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Loess record of the Pleistocene–Holocene transition on the northern and central Great Plains, USA

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Abstract
Various lines of evidence support conflicting interpretations of the timing, abruptness, and nature of climate change in the Great Plains during the Pleistocene–Holocene transition. Loess deposits and paleosols on both the central and northern Great Plains provide a valuable record that can help address these issues. A synthesis of new and previously reported optical and radiocarbon ages indicates that the Brady Soil, which marks the boundary between late Pleistocene Peoria Loess and Holocene Bignell Loess, began forming after a reduction in the rate of Peoria Loess accumulation that most likely occurred between 13.5 and 15 cal ka. Brady Soil formation spanned all or part of the Bølling-Allerød episode (approximately 14.7–12.9 cal ka) and all of the Younger Dryas episode (12.9–11.5 cal ka) and extended at least 1000 years beyond the end of the Younger Dryas. The Brady Soil was buried by Bignell Loess sedimentation beginning around 10.5–9 cal ka, and continuing episodically through the Holocene. Evidence for a brief increase in loess influx during the Younger Dryas is noteworthy but very limited. Most late Quaternary loess accumulation in the central Great Plains was nonglacigenic and was under relatively direct climatic control. Thus, Brady Soil formation records climatic conditions that minimized eolian activity and allowed effective pedogenesis, probably through relatively high effective moisture. Optical dating of loess in North Dakota supports correlation of the Leonard Paleosol on the northern Great Plains with the Brady Soil. Thick loess in North Dakota was primarily derived from the Missouri River floodplain; thus, its stratigraphy may in part reflect glacial influence on the Missouri River. Nonetheless, the persistence of minimal loess accumulation and soil formation until 10 cal ka at our North Dakota study site is best explained by a prolonged interval of high effective moisture correlative with the conditions that favored Brady Soil formation. Burial of both the Brady Soil and the Leonard Paleosol by renewed loess influx probably represents eolian system response that occurred when gradual change toward a drier climate eventually crossed the threshold for eolian activity. Overall, the loess–paleosol sequences of the central and northern Great Plains record a broad peak of high effective moisture across the late Pleistocene to Holocene boundary, rather than well-defined climatic episodes corresponding to the Bølling-Allerød and Younger Dryas episodes in the North Atlantic region.

1. Introduction

Events during the last glacial termination in North America have intrigued Quaternary scientists and provoked controversy for decades. Between 15 and 10 cal ka, many large mammals became extinct, plant and animal assemblages with no modern analogs were widespread, and humans either first arrived in the Americas or became abundant enough to leave an extensive archaeological record (Overpeck et al., 1992; Graham et al., 1996; Williams et al., 2001; Grayson and Meltzer, 2003; Barnosky et al., 2004; Ugan and Byers, 2007; in this paper *cal ka* refers to age estimates in thousands of calendar years before the present). The role of climate change is an essential element in discussions of these events, often focusing on the role played by local expression of the large and rapid climatic changes marking the beginning and end of the Bølling-Allerød (B-A, approximately14.7–12.9 cal ka) and Younger Dryas (YD, 12.9–11.5 cal ka) episodes in the North Atlantic region (Stuiver et al., 1995; Björck et al., 1998). A new and controversial hypothesis also links both YD climate change and megafaunal extinction to an extraterrestrial impact (Firestone et al., 2007).

This paper considers whether loess–paleosol sequences can shed new light on climatic change during the last glacial–interglacial transition in the Great Plains, the setting for both a rich Pleistocene faunal record (Graham et al., 1996) and many well-known Paleoindian sites (Holliday, 2000a). One objective is to improve age constraints on the Brady Soil, a prominent paleosol that formed during the late Pleistocene and earliest Holocene in the central Great Plains, which should clarify its paleoclimatic significance and possible relations with hemispheric or global climatic changes associated with the B-A and YD. To do so, we synthesize many new and previously reported optical and radiocarbon ages, including new high-resolution dating at a well-studied loess section in southwestern Nebraska. A second goal is to explore the spatial extent of the paleoenvironmental conditions that favored Brady Soil development, through investigation of loess in North Dakota, on the northern Great Plains. The loess record on the northern Great Plains may also be helpful in achieving a third objective, which is to reconcile the environmental stability suggested by Brady Soil formation with climatic and vegetation change reconstructed from other evidence. In the northern Great Plains, high-resolution lake sediment records clearly indicate that major changes in climate, vegetation, and hydrology occurred as the Brady Soil developed farther south. Sparser proxy data from the central Plains support a similar conclusion, but comparison between northern Plains loess sections and nearby lake sediment records should provide new insight on this issue.

In the central Great Plains of Nebraska, Kansas, and eastern Colorado (Figure 1), recent work has emphasized the importance of the loess record in reconstructing late Quaternary environmental change (Muhs et al., 1999a; Johnson and Willey, 2000; Mason et al., 2003). Late Pleistocene and Holocene loess of this region, especially in Nebraska where it is thickest and most widespread, is primarily derived from dryland dust sources, with less important local input from glacially influenced river floodplains (Aleinikoff et al., 1999; Mason, 2001; Mason et al., 2003; Muhs et al., in press). Loess accumulation was most rapid immediately downwind of dunefields, which probably acted as conveyers of dust from more distant sources at times when the dunes were active (Mason, 2001). Discrete loess depositional units are separated by buried soils, interpreted as the result of slower deposition and/or more effective pedogenesis (Johnson and Willey, 2000; Miao et al., 2005). Thus, intervals of rapid loess deposition imply dune activity immediately upwind, and record climatic conditions that reduced vegetation cover both on the dunes and in more distant dust sources, while paleosols represent intervals of limited dune activity and dust production. This linkage has been confirmed for the Holocene by correlative ages from loess and dune sand (Miao et al., 2007b). In the Holocene, low effective moisture (precipitation relative to evaporation) is the most likely climatic cause of accelerated dust production (Miao et al., 2007b), but low temperatures and/or accelerated hillslope erosion supplying sediment to floodplain dust sources may have played an important role in the late Pleistocene (Muhs et al., 1999a).

The Brady Soil (Schultz and Stout, 1948), marking the boundary between late Pleistocene Peoria Loess and Holocene Bignell Loess, is one of the best-defined, yet in some respects most enigmatic, features of the central Great Plains loess record. The Brady Soil is characterized mainly by a thick, dark-colored A horizon and a well-structured Bk, Bw, or Bt horizon (Schultz and Stout, 1948; Johnson and Willey, 2000; Jacobs and Mason, 2004). Jacobs and Mason (2007) concluded that clay illuviation occurred during Brady Soil development in

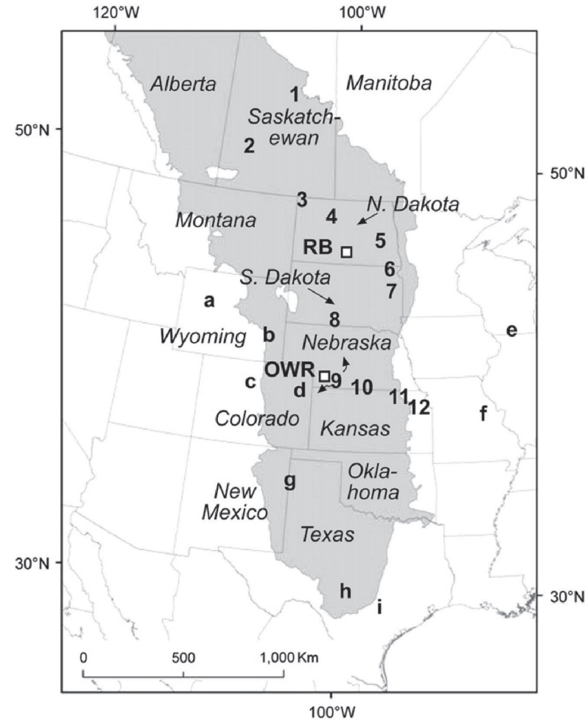

Figure 1. Study sites and other localities discussed in this paper. Gray shade represents Great Plains grassland ecoregions (Commission for Environmental Cooperation, 1997). RB = Rattlesnake Buttes, OWR = Old Wauneta Roadcut. Numbered sites are localities of studies on paleoecology or eolian stratigraphy of northern and central Great Plains: (1) Nisbet and Fort à la Corne dunefields (Wolfe et al., 2006); (2) Clearwater Lake (Last et al., 1998); (3) Kettle Lake (Clark et al., 2001); (4) Rice Lake (Laird et al., 1998); (5) Moon Lake (Valero-Garcés et al., 1997; Laird et al., 1998); (6) Pickerel Lake (Watts and Bright, 1968); (7) Medicine Lake (Radle et al., 1989); (8) Rosebud site (Watts and Wright, 1966); (9) southwestern Nebraska and northwestern Kansas Bignell Loess sections (Johnson and Willey, 2000; Mason et al., 2003; Miao et al., 2005); (10) Jensen, North Cove, and Cortland Canal sites (Fredlund and Tieszen, 1997); (11) Muscotah Marsh (Grüger, 1973); (12) DB loess section (Bozarth, 1998; Johnson and Willey, 2000). Letters indicate localities where evidence of Younger Dryas climate change and/or geomorphic response has been reported: (a) Wind River Range (Gosse et al., 1995); (b) slopewash aprons, Wheatland, Wyoming (Hanson et al., 2004); (c) Colorado Front Range (Menounos and Reasoner, 1997; Reasoner and Jodry, 2000); (d) Beecher Island loess section (Muhs et al., 1999a); (e) northern Illinois/southern Wisconsin lakes (Grimm and Maher, 2002); (f) Onondaga Cave (Denniston et al., 2001); (g) Southern High Plains, including dunes, playas, draws, and Folsom and Clovis archaeological sites (Haynes, 1991; Holliday, 2000b; Balakrishnan et al., 2005); (h) Edwards Plateau cave deposits (Toomey et al., 1993); (i) Medina River alluvium (Nordt et al., 2002).

central Nebraska, but not later in the Holocene. Formation of the Brady Soil has been attributed to climatic conditions, especially high effective moisture, that reduced dust influx and at the same time made pedogenic processes more effective (Johnson and Willey, 2000; Jacobs and Mason, 2004). Numerous radiocarbon ages of organic carbon in Brady Soil A or B horizons in Nebraska and Kansas fall largely between 11,800 and 7900 ^{14}C yr B.P., or ~13,800–8800 cal yr BP (Johnson, 1993; Johnson and Willey, 2000; Muhs et al., in press). The first increment of loess that buried the Brady Soil has yielded optical ages between 8.8 and 10.5 cal ka (Mason et al., 2003; Miao et al., 2005). These age constraints are broadly consistent with the suggestion of Forman et al. (1995), that an interval of relatively high moisture transport to the Great Plains and limited eolian activity occurred between 13,000 and 9000 ^{14}C yr B.P. (~15,600–10,200 cal yr BP).

Existing age control also suggests that major changes in dust production and effective moisture in the central Great Plains did not coincide with the beginning and end of the YD. Radiocarbon ages indicate that Brady Soil organic carbon accumulation had begun by the latter part of the B-A, and both radiocarbon and optical dating indicate that burial of this soil occurred more than 1000 yr after the end of the YD. This appears to conflict with the proposal by Haynes (1991) that a shift from widespread drought to a relatively wet climate occurred across large areas of the western USA, coinciding with both the change from Clovis to Folsom artifact types and the beginning of the YD. The loess record also offers little positive support for the alternative hypothesis that there was episodic drought during the YD, as inferred from increased dune activity at this time in the Southern High Plains (Figure 1; Holliday, 2000b). Radiocarbon ages do indicate an episode of increased loess influx during the YD at one site in eastern Colorado (Beecher Island, Figure 1; Muhs et al., 1999a).

The limited evidence for changes in dust production associated with the YD in the central Great Plains raises the possibility of significant spatial variation in the nature and magnitude of YD climate change across the central and western USA A variety of evidence indicates significant climate change during the YD in the Rocky Mountains and Midwestern USA, to the east and west of the central Great Plains (Figure 1; Gosse et al., 1995; Menounos and Reasoner, 1997; Reasoner and Jodry, 2000; Denniston et al., 2001; Grimm and Maher, 2002; Hanson et al., 2004). In the southern Great Plains, existing data provide a complex and sometimes contradictory picture of climate change during the YD. In south-central Texas and the Southern High Plains (Figure 1), δ^{13}C values from organic matter in buried soils and playa muds and aragonite in land snail shells indicate a large increase in C_4 plant abundance, interpreted as the result of rising temperature that peaked within the YD (Holliday, 2000b; Nordt et al., 2002; Balakrishnan et al., 2005). While Holliday (2000b) inferred frequent drought during the YD, vertebrate fossils in cave deposits on the Edwards Plateau (Figure 1), in the southernmost Great Plains, suggest a trend of increasing moisture between about 13,000 and 10,000 ^{14}C yr B.P. (~15.6–11.5 cal ka) (Toomey et al., 1993).

The presence of thick loess in parts of the northern Great Plains has been known for several decades, but its significance as a paleoenvironmental record remains largely unexplored. Clayton et al. (1976) described loess in North and South Dakota and assigned it to the Oahe Formation, which also includes alluvium. The stratigraphy of the Oahe Formation closely resembles the Peoria Loess–Bignell Loess sequence of the central Great Plains, with the prominent Leonard Paleosol possibly corresponding to the Brady Soil (Clayton et al., 1976; Mason et al., 2003). This correlation remains uncertain because of the limited number of published ^{14}C ages from Oahe Formation loess sections. McFaul et al. (2006) report ages obtained from charcoal indicating that Leonard Paleosol formation at one locality began by 11,186 ± 80 ^{14}C yr B.P. and continued until some time after 10,235 ± 76 ^{14}C yr B.P., within the range of Brady Soil ^{14}C ages.

In this paper, we report the first optical ages and stable carbon isotope data from loess of the Oahe Formation. Field observations of loess stratigraphy and distribution in North Dakota are also described, and are used to evaluate the relative importance of various loess sources, which may complicate paleoclimatic interpretation. Thick Oahe Formation loess occurs mainly along the Missouri River valley (Clayton et al., 1976), suggesting derivation from the floodplain of that river, where dust production could have been influenced by sediment and water input from the Laurentide Ice Sheet and alpine glaciers in the Rocky Mountains. On the other hand Clayton et al. (1976) also hypothesized that Holocene loess in the Oahe Formation might have been derived by local wind erosion of slopewash deposits.

Although it would be easy to interpret the Brady Soil as evidence of a period of climatic as well as geomorphic stability, a variety of data from the central and northern Great Plains indicate ongoing, significant change in climate, vegetation, and lake water salinity as the Brady Soil developed. The stable C isotope composition of organic matter extracted from the Brady Soil and loess just below it consistently records a distinct increase in the abundance of C_4 grasses, culminating in C_4 dominance at many sites just prior to Brady Soil burial (Johnson and Willey, 2000; Feggestad et al., 2004; Miao et al., 2007a). Muhs et al. (1999a, 1999b) report an increase in C_4-derived organic carbon at about the same time, in a northeastern Colorado loess section. A shift toward greater C_4 plant abundance is most plausibly explained as the result of rising temperature, which would give C_4 plants a competitive advantage despite rising atmospheric CO_2 (Ehleringer et al., 1997). Bozarth (1998) found an upward increase in phytoliths from warm-season (C_4) grasses within the Brady Soil at a site in northeastern Kansas (Figure 1, locality 12), but noted a short-term reversal toward greater abundance of cool-season (C_3) grasses, possibly correlative with the YD. A large increase in mean annual temperature from the late Pleistocene to the Holocene was also reconstructed by Fredlund and Tieszen (1997) from change in phytolith assemblages in the central Plains, using regression models calibrated with modern analogs (Figure 1, locality 10). Grüger (1973) interpreted the pollen record from Muscotah Marsh (Figure 1, locality 11) in eastern Kansas as indicating a transition from spruce (*Picea*) forest to deciduous forest and then to open prairie, within the period of Brady Soil development. Sparse age control limits detailed comparison with other sites, but spruce pollen decreases rapidly, accompanied by increasing deciduous tree pollen, just below a radiocarbon age of 11,340 ± 300 ^{14}C yr BP (12,787–13,830 cal yr BP). The inferred transition to open grassland is dated at 9930 ± 300 ^{14}C yr BP (10,577–12,402 cal yr BP), before Brady Soil burial at most localities.

Lake sediment records from the northern Great Plains even more clearly indicate increasing temperature and decreasing effective moisture through the time of Brady Soil formation (Figure 1, localities 2–8; Watts and Wright, 1966; Watts and Bright, 1968; Radle et al., 1989; Valero-Garcés et al., 1997; Laird et al., 1998; Last et al., 1998; Clark et al., 2001; Grimm, 2001; Yansa, 2006). This extensive body of research can be summarized as follows. Pollen and/or macrofossils of spruce in the oldest sediments at most sites indicate either spruce forest (Grimm, 2001) or more open spruce parkland (Yansa, 2006). A major time-transgressive, or at least spatially variable, decline in spruce abundance occurred between about 11,200 and 10,000 ^{14}C yr B.P. (~13,200 to 11,500–11,000 cal yr BP), which predates the renewed loess deposition that ended Brady Soil formation in the central Plains, by at least 1000 yr. Spruce decline was followed by an interval in which deciduous trees were present. Ultimately a shift to largely treeless grassland occurred, generally after 10,000 ^{14}C yr B.P. (~11,500–10,000 cal yr BP), accompanied by rising lake salinity inferred from diatoms or geochemical proxies.

2. Methods

Sampling sites were identified through reconnaissance field work, information provided by earlier investigators, and soil survey interpretation, and included roadcuts and other exposures as well as hand auger holes. Study sites in the central Great Plains include Old and New Wauneta Roadcuts (40° 29′ 59″ N, 101° 25′ 10″ W), Moran Canyon (41° 1′ 29″ N, 100° 21′ 22″ W), Bignell Hill (41° 2′ 28″ N, 100° 36′ 0″ W), County Line Ranch (41° 7′ 56″ N, 100° 14′ 42″ W), and Logan Roadcut (41° 28′ 41″ N, 100° 18′ 35″ W). All have been described in detail in earlier papers (Johnson and Willey, 2000; Mason et al., 2003; Jacobs and Mason, 2004; Miao et al., 2005, 2007a), but this paper reports new optical ages from directly below the Brady Soil at each site, along with new optical and ^{14}C ages obtained at approximately 10 cm intervals through the Brady Soil at the Old Wauneta Roadcut. New study sites in North Dakota are described in more detail below.

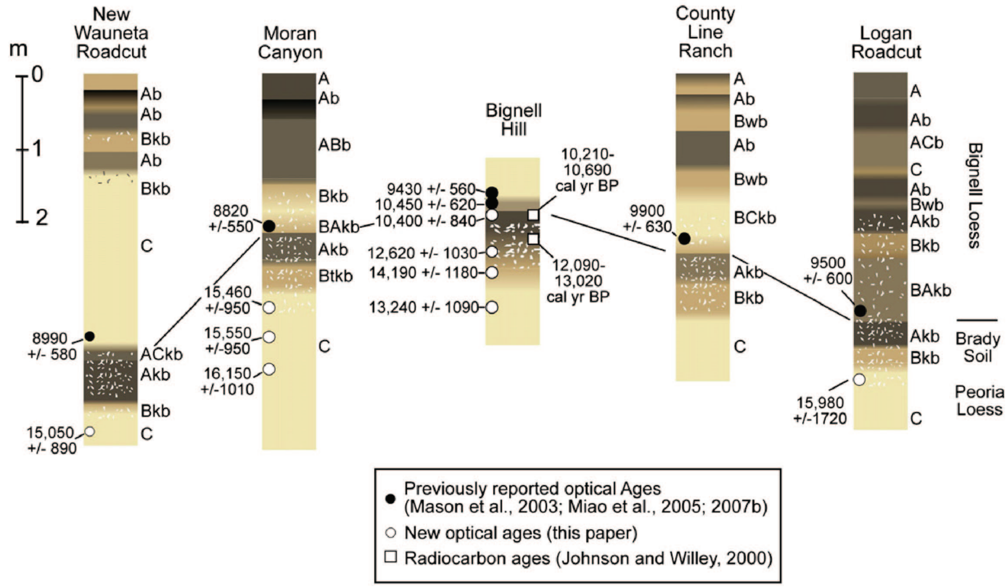

Figure 2. Stratigraphic columns representing five loess sections in southwestern Nebraska, with optical ages bracketing the Brady Soil. Shading represents color variation between darker buried soil horizons and lighter unaltered loess, with major pedologic horizons labeled at right of columns; white speckled pattern indicates secondary carbonates. Sections and their local landscape settings are described in Johnson and Willey (2000), Mason et al. (2003), Jacobs and Mason (2004), and Miao et al. (2005, 2007b)).

Samples for optical dating were collected by driving opaque tubes into cleaned exposure faces. Material for dose rate estimation was collected from within a 30 cm radius of the sample. All new optical ages reported here were determined at the University of Nebraska-Lincoln, using methods described in detail by Goble et al. (2004). In brief, the single-aliquot regeneration method (Murray and Wintle, 2000) was used to determine equivalent dose (D_e) values on Risø model DA 15 and DA 20 TL/OSL readers. In all cases we used quartz grains for dating; in most cases there was adequate 90–150 µm sand, but in some cases 30–90 µm grains were used. Final age estimates were based on a minimum of 17 aliquots, and aliquots were rejected if they had recycling ratios that exceeded 10% or if they showed evidence for contamination with minerals other than quartz. Dose rates were calculated using analyses of U, Th, K, and Rb as measured by ICP-MS. Moisture contents were based on measurements taken from bulk sediment samples taken adjacent to the optical dating tube. The cosmic ray contribution to the dose rate was estimated from burial depth, latitude, and elevation using equations from Prescott and Hutton (1994).

Samples for accelerator mass spectrometer (AMS) radiocarbon dating were collected <5 cm from the closely spaced optical dating samples in the Old Wauneta Roadcut, at the same depths, after cutting back the face to remove vertical contamination. AMS sample tubes were of the same diameter as the optical dating sample tubes. At the University of Kansas, AMS samples were picked for rootlets under magnification, dried, picked again, and pulverized prior to submission to the National Ocean Sciences Accelerator Mass Spectrometry Facility (NOSAMS) where carbonates were removed before dating. All ^{14}C ages were calibrated with Calib v. 5.02 (Stuiver and Reimer, 1993), using the IntCal04 calibration dataset (Reimer et al., 2004).

Samples for stable carbon isotope analysis of organic matter at the Rattlesnake Buttes section in North Dakota were collected as intact blocks from a roadcut face, after excavation to remove surface contamination. In the lab, small subsamples were obtained from the interior of freshly broken blocks, using ethanol and acid-washed tools and avoiding areas near rootlets, and placed in acid-washed glass vials. At the University of Kansas, subsamples were inspected under magnification for any remaining rootlets and were then treated with 1 M HCl at 23 °C for 16 h to remove carbonates, followed by drying at 50 °C and pulverization. Stable C analyses were conducted in the Keck Paleoenvironmental and Environmental Stable Isotope Laboratory, University of Kansas, using a ThermoFinnigan MAT 253 mass spectrometer. Results are expressed as $\delta^{13}C_{org}$ (‰) relative to the VPDB standard.

3. Results

3.1. Age of Brady Soil in the central Great Plains

Optical ages from rapidly deposited loess should represent the time since deposition. In soils, where the dated grains have a long residence time near the ground surface, an apparent age younger than the time of deposition may result if enough grains are exposed to light through some forms of bioturbation, such as rodent burrowing that mounds soil on the surface.

Figure 2 illustrates four stratigraphic sections in which optical ages bracket the time of Brady Soil formation. Dose rates are very similar in paleosols (mean ± one standard deviation = 3.32 ± 0.22 Gy kyr^{-1}) and relatively unaltered Peoria and Bignell Loess (3.41 ± 0.24 Gy kyr^{-1}) at these four sites and the Old Wauneta Roadcut (Table 1). Thus, pedogenesis does not appear to significantly change the dose rate over time, which would add uncertainty to age estimates.

The abundant radiocarbon and optical ages now associated with the Brady Soil at numerous sites are concordant, and they allow us to define the time of soil formation (Tables 1 & 2; Figure 3). Radiocarbon ages from Brady Soil organic carbon fall largely between the two sets of optical ages below and above the Brady Soil, which provide maximum and minimum ages for pedogenesis, respectively. Optical ages below the Brady Soil but still within 1 m of its upper boundary, range between 12.6 and 15.8 cal ka, but their ±2σ errors all include the interval between 13.5 and 14.7 cal ka (Figure 3). Thus, it is possible that the final increment of relatively rapid loess deposition prior to Brady Soil formation occurred within that time window at all sites, implying that Brady Soil formation began during the B-A. This in-

Table 1. Optical ages and supporting data

UNL lab. number	Field number	Depth (m)	U (ppm)	Th (ppm)	K_2O (wt%)	H_2O[a,b] (%)	Dose Rate (Gy ka^{-1})	D_e (Gy) ±1 Std. Err.	Aliquots[c] (n)	Optical Age ±1σ	Reference
Rattlesnake Buttes, ND											
UNL-1560	RB-6	0.2	2.0	8.2	1.8	3.8	2.65 ± 0.13	3.54 ± 0.18	25/33	1340 ± 110	This paper
UNL-1559	RB-5	0.4	1.9	7.9	1.7	3.4	2.54 ± 0.13	4.99 ± 0.31	26/31	1960 ± 180	This paper
UNL-937	RB-3	0.8	1.4	6.8	2.0	3.4	2.51 ± 0.13	7.06 ± 0.71	22/28	2810 ± 340	This paper
UNL-1558	RB-4	1.4	2.1	7.9	1.9	5.4	2.62 ± 0.14	13.38 ± 0.67	26/33	5110 ± 420	This paper
UNL-936	RB-2	2.3	1.9	10.0	2.0	5.2	2.79 ± 0.15	19.68 ± 1.62	27/28	7050 ± 750	This paper
UNL-1561	RB-7	2.5	2.6	10.1	1.9	2.7	2.97 ± 0.15	29.37 ± 0.87	26/27	9870 ± 710	This paper
UNL-935	RB-1	3.0	2.0	8.8	1.6	3.9	2.43 ± 0.12	29.85 ± 1.53	46/49	12,300 ± 1020	This paper
Old Wauneta Roadcut, NE											
UNL-543	WRT-A	5.9	2.8	12.4	2.4	7.5	3.43 ± 0.18	35.16 ± 0.45	23/25	10,250 ± 610	Miao et al. (2005)
UNL-1354	OWRB10	6.2	2.7	12.2	2.5	8.3	3.38 ± 0.19	34.62 ± 1.14	19/22	10,250 ± 720	This paper
UNL-1355	OWRB20	6.3	2.6	12.4	2.5	8.6	3.28 ± 0.18	34.29 ± 0.61	17/20	10,460 ± 680	This paper
UNL-1356	OWRB30	6.4	2.7	11.8	2.6	8.1	3.34 ± 0.18	35.88 ± 1.06	24/27	10,750 ± 740	This paper
UNL-1357	OWRB40	6.5	2.7	12.5	2.5	7.4	3.38 ± 0.18	40.69 ± 0.91	20/20	12,020 ± 790	This paper
UNL-1358	OWRB50	6.6	2.7	12.5	2.6	7.1	3.37 ± 0.18	40.95 ± 0.73	20/20	12,140 ± 780	This paper
UNL-1359	OWRB60	6.7	2.4	11.1	2.4	7.8	3.13 ± 0.17	42.14 ± 0.90	20/20	13,470 ± 890	This paper
UNL-1360	OWRB70	6.8	2.4	11.9	2.4	9.5	3.08 ± 0.17	45.00 ± 1.00	26/30	14,610 ± 980	This paper
UNL-1361	OWRB80	6.9	2.5	12.1	2.5	7.1	3.28 ± 0.17	49.65 ± 0.66	19/20	15,120 ± 960	This paper
UNL-1362	OWRB90	7.0	2.4	11.3	2.3	7.1	3.09 ± 0.17	48.84 ± 1.25	20/24	15,830 ± 1060	This paper
UNL-636	WRT-G	7.1	3.0	14.6	2.6	5.3	3.75 ± 0.20	56.34 ± 0.77	35/46	15,040 ± 900	This paper
UNL-637	WRT-H	7.3	3.0	14.4	2.6	7.6	3.58 ± 0.20	59.15 ± 0.82	30/36	16,540 ± 1020	This paper
New Wauneta Roadcut, NE											
UNL-424	WR-2	3.6	3.1	13.2	2.5	9.4	3.49 ± 0.20	31.41 ± 0.59	27/30	8990 ± 570	Mason et al. (2003)
UNL-423	WR-1	5.1	3.1	12.6	2.6	5.2	3.63 ± 0.19	54.67 ± 0.64	25/30	15,050 ± 890	This paper
Moran Canyon, NE											
UNL-525	5G01C-4	3.2	2.7	15.2	1.9	6.0[b]	3.19 ± 0.17	49.38 ± 0.81	21/28	15,460 ± 950	This paper
UNL-526	5G01C-5	3.5	2.6	14.0	2.0	6.0[b]	3.14 ± 0.16	48.84 ± 0.75	29/32	15,550 ± 950	This paper
UNL-527	5G01C-3	2.1	2.2	13.8	2.6	6.0[b]	3.51 ± 0.19	30.95 ± 0.44	19/22	8820 ± 550	Mason et al. (2003)
UNL-299	5G01C-6	4.0	2.7	14.0	1.9	6.0[b]	3.11 ± 0.16	50.30 ± 0.95	29/38	16,150 ± 1010	This paper
Bignell Hill, NE											
UNL-298	BN-6	2.0	2.9	12.6	2.7	5.9	3.69 ± 0.20	48.88 ± 0.94	34/35	13,240 ± 1090	This paper
UNL-294	BN-7	1.6	2.5	12.0	2.7	5.9	3.56 ± 0.19	50.58 ± 1.11	30/30	14,190 ± 1180	This paper
UNL-303	BN-8	1.2	2.8	13.4	2.6	3.9	3.76 ± 0.19	47.41 ± 0.87	30/30	12,620 ± 1030	This paper
UNL-295	BN-9	0.8	2.8	16.0	2.6	5.2	3.86 ± 0.20	40.12 ± 0.70	27/28	10,400 ± 840	This paper
UNL-296	BN-10	0.6	2.6	14.2	2.1	4.0	3.39 ± 0.17	35.40 ± 0.47	21/30	10,450 ± 620	Mason et al. (2003)
UNL-297	BN-11	0.4	2.7	14.0	1.9	4.0	3.29 ± 0.17	31.04 ± 0.48	24/30	9430 ± 560	Mason et al. (2003)
County Line Ranch, NE											
UNL-531	14G01H	2.4	2.0	8.8	2.3	5.8	2.96 ± 0.16	29.33 ± 0.48	21/27	9900 ± 630	Mason et al. (2003)
Logan Roadcut, NE											
UNL-478	LR-2	3.3	2.4	12.2	2.5	7.8	3.31 ± 0.18	31.49 ± 0.66	20/25	9500 ± 600[f]	Miao et al. (2007b)
UNL-477	LR-1	4.2	2.7	12.8	2.5	9.1	3.33 ± 0.19	53.20 ± 0.88	29/30	15,980 ± 1720	This paper

[a] Assumes 30% error in measurement.
[b] H_2O content based on average of loess samples taken at similar depths.
[c] Accepted disks/all disks.

terpretation is consistent with radiocarbon ages indicating that Brady Soil organic matter accumulation was underway at most sites by 14–12.5 cal ka (Figure 3C & D), optical ages reported by Roberts et al. (2003) indicating deposition of much of the upper part of Peoria Loess between about 15 and 14 cal ka at two sites in Nebraska, and the conclusion that Peoria Loess deposition in northeastern Colorado ended about 12,000 ^{14}C yr B.P. (Muhs et al., 1999a). Alternatively, the beginning of Brady Soil formation may have been time-transgressive across the study sites, ranging from just before the B-A into the early YD. This scenario is also consistent with the optical ages, although radiocarbon ages provide no positive support for it. Radiocarbon ages from the uppermost part of the Brady Soil A horizon, and optical ages from directly overlying loess, both indicate that the Brady Soil was buried by renewed loess sedimentation between 10.5 and 8.8 cal ka, 1000 yr or more after the end of the YD.

Average rates of sediment accumulation between optical ages bracketing the Brady Soil are 0.1–0.2 m kyr^{-1}, far lower than rates of more than 4 m kyr^{-1} estimated for thick proximal Peoria Loess from the data of Roberts et al. (2003) and substantially lower than the rates of 0.72–0.80 m kyr^{-1} estimated by Miao et al. (2005) for early- to middle-Holocene loess in the New and Old Wauneta Roadcuts (Figure 2). This observation

Table 2. AMS radiocarbon ages from soil organic carbon at Old Wauneta Roadcut

NOSAMS Lab. no.	Univ. of Kansas Sample no.	Depth (cm)	Age ^{14}C yr BP	δ^{13}C ‰
OS-61910	OWR 9	605	9370 ± 50	−16.64
OS-61954	OWR 1	620	8720 ± 50	−16.76
OS-61938	OWR 2	630	9450 ± 40	−16.22
OS-61915	OWR 3	641	9950 ± 50	−16.35
OS-61914	OWR 4	652	10,300 ± 60	−16.93
OS-61939	OWR 5	662	11,300 ± 45	−19.97
OS-61940	OWR 6	670	10,850 ± 65	−18.1
OS-61941	OWR 7	682	11,600 ± 60	−19.56
OS-61966	OWR 8	694	12,250 ± 45	−21.34

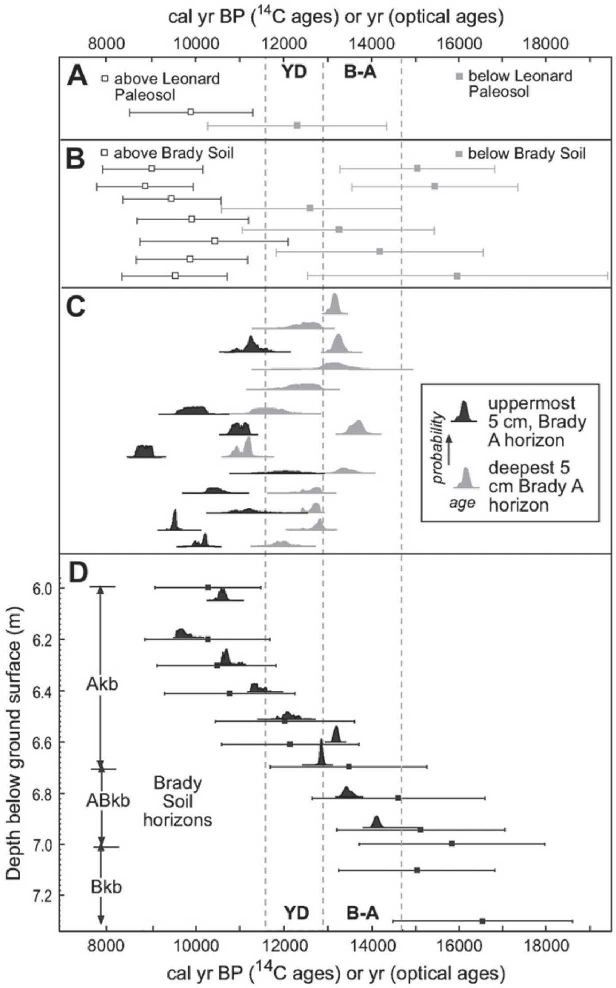

Figure 3. Summary of ^{14}C and optical age control for Brady Soil and Leonard Paleosol, relative to YD and B-A episodes. Optical ages are shown with $\pm 2\sigma$ errors, and ^{14}C ages are represented by small plots representing calibrated ages as probability distributions (see key at right of panel C). Distributions generated with Calib v. 5.02 (Stuiver and Reimer, 1993) using IntCal04 calibration dataset (Reimer et al., 2004). Ages are plotted relative to slightly different "zero ages", ~2002–2005 AD for optical, 1950 AD for ^{14}C, but offset is considered insignificant over this timescale. A. Optical ages bracketing Leonard Paleosol at Rattlesnake Buttes loess section, North Dakota (Figure 5). B. Optical ages bracketing Brady Soil, in Nebraska loess sections shown in Figure 2 (ages of two samples >1 m below top of Brady Soil A horizon at Moran Canyon not included). C. Calibrated ^{14}C ages from organic C in upper and lower Brady Soil A horizons at sites in Nebraska and Kansas; all ages from Table 1 of Johnson and Willey (2000). D. Vertical sequence of optical ages, and calibrated ^{14}C ages from soil organic C, through Brady Soil at Old Wauneta Roadcut (stratigraphy described in Miao et al., 2005); depth below modern ground surface and soil horizons indicated at left.

supports the interpretation of Brady Soil development as, at least partially, the result of a decrease in loess sedimentation rates that allowed more effective pedogenesis.

Closely spaced pairs of optical and calibrated ^{14}C ages through the Brady Soil at the Old Wauneta Roadcut (Figure 3D) follow quite similar age-depth trends to about 60 cm below the soil's upper boundary (670 cm total depth), while below that depth the optical ages are 1000–1500 years older than calibrated ^{14}C ages. In all cases, the ^{14}C ages fall within $\pm 2\sigma$ errors of the optical ages. These ages are consistent with two distinctly different interpretations. One interpretation is that the optical ages accurately represent the time since deposition, and that loess accumulation continued through the period of Brady Soil formation, though slowly and perhaps episodically. Soil organic carbon then accumulated in each increment of loess after it was deposited, resulting in approximately parallel depth trends of optical and ^{14}C ages. In theory, the ^{14}C age of soil organic carbon should be at least slightly younger than the depositional age of the parent material, and that difference should increase over time as long as a soil horizon remains in the zone of active organic matter addition and turnover (Wang et al., 1996). In the Brady Soil at OWR, any such age offset is apparently smaller than random error in dating; however, if the divergence of optical and ^{14}C age trends below about 670 cm is not simply random error, it may result from longer residence time of the lower Brady Soil in the zone of active carbon cycling. One implication of this interpretation is that there may have been a minor episode of renewed loess sedimentation during the YD, recorded by the two optical ages of 12.1 and 12.5 cal ka.

The second possible interpretation is that age-depth trends observed with both dating methods result from post-depositional soil processes. Even in a nonaggrading soil profile, ^{14}C ages of organic matter should increase with depth, because of slower turnover in deep horizons where young organic carbon input is limited and decomposition is slow (Wang et al., 1996). Highly effective, deep bioturbation might also yield a trend of increasing apparent optical age with depth in a soil. We are currently investigating additional evidence for and against such deep bioturbation in the Brady Soil, to distinguish between these interpretations.

3.2. Distribution of Oahe Formation loess in North Dakota

We are not aware of any reported thick sections of Oahe Formation loess containing the Leonard Paleosol that are more than 30 km from the Missouri River valley, and none were found in this study. Extensive exposures of relatively thick loess are present near some segments of the Missouri valley. As reported by Clayton et al. (1976), exposures of silt overlying glacial sediments or bedrock, and containing the Leonard Paleosol, extend for long distances along the wave-cut bluffs around the eastern part of Lake Sakakawea, a large reservoir on the Missouri River (Figure 4A). Upstream of Lake Sakakawea, we found loess 1.4 m thick, with the Leonard Paleosol from 0.6 to 0.8 m, on the east bluff of the Yellowstone River, a few km above its confluence with the Missouri (Rau site; 47° 54′ 0″ N, 103° 55′ 32″ W; Figure 4A).

Many Oahe Formation exposures also occur downstream from Lake Sakakawea, along wave-eroded terrace scarps in the Missouri River valley, near the upper end of the Lake Oahe reservoir (Figures 4 & 5). Here, loess appears to form a blanket of relatively uniform thickness mantling the largely undissected terraces, and also appears to be much thicker on the east side of the river than on the west side (Figures 4 & 5). Along these terrace scarps, the Leonard Paleosol is clearly identifiable, but underlying loess of the Mallard Island member is very thin or absent. The Leonard Paleosol often directly overlies thin fluvial sand and gravel, which in turn rests on a bedrock strath (Figure 5). This is in contrast to upland sections along Lake Sakakawea and at Rattlesnake Buttes (discussed below), where a significant thickness of loess occurs below the Leonard Paleosol (Figure 5). We interpret the stratigraphy in these terrace sections as indicating that the loess-capped terrace was the Missouri River floodplain while the lower, Late Pleistocene part of the Oahe Formation was accumulating on uplands. The Leonard Paleosol then formed soon after the Missouri River began incising to form the present terrace, followed by loess accumulation on the terrace through much of the Holocene.

On the uplands east of these loess-mantled terraces, soil surveys indicate that loess is patchy but widely distributed on gently sloping uplands, or in swales where it may have been

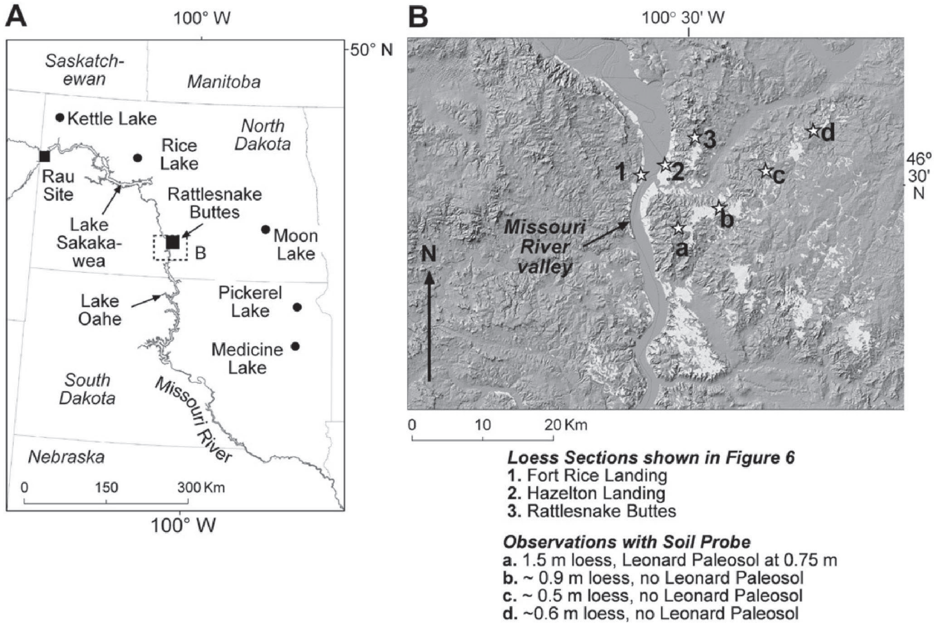

Figure 4. Distribution of Oahe Formation loess in North Dakota, and setting of the Rattlesnake Buttes loess section, North Dakota. A. Regional setting, including Missouri River and major reservoirs (Sakakawea and Oahe), lakes used for paleoecological studies (references in Figure 1 caption), and northwesternmost known Oahe Formation locality (Rau site). B. Shaded relief map of south-central North Dakota (small dashed box in A). Light gray shade indicates soils formed in loess ≥about 1.5 m thick, from National Resource Conservation Service SSURGO data; stars indicate loess sections shown in Figure 5 or observations with hand soil probe.

reworked by slopewash (Figure 4). Areas that lack loess cover are often characterized by steeper slopes and greater dissection. Soil probing across the uplands east of the upper end of Lake Oahe confirmed the presence of loess but the Leonard Paleosol was not distinguishable except at one site near the Missouri River (Figure 4).

In western North Dakota, many upland soils have silty upper horizons that may be loess (noted by Clayton et al., 1976, and supported by our field observations), but the Leonard Paleosol is not identifiable. Areas around small dunefields in North Dakota were also investigated using exposures and hand augering. We found no thick loess at the dunefield margins, even where a north or northwest facing scarp overlooks the dunefield, an especially favorable location for thick Holocene loess in the central Great Plains (Mason et al., 2003).

3.3. Optical ages and stable carbon isotope profile from Oahe Formation loess

We applied optical dating to an upland section of Oahe Formation loess in south-central North Dakota, USA. The Rattlesnake Buttes section (46° 33′ 39″ N, 100° 29′ 17″ W; Figures 4 & 5) is a roadcut exposing silt that contains multiple buried soils, overlying pre-Quaternary bedrock, on the dissected uplands

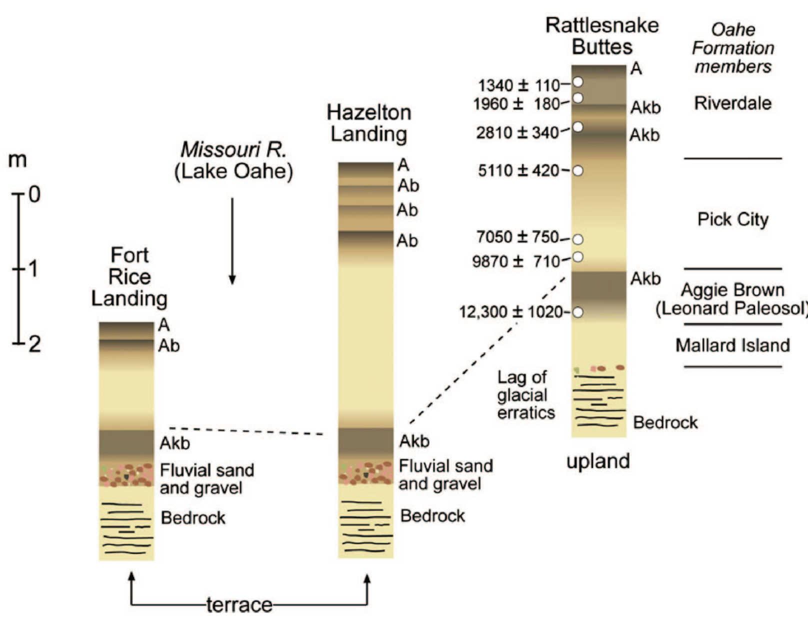

Figure 5. Stratigraphic columns representing loess sections spanning the Missouri River valley, including two terrace sections and upland Rattlesnake Buttes section, with optical ages (locations in Figure 4). Shading represents soil horizons, as in Figure 2, with only buried A horizons labeled; members of Oahe Formation (Clayton et al., 1976) labeled at right.

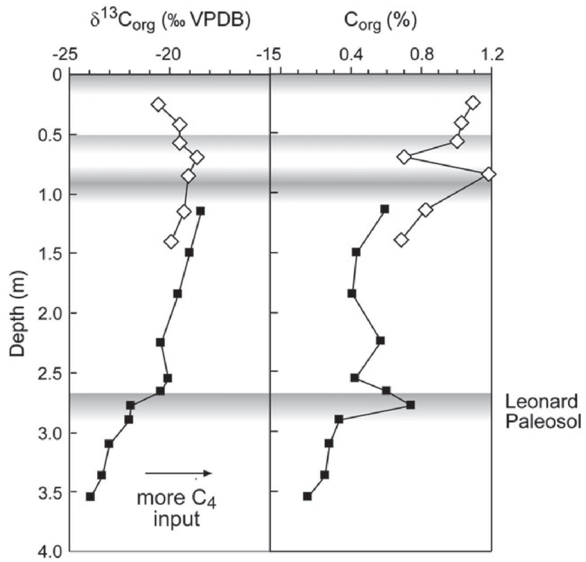

Figure 6. Stable carbon isotope and organic carbon data from within profiles at Rattlesnake Buttes section. Squares and diamonds represent two vertical sequences of samples, overlapping but offset about 4 m laterally along roadcut face. Gray bars mark most distinct surface and buried A horizons within loess section.

east of upper Lake Oahe. The exposure is in a saddle between two higher bedrock buttes, but the silt drapes over local high points in this saddle, where the only plausible mode of deposition is through dustfall. The section is 4 km east of the Holocene Missouri River floodplain, above any identifiable stream terrace. It lies just outside an ice margin that apparently has not been directly dated but is generally assigned to about 20–21 cal ka (Clayton et al., 1980; Mickelson et al., 1983; Dyke and Prest, 1987). The site was glaciated, possibly in the early Wisconsin (Clayton et al., 1980), but the only remaining glacial sediment is a lag of erratics overlying bedrock.

The optical ages support earlier correlations of the Oahe Formation with the loess sequence of the central Great Plains. The basal Mallard Island member at Rattlesnake Buttes is late Pleistocene, similar in age to upper Peoria Loess of the central Plains. Optical ages bracketing the Leonard Paleosol at Rattlesnake Buttes are similar to optical and radiocarbon ages constraining the time of Brady Soil formation on the central Plains (Figure 3). The age of 12.3 cal ka below the Leonard Paleosol is at least slightly younger than any optical age near the base of the Brady Soil, but there is a large overlap between $\pm 2\sigma$ errors of all of these ages. The optical age that dates Leonard Paleosol burial by Holocene loess at Rattlesnake Buttes agrees closely with ages constraining Brady Soil burial, with burial of both soils occurring well after the end of the YD. The lighter-colored Pick City member of the Oahe Formation is similar in age to rapidly deposited early- to middle-Holocene Bignell Loess (Zone 2 of Mason et al., 2003), while the darker-colored and more pedogenically altered Riverdale member was deposited in the late Holocene, like the upper portion of the Bignell Loess (Zone 3, Mason et al., 2003). Ages from Rattlesnake Buttes suggest that late Holocene episodes of rapid loess deposition at that site occurred at different times than those identified in the central Great Plains (Miao et al., 2007b), but more ages from Oahe Formation sections are needed to test this possibility.

Stable carbon isotope analysis of organic matter from the Rattlesnake Buttes section (Figure 6) yielded a profile similar to those obtained from loess sections in the central Great Plains, but with a less pronounced shift toward less negative $\delta^{13}C_{org}$ values across the Pleistocene to Holocene transition. We have not dated organic carbon from this section, but it is reasonable to interpret the change in $\delta^{13}C_{org}$ upward through the Leonard Paleosol as an approximation of the trend in $\delta^{13}C_{org}$ of added organic matter over time, since the mean residence time of carbon tends to become older with depth in most soils (Wang et al., 1996) even without ongoing upbuilding of the soil by sedimentation. In the late Pleistocene loess below the Leonard Paleosol, $\delta^{13}C_{org}$ values record nearly complete dominance of plants using the C_3 photosynthetic pathway, including trees, shrubs, and cool-season grasses. Increased input from C_4 plants, almost entirely warm-season grasses in the Great Plains, is indicated by rising $\delta^{13}C_{org}$ values toward the upper boundary of the Leonard Paleosol.

These interpretations assume that most of the organic C in these samples was added after loess deposition, which is reasonable given the relatively high organic C content of all horizons above the Mallard Island member (Figures 5 & 6). Contamination of bulk lake sediment ^{14}C ages by Paleocene and Eocene lignites or Cretaceous black shale is known to occur in the northern Great Plains (Nambudiri et al., 1980; Grimm and Jacobson, 2004). Harrington et al. (2005) report $\delta^{13}C_{org}$ values averaging −24.5‰ (standard deviation = 1.2‰, n = 53) from Paleocene–Eocene lignites, carbonaceous shales, and mudstones in western North Dakota, and Cretaceous marine shales generally yield $\delta^{13}C_{org}$ values more negative than −24‰ (Arthur et al., 1985). Thus, any minor contamination from these pre-Quaternary carbon sources could have slightly reduced the apparent late Pleistocene to Holocene increase in C_4-derived organic carbon.

4. Discussion

In the central Great Plains, the Peoria Loess–Brady Soil–Bignell Loess sequence represents the full range of variation in the loess system, between rapid dust accumulation at times when dust sources and dunefields were active, to soil development at times of low dust influx and effective pedogenesis. New age control reported here reinforces the conclusion that the loess system remained near one end of this range, with low dust influx and Brady Soil formation, over much of an extended interval that probably started within the B-A and ended well after the YD. As discussed above, we believe it is most likely that this interval began between 13.5 and 14.7 cal ka (or 13.5–15 cal ka in round numbers), consistent with maximum ages for Brady Soil formation, radiocarbon ages of organic C within that soil, and other evidence. Time-transgressive stabilization over a wider timespan is also plausible, however. Ages from Old Wauneta Roadcut suggesting a brief renewal of loess influx during the YD are intriguing, but at the most represent minor dust source reactivation. At Rattlesnake Buttes in the northern Great Plains, changes in loess sedimentation rate and the effectiveness of pedogenesis occurred at about the same time, and in the same direction, as in the central Great Plains.

Interpretation of these results is complicated by both the presence of quite different loess sources in the central and northern Plains, and the nature of eolian system response to climatic change. Loess in North Dakota probably had different sources, and thus may have more complex controls, than loess of the central Great Plains. Although more detailed information on loess grain size and thickness trends is needed, available evidence points to the Missouri River floodplain as the major loess source in North Dakota. Thick loess appears to occur only near the Missouri valley. The greater spatial extent of relatively thick loess on the east side of the Missouri River in south-central North Dakota is similar to the asymmetric distribution of loess thickness across north–south river valley sources in the Midwest (Handy, 1976; Fehrenbacher et al., 1986). Other, more local, sources of dust have apparently produced thin loess mantles on stable upland surfaces in semiarid western North Dakota, but not thick deposits containing the Leonard Paleosol. Loess or clifftop eolian deposits that are of local origin, and not derived from large river floodplains, have been documented nearby in Saskatchewan (David, 1970; Vreeken, 1994) and South Dakota (Rawling and Fredlund, 2003).

The Missouri River floodplain represents a dust source subject to controls that did not affect nonglacial dryland loess sources in the central Great Plains. Most importantly, the Missouri River was located at or near the southwestern margin of the Laurentide Ice Sheet during the last glaciation. Dust production from the Missouri River valley at that time probably resulted from an extensive, sparsely vegetated floodplain, as observed today along rivers affected by high glacigenic sediment loads and highly variable, meltwater-dominated discharge (Trainer, 1961; McGowan et al., 1996; Muhs et al., 2004). The ice sheet readvanced to a margin draining to the Missouri River in South Dakota and probably also upstream in North Dakota, around 12,000 to 12,600 ^{14}C yr B.P., or about 14,100 to 14,500 cal yr BP (Mickelson et al., 1983; Dyke and Prest, 1987). This advance was followed by rapid retreat out of the Missouri River basin before 13 cal ka (Dyke and Prest, 1987). This retreat just preceded the onset of Leonard Paleosol formation at Rattlesnake Buttes (Figure 5). Assuming that the Leonard Paleosol is about the same age on nearby terraces (Figure 5), ice sheet retreat after 14.1–14.5 cal ka may also have resulted in a large reduction in sediment load that triggered incision of the Missouri River and produced the terraces. A lower sediment load and more stable discharge of the Missouri River, after loss of glacial input, could in turn have lowered dust production rates and allowed Leonard Paleosol development.

The accelerated loess influx that buried the Leonard Paleosol at Rattlesnake Buttes around 10 cal ka cannot be attributed to glacial influence, however, since only small alpine glaciers in the Rocky Mountains remained in the Missouri River basin by that time (Mickelson et al., 1983; Dyke and Prest, 1987). The thick Holocene loess along the upper Missouri River valley provides strong evidence that the Missouri River floodplain can be a major dust source even when glacial influence is minimal. Thus, although the specific time at which loess accumulation slowed to allow Leonard Paleosol formation may have been determined by ice sheet retreat, persistence of low loess influx and effective pedogenesis until about 10 cal ka almost certainly reflects more direct climatic control. The most important factor was probably effective moisture. Individual lake sediment records from the northern Plains indicate low salinity before about 9–10 cal ka, along with abundant pollen influx from tree species that were largely absent from the Plains under the drier conditions of the Holocene (Watts and Bright, 1968; Valero-Garcés et al., 1997; Laird et al., 1998; Last et al., 1998; Clark et al., 2001; Grimm, 2001; Yansa, 2006). The same paleoclimatic proxy data indicate that the climate had become much warmer and drier, by the time loess accumulation at Rattlesnake Buttes buried the Leonard Paleosol in the early Holocene. A transition from high effective moisture on the Great Plains at about 11 cal ka to dry conditions in the early Holocene is also evident in a continental-scale synthesis of lake-level data by Shuman et al. (2002).

We speculate that two interacting mechanisms might link effective moisture with the availability (*sensu* Kocurek and Lancaster, 1999) of wind-erodible sediment unprotected by vegetation, on the Missouri River floodplain. First, the Missouri River probably had more variable discharge and was supplied with more sediment when effective moisture was relatively low and the climate was semiarid across much of its basin (Langbein and Schumm, 1958; Walling and Webb, 1996; Molnar et al., 2006). Second, revegetation of areas scoured or buried by sediment during floods may have been slower in a dry climate, leaving sediment exposed to wind erosion for longer periods of time. A possible modern analog is provided by the Cimarron River in western Kansas, where a large flood in 1914 produced a wide barren channel belt that did not begin to narrow again until after the dry years of the 1930s (Schumm and Lichty, 1963).

Based upon this interpretation, both the Brady Soil and the Leonard Paleosol represent an interval of relatively high effective moisture and eolian system stability, extending at least from western Kansas to North Dakota, and lasting several thousand years. The extent to which this stable interval was interrupted by eolian activity during the YD remains an open question. To our knowledge, the only directly dated evidence of a discrete episode of eolian activity in the central Great Plains during the YD, bounded by intervals of soil formation, comes from the Beecher Island loess section (Figure 1) in northeastern Colorado (Muhs et al., 1999a). Three of the optical ages reported here from below the Brady Soil, and one optical age of 13.1 cal ka from dune sand in the Nebraska Sand Hills (Goble et al., 2004), also fall within or just before the early YD. There is, however, no evidence of a preceding interval of reduced loess influx or soil formation at the sites where those ages were obtained. Other suggestions of eolian activity in the central Plains at that time are based on poorly dated and/or indirect evidence. For example, Loope et al. (1995) suggested dune activity in western Nebraska around 11,000 ^{14}C yr B.P., based on a minimum age of 10,600 ^{14}C yr B.P. from peat overlying eolian sand. Madole (1995) noted the occurrence of Clovis and Folsom artifacts within eolian sand in northeastern Colorado, and inferred that eolian activity was underway at about 11,000 ^{14}C yr B.P. (end of Clovis occupation), ending before 9010 ^{14}C yr B.P. (a limiting minimum soil age). In the northern Great Plains, however, Wolfe et al. (2006) identified an interval of dune activity at about 11 cal ka, in central Saskatchewan dunefields about 600 km north of Rattlesnake Buttes (Figure 1). The sites in the Southern High Plains where Holliday (2000b) found evidence of YD drought and dune activity are a similar distance south of Bignell Loess sections in Nebraska.

Overall, the available evidence indicates that eolian activity during the YD in the central Great Plains did not affect large areas and/or was short-lived. In interpreting this observation, it is important to consider the likelihood that eolian systems display a highly nonlinear response to climatic change, with little initial response followed by rapid increases in both dust production and dune activity after a threshold is crossed. This view is supported by the contrasting response of various Great Plains dunefields to historic droughts such as the 1930s "Dust Bowl," with some dunes remaining mostly stable while others in a slightly drier setting became fully active (Muhs and Maat, 1993). Dust production also increased dramatically in historic droughts, although this partly reflected human land use and may not be strictly analogous to prehistoric dry periods. The YD in the central Plains may have been characterized by effective moisture low enough to trigger occasional activity but not crossing the threshold for a large, spatially extensive response. Brief or intermittent dune activity might leave a poorly preserved stratigraphic record in the dunefields, and the increment of loess resulting from a brief increase in dust production could have been completely incorporated into the Brady Soil A horizon. Late Holocene episodes of eolian activity lasting only a few hundred years resulted in clearly identifiable depositional units in Bignell Loess, however (Miao et al., 2007b), and sustained dust production over much of the YD should probably have left a recognizable stratigraphic record.

Glacial advances and a lower treeline during the YD in the Rockies (Gosse et al., 1995; Menounos and Reasoner, 1997; Reasoner and Jodry, 2000) indicate lower temperatures, which might also help explain limited eolian activity nearby in the central Plains: even a large decrease in precipitation might have been offset by lower evapotranspiration in a colder climate. On the other hand, Muhs et al. (1999a) noted that $\delta^{13}C_{org}$ values from loess sections provide no evidence of YD cooling on the plains of northeastern Colorado.

The assumption of a threshold response for dust production can also help address the final objective of this paper, which is to explain why the loess record appears to record a relatively rapid shift toward a drier climate and much greater dust influx around 10.5–9 cal ka, while other proxy data often indicate gradual but significant environmental change well before that time. The increase in loess sedimentation rate at 10.5–9 cal ka need not imply particularly large or abrupt climatic change *at that time*. Instead it may reflect the point at which a longer-term trend toward warmer and drier conditions finally crossed the threshold for a large increase in dust production. From this perspective, it is not surprising that increased dust influx did not bury the Brady Soil/Leonard Paleosol until after C_4 grasses became dominant on the central Plains (Bozarth, 1998; Johnson and Willey, 2000; Feggestad et al., 2004) and well after the decline of spruce in North Dakota (Grimm and Jacobson, 2004; Yansa, 2006).

In addition, the well-defined upper boundaries of the Brady Soil and Leonard Paleosol, often marked by a distinct upward shift to lower organic matter and lighter color (Miao et al., 2007a), may to some extent exaggerate the actual change in loess sedimentation rate at 9–10.5 cal ka. Both pedologic evidence and optical ages from thick Bignell Loess sections in Nebraska indicate that the loess accumulation rate increased substantially from 10 cal ka to about 8 cal ka (Jacobs and Mason, 2004; Miao et al., 2005), consistent with a continuing trend of decreasing effective moisture and increasing dust production. The limited age control now available (Figure 5) suggests a similar scenario for Holocene loess sedimentation in North Dakota.

Overall, the loess record of the central and northern Great Plains indicates that the Pleistocene–Holocene transition was marked by a peak of effective moisture and eolian system stability across a large portion of the Great Plains between about 15 cal ka and 10 cal ka, as proposed by Forman et al. (1995). In our interpretation, this peak of effective moisture gradually gave way to drier conditions in the early Holocene. To the extent that a YD oscillation was experienced on the central Great Plains, it apparently was superimposed on this broader sequence of climatic change.

Explaining the climatic conditions that minimized eolian activity and allowed Brady Soil and Leonard Paleosol development could be an excellent target for future paleoclimatic modeling studies, given the regionally consistent nature of the loess record. Forman et al. (1995) attributed wet conditions during the Pleistocene–Holocene transition to enhanced monsoonal precipitation resulting from enhanced land–sea temperature contrasts, related to both high summer insolation (peaking at 11–10 cal ka) and meltwater-cooled sea-surface temperature in the Gulf of Mexico. The well-known set of experiments by the COHMAP group, using the NCAR CCM1 general circulation model, do not indicate relatively high effective moisture in the central and northern Great Plains at 11 and 14 cal ka. Instead, simulated summer precipitation is somewhat lower than modern, and coupled with higher-than-modern summer temperature due to higher insolation, this results in lower effective moisture than at present (Bartlein et al., 1998).

Several recent modeling studies have focused on a related problem, the explanation of a *dry* climate in the North American interior during the mid-Holocene. Proposed mechanisms linking high summer insolation in the mid-Holocene to aridity in the Great Plains include a "dipole" effect between an enhanced North American monsoon and subsidence creating dry conditions on the Great Plains (Harrison et al., 2003), enhanced anticyclonic circulation over the midcontinent (Diffenbaugh et al., 2006), and the combined effects of a northward shift in the Intertropical Convergence Zone and cold, "La Niña-like" sea-surface temperatures in the eastern tropical Pacific during the early- to mid-Holocene (Moy et al., 2002; Shin et al., 2006). Our interpretation of the loess record implies that none of these mechanisms generated dry conditions at the actual peak of northern summer insolation at 10–11 cal ka, perhaps because of different boundary conditions, including the continued presence of the Laurentide Ice Sheet and/or different sea-surface temperatures than in the early Holocene.

5. Conclusion

Across North America, most late Pleistocene megafaunal extinctions, the disappearance of non-analog vegetation, and the appearance of humans making Clovis-style tools all occurred as loess sedimentation in the central Great Plains slowed and was replaced by Brady Soil development (Overpeck et al., 1992; Graham et al., 1996; Williams et al., 2001; Grayson and Meltzer, 2003; Barnosky et al., 2004; Ugan and Byers, 2007). The Brady Soil represents a broad peak of effective moisture and eolian system stability beginning about 15–13.5 cal ka and ending around 9–10.5 cal ka. Evidence that this stability was interrupted in some areas by dust production and dune activity during the YD is intriguing but quite limited. Additional age control is needed on the Leonard Paleosol of the northern Great Plains, but ages reported here support its correlation with the Brady Soil and suggest that high effective moisture during the Pleistocene to Holocene transition extended at least from Kansas to North Dakota.

These conclusions are directly relevant to understanding the environment in which the Clovis-style tool makers and later Paleoindians lived, and should also be considered in evaluating hypotheses on why the megafauna did not survive on the Great Plains. More importantly, the loess record and other paleoenvironmental data from the Plains represent an important test for hypotheses on changes in atmospheric circulation and precipitation. A satisfactory explanation for high effective moisture in parts of the Great Plains during the period of Brady Soil and Leonard Paleosol formation is likely to provide insight on mechanisms of climate change in that critical time period over much larger areas of North America.

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