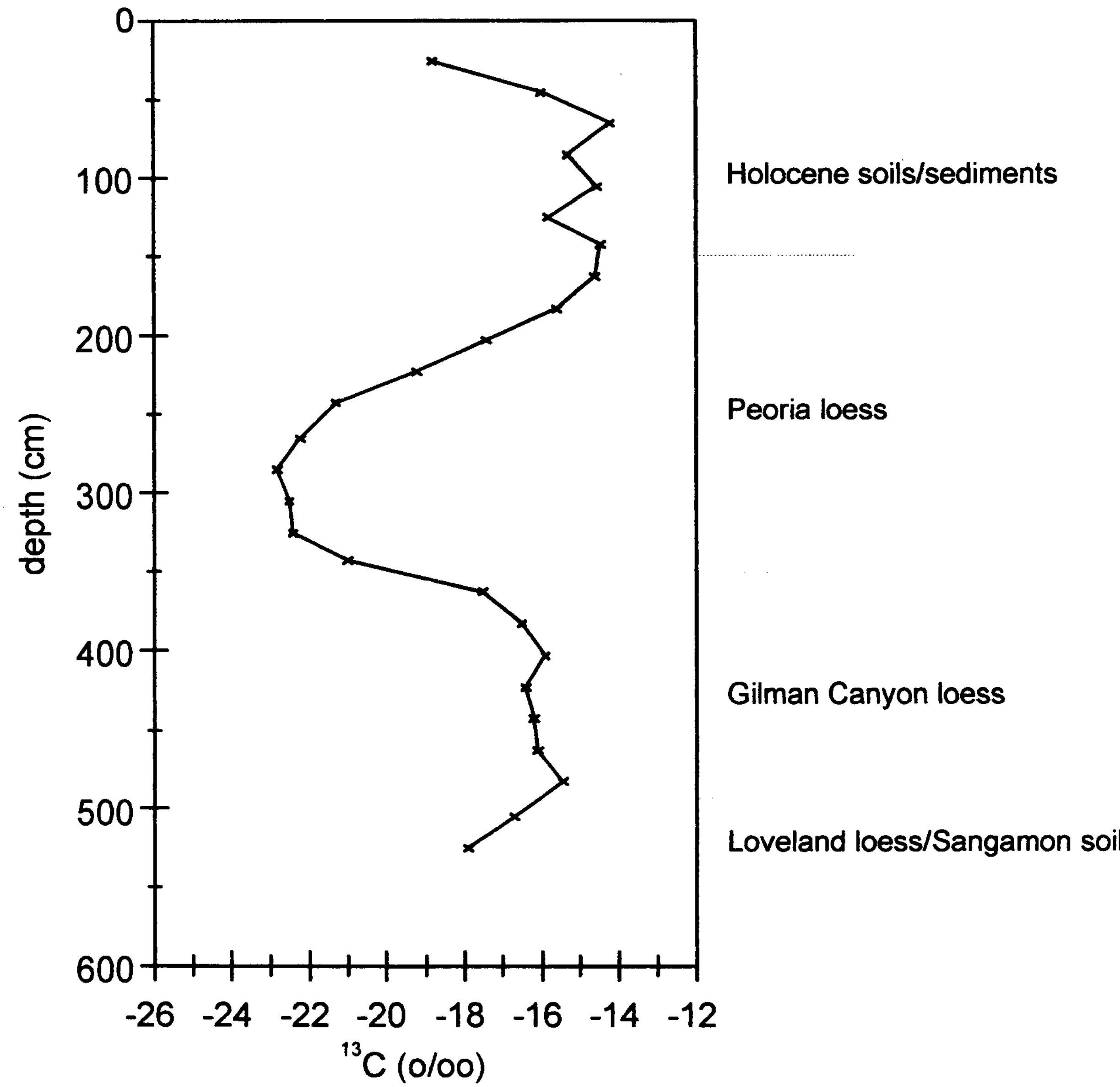
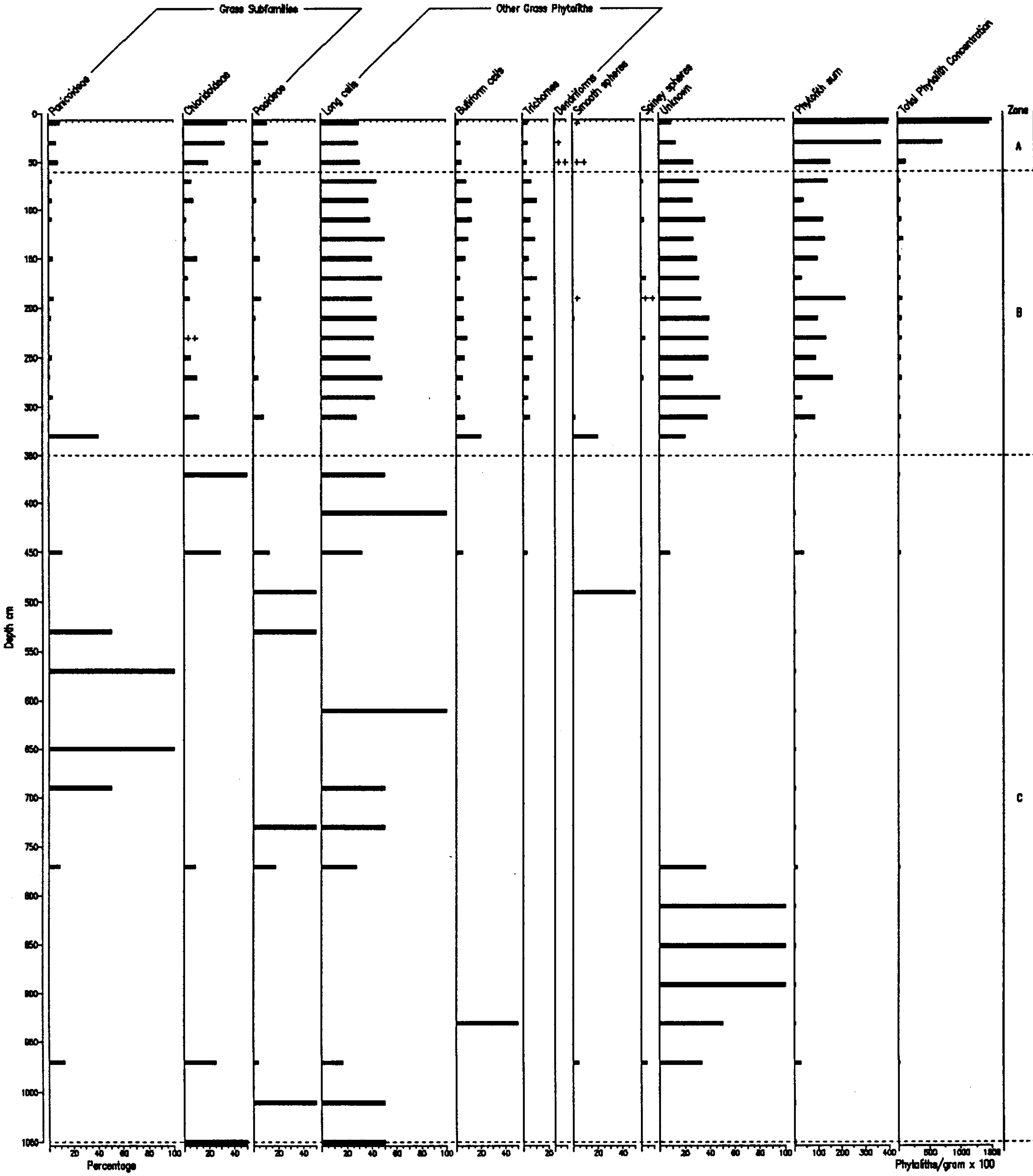


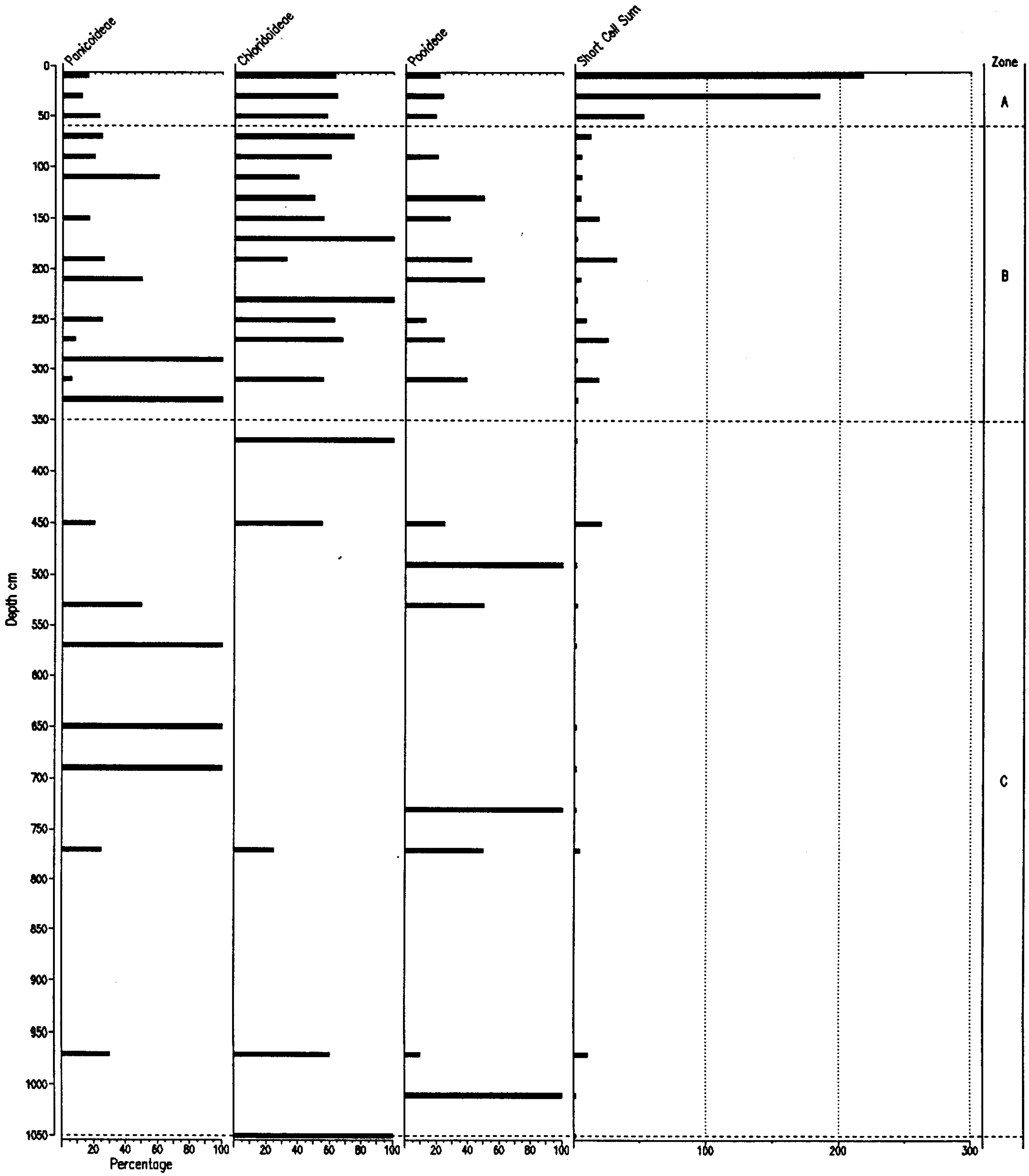
Figure 6. Changes in $\delta^{13}C$ with depth for the Beisel-Seinle site, Russell County, Kansas. (E. Diekmeyer, unpublished data)

Figure 7. Frequency of Phytolith Types



Note: one + = < .5%; two +'s = > .5% < 1.0%

Figure 8. Frequency of Grass Short Cells



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Appendix A

Soil and Stratigraphic Data For Core 21
(from D.L. Johnson, 1994)

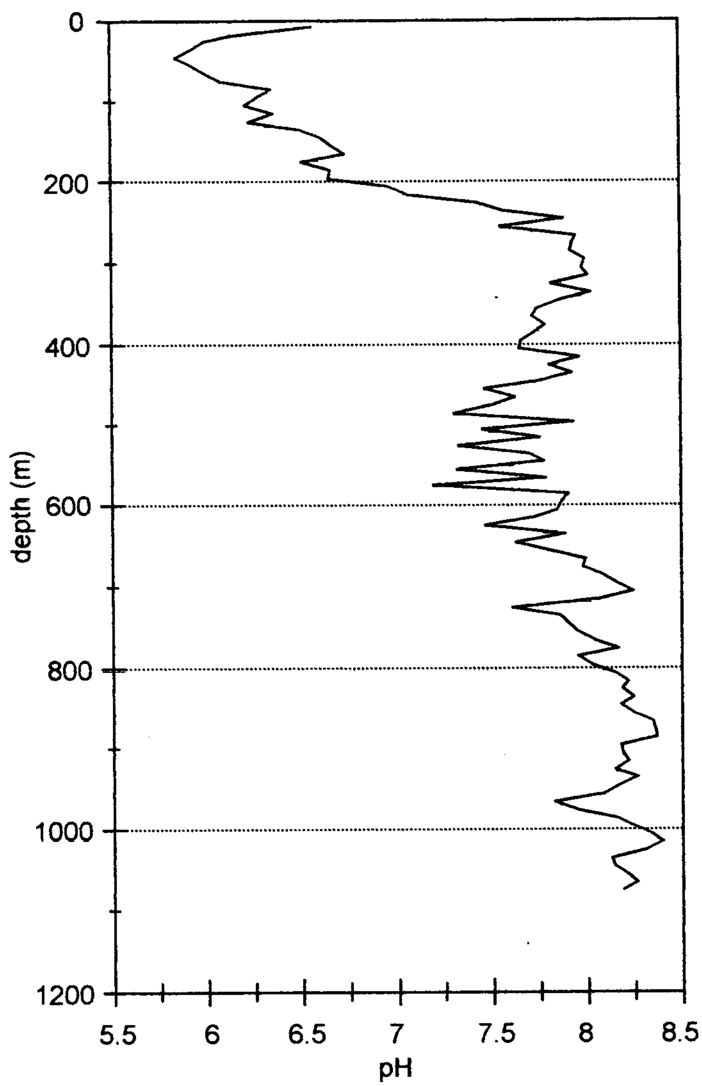
SOIL DESCRIPTION CORE 21

CLASSIFICATION: Kenesaw Series, Typic Haplustolls
LOCATION: Ft. Riley Ks; On Sumner Hill, the highest part of bluff immediately above Camp Funston
GEOGRAPHIC COORDINATES: 39° 6' 26" N. Lat., 96° 44' 8" W. Long.
LANDFORM: Loess-covered high terrace of Kansas River
PARENT MATERIAL: Loess over probable alluvium
SLOPE: 1-2%
ELEVATION: ~1217 ft (371 m)
VEGETATION: Grass
SAMPLED BY: D.L. Johnson and L. R. Abbott, 6/15/93
DESCRIBED BY: D.L. Johnson
EXPOSURE: Shelby tube core

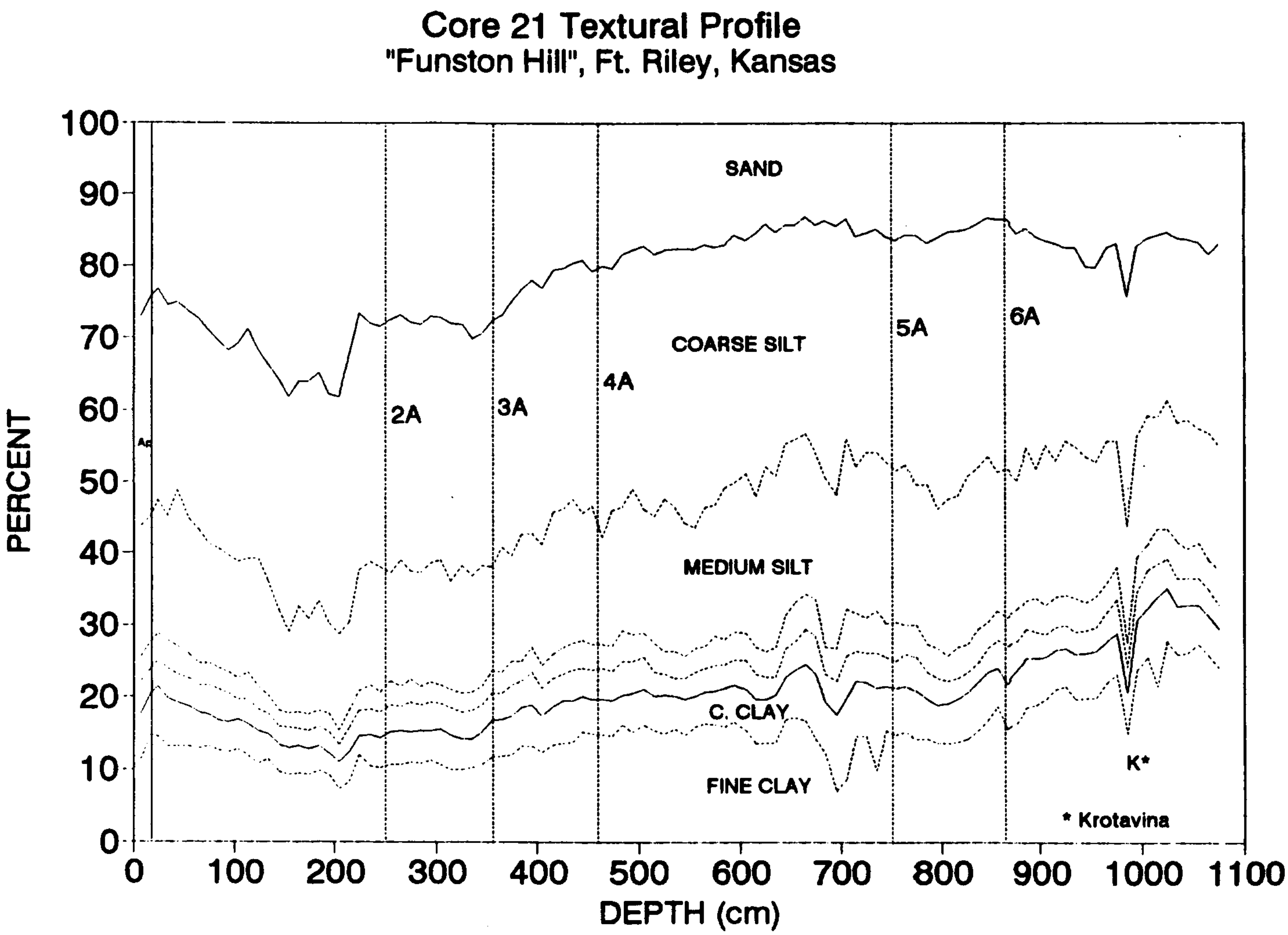
HORIZON	DEPTH (cm)	DESCRIPTION
Ap	0-18	Very dark gray (10 YR 3/1m, 3/3d) silt loam; moderate coarse platy structure; non-sticky when wet, hard when dry; abrupt smooth boundary
AB	18-30	Dark brown (10YR 3/3, 4/3d) silt loam; weak medium crumb structure; non-sticky when wet, hard when dry; clear smooth boundary
BA	30-41	Dark brown (10YR 3/3m, 4.5/3.5d) silt loam; very weak fine and medium subangular blocky structure; non-sticky when wet, hard when dry; gradual smooth boundary
Bwl	41-83	Dark brown (10YR 3/3m, 5/4d) silt loam; very weak fine and medium subangular blocky structure; non-sticky when wet, slightly hard when dry; very few very thin clay films on ped faces and biopores; diffuse clear boundary
Bw2	83-140	Dark brown (10YR 4/3m, 5.5/4d) silt loam; very weak fine subangular blocky structure; non-sticky when wet, hard to very hard when dry; diffuse clear boundary
C	140-201	Brown (10Yr 4.5/3m, 6/4d) silt loam; very weak fine and medium subangular blocky structure; non-sticky when wet, soft to slightly hard when dry; a 15 cm diameter krotovina at 190 cm; abrupt wavy boundary
Ck	201-252	Brown (10YR 4.5/3m, 6/4d) silt loam; weak fine and medium subangular blocky structure; non-sticky when wet, soft to slightly hard when dry; slightly effervescent but no visible carbonates; clear smooth boundary
2Ab	252-280	Dark yellowish brown (10YR 4/3m, 6/4d) silt loam; structureless (massive)to very weak medium subangular blocky structure; sticky when wet, friable when moist, soft to slightly hard when dry; gradual smooth boundary
2ABb	280-310	Brown (10YR 4.5/3m, 6/4d) silt loam; structureless (massive) to very weak medium subangular blocky structure; non-sticky when wet, friable when moist,

		soft to slightly hard when dry; gradual smooth boundary
2Bb	310-358	Brown (10YR 4.5/3m, 6/4d) silt loam; structureless (massive) to very weak medium subangular blocky structure; non-sticky when wet, friable when moist, soft to slightly hard when dry; gradual smooth boundary
3Ab	358-410	Brown (7.5YR 4.5/4m, 6/5d) silt loam; structureless (massive) to very weak medium subangular blocky structure; non-sticky when wet, friable when moist, soft to slightly hard when dry; very few thin clay films in biopores; gradual smooth boundary
3 Bwkb	410-463	Brown (7.5YR 5/4m, 6/6d) silt loam; moderate medium subangular blocky structure; slightly sticky when wet, friable when moist, soft to slightly hard when dry; common thin clay; gradual smooth boundary
4Ab	463-493	Brown (7.5YR 5/4m, 5.5/6d) silt loam; moderate medium subangular blocky structure blocky structure; slightly sticky when wet, friable when moist, hard when dry; gradual smooth boundary
4 Bwb	493-580	Brown (7.5YR 5/4m, 5.5/6d) silt loam; moderate medium subangular blocky structure; slightly sticky when wet, friable when moist, hard when dry; dominantly noncalcareous, but some CaCO3 in isolated vertical biochannels earthworm burrows?); very few faint and very thin clay films on ped faces; gradual smooth boundary
4 Bwkb	580-605	Brown (7.5YR 5/4m, 5.5/6d) silt loam; moderate medium platy structure; slightly sticky when wet, friable when moist, hard to very hard when dry; few thin clay films along ped faces, gradual smooth boundary
4Bwb	605-650	Brown (7.5YR 5/4m, 5.5/6d) silt loam; moderate medium subangular blocky structure; non-sticky when wet, friable to very friable when moist, soft when dry; very few thin clay films along ped faces; abrupt smooth boundary
4Bwkb	650-680	Brown (7.5YR 5/4m, 6/6d) silt loam; moderate medium subangular blocky structure; non-sticky when wet, very friable when moist, soft when dry; very fine faint and thin clay films along ped faces; carbonates occur as blebs and islands but matrix noncalcereous, gradual smooth boundary
4Bwkb	680-755	Brown (7.5YR 5/4m, 6/5d) silt loam; weak medium subangular blocky structure; non-sticky when wet, very friable when moist, soft when dry; very few and faint thin clay films along ped faces; strongly effervescent; abrupt wavy boundary
5(?)Ab	755-807	Brown (7.5YR 5/4m, 6/4d) silt loam; weak medium to fine subangular blocky structure; non-sticky when wet, friable when moist, slightly hard when wet; gradual smooth boundary
5(?)Bwb	807-867	Brown (7.5YR 5/4m, 6/6d) silt loam, weak grading to moderate find and medium subangular blocky structure; slightly sticky when wet, friable when moist, slightly hard when dry; gradual smooth boundary
6Ab	867-902	Brown (7.5YR 5/4m, 5.5/6d) silt loam; moderate fine and medium subangular blocky grading to moderate find and medium angular blocky structure; sticky when wet, friable when moist, slightly hard to hard when dry; gradual smooth

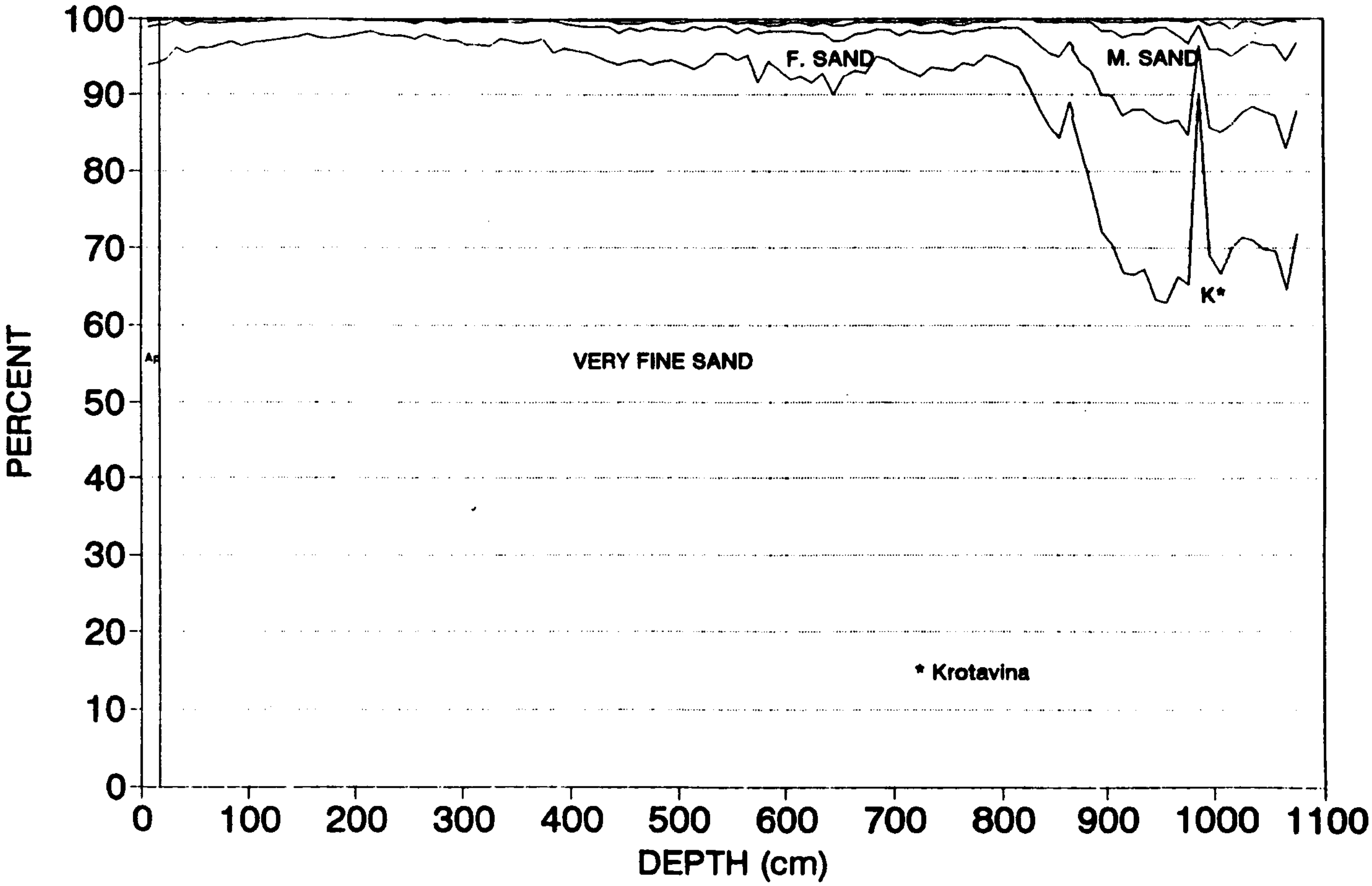
		boundary
6Bwb	902-980	Brown (7.5YR 5/4m, 5/6d) silt loam, moderate fine and medium angular blocky structure (becoming stronger downward); sticky when wet, friable when moist, slightly hard to hard when dry; abrupt smooth boundary
6Bwb	980-990	Dark brown (10YR 3/3m, 4.5/3d) silty clay loam; moderate fine subangular blocky structure; sticky when wet, friable when moist, hard to very hard when dry; abrupt smooth boundary
6Bwkb	990-1030	Brown (7.5YR 5/4m, 5/5d) silty clay loam; moderate to strong, fine and medium subangular blocky structure; sticky when wet, friable to firm when moist, hard to very hard when dry; clear smooth boundary
6Bwb	1030-1080	Brown (7.5YR 5/4m, 5.5/6d) silty clay loam; moderate and strong, fine and medium angular blocky structure; sticky when wet, friable to firm when moist, hard to very hard when dry

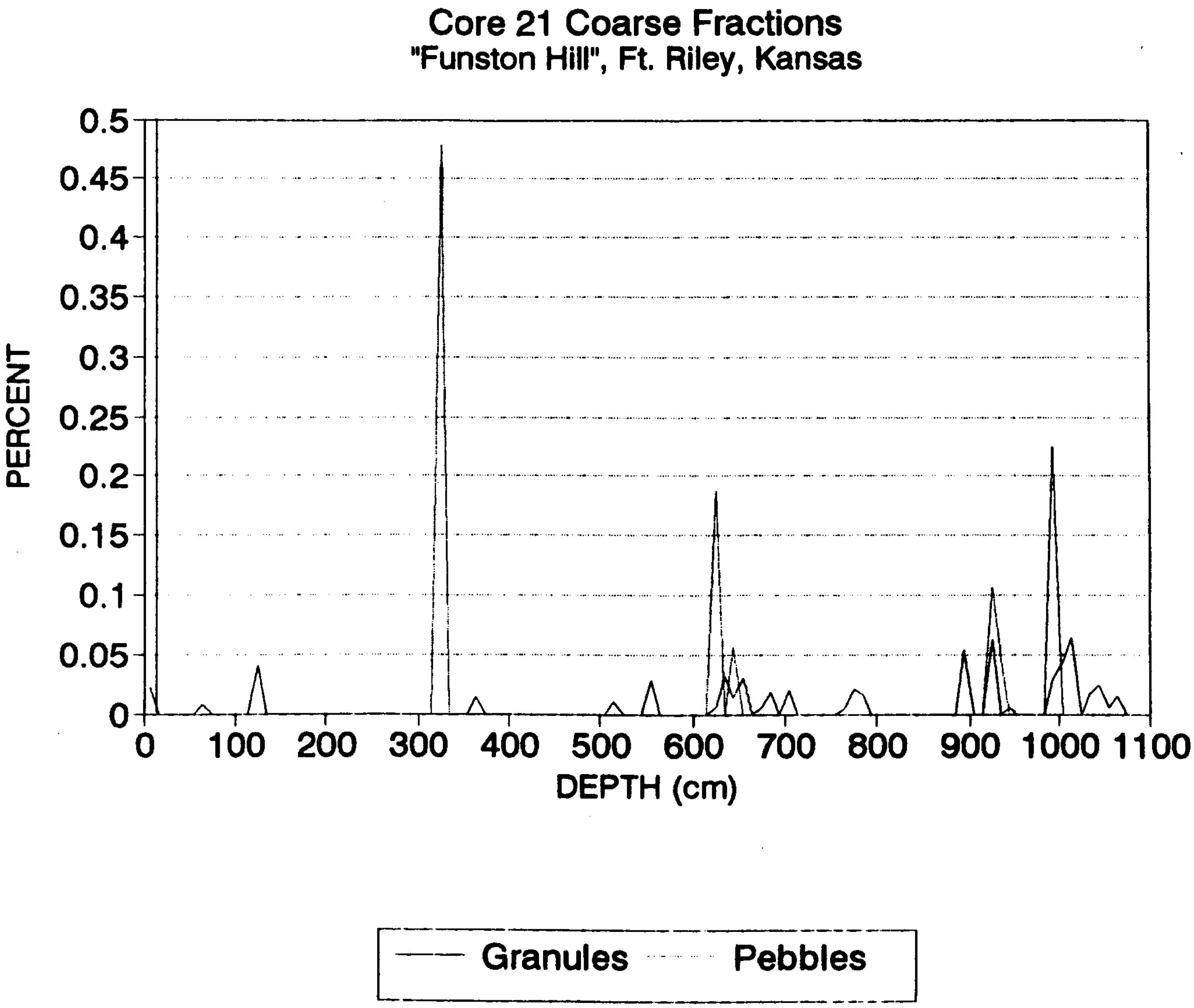


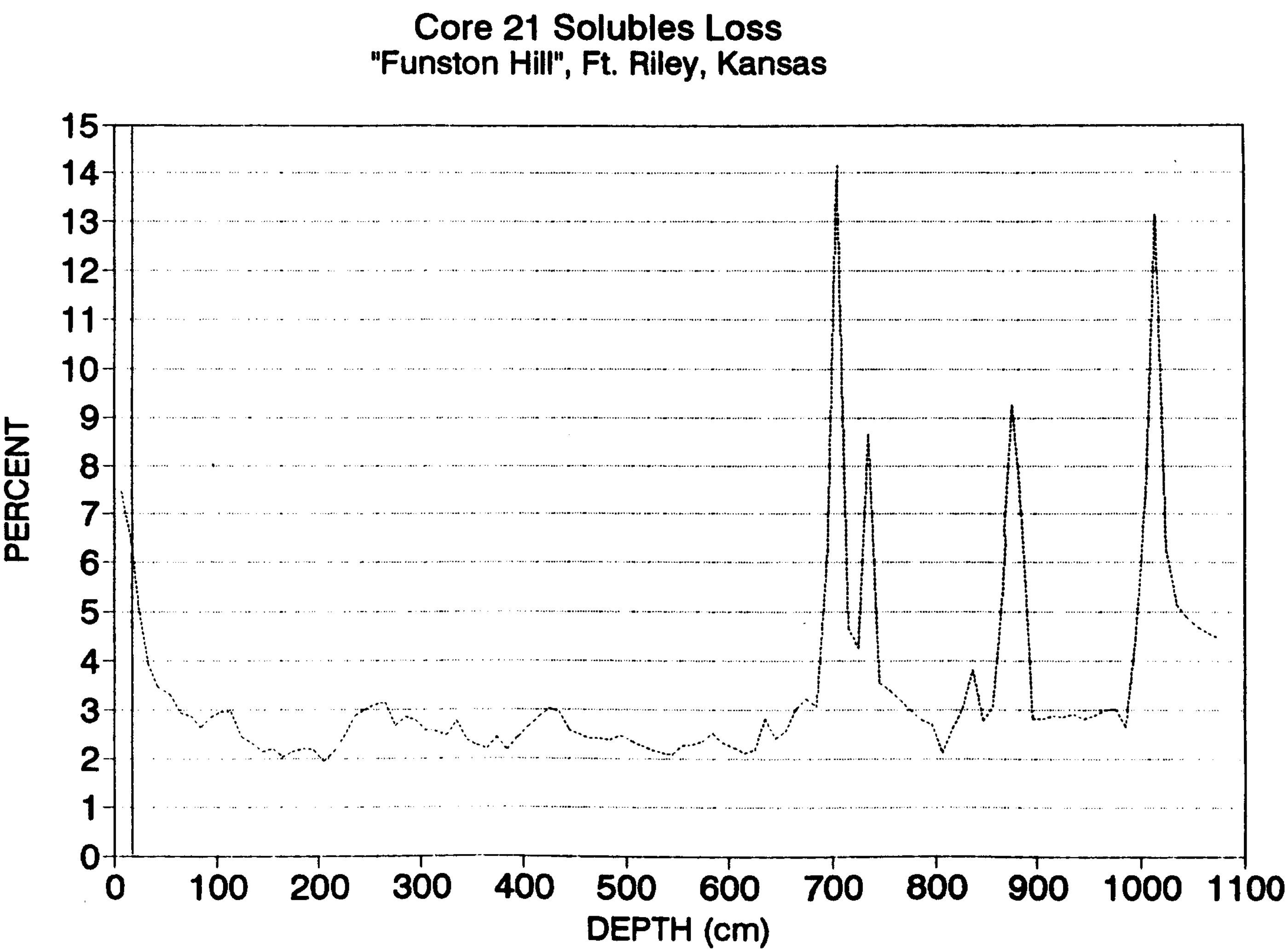
pH-depth relationship for core 21


Core 21 Textural Profile
"Funston Hill", Ft. Riley, Kansas
SAND
COARSE SILT
MEDIUM SILT
C. CLAY
FINE CLAY
2A
3A
4A
5A
6A
K*
* Krotavina
PERCENT
DEPTH (cm)
0
10
20
30
40
50
60
70
80
90
100
0
100
200
300
400
500
600
700
800
900
1000
1100

Core 21 Sand Fractions
"Funston Hill", Ft. Riley, Kansas




Core 21 Coarse Fractions
"Funston Hill", Ft. Riley, Kansas
0.5
0.45
0.4
0.35
0.3
0.25
0.2
0.15
0.1
0.05
0
PERCENT
0
100
200
300
400
500
600
700
800
900
1000
1100
DEPTH (cm)
Granules
Pebbles


Core 21 Solubles Loss
"Funston Hill", Ft. Riley, Kansas
PERCENT
DEPTH (cm)
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
0
100
200
300
400
500
600
700
800
900
1000
1100