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Temporal and Spatial Variations in the Loess Depositional Environment of Central Kansas During the Past 400,000 Years

By

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ABSTRACT

The loess sequences in central Kansas are characterized by variations in their chemical and physical properties. The variations are largely attributable to climatic changes. Based on the physical and chemical properties and radiometric ages, a temporal sequence of climatic change is reconstructed.

The Loveland Loess was deposited from 415 to 95 ka at a consistently low rate (0.03 mm/yr). Four cycles of pedogenic $CaCO_3$ accumulation occurred within the loess: 415-325, 325-250, 250-195, and 195-95 ka. The four $CaCO_3$ peaks correspond chronologically to marine oxygen-isotope stages 11, 9, 7, and 5, respectively. Weathering indices characterizing leaching and oxidization and magnetic susceptibility indicate that weathering of Loveland time was weaker during warm periods than during cold periods.

The early Late-Pleistocene (95-70 ka) was characterized by sand dune activity exhibiting three episodes of weathering. The morphologically defined Sangamon pedocomplex, a reddish soil, underlies the Gilman Canyon pedocomplex and formed from approximately 70 to 30 ka under relatively warm and moist climatic conditions with a very slow rate of deposition (0.016 mm/yr).

The Gilman Canyon pedocomplex, dated at 31-20 ka, was formed under a strong physical weathering regime and a relatively high rate of deposition (0.15 mm/yr), suggesting high effective moisture and relatively low temperature. The overlying Peoria Loess was deposited at a rate of 0.3 mm/yr in central Kansas under dry and cold conditions.

The Brady Soil, dated at 10.5-8.5 ka, has high carbon content, weathering indices and magnetic susceptibility, suggesting that the early Holocene was the optimal time during the last 20,000 years for soil to develop. The overlying Bignell Loess, which was probably deposited during the Altithermal Period (8.5-6 ka), is poorly weathered with a relatively high content of sand.

Sand sequences in the Great Bend Sand Prairie of central Kansas demonstrate that sand activity persisted from the end of Brady time (8.5 ka) to about 5.8 ka, when an accretionary soil developed while sand dunes were relatively stable. Soil development occurred in the sand from 5.8 to 4.1 ka. The sand dunes have been active from 4.1 ka until the present, with four evident hiatuses around 2.9, 1.6, 0.8 ka and the time of surface soil development. Surface soils, probably formed in recent centuries, are presently buried by blow sand, implying historic degeneration of the environment.

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CHAPTER I. INTRODUCTION

Importance of Loess Study

The study of Quaternary environments is without question one of the most complex fields of natural science because inevitably it involves the integration of information from several physical and biological disciplines. When grappling with the complexity of the Quaternary Period, investigators frequently seek to ease frustration by assuming simplicity wherever there is no compelling evidence to the contrary. Such assumptions of simplicity have been made most frequently in the area of climatic reconstruction.

The reconstruction of Quaternary climates relies heavily upon the depositional record. Unfortunately, in many instances, erosion has destroyed deposits preserving climatic imprints, or the information of climatic variation was simply not recorded due to a depositional hiatus . In other instances, the sedimentary processes did not record any climatic information due to a lack of sensitivity or existence of a climatic signal. Loess is being recognized as terrestrial deposits containing a long-term, detailed record of climate.

During the past two decades, development of new dating techniques has solved the problem of ascertaining the age of old sediments (older than 50 ka) and inorganic materials that the ^{14}C method can not accommodate. Advances in dating techniques are creating promising opportunities for the dating of terrestrial Quaternary climatic events with higher resolution and higher sensitivity than even for marine Quatermary climatic events. Also, loess is one of few Quaternary terrestrial deposits that can provide a continuous and reconstructable record of Quaternary climate (Kukla, 1989). Although glacial deposits are often considered to be the most sensitive terrestrial recorder and deep-sea deposits are the most complete recorder of past climate, loess is superor to glacial deposits in conformity of deposition and to deep-sea deposits in the sensitivity of recording climatic information (Liu et al., 1985).

Loess deposits cover more than 10 percent of the land surface of the earth (Smalley, 1978), and most agricultural lands in today's temperate zones and in the marginal semiarid zones of deserts are situated on loess (Goudie, 1983). Thus, the understanding of climatic variation in loess regions is important to human economic practices. More important, a loess study can produce some insight into the history of dust storms and, therefore, can provide a basis for predicting dust storm activity in the future. In fact, climatically controlled dust-storm activity is much more crucial to agriculture in the midcontinental United States than in any other part of the world (Aandahl, 1982). Furthermore, global warming caused by the greenhouse effect will probably increase aridity and induce dust-storm activity in the midcontinental United States (Lorius et al., 1990).

In the western hemisphere, the most extensive thick deposits of loess occur in the midcontinental United States. This region, throughout which loess deposits mantle much of the upland surface, extends for more than 1000 miles east and west and for more than 500 miles north and south. The loess sheets are traceable, essentially without break, across the central Great Plains. The bulk of the loess was deposited during the Wisconsin and Illinoian glaciations (Ruhe, 1977).

Loess deposits in the midcontinental United States provide a rare opportunity to derive information regarding changes in the terrestrial environment and also the climate of North America during the past few hundreds of thousands of years. Because of the nature of its relatively complete record and the sensitivity of its depositional environment, loess sequences can function as a base for correlating terrestrial climatic records with those of ice and deep-sea cores.

Research Area

This study focuses on loess sections exposed in Barton and Pratt Counties of central Kansas in order to explore temporal variation in the loess depositional environment during the past few hundreds of thousands of years. Comparative data regarding the spatial variation of loess deposition are derived from sections in Phillips County, north-central Kansas; Kearny County, southwestern Kansas; and the Great Bend Sand Prairie of central Kansas (northernmost Pratt County). Locations are shown in Figure 1.

The Central Great Plains, where the study area lies, present the following four distinguishing characteristics: (1) flat, (2) grassy, (3) dry, and (4) dusty. These flat grasslands are not only effective traps of dust, but also effective sources of dust during violent storms. These four characteristics are described as follows.

Topography

Streams have been active geomorphic agents in the Central Great Plains. The aggradational processes of the streams have formed the sedimentary plains since the late Tertiary. Furthermore, the loess mantle has been reducing the surface relief in the Central Great Plains since the late Quaternary.

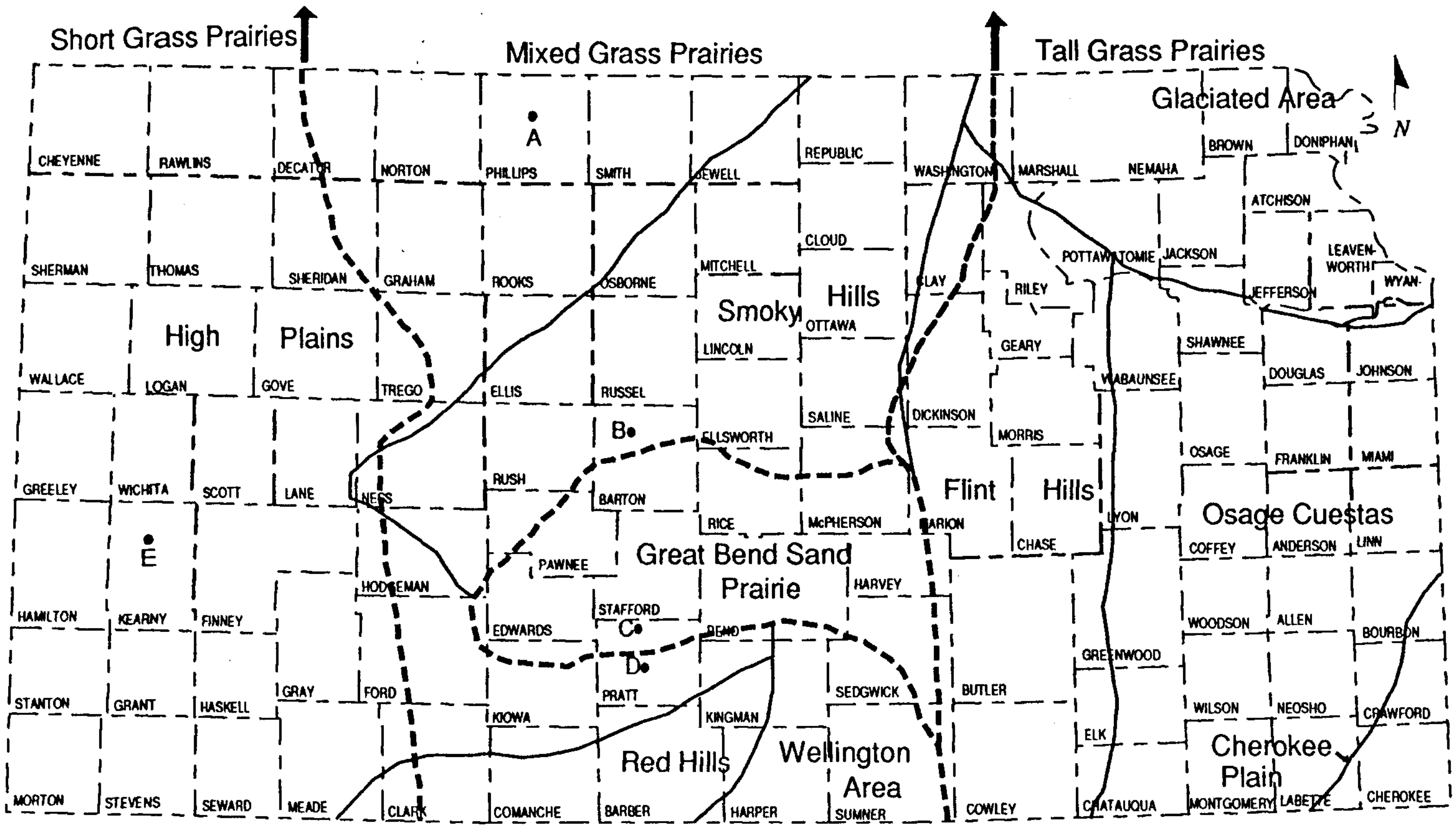


Figure 1. The physiographic provinces (Buchanan, 1984) and the vegetational divisions (Kuchler, 1974) of Kansas. The reseach sites are A (Phillips County site), B (Barton County site), C (central Grea Bend site), D (Pratt County site) and E (Kearny County site).

Loess covers about 50 percent of Kansas. The study area and the area to the west and north are nearly completely mantled by loess with thickness ranging from 3 to 30 meters. Exceptions occur mainly along modern stream valleys, where fluvial deposits dominate, and in the Great Bend Sand Prairie, where sand dunes dominate (Bayne and O'Connor, 1968).

Climate

The climate of the study area is affected by two physical features, the Rocky Mountains to the west and the Gulf of Mexico to the south. The Mountains prevent the importation of moisture from the Pacific Ocean, while the Gulf provides a source for much of the precipitation. Precipitation occurs mainly during the spring. The long and dry summer is a key factor controlling the vegetational patterns and pedogenic processes in the Central Great Plains.

In the study area (38-40°N, 98.5-101.5°W), the annual mean precipitation decreases northwestward from 60 to 45 cm and the potential evaporation decreases northward from 130 to 120 cm. According to Thornthwait's classification of climates, the area is moist subhumid to dry subhumid, with southwestern Kansas being designated as semiarid (Ruller and Bair, 1977; Flora, 1948).

Vegetation and Soil

Vegetation is an index of climate, particularly effective soil moisture. In response to the decline of the annual precipitation from the east to the west in the study area, the grasses are therefore tall in the subhumid east and short in the semiarid west (Kuchler, 1972; Webb, 1959). East of the study area are the tall grass prairies (bluestem prairies) with notable forest islands (Kuchler, 1964). Research sites A (Phillips County section), B (Barton County section) and D (Pratt County section) lie within the mixed grass prairies. Site C is within the Great Bend Sand Prairie. Site E (Kearny County section) is in the short grass Prairies (grama-bluestem-buffalo prairies).

Under these subhumid-semiarid climatic conditions, ecotones are so vulnerable that the vegetational composition can be easily altered when the climate changes (Kuchler, 1972). For example, the drought of the 1930s changed a large portion of the mixed grass prairies into short grass prairies in central Kansas (Collins, 1985).

As an organic body, soil corresponds to natural vegetation, both being governed by the effective soil moisture. Thus, soil is also an index of climate. As a consequence of the climatic interaction between soil and vegetation, Udic Ustolls correspond approximately to the tall-grass prairies, Typic Ustolls to the mixed-grass prairies, and Arid Ustolls to the short-grass prairies. The boundary between calcic and noncalcic soils corresponds approximately to the boundary between Udic Ustolls and Typic Ustolls (Machette, 1985).

Dust Storms

Wind has been an important geomorphic process operating in the study area for at least the past 400,000 years (the time period of this study), and likely much longer. Modern prevailing winds are northwesterly from December to March, southerly or southwesterly from April to November. In most cases the prevailing winds do not produce major dust transporting events; only winds of high velocity have extensive potential for mobilization of silt (loess). During the early spring winds with high velocities blow predominantly from the southwest in the region including southwestern Kansas, northwestern Oklahoma, northwestern Texas, and northeastern New Mexico (Flora, 1948). Not surprisingly, this region was the center of the 1930's dust storm activity (Goudie, 1983).

On the basis of distance and elevation of transportation, dust storms are classified into three types: (a) blowing dust of a local nature by lower-air currents (low intensity and low volume), (b) dust carried a long distance by upper-air currents (high intensity and low volume), and (c) severe dust carried a relatively long distance by lower- and middle-air currents (high intensity and high volume). The latter occurs only during periods of prolonged drought and is most effective in transporting silt. For example, Malin (1946, p.25) quotes a statement of a traveler experiencing a dust storm in the 1930s , "The train could hardly go through it (dust) and the darkness was like going through a tunnel." Historical dust-storm events could be analogues to the geological ones. As a consequence of dust storm activity during at least the past 400,000 years, much of the Central Great Plains has been mantled with loess. Dust deposition (loess) and associated hiatuses (soils) in this sensitive ecological environment provide an outstanding opportunity for reconstructing the past climate.

Objectives of Investigation

Several factors make the study area an ideal place for reconstructing paleoclimatic conditions recorded in the loess deposits. First, the deposits are thick and well exposed by natural and human activities. Second, the study area is located within a relatively sensitive ecological zone. Third, the area has received little detailed attention from Quaternary researchers for nearly forty years. Earlier studies were carried out without the benefit of modern analytical laboratory techniques and access to dating techniques.

As this study developed, many questions emerged, but three deserve detailed attention. First, what have the climatic conditions been during the loess deposition (temporal variation)? Second, what variations occurred in the loess depositional environment throughout the study area (spatial variation)? Third, how does climate reconstructed from loess in the study area compare with the general global climatic records (global comparability)?

Temporal Variation of Loess Deposition

Many Quaternary stratigraphic studies have been conducted in the midcontinental United States, and most have considered the general environmental regime during loess deposition, but few are in-depth detailed studies specifically designed to assess climate during loess deposition. There is, therefore, an opportunity for reconstruction of a detailed climatic history of loess deposition by establishing a detailed chronology and studying the temporal variations in physical and chemical properties within the loess sequences, such a study is undertaken herein.

The Loveland Loess was believed to have been deposited during the Illinoian glaciation (Schultz and Stout, 1948; Frye and Leonard, 1952). But is is not known if it represents only that glacial stage in the Central Great Plains. The last interglaciation was expressed by a well developed soil, Sangamon pedocomplex (a composite soil with several A horizons formed by multiple soil-forming events), which was presumed to have formed from 125,000 to 75,000-70,000 yr BP (Ruhe, 1977). The Gilman Canyon Formation (pedocomplex), which immediately overlies the Sangamon pedocomplex and is a complex of one or more soils, has been dated at 35,000-20,000 yr BP in the Central Great Plains (Johnson, 1990; Johnson et al., 1989). Souders and others (1971) posed the question as to what occurred during the interval from 75,000-70,000 to 35,000-30,000 yr BP. This question is addressed in this study.

The last glaciation is the most extensively studied, but accepted environmental reconstructions are still inadequate. Ruhe (1983) stated that the Peoria Loess was deposited during the last glacial maximum when the climate was cool and moist and conifer forests and woodlands dominated the loess-covered region of the midcontinental United States. Nevertheless, he immediately questioned why sand and silt (loess) were being deposited around and within stands of conifer forests and woodlands, implying the cool-moist loess depostional model should be reevaluated.

Based on the studies in the Northern Great Plains (Barnosky, 1989), in the Central Lowlands (Winkler et al., 1986) and in the Southern Great plains (Holliday, 1989), the dry mid-Holocene climatic model has been well established and popularly accepted. Fluvial sequences in the Central Great Plains support this model (Johnson and Martin, 1987). This study addresses whether or not the loess and sand sequences in the study area confirm this dry mid-Holocene model.

Spatial Variation of Loess Deposition

The importance of large river valleys in contributing loess was emphasized by Swineford and Frye (1951). But, Lugn (1968) affirmed that the Sandhills of Nebraska was a major source of the Peoria loess. Despite disagreement on the primary source of the Peoria Loess, these authors agreed that the northwesterly winds were the most important contributors of Peoria Loess in the Northern and Central Great Plains. But, in the Southern Great Plains, Holliday (1989) indicated that the southwesterly winds contributed the Peoria loess. Furthermore, historic records strongly suggest that southwesterly winds have contributed dust not only to the Southern Great Plains, but also to the Central Great Plains.

Three questions arise with regard to the source of the Peoria Loess as well as Bignell Loess. First, we are not sure whether the loess originated from large river valleys as Swineford and Frye stated (1951) or from the Sandhills of Nebraska as Lugn proposed (1968). Second, historic records suggest that the southwesterly winds have been dominant in contributing loess to the Central Great Plains, but it is not certain what was the importance of the southewesterly winds during the Peoria Loess time. Third, the northwesterly winds are thought to have been dominant in contributing the Peoria Loess (during the the last glacial maximum) of central Kansas, but it is not known if the northwestern winds have been dominated during the Holocene. If the direction of dominant winds contributing loess has changed from the last glacial Maximum to the Holocene, the wind patters must have been changed significantly, especially during the transitional period, in the midcontinental United States. In this context, the variation of loess sources is examined to reconstruct the changes of the wind patterns during the late Quaternary.

Since this study area covers different climatic and vegetational zones and the proximity of each of the five research sites to the proposed loess sources (either the Platte River or the Sandhills of Nebraska) is different, it is expected that the conditions of loess deposition have been different from site to site during the late Quaternary. The temporal and spatial variations of loess deposition in the study area are evaluated to assess spatial differentiation of the loess depositional environment in different times.

Global Climatic Comparison

The loess sequences in Europe (e.g., Kukla, 1975; Pecsi, 1985) and in China (e.g., Liu, 1985; Wang, 1985) have been intensively studied, and the continental climatic records have been successfully correlated with well-dated deep-sea records (Kukla, 1977; Liu et al., 1986). Unfortunately, the potential for obtaining climatic and environmental information from loess sequences in the midcontinental United States has not yet been sufficiently explored. Although Kukla (1987) attempted to correlate the loess sequences of the midcontinental United States with those from other continents and the seas, the correlation is far from satisfactory because of a lack of detailed research in loess sequences of the midcontinental United States. To reconstruct climatic changes of the Quaternary with high temporal and spatial resolution, the loess sequences in the midcontinent deserve special attention.

CHAPTER II. REVIEW OF RELATED LITERATURE

Quaternary Stratigraphy

A comprehensive Pleistocene sequence for Nebraska was established by Lugn in 1935. After Schultz and Stout (1945, 1948) articulated the loess stratigraphy of Nebraska, Condra and Reed (1950) revised the sequences and made spatial correlations among the Pleistocene sequences in Nebraska. At approximately the same time, Quaternary geologists in Kansas (e.g., Frye and Leonard, 1951, 1952; Leonard, 1952; Leonard and Frye, 1954; Swineford and Frye, 1951, 1955) adopted the Pleistocene framework of Nebraska and designated the stratigraphic sequences based on numerous previous studies (e.g., Frye, 1945, 1946; Latta, 1950; McLaughlin, 1943; Smith, 1940).

The stratigraphic classifications in Nebraska (Reed and Dreeszen, 1968) and Kansas (Bayne and O'Connor, 1968; Layton and Berry, 1973), and in the midcontinental United States (Dreeszen, 1970; Frye, 1973), in general, are based primarily upon the classical fourfold glacial scheme. Specifically, in central and western Kansas the Holdrege Formation represents the early Nebraskan stage; Fullerton Formation, the late Nebraskan stage; Afton Soil, the first interglacial stage of the Quaternary; Grand Island Formation and Sappa Formation, Kansan Stage; and the second interglacial stage by Yarmouth Soil (Figure 2). The Illinoian and Wisconsin stages were understood better than the early stages.

The discovery of isotopic variations in deep-sea sediments, which is interpretated as reflecting continental ice volume, challenged the classical fourfold glacial scheme (Shackleton and Opdyke, 1973; 1976). Subsequently, Quaternary geologists began looking at terrestrial deposits, hoping to find evidence to identify the Quaternary climatic signals derived from deep-sea sediments. Easy access to radiometric dating techniques permitted Quaternary geologists to reexamine the reliability of the classical stratigraphic framework. Based on radiometric dating data from volcanic ashes interbeded with glacial tills and loess units in the midcontinental United States, Boellstorff (1978) reexamined the classic stratigraphic scheme. He found that the classical Kansan till is younger than 0.6 ma and classical Nebraskan till is about 1.0 ma. This means that the classical North American Pleistocene stages, as defined in their type areas, represent only last portion of the Pleistocence time, which began about 1.6 ma (Richmond and Fullerton, 1986) or 2.4 ma (Bowen, 1978; Nilsson, 1983; Pecsi, 1985). The recent identification of more than four glaciations in North America (Richmond and Fullerton, 1986) completely

Chronostratigraphic Units	Lithostratigraphic Units	
	Northeastern	Central and Western
Recent Stage	Eolian and Fluvial Deposits	
Wisconsinan Stage	Bignell Formation	Bignell Formation
	Brady Soil	
	Peoria Foramtion	Peoria Formation
	Gilman Canyon Formation	
Sangamonian Stage	Sangamon Soil	
Illinoian Stage	Loveland Formation	Loveland Formation
		Crete Formation
Yarmouthian Stage	Yarmouth Soil	
Kansan Stage	Loess	Sappa Formation
	Cedar Buffs Till	
	Fluvial Deposts	
	Nickerson Till	Grand Island Fm.
	Atchison Fm.	
Aftonian Stage	Afton Soil	
Nebraskan Stage	Loess	Fullerton Formation
	Iowa Point Till	
	David City Fm.	Holdrege Formation

Figure 2. Classification of Pleistocene Series in northeastern area and central and western area of Kansas (after Bayne and O'Connor, 1968).

dismantled the basis on which the classical stratigraphic framework of the Great Plains relied (Frye et al., 1968; Stout, Dreeszen et al., 1965).

Loveland Loess

The Loveland Loess is the oldest recognized loess in Kansas and was named by Shimek (1910). Loess deposits of western Kansas were designated as the Sanborn Formation by Elias (1931). Subsequently, three major loess units (Loveland, Peoria, and Bignell) were recognized as the Sanborn Formation (Frye and Fent, 1947). The loess at the type locality in Kansas was described by Frye and Leonard (1951) as being 9 m of massive, well-sorted, fine sand and silt. The upper 6.6 m is leached of calcium carbonate and the lower 2.4 m is unleached stringers and nodules of calcium carbonate. The upper leached zone has a distinct pink to red-brown tint as a result of weathering. Frye and Leonard (1951) considered this zone as part of the Sangamon pedocomplex.

The Loveland Loess was presumed to have been deposited from 400 to 127 or 125 ka (Catt, 1986; Linebach, 1979; Wright and Frye, 1965) and was regarded as a formation of the Illinoian glacial stage. However, the presence of weakly developed pedocomplex observed within the Loveland Loess of Kansas (Frye and Leonard, 1951) suggests that the loess is a complex of loess units and pedocomplexes.

Schultz and Martin (1970) identified several pedocomplexes within the Loveland Loess in Nebraska (Buzzard's Roost pedocomplexes and Ingham pedocomplex). In Illinois, till and loess sequences suggest three substages during the Illinoian glacial stage: Jubilean; Monican; and Liman (Frye and Willman, 1975). The three tills are separated by three silt units probably representing three glacial retreats . This till-silt sequence of the Illinioian stage, as defined in Illinois, could correspond to the soil-loess (or loess-soil) sequences in the Loveland Loess of Kansas and Nebraska.

Sangamon pedocomplex

After the Illinoian glacial stage, the Sangamon Formation developed during the last interglaciation (125-75 ka). This formation is composed of one or more soils formed by multiple soil-forming events, and therefore, called pedocomplex. This pedocomplex emerges from beneath Wisconsin loess in the midcontinental United States (Frye and Leonard, 1965; Ruhe, 1965; Simonson, 1954; Thorp et al., 1951). In Kansas, the Sangamon pedocomplex is perhaps the most widespread and laterally traceable stratigraphic unit of the Pleistocene. The recognition of the pedocomplex is alway based on the assumption that the first reddish soil in loess sequence is the Sangamon pedocomplex, however, there is not much published data to support it.

The soil is typified by a zone of clay accumulation (0.1-m-thick), reddish brown color, and well-developed structure. The upper 0.5 m of the soil is leached of calcium carbonate, and a zone of caliche has accumulated at the base of the leached zone.

Gilman Canyon pedocomplex

The late Pleistocene (after 75 ka) is regarded as the Wisconsin glaciation (Catt, 1986). Although the lower Wisconsin loess in Illinois (Roxana) has been projected to 75 ka (Frye and Willman, 1975), there is no absolute dating to verify that antiquity. In Iowa, the base of the lower loess is dated at 31 ka (Ruhe, 1977), and ages from the base of the Gilman Canyon pedocomplex (Gilman Canyon Formation) in Nebraska are 31-35 ka (Souders et al., 1971). Recent TL dates confirm that the Roxana Silt was deposited from 30 to 40 ka (Forman et al., 1991), even to 50 ka (Leigh, 1991; McKay, 1979).

The accumulation of humus-enriched silt, the *Citellus* zone of Nebraska (Schultz and Stout, 1945), was given the formal name Gilman Canyon Formation by Reed and Dreeszen (1965). They demonstrated that it was formed under conditions of slow loess accumulation from 35 to 23 ka, based on ^{14}C ages from the top and bottom of this pedocomplex. These ages suggest that the Gilman Canyon

Formation in Nebraska may be equivalent to the Farmdale Silt of Illinois (Willman and Frye, 1970). The Farmdale Silt was originally dated at 22-28 ka (Ruhe, 1977) and recently was dated at younger than 25 ka (Curry, 1989). In both Nebraska and Kansas, much attention has been devoted to dating the Gilman Canyon Formation (Gottula and Souders, 1989; Johnson, 1990, 1991; Johnson et al., 1990; May and Souders, 1988; Souders and Kuzila, 1990). These ages indicate that the Gilman Canyon Formation ranges from approximately 35 ka at the base to 20 ka at the top.

Peoria Loess

The name "Peoria" was first applied to a weathering zone and assumed to represent an interglacial period between the Iowan and Wisconsin glacial stages (Leverett, 1898). Alden and Leighton (1917) found that the Peoria was actually post-Iowan in age, representing a glacial period. The term "Peoria" has been, in turn, used for equivalent deposits across Nebraska (Lugn, 1935) and Kansas (Frye, 1945; Frye and Leonard, 1951).

In Kansas, the Peoria Loess includes loess and locally water-laid silt that is stratigraphically continuous with it (Fry and Fent, 1947). Peoria Loess is strikingly uniform in mineral and chemical composition, general appearance, structure and texture, and faunal assemblages throughout Kansas (Swineford and Frye, 1951) and Nebraska (Condra, Reed and Gordon, 1947; Lugn, 1962). Thickness of the Peoria Loess is more than 30 m along the southern margin of the Sandhills of Nebraska (Lugn, 1962), about 6 m in Phillips County of north-central Kansas, about 8 m in Kearny County of southwestern Kansas (Ransom, 1990, personal communication). The thickness of the loess is only about 2 m in central Kansas.

The Peoria Loess in Kansas is a yellowish tan, buff color, homogeneous, massive, and calcareous silt. It is sudivided into three faunal zones (Leonard, 1951) and correlated with three loess units of Illinois (Leighton and Willman, 1950). A basal zone devoid of faunal fossil material (Figure 3) was correlated with the Farmdale Loess (Roxana Silt in current Illinois classification); a lower faunal zone above the basal zone with Iowan Loess; and upper faunal zone with Tazewell Loess. The lack of any well-developed, buried paleosols or other unconformities suggests that, as in Nebraska (Souders et al., 1971), the Peoria Loess in Kansas represents continuous deposition, and the faunal zonation may reflect a change in the rate of deposition (Welch and Hale, 1987). However, Johnson (1991) argues that, although no stratigraphic break has yet been identified, charcoal has been observed within or near the

<table>
<tr><td colspan="4">Late Pleistocene Series Classification of Kansas</td><td colspan="2">Chronostratigraphic units of the Upper Mississippi Valley</td></tr>
<tr><td colspan="2">(Leonard, 1951)</td><td colspan="2">(Bayne and O'Connor, 1968)</td><td colspan="2">(Willman and Frye, 1970)</td></tr>
<tr><td colspan="2">Eolian and Fluvial</td><td colspan="2">Eolian and Fluvial</td><td colspan="2">Holocene</td></tr>
<tr><td rowspan="7">Sanborn Formation</td><td rowspan="4">Bignell Formation
Brady Soil
Peoria Silt
Upper Zone
Lower Zone
Basal zone</td><td rowspan="5">Wisconsinan</td><td rowspan="5">Bignell Fm
Brady Soil
Peoria Fm
Gilman Canyon Fm</td><td rowspan="5">Wisconsinan</td><td>Valderan</td></tr>
<tr><td>Twocreekan</td></tr>
<tr><td>Woodfordian</td></tr>
<tr><td>Farmdalian</td></tr>
<tr><td rowspan="2">Sangamon Soil</td><td>Altonian</td></tr>
<tr><td colspan="2">Sangamon Soil</td><td colspan="2">Sangamonian</td></tr>
<tr><td>Loveland Formation</td><td colspan="2">Loveland Fm
Crete Fm</td><td colspan="2">Illinoian</td></tr>
</table>

Figure 3. Evolution of the late Pleistocene of Kansas and its comparison with The chronostratigraphy of the upper Mississippi Valley (After Welch and Hale, 1987).

boundary between the lower and upper zones; and the upper zone often seems to have a lower bulk density and, infrequently, fine bedding. The Peoria Loess is bracketed by the Gilman Canyon pedocomplex and the Brady soil.

Brady Soil

The Brady Soil was first named and described by Schultz and Stout (1948) at the type locality in western Nebraska. It is believed that the soil developed within the Peoria Loess and was subsequently overlain by the Bignell Loess (Schultz and Stout, 1945; Frye and Leonard, 1951). A soil equivalent to the Brady of Nebraska has been found in northwestern and west-central Kansas. Without the overlying Bignell Loess, the Brady Soil does not exist or is not identifiable (Caspall, 1970). The soil is typically dark gray or gray-brown and characterized by a moderately developed B horizon and a notable accumulation of carbonate in the C horizon. The profile of the Brady Soil is generally thicker than that of the modern surface soil developed on the Bignell Loess. The contrast may be attributable to the fact that the Brady Soil developed under conditions of poorer drainage than did the modern soil at the same locality. Recent field investigations show that the Brady Soil is ubiquitously distributed on the uplands, suggesting that this soil not only developed under topographically controlled poor drainage conditions, but also under climatically controlled moist conditions.

Bignell Loess

The Bignell Loess was named and described at the same locality as the Brady Soil (Schultz and Stout, 1945). It is typically a gray or yellowish tan, calcareous silt, darker in color than the Peoria Loess. The Bignell Loess, seldom more than 2 m in thickness, does not form a continuous mantle on the Peoria Loess. The loess is only about 0.8 m thick in central Kansas. In the Great Bend Sand Prairie, dune sand occupies the stratigraphic position of the Bignell Loess. The existence of the Bignell Loess in northeastern Kansas is in dispute since the presence of the Brady Soil is questioned by Caspall (1970). Caspall argued that the Brady soil in northeastern Kansas is not a true soil, but a weathering zone developed in the Peoria Loess. However, he did not define the difference between a soil and a weathering zone.

The begining of Bignell Loess deposition marked the end of the Brady pedogenesis. The end of the Bignell Loess deposition is uncertain. Ruhe (1977) speculated that it was deposited during the Valderan substage of the Wisconsinan from 11,000 to 5000 yr BP. Only one age suggests that the Bignell Loess ended about 8000 yr BP (Martin, 1990). This age of the Bignell Loess agrees with that of sand-dune activity in the Sandhills of Nebraska (Swinehart, 1990). Deposition of the Bignell Loess likely occurred simultaneously with the sand-dune activity during the dry and warm period from 8500 to 5800 yr BP (equivalent to the Altithermal Period) in the Great Bend Sand Prairie of Kansas (Feng et al., 1991).

Quaternary Environments

As mentioned earlier, the classical fourfold glacial scheme is no longer valid. The East Anglia sequences of England exhibit at least seven cold phases since 2.4 ma (Zagwijn, 1975). The sequences of Chinese loess (Kukla and An, 1989; Liu, Zhang and Han, 1986), European loess (Kukla, 1977; 1978), and Pacific V28-238 core and V28-239 core (Shackleton and Opdyke, 1973; 1976) reveal that there have been seventeen temperature cycles since 2.4 ma and nine of them have occurred since 900 ka. The magnitude of the temperature variation of eight of these nine cycles is approximately the same, i.e., the last glacial-interglacial cycle has been repeated eight times since 900 ka.

Loveland Time

Loveland Loess was presumed to have been deposited during the Illinoian glacial stage, lasting from 400 to 127-125 ka (Catt, 1986; Lineback, 1979; Wright and Frye, 1965). The presence of weakly developed soils within the

Loveland Loess in Kansas, however, reflects intervals of landscape stability (Leonard and Frye, 1951).

The glacial sequence in Illinois shows that after the Yarmouth Soil (interglacial), Petersburg Silt was deposited; this was, in turn, followed by a till unit. Subsequently, the Pike Soil developed, indicating that the climate was relatively warm and moist. This soil was immediately followed by another silt, the Mulberry Grove Silt. Overlying this silt is another till (glacial advance). The last (third) substage started with the Teneriff Silt and ended with the last glacial advance before 127 ka (Catt, 1986).

Fredlund and others (1985) and Schultz and Stout (1980) found that three to five paleosols within the Loveland Loess of Nebraska expressed fluctuations in the loess depositional environment during Loveland time. Recently established terrestrial and marine glacial-interglacial sequences demonstrate that four interglaciations and three glaciations occurred in the interval from about 400 to 100 ka (classical Illinoian Stage). This glacial-interglacial chronology leads this author to speculate that the three silt units plus the Sangamon pedocomplex in Illinois might represent the four interglaciations, and the three tills represent the three glaciations from 400 to 100 ka. Both the middle silt unit and the Pike Soil which immediately underlies the silt probably represent one glacial retreat interval. In fact, there is no hard evidence to confirm that the Laurentide ice sheet did not disappear during the retreating intervals of the Illinoian Stage (McCoy, 1991, personal communication).

Sangamon Time

The Sangamon pedocomplex can be traced from Illinoian tills in central llinois to the Loveland Loess in the Central Great Plains (Frye et al., 1965). The soil is usually buried by Wisconsin loess and characterized by eluvial E and argillic B horizons. This buried B horizon has redder hues and stronger chromas than the surface soil (Frye and Leonard, 1951).

The profiles of the Sangamon pedocomplex are similar to the present southern reddish forest soils, the Ultisols (Ruhe, 1965). This pedocomplex extends throughout the Central Lowlands and the eastern Great Plains and maintains the southern, reddish forest-soil appearance. An environment comparable to that currently present in the southeastern states may have extended to the northwest as far as southeastern South Dakota during the Sangamon time. If so, the border of the southern reddish forest soils would have been displaced at least 610 km northwest of its present boundary in southern Missouri (Ruhe, 1977). This soil is believed to have been formed from 127-125 to 75-70 ka (Frye and Willman, 1975).

Wisconsin Time

Early Wisconsin. Due to easy access and high reliability of radiocarbon dating techniques, the Quaternary history of the past 50 ky has been intensively studied. Frye and Willman (1963) proposed the following five substages for the Wisconsin Stage based on glacial chronosequences in Illinois: Altonian ((70-28 ka), Farmdalian (28-22 ka), Woodfordian (22-12.5 ka), Twocreekan (12.5-11 ka), and Valderan (11-0.5 ka).

Although the Altonian substage was projected as 70-28 ka, the Roxana Silt deposited during the Altonian substage was ^{14}C dated 37-27 ka in Illinois (Ruhe, 1977) and TL dated 40-30 ka in Iowa (Forman et al., 1991), even to 50 ka in Wisconsin and Illinois (Leigh, 1991; McKay, 1979). Curry (1989) stated that there is no depositional record found in Illinois from 75 ka (Sangamon time) to 25 ka. He denied the presence of Altonian glaciation in Illinois as identified by Wright (1984). In Nebraska and Kansas, no Altonian silt has been identified. The question then is what happened during the so-called Altonian substage (Souders et al., 1971), i.e., there is apparently no depositional record for the interval from 75-70 to 35-30 ka.

Middle Wisconsin. As mentioned earlier, the Farmdalian substage in Illinois dated at 28-22 ka is approximately equivalent to the Gilman Canyon Formation in Nebraska and Kansas dated at 35-20 ka. The planktonic foraminiferal assemblages in the North Atlantic show that the ocean surface temperature at 28 ka was as high as today's (Flohn, 1983; Kipp, 1976), at least apparently warmer than the last glacial maximam (Ruddiman and McIntyre, 1981). Evidence indicates that the Laurentide ice sheet began to advance at about 28 ka, entered the United States about 24 ka, and reached its maximum extent (southernmost) at about 19 ka (Mickelson et al., 1983; Ruhe, 1983).

Faunal studies (Carter, 1985) in the Central Lowlands show that from 28 to 21 ka boreal forest beetle fossils accumulated in eutrophic lakes surrounded by spruce forests, as indicated by abundant plant macrofossil remains and a diverse assemblage of scolytid beetles. Existence of an open-ground beetle fauna of arctic-subarctic affinities has been established along the southern margin of the Laurentide ice sheet in the midcontinent by 20 ka (Schwert and Ashworth, 1988). Seemingly, the boreal forest faunal assemblages match well chronologically with the Farmdalian substage in the Central Lowlands, demonstrating that this period was cool (not cold) and relatively moist.

Contrary climatic evidence was found in several localities in the eastern Great Plains at the Farmdalian-Woodfordian boundary (24-21 ka). The Ozark Spring records, western Missouri (King, 1973; Mehringer et al., 1970) indicate that jack-pine parkland or savanna was overrun by spruce forests during this climatic shift. A similar shift happened at the same time in east-central Iowa (Hallberg et al., 1980; Van Zant et al., 1980). Studies at the Eustis ash pit in south-central Nebraska and at Sander's Well in southeastern Kansas suggest that the Gilman Canyon Formation was formed under conditions of low effective moisture (Fredlund and Jaumann, 1987; Fredlund and Johnson, 1985; Fredlund et al., 1985).

Late Wisconsin. The Peoria Loess was deposited during the last glacial maximum substage, the Woodfordian. The climatic and vegetational changes during this period have been intensively studied. High lake levels and expansion of mountain forests in the American Southwest (Spaulding and Graumlich, 1986) and in the Great Basin (Scott et al., 1983) suggest cool and moist conditions during this time. But, the same evidence was interpreted differently by Galloway and Brakenridge (1978), who stated that lower evaporation rates due to 7-8°C cooling would have sufficed to maintain high lake levels and allow expansions of mountain forests under less precipitation conditions. The climatic models (CLIMAP members, 1981; COHMAP members, 1988) for the last glacial maximum, however, provide an acceptable explanation. The large Laurentide ice sheet split the westerly jet into northern and southern branches over North America. An increase in storms associated with the southern branch promoted expansion of woodlands and caused the lake levels to rise in the Southwest and Great Basin. Anderson (1990) argued, however, that these cool and moist conditions can be explained more reasonably by assuming more cloudliness during that time, not necessarily more precipitation or sufficiently low temperatures.

Pollen assemblages (Jacobson et al., 1987; Wright, 1970), macrofossils of spruce and pine, landsnail and small-mammal faunas (Jacobson et al., 1987; Wells, 1983; Wells et al.,1987; Wright, 1970) indicate that the midcontinental United States might also have been cool and moist during the Woodfordian substage. In the Central Great Plains, at the onset of Peoria Loess deposition or the end of Gilman Canyon pedogenesis (22-20 ka), the regional vegetation rapidly changed towards increased forest cover (Fredlund and Jaumann, 1987). Also well documented is the fact that the range of white spruce during this time extended as far as northern Louisiana (Delcourt and

Delcourt, 1985; Holloway and Bryant, 1984) and as far west as the Nebraska Sandhills (Watt and Wright, 1966).

Appreciable evidence suggests that the Peoria Loess was deposited on a well-vegetated surface. Wells and Stewart (1987) found needle leaves of limber pine and spruce charcoal dated at 14,500 yr BP at two loess localities in central Kansas. Analysis of grass phytolith in loess exposed at the Eustis ash pit, south-central Nebraska, indicates that Peoria Loess was deposited under conditions of high effective moisture; in contrast, the Gilman Canyon soil was believed to have been formed under conditions of low effective moisture (Fredlund et al., 1985). But, based upon pollen data from the Cheyenne Bottom of central Kansas, contrary evidence was found (Fredlund and McClain, 1990), i. e., more woodlands were present from 45,000 to 20,000 and the aridity increased from 20,000 to 12,000 ka. Pollen data from the North Cove site of south-central Nebraska suggest that an aspen parkland existed on uplands between 24 and 12.8 ka (Fredlund and Jaumann, 1987). Ruhe (1983), in a synthesis of the biological evidence within the Peoria Loess, concluded that trees were present during deposition of the loess. It should be noted, however, that in most cases pollen and microfossil samples were derived from peats, lake sediments, floodplains, or archaeological sites, which might only represent local topographically or hydrologically controlled situations, rather than broadscale, climatically controlled conditions. In fact, treeless openings, perhaps a pine parkland, was believed to have existed in the Central Great Plains during the late Pleistocene (Baker and Waln, 1985; Webb et al., 1987). Loess deposits in Iowa dating between 24 and 14 ka yielded small mammal fauna typical of grassland areas (Rhodes and Semken, 1986). Prairie and steppe fauna have been recovered from the Peoria Loess across the Central Great Plains (Goodwin, 1989; Hoffman and Jones, 1970). Discrepancies between conifer forests and prairies during Peoria Loess deposition have caused a great deal of concern in the Quaternary community. Ruhe (1983) stated that it is difficult to explain how sand (dunes) and silt (loess) could have been deposited around and within stands of conifer forests and woodlands, since the forests and woodlands inhibit wind necessary to transport silt (Feng and Thompson, 1987).

After reexamining the published works on the late Pleistocene environments in the Southern Great Plains, Holliday (1987) challenged the cool and moist climatic model for the last glacial maximum substage (Woodfordian) in the region. He concluded that the so-called conifer forest reconstructed from the pollen data is misleading, because pollen types indicative of conifer forest are the most resistant to oxidization, and therefore, best preserved. Consequently, Holliday suggested that the last glacial maximum substage in the Southern Great Plains was characterized by cool and dry conditions, rather than the cool and moist conditions proposed by Wendorf and Hester (1975). He further argued that neither a pedocomplex dated at 27 ka nor younger deposits (about 18 ka) representing the proposed conifer forests have any features of podzolization, the soil-forming process typical of conifer forests. He continued to argue that even pinyon, which prefers to grow in dry climate and thin rocky sediments, will produce soils having podzolic features, including E horizons. Holliday's argument is strengthened by modern analogues. In some modern arid situations, such as the northern Loess Plateau of China, and southwestern Australian Plateau, topographically controlled conifer trees (in form of patches) are only the plants producing wind-dispersed pollen.

The Peoria Loess does not exhibit any features of podzolization, i.e., no characteristic E horizon. Even the well-developed Gilman Canyon pedocomplex does not have any features of podzolization. In fact, the Gilman Canyon pedocomplex appears similar to the modern Pawnee soil (Mesic Aquic Argiudolls) developed in the tall-grass prairies of northeast Kansas in its appearance and structure (Aandahl, 1982).

Faunal investigations in the Central Great Plains have led to the conclusion that the late-Pleistocene biological communities were more complex and diverse than those of the Holocene and less seasonal, featuring cooler summers and warmer winters than those of the Holocene (Martin, 1984; Martin and Neuner, 1978; Martin and Hoffman, 1987; Martin and Martin, 1987). Recent climatic models have largely agreed with these conclusions: for the Central Great Plains, Kutzbach and Wright (1985) postulated drier conditions coupled with slightly cooler January and significantly cooler July temperatures during the late Pleistocene than in the region today.

Holocene Time

Climate, vegetation, and mammalian fauna of the region underwent dramatic changes around 10,500 yr BP (Semken, 1983; Webb et al., 1983; Wright, 1970). Spruce trees had been replaced by deciduous trees in northeastern Kansas, and deciduous trees persisted until about 9,000 yr BP when grass lands expanded (Webb et al., 1983). It is clear that megamammalian extinction and dissolution of disharmonious faunas began about 12,000 yr BP, and the moist conditions under which the Brady soil developed persisted until about 8,000 yr BP, when the modern climate first appeared

(Semken, 1983). These changes in vegetational and faunal assemblages are thought to reflect a shift to warmer and drier conditions with more seasonality (COHMAP Members, 1988).

The Bignell Loess is a major upland Holocene deposit. The loess is darker than the Peoria Loess, which is believed attributable to the Brady Soil's contribution to the loess (Johnson and May, 1990, personal communication). Based on the vegetational history (Webb et al., 1983), the loess was probably deposited during the Altithermal Period from 9,000 to 5,000 yr BP.

During the Altithermal Period, the tall grass prairie species migrated eastward to areas that support mixed deciduous-prairie vegetation presently (Semken, 1983; Van Zant, 1979). The dry middle Holocene was documented in extensive areas of the midcontinent, such as in Iowa (Semken, 1983), in Montana (Barnosky, 1989), in the Southern Great Plains (Holliday, 1989a). Winkler and others (1986) synthesized a variety of geological and biological evidence of the midcontinent and concluded that annual precipitation was reduced about 10 percent, and mean annual temperature reached its maximum around 6500 yr BP.

In the Southern Great Plains, sand dunes were active from 9,000 yr BP to 6,000-4,500 yr BP due to prolonged drought (Holliday, 1989a). Nevertheless, there is no evidence in any Texas local fauna to indicate that climate was either drier or warmer between 7,000 and 4,000 yr BP than at present (Semken, 1983).

Sources of Peoria and Bignell Loesses

Glacial outwash was thought to be an ideal source of loess in Europe, because glacial grinding was considered as the most effective mechanism for producing silt (Smalley, 1978). As a consequence of the importation of the European glacial theory to North America, the outwash of the Laurentide ice sheet was also regarded as the source of the loess in the midcontinental United States. Such parameters as mineralogy, particle size, and thickness change with distance from large river valleys have been used to determine the source of the Peoria loess. Many researchers (Caldwell and White, 1956; Fehrenbaher et al., 1965; Frazee et al., 1970; Johnson and Follmer, 1973; Kleiss, 1989; Kleiss et al., 1973; Putman et al., 1989; Rutledge et al., 1975) have proposed that river valleys such as those of the Ohio, Mississippi, Missouri, Snake River in Idaho, Illinois River in Illinois, and Wabash River in Indiana were sources of the Peoria Loess. The glacial outwash of the Laurentide ice sheet was considered the silt sources for these river valleys.

In the Central Great Plains, particularly Nebraska and Kansas, sources of the Peoria Loess were enthusiastically debated during the first half of this century. The glacial outwash explanation has gained wide acceptance (Frye and Leonard, 1951; 1952; Swineford and Frye, 1951). The principal glacial outwash streams recognized as contributors of the Peoria Loess were the Missouri, Platte, Arikaree, Republican, and Arkansas Rivers.

Studies by Swineford and Frye (1951) show that the median particle size of the Peoria Loess decreases southward and eastward from points near these rivers, and the Arkansas River might have contributed coarse material locally. Their data and diagrams clearly show that the median particle size decreases systematically from north to south and from west to east, with large rivers adding coarse particles only adjacent to the river valleys and on the downwind side.

A regional desert source area for the Peoria Loess was postulated by Lugn (1935; 1960; 1962; 1968). Considering the textural evidence of the Peoria Loess and the synoptic climate of dust storms in the 1930s, Lugn concluded that the Sandhills of Nebraska were the most important source for the Peoria Loess. The loess in this area exhibits the expected reduction in particle size from its hypothesized source, the Sandhills. Goudie and his coworkers (Goudie et al., 1979) provided support for the desert origin of loess. They found that salt-weathering processes in deserts are as efficient as glacial grinding in producing silt.

Reed and Dreeszen (1965) proposed that the Peoria Loess of Nebraska was derived from fluvial deposits originating from the Ogallala Formation. Reed (1968) argued that Lugn's desert hypothesis was not acceptable for explaining the loess source. He believed that, if northern Nebraska was a desert during the Peoria time, the region south of the desert must be even drier, and drier conditions were not favorable for trapping loess. Ahlbrant and others (1983) suggested that the Ogallala Formation, reworked by the early-Quaternary Platte River system, could have served as an important source for both the sand of the Sandhills and the loess south of the Sandhills.

As Welch and Hale stated (1987), a single-source hypothesis does not satisfactorily explain the discontinuity in distribution and the variation in thickness of the Peoria Loess of Kansas. Therefore, a multiple-source hypothesis should be adopted for further research into the origin of the loess.

Feng and Thompson (1987, 1989) proposed a relatively uniform source of the Peoria Loess in the midcontinental United States. Their studies show that the silt percentage and mean size of the silt decrease systematically from

Nebraska and Kansas to Ohio, but the sand percentage is high only in places adjacent to large rivers. This implies that silt in suspension was transported mainly from far west. Large rivers added sand and coarse silt only locally. This conclusion agrees with Beavers (1957) who analyzed clay minerals of the Peoria Loess and found that the clay minerals are quite uniform throughout the loess region. Beavers further concluded that the clay minerals are totally different from those of the tills, which were supposed to be the original source of the loess. This conclusion seems to reject the glacial outwash hypothesis. Frye and Leonard (1951, 1952) rejected the uniform-source hypothesis for the the Peoria Loess, but they also failed to find any local difference in chemical composition of the loess.

As documented by Goudie (1983), Kutzbach and Wright (1985), and McCauley and others (1981), the southwesterly winds have been very important in initiating dust storms in the Southern and Central Great Plains during historical time. The question then is how important the southwesterly winds were prehistorically. Fent (1950) demonstrated that two to three generations of sand dunes in the Great Bend of central Kansas were oriented by southwesterly or southerly winds. Several ^{14}C ages from the sand sequences in the Great Bend Sand Prairie show that the dunes were oriented primarily during the Holocene (Feng et al., 1991; Johnson and Sophocleous, 1991). Particle-size analysis of samples derived from W-E and S-N transects in the Sand Prairie also supports this conclusion (Feng, 1988a). It is likely that southwesterly winds dominated central Kansas during the Holocene as they do at present (Kutzbach and Wright, 1985; Wright, 1984).

During deposition of the Peoria Loess, northwesterly winds were undoubtedly predominant in most parts of Kansas, and southwesterly winds brought coarse particles from the Canadian River valley to southwestern Kansas (Frye and Leonard, 1951). Holliday (1989b) observed that the sand percentage and thickness of the loess decrease significantly and systematically from southwest to northeast in eastern New Mexico and western Texas. He also showed that in the region immediately south of the Canadian River, the loess becomes quite clayey. If the clay was transported in suspension from the southwest, it would have been carried into Kansas, at least into southwestern Kansas.

Summary and Problems

The Loveland Loess was presumed to have been deposited during the Illinoian glacial stage (400-125 ka). The presence of weakly developed soils observed within the loess, however, suggests that the Loveland Loess is actually a loess-pedocomplex, indicating significant climatic variation during that time. Recently established terrestrial and marine glacial-interglacial sequences suggest that four glacial-interglacial cycles occurred during Loveland time, or classical Illinoian Stage. The environments represented by the Loveland Loess still remain unknown.

The Sangamon pedocomplex was thought to have developed during the last interglaciation (127-125 to 75-70 ka) under warm and moist environments in the midcontinental United States. This pedocomplex was immediately overlain by the Gilman Canyon Formation, a pedocomplex formed from 35-28 to 22-20 ka. A crucial question is what occurred between 75-70 and 35-28 ka in the midcontinent, particularly, in the Central Great Plains.

The Gilman Canyon pedocomplex, mainly organic-enriched A horizons in the Central Great Plains, is believed to have been formed under low effective moisture, whereas the Peoria Loess, a rapidly deposited silt, is believed to have been deposited from 22-20 to 10.5 ka under high effective moisture. However, it is difficult to understand why the Gilman Canyon pedocomplex on uplands accumulated so much organic matter in dry environments. Furthermore, it is still uncertain that the Peoria Loess, being deposited rapidly and massively within forests and woodlands, contains no forest soil-forming imprints.

The Brady Soil dated at 10.5-8.5 ka was believed to have developed under extensive poorly drained conditions in the Central Great Plains. But, the climate of Brady time remains unknown. Although the Bignell Loess was believed to have been deposited during the Altithermal Period, age and the climate of Bignell time are not yet certain. Particularly, the climatic sequence of the Holocene is not well established. Fortunately, sand sequences in the Great Bend Sand Prairie of central Kansas provide a high-resolution temporal record of Holocene climatic changes.

The Peoria Loess was believed to have been transported predominantly by northwesterly winds from the river-valley sources. The hypothesis should be reevaluated. The Sandhills of Nebraska and areas to the north and west may have functioned as a source of the loess. Judging from the sand dune orientations in central and southwestern Kansas and and historical dust-storm activity in the Central Great Plains, southwesterly winds may have dominated southwestern and central Kansas during the Holocene. The southwesterly winds may also have contributed some fine particles of loess to the upper portion of Peoria Loess in southwestern and central Kansas during the late Pleistocene.

CHAPTER III. METHODOLOGY

Field investigations were conducted to select research sites and to understand the geomorphic and stratigraphic contexts of the research sites. Based on the contexts, sampling strategies were implemented. Thermoluminescence (TL) and radiocarbon (^{14}C) methods were employed to establish the chronologies. Magnetic measurements, mineralogies, particle size and chemical analyses were adopted to reconstruct environment and climate from the loess sequences in central Kansas.

Field Investigation and Sampling

The five research sites selected for this study and analytical laboratory procedures to be used are described in this chapter. The stratigraphies and their physical and chemical characteristics are discussed in Chapter IV. The specific data for the diagrams and tables of Chapter IV appear in the appendices of this thesis (Appendix 1-5). The five research sites selected for this study are located in Figure 1. This study focuses on sites B (Barton County Landfill section) and D (Pratt County Landfill section). However, supportive information was obtained from sites A (Phillips County Landfill section) and E (Kearny County section), and a Holocene sequence was derived from central Great Bend Sand Prairie (site C). The sites used for supportive information are discussed first, and then the two focal sites (B and D). Finally the Holocene locality is described.

Phillips County Site (A)

The Phillips County landfill site is located at SE1/4, Section 34, R18W, T35S, Phillips County, north-central Kansas. The site is on the loess-mantled divide between Deer Creek and Plotner Creek (a large tributary of Deer Creek), one mile southwest of Phillipsburg. At this site, the sequence is: surface soil (0-0.5 m b.s), Bignell Loess (0.5-1.4 m), Brady Soil (1.4-2.0 m), Peoria Loess (2.0-4.4 m), Gilman Canyon pedocomplex (4.4-5.5 m), Sandy Silt I (5.5-6.4 m), Sangamon pedocomplex (6.5-8.0), Sandy Silt II (8.0-9.6 m), and Loveland Loess (9.6-14 m). Fourteen samples were obtained from stratigraphically unique units for deriving mineralogical, chemical and particle-size information.

Kearny County Site (E)

The Kearny County backhoe-trench site is located at NE1/4, Section 30, R37W, T23S. The site is on the loess-mantled upland interfluve between Matroy Draw and Shirling Creek, north of the Arkansas River in southwestern Kansas. The sequence is composed of only two stratigraphic units: surface soil (0-0.8 m) and Peoria Loess (0.8-5.49 m). Thirteen samples were taken from different horizons of the surface soil and different stratigraphic positions of the Peoria Loess. Samples for this site were collected and provided by M. D. Ranson, Kansas State University. Particle size, mineralogies and chemical properties were derived from these samples.

Pratt County Site (D)

The Pratt County landfilll site is located at NW1/4, Section 21, R13W, T28S, Pratt County, central Kansas. This site is on the loess-mantled divide between Sand Creek (a tributary of the Medicine River) and the Ninnescah River , 3.5 miles south of Pratt, and one mile east of State Highway 281. The landfill trench had been recently excavated at the time of sampling and was 5 m deep. Two meters more were excavated by hand, for a total exposure of 7 m. The upper portion (0-3.4 m) is characterized by alternating occurrences of loess and soils and the lower portion (3.4-5.8 m) by alternating occurrences of carbonate-enriched eolian sandy silt and carbonate-depleted eolian sand. Beneath 6 m there is a fluvial sand unit. The well-preserved surface soil is underlain by a thin loess, the Bignell Loess (0.4-0.8 m). The Brady Soil (0.8-1.4 m) underlies the Bignell Loess and overlies the Peoria Loess (1.4-2.4 m). The Peoria loess, in turn, rests upon the Gilman Canyon pedocomplex (2.4-3.2 m). Underlying the Gilman Canyon pedocomplex is the reddish Sangamon pedocomplex, and beneath it is the Loveland sand-silt sequence (3.4–5.8 m), with four sand layers characterized by carbonate depletion and four silt layers characterized by carbonate enrichment.

Fifty-five samples were taken every other 5 cm from the top to 5.5 m, and ten samples were collected from 5.8 to 7 m every other 15 cm interval. They were collected for derivation of particle size, mineralogy, and chemical properties. One hundred fifty samples were extracted in 1 cm3 boxes at 2 cm interval for magnetic measurements. Two paleosols, the Brady and the Gilman Canyon, were sampled for ^{14}C dating. One sample at the bottom of the eolian silty sand (5.6–5.8 m) was taken for thermoluminescence dating.

Barton County Site (B)

The Barton County landfill site is located at NE1/4, Section 12, R13W, T19S, Barton County, central Kansas. This site is on the loess-mantled divide between the Arkansas River and Cheyenne Bottoms, 3.5 miles north and 3.5 miles east of Great Bend, and a half mile north of the State Highway 156. The eastern face of this landfill exposure is 3 m in height. The western face was approximately 10 m; five more meters were excavated by backhoe to a total depth of 15 m. Eight cores drilled near this site confirm that the total thickness of loess is 18 m (Kansas Biological Survey and Kansas Geological Survey, 1987).

The surface soil was removed above the western face by landfill operation, but was well preserved in the eastern face. From top to bottom, there is the complete late-Quaternary sequence: surface soil, Bignell Loess, Brady Soil, Peoria Loess, Gilman Canyon pedocomplex, Sangamon pedocomplex, Loveland Loess, and a sand unit (4.2-6.6 m) between the Sangamon pedocomplex and the Loveland Loess.

Sixty samples were taken at alternate 10 cm intervals from the top to 12 m, with five samples from 12-15 m taken every other 25 cm interval. These samples were collected for particle size, mineralogical, and chemical analyses. Two hundred sixty samples were collected in 1 cm3-boxes at alternating 2 cm intervals for magnetic measurements. Three organic samples (upper Brady, lower Brady, and the Gilman Canyon) were collected for the ^{14}C dating of soil humates. Four nonorganic samples, the bottom of the Sangamon pedocomplex (4.2 m), the bottom of Barton sand (6.6 m) and at depths of 10 and 12.5 in the Loveland loess, were collected for TL dating.

Central Great Bend Sand Prairie Site (C)

This site was a backhoe-excavated trench in northern Pratt County, central Kansas. It is located at NE1/4, Section 11, R14W, T26S, near recharge well 7 of Groundwater Management District 5.

The upper portion of the sequence (0-1.65 m) is sand, and the lower portion is silt (1.65-2.75 m). The sequence consists of surface blow sand (0-0.2 m), a weakly developed sandy soil (0.2-0.5 m), a sand layer (0.5-0.8 m), another weakly developed sandy soil (0.8-1.0 m), a thick sand unit (1-1.65 m); a strongly expressed paleosol (1.65-1.95), an eolian silt unit (1.95-2.25 m), and another soil (2.25-2.75 m). Nine samples were obtained from unique stratigraphic units for particle size, mineralogical, and chemical analyses.

Chronological Techniques

The most complete records of the Pleistocene climate are from the study of deep-sea sediments in which the stratigraphic record of oxygen isotope ($^{18}O/^{16}O$) variations, reflecting terrestrial ice volume, is the primary recorder of climate. On continents, the available record is far less complete because glacial or other geological events are erosional rather than depositional. The area of loess deposition may be the best place to establish a relatively complete regional chronostratigraphy that is comparable with deep-sea records. For reconstruction of paleoclimate from loess deposits, a complete and relatively detailed chronology is essential. To build a detailed and continuous chronology of loess deposition in central Kasnas, the TL technique was employed for the stratigraphically unique units older than 50 ka, and the ^{14}C technique was adopted for the stratigraphically unique units younger than 50 ka.

Thermoluminescence (TL) Dating

General Principle. The phenomenon of thermoluminescence has been discussed in many publications (e.g., Aitken, 1985; Daniels et al., 1953; Fleming 1971; Huntey and Johnson, 1976; Hutt and Smirnow, 1982; Wintle , 1987; Wintle et al., 1984). The principle of TL dating is that heating (e.g., pottery) or exposure to sunshine (e.g., eolian silt) at a certain time in the past released all energy accumulated by such radioactive impurities as K-40, Th-series, U-series and set the "energy clock" at zero. From that time on, the nuclear radiation continued to cause ionization that begin to refill existing electron traps and progressively increase TL signals. The "TL" age of a sample has been defined by Aiken (1985) as follows: Age (ka) = [Equivalent Dose (GY)]/[Dose Rate(GY/ka)]. The unit of measurement of dose is the rad (radiation absorbed dose), which is defined as the absorption of 100 ergs per gram.

Paleodose Measurement. Since the additive-dose procedure (modified total bleaching procedure) is most applicable for loess deposits which stayed in the atmosphere or on the ground for several weeks before burial (Lu Yanchou et al., 1987a, 1987b) , the method was used in this study. The TL dating was conducted by Dr. Lu Yanchou of Xian Loess Research Laboratory, People's Republic of China. For determing the TL residual, besides exposing the sample to sunshine for 94 hours, each sample is irradiated by using 6-7 β irradiations. Although it can be established that the laboratory bleaching is long enough for only the unbleachable component to remain, there is the risk that exposure to sunshine prior to deposition was not sufficient to reach this level. A check is made by the plateau test in which the paleodose is plotted against the glow-curve temperature. First the natural TL peak is determined. The sample is then irradiated at a heating rate of 5°C per second for a sufficient time until the peak (plateau) nearly completely disappears. The TL signal at the point where the peak disappears is considered as the residual component (Figure 4). Besides the residual level (i.e., zero point), many other factors, such as post-depositional change, specific method used, and type and size of grains, also influence on the TL results. In this study, the influence of these factors has been taken into account, and therefore, the results are reliable. A detailed discussion of these factors is in Appendix 6.

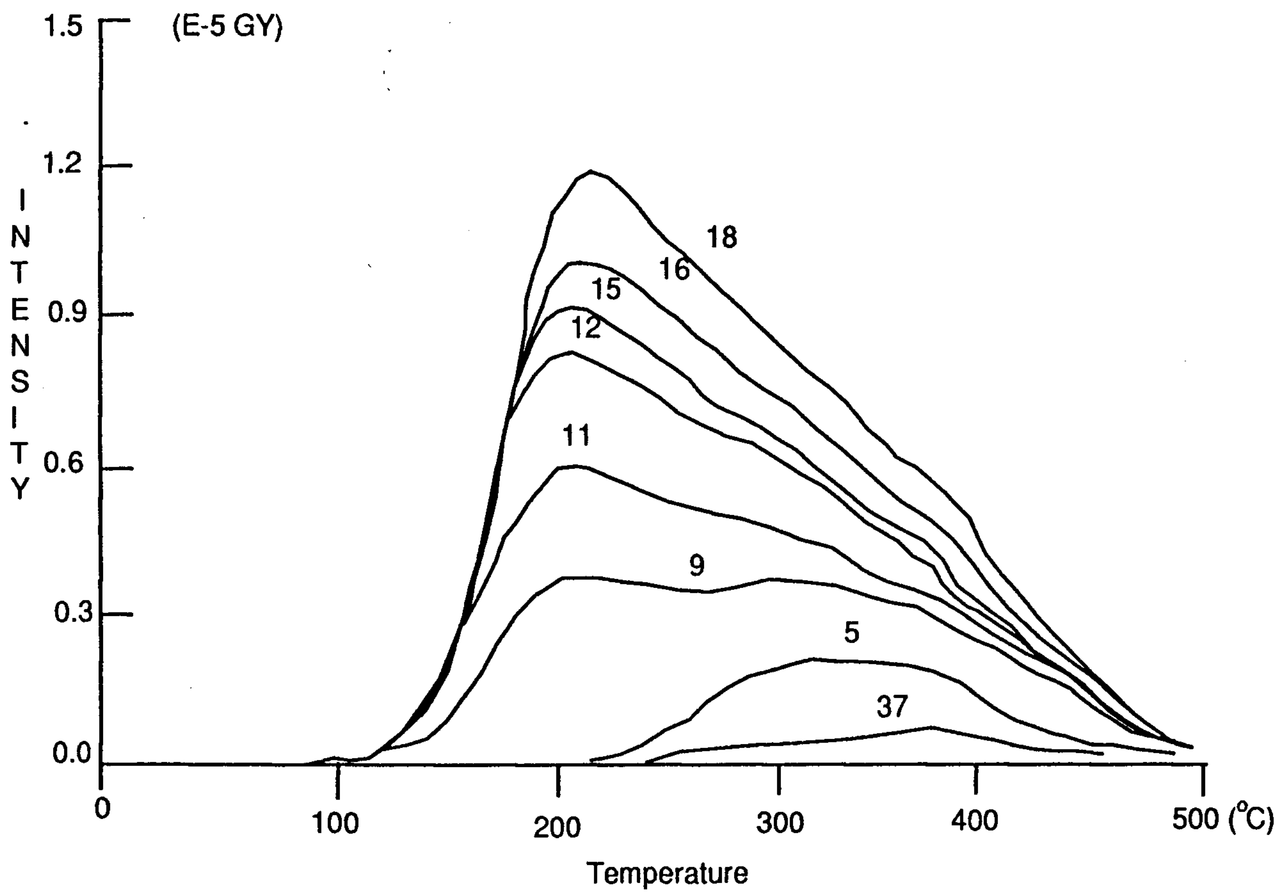


Figure 4. The TL glow curve of fine grains (4-11 μ) from sample XAL-77 (BT4) at a heating rate of 10°C per second. The curve 5 is the natural TL; the curves 9, 11, 12, 15,16 and 18 are the natural TL plus 179, 358, 536, 715, 894 and 1190 Gy β-radiations, respectively. The curve 37 is the residual TL after 20 hour exposure to sunlight.

For this study, about 5 mg of fine-grained (4-11 μ) sample was deposited on each of 16 etched aluminium disks. The sample was evenly distributed over a disk and held in place by a silicon spray. The prepared quartz grain material was bleached under a sunlamp for 94 hours, thus reducing its TL to a minimum. To establish a laboratory-induced TL growth curve, a range of b (^{90}Sr) irradiations were emitted to these disks, up to approximately twice the natural accumulated equivalent dose. Eight disks were used to determine the value of the natural TL, and eight other disks were used to check the possibility of a ensitivity change due to bleaching and irradiation. Following irradiation, at least 24 hours was allowed to elapse prior to glow-out. Sample disks were heated to a temperature of 500° C at a rate of 5°C per second in a high-purity argon atmosphere (Daybreak TL system). In each case there is a prominent TL peak at approximately 350°C. The paleodose can be induced from the ratio of any one of the six b irradiated curves (N+β1......N+β6) to the natural TL (NTL), if all of these seven curves reach their eaks at approximately the same temperature; there is a tendency to level off at about 350°C (Figure 5). By preheating for one minute at 560°C before measurement (both natural TL and natural-plus-artificial TL), a satisfactory plateau

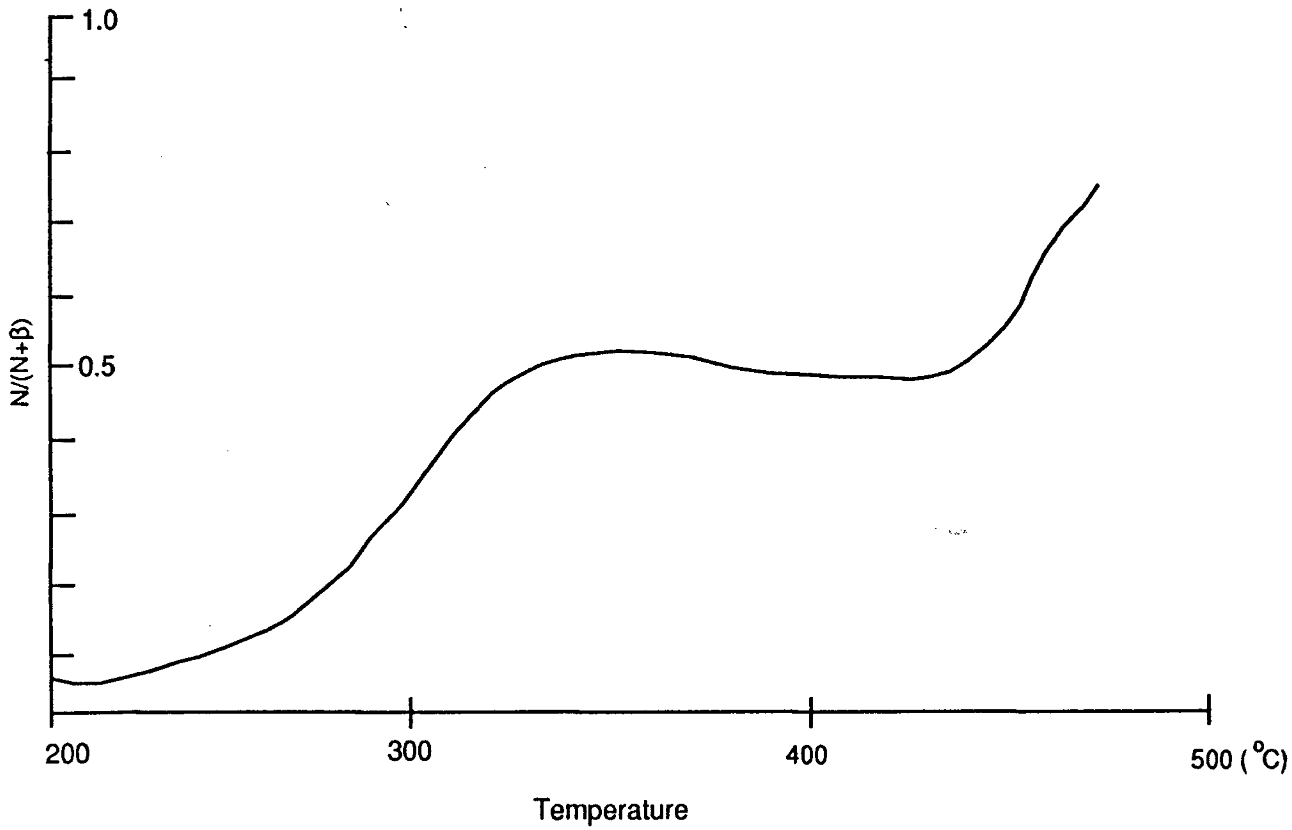


Figure 5. The plot of the ratio of N/[N+ß] (100Gy) versus temperature for sample XAL124 (Pratt County section) showing a ratio plateau for temperature of about 300-450°C.

extending down to about 250°C was established. The onset of satuation for the high temperature peak is in the region of 5000 GY. Then the starting point of the natural TL is determined.

Six to seven natural-plus-artificial TL signals are plotted against the additive dose after measuring the residual TL (G_0). Extending the TL intensity line to intersect with the G_0 line, the equivalent dose Q, the length from the intersection point to zero additive dose point is obtained (Figure 6).

Annual Dose Rate Determination. The contributions of U-series and Th-series to the environmental dose rate were calculated from the mean U and Th contents obtained by both neutron activation analysis and XRF analysis. Thick-source g counting was used to check disequilibrium in the decay series.The K-40 contribution was calculated from the total potassium content determined by atomic absorption spectrum analysis. The dose rate conversion factors of Bell (Aitken, 1985) were used. The cosmic ray contribution was also taken into account (Prescott and Stephan, 1982), as well as the rubidium content. The sample TL age was derived from the mean of the values at 25°C intervals along the TL age versus temperature plateau region.

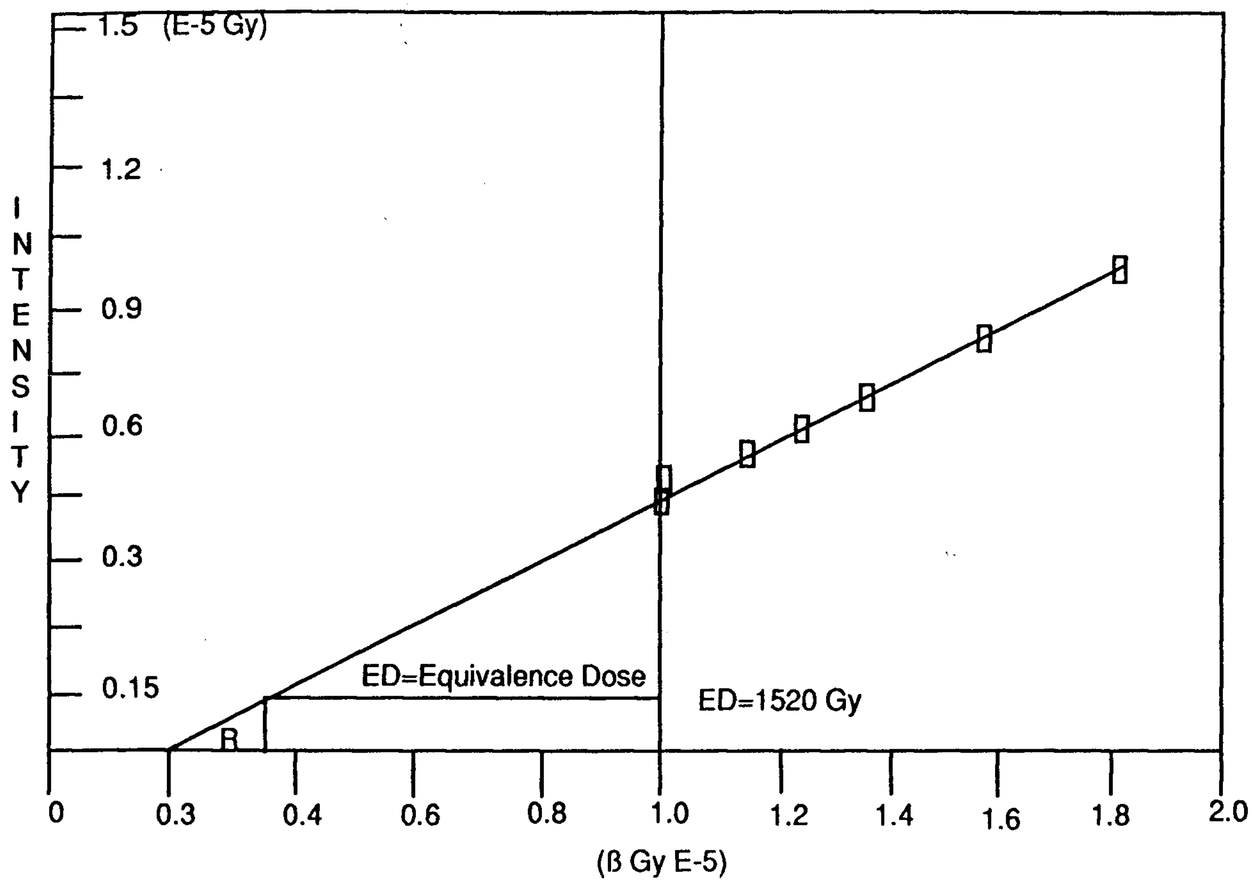


Figure 6. The relationship between the ß radiation intensity (E-5 Gy) and the emitted TL intensity (E-5 Gy) at a temperature of 350°C, showing the Equivalence Dose (the intercept of the curve) and the residual (R).

Radiocarbon (^{14}C) Dating

Stuiver (1982), Groots (1983), and Terasmae (1984) reviewed radiocarbon techniques and their applications in great detail. The following assumptions for radiocarbon dating are: (1) the sample has been in a closed system with respect to carbon exchange with its environment, (2) the atmospheric ^{14}C concentration has been constant through time, and (3) the initial ^{14}C concentration is independent of place and environment. The closed system means that no ^{14}C has been added or released since the plants and animal, the ^{14}C carriers, died. Based on the half-life of ^{14}C (5730+/- 40 years), the age (T) of a sample containing radiocarbon-14 is determined by the ratio of the present ^{14}C activity (A) to the activity (A_0) of ^{14}C in the same specimen at the time the plants were alive: $T = 19.035 \times 1000 \log (A_0/A)$ Years.

There is no challenge to the third assumption. But, the second assumption, the constancy of atmospheric ^{14}C concentration, has been vigorously challenged. Through comparison with tree-ring chronologies, Stuiver (1982) and Stuiver and others (1991) found that ^{14}C content varies with ^{12}C, which has not been constant at least during the past 9000 years. He provided a calibration curve for ^{14}C age to correct to calendar year, according to the tree-ring chronologies. Beyond the 9,000-year limit, Bard and others

(1990) stated that ^{14}C ages are systematically younger than the U-Th ages of corals, with a maximum difference of about 3,500 years at about 20,000 yr BP. The U-Th ages are not affected by changes in the cosmic ray flux, and therefore can be compared directly with tree-ring ages. Investigations show that ^{13}C (a stable carbon isotope) is relatively abundant in the atmosphere and relatively constant through time. ^{14}C is being used to determine the degree of fractionation of ^{14}C relative to ^{14}C in different organisms; therefore it is useful for comparing ages determined on different materials (Faucre, 1983).

Radiocarbon dating of paleosols measures the decay of organic carbon present in soil humates. Because the age of humates varies, age determinations from soil humates measure the mean residence time of the humates in the soil. The soils developed in the loess region of the midcontinental United States are almost exclusively cumulative ones, i.e., they formed as eolian (silt) deposition continued, though at a very slow rate. There was mixing of younger with older humates, but the average residence time became younger and younger until pedogenesis was exceeded by the rate of loess deposition. Once buried, and thus isolated from the surface, the soil no longer receives young humates (Schaetzl and and Sorenson, 1987).

Apart from in situ contamination, the admixture can also be contaminated during sampling, storage, and handling in the dating laboratory. To minimize in situ contamination, ^{14}C samples were derived from freshly excavated faces. The post-collection contamination was minimized through careful sampling, storage, and pretreatment. The sample representing the upper portion of the soil was collected from the upper five centimeters of the Ab horizon, while the sample representing the lower portion of the soil was taken from the lower five centimeters of the Ab horizon. If only one sample was taken for a soil, it was derived from the middle portion. The samples were collected and stored in plastic bags. Before being sent to the laboratory, the samples were ground to fine-sand size, and then visible plant rootlets were picked out. Before separating the insoluble humate fraction from the soil material, microscopic pieces of charcoal and remaining rootlets were removed through floatation. The ^{14}C dating was conducted by S. Valastro at the Radiocarbon Dating Laboratory of the University of Texas at Austin.

Magnetic Measurements

Remnant Magnetic Intensity

A great deal of attention has been devoted to magnetic measurements of loess sections in Europe (Heller et al., 1987) and in China (Liu and Heller, 1982; Liu et al., 1987). Most attention has focused on finding recognizable paleomagnetic events for absolute dating purpose. Previous workers have carefully investigated variations in remnant magnetic intensity to determine what factors control the magnetic intensity. To explain differences in the remnant magnetic intensity between loess and paleosols, several possible factors have been proposed, e.g., differences in the magnetic field intensities between the times of soil-forming and loess deposition, differences in the mechanisms for obtaining the remnant magnetism between processes of soil formation and loess deposition, probable differences in the parent material sources between loess and the soils (Sadajima and Wang, 1983). The remnant magnetic intensity is measured with a minispin magnetometer after routine stepwise thermal demagnitization in a Schonstedt oven at temperatures up to 350°C.

Magnetic Susceptibility

Magnetic susceptibility of loess sequences has been used to document changes in loess deposition through time that have been interpreted as being caused by changes in climate (Kukla et al., 1988; Wang et al., 1990). Researchers propose that susceptibility varies with the degree of pedogenesis and the degree of oxidization, and can, therefore, serve as a proxy indicator of climatic conditions. Zhou and others (1990) confirmed that the susceptility of the Chinese loess-soil sequence is controlled by the size of magnetic particles, i.e., loess has coarse magnetic particles and soil has fine magnetic particles, which have higher susceptibility than coarse particles. Beget and others (1990) interpreted susceptibility changes in Alaska loess as representing changes in wind intensity and competence. The basic principle relating susceptibility to paleoclimate is that pedogenic processes concentrate (leaching) and form (oxidization) sesquioxides. Some oxides, such as Fe_2O_3 (hematite), $FeTiO_4$ (Ulvospinel), $FeTiO_3$ (Ilmenite), FeOOH (mainly geothite), are magnetic.

Susceptibility of samples from the study area was measured by using the bulk susceptibility head of Minisep alternating-current bridge and was expressed with respect to the sample volume, which was generally 8 cm^3 . This bridge consists of two ferrite rings with windings carrying an alternating current, which produces an alternating magnetic field of up to 10 oersteds across gaps cut in the rings. A ferrite plug can be positioned to balance the two circuits so that the specimen placed within one of the gaps unbalances the circuit in secondary windings in proportion to its susceptibility. Dr. D. R. Sprowl did the analyses in his laboratory.

Particle Size Analysis

Particle-size distribution is important in considering soil properties and the dynamics of silt transportation. Clay content in a specific soil profile, particle-size distribution throughout a loess section, and comparison of the particle-size distribution from different loess sections may produce insight into the degree of weathering and soil development (Ruhe, 1983). Particle-size analysis can also provide insight into the past wind patterns, thus, spatial variation of the environment of loess deposition (Liu et al., 1985).

Sand percentage and mean size of silt can serve as proxy of wind intensity (Zhu et al., 1985). Different fractions of particle size are carried by different wind transporting processes, e.g., sand is rolled along the ground, fine sand by saltation mainly within 2 m height from the ground, and silt and clay in suspension at different heights. Thus, traction and windborne fractions should be considered separately in determining the local and long-distance contributions of loess.

Pipette Method

The pipette method (Janitzky, 1987) was used for particle-size analysis. The basic principle of this analysis is that spherical particles will settle in fluid at a rate proportional to their radius (Stoke's Law). Samples were treated to remove organic, soluble salts using 100 cc distilled water and 300 cc hydrogen peroxide (H_2O_2). Carbonate was removed using diluted 6NHCI. The samples were then dispersed in a sodium solution. Sand was separated by wet sieving, and silt and clay were washed into sedimentation cylinders. Twenty cc of buffer dispersing agent (Calgon) was then added and the solution was mixed for five minutes. The solution was then placed in a one-liter settling cylinder. After eight hours, 10 cm solution in the cylinder was pipetted under a room temperature of 20°C. The clay in the pipetted solution was collected by centrifugation. The solution was mixed and pipetted three times to extract clay as completely as possible. Subsequently, the sand, silt, and clay were dried in an oven at 110°C. The sand was dry sieved into different fractions (0, 1, 2, 3.25, 4.25 phi units). The silt and clay were weighed separately. Then the silt and clay were mixed for a detailed particle analysis by using an electronic counter. The pipette analysis was done in the Soils Laboratory of the Department of Geography, University of Kansas.

Electronic Counter

To get a high-resolution curve of particle-size distribution, a dispersed, unseparated silt and clay admixture was analyzed using an Electronic Sensing Zone Coulter Counter in the Byrd Polar Research Center at Ohio State University. This instrument provides rapid, detailed information regarding particle numbers of each particle-size interval defined. The counter analyzed particles sized from 80 to 0.3 μm.

The ultrasonic technique was employed to transfer particles directly into the aqueous electrolyte solution. The membrane filter and electrolyte were placed into a beaker immersed in a ultrasonic bath. The counter provides a flow-metering device, which allows particle concentration to be calculated from the number of particles counted in the metered volume of liquid. The data can be expressed either as percentage of particle numbers of each fraction or as weight calculated from the particle number of each fraction (Graf, 1977).

Mineralogical Analysis

Clay Mineralogy

Clay mineralogy was determined using an X-ray diffractometer. The percentages of quartz and kaolinite in the clay fraction were determined, and the kaolinite/quartz ratio was calculated (Carroll, 1970). The ratio and these percentages can reveal the intensity of soil development

(Birkeland, 1974) and the intensity of loess weathering (Wang, 1982).

The X-ray powder diffraction method was employed to measure the 2θ maxima and their relative intensities. Each 2θ value is converted to a d-spacing by using Bragg's law. A photon detector, typically a scintillation detector, is placed behind a receiving slit and converts the diffracted X-ray photons into voltage pulses. These pulses are integrated in a rate meter to give an analog signal on an x-t recorder (Brown, 1966). To identify the minerals, the x-t patterns were compared with computer-based retrieval patterns (Jenkins, 1988).

In this study, this author was interested in the kaolinite/quartz ratio in clay, because quartz in clay is mainly a product of physical breakage, and kaolinite is formed primarily by chemical weathering processes (Birkeland, 1968, 1984; Carroll, 1970). The kaolinite/quartz ratio, if other factors are constant, can reveal the intensity of soil development (Wang, 1982).

To obtain the kaolinite/quartz ratio in the clay, mixtures of kaolinite and quartz with other known minerals, totaling 0.3 grams, were prepared to establish references for quantitative estimates. The reference samples were: (1) IL (Illite)=25%, Mo (Montmorillonite)=25%, KA (kaolinite)=25, QU (quartz)=25%; (2) IL=25%, MO=25%, KA=15%, QU=35%; and (3) IL=25%, MO=25, KA=35%, QU=15%.

The references were x-rayed, and diffractometer tracings were obtained at the rate 1^{o} 2θ per minute. Based on the energy-counting ratios of IL/KA, IL/QU, MO/KA, MO/QU for the three reference samples, the references were established, i.e., the energy intensities of the kaolinite and quartz relative to other known minerals are realized. The sample to be measured was then x-rayed. The energy-counting comparison between the reference samples and the measured sample estimates the percentages of kaolinite and quartz of the measured sample.

Fine-sand Mineralogy

Similar procedures were employed for fine-sand mineralogy. In some cases chemical parameters are not preserved because of post-depositional alteration, so Ruhe (1974) suggested a mineralogical index, the quartz/felspar ratio of fine sand, to estimate the intensity of weathering. To obtain the ratio of feldspar and quartz in the fine sand fraction, three mixtures (reference samples) of quartz and feldspar with other known minerals, totaling 0.3 gram, were prepared according to the proportions shown below: (1) CH (Chlcedony)=25%, FL (Flint)=25%, FE (Felspar)=25%, QU (Quartz)= 25; (2) CH=25, FL=25%, FE=15%, QU=35%; and (3) CH=25%, FL=25%, FE=35%, QU=15.

The reference samples were x-rayed, and diffractometer tracings were obtained at the rate of 1^{o} 2θ per minute. Based on the ratios of CH/FE, CH/QU, FL/FE, FL/QU for the three reference samples, references were established. The energy counting comparison between the reference samples and the measured sample estimates the percentages of quartz and feldspar and, therefore, the quartz/feldspar ratio of the measured sample (Seitlheko, 1975). All mineralogial analyses were conducted under guidance of Mr. L. M Magnuson in the Analytical Laboratory of the Kansas Geological Survey.

Chemical Analysis

Chemical Composition of Silt and Clay

An x-ray fluorescence spectrometer was used to analyze the chemical composition of the fraction finer than 54 m in loess and the soils developed within it. The chemistry addresses the history of soil development and loess weathering. The following chemical compounds were determined: SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MgO, CaO, Na_2O, K_2O, P_2O_5. By ratioing the stable compounds (Fe_2O_3 and Al_2O_3) with unstable compounds (MgO, CaO, Na_2O, K_2O, and P_2O_5), a leaching index (L.I = $[MgO+CaO+K_2O+Na_2O+P_5]/[Al_2O_3+Fe_2O_3]$) can be obtained (Buol et al., 1980; Wen and Sun, 1981). Also, the ratio of Fe_2O_3/Al_2O_3, oxdization index (O.I), is a indicator of the intensity of oxidization (Araki and Kyuma, 1983).

Quaternary geologists (e.g., Liu, 1985; Liu and Yua, 1987; Pecsi, 1985; Wang, 1982) have been successfully employing chemical indices in reconstructing loess depositional environments, although Ruhe (1974) contended that chemical indices are not reliable for such due to post-depositional alteration. Kukla and others (1988) and Wang and others (1985) demonstrated, however, that post-depositional alteration in Chinese loessial soils is rather limited. Further, Ransom and others (1987) showed that chemical indices are useful in distinguishing environments in loess deposits of Ohio.

To determine the major chemical composition of loess and its soils, an energy dispersive spectrometer was used. The spectrometer consists of only two basic units, the excitation source and the detection system. The detector is typically a Si one, which is a proportional detector of high intrinsic resolution. A multichannel analyzer is used to collect, integrate, and display the resolved pulses. The spectrometer is fitted with multisample handling facilities, and is automated by using a microcomputer. The precision of measurement for major chemical composition is a few tenths of one percent (Genkins, 1988). The spectrometer has a range from 0.4 to 20 Å (40-0.6 KeV), which permits measurement of the K-series from Fluorine (Z=9) to Lutecium (Z=71) and the L-series from Manganese (Z=25) to Uranium (Z=90).

Matrix corrections are made by using the nominal percentage. Matrix correction coefficients have been determined experimentally and agree with theoretical values. Once the coefficients are established, any sample of known composition can be used to calibrate for nominal concentration for a particular oxide of the sample to be measured. To convert each element into oxide form, one has to use a converting factor and an experimental coefficient for a specific element to calculate the oxide based on the relationship between the measured energy intensity and the element concentration (Norrish and Hutton, 1969).

Everything discussed above is set in advance through the calibration processes and by making a standard reference disk. Sample preparation is extremely time-consuming. To obtain good results, one must strictly follow the procedures in preparing sample disks. The specimen for analysis was prepared by mixing 1.5 grams of borate glass (Lanthanum oxide, Lithium tetraborate, and Lithium carbonate) and 0.28 gram of powdered sample (loess). The mixture was heated to 1000°C and kept at that temperature until all the speciman was melted. The melt was stirred in a crucible. An aluminium plunger and a graphile disk were kept on a hot plate at 220°C. A brass ring was put over the disk, and the melt was poured onto the center of the disk. Immediately, the plunger was brought down gently to mold and quench the melt. A uniformly thick and round glass disk was ready for mounting into the X-ray specimen holder. This work was done in the X-Ray Laboratory of the Department of Geology, University of Kansas.

Organic Carbon and Carbonate Determination

Organic matter content was measured by dry combustion through loss on ignition. This method is only an approximation of the total soil organic matter content. First, soil and loess samples were ground into fine-sand size, and then put into an oven to dry at 110°C for 12 hours. Ten grams of each sample were put in a weighed crucible which was then transfered to a muffle furnace and heated up to 600°C for 4 hours. When the furnace was cooled to 100°C, the crucible was transfered to a desiccator and cooled for 20 minutes. Finally, the crucible plus the ignited sample was weighed. The loss (%) on ignition (LOI) is defined as follows (Jackson, 1969):

LOI = (Loss in weight)/(oven dry weight).

A similar technique was used to determine the concentration of carbonate. Having been ignited and weighed for measuring the organic matter, the crucible plus the ignited sample was immediately put back into the furnace and reignited at a temperature of 950°C for 12 hours. Then the furnace was cooled to 100°C. The crucible with sample was then transfered to a desiccator and cooled for 30 minutes. The crucible plus the reignited sample was weighed (Galle and Runnels, 1960, Hill, Galle and Runnels, 1961). The carbonate loss (%) on ignition (CLOI) was calculated as follows:

CLOI = Loss (second ignite)/weight (first ignite).

Using the same principle, the concentration of carbonate was remeasured when making a glass disk for chemical determination by using an x-ray fluorescence spectrometer. Loss on ignition at a temperature 1000°C during fusing process, expressed as LOI, is a double check for measurement of carbonate concentration. This analysis was done in the Soils Laboratory of the Department of Geography, University of Kansas.

Carbon and Oxygen Isotopic Analysis

Carbon and oxygen isotopic analyses have been increasingly attracting attention in identifying modern ecological and soil-forming conditions (e.g., Kelly et al., 1990; Quade and Cerling, 1990; Shlesinger et al., 1988). Studies of modern analogues have created interest in using isotopic analysis to reconstruct paleoenvironments (Cerling et al., 1989; DeNiro, 1982; Gardner et al., 1991; Krishnamurthy, 1990; Ludvigson et al., 1990). The oxygen isotopic composition of pedogenic carbonate is determined by temperature and isotopic composition in the soil water. The latter is controlled, in turn, by the isotopic composition of precipitation entering the soil and the extent to which the soil water is subsequently enriched in Oxygen-18 by evaporation (Gardner, 1984). In general, an increase in soil temperature would reduce the value of $\delta^{18}O$ (more negative); however, in continental interiors an increase of air temperature will cause an increase of $\delta^{18}O$ of precipitation (C.I.A.E.A., 1970). Thus, to produce an increase in the $\delta^{18}O$ of pedogenic carbonate under a regime of progressive aridity associated with rising air and soil temperatures requires an increase in the fraction of precipitation entering the soil. Water is much more important in contributing oxygen-18 to soil carbonate than atmospheric CO_2. Consequently, the $\delta^{18}O$ in soil carbonate has an excellent potential for providing paleo-temperature information (Schlesinger et al., 1989).

The ^{13}C of pedogenic carbonate and soil organic matter depends on the interplay of three factors: (1) the relative proportion of C-3 ($d^{13}C$=-25‰) and C-4 ($d^{13}C$=-15‰) plant biomass, (2) the rate of soil respiration, and (3) the rate of diffusive exchange of CO_2 between the soil and the atmosphere (Hoefs, 1980; Cerling et al., 1989). Air temperature has only a weak effect on this fractionation.

Since soil moisture strongly governs plant production and soil respiration rates in arid regions, increasing aridity should result in greater penetration of atmospheric CO_2 and thus pedogenic carbonate which is enriched in carbon-13 as well as oxygen-18 (Amundson et al., 1989; Cerling and Hay, 1986).

Ten samples of CO_2 extracted from carbonate nodules of the middle portion of the loess (5-12.5 m) at the Barton County section were sent to the isotope laboratory of the University of Missouri at Columbia for carbon-13 and oxygen-18 analyses. All carbonate concretions used in this study were mechanically stabilized by the vacuum injection of epoxide resins, and were collected using a small knife. Powdered samples were loaded into 2-ml borosilicate glass ampules and roasted in vacuum at 200°C for four hours to remove volatile organic contaminants. Both ^{13}C and ^{18}O were determined. Samples were exposed to 100% H_3PO_4 at 25°C for 24 hours to release the CO_2 gas. All samples were analyzed by using a Finnigan Delta-E mass spectrometer in Geochemistry Laboratory of University of Missouri at Columbia by Dr. K. Shelton. Analytical precision for ^{13}C and ^{18}O values are considered to be less than +/-0.05 per ml, and values are reported per mil relative to the PDB standards.

CHAPTER IV. RESULTS AND DISCUSSION

Major advances in the understanding of Quaternary climates have been recently made from studies of deep-sea sediments. Because of nondeposition or erosion in terrestrial depositional sequences, terrestrial Quaternary studies have been ignored for decades. Nevertheless, loess studies in Europe and China have demonstrated that loess is not only a more or less continuous Quaternary deposit, but also a relatively sensitive climatic recorder. Unfortunately, the climatic information contained within the widespread loess deposits in the midcontinental United States has not been sufficiently explored. This study of loess deposits in central Kansas is especially germane to the changes in climate and associated environment of the late Quaternary in the midcontinental United States.

In this chapter, lithostratigraphies of the five research sites are first described, and a comprehensive loess chronostratigraphy of central Kansas is established. Finally, changes in climate and associated environment are reconstructed based on the chronosequence of physical, mineralogical and chemical properties of loess deposits in central Kansas.

Stratigraphy

Unique lithostratigraphic units typically represent unique environmental events. The environmental events of loess regions were controlled mainly by changes in climate (Kukla et al., 1988; Liu et al., 1985). Thus, lithostratigraphy of loess deposits is of great importance for reconstructing paleoclimates. Needless to say, time control of the lithostratigraphy is essential to establishing a temporal sequence of changes in climate and associated environment during loess deposition.

Lithostratigraphy

Of the five research sites, the four loess sections (Phillips, Kearny, Pratt, and Barton) not only exhibit temporal variation, but also spatial variation of loess deposition in central Kansas. The Phillips County section of north-central Kansas displays the most complete sequence of the late Quaternary. The sequence includes the surface soil, Bignell Loess, Brady Soil, Peoria Loess, Gilman Canyon pedocomplex, Sandy Silt I (equivalent to Roxana Silt), Sangamon pedocomplex, Sandy Silt II (equivalent to a sand unit found at the Barton County section), and Loveland Loess (Table 1a). In contrast, the Kearny County section is composed of only two stratigraphic units, the surface soil and Peoria Loess (Table 1b).

The Pratt County section (Table 1c) and Barton County section (Table 1d) of central Kansas are quite comparable. The upper portions (post Loveland) of the two sections have exactly the same stratigraphy (Plates 1, 2 and 3). The lower portions (Loveland Formation) are a little different. The Loveland Formation is characterized by the alternating occurrence of carbonate-enriched clayey layers and carbonate-depleted silty layers in the Barton County section and by the alternating occurrence of carbonate-enriched silt layers and carbonate-depleted sand layers in the Pratt County section.

The Great Bend Sand Prairie section provided an excellent sequence of Holocene deposits. The upper portion (0-1.6 m) is sand and the lower portion (1.6-2.65 m) is silt. Two events of soil formation or organic matter accumulation occurred in the upper portion, and two events of soil formation occurred in the lower portion (Table 1e).

The well-known Buzzard's Roost composite section (Figure 7a) of Nebraska (Schultz and Tanner, 1957) is comparable with the Barton County section (Figure 7b), except that a sand unit underlies the Sangamon pedocomplex in the Barton County section. Based on a high correlation

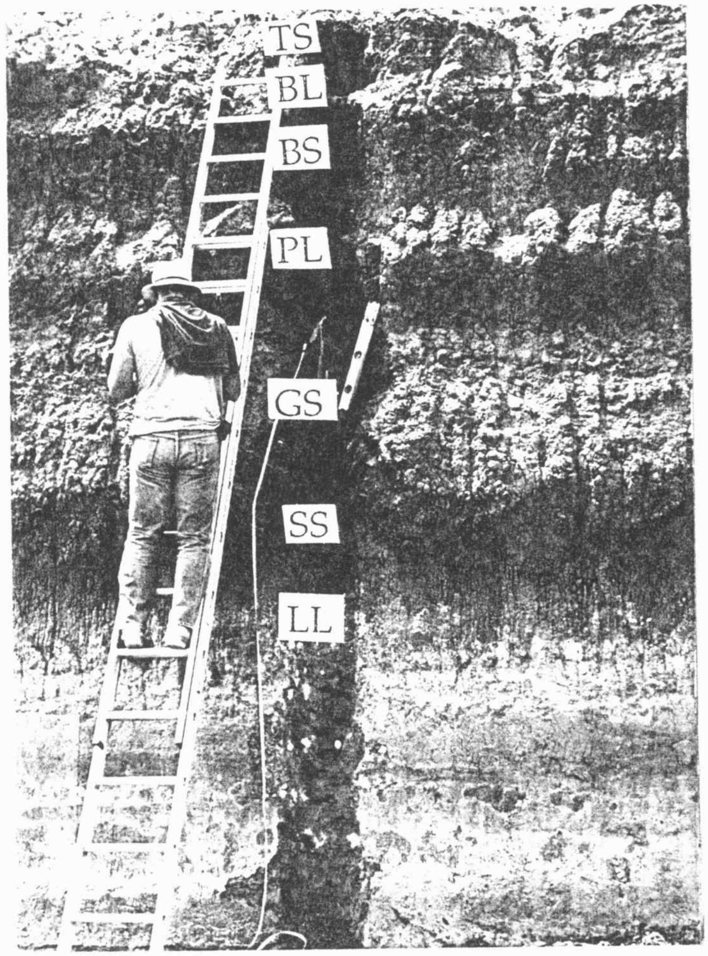


Plate 1. The Pratt County section showing sequence: the surface soil (TS), Bignell Loess (BL), Brady Soil (BS) dated at 6460+/-590 yr BP (TX-6258), Peoria Loess (PL), Gilman Canynon pedocomplex (GS) dated at 20,550+/-590 yr BP (TX-6633) at the top, Sangamon pedocomplex (SS), and Loveland eolian sand(LL) dated 416 ka (XAL-124) at the bottom (5.7 m deep).

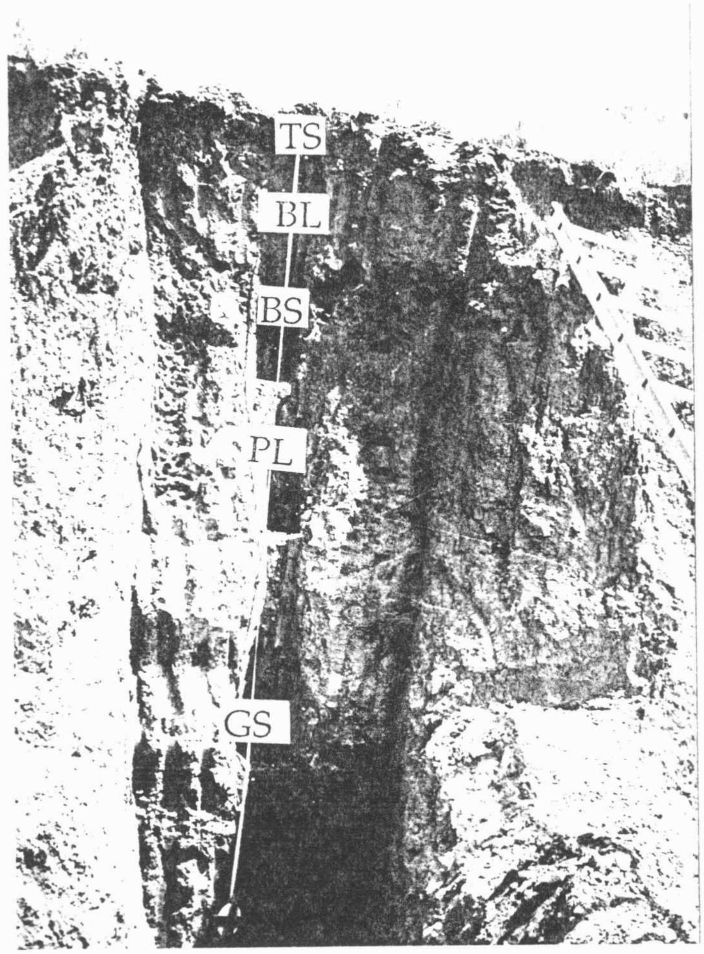


Plate 2. The eastern face of the Barton County section showing the suface soil (TS), Bignell Loess (BL), Brady Soil (BS) dated 9,720+/-110 yr BP (TX-7045) in the upper portion and 10,550+/-150 yr BP (TX-7046) in the lower portion, Peoria Loess (PL), and Gilman Canyon pedocomplex (GS).

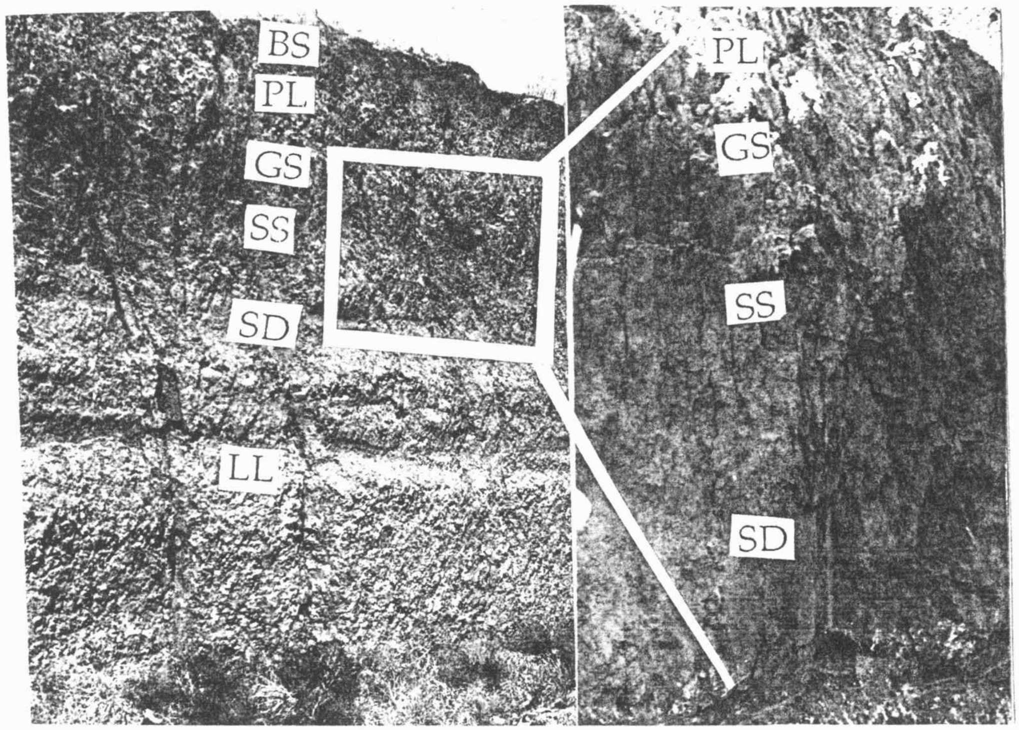
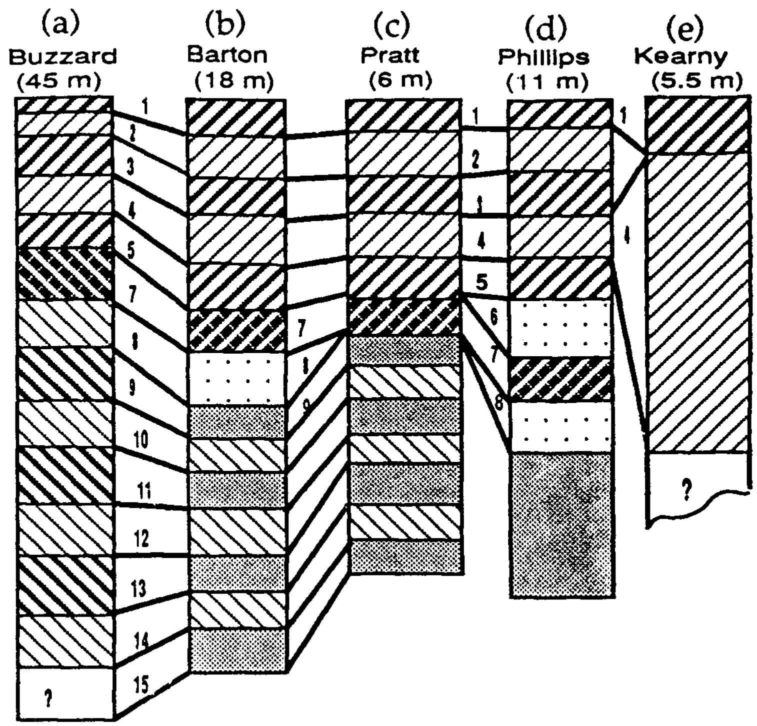


Plate 3. The western face of the Barton County section showing the Brady Soil (BS), Peoria Loess (PL), Gilman Canyon pedocomplex (GS) dated at 24,360+/-570 (TX-5809), Sangamon pedocomplex (SS) dated at 69 ka (XAL-77), Barton sand (a sand unit labeled as SD) dated at 92 ka (XAL-78), Loveland Loess (LL) dated193 ka (XAL-76) at 10 m deep and 260 ka (XAL-75) at 12.5 m deep(this landfill exposes only 10 m, but the depth of the loess is 18 m).

of clay and carbonate, and other evidence, the carbonate-enriched clayey layers within the Loveland Loess in the Barton County section are believed to be pedogenically formed (discussed later in this chapter). If so, the soils at the Buzzard's Roost section are comparable with the carbonate-enriched clayey layers in the Barton County section. The sand unit, or Barton sand as proposed here, is stratigraphically comparable with the sandy loess unit underlying the Sangamon pedocomplex in the Buzzard's Roost section. The Pratt County section (Figure 7c) is stratigraphically comparable with the Buzzard's Roost and Barton sections, except that there is no stratigraphic unit equivalent to the sandy silt in the Buzzard's Roost section or Barton sand in the Barton County section.

In the Phillips County section of north-central Kansas, the stratigraphy corresponds well to that in the Buzzard's Roost and the Barton County sections with two exceptions (Figure 7d): the carbonate-enriched Loveland Loess is not subdivided into carbonate-enriched and carbonate-depleted layers, and in addition to the Sandy Silt II, there is another sandy silt unit (Sandy Silt I) overlying the Sangamon pedocomplex and underlying the Gilman Canyon Soil. This stratigraphically defined sandy silt (Sandy Silt I) is recognized in Nebraska (Ahlbrant and Fryberger, 1980), Iowa and Illinois (Ruhe, 1977).



Notes: **1**: Surface soil, **2**: Bignell Loess. **3**: Brady Soil. **4**: Peoria Loess. **5**: Gilman Canyon Soil Complex. **6**: Sandy Silt I (found at Phillips section). **7**: Sangamon Soil. **8**: Sand (Barton sand), corresponding is a loess unit at the Buzzard's Roost section. **9**: Carbonate-enriched layer I at Barton and Pratt sections or paleosol at Buzzard's Roost section. **10**: Carbonate-depleted loess I or loess unit at Buzzard's Roost section. **11**: Carbonate-enriched layer II. **12**: Carbonate-depleted loess II. **13**: Carbonate-enriched layer III. **14**: Carbonate-depleted loess III. **15**: Carbonate-enriched layer IV (**9-15** consist of the Loveland Loess).

Figure 7. Stratigraphic comparison of the loess sequences in central Kansas with the loess sequence of the Buzzard's Roost section in south-central Nebraska (Schultz and Stout, 1945; Morrison, 1987).

Table 1a. Lithostratigraphy of the Phillips County section.

Depth(m)	Stratigraphy	Descriptions
0.0-0	surface soil:	moderately dark silt, mainly A horizon with slightly developed B horizon.
0.5-1.4	Bignell Loess:	brownish-yellowish, massive structure slightly calcareous silt.
1.4-2.0	Brady Soil:	dark, organic-enriched A horizon with B horizon being relatively well developed, clayey silt.
2.0-5.5	Peoria Loess:	yellow-tan to gray-tan with mottles and streaks of rusty brown, well-preserved charcoal pieces in grass rootlet canals, calcareous silt.
5.5-6.5	Gilman Canyon pedocomplex:	mainly A horizons, dark brownish gray silt, slight angular blocky structure at the very low portion (B horizon).
6.5-6.8	Sandy silt I:	brownish-yellowish, noncalcareous sandy silt.
6.8-8.0	Sangamon pedocomplex:	a slightly reddish silt, calcareous at the base, identifiable B horizon (blocky structure).
8-9.6	Sandy silt II:	slightly reddish to yellowish brown, calcareous sandy silt.
9.6-14	Loveland Loess:	slightly brownish-yellow silt, evident caliche layers and carbonate nodules.

Table 1b. Lithostratigraphy of Kearny County section.

Depth(m)	Stratigraphy	Descriptions
0.0–0.8	surface soi:	dark grayish-brown silt, few fine irregular soft mashes of carbonate at the bottom of the soil.
0.8–5.5	Peoria Loess:	yellowish to slightly brown silt, common very fine tabular pores, common patchy distinct irregular carbonate concretions, prismatic structure, well-preserved charcoal pieces in grass rootlet canals.

Table 1c. Lithostratigraphy of Pratt County section.

Depth (m)	Stratigraphy	Descriptions
0.0-0.2	surface soil:	moderately dark, organic enriched silt.
0.2- 0.6	Bignell Loess:	dark brown, slightly calcareous silt.
0.6-1.2	Brady Soil:	dark grayish-brown silt with light brownish gray mottles, calcareous B horizon with identifiable peds with clay coatings.
1.2-2.4	Peoria Loess:	yellowish silt, soft calcareous mashes, well-preserved charcoal pieces in grass rootlet canals.
2.4-3.0	Gilman Canyon pedocomplex:	a dark brownish-gray silt, noncalcareous, organic carbon-enriched A horizons and identifiablly structured B horizon at the very bottom (subangular blocky structure).
3.0-3.4	Sangamon pedocomplex:	a reddish silt, quite calcareous at the lower portion.
3.4-5.7	Loveland Loess:	eolian sand with four carbonate concentration layers; these carbonate-enriched layers are lower in sand content and high in silt content. Depths of the four carbonate-enriched layers are: 3.4-3.8 m, 4.1-4.4 m, 4.6-4.8, 5.0-5.8 m.

Table 1d. Lithostratigraphy of Barton County section.

Depth(m)	Stratigraphy	Description
0.0-0.2	surface soil:	moderately dark, organic carbon-enriched silt
0.2-0.8	Bignell Loess:	slightly dark yellowish-brown, calcareous silt with secondary calcium concretions
0.8-1.6	Brady Soil:	dark grayish brown silt, mottled with light brownish gray spots, secondary calcareous concretions at the lower half (B horizon), numerous iron-stained channels.
1.6-2.4	Peoria Loess:	yellowish-tan to gray-tan silt, with mottles and streaks of rusty brown, calcareous, well-preserved charcoal pieces in the grass rootlet canals.
2.4-3.4	Gilman Canyon pedocomplex:	a dark browish gray silt, organic carbon-enriched A horizons and slightly structured B horizon.
3.4-4.2	Sangamon pedocomplex:	a reddish silt, noncalcareous, identifiable soil structures.
4.2-6.4	Barton sand:	slightly reddish, calcareous, sorted eolian sand.
6.4-18	Loveland Loess:	pale-yellowish silt, massive structure with four caliche layers. Depths of the four caliche layers are: 6.4-8.2 m, 9.2-11.2 m, 12-14 m, 15.5-18 m.

Table 1e. Lithostratigraphy of Central Great Bend Sand Prairie section.

Depth(m)	Stratigraphy	Descriptions
0.0-0.75	blow sand:	yellow-tan, loose sand with very slight organic carbonate accumulation and abundant grass roots.
0.75-1.05	soil:	loose yellowish-brownish gray silty sand with moderate organic carbon accumulation.
1.05-1.2	sand:	loose sand with abundant grass rootlets.
1.2-1.6	soil:	yellowish-tan, loose, organic matter-enriched.
1.6-1.75	sand:	yellowish-brownish sandy silt, slight organic carbon concentration.
1.75-1.96	sandy silt:	yellowish-tan sandy silt, loose, slight organic carbon accumulation.
1.95-2.65	soil:	high in organic matter, compacted silt.
2.65-3.0	(?)soil:	yellowish brown, compacted.

In the Kearny County section of southwestern Kansas, there are only the surface soil and the Peoria Loess (Figure 7e). Because this site might have been drier than the other three loess sites during the periods of loess deposition or because this site has been closer to the loess source region (eolian erosion probably dominated over eolian deposition at certain times), soils have been neither developed nor preserved.

The central Great Bend section covers the history of sand dune activities during the Holocene. This sequence exhibits several soils at the interface between the silt portion (1.6–2.65 m) and the overlying sand (0–1.6 m) and within the sand, indicating fluctuations of climatic environments during the Holocene.

Time Control

Many of the spatial stratigraphic correlations are misleading without numerical age control. In order to reconstruct a temporal sequence of paleoclimatic change, time control is as, if not more, important as the climatic information itself.

Pratt County Section. Three stratigraphic units were dated at this section (Plate 1). The Brady Soil and Gilman Canyon pedocomplex were dated using the ^{14}C dating technique, and the bottom of the eolian sand was dated by using the TL dating method. The upper portion of the Brady soil was dated at 6460+/-590 yr BP (TX-6258), and the upper portion of the Gilman Canyon pedocomplex was dated at 20,550+/-590 yr BP (TX-6633). The bottom of the eolian sand (5.7 m below the surface) was TL dated at 416+/- 35 ka (XAL-124). Because of its antiquity, this sample was dated three times, and results were consistent.

Barton County Section. At this section, the upper and lower portions of the Brady Soil (Plate 2) and middle of the Gilman Canyon pedocomplex were ^{14}C dated. The following were TL dated (Plate 3): the bottom of the Sangamon pedocomplex (4.2 m below the surface), the bottom of the sand unit (i.e., the top of the Loveland Loess , 6.4-6.6 m), the first carbonate-depleted loess layer (10-10.2 m), and the bottom of second carbonate-enriched clayey loess layer (12.3-12.5 m). The upper and lower 5 cm of the Brady Soil A horizon were ^{14}C dated at 9,820+/-110 (TX-7045) and 10,550+/-150 yr BP (TX 7046), respectively. These ages are consistent with the age of the Brady equivalent in Cheyenne Bottoms, 9,690 yr BP (Fredlund, 1991), and other sites in Kansas and Nebraska (Johnson, 1991). The middle portion of the Gilman Canyon pedocomplex was ^{14}C dated at 24,360+/-570 yr BP (TX-5809), this being consistent with the age of the Gilman Canyon pedocomplex at the Phillips County section (25,500+/-820 yr BP) and other localities in the Central Great Plains (Johnson, 1991). The upper portion of the same soil was dated at 20,550 yr BP in the Pratt County section, and the equivalent to the Gilman Canyon pedocomplex at the Cheyenne Bottoms dated at 30,220 yr BP (Fredlund and McClain, 1990).

The bottom of the Sangamon pedocomplex was TL dated at 69+/-6.4 ka (XAL-077). The bottom of the Barton sand, or the top of the Loveland Loess, was dated at 92+/-7 ka (XAL-078). The TL age of the first carbonate-depleted loess layer is 193+/-22 ka (XAL-076). The boundary between the second carbonate-enriched layer and the second carbonate-depleted layer was dated at 260+/-25 ka (XAL-075). All parameters related to these TL dates are listed in Table 2.

This section reached only 15 m in depth, but data from eight adjacent drill holes indicate that the loess thickness is about 18 m at this site (Kansas Biological Survey and Kansas Geological Survey, 1987). Based on three TL dates (92 ka at 6.4-6.6 m, 193 ka at 10 m and 260 ka at 12.3-12.5 m), it is interpolated that the loess deposition rate from 92 to 260 ka was consistently low, about 0.034 mm/yr. According to the rate of deposition, it is inferred that the bottom of the 18 m thick loess should be 420 ky in age. The bottom of the Loveland eolian sand at the Pratt County section was dated at 416 ka. The inferred age

Table 2. Thermoluminescence dates and their related parameters of the loess sequences at Barton County and Pratt County sections of central Kansas

Samp No	Lab No	Depth (m)	Water (%)	U (ppm)	Th (ppm)	K (%)	Dose Rate (Gy/ka)	Plateau T (°C)	Equivalent Dose (Gy)	Age (ka)
BT5	XAL 77	4.2	7.4	3.6	12.3	2.56	5.5	350-390	379+/-37	69+/-6.4
BT4	XAL 78	6.6	12.9	3.2	9.9	1.76	4.87	320-370	449+/-28	92+/-7
BT2	XAL 76	10	16.3	3.4	13.3	2.37	5.11	300-340	986+/-115	193+/-22
BT1	XAL 75	12.5	16.9	3.3	12.8	2.33	5.33	380-410	1383+/-140	260+/-25
PR1	XAL124	5.7	12.6	2.4	6.8	1.96	3.66	320-360	1520+/-90	416+/-35

(420 ka) of the bottom of the eolian silt at the Barton County section and the TL measured age (416 ka) of the bottom of the eolian sand at the Pratt County section confirm that the loess deposition in central Kansas started about 416 ka. Based on the depositional rate (0.034 mm/yr), 327 ka was inferred for the age of the middle portion of third carbonate-enriched loess layer (15 m deep). The extrapolation of the age of the bottom of the eolian sand at the Pratt County section (416 ky) to the bottom of eolian silt at the Barton County section is also supported by a very good stratigraphic correlation between these two sections (cf. Figure 7).

In ^{14}C dating the Brady Soil both at the Barton County and Pratt County sections, discrepancies were encountered. The soil was initially dated 6460+/-590 (TX-6258) at the Pratt County section and 6560+/-130 (TX-6190) at the Barton County section. These ages of the upland soils encouraged this writer to question the applicability of the so-called warm dry mid-Holocene climatic model to the Central Great Plains. But, the same soil, freshly excavated, at the Barton County section was redated to 10,550-9,820 yr BP. The new ages are consistent with those of the Brady Soil at many other localities in the Central Great Plains (10,500-8,500 yr BP). The original ages of the Brady Soil at the Pratt County section (6460 yr BP) and Barton County section (6560 yr BP) are younger than redated age at the Barton County section (9820-10,550 yr BP) probably because of organic carbon contamination due to long exposure of the sampled sections. Therefore, this writer adopted the new ages and rejected the original ones. Because of limited time and resources, the Brady Soil at the Pratt County section was not redated. However, this writer believes that the younger ages at these two sections should be reconfirmed before they are completely rejected.

The ^{14}C ages from the Gilman Canyon pedocomplex in the Central Great Plains vary significantly from 35 to 20 ka (Johnson, 1990b). All ^{14}C ages of the Gilman Canyon pedocomplex available for the Central Great Plains were plotted against the frequency, demonstrating that there are three age clusters: 20-21 ka, 24-25 ka and 30-31 ka. Stratigraphically, the three clusters represent the top, middle, and bottom of the pedocomplex. As noted earlier, the middle of the Gilman Canyon pedocomplex was dated at approximately 24 ka at the Barton County section, and the upper portion, 20.5 ka at the Pratt County section. Considering these ages, as well as the ages of the Roxana Silt (30-40 ka) in the Central Lowlands, this writer puts the age range of the Gilman Canyon pedocomplex in central Kansas from 20 to 31 ka.

Unexpectedly, the reddish Sangamon pedocomplex, which was presumed to have developed from 127-125 to 75-70 ka, dated only 69 ka at the bottom. The Barton sand was deposited from 69-92 ka. The four cycles of carbonate enrichment occurred at 416-327, 327-260, 260-193 and 193-92 ka. Considering the accuracy level of TL dating (10%), the dates are rounded to 415-325 (for 416-327), 325-250 (for 327-260), 250-195 (for 260-193) and 195-95 (for 193-92) ka for the loess-soil complex of the Loveland Loess, 95 ka for the Barton sand, and 70 for the Sangamon pedocomplex (Figure 8). Figure 9 shows that the rate of loess deposition was quite high during the early Wisconsin time (70-95 ka) and late Wisconsin time (10-20 ka). The rate was low during the Loveland time (95-415 ka) and very low during the middle Wsconsin time (30-70 ka).

Central Great Bend Section. In the central Great Bend Sand Prairies, central Kansas, 2.5-m-thick dune sand overlies a grayish silt unit. The upper portion of the silt unit is a soil, and ages of the soil vary from locality to locality. For example, it is dated at 13,670 yr BP in the northwestern corner of Reno County (active sand dunes there); 810 yr BP in central Stafford County (close to the Rattlesnake Creek, where modern sand dunes are active); 1620 yr BP in northernmost corner of Edwards County (close to the Arkansas River where modern sand

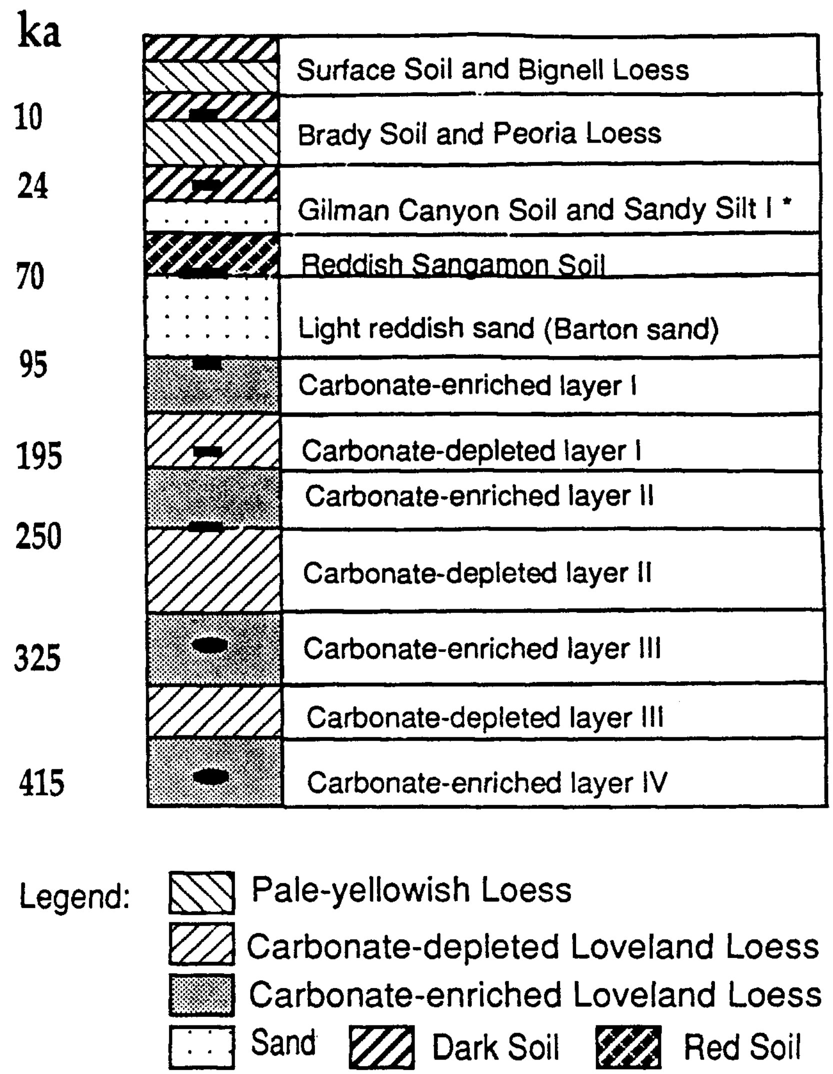


Figure 8. A comprehensive loess chronostratigraphy in central Kansas, based on Barton County landfill section except for the Sandy silt I (*) which was observed at Phillips County landfill section (325 ka is inferred, 415 ka is transplanted from the Pratt County section).

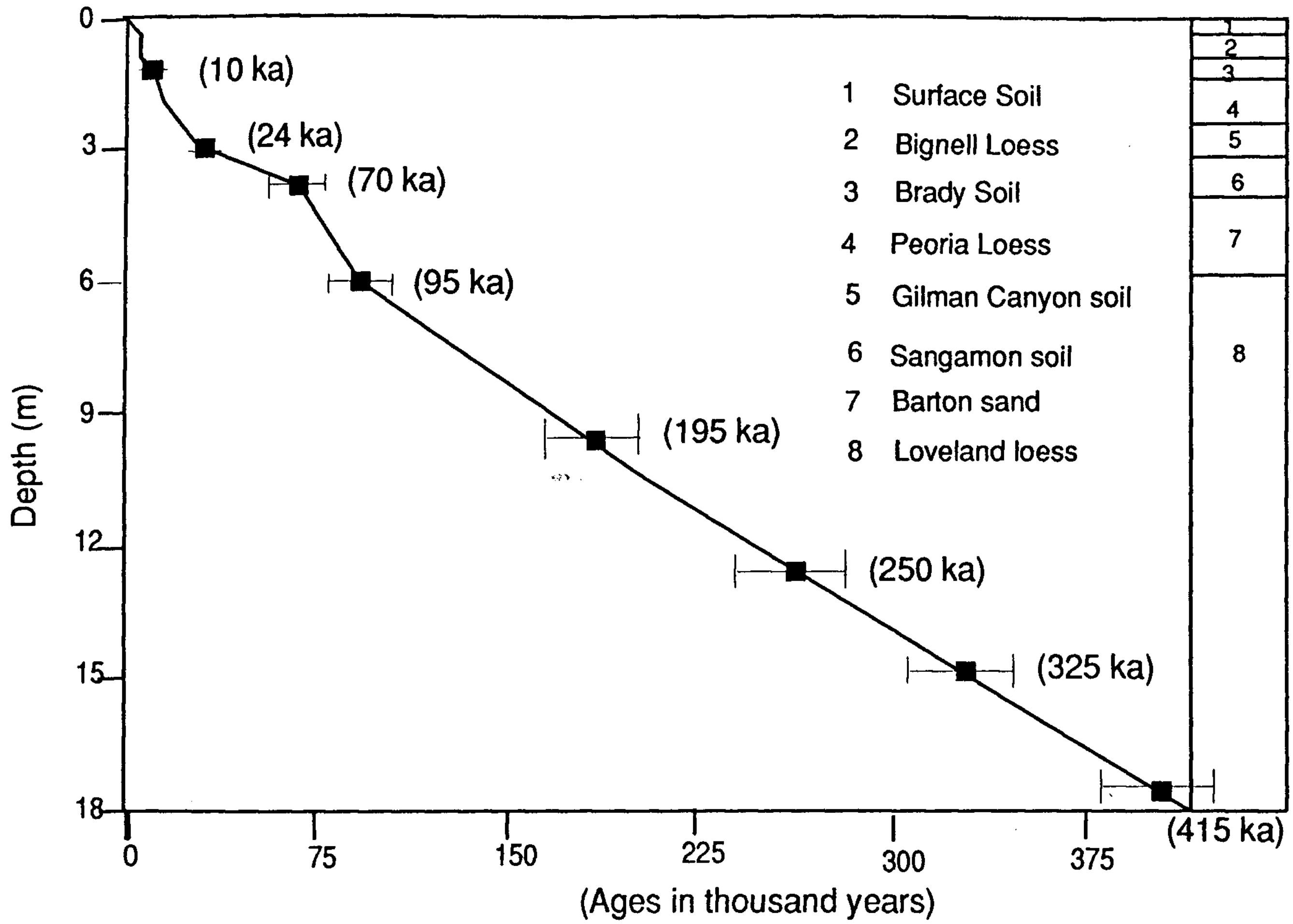


Figure 9. The relationship between the depth (m) and the ages at the Barton County section.

dunes are active). The reasonable interpretation for these age inconsistencies of the same stratigraphic unit may be that the soil ages represent the times when the silt unit was exposed to the surface. Soil was formed when the surface was stable for a relatively long time.

Understandably, when the silt was exposed to the surface and soil was formed in the silt at some localities, some other localities were buried by stabilized sand dunes. If these dunes were subsequently reactivated, the soil developed in the sand, if any, was eroded; consequently, reactivated sand dunes buried the soil developed in the surface of the silt. The ages of soils represent periods of land-surface stabillity.

Based on five, well-dated, sand-silt sequences in the Great Bend Sand Prairie, central Kansas, the geomorphic sequence appears to be as follows: after the 13,670 yr BP soil and the Brady soil (upland soil) dated at 10,500-8500 yr BP, sand dunes were dominant until about 5870 yr BP. From around 5870 to 4100 yr BP there was more or less continuous soil formation. Sand-dune activity persisted from 4100 until today with four major interruptions around 2940, 1620, 800 yr BP, and the time when the surface soil developed. The age of the surface soil is uncertain, and the surface soil has been covered in most areas by blow sand (Figure 10). The surface blow sand suggests a

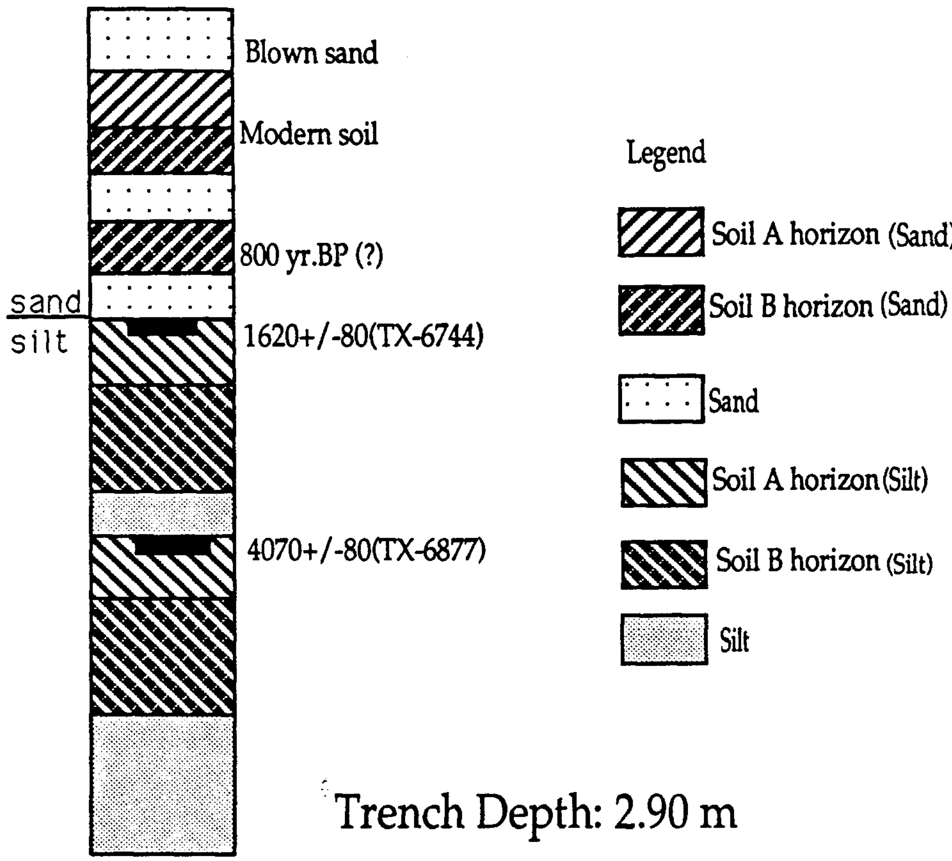


Figure 10. The Holocene sand-soil sequence at the central Great Bend Sand Prairie section (at the site of recharge well 7, Groundwater Management District 5, northernmost Pratt County). The 800 yr soil is inferred based on the soil dated at 810+/-60 at the site of recharge well 2. Another soil found at the site of recharge well 10 was dated at 2940+/-80 (TX-6745).

degeneration of the environment. It is not known whether the degeneration has been caused by climatic deterioration or by a negative human impact (e.g., destroying or modifying vegetational cover).

Chemical and Physical Properties and Their climatic Implications

As discussed earlier, chemical properties of loess can be proxies of the degree of soil development and loess weathering. The physical properties can also provide some insight to the intensity of soil development and loess weathering. Magnetic susceptibility, the kaolinite/quartz ratio in the clay fraction, and the quartz/feldspar ratio in the fine sand fraction can demonstrate the degree of pedogenesis in the loess

Phillips County Section

Because this site is located on a flat upland, topographic and hydrologic influences can be excluded when considering variation in the loess depositional environment. As discussed earlier, the particle-size spectrum of loess may function as an indicator of wind intensity during loess deposition. When environmental conditions are favorable for loess deposition (dry and windy), the dustfall (loess) tends to be coarse, and when loess deposition was slowed under relatively moist and less windy conditions, the land surface becomes sufficiently stable for soil to develop and the dustfall tends to be fine (Liu et al., 1985).

Figure 11 shows that the surface soil, the Brady soil, the Gilman Canyon pedocomplex, and the Sangamon pedocomplex are characterized by low sand content, and high silt and clay content (Table 3). The two sandy silt units both underlying and overlying the Sangamon pedocomplex stand out prominently. The sandy silt I overlying the Sangamon pedocomplex is unique within the study area, although the silt unit has apparent stratigraphic counterparts in southern Nebraska (Ahlbrandt and Fryberger, 1980) and Illinois and Iowa (Ruhe, 1965), suggesting that the environment was dry and windy during the transitional period from Sangamon time to Gilman Canyon time. The Gilman Canyon pedocomplex is high in silt content (about 89%), but not in clay (only 7.2%), and its sand content is only 4% (Table 3). This particle size differentiation demonstrates that clay transformation during the formation of Gilman Canyon pedocomplex was not dominant when the soil developed (Figure 11, Table 3)

The Peoria Loess is characterized by a low clay content and high sand content. The carbonate content in this section is constantly low from the top to 6.5 m deep (bottom of the Gilman Canyon pedocomplex), increases a little bit from the Sandy Silt I to the Sandy Silt II,

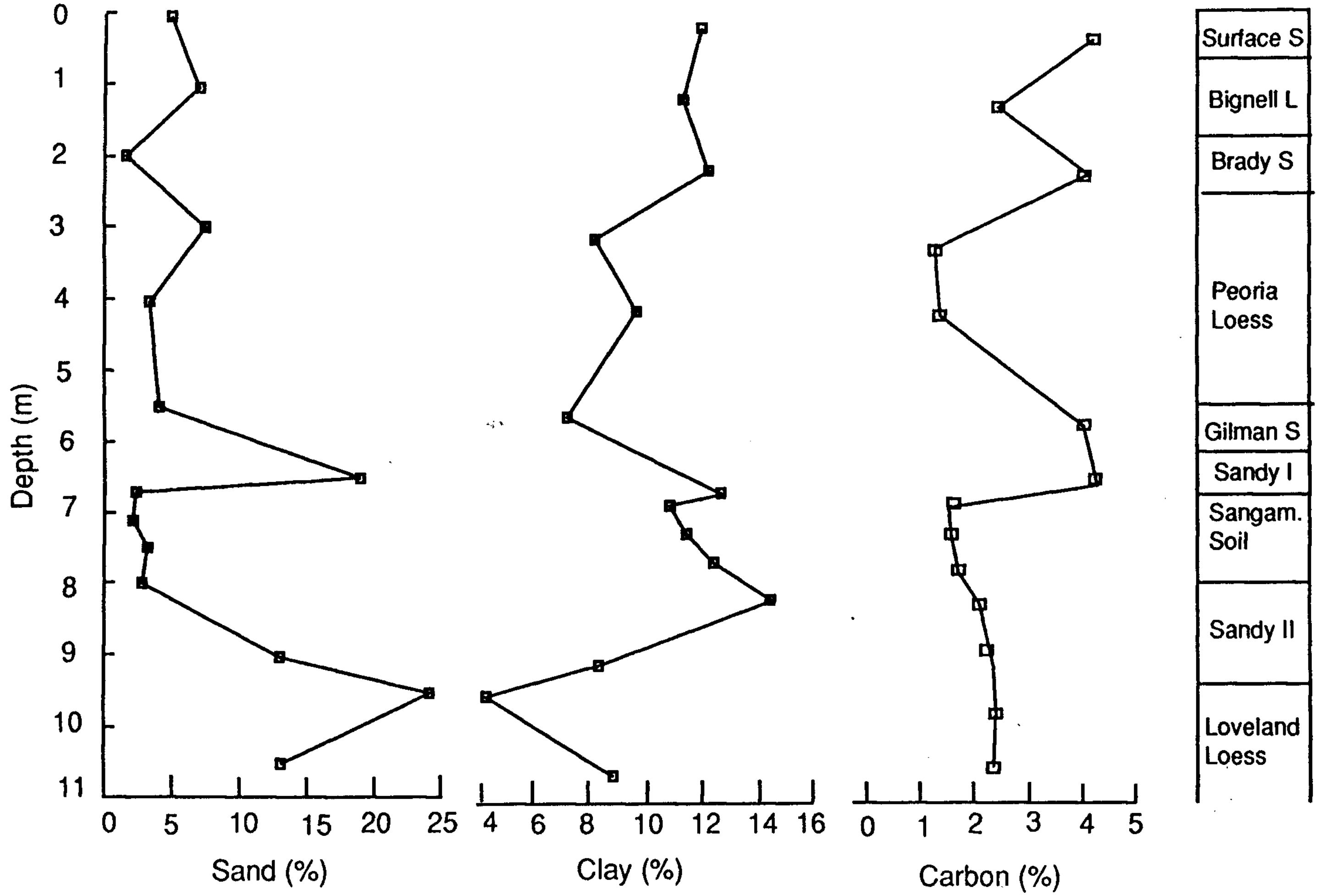


Figure 11. The distributions of sand, clay and organic carbon at the Phillips County section.

Table 3. Average physical and chemical properties of the loess sequence at the Phillips County section of north-central Kansas, based on the stratigraphic units. (L.I: leaching index= {[Fe2O3+Al2O3]/[CaO+P2O5+K2O+Na2O]}; O.I: oxidization index= {Fe2O3/Al2O3}).

Stratigraphy	Depth (m)	Sand (%)	Silt (%)	Clay (%)	CaCO3 (%)	Carbon (%)	L.I	O.I	Fe2O3 (%)	Al2O3 (%)
Surface Soil	0.5	4.0	83.6	12.4	2.0	3.1	1.15	0.27	2.8	10.5
Bignell Loess	1.4	7.0	81.8	11.2	2.0	2.1	1.22	0.25	2.8	10.9
Brady Soil	2.0	1.5	86.7	12.1	2.3	3.6	1.28	0.26	2.8	10.8
Peoria Loess	4.4	7.5	84.4	8.1	2.3	2.0	1.32	0.25	3.5	13.6
Gilman Soil	5.5	4.0	88.8	7.2	2.8	3.1	1.29	0.25	2.8	11.5
Sandy I	6.5	19.1	68.3	12.8	4.5	3.0	0.92	0.27	2.6	9.7
Sangamon Soil	8.0	2.8	82.7	14.5	7.0	1.2	0.85	0.31	3.2	10.5
Sandy II	9.6	24.1	74.4	4.2	18.0	1.1	0.62	0.30	2.7	9.0
Loveland Loess	14.0	10.0	78.5	11.5	20.1	1.0	0.55	0.25	2.8	11.0

i.e., from 6.5 to 9.6 m (5-10%), and increases appreciably in the Loveland Loess, from 9.6 to 14 m (about 20%). Through sequentially comparing the leaching index (L.I) and the oxidization index (O.I), it can be seen from Table 3 that the oxidization is relatively high in the Sangamon pedocomplex (O.I=0.31) and in the Sandy Silt II (O.I=0.30). The leaching index is high for the upper portion (0-5.5) and low in the lower portion (5.5-14 m). This difference of leaching index has only geological implications, rather than climatic meanings (discussed later in this chapter).

Organic carbon content is the best indicator of soil development in the grasslands of the midcontinental United States, i.e., effective carbon accumulation, inefficient carbon decomposition and weak oxidization. Organic carbon content is high in the surface soil (3.1%), the Brady Soil (3.6%) and the Gilman Canyon pedocomplex (3.1%). The carbon-enriched Sandy Silt I (carbon content 3.0%) underlying the Gilman Canyon pedocomplex was either influenced by the soil-forming processes of the Gilman Canyon pedocomplex or was the lower portion of the Pedocomplex.

Table 3 also shows that Fe_2O_3 and Al_2O_3 levels are slightly higher in the Peoria Loess than in any other stratigraphic unit. These high levels of Fe_2O_3 and Al_2O_3

can be attributed to the geological source difference rather than difference in the weathering intensity. It is likely that these highs in Fe_2O_3 and Al_2O_3 resulted from the contribution of less weathered igneous rocks to the loess through fluvial and eolian processes.

Kearny County Section

After discussing the loess section in north-central Kansas (Phillips County), the focus now is shifted to southwestern Kansas (Kearny County). As with the Phillips County section, the particle size spectrum functions as a good indicator of the loess depositional environment in the Kearny County section. The surface soil, 0.78 m deep, is characterized by relatively high silt and clay content (Figure 12 and Table 4). The Peoria Loess has two clay peaks (18% at 1.6 m and 14% at 3.2 m deep), one silt peak (84% at 4 m), and one sand peak (11% at 2 m). Neither the Brady Soil nor the Bignell Loess was identified. Table 4 shows that leaching index (L.I) is low in the surface soil (0.75-0.89), and relatively high in the Peoria Loess (0.94-1.21). Again, this writer believes that the leaching index was geologically generated, rather than climatically generated. As with the leaching index, the oxidization index fails to indicate the soil-forming processes. The reason for the failure of the oxidization index to indicate the intensity of weathering is that the

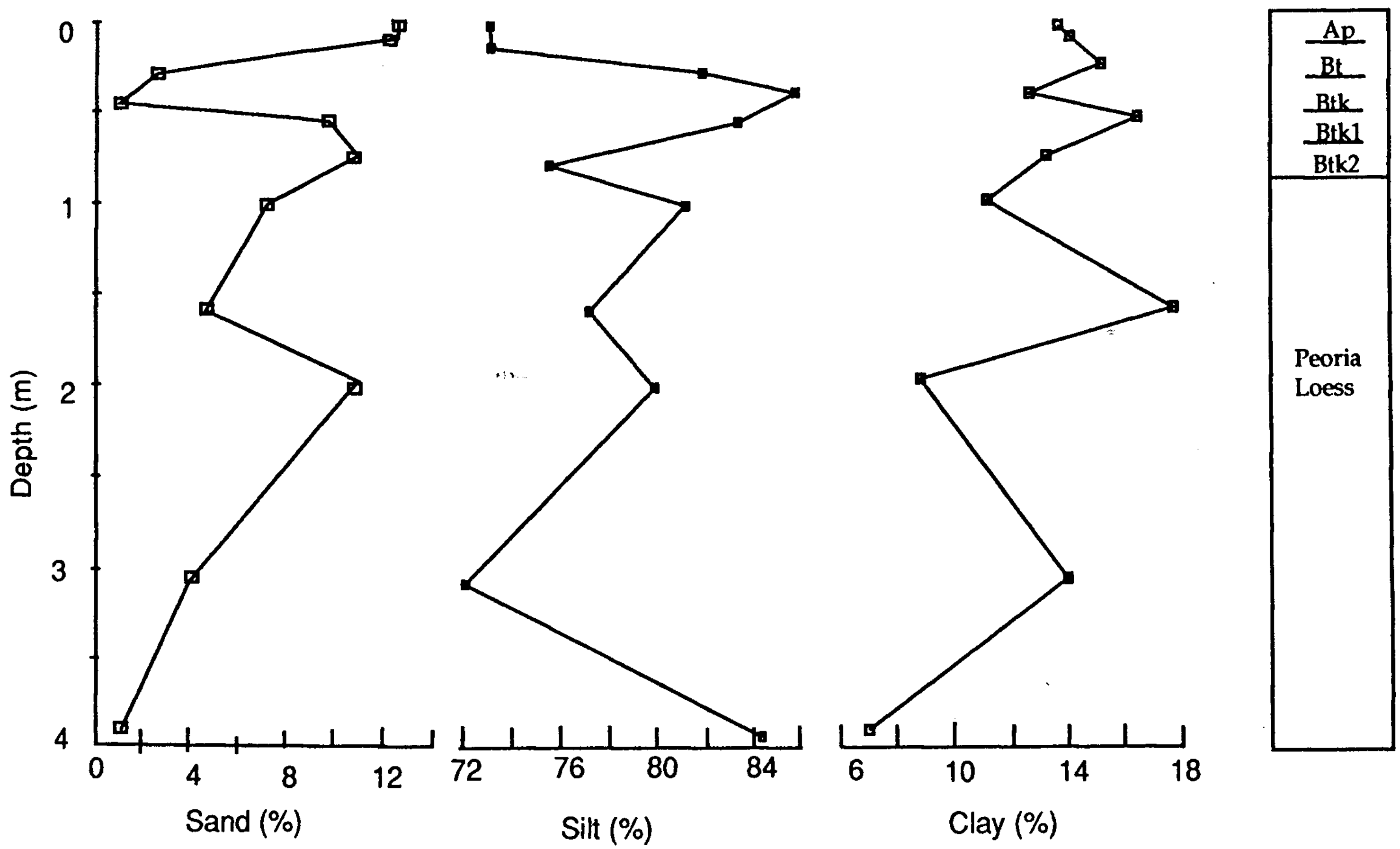


Figure 12. The spectra of particle size distribution at the Kearny County loess section.

Table 4. Average physical and chemical properties of the loess sequence at the Kearny County section of southwestern Kansas, based on stratigraphic units (Q/F: the ratio of Quartz/Feldspar in fine-sand fraction).

Strata	Horizn	depth (m)	Sand (%)	Silt (%)	Clay (%)	Carbon (%)	L.I	O.I	Q/F	Fe2O3 (%)	Al2O3 (%)
Soil	Ap	0.13	12.8	73.2	14.0	2.8	0.89	0.27	7	2.8	10.5
	Bt	0.28	2.8	82.0	15.2	2.7	0.89	0.28	3	3.1	10.9
	Btk	0.43	1.2	86.0	12.8	1.9	0.81	0.28	3	3.1	10.9
	Btk1	0.58	1.0	83.4	16.6	2.0	0.75	0.27	4	2.9	10.6
	Btk2	0.78	11.2	75.6	13.2	1.6	0.85	0.28	6	2.9	10.8
Peoria Loess		1.02	7.4	81.4	11.2		0.83	0.29	3	3.2	11.2
		1.6	4.7	77.3	18.0		1.07	0.29	4	3.4	11.9
		2.0	11.2	80.0	8.8		1.21	0.30	3	3.8	12.7
		3.2	4.0	84.3	7.0						
		3.95	8.7	84.3	7.0		0.97	0.26		3.7	12.1
		5.49	9.6	80.9	9.5		0.94	0.25		3.1	12.4

geological source likely dominated over the soil-forming processes. Table 4 shows that Fe_2O_3 and Al_2O_3 content, especially Al_2O_3, is slightly higher in the Peoria Loess than in the surface soil. This unexpected high level of Al_2O_3 in the Peoria Loess is consistent with data from the Phillips County, Pratt County and Barton County sections.

The clay content does reflect differentiation of soil horizons in the surface soil (cf. Figure 12). But, high clay content in the upper Peoria Loess indicates that a major change in clay content might be controlled by source material and by wind direction during the time of loess deposition, rather than by soil-forming processes. The interpretation is that winds from different directions might have contributed a variable amount of clay due to the availability of clay in the source areas. The Fe_2O_3 /Al_2O_3 ratio and the levels of Fe_2O_3 and Al_2O_3 are relatively high in the Bt and Btk horizons. It appears that the chemical properties work well for interpreting soil-forming processes and the intensity of loess weathering, only if the difference of geological source does not overshadow the soil-forming processes.

As discussed in Chapter III (Methodology), application of chemical parameters in reconstructing weathering history has been successfully applied to loess deposits in other regions. But, Ruhe's precaution (1974) about using chemical parameters needs to be seriously considered. After working on thin loess sequences overlying tills in Iowa, Ruhe (1974) suggested that mineralogical index, the quartz/feldspar ratio in the fine sand fraction, is more reliable than chemical indices. This study shows that the quartz/feldspar ratio in fine-sand fraction is a crude indicator of soils in loess, but, the ratio is not sufficiently precise to diferentiate soil horizons. For example, it is difficult to explain why the vey top horizon has the highest quartz/feldspar ratio (the ratio = 7) and why the Bt and Btk horizons have relatively low values of the ratio (the ratio = 3), as low as that of the Peoria Loess (cf. Table 4).

Chemical parameters are more precise than the mineralogical parameters in interpreting soil horizonation within loess, if the geological source does not overshadow the soil-forming processes. The reason for the failure of the mineralogical weathering index in indicating soil-forming processes could be that weathering agents (either physical or chemical) work mainly on the silt fraction in the dry and rapid deposition environment where loess occurrs. In this situation, soil is mainly cumulative, i.e., soil develops as loess continues to accumulate, but at very slow rate. Under this environment, weathering is not strong enough or does not last long enough to completely alter chemical imprints of the environment in the silt and to produce a detectable impact on the fine-sand fraction of loess. Under strong, extended weathering conditions, all chemical parameters might be easily altered and the climatic information in the chemical parameters could be readily deleted. In this case, fine sand could function as the best recorder of weathering history.

Pratt County Section

Table 5 shows that the silt fraction is dominant from the top to 3.4 m, while eolian sand dominates from 3.4 to 6 m (below 6 m is a fluvial sand unit). Table 6 shows that there are correlations among sand, organic carbon, clay content, and magnetic susceptibility. Understandably, the sand and the organic carbon are negatively correlated, r^2= 0.82. The clay is positively correlated with the organic carbon, r^2=0.81. Not surprisingly, as an indicator of soil development, the magnetic susceptibility (S.I) is positively correlated with organic carbon content (r^2 = 0.64). The Sangamon pedocomplex is an exception: it has a very high magnetic susceptibility but low organic carbon due to strong oxidization (Figure 13). Similarly, clay and magnetic susceptibility have a relatively strong relationship (r^2 = 0.51). Above the Sangamon pedocomplex (0-3.4 m), the sand, clay, carbon, and susceptibility

Table 5. Average physical and chemical properties of the loess at the Pratt County section of central Kansas, based on the stratigraphic units (Ka/Qu: the ratio of kaolinite/quartz in clay fraction. Loveland C1 means the first carbonate-enriched layer; Loveland L1 means the first carbonate-depleted layer, etc).

Stratigraphy	Depth (m)	Sand (%)	Silt (%)	Clay (%)	Suscept. (S.I)	Carbon (%)	CaCO3 (%)	Ka/Qu	Fe2O3 (%)	Al2O3 (%)	L.I	O.I.
Surface Soil	0.2	27.8	67.5	9.8	5.50	4.4	2.9	0.38	2.6	10.3	1.64	0.25
Bignell Loess	0.6	34.3	58.5	7.4	5.72	3.8	2.3	0.40	2.3	9.4	1.74	0.24
Brady Soil	1.2	23.7	62.1	14.2	5.95	3.9	3.1	0.43	2.7	10.8	1.48	0.25
Peoria Loess	2.4	59.5	35.0	6.2	5.32	2.8	1.9	0.37	3.3	11.3	1.56	0.29
Gilman Soil	3.0	40.9	54.5	4.9	5.80	3.3	2.2	0.27	3.1	11.5	1.92	0.29
Sangamon Soil	3.4	37.1	59.7	3.2	5.75	2.2	2.9	0.82	3.6	12.0	1.81	0.30
Loveland C1	3.8	45.2	54.2	3.1	4.52	1.5	5.0	0.73	3.7	12.1	1.66	0.31
Loveland L1	4.1	69.8	28.7	1.5	3.33	1.0	2.7	0.41	3.9	12.6	1.27	0.31
Loveland C2	4.4	51.7	44.6	3.5	3.56	1.7	8.2	0.40	3.5	11.4	0.98	0.31
Loveland L2	4.6	77.4	21.5	1.1	2.50	1.0	2.0	0.39	3.7	12.4	1.42	0.31
Loveland C3	4.8	78.4	19.8	1.4	0.65	1.2	5.7	0.40	4.3	12.9	1.47	0.33
Loveland L3	5.0	80.6	18.0	1.4	0.60	1.0	1.8	0.40	3.4	11.6	1.53	0.29
Loveland C4	5.7	66.4	30.0	0.4	0.75	1.4	7.9	0.40	4.0	12.0	1.38	0.34

Table 6. Correlations of the physical and chemical parameters of the loess sequence at the Pratt County section of central Kansas (The linear regressional relationship is: $Y = A + BX$).

Y item	X item	A	B	r^2
carbon	susecptibility	0.466	0.466	0.64
clay	susceptibility	2.160	0.377	0.51
clay	carbon	1.033	0.272	0.81
sand	carbon	5.153	-0.055	0.82
susceptibility	leaching index	1.312	0.056	0.24
susceptibility	kaolinite/quartz	0.341	0.019	0.49

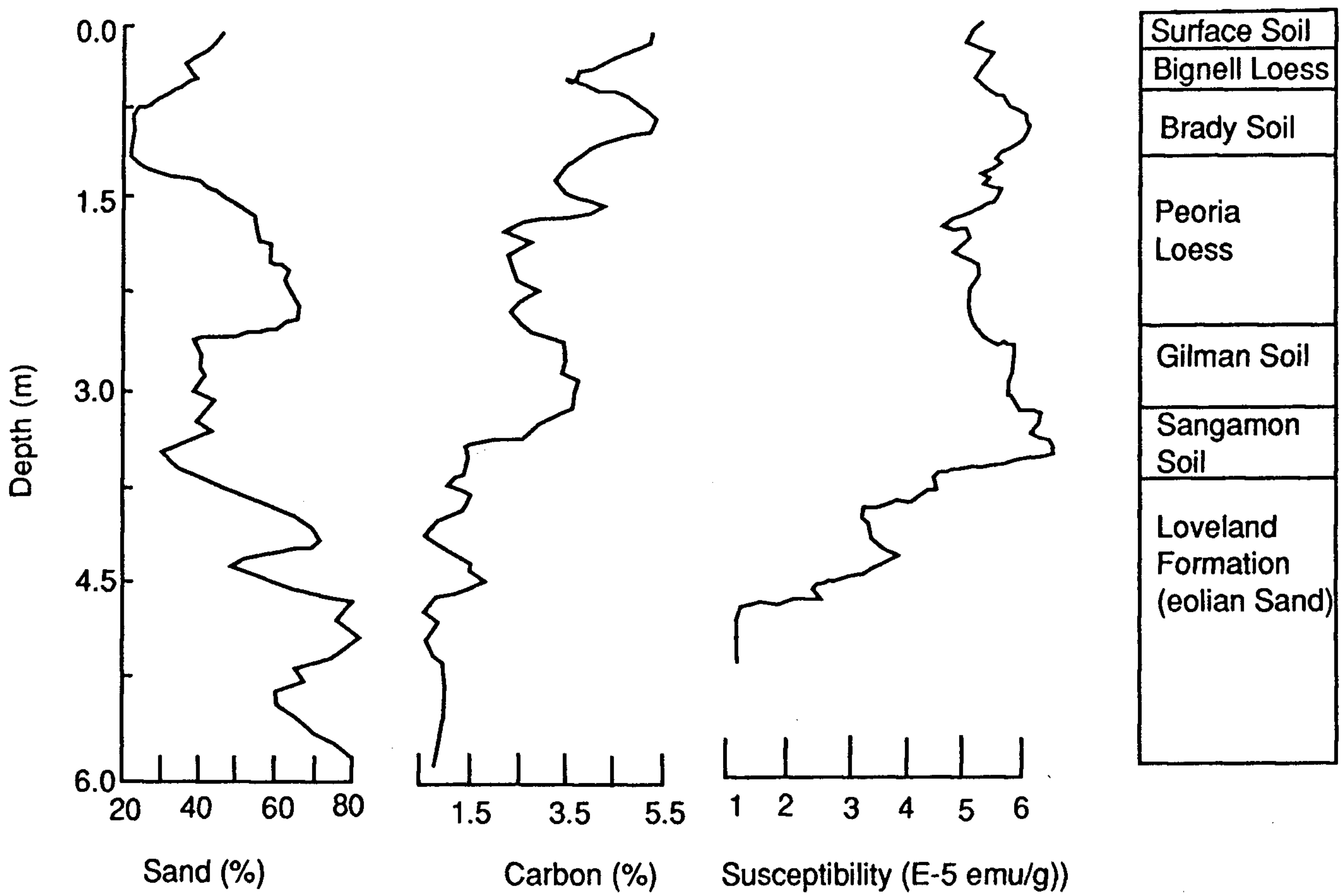


Figure 13. The comparison of the sand percentage, organic carbon content and magnetic susceptibility at the Pratt County section.

express the soil-forming or loess weathering sequences very well, i.e., the environmental history are well represented by these parameters.

The surface soil is characterized by low sand, high carbon, high clay, and high susceptibility. The Bignell Loess, although rather thin, is expressed by low clay, low carbon, and lowered susceptibility in comparison with the surface soil and the Brady Soil. The Brady Soil is typified by a very strong peak in the susceptibility, a very high peak of clay, and a high percentage of organic carbon. The Peoria Loess has low organic carbon and susceptibility values and a high percentage of sand. The clay content is low in the lower part and relatively high in the upper part of the Peoria Loess. The high clay percentage in the upper part may suggest that loess source was different during later Peoria time than during earlier Peoria time. It is unlikely that the clay was preweathered since massive, clay-depleted loess deposition occurred immediately before the clayey loess was deposited. The magnetic susceptibility of Gilman Canyon pedocomplex is as high as that of Brady Soil. It should be pointed out that the Gilman Canyon pedocomplex is characterized by an increase in silt rather than clay content. The Sangamon pedocomplex (3.0-3.4 m) is high in silt and low in sand content, and characterized by the highest susceptibility and the highest kaolinite/quartz ratio in clay throughout the whole section (Figure 14).

Because sand is so dominant, carbonate is the only parameter that can express the environmental sequence below 3.4 m (Loveland Formation). From 3.4 to 6 m there are four carbonate cycles: 3.8 m (5.0%), 4.4 m (8.2%), 4.8 m (5.7%) and 5.7 m(7.9%). The reddish Sangamon pedocomplex (3.0-3.4 m) and the four cycles of carbonate have been established as stratigraphic markers to correlate the Pratt County section with the Barton County section. The kaolinte/quartz ratio in clay and the magnetic susceptibility reflect soil-forming and loess weathering processes quite well (cf. Figure 14). Both the Peoria Loess and the Gilman Canyon pedocomplex are low in the kaolinite/quartz ratio of clay, indicating a low intensity of chemical weathering (Birkeland, 1968). Both the magnetic susceptibility and the kaolinite/quartz ratio of the clay fraction point to the fact that the Sangamon pedocomplex (3.0-3.4 m) was strongly weathered, mainly chemically. The kaolinte/quartz ratio and the susceptibility suggest that the overall intensity of weathering of the Loveland Formation (eolian sand) is rather low.

The chemical parameters do not record the environmental history of eolian deposition, if

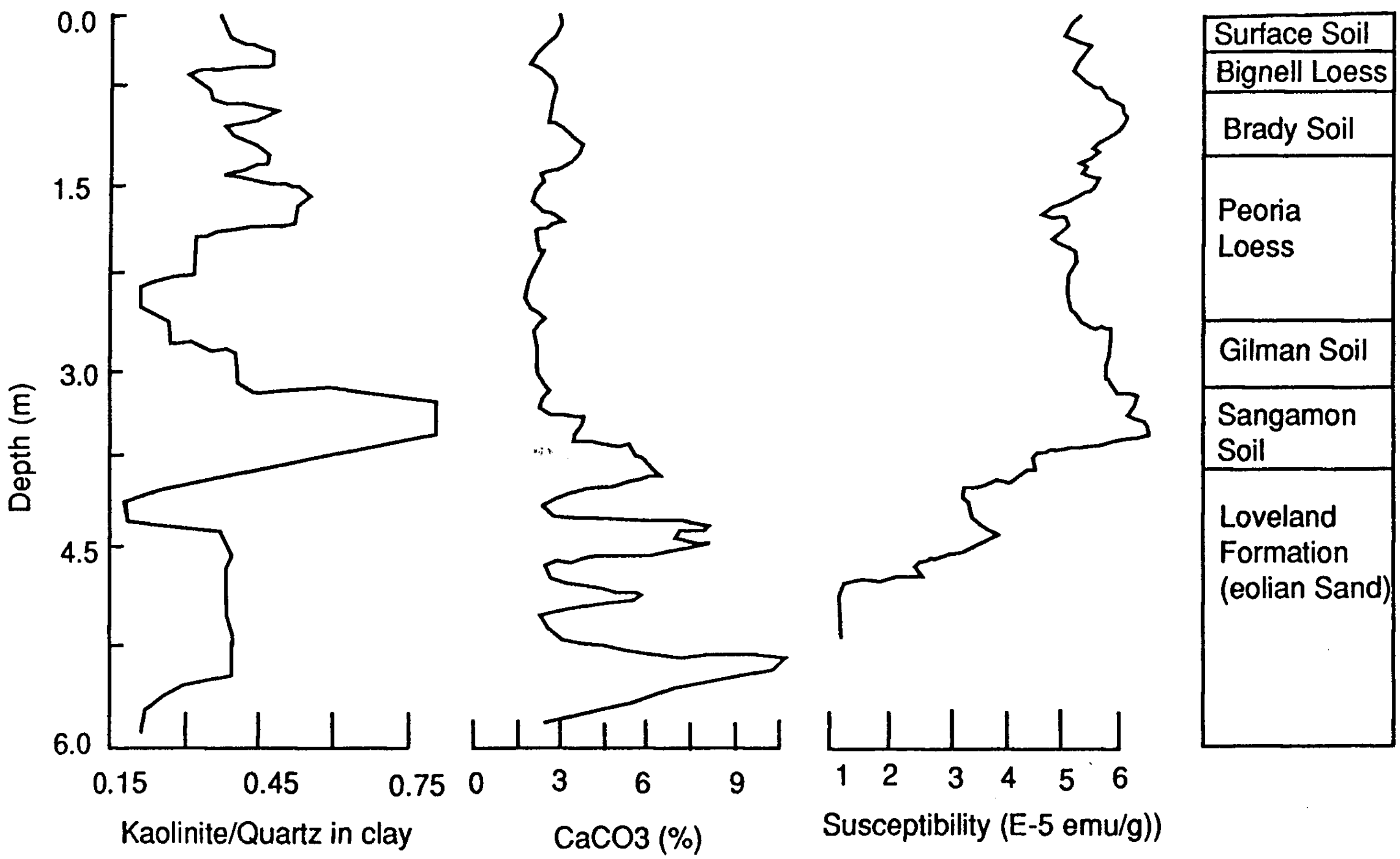


Figure 14. The comparison of the ratio of kaolinite/quartz in clay with the content of CaCO3 and magnetic susceptibility at the Pratt County section.

geologically controlled unweathered particles dominate. This is the case in the Pratt County section. When wind-blown sand dominated over the weathering processes, the chemistry of the fine particles (mainly silt) did not have a chance to be altered to indicate the weathering history. Chemistry of the fine fraction is inherited from larger particles, rather than altered from the particles. The Fe_2O_3 level is quite high in the lower portion of this section, but the magnetic susceptibility is rather low. This inconsistency between the susceptibility and the Fe_2O_3 implies that the latter is not contributing to the magnetic susceptibility.

Quaternary magnetic studies by Thompson and Oldfield (1986) showed that, in less-weathered deposits, the remnant magnetic intensity is quite high if the concentration of fine magnetic particles is high. They also indicated that magnetic susceptibility is more dependent upon coarse magnetic particles than fine magnetic particles. For example, a magnetite particles with a size of 1-2 μm have 1140 emu/cm^3 of magnetic susceptibility; while a magnetite particles with a size of 75-150 μm have 2200 emu/cm^3 of susceptibility (Carmichael, 1989). Studies also show that the remnant magnetic intensity (RMI) is consistent with the magnetic susceptibility (S.I) in most cases, but not all. For example, in comparison with other minerals, titanomagnetite has a relatively high RMI and a relatively low S.I; while hematite has a relatively high S.I and a low RMI (Carmichael, 1989). Kukla (1987) suggested that magnetic susceptibility of the soils developed in loess could be enriched by leaching processes or chemical weathering processes. Because of differences in particle size and mechanisms influencing the remnant magnetic intensity and the magnetic susceptibility, the chemical parameters, particularly Fe_2O_3 and the oxidization index (Fe_2O_3/Al_2O_3), do not indicate the depositional environment of the Loveland Formation (eolian sand) at the Pratt County section, i.e., the chemistry was controlled by the original magnetic properties of the particles, not by weathering processes.

The top portion of this section (Brady Soil, Bignell Loess, and modern soil) is characterized by apparent low values in Fe_2O_3 (2.3-2.7%), Al_2O_3 (9.4-10.8%), and the Fe_2O_3/Al_2O_3 ratio (0.24-0.25). In contrast, all other stratigraphic units have high values in Fe_2O_3 (2.9% or more), Al_2O_3 (11% or more), and the Fe_2O_3/Al_2O_3 ratio (0.29 or more). These differences in chemistry between the upper portion (0-1.2 m) and the lower portion (1.2 m-6 m) may be attributable to differences in geology of source area. For example, the sedimentary rock types of the

Southern Great Plains could have contributed to the loess having less Fe_2O_3 and Al_2O_3; while the eastern piedmont of the Rocky Mountains, which is dominated by igneous rocks, might have contributed more Fe_2O_3 and Al_2O_3 to the Peoria Loess.

Barton County Section

Figure 15 shows that clay content is high in the upper portion of the Peoria Loess, in the Brady Soil (1-2.5 m), and in the bottom portion of this section (13.5-18 m). Sand content is anomalously high in the middle of the section (4.5-7.5 m). The high clay content in the upper portion of the Peoria Loess might not be attributable to soil-forming processes; rather, it might be associated with the geological source, i.e., a clayey source might have made more of a contribution during late Peoria time.

Figure 15 and Table 7 indicate that there are four major peaks of carbonate from 6.4 to 18 m in this section. It should be pointed out that the carbonate content from 15 to 18 m is established based upon the particle-size distribution from 15 to 18 m (data from Kansas Biological Survey and Kansas Geological Survey, 1987) and the linear correlation between the carbonate content and particle size from 6.6-15 m deep at this section. The generation of the carbonate peak (16-17 m) is also supported by the existence of the fourth carbonate peak dated at 415 ka in

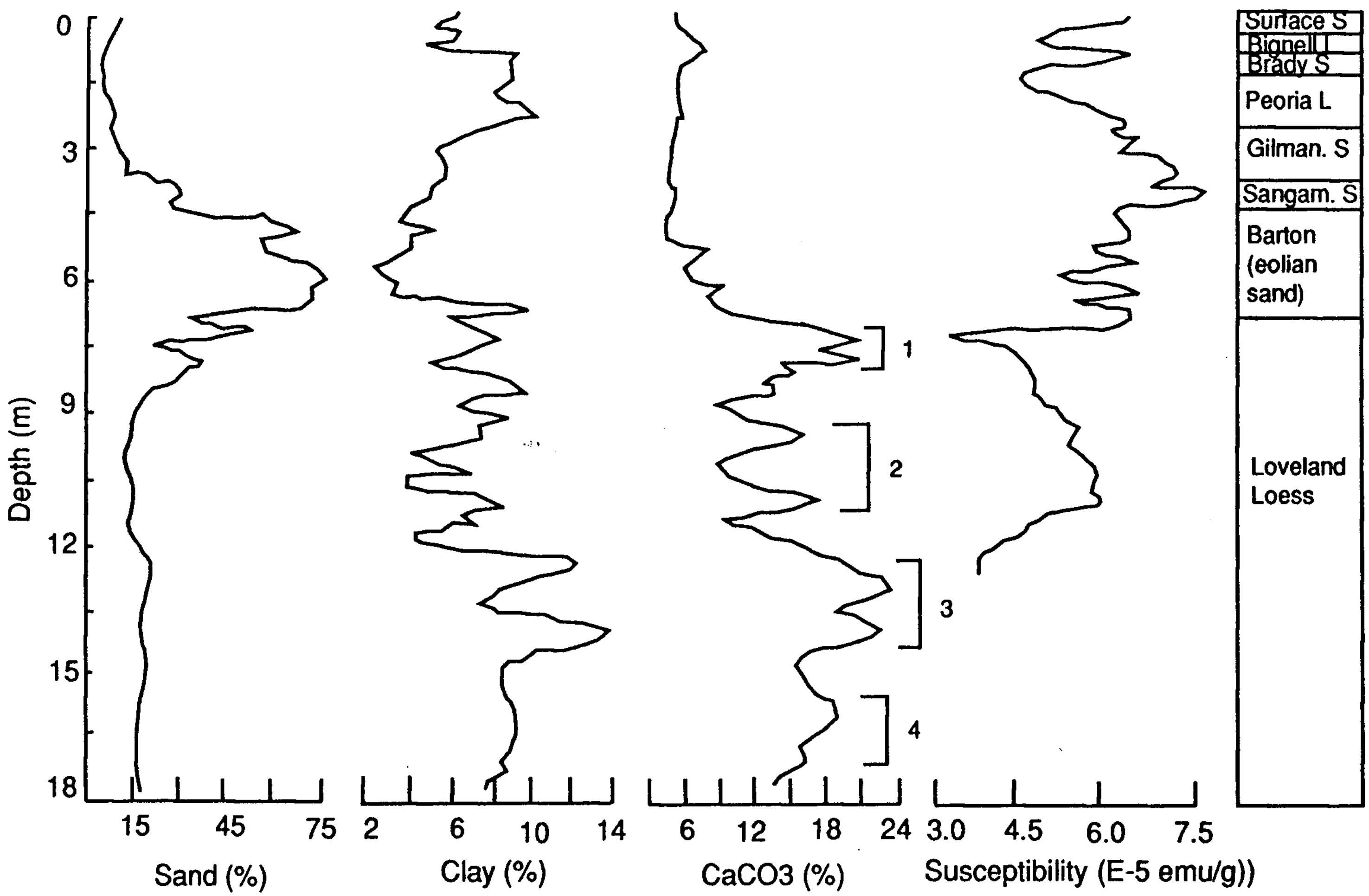


Figure 15. The comparison of the sand, clay, carbonate and magnetic susceptibility at the Barton County section.

Table 7. Average physical and chemical properties of the loess sequence at the Barton County section of central Kansas, based on the stratigraphic units (Al = Al2O3; Fe = Fe2O3)

Stratigraphy	depth (m)	Age (ka)	Sand (%)	Clay (%)	Suscept. (S.I)	CaCO3 (%)	L.I.	Quartz (%)	Kaolin (%)	Ka/Qu	Al (%)	Fe (%)	O.I
Surface Soil	0.2		2.0	10	6.8	2.1	1.64	11.8	4.6	0.38	10.2	2.2	0.22
Bignell Loess	0.8		6.6	8.3	5.6	3.1	1.57	14.0	5.6	0.43	9.4	1.9	0.21
Brady Soil	1.6	10	4.4	16.5	6.6	3.0	1.57	22.0	9.4	0.43	10.5	2.7	0.26
Peoria Loess	2.4		7.1	16.3	4.5	2.3	1.55	16.3	6.5	0.39	11.2	2.7	0.24
Gilman Soil	3.4	24	10	10.1	6.5	1.7	1.75	19.8	8.6	0.42	10.8	2.8	0.26
Sangamon Soil	4.2	70	26	8.1	7.1	1.8	1.71	13.1	6.8	0.58	9.5	2.8	0.30
Sand Unit	6.6	95	63.8	3.3	5.9	2.9	1.28	24.8	11.1	0.46	9.7	2.5	0.25
Loveland C1	8.2		30.6	11.8	4.7	15.8	0.81	7.7	2.7	0.38	8.5	2.3	0.27
Loveland L1	9.2		17.0	15.0	4.8	7.7	1.18	7.4	2.4	0.18	10.4	2.5	0.24
Loveland C2	9.8	195	17.3	9.9	5.9	10.6	1.44	8.8	2.6	0.36	10.9	2.6	0.24
Loveland L2	12.5	250	10.2	8.5	4.5	8.4	1.47	10.5	4.6	0.42	11.1	2.5	0.22
Loveland C3	14.0	325	17.7	20.4		18.4	0.77	5.6	1.1	0.19	9.2	2.3	0.25
Loveland L3	15.5		17.3	13.8		7.1	0.66				8.4	2.1	0.25
Loveland C4	18.0	415	12.0	12.5		11.6							

the Pratt County section. The first carbonate-enriched layer was formed at 95 ka, the first carbonate-depleted layer was deposited at 195 ka, and the second carbonate-enriched layer ended at 250 ka. The second carbonate peak consists of two minor peaks with a carbonate valley in between. The third and fourth carbonate concentrations occurred at 325 and 415 ka, respectively.

Figure 15 and Tables 7 and 8 show that clay and carbonate are positively correlated, especially in the lower portion of this section, i.e., the Loveland Loess from 6.4 to 18 m ($r^2 = 0.51$). The figure and tables also show that both clay and carbonate correlate negatively with the magnetic susceptibility. The coefficient between clay and susceptibility is -0.46 ($r^2 = 0.21$) for the entire section, and increases to -0.63 ($r^2 = 0.39$) for the lower portion (Loveland Loess). Similarly, the coefficient between carbonate and susceptibility is -0.58 for the entire section and increases to -0.79 ($r^2 = 0.62$) for the Loveland Loess. The high correlation between clay and carbonate in the lower portion suggests that the carbonate might have formed pedogenically. To be specific, both clay and carbonate might have been accumulated in the Btk or Bk horizons by soil-forming processes.

The correlation between clay and carbonate may also be explained by geological processes. For example, previously

Table 8. Correlations of the physical and chemical parameters of the loess sequence at the Barton County section of central Kansas (Y = A + BX).

Y item	X item	A	B	r^2	Depth Range(m)
Clay	Carbonate	1.84	0.71	0.51	6.4-18
Clay	Susceptibility	6.81	-0.11	0.21	0-18
Clay	Susceptibility	5.93	-0.09	0.39	6.4-18
Susceptibility	Leaching Index	0.54	0.16	0.39	0-18
Susceptibility	Oxidize Index	0.20	8.55	0.86	0-18
Susceptibility	Fe2O3	1.71	0.15	0.71	0-18
Susceptibibility	Carbonate	30.20	-4.1	0.62	6.4-18
Susceptibility	Kaolinite/Quartz	0.06	0.06	0.37	0-18
Susceptibility	Kaolinite	-5.2	1.95	0.45	0-18
Susceptibility	Quartz	-10.3	4.37	0.58	0-18

existing clayey layers might have functioned as impermeable layers which might stop or slow carbonate translocation. But, if this were the case, one should expect that the clayey layers should be somewhat lower than the corresponding carbonate layers stratigraphically, because the clayey layers would have prevented the carbonate from being leached downward, and the carbonate would have continued to build up above the clayey layers. However, Figure 15 shows that all clayey layers are higher than their corresponding carbonate layers in the stratigraphic positions. This lag in response of clay to carbonate could be a very good indicator of soil-forming processes, i.e., soil Btk horizons underly soil Bt horizons.

On the basis of previous literature (Aandahl, 1982; Gile et al., 1966; Jackson, 1965; Marion, 1989; Reheis, 1987), pedogenic carbonate concentration and morphology are functions of the soil-forming environment and the age of carbonate concentration. All investigators found that, either in gravel or in fine parent material (such as loess), pedogenic carbonate is formed under semiarid and arid environments, accumulated primarily in Btk horizon in semiarid environment (such as grasslands), and in Bk horizons of arid environments (such as marginal zones of deserts). For example, petrocalcic soils of the American

Southwest, where mean annual precipitation ranges from 200 to 500 mm, commonly contain 10-30% carbonate in Bk horizons developed from 30 to 100 cm deep (Marion, 1989).

Another important characteristic of the pedogenic carbonate is that there are varying sizes of clayey lenses included in carbonate nodules or carbonate layers (Gile et al., 1965). A distinguishing feature of pedogenic carbonate is that a platy carbonate horizon overlies a plugged carbonate horizon consisting of vertical carbonate cylinders or nodules, equivalent to Stage III of carbonate accumulation (Gile et al., 1966; Machette, 1985). The coexistence of abundant grass rootlets with the carbonate nodules or carbonate layers is also an important indicator of the pedogenic origin of the carbonate nodules or layers (Quade and Cerling, 1990).

There could be some uncertainty about the origin of the carbonate nodules and carbonate-enriched layers. Carbonate origin is important in interpreting the paleoenvironments of soil formation. The critical question is whether the carbonate concentration is pedogenically formed or groundwater deposited. From the following lines of evidence, the carbonate-enriched layers are clearly pedogenically formed.

First, the existence of carbonate-marked (not filled) plant-root canals and varying sizes of clayey lenses in the carbonate-enriched nodules and layers strongly suggests that the carbonate concentration was pedogenically formed (Plate 4). Second, as discussed earlier, the concurrence of clay and carbonate, as well as the lag in response of clay to carbonate in the stratigraphic positions, is another strong indication of a pedogenic origin of the carbonate concentration in the Bk or Btk soil horizons. Third, the carbonate-enriched clayey layers in the Barton County and Pratt County sections are characterized by the fact that each platy carbonate horizon overlies a plugged carbonate horizon consisting of loose carbonate cylinders or nodules, similar to the carbonate accumulation stage IV described by Gile and others (1966).

Fourth, the analysis of modern groundwater shows that Mg is mg/l and Ca is 80-100 mg/l (data provided by Donald Wittemore of Kansas Geological Survey). In the carbonate-enriched clayey loess, $CaCO_3$ is as high as 20% and MgO is only 2%, although both of them have approximately the same solubility. If the carbonate layers were a product of groundwater evaporation during Loveland time, MgO should also be high because there is no reason to assume that the groundwater chemistry of that time was dramatically different from that of today. Fifth, more than 100 feet of local relief would have prevented groundwater from

Plate 4. Structures of pedogenically originated carbonate-enriched layer. It includes carbonate nodules (CN), clay lense (CL), plant root canals (GR), and worm burrows (WB). The sample was derived from the Barton County section at depth of 7.5 m.

reaching the surface of upland and would have allowed soluble carbonate to be leached away. Local relief might have been even more during Loveland time before the adjacent Cheyenne Bottoms filled to its present level (Fredlund, 1990). Finally, the multiple layers of carbonate concentration could not be formed by evaporation of groundwater. If the bottom carbonate layer was formed first through evaporation, it is impossible for the groundwater carrying carbonate to penetrate the bottom carbonate layer to reach upper positions; if the top carbonate-enriched layer was formed first, it is unlikely that the evaporation processes would reach the lower positions to accumulate carbonate. Carbonate concentrations and their close association with identifiable soil structures within the Loveland Loess in south-central Nebraska (Schultz and Stout, 1980) and in the Southern Great Plains (Holliday, 1989b) support this argument for the pedogenic origin of carbonate.

Since the clay and carbonate, both being positively correlated, are negatively correlated with the magnetic susceptibility, what is the environmental interpretation for this negative correlation? As indicated earlier, the clay content in the upper portion of this section does not carry reliable information about soil formation: clay is unexpectedly high in the upper Peoria Loess, as well as in the Brady Soil. There is very little carbonate in the upper portion of this section (0-6.4 m), so the discussion applies only to the lower portion of this section, i.e., the Loveland Loess. This negative relationship between carbonate content and susceptibility suggests that the pedogenic carbonate-enriched clayey layers were formed during periods when the weathering was weak. These carbonate-enriched layers might be desert-like soils, which were weakly weathered. According to the chronological comparison of the loess sequence with the $\delta^{18}O$ sequence of the Pacific Ocean V28-238 core (discussed in Chapter V), it can be concluded that the carbonate concentrations occurred during warm periods. Weathering was weak during warm periods, probably due to low effective moisture.

Fe_2O_3 and magnetic susceptibility are very well correlated (Figure 16), $r^2 = 0.71$ (cf. Tables 7 and 8). This high correlation supports the interpretation that weathering was weak during warm periods of Loveland time, i.e., the Fe_2O_3 was low in the carbonate-enriched clayey layers, suggesting weaker weathering. The leaching index (L.I) also has a positive correlation with the susceptibility (r^2= 0.39). The highest leaching index occurs in the Gilman Canyon pedocomplex (L.I = 1.75), while the highest values in the oxidization index (0.30), the ratio of kaolinite/quartz of clay (0.58), magnetic susceptibility (7.1), and Fe_2O_3 content (2.8) occur in the Sangamon pedocomplex.

Figure 16 shows that the surface soil is relatively well weathered as indicated by the magnetic susceptibility (6.8) and the leaching index (1.64). The magnetic susceptibility (5.6) and the kaolinite/quartz ratio in clay (0.43) suggest that the Bignell Loess was better weathered than the Peoria Loess. In both the surface soil and the Bignell loess, the Fe_2O_3 content is rather low in comparison with other stratigraphic units below. Accordingly, the oxidization index is low. Development of the Brady Soil is nearly as strong as that of the Gilman Canyon pedocomplex. The poorly weathered Peoria Loess is characterized by lower susceptibility (4.5), kaolinite/quartz ratio in clay (0.39), oxidization index (0.24), and leaching index (1.55). The Fe_2O_3 content is, however, unexpectedly high with regard to the drop in magnetic susceptibility within the Peoria loess, i.e., the linear correlation between magnetic susceptibility and Fe_2O_3 content is nearly perfect except within the Peoria Loess. In particular, Al_2O_3 content is exceptionally high in the Peoria Loess (cf. Table 7). The high leaching index (1.75) and relatively low oxidization index (0.26) of the Gilman Canyon pedocomplex indicate that the pedocomplex

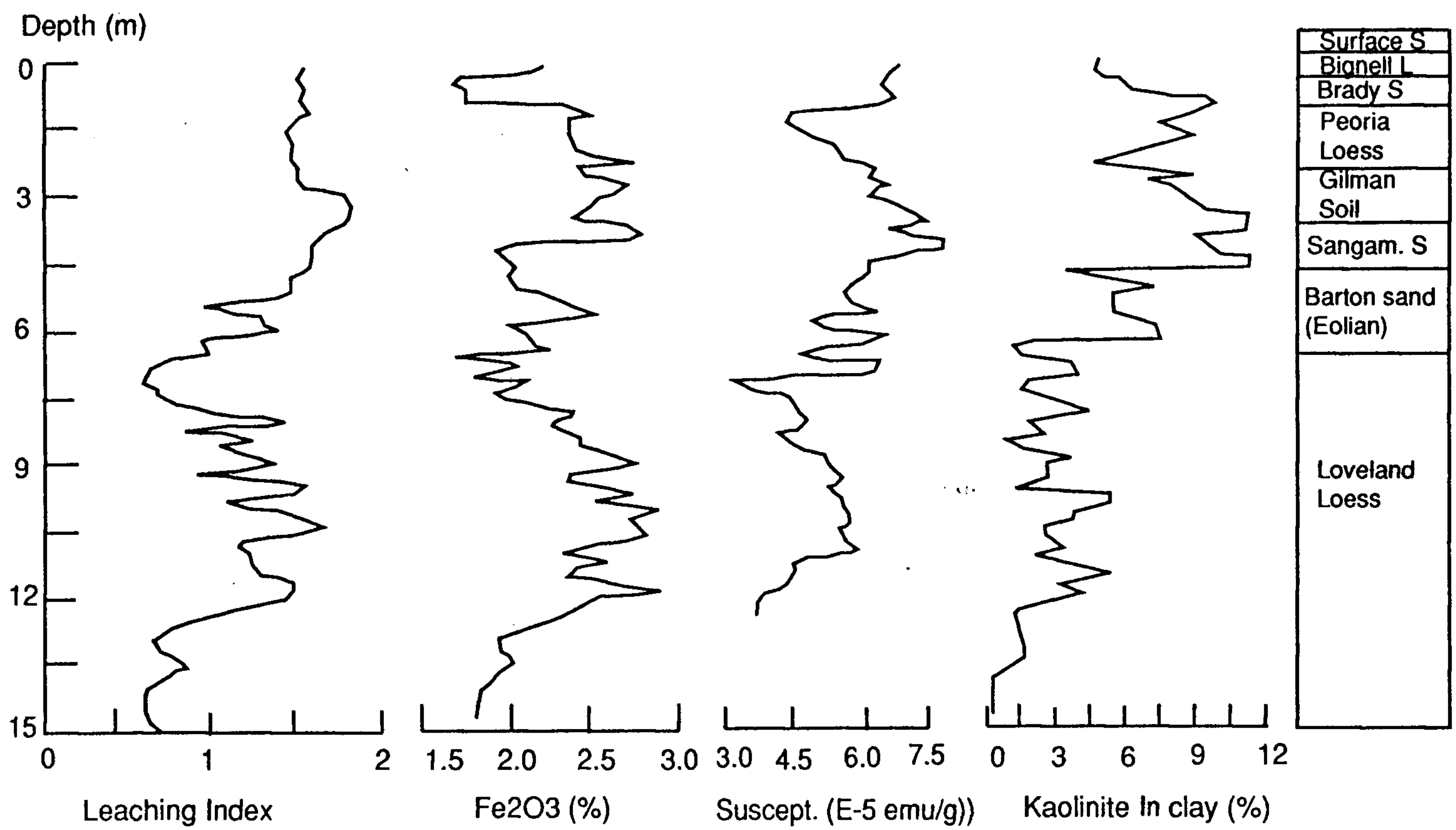


Figure 16. The comparison of the leaching index (L.I), content of Fe_2O_3, magnetic susceptibility and content of kaolinite of clay at the Barton County section.

was dominated by physical weathering rather than chemical weathering.

Kaolinite content and the kaolinite/quartz ratio in clay fraction support the argument that the weathering was the strongest in the Sangamon pedopcomplex and the weakest in the Peoria Loess, as also evidenced by Fe_2O_3 content and susceptibility. The coefficients between the susceptibility and the kaolinite in clay and between the susceptibility and the kaolinite/quartz ratio are 0.67 (r^2 = 0.45) and 0.61 (r^2 = 0.37), respectively.

The comparison between the Al_2O_3 content and the susceptibility shows that the Al_2O_3 content corresponds positively to susceptibility in the lower portion (8-15 m), but a reversed (negative) relationship between these two is apparent in the upper portion (0-8 m) of the section (Figure 17). This reversed relationship between Al_2O_3 and the magnetic susceptibility in the upper portion (r^2 = 0.85) can be attributed to the difference in geological sources at different times of loess deposition. The susceptibility and original curve of the Fe_2O_3/Al_2O_3 ratio are quite comparable, except that the magnitudes of change are different: the background of the Fe_2O_3/Al_2O_3 curve does not correspond to the susceptibility very well, but the change of the Fe_2O_3/Al_2O_3 ratio responds sensitively to the change in the susceptibility.

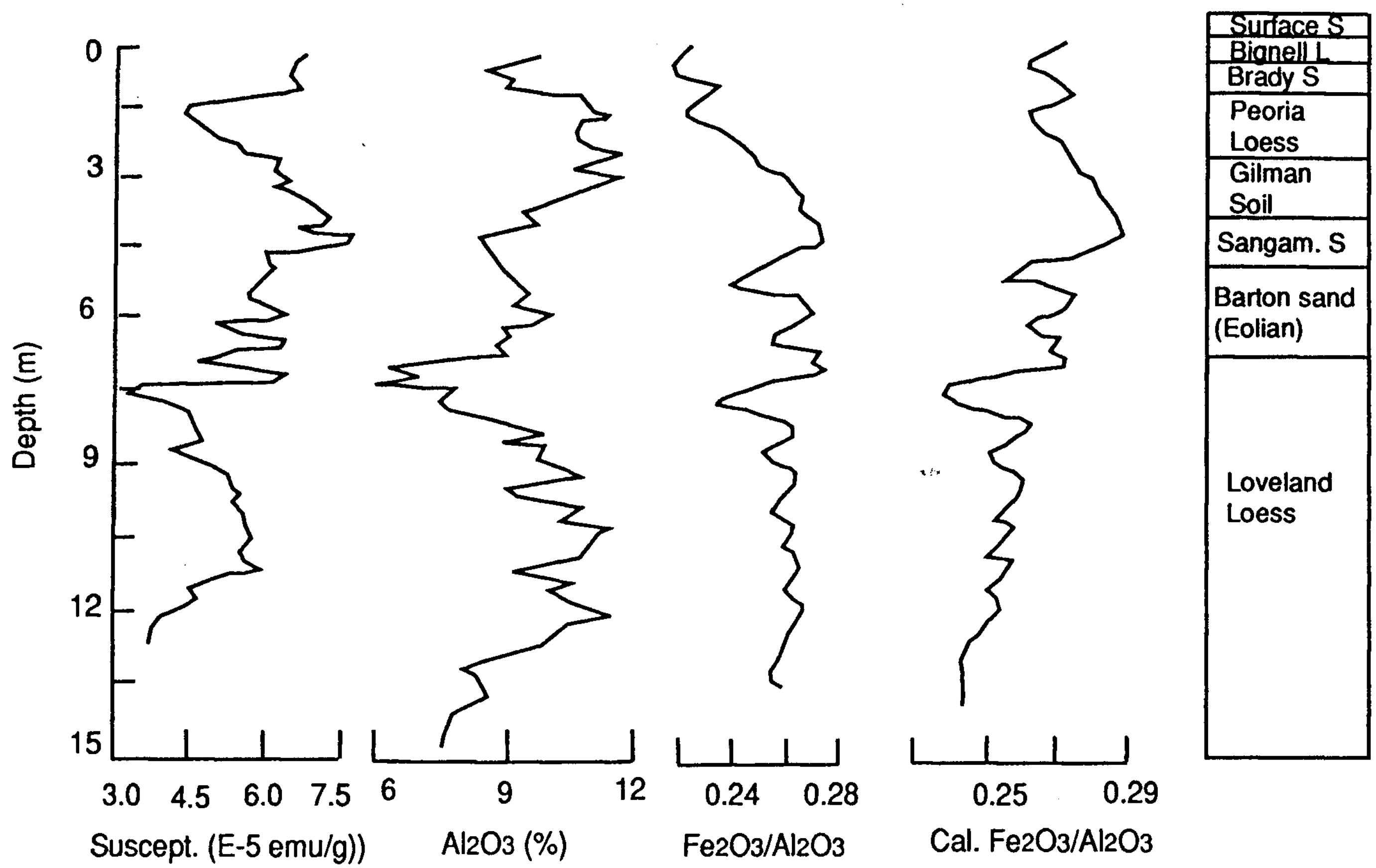


Figure 17. The comparison of the magnetic susceptibility, content of Al2O3, ratio of Fe2O3/Al2O3 and calibrated ratio of Fe2O3/Al2O3 at the Barton County section.

To filter out the first-order change (background) and to obtain the second-order change which responds to the susceptibility, a 7-moving average (i.e., 7 data points) filter was used to smooth the original curve and produce a background curve. The difference was derived by subtracting the smoothed value from the original value for each datum. The difference plus the overall average (0.25) equals a calibrated Fe_2O_3/Al_2O_3 ratio (i.e., calibrated ratio = smoothed ratio - original ratio + 0.25). The calibrated Fe_2O_3/Al_2O_3 ratio corresponds very well to the susceptibility curve (cf. Figure 17), i.e., the two curves are parallel with each other. The correlation coefficient between the calibrated Fe_2O_3/Al_2O_3 ratio and the magnetic susceptibility is 0.93 (r^2=0.86). In brief, the Fe_2O_3/Al_2O_3 ratio works quite well in recording the depositional environment of the loess, if the geological background is filtered out.

The clay mineralogy, chemical parameters, and physical indicators demonstrate that the weathering intensity was indeed low during the warm periods of Loveland time when the pedogenic carbonate-enriched clayey layers were formed. The sand unit (4.2-6.6 m) is characterized by three small weathering peaks; the Sangamon pedocomplex has been most strongly weathered. The Gilman Canyon pedocomplex is typified by strong physical weathering. Although the Al_2O_3 content is high throughout the Peoria Loess and the clay content is high in the upper portion of the Peoria Loess, the magnetic susceptibility, the kaolinite/quartz ratio in clay, and the oxidization index indicate that the loess was poorly weathered. The Brady Soil is well expressed not only by visible features, but also by physical and chemical parameters. The Bignell Loess, though quite thin, is characterized by lower values in physical, mineralogical and chemical parameters. The surface soil does express itself, though not strongly.

The quartz/feldspar ratio in the fine-sand fraction demonstrates the difference in weathering intensities between the loess and its soils in general. The ratio is only 3 for Bignell and Peoria Loesses and higher for soils. For example, the ratio is 30 for the Brady Soil, 5 for the Gilman Canyon pedocomplex, and 7 for the Sangamon pedocomplex. The magnetic susceptiblity peak around 9.5 m deep is corresponded with a high quartz/feldspar ratio, 30. But, the ratio is far from a sensitive indicator of weathering intensity. As discussed earlier, under semiarid-arid environment where loess deposition has dominated, weathering has not been sufficient to alter the fine sand fraction. Thus, the mineralogical weathering index (quartz/feldspar in fine sand) fails to indicate weathering intensity. The fine sand fraction has recorded the weathering history only under a moist and stable land-surface environment where chemical imprints of weathering were altered after deposition .

Stable isotopes, $\delta^{13}C$ and $\delta^{18}O$, have been used to elucidate soil-forming conditions at the time of soil formation (Kelly et al., 1990; Quade and Cerling, 1990; Schlesinger et al., 1988). Isotopic studies of modern pedogenic carbonate have been applied to the interpretation of paleosol environments (Cerling et al., 1989; DeNiro, 1982; Gardner et al., 1991; Krishnamurthy, 1990).

Cerling (1984) found a positive correlation between $\delta^{18}O$ values of rainfall and $\delta^{18}O$ values in soil carbonate in Arizona. He concluded that the level of $\delta^{18}O$ in pedogenic carbonate is influenced by rainfall and temperature. Temperature affects not only the $\delta^{18}O$ values of rainfall, but also the fractionation of $\delta^{18}O$ as carbonate precipitates from the soil solution. The level of fractionation of oxygen isotope in carbonate formation is inversely proportional to temperature. The $\delta^{18}O$ in carbonate is influenced mainly by the oxygen-18 of the water, because there is a large quantity of water and very little CO_2 in the reaction resulting in carbonate. Consequently, the effect of temperature on $\delta^{18}O$

fractionation in carbonate synthesis is much less than its effect on $\delta^{18}O$ in rainfall. Therefore, it is believed that the $\delta^{18}O$ of rainfall is the primary determinant of $\delta^{18}O$ level in pedogenic carbonate (Schlesinger et al., 1988).

Studies on soils of the arid and semiarid regions of American Southwest show that soil carbonate concentration in Bk or Btk horizons is a function of soil moisture (Ruhe, 1970), which is influenced by climate, i.e., the drier the climate, the higher the soil carbonate concentration. If the climate is too dry, there is insufficient moisture to form carbonate; if the climate is too wet, the carbonate is leached away (Gile et al., 1965). In the Great Plains, soil moisture is determined by evaporation, which is controlled primarily by temperature (Aandahl, 1982). Therefore, the carbonate concentration in the Barton County section indicates long-term changes in temperature, i.e., carbonate accumulated more when temperature was high.

The negative correlation of carbonate concentration with degree of weathering suggests that the weathering was weak during carbonate accumulation. Comparing the chronology of loess in central Kansas with that of the Pacific V28-238 core indicates that carbonate-enriched layers in central Kansas were formed during warm periods (discussed in Chapter V). That weak weathering occurred during warm periods indicates carbonate enrichment periods were not only warm, but also dry.

Based on above discussion, the $\delta^{18}O$ of carbonate is primarily influenced by the temperature of carbonate enrichment periods, while soil carbonate concentration in soil Btk or Bk horizons is determined basically by soil moisture at the time of soil formation. In the Great Plains, the soil moisture might have been closely associated with the temperature.

Ten samples of soil carbonate were analyzed for $\delta^{18}O$ and $\delta^{13}C$ (Table 9). Among them, four samples are from the sand unit (Barton sand: 4.2-6.6 m), four from the first carbonate-enriched clayey layer (6.8-8.2 m) and two from the second carbonate-enriched clayey layer (9.8-12.5).

The relationship between carbonate concentration (X) and the $\delta^{18}O$ of the carbonate (Y) is not linesar (Figure 18a), i.e., both low and high values in carbonate concentration correspond to higher $\delta^{18}O$. But, if the carbonate in sand (Barton sand) and the carbonate in loess (Loveland Formation) are considered separately, it can be seen that $\delta^{18}O$ rises with an increase in carbonate concentration of loess (Loveland), and $\delta^{18}O$ falls with an increase in carbonate concentration within the sand (Barton sand). This relationship suggests that during the period of eolian sand deposition (70-95 ka), the

Table 9. Relationships of the carbonate concentration to the $\delta^{13}C$ and $\delta^{18}O$ at the Barton County section (Strata: BF = Barton sand; LLc = Loveland Loess carbonate layer).

Sample (#)	Depth (m)	Strata	Carbonate (%)	$\delta^{13}C$ (PDB) (‰)	$\delta^{18}O$(PDB) (‰)
BT25	-5	Bs	1.18	-0.44	-1.35
BT26	-5.2	Bs	1.34	-0.37	-4.34
BT30	-6.0	Bs	2.56	-0.60	-5.02
BT31	-6.2	Bs	6.46	-1.89	-6.68
BT34	-6.8	LL c1	18.5	-5.36	-4.90
BT36	-7.2	LLc1	13.4	-3.37	-8.52
BT39	-7.8	LLc1	19.1	-3.31	-7.00
BT40	-8.0	LLc1	8.52	-3.68	-6.03
BT56	-11.2	LLc2	13.0	-3.11	-7.85
BT61	-12.5	LLc2	16.2	-3.26	-6.69

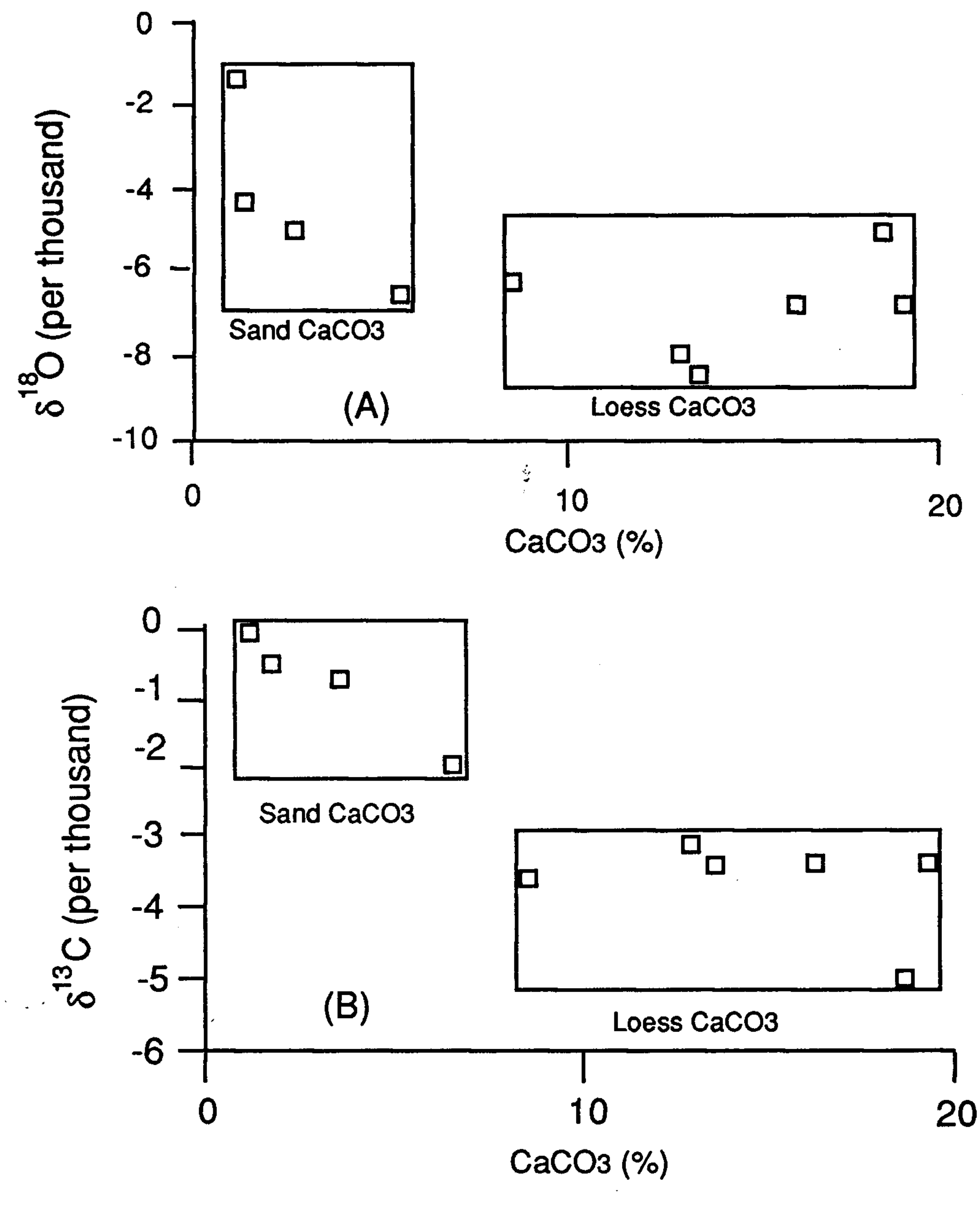


Figure 18. The relationships of the carbonate concentration to the $\delta^{18}O$ (A) and $\delta^{13}C$ (B) at the Barton County section of central Kansas.

temperature was so high and the soil moisture was so low that carbonate did not form effectively. Another possibility is that the duration of carbonate accumulation was too brief for carbonate to concentrate effectively. The negative relation between $\delta^{18}O$ and carbonate concentration in the sand (Barton sand) suggests that more carbonate accumulated as the temperature decreased (i.e., effective soil moisture increased) during sand deposition. On the contrary, a slight positive correlation in the first and second carbonate-enriched layers of the Loveland Loess shows that more carbonate accumulated as the temperature increased (i.e., effective soil moisture decreased) during last two warm periods of Loveland time. In brief, during the sand deposition (early Wisconsin), carbonate concentration was controlled mainly by the effectiveness of carbonate translocation, i.e., availability of soil moisture, while during the deposition of the Loveland Loess, carbonate concentration was controlled mainly by the effectiveness of evaporation. This conclusion is also supported by magnetic susceptibility. The susceptibility is positively correlated with carbonate in the Barton sand, probably indicating that carbonate increased with increasing moisture, and it is negatively correlated with carbonate, suggesting that carbonate increased with decreasing moisture.

Regional controls on the $\delta^{13}C$ in soil carbonate are more complex. The $\delta^{13}C$ in carbonate is dependent upon the ratio of CO_2 in the soil and air and fractionation that occurs between CO_2 and carbonate during the precipitation process. The ratio in soil CO_2 is influenced by the proportion of C3 and C4 plants in the local flora and the exchange with the external atmosphere during soil respiration and diffusion (Rundel et al., 1989).

Research in Arizona demonstrated that the $\delta^{13}C$ of soil carbonate is influenced by the degree of atmospheric mixing of CO_2 during cold periods of the year and by the composition of vegetation during warm period of the year when the plant respiration rate is high (Cerling, 1984; Parada et al., 1983). Since soil moisture strongly governs the vegetation patterns, and thus soil respiration rates in arid regions, increasing aridity should result in greater penetration of atmospheric CO_2 and thus pedogenic carbonate, which is enriched in $\delta^{13}C$ (Amundson et al., 1989; Cerling and Hay, 1986). The study in Yucca Mountain, Nevada (Quade and Cerling, 1990) showed that the $\delta^{13}C$ of modern soil carbonate decreases (more negative) with increasing elevation, i.e., the $\delta^{13}C$ increases with increasing temperature and decreasing moisture.

This study demonstrates a result contrary to that by Quade and Cerling (1990). The carbonate concentration and the $\delta^{13}C$ of the carbonate in the Barton County section are negativly correlated (Figure 18b). The linear correlation coefficient is -0.89, demonstrating that the $\delta^{13}C$ decreased with increasing temperature and/or decreasing moisture, if the carbonate concentration really indicates the temperature and/or the moisture. The discrepancy might be resolved if the sand carbonate and loess carbonate are considered separately. Specifically, high $\delta^{13}C$ values might have indeed occurred in warm and dry conditions when the Barton sand (sand) was deposited, although carbonate did not effectively accumulate probably due to a very dry condition (Figure 19).

If the data points of the sand carbonate were removed, the negative correlation between the carbonate concentration and the $\delta^{13}C$ disappears. In fact, a significant difference between the sand carbonate and the loess carbonate exists. The $\delta^{13}C$ of the sand carbonate ranges from -0.37 to -1.89‰; while the $\delta^{13}C$ of the loess carbonate ranges from -3.11 to - 3.68‰ with an extreme value of -5.36‰ (cf. Table 9), indicating the $\delta^{13}C$ of the loess carbonate does not vary significantly with respect to the carbonate concentration.

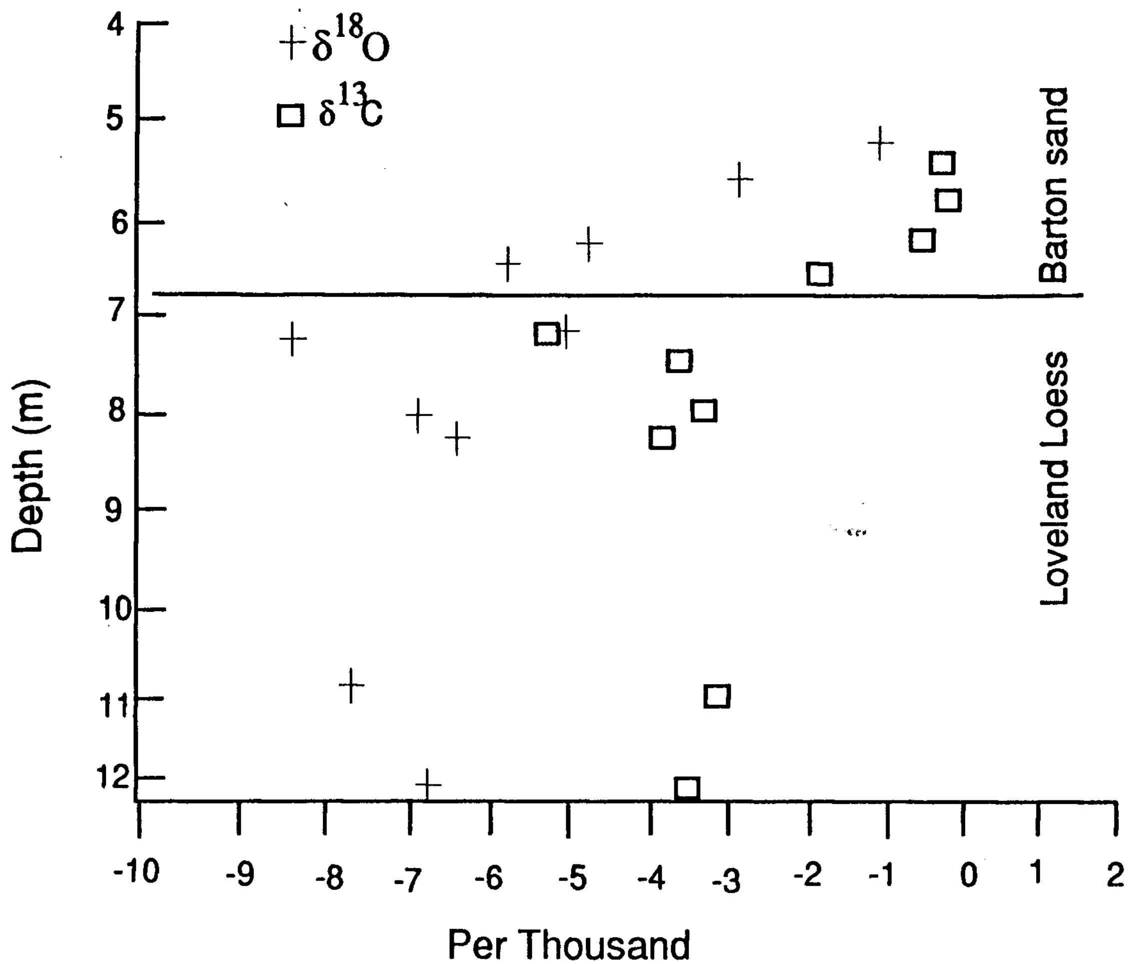


Figure 19. The distribution of $\delta^{18}O$ and $\delta^{13}C$ with depth of loess in the Barton County section.

It is well known that three different photosynthetic pathways exist in plants: C3, C4 and CAM, which produce organic matter with a carbon isotope composition of -24 to -34, -9 to -16, and -9 to -19‰, respectively (Hoefs, 1980). Cerling and others (1989) showed that the $\delta^{13}C$ of soil carbonate is closely correlated with that of soil organic carbon. They found that the two differ systematically by -15‰ in modern soil. Adding the -15‰ of the difference to the $\delta^{13}C$ of soil carbonate and comparing the new values ($\delta^{13}C$ of soil carbonate plus 15‰) with Hoefs' values show that the $\delta^{13}C$ of soil carbonate in the Barton County section falls between C3 and CAM plants (-16 to -21‰), meaning that there were more C3 and CAM plants than C4 plants. The loess carbonate ($\delta^{13}C$ = 15-18‰) has more of a C3 plant cover than the sand carbonate (the $\delta^{13}C$ = 19-21‰).

Although the potential of using the isotopes to analyze the past soil-forming conditions is great, the interpretation of this study needs to be tested further. A severe problem encountered in interpreting the isotopic data is that the processes of carbon incorporation into the carbonate of soil are not clearly understood yet.

Great Bend Sand Prairie Section

Unlike the other sites which have large local relief, the central Great Bend section (Groundwater Management District No. 5, Recharge Well 7, in the northernmost Pratt County) has been strongly influenced by groundwater. Five trenches excavated in the Great Bend Sand Prairie demonstrate that the top 1.5-2.5 m is sand, and that silt extends down at least 3 m (Johnson and Sophocleous, 1991). Ages of soils developed within the silt underlying the dune sand demonstrate that the land surface was probably stable, with soil forming on the silt surface about 13,600 yr BP. The 13,600 yr BP soil may be the equivalent of the sand peat dated at 13,160 yr BP in the Sandhills of Nebraska (Swinehart, 1990), and a charcoal concentration around 14,000 yr BP in loess sequences in central Nebraska and western Kansas (May, 1989; Wells et al., 1987). Local accretionary soils (Jules Soil) within the Peoria Loess in the Central Lowlands dated at 12 to 15 ka (Frye at al., 1974; Follmer, 1983) may also correspond to the 13,600 yr accretionary soil. No soil has yet been found among these exposures from 13,600 yr BP to 5,800 yr BP, but the existence of the Brady Soil dated at 10,500-8,500 yr BP on the uplands nearby indicates another soil development event between 13,000 and 5,800 yr BP. After the Brady Soil, sand dune activity dominated until about

5,800 yr BP, when dune sand was stabilized and a soil was formed in the reexposed silt surface.

Soils from the same stratigraphic position at two localities were dated at 5,800 and 4,600-4,070 yr BP, respectively. Considering the variables operating in ^{14}C dating, this soil might have started at 5,800 yr BP and lasted until about 4,100 yr BP. After that time, sand dunes have dominated the region with four major hiatuses around 2,900 yr BP, 1,600 yr BP, 800 yr BP and the time when the surface soil developed. The surface blow sand burying the surface soil in the Great Bend Sand Prairie demonstrates recent degeneration of the environment.

Clay content is negligible in the sequence, and silt and sand contents are negatively correlated (Figure 20 and Table 10). The organic carbon content is well correlated with silt content ($r = 0.80$, Y [silt] $= 1.214 + 0.027X$ [carbon]). In this section, when the soil dated at 4,070 was formed, silt content reached its maximum (74.9%). Sandy silt extends from 2.16 to 2.56 m. The sand percentage increases from 30% at 2.16 m to 68% at 1.95 m, and subsequently decreases to 46% at 1.75 m, where the 1,600 yr BP soil occurs. The sand increases drastically from 46% to 91% at 1.6 m. Another soil (probably equivalent of 800 yr BP soil dated elsewhere in the Sand Prairies) was encountered at 1.2 m. This soil is indicated

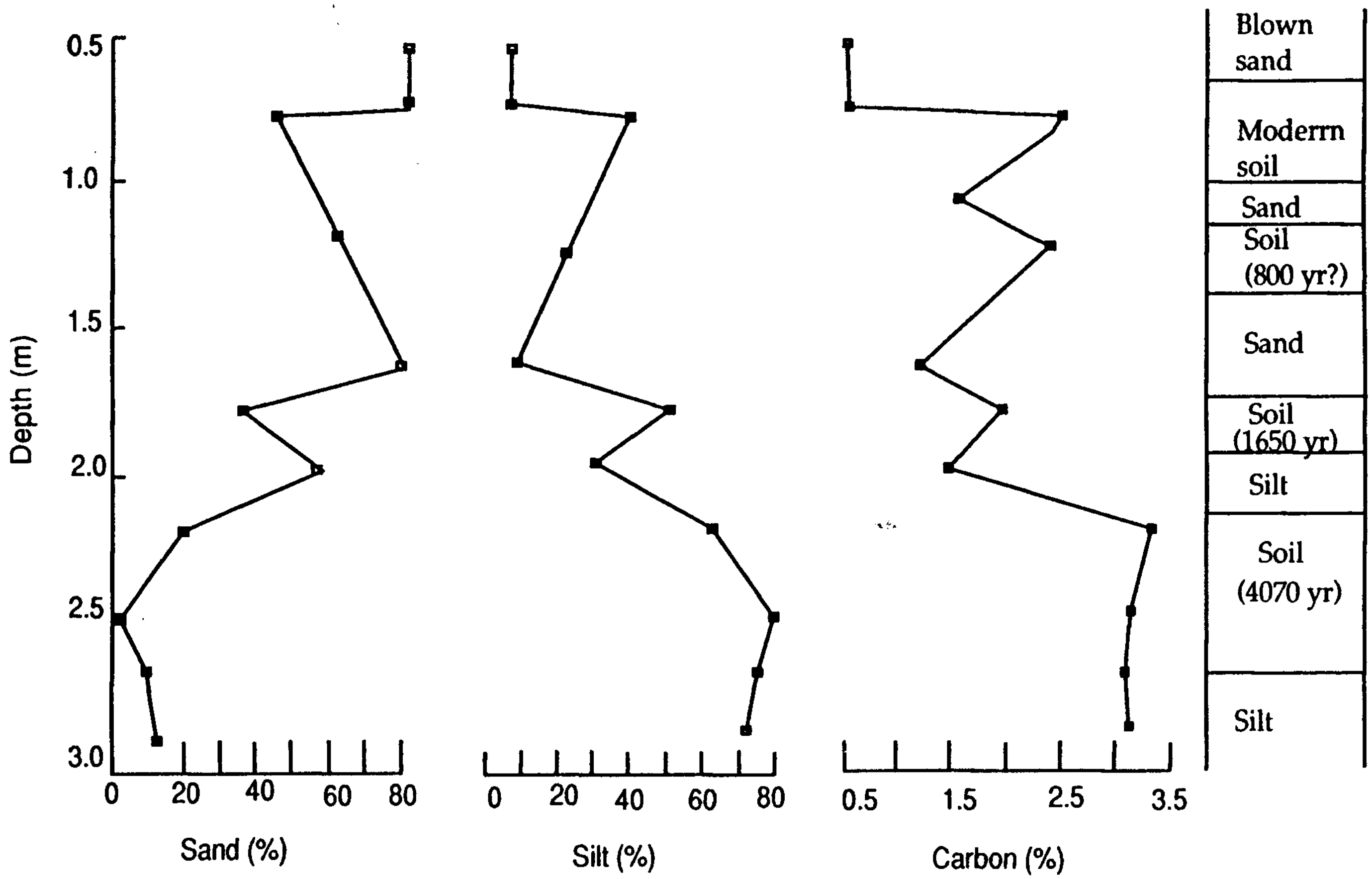


Figure 20. The spectra of the sand, silt and organic carbon distributions at the central Great Bend Sand Prairie section.

Table 10. Physical and chemical properties of the loess sequence in the central Great Bend Sand Prairie section of central Kansas (a Holocene sequence).

Depth (m)	Sand (%)	Silt (%)	Clay (%)	Carbon (%)	L.I.	Quartz (%)	Kaolini (%)	Ka/Qu	Fe2O3 (%)	Al2O3 (%)	O.I
0.2	92.0	7.0	1.0	0.7	1.53	23	8.3	0.36	2.9	10.7	0.27
0.4	92.0	7.0	1.0	0.9	1.53	23	8.3	0.36	2.9	10.7	0.27
0.75	56.4	47.9	5.2	2.5	1.67	25	9.5	0.37	2.8	10.6	0.26
1.05	58.2	36.0	5.8	1.6							
1.20	65.0	28.3	6.7	2.6							
1.6	91.0	8.4	1.6	1.2	1.62	27	10.5	0.39	2.8	10.6	0.26
1.75	46.3	48.1	4.6	1.9	1.56	27	10.5	0.39	1.9	9.6	0.19
1.95	68.4	30.3	1.2	1.5	1.48	27	10.5	0.39	1.6	10.0	0.16
2.16	29.7	55.7	4.6	3.4	1.47	28	14.3	0.51	1.5	10.3	0.14
2.46	11.8	74.9	12.8	3.2	1.29	19	5.5	0.29	1.6	10.4	0.15
2.65	19.4	70.5	10.0	3.1	1.47	23	4.4	0.19	1.2	9.3	0.13

by only a relatively high organic carbon content (cf. Table 10). A sand unit extends from 1.2 to 0.75 m, and the carbon content increases up 2.53% and the sand percentage drops to 56% at 0.75 m. This silty and organic carbon-enriched layer is the surface soil. The top 0.72 m is dominated by sand (92%), i.e., surface-blow sand burying the surface soil (cf. Figure 20).

The same parameters used in the other sites suggest that the sand portion (upper half) is characterized by a high Fe2O3 content (about 2.8%) and a high Fe_2O_3/Al_2O_3 ratio (0.28). These two values are less in the lower silt portion, 1.5% and 0.15, respectively. These contrasts do not provide any climatic or climatically controlled environmental implications. Instead, they only yield local, topographically or hydrologically controlled environmental information. Because the lower portion has long been under reducing conditions, it is not surprising that the Fe_2O_3 content is very low. This reducing and nonleaching environment is also indicated by a high Na_2O content in the lower silt.

Sources of Peoria and Bignell Loesses

Historical records suggest that dust storms originating from the southwestern and western Great Plains and travelling for long distances have been a major source of dust in the midcontinental United States. Dust storms occurring during the late Quaternary must have preserved some information about the sources. The information may indicate the intensity and direction of winds during various times of the late Quaternary.

Modern Dust Storms

Wind is probably the most conspicuous natural phenomenon in the Great Plains. The Great Plains region is the most persistently windy inland area of North America and also has some of the highest average annual and fast wind speeds of any nonmaritime region.

The effect of wind on the geologic evolution of the Great Plains is significant. As a result, many of the Quaternary deposits are eolian. The wind continues to be important geologically, often scouring the surface and producing huge clouds of dust. Indeed, the Southern and Central Great Plains are the dustiest region in North America (McCauley et al., 1981).

The severe dust storms occurring in the late 1890s, 1930s, 1950s and in the year 1977 (Malin, 1946; McCauley et al., 1981; Roserberg, 1979) suggest that the Southern and western Central Great Plains have been the dominant sources of dust storms. It is notable that the center of dust-storm activity lies in a region including southwestern Kansas, northwestern Oklahoma, northwestern

Texas, and northeastern New Mexico (Figure 21a), the region with the highest annual and fastest wind speeds (Goudie, 1983; Ruller and Bair, 1977). Dust storms are associated with prolonged drought and high temperatures, such as those occurring in the 1930s, 1950s, and 1977. They originated primarily from the Southern and western Central Great Plains, moved northeastward, and reached the north-central Atlantic Ocean (Jackson et al., 1973; Smith et al., 1970).

Statistics derived from Marlin (1946) indicate that from 1890 to 1940 about 68% of the Kansas dust-storms occurred in the spring and started in the 1930's dust storm center, the region including northeastern New Mexico, northwestern Oklahoma, northwestern Texas, and southwestern Kansas. About 17% of the dust storms started in eastern Colorado during the spring. The rest occurred in winter and originated from western Nebraska or to the north. Only one of more than 200 dust storms in Kansas from 1890 to 1940 originated from the Platte River valley, which has been considered the most important source of Peoria Loess (Frye and Leonard, 1951). This pattern of dust-storm directions is supported by the fast wind distribution patterns (Kutzbach and Wright, 1985), i.e., northwestern winds dominate in northwestern Kansas, while southwestern winds dominate in southwestern Kansas, and

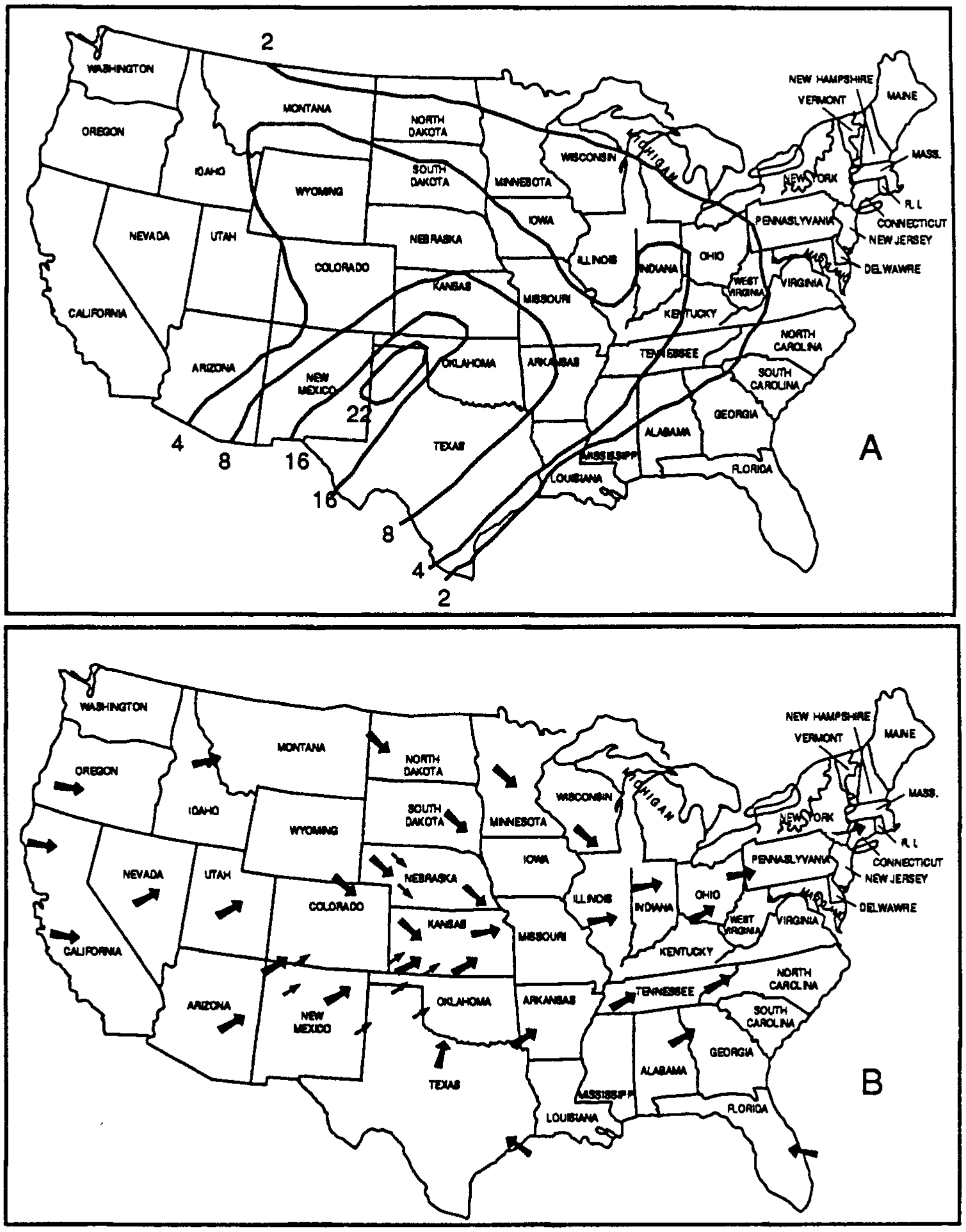


Figure 21. Comparison of the dust storm frequency (map A, days per month) during the March of 1936 (Goudie, 1983) with the directions (large arrows) of fast winds (>6 m/s) and with the trends (small arrows) of active sand dunes (map B, Kutzbach and Wright, 1985; Swinehart, 1990).

the two wind patterns converged in the eastern half of Kansas (Figure 21b). According to a study on dust tracks in the 1950s, it was strongly believed that dust storms were tightly associated with the mid-latitudinal cyclonic systems in the United States, which typically move from the southwest to the northeast and directly toward Greenland and the north-central Atlantic Ocean (Smith et al., 1970).

In regard to efficiency of the dust storms to transport dust, the Central Great Plains region is one of the dustiest areas in the world. For example, the dust-depositional rate in individual dustfall event is 371 tons per square kilometer in Egypt, and 300 tons per square kilometer in Nebraska and Kansas (Goudie, 1978). Except for locally originated dust storms (Smith et al., 1970), most of the dust storms traveled 2500 to 10,000 km in the midcontinental United States (Goudie, 1983; Jackson et al., 1973).

Historical data show that the dust storms in the Great Plains have been associated mainly with droughts. The droughts accompanied a persistence in either meridional or zonal flow (Barry, 1983; Borchert, 1971). This persistent flow led to high temperatures and strong winds, loss of soil moisture, destruction of vegetation cover, and surface sediment susceptible to eolian movement (Holliday, 1987b).

Based on correlation analysis of dust properties of the 1960's dust storms for 37 months at 14 sites east of the Rocky Mountains, Jackson and others (1973) showed that the dust from Ohio, Texas, North Dakota, Nebraska, Colorado, Kansas, Missouri, and Mississippi are correlated very closely chemically and minerologically. The consistency of the dust properties in a vast geographic area east of the Rocky Mountains suggests that the modern dust storms originated from relatively uniform sources and were transported for a long distance from the region including the western Central Great Plains and the Southern Great Plains. This fact may imply that (1) the Peoria and Bignell Loesses might have also been produced by dust storms travelling long-distances, and (2) the popularly accepted sources, local rivers, might have added only coarse particles locally during the Peoria and Bignell Loess deposition.

Meteorological studies of the dust storms in the midcontinental United States (Flora, 1946; Malin, 1946; Ruller and Bair, 1977) demonstrate that there are three types of duststorms: (1) blowing dust of a local nature by lower air currents (low intensity and low volume); (2) dust carried long distances by upper air currents (high intensity and low volume); and (3) severe dust carried relatively long distances by lower and middle air currents (high intensity and high volume). The second type can occur at almost any time. The third type, the most effective one in transporting silt, happens only during prolonged drought periods.

In summary, southwesterly winds have been a major generator of the dust storms in the Central and Southern Great Plains historically. Did the Peoria and Bignell Loesses originate locally or were they transported by long-distance travelling dust storms? Undoubtedly, the northwesterly winds prevailed during the Peoria time. What was the importance of the southwesterly winds during that time? Were the southwesterly winds the dominant source of silt for the Bignell Loess in Kansas? The following discussion addresses these questions through particle size distribution and chemical parameters.

Particle size Distribution

Lugn (1965) demonstrated that the sand percentage of the Peoria Loess of Nebraska decreases southeastward from the Sandhills of Nebraska. The sand percentage is 75% at the southern margin of the Sandhills and 25% at the southern margin of the state (Figure 22a).

Swineford and Frye (1951) showed that sand percentage in the Peoria Loess of Kansas is controlled mainly by

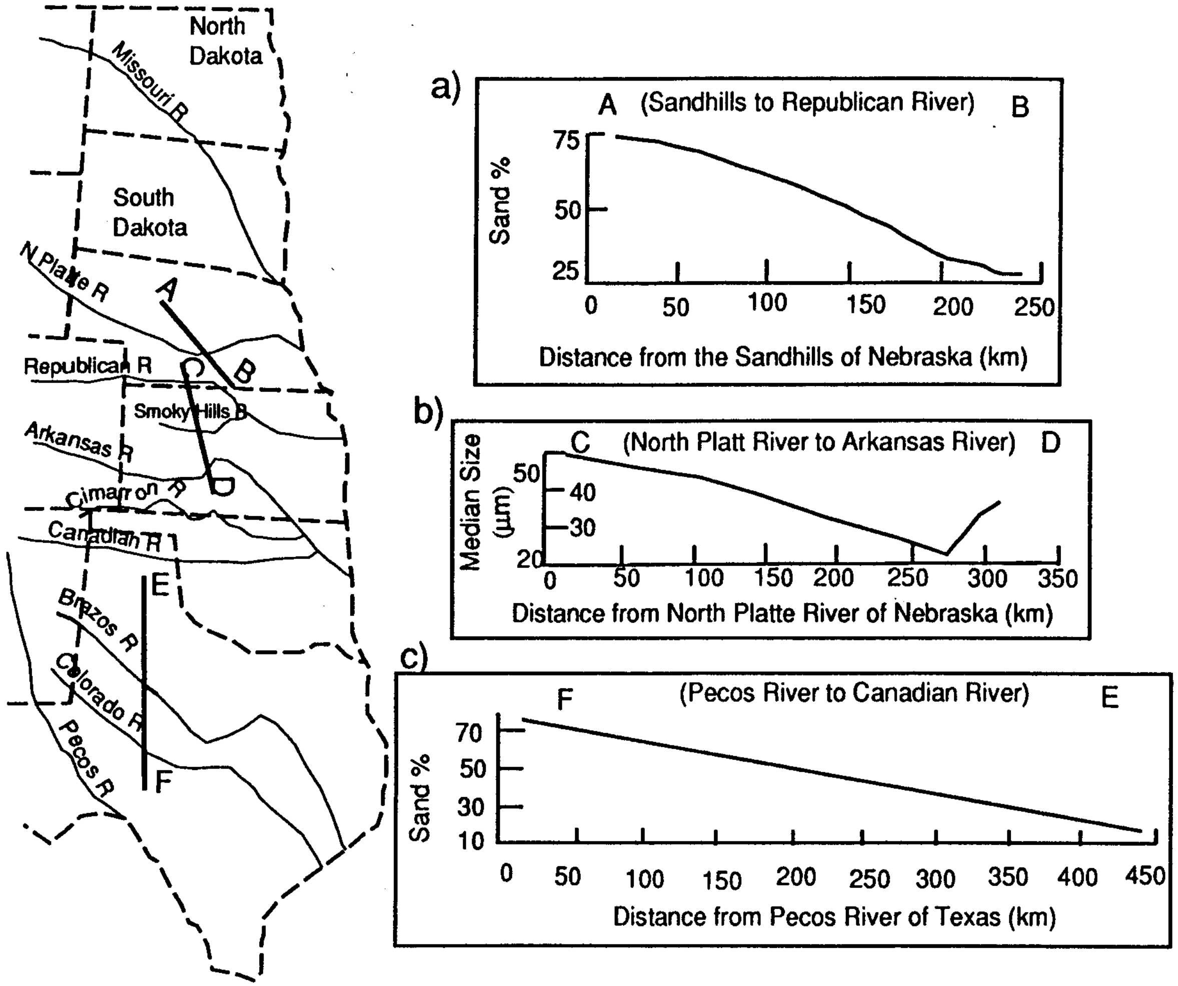


Figure 22. Particle size distribution of loess in the Central Great Plains and the Southern Great Plains. Transects A-B (a) and C-D (b) indicate north-south decreases in sand content and median size of loess (Frye and Leonard, 1951; Lugn, 1960). Transect F-E (c) shows a south-north decrease in sand content of loess (Holliday, 1989b).

proximity to large river valleys, but the median diameter of Peoria Loess decreases systematically from north to south and from west to east. This overall trend is apparently reversed adjacent to the Arkansas River (Figure 22b).

In the Southern Great Plains, Seitlheko (1975) and Holliday (1989b) demonstrated that the sand percentage, the mean sand size and the ratio of sand to silt decrease from the Peco River in southwestern Texas, northeastward to the Canadian River in the panhandle region of northern Texas (Figure 22c). The loess (Peoria in age) in northern Texas is quite clayey.

As Swineford and Frye (1951) indicated, sand percentage and median size of the Peoria Loess exhibits a coarse loess island in the region immediately north of the Canadian River. They attributed this island of coarse Peoria Loess to local contributions from the Canadian River by southerly and southwesterly winds. The distribution of particle size indicates southwestern and central Kansas might have been the convergence zone of two opposing wind systems: northwesterly and southwesterly.

Some authors (e.g., Ahlbrandt et al., 1983; Zhu et al., 1985) noted that, repeatedly shifted and sorted sand is relatively silt-free and almost absolutely clay-free. Observations of sand movement (Bagnold, 1959; Zhu et al.,

1985) and dust-storm occurrence (McCauley et al., 1981; Zhang 1984) indicate that coarse and medium sand is moved by traction on the ground, fine sand by saltation mainly within 2 m height from the ground, and silt by suspension (airborne). Coarse clay could be transported as individual particles in suspension, but fine clay is transported with silt either as clay aggregates or as attachments to silt particles.

As mentioned earlier, the Pratt section consists of three units: clayey silt in the upper part (0-3.4 m), sandy silt in the middle part (3.4-5.8 m), and sand with small pebbles of the lower part (below 5.8 m). The upper and middle parts are eolian in origin and the lower part is fluvial. The percentages of particle number for each particle-size interval of all 65 samples were averaged, and then the averaged percentage of each particle-size interval was plotted against the particle size (Figure 23). There are three peaks of particle-size distribution: 0 phi and larger, 0-5 phi, and 5 phi and smaller. The coarsest peak was formed by fluvial processes; the peak from 0 to 5 phi unit by wind traction and saltation processes; the finest peak (less than 5 phi unit) by windborne transportation. Therefore, mean size of the windborne portion is used to trace the direction of winds carrying the Peoria and Bignell Loesses.

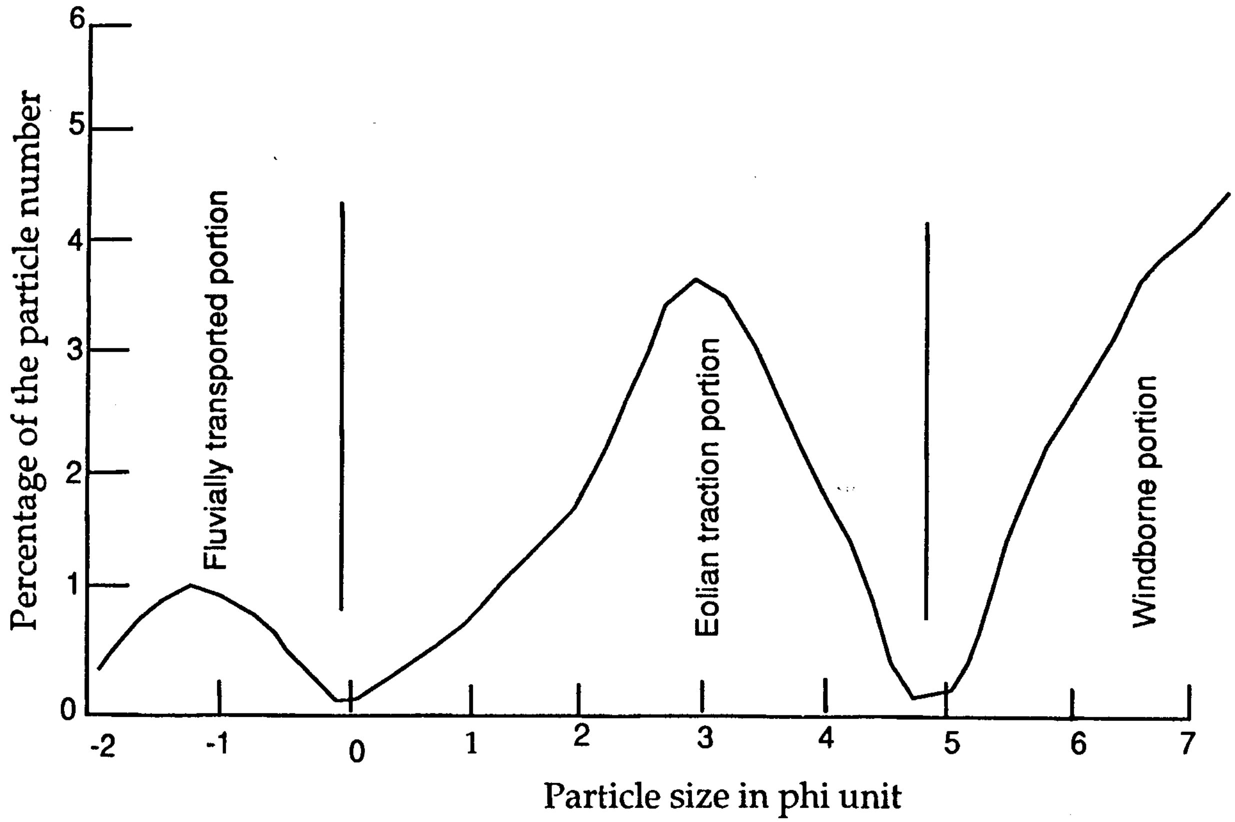


Figure 23. Particle size distribution of the Pratt County section reflecting two thresholds in the tranportation processes: (1) the thresold between eolian and fluvial deposits, (2) the threshold between wind traction and windborne deposits.

Feng and Thompson (1987) showed that the sand percentage of the Peoria Loess in the midcontinental United States is high only in proximity to large rivers. This result is generally in agreement with that of Swineford and Frye (1951). On the contrary, the silt fraction changes systematically from western Nebraska and Kansas, eastward to Ohio, indicating that the Peoria Loess (silt) was transported a long distance.

As discussed earlier, during the late Pleistocene, southwesterly winds were dominant in the Southern Great Plains, at least as far as the southwestern corner of Kansas. At the same time, northwesterly winds were dominant in the Central Great Plains, at least to the Arkansas River of central Kansas. The CLIMAP climatic model supports the argument that the southwesterly winds contributed to the Peoria Loess in southwestern, and probably central Kansas. The model indicates that during the last glaciation the westerly jet was split into two branches: northern and southern. The southern branch prevailed the Southern Great Plains and produced southwesterly winds, and the winds touched the southwestern kansas (Kutzbach and Wright, 1985).

There is a lack of data regarding to the wind patterns since the late Pleistocene in the region south of the

Arkansas River and north of the Cimarron River in southwestern Kansas. To find the temporal and spatial variations of wind patterns since the late Pleistocene in the study area, a parameter, mean size of the suspended fraction (50 μm to 1.9 μm), was used to examine microscale variation of particle size. Table 11 shows that the mean size at the Phillips County site (north-central Kansas) is 6.29 μm for the Bignell Loess and 7.48 μm for the Peoria Loess. The mean size for the Kearny County site (southwestern Kansas) is 7.61 μm for the Bignell Loess (the surface soil is taken as the equivalent of the Bignell Loess because there is no Brady Soil here functioning as base to identify the Bignell Loess) and 8.37 μm for the Peoria Loess. At the Pratt County site (south of the Arkansas River), the mean sizes are 3.11 μm for the Bignell Loess and 3.41 μm for the Peoria Loess. At the Barton County site (north of the Arkansas River), the mean sizes are 2.99 μm for the Bignell Loess and 3.54 μm for the Peoria Loess. The spatial distribution of these mean sizes suggests that the Kearny County site might have been closer to its source than the Phillips County site, probably indicating the importance of westerly winds. The mean size of the Pratt County site (3.11 μm) is larger than that of the Barton County site (2.99 μm) for the Bignell Loess, suggesting that southerly and southwesterly

Table 11. Comparison of particle size of the Bignell and Peoria Loesses at the four research sites: Phillips County (north-central Kansas), Kearny County (southwestern Kansas), Pratt County and Barton County (Central Kansas).

Site	Stratigraphy	50 µm	30 µm	20 µm	10 µm	6 µm	2 µm	1.9 µm	Mean size µm
Phillips	Bignell Loess	0.05	0.30	4.4	10.2	64.7	19.5	0.7	6.29
Kearny	Bignell Loess	0.22	1.70	8.3	13.6	57.4	14.7	1.3	7.62
Pratt	Bignell Loess	1.65	0.05	0.3	2.1	4.1	17.6	74.3	3.11
Barton	Bignell Loess	0.10	0.13	0.7	2.7	5.6	39.0	50.0	2.99
Phillips	Peoria Loess	0.13	1.77	8.7	14.3	56.1	18.5	1.1	7.48
Kearny	Peoria Loess	0.54	3.20	9.6	25.4	60.0	14.9	1.2	8.37
Pratt	Peoria Loess	3.00	0.05	0.4	2.29	6.1	31.6	56.7	3.41
Barton	Peoria Loess	1.30	0.10	0.6	4.2	6.1	45.0	43.0	3.54

winds might have been predominant in central Kansas during the Bignell time (early-middle Holocene). The mean size of the Peoria Loess at the Pratt County site (3.41 μm) is slightly smaller than that of the Barton County site (3.54 μm), suggesting that the Pratt County site could have been a little farther away from the source than the Barton County site during Peoria time. That is, the northerly or northwesterly winds might have dominated central Kansas during Peoria time.

Sand-dune orientations were examined in order to derive Holocene wind directions in the Great Bend Sand Prairie. Ten 7.5 minute topographic maps were selected from places where sand dunes are active and more relief, such as around the towns of Kinsley, Great Bend, and St. John. Thirty sand dunes were measured on each map to derive the orientations. The results show that about 65% of the dunes are oriented SSW-NNE or SW-NE; about 22%, W-E; and the rest, S-N. Results show that SSW-NNE or SW-NE oriented sand dunes dominate the Great Bend Sand Prairie, which is coincident with results reported by Kutzbach and Wright (1985) and Simonett (1960). They noted that sand dunes are strongly oriented by southwesterly winds in the region including southwestern Kansas, panhandles of Oklahoma and Texas, and northeastern New Mexico (cf. Figure 21b).

Some subdued and nonactive sand dunes in the regions away from the Rattlesnake Creek and the Arkansas River are, however, oriented mainly WNW-ESE or W-E. These stable sand dunes were considered as older generations of dunes (Fent, 1950). If these subdued and nonactive sand dunes are late Pleistocene in age as stated by Fent (1950), the WNW-ESE and W-E dune orientations suggest that the northwesterly winds, as well as westerly winds, dominated central Kansas during the late Pleistocene.

The clay content is unexpectedly high from the upper part of the Peoria loess to the Brady Soil in both the Pratt County and Barton County sections. Figure 24 shows that clay content is about 17% from 2.2 to 1.2 m at the Barton County section and about 6.5% from 1.8 to 0.6 m at the Pratt County section. In contrast, the clay content of the Gilman Canyon pedocomplex and the Sangamon pedocomplex is quite low. The reason for this unexpectedly high clay content from the upper part of the Peoria Loess to the Brady Soil could be that the southwesterly winds might have been dominant contributors of fine particles during the late part of the last glaciation. Again it should be emphasized that, since the southwesterly winds brought clayey loess to the panhandle region of Texas (Holliday, 1987b), it must heve also brought some clayey loess into southwestern and central Kansas.

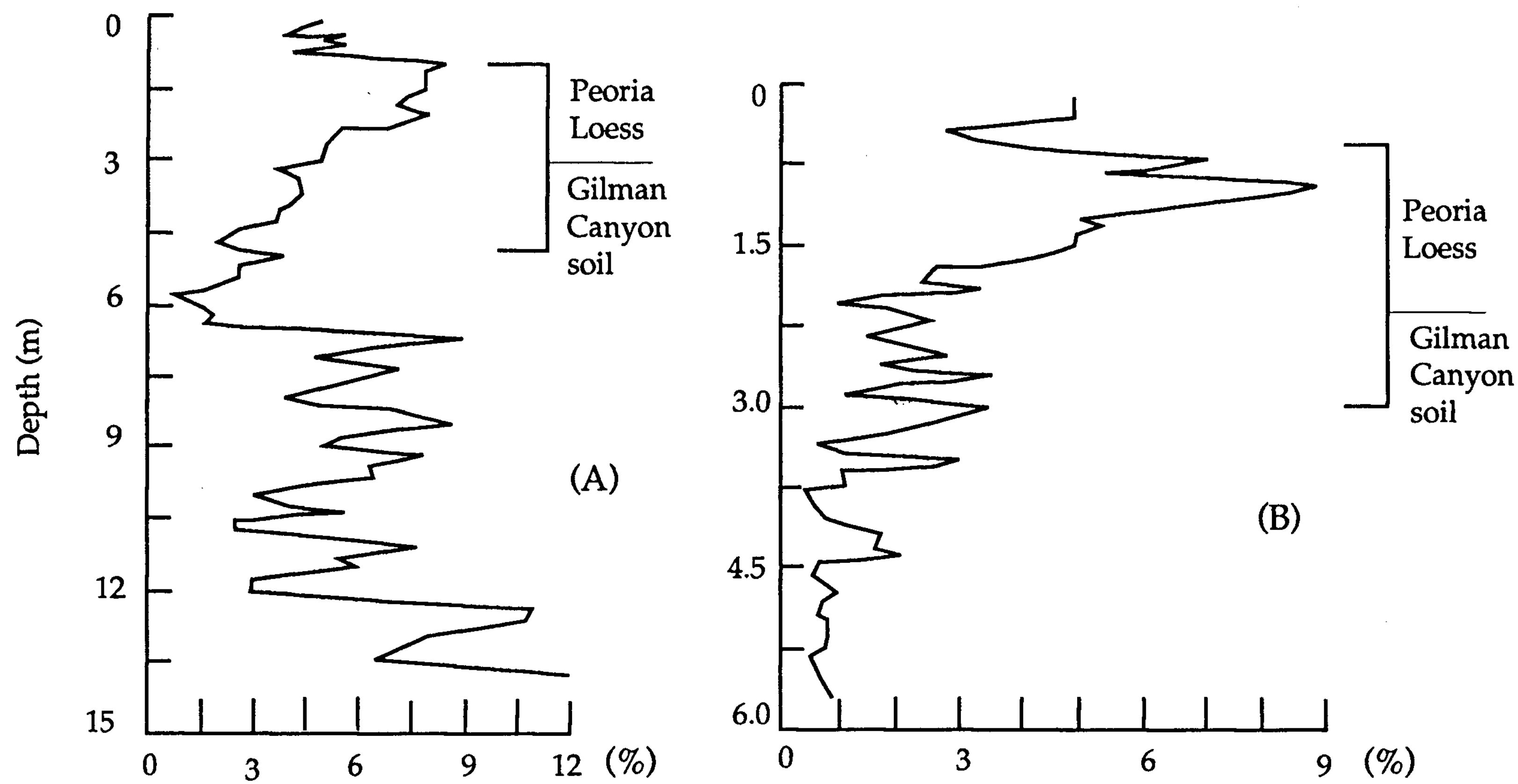


Figure 24. Clay distributions at the Barton County (A) and Pratt County (B) sections indicating high peaks in the upper portion of the Peoria Loess, in comparison with the Gilman Canyon pedocomplex.

Considering the particle size distribution of the Peoria and Bignell Loesses, the synoptic climatic picture for the late Quaternary could have been as follows. After the last glacial maximum (about 20 ka), the Laurentide ice sheet started to retreat and the deposition of Peoria Loess began. The ice continued to recede, the cold high-pressure systems became weaker and the cyclonic jets became stronger than before. As a result, winds from the north started to converge with the winds from the south, just like today's situation during early spring in the midcontinent. Although the winds from the north might have dominated central Kansas, the winds from the south became more and more prevalent and brought fine-textured loess to central Kansas during the late part of the last glaciation.

Chemical Parameters

As discussed earlier, the Fe_2O_3 content in Peoria Loess is unexpectedly high, as high as that in the Gilman Canyon pedocomplex below. Also the Al_2O_3 content of the Peoria Loess is extremely high in comparison with other stratigraphic units (Table 12). High values of Fe_2O_3 and Al_2O_3 are expected in soils because weathering processes accumulated or formed the stable oxides. But high values in the Peoria Loess need a different explanation. Based on

Table 12. Comparison of chemical composition of the four loess sequences in the study area (Fe = Fe2O3, Al = Al2O3; PH = Phillips, KE = Kearny, PR = Pratt, BT = Barton).

Stratigrapgy	Fe(PH)	Fe(KE)	Fe(PR)	Fe(BT)	Al (PH)	Al (KE)	AL (PR)	Al (BT)	RMI (BT) (E-5 emu/g)
Surface Soil	2.82	2.98	2.55	2.21	10.45	10.76	10.34	10.15	0.30
Bignell Loess	2.79		2.25	1.92	10.94		9.43	9.39	0.40
Brady Soil	2.84		2.68	2.69	10.80		10.82	11.23	0.80
Peoria Loess	3.47	3.30	3.20	2.77	13.61	12.01	11.38	11.24	1.75
Gilman Soil	2.83		3.10	2.85	10.17		12.02	10.82	1.05
Sand I	2.59			2.42	11.50			9.94	0.75
Sangamon Soil	3.22		3.60	2.47	9.65		12.14	9.71	0.45
Sand II	2.74						12.16	9.72	0.55

the kaolinite/quartz ratio in clay, the quartz/feldspar ratio in fine sand, and the magnetic susceptibility, the weathering intensity is rather low in the Peoria Loess. Massive and rapid deposition of the Peoria Loess implies that the loess would not have been well weathered. There is no indication of soil-forming impact on the loess, and the high Fe_2O_3 and Al_2O_3 values must be attributable to the difference in loess sources.

Remnant magnetic intensity (RMI) was plotted against depth of the Barton County section (Figure 25). Peaks of RMI occurred both in the Peoria Loess and in the loess at a 12 m depth demonstrate inconsistency between susceptibility (S.I) and RMI. The inconsistency could be caused by the difference in size of the magnetic particles and/or by the difference of magnetic minerals in the parent material (Carmichael, 1989).

As indicated earlier, as a major contributor of the magnetic susceptibility, Fe_2O_3 explains most of the variations of the magnetic susceptibility (S.I) in the Barton County section ($r = 0.844$; Y [S.I] = 1.714 + 0.1467X [Fe_2O_3]). By calculating the difference between the predicted value (percentage) of the Fe_2O_3 (based on the linear relationship between the S.I and the Fe_2O_3 percentage) and the actual Fe_2O_3 percentage at each data point, the goodness of the fit between these two is

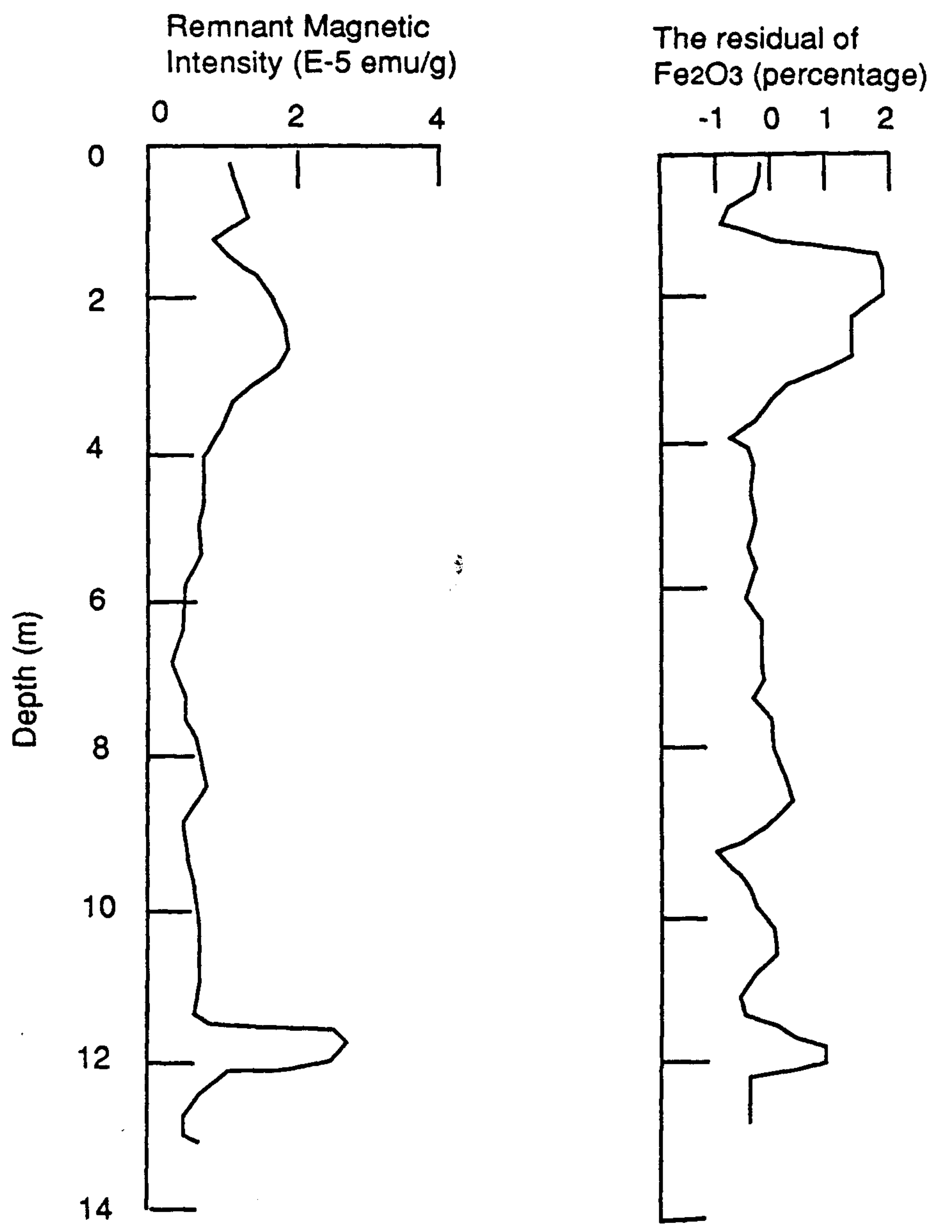


Figure 25. The correspondence of the remnant magnetic intensity to the residual of Fe_2O_3 (based on the correlation between the magnetic susceptibility and the Fe_2O_3 content) at the Barton County section.

plotted: the residual or error term (the difference between the predicted and actual values) was plotted against depth of the Barton County section. A good correlation between the Fe_2O_3 and the susceptibility means that the Fe_2O_3 is accumulated or formed by weathering processes, while a poor correlation, unexpected higher actual Fe_2O_3 than the predicted Fe_2O_3, means that the high Fe_2O_3 is not attributable to the weathering processes. Figure 25 also shows that the residual is extremely high in the Peoria Loess and in the Loveland loess around a 12-m depth.

The RMI curve almost exactly matches the Fe_2O_3 error distribution. The inconsistency between the S.I and RMI and the correlation between the RMI and Fe_2O_3 suggest that the properties of the parent material of the Peoria Loess and the loess at a 12-m depth could be different from those of other stratigraphic units. In other words, the RMI is primarily a magnetic property preserved in parent material of the loess, not a product of weathering processes. It can be said that the Fe_2O_3 which can be explained by the magnetic susceptibility is attributable to the weathering processes, while the Fe_2O_3 which can be explained by the RMI is attributable to the geologic source.

Based on evidence discussed above, one can speculate that the Peoria Loess might have originated from a poorly weathered and Fe_2O_3- and Al_2O_3-enriched source. The loess might not have been well pre-weathered or post-altered. The original parent material might have been derived from Fe_2O_3- and Al_2O_3-enriched igneous rocks in the Rocky Mountains, other than from Fe_2O_3- and Al_2O_3-depleted sedimentary rocks in the Southern Great plains. This genetic association between the remnant magnetic intensity and the Fe_2O_3 content can explain why the Fe_2O_3/Al_2O_3 ratio (oxidization index) does not record the history of weathering at the Pratt County section.

CHAPTER V. GLOBAL CLIMATIC COMPARISON

Investigation of Quaternary glaciation in many continents, particularly in the Alps region of Europe, initiated our interest in Pleistocene climatic and environmental changes (Van Husen, 1984); however, due to disconformity and lack of chronology in terrestrial Quaternary deposits, Quaternary studies have been dominated by deep-sea studies since the late 1960s. Based on an oxygen isotopic analysis of deep-sea core, Shackleton (1967) provided a reliable assessment of continuous paleotemperature changes. He and his colleague's persistent efforts (Shackleton and Opdyke, 1973) also resolved the problems in dating stratigraphy of the deep-sea sediments. Equally important, ice-core studies in north and south polar areas contributed high-resolution information on climatic change during the late Quaternary. Rsearch on deep sea and ice cores articulated the Milankovitch's theory of climatic change and led to a symposium on Milankovitch and Climate (Berger et al., 1984).

Using deep-sea stratigraphy as a correlation tool, Kukla (1970, 1977) reaffirmed his early findings on the existence of cycles of loess deposition in central Europe (Kukla, 1961). Although Kukla's work on terrestrial

Quaternary deposits was highly respected, terrestrial Quaternary studies have been overshadowed by deep-sea Quaternary studies for almost three decades. Only recently, the Quaternary community has started to realize that terrestrial Quaternary studies deserves much more attention than it has received (Kukla, 1989).

Climatic Sequence of Loess Deposition in Central Kansas

There were four carbonate concentration cycles during the Loveland time (Figure 26). Based on radiometric dating data and the depositional rate, the carbonate peaks appeared approximately during 95-130, 200-250, 290-330 and 360-410 ka. As discussed earlier, the magnetic susceptibility and the Fe_2O_3/Al_2O_3 ratio have a reverse relationship with the carbonate in the Loveland Loess, suggesting that the carbonate-enriched layers were formed during the periods when the weathering was weak. Comparing the carbonate curve with the $\delta^{18}O$ curve of the Pacific V28-238 core demonstrates that the carbonate-enriched layers were formed during warm periods. The rate of deposition was very low (0.034 mm/yr) during the period of 95-415 ka.

Figure 26 also shows that from 95 to 70 ka a dune sand unit (Barton sand) was deposited at a very high rate of

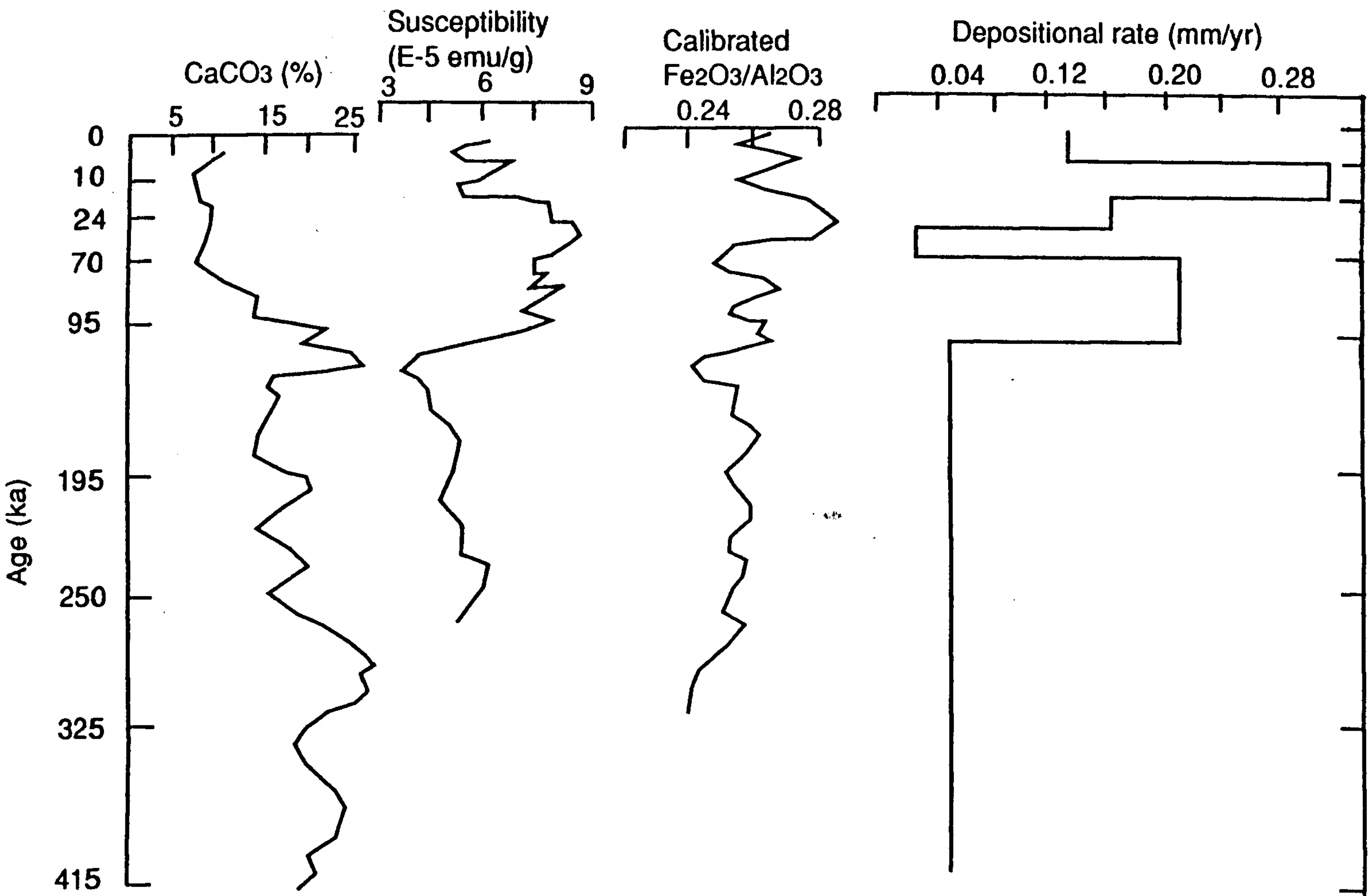


Figure 26. The temporal sequences of $CaCO_3$, susceptibility, calibrated Fe_2O_3/Al_2O_3 and depositional rate of loess at the Barton County section.

deposition (0.22 mm/yr). During this period, three weathering episodes were expressed by three minor peaks in the magnetic susceptibility and the Fe_2O_3/Al_2O_3 ratio. Subsequently, a strongly weathered reddish soil, the Sangamon pedocomplex,developed from 70-31 ka under relatively warm and moist conditions at a very slow depositional rate (only 0.016 mm/yr). Notably, a sandy unit (Sandy Silt I) overlies the reddish Sangamon pedocomplex and underlies the Gilman Canyon pedocomplex in the Phillips County section. The Sandy Silt I, possibly equivalent to the Roxana silt from 30 to 40 ka, even to 50 ka in Iowa and Illinois (Forman et al., 1991; Leigh, 1991; McKay, 1979), suggests that the transitional period from Sangamon to Gilman Canyon times was relatively windy and dry. The Gilman Canyon pedocomplex was formed from about 31 to 20 ka. The pedocomplex is characterized by a high concentration of organic matter, high leaching index, and relatively high depositional rate (0.15 mm/yr), implying that this pedocomplex was formed under cool and relatively moist conditions with a moderate rate of silt (dust) deposition.

Peoria Loess was deposited from 20.5 to 10.5 ka in central Kansas at a rate of 0.3 mm/yr and in central Nebraska at a rate of 3 mm/yr (Martin and Dort, 1987). The early Holocene, i.e., Brady time (10.5-8.5 ka), was characterized by a strongly expressed (both chemically and physically) soil, the Brady Soil. The Brady Soil is better developed than any soils developed since, suggesting that the early Holocene was the optimal time for soil development during the Holocene. The available data suggest that the Bignell Loess was deposited during the Altithermal time, i.e., the dry mid-Holocene.

Sand sequences in the Great Bend Sand Prairie demonstrate that the sand dune activity persisted from the end of the Brady time, about 8,500 ka, to 5,8 00 ka when a soil developed. After a soil-forming event from about 5,800 to 4,100 ka, sand dune activity dominated with four hiatuses: around 2,900, 1,600, 800, and the time when the surface soil developed. Surface blow sand is burying the surface soil, indicating recent deterioration of the environment.

Long-term Comparison (400 ky)

Records of deep-Sea

Since Shackleton's (1967) publication, $\delta^{18}O$ curves from numerous deep-sea cores around the world have been chronologically correlated, and the correlations demonstrate that the climatic information from many localities are comparable (Cline and Hays, 1976). The combination of oxygen isotope and paleomagnetic stratigraphy in the Pacific Ocean cores V28-238

(Shackleton and Opdyke, 1973) and V28-239 (Shackleton and Opdyke, 1976) has provided an excellent framework within which to investigate the history of events in the Pacific during the past 7 million years and to correlate this history with events elsewhere.

The records of changes in the oxygen-isotope composition of the world's oceans have been used as a stratigraphic tool in correlating the Pleistocene deep-sea sediments of all oceans (Hays et al., 1976; Oba, 1969; Shackleton, 1967; Van et al., 1969). Moreover, since the primary mechanism giving rise to these changes is the growth and the retreat of continental ice sheets, the deep-sea records are of considerable value as a basic stratigraphic reference to which the less continuous Pleistocene records from the continents could be effectively correlated (e.g., An Zhisheng and Liu Tungsheng, 1987; Kukla, 1970; 1977; 1989).

The chronological framework of deep-sea stratigraphy is based mainly on the magnetic stratigraphy with consideration of a limited number of radiometric dating data, but high-resolution age control still relies on the assumption that the depositional rate during a certain period was constant (Williams et al., 1988).

Radiometric dating data from the loess sequences of central Kansas were compared with the chronology of

Pacific core V28-238 (Table 13). TL and ^{14}C ages of unique stratigraphic units in the Kansas loess deposits marked important climatic events in comparison with the chronology of oxygen isotope of the Pacific V28-238 core and the chronology of the Chinese loess sequence. The carbonate-enriched layers in central Kansas are comparable with the carbonate-depleted soils in Chinese loess and occurred during warm stages. The sand unit (Barton sand) corresponds to oxygen stage 4 in the Pacific and to the Loess unit I (L1) in China. The reddish Sangamon pedocomplex corresponds to oxygen stage 3 in the Pacific and Soil unit 0 (S0) of Chinese loess. Both the Gilman Canyon pedocomplex and the Peoria Loess are correspond chronologically to oxygen isotop stage 2. The Brady Soil associates with stage 1.

The pattern of changes in $\delta^{18}O$ of Pacific core V28-238 (Figure 27a) is well reflected by the changes of carbonate concentration of the loess section in central Kansas (Figure 27b), i.e., carbonate peaks correspond with peaks of $\delta^{18}O$, suggesting that the carbonate-enriched layers were formed during the interglacial stages (5, 7, 9, 11 stages of oxygen isotope). An exception is that, however, the carbonate of the Holocene soil and loess is not so high as it should be in comparison with the $\delta^{18}O$ peak in the Pacific core.

Table 13. Comparison of the ages of the lower boundary of the oxygen-18 stages in the Pacific V28-238 Core (Shackleton and Opdyke, 1973) with those of loess sequences in central Kansas and China (An and Liu, 1987; Kukla, 1989). Note: S: soil, Lc: carbonate-enriched layer, Ld: carbonate-depleted layer; L: Chinese loess.

V28-238 Core of the Pacific		Loess of central Kansas		Loess of China	
Oxygen-18 stage	Age (ka)	Loess-soil Depth (m)	Age (ka)	Loess-soil	Age (ka)
1	13	1.2(S)	10	S0	10
2	30	3.4(S)	30		
3	65	4.2(S)	70	L1	70
4	75				
5a	87				
5b	92	6.6(Lc1)	95		
5c	110				
5d	118				
5e	128			S1	140
6	185	10.0(Ld1)	195	L2	200
7a	215				
7b	230				
7c	250	12.5(Lc2)	250	S2	250
8	275	(Ld2)		L3	290
9	330	15.0(Lc3)	325	S3	330
10	360	(Ld3)		L4	350
11	420	18.0(Lc4)	415	S4	410

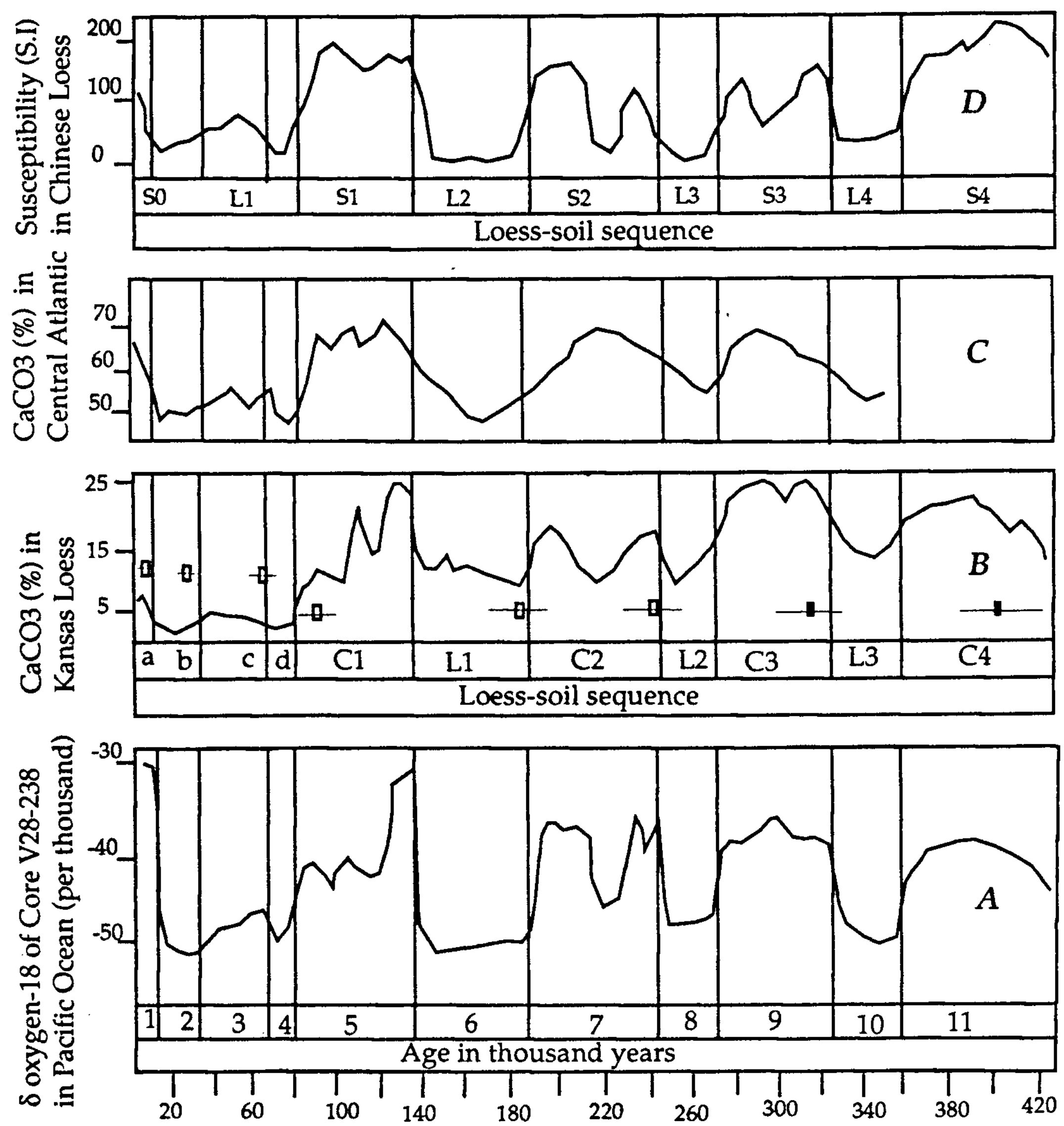


Figure 27. The comparison of δ oxygen-18 curve from the Pacific V28-238 Core (A) with the carbonate concentration of Kansas loess (B), the magnetic susceptibility of Chinese loess (C), and the carbonate concentration of central Atlantic Ocean (Chinese loess: L=loess, S=soil. Kansas loess: a=Holocene sequence, b=Peoria loess and Gilman Canyon Soil Complex, c=Sangamon Soil, d=Sand unit, L=carbonate-depleted Loveland Loess, C=carbonate-enriched Loveland Loess; ▯ age position, ┿ inferred age position).

The $\delta^{18}O$ curve does not indicate a strong signal in stage 3 (interstadial stage dated about 30-70 ka), implying the the temperature of that time was not as high as that of other warm stages. The carbonate curve in central Kansas also does not give a strong signal to express the warm stage, but the well-weathered Sangamon pedocomplex dated at 70 -31 ka manifests the warm stage. The well-weathered pedocomplex was due to relatively warm and moist soil-forming conditions and to a very slow rate of loess deposition. Here, time (duration) and climate (warmth and moisture) were balanced with each other, i.e., a well-weathered soil can form either under a warmer and moister climatic condition for a shorter period or under a cooler and drier condition for a longer period.

In the Atlantic Ocean, analyses of foraminiferal assemblages and $\delta^{18}O$ (Gardner and Hays, 1976; Keigwin and Jones, 1989; Ruddiman and McIntyre, 1976, 1981; Van Donk, 1976) demonstrate the synchroneity of climatic changes between the Pacific and the Atlantic Oceans. In terms of eolian deposits and carbonate accumulation in the seas, different localities have different histories. In lower latitudes of the Pacific Ocean, the eolian depositional rate was high during interglacial stages due to expansion of the subtropical deserts (Rea and Leinen, 1986; Rea and

Leinen, 1988; Rea et al., 1985), whereas in higher latitudes of the Pacific Ocean, the eolian depositional rate was high during glacial stages because of violent dust storm activity in north China (An and Liu, 1987). Both higher and lower latitudes of the Atlantic Ocean have higher carbonate concentration during interglacial stages, but for quite different reasons. High carbonate in higher latitudes was due to a lower terrestrial particle contribution to the ocean (Keigwin and Jones, 1989), whereas high carbonate in lower latitudes was due to stronger evaporation under the dominant influence of intertropical high pressure during interglacial stages (Ruddiman and McIntyre, 1976).

As Gardner and Hays (1976) stated, carbonate concentration in the equatorial Atlantic reflects the effectiveness of evaporation, which is controlled by both the temperature and the precipitation. During interglacial stages, warm temperatures expanded the Intertropical Convergence Zone (ITCZ), leading to high evaporation in seas of the lower latitudes. Carbonate of the loess sequence in central Kansas matches the carbonate of deep sea core sequences in the equatorial Atlantic Ocean (Figure 27c). As discussed earlier, the carbonate concentration in central Kansas is pedogenically formed; the soil-forming conditions during periods of carbonate concentration were similar to those of today in northern Texas and northern New Mexico with regard to effective soil moisture. If this was the case, the carbonate concentration in central Kansas, actually in the Central Great Plains, was probably also due to the northward expansion of the ITCZ during the interglacial stage. or/and to northward displacement of westerly belt, which has been a major source of moisture for the midcontinental United States.

In higher latitudes, the temperature of the sea surface is the only factor affecting the foraminiferal assemblage; therefore, the foraminiferal assemblage in the seas of higher latitudes could function as a proxy of the temperature (Ruddiman and McIntyre, 1976). Ruddiman and McIntyre's work (1976) demonstrated that in the region 45-55°N in the north Atlantic Ocean latitudinal shifts and percentage variations of polar and subpolar faunae reflect the glacial-interglacial cycles rather sensitively, being coincident with $\delta^{18}O$ cycles. The foraminferal assemblages, indicators of temperature, further confirm that the carbonate concentration in central Kansas occurred during warm periods.

Records from Other Continents

Kukla (1970) correlated the loess sequence of central Europe with deep-sea stratigraphy. With development of dating techniques, this correlation has been refined (Kukla, 1977; 1978; 1987). Kukla found that the loess of central Europe was deposited during glacial stages, while forest soils in the loess sequence were formed during interglacial stages. Not only are the last glaciation (oxygen isotope stage 2) and the last interglaciation (oxygen isotope stage 5) well expressed in the loess sequence, but also small oscillations such as oxygen isotope stages 3 and 4 are apparently articulated in the loess sequence.

Based on magnetic stratigraphy, plus other radiometric dating data, Chinese loess is chronologically correlated with the deep-sea stratigraphy (An and Liu, 1987; Burbank and Li, 1985; Kukla et al., 1988; Liu and Han, 1986; Liu and Yuan, 1987; Wang et al., 1985). The correlation shows that soils formed in deciduous forests and tall grass occurring within the Chinese loess sequences developed during interglacial stages, whereas the poorly weathered loess units were deposited under short grass at quite high depositional rates during glacial stages.

Wen and Sun (1981) found that, due to strong leaching and weathering during warm periods (soil-forming periods), carbonate concentration in the paleosols (only 4%) is much lower than in loess units (about 12%). Apparently the carbonate curve of Chinese loess is contrary to that of central Kansas loess, i.e., peaks of carbonate in Chinese loess correspond to glacial stages, while carbonate peaks of central-Kansas loess correspond to interglacial stages. Undoubtedly, paleosols in the Chinese loess sequence were formed during interglacial stages when the Pacific monsoon was much stronger and invaded northward to bring more precipitation to the Loess Plateau of north China (Feng et al., 1985; Li and Feng, 1988).

The curve of magnetic susceptibility from Chinese loess sequences exhibits an excellent correspondence with the $\delta^{18}O$ curve of the Pacific core V28-238 (Kukla et al., 1989). As a proxy of paleoclimate (warmth and moisture), the magnetic susceptibility indicates that interglacial periods were times when paleosols formed under much better vegetational conditions, whereas the loess units were always deposited under poorer vegetational conditions during glacial periods. Figure 27d shows that the strongest weathering of Chinese loess since 400 ka might have occurred during oxygen isotope stage 5 (75-110 ka) and 11 (350-410 ka), being coincident with the two highest peaks of carbonate in loess sequence of central Kansas (Figure 27b). As with the records from loess of central Kansas and $\delta^{18}O$ of the Pacific core, stage 3 (30-70 ka) is not well expressed. Unlike $\delta^{18}O$ of the Pacific core, the susceptibility of Chinese loess does not express the

Holocene soil well and neither does the carbonate of the Holocene soil and loess in central Kansas.

As with evidence from other localities around the world, the magnetic susceptibility of Chinese loess confirms that the oxygen isotope stage 3 was a minor climatic warming event and left a minimal imprint of soil development. Not surprisingly, this stage is always grouped into the last glaciation. To the contrary, the reddish paleosol (Sangamon pedocomplex) was strongly developed during this period; under relatively warm and moist conditions at a slow rate of deposition in the midcontinental United States.

Medium-term Comparison (100 ky)

Records from Ice Cores

Climatic changes during the past 100 ky have been relatively well investigated. Based on studies of terrestrial deposits, deep-sea and ice cores, Quaternary geologists now have confidence in reconstruction of the paleoclimate for the past 100 ky (Nilsson, 1983). As a consequence of these intensive studies, Goudie (1983) was able to compile global paleoclimatic information of the past 100 ky. Through spatial comparisons, Goudie found that, although the climatic changes might have been expressed in such different ways as temperature, moisture, glacial regime, and so on in different geographic locations on the earth, the climatic changes during the past 100 ky have very good comparability through almost all continents and seas, such as Greenland, Atlantic, Caribbean, western and central Europe, Siberia region, North America, and Middle East (Goudie, 1983).

Ice core studies are becoming increasingly important in reconstructing the paleoclimate of the late Quaternary because the studies can provide sensitive climatic information (e.g., $\delta^{18}O$ and dust content) and a reliable chronological control (year-layer counting). It has been proven that the climatic information for the last 100 ky from the Antarctic ice sheets agrees with that from the Greenland ice sheet: they both have changed simultaneously during the past 100 ky (Dansgaard, 1987; Jouzel et al., 1989).

Data from ice cores of Greenland show that during the last glaciation(120-10 ka), three negative $\delta^{18}O$ peaks appeared around 115-105, 75-65, and 30-18 ka (Figure 28a). The latter two cold peaks (75-65 ka and 30-18 ka) in the Greenland ice core are mirrored by two major glacial advances in North America (65-55 and 22-17 ka) with some lag. Apparently the $\delta^{18}O$ variations ranged within 5‰(-40 to -35‰) during the last glaciation except at the very end of the last glaciation from 14 to 9 ka. From 14-9 ka there were two abrupt warming periods (13 ka and 11 ka)

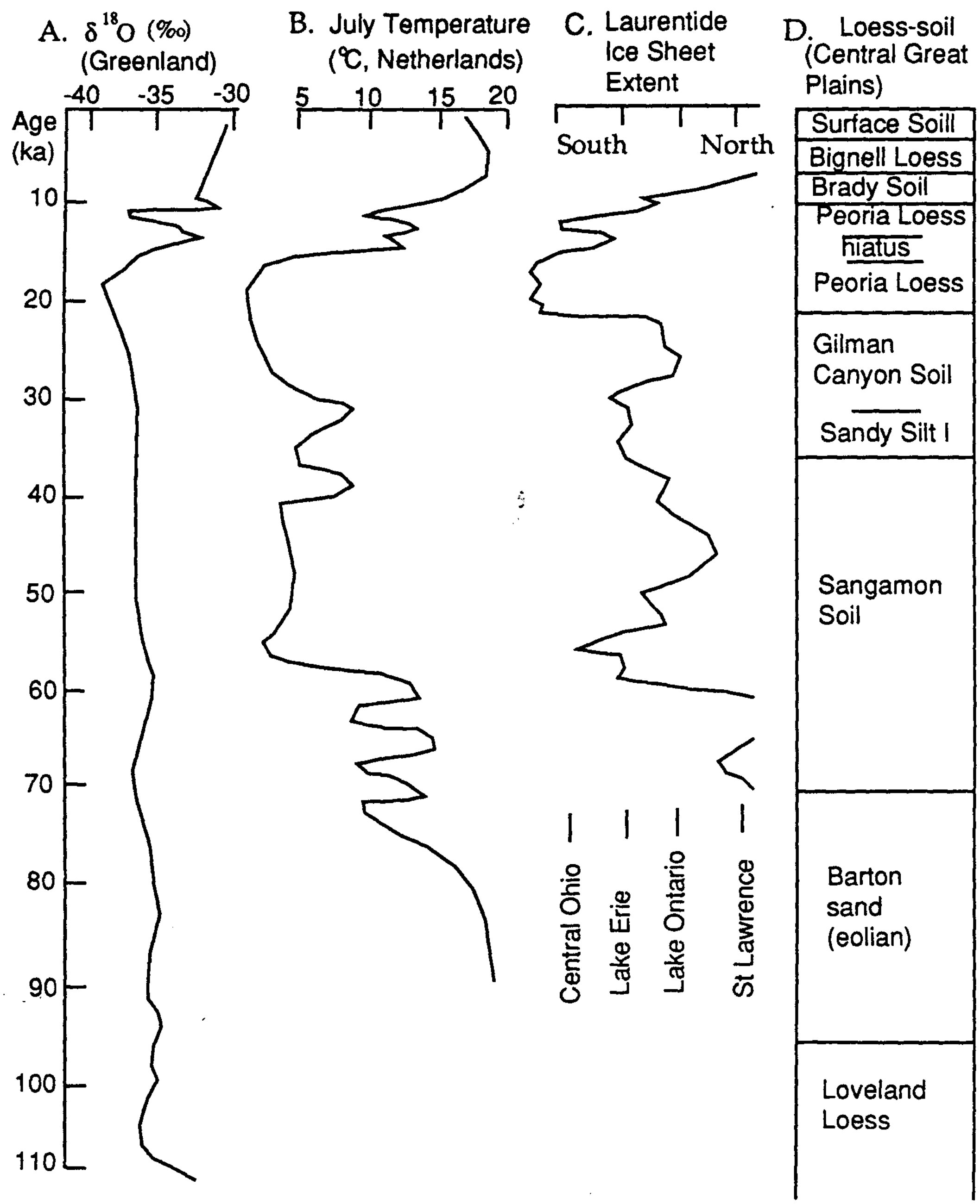


Figure 28. The comparison between the oxygen-18 curve of Greenland ice core (Dansgard et al., 1982), average July temperature of the Netherlands (Nilsson, 1983), southern extent of the Laurentide ice sheet (Wright, 1984) and loess sequence in the Central Great Plains.

interrupted by a cold period. After jumping up and down from 10750 to 9500 yr BP, the temperature started rising persistently to reach the typical Holocene regime around 9000-8500 yr BP (Patterson and Hammer, 1987).

Terrestrial Sequences

The average July temperature reconstructed from beetle fauna and other geological evidence in the Netherlands (Coope, 1975; Nilsson, 1983) was popularly accepted as a continental climatic reference of the late-Quaternary (Figure 28b). Seemingly, the temperature did not experience an abrupt decline until 75 ka. From 75 to 62 ka, three warm and cool cycles occurred with a range from 10 to 15°C of temperature. After 62 ka, the temperature dropped drastically, reaching the point below 5°C for the first time. The average July temperature stayed about 5°C from 55 to 40 ka. Two small temperature peaks (up to 10°C) appeared from 40 to 30 ka with a drop (close to 5°C) in between.

The last glacial maximum started about 30 ka, approaching the climax around 20 ka. From 14 to 10 ka, the temperature rose quite dramatically, with two short periods of temperature decline. The temperature rose after 10 ka and reached the peak around 7.5 ka. The peak period from 7.5 to 5 ka (Goudie, 1983; Lamb, 1975)), the optimal time of the Holocene, is comparable with the warmest and wettest time of the Holocene in China (Chu, 1973). This optimal time corresponds with the Altithermal Period, a warm and dry period recognized in parts of the United States.

In North America, the southern margin of the Laurentide ice sheet reflected the temperature change during the last glaciation. Figure 28c shows that the ice sheet entered the St. Lawrence Lowlands (southeastern Canada) around 70 ka. After a short appearance, the ice retreated from the lowland and reentered the lowland about 60 ka, reaching central Ohio for the first time around 55 ka, the time when the average July temperature in the Netherlands first dropped below 5°C (Coope, 1975; Nilsson, 1983). Seemingly, the southern margin of the Laurentide ice sheet corresponded rather sensitively to the average July temperature of the Netherlands, with an exception at the latter part of the last glacial maximum.

The average July temperature of the Netherlands shows a rapid decline around 30-28 ka, while rapid southward invasion of the Laurentide ice sheet began around 22 ka, reaching its maximum (central Ohio) around 20 ka. When compared to the high frequency changes of $\delta^{18}O$ in the Greenland ice core toward the end of the last glaciation, the southern margin of the Laurentide ice sheet appeared to be more sensitive than the average July temperature of the Netherlands (Wright, 1984). For example, the ice retreated during the two warm periods of 13,000-11,500 and 10,750-9550 yr BP and advanced during the cold period in between (11,500-10,750 yr BP).

The early Wisconsin formation (i.e., Barton sand) which formed at 95-70 ka, reflects warm and dry conditions, comparable with the higher temperatures in the Netherlands. The Sangamon pedocomplex, dated at 70-31 ka, corresponds with a moderate temperature regime. As the Sangamon pedocomplex was forming (Figure 28d), the Laurentide ice sheet stayed primarily in northern Canada without invading the United States. The well-weathered Sangamon pedocomplex was developed under a moderate climatic condition (relatively warm and moist) and a very slow rate of loess deposition.

Although the temperature of the Netherlands started its regime of the last glacial maximum at about 28 ka, the Laurentide ice sheet reached its maximum only about 22-20 ka. The lag between average July temperature of the Netherlands (30-28 ka) and the advance of the Laurentide ice sheet (22-20 ka) was the time when the Gilman Canyon pedocomplex was formed (20-31 ka) in the midcontinental United States. Since the pedocomplex, characterized by a thick cumulative organic-enriched A horizon, required moist conditions, the Laurentide ice sheet would have invaded southward if the temperature were as low as in the Netherlands. Unlike the Netherlands where the temperature during 30-20 ka was as low as during 20-13 ka, the period from 30-28 to 22-20 ka when the Gilman Canyon pedocomplex developed could have been relatively warm, i.e., cool rather than cold in the midcontinental United States. This interpretation is supported by foraminifera evidence from the middle northern Atlantic Ocean (45-55°N). It has been demonstrated that sea surface temperature remained as warm as today's from 75 to 22 ka (Flohn, 1983; Kipp, 1976) or at least apparently warmer than the last glacial maximum (Ruddiman and McIntyre, 1981), and that climates in the Pacific and Atlantic Oceans have changed simultaneously (Gardner and Hays, 1976; Keigwin and Jones, 1989). These facts imply that sea surface of the Pacific Ocean (a major source of vapor for the United States) might also have remained relatively warm from 75 to 22 ka. It was likely that in North America the temperature was relatively low (cool) due to the effect of the ice sheet in the lands of high latitudes and the precipitation was relatively high due to the effect of warm sea surface. Therefore, a cool and moist model for the Gilman Canyon time of the midcontinental United States could be an acceptable one.

The Peoria Loess was deposited during the period 20-10.5 ka after the Laurentide ice sheet reached its maximum. The Brady Soil developed during the warming period immediately after the last glaciation (10.5-8.5). Following the Brady was Bignell Loess deposition during the Altithermal Period (warm and dry). These comparisons seem to indicate that soils, such as the Sangamon pedocomplex, the Gilman Canyon pedocomplex, and the Brady Soil, were formed exclusively during moderate temperatures. Probably the moderate temperature regime was the key to relatively high soil moisture. In term of the depositional conditions, the Bignell Loess and Peoria Loess belong to two categories: cold loess (Peoria) and warm loess (Bignell).

Short-term Comparison (18 ky)

Last Glacial Maximum

The focus is on events that occurred during the last glacial maximum. The most striking fact is that the dust content in the Greenland ice core was as high as 2 mg/kg from 18 to 13 ka, three times higher than that from 70-18 ka (only 0.7 mg/kg) and 10 times higher than that of the Holocene (Patterson and Hammer, 1987). The high dust content of the ice core is in accord with the fact that the rate of loess deposition was highest from 19.5 to 13 ka in south-central Nebraska, 3 mm/yr (Martin and Dort, 1987) and in central Kansas, 0.3 mm/yr (Feng et al., 1991). Martin (1990) indicated that Peoria Loess deposition started about 20 ka and was interrupted by major river entrenchment (possibly indicative of moist conditions) about 13 ka along the Republican River. The Peoria Loess resumed deposition not long after river entrenchment. Keigwin and Jones (1989) showed that in the western North Atlantic, the rate of terrestrial particle contribution was relatively low during the Holocene and the rate was as high as 2 mm/yr during the last glacial maximum (22-11 ka). Terrestrial particle content in the North Atlantic deep sea core and dust content in the Greenland ice cores suggest that the period from 18-13 ka was the dustiest time during the past 100 ky. Likewise, 18-13 ka was also the dustiest time in North America. As discussed earlier, this period was typified by a dry and cold climate in North America and was coincident with the cold-arid phase (18-13 ka) in Asian-African monsoon regions (Fohn, 1983).

It has been shown that modern dust storms in the midcontinent bring dust to the Atlantic and Greenland (Jackson et al., 1973; McCauley et al., 1981). The above comparisons demonstrate that the rate of loess deposition in the midcontinental United States coincides chronologically with the rates of dust deposition in the North Atlantic Ocean and Greenland during the Peoria time, the last glacial maximum. These facts suggest that the midcontinental United States might have functioned as an important source of dust to the North Atlantic Ocean and Greenland.

Pleistocene-Holocene Transition

Towards the end of the last glaciation, dust concentration and $\delta^{18}O$ in the Greenland ice core (Hammer et al., 1985) display an excellent comparison with the $d^{18}O$ record of Switzerland lake sediments (Siegenthaler et al., 1984). The comparison shows that, after a slight warming trend, linear in nature, from 18 to 14 ka, the temperature increased abruptly around 13 ka both in Greenland and Switzerland and was accompanied by an abrupt drop in dust content in the Greenland ice core (the late glacial interstadial, or Bolling-Allerod Period). Accretionary soils dated at about 13 ka in Nebraska and in Kansas corresponded to the river entrenchment (Feng et al., 1991; May 1989). The Jules Soil (dated at 12 to 15 ka) developed within the Peoria Loess in the Central Lowlands (Frye et al., 1974) strongly suggests that moist, probably warm, conditions existed in the midcontinental United States around 13 ka. Cold conditions resumed from 11.5 to 10.75 ka in both Greenland and Switzerland (the Younger Dryas stadial), with dust content abruptly increasing in the ice core (Patterson and Hammer, 1987). After rapid warming around 13 ka and sudden cooling around

11 ka, another rapid warming started at 10.75 ka and was followed by a short and slight cooling. The characteristic Holocene temperature regime was reached about 9000 yr BP (Figure 29a, b).

The climatic record in the midcontinental United States is comparable with those in Switzerland and Greenland. Further, the $\delta^{18}O$ curve of the Greenland ice core is nearly identical to the record of ice advances and retreats in North America (Wright, 1984). The period of low temperature from 18 to 13 ka, when most Peoria Loess deposition occurred, was reflected in stabilization of the Laurentide ice sheet in the Central Lowlands of the United States. Even the Allerod-Bolling late-glacial interstadial period was recorded by a dramatic retreat of the ice. The late-glacial stadial period (Younger Dryas period) was documented by a rapid advance of the ice (Figure 29c). Well-documented macrofossils (mainly spruce charcoal) concentrated in the upper part of the Peoria Loess in central and western Kansas (Wells and Stewart, 1987) and southern Nebraska (May, 1989) were dated at about 14 ka. Organic matter-enriched sediment (peaty accretionary soil) in the sand of the Sandhills in Nebraska (Swinehart, 1990) and organic matter-enriched accretionary soil in the sand of the Great Bend Sand Prairies (Johnson and Sophocleous, 1991) were dated at 13.6 ka. As stated earlier, the Peoria

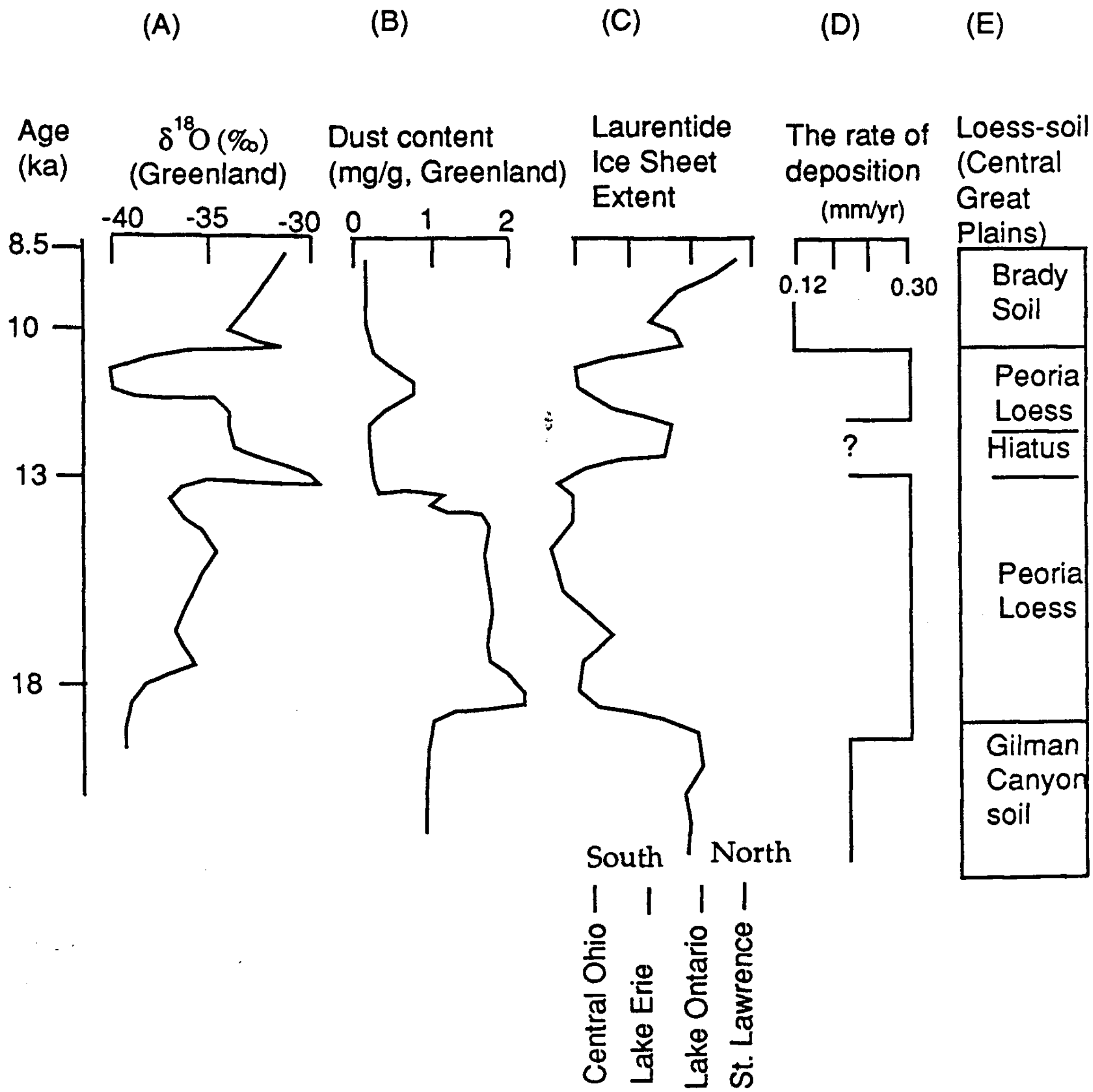


Figure 29. The comparison of the oxygen-18 curve (A) and dust content (B) in Greenland ice core (Patterson and Hammer, 1987) with the extent (C) of the Laurentide ice sheet (Wright, 1984), and loess deposition rate (D) and loess sequence (E) in the Central Great Plains.

loess was interrupted by major river entrenchment around 13 ka in the Republican River system (Martin, 1990). More important, the Jules soil was found at many localities in the Central Lowlands (Frye et al., 1974).

All of this evidence seems to support the argument that a moist (perhaps somewhat warm) climate prevailed around 13 ka in the Central Great Plains. With consideration of the accuracy level of the radiocarbon dating, it is legitimate to assume that the river entrenchment and the spruce tree flourishing in the Central Great Plains were chronologically comparable with the warm period from around 13 ka in the Greenland ice core and Switzerland lake sediment. This interruption of Peoria Loess deposition and major spruce tree growth around 13 ka in the Central Great Plains could be the reason why dust content in the Greenland ice core dropped significantly during the late glacial stadial period (Figure 29b, d, e). The resumption of Peoria Loess deposition in the Central Great Plains corresponds to the Younger Dryas stadial period occurring around 11.5 ka in Greenland. Another piece of evidence to support the augument for two stages of Peoria loess deposition (before and after the 13 ka warm period) is that the clay content of the upper Peoria Loess is unexpectedly high in both the Pratt County and Barton County sections. The increased clay content of the upper Peoria Loess suggests that the importance of the clayey source of the loess became greater during late Peoria time, i.e., more clayey material was brought into this area from the Southern Great Plains. The Brady Soil is definitely corresponding to the early Holocene warming started at 10,750 yr BP. These comparisons suggest that frequent fluctuations toward the end of the last glaciation were global, at least Northern Hemispherical phenomena.

Holocene

In the Northern Great Plains, the transitional stage from the last glaciation to the Holocene started about 12,200 yr BP (Barnosky, 1989). At that time, coniferous trees were replaced by a temperate grassland with the temperate grassland lasting until 9300 yr BP. A xeric grassland persisted from 9300 to 6000 yr BP. After 6000 yr BP, a cooler and moister condition dominated this region.

In the Southern Great Plains (Holliday, 1989a), eolian sedimentation began at least locally between 10,000 and 9000 yr BP. It was episodic but widespread from 9000 until 5500 yr BP, with most of the area affected by 6500 yr BP. Between 5500 and 4500 yr BP, eolian sedimentation occurred at all locations.

In sand sequences of the Great Bend Sand Prairies of central Kansas, after the Brady Soil, no soil had formed until 5800 yr BP. The soil event lasted from 5800 to 4100 yr BP in the Sand Prairies. After this soil-forming event, sand dunes were reactivated and persisted until today with four hiatuses: around 2900, 1600, 800 yr BP and the time when the surface soil developed. The surface soil is buried by surface blow sand, indicating deterioration of the modern environment (Feng et al., 1991; Johnson and Sophocleous, 1991). The chronological sequence of sand-dune activity and soil formation (sand dune stabilization) is supported by the sand-dune activity and soil formation sequences in the Sandhills of Nebraska (Swinehart and Loope, 1991) and Northeastern Colorado (Yuhas, 1991).

The Altithermal Period was from 9300 to 6000 yr BP in the Northern Great Plains and from 10,000-9000 to 4500 yr BP in the Southern Great Plains. In the Central Great Plains, the Altithermal Period started about 8500 yr BP and lasted until 5800 yr BP. The maximum of the Altithermal Period occurred before 6000 yr BP in the Northern Great Plains and about 5500 yr BP in the Southern Great Plains. Based on soil development on river terraces in the Central Great Plains (Johnson and Martin, 1987), the maximum of the Altithermal Period might be around 6000 yr BP, at least before 5800 yr BP, in the Central Great Plains. These chronologies suggest that the duration of the Altithermal Period was shorter (started later and ended earlier) in the North and longer in the South. The maximum of the Altithermal Period has a time-transgressive behavior, being younger southward.

This dry and warm mid-Holocene period might have resulted from an increase in solar radiation from 12000 to 6000 yr BP (COHMAP Members, 1988). It is concluded that the increase of solar radiation from 12000 to 6000 yr BP led to warm and moist conditions in monsoonal regions, such as Africa (COHMAP Members, 1988; Goudie, 1983) and China (Chu, 1973; Feng, 1988c), during the early and middle Holocene. The last 6000 years have been characterized by generally declining temperature in all parts of the world. Many lines of evidence suggest that this general cooling trend was reversed in Europe and Africa at 5300, 2800, and 350 (Lamb, 1975).

As discussed earlier, the interglacial stages during Loveland time (95-415 ka) were characterized by strong weathering in China, but by weak weathering in the midcontinental United States, at least in the Central Great Plains. The reversal occurred during the glacial stages. Surprisingly, similar environmental contrasts between China and the midcontinental United States also existed for the Holocene. Geological and historical evidence shows that in northern China times of high trenperature were always accompanied by high precipitation during the Holocene (Chu, 1973; Feng and Thompson, 1988; Li and Feng, 1988; Zhang, 1984)). During these moist and warm periods of the Holocene, fluvial and loessial soils developed in the Loess Plateau of north China (Feng, 1988b). However, soil-forming periods in the Great Bend Sand Prairies correspond almost exactly to the cold (also dry) times in northern China (Figure 30). The Brady Soil of the Central Great Plains corresponds to cool conditions during the early Holocene within China. The major soil-forming event from 5800 to 4100 yr BP mirrored the lowering of temperature in northern China around 6000-4600 yr BP. The 2900 yr BP soil of the Great Bend Sand Prairie matches the cold episode of 3000 yr BP (Neoglaciation). The 1600 yr BP soil could be a counterpart of the cool period from 1900 to 1500 yr BP. More surprisingly, even the 800 yr BP soil has its temperature counterpart in northern China, i.e., 1100-1200 AD cold interval. The surface soil in the Sand Prairies might mirror the cool period from the 17th to 19th centuries (the little ice age). Warm conditions always seemed to result in more precipitation in China, but warm conditions always led to prolonged drought in the midcontinental United States.

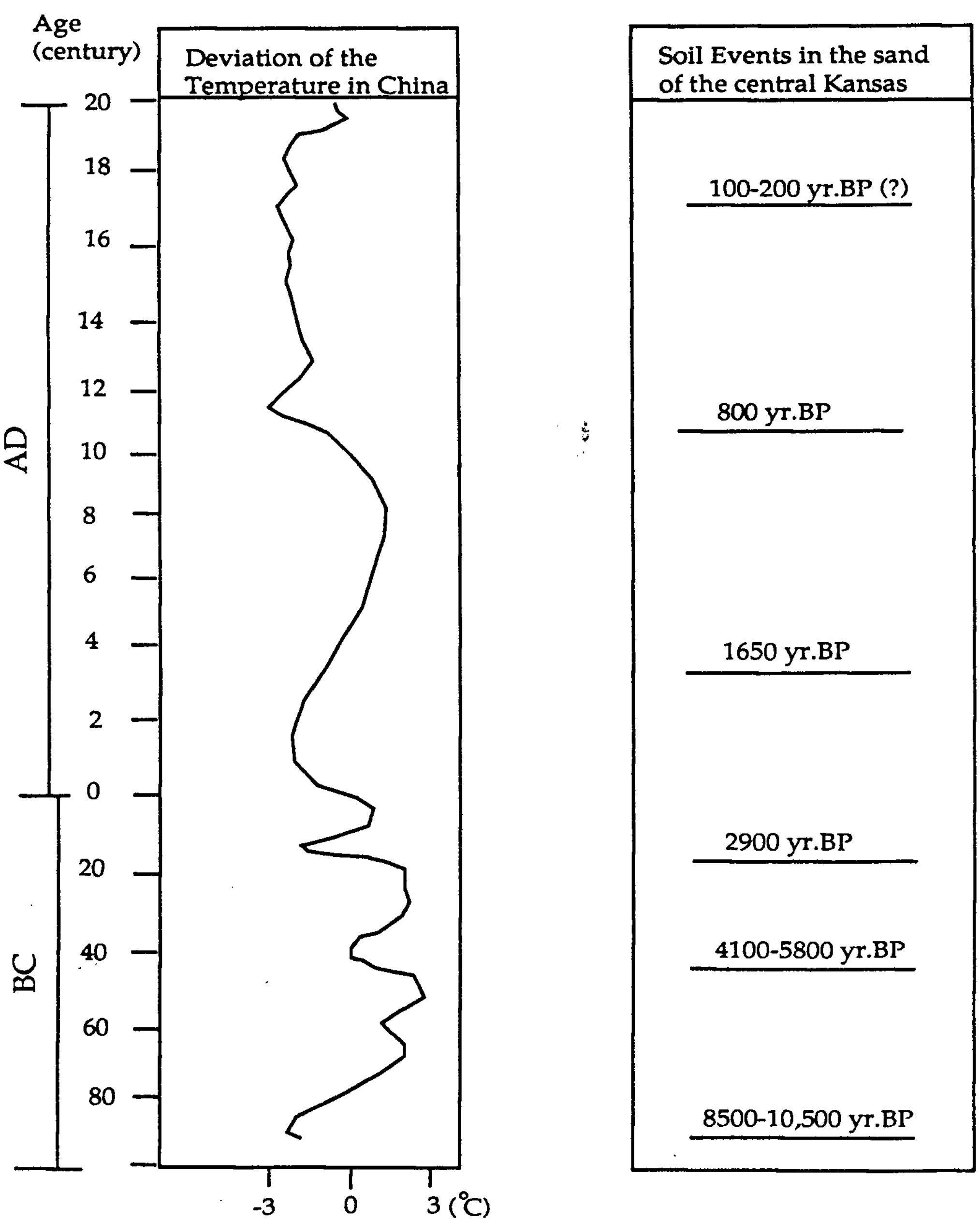


Figure 30. The comparison of the temperature deviation in China with the soil-forming events in the Great Bend Sand Prairie of central Kansas (notice scale change).

A Climatic Hypothesis of for the Late Pleistocene

This writer hypothesizes that when the temperature of the north-central Atlantic Ocean was high (as warm as today) and the average July temperature of northern Europe was low (nearly as cold as during the last glacial maximum) from 30 to 20.5 ka, the middle continental United States was a place where cold air masses from the Laurentide ice sheet and warm air masses from the oceans in the south interacted violently to produce abundant precipitation. At the same time, the anticyclonic system developed on the ice sheet brought a large quantity of silt into this relatively moist region. Consequently, the Gilman Canyon pedocomplex experienced a relatively high depositional rate and a high leaching index. Probably the active cyclonic systems also promoted the growth of the Laurentide ice sheet by bringing more moisture onto the ice sheet and accelerated the advance of the ice about 22 ka.

After the ice sheet exceeded a threshold, sea surface temperature was cooled down, the ice sheet started to retreat due to lack of water vapor supply from the oceans in the South. At that time, the anticyclonic systems on the ice sheet prevailed in the midcontinental United States. At this threshold point (about 20 ka), the anticyclonic systems moved a great deal of silt from the ice-free part of the Northern Great Plains and the northern portion of the Central Great Plains, and the silt (the Peoria Loess) was deposited in the Central Great Plains under dry and cold conditions.

Thomas (1977) hypothesized that when a drastic global warming started about 13 ka, cold water melted from the glacial ice cooled down ocean surface and reversed the global warming trend around 11 ka. The warming trend was not resumed until about 10,750 yr BP when the cooled ocean surface was warmed. This warming led to the disintegration of continental ice sheets by calving processes. This writer believes that the transitional time from the Pleistocene to the Holocene might be characterized by a warm ocean surface at lower latitudes and a cold land surface in higher latitudes due to the existence of ice sheets. This striking contrast of temperature between lower and higher latitudes led to a violent interaction between the dry cold air from the north and warm moist air from the south and produced a great deal of precipitation through active cyclonic systems in the middle latitudes. As a consequence of this interaction, soils developed extensively in middle latitudes, like in the Central Great Plains (Brady Soil).

CHAPTER VI. SUMMARY

The complete stratigraphic sequence in the study area consists of the Loveland Loess, Barton sand (eolian), Sangamon pedocomplex, Sandy Silt I (equvalent to Roxana Silt), Gilman Canyon pedocomplex, Peoria Loess, Brady Soil, and Bignell Loess. The Loveland Loess is characterized by alternating occurrences of carbonate-enriched layers and carbonate-depleted layers in central Kansas. The Barton sand, dated at 95-75 ka in the Barton County section, has its counterpart in the Phillips County section (Sandy Silt II). The Sangamon pedocomplex is strongly expressed in color and structure at the Barton County and Pratt County sections. Another sandy silt unit (Sandy Silt I) lies between the Gilman Canyon pedocomplex and the Sangamon pedocomplex at the Phillips County section. Sandy Silt I in the Phillips County section appears correlative with a similar sandy-silt unit at several localities in southern Nebraska (Reed and Dreeszen, 1965). This sandy silt is likely equivalent to the Roxana Silt in Iowa and Illinois (Frye and Willman, 1963).

TL dates from the Loveland Loess in central Kansas indicate that loess deposition started around 415 ka and ended about 95 ka, implying that the deposition of

Loveland Loess began with the beginning of the classically defined Illinoian stage and ended with the classically defined Sangamon stage (interglacial), which is presumed to have lasted from 127-125 to 75-70 ka. Based on TL data, pedogenic carbonate accumulation occurred four times during Loveland time: 415-325, 325-250, 250-195, 195-95 ka. The four carbonate peaks correspond chronologically to marine oxygen isotope stages 11, 9, 7, and 5, respectively.

Although Roxana Silt (Altonian substage) was thought to have been deposited during the early Wisconsin stage (Frye and Willman, 1963), both ^{14}C and TL ages confirm that the Roxana silt was deposited from 40 to 30 ka in Iowa and Illinois. The sand unit (Barton sand) dated at 95-70 ka at the Barton County section represents the early Wisconsin stage. The magnetic susceptibility and weathering indices show that there were three relatively strong weathering episodes in the sand unit, reflecting climatic fluctuations from 95 to 70 ka. Judging from overall weathering intensity, the slightly reddish dune sand must have been deposited under rather warm, dry and windy conditions, with a relatively high rate of deposition during the early Wisconsin stage.

The middle Wisconsin stage is represented by the classically defined Sangamon pedocomplex. The soil was strongly weathered. A chronological comparison of this soil with climatic information from the Laurentide ice sheet and from other localities around the world indicates that the Sangamon pedocomplex was formed under relatively warm and moist conditions and a very low rate of deposition.

TL dating (70 ka) of the reddish Sangamon pedocomplex, which is supported by dating (70 ka) in Iowa (Forman, 1990), resolved an outstanding issue: What happened from 75-70 (the end of the classic Sangamon time) to 35-28 ka (the beginning of the Gilman Canyon time)? The answer is that a soil was developing on a stable land surface under a moderate climate. The classically defined Aftonian substage (70-28 ka) of Illinois actually consists of two substages in Kansas: Barton substage and Sangamon substage (if the term Sangamon follows the reddish soil, not the classically-defined time scheme). The late-Wisconsin stage is represented by the Gilman Canyon and Peoria substages.

The Gilman Canyon Formation (pedocomplex), a chronostratigraphic equivalent to the Farmdale Silt in Illinois (28-22 ka), was deposited from 31-20 ka. The soil is characterized by a thick A horizon with high organic matter content, relatively low clay content, and strong physical weathering (leaching). Chemical weathering and mineralogical indices are relatively low compared with those of the Sangamon pedocomplex. The loess depositional rate during Gilman Canyon time was relatively high (0.15 mm/yr). High organic matter content and strong leaching, in conjunction with a relatively high deposition rate, suggest that the soil was formed under high effective moisture, contrary to the low effective moisture conditions proposed by Fredlund and others (1985).

The Peoria Loess is characterized by a very rapid rate of deposition, 0.3 mm/yr in central Kansas and 3 mm/yr in south-central Nebraska. The greatest rate of loess deposition during Peoria time is contemporaneous with the time of highest dust content in the Greenland ice core and the highest terrestrial particle level in the northern Atlantic sediments from 18 to 13 ka, when the $\delta^{18}O$ is the lowest. Magnetic susceptibility and the kaolinite/quartz ratio in clay indicate that weathering of the Peoria Loess was quite weak. There is no detectable impact of soil-formation on the Peoria Loess, e.g., no clay transformation and no carbonate accumulation in the calcareous loess. Well-preserved long grass stems in the form of charcoal in the loess suggest that the grasses were buried so rapidly that the charcoal pieces of grass were not oxidized or decomposed under cold and dry climatic conditions.

Toward the end of Peoria time, the climate started to fluctuate dramatically. The late-glacial interstadial interval around 13 ka, indicated by a peak of $\delta^{18}O$ in the Greenland ice core and lake sediment from Switzerland and a low level of dust in the Greenland ice core, is confirmed by a hiatus in Peoria Loess deposition in the Central Lowlands, dramatic river entrenchment along the Republican River in southern Nebraska, and well-developed accretionary soils in central Kansas and central Nebraska around 13 ka. During the transitional period from the Pleistocene to the Holocene, the Brady Soil was formed from approximately 10,500 to 8,500 yr BP in the Central Great Plains. This is obviously a climatically controlled soil: it developed under a relatively moist and warm climatic conditions. This soil is characterized by high organic matter content, high magnetic susceptibility, and high weathering indices, suggesting that this soil was not only strongly physically weathered, but also strongly chemically weathered.

During the Holocene, the Bignell Loess was deposited during the Altithermal Period (about 8,500-5,800 yr BP) in the Central Great Plains. After the Altithermal Period, five soil-forming episodes around 5,800-41,00, 2,900, 1,600, 800 yr BP, and the time of the surface soil development corresponded approximately to cool and dry episodes of China around 6,200-4,600, 3,000, 1,900-1,500 (1000-1500 AD), 700-800 yr BP (1100-1200 AD), and 100-200 yr BP (a cool period of 17-19th centuries).

The supply of silt for the Peoria Loess was provided primarily by northwesterly winds, probably from the Sandhills of Nebraska and the regions to the north and to west. Both dune sand in the Sandhills and the Peoria Loess south of the Sandhills likely originated from the same sources: fluvially-eroded river deposits and weathered surface material in the Northern Great Plains and the northern and western Central Great Plains. During late Peoria time the southwesterly winds were not dominant, but more important in contributing silt than before in the Central Great Plains. During Bignell time, southerly and southwesterly winds prevailed in bringing silt into the southern portion of the Central Great Plains, at least into central Kansas.

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Appendix 1a. Chemical composition of the loess at the Phillips County section of north-central Kansas.

Sam. (#)	Depth (m)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
PH0	-0.5	3.82	0.88	10.5	71.1	2.53	1.0	0.61	2.82	3.34
PH3	-1	3.49	0.79	10.9	71.4	2.48	0.9	0.62	2.79	3.57
PH01	-2	3.35	0.81	10.9	74.9	2.5	0.8	0.6	2.83	3.21
PH5	-3	3.60	1.54	13.6	68.0	2.73	2.0	0.58	3.47	3.02
PH6	-4	4.16	0.82	10.2	66.8	2.49	4.4	0.58	2.68	4.55
PH1	-5.5	3.58	0.87	11.5	72.2	2.61	1.0	0.62	2.83	3.19
PH7	-6.5	4.29	0.72	9.6	61.4	2.19	0.9	0.54	2.59	5.21
PH11	-8	4.37	0.99	10.5	65.5	2.47	3.7	0.58	3.22	4.64
PH13	-9.6	4.06	0.98	9.0	55.2	2.09	12.5	0.49	2.74	4.68
PH14	-14	4.27	0.85	9.5	64.9	2.12	10.5	0.45	2.85	3.21

Appendix 1b. Particle size distribution of the loess at the Phillips County section of north-central Kansas.

Sam. (#)	Depth (m)	50μ (%)	30μ (%)	20μ (%)	10μ (%)	6μ (%)	2μ (%)	1.9μ (%)
PH2	-0.5		0.2	4.0	10.2	65.2	16.5	0.5
PH3	-1		0.3	4.2	10.2	67.4	17.1	0.6
PH4	-2		0.7	4.6	10.3	62.0	22.1	0.4
PH5	-3	0.1	1.8	11.5	18.3	54.0	13.5	0.9
PH6	-4	0.1	2.2	8.0	12.5	54.4	22.6	0.2
PH1	-5.5	0.2	1.3	6.7	12.1	60.0	19.5	1.6
PH7	-6.5	0.2	1.6	7.2	12.3	59.3	17.2	1.2
PH8	-6.7	0.6	1.3	11.3	13.4	58.6	14.3	0.4
PH9	-7.2	0.5	3.4	12.6	15.4	54.4	12.6	1.0
PH10	-7.5	0.1	1.1	6.1	13.7	55.2	17.5	1.9
PH11	-8	0.1	1.1	6.1	8.0	64.7	18.5	1.5
PH12	-9	0.3	1.7	6.8	12.1	57.3	18.4	3.3
PH13	-9.6	0.1	0.5	2.7	7.5	61.1	25.2	2.8
PH14	-10.4	0.1	0.8	3.8	8.5	65.1	21.0	0.5

Appendix 2a. Chemical composition of the loess at the Kearny County section of southwestern Kansas.

Sam. (#)	Depth (m)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
KE1	0.13	4.6	0.85	10.5	69.32	2.5	1.03	0.59	2.81	5.31
KE2	0.28	5.2	0.95	10.9	69.91	2.5	1.02	0.57	3.08	6.02
KE3	0.43	5.8	1.10	10.8	64.91	2.4	1.01	0.55	3.09	7.06
KE4	0.58	4.3	1.36	10.6	63.85	2.5	5.19	0.54	2.89	4.63
KE5	0.78	4.8	1.50	11.3	63.48	2.6	3.77	0.53	3.42	4.73
KE6	1.02	4.2	1.54	11.2	62.91	2.5	4.84	0.56	3.19	4.27
KE7	1.6	3.8	1.46	11.9	66.59	2.7	2.87	0.58	3.44	3.51
KE8	2.0	3.7	1.75	12.7	66.46	2.7	2.93	0.62	3.83	2.69
KE10	3.95	4.3	1.57	12.1	62.79	2.5	3.02	0.58	3.71	4.87
KE11	5.49	5.5	1.05	12.4	67.59	2.5	1.0	0.55	3.06	6.51

Appendix 2b. Particle size distribution of the loess at the Kearny County section of southwestern Kansas.

Sam. (#)	Depth (m)	50μ (%)	30μ (%)	20μ (%)	10μ (%)	6μ (%)	2μ (%)	1.9μ (%)
KE1	0.13	0.2	1.5	6.3	12.9	62.9	15.5	0.6
KE2	0.28	0.1	1.1	8.0	15	58.0	16.7	1.0
KE3	0.43	0.3	2.1	7.8	13.1	59.3	16.5	1.0
KE4	0.58	0.1	2.2	12.7	24.8	45.6	12.5	2.1
KE5	0.78	0.3	2.1	8.1	17.8	57.2	13.6	1.0
KE6	1.02	0.3	1.1	7.0	15.9	61.3	13.2	1.0
KE7	1.6	0.04	1.1	7.2	14.0	59.6	17.7	0.4
KE8	2.0	1.66	8.6	16.3	57.8	60.0	14.0	1.0
KE9	3.1	0.03	1.7	7.3	14.8	60.8	14.1	0.8
KE10	3.95	0.15	1.3	7.6	14.9	60.3	13.9	1.8

Appendix 3a. Physical properties of the loess at the Pratt County section (sand, silt, clay, carbon, kaolinite and quartz in clay, and the ratio of the kaoline/Quartz).

Sample (#)	Depth (m)	Sand (%)	Silt (%)	Clay (%)	Carbon (%)	Qua (%)	Kao (%)	Kao/Qua
PR1	-0.10	23.30	76.90	9.8	4.68	26	9.62	0.37
PR2	-0.20	32.40	57.80	9.8	4.13	16	6.24	0.39
PR3	-0.30	34.50	55.70	9.8	3.66	26	12.5	0.48
PR4	-0.40	41.20	53.80	5.0	3.14	26	12.5	0.48
PR5	-0.50	32.70	61.70	6.6	4.06	25	7.00	0.28
PR6	-0.60	28.90	62.90	8.2	4.41	24	8.40	0.35
PR7	-0.70	17.80	67.80	14.4	4.73	24	8.40	0.35
PR8	-0.80	20.40	69.60	10.0	4.73	28	14.3	0.51
PR9	-0.90	22.10	56.90	21.0	3.86	11	4.07	0.37
PR10	-1.00	19.80	63.20	17.0	3.54	24	9.60	0.40
PR11	-1.10	25.50	61.50	13.0	3.26	23	10.8	0.47
PR12	-1.20	36.50	53.70	9.8	3.03	25	12.0	0.48
PR13	-1.30	45.70	43.30	11.0	3.14	27	9.72	0.36
PR14	-1.40	54.10	36.10	9.8	4.08	26	14.0	0.54
PR15	-1.50	56.00	34.20	9.8	3.02	20	11.2	0.56
PR16	-1.60	54.50	46.50	9.0	2.14	27	14.3	0.53
PR17	-1.70	60.10	34.70	5.2	2.75	27	14.3	0.53
PR18	-1.80	57.80	37.60	4.6	2.23	23	7.36	0.32
PR19	-1.90	66.80	26.20	7.4	2.35	23.	7.36	0.32
PR20	-2.00	62.40	36.40	1.2	2.44	23	7.36	0.32
PR21	-2.10	63.60	31.60	4.8	2.88	23	7.36	0.32
PR22	-2.20	67.00	27.80	5.2	2.32	27	5.67	0.21
PR23	-2.30	67.20	28.90	2.6	2.39	27	5.67	0.21
PR24	-2.40	59.40	36.60	4.0	2.69	27	5.67	0.21
PR25	-2.50	38.20	55.70	6.0	3.24	19	5.13	0.27
PR26	-2.60	41.10	56.10	2.8	3.26	19	5.13	0.27
PR27	-2.70	40.40	53.50	8.0	3.17	19	5.13	0.27
PR28	-2.80	41.80	54.60	3.6	3.50	20	5.20	0.26
PR29	-2.90	38.30	60.20	1.4	3.38	20	5.20	0.26
PR30	-3.00	45.50	47.10	7.6	3.34	20	5.20	0.26
PR31	-3.10	37.80	56.40	5.8	3.50	16	6.56	0.41
PR32	-3.20	44.80	50.80	4.4	2.83	16	6.56	0.41
PR33	-3.30	38.00	61.20	0.8	1.54	16	6.56	0.41
PR34	-3.40	27.80	70.60	1.6	1.64	16	6.56	0.82
PR35	-3.50	35.10	67.60	7.2	1.61	7	3.22	0.82
PR36	-3.60	39.70	58.30	2.0	1.27	9	7.38	0.82
PR37	-3.70	48.50	49.10	2.4	1.68	9	7.38	0.82
PR38	-3.80	57.40	41.80	0.8	1.61	9	7.38	0.82
PR39	-3.90	66.80	32.00	1.2	1.17	9	7.38	0.46
PR40	-4.00	71.00	27.80	1.2	0.84	9	7.38	0.40
PR41	-4.10	71.80	26.20	2.0	1.10	9	2.88	0.25
PR42	-4.20	56.70	39.90	3.6	1.68	7	2.80	0.20

Appendix 3a (continues)

Sample (#)	Depth (m)	Sand (%)	Silt (%)	Clay (%)	Carbon (%)	Qua (%)	Kao (%)	Kao/Qua
PR43	-4.30	48.50	48.50	3.0	1.62	7	2.80	0.40
PR44	-4.40	50.10	45.30	4.6	1.94	7	2.80	0.40
PR45	-4.50	71.40	27.40	1.2	1.08	16	6.24	0.39
PR46	-4.60	83.50	15.50	1.0	0.85	16	6.24	0.39
PR47	-4.70	74.30	23.70	2.0	1.20	16	6.24	0.39
PR48	-4.80	82.50	16.00	1.4	1.03	16	6.24	0.39
PR49	-4.90	83.20	15.60	1.2	0.63	16	6.24	0.39
PR50	-5.00	78.10	20.50	1.6	1.33	12	4.80	0.40
PR51	-5.10	64.00	24.40	1.6	1.33	12	4.80	0.40
PR52	-5.20	68.60	29.80	1.6	1.47	12	4.80	0.40
PR53	-5.30	60.50	39.00	1.0	1.45	12	4.80	0.40
PR54	-5.40	60.50	38.00	1.0	1.55	12	4.80	0.40
PR55	-5.80	79.50	18.40	2.0	1.10			
PR56	-6.20	80.70	19.10	0.2	0.55			

Appendix 3b. Chemical composition of the loess at the Pratt County section of central Kansas.

Sample (#)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
PR1	1.91	2.26	10.7	81.56	2.75	0.85	0.73	2.67	0.28
PR2	1.95	1.79	9.98	84.0	2.78	0.75	0.79	2.43	0.34
PR3	1.83	1.2	9. 8	83.63	2.65	0.65	0.78	2.47	0.45
PR4	2.0	0.32	8.59	85.76	2.57	0.66	0.78	1.92	0.31
PR5	1.82	1.13	9.06	85.49	2.68	0.63	0.78	2.05	0.26
PR6	2.12	1.72	10.26	82.03	2.78	0.67	0.74	2.59	0.38
PR7	1.98	2.3	11.16	81.78	2.76	0.78	0.72	2.84	0.47
PR8	1.99	2.11	10.61	82.82	2.78	0.83	0.72	2.59	0.32
PR9	2.15	1.69	10.36	83.55	2.83	0.85	0.67	2.25	0.42
PR10	2.8	2.52	10.99	78.67	2.66	0.9	0.7	2.77	1.93
PR11	3.0	2.39	10.48	78.33	2.68	0.86	0.71	2.67	2.18
PR12	2.27	2.58	11.34	80.71	2.73	0.76	0.74	3.0	0.94
PR13	2.8	1.59	9.98	79.85	2.56	0.68	0.75	2.7	2.12
PR14	2.54	2.18	10.42	81.28	2.61	0.58	0.74	2.79	1.5
PR15	5.31	2.85	12.55	69.54	2.48	1.21	0.65	3.68	4.39
PR16	2.39	1.66	10.3	82.61	2.66	0.57	0.79	2.86	0.61
PR17	3.65	0.47	10.22	82.05	2.81	0.61	0.81	2.71	0.86
PR18	2.32	2.04	10.99	81.63	2.77	0.57	0.86	3.17	0.43
PR19	2.03	2.15	10.68	82.61	2.78	0.57	0.88	3.11	0.41
PR20	2.42	3.52	12.1	76.71	2.77	0.72	0.88	3.79	1.73
PR21	1.84	2.72	11.19	80.27	2.8	0.59	0.88	3.33	0.54
PR22	2.05	2.82	11.67	80.5	2.72	0.57	0.83	3.48	0.63
PR23	1.76	3.28	12.0	79.86	2.82	0.68	0.89	3.66	0.48
PR24	1.88	3.85	12.95	78.61	2.77	0.67	0.86	3.99	0.64
PR25	1.92	2.02	10.21	83.5	2.61	0.62	0.75	2.74	0.57
PR26	2.0	3.22	12.29	79.5	2.71	0.73	0.81	3.58	0.73
PR27	1.85	1.47	10.13	84.68	2.79	0.71	0.81	2.59	0.33
PR28	1.6	1.62	13.92	71.23	2.59	0.88	0.69	4.46	0.51
PR29	2.19	0.7	10.09	79.65	2.78	0.66	0.74	2.72	0.85
PR30	2.47	1.23	12.51	75.52	2.86	0.74	0.73	3.79	0.53
PR31	2.01	1.38	12.5	73.9	2.81	0.72	0.69	3.75	0.54
PR32	1.93	0.85	10.81	78.3	2.78	0.76	0.75	3.03	0.41
PR33	1.76	1.65	12.9	69.29	2.55	2.95	0.64	4.11	0.70
PR34	1.42	1.48	11.9	66.45	2.38	6.68	0.6	3.67	0.52
PR35	2.11	0.77	10.42	79.27	2.67	0.87	0.66	2.77	0.39
PR36	1.73	1.73	13.59	71.44	2.6	1.05	0.68	4.37	0.72
PR37	1.73	1.61	12.6	66.68	2.48	5.03	0.63	4.05	0.8
PR38	2.07	4.72	11.95	67.11	2.47	5.44	0.64	3.78	0.96
PR39	2.75	5.14	12.67	69.42	2.59	2.25	0.67	4.06	2.18
PR40	2.47	5.43	13.22	69.35	2.63	1.82	0.64	4.24	1.67
PR41	2.26	3.49	11.93	74.91	2.77	1.15	0.68	3.56	1.03
PR42	2.4	4.24	11.81	70.45	2.64	3.51	0.68	3.64	1.38
PR43	2.13	4.53	11.98	70.56	2.64	3.59	0.69	3.69	1.07
PR44	1.71	3.87	10.42	62.05	2.3	9.99	0.59	3.24	0.94

Appendix 3b (continues)

Sample (#)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
PR45	2.14	4.12	12.37	73.59	2.8	1.43	0.68	3.81	0.64
PR46	2.84	3.56	12.39	73.53	2.82	0.77	0.66	3.68	1.54
PR47	2.4	3.76	12.07	73.99	2.81	1.13	0.67	3.61	1.28
PR48	2.71	4.48	13.72	70.33	2.73	0.59	0.71	4.91	1.37
PR49	2.5	2.6	11.23	76.54	2.82	0.66	0.63	3.07	1.09
PR50	2.39	3.12	11.99	74.44	2.74	0.65	0.67	3.68	0.98
PR51	2.19	4.24	13.18	71.24	2.55	1.24	0.67	4.59	1.21
PR52	2.03	3.19	11.99	74.56	2.79	1.5	0.73	3.6	0.71
PR53	1.81	3.42	11.2	62.71	2.34	9.07	0.62	3.69	0.91
PR54	2.03	3.36	11.92	74.33	2.87	1.85	0.76	4.01	0.68
PR55	2.03	3.36	11.92	74.33	2.87	1.85	0.76	4.01	0.68
PR56	1.87	2.98	9.65	61.79	2.4	9.99	0.66	3.59	0.65

Appendix 3c. Weathering indices* and $CaCO_3$ of the loess at the Pratt County section (* stable = Al_2O_3 + Fe_2O_3; unstable = NaO_2 + K_2O + MgO + CaO + P_2O_5).

Sample (#)	Depth (m)	Suscept (emu/g)	Fe2O3 (%)	Stable (%)	Unstable (%)	St/Unst (L.I)	CaCO3 (%)
PR1	-0.1	5.7	2.67	13.37	8.05	1.661	2.96
PR2	-0.2	5.3	2.43	12.41	7.61	1.631	2.74
PR3	-0.3	6.0	2.47	12.27	6.78	1.81	2.14
PR4	-0.4	5.9	1.92	10.51	5.91	1.778	1.82
PR5	-0.5	5.4	2.05	11.11	6.52	1.704	2.46
PR6	-0.6	5.6	2.59	12.85	7.67	1.675	2.78
PR7	-0.7	5.9	2.84	14.0	8.38	1.671	2.68
PR8	-0.8	6.0	2.59	13.2	8.03	1.644	2.6
PR9	-0.9	6.0	2.25	12.61	7.94	1.588	2.48
PR10	-1	6.1	2.77	13.76	10.9	1.266	3.18
PR11	-1.1	5.8	2.67	13.15	11.1	1.184	3.74
PR12	-1.2	5.9	3.0	14.34	9.28	1.545	3.6
PR13	-1.3	5.6	2.7	12.68	9.75	1.301	2.24
PR14	-1.4	5.7	2.79	13.21	9.41	1.404	2.34
PR15	-1.5	5.6	3.68	16.23	16.2	0.999	1.96
PR16	-1.6	5.3	2.86	13.16	7.89	1.668	2.0
PR17	-1.7	5.6	2.71	12.93	8.4	1.539	3.18
PR18	-1.8	4.9	3.17	14.16	8.13	1.742	2.02
PR19	-1.9	5.2	3.11	13.79	7.94	1.737	2.14
PR20	-2	5.2	3.79	15.89	11.2	1.424	2.46
PR21	-2.1	5.1	3.33	14.52	8.49	1.71	1.98
PR22	-2.2	5.4	3.48	15.15	8.79	1.724	1.86
PR23	-2.3	5.1	3.66	15.66	9.02	1.736	1.7
PR24	-2.4	5.1	3.99	16.94	9.81	1.727	1.64
PR25	-2.5	5.2	2.74	12.95	7.74	1.673	2.48
PR26	-2.6	5.9	3.58	15.87	9.39	1.69	1.96
PR27	-2.7	6.0	2.59	12.72	7.15	1.779	2.2
PR28	-2.8	5.9	4.46	18.38	7.2	2.553	2.06
PR29	-2.9	5.8	2.72	12.81	7.18	1.784	2.14
PR30	-3	6.0	3.79	16.3	7.83	2.082	2.14
PR31	-3.1	6.0	3.75	16.25	7.46	2.178	2.68
PR32	-3.2	5.5	3.03	13.84	6.73	2.056	2.1
PR33	-3.3	6.0	4.11	17.01	9.61	1.77	2.76
PR34	-3.4	6.1	3.67	15.57	12.5	1.248	4.92
PR35	-3.5	5.8	2.77	13.19	6.81	1.937	2.18
PR36	-3.6	5.2	4.37	17.96	7.83	2.294	5.06
PR37	-3.7	4.5	4.05	16.65	11.7	1.429	5.94
PR38	-3.8	4.0	3.78	15.73	15.7	1.004	6.5
PR39	-3.9	3.2	4.06	16.73	14.91	1.122	3.5
PR40	-4	3.4	4.24	17.46	14.0	1.245	2.5
PR41	-4.1	3.4	3.56	15.49	10.7	1.448	2.1
PR42	-4.2	3.8	3.64	15.45	14.17	1.09	9.1

Appendix 3c (continues)

Sample (#)	Depth (m)	Suscept (emu/g)	Fe2O3 (%)	Stable (%)	Unstable (%)	St/Unst (L.I)	CaCO3 (%)
PR43	-4.3	3.9	3.69	15.67	13.9	1.122	6.0
PR44	-4.4	3.0	3.24	13.66	18.8	0.726	9.46
PR45	-4.5	2.4	3.81	16.18	11.1	1.454	2.72
PR46	-4.6	2.6	3.68	16.07	11.53	1.394	1.24
PR47	-4.7	0.7	3.61	15.68	11.38	1.378	4.0
PR48	-4.8	0.6	4.91	18.63	11.88	1.568	7.42
PR49	-4.9	0.6	3.07	14.3	9.68	1.477	1.0
PR50	-5	0.6	3.68	15.67	9.88	1.586	1.8
PR51	-5.1	0.7	4.59	17.77	11.43	1.555	2.66
PR52	-5.2	0.8	3.6	15.59	10.22	1.525	6.0
PR53	-5.3		3.69	14.89	17.55	0.848	15.0
PR54	-5.4		4.01	15.93	10.79	1.476	8.22
PR55	-5.8		4.01	15.93	10.79	1.476	2.36
PR56	-6.2		3.59	13.24	17.89	0.74	2.35

Appendix 3d. Particle size distribution (in percentage of particle numbers) of the loess at the Pratt County section.

Sample (#)	50μ (%)	30μ (%)	20μ (%)	10μ (%)	6μ (%)	2μ (%)	1.9μ (%)	<1.9μ (%)
PR1	2.38		0.1	2.1	5	18.4	16.6	57.6
PR2	1.85		0.1	0.7	1.3	15	34	48
PR3	1.62		0.1	2	5.1	17.2	16.6	58.9
PR4	2.61	0.1	0.8	3.6	6.7	19.5	16.1	53
PR5	1.99	0.2	0.5	3.1	6.2	19.1	16	54.7
PR6	1.92	0.07	0.4	3.1	6.7	19.9	17.1	52.5
PR7	1.85		0.1	1	2.5	9.4	7.9	79
PR8	1.09		0.3	1.7	3.8	13	12.1	69
PR9	1.25	0.05	0.3	1.5	4.7	15.2	4	64.3
PR10	1		0.3	1.9	4	14.7	17.8	61.6
PR11	1.12		0.3	1.9	4.5	16.5	16.2	60.4
PR12	1.81	0.04	0.2	2.2	5.2	17.4	15.7	59.2
PR13	2.5	0.1	0.6	4.3	7.8	23.3	17.1	46.4
PR14	2.62		0.1	1.7	4.4	17.1	16.7	59.8
PR15	1.74	0.03	0.3	2.2	5.5	20	19.5	52.4
PR16	3.04		0.3	2.3	5.3	18.1	17.1	56.3
PR17	1.85		0.05	1.5	4.4	18.6	17.9	57.4
PR18	1.65		0.2	1.8	4.1	15.7	16.1	61.8
PR19	0.71		0.2	1.5	3.3	12.2	14.6	68.1
PR20	1.8		0.3	1.9	4.1	16.1	16.9	60.6
PR21	2.52		0.2	2.3	8	49.4	23.6	16.3
PR22	4.28		0.3	2.9	8.1	43.3	29.6	15.5
PR23	4.35		0.3	2.1	6.6	45.9	31.7	13.2
PR24	4.92	0.06	0.3	2.3	7	51.4	27.1	11.7
PR25	3.86	0.2	0.8	3	7.9	48.4	24.2	11.9
PR26	2.87	0.06	0.3	3.5	9.3	50.5	25	11
PR27	3.04		0.4	2.6	8	47.9	25.2	15.7
PR28	2.58	0.1	0.5	2.2	5.7	57.6	22	11.7
PR29	3.29	0.15	1	5.9	11	44.7	27.3	12.6
PR30	3.39	0.1	1	3.6	8.3	47.1	27.3	12.6
PR31	2.73	0.04	0.3	2.9	7.9	45.2	32.9	10.6
PR32	3.82	0.07	0.6	3.2	8.8	48.3	25.9	12.9
PR33	4.34	0.06	1.2	5	10.3	43.8	31.6	7.8
PR34	3.06	0.06	1	3.9	7.8	49	32.2	5.7
PR35	3.59	0.3	1	5.7	9.7	43.1	34.9	5
PR36	3.44	0.14	2.2	6.2	9	46.3	34	1.9
PR37	2.18		0.4	3	6.7	38.4	50	0.9
PR38	3.3	0.04	11	0.8	5	53.7	28.6	0.6
PR39	2.94	0.2	1.5	6.5	10.6	52	28.7	0.5
PR40	3.66	0.06	0.7	3.4	6.5	34.6	53.7	0.8
PR41	2.52	0.07	0.7	2.7	5.3	35.1	54.2	1.7
PR42	2.97	0.1	0.6	3	7.3	54.3	31.9	2.6
PR43	3.34	0.05	1	4.9	9	43.4	19.4	1
PR44	2.2	0.03	0.4	2.7	7.5	57.1	28.8	3.4

Appendix 3d (continues)

Sample (#)	50μ (%)	30μ (%)	20μ (%)	10μ (%)	6μ (%)	2μ (%)	1.9μ (%)	<1.9μ (%)
PR45	1.3		0.5	3.6	8.5	56.3	29	2
PR46	2.25	0.1	0.8	3.3	8.6	53.1	28.3	5.6
PR47	1.78	0.1	0.6	3.2	7.9	52.2	31.7	4.1
PR48	4.9		0.3	1.7	5.2	46.3	44.6	1.7
PR49	1.7		0.2	2	6	35.8	45.9	9.9
PR50	1.53		0.3	3.3	9.2	51.2	27.6	6.7
PR51	2.34	0.03	0.5	2.8	4.5	28.9	62.4	1
PR52	2.7	0.05	0.4	3.8	10.8	55.3	24.4	5.4
PR53	1.96		0.5	4.8	13.7	58.9	19.1	2.8
PR54	2.44	0.04	0.5	5.4	12.7	60	20	1.2
PR55	2.98	0.1	0.5	2.3	7.7	60.3	26	2.8
PR56	2.49		0.2	2.2	9.1	61.2	24.5	2.7

Appendix 4a. Physical properties of loess at the Barton County section.

Sample (#)	Depth (m)	Sand (%)	Silt (%)	Clay (%)	Quartz (%)	Kaoli (%)	Kao /Qua	CaCO3 (%)
BT1	-0.2	2	88.0	10	14	4.2	0.3	2.12
BT2	-0.4	7.7	85.3	7	9.6	4.4	0.46	2.1
BT3	-0.6	6	81.2	12.8	18	5.9	0.33	3.18
BT4	-0.8	6	89	5	15	6.3	0.42	4.02
BT5	-1	3	79.8	17.2	27	13.2	0.49	4.66
BT6	-1.2	6	88.4	15.6	20	8.8	0.44	2.88
BT7	-1.4	4.5	88.7	15.8	18	8.1	0.45	2.31
BT8	-1.6	4	80	16	23	7.4	0.32	2.28
BT9	-1.8	6	80	14	18	9.9	0.55	2.26
BT10	-2	7.5	77.5	15	2	7.2	0.36	2.16
BT11	-2.2	9	73.4	17.6	15	5.3	0.35	2.3
BT12	-2.4	6	75.4	18.6	12	3.7	0.31	2.54
BT13	-2.6	7	80	13	13	9.8	0.75	1.82
BT14	-2.8	8	81	11	18	6.7	0.37	1.82
BT15	-3	11	79.6	10.4	22	8.1	0.37	1.7
BT16	-3.2	13	79.6	7.4	17	8.8	0.52	1.74
BT17	-3.4	11	80.4	8.6	29	9.6	0.33	1.6
BT18	-3.6	21	70.2	8.8	34	13.6	0.4	1.7
Bt19	-3.8	29	62.2	8.8	23	12.2	0.53	1.42
BT20	-4	29	63.4	7.4	19	8.6	0.45	2.16
BT21	-4.2	25	67.6	7.4	23	10.1	0.44	2.06
BT22	-4.4	34	61.6	5.4	23	10.1	0.44	1.74
BT23	-4.6	48.5	47.1	4.4	21	14.9	0.71	1.62
BT24	-4.8	71.7	24.7	3.6	2	1.4	0.71	1.26
BT25	-5	55	36.4	8.6	16	7.7	0.46	1.18
BT26	-5.2	55	40	5	16	7.4	0.46	1.34
BT27	-5.4	69.6	25.2	5.2	9	6.1	0.68	4.3
BT28	-5.6	72.9	23.4	3.8	9	6.1	0.68	3.74
BT29	-5.8	76.6	21.4	1	15	10.2	0.68	2.4
BT30	-6	74.3	23.5	2.2	15	10.2	0.68	2.58
BT31	-6.2	72	24	4	15	10.2	0.68	6.48
BT32	-6.4	72.7	24.3	3	3	1.14	0.38	2.9
BT33	-6.6	40	47	13	3	1.14	0.38	18.6
BT34	-6.8	45	30.6	20	3	2.16	0.72	18.5
BT35	-7	58	33.3	8.6	8	4.56	0.57	10.4
BT36	-7.2	32.6	55.6	11.8	8	2.8	0.35	13.4
BT37	-7.4	11.8	73.2	15	4	0.76	0.19	20.3
BT38	-7.6	29.6	58.2	12.2	8	2	0.25	20
BT39	-7.8	40	49.3	11.2	9	3.24	0.36	19.1
BT40	-8	30	63.2	6.8	23	5.29	0.23	8.52
BT41	-8.2	29	57.8	13.2	4	1.52	0.38	13.3
BT42	-8.4	14.4	68.8	16.8	8	2.64	0.33	7.06

Appendix 4a (continues)

Sample (#)	Depth (m)	Sand (%)	Silt (%)	Clay (%)	Quartz (%)	Kaoli (%)	Kao /Qua	CaCO3 (%)
Bt43	-8.6	17.4	64.6	18	8	0.96	0.12	11
BT44	-8.8	24	64	12	8	1.44	0.18	5.56
BT45	-9	12.7	67.5	9.8	7	4.27	0.61	4.94
BT46	-9.2	16.5	65.1	18.4	6	2.52	0.42	7.54
BT47	-9.4	22	64.6	13.4	9	2.79	0.31	10.3
BT48	-9.6	19	68	13	5	0.2	0.04	13.4
BT49	-9.8	14.5	75.5	10	15	5.4	0.36	10.6
BT50	-10	21	74.4	4.6	15	5.4	0.36	7.97
BT51	-10.2	18.5	73.5	8	11	3.7	0.34	3.14
BT52	-10.4	22.3	65.3	12.4	11	3.74	0.34	3.6
BT53	-10.6	21	74.2	4.8	6	2.46	0.41	7.6
BT54	-10.8	10.7	84.7	4.6	6	2.46	0.41	8.6
BT55	-11	10.7	72.9	12.4	5	3.75	0.75	15
BT56	-11.2	13.6	70	16.4	5	1.6	0.32	13
BT57	-11.4	9.8	79.8	10.4	8	4.0	0.51	5.5
BT58	-11.6	7.6	79.8	12.6	15	6.15	0.41	3.4
BT59	-11.8	10.4	84	5.6	9	2.79	0.31	12.3
BT60	-12	13	81.4	5.6	10	4.5	0.45	12.3
BT61	-12.5	9	66.8	24.2	5	1.25	0.25	16.2
BT62	-13	12.4	72	15.6	6	1.5	0.25	22.6
BT63	-13.5	28	59.4	12.6	7	1.75	0.25	13.4
BT64	-14	21.4	49.4	29.2	5	0.05	0.01	21.4
BT65	-14.5	15	55	30	5	0.05	0.01	12.4
	-15	21		4.8				2.66
	-15.5	16		6.6				6
	-16	15		8.6				15
	-16.5	8.8		13				8.22
	-17	7.6		15				4.36
	-17.5	8.9		16				4.7
	-18	12		16				3.15

Appendix 4b. Chemical composition of the loess at the Barton County section.

Sample (#)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
BT1	1.8	11.99	10.15	80.36	2.71	0.74	0.65	2.21	0.3
BT2	1.99	1.18	9.01	82.99	2.6	0.79	0.7	1.79	0.48
BT3	1.86	1.53	9.72	80.34	2.71	0.84	0.67	2.01	0.32
BT4	1.92	1.47	9.45	81.11	2.67	0.82	0.69	1.96	0.33
BT5	1.71	2.7	11.11	78.67	2.78	0.74	0.7	2.72	0.29
BT6	2.09	2.67	11.28	78.91	2.89	0.8	0.7	2.77	0.42
BT7	2.03	3.14	11.74	77.58	2.85	0.89	0.69	2.65	0.41
BT8	2.66	2.4	11.03	78.61	2.91	0.84	0.68	2.64	0.34
BT9	2.1	2.49	11.03	78.42	2.89	0.93	0.68	2.67	0.31
BT10	2.4	2.54	11.19	77.96	2.91	0.96	0.7	2.7	0.31
BT11	2.47	3.1	11.91	77.48	2.82	1.02	0.71	3.04	0.33
BT12	2.24	2.25	10.83	77.65	2.78	1.03	0.71	2.69	0.22
BT13	2.66	2.91	11.96	77.02	2.84	0.98	0.68	3.08	0.4
BT14	2.05	2.76	11.33	78.44	2.77	0.9	0.69	3	0.3
BT15	1.89	2.13	10.7	80.08	2.7	0.7	0.72	2.81	0.24
BT16	1.85	1.93	10.17	80.74	2.61	0.57	0.77	2.83	0.31
BT17	1.8	1.57	9.91	81.78	2.66	0.56	0.76	2.56	0.24
BT18	1.28	1.89	10.32	79.72	2.7	0.57	0.72	2.88	0.31
Bt19	1.81	1.33	9.35	83.35	2.68	0.71	0.79	2.39	0.23
BT20	1.88	1.35	9.02	82.49	2.58	0.78	0.75	2.2	0.24
BT21	1.86	1.33	9.21	82.97	2.58	0.75	0.71	2.23	0.23
BT22	1.87	1.51	9.43	81.49	2.63	0.75	0.72	2.34	0.23

Sample (#)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
BT23	1.87	1.97	9.49	82.42	2.66	0.78	0.74	2.4	0.29
BT24	2.1	1.87	9.67	81.67	2.71	0.86	0.66	2.23	0.38
BT25	2.05	1.99	9.8	81.53	2.68	0.9	0.71	2.4	0.33
BT26	2.28	2.04	10.14	79.97	2.69	0.91	0.67	2.55	0.71
BT27	1.88	2.58	9.57	72.01	2.41	6.92	0.65	2.65	0.61
BT28	2.02	2.84	10.62	79.23	2.69	1.85	0.71	2.95	0.57
BT29	2.2	1.86	9.46	82.76	2.65	1.17	0.65	2.22	0.73
BT30	1.89	2.07	9.72	82.78	2.64	1.33	0.69	2.4	0.35
BT31	1.78	2.26	9.31	73.63	2.5	6.33	0.68	2.49	0.59
BT32	1.97	2.27	9.68	76.68	2.58	3.88	0.73	2.56	0.51
BT33	1.39	1.27	6.94	63.84	1.96	9.99	0.57	1.8	0.48
BT34	1.53	2.13	8	63.45	2.11	9.3	0.58	2.35	0.41
BT35	1.32	1.56	6.8	59.39	1.94	9.99	0.56	1.88	0.5
BT36	1.46	2.21	8.76	67.33	2.35	9.99	0.6	2.44	0.47
BT37	1.57	1.98	8.24	64.91	2.25	9.99	0.53	2.16	0.56
BT38	1.56	1.57	8.67	73.62	2.43	7.3	0.64	2.34	0.33
BT39	1.61	2.13	9.48	76.36	2.5	5.52	0.71	2.71	0.39
BT40	2.01	1.81	10.41	81.46	2.87	1.02	0.7	2.53	0.32
BT41	1.58	2.65	9.32	67.88	2.32	9.99	0.62	2.66	0.57
BT42	1.87	2.38	10.41	78.39	2.73	2.53	0.67	2.73	0.41
Bt43	2.15	2.4	10.05	74.71	2.6	4.31	0.65	2.71	1.02
BT44	2.19	3.02	10.96	77.65	2.79	1.98	0.7	3.04	0.86
BT45	1.84	3.05	11.14	78.55	2.7	1.45	0.66	3.11	0.5
BT46	2.23	2.75	9.54	70.63	2.4	6.72	0.62	2.74	1.59

Sample (#)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
BT47	1.88	2.24	9.77	77.22	2.61	0.82	0.69	2.58	0.52
BT48	1.94	2.74	11.26	78.5	2.72	1.02	0.69	3.13	0.43
BT49	1.88	2.68	10.43	76.61	2.63	3.43	0.65	2.78	2.56
BT50	1.96	3.65	11.77	77.48	2.79	1.29	0.69	3.27	0.54
BT51	1.96	2.99	11.41	79.39	2.76	0.8	0.69	3.01	0.37
BT52	1.99	2.34	11.32	78.84	2.77	0.85	0.7	3.14	0.45
BT53	1.94	3.5	11.27	79.31	2.68	0.94	0.68	3.18	0.43
BT54	1.79	2.94	10.75	75.66	2.54	4.2	0.66	2.96	0.45
BT55	1.8	2.19	9.58	79.77	2.61	2.51	0.69	2.61	0.26
BT56	1.87	3.02	11.09	77.23	2.74	2.81	0.69	2.98	0.47
BT57	1.91	2.51	10.28	79.86	2.74	2.27	0.67	2.58	0.47
BT58	2.04	2.44	10.97	80.86	2.85	0.95	0.69	2.73	0.36
BT59	1.84	3.48	12.02	75.31	2.71	1.44	0.67	3.32	0.48
BT60	1.9	2.7	11.04	77.54	2.7	1.7	0.69	2.82	0.51
BT61	1.76	3.02	10.25	72.03	2.52	7.24	0.61	2.64	0.56
BT62	1.54	2.39	8.68	66.62	2.36	9.99	0.6	2.18	0.54
BT63	1.7	2.28	9.33	75.89	2.52	5.95	0.68	2.35	0.44
BT64	1.41	2.81	8.37	62.85	2.04	9.99	0.57	2.11	0.56
BT65	1.41	2.81	8.37	62.85	2.04	9.99	0.57	2.11	0.56

Appendix 4c. Weathering indices* of the loess at the Barton County section (* stable = $Al_2O_3 + Fe_2O_3$; unstable = $Na_2O + MgO + K_2O + CaO + P_2O_5$).

Sample (#)	Depth (m)	Suscept (emu/g)	Fe2O3 (%)	Stable (%)	Unstable (%)	St/Unst (L.I)
BT1	-0.2	6.8	2.21	12.36	7.55	1.6371
BT2	-0.4	6.5	1.79	10.8	7.04	1.5341
BT3	-0.6	6.4	2.01	11.73	7.26	1.6157
BT4	-0.8	6.8	1.96	11.41	7.21	1.5825
BT5	-1	6.6	2.72	13.83	8.22	1.6825
BT6	-1.2	4.4	2.77	14.05	8.87	1.584
BT7	-1.4	4.3	2.65	14.39	9.32	1.544
BT8	-1.6	4.6	2.64	13.67	9.15	1.494
BT9	-1.8	5.1	2.67	13.7	8.75	1.5657
BT10	-2	5.5	2.7	13.89	9.12	1.523
BT11	-2.2	5.6	3.04	14.95	9.74	1.5349
BT12	-2.4	6.3	2.69	13.52	8.52	1.5869
BT13	-2.6	6.1	3.08	15.04	9.79	1.5363
BT14	-2.8	6.6	3	14.33	8.78	1.6321
BT15	-3	6.1	2.81	13.51	7.66	1.7637
BT16	-3.2	6.8	2.83	13.0	7.27	1.7882
BT17	-3.4	6.5	2.56	12.47	6.86	1.8178
BT18	-3.6	7.5	2.88	13.2	6.75	1.9556
Bt19	-3.8	7.1	2.39	11.74	6.76	1.7367
BT20	-4	8.2	2.2	11.22	6.83	1.6428
BT21	-4.2	6.1	2.23	11.44	6.75	1.6948
BT22	-4.4	6	2.34	11.77	6.99	1.6838
BT23	-4.6	6.2	2.4	11.89	7.57	1.5707
BT24	-4.8	6.1	2.23	11.9	7.92	1.5025
BT25	-5	5.9	2.4	12.2	7.95	1.5346
BT26	-5.2	5.6	2.55	12.69	8.63	1.4705
BT27	-5.4	5.8	2.65	12.22	14.4	0.8486
BT28	-5.6	6.6	2.95	13.57	9.97	1.3611
BT29	-5.8	4.9	2.22	11.68	8.61	1.3566
BT30	-6	5.3	2.4	12.12	8.28	1.4638
BT31	-6.2	6.8	2.49	11.8	13.5	0.8767
BT32	-6.4	5.3	2.56	12.24	11.2	1.0919
BT33	-6.6	4.5	1.8	8.74	15.1	0.5792
BT34	-6.8	6.5	2.35	10.35	15.5	0.6686
BT35	-7	6.4	1.88	8.68	15.3	0.5669
BT36	-7.2	4.5	2.44	11.2	16.5	0.6796
BT37	-7.4	3.5	2.16	10.4	16.4	0.6361
BT38	-7.6	3.3	2.34	11.01	13.2	0.8347
BT39	-7.8	3.9	2.71	12.19	12.15	1.0033
BT40	-8	4.7	2.53	12.94	8.03	1.6115
BT41	-8.2	4.8	2.66	11.98	17.11	0.7002
BT42	-8.4	4	2.73	13.14	9.92	1.3246

Appendix 4c (continues)

Sample (#)	Depth (m)	Suscept (emu/g)	Fe2O3 (%)	Stable (%)	Unstable (%)	St/Unst (L.I)
Bt43	-8.6	4.5	2.71	12.76	12.5	1.0224
BT44	-8.8	4.9	3.04	14	10.8	1.2915
BT45	-9	5.3	3.11	14.25	9.54	1.4937
BT46	-9.2	5.3	2.74	12.28	15.7	0.7827
BT47	-9.4	5.6	2.58	10.35	6.07	1.7051
BT48	-9.6	5.3	3.13	14.39	8.85	1.626
BT49	-9.8	5.6	2.78	13.21	13.2	1.0023
BT50	-10	5.6	3.27	15.04	10.2	1.4702
BT51	-10.2	5.8	3.01	14.42	8.88	1.6239
BT52	-10.4	5.8	3.14	14.46	8.4	1.7214
BT53	-10.6	5.5	3.18	14.45	9.49	1.5227
BT54	-10.8	5.6	2.96	13.71	11.9	1.1502
BT55	-11	6.0	2.61	12.19	9.37	1.301
BT56	-11.2	5.0	2.98	14.07	10.9	1.2896
BT57	-11.4	4.5	2.58	12.86	9.9	1.299
BT58	-11.6	4.6	2.73	13.7	8.64	1.5856
BT59	-11.8	4.4	3.32	15.34	9.95	1.5417
BT60	-12	3.8	2.82	13.86	9.51	1.4574
BT61	-12.5	3.7	2.64	12.89	15.1	0.8536
BT62	-13		2.18	10.86	16.8	0.6457
BT63	-13.5		2.35	11.68	12.9	0.9061
BT64	-14		2.11	10.48	16.8	0.6234
BT65	-15		2.11	10.48	16.81	0.6234

Appendix 4d. The Particle size distribution (in percentage of particle numbers) of the loess at the Barton County section.

Sample (#)	Depth (m)	50μ (%)	30μ (%)	20μ (%)	10μ (%)	6μ (%)	2μ (%)	1.9μ (%)	<1.9μ (%)
BT1	-0.2	0.4	0.2	1.9	6	9	55.3	11.7	3.4
BT2	-0.4	1.1		0.1	2.1	4.8	38.1	48.5	6.2
BT3	-0.6	1.1	0.6	1.5	5.1	6.5	37.1	39.7	9.1
BT4	-0.8	0.6	0.1	1.4	2.4	4.3	33.4	50.9	7.2
BT5	-1	1.0		0.8	1.7	5.2	40.6	42.2	10.1
BT6	-1.2	0.8	0.1	0.4	2.6	6.7	43.2	35.7	11.1
BT7	-1.4	0.7		0.2	2	6.1	43.3	39.5	8.6
BT8	-1.6	1		0.2	2.5	6	45.8	41.8	3.5
BT9	-1.8	1.2		0.2	2	5.2	44.1	44.5	4
BT10	-2	1.8		1	6	8	60	46	2
BT11	-2.2	0.9		0.4	3.1	5.7	31.3	57.8	2.3
BT12	-2.4	1.2	0.2	1	4.5	8	60.7	20.2	1.2
BT13	-2.6	1.2	0.6	1.6	6.3	3.9	32.5	23	9.1
BT14	-2.8	1.8		0.4	6.4	3.2	29.5	45.1	15.1
BT15	-3	2.1	0.1	0.8	4.5	8	62.5	20.2	0.6
BT16	-3.2	1.6		0.5	6.3	7.7	45.8	33.9	8.9
BT17	-3.4	1.6		0.3	1.6	4.8	38	50.5	5.8
BT18	-3.6	3.8		0.4	2.4	5.7	44.4	39.4	7.9
Bt19	-3.8	2.3		0.5	3.1	8.5	50.4	26	11.4
BT20	-4	3.8		0.5	3.2	7.4	51.3	32	5.4
BT21	-4.2	2.8		0.5	3.7	7	72.7	13	0.7
BT22	-4.4	3.7	0.1	0.8	5.4	8	61	17.4	2.2
BT23	-4.6	5.7	0.2	2	6.7	8	64.5	10.1	0.6
BT24	-4.8	6.3	0.06	0.5	3.2	7	61	23	1.5
BT25	-5	5.2	0.1	0.6	3.2	7.8	49.3	36.7	2.1
BT26	-5.2	4.5	0.06	1	5	9.9	64.9	17.9	1
BT27	-5.4	4.6	0.1	0.5	2.5	8	66.1	21.9	0.6
BT28	-5.6	4.3	0.06	0.6	2.7	7.3	64.6	23.3	1.4
BT29	-5.8	3.1	0.1	0.6	2.7	6.6	67.7	21	0.7
BT30	-6	4.2	0.08	0.6	1.5	5.9	62.1	29.1	0.5
BT31	-6.2	4.8	0.07	0.4	3	7.4	62.9	23.4	2.6
BT32	-6.4	4.4		0.2	1.1	3.1	74.3	7.7	0.7
BT33	-6.6	2.3	0.05	0.4	2.3	7.5	68	22.9	5.2
BT34	-6.8	3.0	0.1	0.3	1.5	5.1	66.4	24	2.3
BT35	-7	3.9	0.06	0.7	3.1	7.9	64.5	20.3	3.2
BT36	-7.2	2.9		0.1	1.1	8	66.3	18.3	2.6
BT37	-7.4	1		0.4	3.5	8	57.8	25.4	2.5
BT38	-7.6	4.1	0.1	0.3	3.1	6.3	18	16.8	55.3
BT39	-7.8	5.4		0.3	2.4	4.5	14.7	9.7	68
BT40	-8	3.4		0.7	3.3	6	20.9	17.6	51.3
BT41	-8.2	4.8		0.07	1.2	3.5	14.1	18	63
BT42	-8.4	1.9	0.5	0.9	3.3	5.3	19.4	17.2	53.2
Bt43	-8.6	2.8	0.15	0.8	2.9	4	16.4	17.6	57.7
BT44	-8.8	4.0	0.16	0.6	4.8	7	24.3	17.1	44.1

Appendix 4d (continues)

Sample (#)	Depth (m)	50μ (%)	30μ (%)	20μ (%)	10μ (%)	6μ (%)	2μ (%)	1.9μ (%)	<1.9μ (%)
BT45	-9	1.9	0.2	1	5.3	7	23	17.7	42.7
BT46	-9.2	2.2		0.6	4.9	7	22.4	18.2	44.9
BT47	-9.4	3.7		0.3	2.1	4.8	17.8	17.1	59.7
BT48	-9.6	3.4	0.05	0.5	2.4	4.9	17.4	16.9	57.4
BT49	-9.8	2.2		0.1	1.7	5.2	18.1	18	56.8
BT50	-10	4.2	0.07	0.4	2.7	5.4	17.2	15	59
BT51	-10.2	3.5	0.06	0.8	4.5	6.8	17.9	17.6	50.7
BT52	-10.4	3.6		0.7	5.4	8.8	21.3	17.9	45.4
BT53	-10.6	3.3		1	6	7	26.7	16.5	39
BT54	-10.8	2.0	0.06	0.5	2.8	7	16.5	15.4	59.1
BT55	-11	1.9	0.1	0.2	2.3	4	13.8	12.2	66.9
BT56	-11.2	2.4	0.06	0.5	3.7	7.2	22	18.6	47.8
BT57	-11.4	1.1		0.2	1.5	3	10.8	10.8	73.6
BT58	-11.6	1.2		0.4	4	7.7	23.6	18.9	45.3
BT59	-11.8	1.7		0.4	3.4	7.8	23.6	18.6	45.9
BT60	-12	1.7	0.05	0.2	3.1	6.8	22.3	18.2	49.2
BT61	-12.5	1.0	0.05	0.5	2.8	6.1	19.6	17.3	53.4
BT62	-13	0.9		0.3	2	4.3	16	14.9	62.3
BT63	-13.5	1.7		0.2	2.3	5.6	19.7	17.2	54.9
BT64	-14	1.8		0.2	1.5	3.6	17.5	20	57
BT65	-14.5	1.0							

Appendix 5a. Chemical composition of sand-soil sequence at the Great Bend Sand Prairie section.

Sam. (#)	Depth (m)	NaO2 (%)	MgO (%)	Al2O3 (%)	SiO2 (%)	K2O (%)	CaO (%)	TiO2 (%)	Fe2O3 (%)	P2O5 (%)
ST1	0.20	1.95	2.42	10.7	77.5	2.77	0.67	0.68	2.87	1.01
ST2	0.75	1.85	2.42	10.7	77.7	2.88	0.65	0.86	2.88	1.01
ST4	1.60	1.86	2.32	10.6	77.8	2.84	0.65	0.85	2.75	0.61
ST5	1.75	1.93	1.54	9.61	81.1	2.76	0.76	0.74	1.87	0.39
ST6	1.95	2.01	1.74	10.0	80.5	2.82	0.94	0.71	1.61	0.36
ST7	2.16	2.09	1.79	10.3	80.2	2.78	1.09	0.64	1.45	0.23
ST8	2.46	2.45	1.82	10.4	79.0	2.79	1.08	0.68	1.60	1.16
ST9	2.65	2.01	1.14	9.13	83.0	2.71	0.80	0.66	1.17	0.38

Appendix 5b. Particle size distribution (in percentage of the particle numbers) of the Great Bend Sand Prairie section of central Kansas.

Sam. (#)	Depth (m)	50μ (%)	30μ (%)	20μ (%)	10μ (%)	6μ (%)	2μ (%)	1.9μ (%)	<1.9μ (%)
ST1	-0.20	0.3	3.3	10.5	64.4	15.6	5.6		
ST2	-0.72	0.9	27	13.1	49.7	8.8	0.7		
ST4	-1.60	0.2	1.9	8.0	17.4	59.7	11.0	1.7	
ST5	-1.75	0.6	1.0	6.4	9.8	71.9	5.9	1.4	
ST6	-1.95	5.25	0.4	2.3	9.5	17.7	60.1	9.1	0.7
ST7	-2.16	2.71	0.1	1.8	8.4	18.7	61.0	7.1	1.9
ST8	-2.46	2.49	0.1	1.0	5.7	38.1	48.0	6.5	0.5
ST9	-2.65	3.65	0.2	2.5	11.6	24.1	52.5	6.2	2.2

Appendix 6. Problems and Qualification of the TL Dating of Quaternary Sediments

The phenomenon of thermoluminescence has been discussed in many publications (e.g., Aitken, 1985; Daniels et al., 1953; Fleming 1971; Huntey and Johnson, 1976; Hutt and Smirnow, 1982; Wintle , 1987; Wintle et al., 1984). The principle of TL dating is that heating (e.g., pottery) or exposure to sunshine (e.g., eolian silt) at a certain time in the past released all energy accumulated by such radioactive impurities as K-40, Th-series, and U-series, thereby setting the "energy clock" at zero. From that time on, the nuclear radiation continued to cause ionization that begin to refill existing electron traps and progressively increase TL signals. The "TL" age of a sample has been defined by Aiken (1985) as follows: Age (ka) = [Equivalent Dose (GY)]/[Dose Rate(GY/ka)]. The unit of measurement of dose is the rad (Radiation Absorbed Dose), which is defined as the absorption of 100 ergs per gram.

If a geological sample has been irradiated, with g rays, for example, and subsequently heated, light is emitted as a function of the temperature. This function curve, called a "glow curve", is characteristic of the material heated and its radiation history, i.e., the older the sample, the higher is the intensity of the glow curve. However, the relationship of peak light intensity to age is not linear, and the peak depends upon various factors. The natural TL is measured by heating the sample and recording the glow curve. The TL per unit dose is determined by subsequently exposing the sample to a known dose from an artificial radioactive source and then measuring the induced glow curve. The dose given per year can be determined by chemically measuring the radioactive content of the sample, and subsequently calculating the dose from the radioactive decay schemes of U-series, Th-series, and K-40. Several factors have to be taken into account in TL dating Quaternary sediments.

1. The Zero Point: The zero point for Pleistocene eolian deposits is not so well defined as with archaeological materials. For a sample exposed to the sun, there is a certain probability per incident photon that an electron will be evicted from its trap so that the rate of eviction will be proportional both to its probability and to the number of trapped electrons, as well as to the intensity of the light. This simple model leads to the expectation that the number of electrons remaining trapped at time t will be exponentially dependent on t; however, experimentation indicates that this is not the case and that, as the "bleaching"

proceeds, the probability of eviction decreases. Thus, there is the concept of the "easy-to-bleach" TL which is harder and harder to bleach until eventually an unbleachable residual component is left.

The time required to remove all except this unbleachable residual depends on the intensity and the spectrum of the bleaching illumination, as well as the susceptibility to bleaching of the mineral present in the sample. The practical question in evaluating paleodose is whether or not, at the time of deposition, the bleaching was sufficient to remove all but this residual component. Lu Yanchou and others (1987a) collected the surface sample where the TL samples were obtained to measure the residual component and to define the zero point. It seems a workable and promising method to handle the zero-point problem (Berger, 1988).

2. Post-depositional Changes: Several changes may occur in the material to be dated after the zero point is obtained. These changes have to be considered when interpreting the dating data; for example, ground-water may reduce or increase the concentration of the radioactivity of the radioactive isotopes and cause fluctuations of the level of the radioactivity. A further complication arises, because the decay scheme of U-series includes the gas radon-222 with a half life of 3.84 days.

Radon is chemically inert, so it may readily diffuse from the sample. As 98 percent of the γ dose rate of U-series comes from members of this series beyond radon, considerable loss of γ radiation may result. This source of error, combined with the influence of water content in the sediment, can be effectively eliminated by in situ measuring water content and properly sampling and packing to determine some of the possible loss of γ radiation (Fleming, 1971).

One uncertainty in TL dating is the moisture content of the sediment during the burial period. Moisture absorbs ionizing radiation; thus a variation in moisture content causes corresponding variation in the dose rate. A moisture content of approximately 15-20 percent was assumed for all dated sediments. This is more or less the case for deeply buried loess (Forman, 1990; Forman et al., 1989).

3. Method: In the regeneration method, all portions of the bleaching illumination spectrum are subjected to long leaching, and then artificial irradiation (β or γ) is administered to regenerate the TL growth characteristic. This method produces only a minimum estimate of the age (Berger, 1988; Forman, 1990). Partial bleaching has been considered a good method to use when the sample was shortly exposed to sunshine at

time of the deposition. In the partial bleaching procedure, the glow curve of some portions is measured after artificial irradiation, as in the additive dose procedure. But, eolian silt (loess) stays in the atmosphere or on the ground surface and is exposed to sunshine for several weeks before burial (Lu Yanchou et al., 1987a). The additive-dose procedure (modified total bleaching procedure) is most applicable (Lu Yanchou et al., 1987b); hence, the additive-dose method was used in this research and is discussed.

The additive-dose method, also called the residual method, can be used to avoid the TL sensitivity change. It is read from the intercept on a horizontal line through G_0, the latter being the TL level remaining after a long laboratory bleaching. Besides exposing the sample to sunshine for 94 hours, each sample is irradiated by using 6-7 β irradiations, i.e., NTL (Natural TL), N+β1, N+β2, N+β3, N+β4, N+β5, N+βb applied to each sample to induce a TL signal and determine the residual TL level. Although it can be established that the laboratory bleaching is long enough for only the unbleachable component to remain, there is the risk that the exposure to sunshine prior to the deposition was not sufficient to reach this level. A check of this is made by the plateau test in which the paleodose is plotted against glow-curve temperature, i.e., first the natural TL peak is determined, then the sample is irradiated at a heating rate of 10°C per second for a sufficient time until the peak (plateau) nearly completely disappears. The TL signal at the point where the peak disappears is considered as the residual component.

4. Type and Size of Grains: Feldspar is another type of TL signal-producing grain quite common in eolian deposits. The TL produced in feldspar, especially volcanic feldspar, by laboratory irradiation is unstable and decays rapidly (Wintle, 1977). Because this decay affects all regions of the glow curve and does not behave as a thermal instability-governed fading, it has been called anomalous fading. Zircon, calcite and other polymineral grains exhibit a similar anomalous fading (Aitken, 1985).

Anomalous fading causes a serious problem for TL dating because the short-lived unstable component is present in the artificial TL having mostly disappeared from the natural TL. Many studies showed that feldspars in loess exhibited 50-20 percent fading over 2-6 weeks (Berger, 1988). In unheated deposits, quartz has long been considered to be the best alternative when felspars exhibit fading; however, the TL of most sedimentary quartz saturates at a relatively low dose (Berger, 1988);

thus, an exponential, rather than a linear, relationship between the equivalent dose and the age can partially resolve this problem if the TL is close to the saturation (Forman, 1990).

Grain size is of great importance for TL dating. The average absorbed dose components to grains from external radioactivity depends on grain size because of absorption of the radiation. For this reason, when measuring the equivalent dose, use of grains having a narrow range of diameter is desirable. Furthermore, because the average range of radiation in unconsolidated sediment is about 20 micrometers, much less than 2 mm and 30 cm for β and γ radiations, respectively (Aitken, 1985), the common practice is to use either fine (2-10 micrometers) or coarse (about 100-300 micrometers) grains for equivalent dose measurements.

5. Relationship between Age and Natural Dose: For loess older than 200 ka, a controversy within the TL dating community presently exists over the appropriate method to be used for measuring the equivalent dose, the laboratory β or γ dose that produces the same TL intensity as the paleodose. Debenham (1985) was unable to obtain TL ages greater than about 120 ka when determining the age of European loess, although the loess was believed to span the age range from about 5 ka to 700 ka.

Berger (1985) dated a minimum TL age of 670 ka for the previously undated Coutlee tephra in southern British Columbia. The last interglacial events (about 100 ka) have been successfully dated by using the TL method (Berger, 1988).

Russian scientists dated terrestrial silts and sea deposits whose ages range from 1 ka to 1.2 ma. Although these dates are supported by paleomagnetic chronologies, because of lack of detailed descriptions of the laboratory procedures, the reliability of the dates has been questioned by western TL researchers (Berger, 1988).

A recent major advance in TL dating is the rigorous statistical treatment of nonlinear TL data. Empirical (Berger et al., 1987; Forman, 1988) and theoretical (Huntley et al., 1987) studies indicate that the TL growth function is most faithfully modeled as a saturating exponential. These new statistical procedures allow the extension of the TL bleaching techniques beyond a certain limit when the TL growth function is distinctively nonlinear. The nonlinear relationship between natural dose and age is in an experimental stage, but it is promising to resolve the younger TL dating problem.