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A Paleo-Climatic Reconstruction using Rock Magnetism and Stable Carbon Isotope: Bignell Hill Case, Lincoln County, Nebraska

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암석의 자장특성과 안정동위원소를 이용한 고기후의 복원

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ABSTRACT

In the loess-paleosol sequences from central Great Plains, U.S.A., variation in magnetic susceptibility, FD, NRM have been proven to be excellent proxy for paleoclimate, and the standard interpretation is that climatic processes have enhanced the rock magnetic intensities. By using mineral magnetic properties, we show the magnetic signal is due to pedogenesis during the warm and possibly wet interglacials and interstadials. other proxy records, such as stable carbon isotope and phytolith, are in good agreements with the magnetic records.

요 약

Loess를 이용한 제4기 연구는 최근에 있어서 미국 중서부 캔자스대학을 중심으로 집중적으로 이루어지기 시작하였는데 그 이유는 그레이트 플레인즈 지역에 광범위하게 퇴적되어 있는 뢰스층때문이다. 중국의 황토고원에서 연구가 진행되면서 미국내에서도 뢰스층을 통하여 제4기 기후환경 변화추적이 시도되어 온 것이다. 그러나 퇴적층내의 꽃가루, 연체동물 등의 화석의 보존 상태가 극히 불량하여 뢰스층내의 지자기와 안정

동위원소인 ^{13}C을 이용한 복원에 중점을 두어 왔으며 본 논문 내용도 그의 일환이다. 36,000년 전 (Interstadial)부터 현세까지 퇴적된 뢰스가 본 연구지역인 네브라스카주 Bignell구릉지대에서 약 80m에 걸쳐 다양한 퇴적상을 보여 주고 있다. 최종간빙기 및 빙기에 형성된 Peoria loess Brady soil에서 현재까지의 퇴적물을 분석하여 기후환경을 복원하였다. 여기에서는 암석자장을 이용한 자성반응과 ^{13}C를 이용한 식생의 복원으로 많은 샘플을 분석하여 미시적 데이타가 추출되었다.

한국과 같이 화분분석이 어려운 실정을 감안한다면 이와같은 연구를 통하여 고기후 복원에 기여할 수 있으리라고 생각된다.

INTRODUCTION

A. Study Goals

We tried to extract a high-temporal-resolution paleomagnetic record from Late Quaternary loess-paleosol sequences in the central Great Plains. High-resolution magnetic data will shed light on long-term and short-term climatic changes in the central Great Plains during the Late Pleistocene (Wisconsin) and Holocene. The paleomagnetic data together with other climatic proxy data from this area will help to reconstruct paleoenvironmental conditions in this region.

The magnetic record of loess deposits constitutes one of the most detailed and complete terrestrial records of Quaternary paleoclimate. Measurements and interpretation of the magnetic properties of loess provide corroboration of the major, long term cyclic changes of Pleistocene paleoclimates as indicated from deep-sea sediments (Kukla *et al.*, 1988). However, loess deposits also have potential to identify shorter-term global climatic features like the Younger Dryas (e.g., An *et al.*, 1992) and to provide details about local climatic conditions through time (e.g., An *et al.*, 1991; Heller *et al.*, 1993).

The loess deposits of the central Great Plains provide the ideal environment to pursue rock magnetic study. First, the central Great Plains is a relatively uncomplicated physiographic region which renders possibility of evaluation of climatic impact on magnetic record. Second, preliminary data suggest that even though there is variability in the magnetic response to climate in loess sections within the region (Grimley and Johnson, 1991; Hayward and Lowell, 1993; Feng *et al.*, 1993; Park *et al.*, 1993), there is evidence from these preliminary studies that the loess magnetic records are dominantly controlled by climate. Third, the rapid rate of loess accumulation has led to the preservation of a high-resolution record of environmental changes on the cen-tral Great Plains which represents one of the thickest and most complete loess accumulations in North America. Finally, we have growing data base about loess sources and regional climatic history recently, with which the magnetic data can be compared.

Specifically, we are conducting a detailed multi-faceted study of loess deposits in order to:

1) have a better understanding of the mechanisms by which these sediments attain climatic signals;
2) have data from climate-sensitive nonmagnetic parameters to compare with that from magnetic parameters; and
3) provide better time-stratigraphic control to assess the importance of the finer-scale variations in the magnetic record.

B. Background Information

The past decade has seen a rash of research activity concerned with magnetic record of Pleistocene loess-paleosol sequences. The magnetic susceptibility variations that can be readily measured in the field and in the laboratory closely resemble the marine oxygen isotope record (e.g., Maher and Thompson, 1991; Heller and Liu, 1986). They are therefore extremely useful in the reconstruction of Quaternary climatic changes and their periodicities. Most of the research, done on the thick deposits of loess plateau in central China, has noted a strong positive correlation between the absolute values of magnetic susceptibility and other rock magnetic parameters and the orbitally tuned oxygen isotopic record of deep sea sediments for the last 2.4 my (Heller *et al*, 1984; Kukla *et al.*, 1987; An *et al.*, 1991). The magnetic susceptibility profiles of loess sequences in China display high values in the interbedded soils, which correlate with warmer and more mesic interglacial intervals, and low values in the loess correlating with cool and dry glacial intervals. Similar positive correlations have been found from New Zealand (Pillans and Wright, 1990). However, magnetic records from other areas do not seem to show a strong relationship to paleoclimate (e.g., Hus and Geeraerts, 1986) or even exhibits an inverted relationship, i.e., the weakest susceptibility is associated with the interglacial intervals and highest susceptibility values with glacial intervals (e.g., Bēget *et al.*, 1990; Hayward and Lowell, 1993). The loess-paleosol deposits in Nebraska and Kansas for which we have derived preliminary magnetic data appear to reflect the major global climatic features of the past 120,000 years (Farr *et al.*, 1993) when compared with the Spectral Mapping Project (SPECMAP) data of Imbrie and others (1984), stable oxygen data and dust content data from the deep ice core from Dye 3 in Greenland (Dansgaard *et al.*, 1985), and Chinese loess data (An *et al.*, 1991).

Presently, the exact origin and nature of the high magnetic susceptibility of the paleosols remains an enigma. However, it is likely that several factors contribute to enhancement such as pedogenic processes, climatic changes, and extraneous detrital sources. A depositional rather than a pedogenic origin was favored by Kukla (1987) and Kukla and An (1989), who concluded that magnetic susceptibility profiles constitute a proxy climatic signal. Their model assumes a constant flux of magnetic particles from remote sources, and that susceptibility changes would result from variations in the accumulation rate of the loess. It is now generally accepted, however, that the enrichment of magnetic minerals in paleosols is due to pedogenic processes (Zhou *et al.*, 1991; Maher and Thompson, 1991; Heller *et al.*, 1991; Banerjee *et al.*, 1993; Verosub *et al.*, 1993; Feng and Johnson, in review). This is evidenced by the fact that the magnetic properties of the loess and paleosols differ, with the paleosols containing a higher percentage of ultrafine-grained ferrimagnetic particles which may be primarily magnetite (Maher and Thompson, 1991, 1992; Liu *et al.*, 1992) or maghemite (Verosub *et al.*, 1993). These particles are similar to magnetic minerals in modern soils (Maher, 1986; Singer *et al.*, 1992) and have been demonstrated to be pedogenic in origin (Banerjee *et al.*, 1993; Heller *et al.*, 1993; Verosub *et al.*, 1993).

Another view, represented by Bēget and others (1989, 1990) and Hayward and Lowell (1993), is the inverted magnetic record of loess deposits in which, as the climate shifted, changes in the extent and location of the source area as well as changes in wind patterns could have produced differences in the unweathered material incorporated in loess. The unweathered loess may therefore be enriched in magnetic minerals relative to the intervening paleosols and thus correlate negatively with the SPECMAP oxygen record. Susceptibility changes in both of these studies largely appear to reflect variability in detrital magnetite content due to changes in wind intensity and competence.

Other factors influence and complicate the magnetic record of Quaternary loess: 1) the level of dilution by nonmagnetic component such as organic matter and carbonate, 2) repeated changes in oxidation-reduction conditions being set by soil pH, climate, soil moisture and organic content (Thompson and Oldfield, 1986), 3) redeposition of loess in both water-lain and eolian loess deposits (Hus and Geeraerts, 1986), and 4) nearly continuous bioturbation which welded or bioturbated individual A horizons.

STUDY SITE

Late Quaternary paleosols and loess are exposed in a roadcut on the south valley wall of the Platte River in southwestern Nebraska, southeast of North Platte in Lincoln County. The view from the top of the roadcut provides a nice vista of the Platte River valley and the southern margin of the Sand Hill, which are believed to be main sources for late Quaternary loess in central Great Plains. The section is the type locality for the late-Pleistocene/early Holocene Brady soil and overlying Holocene Bignell loess (Table..1). Schultz and Stout (1945) originally described the Bignell Section as follows:

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

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

LATE QUATERNARY LOESS DEPOSITS IN THE CENTRAL GREAT PLAINS

The late-Wisconsin/Holocene stratigr aphy of Nebraska and Kansas includes, from oldest to youngest, the Gilman Canyon Formation, Peoria loess, Brady soil, and Bignell loess (Table.1). The Gilman Canyon Formation was first recognized as the *Citellus* zone (Schultz and Stout, 1945) because of its fossiliferous nature and abundant krotovina. Reed and Dreeszen (1965) renamed the zone the Gilman Canyon Formation. Johnson and others (1990) subsequently have recognized it as a geosol (regionally extensive and isochronous; North American Stratigr-aphic Code, 1983). Radiocarbon ages on the formation ranges from about 36 ka at the base to about 20 ka at the top (May and Soders, 1988; Johnson, 1993a). The basal age of 36 ka agrees well with the time set by Richmond and Fullerton (1986) for the beginning of the Late Wisconsin. Limited paleoen-vironmental data are emerging for the Gilman Canyon Formation. Humate-derived $\delta^{13}C$ values indicate that C4 grasses (warm, arid adaptation) were dominant for most of the time of pedogenesis (Johnson,

Table 1. Late Plestocene and Holocene stratigraphy of the central Great Plains. From W.C.Johnson (1993a).

Time Stratigraphic Units			Age (ka)	Rock and Pedostratigraphic Units	
			0		
QUATERNARY SYSTEM	HOLOCDENE SERIES			Eolian sand deposits with soils	Fluvial deposits with soils
			5	Bignell Loess	
				Brady soil (geosol)	
			10[a]		
	PLEISTOCENE SERIES	Wisconsin stage		Peoria formation (loess)	Fluvial deposits
			20[b]		
				Gilman Canyon Formation (loess and geosol)	
			50[b]		
		Sangaman stage		Sangamon soil	
			74/130[d]		
		Illinoian stage		Loveland Formation (loess)	
			190[d]		
		Pre-Illonoian stages			
			1650[f]		

Table 2. Radiocarbon ages from the Bignell Hill section.

Lab NO.	Depth	Uncorr. Age	$\delta^{13}C$	Corrected age	Unit
Tx-7425	1.80-1.85	9,110±110	-17.4	9,240±110	upper Brady soil
Tx-7358	2.15-2.20	10.580±130	-19.3	10.670±130	upper Brady soil
Beta-60,508	6.50	11,880±90	n.a.	n.a.	Peoria loess
Tx-7921	8.45`8.50	pending			Peoria loess
Tx-7922	17.45-17.50	pending			Peoria loess
Tx-7706	49.7	28,130±610	-17.6	28,250±620	GCF(loessal soil-upper)
Tx-7707	50.9-51	30,970±780	-24.3	30,980±790	GCF(loessal soil-upper)
Tx-7708	70.2-70.3	33,750±1110	-24.8	33,760±1110	GCF(alluvial phase)

All ages were determined from total humates, except for Beta-60,508 (charcoal; AMS). Tx represents Radiocarbon Laboratory at The University of Texas at Austin. Beta represents Beta Radiocarbon Lab.

1993a). Stratigr aphic and temporal equivalents of the formation have been recognized elsewhere in the midcontinent: the Roxana silt from Minnesota and Wisconsin to Arkansas and the Pisgah Formation in western Iowa (Bettis, 1990).

The dominant feature at this section is the immense thickness of Peoria loess, about 50m (cf. Eustis ash pit with c. 16 m). The proximity of the locality to the Platte River and the Sand Hills provided a tremendous source of silt and fine sand for transport southward by the north and northwest winds. This valley wall was, of course, the destination for some of that sediment. The overall accumul ation rate for the Peoria loess in this section is quite high, i.e., 50 m/10k yr $\approx$ 0.5 mm/yr. The rate of accumul-ation was, however, not likely uniform. Scant stratigraphic information from the middle and lower Peoria suggest dramatically slowed accumulation and perhaps incipient soil development. Stratigraphic breaks in Peoria loess are recognized in Illinois (Frye *et al.*, 1974) and Iowa (Daniels *et al.*, 1960). However, no paleosols are recognized in Peoria loess in Nebraska or Kansas. Frye and Leonard (1949) observed a partially leached zone lacking mollus-kan faunal remains in Peoria loess in northeastern Kansas. Lenses of plant remains have been observed in Peoria loess in Kansas (Johnson 1993) and have been interpreted as discontinuities and possible periods of landscape stability during Peoria time. The magnetic susceptibility records show that there are some periods of strong intensities, which may represent zones of incipient soil development/pre-burial weathering associated with pedogenesis or landscape stability (Park *et al.*, 1993). Magnetic susceptibility data (discussed below) from the region and Bignell Hill exhibit a high degree of correlation (Fig. 1).

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

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

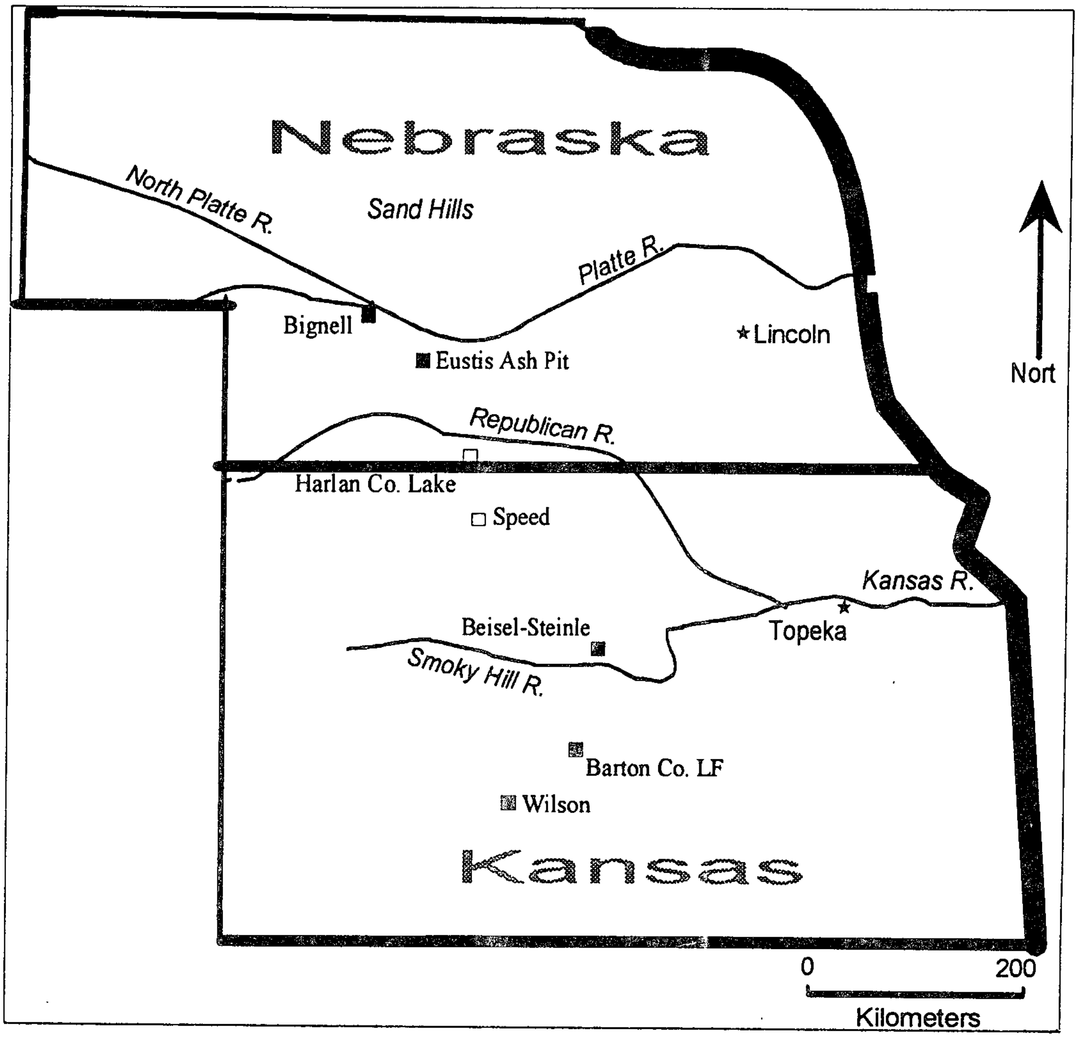

Fig. 1. Study area and adjacent region.

(Frye and Leonard, 1951). Feng (1991) noted that the Brady soil, as expressed at the Barton County Sanitary Landfill, is strongly weathered both physically and chemically.

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

The Brady soil has also been recently dated at other localities in Nebraska and Kansas. Souders and Kuzila (1990) obtained a ^{14}C age of 10,130±140 years B.P. on the Brady soil occurring within the Republican River valley of southcentral Nebraska. Sites along Harlan County Lake have yielded a number of ages, ranging from 10,550±160 to 9,020±95 years B.P., on exposures of the Brady soil (Cornwell, 1987; Johnson, 1989; Martin, 1990; Martin and Johnson, unpub. data). Two ^{14}C ages of 9820±110 (Tx-7045) and 10,550±150 (Tx-7046) years B.P. have been derived from the upper and lower 5 cm, respectively, of the Brady A horizon exposed at the Barton County site (Feng, 1991). The two radiocarbon ages from the Speed roadcut in Phillips County agree reasonably well with those from the type section in adjacent Nebraska and central Kansas (Table. 3; Fig.1).

Although it appears Brady pedoge-nesis occurred from about 10,500 to as recently as 8,500 years B.P., greater refinement of the Brady soil chronology is necessary, but present data clearly indicate it was a product of a major period of landscape stability at a time when widespread climatic shifts were occurring at the end of the Wisconsin. This was the first significant period of soil development since Gilman Canyon time, and represents the climate of the early Holocene. There is an isochronous alluvial soil found throughout the region which is particularly well expressed within the Kansas River basin (Johnson and Martin, 1987; Johnson and Logan, 1990). The two ages of 8274±500 (C-

Table 3. Radiocarbon ages from the Brady soil of Nebraska and Kansas.

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

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

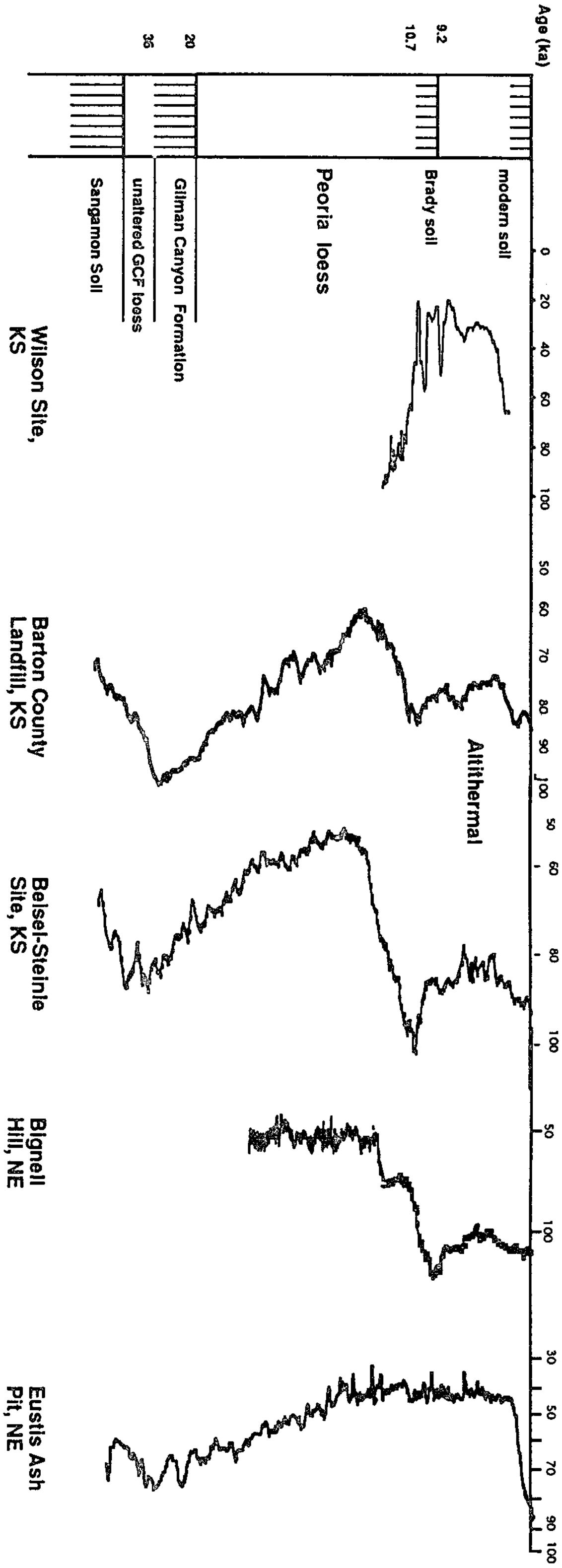

Fig. 2. Comparison of low-field susceptibility along the transect. The Gilman Canyon geosol and Brady soil well represented in this curve. Note that the Eustis ash pit does not have the Brady soil(eroded or welded to the modern soil)

108 a) and 9880±670 (C-471) years B.P. determined from alluvial fill (Fill 2A) at archaeological sites Ft-50 and Ft-41 on Harry Strunk Lake in southwesternNebraska (Schultz *et al.*, 1951; Libby, 1955) were the first radiocarbon determinations on the Brady soil. The soil, occurring in both eolian and alluvial contexts, qualifies, based upon present ^{14}C data, as a *geosol* (regionally extensive and isochronous; North American Stratigraphic Code, 1983) .

Development of the Brady soil correlates well with indicators of regional climatic change. The fossil pollen record at Muscotah Marsh of northeastern Kansas indicates that spruce had essentially disappeared from the region by about 10,500 yr B.P. As this decline occurred, deciduous tree species increased until about 9,000 yr B.P., the time at which grassland expansionz began (Grüger, 1973). On a hemispheric scale, the abrupt decrease in atmospheric dust noted in the Greenland ice core at 10,750 yr B.P. (Paterson and Hammer, 1987) reflects decreased loess deposition and possibly Brady-age pedogenesis associated with relative terrestrial stability. Further, ^{18}O levels within the same core suggest rapid warming about 10,750 yr B.P., with the characteristic Holocene temperature regime being established about 9,000 yr B.P. At least the early and perhaps all of Brady soil forming interval coincide with the Younger Dryas cold interval of the North Atlantic region (Table. 4). Here at the Bignell Hill type section, the beginning of Brady soil forming interval matches well with the oxygen isotope data from Greenland Icecore (Dansgaard *et al.*, 1985). Also, there is a minor, but notable kink in the susceptibility data just below the Brady soil that may indicate climatic degradation comparable to Younger Dryas cold spell (Fig.3). A rapid rate of dust accumul ation, approaching 1 m/1000 yr in some profiles on the Loess Plateau in China has led to the preservation of a high-resolution record of Late Quaternary environmental changes (An *et al.*, 1993). They found a minor fluctuation which is believed to represent the Younger Dryas cold spell.

The Bignell loess, first described at this site, is typically a gray or yellow-tan, massive silt, calcareous and seldom more than 1.5 m thick. Although it is often somewhat less compact and more friable than the underlying Peoria loess, no certain identification can be made without the presence of the Brady soil. The Bignell loess does not form a continuous mantle on the Peoria; instead, it occurs as discontinuous deposits which are most prevalent and thickest adjacent to modern-day

Table 4. Chronology of the Younger Dryas.

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

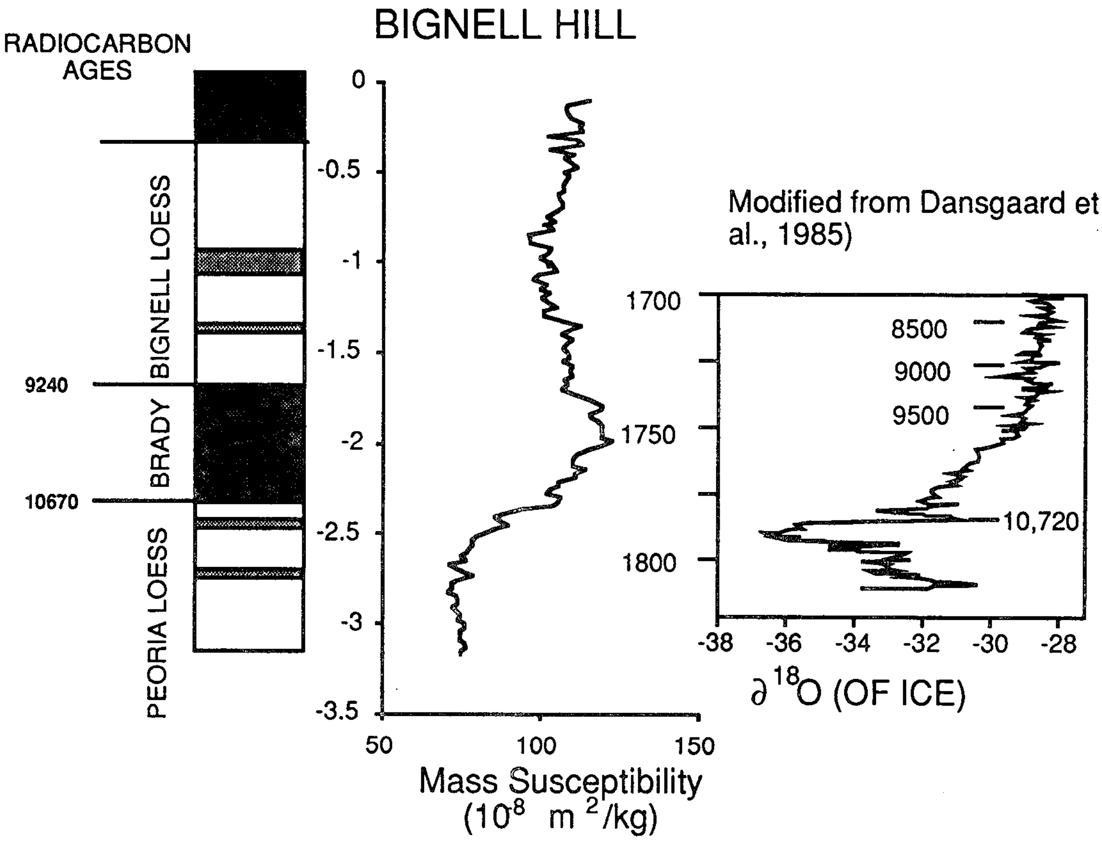

Fig. 3. Mass susceptibility of the uppermost Peoria loess, Brady soil, and Bignell loess at Bignell Hill and the δ^{18}O values from a Greenland ice core.

valleys, particularly the south side, and often within depressions on the Peoria surface, such as here at Bignell Hill. Feng (1991) speculates that the Bignell loess of central Kansas is relatively well weathered because it was derived from a pre-weathered source, the Brady soil surface, perhaps eolian and alluvial phases alike. This is consistent with the earlier interpretation derived in Nebraska that Bignell loess is at least partially comprised of re-worked Peoria loess (Condra *et al.*, 1947, p. 33).

It appears from the ^{14}C-age determ-inations obtained here at the type section in Nebraska and the Speed roadcut that the Bignell loess can be no older than about 8,000 yr B.P. Snails collected by A.B. Leonard from the lower part of the Bignell in Doniphan County, northeastern Kansas, produced ages of 12,500 $\pm$400 (W-231) and 12,700$\pm$300 (W-233) yr B.P. (Frye and Leonard, 1965). Because the shell material had absorbed an indeterminate amount of dead carbonate, Frye and others (1968) proposed an averaged age of approximately 11,000 years. Based upon the age data available for the Brady, the soil humate-derived ages are probably closest to reality.

A pronounced feature of the Holocene climate of the Plains was an extended warm, dry period (Wright, 1970; Benedict and Olson, 1978; Barry, 1983), identified as the Altithermal

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

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

ROCK MAGNETISM AND CLIMATIC IMPLICATION

Iron oxides, iron sulphide and manganese oxides are the main carriers of magnetism, typically make up less than 5% by mass of unconsolidated geologic materials. The significance of magnetic materials, however, is out of all proportion to their frequency of occurrence. Magnetic minerals are near-ubiquitous in the natural environment and highly sensitive to changes in environmental conditions. They are therefore extremely useful paleoenvir onmental indicators. Unfortunately, it is difficult to separate magnetic minerals in a pure form for investigation, particularly if they are fine-grained. An alternative procedure, therefore, is to study the bulk magnetic properties of minerals. Perhaps the most useful of these bulk properties, and certainly the most widely deter mined, is magnetic susceptibility. A remnant magnetization is induced in materials in strong applied fields. Susceptibility is therefore measured in very weak fields (less than 1 m Tesla field strength and less than 1 kHz). However, susceptibility also varies with the frequency of the applied field.

A great deal of attention has been given to paleomagnetic measurements of loess-paleosol sequences, particularly in China and in Europe. Although much research has focused on paleoma gnetic events such as excursions, magnetic data have been used to successfully reconstruct the climatic sequences. For a comprehensive understanding of the magnetic charac teristics of the paleosols and loessal parent material, three bulk magnetic parameters have been selected: magnetic susceptibility, frequency depend ence, and natural remanence magneti zation. Contiguous samples were coll ected (40-50samples/m) from the upper 18 m of the Bignell road cut using demagnet ized, plastic cubes. Bulk magnetic characteristics were measured using a Bartington MS-2 type dual frequency susceptibility sensor, pulse magnetizer, and cryogenic magnetometer (using liquid helium).

Magnetic susceptibility measures magnetization temporarily induced in a rock by an artificially applied, low-amplitude magnetic field. Le Borgne (1955) was able to show, and many subsequent studies have confirmed, that the magnetic susceptibility of topsoil is often higher than that of the underlying material. Le Borgne ascribed this to the formation of secondary ferrimagnetic oxides within the clay size fraction of the soil. The processes contributing to this may be considered under the general heading of magnetic enhancements. Common to all the processes is the conversion of iron from non-ferrimagnetic to ferrimagnetic forms.The strength of the susceptibility signal depends on the concentration and grain size of the magnetic minerals. Magnetic susceptibility measured at Bignell Hill indicates that magnetic intensities are strong in the Brady Soil, in which susceptibilities are nearly twice higher than those of the unweathered Peoria loess and higher than in the modern soil. Susceptibility plots from the Eustis ash pit and Bignell Hill correlate well despite the distance from each other (Fig. 2), thereby reflecting the regional climatic control. The exact origin and nature of the high magnetic susceptibility (χ of the paleosols remains an enigma. It is very likely that several factors contribute to the enhancement such as soil-forming processes, climatic changes, extraneous sources etc. and the main problem is to evaluate each of them or at least to single out the most important ones.

Along with most authors of recent rock magnetic studies of Chinese loess and paleosol, we subscribe to the view that the observed susceptibility fluctuations are not due to a mere concentration of a constant, uniformly-sized magnetite "rain" from the troposphere during warm interglacial or interstadials-periods that were characterized by reduced inputs of less-magnetic wind-borne loess (Kukla, 1988). Rather we accept the overwhel-ming evidence (Hus and Han, 1992; Maher and Thompson, 1991: Maher and Thompson, 1992; Zhou et al, 1990) that during the interglacials and interstadials, higher temperature, and higher rainfall are responsible for the alteration of non-magnetic (paramagnetic) iron-bearing silicate minerals such as feldspars and clays to strongly magnetic (ferrimagnetic) ultrafine grains of magnetite (Fe_3O_4) or slightly oxidized cation-deficient magnetite, compositionally intermediate between magnetite and maghemite (γFe_2O_3) (Singer and Fine, 1989).

Frequency dependence measures the presence of grains lying at the stable single domain/superparamagnetic size boundary (c.a. 0.03 μ dia.). This measure can effectively identify weathering and soil profile development, which gives rise to increased concentrations of extremely fine grained ferrimagnetic crystals. Mass susceptibility data

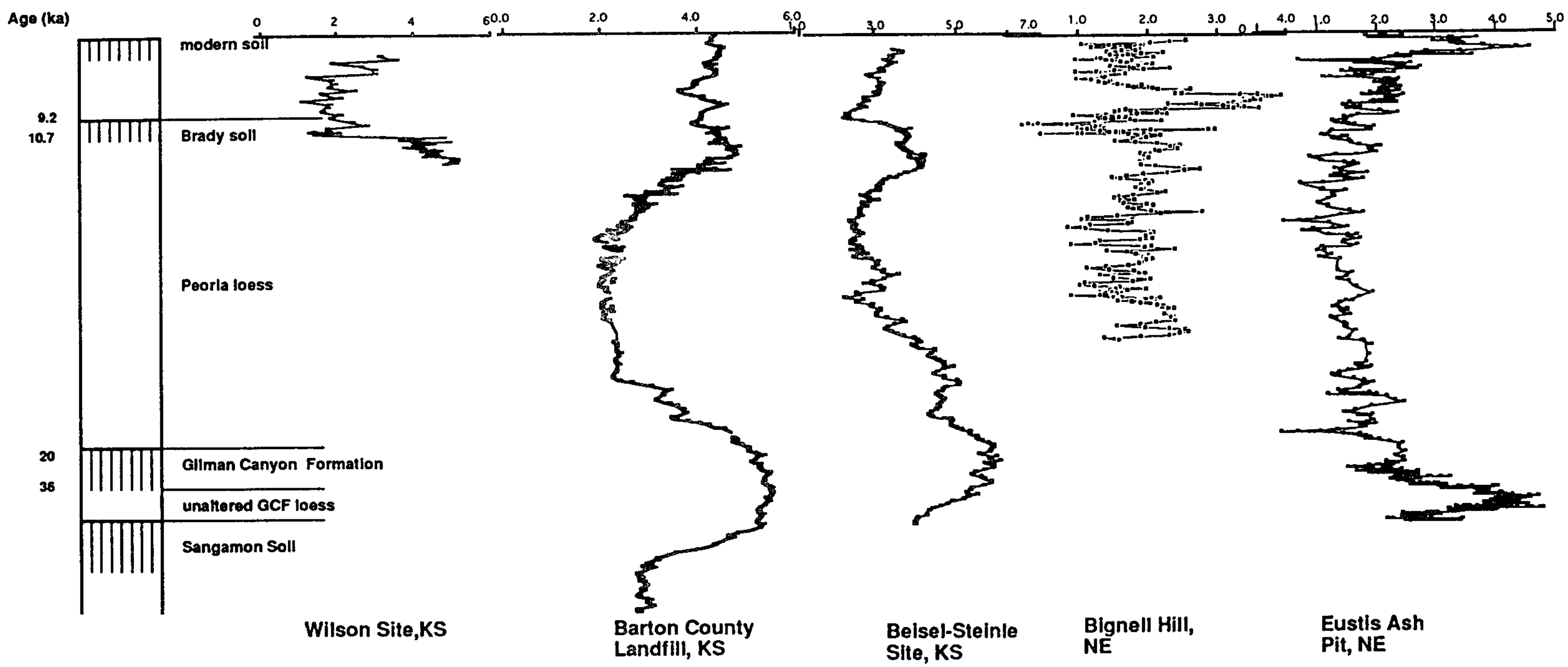

Fig. 4. Frequency dependence along the transect. The Gilman Canyon geosol exhibits the highest proportion of superparamagnetic grains. Brady soil exhibits lower values than Gilman Canyon geosol in spite of its soil development. The Peoria loess, exhibits lowest values.

correlate well with frequency dependence (Fig. ?). From the magnetic point of view, the Bignell loess is closer to Brady soil than Peoria loess. Susceptibility intensities are similar to the Brady soil, and there are no major differences in the frequency dependence data. The Brady soil can be differentiated from Sangamon soil or Gilman Canyon Formation soil by examining the frequency dependence and high-frequency susceptibility (Table. 5). Susceptibility intensities of the Brady soil and the Sangamon soil are similar to those of the modern soil, but the frequency dependence of the Sangamon soil is that of the unweathered loess. The dominant cause of large changes in susceptibility with frequency is instead likely to be associated with the viscous effects of ferri- and ferromagnetic grains lying close to the superparamagnetic/stable single-domain boundary. As the frequency of the applied field increases, the superparamagnetic/stable single-domain boundary shifts to smaller volumes . Thus, at higher frequencies, a certain proportion of grains will no longer contribute to susceptibility as superparamagnetic grains but as stable single-domain grains. Susceptibility will therefore be lower when measured in higher-frequency applied fields. Changes in susceptibility with frequency will thus provide information on the grain sizes of the magnetic carriers in samples. If there is no change in susceptibility with frequency, then none of the magnetic carriers in the sample can lie close to the superparamagnetic/stable single-domain boundary; in other words, the magnetic grains are likely to be relatively coarse and of single domain and multi-domain type. On the other hand, if there is a large change in susceptibility with frequency, then a large proportion of the magnetic carriers in the sample must lie close to the superpara-magnetic/single-domain boundary; in other words, the magnetic grains are likely to be rather fine.

Weathering and pedogenesis are best expressed in the NRM (Natural Reman-ence Magnetization) data (Fig. 5). NRM measures the residual magneti-zation possessed by rocks, sediments and soils. NRM intensities in Brady soil are 3-4 times larger than in the unweathered loess. The intensities of NRM

Table 5. Composite low frequency susceptibility and frequency dependence mineral magnetic data.

Start. Unit	Low Field Susceptibility	L.F Dependence	Frequency Dependence	FD stand. dev.
Modern Soil	70.11	10.85	3.01	1.19
Bignell Loess	104.00	4.11	1.58	0.63
Brady Soil	112.75	5.90	3.14	0.75
Peoria Loess	48.28	5.41	1.93	0.87
GCF	74.82	5.37	3.32	0.76
Unaltered GCF	70.35	5.77	2.12	0.62
Sangamon	90.02	9.05	1.57	0.4
Illinoian Complex	91.12	5.58	1.34	0.2

Values for the modern soil, Bignell loess and Brady soil are from Bignell Hill section, and Peoria loess and underlying units from the Eustis ash pit.

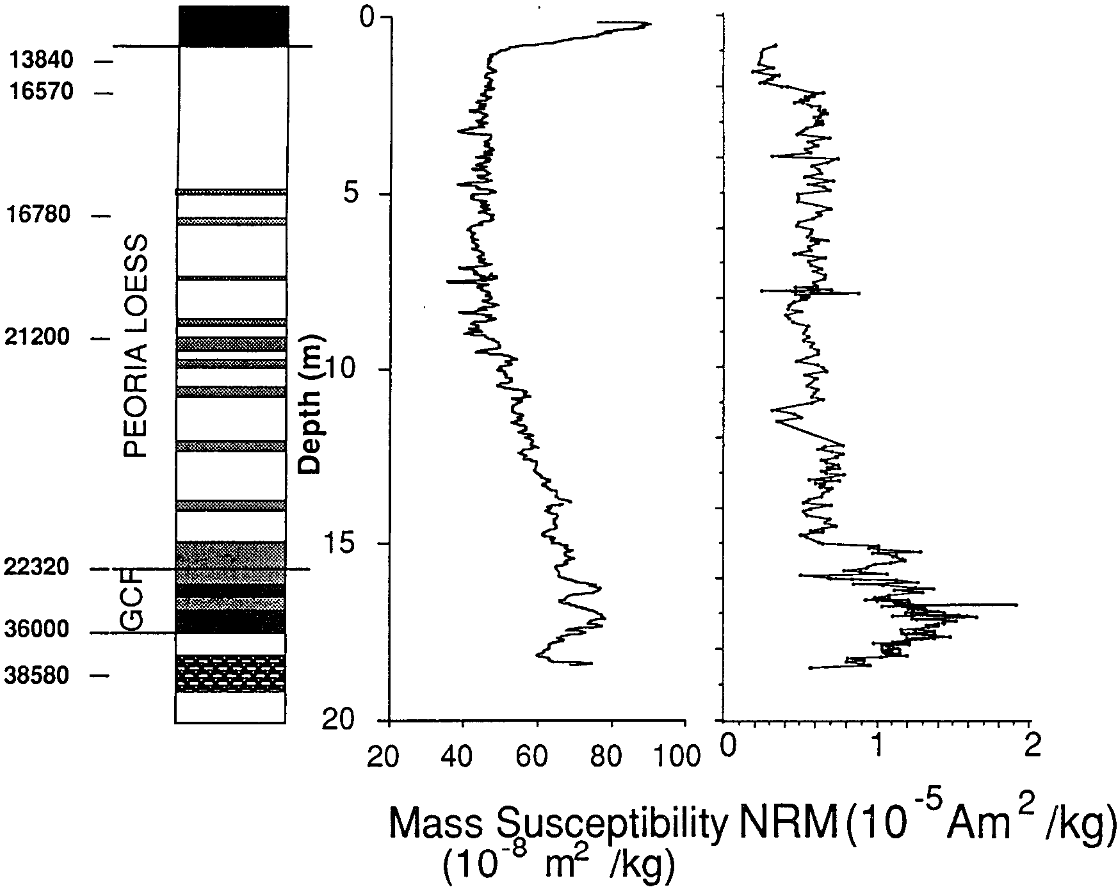

Fig. 5. Mass susceptibility and natural remanence magnetization (NRM) for the uppermost Peoria loess, Brady soil, and Bignell loess.

and low field susceptibility vary by 2-3 orders of magnitude throughout the section, being closely correlated to one another. Higher intensities characterize the Brady soil, whereas low values are typical of the unweath-ered Peoria loess. Also, NRM data differentiate between the Brady soil and Bignell loess (Fig. 5). These features are ascribed to the pedogenic processes, which promote magnetic mineral enrichment in the soil due to alteration in a relatively warm climate (Heller and Liu, 1982; Liu *et al.*, 1992). The exact acquisition mechanism of the natural remanent magnetization (NRM) in loess deposits and exact origin of the enhanced magnetic properties of the interbedded paleosols remain unclear. While Heller and Liu, are in favor of chemical origin for the most stable or characteristic NRM component, others like Sasajima *et al.*(1984) think that the NRM is a depositional or post-depositional remanence (DRM or PDRM).

We have conducted modified Lowrie-Fuller and Cisowski tests on 10 samples. Acquisition of IRM to a maximum field of 1.2 tesla is followed by sequential AF demagnetization of IRM. The intersection point of two curves projected to X axis is considered to be an approximation of the coercive force (Hrc) field. This experiment indicates that the grain sizes from both paleosols and loess ranges between correlate well with frequency dependence (Fig. ?). From the magnetic point of view, the Bignell loess is closer to Brady soil than Peoria loess. Susceptibility intensities are similar to the Brady soil, and there are no major differences in the frequency dependence data. The Brady soil can be differentiated from Sangamon soil or Gilman Canyon Formation soil by examining the frequency dependence and high-frequency susceptibility (Table. 5). Susceptibility intensities of the Brady soil and the Sangamon soil are similar to those of the modern soil, but the frequency dependence of the Sangamon soil is that of the unweathered loess. The dominant cause of large changes in susceptibility with frequency is instead likely to be associated with the viscous effects of ferri- and ferromagnetic grains lying close to the superparamagnetic/stable single-domain boundary. As the frequency of the applied field increases, the superparamagnetic/stable single-domain boundary shifts to smaller volumes . Thus, at higher frequencies, a certain proportion of grains will no longer contribute to susceptibility as superparamagnetic grains but as stable single-domain grains. Susceptibility will therefore be lower when measured in higher-frequency applied fields. Changes in susceptibility with frequency will thus provide information on the grain sizes of the magnetic carriers in samples. If there is no change in susceptibility with frequency, then none of the magnetic carriers in the sample can lie close to the superparamagnetic/stable single-domain boundary; in other words, the magnetic grains are likely to be relatively coarse and of single domain and multi-domain type. On the other hand, if there is a large change in susceptibility with frequency, then a large proportion of the magnetic carriers in the sample must lie close to the superpara-magnetic/single-domain boundary; in other words, the magnetic grains are likely to be rather fine.

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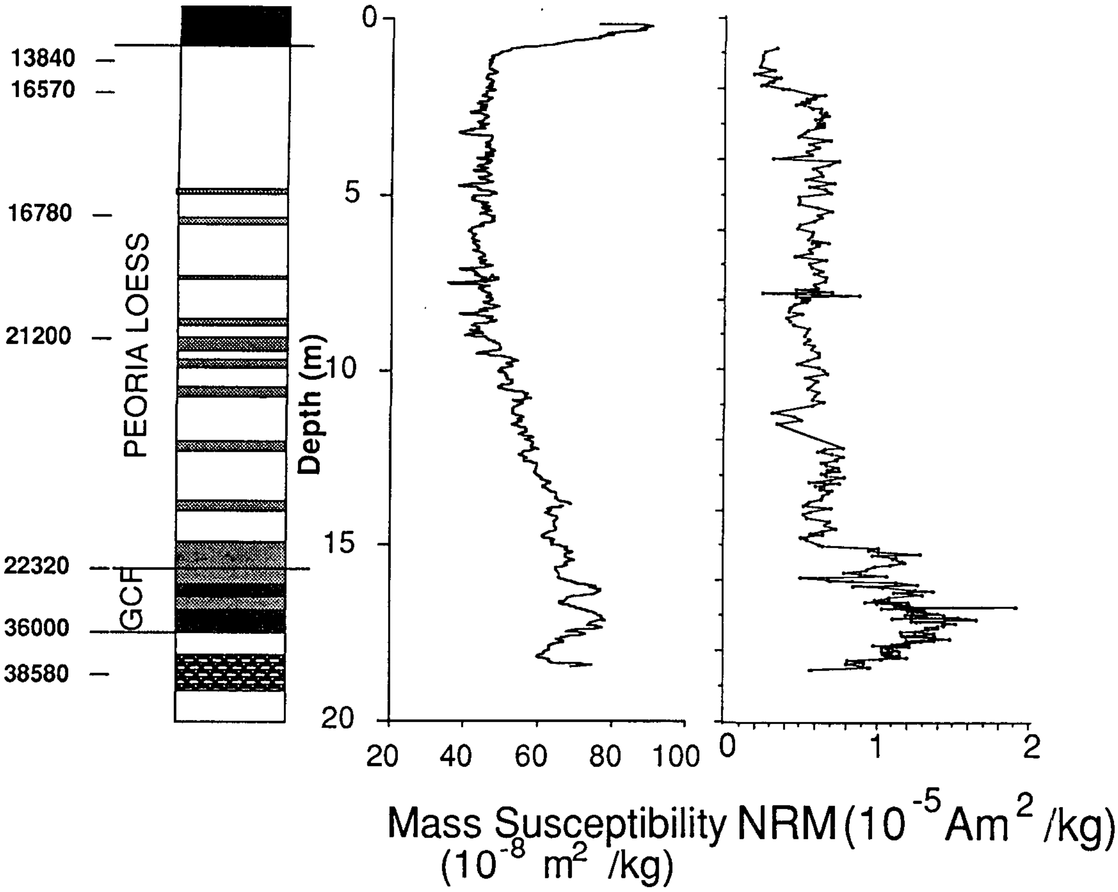

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If it can be shown that the magnetic iron oxides in common with other iron forms are responding to and reflective of soil-forming processes, mineral magnetic techniques may provide a new means of access to this pedogenic information. In Chinese loess deposits, magnetite, maghemite and hematite are present to varying extent in both loess and paleosols. In this study, we will differentiate the lower coercivity phases magnetite and maghemite from the high coercivity, high blocking temperature phase hematite using two techniques. Both methods use the isothermal remanence magnetization (IRM) which is the magnetic remanence acquired from the direct application of a strong magnetic field.

The IRM acquisition of loess and paleosol samples typically show steep acquisition to a value of 0.1 T and complete saturation at about 0.25 T (Fig. 6). A crude estimation of high coercivity material (predominantly hematite which only acquire magnetic remanence at high applied fields.) will be determined by measuring S ratios (IRM300mT/SIRM; Maher and Thompson, 1991: They used 100 mT instead of 300 mT. The bulk of the higher coercivity portion of IRM is from magnetite and maghemite, not hematite. Therefore, S ratios with a 100 mT measurement are not particul-arly useful to differentiate hematite contribution.). Lower S ratios indicate increasing proportion of imperfect antiferromagnetic materials which only acquire magnetic remanence at high applied fields.The examples given (Fig. 6) have S ratios of 0.76 and 0.82 respectively, which are very similar to Chinese data.

The more accurate determination of hematite contribution will be made for selected samples using thermal demagnetization characteristics of IRM using the technique of Lowrie (1990). Three different IRMs are imparted to specimens along three mutually perpendicular directions, each direction essentially corresponding to a different coercivity spectrum. In both loess and paleosol specimens analyzed so far, suggest the presence of magnetite, and maghemite. The high coercivity mineral remaining beyond 600 ℃ is likely natural hematite. The low and intermediate coercivity fractions show significant decrease between 250℃ and 400℃, possibly indicating the conversion of maghemite to hematite, or perhaps indicating the presence of magnetite with a low-blocking temperature spectra. The drop in IRM between 250℃ and 400℃ is more evident in paleosols suggesting that maghemite may be more important in the paleosols.

This phenomenon is most noticeable at Brady soil and Bignell loess. Maghemitization may have destroyed Superparamagnetic (SP) grains. Heller and others (1991) report a similar drop in IRM between 200℃ and 400℃ in Chinese samples and concluded that maghemite was insignificant.

We do not know whether the ultra-fine grained ferromagnetic mineral in the loess and paleosol sequence is magnetite or maghemite. However, Banerjee and others (1993) have shown that this component is definitely pedogenic in origin. Mössbauer spectro-metry is the only direct way of discriminating between magnetite and maghemite (Verosub, 1993). However, selective extraction technique using citrate-bicarbonate-dithionite (CBD) has been proved to be an effective way of removing pedogenic ferrimagnetic grains (primarily maghemite) and that it leaves untouched essentially all of the ferrimagnetic grains inherited from the soil parent material which is

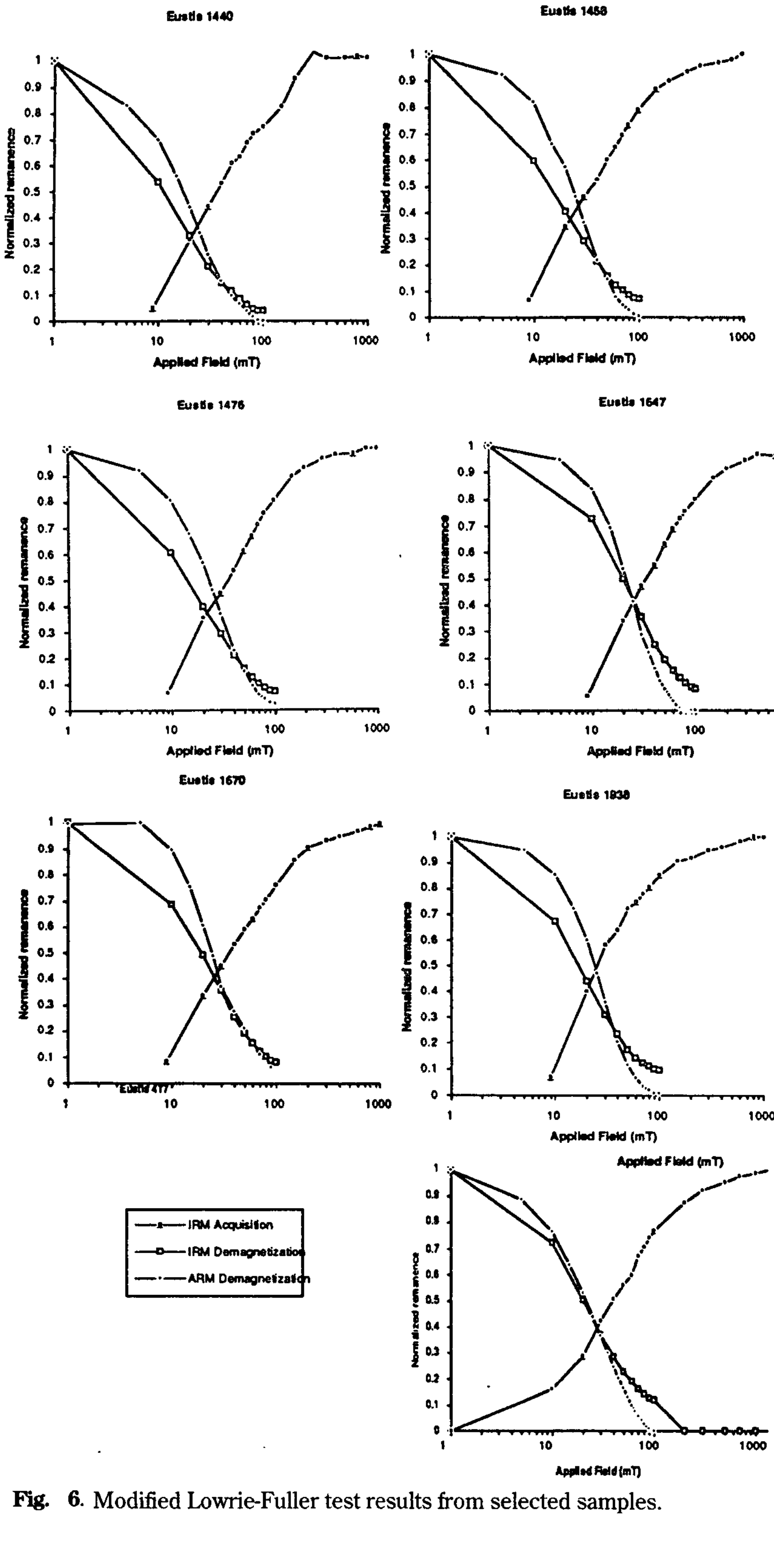

Fig. 6. Modified Lowrie-Fuller test results from selected samples.

primarily magnetite (Fine and Singer, 1989; Fine *et al.*, 1989; Singer *et al.*, 1992; Verosub *et al.*, 1993). The susceptibility measurements from pre- and post-CBD extraction method will be used as an indicator of the amount of maghemite. This test and Mössbauer spectrometry for selected samples will be done later.

CARBON ISOTOPE RECORDS

There are several quantitative techniques in use today for paleoeco logical reconstructions in terrestrial depositional systems. However, one recently developed approach to such reconstructions which appears to have extraordinary application to loess deposits of the central Great Plains is that of estimating the proportion of C3 (cool, moist adaptation) to C4 (warm,dry adaptation) plants once present at a site using carbon isotope values from paleosol humates. It is well established that there are three different photosynthetic pathways among plant species: C3 (Calvin-Benson cycle), C4 (Hatch-Slack cycle) and CAM (Crassulacean Acid Metabolism). These produce organic matter with carbon isotope composition ($^{13}C/^{12}C$) of - 24 to -34, - 9 to -16, and - 9 to - 19 ‰respectively. Diagenesis does not seem to have altered the original isotopic composition of organic matter and pedogenic carbonate, thus confirming their use in paleoecological reconst-ruction (Kelly *et al.*, 1989). Isotopic composition of soil organic matter and pedogenic carbonate in high-respiration rate soils is a direct indicator of the fraction of the biomass using the C3 or C4 photosynthetic pathways. Paleosol humus probably represents organic matter form the last few hundred years before burial, given the short residence times typical for humus in most modern soils (Birkeland, 1984).

The percent C4 plants seems to be a good climatic indicator based on modern studies. According to Teeri and Stowe (1976), percent C4 is strongly correlated with the July minimum temperature in continental United States. The normal July minimum temperature was a better predictor of the percent C4 than either the normal July average temperature or the normal July maximum temperature. Based on their analysis, the percent of C4 species in a grass flora in the continental United States is most accurately predicted by a linear combination of the normal July minimum temperature, mean annual degree-days and the log of the length of the freeze-free period.

Though $\delta^{13}C$ values are a potential source of proxy data for vegetation type and hence climate (Krishnamurthy *et al.*, 1982), studies employing $\delta^{13}C$ determinations of soil organic matter for paleoclimatic reconstruction are few. When determinations are derived from the organic fractions of the soil, they reflect inputs by the plants, particularly grasses, growing on those surfaces. $\Delta^{13}C$ values acquired in association with ^{14}C ages for the Peoria loess in Kansas and Nebraska indicate that C3 plants were dominant for most of Peoria time (Johnson, 1993a.). This reflects the cooling associated with the glacial maximum occurring in Peoria time at about 18 ka. Conversely, C4 plants were dominant for most of the Gilman Canyon time of pedogenesis (36-21 ka), and during the latter develop- ment of the Brady soil and the entire Holocene (Fig. 7). This indicates that vegetation and thus climate during Gilman Canyon time was similar to the present warm, subhumid to semiarid conditions of the central Great Plains today (Johnson, 1993a). Isotopic data seem to agree with limited phytolith data from the same area (Fredlund *et al.*, 1985; Johnson *et al.*, 1993).

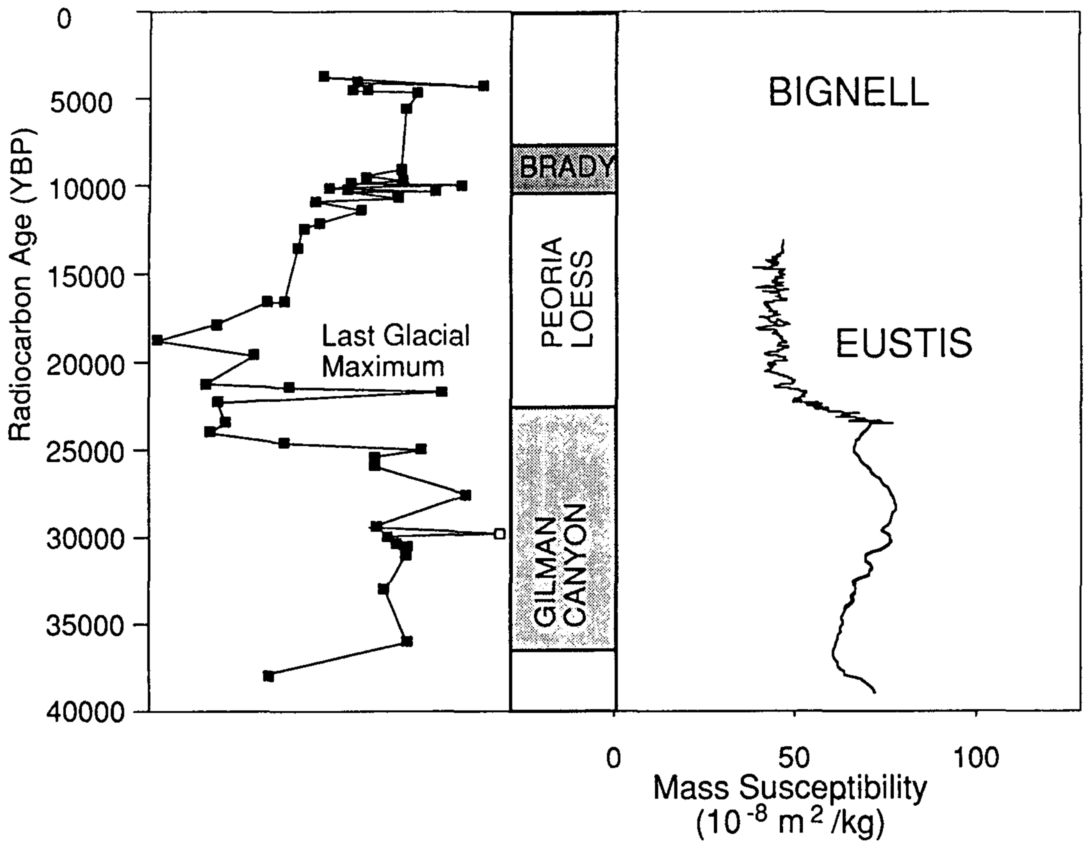

Fig. 7. $\Delta^{13}C$ shifts during the Late Quaternary(Wisconsin and Holocene) from W.C Johnson,(unpublished data).

SUMMARY

A detailed rock magnetic record has been collected from Bignell Hill,Nebraska. The stratigraphy consists of the Sangamon soil, Gilman Canyon Formation, Peoria loess, Brady soil, Bignell loess, and modern soil. IRM (Isothermal Remanence Magnetization) data indicate a mixture of magnetite with minor maghemite in both the loess and major paleosols. Magnetic susceptibility ranges from 70 to 130 x $10^{-8}m^3/kg$ in paleosols. Susceptibility intensities in the Peoria loess are weak due to dilution by the presence of pedogenic carbonates and/or to high accumulation rates. Good correlation between frequency dependence and mass susceptibility is obtained only above the Sangamon soil. Frequency dependence is highest in the Gilman Canyon Formation soil (6%) and lowest in the relatively unweathered loess (2%). Below the Sangamon soil, frequency dependence is uniformly low (<3.5%) suggesting either diagenetic destruction of superparamagnetic magnetite or the lack of original superparamagnetic grains in soils developed prior to the Gilman Canyon Formation soil.

The Bignell loess and Brady soil have similar bulk magnetic characteristics, which suggests that Bignell loess is reworked Brady soil. Susceptibility intensities increase steadily after the Last Glacial maximum. At two sites (Barton County landfill and Bignell Hill), however, minor but

distinct drops in susceptibility intensity during the Late Pleistocene interrupt the inferred general warming trend. This may indicate climatic deterioration comparable to Younger Dryas cold spell of Amphi-Atlantic region. The magnetic susceptibility and NRM record in the central Great Plains correlates well with the marine oxygen isotope record, as well as the magnetic susceptibility record of Chinese loess for the past 150,000 years.

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