BOREAS An international journal of Quaternary research

Luminescence dating of sand wedges constrains the Late Wisconsin (MIS 2) permafrost interval in the upper Midwest, USA

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Schaetzl, R. J., Running IV, G., Larson, P., Rittenour, T., Yansa, C. & Faulkner, D.: Luminescence dating of sand wedges constrains the Late Wisconsin (MIS 2) permafrost interval in the upper Midwest, USA. *Boreas*. https://doi.org/10.1111/bor.12550. ISSN 0300-9483.

Large parts of the upper Midwest, USA were impacted by permafrost during the Last Glacial Maximum (LGM). Even though permafrost persisted as the Laurentide Ice Sheet began to recede, direct age control of this interval is largely lacking. To better temporally constrain the permafrost interval in western Wisconsin, we identified two sites, outside the Late Wisconsin (MIS 2) glacial limit, that contain relict, ice-wedge pseudomorphs, initially interpreted to be sand wedges, hosted within well-drained outwash deposits. The pre-Wisconsin (>MIS 5) host material commonly displays up-turned bedding near the contact with the wedges, indicative of well-formed features. The wedges are filled with well-sorted, gravel-free, medium and fine sands, and lack evidence of post-formational disturbance, pointing to an aeolian sand infill and confirming them as sand wedges. Ventifacts on nearby uplands attest to windy conditions here in the past. Optically stimulated luminescence (OSL) ages on five sand wedges indicate that they filled with sand between *c.* 19.3 and 18.3 ka at the southerly site and between *c.* 15.1 and 14.7 ka at the northerly site, which is closer to the LGM margin. Sand wedges at the latter site were wider and had more complex morphologies, possibly suggesting a longer interval of formation and/or more intense permafrost. We also examined a site along a ridge crest, between the two wedge sites, which displayed interbedded loess and sand, dated by OSL to 12.7 ka. Together, these results point to dry, cold, windy conditions in west-central Wisconsin, within 100 km of the LGM limit. At this time, aeolian sands were being transported across a landscape with (at least scattered) permafrost. The OSL results suggest multiple phases, or perhaps time-transgressive, sand-wedge formation, associated with permafrost between *c.* 19 and 15 ka, with dry, windy (and likely, cold) conditions persisting until at least 12.7 ka.

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During the Last Glacial Maximum (LGM), the Driftless Area of southwestern Wisconsin and parts of the upper Midwest, USA were sometimes within a few hundred kilometres from the ice front (Fig. 1). Prominent lobes of the Laurentide Ice Sheet (LIS) included the Green Bay lobe on the eastern margins of the Driftless Area (Maher & Mickelson 1996; Colgan 1999; Winguth *et al.* 2004; Attig *et al.* 2011), the Wisconsin Valley, Chippewa and Superior lobes to the north (Cahow 1976; Attig *et al.* 1985; Attig & Clayton 1993; Heath *et al.* 2018; Fig. 1) and, later, the Des Moines lobe to the west (Wright *et al.* 1973; Bettis *et al.* 1996; Patterson 1997; Heath *et al.* 2018). Cold, persistent, katabatic winds are documented for some areas near the ice margin (Schaetzl & Attig 2013), even as strong west-northwesterly and northwesterly winds were common across other parts of the region (Mason *et al.*, 1994a,b; Zanner 1999; Zanner & Nater 2000; Millett *et al.* 2018; Schaetzl *et al.* 2018; Mataitis *et al.* 2019). Because the Driftless Area was at times almost surrounded by ice, cold conditions during the last glaciation (Late Wisconsin, Marine Isotope Stage, MIS 2) must have been persistent here, hence its nickname 'Up in the refrigerator' (Mason 2015). Indeed, during the LGM and well into the postglacial period, permafrost was widespread in the Driftless Area (Clayton *et al.* 2001; Carson *et al.* 2019).

Early work suggested that permafrost had been widespread in northern and central Wisconsin between *c.* 30 and 12.5 ka (Black 1964; Fig. 2), even as later work concluded that continuous permafrost persisted until *c.* 13 ka (Attig & Clayton 1986; Fig. 2). Based on the distribution of ^{14}C-dated wood in Pleistocene deposits across Wisconsin (assuming that forest only occurs on landscapes that lack permafrost), Clayton *et al.* (2001) constructed a time-space diagram of permafrost that suggested that permafrost was present between *c.* 23 and 14 ka in southern Wisconsin, and lasted until 10 ka in northern Wisconsin (Fig. 2); later work was based on dated features formed by permafrost conditions, even though both sand wedges and ice-wedge pseudomorphs have been widely documented in the area, particularly in the Chippewa River Valley (Black 1964, 1965, 1976; Holmes & Syverson 1997) and areas nearby (Johnson 1990; Attig 1993; Mason *et al.*, 1994a,b; Zanner 1999; Clayton *et al.* 2001).

Additional geomorphic evidence has been used to constrain the timing of permafrost. Colluvial deposits on

DOI 10.1111/bor.12550

Fig. 1. Map of the locations of the study sites, as they relate to the maximum extent of the major glacial lobes (light grey) during the Late Wisconsin (MIS 2). Note that at various times ice nearly surrounded the Driftless Area. The boundary showing the extent of the pre-MIS 2 (older drift) deposits largely follows Wisconsin Geological and Natural History Survey (2011).

the footslopes of many bedrock uplands (Mason & Knox 1997; Mason 2015; Mason *et al.* 2019), as well as blockfields and talus (Smith 1949; Cahow 1976), in and near the Driftless Area have been taken as evidence of slope instability and periglacial conditions (Carson *et al.* 2019; Fig. 1). Using stratigraphical relationships and radiocarbon dating of wood and gastropod shells below, within and above colluvial and related fluvial deposits, Mason & Knox (1997) were able to constrain the interval of colluvial activity, and hence, possible permafrost conditions, on the Driftless Area landscape. Radiocarbon ages from wood beneath the colluvium ranged between 28.9 and 20.3 ka, whereas fluvial sediment inset into the colluvium on footslope positions contained materials dated between 12.5 and 9.8 ka (Fig. 2). Age control on samples from within the colluvium fell between these outer windows, at *c.* 18.6–12 ka. Thus, their work points to an interval of slope instability that is (conservatively) between 19 and 13 ka, with indications that it could have begun even earlier, while recognizing that slope instability at the end of this interval may have been due to slope adjustments occurring after permafrost had thawed (Mason & Knox 1997). Radiocarbon ages on colluvium from northeastern Iowa and southwestern Wisconsin (Hallberg & Bettis 1985) indicate that colluvial activity in these regions (presumably initiated by permafrost) began between *c.* 34 and 24.5 ka (Fig. 2), correlating with the MIS 2 ice advance (Black 1965, 1976). Other evidence for extreme cold across the study area during the Late Pleistocene comes from relict patterned ground and related features such as ice-wedge casts (Black 1965; Attig & Clayton 1992; Walters 1994; Johnson 2000; Syverson 2007; Carson *et al.* 2019; Fig. 3). Similar patterned ground has been documented in other areas of the upper Midwest – at locations just beyond the ice margin (Johnson 1990; Walters 1995; Schaetzl 2008; Lusch *et al.* 2009).

Key indirect evidence for permafrost in the region comes from the lack of published radiocarbon dates on wood in areas overridden or incorporated into by MIS 2 ice (Attig *et al.* 1989; Clayton *et al.* 2001). These data have been interpreted to mean that the landscape was in tundra, and hence, unforested, due to cold conditions and permafrost, as supported by palynological data (Birks 1976; Baker *et al.* 1999; Williams *et al.* 2015).

Important new data shed more light on the timing of permafrost. Batchelor *et al.* (2019) reported on speleothem growth in Cave of the Mounds, only 18 km outside the LGM border, in the southern Driftless Area. They used a hiatus in speleothem growth in the cave, between 33 and 15 ka, as indirect evidence of the presence of permafrost for that 18 000-year interval.

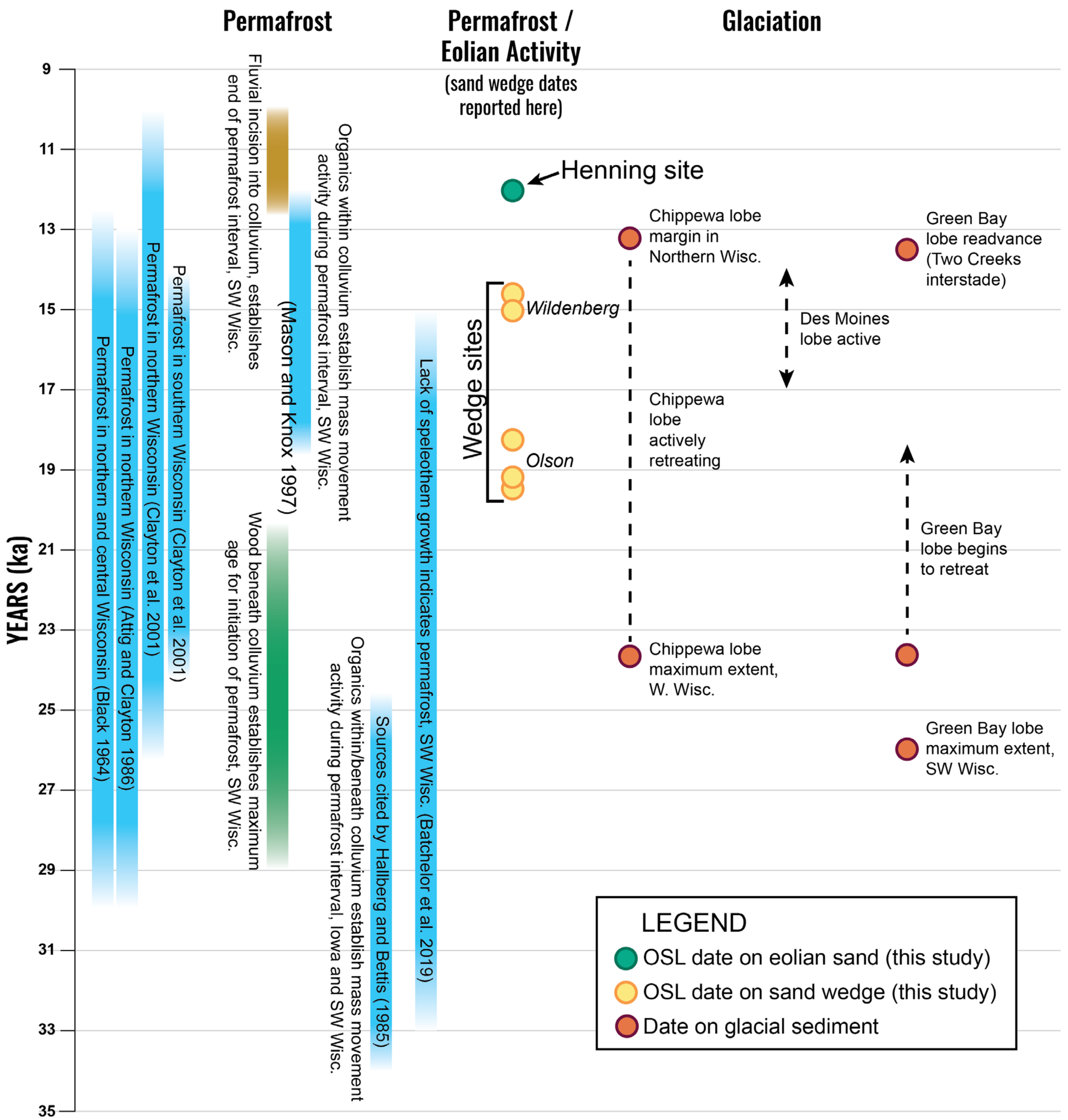

Fig. 2. Literature-derived timelines of permafrost and glacial activity in the region, as they correlate to the OSL results for the sand wedges and the Henning site. See Table 1 for OSL ages and errors.

Similar data from caves in southeastern Minnesota point to an interval of cold climate and minimal speleothem growth from *c*. 35 to 13 ka (Lively 1983).

Although the existence of an extended permafrost interval across western and southwestern Wisconsin is well established, dates on features directly associated with this interval, which could help to constrain its beginning and/or end, are sparse. Such data may help others who are assessing the timing, extent and demise of permafrost in other midlatitude regions that were glaciated (or near the glacial margin) in MIS 2 Vandenberghe *et al.* (2014). Knowing that ice-wedge pseudomorphs occur in the Chippewa River Valley (Holmes & Syverson 1997), we identified freshly exposed pseudomorphs at two gravel quarries and sampled them for grain-size analyses, luminescence dating (Fig. 1), and plant macrofossil analysis. Past work has demonstrated that such features can be dated successfully using

Fig. 3. Typical images (aerial photographs) of landscapes within the Chippewa Valley that retain evidence of weakly formed patterned ground. A. A landscape ~15 km southwest of the Wildenberg gravel quarry. B. A landscape ~3.5 km southwest of the Wildenberg site. Both images are aligned with north at the top.

luminescence techniques (Bateman 2008; Andrieux *et al.* 2018). Here, we report luminescence ages from five, relict, sand-filled, ice-wedge pseudomorphs at two sites in western Wisconsin, so as to better constrain the interval of permafrost conditions. Our subsidiary goal is to use morphological information from these pseudomorphs, in conjunction with new age control, to gain new insights into postglacial palaeoenvironments near the LGM margin.

Background

Our study sites in western Wisconsin, USA are well outside of the MIS 2 ice margin but not, technically, within the Driftless Area (Fig. 1). This landscape consists of rolling hills of nearly flat-lying, friable sandstone bedrock (Martin 1965; Michelson & Dott 1973; Mudrey *et al.* 1982; Havholm 1998; WGNHS 2013; Fig. 4). The Chippewa River carried large quantities of glacial outwash before, during and after the LGM, and thus, its valley is marked with several broad, high, sandy terraces (Andrews 1965; Faulkner *et al.* 2016). Most of the soils in the valley have formed in coarse-textured parent materials, e.g. outwash, sandstone residuum, sandy colluvium, or thin deposits of aeolian sand (Fig. 5B). Water tables are often deep and most of the landscape is well drained; the landscape as a whole has only a few wetlands in some of the low order stream valleys or when dammed by aeolian deposits in upland locations (Fig. 5A; Schaetzl *et al.* 2018). To the east, outside the valley proper, the landscape and soils are generally wetter, having formed in loess overlying finer-textured glacial tills (Mode & Attig 1988; Attig & Muldoon 1989; Stanley & Schaetzl 2011).

The study area, south of the Late Wisconsin terminal (Chippewa) moraine, was not glaciated in MIS 2. Instead, older glacial deposits cover many of the uplands here; these are River Falls Formation sediments (Syverson *et al.* 2011). These reddish-brown sediments are sandy loam till or outwash, likely deposited by an ice advance prior to 130 ka (>MIS 5; Syverson 2007; Syverson *et al.* 2011). Since then, most of the River Falls Formation has been eroded from the landscape, such that it is only infrequently present on the modern surface – usually on stable upland sites (Syverson 2007).

A variety of aeolian sediments mantle much of the study area. Loess data from sites in southeastern Minnesota support the notion of westerly and northwesterly winds for the primary period of loess deposition in the region, between *c.* 26 and 20 ka (Mason *et al.*, 1994a,b; Zanner 1999; Mataitis *et al.* 2019). Loess, derived from these western sources, including the Mississippi River Valley, still covers many of the broad, stable uplands here (Hole 1950; Scull & Schaetzl 2011), but is usually absent from many areas in the central Chippewa Valley, where it appears to have been transported out of the valley by west-northwesterly winds (Schaetzl *et al.* 2018). Evidence for re-entrainment (or lack of initial deposition) of the loess is derived from data on loess distribution; within and to the immediate east (downwind) of the valley, loess is usually absent except for areas on the east and southeast of large, isolated, sandstone ridges (Fig. 5). Schaetzl *et al.* (2018) interpreted these patterns to indicate that sands, driven by strong northwesterly winds, helped to deflate much of the loess that may have initially been deposited here, transporting it out of the main valley. Recent work on extensive sandy aeolian deposits, e.g. sand stringers,

Fig. 4. Typical landscapes on the rolling sandstone hills of the Chippewa Valley. Photographs by RJS.

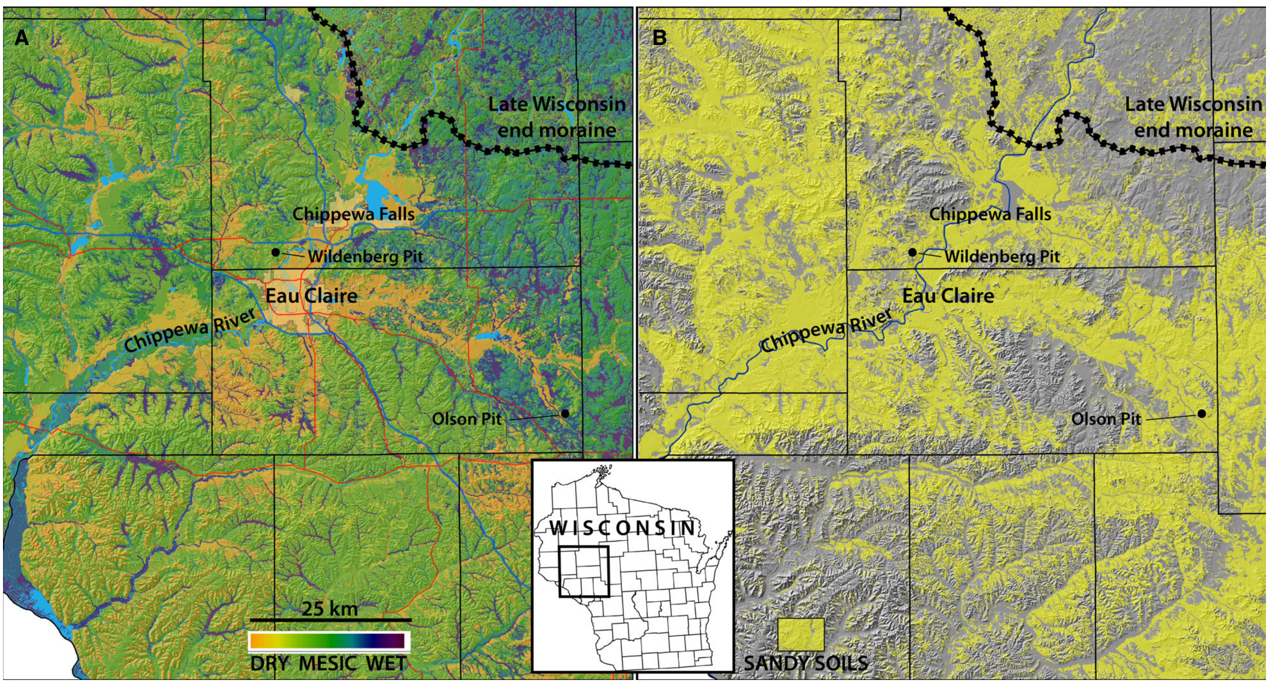

Fig. 5. Thematic maps of the study area, focusing on the Chippewa Valley. A. Map of the long-term wetness–dryness of the soils, as quantified by the Drainage Index of Schaetzl *et al.* (2009). B. Map of soils with sandy surface horizons.

parabolic dunes, sand ramps, climbing dunes and sand sheets, throughout the Chippewa Valley and neighbouring landscapes corroborates this interpretation (Millett *et al*. 2018; Mataitis *et al*. 2019; Shandonay *et al*. 2020). In the lee of these large ridges, loess deposits are protected from erosion, and/or loess from sites farther west has been secondarily deposited on top of them (Schaetzl *et al*. 2014). As a result, at some sites in the immediate lee of large sandstone ridges, loess deposits can commonly exceed 4 m, thinning rapidly to the east (Fig. 5). This aeolian record is pertinent to our study because it documents the strong winds that existed in the area during the immediate postglacial period.

The widespread occurrence of well-formed ventifacts in the Chippewa Valley area but especially on uplands associated with the River Falls Formation (Fig. 6; Cahow 1976; Holmes & Syverson 1997; Johnson 2000) provides further evidence of prolonged and intense aeolian activity. Most likely, this type of erosion was facilitated by a landscape with minimal vegetation cover and/or tundra vegetation, where sands were widely available for saltation and erosion. The ventifacts are often exceptionally well formed, even in resistant rocks such as quartzite (Fig. 6).

Under periglacial conditions, thermal contraction of the frozen soil in winter forms cracks that can then fill in with water, snow, and/or sediment (Zanner & Nater 2000). Repeated cracking and infilling over time forms vertically oriented, downward tapering wedge-like structures of various kinds (ice wedges, sand wedges or composite wedges) that may form polygonal networks of patterned ground (Black 1976). Cracking may be most pronounced in areas that lack persistent snow cover, not unlike what might have been the wind-swept surfaces of the study area (Schaetzl *et al*. 2018). Strata in the surrounding sediment, deformed by the lateral pressures exerted by the expanding wedge, can become upturned. Pseudomorphs of ice wedges then form as the ice melts and the resulting void space fills with sediment, washed and/or blown into open cracks.

Sand wedges, a unique type of ice-wedge pseudomorph, form as aeolian sand fills the open wedge. They tend to develop in drier environments where the frozen ground may not be as ice-rich, and on dry, sandy landscapes where contraction cracks remain open longer (Black 1976; Murton 1996). Characteristically, sand wedges form under cold, arid, windy conditions, i.e. in wind-swept environments, usually with sparse snow cover (Murton 1996; Nai'ang *et al*. 2003; Wolfe *et al*. 2018).

Fig. 6. Ventifacts and wind-abraded rocks from the Chippewa Valley. Most ventifacts here are formed in resistant quartzite or granite. Photographs by RJS.

Material and methods

Field methods

Using aerial photographs, satellite imagery, and LiDAR topographic data, we identified 67 sites in the Chippewa Valley area that appeared to be excavations that might expose potential ice-wedge pseudomorphs. Three of these had been mapped by Holmes & Syverson (1997) in their fig. 2, which showed the locations of nine sites in the Chippewa Valley that contained ice-wedge casts. Unfortunately, we were unable to utilize any of these sites; most had since degraded. Nonetheless, at two of the 67 sites initially identified, our field survey found excavations that exposed well-formed ice-wedge pseudomorphs: (i) a gravel quarry near the city of Eau Claire, *c.* 2.5 km south of the intersection of State Hwy. 29 and County Hwy. F (hereafter: Wildenberg Quarry), and (ii) a gravel quarry *c.* 5.5 km north of the village of Fairchild (hereafter: Olson Quarry) (Fig. 1). Both quarries are cut into the side of a low hill, exposing well-stratified and highly weathered sediments (Fig. 7). We examined, described and photographed many of these features before collecting samples from five representative ice-wedge pseudomorphs for further analysis.

Grain-size analysis

Grain-size analyses were performed on samples from five ice-wedge pseudomorphs, which we further sampled for optically stimulated luminescence (OSL) dating (see below). We also obtained 27 additional samples from an assortment of 10 additional, representative wedges exposed in the quarry faces. Lastly, we obtained 24 samples of the surrounding host sediment, well below the depth of the soil profile, from the two quarries. Six additional samples were recovered from a third quarry, the Menard Quarry, located *c.* 3 km north of the Wildenberg Quarry; this site lacked any wedge forms but did expose similar, pre-Wisconsin-aged sediments, thereby providing a more robust data set.

Grain-size analyses were performed using a Malvern Mastersizer 2000E laser particle-size analyser, on the fine-grained fraction (<2 mm) that had been dispersed for 15 min in a weak solution of $(NaPO_3)_{13}{\cdot}Na_2O$. For these data, we implemented a quality control protocol developed by Miller & Schaetzl (2012), in which multiple measurements from each sample were compared and evaluated against each other, to arrive at the most representative data.

Luminescence dating

On the assumption that the wedge features at the two study sites are sand wedges, and hence filled with aeolian sand, we felt confident that OSL dating could be used to successfully date them (Buylaert *et al.* 2009; Vandenberghe *et al.* 2019). We sampled five wedges for luminescence dating – two at the Wildenberg Quarry and three at the Olson Quarry. We chose the most representative features for sampling – deep and wide wedges filled with clean sand that had been minimally disturbed by pedoturbation, root proliferation, and/or soil development, as indicated by their undisturbed stratigraphy. Samples were collected by driving a steel pipe horizontally into the sand, as deep as possible within the sand infill, but not so deep that the wedge was narrower than about 3× the core diameter, so as to ensure that the sample was derived

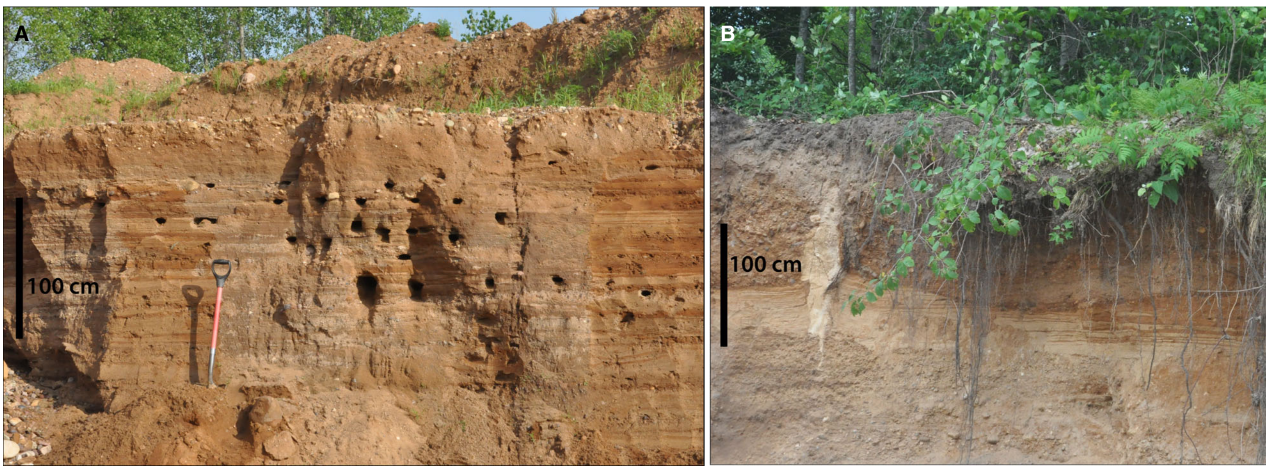

Fig. 7. Horizontally bedded sand and gravel deposits at the (A) Wildenberg gravel quarry and (B) Olson gravel quarry, showing the host sediment – River Falls Formation outwash. The sand wedge in B was sampled as Olson Quarry-2. Holes represent nesting sites holes created by cliff swallows. Photographs by RJS.

Table 1. Luminescence dates for the sand wedges and the Henning site.

Sample name	Sample number	Sample depth (cm)	No. aliquots[1]	Dose rate (Gy ka^{-1})[2]	Equiv. dose±2σ (Gy)[3]	Age±1σ (ka)
Sand wedge sites						
Wildenberg-1	USU-2900	105	21 (28)	1.25±0.05	18.35±1.71	14.69±1.38
Wildenberg-2	USU-3109	110	23 (44)	1.41±0.06	21.30±2.69	15.07±1.54
Olson gravel quarry-1	USU-3107	122	17 (49)	0.89±0.04	16.38±2.49	18.31±2.08
Olson gravel quarry-2	USU-2901	149	21 (28)	0.86±0.04	16.43±1.76	19.12±1.92
Olson gravel quarry-3	USU-3108	100	16 (44)	0.70±0.04	13.42±1.87	19.30±2.19
Loess site						
Henning	USU-2876	90	21 (28)	2.09±0.08	26.58±2.29	12.7±1.15

[1]Age analysis using the single-aliquot regenerative-dose procedure of Murray & Wintle (2000) on 2-mm aliquots of 125–212 μm quartz sand. Number of aliquots used in age calculation, with number of aliquots analysed in parentheses.
[2]Dose rate calculations use 5% water content. Samples from the Olson site use the weighted average of gamma-dose contribution from sediment within the wedge (90% contribution) and host gravel (10% contribution) due to narrower sand wedges (<40 cm). See Table 2 for raw data, Table S1 for proportional contribution to the dose rate and Fig. S1 for individual sample and wedge geometry.
[3]Equivalent dose (D_E) calculated using the central age model (CAM) of Galbraith & Roberts (2012).

entirely from the sand infill (Table 1). Two dose rate samples were recovered from each wedge site: one from the sands within the wedge and one from the surrounding host material (see Fig. S1).

A sixth OSL sample was collected from sand interbedded with loess at a the Henning site, where thin strata of fine sand of presumed aeolian origin interfinger with loess, across the upper metre of a 5-m-thick loess stack (Schaetzl *et al.* 2014; Fig. 5). Thick, silty loess lies to the south and east of this sharp ridge crest, whereas on the west and northwest sides of the ridge, loess is generally absent and instead, a thin (<50 cm) deposit of mixed sand and silt overlies sandstone bedrock. This type of sediment distribution has been generally interpreted, for this area, to indicate aeolian erosion of a pre-existing loess cover on the west and northwest slopes, assisted by saltating sand (Schaetzl *et al.* 2018). We targeted this site, based on the assumption that it may capture the end (or intersection) of a period of aeolian erosion, which in turn may have coincided with the aeolian interval that led to infilling of the wedges at the Wildenberg and Olson Quarries. Working in a 1.5-m-deep, hand-dug pit, we described the soil and sampled a 2E′ horizon at 90-cm depth, with a loamy sand texture, for luminescence dating.

Sample processing and OSL analyses were conducted at the Utah State University Luminescence Laboratory (USU-LL). Samples were processed to purify the quartz fine sand fraction (125–212 μm), following standard procedures involving wet sieving, gravity separation using sodium polytungstate ($Na_6O_{39}W_{12}$, 2.72 g cm^{-3}) and acid treatments with HCl and HF to isolate the quartz component. Samples were analysed following the latest single-aliquot regenerative-dose (SAR) procedures (Murray & Wintle 2000) on 1-mm-diameter (~20–40 grains per disc) aliquots. Ages were calculated using the central age model (CAM) of Galbraith & Roberts (2012). Data quality criteria included rejection of aliquots with repeat point signals outside of unity (>±10%), signal response in the zero-dose step (>10% of natural signal), and feldspar contamination as indicated by response to infrared stimulation (Rittenour *et al.* 2007). Data from rejected aliquots were not included in equivalent dose, overdispersion, or age calculations.

Luminescence ages, reported at 1σ standard errors (Table 2), were calculated by dividing the equivalent dose of radiation the sample received during burial (D_E) by the radioactivity of the surrounding sediment (dose rate). Dose rate values for each sample were calculated from U, Th, K and Rb contents, measured using ICP-MS and ICP-AES techniques, and conversion factors from Guérin *et al.* (2011). Contributions of cosmic radiation to the dose rate were calculated using sample depth, elevation, and latitude/longitude data, following Prescott & Hutton (1994). Dose rates assumed 5% long-term water content, based on data obtained from samples recovered in the field. Due to the complex geometry of the wedges and lack of *in-situ* field gamma spectrography measurements, we calculated the total dose rate for each sample using the wedge geometry and the dose rate contribution of the sand inside the wedge, as well as the host sediment outside the wedge (Table 2). Gamma dose rates were based on a distance weighted function. Due to the more localized influence of beta radiation, the beta dose rate calculations only used the radioactivity of the sediment from the wedge sediment, i.e. where the samples were collected. See the Supporting Information for further details.

Results

Host sediment

Previous research in the area indicated that most of the ventifacts and ice-wedge pseudomorphs are hosted in sediments that pre-date the Late Wisconsin advance

Table 2. Dose rate information for the sand wedges and the Henning site.

Sample name, lab. number	Dose rate fraction[1]	K (%)[2]	Rb (ppm)[2]	Th (ppm)[2]	U (ppm)[2]	Cosmic dose (Gy ka^{-1})[3]
Wildenberg-1	Sand wedge	0.86±0.02	21.5±0.9	1.7±0.2	0.6±0.1	0.20±0.02
USU-2900	Host gravel[4]	1.32±0.03	45.1±1.8	5.1±0.5	1.4±0.2	
Wildenberg-2	Sand wedge	0.93±0.02	27.0±1.1	2.0±0.2	0.6±0.1	0.19±0.02
USU-3109	Host gravel (sand)[4]	1.16±0.03	43.0±1.7	4.6±0.4	1.3±0.2	
	Host gravel (pebbles)[4]	1.66±0.04	44.6±1.8	6.0±0.5	2.5±0.3	
	Host gravel (cobbles)[4]	0.92±0.02	36.2±1.4	2.6±0.2	1.1±0.1	
Olson-1	Sand wedge	0.48±0.01	11.6±0.5	1.4±0.1	0.5±0.1	0.19±0.02
USU-3107	Host gravel (sand)	1.03±0.03	31.4±1.3	3.3±0.3	1.1±0.2	
	Host gravel (pebbles)	2.06±0.05	58.4±2.3	6.2±0.6	1.9±0.3	
Olson-2	Sand wedge	0.52±0.01	11.0±0.4	1.1±0.1	0.4±0.1	0.18±0.02
USU-2901	Host gravel	0.91±0.02	25.9±1.0	3.4±0.4	0.9±0.2	
Olson-3	Sand wedge	0.30±0.01	7.6±0.3	1.2±0.1	0.4±0.1	0.20±0.02
USU-3108	Host gravel (sand)	0.96±0.02	43.0±1.7	4.6±0.4	1.3±0.2	
	Host gravel (pebbles)	1.66±0.04	44.6±1.8	6.0±0.5	2.5±0.3	
	Host gravel (cobbles)	0.92±0.02	36.2±1.4	2.6±0.2	1.1±0.1	
Henning	Sand	1.57±0.04	40.0±1.6	3.2±0.3	0.9±0.1	0.20±0.02
USU-2876						

[1]See Table S1 for proportional contribution of each sediment unit and grain-size fraction to the total dose rate for each sample.
[2]Elemental concentrations determined using ICP-MS and ICP-AES techniques.
[3]Cosmic dose rate calculated using sample depth, latitude and elevation following Prescott & Hutton (1994).
[4]Host gravel >30 cm from the OSL sample and not included in the dose rate.

and lie outside of the Late Wisconsin moraine (Black 1964; Fig. 1). Across the study area, the main pre-Late Wisconsin sediment is the River Falls Formation, which is typically a reddish brown till or outwash deposit with sandy loam to sandy clay loam textures (Syverson *et al.* 2011). Although typically interpreted as a glacial till, the three sites at which we sampled pre-Late Wisconsin sediments (as the host material for the wedges) exhibited stratified, horizontally bedded, and highly weathered sand and gravel, with reddish hues (Fig. 7). Field-estimated gravel contents of this sediment ranged from ~5 to 20%. Data from 32 samples in the three quarries had a mean texture class (for the <2-mm fraction) of coarse sand and a mean sand:silt:clay ratio of 89:8:3, coarser than is typical for the River Falls Formation as reported by Syverson *et al.* (2011) (sand:silt:clay ratio of 60:15:25). The most common sand fractions were medium sand and coarse sand (see Table S1). Between-pit variability in textures was also low and showed a fair amount of textural uniformity, with textural modes commonly between 536 and 540 μm (coarse sand), and mean weighted particle sizes between 409 and 412 μm (medium sand) (Fig. 8, Table S1). Many pebble and cobble clasts within the formation are deeply weathered and easily broken by hand, supporting the interpretation of a pre-Late Wisconsin origin.

Thus, based on the clearly stratified nature of the sediment, its location on hilltops, its textures and its red colours, we conclude that the host material in the three sampled quarries is River Falls Formation outwash, typically with 5–20% gravel and abundant cobbles, and with broad but clear zones of clay formation and/or illuviation. This sediment would have covered the uplands at the locations where the wedges are located, and along with weathered sandstone bedrock and Late Wisconsin outwash in the Chippewa River valley, could have provided ample sand for deflation and transport.

Wedge morphology and sediments

The wedge structures in the Olson and Wildenberg sites are typically between 0.9 and 2.0 m deep and (maximally) 25–40 cm wide, with spacings of 3–15 m (Figs 9, 10). Not all are wedge-shaped (widest at the top); some are widest partway down, narrowing toward the top. Most are generally vertically oriented, although some are angled off the vertical. Many of the wedge sediments exhibit fine, vertically oriented strata and lamellae within the allochthonous, well-sorted, light-coloured sands (Fig. 10). Upturned strata in the host material are common near the margins of wedges (Fig. 11). Because both sites had been ploughed and disturbed in the past, we were unable to determine whether the wedges had originally extended to the surface.

Texture data from 32 samples, recovered from 15 different wedges, confirm that they are composed of well-sorted sandy sediment that lacks coarse (>2 mm) fragments (Fig. 8). The most typical texture is sand, the two most common sand fractions are medium and fine sand, and the typical particle size mode is between 290 and 350 μm (medium sand) (Table S1). The consistency among particle size distributions attests to their likely aeolian origin (Fig. 8). Samples of outwash of the River Falls Formation are much more variable in grain-size distributions and have wider ranges of grain sizes (Fig. 8, Table S1). The amounts of the finer sand fractions, vs. other sand fractions, in the wedge infill sediments is particularly noteworthy (Fig. 8). The typical fine+very

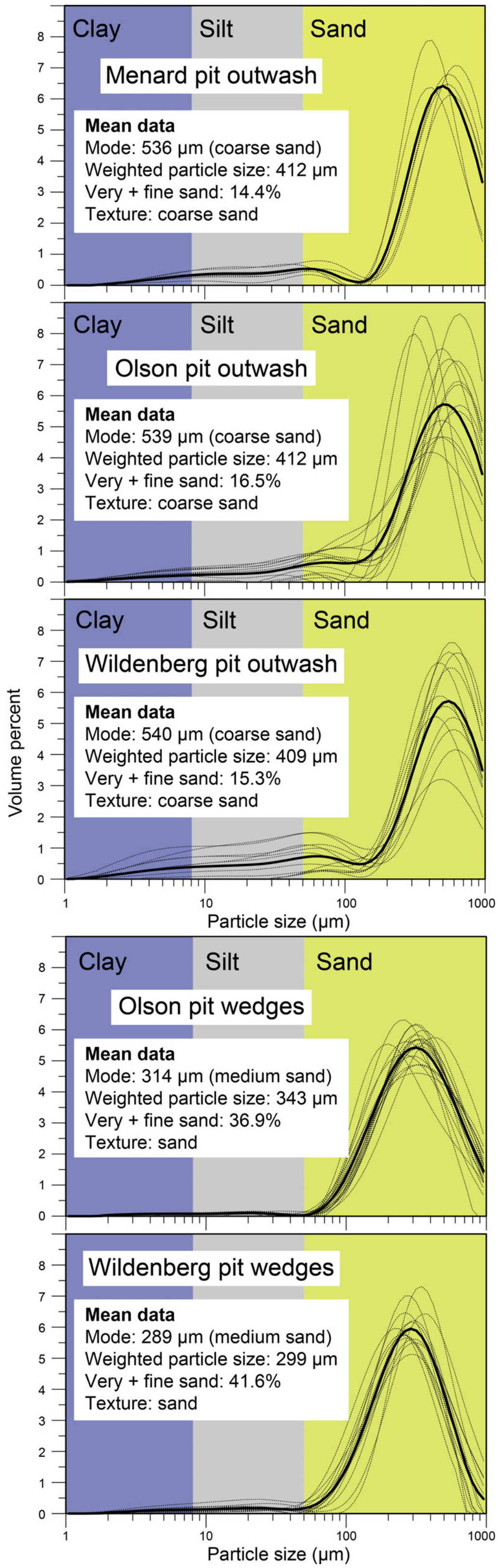

Fig. 8. Continuous texture curves (thick line: mean texture data) and other representative grain-size data for samples taken from the three gravel quarries formed in River Falls Formation outwash, and the wedge samples.

fine sand content of the host outwash ranges between 14 and 17%, whereas the wedge sands have between 36 and 42% fine+very fine sand. The increase in finer sands in the wedge infill, as well as its mean sand:silt:clay ratio of 97:2:1, clearly points to an aeolian origin.

Chronology

Luminescence age estimates from the five sand wedges support an interpretation of permafrost – either widespread or in preferred locations – between *c.* 19 and 15 ka (Table 1, Fig. 2). Low overdispersion in the OSL results (10–24% scatter, see Fig. S2) indicates that the wedge sediments were well bleached at the time of deposition, as is typical of aeolian sands. The OSL results in Table 1 should be interpreted to represent two locations where aeolian sand transport coincided in time with permafrost and periglacial processes. These data do not contradict previous studies (Black 1964; Attig & Clayton 1986; Mason & Knox 1997; Clayton *et al.* 2001; Batchelor *et al.* 2019; Mason *et al.* 2019), but instead, support and tighten up the chronology for the permafrost interval in the upper Midwest, USA, during the waning stages of MIS 2. Transport of aeolian sand continued through and beyond this ~4000-year time window, as the OSL age for the aeolian sand at the Henning site is 12.7 ka (Table 1). It is also important to note that the three ages from the Olson Quarry are considerably older, clustering at 19.3–18.3 ka, whereas the ages from the Wildenberg site are slightly younger and date to 14.7–15.1 ka.

Interpretations

Wedge origin

Based on their morphology (downward-tapering features, lack of collapse structures around the wedges or infill by surrounding host sediments, upturned stratigraphy in the host sediment, and with depths sufficient to have extended into permafrost) and sedimentology (primary infilling of well-sorted, fine sands, lacking coarse fragments), as well as the inferred palaeoenvironmental conditions of the Chippewa Valley in the immediate postglacial period, we interpret the features exposed in the Wildenberg and Olson Quarries to be prototypical sand wedges. Murton (1996) described sand wedges as having simple V-shapes, but often also having slightly curved sides that are convex outward, similar to many of the pseudomorphs we studied (Figs 9, 10). According to Murton (1996), sand wedges are typically filled with well-sorted fine sands, along with some silt. The sand wedges under study here are almost silt-free, but are dominated by fine and very fine sands (Fig. 8). Bateman (2008) noted that sand wedges may also contain vertically aligned veins of sand, which occur in the sand wedges at some of our sites (Fig. 10). The ice and sand in

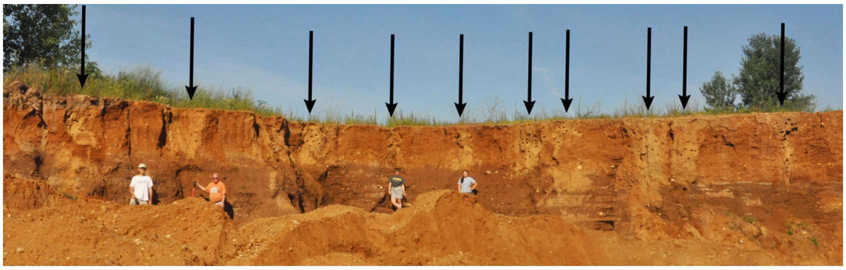

Fig. 9. Wide view of the main exposure at the Wildenberg gravel quarry in 2018. Arrows indicate wedges exposed in the quarry face.

Fig. 10. Images of the five sand wedges sampled in this study. Attempt has been made to scale each image similarly. Tape units are in cm.

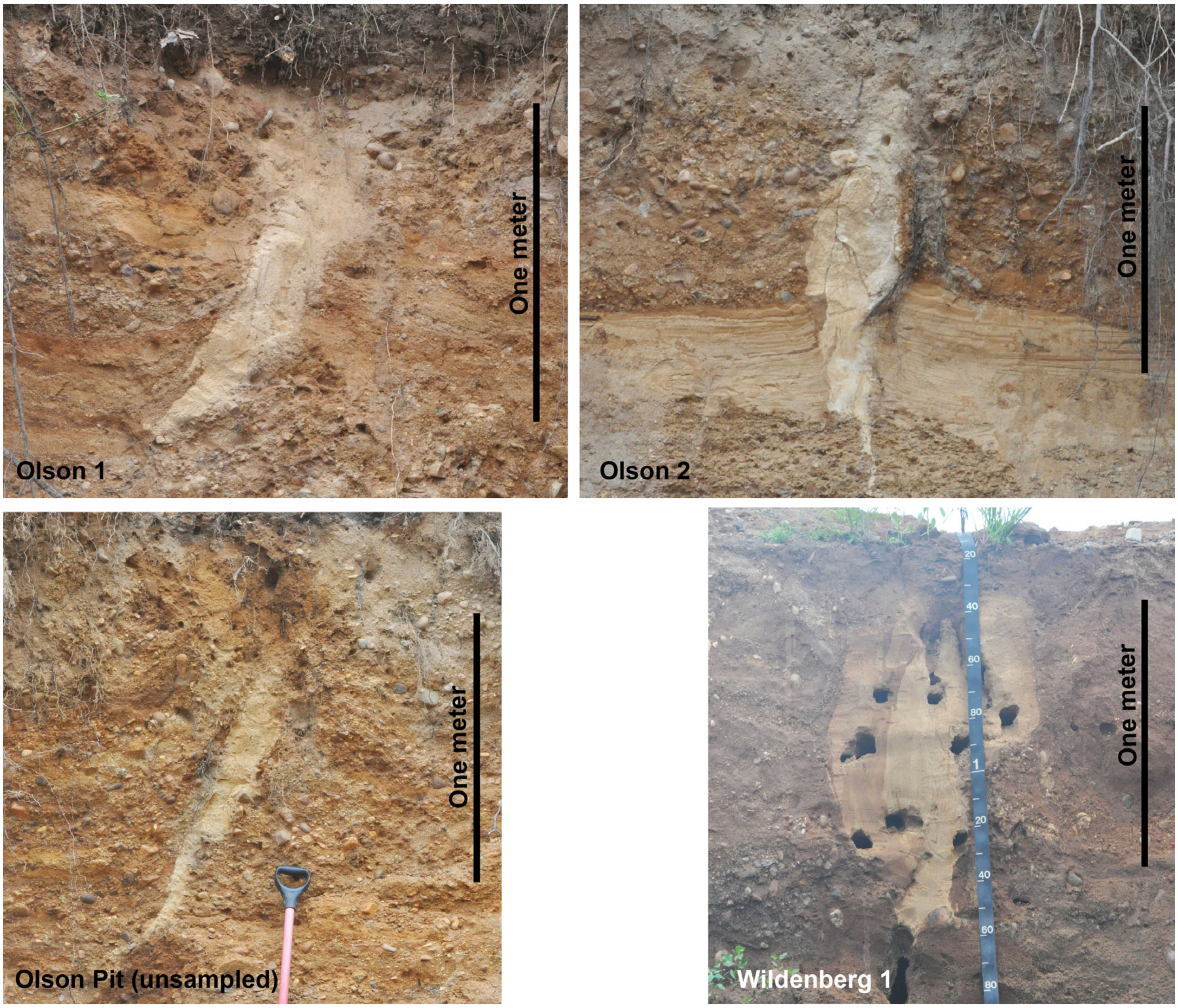

Fig. 11. Examples of upturned strata at the margins of sand wedges at the study sites.

these types of pseudomorphs commonly deform the surrounding sediment, leading to upturned strata (Murton & French 1994; Fig. 11).

Based on the landscape associated with sand wedges forming in the Northwest Territories of Canada, Wolfe *et al.* (2018) identified the optimal environmental conditions for their formation: sandy surfaces with thin snow cover and high thermal conductivities, which promote cold ground temperatures with rapid cooling, along with actively migrating aeolian sands. Similar conditions were invoked for sand wedges in Belgium (Buylaert *et al.* 2009). The lack of loess cover across much of the Chippewa Valley suggests that similar conditions may have been present in the immediate postglacial period, i.e. saltating sand effectively deflated much of the pre-existing loess cover (Schaetzl *et al.* 2018).

The sand wedges at the two study sites differ from syngenetic sand veins, described by Murton & Bateman (2007), which get increasingly narrow upward and which form on aggrading surfaces of aeolian sand. Instead, these sand wedges often exhibit distinctively wide tops, and therefore appear to be anti-syngenetic features, having formed and grown downward during deflation of an otherwise stable ground surface (Murton & Bateman 2007).

Glacial history

Permafrost in the study area likely started to form during the advance of MIS 2 ice, and reached its maximum extent at the LGM (Fig. 2). OSL dates reported by Attig *et al.* (2011) and Carson *et al.* (2012) indicate that the Green Bay Lobe had reached its maximum extent at *c.* 26 ka, with an initial retreat between 23.5 and 18.5 ka, depending on location. Retreat of the Green Bay Lobe was episodic, but with a notable re-advance to within 200 km of the study sites during the Two Creek interstade, at *c.* 13.5 ka (Leavitt & Kalin 1992; Rech *et al.* 2012). To the west, an early advance of the Des Moines Lobe is known but poorly constrained, with its furthest advance (shown in Fig. 1) at *c.* 16.5–17 ka. Ice remained on the Iowa landscape, but in reduced extent, until about 15 ka (Ruhe 1969; Clayton & Moran 1982; Hallberg & Kemmis 1986; Bettis *et al.* 1996; Rittenour *et al.* 2015; Heath *et al.* 2018). The retreat chronology of the ice lobes that border the study area to the north, especially the Chippewa Lobe (Fig. 1), is less well constrained. However, a recent study by Ullman *et al.* (2015) concluded that initial retreat of the Chippewa Lobe began by *c.* 23.7 ka, and that ice remained in northern Wisconsin and the western Upper Peninsula of Michigan until perhaps as late as 13.2 ka.

Permafrost and aeolian activity

Previous work suggests that an ice margin was as close as 24 km to, and never farther than 200 km from, the two quarries between *c.* 26 and 13–14 ka. As shown in Table 1, data constraining the permafrost interval across the wider region, beyond the MIS 2 ice margin, provide a similar, but much less well-constrained, interval: between *c.* 25–24 and 14–12 ka. The OSL ages on sand wedges from the Chippewa Valley area provide an additional window into not only the permafrost interval, but a period of intense aeolian activity, and they show good correspondence to this established time frame (Table 1).

If we assume that the sand wedges formed closer to the end than the beginning of the permafrost interval, then our OSL results point to the demise of permafrost coverage in the Chippewa Valley region between *c.* 19–15 ka (Table 1). The dates from the more northerly Wildenberg site appear to represent the endpoint of permafrost in this area, at *c.* 15 ka, whereas dates from the Olson site indicate that permafrost was present in the north-central Driftless Area until at least 19 ka (Fig. 2, Table 1). The Olson site is 63 km southeast of the Wildenberg site and >40 km from the ice margin (at its closest point). The more complex (wider, more contorted) morphologies of the sand wedges at the Wildenberg site also suggest that permafrost here may have been more intense and/or more continuous. At its closest, the MIS 2 ice margin would have been only 24 km from the Wildenberg site. Nonetheless, with only five OSL dates from two sites, our ability to infer any amount of detail on permafrost distribution or the time window within which it existed, is limited.

Aeolian sand deposition – driven upslope and deposited at the crest of the Henning site at 12.7 ka – indicates that dry and windy conditions, ample for the transport of aeolian sand, continued in the Chippewa Valley at least until this time.

Discussion

OSL ages on sediments from sand wedges in the study area indicate that conditions were cold enough to form contraction cracks under permafrost conditions between *c.* 19 and 15 ka. They provide limiting dates for permafrost conditions within this region; permafrost likely did not persist much longer than these dates. Concomitantly, aeolian sand was migrating, at least locally, across the landscape at that time, and likely longer.

We suggest that the sand wedges capture a palaeoenvironmental signature of widespread (but possibly locally thawing) permafrost, coupled with windy conditions on a dry, treeless landscape. Because each site is located on a low hill, it captures the regional climate better than might other types of ice-wedge pseudomorphs, such as those in lowlands, which may be more reflective of local conditions.

Few palynological data exist for the Chippewa Valley to provide a picture of the regional conditions during the interval within which our sand wedges formed. Nonetheless, data for sites upwind (west) and outside of the Chippewa Valley within (and slightly earlier than) this general time period can provide some insight. At a site in southeastern Minnesota, Baker *et al.* (1999) concluded that the landscape supported sub-arctic to arctic vegetation at 22.6 ka. Their pollen spectrum indicated tundra-like conditions with permafrost, and local environments reconstructed from both physical and palaeobotanical evidence pointed to wind-swept ridge tops with thin loess along with bedrock outcrops on valley walls. At the Wolf Creek site in Minnesota, northwest of the Chippewa Valley, Birks (1976) reported that tundra barrens were present from 24.7 to 17.9 ka, with a warming climate thereafter. Here, the earliest sediment influx is sand, which drops off at *c.* 14 ka. The very low accumulation rates at the Wolf Creek site suggest that that the sediments here did not originate from active solifluction and mass movement, but from aeolian activity. Birks (1976) concluded that the landscape near the Wolf Creek site was virtually treeless between 24.7–17.9 ka.

Based on the distribution of sand dunes and loess, Schaetzl *et al.* (2018) concluded that strong west-northwesterly winds were a common occurrence in the Chippewa Valley in the immediate postglacial period. OSL ages on several of the hundreds of parabolic dunes, sand sheets, sand ramps, and sand stringers in the Chippewa Valley roughly bracket the OSL date from the Henning site, indicating that aeolian activity was ongoing between *c.* 13 and 9 ka (Schaetzl *et al.* 2014, 2018; Millett *et al.* 2018; Mataitis *et al.* 2019; Millett 2019). Recall that the OSL ages of these sandy aeolian landforms represent the waning phase and stabilization of sand transport, as climate warmed and vegetation cover and soil moisture increased across the Chippewa Valley. The beginning of this extended period of aeolian activity is less well constrained. Regardless, sand transport was extensive across the Chippewa Valley in the postglacial period, and the Henning date supports the conclusions of Schaetzl *et al*. (2018), i.e. by mobilizing sand, strong winds in the Late Pleistocene were able to deflate some pre-existing loess deposits and transport the loess downwind.

Taken further, sediment characteristics and the OSL date from the Henning site (Figs 1, 12) suggest an important link between the widespread aeolian activity in the Chippewa Valley and permafrost conditions. The Henning site lies at the border between thick (>5 m) loess to the immediate southeast (downwind) of the ridge crest, and a bedrock surface with only a thin cover of aeolian sand and residuum, to the north and west (Fig. 12). On the ridge crest, the upper metre of loess is intercalated with aeolian sand, presumably driven up the windward slope and into the loess that was pre-existing there. Within a few tens of metres to the south and east, sand is absent in the thick loess mantle (Schaetzl *et al*. 2014). Sand at the Henning site dates to 12.7 ka, falling near the end of the permafrost interval (Table 1, Fig. 2). Even though sand deposition at this site may signal the waning stages of aeolian activity across the Chippewa Valley, aeolian activity was still strong enough to transport sand upslope. By this time, sand wedges in the valley were filled with aeolian sand.

Given the age distributions of the sand wedge sediment and the aeolian sand at the Henning site, we conclude that the widespread aeolian activity in the Chippewa Valley appears to have coincided with a sandy landscape underlain by permafrost, and later, the beginning of permafrost degradation. The end of this interval was probably coincident with warming conditions associated with the continued retreat of the Late Wisconsin ice.

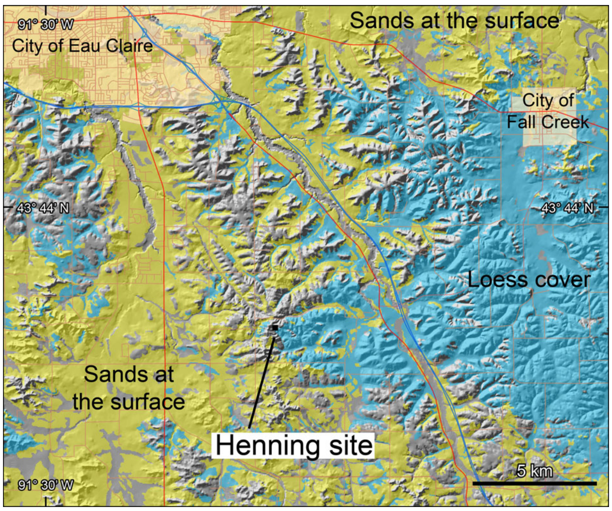

Fig. 12. Shaded relief map of the landscape near the Henning site. Note that the loess cover is generally restricted to areas east and south of the largest ridges, and that the Henning site is at the intersection of the thick loess landscapes to the east, and the loess-poor, sandy landscapes to the west. Areas in grey generally reflect nearly barren bedrock surfaces.

Conclusions

Five OSL dates are provided from two sites in western Wisconsin. Both sites have abundant sand wedges, hosted in River Falls Formation outwash, which pre-dates the LGM. Wedge morphology and sedimentology indicate that they had formed when the landscape was covered with permafrost, and are infilled with aeolian sand. Thus, the dates reflect cold, windy conditions in the Chippewa Valley and adjoining areas. Using the 1σ uncertainties on the five dates presented in Table 1, the age range for sand wedge formation (and hence, permafrost) and aeolian activity across the study area could have been as wide as 21.5–13.3 ka or as narrow as 17.1–16.1 ka. The most likely end of permafrost in this area, based on OSL dates from the sand wedges, is between *c*. 19.4 and 14.7 ka, but later in the northern parts of the valley (nearer the Wildenberg site).

Aeolian activity was ongoing in the valley until *c*. 12 ka (13.9–11.6 ka, using 1σ uncertainties), based on an OSL date from the Henning site on a ridge crest immediately downwind from the Chippewa River. Aeolian sand here was driven up a long, steep slope, where it is intercalated with loess. Numerous other dates, published on aeolian landforms in and near the Valley, suggest that aeolian activity was ongoing in the Chippewa Valley, at least intermittently, until 9 ka.

Our study affirms the utility of dating sand wedges with OSL, and thus, provides the first dates on features directly associated with periglacial activity in and near the Chippewa Valley. Although our data cannot ascertain the timing of the onset of permafrost, they clearly indicate that uplands were often affected by both permafrost and strong winds, dramatically shaping the landscape.

Acknowledgements. – We thank Doug Wildenberg and William Olson for access to their quarries, Joann and Roger Henning for access to their ridgetop site, and Todd Meyer for arranging permissions and access. We also thank Nicole Marcou and Adria Slade, undergraduate students at UW-Eau Claire who helped identify possible research sites, as well as field assistance. Jarrod Knauff and Chris Baish performed many of the lab analyses. Samantha Kaplan provided grain size and mineralogy data on sediment from several of the sand wedges, when the project was just starting. We acknowledge the support given to this project from the UW-Eau Claire Office of Research and Sponsored Programs, and Minnesota State University, Mankato. Steve Wolfe, Kent Syverson and John Attig kindly reviewed an earlier draft of this paper. Lastly, we acknowledge the helpful comments made on the paper during the review process by Mark Bateman and an anonymous reviewer.

Author contributions. – RS oversaw the research, performed the particle size analyses, and wrote the paper, with input from all authors.

GR located the sites and assisted in sampling. PL and DF assisted in sampling in the field. TR performed the OSL analyses, and CY performed the plant macrofossil analyses. The authors declare that there are no conflicts of interest with third parties.

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Supporting Information

Additional Supporting Information may be found in the online version of this article at http://www.boreas.dk.

Data S1. Optically stimulated luminescence (OSL) methods.

Fig. S1. Locations of OSL samples within sand wedges and geometry of contribution areas within and outside of sand wedges, for dose rate calculations.

Fig. S2. Equivalent dose (D_E) distributions and overdispersion (OD) values for each sample.

Table S1. Beta, gamma and cosmic contributions to total dose rate.

Table S2. Summary particle size data for the sediments studied.