Late Quaternary loess in northeastern Colorado: Part I—Age and paleoclimatic significance

Daniel R. Muhs* *U.S. Geological Survey, M.S. 980, Box 25046, Federal Center, Denver, Colorado 80225*

John N. Aleinikoff *U.S. Geological Survey, M.S. 963, Box 25046, Federal Center, Denver, Colorado 80225*

Thomas W. Stafford Jr.[†], **Rolf Kihl** *Institute of Arctic and Alpine Research, Campus Box 450, University of Colorado, Boulder, Colorado 80309*

Josh Been, **Shannon A. Mahan**, **Scott Cowherd** *U.S. Geological Survey, M.S. 980, Box 25046, Federal Center, Denver, Colorado 80225*

ABSTRACT

Loess in eastern Colorado covers an estimated 14 000 km^2, and is the westernmost part of the North American midcontinent loess province. Stratigraphic studies indicate there were two periods of loess deposition in eastern Colorado during late Quaternary time. The first period spanned ca. 20 000 to 12 000 ^{14}C yr B.P. (ca. 20–14 ka) and correlates reasonably well with the culmination and retreat of Pinedale glaciers in the Colorado Front Range during the last glacial maximum. The second period of loess deposition occurred between ca. 11 000 and 9000 ^{14}C yr B.P. This interval may be Holocene or may correlate with a hypothesized Younger Dryas glacial advance in the Colorado Front Range. Sedimentologic, mineralogic, and geochemical data indicate that as many as three sources could have supplied loess in eastern Colorado. These sources include glaciogenic silt (derived from the Colorado Front Range) and two bedrock sources, volcaniclastic silt from the White River Group, and clays from the Pierre Shale. The sediment sources imply a generally westerly paleowind during the last glacial maximum. New carbon isotope data, combined with published faunal data, indicate that the loess was probably deposited on a cool steppe, implying a last glacial maximum July temperature depression, relative to the present, of at least 5–6 °C. Overall, loess deposition in eastern Colorado occurred mostly toward the end of the last glacial maximum, under cooler and drier conditions, with generally westerly winds from more than one source.

INTRODUCTION

In recent studies of Quaternary climate change there has been a renewed interest in loess. Much of this attention stems from new studies of long, possibly continuous, loess sequences that contain detailed records of Quaternary glacial-interglacial cycles, thought to be a terrestrial equivalent to the foraminiferal oxygen isotope record in deep-sea sediments (e.g., Hovan et al., 1989). Loess is also a direct record of atmospheric circulation, and identification of loess paleowinds in the geologic record can test atmospheric general circulation models (Broccoli and Manabe, 1987; COHMAP Members, 1988; Kutzbach et al., 1993, 1998; Bartlein et al., 1998).

Most previous work on North American loess has focused on deposits adjacent to valleys that drained the Laurentide ice sheet (Ruhe, 1983; Follmer, 1996). Fewer studies have been conducted on the origin, stratigraphy, and age of loess in the subhumid to semiarid Great Plains area of Colorado, Nebraska, and Kansas, found to the west of glaciated terrain (Fig. 1). Within the Great Plains region, loess in northeastern Colorado has received the least attention (see review in Madole, 1995), despite the fact that the area covered by this sediment is ~14 000 km^2 (Fig. 2). In this paper we present new data on the stratigraphy, geochronology, sedimentology, geochemistry, stable isotope composition, and paleoclimatic significance of northeastern Colorado loess.

PREVIOUS STUDIES

Three late Quaternary loess units, from oldest to youngest, the Gilman Canyon Formation, and the Peoria and Bignell Loesses, have been identified and correlated on the Great Plains (Schultz and Stout, 1945; Frye and Leonard, 1951). The Gilman Canyon Formation is thin (usually <2 m) and typically has an organic-rich soil developed throughout. It is overlain by Peoria Loess, which is the thickest (to ~50 m) and areally most extensive of the Great Plains loess units. A dark, organic-rich buried soil, referred to as the Brady soil, caps the upper part of the Peoria Loess, separating it from the overlying Bignell Loess. The Bignell Loess is usually no more than ~2 m thick, and occurs sporadically.

Most of the recent age estimates of Great Plains loesses have been from localities in Nebraska. In this paper, all radiocarbon ages are given in ^{14}C years B.P. and all thermoluminescence ages are given in thousands of calendar years (ka). Based on radiocarbon ages of soil organic matter reported by Martin (1993), May and Holen (1993), and Maat and Johnson (1996), and thermoluminescence (TL) analyses by Pye et al. (1995) and Maat and Johnson (1996), the age of the Gilman Canyon Formation is ca. 36 000 to ca. 22 000 ^{14}C yr B.P. Charcoal from spruce (*Picea*), as well as bone, snails, and detrital organic matter found within Peoria Loess give ages ranging from about 21 000 to 10 000 ^{14}C yr B.P. (Wells and Stewart, 1987; Martin, 1993; May and Holen, 1993; Feng et al., 1994; Maat and Johnson, 1996). Direct dating of Peoria Loess in Nebraska, using TL methods, gives ages ranging from ca. 24 to 12 ka (Pye et al., 1995; Maat and Johnson,

*E-mail: dmuhs@usgs.gov.
[†]Present address: Stafford Research Laboratories, Inc., 5401 Western Avenue, Suite C, Boulder, Colorado 80301.

GSA Bulletin; December 1999; v. 111; no. 12; p. 1861–1875; 15 figures; 1 table.

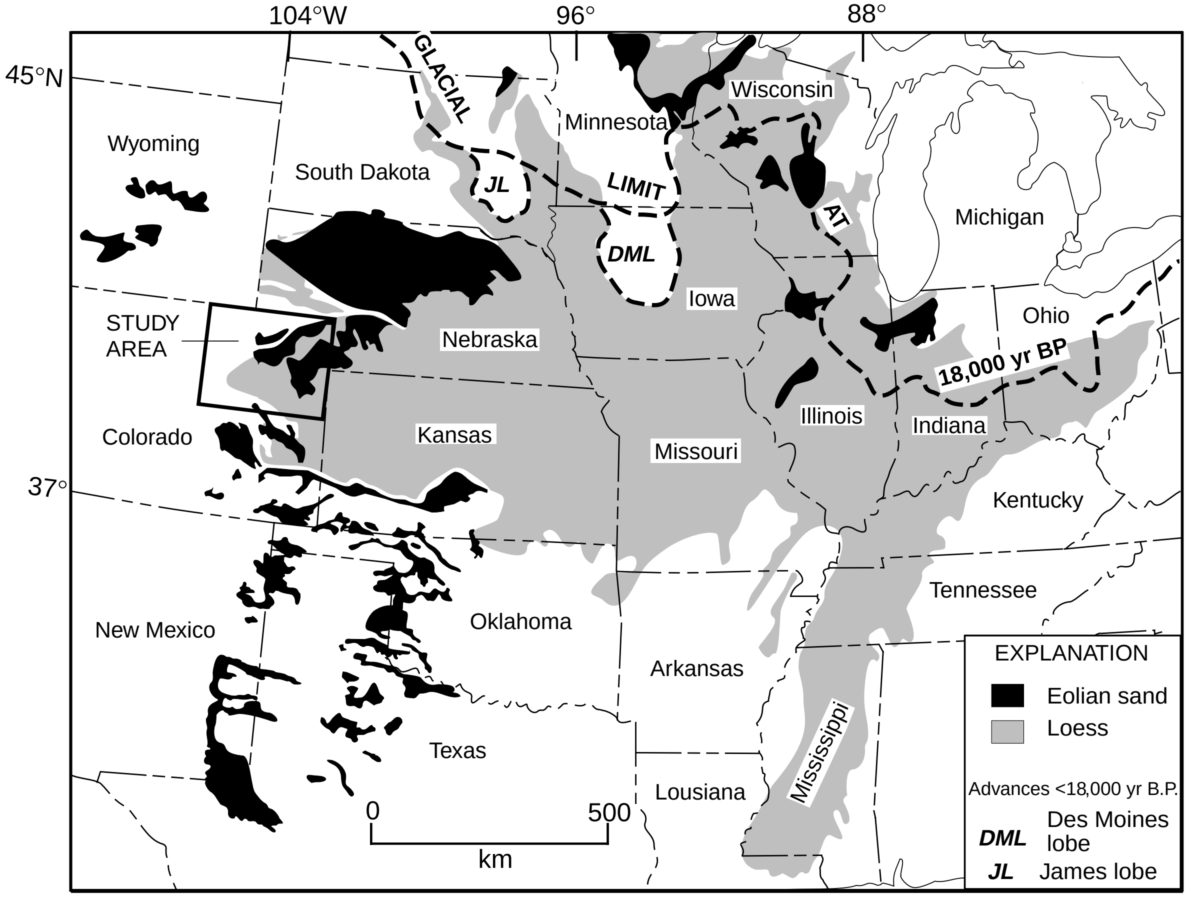

Figure 1. Map showing the distribution of late Pleistocene loess and eolian sand in the North American midcontinent, approximate limit of the Laurentide ice sheet during the last (Wisconsin) glacial period, and study area in eastern Colorado. Loess distribution and eolian sand east of the Mississippi River are from Thorp and Smith (1952); eolian sand distribution west of the Mississippi River is from Muhs and Holliday (1995).

1996). Direct dating of probable Peoria Loess at a locality in eastern Colorado using TL methods gives ages ranging from ca. 20 to 15 ka (Forman et al., 1995). Maximum-limiting ages of the Bignell Loess are based on radiocarbon ages of organic matter from the Brady soil, and range from ca. 11 000 to 8000 ^{14}C yr B.P. (Martin, 1993; Maat and Johnson, 1996). Direct dating of the Bignell Loess using TL gives ages ranging from ca. 9 to 3 ka (Pye et al., 1995; Maat and Johnson, 1996).

The origin of loess on the Great Plains has been debated for more than 50 yr. The uncertainty of loess origins in this region stems from the fact that although eolian silt is widespread (Fig. 1) and often thick (to ~50 m), it is not directly tied to the Laurentide ice sheet. Although local occurrences of thin loess in parts of the Great Plains have clear links to smaller drainages (e.g., Reheis, 1980), large rivers that drained the Laurentide ice sheet occur mostly to the east of the Great Plains. Welch and Hale (1987) reviewed all of the evidence for loess origins in Kansas, and thought that loess in that state (and by extension, probably elsewhere in the Great Plains) was derived from some combination of Rocky Mountain glacial outwash, dune sand, and deflation of bedrock sources such as the Ogallala Formation.

METHODS

Loess in northeastern Colorado was examined in road cuts, natural exposures, and auger borings at ~100 localities (Fig. 2). Samples for particle size, carbonate, mineralogical, and geochemical analyses were taken well below the zone of pedogenesis, usually at depths of ~2 m. At two localities in Colorado, and at Bignell Hill, Nebraska, samples of buried soils were collected for radiocarbon dating and carbon isotope analyses. Accelerator mass spectrometry (AMS) radiocarbon ages were determined on humic acid extractions from organic matter in buried soils (Table 1). Detailed descriptions of the radiocarbon sample extraction steps were given in Abbott and Stafford (1996); radiocarbon abundance was measured by AMS at Lawrence Livermore National Laboratory. For carbon isotope analyses, splits of bulk buried soils or loess samples were leached of carbonates with 6N HCl, then washed and dried. The samples were then combusted at 850 °C for 4 hr in the presence of CuO in vacuo in individual ampules. Resulting contaminating gases were removed through an automated trapping box, and the purified CO_2 was analyzed on a triple-collecting gas-source mass spectrometer. Results are reported in delta (δ) notation relative to Peedee belemnite (PDB). Particle size analysis of most samples was done by a sedigraph particle size analyzer after removal of organic matter and carbonates and dispersion with Na-pyrophosphate. A few samples were analyzed by the hydrometer method but with similar pretreatments. Mineralogical and geochemical studies of loesses and possible source sediments were conducted using X-ray diffraction and energy-dispersive X-ray fluorescence methods, respec-

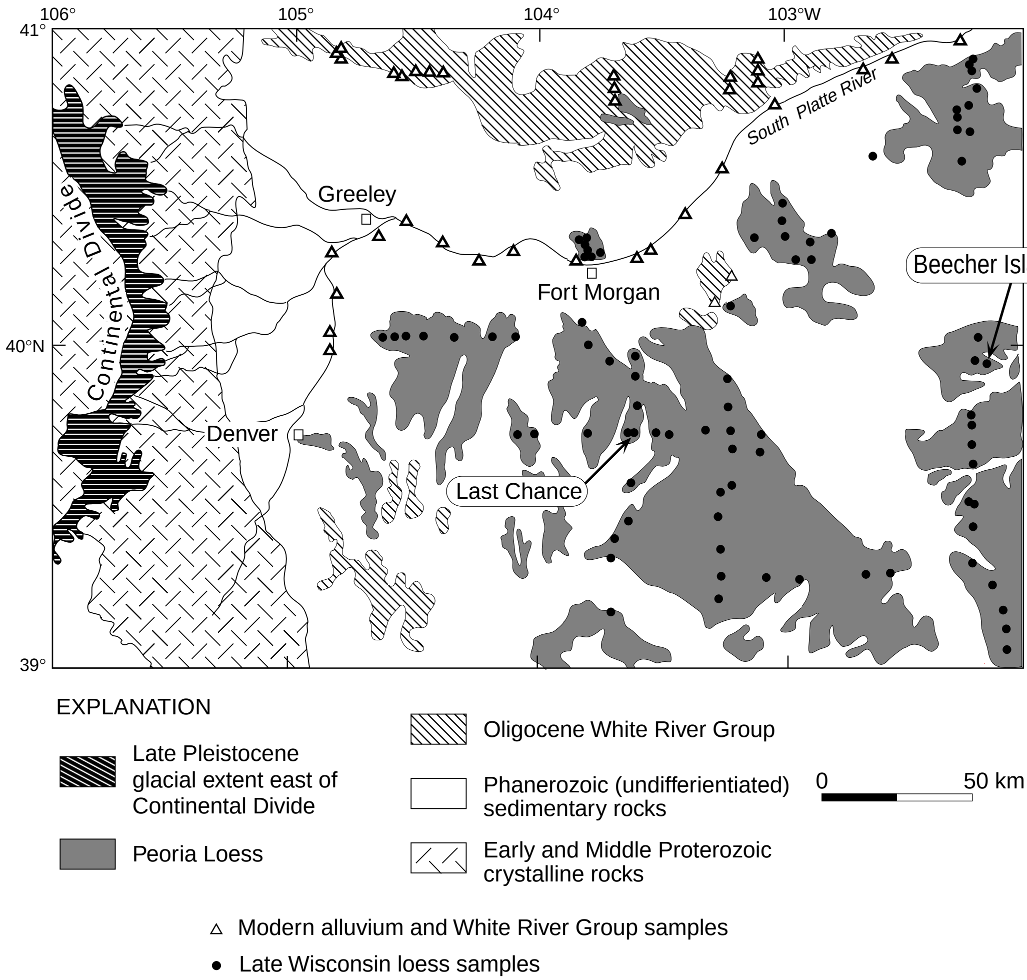

Figure 2. Generalized geologic map of northeastern Colorado, distribution of late Pleistocene glaciers east of the continental divide, and sample locations. Compiled from data in McGovern (1964), Weist (1964), Scott (1978), Crabb (1980), Sharps (1980), Bryant et al. (1981), and Madole et al. (1998); we field checked the loess distribution.

TABLE 1. RADIOCARBON AGES OF HUMIC ACID FRACTIONS OF PALEOSOLS IN LOESS

Section	Field number	Unit	Depth (m)	University of Colorado number	Lawrence Livermore National Laboratory number	Radiocarbon age (^{14}C yr B.P.)	Calendar age* (yr B.P., 2σ)
Last Chance, Colorado	LI-12-9	Buried soil	3.55	NSRL-2756	CAMS-23133	21 060 ± 100	—
	LI-12-10	Buried soil	4.05	NSRL-2757	CAMS-23134	22 090 ± 100	—
	LI-12-11	Buried soil	4.55	NSRL-2758	CAMS-23135	22 940 ± 120	—
Beecher Island, Colorado	LI-201	Modern soil A horizon	0.07	NSRL-2751	CAMS-23128	1480 ± 60	1284–1513
	LI-202	Modern soil Bw1 horizon	0.22	NSRL-2752	CAMS-23129	6680 ± 60	7394–7585
	LI-203	Modern soil Bw2 horizon	0.43	NSRL-2753	CAMS-23130	8160 ± 60	8955–9361
	LI-204	Modern soil BC horizon	0.70	NSRL-2754	CAMS-23131	9250 ± 60	10 036–10 372
	LI-207	Buried soil A1 horizon	1.63	NSRL-2072	CAMS-17300	11 090 ± 60	12 849–13 159
	LI-208	Buried soil A2 horizon	1.88	NSRL-2073	CAMS-17297	11 810 ± 50	13 555–14 024
	LI-229	Buried soil Btk1 horizon	12.24	NSRL-2755	CAMS-23132	20 520 ± 90	—
Bignell Hill, Nebraska	BH-5	Modern soil A horizon	0.06	NSRL-2954	CAMS-26399	1360 ± 60	1152–1366
	BH-6	Modern soil A horizon	0.19	NSRL-2955	CAMS-26400	1400 ± 50	1188–1393
	BH-1	Brady soil, upper part	1.94	NSRL-2804	CAMS-24344	10 070 ± 60	11 007–12 054
	BH-2	Brady soil, lower part	2.17	NSRL-2805	CAMS-24345	10 490 ± 60	12 176–12 590
	BH-7	Gilman Canyon Formation	50	NSRL-2956	CAMS-26401	30 770 ± 210	—
	BH-3	Gilman Canyon Formation	51.5	NSRL-2806	CAMS-24346	40 600 ± 1100	—

*Calculated using conversion programs of Pearson and Stuiver (1993) and Stuiver and Pearson (1993).

tively. Carbonate content was measured by either manometric or gravimetric methods.

STRATIGRAPHY AND GEOCHRONOLOGY OF LOESS IN NORTHEASTERN COLORADO

Loess in northeastern Colorado is the westernmost part of an almost continuous loess blanket in the North American midcontinent (Fig. 1). Loess is distributed widely but discontinuously to the southeast of the South Platte River, with only isolated occurrences to the north of this major drainage (Fig. 2). Interpretation of limited well log data indicates that Colorado loess may be as thick as ~40 m (Weist, 1964), but the thickest deposit observed in this study, at Beecher Island (Fig. 3), is ~12 m. Thicknesses of 2–5 m are more typical, and in the northeasternmost part of the study area, thin loess occurs in a patchy distribution on the surface of Ogallala Formation bedrock.

Although there are few exposures of loess in eastern Colorado, two road cuts, near Last Chance and Beecher Island, give some insight into loess deposition history in the region (Figs. 2 and 3). Surface loesses in eastern Colorado and western Nebraska are underlain by buried soils, which can be identified on the basis of high organic matter and low carbonate contents in surface (A) horizons and relatively high clay and carbonate contents in subsurface (Bt, Btk, or Bk) horizons (Figs. 4 and 5). Humic acid extractions from organic matter in buried soils provide the basis for the radiocarbon ages reported here.

We recognize the limitations of AMS ^{14}C age estimates of humic acid extractions. Some of the problems with interpretation of radiocarbon analyses of buried soil organic matter were summarized by Martin and Johnson (1995), Wang et al. (1996), and Abbott and Stafford (1996). In studies conducted by Martin and Johnson (1995), wherein total, humic acid, and residue extractions of several buried soils were analyzed, differences are reported for both extractions in the same laboratory and similar extractions from different laboratories. Unfortunately, no systematic relation emerged from these studies that could indicate the most reliable organic matter fraction. Modeling studies

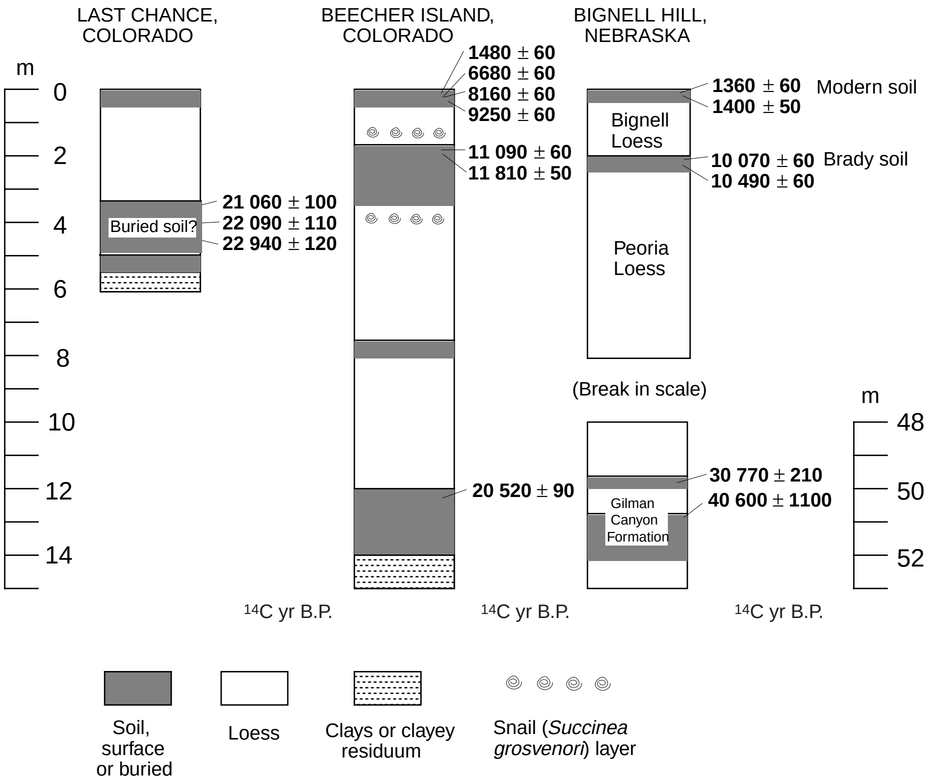

Figure 3. Stratigraphy and accelerator mass spectrometry radiocarbon ages of two loess sections in eastern Colorado (see Fig. 2 for locations) and Bignell Hill, Nebraska. Stratigraphy of Bignell Hill is from Maat and Johnson (1996). Locations of Last Chance and Beecher Island are in Figure 2; location of Bignell Hill is in Figure 12.

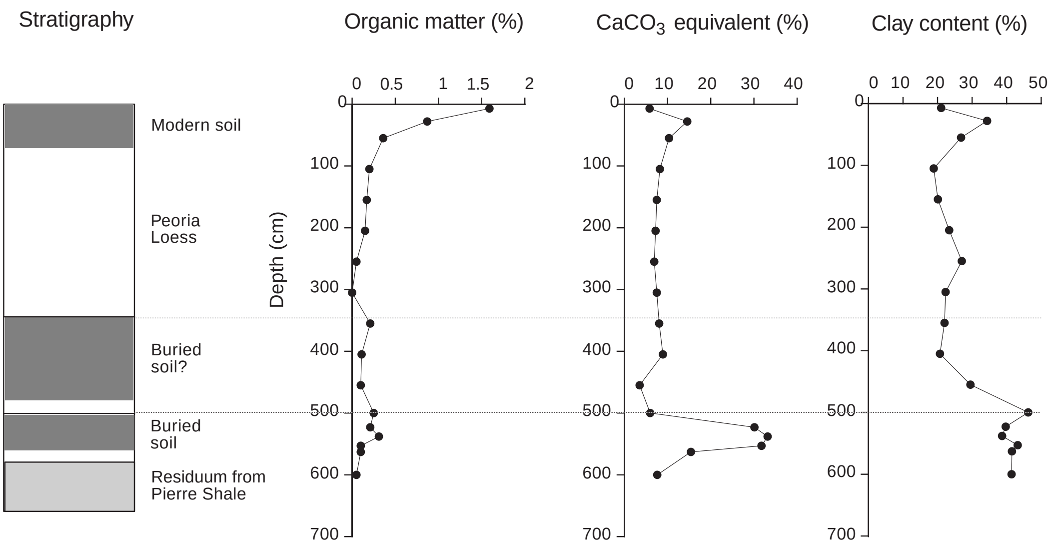

Figure 4. Stratigraphy, organic matter, $CaCO_3$ equivalent, and clay content with depth in the loess section at Last Chance, Colorado.

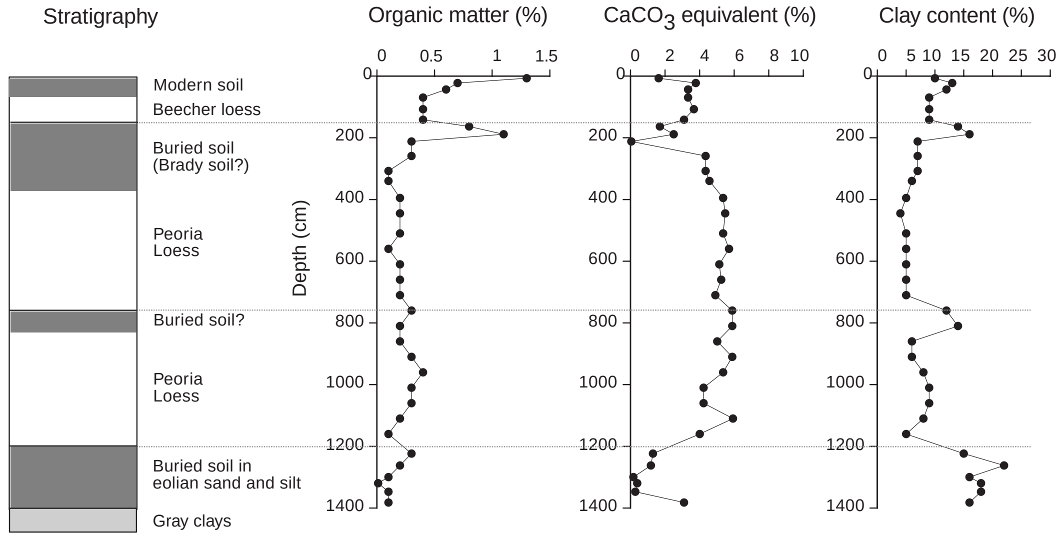

Figure 5. Stratigraphy, organic matter (%), $CaCO_3$ equivalent, and clay content with depth in the loess section at Beecher Island, Colorado.

by Wang et al. (1996) indicate that radiocarbon ages of buried soil organic matter could overestimate the time of burial. This hypothesis is supported by our own studies and those of Abbott and Stafford (1996). Analysis of humic acids extracted from the upper parts of modern soil A horizons at two localities gives apparent AMS ^{14}C ages of ca. 1400 yr B.P. (Fig. 3). These data indicate that buried soil radiocarbon ages may overestimate the time of initiation of overlying loess deposition by as much as 1–2 k.y. However, a thorough AMS ^{14}C study of lake sediments and adjacent watershed soils and peats led Abbott and Stafford (1996) to conclude that humic acid extractions are the best alternative to plant macrofossils for AMS ^{14}C dating because: (1) this fraction is usually present in large enough amounts for dating; (2) unlike humin compounds, humic acids are formed from lower molecular weight organic compounds that are less likely to be composed of reworked refractory carbon; and (3) chemical separation of humic acids from refractory organic matter and carbon-bearing clays is possible. We take a conservative approach in this study, where radiocarbon ages of buried soils are interpreted to be maximum limiting ages for any overlying deposits, and minimum limiting ages for the parent loess and any underlying deposits.

Loess of moderate thickness (~3–5 m) is typical of the area near the town of Last Chance, Colorado (Fig. 3). In a road-cut exposure, a possible clayey residuum from the Pierre Shale is overlain by a clayey silt deposit in which a soil with a Bt/Btk/Bk profile developed (Fig. 4), perhaps mixed with part of the clayey residuum. This buried soil is overlain by another possible buried soil that is not easily seen in the field, but that may be recognized on the basis of subtle organic matter and clay content trends (Fig. 4). Loess overlies this possible weak buried soil, and the modern soil developed in this loess has an A/Btk1/Btk2/C profile. Humic acids from the weak buried soil at a depth of ~3.5 m give stratigraphically consistent AMS radiocarbon ages ranging from ca. 23 000 to 21 000 ^{14}C yr B.P. (Fig. 3), suggesting that loess deposition began sometime after ca. 21 000 ^{14}C yr B.P.

Thicker loess is found to the east, near the Kansas and Nebraska state lines. At Beecher Island, gray-green calcareous clays of unknown origin are overlain by eolian(?) sands and silts in which a strongly expressed buried soil developed (Fig. 3). This buried soil has a subangular blocky to angular blocky structure with well-expressed clay films, and as much as 22% clay in the Bt horizon (Fig. 5). About 10 m of what is interpreted to be Peoria Loess overlies this buried soil, although a possible thin buried soil, with strong, coarse prismatic structure and 12%–14% clay, is found at a depth of ~8 m. Between a depth of ~1.5 m and 3.5 m, there is a thick, though minimally developed, buried soil with an A1/A2/AB/Bw1/Bw2/C profile (Fig. 5). This buried soil is overlain by stratified eolian silt and sand with a modern soil, characterized by an A/Bw1/Bw2/C profile, in its upper part. Land snails were found in both the upper part of Peoria Loess and in the younger loess above the uppermost buried soil. All snail specimens were identified by G. Goodfriend (Carnegie Institution) as *Succinea grosvenori* Lea.

Humic acids from the upper part of the lowermost buried soil at Beecher Island give a radiocarbon age of 20 520 ± 90 ^{14}C yr B.P. This age indicates that most loess deposition at Beecher Island may have occurred at about the same time as that at Last Chance. Humic acids from the A1 and A2 horizons of the buried soil between 1.5 and 3.5 m depth at Beecher Island give ages of 11 090 ± 60 and 11 810 ± 50 ^{14}C yr B.P., respectively. Collectively, the radiocarbon ages indicate that most (Peoria) loess deposition occurred between about 20 000 and 12 000 ^{14}C yr B.P. (Fig. 3). Radiocarbon age determinations were also made on humic acids from the A, Bw1, Bw2, and upper C horizons of the modern soil at Beecher Island, in order to provide a minimum age for the youngest sandy loess at this locality. These ages are consistently younger upward through the modern soil profile, and suggest that the youngest loess was deposited between about 11 000 and 9000 ^{14}C yr B.P.

On the basis of the stratigraphic position of the youngest loess at Beecher Island, it was expected that the bracketing buried soil radiocarbon ages would suggest a correlation with the Bignell Loess of Nebraska, recently shown by TL and radiocarbon methods to be Holocene age (Pye et al., 1995; Maat and Johnson, 1996). In order to test whether interlaboratory differences in dating methods might account for the differences in ages, we also dated humic acids from organic matter in Gilman Canyon buried soils, the Brady soil, and the modern soil at Bignell Hill, Nebraska, the type locality for the Brady soil and the Bignell Loess (Fig. 3). The results are in good agreement with both conventional radiocarbon and TL ages reported by Maat and Johnson (1996) for this locality. The new radiocarbon results from Bignell Hill indicate that Peoria Loess deposition in western Nebraska may have begun earlier (ca. 30 000 ^{14}C yr B.P. as opposed to ca. 20 000 ^{14}C yr B.P.) and ended slightly later (ca. 10 500 ^{14}C yr B.P., compared to ca. 11 800 ^{14}C yr B.P.) than in Colorado. In addition, the new AMS radiocarbon ages confirm that the Bignell Loess is Holocene age. Therefore, it appears that the youngest loess at Beecher Island could be truly older than the Bignell Loess at its type locality. An alternative explanation is that the organic matter that gave ages of ca. 9000 and 8000 ^{14}C yr B.P. from the Bw2 and C horizons of the modern soil is not derived from in situ organic matter accumulation during pedogenesis, but rather is detrital organic matter transported to the site during loess fall. In this interpretation, the upper loess could be of early or mid-Holocene age and correlative with the Bignell Loess. However, the humic acid extraction method used minimizes the potential danger associated with this kind of problem (Abbott and Stafford, 1996). Another explanation is that the uppermost loess at Beecher Island is correlative with the Bignell Loess in Nebraska, but loess deposition was time transgressive. Because of these alternative explanations, the older loess at Beecher Island, bracketed by ages of ca. 20 000 and 11 810 ^{14}C yr B.P., is correlated with the Peoria Loess of Nebraska and Kansas, but the youngest loess at Beecher Island is informally referred to as the Beecher loess.

A common soil-mapping unit in loess deposits of the western Great Plains is the Kuma series, a Pachic Argiustoll. Typical descriptions of the Kuma series indicate that surface horizons are developed in loess, but that the lower B horizon of the soil is developed in an older loess. These descriptions led Welch and Hale (1987) to speculate that in Kansas, the soil is developed in thin Bignell Loess over thicker Peoria Loess. The Kuma series is also mapped in Colorado and western Nebraska. Plots of organic carbon as a function of depth in Kuma soils from eastern Colorado and Nebraska and Kansas counties immediately to the east of Colorado show two maxima, one at the surface and the other at shallow depths (Fig. 6). It is suggested that the Kuma series in Colorado and adjacent parts of Nebraska and Kansas may represent the two periods of loess deposition, as proposed by Welch and Hale (1987).

SEDIMENTOLOGY OF LOESS IN NORTHEASTERN COLORADO

Loess of northeastern Colorado has sedimentologic characteristics that are distinct from other eolian sediments in the Great Plains. The average particle size is fine silt, as opposed to an average particle size of medium silt for loess of adjacent Kansas (Fig. 7). However, Colorado loess shows a much wider range of mean particle sizes, and individual samples are much less well sorted than Kansas loess. Colorado eolian sands and loess deposits also show no overlap in particle size or degree of sorting, suggesting that they are not different facies of the same deposit, but rather distinct sedimentary bodies with different origins.

Pye (1987) summarized loess particle size data from many parts of the world and most loesses show a prominent mode in the 20–30 µm range with a tail to the fine side. Particle size histograms of Colorado loess commonly show a mode in the same size range, but the clay (<4 µm) fraction also shows a prominent mode (Fig. 8). Such a distinct mode following a more typical fine tail on particle size histo-

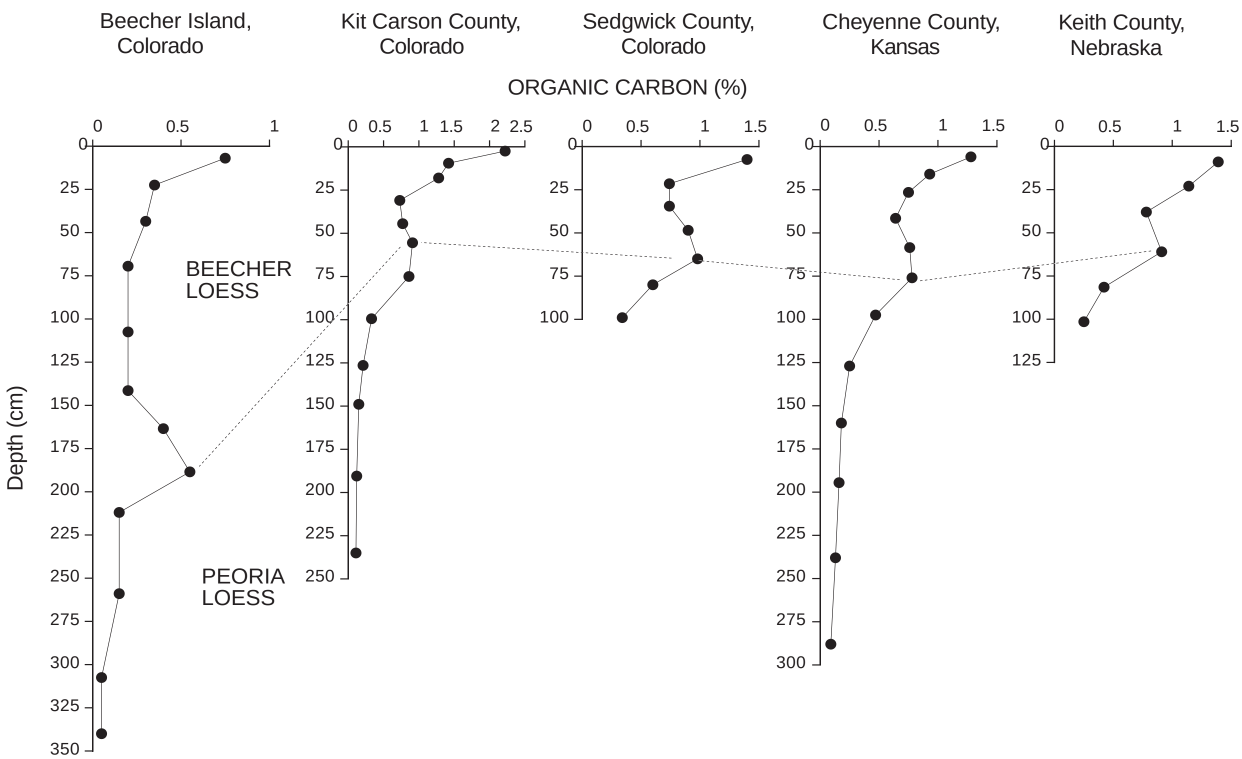

Figure 6. Plots showing organic carbon contents (%) with depth at Beecher Island, Colorado, and four representative localities in Nebraska, Kansas, and elsewhere in eastern Colorado where Kuma soils have been mapped. Dashed line connecting subsurface organic carbon (%) maxima suggests a possible contact between lower (Peoria) loess and upper (Beecher) loess. Organic carbon data are from the Natural Resources Conservation Service soil database, Lincoln, Nebraska.

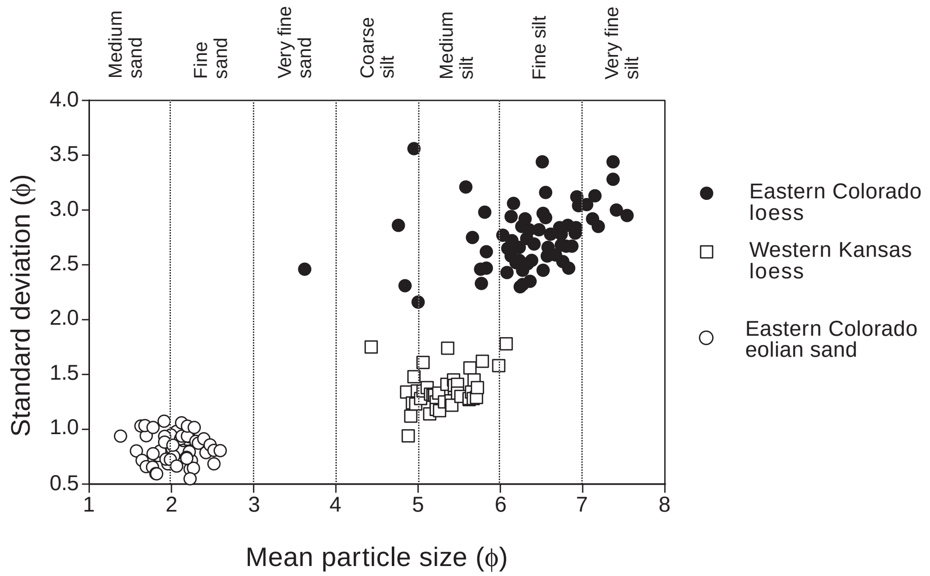

Figure 7. Plot showing mean particle size and apparent sorting for northeastern Colorado loesses compared to Kansas loesses and northeastern Colorado eolian sands. Colorado loess data are from this study, Kansas loess data are from Swineford and Frye (1951), and Colorado eolian sand data are from Muhs et al. (1996).

grams suggests a clay-rich sediment source different from that of the silt fraction. This interpretation is supported by the spatial distribution of clay in eastern Colorado loess, which shows a distinct northeastward-diminishing trend (Fig. 9). The clay content is generally <10% in the northeastern part of the study area, and is generally >30% in the southwestern part. This spatial pattern suggests a clay-rich source in the southwestern or western parts of the study area, and is consistent with the relatively low clay contents reported by Swineford and Frye (1951) in western Kansas loess. It is possible, even likely, that the clay-sized particles were transported as silt-sized aggregates of clay, in much the same manner as suggested for the clay-rich eolian parna deposits of Australia (Dare-Edwards, 1984).

MINERALOGY AND GEOCHEMISTRY OF LOESS AND SOURCE SEDIMENTS

Mineralogical and geochemical studies of eastern Colorado loess were conducted in order to make comparisons to Nebraska loess, Colorado dune sands, and possible source sediments. Loess in eastern Colorado could have its origins in glacial silt derived from Front Range glaciers, which were more extensive during late Pinedale (late Wisconsin) time (Fig. 2). Meltwaters from these glaciers were drained by the South Platte River and its tributaries (Madole et al., 1998). Another possible source sediment for eastern Colorado loess is the White River Group (Eocene-Oligocene), found over much of northern Colorado (Fig. 2), as well as southwestern South Dakota, western Nebraska, and southeastern Wyoming. Denson and Bergendahl (1961) reported that the White River Group, which is mostly volcaniclastic, contains between 63% and 86% silt in the noncarbonate fraction, and has carbonate contents of 20%–30%. If either glacial silt carried by the South Platte River or volcaniclastic sediments from the White River Group were sources, northerly or northwesterly paleowinds would be required, because these sources are situated mostly to the north of loess occurrences. However, the high clay content of Colorado loess argues for contributions from a clay-rich source, such as residuum from the Pierre Shale. This rock unit is found at the surface over a wide area immediately to the west and southwest (and over lesser areas to the north) of the westernmost loess in the study area (Scott, 1978; Sharps, 1980), and would therefore require a westerly or southwesterly wind.

Mineralogical analyses indicate that quartz, K-feldspar, plagioclase, calcite, and dolomite are all important components in eastern Colorado loess. Quartz is more abundant in loess compared to eolian sand in eastern Colorado. Carbonates, minimal in eolian sand in this region, are abundant in loess, as determined by both manometric and X-ray diffraction methods. Total carbonate (calcite plus dolomite) contents range from 3% to 12%, although most samples analyzed have values between 7% and 9%. Calcite-to-dolomite values range from 1 to 4; most values are between 1.2 and 1.8. Modern fine-grained (silt and clay rich) overbank alluvium of the South Platte River and sediments of both the White River Group and the Pierre Shale all contain carbonates, so the presence of carbonates says little about source sediments.

Geochemical data reinforce the sedimentological evidence indicating that eolian sand and loess in eastern Colorado are distinctly different sedimentary bodies. Loess has much greater abundances of Ca and Sr compared to eolian sand (Fig. 10). The much higher Ca/Sr values in loess compared to eolian sand also indicate that these elements are found in different minerals in the two sediments. Carbonates are important minerals in loess but not in eolian sand, so much of the Ca and Sr in eolian sand are most likely found in noncarbonate minerals. K and Rb have much lower concentrations in loess compared to eolian sand, but the two sediments have similar K/Rb values. This observation suggests that K and Rb are found in the same mineral, likely K-feldspar, but that K-feldspar is more abundant in eolian sand. X-ray diffraction analyses confirm that loess has lower K-feldspar contents than eolian sand in eastern Colorado, supporting the K and Rb data. Ti and Zr concentrations, and Ti/Zr values, are higher in loess than in eolian sand. Collectively, the data indicate that eastern Colorado loess has a different mineralogy than eolian sand found in the same region.

Muhs et al. (1996), using mineralogical, geochemical, and sedimentological data, suggested that the most likely source of most eolian sand in eastern Colorado is the South Platte River. In contrast, fine-grained overbank alluvium from the South Platte River, consisting of fine sand, silt, and clay, has geochemical characteristics distinct from those of eastern Colorado loess (Fig. 11). Colorado loess is higher in Ca, and lower in Rb, Ti, and Zr concentrations than South Platte River alluvium. Samples of the White River Group collected in northern Colorado show a wide range of Ca, Sr, K, and Rb values, but overlap the range of values for eastern Colorado loess (Fig. 11). Ti and Zr concentrations of eastern Colorado loess are intermediate between White River Group sediments and South Platte River alluvium. We conclude from these data that sediments from both the South Platte River and the White River Group must be considered as possible source sediments for eastern Colorado loess.

Although loess appears to form a nearly continuous blanket from Ohio to Colorado (Fig. 1), few studies have attempted interregional comparisons of loess composition. Peoria Loess

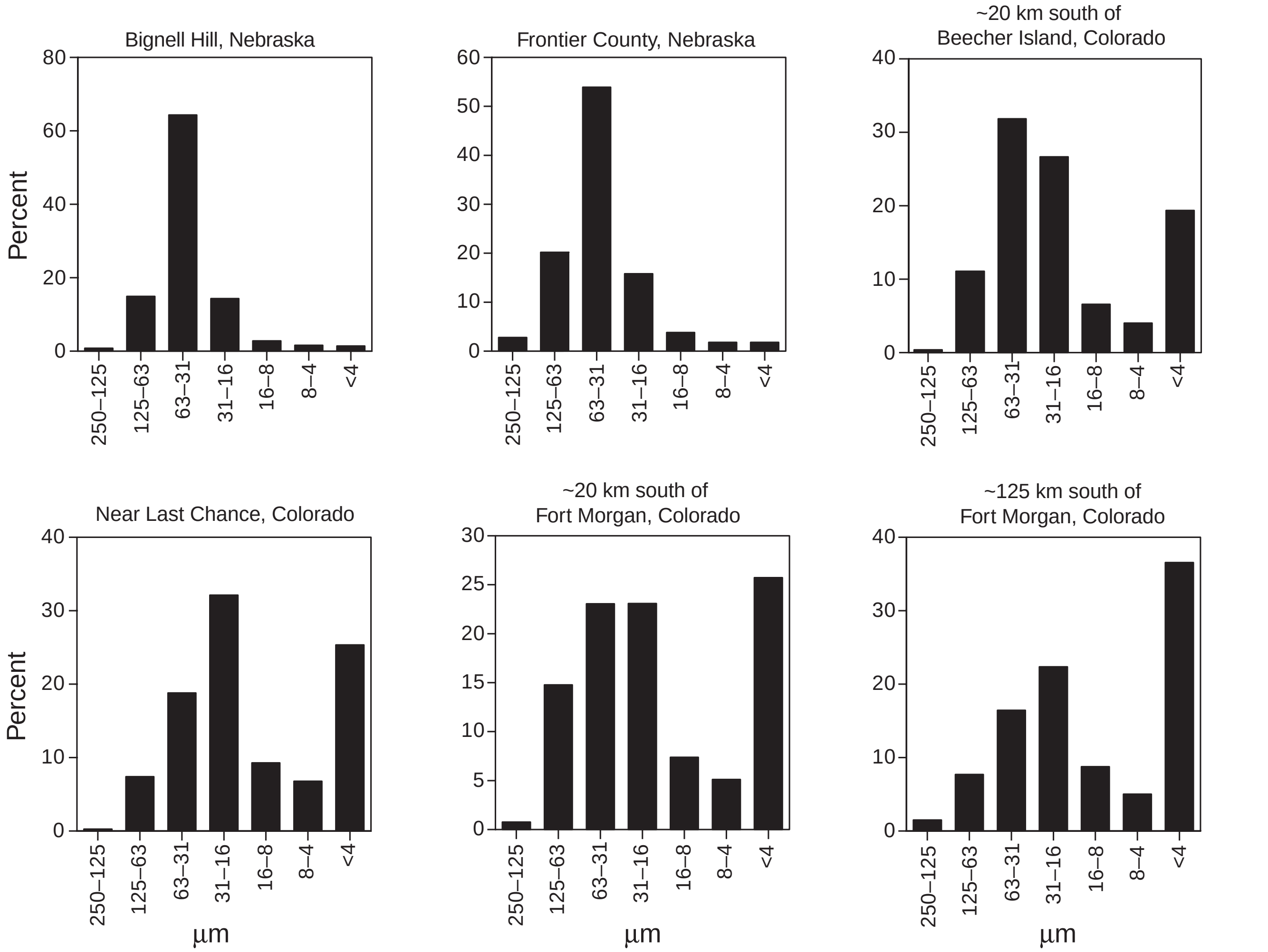

Figure 8. Histograms showing particle size class distribution in selected unweathered Peoria Loess samples of eastern Colorado and Nebraska. Nebraska data are from Swineford and Frye (1951). Note bimodal distribution for Colorado samples with secondary peak in clay fraction.

north and south of the Platte River in Nebraska and adjacent northern Kansas was sampled (Fig. 12) to compare with eastern Colorado loess. Results of geochemical analyses support the findings of Winspear and Pye (1995) that there is no significant difference in composition between loess found north and south of the Platte River in Nebraska and adjacent Kansas (Fig. 13). Ca and Sr (which largely reflect carbonate contents), K and Rb (found in mica and K-feldspar), and Ti and Zr (found in heavy minerals) show no significant differences on either side of the Platte River. Similar findings for other elements in Nebraska loess reported by Winspear and Pye (1995) led them to hypothesize the existence of an ancestral North Platte River to the north of the northernmost occurrences in Nebraska that served as the source for all the loess found to the south. However, the presence of loess-mantled, probable late Wisconsin river terraces in the North Platte, South Platte, and Platte River valleys (Swinehart et al., 1994), suggests that all three rivers were situated fairly close to their present positions during the time of Peoria Loess deposition, and would require a loess source located elsewhere.

Colorado loess differs only slightly from Nebraska loess in its chemical composition. Concentrations of Ca and K, and therefore Ca/Sr and K/Rb values, are somewhat higher in Colorado than in Nebraska (Fig. 13). Ti and Zr values are not significantly different between Colorado and Nebraska, and indicate roughly similar heavy mineral compositions. Despite the minor differences in Ca and K concentrations, however, the overall compositions of Colorado and Nebraska loess are similar, and suggest some overlap in source sediments for the two regions.

CARBON ISOTOPE COMPOSITION OF ORGANIC MATTER IN LOESS AND SOILS

Recent studies have shown that carbon isotopes in organic matter in loess and buried loess-derived soils can provide data for paleoclimatic interpretations over the past interglacial-glacial cycle (Wang et al., 1997; Hatté et al., 1998). Two types of photosynthetic pathways, C_3 and C_4, fractionate carbon isotopes in distinctly different ways. C_3 plants, which include cool-season grasses and all trees, have $\delta^{13}C$ values that range from about –22‰ to about –34‰ and average about –26‰, whereas C_4 plants, which include most warm-season grasses, range from about –9‰ to about –20‰ and average about –12‰ (O'Leary, 1988). Teeri and Stowe (1976) found that the relative pro-

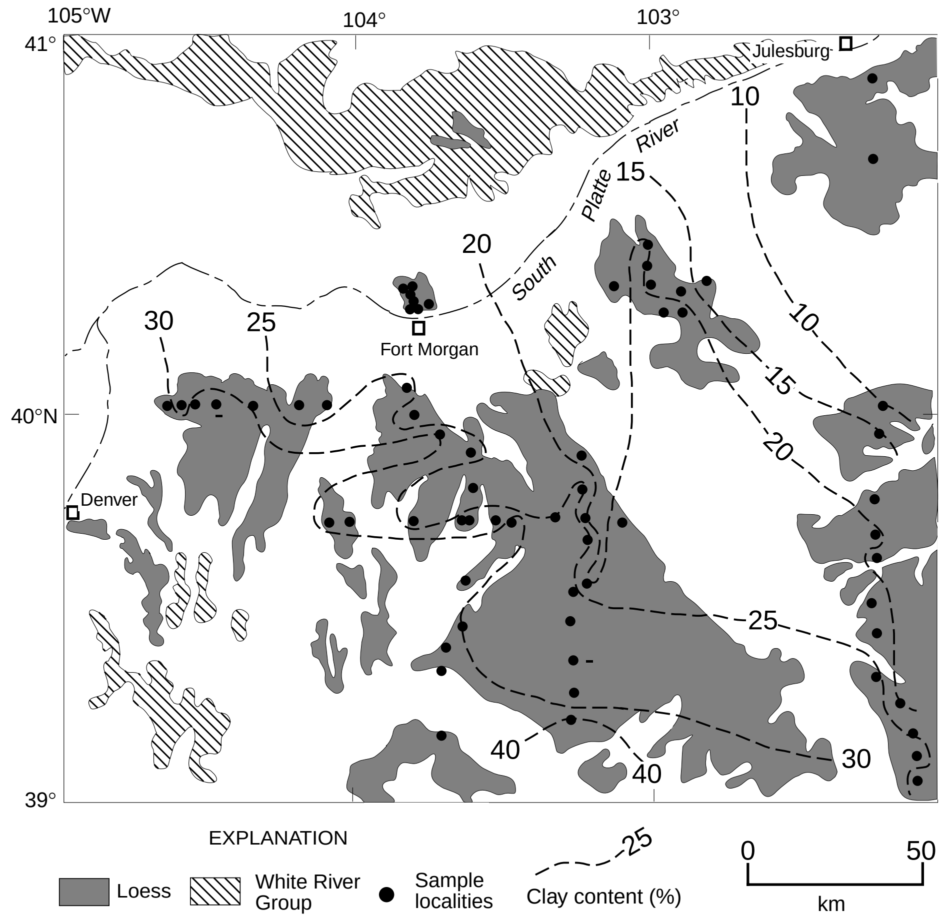

Figure 9. Map showing contours of clay (particles <4 µm) content (%) in eastern Colorado loess. Minus symbols indicate samples with clay contents lower than the contoured value. Only included are those localities from Figure 2 that, through degree of soil development, are correlated as Peoria Loess, are relatively thick, and are not eroded.

portion of C_4 grasses in North America is positively correlated with summer temperatures and length of growing season. Fredlund and Tieszen (1997a) and Follett et al. (1997) measured the carbon isotopic composition of soil organic matter from native grassland sites throughout the Great Plains and found that $\delta^{13}C$ values are in general positively correlated with mean July temperature. The northern Great Plains area is dominated by cool-season C_3 grasses, and the central and southern Great Plains region, including eastern Colorado, is dominated by warm-season C_4 grasses.

Carbon isotopic composition of in situ derived organic matter in buried soils and in situ or detrital organic matter in loess show major changes in dominant vegetation types over the last glacial-interglacial cycle. At Last Chance, the lowermost buried soil underlying the loess has $\delta^{13}C$ values of about –22‰, indicating a probable mixture of C_3 and C_4 vegetation (but with more C_3 vegetation), and the buried soil above this, dated as ca. 23 000–21 000 ^{14}C yr B.P., has $\delta^{13}C$ values of about –24‰, indicating a dominance of C_3 vegetation (Fig. 14). In contrast, the modern soil has values of about –17‰ in the Bk horizon and as high as about –14‰ in the A horizon, indicating a dominance of C_4 vegetation, such as covered the region in presettlement time. A similar sequence of changes is evident at Beecher Island. The lowermost buried soil, dated as ca. 20 000 ^{14}C yr B.P., has $\delta^{13}C$ values of –23‰ to –19‰, indicating a probable mix of C_3 and C_4 vegetation (Fig. 14). In the overlying Peoria Loess, detrital organic matter has $\delta^{13}C$ values ranging from about –23‰ to –25‰, indicating a dominance of C_3 vegetation at the time of loess fall. However, the A horizon of the ca. 11 000 ^{14}C yr B.P. buried soil and the loess and modern soil above it have $\delta^{13}C$ values of –16‰ to –17‰, indicating a dominance of C_4 vegetation. Overall, the carbon isotopic compositions indicate a dominance of C_3 vegetation from about 23 000–12 000 ^{14}C yr B.P., and a dominance of C_4 vegetation after ca. 12 000 ^{14}C yr B.P.

DISCUSSION

Timing of Loess Deposition

The ages of Peoria Loess in eastern Colorado and western Nebraska reported here suggest that although there is a general agreement in the timing of last-glacial loess deposition in various parts of the midcontinent, there are differences in detail. In central and western Illinois, Peoria Loess deposition probably began sometime after about 25 000 ^{14}C yr B.P. (Curry and Follmer, 1992), and was still in progress ca. 12 000 ^{14}C yr B.P. (Grimley et al., 1998). To the west in Iowa, Peoria Loess deposition began sometime after about 24 000–19 000 ^{14}C yr B.P., depending on locality, but was completed by about 14 000 ^{14}C yr B.P. (Ruhe, 1983). Loess deposition in Illinois, Iowa, and areas eastward was to a great extent a function of source sediment availability from the Laurentide ice sheet via the Mississippi and Missouri Rivers, and the timing of loess deposition closely followed the history of movement of the ice sheet. Based on the AMS radiocarbon ages from Bignell Hill reported here, and other ages reported by May and Holen (1993), Martin (1993), Feng et al. (1994), and Maat and Johnson (1996), Peoria Loess deposition in western Nebraska could have begun earlier (sometime after ca. 30 000 ^{14}C yr B.P.) and continued until significantly later (ca. 10 500 ^{14}C yr B.P.), suggesting that sources of loess were unrelated to the specific dynamics of the Laurentide ice sheet. The Laurentide ice sheet advanced to just north of the present position of the Missouri River at the Nebraska–South Dakota state line during the last glacial (Fig. 1), but it is unlikely that the ice sheet was anywhere near the drainage of the Missouri River ca. 30 000 ^{14}C yr B.P. (see Andrews, 1987, p. 18), and was probably north of the Canadian border by ca. 10 500 ^{14}C yr B.P. (Dyke and Prest, 1987).

Loess deposition in eastern Colorado apparently began after ca. 20 000 ^{14}C yr B.P., but may have continued until ca. 12 000 ^{14}C yr B.P. Thus, eolian sedimentation in Colorado may have lagged that in Nebraska by as much as ~10 k.y., but continued ~2 k.y. later than in Iowa. A pertinent question is whether the timing of loess deposition in Colorado was synchronous with last-glacial (Pinedale) history in the Front Range. An overall summary of the Rocky Mountain glacial record suggests a culmination of glacial advances ca. 23 000 ^{14}C yr B.P. (Porter et al., 1983). The last Pinedale advance in the Front Range is estimated to have occurred from ca. 30 000 to 12 000 ^{14}C yr B.P., with a maximum advance ca. 22 000 ^{14}C yr B.P. (Nelson et al., 1979; Madole, 1986). Somewhat younger ages for the time of maximum advance and deglaciation are estimated from ^{10}Be dating of Pinedale moraines in the Wind River Range of Wyoming by Gosse et al. (1995a). The ^{10}Be age estimates are, however, dependent upon calibration of cosmogenic radiation flux rates that are not known with certainty. If all the age estimates of glacial events are approximately correct, then the stratigraphic data and radiocarbon ages in this study indicate that loess deposition began near the culmination of, and occurred mostly during, the recession of Rocky Mountain glaciers.

Other evidence suggests that the youngest (Beecher) loess in eastern Colorado may also be related to glacial events. Gosse et al. (1995b) dated a slightly younger (i.e., younger than the main part of the Pinedale) glacial advance in the Wind River Range that they correlated to the Younger Dryas event of Europe. Menounos and Reasoner (1997) reported stratigraphic and radiocarbon evidence of a possible Younger Dryas event in the Front Range of Colorado. The latter workers bracketed this event between 11 070 ± 50 and 9970 ± 80 ^{14}C yr B.P., ages that are remarkably close to the bracketing ages for the Beecher loess (their samples were also humic acids extracted using the same methods and in the same laboratory as those from Beecher Island). However, establishment of whether the hypothesized Younger Dryas events in the Rocky Mountains and Great Plains are real will require more well-dated sequences. We note that the carbon isotope data from Beecher Island do not suggest a return to a cooler, C_3 vegetation type during the hypothesized Younger Dryas period (Fig. 14).

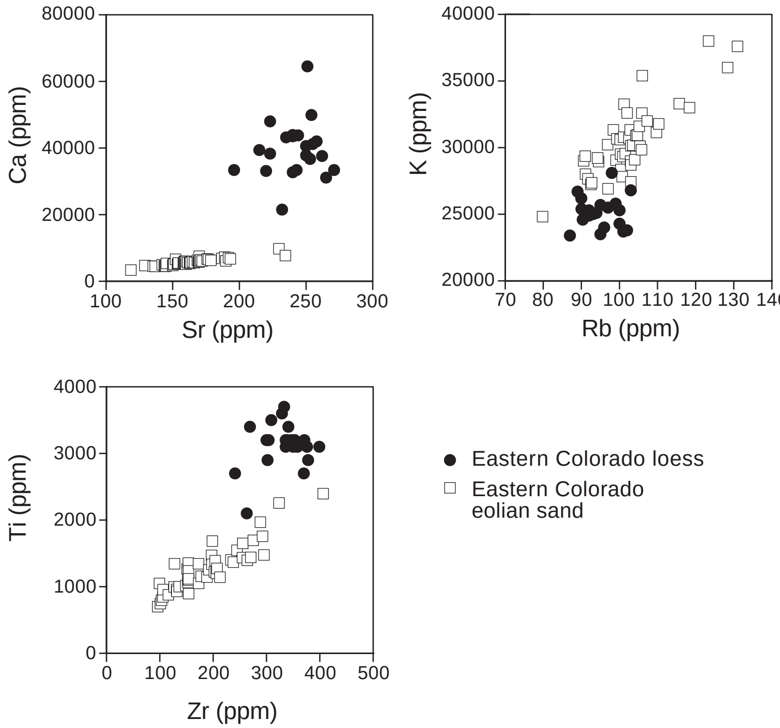

Figure 10. Plots showing concentrations of selected elements in eastern Colorado loess compared to eolian sand in eastern Colorado. Eolian sand data are from Muhs et al. (1996).

Sources of Loess

Particle size, mineralogical, and geochemical data suggest that eastern Colorado loess probably had at least two and possibly three sources. Both glacially derived silt from the Front Range, carried to the Great Plains by the South Platte River, and volcaniclastic silt from the White River Group are potential sources for eastern Colorado loess. Pye (1987) pointed out that fluvial systems provide a very efficient means by which silt can be segregated from clay, which suggests that the South Platte River, carrying glaciogenic silt from the Front Range, could be the most important source. This is reinforced by studies that suggest that sediment yield is greatly increased (see review by Hallet et al., 1996) with expanded glaciers, such as those in the Front Range during Pinedale time. However, the White River Group has a very high silt content and chemical composition that overlap that of eastern Colorado loess. A clay-rich source must be invoked to explain the anomalous clay modes in the particle size distribution of Colorado loess, and the best candidate to explain the generally higher clay contents is the Pierre Shale. Colton (1978) mapped deflation basins to the northwest of the loess belt in northeastern Colorado. He interpreted these features to have formed from wind erosion, and many of them are found in the Pierre Shale. Thus, it is possible that some of the clay-sized material in northeastern Colorado loess was derived from such deflation basins.

We conclude that the methods in this study cannot be used to determine which source, if any, was dominant during loess deposition. However, isotopic methods (Aleinikoff et al., 1999, companion paper in this volume) provide a means by which to make this determination. Given that multiple sources are potentially responsible for eastern Colorado loess, it is not possible to generate the classical loess distance-decay equations summarized by Ruhe (1983). However, if all three eastern Colorado source sediments made at least some contribution to late Pleistocene loess, paleowinds were most likely from the northwest, west, and southwest (Fig. 2).

ENVIRONMENT OF LOESS DEPOSITION

The climatic and ecological conditions accompanying loess fall in the central Great Plains have been studied by several workers. Wells and Stewart (1987) suggested, on the basis of fossil snails and *Picea* macrofossils, that during the period of Peoria Loess deposition in western Nebraska and Kansas there was a mix of boreal forest and aspen parkland. They did not envision a loess steppe during the main period of loess deposition in late Pleistocene time. In contrast, Rousseau and Kukla (1994) interpreted snail assemblages in western Nebraska to indicate a progressive cooling and drying through the period of loess deposition and a cold steppe vegetation, with only scattered, if any, trees. A cool steppe ca. 18 000 ^{14}C yr B.P. is also inferred for the Colorado Piedmont near Denver based on fossil beetle assemblages (Elias and

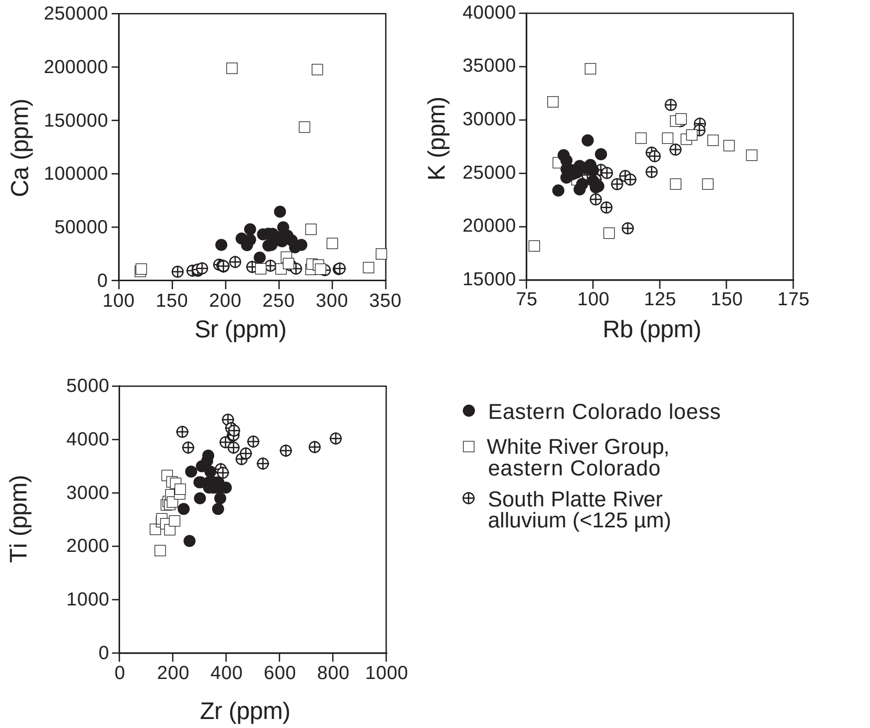

Figure 11. Plots showing concentrations of selected elements in eastern Colorado loess compared to fine-grained (<125 µm) overbank alluvium of the South Platte River and White River Group rocks found in eastern Colorado (localities shown in Fig. 2).

Toolin, 1990). Fredlund and Tieszen (1997b) reported phytolith and some pollen data from several loess sites in Nebraska and Kansas, ranging in age from ca. 14 500 to ca. 10 400 ^{14}C yr B.P. Although pollen from a 12 600 ^{14}C yr B.P. locality is dominated by *Picea*, there are significant amounts of grassland pollen types, and phytolith data support the interpretation that a parkland-type of environment existed in the region. Phytolith data indicate that C_3 grasses were dominant, indicating relatively cool conditions compared to present. Using both phytolith and carbon isotope data, Fredlund and Tieszen (1997a) reported that near Wind Cave in the southern Black Hills area of South Dakota, C_3 grassland was dominant until ca. 11 000 ^{14}C yr B.P., at which time C_4-dominated grassland became prominent and persisted until the present.

We hoped to use the snails found in both the upper Peoria Loess and Beecher loess to reconstruct the environment at the time of loess deposition, following the approach of Rousseau and Kukla (1994). Unfortunately, the only taxon present, *Succinea grosvenori*, apparently has a very wide modern geographic distribution (Pilsbry, 1948), and therefore says little about paleoclimate. However, two faunal localities in Peoria Loess very close to Beecher Island studied by Graham (1981) provide a more detailed picture of the environment at the time of loess deposition. Fossil ungulates and rodents were recovered from Peoria Loess at these localities; the ungulates (*Camelops, Platygonus, and Equus*) are all grazers, and the rodents (*Spermophilus richardsoni, Spermophilus* sp., *Cynomys ludovicianus*, and *Thomomys talpoides*) are indicative of a grassland environment. On the basis of faunal data, Graham (1981) concluded that grassland was the predominant vegetation in eastern Colorado during the time of Peoria Loess deposition. Carbon isotope data from the present study indicate that this vegetation was dominated by cool-season grasses.

The vegetation reconstruction from this study is consistent with evidence from other workers that latest Pleistocene time was significantly cooler than present in eastern Colorado. If our reconstruction of a cool steppe vegetation is correct, Fredlund and Tieszen's (1997a) correlation of grassland soil carbon isotopic composition with July temperature can be used to calculate a minimum-limiting last glacial temperature depression compared to present. Mean July temperatures near Beecher Island are ~23.7 °C, but the closest locality with modern carbon isotopic compositions similar to those of last glacial time in eastern Colorado is Stavely, Alberta, which has a modern mean July temperature of 18.4 °C (Fredlund and Tieszen, 1997a). This comparison indicates a last glacial July temperature depression of at least 5–6 °C, and is consistent with fossil evidence (Elias, 1996) and glacial modeling estimates of summer temperature depressions (Leonard, 1989) from elsewhere in Colorado. In nearby southeastern Wyoming, fossil ice wedge and sand wedge casts indicate mean annual temperature depressions of 10–16 °C during glacial maxima (Mears, 1997).

The carbon isotope data and reconstructed cool steppe presented here for eastern Colorado

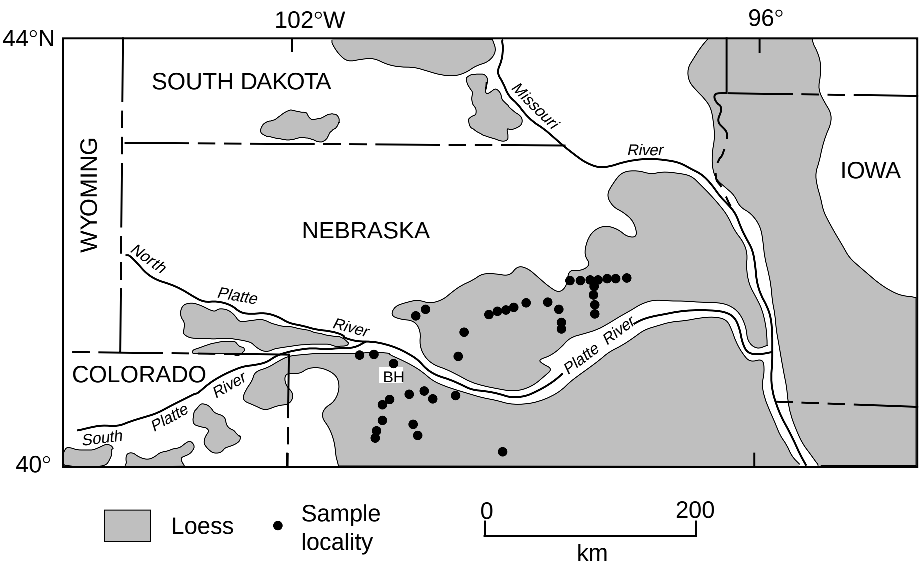

Figure 12. Map showing the distribution of Peoria Loess (shaded) in Nebraska and adjacent states and geochemical sampling localities (filled circles). A few samples (not shown) were also collected in northern Kansas. Loess distribution is from Hallberg et al. (1991), Swinehart et al. (1994), and this study. BH—Bignell Hill.

and elsewhere in the Great Plains have interesting implications for the effects of lower CO_2 on vegetation during the last glacial maximum. During the last glacial maximum, atmospheric CO_2 concentrations were on the order of 185–205 ppmv (parts per million volume) (Neftel et al., 1988), compared to modern, preindustrial concentrations of 270–280 ppmv. Street-Perrott et al. (1997) pointed out that lower atmospheric CO_2 values favor the spread of C_4 plants, because these taxa have a CO_2-concentrating mechanism. These workers reported evidence from high-altitude lake sediments in East Africa that C_4 plants, including grasses, increased in abundance and range during the last glacial maximum, and attributed this increase to a lower CO_2 concentration in the atmosphere. Peng et al. (1998) also suggested that C_4-dominated grasslands may have expanded at the expense of forest during the last glacial maximum due to lower CO_2 concentrations. However, data reported here, as well as from South Dakota and Texas, indicate that C_3 grasses were dominant over much of the western part of the Great Plains during the last glacial maximum (Fig. 15). Apparently, the lower temperatures on the Great Plains during the last glacial maximum were more important than the lower concentrations of CO_2 in the determination of which grass taxa would be dominant.

Although the sediments of bedrock loess sources—such as the White River Group and the Pierre Shale—are widely exposed at the surface

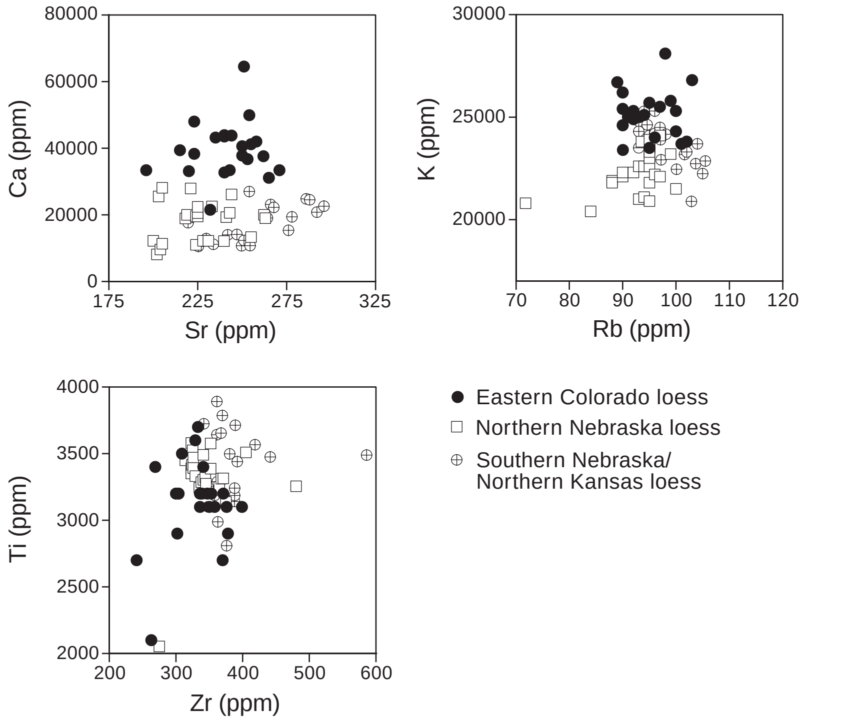

Figure 13. Plots showing concentrations of selected major and trace elements in eastern Colorado loess compared to loess in Nebraska and adjacent Kansas. Northern Nebraska loess includes all localities shown in Figure 12 that are north of the Platte River; southern Nebraska–northern Kansas loess includes all localities shown in Figure 12 that are south of the Platte River.

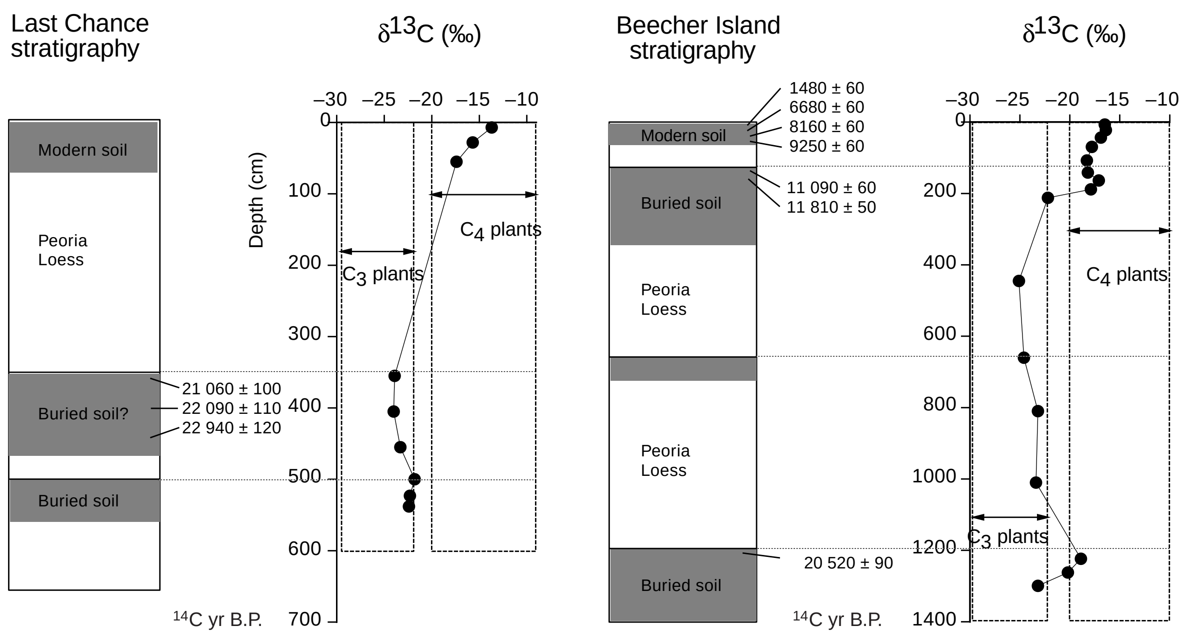

Figure 14. Carbon isotopic composition of soil and loess organic matter at the Last Chance and Beecher Island localities.

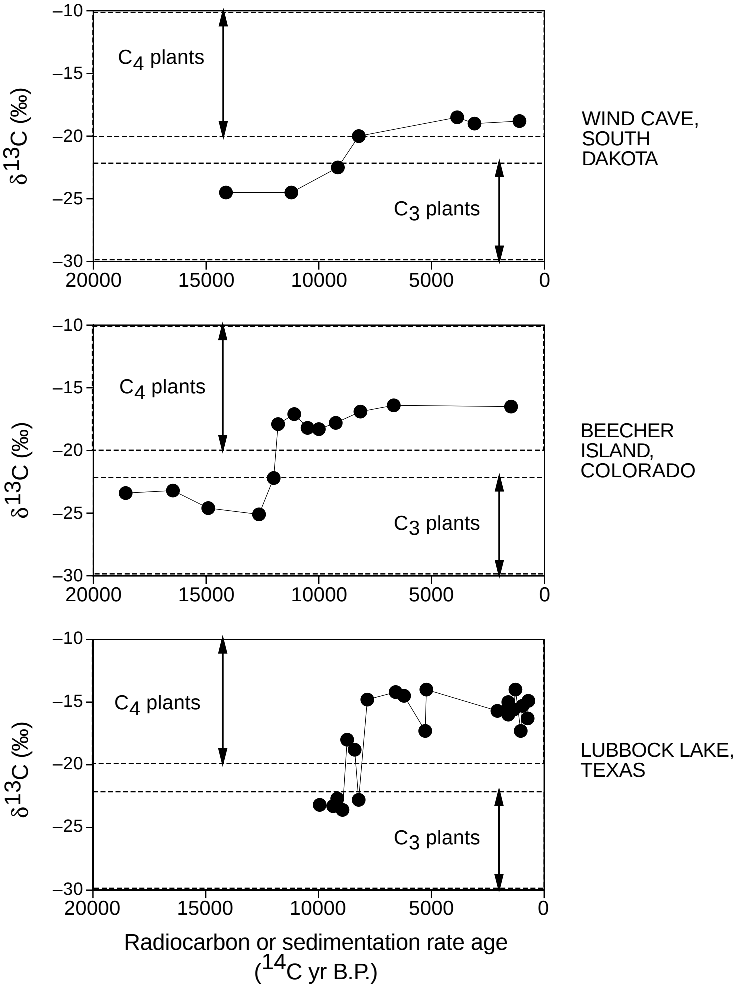

Figure 15. Carbon isotopic composition of soil or sediment organic matter as a function of apparent radiocarbon age from three localities in the Great Plains. Wind Cave data are from Fredlund and Tieszen (1997a), Beecher Island data are from this study, and Lubbock Lake data are from Holliday (1995). The ages of loess from Beecher Island are based on an assumed linear vertical sedimentation rate of ~96 cm/1000 yr calculated from loess thickness and ages of bracketing buried soils.

over northeastern Colorado (Scott, 1978; Sharps, 1980), there is, at present, little or no eolian movement of silt from these sediments. An argument could be made that the present stability of these sediments is due to a lower degree of wind erosive power compared to the last glacial maximum. Persistence of stronger winds during glacial periods compared to interglacial periods was cited by Phillips et al. (1993) as a rationale to explain greater dust flux from loess source areas in China during the last glacial maximum. Based on daily wind measurements at a height of 3 m from 1984–1991 records, mean wind velocity at Akron, Colorado, is ~4 m/s; however, winds during some springs (e.g., 1988) have mean values as high as ~5 m/s, and often are in excess of 9 m/s. During the drought of the 1950s, winds that generated dust storms in eastern Colorado had (at heights of ~2.5 m) velocities of 6–14 m/s (Chepil and Woodruff, 1957). These values agree closely with wind velocities of 4–14 m/s that have been recorded in modern dust storms in periglacial environments in Alaska, Canada, Iceland, and New Zealand (Péwé, 1951; Nickling, 1978; Ashwell, 1986; McGowan et al., 1996). We conclude that stronger winds, if they existed during the last glacial maximum, could have enhanced the potential for eolian entrainment of silt in eastern Colorado. However, such winds were probably not the major factor for loess deposition in eastern Colorado, because present winds are strong enough to entrain silt, at least seasonally.

Erosion of the White River Group and Pierre Shale by the wind during the last glacial maximum could imply that the vegetation cover was absent or greatly diminished, which implies conditions drier than present. Alternatively, if temperatures were much lower, increased frost action may have generated greater sediment delivery to streams without drier-than-present conditions. In either case, it is unlikely that all sediment was eroded directly by wind from bedrock sources during the last glacial maximum; many particles were probably first delivered to tributaries of the South Platte River by hillslope erosion via overland flow or frost action over short distances, followed by eolian transport from stream valleys. However, the presence of eolian deflation basins to the west of loess occurrences in northeastern Colorado (Colton, 1978) indicates that some direct eolian erosion of bedrock residuum may have occurred.

Collectively, the sedimentological, geochemical, and carbon isotope data indicate that during the last glacial maximum, eastern Colorado was probably both colder and drier than present. A likely vegetation type might be a sparsely vegetated cool steppe, similar to that found today in parts of the prairie provinces of Canada. Colder and drier conditions during the last glacial maximum are in agreement with atmospheric general circulation model reconstructions of temperature and precipitation for the region summarized by Thompson et al. (1993), Kutzbach et al. (1993, 1998), and Bartlein et al. (1998).

CONCLUSIONS

1. Loess, previously not well studied in eastern Colorado, covers an estimated 14 000 km^2, and is the westernmost part of the North American midcontinent loess belt.

2. Stratigraphic studies indicate there were two periods of loess deposition in eastern Colorado during late Quaternary time. The first spanned ca. 20 000 to 12 000 ^{14}C yr B.P., and correlates reasonably well with the culmination and recession of Pinedale glaciers in the Colorado Front Range. This period of loess fall also correlates with the main period of Peoria Loess deposition in Illinois, but lasted longer than in Iowa, and not as long as in Nebraska. The second period of loess deposition occurred between ca. 11 000 and ca. 9000 ^{14}C yr B.P., and may be Holocene age or correlate with a hypothesized Younger Dryas glacial advance in the Colorado Front Range. However, carbon isotope data argue against cooler conditions in eastern Colorado during the hypothesized Younger Dryas period.

3. Sedimentologic, mineralogic, and geochemical data suggest that as many as three sources could have supplied loess in eastern Colorado, including glaciogenic silt, volcani-

clastic silt from the White River Group, and probably clay from the Pierre Shale. If these were the sources of loess, then northwesterly, westerly, and southwesterly paleowinds during full-glacial time are implied. Bedrock residuum sources for eastern Colorado loess may also imply a lesser vegetation cover and therefore drier conditions during the last glacial maximum.

4. Carbon isotope data from soil and loess organic matter, combined with published faunal data, imply a cooler, C_3-dominated vegetation type, most likely cool steppe, during the time of loess deposition. This vegetation reconstruction suggests that temperature, rather than concentration of carbon dioxide, is a more important control over the type of grassland community in this region, in contrast to the tropics, where C_4 vegetation expanded during the last glacial period.

ACKNOWLEDGMENTS

This study was supported by the Global Change and Climate History Program of the U.S. Geological Survey. We thank Glenn Goodfriend (Carnegie Institution) for identifying the snails at Beecher Island. Jim Burdett (Cornell University) conducted the carbon isotope analyses. The High Plains Climate Center (Lincoln, Nebraska) kindly supplied wind data for Akron, Colorado, and the Natural Resources Conservation Service (Soil Survey Laboratory, Lincoln, Nebraska) provided organic carbon data for soil profiles in Nebraska, Kansas, and Colorado. Pete Birkeland (University of Colorado), Vance Holliday (University of Wisconsin), Jay Quade (University of Arizona), Ken Pierce (U.S. Geological Survey), and Art Bettis (University of Iowa) made helpful suggestions on an earlier draft of the paper.

REFERENCES CITED

Abbott, M. B., and Stafford, T. W., Jr., 1996, Radiocarbon geochemistry of modern and ancient Arctic lake systems, Baffin Island, Canada: Quaternary Research, v. 45, p. 300–311.

Aleinikoff, J. N., Muhs, D. R., Sauer, R. R., and Fanning, C. M., 1999, Late Quaternary loess in northeastern Colorado: Part II—Pb isotopic evidence for the variability of loess sources: Geological Society of America Bulletin, v. 111, p 1876–1883.

Andrews, J. T., 1987, The Late Wisconsin glaciation and deglaciation of the Laurentide Ice Sheet, *in* Ruddiman, W. F., and Wright, H. E., Jr., North America and adjacent oceans during the last deglaciation: Boulder, Colorado, Geological Society of America, Geology of North America, v. K-3, p. 13–37.

Ashwell, I. Y., 1986, Meteorology and duststorms in central Iceland: Arctic and Alpine Research, v. 18, p. 223–234.

Bartlein, P. J., Anderson, K. H., Anderson, P. M., Edwards, M. E., Mock, C. J., Thompson, R. S., Webb, R. S., Webb, T., III, and Whitlock, C., 1998, Paleoclimate simulations for North America over the past 21,000 years: Features of the simulated climate and comparisons with paleoenvironmental data: Quaternary Science Reviews, v. 17, p. 549–585.

Broccoli, A. J., and Manabe, S., 1987, The effects of the Laurentide ice sheet on North American climate during the last glacial maximum: Géographie Physique et Quaternaire, v. 41, p. 291–299.

Bryant, B., McGrew, L. W., and Wobus, R., 1981, Geologic map of the Denver 1° × 2° quadrangle, north-central Colorado: U.S. Geological Survey Miscellaneous Investigations Series Map I-1163, scale 1:250 000.

Chepil, W. S., and Woodruff, N. P., 1957, Sedimentary characteristics of dust storms: II. Visibility and dust concentration: American Journal of Science, v. 255, p. 104–114.

COHMAP Members, 1988, Climatic changes of the last 18,000 years: Observations and model simulations: Science, v. 241, p. 1043–1052.

Colton, R. B., 1978, Geologic map of the Boulder–Fort Collins–Greeley area, Colorado: U.S. Geological Survey Miscellaneous Investigations Series Map I-855-G, scale 1:100 000.

Crabb, J. A., 1980, Soil survey of Weld County, Colorado, southern part: Washington, D.C., U.S. Government Printing Office, 135 p.

Curry, B. B., and Follmer, L. R., 1992, The last interglacial-glacial transition in Illinois: 123–25 ka, *in* Clark, P. U., and Clark, P. D., eds., The last interglacial-glacial transition in North America: Geological Society of America Special Paper 270, p. 71–88.

Dare-Edwards, A. J., 1984, Aeolian clay deposits of southeastern Australia: Parna or loessic clay: Institute of British Geographers Transactions, new series, v. 9, p. 337–344.

Denson, N. M., and Bergendahl, M. H., 1961, Middle and upper Tertiary rocks of southeastern Wyoming and adjoining areas: U.S. Geological Survey Professional Paper 424-C, p. C-168–C-172.

Dyke, A. S., and Prest, V. K., 1987, Late Wisconsinan and Holocene history of the Laurentide Ice Sheet: Géographie Physique et Quaternaire, v. 41, p. 237–263.

Elias, S. A., 1996, Late Pleistocene and Holocene seasonal temperatures reconstructed from fossil beetle assemblages in the Rocky Mountains: Quaternary Research, v. 46, p. 311–318.

Elias, S. A., and Toolin, L. J., 1990, Accelerator dating of a mixed assemblage of late Pleistocene insect fossils from the Lamb Spring site, Colorado: Quaternary Research, v. 33, p. 122–126.

Feng, Z., Johnson, W. C., Lu, Y., and Ward, P. A., III, 1994, Climatic signals from loess-soil sequences in the central Great Plains, USA: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 110, p. 345–358.

Follett, R. F., Paul, E. A., Leavitt, S. W., Halvorson, A. D., Lyon, D., and Peterson, G. A., 1997, Carbon isotope ratios of Great Plains soils and in wheat-fallow systems: Soil Science Society of America Journal, v. 61, p. 1068–1077.

Follmer, L. R., 1996, Loess studies in central United States: Evolution of concepts: Engineering Geology, v. 45, p. 287–304.

Forman, S. L., Oglesby, R., Markgraf, V., and Stafford, T., 1995, Paleoclimatic significance of late Quaternary eolian deposition on the Piedmont and High Plains, central United States: Global and Planetary Change, v. 11, p. 35–55.

Fredlund, G. G., and Tieszen, L. L., 1997a, Phytolith and carbon isotope evidence for late Quaternary vegetation and climate change in the southern Black Hills, South Dakota: Quaternary Research, v. 47, p. 206–217.

Fredlund, G. G., and Tieszen, L. L., 1997b, Calibrating grass phytolith assemblages in climatic terms: Application to late Pleistocene assemblages from Kansas and Nebraska: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 136, p. 199–211.

Frye, J. C., and Leonard, A. B., 1951, Stratigraphy of the late Pleistocene loesses of Kansas: Journal of Geology v. 59, p. 287–305.

Gosse, J. C., Klein, J., Evenson, E. B., Lawn, B., and Middleton, R., 1995a, Beryllium–10 dating of the duration and retreat of the last Pinedale glacial sequence: Science, v. 268, p. 1329–1333.

Gosse, J. C., Evenson, E. B., Klein, J., Lawn, B., and Middleton, R., 1995b, Precise cosmogenic ^{10}Be measurements in western North America: Support for a global Younger Dryas cooling event: Geology, v. 23, p. 877–880.

Graham, R. W., 1981, Preliminary report on late Pleistocene vertebrates from the Selby and Dutton archeological/paleontological sites, Yuma County, Colorado: University of Wyoming Contributions to Geology, v. 20, p. 33–56.

Grimley, D. A., Follmer, L. R., and McKay, E. D., 1998, Magnetic susceptibility and mineral zonations controlled by provenance in loess along the Illinois and central Mississippi River valleys: Quaternary Research, v. 49, p. 24–36.

Hallberg, G. R., Lineback, J. A., Mickelson, D. M., Knox, J. C., Goebel, J. E., Hobbs, H. C., Whitfield, J. W., Ward, R. A., Boellstorf, J. D., Swinehart, J. B., and Dreeszen, V. H., 1991, Quaternary geologic map of the Des Moines 4° × 6° quadrangle, United States: U.S. Geological Survey Miscellaneous Investigations Series Map I-1420 (NK-15), scale 1:1 000 000.

Hallet, B., Hunter, L., and Bogen, J., 1996, Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications: Global and Planetary Change, v. 12, p. 213–235.

Hatté, C., Fontugne, M., Rousseau, D.-D., Antoine, P., Zöller, L., Tisnérat-Laborde, N., and Bentaleb, I., 1998, δ^{13}C variations of loess organic matter as a record of the vegetation response to climatic changes during the Weichselian: Geology, v. 26, p. 583–586.

Holliday, V. T., 1995, Stratigraphy and paleoenvironments of late Quaternary valley fills on the Southern High Plains: Geological Society of America Memoir 186, 136 p.

Hovan, S. A., Rea, D. K., Pisias, N. G., and Shackleton, N. J., 1989, A direct link between the China loess and marine δ^{18}O records: Aeolian flux to the north Pacific: Nature, v. 340, p. 296–298.

Kutzbach, J. E., Guetter, P. J., Behling, P. J., and Selin, R., 1993, Simulated climatic changes: Results of the COHMAP climate-model experiments, *in* Wright, H. E., Jr., Kutzbach, J. E., Webb, T., III, Ruddiman, W. F., Street-Perrott, F. A., and Bartlein, P. J., eds., Global climates since the last glacial maximum: Minneapolis, University of Minnesota Press, p. 24–93.

Kutzbach, J. E., Gallimore, R., Harrison, S. P., Behling, P., Selin, R., and Laarif, F., 1998, Climate simulations for the past 21,000 years: Quaternary Science Reviews, v. 17, p. 473–506.

Leonard, E. M., 1989, Climatic change in the Colorado Rocky Mountains: Estimates based on modern climate at late Pleistocene equilibrium lines: Arctic and Alpine Research, v. 21, p. 245–255.

Maat, P. B., and Johnson, W. C., 1996, Thermoluminescence and new ^{14}C age estimates for late Quaternary loesses in southwestern Nebraska: Geomorphology, v. 17, p. 115–128.

Madole, R. F., 1986, Lake Devlin and Pinedale glacial history, Front Range, Colorado: Quaternary Research v. 25, p. 43–54.

Madole, R. F., 1995, Spatial and temporal patterns of late Quaternary eolian deposition, eastern Colorado, U.S.A.: Quaternary Science Reviews, v. 14, p. 155–177.

Madole, R. F., VanSistine, D., and Michael, J. A., 1998, Glaciation in the upper Platte River drainage basin, Colorado: U.S. Geological Survey Geologic Investigations Series I-2644, scale 1:300 000.

Martin, C. W., 1993, Radiocarbon ages on late Pleistocene loess stratigraphy of Nebraska and Kansas, central Great Plains, U.S.A.: Quaternary Science Reviews. v. 12, p. 179–188.

Martin, C. W., and Johnson, W. C., 1995, Variation in radiocarbon ages of soil organic matter fractions from late Quaternary buried soils: Quaternary Research, v. 43, p. 232–237.

May, D. W., and Holen, S. R., 1993, Radiocarbon ages of soils and charcoal in late Wisconsinan loess, south-central Nebraska: Quaternary Research, v. 39, p.55–58.

McGovern, H. E., 1964, Geology and ground-water resources of Washington County, Colorado: U.S. Geological Survey Water-Supply Paper 1777, 46 p.

McGowan, H. A., Sturman, A. P., and Owens, I. F., 1996, Aeolian dust transport and deposition by foehn winds in an alpine environment, Lake Tekapo, New Zealand: Geomorphology, v. 15, p. 135–146.

Mears, B., Jr., 1997, Glaciation of the Wyoming mountains and Pleistocene permafrost in the basin floors: Laramie, Wyoming State Geological Survey Public Information Circular, v. 38, p. 107–114.

Menounos, B., and Reasoner, M. A., 1997, Evidence for cirque glaciation in the Colorado Front Range during the Younger Dryas chronozone: Quaternary Research, v. 48, p. 38–47.

Muhs, D. R., and Holliday, V. T., 1995, Evidence of active dune sand on the Great Plains in the 19th century from accounts of early explorers: Quaternary Research, v. 43, p. 198–208.

Muhs, D. R., Stafford, T. W., Jr., Cowherd, S. D., Mahan, S. A., Kihl, R., Maat, P. B., Bush, C. A., and Nehring, J., 1996, Origin of the late Quaternary dune fields of northeastern Colorado: Geomorphology, v. 17, p. 129–149.

Neftel, A., Oeschger, H., Staffelbach, T., and Stauffer, B., 1988, CO_2 record in the Byrd ice core 50,000–5,000 years B.P.: Nature, v. 331, p. 609–611.

Nelson, A. R., Millington, A. C., Andrews, J. T., and Nichols, H., 1979, Radiocarbon-dated upper Pleistocene glacial sequence, Fraser Valley, Colorado Front Range: Geology, v. 7, p. 410–414.

Nickling, W. G., 1978, Eolian sediment transport during dust storms: Slims River Valley, Yukon Territory: Canadian Journal of Earth Sciences, v. 15, p. 1069–1084.

O'Leary, M. H., 1988, Carbon isotopes in photosynthesis: BioScience, v. 38, p. 328–336.

Pearson, G. W., and Stuiver, M., 1993, High-precision bidecadal calibration of the radiocarbon time scale, 500–2500 BC: Radiocarbon, v. 35, p. 25–33.

Peng, C. H., Guiot, J., and Van Campo, E., 1998, Estimating changes in terrestrial vegetation and carbon storage: Using palaeoecological data and models: Quaternary Science Reviews, v. 17, p. 719–735.

Péwé, T. L., 1951, An observation on wind-blown silt: Journal of Geology, v. 59, p. 399–401.

Phillips, F. M., Zreda, M. G., Ku, T.-L., Luo, S., Huang, Q., Elmore, D., Kubik, P. W., and Sharma, P., 1993, ^{230}Th/^{234}U and ^{36}Cl dating of evaporite deposits from the western Qaidam Basin, China: Implications for glacial-period dust export from Central Asia: Geological Society of America Bulletin, v. 105, p. 1606–1616.

Pilsbry, H. A., 1948, Land mollusca of North America (north of Mexico): Academy of Natural Sciences of Philadelphia Monograph 3, v. II, part 2, p. 521–1113.

Porter, S. C., Pierce, K. L., and Hamilton, T. D., 1983, Late Wisconsin mountain glaciation in the western United States, *in* Wright, H. E., Jr., and Porter, S. C., eds., Late-Quaternary environments of the United States, Volume 1, The late Pleistocene: Minneapolis, University of Minnesota Press, p. 71–111.

Pye, K., 1987, Aeolian dust and dust deposits: London, Academic Press, p. 334.

Pye, K., Winspear, N. R., and Zhou, L. P., 1995, Thermoluminescence ages of loess and associated sediments in central Nebraska, USA: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 118, p. 73–87.

Reheis, M. C., 1980, Loess sources and loessal soil changes on a downwind transect, Boulder-Lafayette area, Colorado: The Mountain Geologist, v. 17, p. 7–12.

Rousseau, D.-D., and Kukla, G., 1994, Late Pleistocene climate record in the Eustis loess section, Nebraska, based on land snail assemblages and magnetic susceptibility: Quaternary Research, v. 42, p. 176–187.

Ruhe, R. V., 1983, Depositional environment of late Wisconsin loess in the midcontinental United States, *in* Wright, H. E., Jr., and Porter, S. C., eds., Late-Quaternary environments of the United States, Volume 1, The late Pleistocene: Minneapolis, University of Minnesota Press, p. 130–137.

Schultz, C. B., and Stout, T. M., 1945, Pleistocene loess deposits of Nebraska: American Journal of Science, v. 234, p. 231–244.

Scott, G. R., 1978, Map showing geology, structure and oil and gas fields in the Sterling 1° × 2° quadrangle, Colorado, Nebraska and Kansas: U.S. Geological Survey Miscellaneous Investigations Series Map I-1092, scale 1:250 000.

Sharps, J. A., 1980, Geologic map of the Limon 1° × 2° quadrangle, Colorado and Kansas: U.S. Geological Survey Miscellaneous Investigations Series Map I-1250, scale 1: 250 000.

Street-Perrott, F. A., Huang, Y., Perrott, R. A., Eglinton, G., Barker, P., Khelifa, L. B., Harkness, D. D., and Olago, D. O., 1997, Impact of lower atmospheric carbon dioxide on tropical mountain ecosystems: Science, v. 278, p. 1422–1426.

Stuiver, M., and Pearson, G. W., 1993, High-precision calibration of the radiocarbon time scale, AD 1950–500 BC and 2500–6000 BC: Radiocarbon, v. 35, p. 1–23.

Swineford, A., and Frye, J. C., 1951, Petrography of the Peoria Loess in Kansas: Journal of Geology, v. 59, p. 306–322.

Swinehart, J. B., Dreeszen, V. H., Richmond, G. M., Tipton, M. J., Bretz, R., Steece, F. V., Hallberg, G. R., and Goebel, J. E., 1994, Quaternary geologic map of the Platte River 4° × 6° quadrangle, United States: U.S. Geological Survey Miscellaneous Investigations Series Map I-1420 (NK-14), scale 1:1 000 000.

Teeri, J. A., and Stowe, L. G., 1976, Climatic patterns and the distribution of C_4 grasses in North America: Oecologia, v. 23, p. 1–12.

Thompson, R. S., Whitlock, C., Bartlein, P. J., Harrison, S. P., and Spaulding, W. G., 1993, Climatic changes in the western United States since 18,000 yr B.P., *in* Wright, H. E., Jr., Kutzbach, J. E., Webb, T., III, Ruddiman, W. F., Street-Perrott, F. A., and Bartlein, P. J., eds., Global climates since the last glacial maximum: Minneapolis, University of Minnesota Press, p. 468–513.

Thorp, J., and Smith, H. T. U., 1952, Pleistocene eolian deposits of the United States, Alaska, and parts of Canada: National Research Council Committee for the Study of Eolian Deposits: New York, Geological Society of America, scale 1:2 500 000.

Wang, H., Ambrose, S. H., Liu, C.-L. J., and Follmer, L. R., 1997, Paleosol stable isotope evidence for early hominid occupation of East Asian temperate environments: Quaternary Research, v. 48, p. 228–238.

Wang, Y., Amundson, R., and Trumbore, S., 1996, Radiocarbon dating of soil organic matter: Quaternary Research, v. 45, p. 282–288.

Weist, W. G., Jr., 1964, Geology and ground-water resources of Yuma County, Colorado: U.S. Geological Survey Water-Supply Paper 1539-J, 56 p.

Welch, J. E., and Hale, J. M., 1987, Pleistocene loess in Kansas—Status, present problems, and future considerations, *in* Johnson, W. C., ed., Quaternary environments of Kansas: Kansas Geological Survey Guidebook Series 5, p. 67–84.

Wells, P. V., and Stewart, J. D., 1987, Spruce charcoal, conifer macrofossils, and land snail and small-vertebrate faunas in Wisconsinan sediments on the High Plains of Kansas, *in* Johnson, W. C., ed., Quaternary environments of Kansas: Kansas Geological Survey Guidebook Series 5, p. 129–140.

Winspear, N. R., and Pye, K., 1995, Textural, geochemical and mineralogical evidence for the origin of Peoria loess in central and southern Nebraska, USA: Earth Surface Processes and Landforms, v. 20, p. 735–745.

Manuscript Received by the Society August 3, 1998
Revised Manuscript Received February 15, 1999
Manuscript Accepted March 13, 1999

Printed in U.S.A.