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Paleoclimatic Reconstruction in the Central Great Plains Using Environmental Magnetism and Stable Isotope

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자성과 동위원소를 이용한 중부대평원의 고기후 복원

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Abstract : The magnetic record of loess deposits may be one of the most detailed and useful records of Quaternary climate change on the continents. Stratigraphic variations of magnetic parameters define alternating zones of high and low concentrations of magnetic minerals. All the concentration-sensitive magnetic parameters show an increase within the interstadial Gilman Canyon Formation and interglacial Brady soil and a systematic decrease within the Wisconsinan Peoria loess. The influence of climate change on magnetic records is confirmed by a high correlation between the magnetic parameters and biological proxies. Rock magnetic data appear to be better correlated with temperature-sensitive biological proxies than does a precipitation-sensitive index such as the aridity index derived from opal phytoliths. Simultaneous, higher resolution sampling of magnetic and biological proxies proved to be a better sampling tactic, and enhanced the feasibility of rock magnetic parameters as independent climate proxies.

Key Words : Environmental magnetism, Paleoclimate, Loess, Stable isotope, Proxy

요약 : 화분분석이나 연륜분석 같은 전통적인 고기후 연구 방법론이 이용되기 어려운 육상 환경하에서, 토양 내 자성 특성은 제4기 기후 연구의 수단중 가장 좋은 수단 중 하나가 되었다. 본 연구에서는 미국 중부 평원지역에서 제4기말 위스컨신 빙하기와 홀로세에 퇴적된 고토양/뢰스 층에서 자성변수들의 층서에 따른 대자율과 잔류자기의 특성을 연구하였다. 고토양에는 자성물질 집적이 많고, 잔류자기의 특성도 토양화의 영향으로 크기가 작은 2차 자성광물의 존재를 보이는데 비하여, 빙하기의 풍화정도가 약한 뢰스에서는 집적량이 적고 입자가 큰 1차 광물이 많은 것으로 나타났다. 자성변수들은 기후에 민감한 것으로 알려진 다른 생물학적 지표들과의 높은 상관관계를 통해 기후변화를 잘 반영하는 것으로 나타났다.

주요어 : 환경자기, 고기후, 뢰스, 안정동위원소, 기후 지표

1. Introduction

Less is known about environmental conditions in the central Great Plains during the Late Pleistocene and Holocene than about many other regions of North America. This is probably because wide application of the more traditional investigative tools such as palynology and dendroclimatology is very difficult. Consequently, what is known has been and is being derived using some of the newer approaches to late-Quaternary environmental reconstruction such as stable isotope analysis, opal phytolith analysis, and environmental magnetism. Those sedimentological contexts being explored include loess, eolian sand sheets and dunes, alluvial fills, and to a lesser extent isolated lake and peat deposits. The loess deposits of the region, which represent some of the thickest and most complete loess accumulations in North America, hold the potential to provide a particularly promising avenue for the pursuit of the paleoenvironmental record.

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Due to its relative youth, the Late Wisconsinan Stage has the greatest chronostratigraphic resolution. Based on the chronology from Illinois, five substages of the Wisconsin have been traditionally recognized: the Altonian (70,000-28,000 yr B.P.), Farmdalian (28,000-22,000 yr B.P.), Woodfordian (22,000-12,500 yr B.P.), Twocreekan (12,500-11,000 yr B.P.), and Valderan (11,000-5,000 yr B.P.) (Willman and Frye, 1970). This chronology of substages has, however, limited stratigraphic application in Nebraska, Kansas and eastern Colorado, and has therefore not been adopted literally.

In Nebraska, Reed and Dreeszen (1965) identified four Wisconsinan units: the Gilman Canyon Formation (an upland loess with soil development), Peoria Formation (fluvial sand and silt in valleys and loess on the uplands), Brady Interstadial soil, and Bignell Formation (dune sand and loess). For the Wisconsin of Kansas, Frye and Leonard (1952) recognized early Wisconsinan alluvial deposits and the Sanborn Formation. The late Wisconsinan units of the latter include the Peoria loess, Brady soil and Bignell loess (Figure 1). Since these early statements of stratigraphic succession, the Bignell loess has been assigned to the Holocene.

During the 1960s, the record of past climate was based primarily on continental deposits, but these were rarely continuous sedimentary records, and consequently the picture of past climatic variations that developed was incomplete (Bradley, 1985). In the next decade, studies of marine sediments revolutionized our understanding of climatic

Time-stratigraphic units	Rock-stratigraphic units					
	Northeastern area		Southeastern area		Central and Western area	
Recent stage	Eolian and fluvial deposits					
Wisconsinan Stage	Bignell Formation	Fluvial deposits	Bignell Formation	Fluvial deposits	Bignell Formation	Fluvial deposits
	Brady Soil					
	Peoria Formation	Fluvial deposits	Peoria Formation	Fluvial deposits	Fluvial deposits	Fluvial deposits
	GCF					
Sangamonian Stage	Sangamon Soil					
Illinoian stage	Loveland Formation	Fluvial deposits	Loveland Formation	Fluvial deposits	Loveland Formation Crete Formation	
	Yarmouth Soil					
Pre-Illinoian	Loess	Fluvial deposits	Fluvial deposits		Sappa Formation	
	Cedar Bluffs Till					
	Fluvial deposits				Grand Island Formation	
	Nickerson Till					
	Atchinson Formation					
	Atfon Soil					
	Loess	Fluvial deposits	Fluvial deposits		Fullerton Formation	
	Iowa point Till				Holdredge Formation	
	David City Formation					

Figure 1. Late-Quaternary stratigraphic succession in Kansas(Bayne and O'Connor, 1968).

variations and enabled models of the causes of climatic changes to be tested. Undoubtedly, studies of marine sediments have provided data bases which continue to expand in quantity and quality (Ruddiman, 1985). However, the 1980s have seen a renewed focus on continental records of climate, which complement the perspective provided by marine sediment (COHMAP members, 1988). Continental deposits often provide more detailed information about short-term (high-frequency) changes of climate than do most marine records.

2. Loess Stratigraphy and Environment

Late Wisconsinan loess deposits form a mantle over much of the upland surface of the region, covering the central Great Plains and provide a terrestrial record of late Quaternary climate. From Nebraska, loess deposits in the central Great Plains extend west across eastern Colorado and south across most of Kansas (Figure 2). The thickest deposits of loess are adjacent to and underlie the Nebraska Sand Hills (Kollmorgen, 1963; Ahlbrandt *et al.*, 1980). The oldest laterally extensive loess unit in the region is the Loveland loess, found as far west as the Colorado/Kansas state line. In Kansas, exposures of Loveland loess are patchy and are found mostly on ridges near drainages (Welch and Hale, 1987). The Peoria loess is the thickest and most laterally continuous loess deposit (Frye and Leonard, 1951), whereas the overlying Holocene Bignell loess is found discontinuously in the central Great Plains and is not identified east of the Missouri River.

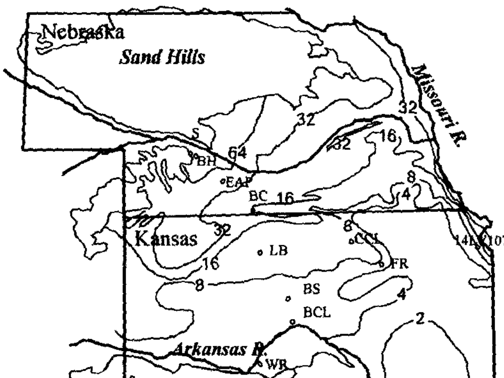

Figure 2. Loess isopatcheous map and locations of magnetic sampling sites

(SS=Sargent, BH=Bignell Hill, EAP=Eustis ash pit, BC=Bone Cove, LB=La-Barge ash pit, CCL=Cloud County landfill, BS=Beisel-Steinle, BCL=Barton County landfill, WR=Wilson Ridge, FR=Fort Riley, 14LV1071).

Until recently, little age control existed for the timing of loess deposition in the central Great Plains. Age assumptions were based largely on the classical continental glaciation sequence, similar to that used for the loess stratigraphy of the Mississippi and Missouri River Valleys. However, younger loesses and associated buried soils exposed at a number of localities in Kansas and Nebraska have been systematically radiocarbon and, to a lesser extent, thermoluminescence-dated (*e.g.*, Souders and Kuzila, 1990; Johnson, 1993; Martin, 1993; May and Holen, 1993; Feng *et al.*, 1994a, b; Maat and Johnson, 1996). According to these age estimates, the Gilman Canyon Formation was deposited from at least 40 ka to about 20 ka, Peoria loess about 20 ka to 10.5 ka, and Bignell loess from about 9 ka to about 5.5 ka.

1) Peoria loess

Late Wisconsinan loess deposits form a mantle over much of the upland surface of the region, covering the central Great Plains and provide a terrestrial record of late Quaternary climate. Thickest deposits lie adjacent to the Missouri River and its major tributaries (Ruhe, 1983).

The Peoria loess is typically eolian, calcareous, massive, light yellowish-tan to buff silt that overlies the Loveland loess or an approximate equivalent of the Gilman Canyon Formation. Based on conventional and accelerator radiocarbon ages, deposition of the late-Wisconsinan Peoria loess in Kansas and Nebraska began about 19.5-21 ka. During the late Wisconsin, loess accumulated relatively fast. At the Bignell Hill type section in southwestern Nebraska, for example, the accumulation rate for the late-Wisconsinan Peoria loess averaged 5.7 mm/year. This rapid accumulation rate seems to be comparable to marine records and rapid enough to preserve high-resolution data for the late-Wisconsinan environmental changes. The rate of accumulation for the Peoria loess was certainly variable, but apparent annual laminae are present at many localities near the Platte River valley of Nebraska, including the Bignell Hill and the Eustis ash pit sites (Johnson, 1993). Loess accumulation rates decreased as the regionally-expressed Brady soil began developing between 10.6 and 10.1 ka.

The lack of any well developed buried soils or other unconformities suggests that Peoria loess in the region represents a continuous deposit and that the faunal zonation reflects a change in the rate of deposition. Evidence that Peoria loess deposition was episodic has emerged from western Iowa (Daniels *et al.*, 1960; Ruhe *et al.*, 1971), central Kansas (Arbogast, 1995) and southwestern Illinois (McKay, 1979), where deposits exhibit dark, organic-rich bands that are thought to represent incipient soils formed during periods of slower deposition. Differential abundance and preservation of fossil mollusks in the loess have also been cited as evidence of episodic loess deposition (Frankel, 1957).

The loess thickness decreases gradually with distance to the south and southeast of the Platte River valley. Except for the loess of the Loess-Drift Hill area in southeast Nebraska, loess south of the Platte was deposited rather evenly on a nearly level surface of old alluvial sands and gravels. In the Loess-Drift Hill area of southeast Nebraska and in most of the area north of the Platte River, the loess mantles a previously dissected and hilly topography.

Despite the attention given to the Peoria loess in central Great Plains, the source of the silt is not completely certain. From their review of available data, Welch and Hale (1987) concluded that a single source was not likely for all loess deposits in Kansas and that the loess was derived from a combination of three sources: glacial outwash river flood plains, present sand dune areas, and fluvial and eolian erosion of the Ogallala Formation. The Platte River undoubtedly contributed massive quantities of loess during glacial stages, as presumed earlier by Swineford and Frye (1951). Loess is thickest immediately south of the Platte River Valley, which suggests that the alluvium in the valley was the source of the loess, at least for those deposits adjacent to the valley (Kollmorgen, 1963). Some local thickening of loess occurs to the southeast of the Platte River wherever streams enter from the Sand Hills to the northwest. With prevailing northwesterly winds, these locally thick deposits are probably partially derived from alluvium brought into this valley by these streams. In addition, nonglacial rivers in western Kansas and Nebraska probably contributed substantially more to the volume of loess in the area. Local loess deposits in excess of 23 m have been measured along the southeastern bluffs of the Arikaree and Republican Rivers (Swineford and Frye, 1951). Swineford and Frye (1951) concluded that the Arkansas River carried too sandy a sediment load to act as a major loess source and suggested that most of the loess deposited south of the Arkansas River in southwest Kansas was derived from northern sources.

In Nebraska and Kansas, radiocarbon and thermoluminescence dating indicates that Peoria loess in those areas correlates temporally with the Peoria loess of Iowa, Illinois, and Indiana (*e.g.*, Johnson *et al.*, 1993; May and Holen, 1993; Martin, 1993; Maat and Johnson, 1996). However, much of the loess in Kansas and Nebraska occurs upwind of or distant from late-Wisconsinan continental glacial outwash sources. In addition, some of the thickest deposits of loess in Nebraska occur upwind of the Platte River (Swinehart, 1990). Flint (1971) pointed out that the volume of loess on the Great Plains is surprisingly high if it was all generated from glacial outwash derived from the Rocky mountains. At the present time, the source of loess in the Kansas and Nebraska portion central Great Plains is unknown, and more than one source may be involved (Welch and Hale, 1987).

Leonard (1952) subdivided the Peoria loess of Kansas into four zones on the basis of the molluskan fauna assemblages present. The *basal zone* is equivalent to a leached interval above the Gilman Canyon Formation and is void of molluskan material. The *lower molluskan zone*, or Iowan, produced an assemblage containing 14 species, 2 of which are diagnostic of the zone. A *transitional zone*, located between the upper and lower faunal zones contains elements of both assemblages and does not imply any abrupt changes in the depositional environment, although the depositional rate may have slowed somewhat. The *upper molluskan zone*, or Tazewellian, contains 26 species, 14 of which do not occur in the lower zone. Because of the relative youth of the Peoria loess, little of the upper zone has been removed from the upland.

2) Pleistocene/Holocene Transition: Brady soil

Classically, the Brady soil was associated with the upland loess deposits, but recent investigations have identified a contemporaneous soil in upland eolian sands and in alluvial valley fill (Johnson and May, 1992). It therefore appears that Brady soil development represents a time of extensive, broad-scale landscape stability. The Brady soil represents the most important break in sedimentation recorded since development of the cumulic soil of the Gilman Canyon Formation, and also marks the position of a distinct faunal discordance (Frye and Leonard, 1955). At least the early and perhaps all of the Brady soil-forming interval coincides with the Younger Dryas cold interval of the North Atlantic region.

Until recently the age of the Brady soil had been uncertain, even at the type section: Dreeszen (1970) reported two ages of 9160 and 9750 yrs B.P., both of which were believed to be too young because of contamination. Luttenegger (1985) reported an age of 8080 yrs B.P. without any stratigraphic context. Souders and Kuzila (1990) dated a core at a site in the Republican River valley and reported an age of 10,130 yrs B.P. Johnson (1993) reported two ages of 10,670 and 9240 yrs B.P. on the lower and upper 5 cm, respectively, of the Brady A horizon at the type section. Similar ages from the eolian phase have been obtained in south-central Nebraska and north-central and central Kansas. Ages of the alluvial phase of the Brady from Nebraska and Kansas closely correspond with the ages of the eolian phase. According to age data, soil development began at about 10.5 ka and ended 9-9.5 ka, suggesting a soil forming interval of greater than 1000 years.

3) Holocene

As the Laurentide ice sheet continued to waste during the early Holocene, the steep north-south temperature gradient which had been present during the late-Wisconsin continue to weaken, promoting further zonal flow. These factors triggered the generally warm and dry conditions of the Altithermal that prevailed in the central North America from about 8000 to 5000 yrs B.P. (Knox, 1988; COHMAP Members, 1988).

The Bignell loess appears to be no older than about 8000 yrs B.P. from the ^{14}C ages for the type section in Nebraska and for the Speed roadcut (Johnson, 1993). Feng (1991) speculated that the

Bignell loess is relatively well weathered because it was derived from the pre-weathered Brady soil surface, perhaps eolian and alluvial phase alike.

3. Methodology

To test the hypotheses formulated, rock magnetic parameters were measured at close interval. In order to differentiate soils from unweathered loess, rock magnetic parameters were plotted against depth and pedo-stratigraphy. Furthermore, biparametric plots were produced to compare soils with unweathered loess. Stable isotope and opal phytolith data were plotted against depth and compared with rock magnetic parameters where available. Correlation coefficients were measured to test the feasibility of the rock magnetic parameters to be an independent climate proxy.

1) Sampling

Samples for magnetic analyses were collected in the field from freshly exposed or cleaned profiles, or from cores extracted and transported to the laboratory in clear carbonate plastic liners. The individual magnetic samples were collected in numbered, demagnetized $8cm^3$ plastic cubic containers with lids. The sample interval varied slightly, but averaged 40 per meter. These cubes were pressed by hand or driven with a rubber-coated, dead-blow hammer into the exposure or core to obtain the required amount of sediment. In the laboratory, the cubes were cleaned, sorted, air dried, weighed, and placed in wooden trays prior to measurement.

Samples for stable carbon isotope were collected in the field from 10-cm intervals of the cores or exposures of some of the study sites. 300 to 400-gram samples were prepared for $\delta^{13}C$ analysis. Samples for opal phytolith analysis were done either from a core extracted with Giddings drilling machine or from exposures created by backhoe trenching. In both cases, sampling interval was 10cm.

2) Measurements

Various rock magnetic parameters including susceptibility, frequency dependence, ARM, and SIRM were measured at each site. Stable isotopes data from the Fort Riley are pending except the Pump House Canyon site. Opal phytolith data are available for several sites.

(1) Rock magnetic parameters

In order to differentiate authigenic magnetic grains from exogenic detrital input of magnetic grains, it was necessary to gain as much information as possible about the abundance, mineralogy, and grain size characteristics of magnetic minerals. Susceptibility and FD measurements were obtained using a Bartington magnetic measurement system consisting of a Model MS2 susceptibility meter and a dual-frequency sensor (MS2B). ARMs and IRMs were imparted using a Schoenstedt GSD-5 model and an ASC Scientific IM10 impulse magnetizer, and obtained with a Molspin Spinner magnetometer interfaced with a McIntosh computer. Those instruments are housed in Paleomagnetic Laboratory within the department of Geology at the University of Kansas.

The abundance of magnetic minerals is indicated by parameters such as susceptibility, SIRM and ARM. The major factor controlling the susceptibility of samples in this range is the concentration of ferrimagnetic minerals, *i.e.,* the total volume of ferrimagnetic crystals (Dearing, 1994). The concentration of magnetite in a sample is estimated by dividing the bulk susceptibility value of the sample by the susceptibility of the assumed or known magnetite type or size.

There are no completely time-efficient, satisfactory methods to quantitatively discriminate ferrimagnetic minerals from other types of minerals, discern magnetite from maghemite within the same type of minerals. Banerjee *et al.* (1993) have shown that the ultrafine-grained ferrimagnetic minerals are definitely pedogenic in origin. However, it is not yet clear whether this component is magnetite (*e.g.*, Heller and Liu, 1984; 1986) or maghemite (Verosub *et al.*, 1993). Since magnetite and maghemite, both strongly ferrimagnetic minerals, are identified as the dominant ferrimagnetic minerals by most studies both in China and North America, the type of iron oxides which occur in samples was not rigorously sought in this study. The abundance of high-coercivity minerals (predominantly hematite) was determined by the $IRM_{0.3T}/IRM_{1.1T}$, which is the ratio of the isothermal remanent magnetization from an applied field of 0.3 Tesla to the IRM form an applied field of 1.1 Tesla. Samples without high-coercivity material have a value of 1, while samples with hematite as the only magnetic phase should have a very low values (< 0.4) parameter on about 50 % of the samples from several sites including Eustis ash pit, DB site, Beisel-Steinle site, and Barton County Landfill.

In China and some other parts of the world, it has been known that there were substantial differences in magnetic grain-sizes between the loess and paleosol samples, with the latter selectively enriched in the finest grains (Maher and Thompson, 1991). Results showed that the grain size assemblages at each depth along the measured section are very sensitive to stratigraphic correlations. ARM and FD were the properties most closely linked to the presence of pedogenic fine grains (approximately 0.015–0.1μm). Comparing ARM/SIRM ratios with FD identified the fine grained tail from the detrital coarse mineral. Assembling this information on the cross-plot provided qualitative rather than quantitative grain size information.

Recent measurements on natural and synthetically produced magnetites have helped to define the major change in susceptibility with crystal size (Maher, 1988). It can be seen that there are two size ranges in which susceptibility reaches high values. At the coarse end of the scale, between 20 and 100μm, values exceed $60{,}000 \times 10^{-8}$ m^3kg^{-1}, and in crystals smaller than 0.03μm, values exceed $100{,}000 \times 10^{-8}$ m^3kg^{-1}. These two peaks effectively divide the data into three ranges which correspond to three quite different states of ferrimagnetic behavior. Above 10-20μm the crystals are divided up into different regions or cells of magnetization, known as domains, and are referred to as having multidomain behavior. In smaller crystals, normally <1μm, the restricted volume allows only one domain to form, and hence these show single domain behavior. In ultrafine crystals below 0.035μm, the magnetization is strong but unstable, and thermal energy counteracts induced magnetization very quickly after a magnetic field is removed. This behavior is similar to para-magnetism, but with a much greater susceptibility. Hence, it is termed superparamagnetic behavior.

For materials in which magnetite is the dominant magnetic species, magnetic methods offer the potential for rapid and non-destructive granulometry, if the grain size dependence of the measured parameters (and/or interparametric ratios) can be shown to be sufficiently size-diagnostic. Thompson and Oldfield (1986) noted that the ratio of SIRM to MS can be used as a rough estimate of magnetic grain size. However, mixtures of SP and SD grains give rise to similar SIRM/MS ratios as MD grains. An alternative method of granulometry was proposed by King *et al.* (1982), employing a comparison of χ_{ARM}, rather than SIRM, with χ. This method depends on χ_{ARM} showing particular sensitivity to the SD and small PSD grain sizes, and χ to the larger PSD and MD grains. χ_{ARM}/χ model is, however, complicated by the possible presence of SP magnetite grains. Firstly, a natural assemblage containing predominantly SP magnetite would be characterized by χ_{ARM}/χ values plotting in a similar location to those of very coarse magnetite. Secondly, if the proportion of SP to SD magnetite in a natural sample were to increase, the χ_{ARM} vs χ method would interpret this change as an apparent increase in grain size rather than as an actual decrease.

In view of these difficulties, a modified granulometric method of Maher (1988) was used in this study. The initial stage compared ARM not with χ (as in King et *al.'s* model (1983) but with SIRM. This parameter is more size-sensitive within the MD range, and both ARM and SIRM, as remanence parameters, seem to be unaffected by paramagnetic contributions. In the second stage, the ARM/SIRM was plotted against FD. The latter effectively separates the fine tail form the coarse end of the grain size-spectrum.

These relationships, however, are complicated by 1) variations in magnetic mineralogy (Verosub *et al.*, 1993) which may include SD and MD ferrimagnetic grains, and paramagnetic and canted antiferromagnetic minerals or by 2) magnetic grain interactions in SP grains which may cause groups of SP grains to show the magnetic behavior of coarser SD grains (Cisowski, 1981; Maher, 1988; Dearing *et al.*, 1996). Furthermore, ARM/SIRM is insensitive to the presence of true SP grains, because SP grains are unable to maintain remanence at room temperature due to thermal reorientation.

(2) Stable carbon isotope

The procedure utilized was identical to that used by our laboratory for the preparation of soil and sediment samples for ^{14}C humate dating, which renders the results compatible with those obtained in the course of age correction for the effects of isotopic fractionation (Johnson and Valastro, 1994). Samples were first disaggregated in 4-liter beakers filled with distilled water. They were then skimmed with a 60-mesh screen to remove floating organic debris. Next, the samples were washed through a 230-mesh screen with distilled water into a second beaker in order to remove the sand and coarse silt fractions; the fine fraction remaining is assumed to contain the adhering organic carbon. The samples were then treated with concentrated HCl in order to remove the inorganic carbon contained within the carbonate. This step is particularly important because of the significant amounts of carbonate precipitated within the loess. Following distilled water washes and oven-drying (100°C) in 4-liter beakers, the samples were pulverized and packaged. They were then submitted to the University of Texas Radiocarbon Laboratory for stable carbon isotope ratio analysis.

As discussed earlier, the value for a carbon isotope in soil organics is defined as:

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

However, soil formation and other processes may alter the isotopic signature imparted by vegetation. For example, two per mil (‰) isotopic enrichment may occur during litter decomposition (Melillo *et al.*, 1989; Wendin *et al.*, 1995). The ratio of C_3 plant contribution is calculated using the modified formula considering enrichment during the pedogenesis.

$$(^{13}C\ soil\text{-}2\,) = (^{13}C_{C3})\,(x) + (^{13}C_{C4})(1 - x),$$

where $^{13}C_{C4}$ is the average of ^{13}C values of C_4 plants (-13‰); $(^{13}C_{C3})$ is the average of ^{13}C values of C_3 plants (-27‰); and x is the proportion of carbon from C_3 plant sources. The resulting C_3 plant contribution is compared with the magnetic parameters and the pooid ratio from the short cell phytolith counting where available.

4. Results

In many different settings, the abundance, types, and magnetic grain sizes of magnetic minerals incorporated in sediments have varied in response to climate changes. High resolution climatic proxy records could be reconstructed by using magnetic and biological records of the last 40,000 years.

Climatic proxies within the Gilman Canyon loess indicate cool, and moist conditions. Warm, dry conditions prevail during mid-Gilman Canyon time, while late Gilman Canyon time is characterized by trends toward cooler and moister conditions after formation of the Gilman Canyon soil. Climate conditions from the Peoria loess indicate cool, moist environment at the time of loess deposition. This condition continues upward through the Peoria loess profile at the Eustis ash pit and 14LV1071 site until Brady soil is encountered at about 1.5 meters below the surface. At this Brady time, Peoria loess comes under the influence of warm, dry conditions dominating the modern surface.

The magnetic and isotopic proxies from the 14LV1071 site and the regional climate history will be discussed in the following section.

1) 14LV1071 Site

Magnetic samples were collected from several archaeological test excavations which had been deepened by backhoe. Two profiles were selected for this study: one was taken directly from the face of test excavation 14LV1071, and another (14 m) was taken at the laboratory from the soil column taken by a Giddings drill machine. The stratigraphy of the excavation site consists of Peoria loess, Brady soil, Bignell loess, and modern soil. The core, extending to about 13 m, consists of a basal layer of Gilman Canyon Formation loess and soil, Peoria loess, Brady soil, Bignell loess and modern soil.

(1) Core: Rock Magnetic Records

There is no ^{14}C age control available for the core; stratigraphic control is based solely on magnetic measurements. The results of rock magnetic measurements from the core site are summarized in Table 1. MS values range from 35 to 101 $\times 10^{-8}m^3/kg$ and exhibit the systematic differences between loess and soils. MS values of the supposedly Gilman Canyon soil (*c.* $90 \times 10^{-8}m^3/kg$) are higher than those of the underlying unweathered Gilman Canyon loess (*c.* $60\text{-}70 \times 10^{-8}m^3/kg$) and overlying Peoria loess (*c.* $40 \times 10^{-8}m^3/kg$). Once again, bimodal peaks within the Gilman Canyon Formation are clearly shown in the MS-depth profile (Fig. 3). Overlying the Gilman Canyon Formation lies the Peoria loess. MS values gradually decrease until they reach their minimum at 7 m below the surface. In the absence of ^{14}C age control, it is hard to pinpoint at what point the MS values started to bounce back to the level of the Brady soil (1 m below the surface). However, comparisons to other MS profiles from the region reveal a striking resemblance.

ARMs and SIRMs were measured more systematically at this site than were the samples from the Eustis ash pit. SIRMs correlate well with the MS variations ($r^2 = 0.968$) and are greater in the Gilman Canyon soil relative to the overlying Peoria loess by a factor of two (Fig. 3). SIRM variations

Table 1. Descriptive statistics on various magnetic parameters from the DB site.

	No.	Minimum	Maximum	Mean	Std. dev.
ARM/SIRM	170	0.27	3.04	1.47	0.70
ARM	170	0.9	17.1	6.76	0.46
FD	170	0.00	8.55	2.71	2.13
MS	170	34.63	101.84	57.79	19.45
SIRM	170	2.76	6.49	4.18	1.04

also exhibit bimodal peaks within the Gilman Canyon Formation. ARMs also correlate well with MS profiles ($r^2 = 0.965$) and are more concentrated by a factor of eight compared to the overlying Peoria loess. It should be noted that ARMs are two orders smaller than SIRMs, so the absolute amounts of variation of ARMS between the stratigraphic units are quite small compared to SIRM variations. Bimodal peaks within the Gilman Canyon Formation are evident at the DB site, something not seen at the Eustis ash pit. The almost linear behavior of SIRM and ARM with MS (Tab. 2)

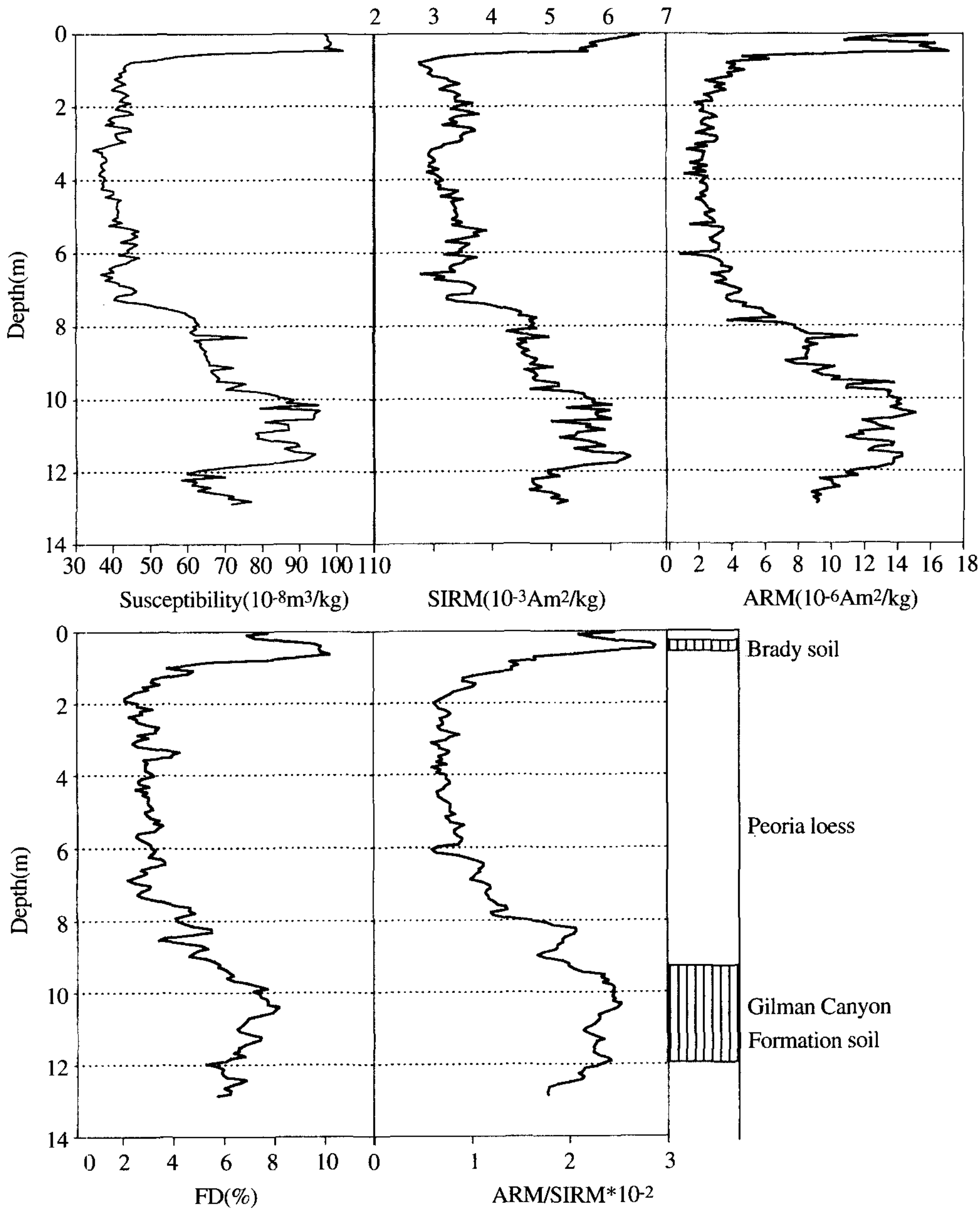

Figure 3. MS, SIRM, ARM, FD, and ARM/SIRM data and stratigraphy for the profile sampled at the 14LV1071 site(core).

Table 2. Correlation coefficient and significance level between the magnetic parameters and biological parameters at the 14LV1071 site.

	AI@	FD	$\delta^{13}C$	MS	Pooid
AI	1.000	-0.257	0.048	-0.290	0.321
FD	-0.257	1.000	-0.846**	0.910**	-0.601**
$\delta^{13}C$	0.048	-0.846**	1.000	-0.769**	0.608*
MS	-0.290	0.910**	-0.769**	1.000	-0.647**
Pooid	0.321	-0.601**	0.608*	-0.647**	1.000

@AI = chloridoid/chloridoid+panicoid.

** Correlation is significant at the 0.01 level (2-tailed).

* Correlation is significant at the 0.05 level (2-tailed).

suggests that MS may be controlled by ferrimagnetic concentration (Eyre and Shaw, 1994).

The FD data obtained by the MS2B dual frequency bridge shows the presence of an ample amount of SP materials only in the two mature soils, Gilman Canyon soil and Brady soil. FD values range from 0 to 6 percent at the DB site. FDs from the Gilman Canyon soil (*c.* 6 percent) are higher than in the Peoria loess and underlying loess. These FD values seem to be quite low compared to Chinese loess, but are high enough to contain a significant amount of SP materials (Dearing *et al.*, 1996). FDs up to >11 percent have been observed in the loess/soil sequence at Xifeng, (Heller *et al.*, 1991), and FDs up to >15 percent are reported from Luochuan, China (Forster *et al.*, 1994).

The relative variations in the amount of SD and PSD grains can be approximated by normalizing ARM values with SIRM values to obtain ARM/SIRM. This ratio also shows a sharp increase in the Gilman Canyon Formation and Brady soil, confirming a previous study (Hus and Han, 1992) which found that not only SP grains but also SD and PSD grains are relatively common in these soils. But considering that the ARM acquired per unit field is approximately five times more effective than SIRM (Hunt *et al.*, 1995), and that normalized ARM/SIRM values are only 1-3 percent, the observed two- to three-fold increase in ARM/SIRM corresponds to only a small increase in the proportion of SD and PSD grains in the soil layers.

Figure 4 shows the biparametric plots of FD (percent) vs MS and ARM (normalized to SIRM) data obtained from the DB site; compare this with the published data for synthetic magnetites of controlled grain sizes (Maher, 1988). The ARM/SIRM values for the Gilman Canyon soil vary within the range occupied by those of synthetic grains of SD size or finer. Their FD values are closely correlated with synthetics in the size range ~20-30 nm. Hence, these data identify distinctive grain-size contrasts between the Peoria loess and soil layers. The soil horizons are characterized by the presence of ultrafine-grained magnetic minerals of SD and SP size, whereas the Peoria loess layers contain little SD and a small amount of SP minerals.

The variations in SIRM also represent changes in absolute MD content. This parameter again shows some increase in the Gilman Canyon Formation and modern/Brady soil. Similar variations in SIRM were also observed at Xifeng in the central Loess Plateau (Liu *et al.*, 1992) and at Xining, in the western Loess Plateau of China (Hunt *et al.*, 1995).

2) Trench

The influence of climate change on magnetic records is demonstrated by comparing magnetic parameters to biological proxies from an archaeological excavation site (Fig. 5). Phytolith

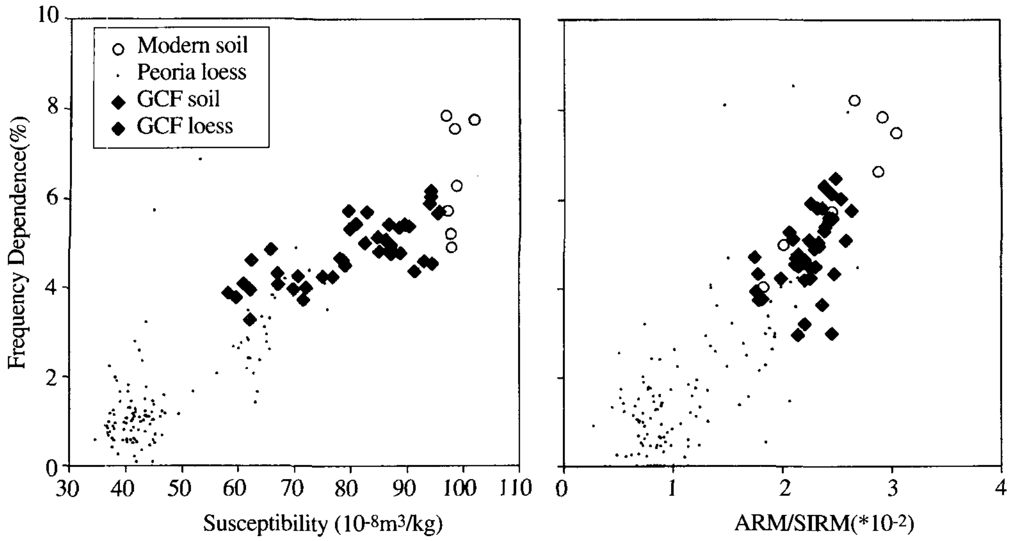

Figure 4. Biparametric plots of MS versus FD and the ratio ARM/SIRM versus FD for the samples from the 14LV1071(DB) site.

counting data have been provided by S. Bozarth.

MS variations and FD variations are similar to core samples from the same area. MS and FD values from the Peoria loess (*c.* $40 \times 10^{-8} m^3/kg$;*c.* 1 percent) are lower than in the overlying Brady soil and Bignell loess (Fig. 5). Both parameters exhibit a sudden increase at top 1 m. $\delta^{13}C$ values from the Peoria loess indicate a dominance of C_3 plants, indicative of cool and moist conditions during the period. $\delta^{13}C$ values are increasingly heavy toward the top of the profile, indicating the presence of more C_4 plants toward the end of the Pleistocene. Both magnetic parameters and $\delta^{13}C$ variations correlate well. MS and FD values start to exhibit the characteristic Brady signal at about 1.2 m below the surface. Simultaneously, $\delta^{13}C$ values indicate more C_4 plants at the end of the Peoria loess accumulation. Pooid ratio exceeds 80 percent within the Peoria loess but shows little change during the Brady time and Holocene. Correlation coefficients between the magnetic parameters and biological parameters are summarized in table 2. MS and FD values exhibit a very strong negative correlation with the pooid ratio and C_3 plant ratios derived from the mass balance formula. If variations in $\delta^{13}C$ of soil organic matter reflect the relative abundance of C_3 and C_4 grasses and summer temperature as recent research confirms (Fredlund and Tieszen, 1997; Tieszen *et al.*, 1997), magnetic parameters reflect temperature rather than moisture. The high percentage of panicoid indicates that moisture levels stayed relatively high. The decline in available moisture is pronounced at the top of the profile (Fig. 5).

5. Regional Magnetic Records

As was discussed in Chapter 4, climatic fluctuations manifest themselves in terms of alternating loess and soils, unaltered loess corresponding to a weaker magnetic signal and soils reflecting a stronger magnetic signal. Not only concentration-dependent parameters such as MS and SIRM, but also grain size-dependent parameters including FD and ARM/SIRM correspond to the alternating stratigraphic units. Also shown was the strong correlation between the

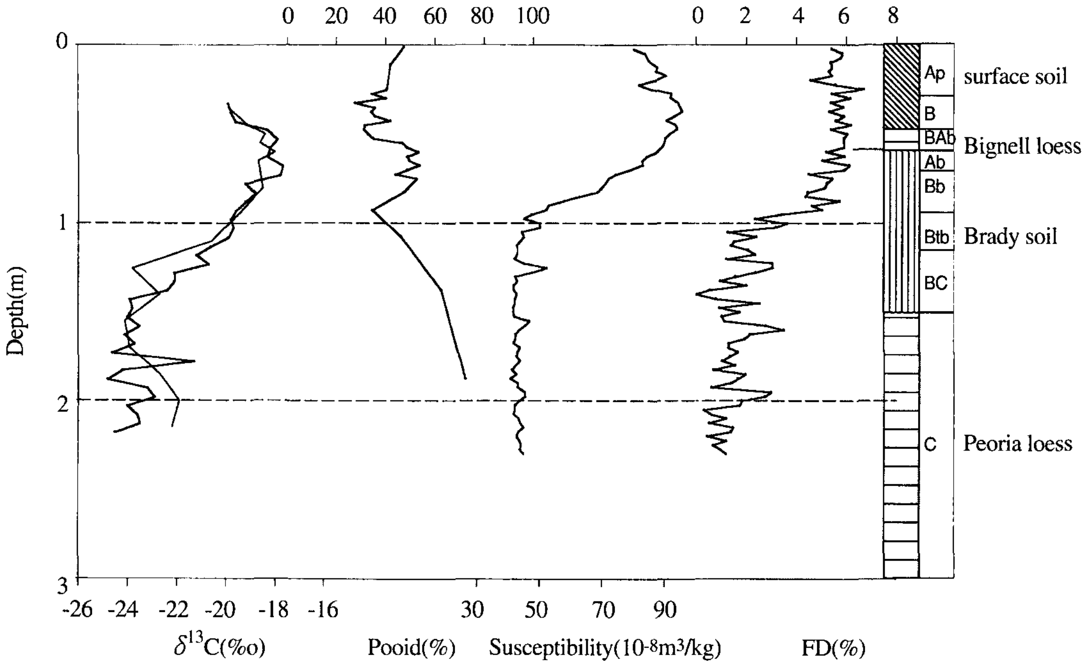

Figure 5. $\delta^{13}C$ **data from isotopic analysis(close interval) and** ^{14}C **dating correction, pooid percentage, MS and FD with stratigraphy for the profile sampled at a trench in the 14LV1071 site.**

climate-sensitive proxies, including $\delta^{13}C$ from soil humates, opal phytolith data and the magnetic parameters. By examining regional magnetic records, this study identified a regional climatic effects on magnetic records, and also discerned the influence of local climatic conditions on the degree of pedogenesis and the evolution of magnetic minerals.

The first common characteristic observed was that fairly low coercivity minerals such as magnetite are an important constituent of all the samples investigated. This was indicated by the limited number of Curie temperature data and the $IRM_{0.3T}/IRM_{1.1T}$ data. There was also evidence of significant amounts of maghemite, indicated by CBD treatment on a limited number of samples and There may well be other magnetic phases present (such as hematite), but this study agrees with Maher and Thompson (1992) in recognizing the significance of the magnetite contribution.

Second, intersite variability in soils is smaller than what it is in Peoria loess. Intersite variability in soils from the Chinese Loess Plateau is about twice what it is in loess, suggesting that pedogenesis plays a significant role in determining magnetic characteristics (Evans and Heller, 1994). This larger variability in soils is not detected in the central Great Plains, which indicates that the climatic conditions prevailing during formation of the soils were comparable despite the hundreds of kilometers between the sites.

Third, MS measurements available from the eleven sites studied thus far indicate a widespread MS minimum of $30 \times 10^{-8} m^3/kg$. There may be a MS property which supports the ground level concept of Heller *et al.* (1991).

Fourth, the soil samples have ARM/SIRM ratios in the range of 2-3 × 10-2 and FD in the range of 4-6 %. Unaltered Peoria loess samples exhibit ARM/SIRM ratios in the range of < 2 × 10-2 and FD in the range of 0-3 %, suggesting that the magnetic properties of soils are due to ultrafine magnetite grains with typical dimensions of significantly less than 100 nm (Maher, 1988).

This study accepts the overwhelming evidence (*e.g.*, Zhou *et al*, 1990; Maher and Thompson, 1991; Hus and Han, 1992; Maher and Thompson, 1992) that during the interglacials and interstadials, warmer and more humid climates were responsible for the alteration of non-magnetic (paramagnetic), iron-bearing clay minerals to strongly magnetic (ferrimagnetic), ultrafine grains of magnetite or maghemite. In all sections analyzed thus far, MS values are highest within the soils, including the Gilman Canyon geosol and Brady soil. The intervening Peoria loess has the lowest values.

A comparison among the four sites yields comparable magnetic climate proxies which provide an average picture of the climate proxy in the central Great Plains. The MS intensities of the central Great Plains loess were plotted with the regional loess/soil stratigraphy. A simple linear interpolation technique was applied between ^{14}C ages, or between stratigraphic boundaries when ^{14}C ages were unavailable (14LV1071 site).

Peoria loess deposition occurred at a relatively rapid rate so as to preclude any significant pedogenesis. At more than one site, however, the increase in MS and FD within the Peoria loess suggest one or more brief periods of landscape stability. The absolute timing of this landscape stability is unknown, but is probably centered around the Last Glacial Maximum; a moderately developed glacial maximum soil has been identified elsewhere in the region (Arbogast, 1995).

It is interesting to note that one of the most recent 3-D simulation models suggests that a high glacial dust loading may have caused a significant, episodic regional warming of over 5° C downwind of major ice-margin dust sources (Overpeck *et al.*, 1996). If this model is accurate, then one or more brief soil-forming periods may have existed during the Peoria period.

At each site, MS exhibits a decreasing trend until it reaches its minimum values, which seem to represent the Last Glacial Maximum. The weaker MS within the Peoria loess could be explained either by the difference in loess accumulation rates or by the different phase of pedogenesis among the stratigraphic units. Considering the fact that $\delta^{13}C$ values acquired from the Peoria loess indicate the dominance of C_3 plants, which demonstrates that cooling associated with the Wisconsinan glacial maximum (c. 20-18 ka) occurred in early and middle Peoria time, decreased pedogenesis/weathering during the cold interval seems the better explanation for the weaker MS signal.

Several studies in China indicate that mean MS values of loess deposits increase linearly from low values in the drier western parts of the loess plateau to higher values in the more humid central sites (Heller *et al.*, 1993). In the central Great Plains, the mean susceptibilities of Peoria loess show a systematic decrease from south to north (Fig. 6.3; Tab. 6.1).

FD values provide information on grain sizes of the magnetic carriers. Apparently, much of the pedogenic magnetic material in modern soils is at or near the superparamagnetic grain sizes (SP, 30 nm to zero). Therefore, FD is considered to be a good indicator of pedogenic processes. FD, as measured conventionally using two frequencies, is only sensitive to the presence of magnetite grains between 18 and 20 nm, not to all SP (Banerjee *et al.*, 1993). At Eustis, FD correlates very well with another grain size indicator, the ARM/SIRM parameter. Correlation between these two parameters is reasonably high at every site, suggesting that there are pedogenic magnetic grains which span the SP/SD boundary. Because a magnetic method which provides a truly quantitative measure of all SP grains produced by pedogenesis is not available, the use of FD

Table 3. Magnetic MS statistics from the region.

(unit: $10^{-8}m^3/kg$)

	Bignell Hill	Sargent	Eustis ash pit	Beisel-Steinle	Barton Co.
GCF			74.8	76.5	88.5
Peoria loess	50.03	52.90	48.3	57.7	68.8
Brady soil	70.71	72.96		90.3	76.2
Bignell loess	70.22	74.12		86	78.2
Modern soil	72.1	78.03	90.3	84.4	84

Table 4. FD statistics from the region.

(unit : %)

	Bignell Hill	Sargent	Eustis ash pit	Beisel-Steinle	Barton Co.
GCF			5	5	4.5
Peoria loess	1.5	1.1	3.3	2.8	2.8
Brady soil	3.1	3.7		3.3	5.1
Bignell loess	1.6	2.4		3.1	
Modern soil	1.8	2.2	3.3	3.2	3.4

fluctuations and ARM/SIRM could be a good measure of paleoclimate variation in the central Great Plains, at least until additional grain size-sensitive parameters are measured.

6. Future Research

Concentration sensitive magnetic parameters including magnetic susceptibility values are highest within the interstadial and interglacial soils. The intervening unweathered loess has the lower values. Magnetic records, therefore, provide an average climate proxy in the central Great Plains. MS-age plot could identify not only longer-term climate variation but also shorter-term climate fluctuations including Younger Dryas and Altithermal period.

Magnetic parameters also exhibit good correlations with the biological proxies including $\delta^{13}C$ values from humates and opal phytolith data from the region. However, much remains to be done in the central Great Plains before a comprehensive magnetic model for predicting climate change can emerge. Specifically, future study in the region should focus on 1) a quantitative understanding of the origin of SP grains through pedogenesis, and 2) developing a method for a truly quantitative measure of all the SP grains produced by pedogenesis.

Although close-interval magnetic analyses were conducted in this study, the precise mode of formation of these enhanced ultrafine-grained magnetic minerals remains unclear. Concentration-sensitive parameters correspond highly to the stratigraphic units, but size-sensitive parameters do as well. Although a number of biological proxies available from the region, including $\delta^{13}C$ and phytolith analysis, and non-biological proxies such as particle size analysis confirm the possible causal relationship between climate and magnetic parameters, there is no direct transfer function comparable to the one derived from pollen analysis elsewhere. Future research should attempt to quantify the origin of ultrafine grains, and therefore an attempt should be made to derive a closer interval $\delta^{13}C$ signal from the soil humate. An attempt to tighten the chronology of loess deposition would be useful, as regards minor climate variations.

Alternative experiments should be attempted to develop a method to measure the amount of magnetics that is of truly pedogenic origin. MS and FD zonations can be correlated from southwestern Nebraska and beyond, while MS zones have a different character at the Bone Cove site and Sumner Hill site. MS signals in the Peoria loesses from the sites are comparable to those of well-developed soils, but FD signals do not indicate the presence of finer magnetic minerals. Differences in MS zones probably reflect the addition and mixing of sediments from different source areas. Also, archaeological remains such as hearths, produce a sharp increase in MS intensity and virtually no change in FD, complicated the investigation. A quantitative measure of truly pedogenic magnets would establish better correlation between climate and magnetic parameters.

Finally, the results of this study on loess should be integrated with other studies, such as sand dunes studies both in the Sand Hills of Nebraska and Great Bend, Kansas, and alluvial stratigraphy in the region.

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