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Abstract

The conterminous United States contains an extensive and generally well-studied record of Last Glacial loess. The loess occurs in diverse physiographic provinces, and under a wide range of climatic and ecological conditions. Both glacial and non-glacial loess sources are present, and many properties of the loess vary systematically with distance from loess sources. United States' mid-continent Last Glacial loess is probably the thickest in the world, and our calculated mass accumulation rates (MARs) are as high as 17,500 g/m^2/yr at the Bignell Hill locality in Nebraska, and many near-source localities have MARs greater than 1500 g/m^2/yr. These MARs are high relative to rates calculated in other loess provinces around the world. Recent models of Last Glacial dust sources fail to predict the extent and magnitude of dust flux from the mid-continent of the United States. A better understanding of linkages between climate, ice sheet behaviour, routing of glacial meltwater, land surface processes beyond the ice margin, and vegetation is needed to improve the predictive capabilities of models simulating dust flux from this region.

1. Introduction

Instrumental records of global warming over the past century have raised concerns over the role of anthropogenic greenhouse gas additions in climate change. Recently, atmospheric scientists are also considering that aerosolic dust may be an important component in the Earth's radiative transfer processes (Tegen et al., 1996; Harrison et al., 2001). Ice core and marine records show that atmospheric dust loading occurs in response to climatic and environmental changes (Petit et al., 1990; Rea, 1994). Atmospheric dust loading also affects the radiative properties of the atmosphere and the impact of these feedbacks could have strong effects in global, regional, and local palaeoclimate records (Arimoto, 2001). There has been a growing appreciation of the palaeoclimatic significance of loess sequences and their intercalated palaeosols (Hovan et al., 1989; Begét, 1991; Wang et al., 1998, 2000; Muhs and Bettis, 2000, 2003; Muhs and Zárate, 2001; Grimley et al., 2003). Some of these records are considered the best continental analog of the deep-sea oxygen isotope record as a palaeoclimate record (Hovan et al., 1989; Rea, 1994).

The conterminous United States of America contains a subcontinent-scale late Quaternary loess record that covers more than 4.5 million km^2 across 42° of longitude from the Cascades eastward to the Appalachians, and 20° of latitude from Minnesota southward to Louisiana (Fig. 1). This area encompasses a number of distinct physiographic and geologic provinces, and a wide range of climates and biomes (Fenneman, 1931, 1938; Bailey et al., 1994). In this paper we discuss the stratigraphy, lithology, chronology and mass accumulation rates (MARs) of loess dating to the Last Glacial period (Marine Oxygen Isotope Stage 2, MIS 2) in order to increase the spatial extent of records that may be used to evaluate models of the role of dust in climate change. We present stratigraphic, chronologic, and mineralogical data gleaned from the literature, as well as data from unpublished studies. The sedimentation system of loess in various parts of the country, including changing source areas through time, is also considered in this paper because of the impact on regional and local geographic and geochemical patterns that may influence atmospheric loading rates and radiative effects (Kohfeld and Harrison, 2000; Harrison et al., 2001).

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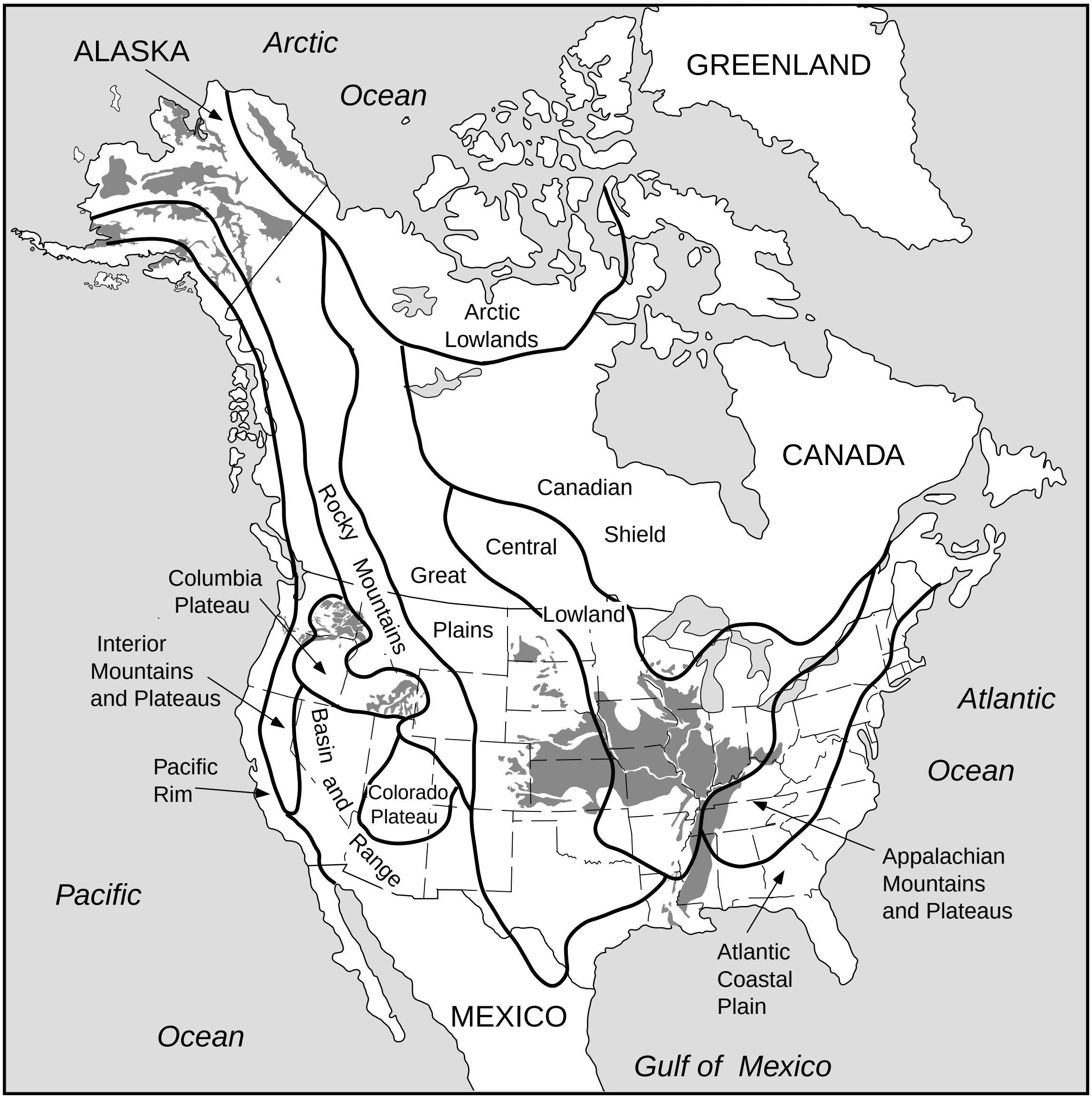

Fig. 1. Physiographic provinces of North America, modified from Fenneman (1931) and distribution of loess (shaded area). Loess distribution compiled from Péwé (1975), Lewis and Fosberg (1982), Busacca and MacDonald (1994), and sources given in Fig. 2.

2. Distribution and genesis of loess in the mid-continent

The mid-continent of North America includes the Great Plains and Central Lowland physiographic provinces (Fenneman, 1931, 1938; Fig. 1). Climate and vegetation are diverse across this enormous region. Mean annual precipitation increases from 350 to 1000 mm from west to east, and to 1500 mm in the southern part of the region (Webb et al., 1993). Vegetation patterns reflect these climatic variations. Steppe, short grass, and mixed grass prairies of the Great Plains give way to tall grass prairie, savanna, and cool-temperate deciduous woodlands of the Central Lowland. Cold-tolerant vegetation of the north gives way to warm temperate deciduous and mixed coniferous-deciduous woodlands of the south. The mid-continent contains the source areas of the Mississippi River Basin's loess sequence, and a diverse assemblage of physiographic regions and geologic terrains that all share a legacy of direct or indirect impact of Quaternary glaciation and related climate change.

The Peoria Loess (also known as Peoria Silt, Peoria Formation, Peorian loess, or Wisconsin loess), deposited during the Last Glacial maximum (LGM) in North America, is probably the thickest LGM loess in the world. Peoria Loess is more than 48 m thick at Bignell Hill in central Nebraska (Dreeszen, 1970; Maat and Johnson, 1996; Bettis et al., 2003), 41 m thick at the Loveland paratype locality in western Iowa (Muhs and Bettis, 2000), 10–20 m thick at many sections along the

Mississippi Valley (Willman and Frye, 1970; Miller et al., 1988; Markewich et al., 1998) and 10 m thick along the lower Wabash Valley in southeastern Indiana (Olson and Ruhe, 1979). Even in eastern Colorado, distant from the Laurentide Ice Sheet, Peoria Loess is as thick as 10 m (Muhs et al., 1999a). By way of comparison, LGM loess in China is at most 41 m thick (Mangshan section) and typically is 2.5–1.4 m thick (Sun et al., 2000). The thickest Last Glacial loess in Europe is about 20 m. In the subtropical region of Argentina the combined thickness of Late Pleistocene and Holocene loess ranges from an average of 10–15 m in the eastern Pampas to a maximum of about 40 m in the southern Buenos Aires Sierras (Sayago et al., 2001).

Peoria Loess has silt loam texture with local occurrences of sandier loess adjacent to valley sources, regionally extensive erosion surfaces, and outwash plains. Less commonly, such as in eastern Colorado, the Peoria is clay rich. As noted above, loess becomes finer grained and exhibits a number of geochemical decay functions with distance from source areas. Lower portions of the Peoria Loess were subjected to varying degrees of reworking very shortly after initial aeolian deposition. Evidence of secondary sedimentation is most pervasive in the relatively narrow belts of thick loess along the Missouri, Mississippi, and Platte River valleys, where the silt fell on deeply dissected landscapes. The great local relief in these areas fostered gravity-driven processes such as large-scale slumping and creep, as well as erosion by rainsplash, overland flow, and channelized flow. The products of these processes, overthickened stratigraphic sections and stacked palaeosol sequences, complicate interpretations of near-source-bluffline stratigraphic sections. Evidence for reworking ranges from laminae and thin planar bedding to discontinuous streaks and thin beds composed of transported soil material that are sometimes overturned into recumbent folds (Bettis, 1994; Mandel and Bettis, 1995). The transported and folded soil material and the presence of coarse horizontal parting and "lensoid" structure in the lower part of Peoria Loess have been attributed to the occurrence of solifluction and the presence of permafrost during the early phases of loess accumulation (ca 21,000–16,500 ^{14}C yr BP, Johnson, 1990; Bettis, 1994; Mandel and Bettis, 1995; Bettis et al., 2003).

In order to compare the Last Glacial terrestrial dust flux in the loess record of the conterminous United States with that of marine sediments, we calculated loess MARs in units of g/m^2/yr. We chose localities that spanned a wide range of total loess thickness and where reliable numerical age control exists (Fig. 2).

The MAR of loess is estimated using the equation:

$$\mathrm{MAR_{eol}(g/m^2/yr)} = \mathrm{AR(m/yr)} f_{\mathrm{eol}} \mathrm{BD(g/m^3)}, \qquad (1)$$

where AR is the accumulation rate, f_{eol} is the mass concentration of aeolian material within the sample (for loess we assume that all the material is aeolian; therefore $f_{\mathrm{eol}} = 1$), and BD is the bulk density. Assumptions regarding maximum and minimum ages of loess accumulation and section completeness and representativeness are outlined in the following discussion. Ages used to calculate the accumulation rate were obtained by luminescence and radiocarbon techniques. Luminescence ages are in calendar years, while radiocarbon ages require correction to obtain calendar years. Radiocarbon ages younger than 20,265 ^{14}C yr BP were calibrated to calendar years using Version 4.1 of the CALIB program (Stuiver and Reimer, 2000). Radiocarbon ages older than 20,265 ^{14}C yr BP were calibrated to calendar years by selecting the closest-age marine calibration point presented in Voelker et al. (2000). In the following discussion, uncorrected radiocarbon ages are designated as "^{14}C yr BP" while calendar-year corrected radiocarbon ages are designated as "cal yr BP" and luminescence ages are given in yr BP.

2.1. Loess of the Central Lowland province

In this paper the Central Lowland includes Minnesota, Iowa, Missouri, Arkansas, Wisconsin, Illinois, Indiana, and western Ohio (Fig. 1). Loess in northern and western Kentucky, western Tennessee, Mississippi and northeastern Louisiana is genetically linked with loess in the Central Lowland and is included in this discussion. This is one of Earth's most agriculturally productive areas, and loess is its most areally extensive surficial deposit. Loess of varying thickness forms an extensive sediment blanket over upland parts of the landscape and records multiple episodes of aeolian sedimentation, separated by buried soils. Three middle-to-late Pleistocene loess units (from oldest to youngest, Loveland Loess, Roxana Silt/Pisgah Fm. loess, and Peoria Loess) have been identified and correlated in this region (Shimek, 1902, 1909; Wascher et al., 1948; Leighton and Willman, 1950; Krinitzsky and Turnbull, 1967; Snowden and Priddy, 1968; Ruhe, 1969; Willman and Frye, 1970; McKay, 1979a, b; Pye and Johnson, 1988; Johnson and Follmer, 1989; Autin et al., 1991; Forman et al., 1992; Oches et al., 1996; Rodbell et al., 1997; Markewich et al., 1998; Forman and Pierson, 2002). Remnants of older middle Pleistocene loess sheets are present, but their stratigraphic relationships are not well understood (Jacobs and Knox, 1994; Markewich et al., 1998; Bettis et al., 2003). Loveland Loess is usually no more than a few metres thick and is commonly the stratigraphically oldest loess exposed at most localities. Loveland Loess has the last interglacial (MIS 5 and part of 4) Sangamon Soil developed in its upper part. This buried soil is usually thick (2–3 m), exhibits 7.5YR or

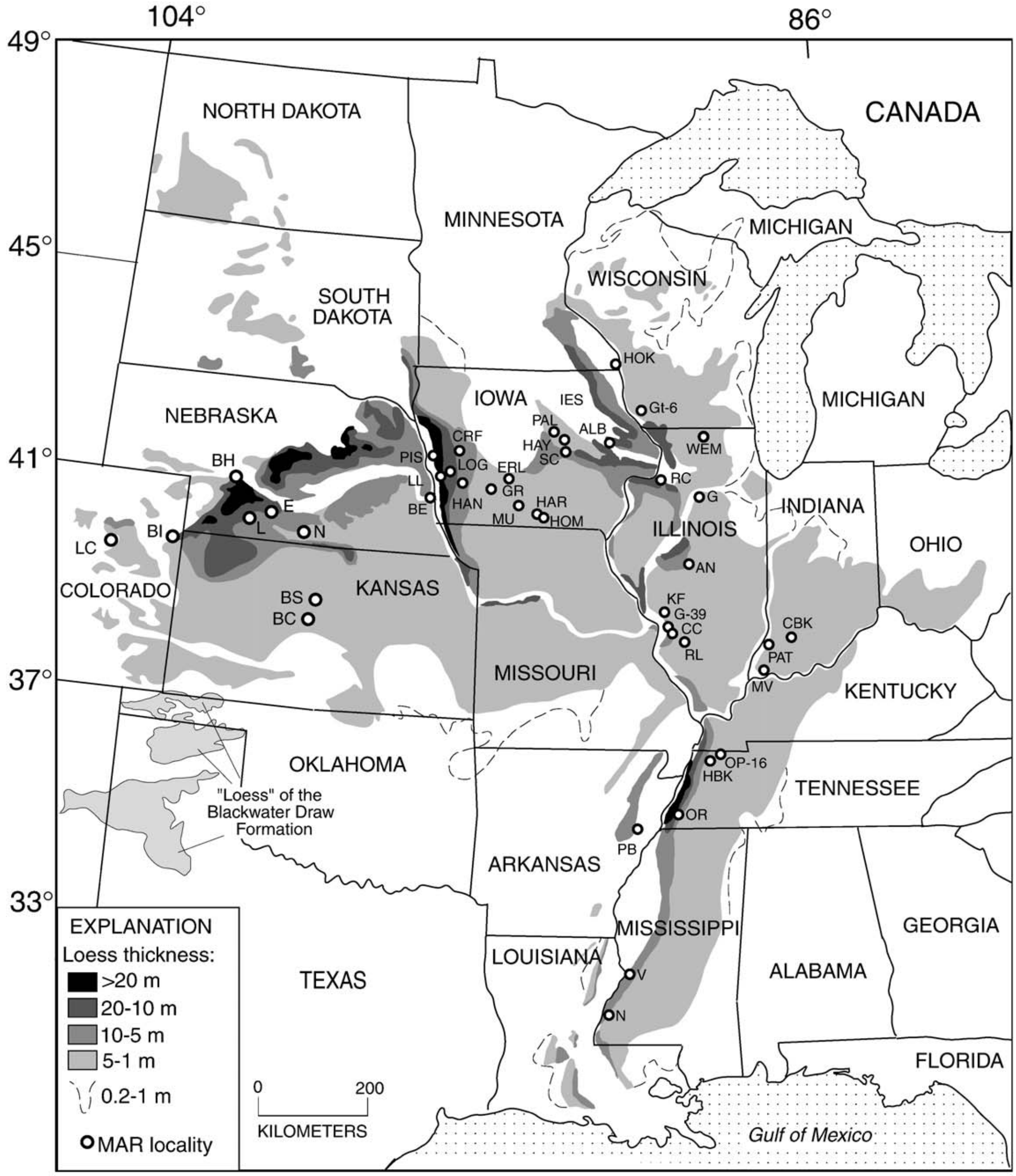

Fig. 2. Map showing the distribution and thickness of Last Glacial loess (Peoria Loess) in the mid-continent of the United States (Central Lowland and Great Plains physiographic provinces). Localities where MARs of Peoria Loess were calculated are indicated. See text for locality abbreviations. Compiled from Wascher et al. (1948), Thorp and Smith (1952), Willman and Frye (1970), Clayton et al. (1976), Hole (1976), Lineback et al. (1983), Welch and Hale (1987), Miller et al. (1988), Holliday (1989), Hallberg et al. (1991), Denne et al. (1993), Swinehart et al. (1994), and Muhs et al. (1999a).

5YR hues, has well developed angular blocky or prismatic soil structure with continuous clay films in the Bt horizon, and is leached of carbonates. The Loveland Loess is typically reddened beneath the Sangamon Soil and unaltered sections are rare.

Two Wisconsin (Last Glacial) loess sheets are present in the mid-continent. The upper unit, the Peoria Loess, accumulated during and after the LGM (equivalent to MIS 2) and is the focus of this paper. The older Wisconsin loess is the Roxana Silt (Willman and Frye, 1970; Leigh and Knox, 1993) and its stratigraphic equivalent, the Pisgah Formation loess of Iowa (Bettis, 1990). The upper portion of the Roxana/Pisgah (R/P) loess is pedogenically modified, forming the Farmdale Soil, which is the stratigraphic horizon that marks the base of the Peoria Loess. This loess has also been

referred to as the Late Sangamon Loess (Smith, 1942) and Farmdale Loess (Leighton and Willman, 1950; Olson and Ruhe, 1979). R/P loess buries the last interglacial (MIS 5 and perhaps part of Stage 4) Sangamon Soil that is developed in a variety of older deposits, including the Loveland Loess.

R/P loess has a distribution pattern similar to, but not as areally extensive as, the overlying Peoria Loess (Olson and Ruhe, 1979; Fehrenbacher et al., 1986; Johnson and Follmer, 1989). The R/P loess is significantly thinner than the Peoria Loess and cannot be traced as far from its likely source areas. It attains a maximum thickness of about 4 m east of the Missouri River Valley source in western Iowa (Bettis, 1990) and along the Lower Illinois River Valley (ancient Mississippi Valley source) in western Illinois (Leighton, 1965; Fehrenbacher et al., 1986). The distribution pattern and lithology of R/P loess suggests that it has source areas similar to the overlying Peoria Loess (Smith, 1942; Frye et al., 1962; Glass et al., 1968; Fehrenbacher et al., 1986; Johnson and Follmer, 1989; Leigh, 1994; Grimley, 2000).

Most R/P loess has been modified by pedogenesis. Thick sections near valley source areas in Illinois contain multiple buried soils (Willman and Frye, 1970). Unaltered R/P loess is found in thick sections only along the Missouri River Valley in western Iowa and along the Mississippi River Valley and Lower Illinois River Valley (Ancient Mississippi River Valley) in Illinois. In most sections, however, pedologic features associated with the Farmdale Soil, such as organic matter and iron accumulation, weak soil structure and matrix carbonate removal, extend through the R/P loess.

The contact between the Farmdale Soil, developed in the upper part of the R/P loess, and the overlying Peoria Loess is usually gradational over a few decimetres. The gradational contact is a product of initial slow Peoria Loess accumulation accompanied by mixing (pedoturbation) and syndepositional weathering. In some sections, the contact has been altered by congeliturbation associated with periglacial conditions during the LGM (Baker et al., 1991; Woida and Thompson, 1993; Bettis, 1994; Mandel and Bettis, 1995; Bettis et al., 2003).

R/P loess was deposited between about 55,000 and 23,000 ^{14}C yr BP (Fig. 3; Leigh and Knox, 1993). The basal age of R/P loess decreases with distance from source areas (Kleiss, 1973; Ruhe, 1976; Johnson and Follmer, 1989). Leigh and Knox's (1993) estimate of 55,000 yr BP for the beginning of R/P loess accumulation along the Upper Mississippi River Valley source is supported by TL (thermoluminescence) and IRSL (infrared stimulated luminescence) dating at the Bonfils Quarry site in St. Louis, Missouri (Forman and Pierson, 2002). This is considerably earlier than initiation of R/P

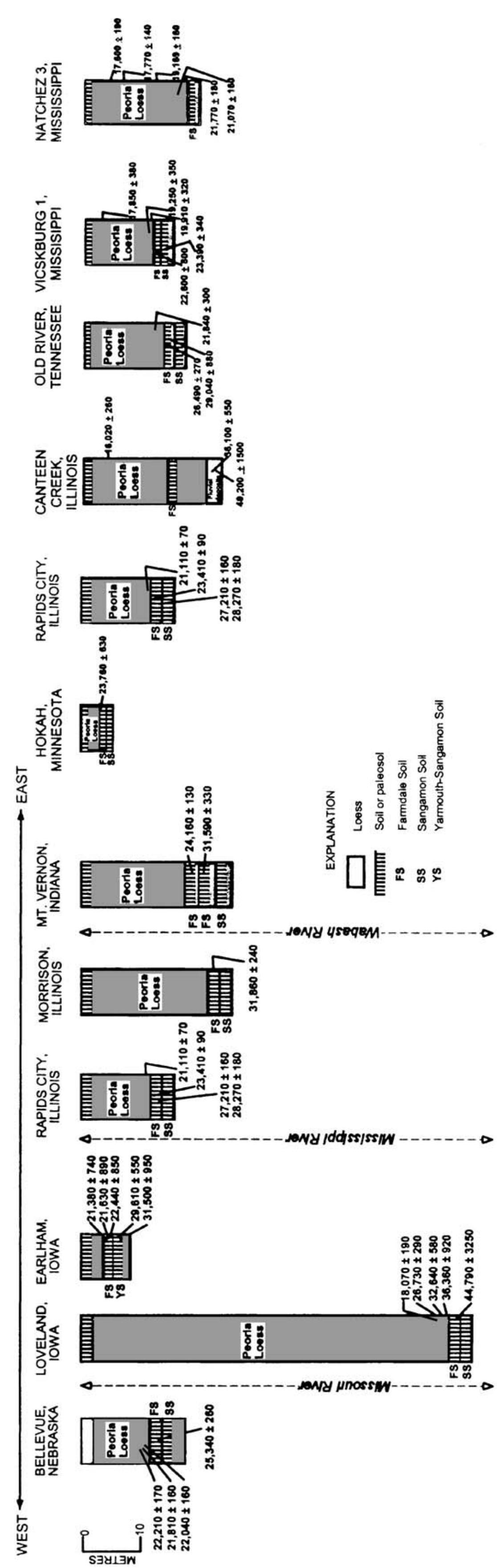

Fig. 3. Selected loess stratigraphic sections in the Central Lowland. Ages shown are ^{14}C yr BP. See Table 3 for stratigraphic and radiocarbon data sources. New AMS radiocarbon ages from Mt Vernon Indiana are previously unpublished data of the authors.

loess accumulation along the Missouri River Valley source (45,960 ± 8310 yr BP according to Norton and Bradford, 1985; 46,300 ± 3900 yr BP according to Forman and Pierson (2002) at Loveland, Iowa, Fig. 3). Minimum limiting ages for R/P loess are based primarily on radiocarbon ages on soil organic matter (SOM) from the upper horizons of the Farmdale Soil. These indicate a minimum age between 25,000 and 27,000 ^{14}C yr BP for R/P loess sourced from the Missouri River Valley (Ruhe, 1969; Ruhe et al., 1971; Bettis, 1990; Forman et al., 1992; Muhs and Bettis, 2000; Forman and Pierson, 2002; Bettis et al., 2003), the Upper Mississippi and Wabash River Valleys (Ruhe and Olson, 1978; Follmer, 1983; Ruhe, 1983; Johnson and Follmer, 1989; Leigh and Knox, 1993; Leigh, 1994), and the Lower Mississippi River Valley (Snowden and Priddy, 1968; Otvos, 1975; Pye and Johnson, 1988; Oches et al., 1996; Rodbell et al., 1997; Markewich et al., 1998). The limited numbers of available luminescence ages from R/P loess generally support this minimum limiting age (Norton and Bradford, 1985; Pye and Johnson, 1988; Forman et al., 1992; Oches et al., 1996; Markewich et al., 1998; Forman and Pierson, 2002; Bettis et al., 2003). Minimum limiting ages for the R/P loess serve as maximum limiting ages for the overlying Peoria Loess.

The youngest loess in the Central Lowland is the Peoria Loess, which, together with Peoria Loess in the adjacent Great Plains, forms the largest contiguous loess sheet in North America. Peoria Loess > 10 m thick (Fig. 2) occurs east or southeast of major river valleys that drained meltwater from the Laurentide Ice Sheet (Smith, 1942; Wascher et al., 1948; Thorp and Smith, 1952; Caldwell and White, 1956; Fehrenbacher et al., 1965; Snowden and Priddy, 1968; Kleiss, 1973; Worcester, 1973; Olson and Ruhe, 1979; Lineback et al., 1983; Ruhe, 1983; Miller et al., 1984; Fehrenbacher et al., 1986; Hallberg et al., 1991; Denne et al., 1993; Swinehart et al., 1994), and east and south of the "Iowan Erosion Surface" (IES) in northeastern Iowa and southeastern Minnesota (Hallberg et al., 1978; Mason et al., 1994, 1999).

Peoria Loess has been altered to varying degrees by the development of soils and weathering profiles. Radiocarbon ages of SOM and organic remains in the loess provide the regional chronology for the commencement and early phases of Peoria Loess deposition, as well as upper-limiting ages on the terminal phase of deposition. Luminescence dating has yielded ages in general agreement with the established radiocarbon chronology (Norton and Bradford, 1985; Forman et al., 1992; Rodbell et al., 1997; Markewich et al., 1998; Forman and Pierson, 2002; Bettis et al., 2003). The Central Lowland's Peoria Loess has traditionally been viewed as sourced from river valleys that carried glacial meltwater from the Laurentide Ice Sheet (Ruhe, 1983). The loess is thick adjacent to the major valley train sources and thins and becomes finer grained downwind of these linear sources (Smith, 1942; Hutton, 1947; Wascher et al., 1948; Fehrenbacher et al., 1965; Snowden and Priddy, 1968; Ruhe, 1969, 1983; Handy, 1976; Fig. 2). Recent geochemical, isotopic, and grain-size studies suggest distant non-glacial Great Plains sources for some of the loess in the Central Lowland (Muhs and Bettis, 2000; Bettis et al., 2003).

The thickest Peoria Loess in the Central Lowland (> 40 m) lies east of the Missouri River Valley in western Iowa. Loess thickness decreases systematically with distance from inferred sources (Smith, 1942; Hutton, 1947; Simonson and Hutton, 1954; Ruhe, 1954, 1983; Frazee et al., 1970; Handy, 1976; Olson and Ruhe, 1979; Fehrenbacher et al., 1986; Mason et al., 1994; Fig. 2). Although the loess thins to the east and southeast of these valley sources, a narrow band of westward-thinning loess also occurs on the west side of the valleys (Handy, 1976; Mason, 2001; Fig. 2).

Many physical and geochemical properties of the loess exhibit distance-decay trends from source valleys; these trends are considered to be the result of downwind winnowing of coarse particles and heavy minerals as well as syndepositional leaching of carbonates (Ruhe, 1954; Muhs and Bettis, 2000). Sand and coarse silt content decrease while fine silt and clay content increase with distance from source valleys (Smith, 1942; Snowden and Priddy, 1968; Kleiss, 1973; Rutledge et al., 1975; Ruhe, 1983; Miller et al., 1984; Putman et al., 1988). Total aluminium and iron content correlate closely with clay content, and the content of both increases with distance from source areas (Ebens and Connor, 1980; Muhs and Bettis, 2000).

Clay and related iron and aluminium content also vary between sources, with a general increase to the west across the Central Lowland into the Great Plains (Fig. 4a). Mississippi and Missouri Valley source loesses contain matrix carbonate derived from Palaeozoic marine limestone, dolostone, and calcareous shale. Carbonate content and carbonate mineralogy of the loess reflects the source sediments (McKay, 1977). Loess with the highest carbonate content is sourced from the Middle Mississippi and Wabash valleys (Lake Michigan, Green Bay, and Erie lobes of the Laurentide Ice Sheet), while loess from glacial and non-glacial sources to the west has lower matrix carbonate content (Table 1). The mineralogy of the carbonate fraction also varies with source region. The general decrease in carbonate contents westward through the Central Lowland is reflected in the bulk CaO and MgO content of the loess (Fig. 4b). Illinois and Wabash Valley source loess contains a relatively high proportion of dolomite compared to Upper Mississippi and Missouri Valley source loess and sources to the west (McKay, 1977; Ruhe and Olson, 1980). Although all Peoria Loess in the

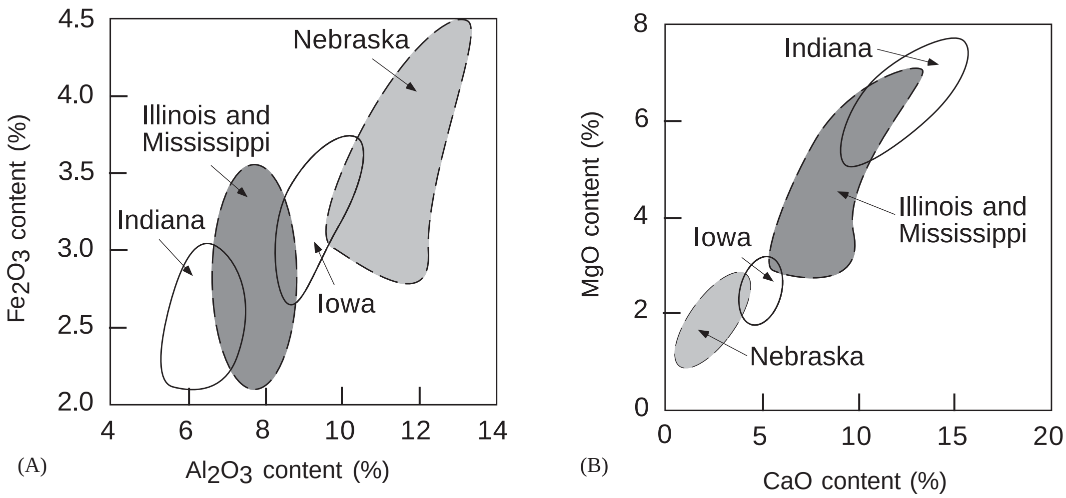

Fig. 4. (a) Iron and aluminium content of Peoria Loess generally increases east to west across the Central Lowlands and Great Plains provinces; (b) the mineralogy of the carbonate fraction of Peoria Loess reflects bedrock lithologies in source areas. Calcite and dolomite content (reflected in the CaO and MgO percentages, respectively) generally decrease from east to west across the Central Lowland and Great Plains. Chemistry by wavelength-dispersive X-ray fluorescence. Data from Pye and Johnson (1988), Muhs and Bettis (2000) and Muhs et al. (2001).

Table 1
Carbonate content of mid-continent United States loess

Source area	Carbonate (%)	Reference
North/South Platte Valley	3–12[a]	Muhs et al. (1999a)
White River Group	10–18[a]	Swineford and Frye (1951)
Missouri Valley	12–14[a]	Ruhe (1969)
Middle Mississippi Valley	26–30[b]	Grimley et al. (1998)
Lower Mississippi Valley	15–30[b]	Miller et al. (1984)
Wabash Valley	25–29[a]	Ruhe and Olson (1978)

[a] Bulk loess, determined with Chittick Apparatus.
[b] 8–63 μm, determined by X-ray diffraction.

region initially contained detrital carbonate (calcite and dolomite), syndepositional and post-depositional weathering has removed these minerals where the loess is thin (Smith, 1942; Muhs et al., 2001). In many thicker sections in the eastern and southern part of the Central Lowland, where precipitation and effective weathering are highest, carbonates are also depleted (Smith, 1942; Pye and Johnson, 1988; Markewich et al., 1998).

The clay mineralogy of the Peoria Loess also reflects the lithology of the source sediments. Ruhe (1984a) defined four mid-continent clay mineral provinces of the Peoria Loess on the basis of differences in kaolinite-illite-expandable mineral compositions and vermiculite-montmorillonite (smectite) compositions (Table 2). These clay mineral provinces reflect the relative contribution of source sediments to the clay fraction of the loess. The largest province, the combined Missouri/Upper Mississippi River Basin, dominates the clay mineralogy of loess in the Lower Mississippi River Valley and is characterized by relatively large amounts of expandable clays. Up-section trends in clay and carbonate mineral composition and magnetic susceptibility, both related to source area shifts, have been documented for the Peoria Loess in Illinois (Frye et al., 1968; McKay, 1977, 1979a, b; Grimley et al., 1998; Grimley, 2000). Up-section changes in particle size and bulk chemistry have also been attributed to source area shifts through time (Muhs and Bettis, 2000; Bettis et al., 2003).

The direction of sediment-transporting palaeo-winds can be inferred using distance-from-source trends in loess thickness, grain size, and geochemistry (Ruhe, 1969; Muhs and Bettis, 2000). Most published loess studies indicate that south of the Laurentide Ice Sheet from Colorado east to Ohio, and southward from Minnesota to Louisiana, sediment-transporting winds over the North American mid-continent were dominantly from the west or northwest (Muhs and Bettis, 2000). A few areas show differences from the overall westerly palaeo-wind data that seem to be characteristic of the area. In addition, valley sources such as the Missouri, Upper Mississippi, Illinois, and Wabash all have narrow bands of loess on their west sides that thin westward. These westward-thinning wedges of loess have been interpreted as resulting from variable winds (Fehrenbacher et al., 1965; Handy, 1976) or anticyclonic circulation (katabatic winds) off the nearby Laurentide Ice Sheet (Hobbs, 1943; Fehrenbacher et al., 1965). Peoria Loess is thicker on the west side of the Mississippi River Valley than on the east side south of

Table 2
Composition of loess clay mineral provinces of the Central Lowland. From Ruhe (1984a). Expandables = vermiculite plus montmorillonite

Province	Kaolinite (%)	Illite (%)	Expandables (%)	Vermiculite (%)	Montmorillonite (%)
Upper Ohio Valley	13–28	19–37	35–63	27–52	3–29
Wabash Valley	10–22	42–61	27–45	7–24	16–34
Lower Mississippi Valley	3–13	20–45	45–76	2–12	41–75
Upper Mississippi-Missouri Valley	3–11	15–24	66–80	0–10	58–79

the Red River confluence in central Louisiana (Fig. 2; Miller et al., 1984). This reversal of the general mid-continent eastward-thinning trend may have been the result of (1) a shift in palaeo-winds over the Gulf of Mexico (Emerson, 1918), (2) presence of the loess source, the Mississippi River braid plain, near the west side of the Mississippi Valley during accumulation of the Peoria Loess (Miller et al., 1984), or (3) the Red River Valley acting as a local source of Peoria Loess in this area. At Sicily Island, located west of the Mississippi Valley near the junction of the Red and Mississippi River valleys in north central Louisiana, Peoria Loess contains more total iron, higher percentages of kaolinite and micaceous clay minerals, and lower percentages of smectite than Peoria Loess at Vicksburg on the east side of the Mississippi Valley north of the Red River Valley junction (Tables 32, 34, 40, and 42 in Miller et al., 1984). These compositional variations suggest that the Red River Valley may be a local source for the thick loess deposits west of the Mississippi River Valley in the southernmost part of the mid-continent loess belt. If the Red River Valley is a loess source, then the loess distribution pattern west of the Mississippi River Valley in central Louisiana may also reflect a dominance of westerly or northwesterly sediment-transporting winds.

A large body of radiocarbon age data exists for the lower one-third of the Peoria Loess, and a smaller number of radiocarbon and TL ages provide limited chronological control for the upper portion of the unit. A radiocarbon date provides the age of organic material buried by, incorporated in, or developed on, the inorganic constituents of the loess sediment, and therefore does not provide a direct age of loess accumulation. Nevertheless, radiocarbon ages do provide limiting ages on periods of loess deposition. Wood and charcoal, the preferred materials for radiocarbon dating, are not often preserved in the region's loess sections and therefore most published ages are derived from analysis of SOM associated with buried soil A horizons. Several problems are inherent in interpretation of radiocarbon ages of SOM. SOM is a pool of organic carbon that accumulated over a period of time. The bulk age of this carbon pool depends on the rates of organic matter addition and decomposition and provides a mean residence time for the soil carbon pool. Bulk ages of SOM increase with depth in a soil, and different fractions of the SOM have different mean residence times. Studies by Abbott and Stafford (1996) and McGeehin et al. (2001) have shown that radiocarbon ages of bulk soil carbon overestimate the "true" age as determined by AMS-radiocarbon ages of associated plant macrofossils, and that the $<63\,\mu m$ humic acid fraction of the SOM most closely matches the age of those macrofossils. Incorporation of older organic carbon from source sediments into the SOM pool can also occur and will increase the bulk SOM age. Contamination of the SOM pool of buried soils with younger organic carbon can also take place by mixing or root penetration. Generally speaking, deep burial, such as occurs with most of the buried soils in the mid-continent loess sequence, lessens the likelihood that this source of error is significant. If the body of radiocarbon ages from SOM in the loess sequence is in error, it is most likely that it overestimates the age of the lower portion of the loess.

Luminescence techniques date the time that loess mineral grains were buried by subsequent loess accumulation and so were shielded from light. These grains are very suitable for luminescence dating because, being transported through the atmosphere, the light-sensitive part of the luminescence signal is depleted to a low residual value at the time of deposition. Sediment mixing, during the original fluvial and subsequent atmospheric transport, homogenizes the sediment's mineralogy and gives rise to a uniform dose rate for most sections. Most comparisons of luminescence ages have been with radiocarbon ages. However, it should be noted that many of the early comparisons were made using uncalibrated radiocarbon ages, which for the LGM are up to 3000 yr younger than the calendar age (Stuiver et al., 1998; Voelker et al., 2000). Most published luminescence ages from the Peoria Loess use TL measurements on the fine grained (4–11 μm) poly-mineral fraction. As discussed by several authors, it has been difficult to decide on the most appropriate bleaching procedure to determine the most likely signal level that would have remained in the sediment grains at deposition. Bleaching experiments are conducted using grains deposited in a monolayer on the carrier disc. These grains have been dispersed into their component grain size, breaking down any agglomerates prior to deposition on the disc. It is likely that some of these very fine grains were transported relatively short distances

from the source as aggregates of coarse silt size (50–30 μm). Thus, the bleaching conditions for these grains are very different from those experienced by grains in the natural environment. Because the larger aggregates may not have been broken up until they were dispersed in the laboratory, the finer grains being dated may not have been thoroughly exposed to light prior to deposition. Thus, the laboratory dating procedures, and particularly the total bleach method, may give rise to overestimation of true age. Evidence for this is evident in the data obtained by all laboratories that have dated Peoria Loess, e.g. USGS laboratory (Maat and Johnson, 1996; Markewich et al., 1998), Cambridge laboratory (Pye et al., 1995) and Chicago laboratory (Forman et al., 1992; Rodbell et al., 1997). Ages from earlier publications are likely to be similarly affected. Rodbell et al. (1997) discussed this problem and also the effect of chemical removal of pedogenic iron oxide coatings on grains, prior to their deposition on the discs for measurement. However, the TL results generally show significant age underestimation when compared with radiocarbon ages at the sites, and this has been attributed to anomalous fading (Rodbell et al., 1997). The latter authors also demonstrated that, even in their cleaned condition, samples bleached very slowly, and about 40% of the TL remained even after 64 h of exposure to sunlight. Such behaviour makes it very difficult to assign a residual TL level. In summary, three problems limit the reliability of available TL ages on Peoria Loess: (1) poor bleaching during transport, (2) intrinsic poor bleaching response and (3) possible anomalous fading of the TL signal. The last two problems can be overcome by the use of optically stimulated luminescence (OSL) signals obtained from pure quartz extracts for determination of luminescence ages (e.g. Roberts et al., 2003). Quartz does not suffer from anomalous fading and the OSL signal, because of the way in which it is measured (i.e. by light exposure) is the most light sensitive luminescence signal. Selection of a grain size more representative of the total loess composition, around 30–50 μm, rather than 4–11 μm as in all the studies discussed, would help to select grains that did not travel as aggregates, and thus avoid some of the problems associated with poor bleaching during transport.

Radiocarbon ages indicate that accumulation of Peoria Loess was diachronous across the Central Lowland region. It began to accumulate in the Missouri River Valley source area between 23,000 and 22,000 ^{14}C yr BP (Bettis, 1990). In the Upper Mississippi and Lower Illinois River Valley (ancient Mississippi) source areas and in the Lower Mississippi River Valley loess began to accumulate at about 25,000 ^{14}C yr BP (McKay, 1979a, b; Hansel and Johnson, 1996; Rodbell et al., 1997; Markewich et al., 1998). These regional differences in the timing of Peoria Loess accumulation may reflect non-synchronous behaviour of the various lobes of the Laurentide Ice Sheet.

Loess accumulation was also time transgressive with respect to distance from individual source areas. In western Iowa the base of the Peoria Loess is as much as 5500 radiocarbon years older near the Missouri River Valley than at localities to the east, while in Illinois the Peoria's base is 4000 radiocarbon years older near the Mississippi River Valley source than it is 20 km away (Kleiss and Fehrenbacher, 1973; Ruhe, 1983). This distance-from-source age variation reflects progressive overlap of older by younger loess increments (Olson and Ruhe, 1979; Ruhe, 1983). The basal age of the Peoria also varies on a local scale because of differences in the stability and age of landscape elements on which the loess accumulated (Hallberg et al., 1978; Martin, 1993).

Accumulation of the Peoria Loess was not continuous. Radiocarbon ages indicate as many as five cycles of loess accumulation and soil formation in the period from 20,710 to 17,630 ^{14}C yr BP in western Iowa (Daniels et al., 1960; Ruhe et al., 1971). At the Keller Farm section near East St. Louis, Illinois (Fig. 2, KF) at least 20 accumulation cycles are evident within the Peoria Loess (Wang et al., 1998, 2000). Magnetic susceptibility variations within Peoria Loess in Indiana have also been interpreted as indicating cycles of loess accumulation punctuated by pedogenesis during the Last Glacial period (Hayward and Lowell, 1993).

The end of Peoria Loess accumulation in the Central Lowland has been difficult to date because of a general absence of organic material in the upper part of the loess, introduction of Holocene organic matter to the surficial unit during formation of the modern soil, and biological mixing of surface soils formed in the loess that frustrates luminescence dating. Nevertheless, estimates for the end of the deposition period have been obtained by radiocarbon dating of terrestrial gastropod shells in the upper part of the loess, by dating organic matter associated with buried and surface soils, and by obtaining luminescence ages just beneath the surface soil horizons. These age estimates suggest that Peoria Loess accumulation ended throughout the region some time between about 12,000 and 11,000 ^{14}C yr BP (Daniels et al., 1960; Norton and Bradford, 1985; Forman et al., 1992; Curry, 1998; Grimley, 2000; Wang et al., 2000; Bettis et al., 2003). Regional stratigraphic and geomorphic relationships also provide constraints on the timing of loess deposition. Stratigraphic relationships between Peoria Loess, tills, and alluvium across Iowa and Illinois suggest Peoria Loess deposition ceased after 13,900 ^{14}C yr BP. In central Iowa, the Des Moines Lobe of the Laurentide Ice Sheet buried a spruce forest growing on Peoria Loess near the southern margin of the lobe at 13,900 ^{14}C yr BP (Bettis et al., 1996). Loess is absent from the Des Moines Lobe till plain, which was ice-free by about 12,600 ^{14}C yr BP. Alluvium with

associated radiocarbon ages of 12,000 ^{14}C yr BP in Iowa, eastern Nebraska, Missouri, and Illinois is loess-free, while alluvium older than about 14,000 ^{14}C yr BP has a loess mantle (Bettis, 1990; Hajic et al., 1991; Mandel and Bettis, 1995; Mandel and Bettis, 2001). In Illinois, the Lake Michigan Lobe of the Laurentide Ice Sheet advanced to its terminal position about 20,000 ^{14}C yr BP and deposited till of the Wedron Group (Hansel and Johnson, 1996). The advance occurred after Peoria Loess deposition was underway, and Wedron Group till buried Peoria Loess (the Morton Tongue of the Peoria Silt) that had accumulated on the landscape prior to the advance. The presence of Peoria Loess on all but the youngest Lake Michigan Lobe tills indicates that loess deposition continued until at least 11,800 ^{14}C yr BP. Numerical ages are not available from the upper metre of the Peoria Loess, and Holocene loess, which provides an upper age limit in the Great Plains province, has not been unequivocally identified east of the Missouri River valley. The youngest reliable numerical age for Peoria Loess in the Central Lowlands is 11,350 ^{14}C yr BP (13,250 cal yr BP) from the Keller Farm site near East St. Louis, Illinois (Wang et al., 2000). This age was obtained from a weak soil 1.5 m below the top of the Peoria Loess. Although Holocene loess has not been identified east of the Missouri River Valley, evidence exists for significant contributions of postglacial dust to the profiles of modern soils in the Central Lowland (Ruhe, 1984b; Mason and Jacobs, 1998; Muhs et al., 2001).

2.2. MARs of Peoria Loess in the Central Lowland

MAR calculations for the Last Glacial period in the Central Lowland are based on the following assumptions: (1) distance-from-source basal-age models for each Central Lowland source area, based on radiocarbon ages from the upper horizon of the Farmdale Soil, provide a maximum-limiting age for the initiation of Peoria Loess deposition; (2) calendar-year-converted radiocarbon ages of buried SOM from the Keller Farm section in Illinois provide a minimum-limiting age for the termination of Peoria Loess deposition; (3) Peoria Loess has an average bulk density of about 1.33–1.53×10^6 g/m^3; and (4) no significant erosion or deposition of non-loess sediment has occurred at the site. Errors arise in this method because maximum-limiting radiocarbon ages of underlying SOM, upon which the basal age model is based may be older than the true age. The minimum-limiting age we use (12,999 cal yr BP–11,000 ^{14}C yr BP) from a buried soil in the upper portion of the loess is a conservative under-estimate for the end of loess accumulation. In this regard, our calculated MARs are conservative. In those sections where datable material has been found at two or more depths within Peoria Loess, it is possible to generate more realistic MARs for the dated intervals. Other errors arise with the use of the assumed bulk density of the sediments. Published and unpublished bulk density values for the Peoria Loess in this region range from 1.33 to 1.53×10^6 g/m^3 (Davidson and Handy, 1952; Daniels and Jordan, 1960; Lutenegger and Hallberg, 1988; Markewich, 1993; Handy and Ferguson, 1994; Lutenegger, pers. comm., 2001) with no apparent relationships between bulk density values, depth below the base of the modern solum, loess thickness, or source area. Based on these data, we consider an oven-dry bulk density value for unaltered loess of 1.45×10^6 g/m^3 to be reasonable; the use of an average bulk density value affects the MAR calculations by no more than $\pm 6\%$.

We used maps depicting loess thickness for the Peoria Loess to choose localities where MARs for Last Glacial loess could be determined and extrapolated to loess of similar thickness from the same source (Fig. 2). We chose to use only radiocarbon ages for determining MAR in the Central Lowland because a considerable number of such published ages exist and can be consistently compared across the region.

2.2.1. Missouri River Valley source

Localities used to calculate MARs for Peoria Loess from this source include Loveland (Fig. 2—LL; loess thickness = 41 m), Pisgah (PIS)/Woodbine/Dunlap/Logan (LOG; loess thickness = 21–19 m), Crawford (CRF)/Hancock (HAN; loess thickness = 10–9.1 m), Greenfield (GR; loess thickness = 4.6 m), Earlham (ERL; loess thickness = 3.7 m) and Humeston (HOM)/Harvard (HAR; loess thickness = 2.8–2.5 m). The Bellevue site (BE; loess thickness = 8.5 m) is used to calculate MARs for the narrow band of thick loess west of the Missouri River Valley. These localities contain loess from glacial and non-glacial sources in the northern Great Plains and the northwestern portion of the Central Lowland (Fig. 2; Muhs and Bettis, 2000; Bettis et al., 2003).

The basal age of the Peoria Loess decreases to the east away from the Missouri River Valley (Ruhe, 1969, 1976). Because there is both a systematic decrease in Peoria Loess thickness and basal age with distance from the Missouri River Valley, we assigned basal ages on the basis of loess thickness. The following Peoria Loess thickness/basal age relationships are used (Table 3): >20 m 24,500 ^{14}C yr BP (27,500 cal yr BP), 20–15 m 23,900 ^{14}C yr BP (26,900 cal yr BP), 15–5 m 22,800 ^{14}C yr BP (25,640 cal yr BP), <5 m 19,300 ^{14}C yr BP (22,888 cal yr BP). The narrow band of thick loess east of the Missouri River Valley in Nebraska and Kansas is adjacent to the loess source and so we use the basal age assigned to near-source thick loess (27,500 cal yr BP) to calculate MARs for this area.

The regionally averaged MAR for Missouri River Valley source Peoria Loess is 1273 g/m^2/yr (Table 4),

Table 3
Radiocarbon ages associated with Peoria Loess in the Central Lowland Province

Locality name (see Fig. 2 for location)	Peoria Loess thickness (m)	Height above base of Peoria (m)	Location	Lab number	^{14}C age ± 1-σ	Calibrated age range			Calibration method	Age used in this paper	Material dated	Reference
						Min.	(cal. age)	Max.				
Missouri Valley source												
Loveland, IA (LL)	41	0.1	41.5N,95.88W	FRL-1443	24,100±160		27,100		[b]	27,100	Disseminated carbon	Forman et al. (1992)
Loveland, IA	41	9		AA-4828	20,540±200		23,140		[b]	23,140	Gastropod shell	Forman et al. (1992)
Pisgah, IA (PIS)	21.1	−0.3	41.82N,95.63W	I-3870	24,750±700		27,750		[b]	27,750	Disseminated carbon	Ruhe et al. (1971)
Woodbine, IA (LOG)	21	3.35	41.82N,95.73W	I-4211	22,350±700		25,210		[b]	25,210	Disseminated carbon	Ruhe et al. (1971)
Logan 2, IA (LOG)	19.8	−0.3	41.83N,95.63W	I-4619	24,400±700		27,400		[b]	27,400	Disseminated carbon	Ruhe et al. (1971)
Dunlap, IA (LOG)	19	15.7	41.83N,95.65W	I-4029	15,300±300	17,854	18,285	18,736	[c]	18,285	Disseminated carbon	Ruhe et al. (1971)
Panama, IA	13.1	1.06	41.63N,95.57W	I-3703	19,250±400	22,257	22,831	23,430	[c]	22,831	Disseminated carbon	Ruhe et al. (1971)
Bentley, IA	13.1	0.5	41.37N,95.58W	I-1023	21,360±850		23,990		[b]	23,990	Spruce wood	Ruhe (1969)
Bentley, IA	13.1	−0.5		I-1420	23,900±1100		26,900		[b]	26,900	Disseminated carbon	Ruhe (1969)
Crawford, IA (CRF)	10	2.9	41.90N, 95.67W	I-5106	17,170±280	19,988	20,437	20,893	[c]	20,437	Disseminated carbon	Ruhe (1976)
Hancock, IA (HAN)	9.1	−0.3	41.35N, 95.42W	W-141	24,500±800		27,500		[b]	27,500	Larch wood	Ruhe (1969)
Hancock, IA	9.1	−0.15		I-3945	23,200±600		26,200		[b]	26,200	Disseminated carbon	Ruhe et al. (1971)
Hancock, IA	9.1	0.15		I-3944	22,750±600		25,590		[b]	25,590	Disseminated carbon	Ruhe et al. (1971)
Hancock, IA	9.1	0.3		I-3943	22,200±500		25,060		[b]	25,060	Disseminated carbon	Ruhe et al. (1971)
Greenfield, IA (GR)	4.57	0.15	41.37N,94.45W	I-1411	18,700±700	21,323	22,198	23,085	[c]	22,198	Disseminated carbon	Ruhe (1969)
Earlham, IA (ERL)	3.94	−0.1	41.48N, 94.10W	TX-6619	21,380±740		24,050		[b]	24,050	Disseminated carbon	Woida and Thompson (1993)
Murray, IA (MU)	3.35	0.2	41.03N,93.95W	I-1410	20,900±1000		23,900[a]		[b]	23,900[a]	Disseminated carbon	Ruhe (1969)
Humeston, IA (HOM)	2.8	0.15	40.75N,93.52W	I-1419	16,500±500	19,020	19,666	20,324	[c]	19,666	Disseminated carbon	Ruhe (1969)
Harvard, IA (HAR)	2.53	0.1	40.70N,93.27W	I-1408	19,200±900	21,687	22,773	23,884	[c]	22,773	Disseminated carbon	Ruhe (1969)
West of valley												
Bellevue, NE (BE)	8.5	0.32	41.13N,95.97W	CAMS-10186	22,040±160		24,900		[b]	24,900	Humic acids	Mandel and Bettis (1995)
Bellevue, NE	8.5	0.36		CAMS-10187	20,810±160		23,480		[b]	23,480	Humic acids	Mandel and Bettis (1995)
Bellevue, NE	8.5	0.38		CAMS-10188	22,210±170		25,070[d]		[b]	25070[d]	Humic acids	Mandel and Bettis (1995)
Iowan Erosion Surface												
Alburnett Paha, IA (ALB)	12.9	12.9	42.17N,91.63W	I-2332	20,700+500		23,370		[b]	23,370	Organic carbon	Ruhe (1969)
Salt Creek, IA (SC)	4.57	4.5	42.10N,92.52W	W-1687	18,300+500	20,425	21,738	23,105	[c]	21,738	Organic carbon	Ruhe (1969)
Palermo, IA (PAL)	3.66	3.67	42.32N,92.87W	W-1681	21,600±600		24,230		[b]	25,440	Organic carbon	Ruhe (1969)
Haywards Paha, IA (HAY)	3.66	3.44	42.22N,92.30W	I-1409	20,300±400		22,970		[b]	22,970	Organic carbon	Ruhe (1969)
Hokah, MN (HOK)	3	−0.1	43.75N,91.3W	B-14886	23,760±630		26,600		[b]	26,600	Bulk soil carbon	Lively et al. (1987)
Upper Mississippi Valley												
Keller Farm, IL (KF)	16	14.45	38.75N,90.0W	ISGS-4137B	11,350±100	13,023	13,250	13,792	[c]	13,250	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	12.95	38.75N,90.0W	ISGS-4131	13,670±250	15,690	16,410	17,162	[c]	16,410	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	11.75		ISGS-4127	16,040±130	18,503	19,137	19,827	[c]	19,137	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	10.65		ISGS-4095	17,380±180	19,931	20,679	21,458	[c]	20,679	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	10.5		ISGS-3833	17,630±240	20,130	20,967	21,836	[c]	20,967	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	7.95		ISGS-3906	18,770±220	21,449	22,279	23,179	[c]	22,279	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	6.75		ISGS-3939	19,640±220	22,419	23,280	24,201	[c]	23,280	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	6.15		ISGS-3842	20,420±330		23,090		[b]	23,090	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	5.25		ISGS-3980	20,710±220		23,809		[b]	23,809	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	3.25		ISGS-4143	22,440±250		25,280		[b]	25,280	Pyrolysis-volatile SOM	Wang et al. (2000)
Keller Farm, IL	16	2.45		ISGS-4147	22,810±210		25,650		[b]	25,650	Pyrolysis-volatile SOM	Wang et al. (2000)
Canteen Creek, IL (CC)	15.5	10.9	36.64N,90.02W	ISGS-421	16,020±260	18,304	19,114	19,981	[c]	19,114	Organic carbon	McKay (1977, 1979a)
Core G39, IL (G-39)	9.5	3	38.72N,90W	ISGS-412	20,910±520		23,580		[b]	23,580	Wood	McKay (1977, 1979b)
Core G39, IL	9.5	1.1		ISGS-413	23,110±810		25,951		[b]	25,951	Wood	McKay (1977, 1979b)
Rapids City, IL (RC)	9.2	0.6	41.57N,90.37W	WW-2170	21,110±70		23,780		[b]	23,780	Conifer needles	Muhs et al. (2001)

Table 3 (*continued.*)

Locality name (see Fig. 2 for location)	Peoria Loess thickness (m)	Height above base of Peoria (m)	Location	Lab number	^{14}C age $\pm$1-σ	Calibrated age range			Calibration method	Age used in this paper	Material dated	Reference
						Min.	(cal. age)	Max.				
Rapids City, IL	9.2	−0.2		WW-2153	23,410±90		26,250		b	26,250	Humic acids	Muhs et al. (2001)
Ruby Lane, IL (RL)	4.6	1.1	38.62N,90W	ISGS-294	21,910±270		25,320		b	25,320	Peat	McKay (1977)
Ruby Lane, IL	4.6	0.4		ISGS-307	23,390±280		26,230		b	26,230	Organic silt	McKay (1977)
Core GT-6, WI (Gt-6)	4.57	0.27	42.68N,90.83W	GX-15888	24,250±970		27,550		b	27,550	Snail shell	Leigh and Knox (1993)
Athens Quarry N., IL (AN)	3.28	1.03	40N,89.7W	ISGS-534	22,170±450		25,030		b	25,030	Wood	Follmer et al. (1986)
Gardena, IL (G)	2.0[e]	2.00	40.67N,89.47W	ISGS-532	19,680±460	22,689	23,326	23,991	c	23,326	Moss	Follmer et al. (1979)
Gardena, IL	2.0[e]	0		ISGS-531	25,370±310		28,670		b	28,670	Wood	Follmer et al. (1979)
Wempletown SE, IL (WEM)	1.87	0.1	42.32N,89.14W	ISGS-1302	20,150±500		22,820		b	22,820	Humic material	Berg et al. (1985)
Wabash/Ohio River Valley source												
Mt. Vernon, IN (MV)	13.3	−0.5	37.91N,87.96W	NSRL-2744	24,160±130		27,460		b	27,460	Humic acids	This paper
Mt. Vernon, IN	13.3	−1.1		NSRL-2749	31,590±330		34,510		b	34,510	Humic acids	This paper
Middle and Lower Mississippi Valley source												
Phillips Bayou, AR (PB)	22	−0.7	34.63N,90.63W	W-6437	28,980±800		32,810		b	32,810	Charcoal	Markewich (1993)
Phillips Bayou, AR	22	5.5		WW-102	21,070±230		23,410		b	23,410	Gastropod shells	Markewich (1993)
West Helena, AR (Phillips Bayou)	22	8.2	34.63N,90.63W	GX-14657	23,485±490		26,325[d]		b		Gastropod shells	Oches et al. (1996)[f]
West Helena, AR (Phillips Bayou)	22	15.8		AA-4391	17,850±145	20,503	21,220	21,973	c		Gastropod shells	Oches et al. (1996)[f]
Natchez 3, MS (N)	18.2	0.25	31.58N,91.9W	B-53963	21,770±180		24,400		b	24,400	Gastropod shells	McCraw and Autin (1989)
Natchez 3, MS	18.2	2.8		B-53962	21,070±160		23,740		b	23,740	Gastropod shells	McCraw and Autin (1989)
Natchez 3, MS	18.2	5.2		B-53961	19,160±160	21,961	22,727	23,585	c	22,727	Gastropod shells	McCraw and Autin (1989)
Natchez 3, MS	18.2	7		GX-14563	21,360±425		24,030[d]		b		Gastropod shells	Oches et al. (1996)
Natchez 3, MS	18.2	8		GX-14568	20,740±385		23,080[d]		b		Gastropod shells	Oches et al. (1996)
Natchez 3, MS	18.2	11.3		B-53957	17,700±140	20,338	21,047	21,790	c	21,047	Gastropod shells	McCraw and Autin (1989)
Natchez 3, MS	18.2	14.2		B-53956	17,600±190	20,166	20,932	21,730	c	20,932	Gastropod shells	McCraw and Autin (1989)
Vicksburg 2, MS (V)	14.7	0	32.38N,90.87W	UM-2575	20,800±210		23,470		b		Gastropod shells	Pye and Johnson (1988)
Vicksburg 2, MS	14.7	5.2		UM-2573	17,130±200	19,625	20,391	21,189	c		Gastropod shells	Pye and Johnson (1988)
Vicksburg 2, MS	14.7	9		UM-2572	16,620±130	19,138	19,804	20,508	c		Gastropod shells	Pye and Johnson (1988)
Old River, TN (OR)	14	−0.15	35.42N,89.97W	WW-48	26,490±270		29,120		b	29,120	Charcoal	Markewich (1993)
Old River, TN	14	−0.1		CAMS-3278	26,500±300		29,340		b	29,340	Gastropod shells	Oches et al. (1996)
Old River, TN	14	0.3–2.0		WW-102	21,070±230		23,410		b		Gastropod shells	Markewich (1993)
Vicksburg 1, MS (V)	12.8	−0.3	32.38N,90.87W	UM-2576	23,390±340		26,230		b	26,230	Gastropod shells	Pye and Johnson (1988)
Vicksburg 1, MS	12.8	−0.1		I-1681	22,600±800		25,440		b	25,440	Gastropod shells	Snowden and Priddy (1968)

Vicksburg 1, MS	12.8	0.1		UM-3004	19,910±320	22,571	23,590	24,593	c		Gastropod shells	Pye and Johnson (1988)
Vicksburg 1, MS	12.8	1.8		I-1904	19,250±350	21,785	22,831	23,969	c		Gastropod shells	Snowden and Priddy (1968)
Vicksburg 1, MS	12.8	4.8		OX-184	19,200±420	21,602	22,773	24,028	c		Gastropod shells	Snowden and Priddy (1968)
Vicksburg 1, MS	12.8	8.7		UM-3003	13,660±200	15,767	16,398	17,062[d]	c		Gastropod shells	Pye and Johnson (1988)
Vicksburg 1, MS	12.8	9.3		OX-185	17,850±380	20,143	21,220	22,337	c		Gastropod shells	Snowden and Priddy (1968)
Natchez 1, MS (N)	10.1	−0.2	31.58N,91.9W	UM-2578	18,620±350	21,074	22,106	23,204[d]	c		Gastropod shells	Pye and Johnson (1988)
Natchez 1, MS	10.1	0.1		UM-2579	19,300±360	21,824	22,888	24,039	c		Gastropod shells	Pye and Johnson (1988)
Natchez 3, MS (N)	6.5	−0.5	31.58N,91.9W	UM-2580	21,400±390		23,740		b		Gastropod shells	Pye and Johnson (1988)
Hornbeak, TN (HBK)	4.4	4.4	36.37N,89.3W	GX-17725	23,215±485		26,055		b	26,055	Charcoal	Rodbell et al. (1997)
Core OP-16, Union City, TN (OP-16)	2.98	0.48	36.42N,89.0W	GX-17029	19,900±230	22,692	23,579	24,472	c		Wood	Rodbell (1996)

[a] Age not used—older than age model for Peoria base.
[b] Nearest calibration point derived from core PS2644 (Voelker et al., 2000).
[c] CALIB version 4.3 (Stuiver and Reimer, 2000).
[d] Rejected—age inversion.
[e] Loess buried by Wedron Fm. glacial till, complete Peoria Loess section not present.
[f] This section, referred to as Natchez 1 in Oches et al. (1996) is not the Natchez 1 section of Pye and Johnson (1988). The Natchez 1 section of Oches et al (1996) is the Natchez section described in McCraw and Autin (1989), and is referred to as Natchez 3 in the present paper.

nearly 3 times that calculated for MIS 2 loess of the Chinese Loess Plateau (Sun et al., 2000). The section-averaged MAR of the Peoria Loess ranges from a high of 4216 g/m^2/yr adjacent to the Missouri River Valley source (Loveland; LL), to a low of 371 g/m^2/yr at a distance of about 230 km to the east-southeast (Fig. 2, Harvard site; HAR). The Peoria Loess at the site with the lowest MAR (Harvard) is the thinnest typically found on stable uplands in the Missouri River Valley source region. The MAR of the Peoria loess on the west side of the source (Bellevue; BE) is much less than that on the east side (Loveland).

Muhs and Bettis (2000) identified three zones in the Peoria Loess at Loveland that differed in grain-size and geochemistry. They concluded that the different zones represented (1) initial relatively slow accumulation of Missouri River Valley-source loess followed by (2) a middle zone representing rapid accumulation of relatively coarse glaciogenic-source Missouri River Valley loess overlain by (3) an upper zone of finer grained glacogenic loess derived from the Missouri River Valley as well as western, non-glaciogenic sources.

The presence of weakly expressed soils ("dark bands"—Daniels et al., 1960; Ruhe et al., 1971) and apparent syndepositional weathering of carbonate minerals in the lower part of the Peoria Loess at other Missouri River Valley source sections (Ruhe, 1969, 1983; Kleiss, 1973; Muhs and Bettis, 2000) suggest that MARs varied significantly during the period of accumulation.

2.2.2. *Iowan erosion surface (IES)*

The IES occupies portions of southeastern Minnesota, and northeastern Iowa east of the Des Moines Lobe of the Laurentide Ice Sheet (Fig. 2). The precise mechanisms by which the IES developed are poorly understood (Hallberg et al., 1978) but it is known that significant mass wasting, and aeolian and fluvial erosion took place under periglacial conditions during the LGM (Bettis and Kemmis, 1992; Walters, 1994; Mason and Knox, 1997). Much of the erosion surface is loess-free, but 1–1.5 m of Peoria Loess mantles the eastern and southern margins. A limited number of radiocarbon dates, and stratigraphic relationships between the erosion surface and adjacent loess-mantled areas, indicate that the IES developed between about 22,000 and 17,000 ^{14}C yr BP where it is loess mantled, and continued to develop until after deposition of the Peoria Loess (ca 11,000 yr BP) where it is loess-free (Hallberg et al., 1978). A band of thick loess forms the eastern and southern margin of the IES, and it has been suggested that the erosion surface was a loess source or a transport zone during accumulation of the Peoria Loess (Ruhe et al., 1968; Ruhe, 1969; Hallberg et al., 1978; Mason et al., 1994, 1999). Thick loess is also found on "paha"—erosional inliers inside the margins of the

Table 4
Section information, ages and mass accumulation rates of Last Glacial and Holocene Loess in the Great Plains and Central Lowland Provinces

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate (g/m^2/yr)	Notes
LAST GLACIAL PERIOD												
Eastern Colorado (Platte Valley and White River Group sources)												
Last Chance, CO	3	39.7N	103.6W	Upland	0.3 (3)	13,555–14,024	23,000?	9000	0.0003	1.45×10^6	435	Min. age from Beecher Is.
Beecher Island, CO	10.5	39.9N	102.1W	Upland	1.5–12.0 (10.5)	13,555–14,024	23,000?	9000	0.0012	1.45×10^6	1692	
Central Nebraska and Kansas (Platte Valley and White River Group sources)												
Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate (g/m^2/yr)	Notes
Bignell Hill, NE	48	41.0N	101.5W	Upland	2–50 (48)	12,478	34,240	21,762	0.0022	1.45×10^6	3198	Based on bracketing Brady/GC ^{14}C ages
Bignell Hill, NE				Upland	3–49 (46)	17,900	21,700	3800	0.0121	1.45×10^6	17,552	Based on TL ages within Peoria Loess
Bignell Hill, NE				Upland	0–1 (1)	12,478	13,800	1322	0.0008	1.45×10^6	1097	Based on ^{14}C and OSL ages within Peoria Loess
Bignell Hill, NE				Upland	1–14.9 (13.9)	13,800	16,600	2800	0.0050	1.45×10^6	7198	Based on OSL ages within Peoria Loess
Bignell Hill, NE				Upland	14.9–40.65 (25.75)	16,600	18,900	2300	0.0112	1.45×10^6	16,234	Based on OSL ages within Peoria Loess
Bignell Hill, NE				Upland	40.65–48 (7.35)	18,900	25,100	6200	0.0013	1.45×10^6	1719	Based on OSL ages within Peoria Loess
Bignell Hill, NE				Upland	0–48 (48)	12,478	25,100	12,622	0.0038	1.45×10^6	5514	Based on ^{14}C and OSL ages within Peoria Loess
Eustis, NE	16.3	40.63N	100.07W	Upland	0–16.3 (16.3)	12,478	25,000	12,522	0.0013	1.45×10^6	1887	Based on GC ^{14}C age and assumed Brady age for minimum
Eustis, NE				Upland	6.0–16.3 (10.3)	14,305	25,000	10,695	0.0010	1.45×10^6	1396	Based on GC ^{14}C age and ^{14}C age within Peoria Loess
Eustis, NE				Upland	1.5–14.5 (13)	18,900	24,400	5500	0.0024	1.45×10^6	3427	Based on TL ages within Peoria Loess
Eustis, NE				Upland	0–2.5 (2.5)	11,600	14,200	2600	0.0010	1.45×10^6	1394	Based on ^{14}C and OSL ages within Peoria Loess
Eustis, NE				Upland	2.5–5.5 (3)	14,200	14,900	700	0.0043	1.45×10^6	6214	Based on OSL ages within Peoria Loess
Eustis, NE				Upland	5.5–9 (3.5)	14,900	15,800	900	0.0039	1.45×10^6	5639	Based on OSL ages within Peoria Loess
Eustis, NE				Upland	9–13.15 (4.15)	15,800	18,600	2800	0.0015	1.45×10^6	2149	Based on OSL ages within Peoria Loess
Eustis, NE				Upland	13.15–16 (2.85)	18,600	20,700	2100	0.0014	1.45×10^6	1968	Based on OSL ages within Peoria Loess
Eustis, NE				Upland	0–16 (16)	11,600	20,700	9100	0.0018	1.45×10^6	2549	Based on ^{14}C and OSL ages within Peoria Loess
La Sena, NE	6.2	40.4N	100.24W	Upland	0–6.2 (6.2)	12,478	24,080	11,602	0.0005	1.45×10^6	775	Based on GC ^{14}C age and assumed Brady age for minimum
La Sena, NE				Upland	3.4–6.2 (2.8)	22,395	24,080	1685	0.0017	1.45×10^6	2409	Based on GC ^{14}C age and ^{14}C age within Peoria Loess
Naponee, NE	9.3	40.09N	99.09W	Upland	4.67–9.3 (4.63)	11,696	23,430	11,734	0.0004	1.45×10^6	572	Based on bracketing Brady/GC ^{14}C ages
Naponee, NE				Upland	4.67–8.85 (4.18)	11,696	21,255	9559	0.0004	1.45×10^6	634	Based on Brady and lower Peoria ^{14}C age
Beisel-Steinle, KS	2.25	38.8N	98.5W	Upland	1.25–3.50 (2.25)	13,250	25,620	12,370	0.0002	1.45×10^6	264	Based on bracketing Brady/GC ^{14}C ages
Barton County, KS	1.6	38.4N	98.7W	Upland	0.8–2.4 (1.6)	12,750	29,000	16,250	0.0001	1.45×10^6	143	Based on bracketing Brady/GC ^{14}C ages

HOLOCENE:

Locality	Bignell Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Bignell (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate ($g/m^2/yr$)	Notes
Bignell Hill, NE	4	41.04N	101.5W	Upland	0–4 (4)	0	10,462	10,462	0.0004	1.45×10^6	554	Based on age of uppermost part of Brady Soil; thickness is a maximum
Naponee, NE	4.67	40.09N	99.09W	Upland	0–4.67 (4.67)	0	11,696	11,696	0.0004	1.45×10^6	579	Based on bulk age of Brady Soil
Peters/Speed, KS	1.3	39.7N	99.4W	Upland	0–1.3 (1.3)	0	10,091	10,091	0.0001	1.45×10^6	187	Based on age of uppermost part of Brady Soil
Beisel-Steinle, KS	1.25	38.8N	98.5W	Upland	0–1.25 (1.25)	0	13,250	13,250	0.0001	1.45×10^6	137	Based on bulk age of Brady Soil
Barton County, KS	0.8	38.4N	98.7W	Upland	0–0.8 (0.8)	0	11,202	11,202	0.0001	1.45×10^6	104	Based on age of uppermost part of Brady Soil

Western Iowa/Eastern Nebraska (mixed glacial and non-glacial source)

LAST GLACIAL PERIOD

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate ($g/m^2/yr$)	Notes
Loveland, IA	41	41.5N	95.88W	Upland	0–0.1 (0.1)	27,100	27,500	400	0.0003	1.45×10^6	362	Basal age from age model
					0.1–9.0 (8.9)	23,140	27,100	3960	0.0022	1.45×10^6	3259	
					9.0–41 (32)	12,999	23,140	10,141	0.0032	1.45×10^6	4575	
					0–41 (41)	12,999	27,100	14,101	0.0029	1.45×10^6	4216	Basal age from age model
Woodbine, IA	21	41.82N	95.73W	Terrace	0–3.35 (3.35)	25,210	27,500	2290	0.0015	1.45×10^6	2121	Basal age from age model
					3.35–21 (17.5)	12,999	25,210	12,211	0.0014	1.45×10^6	2078	
					0–21 (21)	12,999	27,500	14,501	0.0014	1.45×10^6	2100	Basal age from age model
Dunlap, IA	19	41.83N	95.65W	Terrace	0–15.7 (15.7)	18,285	26,900	8615	0.0018	1.45×10^6	2642	Basal age from age model
					15.7–19 (3.5)	12,999	18,285	5286	0.0006	1.45×10^6	960	
					0–19 (19)	12,999	26,900	13,901	0.0014	1.45×10^6	1982	Basal age from age model
Panama/Bentley, IA	13.1	41.37N	95.58W	Upland	0–0.5 (0.5)	23,990	25,640	1650	0.0003	1.45×10^6	439	Basal age from age model
					0.5–1.06 (0.56)	22,831	23,990	1159	0.0005	1.45×10^6	701	
					1.06–13.1 (12.04)	12,999	22,831	9832	0.0012	1.45×10^6	1776	
					0–13.1 (13.1)	12,999	25,640	12,641	0.0010	1.45×10^6	1502	Basal age from age model
Crawford, IA	10	41.90N	95.67W	Upland	0–2.9 (2.9)	25,640	20,473	5167	0.0006	1.45×10^6	814	Basal age from age model
					2.9–10 (7.1)	12,999	20,473	7474	0.0010	1.45×10^6	1377	
					0–10 (10)	12,999	25,640	12,641	0.0008	1.45×10^6	1147	Basal age from age model
Hancock, IA	9.1	41.35N	95.42W	Upland	0–0.15 (0.15)	25,590	25,600	10	0.015	1.45×10^6	21,750	Basal age from age model
					0.15–0.3 (0.15)	25,060	25,590	530	0.0003	1.45×10^6	410	
					0.3–9.1 (8.8)	12,999	25,060	12,061	0.0007	1.45×10^6	1058	
					0–9.1 (9.1)	12,999	25,600	12,601	0.0007	1.45×10^6	1047	Basal age from age model
Greenfield, IA	4.57	41.37N	94.45W	Upland	0–0.15 (0.15)	22,198	22,888	690	0.0002	1.45×10^6	315	Basal age from age model
					0.15–4.57 (4.42)	12,999	22,198	9199	0.0005	1.45×10^6	697	
					0–4.57 (4.57)	12,999	22,888	9889	0.0005	1.45×10^6	670	Basal age from age model
Earlham, IA	3.94	41.48N	94.10W	Upland	0–3.94 (3.94)	12,999	22,888	9889	0.0004	1.45×10^6	578	Basal age from age model
Murray, IA	3.35	41.03N	93.95W	Upland	0–3.35 (3.35)	12,999	22,888	9889	0.0003	1.45×10^6	491	Basal age from age model
Hummeston, IA	2.8	40.75N	93.52W	Upland	0–0.15 (0.15)	19,666	22,888	3222	0.00005	1.45×10^6	68	Basal age from age model
					0.15–2.8 (2.65)	12,999	19,666	6667	0.0004	1.45×10^6	576	
					0–2.8 (2.8)	12,999	22,888	9889	0.0003	1.45×10^6	410	Basal age from age model
Harvard, IA	2.53	40.70N	93.27W	Upland	0–0.1 (0.1)	22,773	22,888	115	0.0009	1.45×10^6	1261	Basal age from age model
					0.1–2.53 (2.43)	12,999	22,773	9774	0.0003	1.45×10^6	360	
					0–2.53 (2.53)	12,999	22,888	9889	0.0003	1.45×10^6	371	Basal age from age model

Table 4 (*continued*)

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate ($g/m^2/yr$)	Notes
West of valley												
Bellevue, NE	8.5	41.13N	95.97W	Upland	0–0.32 (0.32)	24,900	27,500	2600	0.0001	1.45×10^6	178	Basal age = Peoria maximum age
					0.32–0.36 (0.04)	23,480	24,900	1420	0.00003	1.45×10^6	41	
					0.36–8.5 (8.14)	12,999	23,480	10,481	0.0008	1.45×10^6	1126	
					0–8.5 (8.5)	12,999	27,500	14,501	0.0006	1.45×10^6	850	Basal age = Peoria maximum age

Iowan Erosion Surface (mixed glacial and non-glacial source)

LAST GLACIAL PERIOD

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate ($g/m^2/yr$)	Notes
Alburnett Paha, IA	12.9	42.17N	91.63W	Upland	0–12.9 (12.9)	12,999	24,860	11,861	0.0011	1.45×10^6	1577	
Salt Creek, IA	4.57	42.1N	92.52W	Upland	0–0.07 (0.07)	18,300	24,860	6560	0.00001	1.45×10^6	16	
					0.07–4.57 (4.5)	12,999	18,300	5301	0.0008	1.45×10^6	1231	
					0–4.57 (4.57)	12,999	24,860	11,861	0.0004	1.45×10^6	559	
Palermo, IA	3.66	42.32N	92.87W	Upland	0–3.66 (3.66)	12,999	24,860	11,861	0.0003	1.45×10^6	447	
Haywards Paha, IA	3.66	42.22N	92.30W	Upland	0–0.22 (0.22)	24,270	24,860	590	0.0004	1.45×10^6	541	
					0.22–3.66 (3.44)	12,999	24,270	11,271	0.0003	1.45×10^6	443	
					0–3.66 (3.66)	12,999	24,860	11,861	0.0003	1.45×10^6	447	

Upper Mississippi River Valley (glacial source)

LAST GLACIAL PERIOD

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate ($g/m^2/yr$)	Notes
Keller Farm, IL	16	38.75N	90.0W	Upland	0–1.55 (1.55)	25,650	28,300	2650	0.0006	1.45×10^6	848	
					1.55–3.25 (1.7)	25,280	25,650	370	0.0046	1.45×10^6	6662	
					3.25–5.25 (2)	23,809	25,280	1471	0.0014	1.45×10^6	1971	
					5.25–6.75 (1.5)	23,280	23,809	529	0.0028	1.45×10^6	4112	
					6.75–7.95 (1.2)	22,279	23,280	1001	0.0012	1.45×10^6	1738	
					7.95–10.5 (2.55)	20,967	22,279	1312	0.0019	1.45×10^6	2818	
					10.25–10.65 (0.4)	20,679	20,967	288	0.0014	1.45×10^6	2014	
					10.65–11.75 (1.1)	19,137	20,679	1542	0.0007	1.45×10^6	1034	
					11.75–12.95 (1.2)	16,410	19,137	2727	0.0004	1.45×10^6	638	
					12.95–14.54 (1.59)	13,250	16,410	3160	0.0005	1.45×10^6	729	
					14.54–16.09 (1.55)	12,999	13,250	251	0.0062	1.45×10^6	8954	
					0–16.09 (16.09)	12,999	28,300	15,301	0.0011	1.45×10^6	1525	
Canteen Creek, IL	15.5	38.64N	90.02W	Upland	0–10.9 (10.9)	19,114	28,300	9186	0.0012	1.45×10^6	1721	
					10.9–15.5 (4.6)	12,999	19,114	6115	0.0008	1.45×10^6	1091	
					0–15.5 (15.5)	12,999	28,300	15,301	0.0010	1.45×10^6	1469	
Core G39, IL	9.5	38.72N	90W	Upland	0–1.1 (1.1)	25,951	28,300	2349	0.0005	1.45×10^6	679	
					1.1–3 (1.9)	23,580	25,951	2371	0.0008	1.45×10^6	1162	

					3–9.5 (6.5)	12,999	23,580	10,581	0.0005	1.45×10^6	891	
					0–9.5 (9.5)	12,999	28,300	15,301	0.0006	1.45×10^6	900	
Rapids City, IL	9.2	41.57N	90.37W	Upland	0–0.6 (0.6)	23,780	28,300	4520	0.0001	1.45×10^6	192	
					0.6–9.2 (8.6)	12,999	23,780	10,781	0.0008	1.45×10^6	1157	
					0–9.2 (9.2)	12,999	28,300	15,301	0.0006	1.45×10^6	872	
Ruby Lane, IL	4.6	38.62N	90W	Upland	0–0.4 (0.4)	26,230	28,300	2070	0.0002	1.45×10^6	280	
					0.4–1.1 (0.7)	25,320	26,230	910	0.0007	1.45×10^6	1115	
					1.1–4.6 (3.5)	12,999	25,320	12,321	0.0003	1.45×10^6	412	
					0–4.6 (4.6)	12,999	28,300	15,301	0.0003	1.45×10^6	436	
Core GT-6, WI	4.57	42.68N	90.83W	Upland	0–0.27 (0.27)	27,550	28,300	750	0.0004	1.45×10^6	522	
					0.27–4.57 (4.3)	12,999	27,550	14,551	0.0003	1.45×10^6	428	
					0–4.57 (4.57)	12,999	28,300	15,301	0.0003	1.45×10^6	433	
Athens Quarry N., IL	3.28	40N	89.71W	Upland	0–1.03 (1.03)	25,030	28,300	3270	0.0003	1.45×10^6	457	
					1.03–3.28 (2.25)	12,999	25,030	12,031	0.0002	1.45×10^6	271	
					0–3.28 (3.28)	12,999	28,300	15,301	0.0002	1.45×10^6	311	
Hokah, MN	3	43.75N	91.3W	Upland	0–3.0 (3.0)	12,999	24,860	11,861	0.0003	1.45×10^6	367	
Wempletown SE, IL	1.87	42.32N	89.14W	Upland	0–0.1 (0.1)	22,820	28,300	5480	0.000002	1.45×10^6	26	
					0.1–1.87 (1.77)	12,999	22,820	9821	0.0002	1.45×10^6	261	
					0–1.87 (1.87)	12,999	28,300	15,301	0.0001	1.45×10^6	177	

Wabash River (glacial source)

LAST GLACIAL PERIOD

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate (g/m^2/yr)	Notes
Mt. Vernon, IN	13.3	37.91N	87.96W	Upland	0–13.3 (13.3)	12,999	27,460	14,461	0.0010	1.45×10^6	1484	Mixed Wabash and Ohio River source
Patoka, IN	9.83	38.39N	87.6W	Upland	0–9.83 (9.83)	12,999	27,460	14,461	0.0007	1.45×10^6	986	Assumed basal age similar to Mt. Vernon; aeolian sand 1.45-1.6 m (23 cm)
Cumback, IN	2	38.54N	87.13W	Upland	0–2.0 (2.0)	12,999	27,460	14,461	0.0001	1.45×10^6	201	Assumed basal age similar to Mt. Vernon; 1.3 m of aeolian sand overlies Peoria

Middle and Lower Mississippi Valley (integrates all sources above)

LAST GLACIAL PERIOD

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate (g/m^2/yr)	Notes
Phillips Bayou, AR	22	34.63N	90.63W	Upland	0–5.5 (5.5)	23,410	28,300	4890	0.0011	1.45×10^6	1631	Basal age assumed
	22			Upland	5.5–22 (16.5)	12,999	23,410	10,411	0.0016	1.45×10^6	2298	
	22			Upland	0–22 (22)	12,999	28,300	15,301	0.0014	1.45×10^6	2085	Basal age assumed
West Helena, AR	22	34.63N	90.63W	Upland	0–15.8 (15.8)	21,220	28,300	7080	0.0022	1.45×10^6	3236	Basal age assumed
	22			Upland	15.8–22 (6.2)	12,999	21,200	8201	0.0008	1.45×10^6	1096	
	22			Upland	0–22 (22)	12,999	28,300	15,301	0.0014	1.45×10^6	2085	Basal age assumed
Natchez 3, MS	18.2	31.58N	91.29W	Upland	0–0.25 (0.25)	24,400	28,300	3900	0.0001	1.45×10^6	93	Basal age assumed
	18.2			Upland	0.25–2.8 (2.55)	23,740	24,400	660	0.0039	1.45×10^6	5602	

Table 4 (*continued*)

Locality	Peoria Loess thickness	Latitude	Longitude	Geomorphic setting	Depth interval within Peoria (thickness in m)	Calendar minimum age (yr)	Calendar maximum age (yr)	Calendar age time interval (yr)	Accumulation rate (m/yr)	Assumed bulk density (g/m^3)	Mass accumulation rate ($g/m^2/yr$)	Notes
	18.2			Upland	2.8–5.2 (2.4)	22,727	23,740	1013	0.0024	1.45×10^6	3435	
	18.2			Upland	5.2–11.3 (6.1)	21,047	22,727	1680	0.0036	1.45×10^6	5265	
	18.2			Upland	11.3–14.2 (2.9)	20,932	21,047	115	0.0252	1.45×10^6	36,565	
	18.2			Upland	14.2–18.2 (4.0)	12,999	20,932	7933	0.0005	1.45×10^6	731	
	18.2			Upland	0–18.2 (18.2)	12,999	28,300	15,301	0.0012	1.45×10^6	1725	Basal age assumed
Vicksburg 2, MS	14.7	32.38N	90.67W	Upland	0–5.2 (5.2)	20,391	23,470	3079	0.0017	1.45×10^6	2449	
	14.7			Upland	5.2–9 (3.8)	19,804	20,391	587	0.0065	1.45×10^6	9387	
	14.7			Upland	9–14.7 (5.7)	12,999	19,804	6805	0.0008	1.45×10^6	1215	
	14.7			Upland	0–14.7 (14.7)	12,999	28,300	15,301	0.0010	1.45×10^6	1393	Basal age assumed
Old River, TN	14	35.42N	89.97W	Upland	0–1.2 (1.2)	23,410	28,300	4890	0.0003	1.45×10^6	356	Basal age assumed
	14			Upland	0–14 (14)	12,999	28,300	15,301	0.0009	1.45×10^6	1327	Basal age assumed
Vicksburg 1, MS	12.8	32.38N	90.67W	Upland	0–0.1 (0.1)	23,590	25,440	1850	0.0001	1.45×10^6	78	
	12.8			Upland	0.1–1.8 (1.7)	22,831	23,590	759	0.0022	1.45×10^6	3248	
	12.8			Upland	1.8–4.8 (3)	22,773	22,831	58	0.0517	1.45×10^6	75,000	
	12.8			Upland	4.8–9.3 (4.5)	21,220	22,773	1553	0.0029	1.45×10^6	4202	
	12.8			Upland	9.3–12.8 (3.5)	12,999	21,220	8221	0.0004	1.45×10^6	617	
	12.8			Upland	0–12.8	12,999	28,300	15,301	0.0008	1.45×10^6	1213	Basal age assumed
Natchez 1, MS	10.1	31.58N	91.9W	Upland	0–0.1 (0.1)	22,888	28,300	5412	0.00002	1.45×10^6	27	Basal age assumed
	10.1			Upland	0.1–10.1 (10)	12,999	22,888	9889	0.0010	1.45×10^6	1466	
	10.1			Upland	0–10.1 (10.1)	12,999	28,300	15,301	0.0007	1.45×10^6	957	Basal age assumed
Hornbeak, TN	4.4	36.37N	89.3W	Terrace	0–4.4 (4.4)	12,999	26,055	13,056	0.0003	1.45×10^6	489	
Core OP-16, Union City, TN	2.98	36.42N	89.0W	Terrace	0–0.48	21,630	28,300	6670	0.0001	1.45×10^6	104	Basal age assumed
	2.98			Terrace	0.48–2.98 (2.5)	12,999	21,630	8631	0.0003	1.45×10^6	420	
	2.98			Terrace	0–2.98 (2.98)	12,999	28,300	15,301	0.0002	1.45×10^6	282	Basal age assumed

IES where local conditions promoted loess accumulation (Ruhe, 1969; Hallberg et al., 1978). Valleys that carried Last Glacial outwash from the Des Moines Lobe acted as local sediment sources that enhanced atmospheric dust loading along the southern boundary of the IES.

Four localities were used to calculate MARs for Peoria Loess associated with the IES (Fig. 2, Tables 3 and 4). Thickness of the Peoria Loess at these localities ranges from 12.9 m at the Alburnett Paha (ALB) to 3.66 m at Haywards Paha (HAY) and Palermo (PAL; Fig. 2, Table 4). At all the IES MAR localities, the Peoria Loess buries an erosion surface cut across older deposits and soils, including the Farmdale Soil. The erosion surface must therefore be younger than the upper limiting ages on the Farmdale Soil elsewhere in the region. We chose 22,000 ^{14}C yr BP (24,800 cal yr BP) for the basal age of the IES-source Peoria Loess because the basal age of the Peoria Loess that buries an uneroded Farmdale Soil in the 5–1 m isopach area in central Iowa is 22,750 ^{14}C yr BP (25,590 cal yr BP). Our choice of basal age is a very conservative maximum age and results in conservative MAR estimates.

Peoria Loess MARs at IES-source localities range from 1577 to 443 g/m^2/yr, with a regional average of 781 g/m^2/yr (Table 4). These rates are slightly higher than those associated with Peoria Loess of similar thickness from the Missouri River Valley source.

2.2.3. Upper Mississippi River Valley

Nine localities were selected for MAR calculations in the Upper Mississippi River Valley source area. All except the Gardena section are beyond the margin of the late Wisconsin Lake Michigan Lobe glaciation and contain a complete Peoria Loess succession. At the Gardena section, located 1 km southeast of the Farm Creek section near Peoria (Illinois), early increments of Peoria Loess (Morton Tongue of the Mason Group's Peoria Silt) are buried by Wedron Group till (Delevan Till Member) deposited by the Lake Michigan Lobe (Follmer et al., 1979). Peoria Loess in the Upper Mississippi Valley was ultimately sourced from several lobes along the southern margin of the Laurentide Ice Sheet (Grimley, 2000). Diversion of the Ancient Mississippi River about 22,690 cal yr BP (20,350 ^{14}C yr BP) during the early phases of Peoria Loess accumulation, changed the location of the river from its pre-diversion Green River Valley/Lower Illinois Valley course to the present course along the western side of the state (Anderson, 1968; McKay, 1979a, b; Curry, 1998). The former course of the Ancient Mississippi is marked by an arc of thick loess through west central Illinois (Fig. 2). This change in the river's course shifted the relative contribution of outwash from various lobes of the Laurentide Ice Sheet in the Mississippi and Illinois River Valleys, and produced mineralogical changes in the Peoria Loess deflated from those valleys (Glass et al., 1964; McKay, 1977; Grimley et al., 1998; Grimley, 2000).

The Missouri River joins the Mississippi River near St. Louis, Missouri (Fig. 2), and localities below the junction record loess accumulation from the combined source. Isopachs of Peoria Loess do not show significant changes below the junction, suggesting that the glacial lower Missouri River had little effect on the loess sedimentation system at and below its junction with the Mississippi River. This conclusion is supported by geochemical, clay mineral and magnetic studies that show little influence of Missouri River-source sediment on the composition of Peoria Loess downstream of the Mississippi confluence (Grimley, 2000; Muhs et al., 2001).

Peoria Loess at the localities used to calculate MARs from the Upper Mississippi River Valley source ranges in thickness from a maximum of 15.5 m at the Keller Farm section (Fig. 2; KF) near East St. Louis, Illinois to a minimum of 1.87 m at the Wempletown SE section (WEM) in northwestern Illinois (Berg et al., 1985). We use a basal age of 25,000 ^{14}C yr BP (28,300 cal yr BP) for Peoria Loess in this source area on the basis of several maximum limiting ages obtained by a number of investigators for the upper portion of the Farmdale Soil developed in R/P loess (McKay, 1979a, b; Follmer et al., 1986; Leigh and Knox, 1993, 1994; Hansel and Johnson, 1996; Muhs et al., 2001). We use a minimum limiting age of 11,000 ^{14}C yr BP (12,999 cal yr BP) on the basis of a radiocarbon age of 11,250 $\pm$ 100 ^{14}C yr BP on the pyrolosis-volatile fraction of the SOM from a buried soil 2.45 m below the top of the Peoria Loess at the Keller Farm site (Wang et al., 2000; Table 3).

Calculated MARs for the Upper Mississippi River Valley source range from a high of 1525 g/m^2/yr where the loess is 16 m thick adjacent to the valley at the Keller Farm site to a low of 177 g/m^2/yr at a distance of about 90 km from the source at the Wempletown site where the Peoria is 1.87 m thick (Table 4). These values are, in general, lower than those in the IES and Missouri River Valley source areas. Peoria Loess began to accumulate slightly earlier in the Upper Mississippi River Valley source area than in the Missouri River Valley and IES source areas for two reasons. First, the Lake Michigan Lobe advanced over the continental divide earlier than lobes to the west (Mickelson et al., 1983), fostering valley train conditions that favoured loess generation at an earlier date in the Mississippi River Valley. Second, extensive development of the IES, and generation of loess from the erosion surface, was promoted by development of periglacial conditions during early phases of the advance of the Lake Michigan Lobe. Development of the IES therefore lagged behind outwash input from the Lake Michigan Lobe into the Mississippi River Valley.

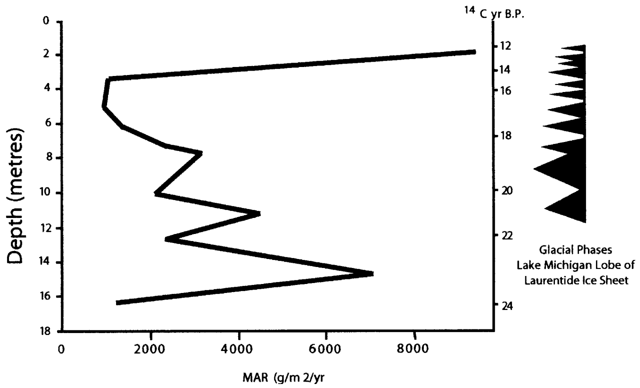

Fig. 5. Changes in MAR of the Peoria Loess at the Keller Farm site near East St. Louis, Illinois. Also shown is the timing of advances of the Lake Michigan Lobe of the Laurentide Ice Sheet correlated with the loess section on the basis of radiocarbon chronology. Radiocarbon data and Lake Michigan Lobe chronology adapted from Wang et al. (2000).

A sequence of 11 radiocarbon ages in vertical succession though the Peoria Loess at the Keller Farm site near East St. Louis, Illinois (Wang et al., 2000) provides a data set suitable for examining temporal variations in MAR during accumulation of the loess (Tables 3 and 4). During the early phases of accumulation, before $\sim$23,000 ^{14}C yr BP, the MAR was low as the Laurentide Ice Sheet advanced southward into the Lake Michigan Basin. Later periods with high MARs coincide with advances of the Lake Michigan Lobe that deposited the Tiskiwa and Lemont formations of the Wedron Group (Fig. 5). After $\sim$17,000 ^{14}C yr BP the MAR of the Peoria Loess dropped significantly and varied less than it had before that time. During this period the Lake Michigan Lobe was much reduced in size and moraine-dammed proglacial lakes may have acted to trap a significant amount of silt before it entered Mississippi River Valley tributaries. The highest MAR in the Peoria Loess occurred after 11,350 ^{14}C yr BP. This peak may differ in magnitude depending on the minimum age assumed for the loess but, even so, the MAR does increase significantly in the upper part of the section. This increase may be related to a period of proglacial lake drainage (Teller, 1990) that delivered large amounts of silt and formed intermittently flooded lowland areas that that were very conducive to dust entrainment.

2.2.4. *Wabash River source*

The Wabash River carried outwash from the southeastern portion of the Lake Michigan Lobe (Decatur Sublobe) and the southwestern portion of the Lake Huron Lobe (East White Sublobe) of the Laurentide Ice Sheet during the Last Glacial period. The Wabash, and to a lesser extent the Ohio River Valley, which drained the Lake Huron and Lake Erie Lobes, were significant sources of Peoria Loess for the eastern portion of the Central Lowland and the northwestern margin of the Appalachian Mountains and Plateaus physiographic regions (Figs. 1 and 2). Peoria Loess thins systematically to the east and northeast from the confluent riverine sources of the Wabash and Ohio valleys (Hall, 1973; Olson and Ruhe, 1979; Fig. 2). Previously discussed basal age/distance from source relationships in Iowa and Illinois indicate that the basal age of the Peoria Loess should decrease with distance from the valley sources, but the lack of chronologic control on the base of the loess along this thickness trend in Indiana does not permit an estimate of the age of the base of the loess along the transect. We use a basal age of $24{,}160 \pm 130$ ^{14}C yr BP (27,460 cal yr BP) for the Peoria Loess in this source area. This is a new AMS radiocarbon age of our own on the $<63\,\mu$m humic acid fraction of the uppermost horizon of the Farmdale Soil at the Mt. Vernon section (Fig. 2, MV; see also Fig. 3 and Table 3), and provides a conservative maximum age for the Peoria Loess in this area.

MARs were calculated for three localities in the Wabash/Ohio source area (Fig. 2); Mt. Vernon (MV; Peoria Loess thickness 14.8 m), Patoka (PAT; Peoria Loess thickness 9.83 m), and Cumback (CBK; Peoria Loess thickness 2.0 m). Data for these localities come from Hall (1973), Olson and Ruhe (1979), Ruhe and Olson (1980) and Hayward and Lowell (1993). We also described and measured the Mount Vernon, Indiana section for this study. MARs range from a high of 1484 g/m^2/yr near the Wabash and Ohio River Valley junction at Mt. Vernon to a low of 2001 g/m^2/yr at the Cumback site located the farthest from the Wabash and Ohio River Valley sources (Table 4). The MAR at Patoka, located on the eastern margin of the Wabash River Valley, is intermediate between that at Mt. Vernon and Cumback. These values are comparable to MARs at sites of similar loess thickness in the Upper Mississippi River Valley source area.

2.2.5. *Lower Mississippi Valley source*

This source extends from the junction of the Ohio River Valley with the Mississippi River Valley south to the head of the Mississippi River delta in Louisiana (Fig. 2). The Lower Mississippi River Valley ranges in width from 50 to 170 km, generally decreasing in width southward. Several distinct Last Glacial braid plains are evident in the valley (Saucier, 1994; Blum et al., 2000) and the location of the Peoria Loess source (the active braid plain) shifted during the period of loess accumulation on the upland. The Arkansas and Red River Valleys drain portions of the central and southern Great Plains and contributed water and sediment to the Lower Mississippi River Valley during the Last Glacial period. The Arkansas drained glaciers in the Rocky Mountain Front Range, but the Red River did not carry glacial outwash. Loess and aeolian sand deposition occurred in

the upper portion of both river basins during the Last Glacial period (Holliday, 1989; Olson et al., 1997; Olson and Porter, 2002).

Peoria Loess extends in a band 20–30 km wide from western Kentucky south to Baton Rouge, Louisiana, and is continuous across Crowley's Ridge, an intra-valley Tertiary alluvial remnant in the northern part of the lower valley (Fig. 2; Autin et al., 1991). Loess is present on both sides of the valley but attains its greatest thickness and continuity on the east. Peoria Loess on Crowley's Ridge was deflated from valley trains in the Western Lowlands to the east of the ridge (West et al., 1980; Rutledge et al., 1985, 1996).

The Lower Mississippi River Valley integrates all the loess sources of the region east of the Rocky Mountains and west of the Appalachians. Peoria Loess derived from this area is geochemically similar to Peoria Loess of the Upper Mississippi River Valley (Muhs et al., 2001). This suggests that the Upper Mississippi River Valley contributed the bulk of the total silt load to the lower valley during the Last Glacial period. Loess in the lower valley has been leached of carbonates to a greater depth than that in the Upper Mississippi River Valley due to higher annual precipitation in the southern mid-continent (Ruhe, 1984b; Muhs et al., 2001).

Peoria Loess in the Lower Mississippi River Valley has not been as extensively studied as it has in the upper valley. Eastward thinning of the loess has been recognized since the early part of the 20th century (Emerson, 1918). Wascher et al. (1948) presented the first map of loess isopachs for the lower valley. Snowden and Priddy (1968) elaborated on Peoria Loess thickness trends in Mississippi and presented a graph showing thickness trends relative to distance from the eastern bluff of the Mississippi River Valley.

Chronologic control on the Peoria Loess in the lower valley is restricted to near-source localities. Information regarding the age of the base of the Peoria with distance from the Mississippi River Valley source does not exist in the lower valley. Radiocarbon ages on charcoal in the upper part of the Farmdale Soil at Phillips Bayou (at the southern end of Crowley's Ridge) and Old River (Tennessee) suggest that the base of the Peoria Loess is less than 29,000 cal yr BP in the lower valley (Markewich et al., 1998). We use a conservative maximum age estimate of 28,300 cal yr BP for calculating Peoria Loess MARs in the lower valley. All except two of the remaining radiocarbon ages from within the Peoria Loess of the lower valley were determined on the tests of terrestrial gastropods (Table 3). As with all the other sources, we used an upper limiting age of 12,999 cal yr BP for the Peoria Loess. Total-section MARs range from about 2085 g/m^2/yr at 22 m-thick sections on Crowley's Ridge (Phillips Bayou and West Helena; Fig. 2, PB) to 282 g/m^2/yr at a 2.98 m-thick section near Union City, Tennessee (Fig. 2, OP-16; Table 4). MARs of Peoria Loess in the lower valley are comparable to those at sections with similar loess thickness in the Upper Mississippi River Valley. The Natchez 3 (N) and Vicksburg 1 (V) sections in Mississippi (McCraw and Autin, 1989) have a sufficient number of dated horizons to examine variations in MAR through the lower two-thirds of the Peoria Loess. All three sites show high MARs prior to about 20,000 cal yr BP, and much lower MARs thereafter. This trend is consistent with MARs at Keller Farm, Illinois, the only site in the Upper Mississippi Valley with a comparable number of ages (Table 4).

2.3. Loess of the Great Plains province

In this paper, the central Great Plains includes Nebraska, Kansas, and eastern Colorado, where late Quaternary loess is the most areally extensive surficial sediment (Figs. 1 and 2). Central Great Plains loess has many similarities to loess of the Central Lowland region in that it forms an extensive sediment blanket of varying thickness over upland parts of the landscape and records multiple episodes of aeolian sedimentation, separated by buried soils. Four middle-to-late Quaternary loess units have been identified and correlated on the Great Plains, from oldest to youngest, namely Loveland Loess, the Gilman Canyon Formation, Peoria Loess and the Bignell Loess (Schultz and Stout, 1945; Frye and Leonard, 1951; Leonard, 1951; Johnson, 1993; Muhs et al., 1999b). Loveland Loess is usually no more than a few metres thick and is the oldest loess unit exposed at many localities. In places, it appears to have an aeolian sand facies. Loveland Loess is identifiable by the presence of the last interglacial Sangamon Soil in its upper part. This palaeosol is usually thick (1–2 m), frequently exhibits 7.5YR hues, and has well-developed prismatic or subangular blocky structure with clay films in the upper part of the B horizon and carbonate coatings or nodules in the lower part of the B horizon. The Gilman Canyon Formation is thin (usually <2 m) and typically has an organic-rich soil developed in it. It was referred to as the "*Citellus* zone" in the older literature (Condra et al., 1947; Frye and Leonard, 1951) because of the presence of krotovina and fossils from small, burrowing mammals. Commonly, the soil developed in the Gilman Canyon Formation is welded to the upper part of the Sangamon Soil. In places, the Gilman Canyon Formation has two buried soils, such as those studied by Maat and Johnson (1996) and Muhs et al. (1999a, b) at Bignell Hill, Nebraska (BH, Fig. 2). The Gilman Canyon Formation is overlain by Peoria Loess, which is the thickest (up to $\sim$48 m) and areally most extensive of the Great Plains loess units (Fig. 6). Peoria Loess dates, at least approximately, to the Last Glacial period and its precise age is discussed in more detail later. A dark, organic-rich buried soil, referred to as the

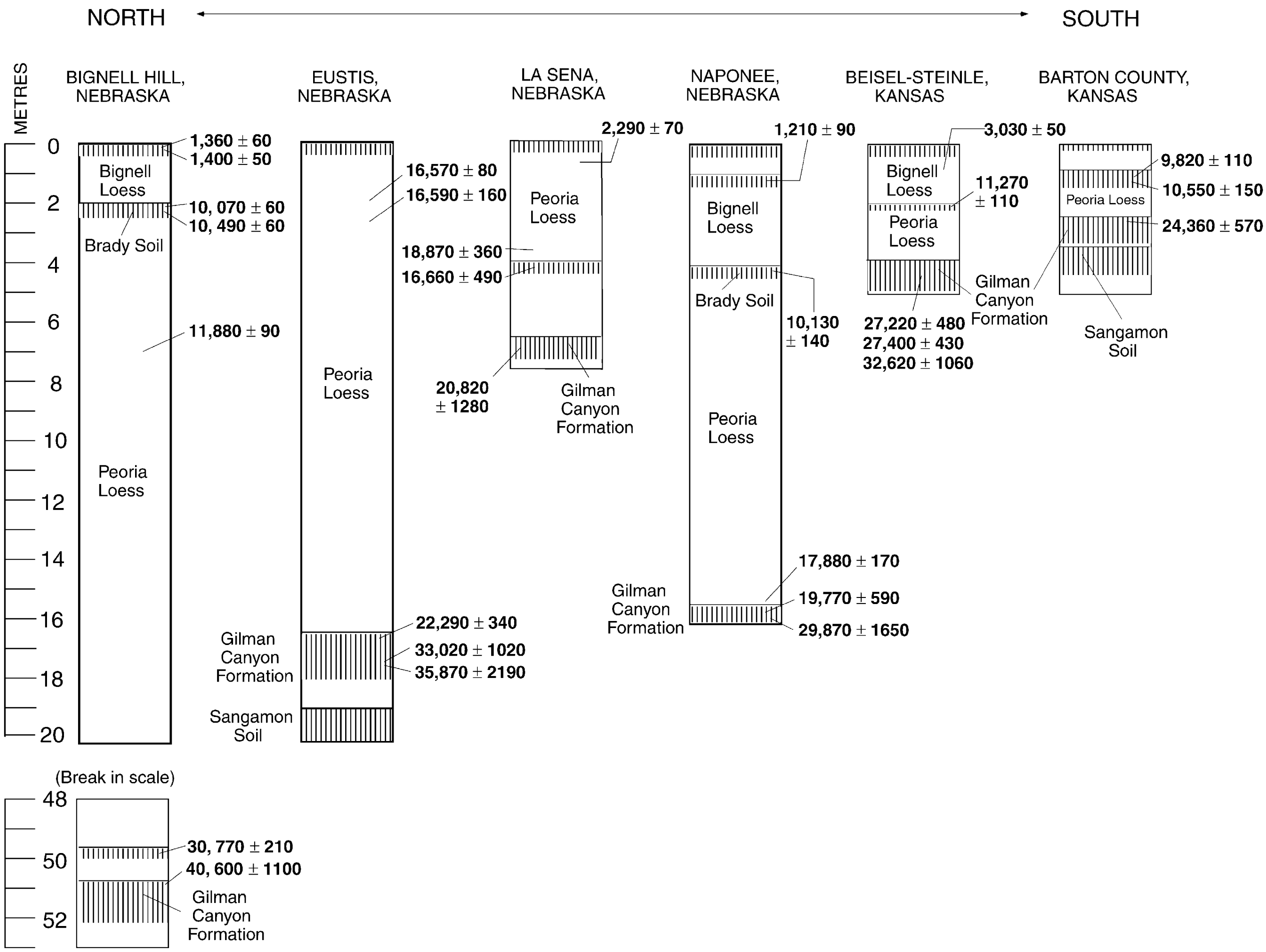

Fig. 6. Stratigraphic profiles of selected loess localities in the Great Plains province. Ages shown are ^{14}C yr BP. Deposits beneath the Last Interglacial Sangamon Soil are not shown. Stratigraphic and age data from Johnson (1993), May and Holden (1993), Martin (1993), Feng et al. (1994a, b), Maat and Johnson (1996), Muhs et al. (1999a), and Johnson and Willey (2000).

Brady Soil, caps the upper part of the Peoria Loess, separating it from the overlying Bignell Loess. Bignell Loess is usually no more than ~2 m thick and has a patchy distribution. The thickest (up to 6 m) Bignell loess occurs along the top of table-edge escarpments facing dune fields and sand sheets (Mason et al., 2002, in press). Bignell Loess has been found in Nebraska, Kansas and Colorado (Frye and Leonard, 1951; Caspall, 1972; Muhs et al., 1999a; Johnson and Willey, 2000; Bettis et al., 2003), but has not been reported east of the Missouri River. In southeastern Nebraska, Bignell Loess is so thin that it cannot be identified in the field but can be detected mineralogically in the upper horizons of modern soils (Kuzila, 1995).

The Great Plains province has some of the more areally extensive and thick loess deposits in the world (Fig. 2). Large areas of Nebraska have loess thicknesses in excess of 20 m and Bignell Hill, Nebraska has almost 50 m of Peoria Loess (Maat and Johnson, 1996; Muhs et al., 1999a, b). Loess thicknesses of up to 20 m occur in northwestern Kansas, but thicknesses of 1–5 m are more common. In the Great Plains, loess thins southeastward from southwestern and central Nebraska to central and southern Kansas (Fig. 6). Based on studies at many localities in Nebraska and Kansas, the thickest unit by far is the Peoria Loess. Thinner loess that may date to the latest part of the Last Glacial period (as opposed to the full-glacial period) or to the Holocene covers a large area of southwestern North Dakota, but is generally only 1–2 m thick (Clayton et al., 1976). In the panhandle areas of Oklahoma and Texas, there are extensive occurrences of upland aeolian deposits that have been referred to as loess by Allgood et al. (1962), Reeves (1976) and Denne et al. (1993). Using stratigraphic and sedimentologic data, Gustavson and Holliday (1988) and Holliday (1989) showed that these deposits are probably a fine-grained facies of the Blackwater Draw Formation, which occurs as an aeolian sheet sand over much of the panhandle of Texas (Fig. 2). However, particle-size analyses of this sediment in parts of the

Oklahoma panhandle by Allgood et al. (1962) indicate that it contains 41–54% silt (2–53 μm) and therefore can probably be considered as at least a ''loessial'' or ''loessic'' sediment (Fig. 2).

Particle sizes of Peoria Loess in the Great Plains are quite variable. In Kansas, it has a mean particle size of medium silt (5–6 phi, or 31–15 μm) (Swineford and Frye, 1951, 1955). In contrast, eastern Colorado loess has a mean particle size of fine silt (6–7 phi, or 15–7 μm) (Muhs et al., 1999a). The finer mean particle size of Colorado loess is in part a function of exceptionally high clay contents (up to 35%), thought to be derived from local bedrock sources such as the Pierre Shale. For Nebraska loess, Winspear and Pye (1995) reported a mean particle size of coarse silt (4.25–4 phi or 53–57 μm), but their measurements were made using a Coulter laser granulometer and the authors did not specify whether organic matter and carbonates were removed prior to analysis. If these components were not removed, then it is very likely that the apparently coarser textures of Nebraska loess reported by Winspear and Pye (1995) are due to flocculated grains and so may overestimate the mean particle size.

Carbonate contents of loesses in the Great Plains are variable but are lower, on average, than those for loesses of the Central Lowland province. Loess in Kansas has carbonate contents of 6–21%, with most being between 10% and 18% (Swineford and Frye, 1951). Loess in Colorado has carbonate contents in the range 3–12%, with most values between 7% and 9% (Muhs et al., 1999a). X-ray diffraction indicates that both calcite and dolomite are present in eastern Colorado loess. The generally lower carbonate contents in the Great Plains loess compared to that in the Central Lowland are reflected in lower bulk CaO and MgO contents (Fig. 4b). In the Central Lowland, much of the calcite and dolomite are derived from Palaeozoic carbonate rocks that were incorporated into tills deposited by the Laurentide Ice Sheet. In contrast, Al_2O_3 and Fe_2O_3 contents are generally higher in Great Plains loesses (Fig. 4a), due in part to lesser dilution by carbonates, but also arising from a dominance of smectitic clay minerals, which are richer in Fe_2O_3 than are the dominant illitic and kaolinitic clay minerals in the Central Lowland loess (Ruhe, 1984a).

Numerous studies over the past decade have shown that Peoria Loess in the Great Plains region has maximum-limiting radiocarbon ages of ~30,000–20,000 ^{14}C yr BP and minimum-limiting ages of ~10,000 ^{14}C yr BP (Wells and Stewart, 1987; Johnson, 1993; May and Holen, 1993; Martin, 1993; Kuzila and Lewis, 1993; Feng et al., 1994a, b; Maat and Johnson, 1996; Muhs et al., 1999a, b; Mason and Kuzila, 2000; Johnson and Willey, 2000). Maximum-limiting ages come mostly from analyses of organic matter from the upper part of the Gilman Canyon Formation and minimum-limiting ages are mostly from organic matter from the lower part of the Brady Soil. Nevertheless, at some localities, Peoria Loess contains plant macrofossils, charcoal or gastropod tests that allow direct dating of the period of aeolian sedimentation. Charcoal within Peoria Loess, usually identified as *Picea*, has been found at several localities in Nebraska and Kansas. Radiocarbon ages of charcoal or organic-rich sediment within Peoria Loess range from 21,250±530 to 11,880±90 ^{14}C yr BP (Wells and Stewart, 1987; Souders and Kuzila, 1990; Martin, 1993; May and Holen, 1993; Feng et al., 1994a; Maat and Johnson, 1996). These ages are in good agreement with the maximum-limiting and minimum-limiting radiocarbon ages of the Gilman Canyon Formation and Brady Soil, respectively.

Unlike the loess of the Central Lowland, there is considerable uncertainty about the origin of the Great Plains loess. As discussed earlier, Central Lowland loess has a clear glaciogenic origin, because, during the LGM, continental ice had entered the headwaters of the Missouri, Mississippi, Illinois, and Ohio rivers. Fine-grained particles from silt-rich outwash in these valley trains were deflated by northwesterly and westerly winds and deposited as loess over much of the Central Lowland province. Loess distribution, thickness, particle size and geochemistry have clear down-wind trends that support this model (Ruhe, 1983; Muhs and Bettis, 2000). In contrast, continental glaciers were not present in the Great Plains except for the western portions of the Laurentide Ice Sheet, which reached northeastern Montana and parts of the Dakotas. Only much smaller alpine glaciers were present on the eastern side of the continental divide in the Front Range of Colorado (Madole et al., 1998), and little or no glaciation occurred in the mountains of eastern Wyoming (Porter et al., 1983). The lack of a clear glaciogenic link to Great Plains loess has generated debate about its origin for more than 50 years. Some investigators have favoured a glacial outwash origin, suggesting that the Platte and Arkansas Rivers, which head in the Rocky Mountains of Colorado, were major sources (Bryan, 1945; Frye and Leonard, 1951; Swineford and Frye, 1955). Other workers minimized (but did not exclude) the importance of glacial outwash as a source and emphasized additional sources such as non-glaciogenic alluvium, old till sheets, silt-rich Tertiary bedrock such as the White River Group, and aeolian sand seas, such as the Nebraska Sand Hills (Condra and Reed, 1950; Lugn, 1968). Flint (1971) also questioned the origin of Great Plains loess from glacial outwash alone, because of the relatively small size of Pleistocene alpine glaciers in Colorado compared to the volume of loess in the Great Plains. Based on revised mapping of loess in Kansas, Welch and Hale (1987) concluded that Kansan loess probably had multiple sources, including glacial

outwash, aeolian sand, and the Tertiary Ogallala Formation.

The authors cited above did not present any empirical data in support of their hypotheses of loess origin. With regard to aeolian sand seas as a possible loess source, Muhs and Zárate (2001) showed that aeolian sand and loess in both Colorado and Nebraska have distinctive Ti, Zr, Ca, and Sr concentrations and Ti/Zr and Ca/Sr ratios. These data indicate that loess and aeolian sand of the Great Plains have different sources. Glacial outwash is certainly a candidate for a loess source in eastern Colorado, given the proximity of the region to late Wisconsin (Pinedale) alpine glaciers in the Front Range. Another possible source sediment is the White River Group of Oligocene age, which is derived dominantly from volcanic ash. The White River Group is attractive as a possible loess source because it is physically and chemically weathered, contains 65–85% silt and 20–30% $CaCO_3$ (Denson and Bergendahl, 1961), commonly has little or no protective soil cover, and has a broad distribution upwind of most loess deposits in Nebraska and eastern Colorado. Detailed Pb-isotopic analyses of K-feldspars show that both Front Range-derived alluvium and White River Group sediments were important sources of loess in eastern Colorado (Aleinikoff et al., 1999). This conclusion is supported by U-Pb analyses of detrital zircons, which show that eastern Colorado loess contains sediments derived from Proterozoic (1.7, 1.4, and 1.0 Ga) Front Range rocks and Tertiary (34 Ma) White River Group sediments. Furthermore, the relative importance of these two sources varied through the Last Glacial period (Aleinikoff et al., 1999). Preliminary Pb-isotopic (K-feldspars) and U-Pb (zircons) analyses of Nebraska loesses indicate that the White River Group is a dominant source in that region (Aleinikoff et al., 1998).

The significance of the isotopic results for Nebraska and Colorado is that they demonstrate the importance of a non-glaciogenic source for the thickest and areally most-extensive loess deposits in North America. This finding is not in agreement with models that simulate little or no dust generation in this region during the Last Glacial period (Mahowald et al., 1999; Kohfeld and Harrison, 2000). Loess that dates to the Last Glacial period is as thick as 32 m at Devil's Den, Nebraska (north of the Platte River) and is 48 m thick at Bignell Hill, Nebraska (south of the Platte River). As already mentioned, Peoria Loess at Bignell Hill may be the thickest documented occurrence of Last Glacial loess in the world.

2.4. MARs of Great Plains Late Glacial and Holocene loess

We calculated MARs (see Eq. (1)) for localities that spanned a wide range of loess thickness and where reliable radiocarbon age control exists (Figs. 2 and 6; Table 3). The calculations are based on the following assumptions: (1) calendar-year-converted radiocarbon ages of uppermost Gilman Canyon Formation organic matter provide a maximum-limiting age for the initiation of Peoria Loess deposition; (2) calendar-year-converted radiocarbon ages of lower Brady SOM provide a minimum-limiting age for the termination of Peoria Loess deposition; (3) the bulk density of loess is 1.45×10^6 g/m^3 (Natural Resources Conservation Service, file data; Gibbs and Holland, 1960). Errors related to these assumptions are similar to those outlined in the discussion of the Central Lowland region.

Because of the large range in Peoria Loess thickness, MARs for this region show a range spanning two orders of magnitude, with a systematic southward decrease in rates (Table 3; Fig. 7). The highest value for the Great Plains is Bignell Hill, Nebraska (Fig. 2, BH), where the calculated MAR exceeds 3000 g/m^2/yr. Southern Nebraska has lower MARs of 600–2000 g/m^2/yr, based on results for the Eustis (E), Naponee (N) and La Sena (L) localities. The calculation for Eustis is based on an assumed termination age for Peoria Loess, because the Brady Soil and Bignell Loess, if they exist at this locality, are incorporated in the modern surface soil. In central Kansas, MARs for Peoria Loess are more than an order of magnitude lower than those from central Nebraska.

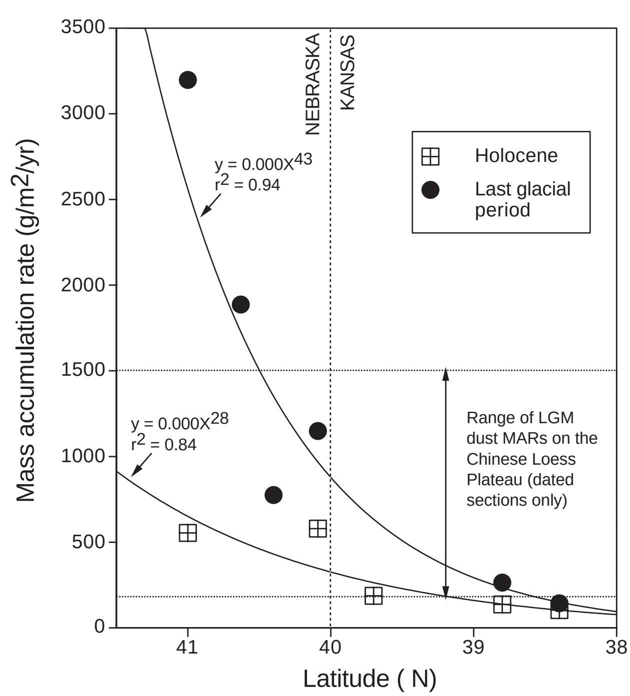

Fig. 7. Plot showing the systematic decrease in MAR of Last Glacial and Holocene loess from central Nebraska southward to central Kansas. Note that Last Glacial MARs are much higher than Holocene rates in central Nebraska, but that the rates converge southward with distance from the source area. Chinese Loess Plateau data from Sun et al. (2000).

Based on OSL and TL ages from samples collected *within* Peoria Loess, Last Glacial MARs may have been much higher than the estimates made using bracketing radiocarbon ages (Table 4). For example, OSL ages reported by Roberts et al. (2003) indicate average MARs of $\sim$11,500 g/m^2/yr at Bignell Hill and $\sim$3500 g/m^2/yr at Eustis for the period between 18,000 and 14,000 yr BP. Using the OSL ages from the top and bottom of the Peoria Loess units, the MARs calculated are 6108 g/m^2/yr for Bignell Hill, and 3011 g/m^2/yr for Eustis, compared to values shown in Table 4 of 3198 g/m^2/yr (Bignell Hill) and 1396–1887 g/m^2/yr (Eustis) for MARs calculated using bracketing radiocarbon ages. This comparison suggests that the use of radiocarbon ages from bracketing palaeosols may be too conservative in assessing the true flux of Last Glacial loess.

Last Glacial MARs for the Great Plains can be compared with Holocene MARs for the same region (Fig. 7). Holocene MARs were calculated on the basis of the *youngest* radiocarbon ages (converted to calendar-year ages) of the uppermost Brady Soil, considered to be the best maximum-limiting age for the Bignell Loess. Because Bignell Loess deposition could have been initiated some time after this assumed age, our MAR calculations for the Holocene, as with the Last Glacial period, are conservative. As noted for the Last Glacial loess, Holocene MARs show a systematic southward decrease. All Holocene MARs are lower than Last Glacial rates at the same or nearby localities; however, the difference in the flux rates decreases in a southward direction. At Barton County, Kansas, the southernmost locality, Last Glacial and Holocene rates are not significantly different.

Great Plains MARs for the Last Glacial period can also be compared to those calculated for other regions. Based on sections where there is radiocarbon and/or TL age control, the Chinese Loess Plateau has Last Glacial MARs that range from 1449 to 105 g/m^2/yr, with an average of about 400 g/m^2/yr (Sun et al., 2000) (Fig. 7). Overall, these rates are lower than those of the Great Plains. Moreover, Last Glacial MARs at dated localities in China are, on average, *lower* than Holocene rates, in contrast to the Great Plains. However, MARs have also been calculated for Chinese loess sections that are undated, but with sediments correlated to the Holocene and Last Glacial period based on soil stratigraphy. These sections show lower Holocene MARs and higher Last Glacial MARs compared to those sections with radiocarbon and TL age control. We conclude from this comparison that more directly dated loess sections are needed in both China and in the Great Plains of North America.

3. Loess of the Colorado Plateau province

Both geologic and soil mapping have revealed that there are extensive aeolian silts, sandy silts and sandy clay loams on upland bedrock areas of the Colorado Plateau (Fig. 8). Thorp and Smith (1952) mapped an extensive area of thin loess in southwestern Colorado and southeastern Utah. Arrhenius and Bonatti (1965) referred to this material as the "Mesa Verde loess." The range and average thickness of aeolian sediment in this region is uncertain, but soil surveys and data in Price et al. (1988) suggest that some may be more than 2 m thick, although the thickness of much of it may not exceed 1 m. Sample transects in southwestern Colorado and southeastern Utah show that silt content (calculated on a clay-free basis) varies from about 20% to more than 60%, indicating that much of the material is sedimentologically similar to true loess found in the Central Lowland and Great Plains provinces. Furthermore, sand content decreases and silt content increases with increasing distance northeast of the San Juan River (Fig. 8), a probable contributing source according to Price et al. (1988). Recent studies in the Canyonlands National Park area of Utah have demonstrated aeolian silt additions to soils on upland sandstone surfaces using mineralogic, magnetic and geochemical data (Reynolds et al., 2001).

The age of Colorado Plateau loess is not certain. Hunt (1956) recognized the occurrence of this material in the region, but thought that it pre-dated the Last Glacial period. In contrast, Arrhenius and Bonatti (1965) thought that the Mesa Verde loess was deposited in two periods of the Holocene. However, Price et al. (1988) reported a radiocarbon age of 15,970 $\pm$ 155 ^{14}C yr BP for a muskox (*Ovobos* sp.) bone buried below the zone of pedogenesis in aeolian material north of Cortez, Colorado. On the basis of this age and the degree of soil development, these workers inferred that aeolian sedimentation was complete by the beginning of the Holocene. A single radiocarbon age and a lack of well-studied stratigraphic sections preclude calculations of MARs for this region. However, the widespread occurrence of this sediment suggests that its importance has been overlooked and invites further study.

4. Loess of the Columbia Plateau province

Loess is extensive over parts of the northwestern United States, including a small part of western Wyoming, a large part of southeastern Idaho, a small part of northern Oregon, and much of eastern Washington (Fig. 9). Loess in this region has not been studied for as long as loess in the Central Lowland and Great Plains provinces. However, recent work indicates that a long Quaternary stratigraphic record is present and that loess origins in parts of the region differ from those found elsewhere in North America.

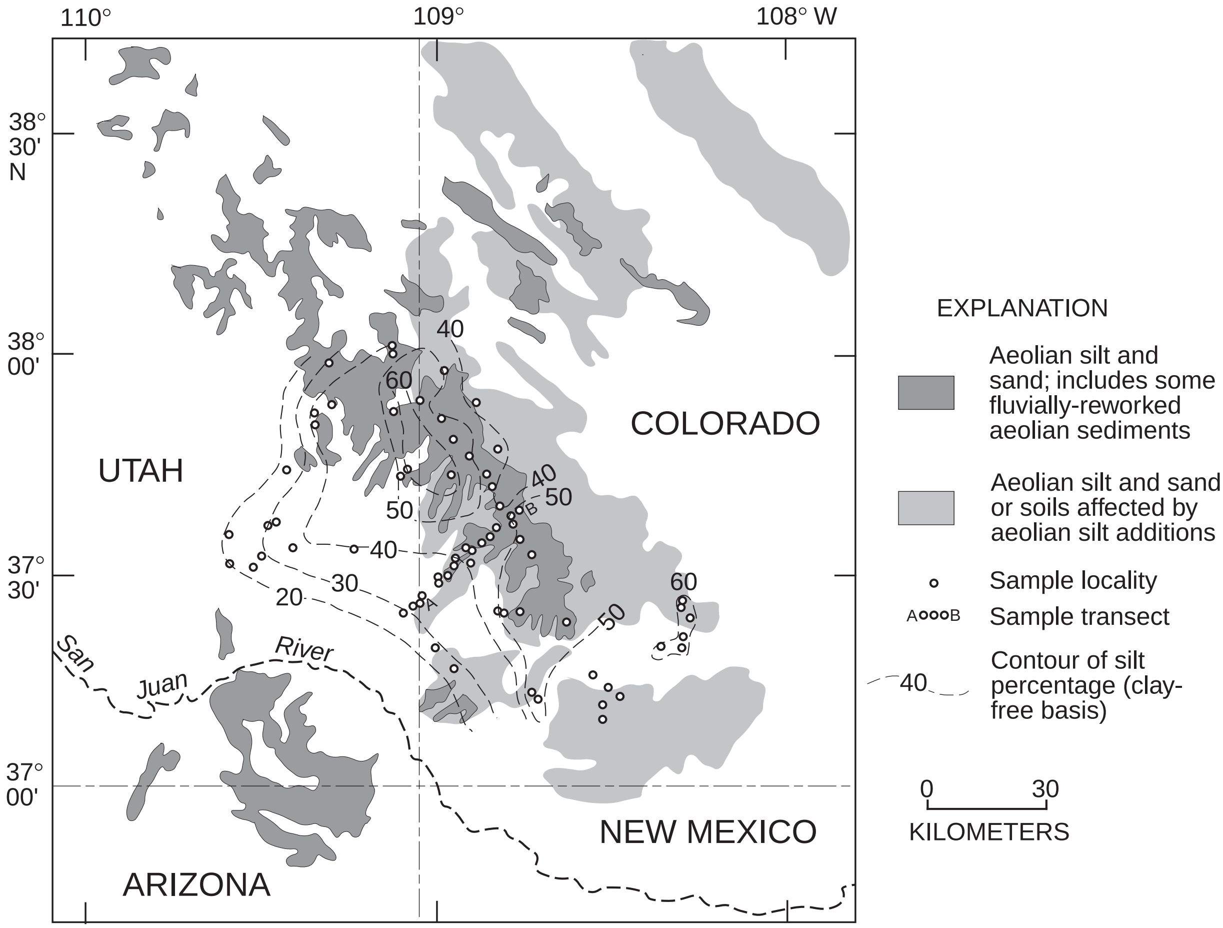

Fig. 8. Map showing the distribution of aeolian deposits and soils affected by aeolian additions in the Colorado Plateau province. Redrawn from Price et al. (1988).

4.1. Loess in the vicinity of the Snake River Plain

In Idaho and adjacent parts of western Wyoming, loess covers large areas to the north and south of the Snake River Plain, which is probably one of its major sources (Lewis et al., 1975; Pierce et al., 1982; Scott, 1982; Glenn et al., 1983). Loess in this region has a thickness of up to 12 m in places, but most is 2 m or less (Lewis et al., 1975). Thickness and particle size data for localities south of the Snake River (Lewis et al., 1975; Lewis and Fosberg, 1982; Pierce et al., 1982; Glenn et al., 1983) suggest that loess deposition took place under northwesterly or westerly winds (Fig. 9). However, loess (though generally thinner: see Scott, 1982) also occurs north of the Snake River, which would not be expected if winds had been consistently from the northwest. To our knowledge, there have been no systematic studies of the source sediments of loess in Idaho using geochemical or isotopic methods, but such work could yield valuable information on Last Glacial palaeo-winds.

Stratigraphic studies of loess in Idaho have been conducted by McDole et al. (1973), Scott (1982), Pierce et al. (1982), and Forman et al. (1993). McDole et al. (1973) showed that Idaho loess has silt contents similar to mid-continent loess and that there are at least four loess units of Quaternary age separated by buried soils (Fig. 10). The soils are easily identified by Bt horizons with high clay content in their upper parts and Bk horizons with high carbonate content in their lower parts. Pierce et al. (1982) summarized the loess stratigraphy at numerous localities (including the locality studied by McDole et al., 1973) over a ~400-km-long transect from western Wyoming to central Idaho and informally designated the youngest two units as loess "A" and loess "B." The two loesses are separated by a well-developed palaeosol. Based on degree of soil development and stratigraphic relations with K/Ar-dated lava flows and well-dated Lake Bonneville flood deposits, they suggested that loess A was deposited during the Last Glacial period and loess B during the penultimate glacial period. Forman et al. (1993) studied loess at two closely spaced sites in eastern Idaho and reported TL data that supports correlation of loess A with the Last Glacial period (late Wisconsin time). However, these workers suggested, on the basis of TL data, that loess B was also deposited during the

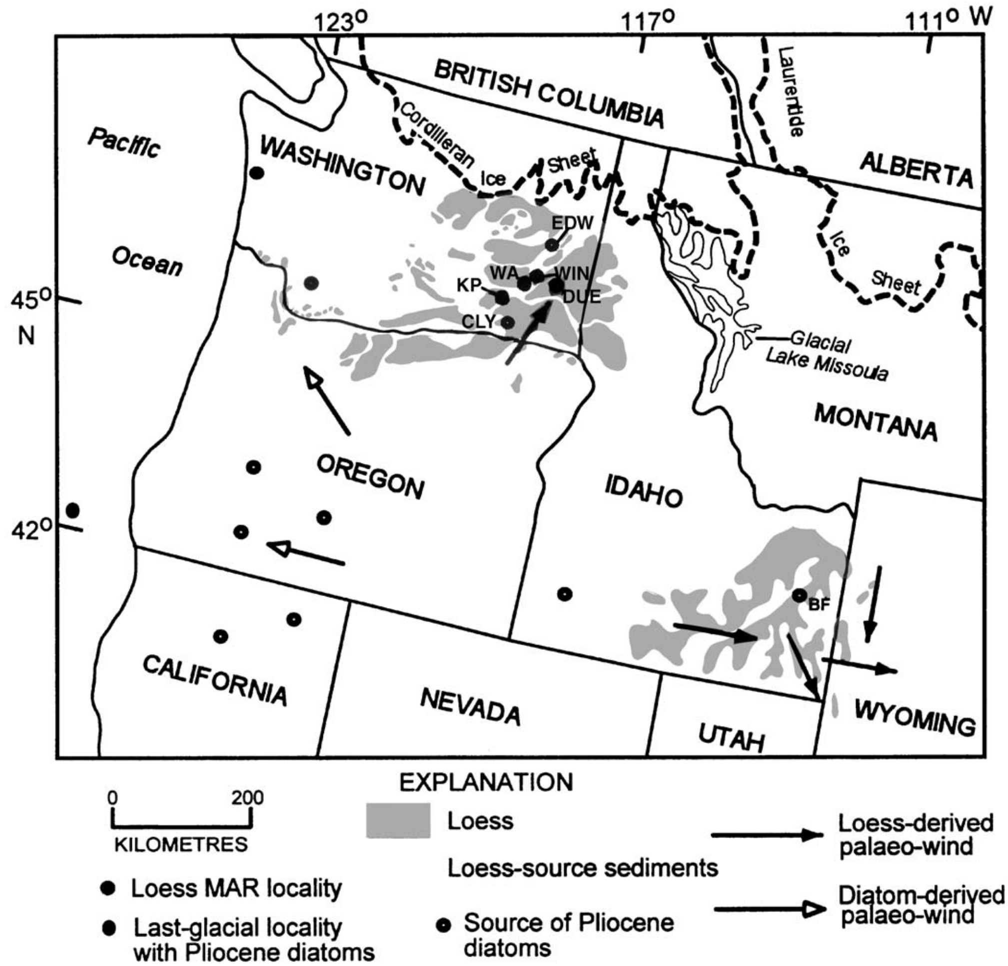

Fig. 9. Distribution of loess and inferred palaeo-winds in the Columbia Plateau province. Loess distribution, inferred palaeo-winds, glacial extent, and Lake Missoula compiled from Lewis and Fosberg (1982) and Busacca and McDonald (1994). Full names and stratigraphy of MAR localities given in Fig. 11, except Blackfoot, Idaho (BF), which is shown in Fig. 10.

Last Glacial period (early Wisconsin time). At present, probably too few data exist to determine which age estimate is correct for loess unit B.

A careful sifting of data yields some constraints on the beginning and ending of loess sedimentation for loess unit A, which all workers agree was probably deposited during the Last Glacial period. Forman et al. (1993) found that the lower parts of loess unit A may date to ~25,000–28,000 ± 3000 yr BP based on TL age estimates, which are minimum ages for the initiation of loess accumulation. Pierce et al. (1982) reported a site where local accumulations of loess unit A post-date the Lake Bonneville flood, which is radiocarbon dated to ~17,000–18,400 cal yr BP, indicating that loess accumulation was still in progress at that time. These workers also thought that the degree of development of modern soils in unit A would require at least 10,000 years. They and other workers (Scott, 1982; Kuntz et al., 1986; Malde, 1991) also pointed out that the Holocene basalts in the region lack a loess cover. Kuntz et al. (1986) analysed charcoal and heated organic matter that occurs in A horizons of soils developed in the upper part of loess in the Craters of the Moon area and elsewhere. Loess in this area was covered by Holocene basalt flows that lack a loess cover. The oldest radiocarbon ages of organic matter and charcoal from these basalt-covered A horizons in loess thus provide limiting ages for the termination of loess deposition. With one exception, the oldest ages considered reliable by Kuntz et al. (1986) range from ~10,200 to 13,900 ^{14}C yr BP (~12,000–16,700 cal yr BP). We conclude from all these data that accumulation of loess A may have begun no later than 28,000–25,000 yr BP and terminated around 12,000–16,000 cal yr BP.

Based on the stratigraphic studies of McDole et al. (1973) and Pierce et al. (1982), loess unit A has a thickness ranging from around 1 m (near Driggs, Idaho) to as much as 6 m (near Blackfoot, Idaho; see Figs. 9, BF and 10). If we assume an average loess bulk density of 1.45×10^6 g/m^3 and a period of loess accretion from

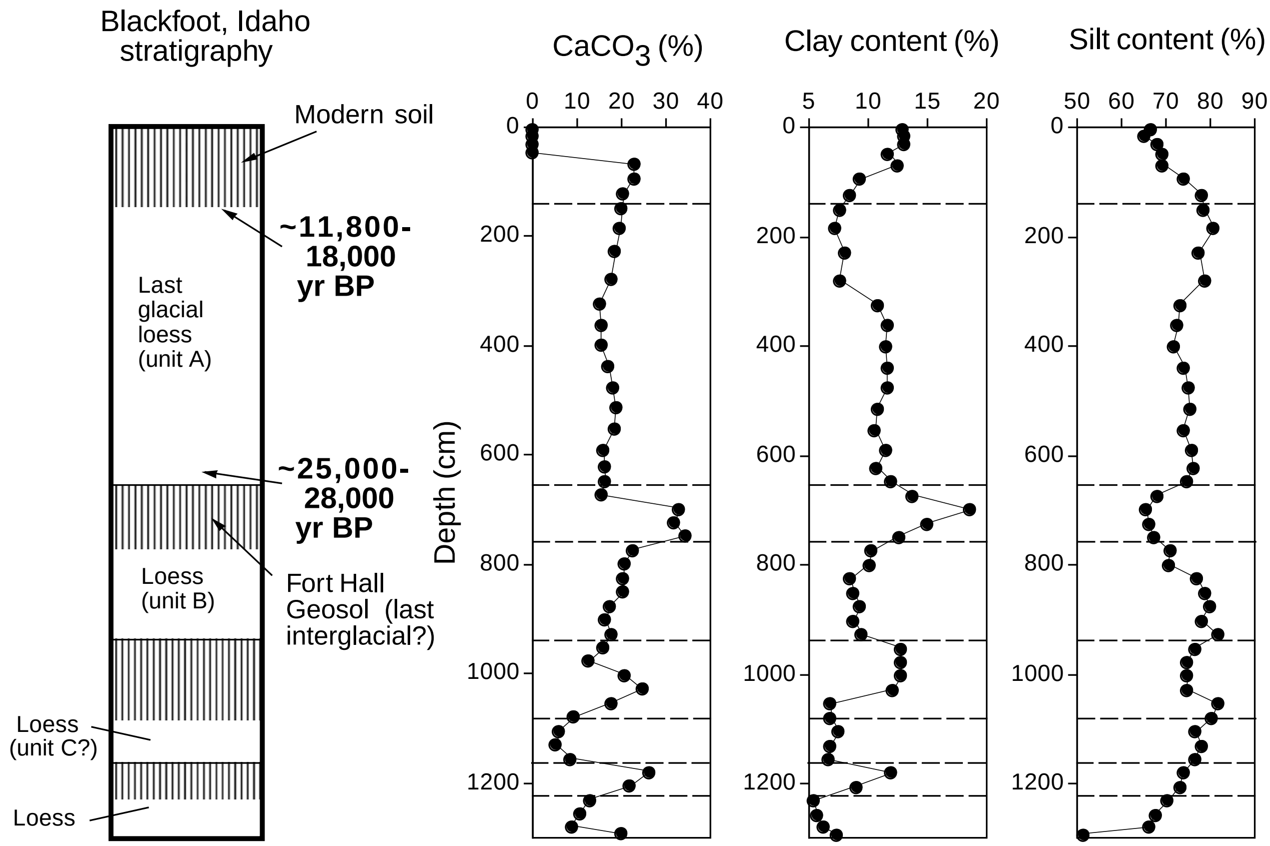

Fig. 10. Loess stratigraphy, carbonate content, and grain size variations at Blackfoot, Idaho in the Snake River Plain of the Columbia Plateau. Redrawn from McDole et al. (1973). Age estimates for loess unit A are from Pierce et al. (1982), Kuntz et al. (1986) and Forman et al. (1993).

~28,000 to ~12,000 cal yr BP, MARs range from about 90 to 540 g/m^2/yr for the Last Glacial period. Use of a shorter interval of loess deposition permitted by the data (from ~25,000 to 16,000 cal yr BP) yields MARs of about 160–960 g/m^2/yr. These rates are comparable to many of those discussed earlier for the Central Lowland and Great Plains provinces.

4.2. Loess in the Palouse region of eastern Washington

In eastern Washington and adjacent parts of northern Oregon and western Idaho, detailed studies by Busacca and colleagues (Busacca, 1991; Busacca et al., 1992; McDonald and Busacca, 1992; Busacca and McDonald, 1994; Berger and Busacca, 1995; Richardson et al., 1997) have added substantially to our knowledge of loess history in this region. Loess in eastern Washington and adjacent parts of Idaho and Oregon may cover as much as 50,000 km^2 (Fig. 9; McDonald and Busacca, 1992). It is as thick as 75 m (Ringe, 1970) and Busacca (1991) estimated that loess deposition may have begun as early as 2 Ma. There are dozens of palaeosols within eastern Washington loess, indicating many periods of non-deposition and surface stability between times of loess accumulation (Fig. 11). The loess itself is thought to be derived primarily from fine-grained slackwater sediments which are derived, in turn, from cataclysmic floods of proglacial Lake Missoula (McDonald and Busacca, 1992).

Busacca and McDonald (1994) designated the two youngest loess units in eastern Washington "L1" (upper) and "L2"(lower). Well-dated Holocene and Late Glacial tephras from the volcanically active Cascade Range to the west are found within eastern Washington loess and provide valuable time lines for dating and correlation (Busacca et al., 1992; Richardson et al., 1997). In addition, TL ages by Berger and Busacca (1995) and TL and IRSL (infrared stimulated luminescence) ages by Richardson et al. (1997) confirm correlations inferred from tephra data that L1 is of late Wisconsin-to-Holocene age and L2 is of early to-middle Wisconsin age. Agreement between TL and IRSL ages is good back to ~70,000 yr. There are differences between these two luminescence studies, however. TL ages reported by Richardson et al. (1997) are generally younger than those reported by Berger and Busacca (1995) for the same sections.

Stratigraphic and geochronologic data show that loess accumulation took place in eastern Washington late in the glacial cycle. The loess is derived primarily

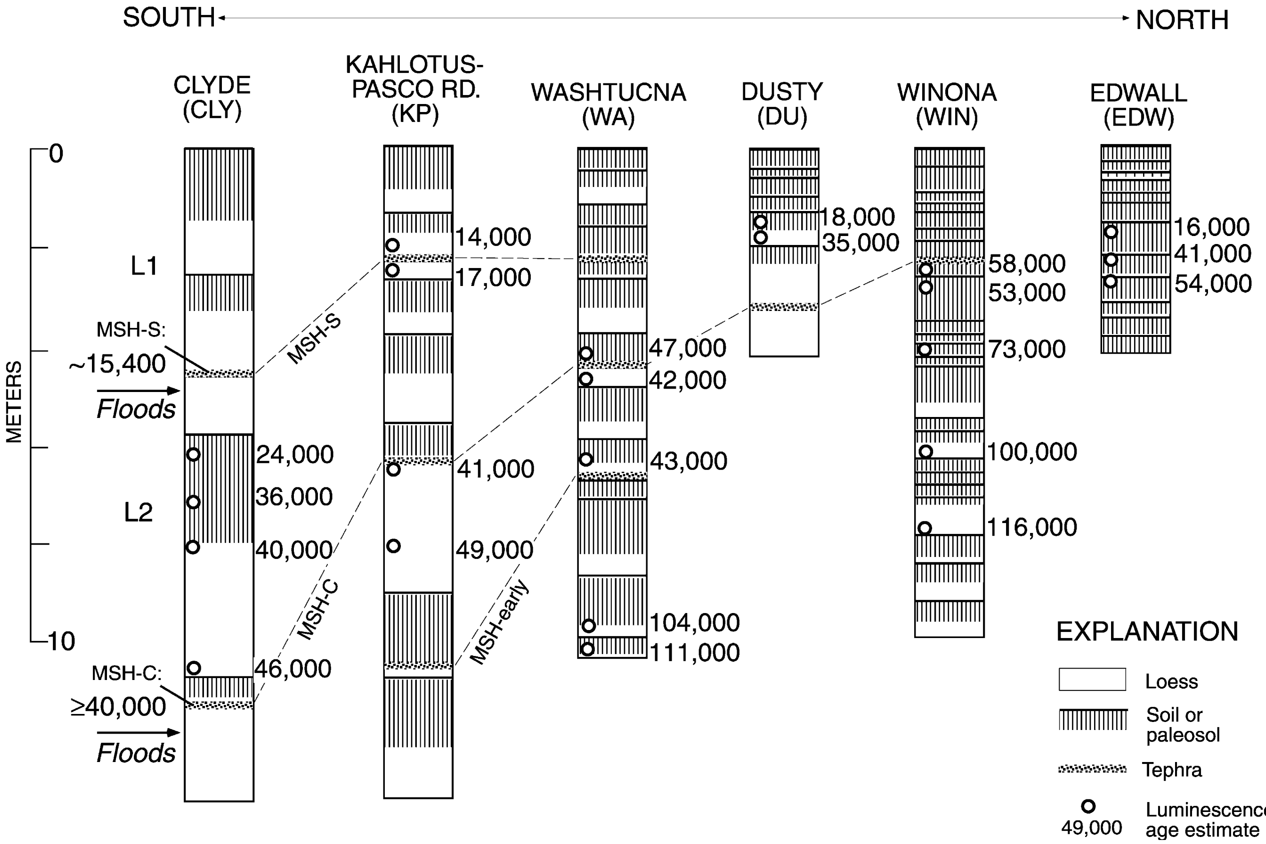

Fig. 11. Loess stratigraphy and luminescence chronology at selected localities in the Palouse region of the Columbia Plateau. Localities shown in Fig. 9. MSH, Mount St. Helens tephras. MSH-C is estimated to be 36,000–38,000 ^{14}C yr BP ($\sim$41,000 cal yr BP) and MSH-S is 12,000–14,000 ^{14}C yr BP (13,680–16, 934 cal yr BP) (Crandell et al., 1981; Mullineaux, 1986). Re-drawn from Richardson et al. (1997).

from sediments derived from cataclysmic floods of proglacial Lake Missoula (Fig. 9). Because the Cordilleran Ice Sheet had to reach its southernmost limit for Glacial Lake Missoula to form, outburst flooding apparently did not begin until sometime after the glacial maximum. This interpretation is supported by the fact that thin loess underlies the flood sets at one locality (Clyde; A. Busacca, pers. comm., 2002) and that Mount St. Helens "S" tephra (calendar-year age of $\sim$15,300 yr) lies slightly above, on top of, or in the upper part of a well-developed buried soil called the Washtucna Soil (Fig. 11; Busacca and McDonald, 1994). The tephra also occurs in the flood deposit silts (Bjornstad et al., 1991). Thus, there may have been little loess accumulation during glacial maximum conditions, and Busacca and co-workers interpret full glacial time to be a period primarily of soil formation. Loess accumulation may have taken place rather rapidly in Late Glacial time, around or shortly after $\sim$15,000 yr BP. Another soil, called the Sand Hills Coulee Soil, is weakly developed and occurs in loess above Mount St. Helens "S" tephra. In addition to Late Glacial loess accumulation, substantial deposition apparently occurred in the Holocene. This interpretation is based on observations of up to 1 m of loess overlying Mount Mazama tephra ($\sim$7700 yr), which, in turn, lies above the Sand Hills Coulee Soil (Busacca, 1991; Busacca and McDonald, 1994). Holocene loess at these localities is almost half as thick as the underlying Late Glacial loess. Given the relatively low trapping efficiency of surfaces in the Palouse and Snake River Plain, this may suggest that MARs for Holocene loess in the region may be similar to those for the Late Glacial. Current studies suggest that the glacial slackwater sediment sources continued to feed the loess sedimentation system during the Holocene (Alan Busacca, pers. comm., 2002). Eastern Washington loess becomes thinner and finer-grained from southwest to northeast, indicating that mid-Wisconsin and latest Wisconsin winds were from the southwest, similar to present winds (Busacca, 1991; Busacca and McDonald, 1994; Fig. 9).

5. Aeolian silt and clay in the Basin and Range province

Although thick loess has not been reported in the semi-arid and arid Basin and Range province, detailed soil and stratigraphic studies have shown that significant amounts of aeolian silt and clay have been deposited in

the region. Silt-loam or loam vesicular A horizons, interpreted as aeolian mantles, have been described from numerous localities in California and Nevada (Muhs, 1983; Wells et al., 1985, 1987; McFadden et al., 1987; McFadden, 1988; Reheis et al., 1989, 1995; Chadwick and Davis, 1990). However, it has only been in the past couple of decades that many investigators have agreed that fine-grained aeolian inputs are important in deserts of the southwestern United States, and this conclusion, to a great extent, is based on soil studies.

In the Basin and Range province, many investigations have suggested that there were major fluxes of aeolian silt and clay shortly after the disappearance of lakes that were extensive in the region during Late Glacial time. Studies adjacent to Lake Lahontan (Chadwick and Davis, 1990) and smaller lakes in the Mojave Desert (Wells et al., 1985, 1987; McFadden et al., 1987; McFadden, 1988; Reheis et al., 1989, 1995) show that the abundance of silt, clay and carbonate in soils can best be explained by pulses of aeolian sediment input derived from newly exposed lacustrine sediments. The maximum extent of many of the lakes of the Basin and Range province is now known to date from Late Glacial (as opposed to full-glacial) time (Benson et al., 1990; Thompson et al., 1990; Oviatt et al., 1992). Thus, exposure of the sediments to deflation must have occurred either in latest glacial or early Holocene time. Furthermore, some lakes receded in stages and it is probable that there were multiple pulses of aeolian sediment as recession took place. For example, the lacustrine record from Lake Bonneville indicates that there could have been pulses of sediment after recession from the Bonneville, Provo and Gilbert high stands (going from oldest to youngest), all in latest Pleistocene time (Oviatt et al., 1992).

Detailed studies by Wells et al. (1985) and Reheis et al. (1995) have generated estimates of MARs of aeolian silt, clay and carbonate in the Mojave Desert portion of the Basin and Range province. Study of aeolian mantles on a $\sim$17,000 yr old volcanic flow and under an early Holocene alluvial fan in the Cima volcanic field in California yields MAR estimates of 50 g/m^2/yr (if material was deposited over the past $\sim$17,000 yr) to 800 g/m^2/yr (if deposited in a short period of $\sim$1000 yr after adjacent lake desiccation). A comprehensive study of several soil chronosequences and modern dust fluxes in the Mojave Desert by Reheis et al. (1995) has given the most detailed picture thus far of aeolian inputs to this region. These investigators also conclude that much of the silt, clay and carbonate in Basin and Range province soils is of aeolian origin, based on similarities in major element chemistry between soils and dust collected in traps. Furthermore, study of the abundance of aeolian silt, clay and carbonate in soil chronosequences shows that rates of dust flux varied through the late Quaternary. The last full-glacial period was a time of relatively low dust flux except during latest glacial time and into the early Holocene, when greater aridity and lake desiccation increased MARs. Modern MARs in the region (based on dust trap collections) range from 7 to 28 g/m^2/yr, based on the sum of aeolian silt, clay and carbonate. In those areas near former lakes, late Holocene MARs, calculated from the amount of silt, clay and carbonate in soils, range from 3 to 31 g/m^2/yr. In the same areas, latest Pleistocene to mid-Holocene rates range from 4 to 25 g/m^2/yr, but are, on average, higher than the late Holocene rates, reflecting the increased input from newly exposed lacustrine sources. Overall, however, MARs of the Basin and Range province are low compared to those of the Central Lowland, Great Plains, and Columbia Plateau provinces, in agreement with the modelling results of Mahowald et al. (1999).

6. Discussion

The MARs of Last Glacial loess varied regionally across the conterminous United States. The lowest calculated rates are from the Great Basin province where source availability and vegetation conditions did not favour significant dust production until the latest glacial and early Holocene. Loess in the Colorado Plateau accumulated during the Last Glacial period at rates higher than in the Great Basin, but at rates generally lower than loess in the Great Plains and Central Lowland. Great Plains loess is very extensive, and MARs in this region range from 3198 g/m^2/yr in thick sections in central Nebraska (Bignell Hill) to 104 g/m^2/yr in central Kansas (Barton Country), and decrease systematically to the south (Table 4). The highest Late Glacial MAR calculated for the conterminous United States (4216 g/m^2/yr) is from the Loveland paratype section along the Missouri River Valley in the Central Lowland province. These MARs are based on previously published radiocarbon dates (Table 3). It should be noted, however, that if the basal age of Last Glacial loess at thick sections in central Nebraska is younger than assumed in this paper, the highest Last Glacial loess MARs may be in the thick loess region of Nebraska. Recent OSL dating of Peoria Loess in Nebraska indicates that accumulation rates fluctuate by as much as an order of magnitude over the Last Glacial period, with extremely high average MARs ($>$10,000 g/m^2/yr) over the period from about 18,000 to 14,000 yr BP (Roberts et al., 2003). MARs calculated using OSL ages from the top and bottom of Peoria Loess sections at Bignell Hill and Eustis are approximately twice those calculated using radiocarbon dates from palaeosols that bracket the Peoria Loess. This suggests that MARs calculated using radiocarbon dates may be underestimating the true accumulation rate, in

addition to being insensitive to fluctuations in MARs during the Last Glacial period. Last Glacial MARs in the Central Lowland province were high compared to other areas in the United States, were highest near linear valley-train sources, and decreased systematically to the east/southeast from those sources.

Many physical, mineralogical, and chemical properties of the loess show distance-from-source decay functions that have been attributed to northwesterly loess-transporting winds during the Last Glacial period (Muhs and Bettis, 2000). These authors point out that a dominance of northwesterly winds is not consistent with the palaeo-wind direction under the glacial anticyclonic circulation predicted by AGCMs (Fig. 12) (e.g. COHMAP Members, 1988; Bartlein et al., 1998). Muhs and Bettis (2000) suggest that large-scale dust entrainment may have been restricted to periods when strong low-pressure systems passed through the region during infrequent northward shifts of the jet stream. Thus, some models fail to reconstruct the dominant wind fields and dust-transporting winds. As a result, these models may lead to significant underestimation of dust delivery to eastern North America, Greenland, and the North Atlantic.

MARs of Last Glacial loess in the Great Plains are two to three times higher than those of Holocene loess in central Nebraska, but the magnitude of the difference between Last Glacial and Holocene flux rates decreases southward into central Kansas. Although low compared to Last Glacial dust flux in other dust-producing regions of North America, the latest glacial and early Holocene dust fluxes in the Great Basin were, on average, higher than late Holocene rates. Holocene loess has not been recognized in the Central Lowland, but physical and chemical properties of modern soils in the northern part of the area suggest that dust has had significant impacts on the soils (Mason and Jacobs, 1998; Muhs et al., 2001).

The flux of dust from the North American mid-continent (Great Plains and Central Lowland) during the Last Glacial period has been underestimated by existing model simulations (e.g., Mahowald et al., 1999; Kohfeld and Harrison, 2000). The MARs of Last Glacial loess in the mid-continent are significantly higher than those documented for the Chinese Loess Plateau (Sun et al., 2000). A number of possibilities could account for this, including differences between the regions in silt availability, climatic/meteorological

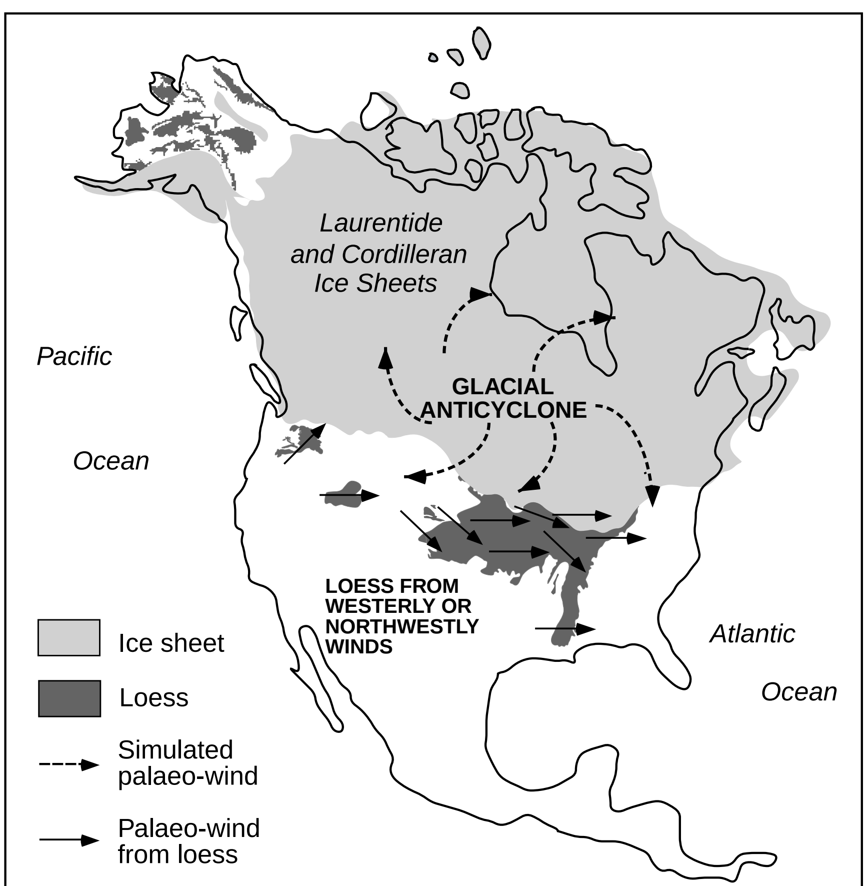

Fig. 12. Map of North America showing extent of the Laurentide Ice Sheet at the Last Glacial maximum and wind patterns derived from a modelled glacial anticyclone (COHMAP Members, 1988) compared to those derived from loess distributions. From Muhs and Bettis (2000).

conditions, trapping efficiencies related to vegetation and/or topography, and ice sheet effects. The non-glacial loess in both the Great Plains and in the Central Lowland (associated with the IES) is not represented in a recent linked AGCM-biome-atmospheric transport model simulation of Last Glacial dust sources (Mahowald et al., 1999; Kohfeld and Harrison, 2000). The amount of dust (loess) generated from these sources far exceeds that recorded in other North American source areas (e.g., southwestern USA, see Reheis et al. (1995) and Alaska, see Muhs et al. (this issue)) and simulated by the model of Mahowald et al. (1999). Also, available simulations do not accurately predict the magnitude of the dust flux from mid-continent valley-train loess sources. Data presented above show that valley trains were very significant loess sources over a large part of the central United States during the Last Glacial period.

It has been demonstrated that loess from different source areas, as well as loess generated by both glacial and non-glacial mechanisms, is present in the United States. Glacial loess sources include valleys carrying valley-train outwash, such as the Missouri, Mississippi, Wabash, and Ohio River Valleys in the Central Lowland, as well as slackwater basins influenced by glacial lake outburst floods, such as the Palouse region in the Columbia Plateau. Improved understanding of dust flux from these sources will require more refined chronologies for the various lobes of the Laurentide and Cordilleran ice sheets and a better understanding of silt transport pathways in glaciolacustrine environments around the ice sheet margin, and in proximal and distal glaciofluvial environments. Sites along the Mississippi River Valley record high MARs between about 25,000 and 20,600 cal yr BP with a significant decline thereafter, even though the meltwater input from the Laurentide Ice Sheet into the valley was probably greatest after 19,000 cal yr BP (Fairbanks, 1989; Teller, 1990). Silt trapping in extensive proglacial lakes that formed along the ice sheet margin during deglaciation (Teller, 1987, 1990; Johnson et al., 1999) may have decreased the amount of silt delivered to valley-train sources, thus lowering loess MARs after 20,000 cal yr BP. The thickest and most areally extensive belts of Last Glacial loess along the Missouri and Upper Mississippi River Valleys occur east and southeast of northwest-southeast-trending valley segments. This valley orientation afforded a long fetch that increased the effectiveness of northwest sediment-transporting winds during the Last Glacial period.

Bioclimatic factors may also have had impacts on the production and availability of silt in the Last Glacial landscape. A variety of physical evidence indicates that periglacial processes became active north of about 43°N, from the Rocky Mountains to the Appalachians during the full-glacial period, ca 24,000–19,000 cal yr BP (Péwé, 1983; Bettis and Kemmis, 1992; Walters, 1994; Mason and Knox, 1997). Ice wedge casts, solifluction deposits and thick colluvial deposits along valley margins date to this full-glacial period, and suggest that local mass wasting induced by cold conditions may have contributed to landscape instability and silt delivery to valley sources. Erosion surfaces that developed in the northern Great Plains and Central Lowland during the full-glacial period have a much thinner loess cover than adjacent landscapes where permafrost was discontinuous or absent. Pollen and plant macrofossil records indicate that boreal forest was replaced by a *Picea-Larix* krummholz with extensive tundra openings in the northern part of the Central Lowland at the onset of the full glacial about 24,000 cal yr BP (Baker et al., 1986, 1989; Gary et al., 1990). The dust (loess) and sand trapping efficiency of the tundra vegetation on the active erosion surfaces was probably lower than that of adjacent boreal forests. Thus, areas of sparse tundra vegetation were likely important sources of aeolian sediment production during the Last Glacial period. As the climate ameliorated about 19,000 cal yr BP, tundra was replaced by boreal forest (Webb et al., 1993; Schwert et al., 1997; Bartlein et al., 1998) and Late Glacial loess began to accumulate in some of these areas.

The Central Lowlands, IES and Great Plains became aeolian sediment sources during the Last Glacial period, probably as a result of vegetation cover reduction and surface disturbance under periglacial conditions. As mentioned earlier, existing models do not simulate conditions indicating that these areas acted as dust sources during the Last Glacial period (Mahowald et al., 1999; Kohfeld and Harrison, 2000). It is possible that the reduction of boreal forest as ice advanced into the upper mid-continent enhanced the cooling caused by glacial boundary conditions (Kohfeld and Harrison, 2000). This, in turn, produced a positive feedback that activated these loess source areas through vegetation reduction and enhanced mass wasting. Poor chronologic control in most areas limits our understanding of important linkages between climatic, biotic, and land surface characteristics in Last Glacial dust production and loess accumulation. A general lack of palaeoenvironmental information for the period of the onset of Last Glacial loess accumulation during full-glacial conditions in the Great Plains frustrates attempts at better understanding of the linkages between climate, vegetation, surface processes, and dust production across a significant portion of the mid-continent United States.

MARs in the North American mid-continent, Alaska (see Muhs et al., 2003) and the Chinese Loess Plateau are significantly higher than MARs of aeolian sediments in the world's oceans. Compilations of Last Glacial aeolian MARs derived from marine sediments have recently been published by Mahowald et al. (1999) and

Kohfeld and Harrison (2000, 2001). The highest rates, off the coast of Oman in the Arabian Sea, are 100–330 g/m^2/yr. However, rates in most other oceans are much lower. Aeolian sediments in the northwestern Pacific Ocean, just east of Japan, are presumably derived at least in part from the Chinese Loess Plateau, but have MARs of only 2–13 g/m^2/yr. Dust in the Tasman Sea, probably derived from Australia, has MARs of 0.6–8 g/m^2/yr. In the North Atlantic Ocean, dust derived from the Sahara and Sahel regions of Africa has been an important part of sediment flux to the western hemisphere for much of the Quaternary (Muhs et al., 1990). Last Glacial aeolian MARs for the North Atlantic Ocean range from about 5 g/m^2/yr near the equator to as much as 80 g/m^2/yr between about 20 and 25°N (Kohfeld and Harrison, 2001). Compared to the MARs for the interiors of the North American and Asian continents, these data suggest that the flux of sediment decreases significantly from the continents to the oceans. Fine-grained components of the sediment flux from the continents to the oceans may, however, have significant indirect effects on the atmosphere. Dust may act as a source of some limiting micronutrients, such as iron, and thus increase primary oceanic productivity; this, in turn, may draw down levels of atmospheric CO_2 (Harvey, 1988; Martin, 1990; Boyd et al., 2000; Watson et al., 2000). Loess from the central United States contains significant amounts of available iron (Allen, 1971; Ruhe and Olson, 1978; Miller et al., 1984), and downwind components of the loess could have had significant effects on the productivity of the North Atlantic Ocean during the Last Glacial period.

7. Conclusions

1. Last Glacial loess is widespread in the conterminous United States and is thicker, on average, than at most other localities in the world.
2. MARs for Last Glacial loess in the United States are higher than in other loess regions of the world.
3. MARs of Holocene loess are an order of magnitude lower than those of Last Glacial loess in the mid-continent, but are similar to, or slightly less than, Last Glacial rates in the Great Basin, Colorado Plateau and Columbia Plateau. Lower trapping efficiencies in the drier parts of the continent may be a source of significant underestimation of the Holocene dust flux.
4. Last Glacial loess occurs in a wide range of physiographic provinces across the United States, and was derived from both glacial and non-glacial sources.
5. Interactions between glaciers, climate, vegetation, and land surface processes controlled the rate of silt production, availability of silt for deflation, and loess accumulation.
6. Last Glacial loess in the mid-continent accumulated over a period of 10,000–14,000 years through the Last Glacial cycle. In the northwestern United States, Last Glacial loess accumulated after episodic glacial lake outburst events late in the glacial cycle and may have continued through the Holocene. In the Great Basin, loess accumulated at the end of the Last Glacial period and into the early Holocene when declining lake levels resulted in mineral grains being made available for transport.
7. At present, chronological control is inadequate for the evaluation of regional variations in loess MARs during the Last Glacial period across the United States. The existing chronology for the loess is based primarily on radiocarbon ages from SOM, plant remains and gastropod shells. A number of uncertainties limit the usefulness of existing luminescence ages on the loess, although new developments in OSL dating are beginning to provide high-resolution information of Last Glacial loess MARs.
8. Existing models of Last Glacial dust production fail to predict the location and magnitude of significant mid-continent United States loess sources. Loess sources in the Columbia Plateau, Great Plains, and Central Lowland produced an enormous volume of Last Glacial loess that could have resulted in significant downwind radiative and biogeochemical effects.

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