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# Optical ages on loess derived from outwash surfaces constrain the advance of the Laurentide Ice Sheet out of the Lake Superior Basin, USA

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ABSTRACT

We present textural and thickness data on loess from 125 upland sites in west-central Wisconsin, which confirm that most of this loess was derived from the sandy outwash surfaces of the Chippewa River and its tributaries, which drained the Chippewa Lobe of the Laurentide front during the Wisconsin glaciation (MIS 2). On bedrock uplands southeast of the widest outwash surfaces in the Chippewa River valley, this loess attains thicknesses >5 m. OSL ages on this loess constrain the advance of the Laurentide ice from the Lake Superior basin and into west-central Wisconsin, at which time its meltwater started flowing down the Chippewa drainage. The oldest MAR OSL age, 23.8 ka, from basal loess on bedrock, agrees with the established, but otherwise weakly constrained, regional glacial chronology. Basal ages from four other sites range from 13.2 to 18.5 ka, pointing to the likelihood that these sites remained geomorphically unstable and did not accumulate loess until considerably later in the loess depositional interval. Other OSL ages from this loess, taken higher in the stratigraphic column but below the depth of pedoturbation, range to nearly 13 ka, suggesting that the Chippewa River valley may have remained a loess source for several millennia.

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## Introduction

Loess deposits help constrain the timing of the geomorphic events that led to their formation (Roberts et al., 2003; Sweeney et al., 2007), particularly the paleoclimatic conditions that promoted their production (Muhs and Bettis, 2000; Muhs et al., 2008). Optical dating is now well established as viable means of establishing the age of loess deposits worldwide (Forman et al., 1992; Singhvi et al., 2001; Forman and Pierson, 2002; Bettis et al., 2003; Roberts et al., 2003; Timar-Gabor et al., 2011; Brown and Forman, 2012). A key advance that has recently emerged, however, is the ability to link loess deposits to small- and regional-scale glacial deposits and events, e.g., active outwash plains, thawing end moraines, or recently drained proglacial lakes (Schaetzl and Loope, 2008; Stanley and Schaetzl, 2011; Luehmann et al., 2013; Schaetzl and Attig, 2013).

In this context, successful dating of loess can help constrain the timing of the related glacial activity, and the goal of our study is to provide one of the first examples of just such an application: the timing of the advance of the Laurentide Ice Sheet in Wisconsin. Considerable debate exists as to the dynamics and synchroneity of the late Wisconsin (MIS 2) ice margin in the central United States (Attig et al., 1985; Eschman and Mickelson, 1986; Carson et al., 2012; Kehew et al., 2012). For example, the Des Moines Lobe advanced several millenia later than many other Midwestern ice lobes (Patterson, 1997). This example stands in contrast with the Saginaw Lobe in Michigan, which stagnated early, allowing its ice and debris to be overridden by bounding lobes (Kehew et al., 2005, 2012). In Wisconsin, this debate is particularly difficult to resolve, because several major ice lobes flowed into the region, each with a unique history, bed topography and bed conditions. These circumstances dramatically affected flow rates and directions, as well as stagnation and possible streaming (Mickelson et al., 1983; Clayton et al., 1985; Attig et al., 1989; Clark, 1992; Lundqvist et al., 1993; Colgan and Mickelson, 1997; Patterson, 1998; Cutler et al., 2001; Bauder et al., 2005; Syverson and Colgan, 2011).

Unfortunately, because of widespread permafrost near the last glacial maximum (LGM) ice margin in Wisconsin, and consequently the lack of forests and buried wood, carbon-rich materials for $^{14}C$ dating are difficult to obtain in the glacial deposits (Holmes and Syverson, 1997). Thus, ice margin dynamics and chronology within the upper Midwest USA are generally poorly constrained (Clayton et al., 2001), with a few exceptions, and these generally occur later in the deglacial sequence (Blewett et al., 1993; Kaiser, 1994; Larson et al., 1994). Fortunately, other geochronometric dating methods have the potential to constrain the timing of the ice advances in the Midwest (Colgan et al., 2002; Schaetzl and Forman, 2008; Attig et al., 2011b; Ullman et al.,

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2011; Carson et al., 2012). Many of these methods are now being applied to inform the debate on the apparent asynchroneity of some of the major Midwestern ice lobes.

Our research goal was to better constrain the timing of the advance of the Chippewa Lobe in west-central Wisconsin (Fig. 1), through luminescence dating of eolian deposits (loess) that were likely derived from its outwash. Meltwater from the Chippewa Lobe of the Laurentide Ice Sheet flowed through the study area, to the Mississippi River, beginning when the ice advanced into the Chippewa River basin and continuing until it had receded back into the Lake Superior basin. During this interval, including when the ice was at its maximum extent, we hypothesize that silt was being deflated from the broad, sandy, outwash deposits, i.e., valley train, of the Chippewa River and its tributaries. Preliminary fieldwork confirmed that thick loess occurs on most uplands near this meltwater system. We dated the basal loess from five such sites using luminescence techniques, with the goal of using these ages to constrain the advance of the ice, southward into the Chippewa drainage basin, and thus provide minimum-limiting ages for advance of the ice into the basin. Although absolute ages on this terminal moraine are emerging for sites in southern Wisconsin (Attig et al., 2011b; Ullman et al., 2011; Carson et al., 2012), our data provide the first age control for the ice in western and west-central Wisconsin, and for the Chippewa Lobe in particular.

## Study area

### *Quaternary history*

In Wisconsin, the Chippewa River and its tributaries (Figs. 1, 2) were one of the major meltwater systems draining the southern margin of the Laurentide Ice Sheet. This system began functioning as meltwater drainageways when the southern margin of the ice sheet crossed the drainage divide marking the southern margin of the Superior basin, and flowed into the northern part of the Chippewa River drainage. The Chippewa River system carried meltwater throughout the advance of the Chippewa Lobe to its maximum extent, ceasing only when the Chippewa Valley Lobe again receded north of the divide. Although the glacial geomorphology of this landscape is reasonably well understood (Syverson, 2007; Syverson and Colgan, 2011), the local glacial chronology is poorly constrained (Clayton and Moran, 1982; Mickelson et al., 1983; Attig et al., 1985, 2011a) because of the lack of closely controlling radiocarbon dates. Regional correlations indicate that the southern margin of the Laurentide Ice Sheet probably advanced southward out of the Superior basin by about 30,000 years ago, reached its maximum extent in the Chippewa River lowland prior to about 22,000 years ago, and that its margin had receded back into the Lake Superior basin by about 17,000 years ago. Dates that could substantiate or refine this general chronology are lacking.

Loess deposits are widespread, although not continuous, throughout Wisconsin (Scull and Schaetzl, 2011). Loess on the uplands near the Mississippi River valley is thick and was likely derived from the valley, while it functioned as a major meltwater drainageway (Leigh and Knox, 1993; Bettis et al., 2003). Loess deposits far from this valley are thinner and spatially disjunct (Scull and Schaetzl, 2011), suggesting that they were derived not from the Mississippi River valley proper, but from other, often more localized, source areas. For example, Stanley and Schaetzl (2011) concluded that the late Wisconsin moraine in central Wisconsin, with its abundant ice-walled lake plains, was a major loess source for the thin loess deposits to its immediate south. Schaetzl and Attig (2013) were able to link the loess deposits covering the drumlins of northeastern Wisconsin to outwash plains on either side of the drumlin field. Luehmann et al. (2013) took this type of

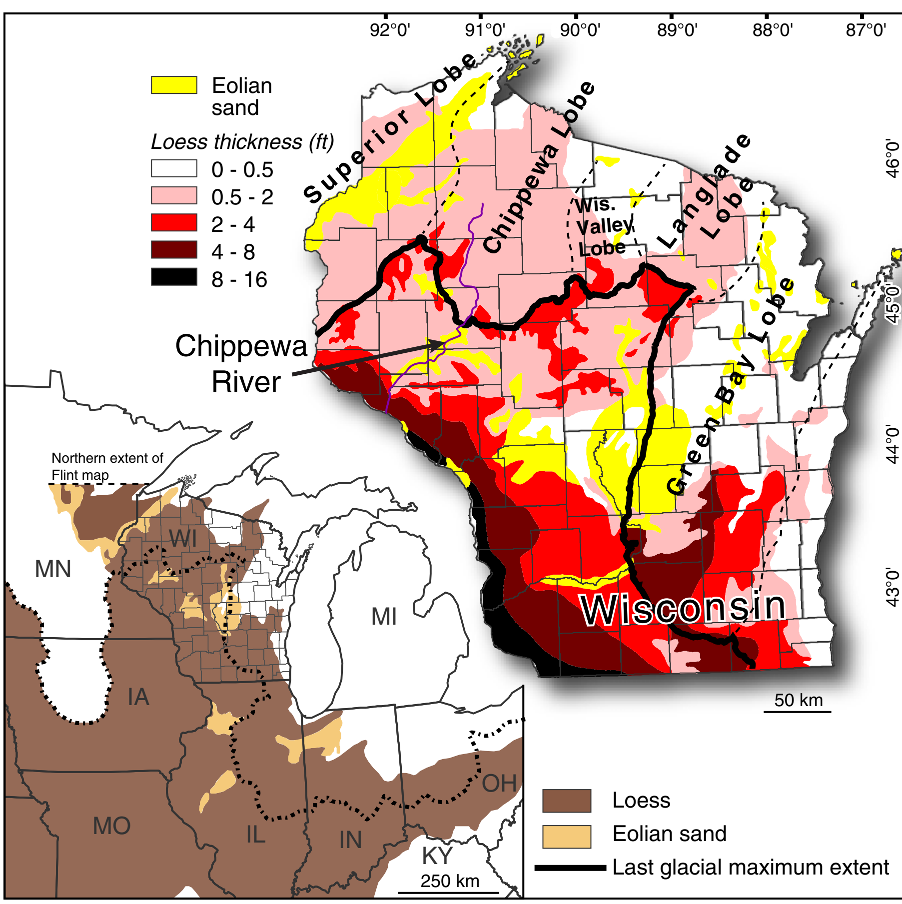


**Figure 1.** Distribution of loess in the Midwest and in Wisconsin, based on Hole (1950) and Flint (1971), both of which were from a small-scale map by Thorp and Smith (1952).

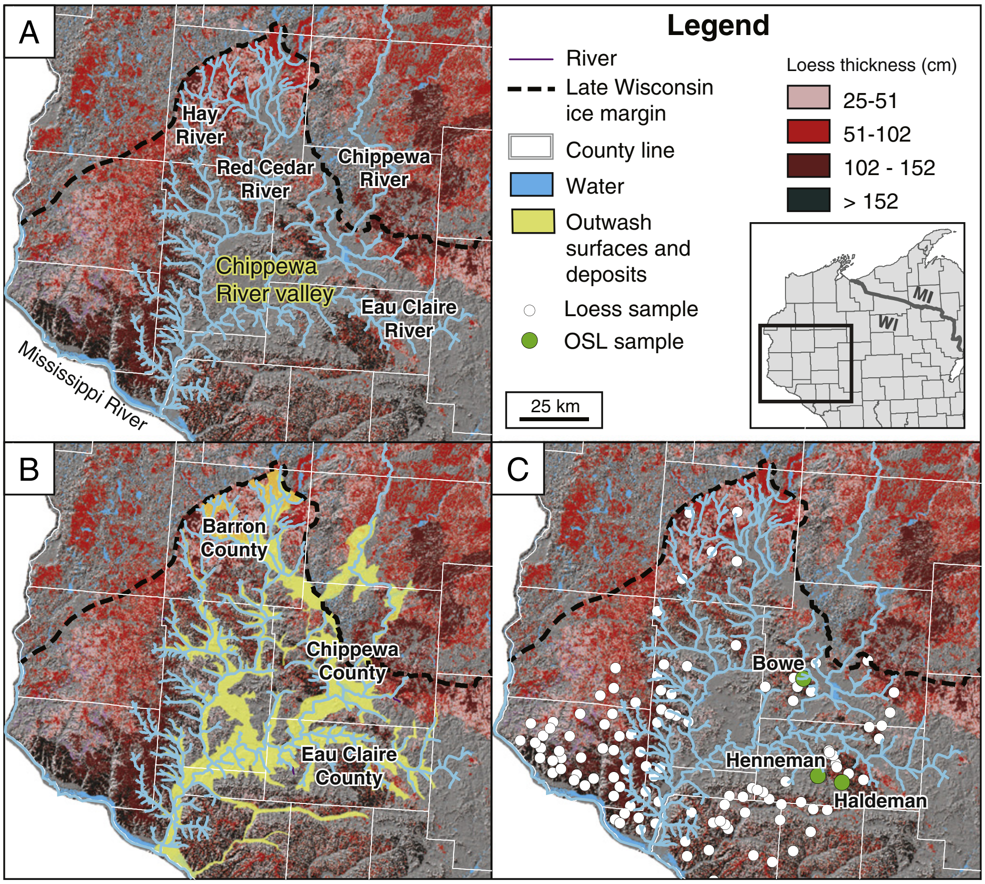


**Figure 2.** Study area maps, showing topography (as depicted by a hillshade DEM), the Chippewa River and its major tributaries, and the LGM extent of the late Wisconsin (MIS 2) ice. A. Loess distribution and thickness, as derived from NRCS county soil surveys. B. Areas of outwash and other sandy soils in the valleys, as derived from NRCS soil surveys and statewide (Hadley and Pelham, 1976) and county-wide Quaternary maps (Syverson, 2007). C. Locations of loess (texture and thickness) and OSL sample sites. OSL sites were also sampled for loess texture and thickness.

analysis even further by stressing the importance of local loess sources and demonstrating that the loess derived from them has clear and interpretable distance-decay trends in texture and thickness. Although some work has been accomplished in Wisconsin regarding loess distributions and source areas, no work has yet been published on loess in western parts of the state, i.e., in our study area. Therefore, this research makes a contribution to our understanding of the distribution, source and character of the loess in western Wisconsin.

*Landscapes, sediments and geomorphology*

The study area is broadly delimited by the watershed of the Chippewa River, as bounded by the late Wisconsin terminal moraine on the north and northeast, and by the Mississippi River valley on the southwest (Figs. 1, 2). The moderate-relief landscape is characterized by bedrock-controlled, rolling and sloping terrain, often with wide, flat valley bottoms. Dissected, Cambrian sandstone- and Ordovician dolomite-cored uplands surround lowlands with thick deposits of residuum and slopewash (Weidman, 1907; Martin, 1965; Michelso and Dott, 1973; Brown, 1988; Syverson, 2007; Fig. 3A). Protected areas on the uplands, as well as some isolated valley locations, may contain a thin cover of pre-Wisconsin glacial sediment, typically weathered till and outwash, resting directly on bedrock (Baker et al., 1983; Jakel and Dahl, 1989; Syverson, 2007; Syverson and Colgan, 2011).

The presence of loess in this area has been known for over a century (Chamberlin, 1897), but its precise distribution, origin and age had not been explicitly studied. Previous statewide maps of aeolian silt and sand deposits (Fig. 1) identified loess on bedrock uplands throughout the study area, as did the more recent map produced by Scull and Schaetzl (2011), derived from NRCS county-scale soil survey data (Fig. 2). Syverson (2007) noted that the loess in Chippewa County is generally <50 cm thick, but he does describe few locations where loess is ≥3 m thick.

Soils that have formed in this loess are mainly weakly developed Mollisols and Alfisols, which have developed under prairie and savanna vegetation. Most of the soil profiles observed (and sampled) have minimally developed argillic horizons, or lack evidence of clay translocation entirely.

Although the loess is usually restricted to bedrock uplands (Hole, 1976), thick deposits of silty, colluvial materials—presumably reworked loess—are common in footslope locations (Jakel and Dahl, 1989; Mason and Knox, 1997; Bettis et al., 2003). Loess thicknesses on uplands appear to be mainly a function of (1) landscape position and hence, site stability, and (2) distance from the Chippewa River valley. That is, locations on steeply sloping uplands typically lack loess, which we attribute to post-depositional erosion, rather than lack of initial loess deposition. Scull and Schaetzl (2011) informally referred to this loess in this area as the Chippewa River valley loess deposit, after its likely source area.

The Chippewa River and many of its tributaries drained the late Wisconsin ice front (Syverson, 2007). Today, the river heads well behind the former ice margin, in Lake Chippewa, in central Sawyer County, Wisconsin. It then flows south across recently glaciated terrain, crossing the late Wisconsin terminal moraine about 13 km NW of the city of Chippewa Falls (Fig. 2). From there, it flows for approximately another 107 km to its junction with the Mississippi River. The main tributaries of the Chippewa River, the Red Cedar and Hay Rivers, flow southerly, originating distal to a major reentrant area along the moraine front (Fig. 2). South of the moraine, the Chippewa River valley and its

**Figure 3.** Photos of study area landscapes. A. Rolling sandstone uplands in the central part of the study area; the hilltops are capped by thin deposits of loess. B. Broad outwash deposits of the lower Chippewa River, with bedrock uplands in the distance. Note the irrigation system in use on the sandy soils of the valley bottom.

tributaries have extensive areas of broad, sandy outwash terraces and floodplains, often attaining widths >6 km (Figs. 2B, 3B). Many of these sandy lowlands and terraces in the main valley are graded to broad ramps of sandy sediment that extent far up into tributary valleys. This sediment may have originated during periods of accelerated slope erosion resulting from permafrost that occurred here during the late Wisconsin. Permafrost likely led to instability on the slopes of the sandstone-cored uplands, allowing them to shed copious amounts of sand into the valley heads (Black, 1965; Holmes and Syverson, 1997; Mason and Knox, 1997; Stanley and Schaetzl, 2011).

## Methods

### *Loess and surficial sediment data acquisition and mapping*

Following methods outlined in previous work (Stanley and Schaetzl, 2011; Luehmann et al., 2013; Schaetzl and Attig, 2013), county soil survey data for the study area were downloaded from the NRCS's Soil Data Mart (http://soildatamart.nrcs.usda.gov/), imported into a GIS, rasterized and merged into a coverage that spanned the study area. To make the data more useful for mapping and sampling, we determined the parent material(s) for most of the soil series from the official series descriptions (OSD) on the NRCS web site (http://soils.usda.gov/technical/classification/osd/index.html). For soils developed in loess, we also determined its thickness from the OSD, entered the value into the GIS attribute table, and coded the map unit symbology in the GIS coverage accordingly (Fig. 2). Using data obtained from the OSDs, soils formed in outwash sand were also coded into the GIS, as was information on surface soil textures. The GIS data were then loaded onto a laptop computer, equipped with built-in GPS capability, facilitating field navigation to predetermined sites for sampling.

### *Field methods: loess sampling*

Our field sampling goal was to obtain loess samples and thickness data from numerous broad, stable upland sites, so as to map its textural and thickness attributes across the study area. Although our goal was to sample the loess uniformly across the uplands of the study area, it became apparent from the NRCS data (and later confirmed by field-checking) that many of the narrower ridgetops, especially in the center of the study area, lacked loess or the loess there was intimately mixed with sandy sediment from the residuum (Fig. 2). Flatter, more stable, ridgetops usually had the thickest and most silty loess.

Preliminary data from the Chippewa River valley and other sites in Wisconsin indicated that OSL dates from loess sampled from depths shallower than 3 m are often too young, a characteristic that we attributed to solar resetting during post-depositional pedoturbation (Bateman et al., 2007). Therefore, we made every effort to seek out and identify hilltop sites where the loess was potentially thick enough (>3 m) for OSL sampling.

We sampled loess and measured its thickness at each of 125 sites on stable uplands across the study area (Fig. 2C), using a 195-cm-long hand auger. Loess thicknesses discussed in this document should be viewed as maximum values, because at these sites the loess should have been optimally preserved. On sloping and other, less stable sites, the loess is thinner. At sites where the loess was thicker than our hand auger, the thickness was noted as 195 cm in the GIS attribute table. We recognize, therefore, that on our loess thickness maps, areas mapped as having loess thicknesses >195 cm could (and sometimes do) have loess that is considerably thicker. Most of these sites are, however, at the far SW margins of the study area. Loess samples were taken within or below the soil profile, but at least ≈30 cm from any underlying bedrock. Our goal was to obtain an amalgamated sample of loess that was representative of the entire loess column.

### Lab analyses

Loess samples were air dried, lightly ground to pass a 2-mm sieve, and passed through a sample splitter three times in order to achieve the high level of homogeneity necessary for analysis on a Malvern Mastersizer 2000E laser particle size analyzer. Removal of carbonates and organic matter from the samples was unnecessary because the loess was not originally calcareous and had virtually no organic matter. From each sample, a 2-g subsample was removed and dispersed in a water-based solution of $(NaPO_3)_{13} \cdot Na_2O$, after shaking for 2 h. As discussed in Miller and Schaetzl (2012), the small subsamples analyzed in laser particle size analyzers may not be representative of the larger sample. Thus, in order to optimize the quality of our particle size data, we analyzed two subsamples from each loess sample and compared the data. When the suite of particle size data (the Mastersizer produces 105 discrete "slices" or bins of data) was sufficiently similar statistically, we used the mean values for all subsequent analyses. However, in cases where the data from the two runs were sufficiently dissimilar (see Miller and Schaetzl (2012) for details), a third, or sometimes even a fourth subsample was run, and the two most comparable samples were used to generate the mean values used in subsequent analyses. In the end, this procedure resulted in a robust and highly representative data set for the loess in the study area.

### Data analyses

As discussed elsewhere (McSweeney et al., 1988; Luehmann et al., 2013; Schaetzl and Attig, 2013; Schaetzl and Luehmann, 2013), loess deposits can become intermixed with underlying sediment, especially in areas known to have had permafrost. In our study area, mixing of the deepest loess with sandstone residuum below is common, as evidenced by bimodality in the particle size curves of loess near the bedrock contact. Therefore, we followed the practice of Luehmann et al. (2013) and "filtered" the particle size data, i.e., adjusting the particle sizes by removing the sediment that composes the coarser "peak" and recalculating the remaining textural data. The goal of the filtering process is to restore the particle-size data as close as possible to its presumed original composition. Given that most of our loess samples lacked a second (sand) peak, the filtering process left most of our original particle-size data unchanged. Nonetheless, for sites shallow to sandstone residuum, the filtering process performs an important function—restoring the particle size data of the loess to a condition closer to its original, unmixed state. Lastly, all particle size data were converted to a clay-free basis, so as to negate the effects of pedogenesis.

### GIS analyses

We kriged the filtered loess particle size and thickness data to create maps of loess characteristics for the study area, using the geostatistical wizard module of ArcGIS. We symbolized the data in isoline format and clipped the isolines to the approximate extent of the data. Normally, we set the default parameters in the geostatistical wizard to 15 and 12 maximum and minimum neighbors, respectively, while also adjusting the number of isolines and their spacing (equal interval vs. geometric interval) to maximize interpretability.

### OSL sampling and analyses

Samples for OSL dating were recovered from five different locations where the loess mantle was >3 m thick. All of these sites are on stable bedrock uplands and located <25 km from the center of the Chippewa River valley (Fig. 2C). Three of the five locations are on a single farm, but for these three, each sampling site is on a different ridgetop, at least 200 m apart.

At each site, a Geoprobe unit (©Geoprobe Systems, Salina, KS) was used to core through the loess and into the soft sandstone below. The Geoprobe recovered a 5-cm loess core, housed in an opaque black plastic tube. After transporting the sealed tubes to the Luminescence Dating Research Laboratory at the University of Illinois, Chicago, samples were recovered for subsequent OSL analyses by cutting the tubes every 10–25 cm and quickly sealing the ends with opaque tape. The deepest sample from each core was recovered <10 cm from the sandstone below. Table 1 provides data on sample depths for the five locations. Samples from duplicate cores, taken a few cm laterally from the original, were also recovered, opened on site and sampled incrementally for particle-size analyses.

### Optically stimulated luminescence dating

Over the past decade there have been significant advances in dating late Pleistocene sequences of eolian sediments, such as loess, by optical dating (Duller, 2004; Lian and Roberts, 2006; Wintle, 2008; Duller and Wintle, 2012). In our research, we purposely did not use the single-aliquot regeneration (SAR; Murray and Wintle, 2003) approach to date the loess for two reasons: (1) the first preheat (240°C) yielded a disproportionate response of the natural and subsequent test dose, and (2) also showed an elevated slow component (1000–3000 photon counts/0.4 s) which constitutes 10–20% of the total luminescence emissions. These results violated the tenets of the SAR test dose (≈7 Gy) shine down curve protocols and often lacked a fully dominant fast component, as demonstrated by ≥50% aliquots with fast to medium component ratios of <10 (Madsen et al., 2009).

As a result, we instead used a modified multiple-aliquot regeneration (MAR) procedure with component dose normalization to generate optical ages on purified extracts of quartz grains from this loess (Table 2) (Jain et al., 2003). This analytical approach has been proven successful for dating eolian sediments in other contexts, with "zero" ages on modern eolian sand, a high fidelity for recovering a known laboratory dose, and good concordance with other dating methods (Bright et al., 2010; Londono et al., 2012; Shanahan et al., 2013). An important tenet of the MAR protocol (Table 1) is that solar resetting of quartz aliquots (8 hours exposure from a sun lamp) effectively resets fast, medium and slow luminescence components. We contend that this laboratory-induced solar resetting mimics what occurred during the original loess deposition, prior to its burial. This laboratory solar resetting is equivalent to about a 16-hour sunlight exposure (Forman and Ennis, 1991), a likely minimum light exposure time for loess.

**Table 1**
Optically stimulated luminescence ages on quartz grains from loess at five upland locations in the study area, where loess thickness is >3 m.

| Field sample location | Depth (cm) | UIC lab no. | Grain size (μm) | Preheat ratio[a] | Equivalent dose (Gy)[b] | U (ppm)[c] | Th (ppm)[c] | $K_2O$ (%)[c] | Cosmic dose (mGy/yr)[d] | Dose rate (mGy/yr)[e] | Optical age (yr)[f] |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Henning 1 | 300 | 2994 | 63–100 | 0.98 ± 0.03 | 21.75 ± 1.33 | 2.4 ± 0.1 | 7.9 ± 0.1 | 2.41 ± 0.02 | 0.14 ± 0.01 | 2.18 ± 0.14 | 9960 ± 925 |
| Henning 1 | 350 | 2991 | 63–100 | 1.02 ± 0.03 | 29.14 ± 1.89 | 1.9 ± 0.1 | 6.5 ± 0.1 | 2.16 ± 0.02 | 0.13 ± 0.01 | 2.35 ± 0.15 | 12,405 ± 960 |
| Henning1 | 400 | 2992 | 63–100 | 0.97 ± 0.03 | 32.24 ± 2.10 | 2.1 ± 0.1 | 7.1 ± 0.1 | 2.10 ± 0.02 | 0.13 ± 0.01 | 2.31 ± 0.15 | 13,980 ± 1060 |
| Henning 1 | 450 | 2988 | 63–100 | 0.99 ± 0.03 | 40.99 ± 2.70 | 2.1 ± 0.1 | 6.6 ± 0.1 | 2.13 ± 0.02 | 0.12 ± 0.01 | 2.37 ± 0.15 | 17,310 ± 1325 |
| Henning 1 | 500 | 2995 | 63–100 | 1.04 ± 0.03 | 38.70 ± 2.60 | 1.6 ± 0.1 | 4.9 ± 0.1 | 1.85 ± 0.02 | 0.11 ± 0.01 | 1.96 ± 0.10 | 19,690 ± 1460 |
| Henning 1 | 550 | 2990 | 63–100 | 0.96 ± 0.03 | 41.92 ± 2.74 | 1.5 ± 0.1 | 4.4 ± 0.1 | 1.63 ± 0.02 | 0.11 ± 0.01 | 1.76 ± 0.99 | 23,820 ± 1820 |
| Henning 2 | 490 | 3250 | 100–150 | 0.99 ± 0.03 | 32.26 ± 2.12 | 2.1 ± 0.1 | 6.6 ± 0.1 | 2.05 ± 0.02 | 0.11 ± 0.01 | 2.23 ± 0.11 | 14,480 ± 1060 |
| Henning 2 | 505 | 3254 | 150–250 | 0.98 ± 0.03 | 27.59 ± 1.79 | 2.1 ± 0.1 | 6.4 ± 0.1 | 2.05 ± 0.02 | 0.11 ± 0.01 | 2.18 ± 0.11 | 12,660 ± 965 |
| Henning 3 | 505 | 3252 | 100–150 | 1.01 ± 0.03 | 42.04 ± 2.85 | 1.5 ± 0.1 | 4.9 ± 0.1 | 2.48 ± 0.02 | 0.11 ± 0.01 | 2.28 ± 0.11 | 18,430 ± 1440 |
| Henning 3 | 520 | 3249 | 150–250 | 1.03 ± 0.03 | 32.33 ± 2.10 | 1.4 ± 0.1 | 4.5 ± 0.1 | 1.76 ± 0.02 | 0.11 ± 0.01 | 1.74 ± 0.09 | 18,540 ± 1380 |
| Haldeman | 330 | 3251 | 150–250 | 0.95 ± 0.03 | 33.99 ± 2.26 | 2.3 ± 0.1 | 7.5 ± 0.1 | 2.41 ± 0.02 | 0.14 ± 0.01 | 2.54 ± 0.13 | 13,395 ± 980 |
| Haldeman | 345 | 3255 | 150–250 | 0.96 ± 0.03 | 35.60 ± 2.45 | 2.1 ± 0.1 | 7.3 ± 0.1 | 2.14 ± 0.02 | 0.13 ± 0.01 | 2.31 ± 0.12 | 15,405 ± 1120 |
| Bowe | 300 | 3253 | 150–250 | 0.97 ± 0.03 | 31.79 ± 2.07 | 2.2 ± 0.1 | 7.3 ± 0.1 | 2.25 ± 0.02 | 0.14 ± 0.02 | 2.41 ± 0.12 | 13,205 ± 960 |

[a] Preheat ratio which is a measure of the stability of the natural luminescence emission reflected in preheat ratio $= \frac{L_{n(\text{nopreheat})}}{T_n} \div \frac{L_{n(\text{preheat})}}{T_n}$.
[b] Quartz fraction analyzed with blue-light excitation (514 ± 20 nm) by multiple aliquot regeneration protocols (Jain et al., 2003).
[c] U, Th and $K_2O$ contents analyzed by inductively coupled plasma–mass spectrometry analyzed by Activation Laboratory LTD, Ontario, Canada.
[d] From Prescott and Hutton (1994).
[e] Includes an assumed moisture content of 20 ± 5% for the burial period.
[f] All errors include systematic and random errors, and are at 1 sigma. Ages calculated from the reference year AD 2010.

At least 24 aliquots were used in the MAR analysis (Fig. 4), with quadruplicate measurement for the natural and regenerative doses (Cox et al., 2007; Bright et al., 2010). Each aliquot contained ~200 to 2000 quartz grains, depending on grain size (63–100, 100–150, or 150–250 μm; see Table 1). The grains were adhered with silicon spray to an approximately 2-mm diameter area on an aluminum disc. Eolian sediments from the Chippewa area are mineralogically mature, with $SiO_2$ contents of between 65 and 87%. This (quartz) fraction was isolated by density separations using the heavy liquid sodium polytungstate. Subsequently, grains were immersed for 40 min in reagent-grade HF, which etched the outer ~10 μm of grains affected by alpha radiation (Mejdahl and Christiansen, 1994). Finally, the quartz grains were rinsed in 10% HCl to remove any insoluble fluorides. The purity of quartz separate was evaluated by point counting of a representative aliquot under a petrographic microscope. Samples that showed >1% of non-quartz minerals were retreated with HF and rechecked petrographically. The purity of quartz separates was subsequently tested by exposing aliquots to infrared excitation (1.08 W from diode at 845 ± 4 nm), which preferentially excites feldspar minerals. The samples measured showed weak emissions (<300 counts/s), at or close to background counts, and a ratio of emissions from blue to infrared excitation of >20, indicating a spectrally pure quartz extract (Duller et al., 2003). Luminescence was measured using a Risø Model TL/OSL-DA-15 System containing light-emitting diodes capable of either infrared (875 ± 30) or blue (470 ± 20) excitation. The resulting luminescence passes through Hoya U-340 filters (>10% transmission >380 nm) prior to detection within the system's Thorn-EMI 9235 QA photomultiplier tube.

Solar resetting of aliquots prior to MAR analysis was accomplished by 8 h illumination from a 275W General Electric Mercury Vapor Sunlamp, removing any pre-existing electrons within photosensitive traps, while also inducing minimal dose sensitivity changes (Richardson, 1994). The MAR protocols employ a test dose (~30 Gy β) to compensate for laboratory-induced sensitivity changes. The test dose is measured first; thus all aliquots have the same heating and irradiation history prior to each regenerative dose. This approach obviates sensitivity changes with increasing regenerative dose. Two heating treatments were employed to eliminate electrons residing in traps that are thermally unstable over geologic time (Table 2). The first preheat is stored at 150°C for 1 h immediately following each laboratory irradiation, and the second heating is 125°C during excitation by blue diodes (Wintle and Murray, 2000). The efficacy of the first preheat treatment for the normalization and regenerative dose was evaluated by comparing curve shape (trap distribution) between the natural and subsequent dose (Bailey et al., 2003). A similar dose response is indicated by zero or low slope (<0.1) between the luminescence for the initial and secondary dose. The natural luminescence emission is derived from electrons residing in time-stable traps. However, to test for stability of the natural luminescence, the natural unpreheated and preheated (150°C for 1 h) are compared as a ratio (Table 1). A ratio of unity (within 2 sigma errors) denotes that there is no change in natural emissions with preheating, which is the case for all samples (Table 1), and hence, indicates a stable natural signal. A sequential regenerative dose of up to ~130 Gy was applied to each sample that exceeded the corresponding natural luminescence; this dose response is unsaturated (Fig. 4). The equivalent dose was calculated for at least the first 40 s of excitation, dependent on background counts (<100 photons/s), as a weighted mean.

To render an optical age, an estimate of environmental dose rate is needed for the burial period. This value reflects exposure to ionizing radiation from the decay of the U and Th series and $^{40}K$, and cosmic sources (Table 2). The $^{40}K$, U and Th contents of the loess, assuming secular equilibrium in the decay series, were determined by inductively coupled plasma–mass spectrometry (ICP–MS) analyzed by Activation Laboratory LTD, Ontario, Canada. A cosmic ray component of 0.11 to 0.14 mGy/yr was included in the estimated dose rate, varied as a function of sample depth (Prescott and Hutton, 1994). Moisture content (by weight) for the dated sediment reflects current conditions, which

**Table 2**
Protocols for Multiple Aliquot Regenerative (MAR) dating of quartz extracts.

| Regenerative growth curve | Natural |
|---|---|
| 1. Solar resetting using 8 h of sunlight exposure. | 1. Photo-stimulation with blue diodes and data collection at 125°C. |
| 2. Test dose of 30 Gy. | 2. Solar resetting using 8 h of sunlight exposure. |
| 3. Preheat 150 °C for 1 h. | 3. Test dose of 30 Gy. |
| 4. Photo-stimulation with blue diodes and data collection at 125°C. | 4. Preheat 150°C for 1 h. |
| 5. Optical bleaching using 8 h of UV. | 5. Photo-stimulation with blue diodes and data collection at 125°C. |
| 6. Regenerative doses, e.g. 33 Gy. | |
| 7. Preheat 150°C for 1 h. | |
| 8. Photo-stimulation with blue diodes and data collection at 125°C. | |

in mesic central Wisconsin was assumed to be 20 ± 5%, based on field data for similar sediments recovered and analyzed between 2006 and 2010.

## Results and discussion

### *Loess sources, textures, distributions and thickness patterns*

Elevation data, used in conjunction with data on surface texture and the extent of soils developed in outwash parent materials, enabled us to map the outwash surfaces of the Chippewa River and its tributaries (Fig. 2B), conforming largely to the mapping efforts of Syverson (2007). Broad, sandy flats, outwash surfaces and terraces are common throughout the basin and south of the late Wisconsin terminal moraine, but are particularly wide and prominent below the City of Chippewa Falls, and continuing for ca. 66 km downstream from there. The lower 35 km of the Chippewa River valley is considerably narrower, being constrained between bedrock uplands.

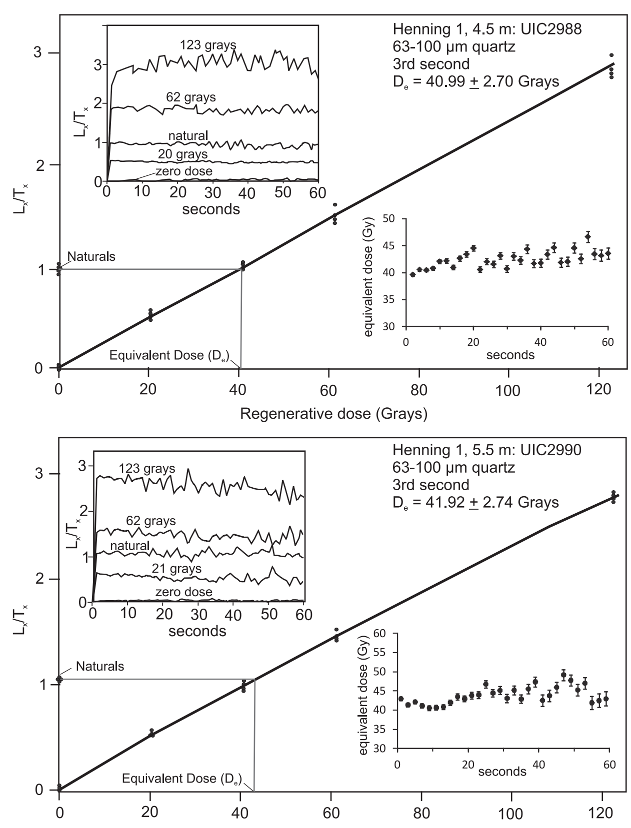


**Figure 4.** Regenerative dose response curve for sample UIC2988 and UIC 2990 for the 63–100 µm quartz fraction. Upper inset figures show natural luminescence normalized shine down curves and associated regenerative dose response. Lower inset figures show equivalent dose for multiple light exposure times (plateau plot), including after 2 s, which is depicted in the main plot.

The loess in this region is very silty and remarkably consistent in textural attributes. Our "filtered" loess data confirmed that all 125 samples were silt loam textured, with an average particle size mode of 36.1 µm (medium-coarse silt). The average silt/sand ratio for this loess is 3.14 ± 0.9, attesting to the high degree of sorting and siltiness of these deposits. Clay, silt and sand contents, respectively, averaged 12.7 ± 1.7, 64.9 ± 5.6, and 22.4 ± 6.4%, with most of the sands falling in the very fine sand category (50–125 µm; 18.6 ± 3.8%).

Although the median loess thickness at sampled sites in the region was 160 cm, loess distribution and thicknesses in Figures 2 and 5 illustrate that many of the highly incised uplands near the valley lack appreciable loess because of their narrow ridgetops and steep slopes. Nonetheless, many of the more gently-sloping, broader uplands do have a more-or-less continuous and thick loess cover, although thickness varies as a function of slope gradient. For example, on the broad, flat bedrock uplands that lie west of the Chippewa River, near its junction with the Mississippi, loess is particularly well preserved, existing as a nearly continuous cover that ranges from 120 to 190 cm thick. On many other uplands, loess cover varies greatly in thickness, even across short distances. For this reason, the loess thicknesses shown in Figure 5A represent maximum potential thicknesses for stable, upland sites. Across the study area, loess is usually absent in valley bottoms and on steep slopes.

Loess thickness trends show clear spatial trends that suggest that the outwash surfaces of the Chippewa River were the main sources of the loess for most—and certainly for the central part—of the study area. Although the Mississippi River has long been known to have been a prodigious loess source along many of its reaches (Fehrenbacher et al., 1986; Leigh and Knox, 1993; Bettis et al., 2003; Scull and Schaetzl, 2011), loess thickness trends suggest that its influence here may have extended inland <20 to 25 km. Beyond this distance, loess thickness trendlines are no longer parallel to the Mississippi River valley, but instead, relate more so to the Chippewa River valley (Fig. 5A). Isolines associated with the thickest loess parallel the Mississippi River valley in the southwestern part of the study area; they undoubtedly indicate that loess here originated from the Mississippi River valley. Farther inland, however, the loess may have had dual sources, and we suggest that much of the loess farther than ca. 40 km from the Mississippi River valley was derived from the Chippewa River valley. The restricted extent of Mississippi River valley loess in this region is best explained by geography, as the valley is oriented NW–SE, which may have paralleled the prevailing westerly/northwesterly winds during this period (Mason et al., 1994; Muhs and Bettis, 2000; Bettis et al., 2003; Stanley and Schaetzl, 2011), limiting the extent to which loess could have been transported to sites far up the Chippewa River valley to the north and northeast.

With the exception of areas immediately adjacent to the Mississippi River valley, loess in the study area is the thickest on uplands that lie 15–20 km southeast of the central Chippewa River valley (Fig. 5A), where at some sites it exceeds 5 m in thickness. Perhaps not coincidentally, this area lies immediately southeast of the widest part of the Chippewa River valley (Fig. 2B). This thick loess area is east of a long (>40 km) section of the valley in which the river flows mainly east-to-west. If winds had been dominantly from the west or northwest during the period of loess deposition, as has been suggested (Mason et al., 1994; Muhs and Bettis, 2000; Stanley and Schaetzl, 2011), this upland would have been immediately downwind from an exceptionally broad reach of the Chippewa River valley (Figs. 2B, 5A). Near the Illinois River valley in northern Illinois, Putman et al. (1988) also found that loess thickness on uplands is directly correlated to valley width. Conversely, loess is notably thin (<80 cm) in the southeastern part of the study area, probably because this area has a large proportion of narrow ridgetops that did not provide stable surfaces for loess deposition and retention. Equally importantly, however, this area lies well north of the direct flow-stream of loess coming from the NW–SE trending reach of the Mississippi River valley and is downwind from only a very narrow reach of the Chippewa River valley (if one assumes westerly/northwesterly winds during loess the transportation period) (Figs. 2B, 5A).

In summary, loess thickness trends across the study area show thinning trends away from the Chippewa and Mississippi River valleys (Fig. 5A). Areas of thick loess in the central parts of the study area occur near wide sections of the Chippewa River valley, and this loess is considerably and predictably thinner toward the margins of the study area, farther from the valley. These general trends are only interrupted by the steep thickening trends associated with the Mississippi River valley in the southwestern part of the study area. Because loess tends to decrease in thickness, predictably, away from source regions, we interpret Figure 5A as strong evidence for the Chippewa River valley as the main loess source for this region.

In northeastern Wisconsin, Schaetzl and Attig (2013) reported a fine sandy loess that was derived from outwash and deposited on nearby uplands. The loess in the Chippewa River valley area is much siltier than the near-source loess in NE Wisconsin, even though it had a similar (glacial outwash) source. We interpret the high silt content of this loess, even on uplands near to the source valley, to the high elevations of the upland sites where it was sampled. Saltating sands—especially fine and medium sands—were apparently unable to be transported to high upland sites in the Chippewa study area. Note, for example, the elevated silt/sand ratios on the high, bedrock uplands just west of the central Chippewa River valley (Fig. 5D).

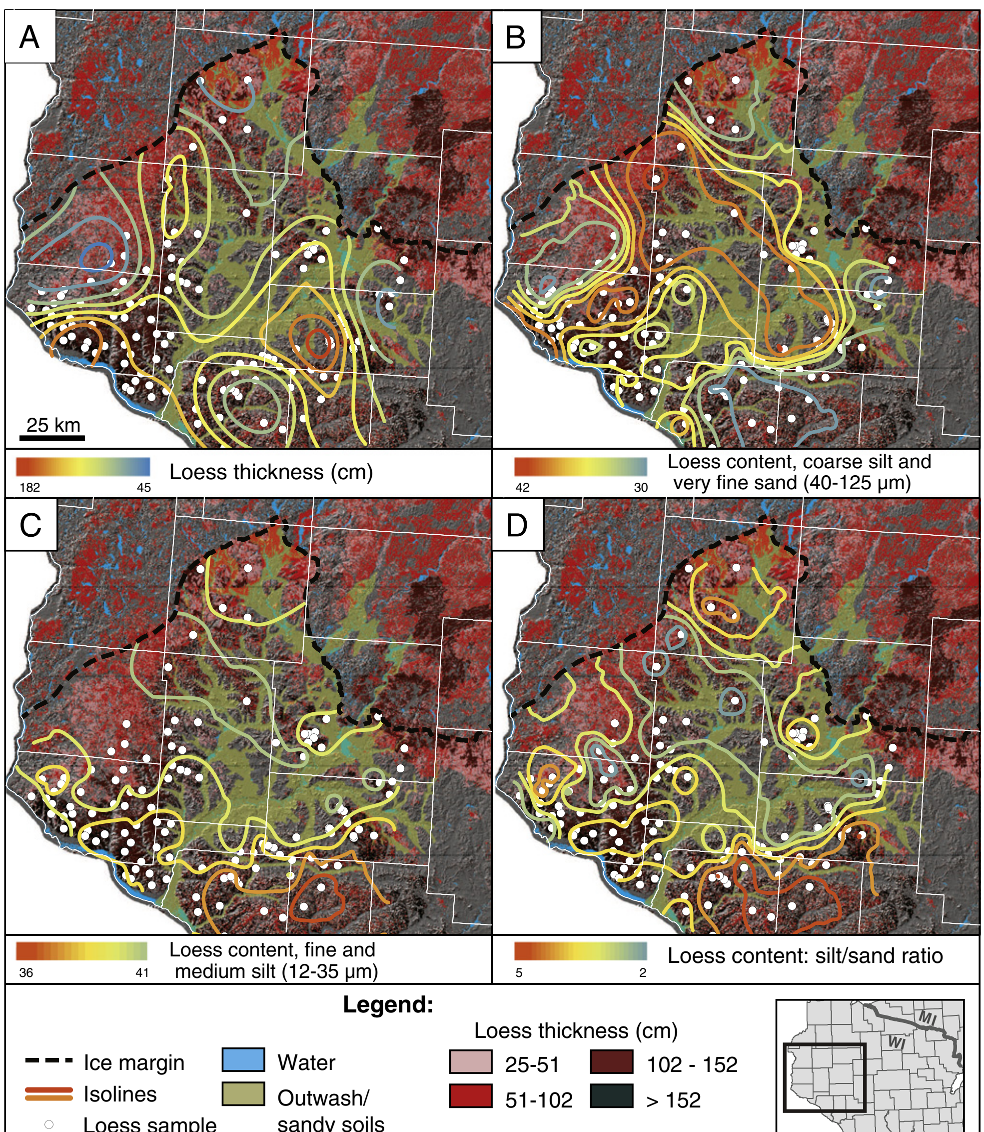


**Figure 5.** Kriged isoline maps of loess characteristics across the study area, set on a gray hillshade background and showing the Chippewa River outwash surfaces. Isoline values do not occur at equal intervals, as per the default mapping routine in ArcGIS. A. Loess thickness (cm). B. Content of coarse silt and very fine sand (40–125 μm). C. Content of fine and medium silt (12–35 μm). D. Ratio of silt/sand contents in the loess.

*Loess textural patterns*

Coarser particle size fractions in loess normally are maximal at sites near to source areas (Fehrenbacher et al., 1965; Rutledge et al., 1975; Ruhe, 1984; Stanley and Schaetzl, 2011; Schaetzl and Attig, 2013). Therefore, we examined patterns of coarse silt through very fine sand contents (40–125 μm) in loess in the study area, as an indicator of potential source areas (Fig. 5B). Loess with the largest contents of these fractions occurs mainly in areas near to the central Chippewa River valley. Contents of this fraction diminish rapidly with distance away from the central valley in all directions except to the southwest, an area that follows the downstream reaches of the river, and one that may also have been supplying some loess to the sedimentation system (Fig. 5B). Conversely, contents of fine and medium silt—presumably the highest in areas that are far from loess source areas—are minimal in the central Chippewa River valley and attain maximum values at the various margins of the study area (Fig. 5C). Lastly, mapped silt/sand ratios show minimal values in the central Chippewa River valley region and increase toward the margins of the study area (Fig. 5D). Collectively, these spatial trends indicate that the outwash surfaces of the Chippewa River, particularly the central parts where they are widest, were the sources for much of the loess on the surrounding bedrock uplands. This conclusion allowed us to tie OSL ages from the Chippewa River valley loess to the event that led to the meltwater itself—the advance of the Chippewa Lobe into the Chippewa River drainage (Fig. 2).

*OSL dating and analyses*

The bedrock at all of five the loess sample sites is Cambrian Mount Simon Formation, a weakly cemented sandstone and shaly sandstone. Loess at the three Henning sites (Fig. 2C) rests directly on very clean sandstone. At the Haldeman site, loess rests conformably on ≈1 m of sandy clay loam diamicton with some gravel (Fig. 6), interpreted to be till of the River Falls Formation based on the presence of pebbles of granite and Flambeau quartzite (Syverson, 2007). We cannot rule out the possibility that some of this sediment is residuum. Although loess appeared to have been mixed into the upper part of this till unit, and sandstone into the lower parts of the same unit, relatively unmixed till appears to exist in the center. Evidence for a few (<20) cm of till or residuum between the loess and the bedrock exists at the Bowe site as well, although here loess and sand grains from the bedrock are intimately mixed into the till, diluting its textural signature.

Incremental depth plots (Fig. 6) show that the loess (excluding the bedrock, till or residuum) at every site is generally uniform in texture throughout the column. The loess is consistently silt loam in texture. Modal particle-size values do tend to slowly and progressively get larger nearer the surface. This modal grain-size trend is gradual, and although it may suggest slightly stronger winds later in the loess depositional period, it does not imply a change in wind direction or a different mode of origin for the loess. Although we cannot explain the additional sandiness shown in the upper meter of loess at the Hennning 1 site, it may be related to localized bioturbation or some other type of disturbance. Nonetheless, this increment of loess was not sampled for OSL and is not relevant to our interpretation that the sediment sampled was pure and uniform loess.

Particle-size data provide no indication that a lower Roxana silt unit exists here, even though it is commonly found in the Upper Mississippi River valley beneath Peoria Loess (Johnson and Follmer, 1989; Leigh and Knox, 1993; Leigh, 1994; Jacobs et al., 1997; Grimley, 2000). Rather, particle-size data point only to mixing between the underlying sediment and the overlying loess (Fig. 6). OSL ages on loess from the bedrock uplands in and near the Chippewa River valley, reported in Table 1, also support the argument that Roxana Silt is not present here. The oldest age we report is 23,820 ± 1820 yr. In his review paper, Grimley (2000) reported that Roxana Silt was deposited in the Upper Mississippi River valley, ca. 55 to 28 ka. We also did not see any of the characteristic pink hues associated with the Roxana in any of the cores. Thus, we argue that (1) the entire loess column in the Chippewa River valley is associated with the last episode of loess accumulation, (2) that episode can be tied directly to the Chippewa River and (3) it, in turn, was derived from meltwater coming from the Laurentide Ice Sheet margin in western and northwestern Wisconsin.

Deposition of Peoria Loess in the upper Midwest, USA, is thought to have started about 25 ka and continued through the peak of the last glaciation (Grimley, 2000). Our OSL ages fall within this range (Table 1). Our oldest age, 23,820 ± 1820 yr (Fig. 6), suggests that meltwater was flowing down the Chippewa River valley by 24 ka and probably before, as the Laurentide Ice Sheet crossed the drainage divide in northwestern Wisconsin, advancing out of the Lake Superior basin. This divide is considerably ≈150 km north–northeast of the Wisconsin terminal moraine (Figs. 1, 2), only a few tens of km from the Michigan border. It is unclear exactly when the first meltwater from the advancing (and probably oscillating) ice front would have entered the Chippewa drainage, but our OSL ages suggest that meltwater began flowing down the Chippewa River valley and depositing outwash in quantities large enough to result in loess deposits on the nearby uplands by about 24 ka. It is likely that the very earliest loess fell onto a landscape underlain by permafrost, and was largely not retained. We suggest that, only later, as the climate began to ameliorate and vegetation thickened, did loess accumulate in measurable amounts on uplands. At present, there is no way to know the amount of time that elapsed between initial loess production from the Chippewa system and the first instances of loess retention on uplands in the study area. Loess retention and erosion during this time were likely episodic.

The ages on the deepest loess at the other four sites are 13,205 ± 906 (Bowe), 15,405 ± 1120 (Haldeman), 18,540 ± 1380 (Henning 3), 12,660 ± 965 (Henning 2) and 23,820 ± 1820 (Fig. 6). Together, these ages point to a relatively long period of loess deposition and erosion between 24 and 13 ka. We again emphasize that loess probably began falling onto this landscape well before these dates, i.e., we interpret these dates on basal loess as minimum-limiting dates for the loess deposition event in the Chippewa River valley. Perhaps more correctly, they represent the earliest instances of loess retention and accumulation on the permafrost-cored uplands of the study area. The majority of the remaining OSL ages suggest that the bulk of the loess deposition had ended by 12 ka, perhaps because the ice margin had retreated too far north to have contributed meltwater to the Chippewa drainage systems.

Only one radiocarbon date exists that can constrain the advance and recession of the southern margin of the Laurentide Ice Sheet in western Wisconsin. Black (1976) reported a date of 26,060 ± 800 $^{14}$C yr BP for a spruce log buried below 60 m of outwash in the lower Hay River valley (Fig. 2). Using the calibration curve of Fairbanks et al. (2005), this date calibrates to 31,270 ± 920 cal yr BP. The outwash that buried the log would have been associated with the MIS 2 advance into northwestern Wisconsin. The age reported by Black (1976) probably records the earliest accumulation of outwash within the valley, perhaps reflecting a period when the ice sheet had first advanced over the divide and outwash had begun accumulating in the valleys of the Chippewa River and its tributaries. This radiocarbon age, therefore, has similar implications and interpretations to our OSL ages on loess on the nearby uplands, and is in general agreement with our reported ages on loess. The slightly earlier age for this log seems logical, because outwash would have been accumulating for a considerable period of time prior to loess deposition and accumulation on the uplands.

In summary, the OSL ages we report here for the onset of loess accumulation on uplands in the study area agree with this one reported $^{14}$C age. They are also in general agreement with the timing of advance and recession of the MIS 2 ice, as indicated by the regional correlation of ice margin positions in the midcontinent area (Clayton and Moran, 1982; Mickelson et al., 1983; Attig et al., 1985; Maher and Mickelson, 1996; Attig et al., 2011a; Syverson and Colgan, 2011; Carson et al., 2012).

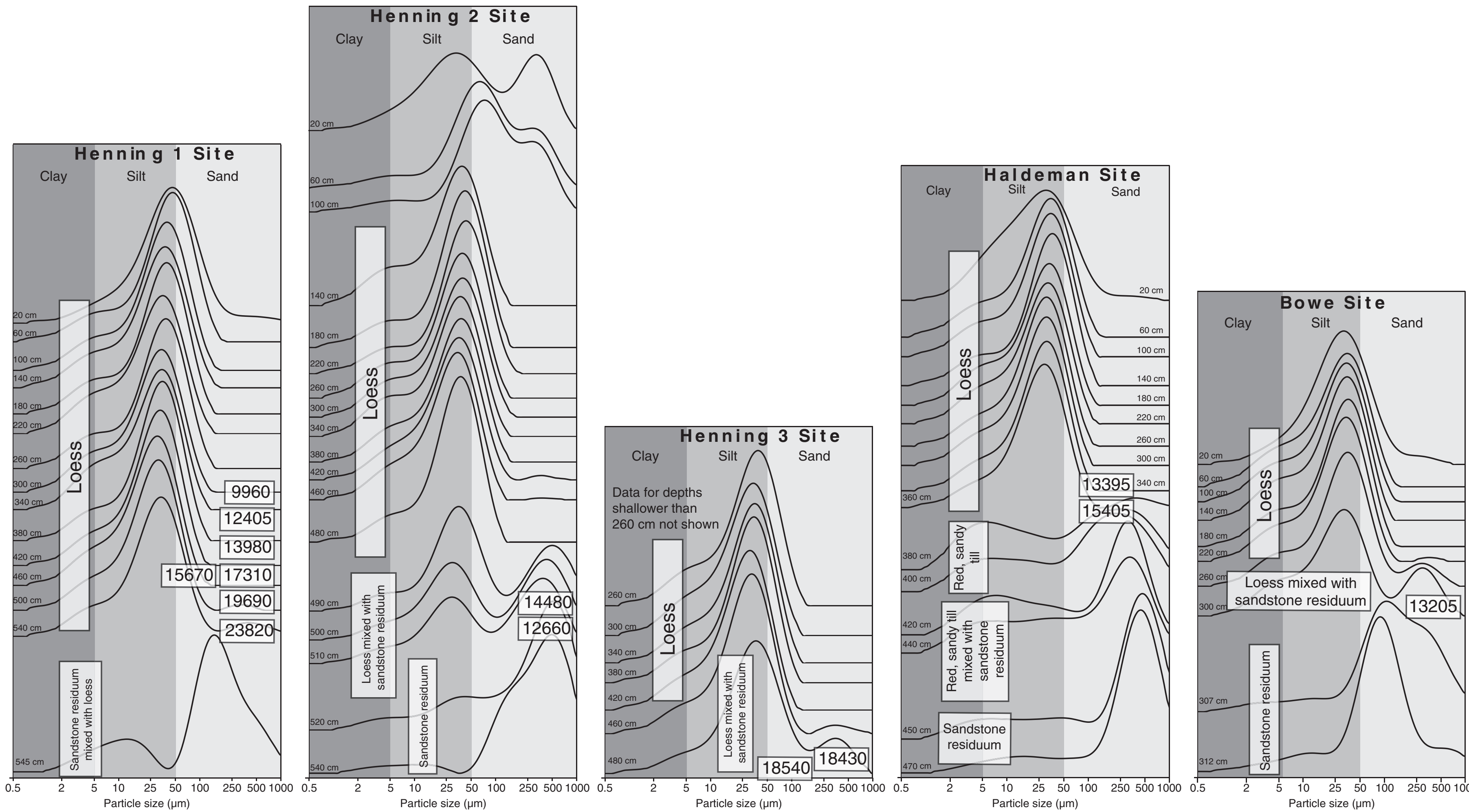


**Figure 6.** Depth plots of continuous particle size curves for the loess and underlying sediment at each of the five sites sampled for OSL dating. Note that, because of stacking mismatches, the vertical distances between each curve may not correspond exactly to the sampling increment used in the field. Also shown are sediment/stratigraphy interpretations, and the OSL dates obtained from specific sites and depths. OSL ages are shown without error terms, because of space limitations. For details, see Table 1.

Our OSL ages on loess also support the findings of studies that have used OSL dating of ice-marginal lake sediment to constrain the timing of the onset of deglaciation in the Baraboo Hills in southern Wisconsin (Attig et al., 2011b; Carson et al., 2012), and the use of cosmogenic radionuclides to date boulders on moraines in several areas of Wisconsin (Ullman et al., 2011). These studies indicate that the onset of deglaciation was about 18,500 and 22,000 yr, respectively, for the southern Green Bay Lobe. For comparison, the OSL ages we report for loess in the Chippewa River valley suggest that the onset of deglaciation in west-central Wisconsin was generally coincident with the chronology that is emerging for southern Wisconsin, but that the overall period of deglaciation (and its associated meltwater) may have lasted longer in western and northwestern Wisconsin.

## Conclusions

Our data strongly suggest that the loess in our study area was derived from the outwash surfaces of the Chippewa River, as meltwater from the late Wisconsin ice sheet flowed down this system and into the Mississippi River. Loess was deflated from the wide, sandy outwash surfaces, deposited on nearby bedrock uplands, and retained particularly well on those that were geomorphically stable. Loess thicknesses on the uplands in the study area occasionally exceed 5 m and commonly are >2 m thick, depending on the local configuration of the hilltop. Flatter areas tend to have thicker loess, as do areas south and east of the central part of the Chippewa River valley. Meltwater probably began flowing down the Chippewa River system, initiating loess production, after the advancing Laurentide ice crossed the drainage divide in northern Wisconsin. Therefore, although loess deposition in the study area cannot be definitively linked to the maximum extent of the late Wisconsin ice, OSL ages on this loess do help constrain the overall glacial chronology of the region, for which we otherwise have little information.

OSL ages on this loess confirm that its deposition began by at least ca. 24 ka, and probably slightly earlier, and continued for at least 10,000 yr. Based on a previously reported $^{14}C$ age on a deeply buried log in an outwash deposit, it seems likely that outwash had been accumulating in the Chippewa River valley area since ca. 26 $^{14}C$ yr BP, and within 2000 yr loess had begun accumulating on the nearby uplands. Loess was probably being generated well before 24 ka, but was not retained on the steep slopes of the bedrock uplands, which were underlain by permafrost at the time. Basal ages from loess on some of our sites were considerably younger than 24 ka, suggestive of slope instability, rather than loess deposition with a later period of onset. In short, the optical ages we report are in general agreement with current understanding of the timing of the ice advance at other locations in the upper Midwest in general, and in Wisconsin in particular.

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