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Unprecedented last-glacial mass accumulation rates determined by luminescence dating of loess from western Nebraska

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Abstract

A high-resolution chronology for Peoria (last glacial period) Loess from three sites in Nebraska, midcontinental North America, is determined by applying optically stimulated luminescence (OSL) dating to 35–50 μm quartz. At Bignell Hill, Nebraska, an OSL age of 25,000 yr near the contact of Peoria Loess with the underlying Gilman Canyon Formation shows that dust accumulation occurred early during the last glacial maximum (LGM), whereas at Devil's Den and Eustis, Nebraska, basal OSL ages are significantly younger (18,000 and 21,000 yr, respectively). At all three localities, dust accumulation ended at some time after 14,000 yr ago. Mass accumulation rates (MARs) for western Nebraska, calculated using the OSL ages, are extremely high from 18,000 to 14,000 yr—much higher than those calculated for any other pre-Holocene location worldwide. These unprecedented MARs coincide with the timing of a mismatch between paleoenvironmental evidence from central North America, and the paleoclimate simulations from atmospheric global circulation models (AGCMs). We infer that the high atmospheric dust loading implied by these MARs may have played an important role, through radiative forcing, in maintaining a colder-than-present climate over central North America for several thousand years after summer insolation exceeded present-day values.

Keywords: Dust flux; Peoria Loess; Optically stimulated luminescence dating; OSL; Quartz; Coarse-silt-sized grains; Mass accumulation rate (MAR); Climate forcing

Introduction

Midcontinental North America has the thickest deposits of last-glacial loess in the world, referred to as Peoria Loess. These loess deposits contain important records of variations in atmospheric dust, providing evidence which may be used to assess the role of dust in climate change. On the basis of thickness alone, the accumulation rates for these deposits are expected to be high, but little is actually known of the fluctuation in dust accumulation rates both during the last glacial period (~70,000–10,000 yr ago), and spatially across North America. A reliable, high-resolution chronology is fundamental to any study of past dust accumulation. However, the existing chronology available for the loess deposits is based primarily on radiocarbon dating of paleosols that bracket all or most of the period of dust accumulation. Using the radiocarbon ages from paleosols gives only an average mass accumulation rate for the package of loess that they bracket, and any dust flux calculation based on these ages implicitly assumes that the accumulation rate has remained essentially constant over the period of deposition. In order to examine the temporal variation in dust accumulation, it is necessary to determine a chronology for the mineral grains that make up the loess deposit.

Luminescence dating is ideally suited to the investigation of the records of dust accumulation contained in loess as it is applied directly to the mineral component of the sediment

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itself. In contrast, radiocarbon dating must be applied to organic matter and relies upon sufficient material being preserved in the sediment. Very few luminescence ages have been published for midcontinental North America, and worldwide the majority of published luminescence ages for loess deposits are based on measurements of thermoluminescence (TL). However, since the event being dated is the last exposure of the sediment grains to light, optically stimulated luminescence (OSL-also referred to as optical dating (Aitken, 1998)) can provide higher precision ages for loess.

The chronologic information available for midcontinental North American loess deposits has been summarized by Bettis et al. (2003), who calculated mass accumulation rates (MARs) where possible, based largely on radiocarbon dating of paleosols. Eolian mass accumulation rates (MARs) can be calculated for the last glacial period based on data from loess deposits, dust in ice cores, and the eolian component of deep-sea sediments (see summaries in Kohfeld and Harrison, 2000, 2001; Mahowald et al., 1999; Sun et al., 2000). Bettis et al. (2003) found MARs of >3000 g m^{-2} yr^{-1} for midcontinental North America, a value that is high compared with those for most other continents e.g. >1000 g m^{-2} yr^{-1} for the Chinese Loess Plateau (Sun et al., 2000). The values calculated represent average accumulation rates for last-glacial dust deposition; if dust deposition rates varied over this time period, then specific MARs could actually be even higher. This study targets three of the thickest last-glacial loess deposits in the North American mid-continent, using high-resolution luminescence dating to determine a chronology for the loess, thus enabling the calculation of more accurate MARs, to examine the temporal variations in last-glacial dust flux.

Study area

Loess is areally the most important surficial deposit in central North America. Although loess is typically associated with areas close to rivers that drained the Laurentide Ice Sheet, it is also distributed widely over the central Great Plains of Nebraska, Kansas, and Colorado (Fig. 1) where much of it is derived from nonglacial sources (Aleinikoff et al., 1998, 1999). Three late Quaternary loess units, from oldest to youngest, the Gilman Canyon Formation, and the Peoria and Bignell Loesses, have been identified and correlated on the Great Plains. The Gilman Canyon Formation (dated to ~40,000–22,000 cal yr B.P., Johnson, 1993; Maat and Johnson, 1996; Martin, 1993; Muhs et al., 1999) is overlain by Peoria Loess. Peoria Loess was deposited during the late Wisconsin period and is the thickest (up to ~48 m) and areally most extensive North American loess unit. On the central Great Plains it is capped by a dark, organic-rich buried soil (the Brady Soil) separating it from the overlying Holocene Bignell Loess. In this study, three thick, last-glacial loess exposures in Nebraska were investigated (Fig. 1): Bignell Hill (48 m Peoria Loess, with ~25 m exposed); Devil's Den (30 m Peoria Loess exposed); and Eustis (16 m Peoria Loess exposed).

Methods

Sample collection and preparation

At each site, samples were taken for luminescence dating using thin-walled cylindrical corers driven into the face of the section, or by carving out loess blocks. Samples were taken from the Peoria Loess, close to the paleosols that bracket this deposit, and throughout the deposit itself.

In the laboratory, 13 samples were prepared for luminescence dating using standard procedures (e.g., Roberts and Wintle, 2001), except that the 35- to 50-μm fraction was isolated by wet sieving. This grain-size fraction was selected as being more appropriate than the 4- to 11-μm grain size typically used for dating loess deposits since the Peoria Loess in the central Great Plains contains abundant coarse silt (Swineford and Frye, 1951; Muhs and Bettis, 2000; Muhs et al., 1999).

The mineral selected for optically stimulated luminescence dating was quartz, as this avoids a type of signal instability called anomalous fading, which is associated with feldspars (e.g., Aitken, 1998). To remove feldspars and isolate pure quartz, the 35- to 50-μm fraction was treated for 7 days using fluorosilicic acid in a ratio of 40 ml 35% H_2SiF_6:1 g sediment (Berger et al., 1980), followed by a 45-min treatment with hydrochloric acid (37%). The H_2SiF_6-treated material was then resieved at 35 μm. This fluorosilicic acid treatment produced pure quartz of 35–50 μm in diameter, suitable for luminescence dating. This quartz fraction was then dry-sieved through 50-μm mesh onto 1-cm-diameter aluminum discs having a light cover of Silkospray.[1]

Luminescence measurement

The equivalent dose (D_e) was determined using a modified single-aliquot regenerative-dose (SAR) procedure (Banerjee et al., 2001; Roberts and Wintle, 2001) applied to quartz grains of 35–50 μm in diameter. This procedure involves using thermal pretreatments from 160 to 300°C for 10 s (to isolate stable OSL signals), and stimulation with an infrared (IR)-emitting laser diode at 830-nm wavelength to deplete the OSL signal from any feldspar remaining following chemical treatment, prior to stimulation with blue light-emitting diodes (LEDs) at 470 nm to obtain the [post-IR] OSL signal from quartz. This SAR procedure corrects for any luminescence sensitivity changes that may have occurred during burial, or caused by laboratory thermal pre-

[1] Use of trade names is for descriptive purposes only and does not imply endorsement by the U.S. Geological Survey.

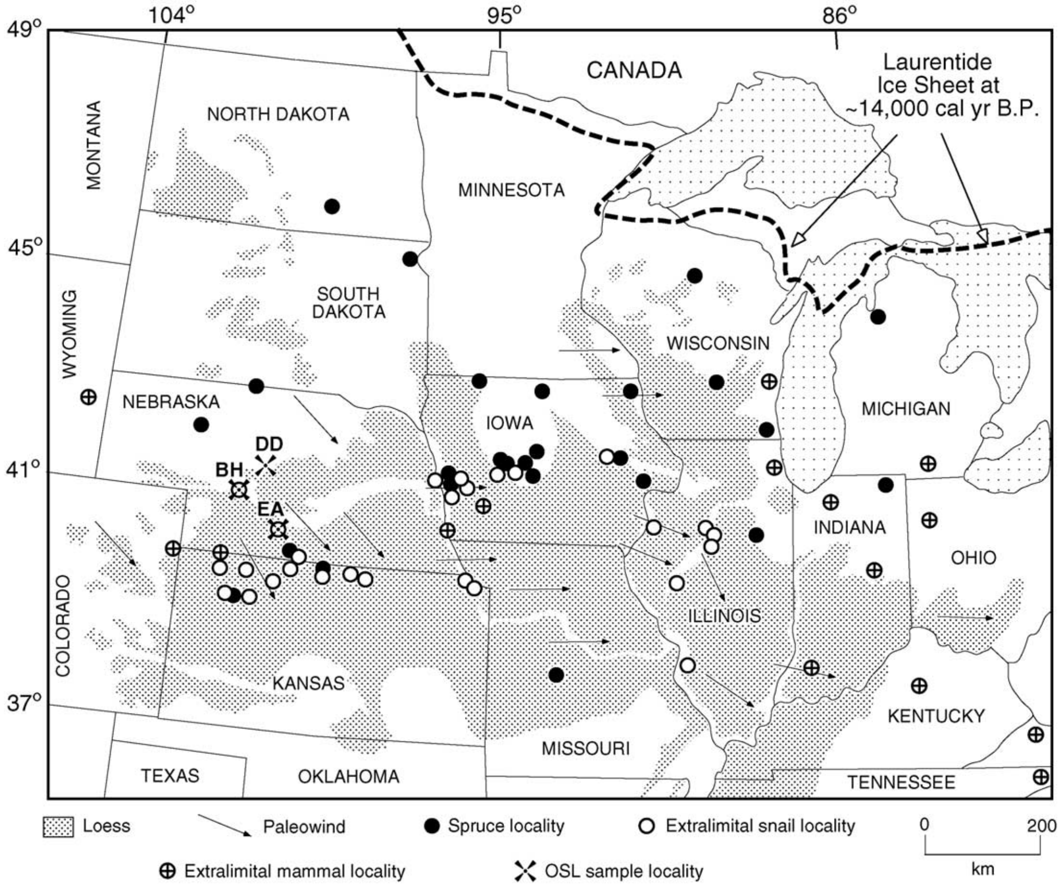

Fig. 1. Map showing the distribution of loess and inferred paleowinds in central North America and OSL sample localities (DD, Devil's Den; BH, Bignell Hill, EA, Eustis). Also shown for the late-glacial period are the estimated position of the Laurentide Ice Sheet (Licciardi et al., 1998), late-glacial localities with spruce macrofossils or pollen records dominated by spruce (Baker et al., 1992; Ruhe, 1969; Swinehart et al., 1994; Watts, 1983; Webb et al., 1983; Wells and Stewart, 1987), late-glacial loess localities with extralimital northern and western (i.e., cool climate) fossil land snails (Baker et al., 1986; Frye et al., 1974; Leonard, 1951; Leonard and Frye, 1960; Rousseau and Kukla, 1994; Ruhe, 1969; Wells and Stewart, 1987), and localities with extralimital northern mammals (FAUNMAP, 1994). Northern and western extralimital land snails include the species *Vertigo modesta, Columella alticola,* and *Pupilla blandi,* all of which do not occur in the central Great Plains region today, but have Cordilleran-boreal affinities (Lauriol et al., 2002). Extralimital northern mammals include the species *Sorex arcticus, Rangifer tarandus, Tamias minimus, Microtus xanthognathus, Phenacomys intermedius,* and *Synaptomys borealis,* all of whose present distributions are limited to tundra, boreal forest, or Cordilleran regions (FAUNMAP, 1994).

treatments. Luminescence measurements were made using an automated Risø TL/OSL reader, equipped with a beta source for irradiations, and blue LEDs (470 nm) providing approximately 17 mW cm^{-2} power density. The OSL measurements were made at 125°C and detected using three 3-mm Hoya U-340 filters (275–390 nm at 1% cut, based on manufacturer's specifications).

Dose–rate determinations

Both field and laboratory measurements of uranium, thorium, and potassium were conducted. In situ radioactivity measurements were made for each sample using a portable gamma spectrometer. In the laboratory, thick-source alpha counting (TSAC) and beta counting using Risø GM-25-5 equipment were undertaken on dried bulk powder. The cosmic-ray dose rate was estimated for each sample as a function of depth, altitude, and geomagnetic latitude (Prescott and Hutton, 1994). The water content was assessed in the laboratory using sealed field samples, taken both from the face of the exposure and at depth from deep coring. Alpha and beta contributions to the dose rate were corrected for grain-size attenuation (Aitken, 1985).

Chronology and dust deposition rates

The OSL ages of the three loess sections are consistent with previous age determinations (Fig. 2). For each of the 13 samples, a minimum of 17 aliquots (each aliquot giving rise to an independent determination of D_e) was used to determine the mean estimate of D_e used in the calculation of the final OSL age (Table 1). The uncertainty on D_e was calculated as the standard error of the mean. Within these uncertainties, the OSL ages are stratigraphically consistent (Fig. 2) and agree with previous radiocarbon ages for the

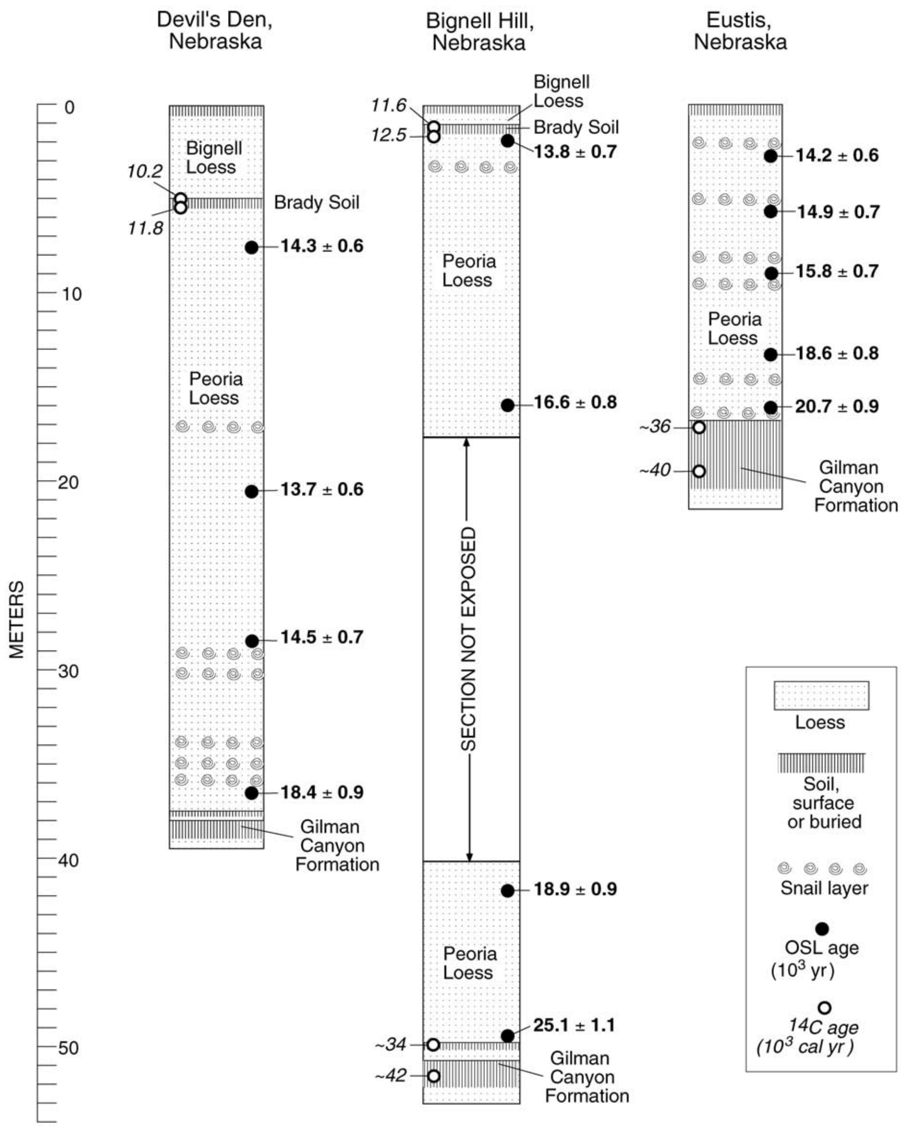

Fig. 2. Stratigraphy and OSL ages (on right side of each column) from the present study of three loess sections from the central Great Plains. Previously reported radiocarbon ages (Muhs et al., 1999; Maat and Johnson, 1996; Johnson and Willey, 2000) have been calibrated to "calendar" years (Stuiver et al., 1998; Voelker et al., 2000).

bracketing Brady Soil and Gilman Canyon Formation (Muhs et al., 1999; Maat and Johnson, 1996; Johnson and Willey, 2000). The OSL ages are also in agreement with previous TL ages (Maat and Johnson, 1996; Pye et al., 1995), but have much greater precision.

At Bignell Hill, an OSL age of 25,000 yr near the contact of Peoria Loess with the underlying Gilman Canyon Formation shows that dust accumulation began early in the last-glacial maximum. This is much earlier than is recorded at Eustis or at Devil's Den, where basal OSL ages are 21,000 and 18,000 yr, respectively (Fig. 2). The Brady Soil, which marks the end of Peoria Loess deposition and separates it from the overlying Holocene Bignell Loess, is not found at Eustis. However, at all three sites, the uppermost OSL ages show that Peoria Loess deposition effectively ceased shortly after ~14,000 yr ago (Fig. 2).

From Figure 2, it can be seen that the maximum dust accumulation occurred between ~18,000 to 14,000 yr ago We can calculate mass accumulation rates (MARs) for the loess deposits over this time interval, which enables comparison between the three sites. MARs were calculated based on the central values for the OSL ages and assuming a typical bulk density for Peoria Loess of 1.45 g cm^{-3} (Bettis et al., 2003). The average mass accumulation rates over the period from ~18,000 to 14,000 yr ago are extremely high, being ~ 11500 g m^{-2} yr^{-1} for Bignell Hill, ~ 10500 g m^{-2} yr^{-1} for Devil's Den, and ~ 3500 g m^{-2} yr^{-1} for Eustis. These values are one to two orders of magnitude higher than those calculated for last-glacial eolian silt in New Zealand and Europe where average values range from ~ 60 to 300 and 300 to 1400 g m^{-2} yr^{-1}, respectively (calculations are based on data from Antoine et al., 2001;

Table 1
Equivalent dose (D_e), dose rates, and luminescence ages

Sample No.[a]	Depth (m)	D_e (Gy)	n[b]	U (ppm)[c]	Th (ppm)[c]	K (%)[c]	Infinite α dose rate[d]	Infinite β dose rate[d]	External α dose rate "wet"[d]	External β dose rate "wet"[d]	External γ dose rate "wet"[d]	Cosmic[d]	Total dose rate[d]	Age (10^3 yr)[e]
*58*DD1	7.30	49.4 ± 0.6	21	3.11 ± 0.10	9.48 ± 0.3	1.95 ± 0.08	15.6	2.24	0.277	1.92	1.15	0.118	3.46 ± 0.14	14.3 ± 0.6[f]
*58*DD2	20.30	50.4 ± 0.6	20	3.36 ± 0.11	10.3 ± 0.3	2.11 ± 0.08	16.9	2.42	0.301	2.08	1.24	0.047	3.67 ± 0.15	13.7 ± 0.6
*58*DD3	28.50	50.1 ± 0.9	22	3.46 ± 0.11	10.1 ± 0.3	1.87 ± 0.07	17.0	2.24	0.303	1.93	1.19	0.031	3.45 ± 0.14	14.5 ± 0.7
*58*DD4	36.55	66.1 ± 1.5	23	3.26 ± 0.10	10.8 ± 0.3	2.03 ± 0.08	17.0	2.36	0.303	2.02	1.24	0.021	3.59 ± 0.15	18.4 ± 0.9[f]
*44*BH47	1.90	49.9 ± 1.0	37	2.94 ± 0.12	10.0 ± 0.4	2.03 ± 0.10	15.5	2.29	0.276	1.97	1.17	0.196	3.61 ± 0.15	13.8 ± 0.7[f]
*44*BH60	15.90	58.0 ± 0.9	38	3.18 ± 0.13	10.1 ± 0.4	1.97 ± 0.10	16.2	2.28	0.289	1.96	1.18	0.063	3.49 ± 0.15	16.6 ± 0.8
*44*BH46	41.65	70.7 ± 1.2	56	3.83 ± 0.14	10.8 ± 0.4	2.04 ± 0.10	18.56	2.45	0.330	2.10	1.30	0.017	3.75 ± 0.16	18.9 ± 0.9[f]
*44*BH37	49.50	87.6 ± 1.7	34	3.13 ± 0.09	11.8 ± 0.3	1.89 ± 0.07	17.4	2.26	0.309	1.94	1.24	0.013	3.50 ± 0.14	25.1 ± 1.1
*57*EA1	2.50	55.2 ± 0.8	40	3.47 ± 0.11	11.1 ± 0.3	2.09 ± 0.08	17.78	2.44	0.316	2.10	1.28	0.182	3.88 ± 0.16	14.2 ± 0.6[f]
*57*EA2	5.50	56.8 ± 1.2	19	3.55 ± 0.11	10.8 ± 0.3	2.07 ± 0.08	17.8	2.43	0.317	2.09	1.27	0.138	3.82 ± 0.15	14.9 ± 0.7
*57*EA3	9.00	58.9 ± 0.7	17	3.69 ± 0.12	10.6 ± 0.3	2.00 ± 0.08	18.0	2.39	0.320	2.05	1.26	0.102	3.74 ± 0.15	15.8 ± 0.7
*57*EA4	13.15	69.1 ± 1.1	20	3.46 ± 0.11	10.9 ± 0.3	2.05 ± 0.08	17.6	2.41	0.313	2.06	1.26	0.075	3.72 ± 0.15	18.6 ± 0.8[f]
*57*EA5	16.00	77.6 ± 1.4	46	3.36 ± 0.10	11.6 ± 0.4	2.06 ± 0.08	17.8	2.42	0.318	2.08	1.29	0.061	3.74 ± 0.15	20.7 ± 0.9

[a] Aberystwyth Luminescence Laboratory sample codes: *Aber/58*DD, Devil's Den; Nebraska; *Aber/44*BH, Bignell Hill, Nebraska; *Aber/57*EA, Eustis, Nebraska.

[b] n is the number of D_e determinations.

[c] Concentrations of U, Th, and K were determined from in situ measurements using a portable gamma spectrometer and are shown to 3 significant figures.

[d] Dose-rate values (Gy/10^3 yr) have been rounded to 3 significant figures, but the total dose rates and ages have been calculated using values prior to rounding. Dose rates were calculated assuming a water content (expressed as % dry mass) of 10 ± 5%, and using an a value of 0.040 ± 0.002 (Rees-Jones, 1995). Central values are given for dose rates—errors are incorporated into that given for the total dose rate.

[e] Luminescence ages are expressed as thousands of years before A.D. 2000, and calculated to 1 decimal place.

[f] Denotes the luminescence ages used to calculate mass accumulation rates (MARs) discussed in the text.

Eden et al., 1992; Hatté et al., 1998; Musson and Wintle, 1994; Nawrocki et al., 1999; Palmer and Pillans, 1996; Pillans et al., 1993; Sümegi and Rudner, 2001; Wintle et al., 1984). These mid-continental North American rates are even higher than those calculated for the Chinese Loess Plateau, where MARs range from 50 to $>$1000 g m^{-2} yr^{-1} (Sun et al., 2000).

The only other locality of which we are aware that has a MAR comparable to those we calculated for Nebraska is near the Delta River in central Alaska. At a single locality, Péwé (1975) reported a very high rate of dust accumulation during the Holocene, giving a MAR of 2900 g m^{-2} yr^{-1}. However, detailed mapping in this area shows that loess thicknesses diminish very rapidly to between approximately 1 and 3 m within 1–2 km downwind from the Delta River valley, the dust source (Lindholm et al., 1959; Péwé and Holmes, 1964). Thus, the high rate of dust accumulation in central Alaska is very localized. In contrast, the region of thick (10–40 m) loess in the Great Plains extends from eastern Colorado to western Iowa, an east–west distance of 500–600 km, with a north–south extent of 100–150 km (Hallberg et al., 1991; Swinehart et al., 1994; Muhs et al., 1999). Furthermore, detailed isotopic studies show that the dust of the Nebraskan loess deposits originates mainly from a nonglaciogenic source, the Tertiary White River Group siltstone, rather than strictly from the local river systems (Aleinikoff et al., 1998, 1999). This siltstone crops out in eastern Wyoming and Colorado, northern Nebraska and southern South Dakota, several hundred kilometers north of the study localities, suggesting that much of the dust is far traveled.

Discussion

The high-resolution OSL chronology established at each site allows the calculation of MARs throughout the 25,000–14,000 yr period. Thus, we can estimate variations in dust flux. Despite differences in the thicknesses of the Peoria Loess for the three sections examined in this study, the timing of the maximum MARs during the last glacial period occurs during the same time interval for all three sites. MARs reach unprecedented levels between ~18,000 and 14,000 yr B.P., with maximum MAR values being more than 50% higher than the average MARs. Luminescence dating of three loess sections in Nebraska shows that the main period of dust accumulation during the last-glacial period occurred from about 20,000 to 14,000 yr ago, with extremely high accumulation rates in the interval between about 18,000 and 14,000 yr ago.

This 4000 yr period of high dust flux coincides with the timing of one of the largest discrepancies between paleoenvironmental data for central North America and the simulated paleoclimate from atmospheric general circulation models (AGCMs) (Bartlein et al., 1998). In North America, the last glacial period was much cooler than present and

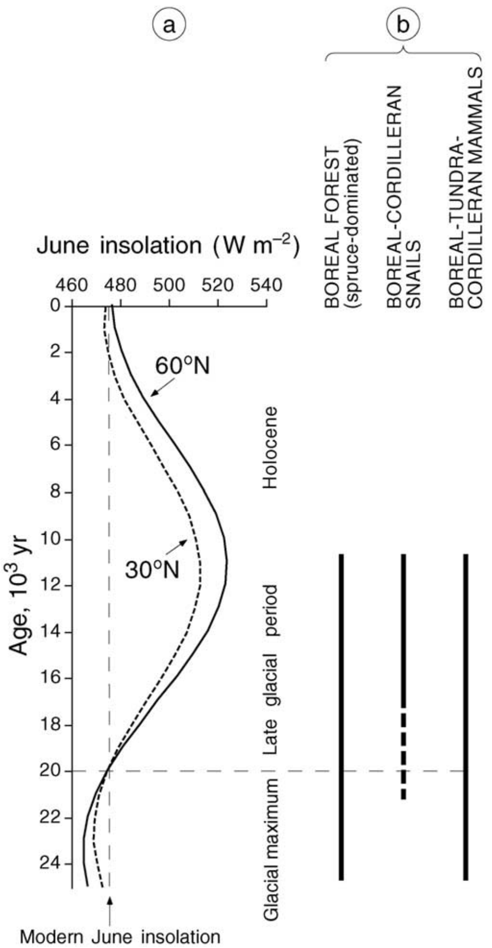

Fig. 3. (a) June insolation at the top of the atmosphere at 30°N and 60°N for the last 25,000 cal yr B.P. (Berger and Loutre, 1991), (b) times of colder-than-present climate inferred from the presence of spruce macrofossils or pollen, extralimital northern and western land snails, and extralimital northern mammals (data compiled from the following sources: Baker et al., 1986; 1989; 1992; FAUNMAP, 1994; Frye et al., 1974; Leonard, 1951; Leonard and Frye, 1960; Licciardi et al., 1998; Rousseau and Kukla, 1994; Ruhe, 1969; Schwert et al., 1997; Watts, 1983; Webb et al., 1983; Wells and Stewart, 1987; Wright, 1981).

culminated in the buildup of the Laurentide and Cordilleran ice sheets, as well as extensive alpine glaciers in mountainous regions. The last glacial maximum (LGM) coincided with the last minimum in northern hemisphere summer insolation around 24,000 to 22,000 cal yr (Fig. 3a). There is abundant evidence for a colder climate in the region to the south of the Laurentide Ice Sheet at that time. This evidence includes periglacial features (solifluction, ice-wedge casts, and patterned ground), and both vegetation communities (boreal forest, tundra, or cool steppe) and fauna (beetles, land snails, and mammals) that now occur only in higher latitude or higher altitude regions (Fig. 3b). However, during the late glacial period (~20,000 to 12,000 cal yr B.P.), summer insolation increased to values above those of today (Fig. 3a), yet many cold-climate-adapted plant and animal communities persisted through the period (Figs. 1 and 3b). These data conflict with the output of AGCMs that have consistently modeled the late-glacial period as being significantly warmer than present in central North America (e.g., Bartlein et al., 1998).

Several mechanisms could be offered to explain the discrepancy between the modeled scenario and the late-glacial paleoclimatic record of central North America. One explanation is that cool ocean waters from the LGM persisted and cooled the adjacent land masses; however, this is unlikely to have affected the study area because central North America is far from any ocean and today has an extremely continental setting. Advection of cool air off the Laurentide Ice Sheet would also have modified the local climate, but AGCMs which incorporate this process (Bartlein et al., 1998; Kutzbach et al., 1998) still fail to predict the magnitude of cooling found in the geological record. Furthermore, the loess record itself shows that winds off the Laurentide Ice Sheet were probably of minimal strength or were limited to a small area south of the ice sheet (Muhs and Bettis, 2000). Given the unprecedented dust accumulation rates identified in this study, a more likely explanation for the discrepancy may relate to the role of dust in atmospheric processes. Traditionally, high rates of dust (loess) deposition were believed to be only the result of climate change, specifically the shift to a dry, cold, windy glacial period with abundant sediment supply and little vegetation cover. Recent climate models have presented an expanded concept, suggesting that increased levels of atmospheric dust—however they are brought about—can also cause climate change (Overpeck et al., 1996; Tegen et al., 1996; Harrison et al., 2001).

We hypothesize that south of the Laurentide Ice Sheet, the effect of increased summer insolation in late-glacial time may have been offset by the cooling effects of a high dust flux. The effect of airborne dust on radiative transfer processes is a function of particle size, mineralogy, the height of the dust column, and the albedo of the surface over which the particles are transported. Over relatively high-albedo surfaces (e.g., ice sheets, snow-covered landscapes, dune fields, or light-colored bedrock), dust in transport has the effect of warming the lower atmosphere. In contrast, over relatively low-albedo surfaces (ocean, grassland, forest, tundra), radiative scatter from dust cools the lower atmosphere (Tegen et al., 1996). Modeling experiments show that in areas of high dust flux, such as over the Arabian Sea, solar radiation at the Earth's surface can be reduced by as much as 25 W m^{-2} (Harrison et al., 2001; Overpeck et al., 1996; Tegen et al., 1996). This amount of negative forcing is similar to the amplitude of summer-insolation variation since the LGM (Fig. 3a). At the LGM, landscapes south of the Laurentide Ice Sheet were probably covered by tundra, boreal forest, or boreal grassland (Baker et al., 1989; Muhs et al., 1999; Schwert et al., 1997; Wells and Stewart, 1987; Wright, 1981), all of which would have relatively low albedos in summer (McFadden and Ragotzkie, 1967). High atmospheric dust loading over this region may, therefore, be expected to result in cooling.

Particle size is a key variable in the amount of radiative forcing by airborne dust, as smaller particles are more effective in scattering radiation than are large particles. When modeling the effects of dust over the Arabian Sea (Tegen et al., 1996), particle diameters were typically <20 μm and much of the radiative scatter causing the reduction of insolation by 25 W m^{-2} came from particles <2 μm in diameter. Thus, when considering the role of dust in climate forcing for central North America, a simple comparison of MARs between distal and proximal sites would be misleading, because grain-size distribution needs to be taken into account. Both grain size and loess thickness change with distance from a source. In southeastern Nebraska, distant from the probable loess source, the <2-μm fraction typically is ~19% and the <20-μm fraction is ~43%. In contrast, at Bignell Hill and Devil's Den, closer to the source, these values are 7% (<2-μm fraction) and 21% (<20-μm fraction). Nevertheless, for Bignell Hill and Devil's Den, MARs recalculated using only the <20-μm fraction still give values higher than those reported anywhere else in the world, ~2000 g m^{-2} yr^{-1}.

Thus, through radiative forcing, dust may have been a major factor in atmospheric cooling across central North America during the latter part of the last glacial period. Dust-transporting paleowinds were from the west and northwest (Mason, 2001; Muhs and Bettis, 2000) across a low-albedo vegetated landscape and, therefore, a large region to the south of the Laurentide Ice Sheet could have been affected by radiative backscatter from dust in the atmosphere. Radiative backscatter could have offset the orbitally driven increase in summer insolation following the LGM. As a result, a colder-than-present climate could have been maintained over central North America for several thousand years after summer insolation exceeded present-day values. Our study permits the first direct calculation of stratigraphic variations in dust flux during the last glacial period, revealing unprecedented mass accumulation rates and highlighting the need for AGCMs to incorporate the effect of atmospheric dust loading at a regional scale.

Conclusions

Optically stimulated luminescence dating of quartz has been applied to three thick last-glacial loess deposits in western Nebraska, midcontinental North America, giving accurate and highly precise ages. These ages show that the onset of Peoria Loess deposition began early during the last glacial maximum at Bignell Hill (~25,000 yr ago), but later at Devil's Den (~18,000 yr ago) and Eustis (~21,000 yr ago). At all three sites, Peoria Loess deposition effectively ended some time after ~14,000 yr ago. Based on the OSL ages, average MARs for the last glacial period at all three sites are higher than those reported for any other pre-Holocene location in the world. Furthermore, within the sections, calculated MARs show that dust accumulation rates reached unprecedented values at all three sites between 18,000 and 14,000 yr ago. This period of extraordinary dust flux coincides with a mismatch between paleoenvironmental evidence (cool climate) and the output of AGCMs (warm climate). We hypothesize that the unprecedented dust fluxes may be responsible for climatic forcing in central North America at this time, maintaining a colder-than-present climate in the region for several thousand years after summer insolation values exceeded those of the present day. This study highlights the need for dust to be incorporated into climate models.

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