MAGNETIC AND STABLE ISOTOPE (13C) PARAMETERS AS INDICATORS OF LATE-QUATERNARY ENVIRONMENTS ON FORT RILEY, KANSAS

by

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Introduction

The sensitivity of grassland community composition and its boundaries to short-term climatic variation during the historical period is well documented for the central Great Plains (Tomanek and Hulett, 1970; Küchler, 1972). Similarly, long-term prairie expansion and contraction, in response to climatic variation, is documented for the prehistoric time scale (*e.g.*, Watts and Wright, 1966; Grüger, 1973; Bradbury, 1980; Webb *et al.*, 1983). Despite the many published studies little is known, however, about the environmental conditions prevailing in the central Great Plains during the late Pleistocene and Holocene. Because the region has few bogs and natural lakes, there is a paucity of good sites for preservation of fossil pollen and botanical macrofossils. Therefore, climatic reconstruction in this region has traditionally relied heavily on the synthesis of other proxy sources such as vertebrate fauna, snails, and occasional botanical macrofossils. As a result, interpretations of late-Pleistocene vegetation of the region range, for example, from continuous taiga-like forest (*e.g.*, Wells and Stewart, 1987) to grassland or steppe (e.g., Graham, 1987)

The quasi-continuously deposited loess of the central Great Plains represents some of the thickest and most complete loess deposits in North America and provides a largely untapped potential for reconstructing past climates. Ongoing research by this investigator and colleagues on magnetic records from these deposits indicates a tremendous potential for environmental reconstruction. For example, lower magnetic susceptibility of the loess associates with times of rapid accumulation, *i.e.*, the cooler glacial intervals, whereas higher susceptibility associates with the intercalated paleosols and weathering zones, which represent times of landscape stability and/or reduced accumulation rates. In addition to correlation with long-term, regional cyclic climatic changes, the relatively rapid accumulation rates associated with loess deposits also permit the resolution of short-term changes in climate (*e.g.*, Zhou *et al.*, 1994; Heller *et al.*, 1993; An *et al.*, 1992).

Recent research using nonmagnetic parameters has also been particularly rewarding. Two approaches that are now being employed include stable isotope ratio analysis (SIRA), especially that of carbon, and biogenic opal analysis, namely that of phytoliths; these techniques have been very effective in yielding unique environmental information. The coincidental use of these three proxies of past environments provides a more comprehensive picture of past environments than use of only one parameter. Magnetic information, e.g., susceptibility, indicates times of weathering and soil development in the stratigraphic sequence; SIRA of carbon provides an overall impression of the type of climate to which the plants have adapted; and opal phytolith morphology permits identification of specific groups of plants.

Fort Riley

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 late-Pleistocene and Holocene landscape evolution. Research began with an overview study by D.L. Johnson (1992). On-site work by D.L. Johnson, an assistant and the contractor in June, 1993 involved further reconnaissance, subsurface exploration with a vehicular-mounted coring rig, and documentation 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 (1994) and subcontractors. 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. Additional study has described the distribution of the loess mantle on the upland and refined the stratigraphy at the Sumner Hill locality, i.e., core site 21 (D.L. Johnson, 1996).

The first phase of the paleoenvironmental research was a test of the potential for SIRA (carbon) and opal phytolith analysis to provide a proxy time series of climate for the reservation (W.C. Johnson et al., 1994). Study was conducted on a core extracted from the Sumner Hill locality. Isotopic analysis of the carbon ($\delta^{13}C$) provided a time series that compares well with a composite of regional carbon isotope data, whereas the opal phytolith record was largely uninterpretable below about 3.5m depth, i.e., for the lower 8.5m of the core. On the basis of a recently derived chronostratigraphy of the site (D.L. Johnson, 1996; this study), the viable portion of the phytolith record does, however, include the time interval related to cultural occupation, i.e., the last 13,000 years or so.

A second phase of environmental reconstruction was designed to consider the record from several sites distributed over the reservation in upland and lowland environments. Because magnetic analyses, particularly susceptibility, have been recently demonstrated to contain a climatic/ environmental signal (e.g., Maher and Thompson, 1994b), this analytical approach was adopted to merge with the SIRA and phytolith data sets. This report summarizes results of the magnetic analyses and SIRA conducted on the reservation over the last year.

Regional Late-Quaternary Stratigraphy

An articulation of the regional loess stratigraphy is necessary for the appreciation of the late-

Quaternary record preserved at Fort Riley. Considerable research has been conducted on these deposits in recent years, with an emphasis on development of the chronostratigraphy. Research results from Fort Riley are, however, making a major contribution to the data base and to the understanding of paleoenvironmental conditions for the central Great Plains.

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 United States (Fullerton and Richmond, 1986).

Because the glacial stages "Nebraskan" and "Kansan" and the 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 stratigraphic 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, 1978b; 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) 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 (1986a) 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, 1986b). 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 sanitary 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 south-central Nebraska and found that the soil-forming periods were warmer and periods of dust accumulation cooler.

Fort Riley. No pre-Illinoian Quaternary sediments have yet been recognized on Fort Riley as a result of this series of CERL-funded studies or are reported in the literature. It appears that major entrenchment and denudation preceded the Illinoian because exposures and cores to date have exposed only what is presumed to be Illinoian-age material overlying bedrock. For example, the

Illinoian Sangamon soil, loess, and alluvium rest upon a strath, or bedrock-defended terrace at the Sumner Hill locality; conversely middle Pleistocene loess rests on an erosion surface developed on limestone at the Bala cemetery site.

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 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 north-central 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 and others (1988) reported thermoluminescence (TL) ages of 136 ka and 130 ka for the upper part of the presumed Loveland loess exposed in an abandoned quarry near the town of Milford, immediately west of Fort Riley. TL age data from this and others sites in Kansas indicates that the Loveland loess began accumulating sometime before 130 ka (Feng *et al.*, 1994; Johnson and Muhs, 1996). Recently, Maat and Johnson (1996) 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 ka. (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 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 suggested by oxidation, apparent deep leaching, and high clay accumulation. A major 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 Wisconsinan loess is typically weathered and forms a pedological continuum with the underlying Sangamon soil (Follmer, 1983), and early investigators 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 two or more paleosols welded together to form a complex that reflects significant spatial and temporal variation in environmental conditions and an appreciable time span. Although laboratory data from exposures in central Kansas indicate the Sangamon soil was strongly weathered chemically, presumably under a warm, moist climate (Feng, 1991), recent data from the Eustis ash pit in south-central Nebraska indicate that the Sangamon soil is not very mature geochemically (Muhs and Johnson, 1996).

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, 1993a, b) provide a minimum age of about 45 ka for the Sangamon soil. Also, Forman (1990b) and Forman and others (1992) 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 were apparently 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.

Although relatively little is known of the climate prevailing during Sangamon time, some recent research provides a first indication. In a regional examination of the stratigraphy representing Sangamon time, Dreimanis (1992) noted that for the midwestern and eastern United States (the southeastern margin of the Laurentide ice sheet of North America) temperatures of the Sangamon climatic optimum were several degrees warmer and precipitation was less than at present. This time of maximum temperature and minimum precipitation, likely occurring during oxygen isotope stage 5e (c. 120ka), was followed by gradual cooling characterized by oscillations between cool and cold climates. Through global circulation model experiments, Harrison and others (1995) simulated regional climate and associated biome changes in North America for the two extremes in orbital parameters during isotope stage 5e. Maximum solar insolation, at about 125-126ka, produced midcontinental summer temperatures about 8°C warmer than today and a corresponding expansion of the warm season grasses. Conversely, minimum solar insolation, at about 115ka, depressed the summer temperatures 5°C below those of today, and expanded the cool season grasses and forests. Pedologic data obtained by Muhs and others (1996) support the notion of a climatic regime warmer and/or drier than at present.

Fort Riley. Thus far, investigations on the reservation have indicated that Illinoian age sediments are ubiquitous but relatively thin, and expressed primarily as the Sangamon soil developed within them. Best known expression of the Sangamon soil occurred at the Sumner Hill locality in a thin layer of Loveland loess overlying alluvium.

Wisconsinan Stage

As the most recent glacial episode, the Wisconsin, beginning between 79-70ka, has the greatest resolution, and has been traditionally subdivided into five substages. 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) referred 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 Wisconsinan (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). As noted above, the formation 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 south-central 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) supported the contention that the leached, or basal zone was slowly accumulating, early Peoria loess experiencing pedogenesis through inference that gastropods were originally present, but subsequently destroyed during weathering of the loess. Above the leached zone, the rate of accumulation of Peoria loess was sufficiently rapid (c. 0.6mm/yr) as to preclude any soil development.

McKay (1979a, b) noted in Illinois that radiocarbon ages on organic materials from within early 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 (1978a) 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 40 ka at the base to 20 ka at the top (May and Souders,1988; Johnson *et al.*, 1993a). The basal age of 40 ka agrees well with the time set by Richmond and Fullerton (1986a) for the beginning of the late Wisconsin. While Nebraska has many dated locations forming an arcuate pattern around the eastern and southern sides of the Sand Hills, data from Kansas are relatively limited. The ages in Kansas do show, however, good agreement from the south-central to the north-central 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 ‰ and C_4 (warm-season/arid) species -13 ‰, 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 (D.W. 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 (Grüger, 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 0.6 m in discontinuous patches. Any accumulation less than 0.6 m is presumed unrecognizable in the field because it has become incorporated into the existing surface soil. Peoria loess typically rests conformably upon the Gilman Canyon Formation.

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 flood plains, present sand dune areas, and fluvial and eolian erosion of the Ogallala Formation. Research on trace element concentrations in loess (Johnson and Muhs, 1996) 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 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 south-central Nebraska at about 13 ka, after which valleys were filled with late Peoria Loess.

Fort Riley. The Gilman Canyon Formation sediments and soil were present at all sites sampled for magnetic and isotopic analyses, indicating that the formation has both upland (loess) and valley (alluvium) phases preserved at Fort Riley, as elsewhere in the region. Radiocarbon ages on the formation range from about 19ka to over 23ka (table 1); these relatively young ages represent the most recent period of Gilman Canyon time pedogenesis; the relatively unaltered loess and alluvium was not sampled for dating.

Coring and backhoe trenching indicates that Peoria loess is typically relatively thin or undetectable. Only one of the magnetic and isotopic study localities exhibited clearly identifiable unaltered Peoria loess below the Brady soil.

Holocene Series

The beginning of the Holocene, about 10 ka (Hopkins, 1975), was 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 (Mörner, 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 soil 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 A.B. Leonard, 1949, 1951). 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 or may have developed in Peoria loess subsequent to the erosion of the Brady soil and Bignell loess. The Brady soil is typically dark gray to gray-brown and better developed than the overlying surface soil within the Bignell loess. 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 (1993b). 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, 1993b). The results indicate a soil forming interval lasting a minimum of 1200 radiocarbon years; this investigator's 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 no 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 has since been secured by this investigator: 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 south-central 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, 1995). 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 8,274±500 (C-108a) and 9,880±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, appears to qualify, 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 disappeared 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 (Grüger, 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 loess, no certain identification can be made without the presence of the Brady soil. The Bignell loess does not form a continuous mantle on the Peoria; instead, it occurs as discontinuous deposits which are most prevalent and thickest adjacent to modern-day valleys, particularly the south side, and often within depressions on the Peoria surface. 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 ages obtained at the type section in Nebraska and the Speed roadcut that the Bignell loess can be no older than 8,000 to 9,000 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 Great 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 loess 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 (Grüger, 1973). Fossil pollen data from Cheyenne Bottoms suggest consistently lower water levels in the marsh during the middle Holocene (Fredlund, 1991, 1995). 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 sheet 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.

Fort Riley. The Brady soil and Bignell loess are prominent elements of the late-Quaternary stratigraphy on the reservation. Together with the surface soil, they typically comprise the approximately upper 2m of most sites documented (D.L. Johnson, 1996), including many of the sites analyzed for this report. Therefore, it appears that appreciable loess was deposited during the Holocene. This is not surprising given the proximal juncture of two large stream systems, which formed a significant loess source (dust from a relatively wide valley floor), and the situation of the sites on the north side of the river valley with southerly winds.

Paleoclimatic History during the Wisconsinan Glacial Stade

As the most recent glacial episode, the Wisconsin has the greatest chronostratigraphic resolution. However, existing knowledge of the climatic environment for this time interval is relatively limited and inconsistent for the central Great Plains. To date, either pollen records from peripheral areas or synthesis of various types of proxy data have been used to reconstruct climate history of the region.

The insolation record exhibits two relatively warm peaks (c. 50 and 30 ka) during marine isotope stage 3. Each peak is followed by a relatively minor and gradual decrease, culminating in the decrease to the glacial maximum (c. 18ka). In contrast to the gradual insolation changes, most paleoclimatic records from this interval indicate rapid alternations of warm and cold events, which in frequency and timing appear to be unrelated to the Milankovitch forcing (Curry *et al.*, 1993). Dansgaard *et al.* (1984) showed that rapid and extreme fluctuations in stable isotopes (signifying air-temperature differences of 5°C) from two sites in Greenland appear to correlate over the past 50,000 years . These fluctuations are in phase with changes in CO_2 and dust content. Rapid and substantial temperature fluctuations recorded in ice core segments during stage 3 also correspond with oscillations in the North Atlantic marine sediment record of species abundance and ice rafting (Heinrich, 1988; Keigwin and Lehman, 1994). Using accelerator ^{14}C ages on a planktonic polar species, Broecker *et al.* (1988) identified four rapid climatic oscillation between 40 and 22 ka in the North Atlantic.

The regional upland vegetation, as inferred from the pollen record from Cheyenne Bottom in Kansas (Fredlund, 1995), appears to have been nearly treeless throughout the Farmdalian (ca. 30,000 to 24, 000 yrs B.P.). The pollen record suggests that, although regional tree and shrub populations were higher and more diverse than in Holocene, they were a secondary component of the overall vegetative structure. Grassland-sage steppe dominated the regional uplands surrounding the Cheyenne Bottoms basin throughout the Farmdalian period. The rise in Cheno-Am pollen percentages and an influx of sand beginning at about 25 ka probably mark the rapid onset of a cycle of aridity. Immediately after the onset of aridity, the most noticeable changes are declining percentages in the Cheno-Am types and rising percentages of *Pinus*. The increase in *Picea* and other arboreal pollen may signal a climatic shift toward cooler climatic conditions at ca. 24 ka. Unfortunately, the Woodfordian substage of the Wisconsin is missing from the Cheyenne Bottoms record.

Limiting ^{14}C dates in North America indicate that glaciation commenced about 25-27 ka and thus allow less than 10,000 years for ice buildup prior to 18 ka (Andrews, 1987). The structure of deglaciation is uncertain. There is evidence supporting: (1) a smooth deglaciation model with fastest ice wastage centered on 11 ka; (2) a two-step deglaciation model with rapid ice wasting from 14 to 12 ka and 10 to 7 ka, and a mid-deglacial pause with little or no ice disintegration from 12 to 10 ka; and (3) a Younger Dryas deglaciation model with two rapid deglacial steps as in (2) above, interrupted by a mid-deglacial reversal with significant ice growth from 11 to 10 ka.

The data supporting the smooth deglaciation model are maps of Laurentide ice area based on ^{14}C-dated glacial deposits (Andrews, 1987). Although there are subtle suggestions of more rapid retreat at or near the time of the two steps mentioned above, these curves indicate a steady progressive retreat of North America ice, with significant oscillations in retreat rate only at local spatial scales. Some marine $\delta^{18}O$ curves also show a smooth progressive decrease toward Holocene values.

The step deglaciation model is also supported by some marine $\delta^{18}O$ records (Mix, 1987). In addition, the distinctive patterns of change in sea-surface temperature of the North Atlantic Ocean and in Greenland ice-core $\delta^{18}O$ values also show abrupt step-like warming at 10 ka and about 13 ka; these warming might be associated with step-like decreases in Laurentide ice volume. Regionally integrated rates of pollen change in eastern and central North America also show a rapid change in centered on 13.7 and 12.3 ka. (Ruddiman, 1987).

The Younger Dryas (*e.g.*, Osborn *et al.*, 1995: Bard *et al.*,1993; Peteet, 1993) deglaciation model is suggested by the strong signal of sea-surface temperature cooling between 11 and 10 ka in the North Atlantic Ocean. At least early and perhaps all of Brady pedogenesis coincides with an abrupt and brief cool interval correlative with the classic Younger Dryas cold interval of the North Atlantic region.

By the middle Holocene, drying had reached a maximum according to most studies. Northwest Texas was experiencing conditions of maximum temperatures, minimum precipitation, and eolian activity between 6000 and 4500 yrs B.P. (Holliday *et al.*, 1983; Holliday, 1985; 1989; Johnson, 1987; Pierce, 1987). This episode coincides with $\delta^{13}C$ values from soil organic matter from the same area revealing a shift from -23‰ in the early Holocene to -15‰ in the middle Holocene (Haas *et al.*, 1986). These results were interpreted to represent a shift from cool-season C_3 grasses to warm-season C_4 grasses. Based on enriched $\delta^{13}C$ values in soil carbonate from northwest Texas, Humphrey and Ferring (1994) also show a middle Holocene xeric episode, although the $\delta^{18}O$ values from these same carbonates do not indicate a significant temperature change.

A noticeable shift back to cooler and/or wetter conditions was detected in many areas shortly after 5000 yr B.P. The Great Bend Sand Prairie transformed to conditions much like present (Arbogast, 1994). According to Humphrey and Ferring (1994), the return to mesic conditions after 5000 yrs B.P. was interrupted in north-central Texas by a brief warming and drying episode between 2000 and 1000 yrs B.P. Based on depositional environments, they concluded that cooler and wetter conditions returned after 1000 yrs B.P.

Study Localities

Four sites were selected for magnetic and isotopic study on the basis of spatial coverage of the reservation, landscape position, and absolute age control resulting from studies by D.L. Johnson (1994,1996). Sumner Hill (core site 21; D.L. Johnson, 1994) and Bala Cemetery (core site 19; D.L. Johnson, 1994) are upland sites. The Sumner Hill site is located along the bluff line above Camp

Funston and consists of approximately 12m of Quaternary sediment over a strath cut into the Permian Wreford Limestone. A basal gravelly alluvium is overlain by over 11m of loess, *i.e.*, the Loveland loess and Sangamon soil up through Bignell loess capped by the surface soil. Due to the analytical constraints imposed by the sampling of cores, a series of six large trenches were excavated in the valley wall slope to form a stair-step exposure of the sediments in order that the entire sequence could be clearly viewed and thoroughly and accurately sampled. The other upland sample site is located adjacent to the Bala cemetery, a location nearly 40km northwest of the Kansas River valley and about 10km east of the Republican River valley. The site consists of approximately 2.75m of loess overlying the Permian Winfield Limestone, a much thinner loess mantle than at the former site located adjacent to the river valley.

The Pump House Canyon site consists of a loess deposit situated on a strath located within a small tributary valley to the Kansas River valley. Sampling was conducted in a 3.3m deep backhoe trench. Although the trenching did not perforate the loess layer, its total thickness is likely less than 5m because the upper part of the Gilman Canyon Formation was exposed. The Manhattan Airport site (core site 20 & 33), located approximately 1.5km northwest of the airport, consists of clay-rich late-Pleistocene and Holocene alluvium. The site was backhoe trenched to expose the upper 3.8m of fill, which included alluvial phases of the Sangamon soil (?), Gilman Canyon Formation, and Brady soil, which was overlain by Holocene alluvium.

Methodology

Magnetic Analyses

The primary carriers of magnetism (iron oxides, iron sulphides and manganese oxides) usually comprise less than 5% of the sediment mass. Magnetic minerals are, however, common in terrestrial materials and extremely sensitive to environmental conditions. Since it is extremely difficult or near impossible to separate out these minute magnetic minerals in order to study them, the bulk magnetic characteristics of the sediments are usually characterized by one of the bulk properties, magnetic susceptibility, which is measured using a non-destructive technique. Magnetic susceptibility (χ) is a measure of the extent to which a sample becomes more strongly magnetized when a small alternating magnetic field is applied, or simply the ratio of the induced magnetism to the strength of the applied field.

Whereas susceptibilty provides information on magnetic concentration, a related parameter, frequency dependence of magnetic susceptibility (FD), provides information on the magnetic grain size. FD is the percent difference between susceptibility measured at a low-frequency applied field compared to its measurement at an applied field with a higher frequency. Unfortunately, FD measured using only fixed low and high frequencies, as is the case with existing instrumentation, will discriminate only a portion of the total superparamagnetic (very small magnetic material) population. With the instrumentation currently available, only the presence of grains between approximately 18 and 20 nm in diameter (very fine clay size) can be detected. Even so, FD values are typically much higher in soils than in intervening loess intervals, reflecting abundant pedogenic material.

Although susceptibility is largely a product of magnetic mineralogy and concentration, other factors come into play. Some of these factors include size and shape of the magnetic grains, frequency of the applied field, and sample size and shape. By controlling for the latter variables, controls on susceptibility reduce to magnetic grain size and shape, in addition to magnetic mineralogy. Magnetic grains fall largely into three groups, but not exclusively on the basis of size: multidomain, single domain, and superparamagnetic (largest to smallest). Since grain shape has so little influence on the susceptibility, variations in shape are easily accommodated in the algorithms employed. Magnetic susceptibility is controlled mainly by the volume of ferrimagnetic minerals in the sediments being analyzed. Magnetite is almost always the most important of the magnetic minerals.

The potential sources of the magnetic minerals in the sediments include the geologic materials (in this study, late-Quaternary loess), organic matter, atmospheric deposition, and groundwater. Unlithified Quaternary sediments carry with them the magnetic signature of the original bedrock source(s). For example, unweathered sediments (*i.e.*, containing primary ferrimagnetic minerals) will usually consist of large multidomain grains and thereby produce very low FD values (*e.g.*, <1%). Organic matter has an impact through dilution of the magnetic mineral concentration and other yet not fully understood processes, all of which reduce the susceptibility, but have little effect on FD. Calcium carbonate accumulations have much the same effect on the former magnetic parameter. Through dust (dry or wet) deposition or precipitation of dissolved material, the atmosphere provides a source of magnetic material. Tephra deposition from a volcanic eruption is a common example found in the Quaternary record, and, more recently, airborne microscopic debris resulting from industrial and urban activity. Groundwater may precipitate iron (*e.g.*, groundwater laterites), and, under proper pH conditions, iron minerals may precipitate in lakes to become incorporated into the sedimentary record.

Weathering and pedogenesis bring about dramatic changes in the magnetic character of sediments, the processes that make application to this study possible. Chemical and biochemical changes in unconsolidated sediments affect magnetic properties through the release, via weathering, of magnetic grains from previously existing sediments, the release of iron in ionic form from iron-bearing minerals, modification of the amount of diluting substances such as calcium carbonate and some clays, and formation of magnetic through the activities of bacteria and algae, especially magnetotactic bacteria. Further, physical weathering through its mechanical change in size, shape, and associated sorting of magnetic grains may affect bulk magnetic characteristic under certain circumstances. Fire has an appreciable effect on the magnetic susceptibility in that hematite is variably altered to magnetite and maghemite during combustion; buried burned surfaces, for example, show up dramatically in the susceptibility measurements, as does the related phenomenon, lightening strikes, albeit rarely. Pedogenesis, which involves chemical and physical weathering, has a major impact on the susceptibility and FD. All well-drained soils tend to exhibit a high susceptibility signal, whereas poorly drained/gleyed soils usually have low susceptibility values due to dissolution of the ferrimagnetic minerals under the reducing conditions. The array of susceptibility and FD patterns in soils is varied, *e.g.*, some exhibit a general bulge in susceptibility over the entire soil and a high in FD within the B horizon where the fine secondary clay minerals are concentrated.

Magnetic susceptibility, along with FD, has a broad application to Quaternary paleoenvironmental studies. It may be used as a correlation tool, *i.e.*, to construct and compare magnetostratigraphies. Also, provenance of a sediment sample or body may be determined by comparing its magnetic signature with that of the suspected source. Of primary importance to the research at Fort Riley is the application to detecting weathering zones and pedogenesis, *i.e.*, buried weathering zones and soils within the loessal and alluvial record. Through the use of susceptibility and FD, periods of soil development can be identified even when too subtle to be observed in cores or exposure profiles.

Methods. Samples were collected in the field from freshly exposed or cleaned profiles or from cores extracted and transported to the laboratory in clear carbonate plastic liners. The individual magnetic samples were collected in numbered, demagnetized, 4-cm^2 plastic cubic containers with lids. The sample interval varied slightly, but averaged 40 per meter. These cubes were pressed by hand or driven with a rubber-coated, dead-blow hammer into the exposure or core to obtain the required amount of sediment. In the laboratory, the cubes were cleaned on the outside, sorted, air dried, weighed, and placed in wooden trays prior to measurement.

Susceptibility and FD measurements were obtained using a Bartington magnetic measurement system consisting of a Model MS2 susceptibility meter and a 36mm-cavity, dual-frequency sensor (MS2B). As each sample was measured, data were entered into a database program (Microsoft Excel) for subsequent analysis and presentation. Specifics of the procedure are presented in the literature provided with the instrumentation package (Bartington Instruments, 1995) and in Gale and Hoare (1991). The facility is located in this investigator's laboratory within the Department of Geography at the University of Kansas.

Stable Isotope Ratio Analysis (Carbon)

There are few quantitative techniques in use today for paleoecological reconstructions in terrestrial depositional systems. One approach to quantitative reconstructions is to estimate the proportion of C_3 (mesic, cool-season) to C_4 (xeric, warm-season) plants once present at a site using carbon isotopes from humates contained within loess and intercalated soils.

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).

The isotopic data are expressed as the difference, or delta value (δ), between the sample or standard (wrt PDB in ‰). The δ value for a carbon isotope in soil is defined as

$$\delta^{13}C\ \text{soil} = (\delta^{13}C_{C4})(x) + (\delta^{13}C_{C3})(1-x),$$

where $\delta^{13}C_{C4}$ is the average of $\delta^{13}C$ values of C_4 plants (-13‰), ($\delta^{13}C_{C3}$) is the average of $\delta^{13}C$ values of C_3 plants (-27‰), and x is the proportion of carbon from C_4 plant sources. Isotopic composition of soil organic matter is a direct indicator of the fraction of the biomass using the C_3 or C_4 photosynthetic pathways. Paleosol humus probably represents organic matter from the last few hundred years before burial, given the short residence times typical for humus in most modern soils (Birkeland, 1984).

Analyses have been performed on pedogenic carbonate (Cerling, 1984; Cerling and Hays, 1986; Cerling *et al.*, 1989; Gu *et al.*, 1991; Humphrey and Ferring, 1994), lacustrine carbonate (Humphrey and Ferring, 1994), on alluvial and eolian sediments (Jasper and Gagosian, 1989; Lin *et al.*, 1991); Aucour *et al.*, 1994; Nordt *et al.*, 1994), on soil organic matter (Krishnamurthy *et al.*, 1982; DeLaune, 1986; Schwartz *et al.*, 1986; Guillet *et al.*, 1988; Schwartz, 1988; Ambrose and Sikes, 1991), and on opal phytoliths (Kelley *et al.*, 1991; Fredlund, 1993).

Total or bulk organic carbon of the sample is assayed to obtain the $\delta^{13}C$ value. These data are, in turn, average values of the isotopic compositions of all the individual components within the sample. Now, however, the isotopic composition of individual components can be determined directly via gas chromatography-mass spectrometry (GCMS) and tandem mass spectrometry (GCMSMS). These organic geochemical approaches have led to the identification of a wide array of extremely complex molecules in a diverse range of environments throughout the geologic record (Precambrian to present). A recent example of application of this approach to Quaternary stratigraphic problems is the study by Huang and others (1995), a molecular and isotopic study of a glacial/interglacial sequence from a freshwater lake in Kenya.

P. Philip and colleagues at the University of Oklahoma, using gas chromatography-isotope ratio mass spectrometry (GCIRMS), have differentiated between C_3 and C_4 plant signals in sediment samples on the basis of the isotopic composition of the *n*-alkanes (lipids) that are present in the plant waxes, which represent primary products of photosynthesis. In general, C_4 plants typically have *n*-alkanes with $\delta^{13}C$ values in the range of -18.5 to -24.2‰, whereas values from C_3 plants range from -29.4 to -33.2‰. Preliminary data have been obtained by this investigator and P. Philp from the Eustis ash pit for each of the major late-Quaternary stratigraphic units (*e.g.*, Sangamon soil, Peoria loess): individual C_{27}, C_{29}, and C_{31} alkanes (indicative of higher plant waxes) exhibit good agreement among themselves and with total (bulk) organic carbon samples. Plant wax isotopic signals provide a better ("cleaner") indication of the site C_3 and C_4 constituents. Other researchers are now beginning to focus on the isotopic composition of single plant components for paleoenvironmental reconstruction (Krishnamurthy *et al.*, 1995).

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 (Johnson and Valastro, 1994). One hundred fifty 300 to 400-gram samples collected from 10-cm intervals of the cores or exposures from the study sites were prepared for $\delta^{13}C$ analysis. Samples were first disaggregated in 4-liter beakers 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 beaker 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 calcium carbonate transported in with the loess. Following distilled water washes and oven-drying (100°C) in 4-liter beakers, the samples were pulverized and packaged in sterile polyethylene bags. They were then submitted to the University of Texas Radiocarbon Laboratory for stable carbon isotope ratio analysis.

Results

Magnetic Analyses

Sumner Hill. Susceptibility data for the Sumner Hill site (fig. 2) indicate only minor changes in the upper 6m, but an increasing trend and pronounced maxima characterize the lower 6m of the section. The magnitude of the susceptibility signal from about 6m to the surface is consistent with that observed for the Gilman Canyon Formation, Brady soil and Bignell loess elsewhere in the region; the Peoria loess is typically much lower with values around 40 to 50 (Johnson et al., 1993a, b).

Frequency dependence data, in contrast to those of susceptibility, exhibit marked variability within the section. The Sangamon soil contains a bimodal susceptibility signal at the top and bottom, and a large FD mode at the top. This composite signal is characteristic of regional Sangamon soil expression; as noted above, the term *pedocomplex* (6Ab + 7Ab) characterizes the Sangamon soil in that it appears to have a had a very complex pedogenic history over a very long period of time (>15k yrs?).

The Gilman Canyon Formation is particulary well expressed at the site. The ubiquitous soil that dominates the upper part of the loess body (4Ab) is apparent in the FD data but not in the susceptibility, indicating that the overall concentration of magnetic material is relatively low, but much of that which is present occurs within the size window detected by FD, *i.e.*, perhaps fine pedogenic clays. Soil development manifest in both susceptibility and FD occurs in that which is presumed to be part of the lower Gilman Canyon Formation. Between these two pedogenic maxima is the less weathered Gilman Canyon loess. Consequently, the relatively thick 5Ab horizon actually consists of the somewhat less weathered loess and an underlying soil.

The Brady soil (3Ab) is expressed only in FD, suggesting pedogenic production of finely divided secondary clays, but not large overall quantities of magnetic minerals. No unaltered Peoria loess is indicated within the section. Approximately 2.5m of Bignell loess exists at the site, with surface soil development dominating the upper 1.75m. The latter is represented well in FD and moderately in susceptibility. The weakly developed 2Ab horizon recognized by D.L. Johnson (1996) is unrecognizable in the magnetic data as a discrete entity; it is likely welded to the surface soil, thereby giving the surface soil an exaggerated thickness.

Bala Cemetery. A core extracted approximately 200m south of the Bala Cemetery from the thin upland loess mantle represents a second upland site sampled. Susceptibility exhibits little variation in the upper meter, but contains exaggerated variations in the lower 2m of the core (fig. 3). Conversely, FD data exhibit high variance. When smoothed, the magnetic data, particularly FD, are more easily interpreted (fig. 4).

The Gilman Canyon soil and a thin zone of relatively unaltered loess are apparent in both magnetic parameters, as is their contact with the underlying limestone bedrock. The ^{14}C age of 19,070 yr BP located at the top of the Gilman Canyon Formation is a terminal age for the formation regionally. Peoria loess has the characteristically low values for both magnetic parameters, indicating relatively low levels of weathering. The susceptibility values indicate, in the absence of ^{14}C age control, that the Brady soil and Bignell loess do not exist at this site. Age data from core site 28 (table 1) to the south of Bala Cemetery also indicates that the Brady soil is absent and that little or no Bignell loess is present. Bignell loess typically produces susceptibility values in excess of 85.

From a minimum value at about 0.9m, FD increases to a plateau that is likely a buried soil (*c.* 0.75-0.5m) welded to the surface soil. The lack of surface soil definition in the susceptibility data may be due to somewhat poor drainage at the site, but FD clearly signifies the surface soil.

Pump House Canyon. Although the site is believed to be a loess-mantled strath, backhoe trenching did not reach bedrock. Like the Sumner Hill site, the Pump House Canyon site contains an expanded Holocene portion of the section, thereby yielding a high-resolution record of Holocene deposition and environmental change (fig. 5). The ^{14}C ages of 18,830 and 23,010 yr BP bracket the end of Gilman Canyon time, e.g., the soil appearing at the bottom of the backhoe trench.

The first soil above the presumed Gilman Canyon Formation is expressed well in both the susceptibility and FD. The offset in data from the two parameters is common in other magnetic records obtained from the central Great Plains by this investigator: the susceptibility response likely represents the A and AB horizons and the FD the B (Bt) and BC horizons. The soil's position between the ^{14}C ages, magnetic signature, and physical appearance suggest that it is the Brady soil. At least one weakly developed soil occurs above the presumed Brady soil, based upon FD data. The surface soil is obvious from the values of both parameters; the characteristic reduction of both parameters in the upper 25cm or so is due to the dilution effect of organic matter in uppermost A horizon.

Manhattan Airport. This site consists of Pleistocene and Holocene alluvial fill within a high terrace of the Kansas River valley. The majority of the fill is Pleistocene, but, on the basis of a single Holocene ^{14}C age, the upper 1 to 1.5 m is Holocene (fig. 6). The Holocene fill may have originated from an unnamed tributary entering from the west, rather than from the Kansas River proper; contributions of Bignell loess may also be present. Soils present presumably represent the alluvial, or valley phases of those expressed regionally (e.g., Sangamon, Gilman Canyon, Brady).

Physical attributes from about 2.75m to the bottom of the trench indicated the Sangamon soil; the relatively weak magnetic signal, particularly for susceptibility, is probably due to poor drainage conditions of the alluvial plain during Sangamon time. The Gilman Canyon soil does not appear magnetically; the 19,990 yr BP age, a terminal Gilman Canyon age, should identify the top of the soil. The pronounced central bulge in both parameters is likely the Brady soil, based on the ^{14}C ages. The steady upward decline in susceptibility above 1m depth may be indicative of poor drainage, but is also related to relatively rapid alluviation. The three modes within FD for the same interval are not clearly interpretable, but may relate to episodic alluviation during the Holocene.

Isotopic analyses

Previous Investigations. In a 1994 study of the Sumner Hill locality, W.C. Johnson and others (1994) reported results of carbon isotope analyses on samples from a core extracted at Sumner Hill (fig. 7a). At that time there was a lack of detailed stratigraphic information and absolute time control, resulting in speculation. Extensive backhoe excavations at the locality by W.C. Johnson and D.L. Johnson provided the profiles necessary for detailed stratigraphic description and collection of material for radiocarbon dating (D.L. Johnson, 1996).

On the basis of the recent dating and stratigraphic definition, the interpretation has changed somewhat in that the Peoria loess has been collapsed, the Gilman Canyon moved upward, and the Sangamon soil added at the base of the stratigraphic sequence (fig. 7b). With absolute time control in the upper part of the sequence, the elevated $\delta^{13}C$ values at the base clearly represent the Sangamon soil. Similarly, the loess and associated soil of the Gilman Canyon Formation overly the Sangamon soil and are relatively thick as at the other localities investigated. Isotopic values increase upward in the Gilman Canyon formation. The Brady soil, well expressed in FD at about 3 to 4m, yields isotopic values second only to the surface soil.

New Investigations. Samples were collected and processed from backhoe trench profiles at Sumner Hill (previously unsampled basal portion), Pump House Canyon and Manhattan Airport sites on the south-southeastern side of the reservation, and from cores extracted at the Bala Cemetery site and site 28 on the northwest side. After pretreatment for analysis, the samples were submitted to the Radiocarbon Laboratory at the University of Texas in Austin. The analyses are yet pending due to an unanticipated backlog at the isotope facility. As soon as the data are received, they will be analyzed and incorporated into this report. Data from these sites will provide a spatial perspective on the paleoenvironmental record of the reservation. A limited number of $\delta^{13}C$ values are, however, available as a result of ^{14}C dating at the sites, resulting in a data set that is consistent with the Sumner Hill data (Johnson et al., 1994) and those data assembled from elsewhere in the central Great Plains.

Discussion

Landscape Evolution

A vision of how the upland landscape of Fort Riley evolved during the last 120,000 years or so is now beginning to emerge. Such a perspective provide a useful context for viewing environmental reconstructions. Evolution of the uplands is, of course, not a process isolated from the valley environment; the alluvial setting is extremely sensitive to environmental change and intimately linked to the upland change. Reconstructions of the alluvial history is presently underway and will likely provide data of greater temporal resolution than that of the uplands, at least for the Holocene.

During pre-Illinioan time, the ancestral Kansas River was cutting a strath in the limestones of the present valley walls, tens of meters above its present elevation. This bedrock surface and it covering of alluvial sands and gravels were subsequently mantled with Loveland loess, at a time when the river had entrenched to near its present level. Loess fall was relatively limited, which indicates that the source was either the alluvial plains to the northwest or the Missouri River valley to the east; either source is sufficiently removed spatially or meteorologically such that little loess fall would occur in this distal location. At the end of Illinoian time, the Sangamon soil developed in this thin mantle of loess. The alluvial phase of the Sangamon soil observed at the Manhattan airport site indicates the river was certainly no higher than this high terrace during the last interglacial.

From the ubiquity of the Sangamon soil in the study area as well as throughout the region, it is apparent that erosion was limited following Sangamon time, *i.e.*, during the early Wisconsinan glacial stage. The Sangamon soil appeared to have survived and continued to dominate the landscape until the middle Wisconsin. At this time another loess unit was being deposited, *i.e.*, the Gilman Canyon loess. The source of this loess is unknown, but it appears to different than that of the Loveland loess because it maintains a uniform thickness throughout the central Great Plains, rather than exhibiting a distance decay function as with the Loveland and Peoria loesses. Gilman Canyon loess fall was, however, exceedingly slow: the meter or so of loess accumulated over approximately 20,000 years, *i.e.*, from before 40ka to about 20ka. Other than the relatively thin nature of the loess unit, the occurrence of relatively low depositional rates (surface stability) is indicated by the presence of two or more periods of soil development. The pedogenesis occurred primarily during the latter part of Gilman Canyon time (Farmdalian substage?) in that soil development typically dominates the upper half of the formation.

The contact between the Gilman Canyon Formation and the overlying Peoria loess is transitional. It appears that Gilman Canyon pedogenesis waned as the late-Wisconsinan loess began to fall such that accumulation rates eventually exceeded the rate of pedogenesis. Peoria loess deposition was, like the Loveland loess, relatively low in the study area. The loess unit thickens to the west and northwest in proximity to the sources, *e.g.*, the Sand Hills of Nebraska, Platte River valley, and to a lesser extent adjacent to the other large river valleys crossing the central Great Plains, *e.g.*, the Republican and Kansas. Accordingly, Peoria loess is thickest in the northwestern part of the reservation; over 2m appear to be present at the Bala Cemetery site.

Although no direct evidence has yet been found, some of the Peoria loess, as with other loess units, was probably lost, at least locally, during erosional maturation of the late-Wisconsinan landscape. Clearly, sites to the northwest such as Bala Cemetery have realized erosion of the Peoria loess and overlying Brady soil, assuming it developed in the area. At some localities, such as Sumner Hill, the total thickness of Peoria loess present at the end of the Pleistocene was completely involved in Brady soil development.

The next major pedogenic period (after the Gilman Canyon) occurred during the Pleistocene/Holocene transition. The Brady soil, developed in the upper Peoria loess as deposition ceased, is, on the basis of excavations and coring, well represented in the southern part of the reservation. Consequently, the upland surface was largely stabilized during this time of environmental shift from the Wisconsinan glacial stage of the Pleistocene into the Holocene. Deposition of the Bignell loess, beginning during the early Holocene, buried the Brady soil. The southerly wind flow off the Kansas

River valley apparently transported silts and fine sands from the alluvial bottom up onto the uplands. The rapid distance decay factor is suggested by the lack of Bignell loess at the Bala Cemetery site to the northwest of the Sumner Hill and Pump House Canyon sites. The contrast may be due to differential erosion, but this is deemed unlikely given that the two sites to the south adjacent to the Kansas River valley are in equally or more erosion-prone settings. Since the Brady soil is presumed to have developed to the northwest but is now apparently gone (*e.g.*, Bala Cemetery), a thin layer of the Bignell loess was probably present but quickly eroded; now the Peoria loess is now exposed with its associated surface soil.

Throughout the reservation, surface soil development appears relatively recent, probably within the last 2,000 to 1,000 years and even less locally. This relative surface youth indicates that either erosion has occurred recently or that Bignell loess has been episodically or slowly accumulating on the landscape until relatively recently. The latter scenario appears correct, at least for areas along the southern margin of the reservation adjacent to the Kansas River valley, in that the Holocene upland surface has been periodically stable (slow or zero loess accumulation) such that soil formation could occur. These short periods of stability are manifest as weak or incipient soils developed in the Bignell loess, such as those indicated at the Pump House Canyon site. The preservation of cultural sequences ranging from Clovis to Woodland and Protohistoric, beginning in the Brady soil and extending up through the Bignell loess into the surface soil at archaeological sites in these setting within the region (*e.g.*, Johnson, 1996) also indicates that there has been little erosion of the Bignell loess, at least locally.

Given the preservation and distribution of late-Quaternary deposits on the uplands, Fort Riley is an ideal study area for modeling landscape evolution in the context of cultural history. Further, the extraction and interpretation of proxy data permits reconstruction of past environments.

Paleoenvironmental and Climatic Interpretations

The Regional Magnetic Record. The magnetic record of loess deposits may prove to be one of the most detailed and useful records of Cenozoic climate change on the continents. At a minimum it appears that the magnetic record from loess deposits can provide corroboration of the major, long-term (>5000 yrs) features of the global climate record indicated from deep-sea sediments (*e.g.*, Heller and Liu, 1984; Kukla *et al.*,1988). However, loess deposits also have the potential to help identify shorter-termed global climatic features (<500 yrs) and to provide details about local climate conditions through time (*e.g.*, Heller *et al.*, 1991; An *et al.*, 1993).

Most of the research thus far has focused on the thick loess sequences of central China where a strong positive correlation exists between rock magnetic parameters such as magnetic susceptibility and the orbitally-tuned oxygen isotopic record of deep-sea sediment for the past 2.4 my (*e.g.*, Heller and Liu., 1984; Kukla *et al.*, 1988; Maher and Thompson, 1992). The loess intervals of these sequences generally have low magnetic susceptibilities correlating with inferred cool glacial intervals, and the associated paleosols contain high magnetic susceptibilities which correlate with inferred warmer interglacial intervals. It is now generally accepted that the enrichment of magnetic material in the Chinese paleosols is due to pedogenic processes (*e.g.*, Zhou et al., 1990; Maher and Thompson, 1991, 1992; Heller *et al.*, 1991; Banerjee *et al.*, 1993a; Liu *et al.*, 1993). The primary lines of evidence for this include differences in the magnetic properties of the loess and buried soils. These soils contain a higher percentage of ultrafine magnetite (*e.g.*, Maher and Thompson, 1992) or maghemite (Verosub *et al.*, 1993) which are similar in grain size and morphology to magnetic minerals from modern soils (Maher, 1986; Singer and Fine, 1989; Fassbinder *et al.*, 1990). Recent work which compares ^{10}Be data with magnetic susceptibility gives independent support for the idea that the magnetic material is largely pedogenic (Heller *et al.*, 1993; Beer *et al.*, 1993).

Positive correlations between magnetic susceptibility and the oxygen isotope record have also been reported from New Zealand (Pillans and Wright, 1990) and from the some areas of the U.S. Midcontinent (*e.g.*, Feng, 1991; Grimley and Johnson, 1991; Johnson *et al.*, 1993a; Grimley, 1996). However, in studies from other areas, much or all of the magnetic record of loess sequences does not seem to show a strong relationship to paleoclimate (*e.g.*, Hus and Gaeerhart, 1986). In fact, some loess deposits such as those in Alaska have higher susceptibilities in the loess intervals than the associated paleosols and thus correlate negatively with the deep-sea sediment oxygen isotope record (Beget *et al.*, 1990). The magnetic record of these deposits appears to have originated not through pedogenic processes, but rather by increased efficiency of transportation and deposition of detrital magnetite under the stronger wind conditions that prevailed during the glacial intervals. In support of a detrital origin, the magnetic grains in these deposits are much coarser than the pedogenic material found in the Chinese loess deposits (Beget *et al.*, 1990).

Although the intense research activity centered on the Chinese loess deposits has been a stunning success in developing a continental-based proxy climatic record for that particular region, there appear to be significant problems in applying this conceptual model to loess magnetic studies elsewhere. Given the exciting results from the Chinese loess studies, new studies must be directed toward the magnetic character of loess in broader paleoclimatic, paleogeographic and pedogenic settings. In order to isolate the effect of climate from other factors, new studies must incorporate a more multidisciplinary approach to gain independent verification of local climatic history using nonmagnetic parameters (Banerjee *et al.*, 1993b), such as the use of stable carbon isotope and opal phytolith analyses.

Magnetic records from loess deposits of the central Great Plains are indicating the presence of a strong climatic signal. Preliminary study of loess deposits in southwestern Nebraska by this investigator and colleagues (Feng, 1991; Feng *et al.*, 1994; Farr *et al.*, 1993) has yielded magnetic records which match the major features of the global climatic record over the past 120 ka as represented by the SPECMAP oxygen isotope record of deep-sea sediments as well as the magnetic record of Chinese loess deposits (fig. 8). However, in the preliminary magnetic study of these deposits, this investigator and colleagues have noted considerable variability in loess magnetic records elsewhere in the Midcontinent (Grimley and Johnson, 1991; Hayward and Lowell, 1993; Rodbell *et al.*, 1993; Grimley, 1994), including one study from Indiana in which the magnetic susceptibility record of a portion of the Peoria loess is inverted with respect to the oxygen isotope record in the same manner as the data from Alaska (Begét *et al.*, 1990). Yet there is evidence from these studies that the loess magnetic records are dominantly controlled by climate. The climatic and physiographic variability of the Midcontinent, as well as the varied magnetic records of loess deposits within the region make it highly attractive for detailed study.

Preliminary susceptibility from three sections along the north-south transect illustrate that correlations between the major loess and paleosol units can be easily made (Farr *et al.*, 1996). These data show a systematic decrease in magnetic susceptibility within the Peoria loess from the south to the north along this transect (fig. 9). Possible explanations for this trend include: l) a decrease in loess accumulation rate versus magnetite influx southward (based on the model of Kukla *et al.*, 1988), or 2) an increase in weathering intensity within the Peoria southward in a manner suggested by Heller and others (1991) to explain systematic decreases in susceptibility within individual loess and paleosol units within China. Interestingly, no systematic trends in susceptibility for any of the paleosols have yet been identified. Frenzel and others (1992) show increased temperature gradients along this transect during the last glacial period due to the circulation effects of the Wisconsinan ice sheet to the north. During the interglacials, including today, there was less temperature difference between the north and south end members. Thus, the increase in susceptibility southward probably reflects more intense weathering due to increased temperature to the south during Peoria loess deposition. Johnson and Farr (1996) are, however, currently testing the two hypotheses and any others that may emerge by comparing these results with regional variation in $\delta^{13}C$ within the Peoria and with parameters such as grain size variability and sedimentation rates to determine if there is a detrital influence on the regional trends.

Detailed magnetic analyses coupled with opal phytolith, SIRA, and sediment analyses such as organic matter content have revealed the unique signals of each of the major late-Quaternary stratigraphic units. The Eustis ash pit is one of the sites receiving intense study (fig. 10). Susceptibility and FD are represented as well as four other magnetic parameters. Saturation isothermal remanent magnetization (SIRM), is, like susceptibility, an indication of magnetic mineral concentration and indicates the extent to which the extent to which the susceptibility is contributed by ferrimagnetic minerals such as magnetite and/or maghemite. The paleosols such as the Sangamon soil, the overlying Gilman Canyon soil, and the Brady soil generally have the highest susceptibility and SIRM values, while the intervening loess units have lower values of both parameters. Magnetic mineralogy is indicated by the "S" parameter, which is derived from other parameters and indicates the relative role of hematite versus magnetite and/or maghemite. The relatively high values demonstrate that hematite has a relatively small role; if hematite was important then the magnetic signal may be due to processes other than pedognesis. In none of the many sites studied in the region has hematite had a significant role. Like FD, anhysteretic remanence magnetization (ARM) and its ratio with SIRM are sensitive to the grain size of the magnetic carriers. The Gilman Canyon soil exhibits a strong signal (fig. 10) as does the Brady soil (undepicted), indicating the creation of fine magnetic material during pedogenesis. The Sangamon soil is, however, problematic; this is not true at all sites which suggests that the Sangamon soil is not as weathered and/or it has been effected by post-pedogenic processes. The Sangamon soil at Fort Riley (Sumner Hill) does, however, seem to be well weathered and void of any major diagenetic alteration as indicated by coincidence of relatively high susceptibility and FD, relative to the Eustis ash pit data, for example (fig. 11).

Several recent studies have attempted to use rock magnetic parameters measured on loess and associated paleosols to quantitatively reconstruct past precipitation and/or temperature (*e.g.*, Heller *et al.*, 1991; Maher and Thompson, 1994a; Liu *et al.*, 1995). Liu and others (1995), for example, have found a strong correlation (r=0.948) between SP estimates in the most recent paleosol and current precipitation in a northwest-southeast transect in central China. Should similar regional trends in rock magnetic parameters (particularly SP and FD) in the central Great Plains which match the expected west to east increase in precipitation and nonmagnetic parameters of weathering intensity in soils, one could postulate a strong climatic control on the magnetic behavior of these units. On the other hand, a lack of correlation would imply other factors such as proximity to detrital sources of magnetic particles may be important in influencing the magnetic record.

At this point it is unknown as to what extent a quantitative link can be established between the amount of magnetic material and precipitation and/or temperature. The rock magnetic data from modern soils must first be examined for relationships with present precipitation, *i.e.*, to calibrate, if possible, the record from magnetic parameters using the present relationship between precipitation (and temperature?) and rock magnetic data (susceptibility, etc.) of the modern soils. Attention can then be focused upon the comparison with loess paleosol data.

The Regional Isotopic Record. Since the use of SIRA of carbon for paleoenvironmental reconstruction is fairly new, particularly its application to the late-Quaternary record of the central Great Plains, few data exist for the region. Values of $\delta^{13}C$ have, however, been determined for several samples from the Eustis ash pit and other loess exposures in the central Great Plains, primarily but not exclusively as a consequence of total humate ^{14}C age correction for the effects of isotopic fractionation (fig.12; Johnson *et al.*, 1993a, b).

The time of Gilman Canyon Formation soil development (c. 36-21 ka) had a larger proportion of C_4-type (warm, dry adapted) vegetation, comparable to the Pleistocene/ Holocene Brady soil and modern vegetation. Soil development and increased aridity are demonstrated by the elevated organic matter concentration, $\delta^{13}C$ values, and opal phytolith assemblages as expressed by an aridity index (figs. 8, 10). Therefore, vegetation and associated climate during Gilman Canyon time, an interstade, were similar to the present warm, semiarid/subhumid conditions of the central Great Plains. Transitions from the Gilman Canyon soil to periods of increased loess accumulation above and below it were clearly dominated by C_3-type (cool, moist- adapted) plants.

$\delta^{13}C$ values acquired from Peoria loess indicate that C_3 plants were dominant for most of the time of Peoria loess deposition, which demonstrates that cooling associated with the glacial maximum clearly occurred in early and middle Peoria time. During the glacial maximum (*c.* 20-18 ka, Peoria time) C_3-type vegetation dominated, followed by a gradual increase in C_4-type plants, culminating in the Brady soil. The trend in isotopic values during Brady time shows an increase, *i.e.*, early Brady soil development was characterized by more C_3-type conditions, whereas values increase (become more C_4 like) as the soil continues to develop into the early Holocene. Between about 12,000 and 9,000 yrs B.P., the climate and vegetation of central North America underwent dramatic changes (*e.g.*, Wright, 1970; Webb *et al.*, 1983). Spruce trees had been replaced by deciduous trees in northeastern Kansas, and deciduous trees persisted until about 9,000 yrs B.P. when grasslands expanded. It is clear that megafaunal extinction and dissolution of disharmonious faunas began about 12,000 yrs B.P., and the mesic conditions under which the Brady soil developed persisted until about 8000 yrs B.P., when the modern climate first appeared (Fredlund, 1995). Based on slight decrease in the $\delta^{13}C$ values from six sites in the region, climatic conditions shifted to more xeric conditions during the Brady time, a period of major landscape stability and pedogenesis. Further, changes in vegetation and faunal assemblages at this time reflect a shift to warmer and drier conditions with increased seasonality (COHMAP Members, 1988) and stronger zonal air flow at the surface (Kutzbach, 1987).

Isotopic values derived from middle Holocene soils (*e.g.*, c. 4-5ka) continue to reflect a higher proportion of C_4-type grasses. Data from the central Great Plains are, however, too limited to discern an Altithermal period as documented on the Southern High Plains (Holliday, 1989, Nordt *et al.*, 1994; Meltzer, 1996). If a middle Holocene drought period were environmentally expressed and recorded in the loessal sequences, then isotopic values should increase from the late-Pleistocene/early Holocene and then decrease from about 3ka or 4ka to the present. The most detailed $\delta^{13}C$ record of the entire Holocene to date comes from the Sargent site, located on the loess-mantled bluff along the south side of the Platte River valley in southwestern Nebraska (Bozarth and Dort, 1996). These data are, however, inconclusive due to high variance and low resolution; a high-resolution resampling and analysis is scheduled for late 1996 or early 1997.

Based on the composite regional data, changes in plant $\delta^{13}C$ values are sufficiently large as to clearly demonstrate major shifts of C_3/C_4 ratios in soil organic matter during the late Wisconsin and

Holocene. There is, however, a need for high-resolution late-Pleistocene and Holocene SIRA of carbon from high-quality sites in the region, and Fort Riley offers the potential for such sites.

The Fort Riley Record. The composite magnetic record for the reservation is consistent with the regional records. During the Last Interglacial, *i.e.*, Sangamon time, a pronounced period of soil development occurred on the reservation, as elsewhere throughout the region. Within the Sangamon soil, as exemplified at the Sumner Hill site (fig. 2), susceptibility values are relatively high (*c.* 150); FD values are even more pronounced and better define the stratigraphic extent of Sangamon pedogenesis. SIRA of bulk carbon content from the Sumner Hill site (fig. 7) indicates elevated values for the Sangamon soil, suggesting an increased role of C_4 plants, presumably due to increased temperature and/or moisture stress. Isotopic values for the Sangamon period are not as high as for the subsequent major soil forming periods; it is unknown if this is the product of real environmental differences, of contamination by overlying sediments, or of other as of yet unappreciated processes.

Soil formation during Gilman Canyon time is well expressed in the magnetic record at all sites except Manhattan Airport, an alluvial rather than eolian sequence. On the basis of magnetic signals from the Bala Cemetery and Pump House Canyon sites, the Gilman Canyon soil is better developed than the surface soil. The Gilman Canyon soil is, as noted above, best described as a composite geosol in that pedogenesis was diminished somewhat during formation of the soil by an acceleration in the rate of loess fall, albeit still very low. As a result, the magnetic signal (susceptibility and FD) typically have a bimodal character; this bimodality is present at the Sumner Hill and Bala Cemetery sites, but not at the Pump House Canyon site because the sampling apparently stopped (trench bottom) in the minimum between the two pedogenic modes. Pedostratigraphy is somewhat more complex at the Sumner Hill site: soil development is evident between 6 and 8m on the basis of susceptibility and particle size (fig. 2). This appears to be a function of an expanded Gilman Canyon soil, *i.e.*, a magnetic mode between 4 and 5m and another between 6 and 7m. An alternative, but less likely explanation is that the Sangamon soil is extremely thick, extending from about 11m to 6m. A detailed magneto- and chronostratigraphy from the Eustis ash pit indicates that the increased rate of loess accumulation/reduction in pedogenic intensity occurred from about 25ka to 30ka, an interval of about 5,000 years (Johnson et al., 1993a). Without detailed close-interval absolute age data from Fort Riley, it is uncertain if the interval occurred at the same time as in Nebraska. SIRA data from Sumner Hill suggest that Gilman Canyon time was about the same as Sangamon time in terms of vegetation composition (C_3/C_4 ratio) and not as warm and/or dry as any of the Holocene, including the present. The bimodality of the Gilman Canyon SIRA data corroborates the magnetic data.

Peoria loess deposition occurred at a relatively rapid rate as to preclude any significant pedogenesis. Low susceptibility and FD values (c. 40 and 2, resp.) at the Bala Cemetery site reflect the rapid rate of deposition and lack of significant pedogenesis. The increase in susceptibility and FD that occurs in the Peoria loess at about 1-1.5m depth does, however, suggest one or more brief periods of soil development. The absolute timing of the pedogenesis/surface stability is unknown but is probably centered around the Last Glacial Maximum (*c.* 18ka); a moderately developed glacial maximum soil has been identified elsewhere in the region (*e.g.*, Arbogast, 1995; Arbogast and Johnson, 1996). Unfortunately the Peoria loess is not preserved in an unweathered state in any other of the sites examined, thereby precluding a local corroboration of the glacial maximum soil.

The Brady soil, while not expressed at the Bala Cemetery site, is well exhibited at the Sumner Hill, Pump House Canyon, and Manhattan Airport sites. Of the three sites, Sumner Hill shows the weakest magnetic expression and Manhattan Airport, the strongest. At the former, the loess accumulation rate on this bluff line adjacent to the Kansas River valley was probably too rapid to permit strong soil development, whereas soil development in the alluvial environment of the latter site progressed relatively rapidly in this presumably well-watered and high-biomass location, *i.e.*, both magnetic parameters are elevated. From the magnetic signal, Brady soil development at Pump House Canyon appears intermediate in development, *i.e.*, approximately equivalent to the modern surface soil, with a moderate to strong response in both magnetic parameters (fig. 5).

Deposition of Bignell loess occurred at a faster rate than that of the Peoria loess and reflects more the microclimatology of a particular region, principles well illustrated at Fort Riley. As the southerly winds of the Holocene evolved, loess was lofted from the Kansas and Republican River flood plains onto the adjacent uplands. Consequently, the Sumner Hill and Pump House Canyon sites contain a Bignell loess component; Bala Cemetery site is too far removed from riverine environments to the south, and Manhattan Airport site has experience the accumulation of Holocene alluvium, not loess. Both the Sumner Hill and Pump House Canyon sites have 2-3m of Bignell loess, with the loess being slightly coarser at the Pump House Canyon site due to its lower elevation, *i.e.*, closer vertical proximity to the alluvial bottoms.

Evidence of the Altithermal, or middle Holocene dry episode appears at both sites, but is best articulated at Pump House Canyon. The depressed susceptibility and FD values during the middle Holocene are indicative of an increased rate of deposition and decreased intensity and/or time of weathering. SIRA values show a discernable but less developed "sag" in the middle Holocene, indicating a decrease in the C_3/C_4 ratio (fig.7). Evidence of pedogenesis still appears, particularly at Pump House Canyon. One of these middle Holocene soils has been dated to 5,810 yr B.P. (table 1), a time of major Holocene soil development in both the alluvial and upland environments elsewhere in the region (*e.g.*, Johnson and Martin, 1987; Johnson and Logan, 1990; Mandel, 1995; Johnson *et al.*, 1996a, b; Arbogast and Johnson, 1996). Although probable, this interpretation is not absolute in that neither site has the resolution of ^{14}C dating needed for a reasonable level of confidence, *e.g.*, the Brady soil identified at the Pump House Canyon site is an interpretation based on physical soil attributes and the intensity of the magnetic signal, but not on absolute age data.

Despite shortcomings in the data, it is apparent that the Holocene environment associated with

Fort Riley was significantly different from the Pleistocene. Magnetic data (especially FD) and SIRA data from Sumner Hill mirror one another, *i.e.*, magnetic and nonmagnetic approaches representing disparate responses to environmental conditions produce similar patterns (fig. 13). In sum, when the loess deposition rate is relatively high, weathering (magnetic carrier concentration) and SIRA values are relatively low. Conversely, when loess depositional rates are relatively low, weathering/ pedogenesis (*i.e.*, magnetic parameters) and SIRA values are elevated.

An examination of the available data also seems to indicate that at most sites soil development began gradually, as reflected in the relatively slow increase in the magnetic and SIRA curves. The cessation of pedogenesis, however, occurred abruptly, *i.e.*, the magnetic and SIRA curves tend to drop in value relatively rapidly. Moreover, the rate of loess deposition decreased more slowly than it increased. This behavior is consistent with contemporary climatic and geomorphic response models, and with regional magnetic and SIRA data.

When SIRA data become available from the University of Texas Radiocarbon Laboratory, the interpretations will most certainly be enhanced. The existing Sumner Hill SIRA data curve (Johnson *et al.*, 1994) will be extended to include the lower Sangamon soil and thin underlying layer of alluvium overlying the strath. Further, data from Pump House Canyon, Bala Cemetery, and Manhattan Airport will be correlated with the magnetic data to ascertain the extent to which the pattern established at Sumner Hill persists and relates to the former.

Although few in number and at scattered stratigraphic levels, $\delta^{13}C$ values determined for the purpose of correcting ^{14}C ages for isotopic fractionation (table 1) provide a limited insight into the variation in climatic conditions as reflected in the vegetation. Late-Pleistocene ^{14}C ages tend to have smaller isotopic values (i.e., more C_3 influence) than do those derived from Holocene age sediments, *e.g.*, -15.6 ‰ for the middle Holocene but -16.5‰ or less for the late Pleistocene at Manhattan Airport.

Summary

Climatic change of the late-Quaternary period has been recorded in the loess deposits of the central Great Plains, and the record of such change is extractable using a number of approaches and parameters. The stratigraphy of loess deposits which have been investigated on Fort Riley exhibits the same sequence of loess units and intercalated buried soils as is found elsewhere in the region, but adds detail unique to the paleoenvironments of the fort. Upland late-Quaternary composite stratigraphy preserved on the reservation consists of the basal Sangamon soil of the Last Interglacial (Illinoian age; *c.* 120-110ka), Gilman Canyon Formation (a pedogenically-altered, middle Wisconsinan loess; *c.* >40-20ka), Peoria loess (late-Wisconsin; *c.* 20-10ka), Brady soil (Pleistocene/Holocene transition; *c.* 11-10ka), Bignell loess (Holocene; *c.* 9-?ka), and modern surface soil.

The loess sequences at Fort Riley represent a nearly continuous time series of climatically-forced environmental change for the late Quaternary. Application of two analytical techniques, magnetics and SIRA has provided proxy data sets that represent a time series of climatically regulated pedogenesis/weathering and botanical composition.

Study localities included two upland sites, Sumner Hill and Bala Cemetery; a valley-wall site, Pump House Canyon; and a valley-bottom site, Manhattan Airport. Sumner Hill, a bluff top site on the north side of the Kansas River valley, has the thickest loess accumulation of all sites examined (*c.* 13m). It extends from the Sangamon soil to the surface soil, whereas Bala Cemetery site, in the extreme northwestern part of the reservation, consists of less than 3m of loess above the limestone bedrock. Pump House Canyon site has a relatively thick accumulation of loess, probably in excess (Gilman Canyon and Peoria loesses) of 5m. Sediments at the Manhattan Airport are alluvial, not eolian, and exhibit the alluvial phase of two or more of the soils found in the upland record (Brady and Sangamon and/or Gilman Canyon).

Magnetic data have yielded an impression of the variation in climate from Sangamon time to the late Holocene through a reconstruction of the history of pedogenesis/weathering. Although the SIRA analyses for this study are yet pending with the contracted laboratory, the SIRA curve obtained by Johnson and others (1994) from a core extracted at Sumner Hill has been temporally calibrated with C^{14} ages. These data portray the climatic and environmental record as shifts between cool season (C_3) and warm season (C_4) plants, primarily grasses.

Sangamon soil formation dominated the reservation during the Last Interglacial as indicated by magnetic parameters; moreover, SIRA reveals a relatively important role of C_4 plants. During Gilman Canyon time, loess influx was usually sufficiently slow as to permit pedogenesis, which appears to have been at a maximum twice during that time. Warm season grasses were important during soil formation, but diminished in importance during the periods of more rapid loess fall, which were cooler and perhaps wetter. Peoria loess fall, a function of the deterioration of climate during the Last Glacial Maximum, thinly blanketed the reservation, with thickest accumulations occurring to the northwest (*e.g.*, Bala Cemetery site), proximal to the source region. Long-term surface stability did not apparently occur within Peoria time, but short-term stability (reduced rate of loess fall) may be indicated by the presence of thin weathering zones (incipient soils) in the Peoria loess. Regional landscape stability prevailed during the environmental shift at the Pleistocene/Holocene transition, resulting in formation of the well expressed Brady soil. As Brady time progressed, warm season grasses became increasingly important, thereby setting the stage for Holocene grassland expansion. Bignell loess was apparently deposited by the southerly winds of the Holocene and is thickest in closest proximity to the source, *i.e.*, the Kansas River valley. One or more weak soils did, however, develop in the Bignell loess as it accumulated. A notable feature of the Bignell loess is the appearance of the Altithermal dry period: the loess experienced little weathering and was dominated by warm season grasses until the latter part of the Holocene.

Recommendations

Based on the results of this and past studies, it is apparent that four additional research initiatives should be conducted in order to realize the potential of the upland paleoenvironmental record and to generate a geoarchaeological model of sufficient temporal and spatial resolution to provide the basis for development and/or testing of the computer-based geoarchaeological model under development by R. Bras and colleagues at MIT. The recommended research projects include the following:

1) In that the latest Pleistocene and entire Holocene are of primary interest in development of a cultural resources and landscape evolution model, further emphasis should be placed on this portion of the upland stratigraphic record. The environmental transition at the late-Pleistocene/Holocene boundary needs to be better characterized, as does the detail of the Holocene (*e.g.*, the Altithermal, soil-forming periods). Specifically, high-resolution (*e.g.*, 2cm contiguous) sampling should be undertaken on two or more new sites with expanded Holocene sequences and on cores from the existing sites of Sumner Hill and Pump House Canyon for the purpose of SIRA of carbon. Further, these sites should be sampled at close interval (*c.* 25cm) for ^{14}C dating; subdividing the samples in order to individually date the total humates and humic acid and humin fractions would be helpful in deciphering the chronostratigraphy (Johnson and Valastro, 1994). Magnetic analyses (susceptibility and frequency dependence) should also be conducted at the new sites selected.

2) The analysis of trace element content has been used in a number of applications related to the reconstruction of late-Quaternary environmental conditions (*e.g.*, Muhs *et al.*, 1990, 1994; Reheis, 1990; Diekmeyer, 1994). Trace elements (e.g., cerium, strontium, yttrium, zirconium) contained with in the loess and intercalated soils provide a "fingerprint" of the sediment and provide information regarding paleowind direction and loess source region. Data obtainable include not only presence or absence, but also the concentration of individual trace elements. Recent research by Johnson and others (1993) have documented the difference in source regions for loess deposits of eastern Colorado versus those of southern Nebraska and Kansas. For Fort Riley, two levels or orders of information can probably be derived from the loess sequences. First, trace element analysis can indicate the stratigraphic level (and time, with sufficient ^{14}C control) at which the winds shifted from the north-northwest to the south-southwest, *i.e.*, determine whether the trace element signature is that of the Platte River sediments or of the Kansas River sediments. The second order variations in the data will document the more subtle changes within the trace element time series. These variations will indicate small order changes in wind intensity and direction at a given location, as well as define the

distance-decay function along a sampled transect. The stronger the winds, the higher the concentration of the trace elements, since they are not only rare but also represent the heavy mineralogic fraction. Short-term changes in direction will also show up in the signature. Trace element analyses should be conducted on two or more of the sites referred to in the first recommendation, with sampling being conducted at close interval (*c.* 2cm). Previous research (Johnson et al., 1993b, Diekmeyer, 1994, Muhs and Johnson, 1996) has indicated that a suite of eleven trace elements contains the optimal amount of information for examining Pleistocene and Holocene loess of the central Great Plains. Facilities for preparation of samples for analysis exist within this investigator's laboratory, and those for analysis exist both in the Department of Geology, University of Kansas, and at the U.S. Geological Survey, Denver, Colorado (D. Muhs' laboratory).

3) Extensive archaeological surveys currently being conducted on the reservation are yielding new upland prehistoric sites, one or more of which have been found to contain evidence for Clovis (latest Pleistocene/earliest Holocene) occupation. Selected archaeological sites should be analyzed to articulate the specific paleoenvironmental record. As a corollary to the first recommendation, these sites, depending on their location relative to the nonarchaeological sites, should receive high-resolution magnetic, isotope, opal phytolith and perhaps trace-element analyses.

4) Results of D.L. Johnson's survey of the upland loess mantle (D.L. Johnson, 1996) should be used as base line data in developing a detailed map of surface materials in order to provide the precision of information needed to articulate a detailed sequence of landscape evolution on Fort Riley. A combination of shallow coring and ground-penetrating radar surveys can be employed to provide the information for input into a GIS in order to create a detailed map of surficial geology and the spatial variation in thickness of the various map units. Shallow coring can be conducted using the trailer-mounted Giddings soils probes operated by the Department of Geography, University of Kansas. Ground-penetrating radar, conducted by dragging a skid-mounted radar system behind an all-terrain vehicle, and personnel are available through both the Department of Geology, University of Kansas and the Kansas Geological Survey.

Research conducted to date by this researcher (W.C. Johnson *et al.*, 1994; this report) and by D.L. Johnson (D.L. Johnson, 1992, 1994, 1996) has defined the late-Quaternary lithostratigraphy of the reservation and documented the first-order trends in the chrono-, magneto-, and biostratigraphy (SIRA, opal phytoliths). Now that the potential for reconstruction of the late-Quaternary environmental record has been realized and broadly defined, high-resolution study needs to be undertaken at selected existing and new sites in order to extract the interpretable detail that exits.

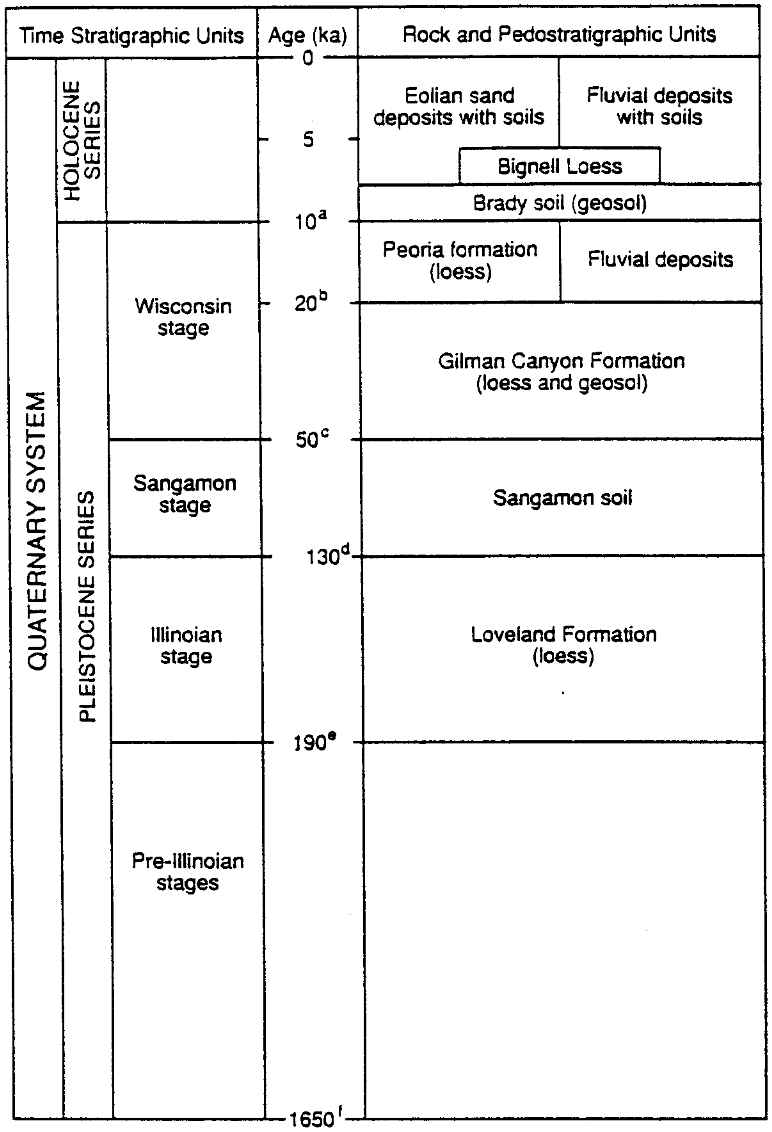

Figure 1. Late-Quaternary stratigraphic succession for the nonglaciated part of Kansas. Age designation rationale consists of the following: Pleistocene-Holocene boundary after Hopkins (1975); (b) the uppermost Gilman Canyon Formation ^{14}C dates at about 20ka (Johnson, 1993a, b); (c) the age of 50ka is an estimate for the beginning of Gilman Canyon time, not the end of Sangamon soil development, *i.e.*, the boundary is believed to be an unconformity; (d) 130ka is consistent with the sea-level record and the age of marine isotope boundary 5e/6, the presumed end of the Illinoian glacial stage and with recent themoluminescence age determinations in southcentral Nebraska and central Kansas (Maat and Johnson, 1996; Johnson and Muhs, 1996); (e) age of the marine isotope boundary 6/7 as determined by Martinson and others (1987); and (f) age of the Pliocene-Pleistocene boundary at the Virca, Italy section (Aguirre and Pasini, 1985).

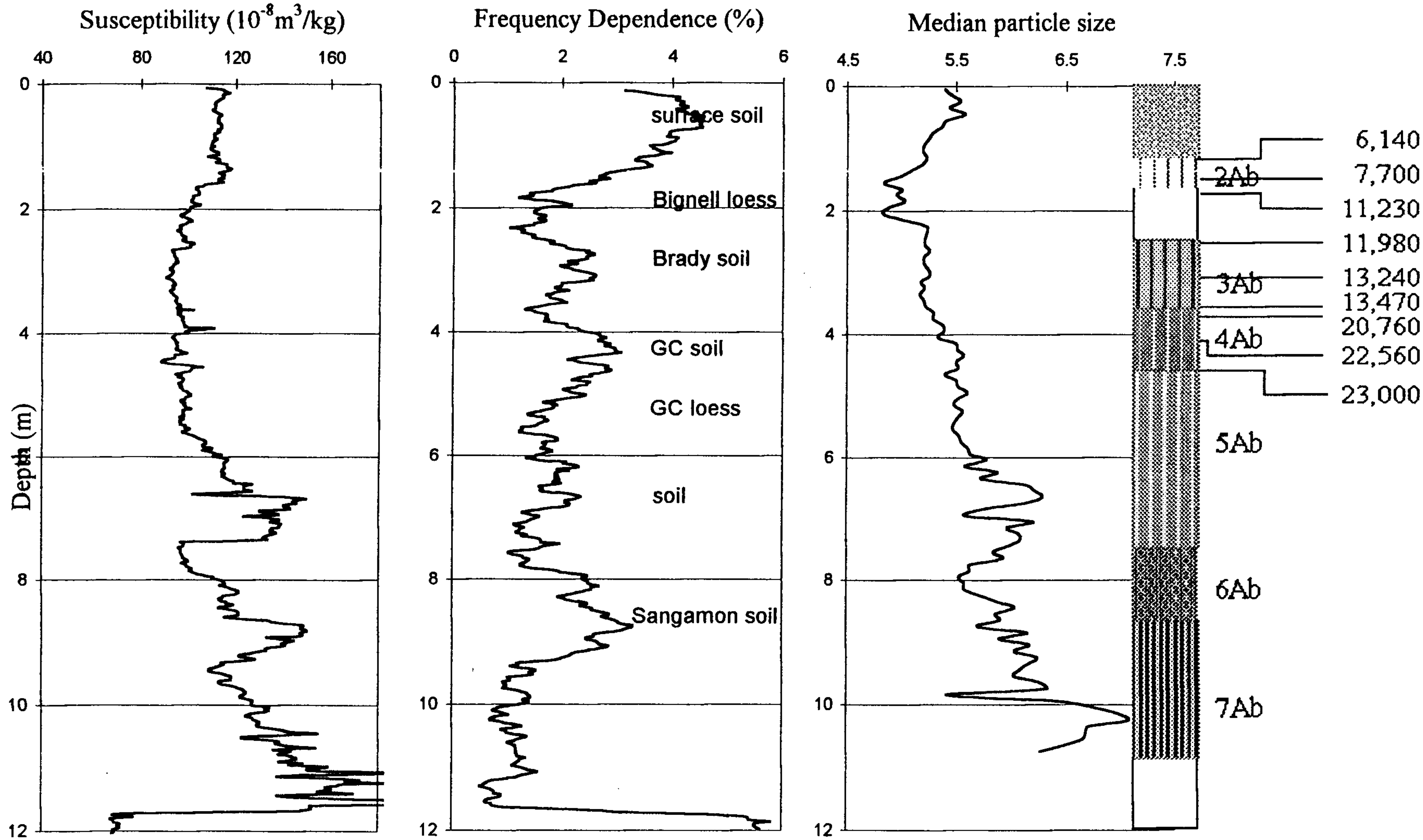

Figure 2. Magnetic susceptibility and frequency dependence data and median particle size data (ϕ scale; fining to right) for the profile sampled at the Sumner Hill site (no. 21).

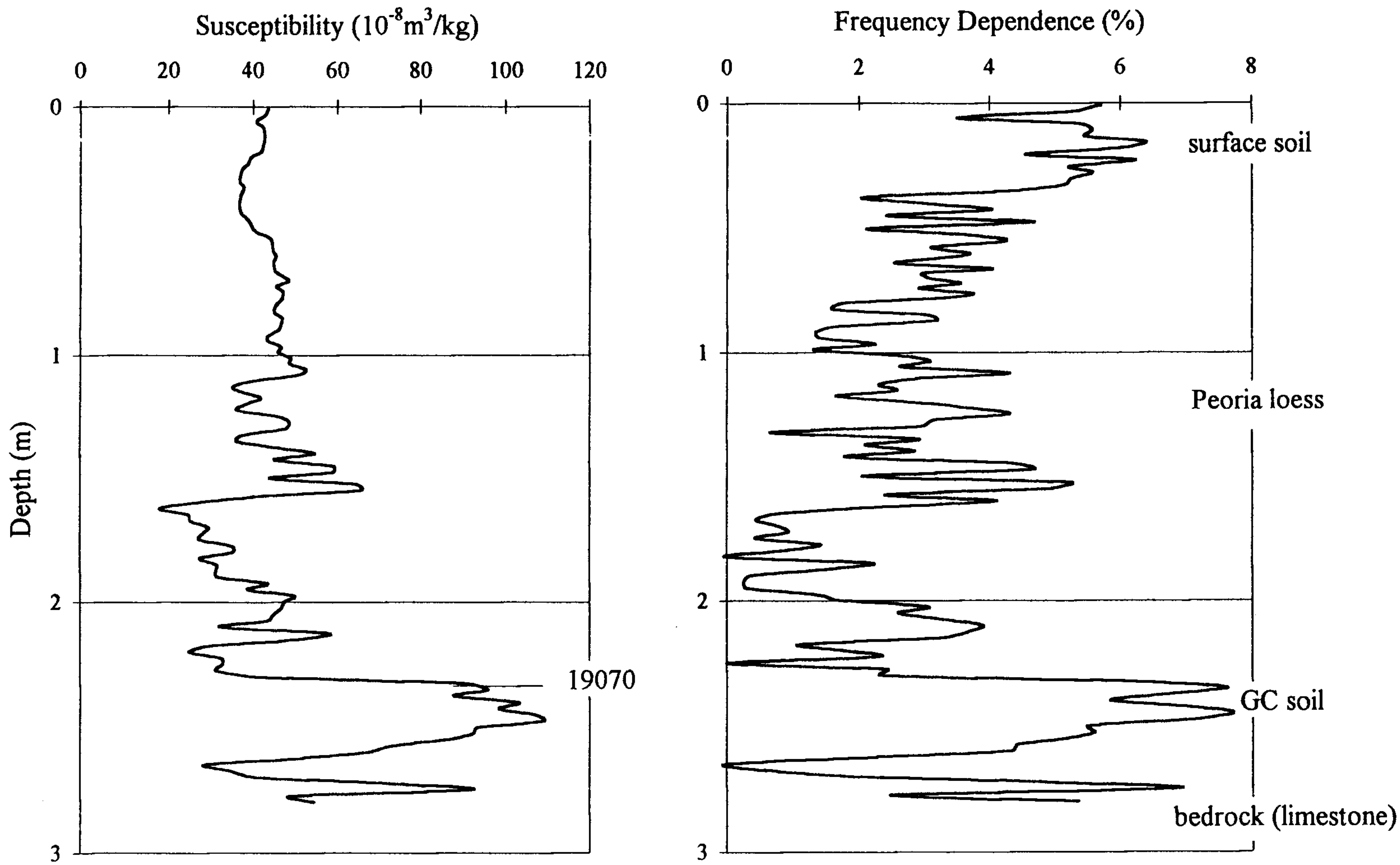

Figure 3. Magnetic susceptibility and frequency dependence data from the core extracted at the Bala Cemetery site (no. 19).

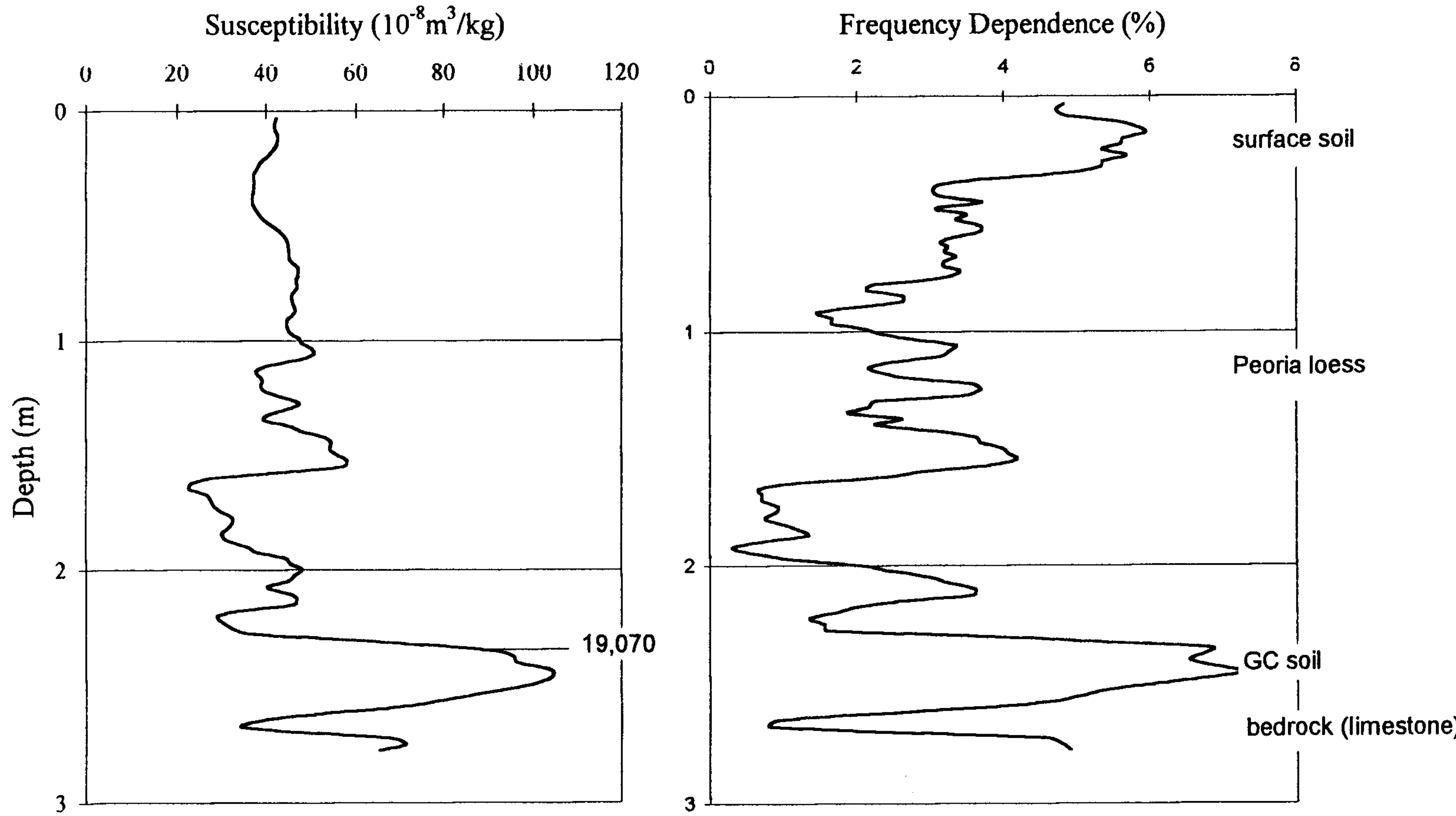

Figure 4. Magnetic susceptibility and frequency dependence data from the core extracted at the Bala Cemetery site (no. 19); data have been smoothed.

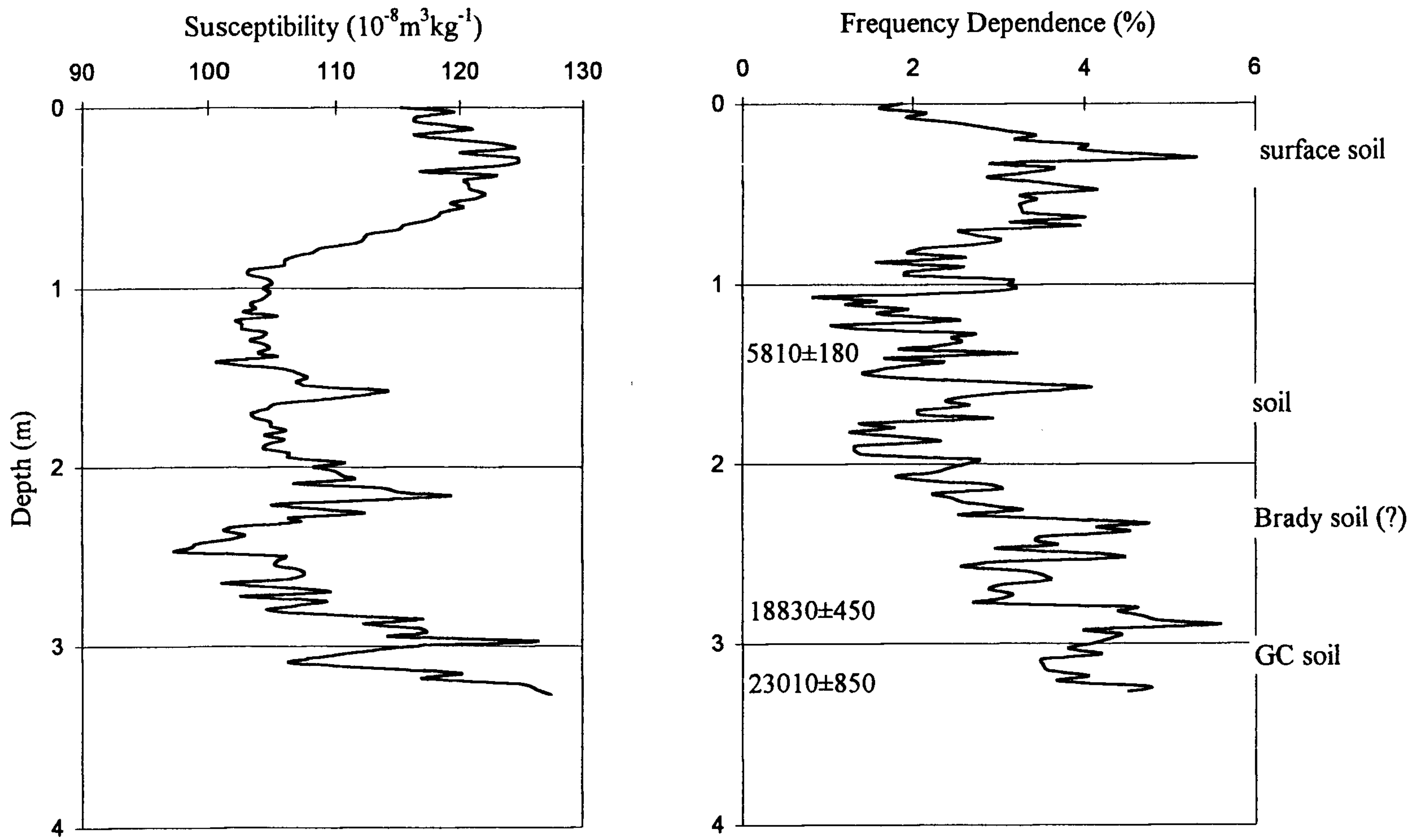

Figure 5. Magnetic susceptibility and frequency dependence data from the trench profile at the Pump House Canyon site.

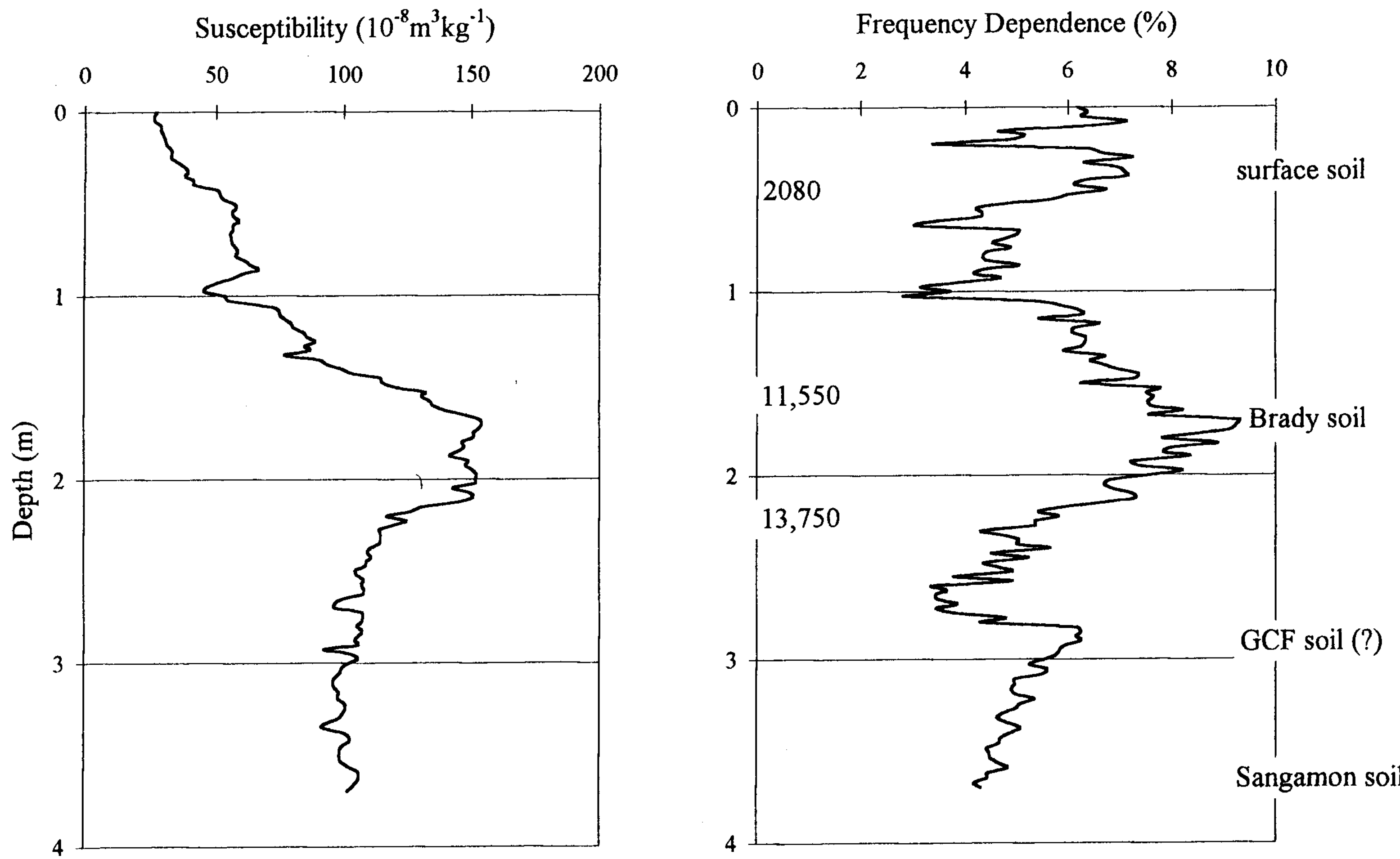

Figure 6. Magnetic susceptibility and frequency dependence data from the trench profile at the Manhattan Airport site (no. 33).

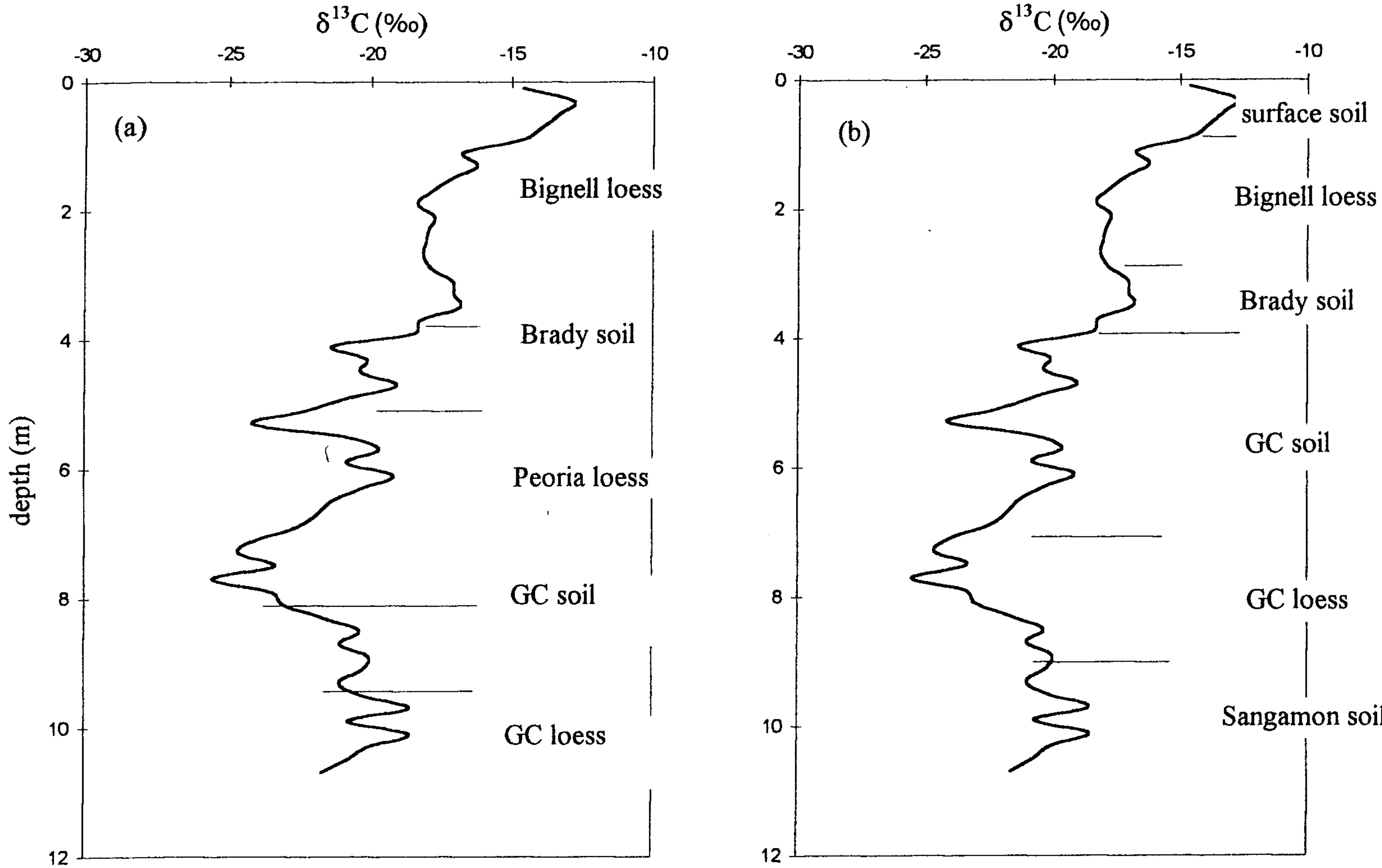

Figure 7. The $\delta^{13}C$ (‰) curve assembled from the core extracted at the Sumner Hill site (no. 21); *a:* with the stratigraphic interpretation made prior to ^{14}C dating of the site (Johnson et al., 1994); and *b:* with the revised stratigraphic interpretation (this study).

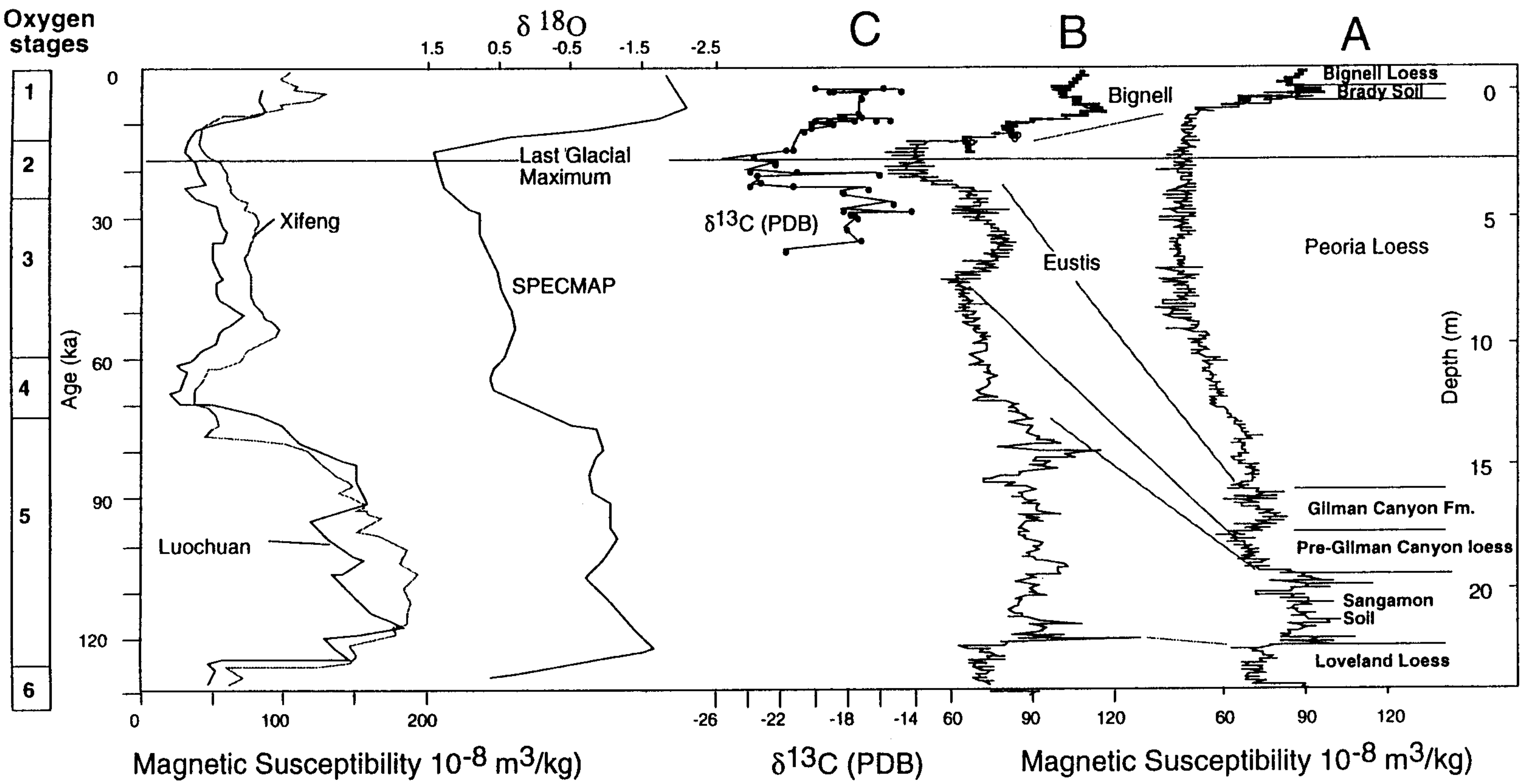

Figure 8. Comparison of the stratigraphic depth variation (A) and interpolated time variation (B) of magnetic susceptibility at the Eustis ash pit and Bignell Hill section in southwestern Nebraska showing positive correlation with SPECMAP data of Imbrie and others (1984). The time curve (B) for the Gilman Canyon through Bignell loess was constructed from ^{14}C ages at the two sites. Temporal data for the lower half of the curve were based on thermoluminenescence ages from the Sangamon soil as well as on the apparent fit with the SPECMAP curve. $\delta^{13}C$ values (C) from humates versus age for various Kansas and Nebraska sites compare well to magnetic susceptibility versus age for the Eustis ash pit.

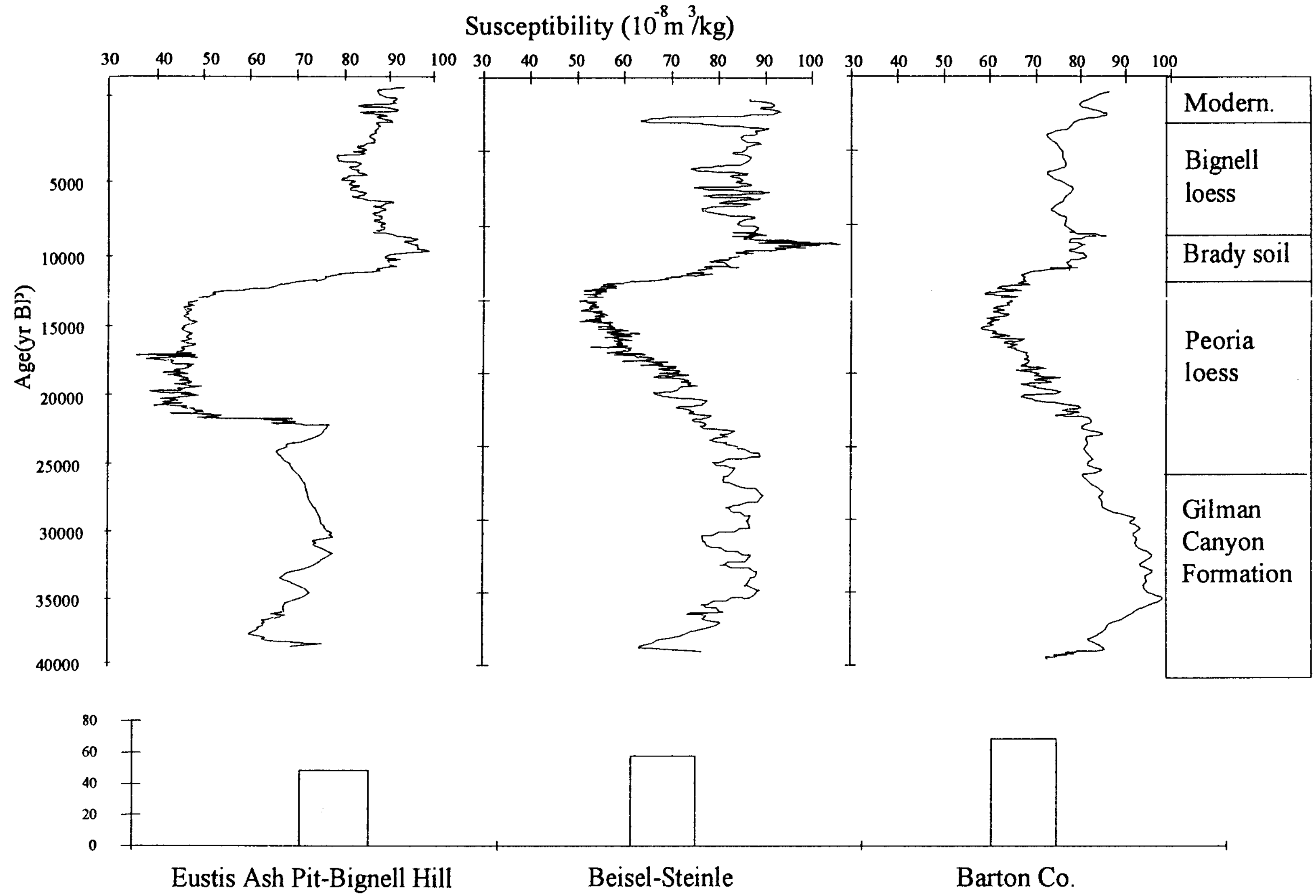

Figure 9. Comparison of magnetic susceptibility data from the Eustis ash pit-Bignell Hill composite section (southwestern Nebraska), Beisel-Steinle site (Russell County, Kansas), and Barton County (Kansas) sanitary landfill section (top), and regional variation of mean susceptibility within the Peoria loess (bottom). Note the systematic southward decrease in susceptibility evident in the latter.

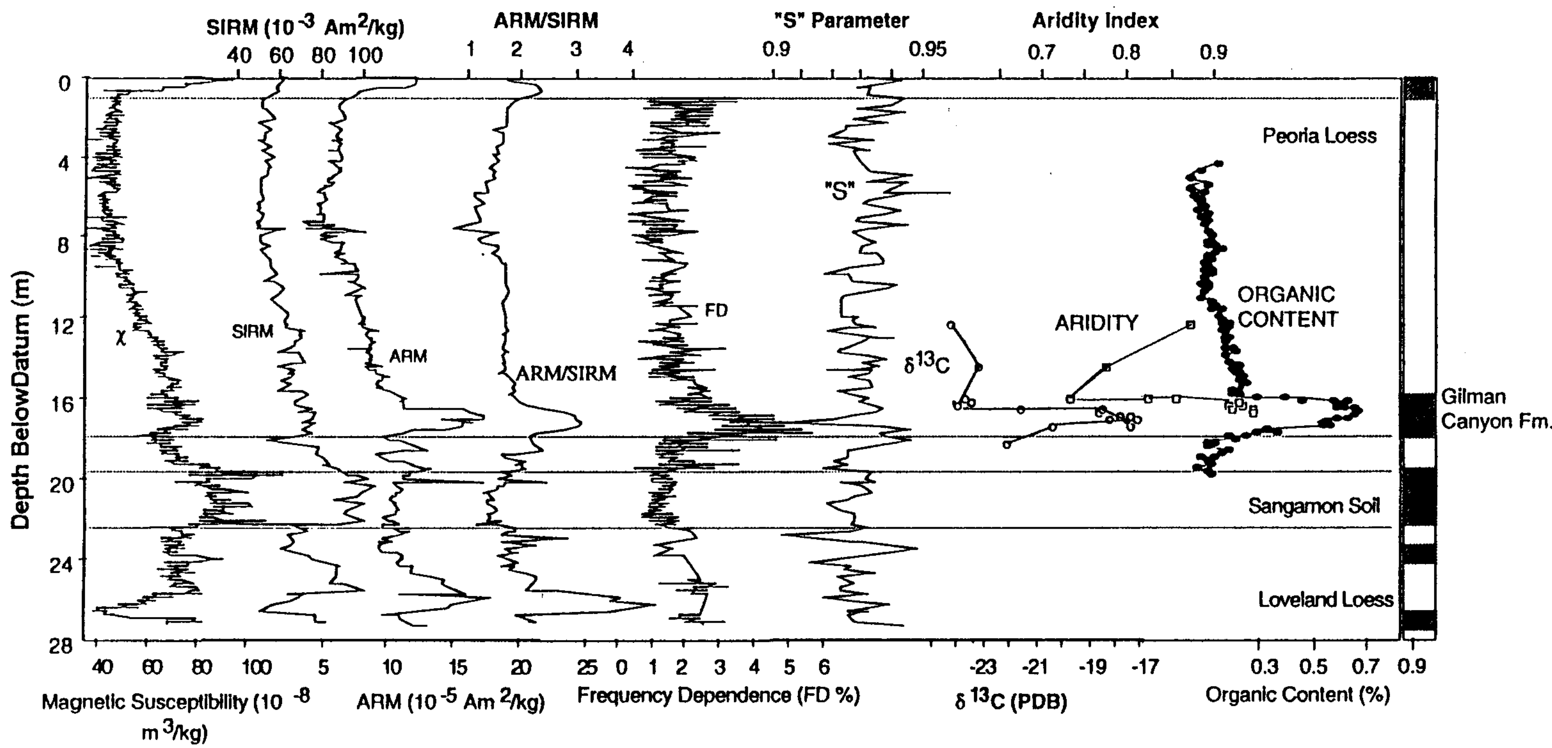

Figure 10. Stratigraphic plots of magnetic parameters (χ, SIRM, ARM, ARM/SIRM, FD, and S) compared with stratigraphic plots of $\delta^{13}C$ values derived from bulk organic carbon, an aridity index derived from opal phytolith data, and percent organic matter, all from the Eustis ash pit in southwestern Nebraska.

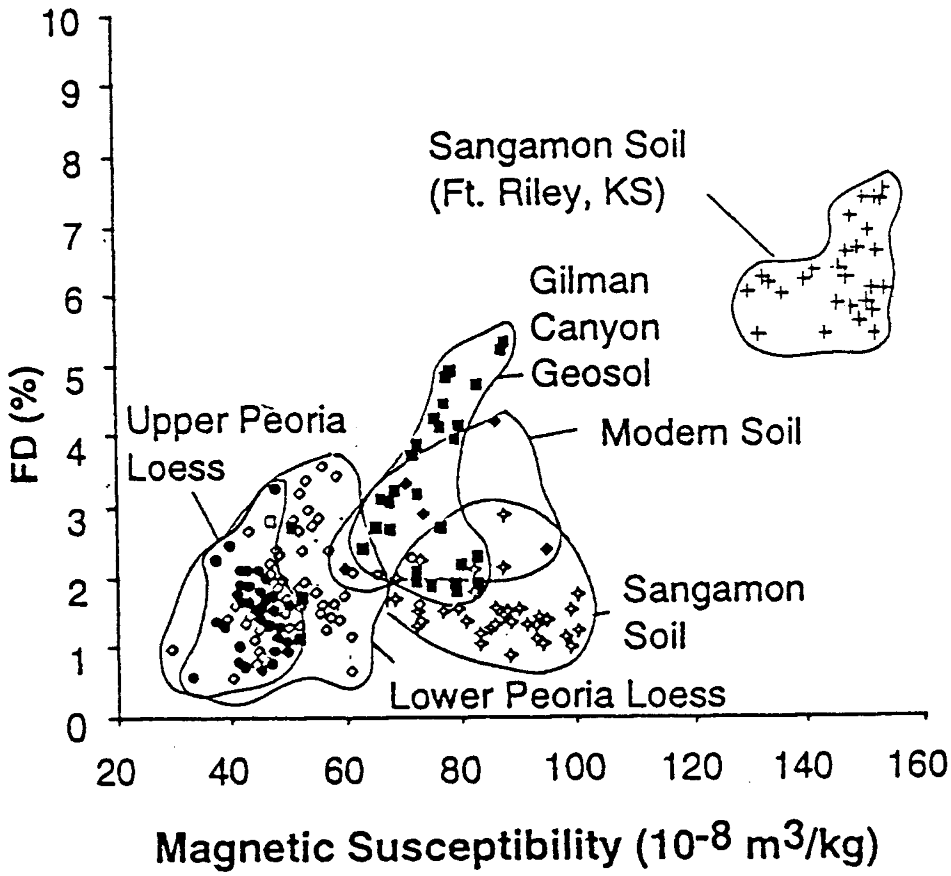

Figure 11. Magnetic susceptibility versus FD by stratigraphic unit for samples from the Eustis ash pit and for the Sangamon soil at Sumner Hill, Fort Riley. The Sangamon soil at the Eustis ash pit is anomalously low in FD, whereas samples from Fort Riley indicate elevated susceptibility and FD.

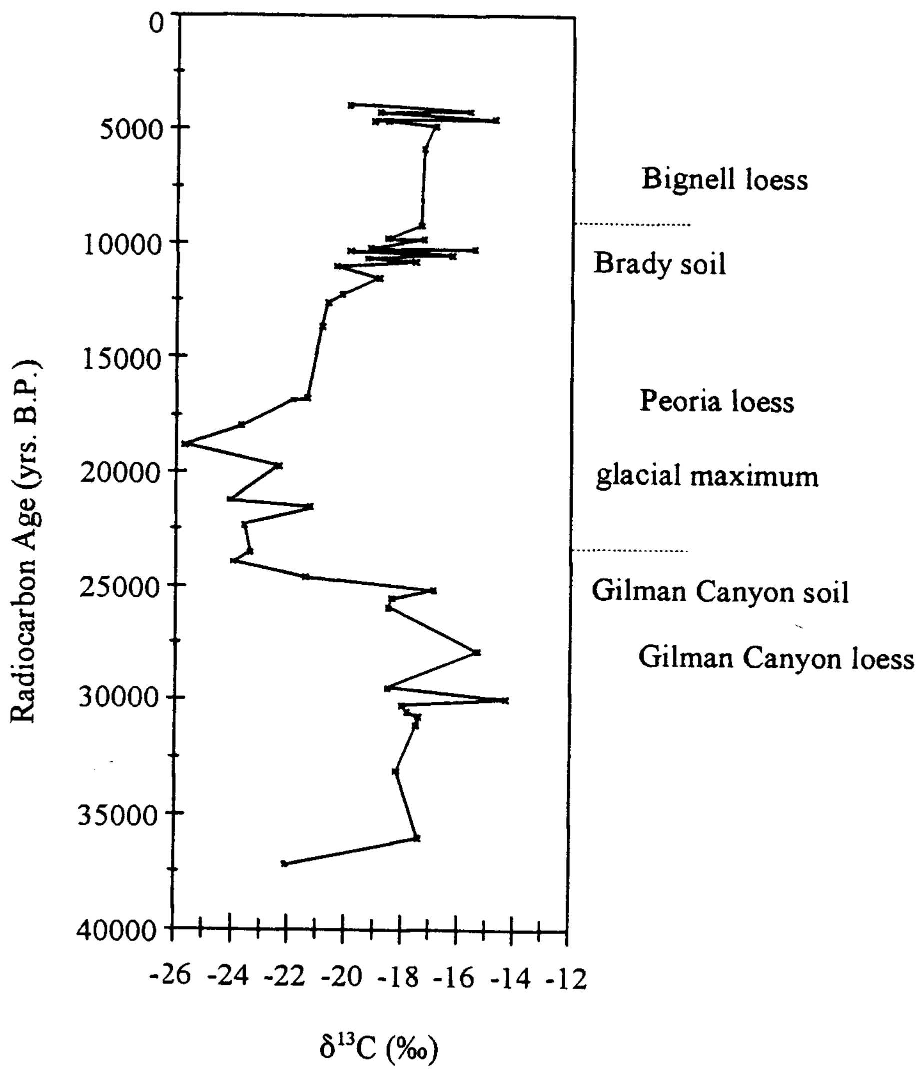

Figure 12. Composite $\delta^{13}C$ curve for the central Great Plains (Johnson et al., 1993a, 1994).

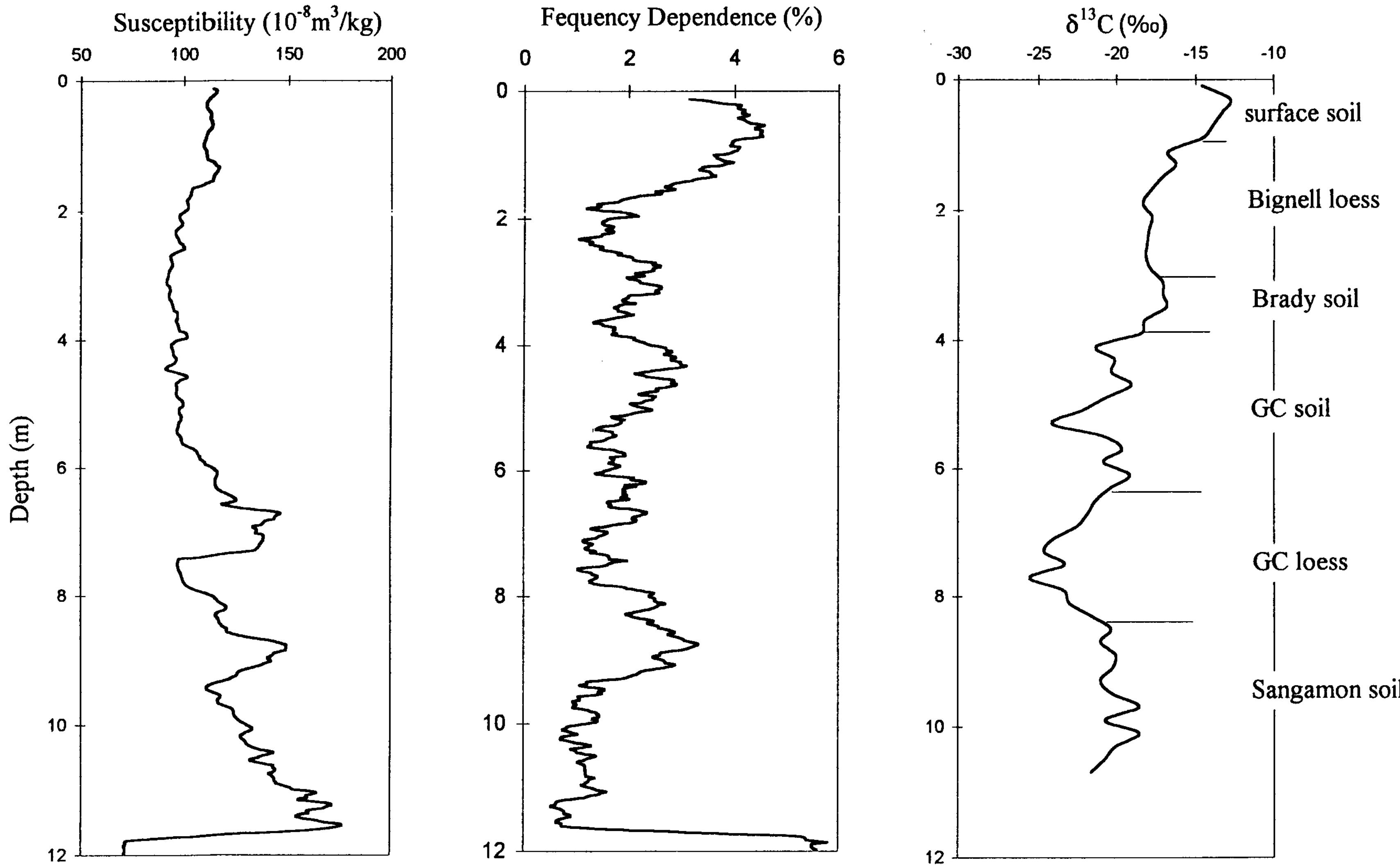

Figure 13. Magnetic susceptibility, frequency dependence, and $\delta^{13}C$ curves, and stratigraphic units for the Sumner Hill site (no. 21).

Table 1. Radiocarbon (^{14}C) Ages.

Site Name (Number)	Depth(cm)	ISGS no.[1]	$\delta^{13}C$ (‰)	Corrected Age (yr BP)[2]
Sumner Hill (21)[3]	150-160/pit 6	3101	-16.2	6,140±130
	170-180/pit 6	3140	-17.1	7,770±220
	200-210/pit 6	3142	-16.8	11,230±310
	240-250/pit 6	3131	-16.2	11,980±270
	260-270/pit 6	3143	-15.9	13,240±310
	280-290/pit 6	3146	-15.6	13,470±430
	215-225/pit 2	3164	-15.5	19,170±650
	310-320/pit 5	3159	-16.0	20,760±750
	255-265/pit 2	3160	-15.4	21,580±740
	235-245/pit 2	3163	-15.1	21,610±590
	330-340/pit 5	3158	-17.0	22,560±990
	350-360/pit 5	3157	-17.2	23,000±1,100
Bala Cemetery (19)	150-220	2622	-17.3	19,070±280
Pump House Canyon	130-140	3165	-15.6	5,810±180
	290-300	3167	-21.6	18,830±450
	330-340	3166	-18.4	23,010±850
Manhattan Airport (33)	40-48	2996	-14.4	2,080±70
	163-168	2997	-20.4	11,550±440
	525-528	3004	-16.5	11,580±570
	228-233	3003	-18.3	13,740±540
(20)	168-240	2623	-19.4	19,990±450
No name (28)	23-28	3005	-15.5	1,000±100
	28-33	3006	-15.4	1,730±110
	147-152	3008	-19.0	14,140±770
	202-207	3007	-17.3	15,110±590

1 Illinois State Geological Survey Radiocarbon Laboratory number.
2 Radiocarbon age corrected for the effects of isotopic fractionation (Taylor, 1987).
3 Pits 2 and 5 overlapped stratigraphically, and the same unit (Gilman Canyon Formation) was dated in both pits, but at slightly different subunit stratigraphic positions.

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