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Invited Review

Up in the refrigerator: Geomorphic response to periglacial environments in the Upper Mississippi River Basin, USA

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ABSTRACT

James C. Knox was best-known for his work on Holocene and historical changes in fluvial systems, but he also had a long-standing interest in the effects of late Pleistocene periglacial environments on landscape evolution in parts of the Upper Mississippi River basin that were just outside the Laurentide Ice Sheet margin, or as Knox put it, 'up in the refrigerator.' Knox and others in the Quaternary community of the Midwestern U.S. often suggested that hillslope erosion was accelerated under periglacial conditions, so that glacial periods have had a dominant effect on the landscape we see today. This paper reviews the evidence and reasoning supporting that view in a study area of the Upper Mississippi basin bordered on three sides by ice margins of the last glaciation, including the Wisconsin Driftless Area and adjacent landscapes. Sparse but compelling paleoecological data and relict ice- or sand-wedge polygons provide clear evidence for a cold climate and widespread permafrost around the peak of the last glaciation. In highly dissected, relatively high-relief parts of the study area, the loess and soil stratigraphy on ridgetops and the colluvial mantles on steeper slopes are best explained by highly effective hillslope erosion, including solifluction, during and just after the Last Glacial Maximum. Knox used the post-depositional truncation of a loess unit to quantify contrasting late Pleistocene and Holocene sediment yields from a small Driftless Area watershed. While the late Pleistocene yield indicates accelerated erosion, it is still lower than modern sediment yields in many tectonically active or semiarid landscapes, and it may reflect deposition of highly erodible loess as well as effects of periglacial conditions.

The views of Knox and other Midwestern geomorphologists on landscape evolution through glacial–interglacial cycles were highly influenced by the work of Robert V. Ruhe. Ruhe proposed that an episode of widespread erosion during and just after the Last Glacial Maximum can explain enigmatic aspects of Quaternary stratigraphy and the soil landscape on the Iowan Erosion Surface, a very low relief landscape of the study area. Ruhe's key evidence is still valid, though it needs to be separated from an implausible model of landscape evolution. Interpretation of the Iowan Erosion Surface and other low-relief landscapes just outside the ice margin also requires recognition of the profound effect of eolian processes on those landscapes under periglacial conditions. Many new insights on landscape evolution in the study area could result from wider application of cosmogenic nuclide-based methods to assess glacial–interglacial changes in basinwide rates of erosion and residence time of soils. Just as important, a need exists for much more field-based characterization of hillslope, fluvial, and eolian sediments for comparison with those of modern permafrost regions and past periglacial environments in Europe.

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1. Introduction

Throughout his long career as a geomorphologist, James C. Knox (1941–2012) worked in the tectonically stable, low- to moderate-relief landscapes of the North American midcontinent, especially in the Driftless Area of southwestern Wisconsin, USA (Fig. 1). While the relative contribution of such landscapes to global sediment flux is debated (Willenbring et al., 2013; Larsen et al., 2014b), they clearly represent an opportunity to study the effects of climate change on sediment yield, especially where they have not recently been affected by glaciation or reorganization of the drainage network (Hidy et al., 2014). Knox is best known for his work on the response of fluvial systems to Holocene climate change and human impacts, but he had a long-standing interest in effects of glacial–interglacial climatic change on rates of erosion, sediment yield, and overall patterns of landscape evolution. On field trips through the Driftless Area (Fig. 2), Knox would almost always reserve time to talk about the hillslopes of that region, and their response to periglacial conditions when the region was 'up in the refrigerator' during the last glaciation. He interpreted hillslope deposits in the Driftless Area as evidence of accelerated hillslope erosion in the late Pleistocene when the region was just outside the margin of the Laurentide Ice Sheet (Knox and Maher, 1974; Knox, 1983, 1989; Leigh and Knox, 1994; Mason and Knox, 1997). Whereas Knox was not the first to argue for a strong influence of past periglacial environments on the Driftless Area (Smith, 1949, is an early example of interest in the topic), his prominence in the Midwestern Quaternary community made him an especially important advocate of that view, shared by many of his colleagues in the region.

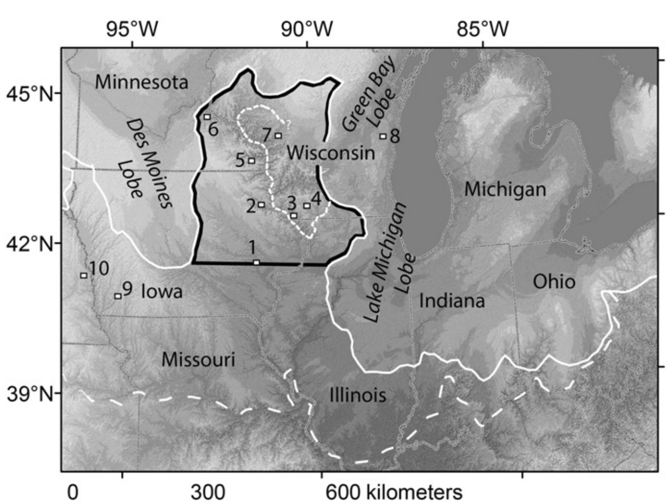

Fig. 1. Location of the study area (solid black line) of this paper in the central U.S., relative to the outermost ice margins of the last glaciation (solid white line) and of earlier late Cenozoic glaciations (long-dashed white line). Wisconsin Driftless Area, within the larger study area, outlined by short-dashed white line. Background gray shades represent elevation. Three ice-sheet lobes mentioned in text are labeled. Sites mentioned in text or shown in figures are marked by numbers: 1. Conklin Quarry, 2. Elkader site, 3. Kieler site, 4. Moscow Fissure, 5. Houston County area, Minnesota (Jore site, Lehman site, Caledonia roadcut, BE Site), 6. Hampton Town Hall site, 7. Sparta ice-wedge cast site, 8. Valders Quarry, 9. Study area of Ruhe et al. (1967) in southwestern Iowa, 10. Study area of Mason et al. (2007) in eastern Nebraska. The Wolf Creek paleoecological site (Birks, 1976) is just off north edge of map.

This paper first reviews the evidence and reasoning that underlie this view of landscape response to late Pleistocene periglacial conditions in the Driftless Area and surrounding regions. I use the term periglacial to encompass cold-climate environments in which ground freezing and thawing strongly affect geomorphic processes. As noted by French (2007, p. 5), 'permafrost is a central, but not defining, element' of periglacial geomorphology; and deep seasonal frost can potentially explain some, though not all, of the landforms and sediments that have been attributed to past periglacial environments. Therefore I begin by reviewing paleoecological and geomorphic evidence on the specific nature of the periglacial environment in the region of interest, including clear evidence for widespread permafrost. This is followed by a review of hillslope and fluvial stratigraphy in the Driftless Area and similar landscapes, interpreted by Knox as the result of accelerated erosion under periglacial conditions. I then turn to the work of Robert V. Ruhe (1918–1993), focusing on his study of a distinctive low-relief landscape outside the late Pleistocene glacial limits in Iowa that was also of great interest to James Knox. Though rarely cited in contemporary geomorphological literature, Ruhe was highly influential from the 1960s through the 1980s among Midwestern Quaternary geologists and pedologists. Knox worked with him in the field while in graduate school and continued to cite Ruhe's ideas throughout his career (e.g., Knox,

Fig. 2. James C. Knox (second from right) at an exposure of Holocene alluvium, on a field trip to the Driftless Area in 2011.

1996). Ruhe's key insights need to be clearly separated, however, from his uniquely dogmatic conceptual model of landscape evolution and his reluctance to fully acknowledge the importance of climate as a control on landscape evolution, particularly where past periglacial conditions are involved. I also emphasize how recognition of the profound impact of eolian processes during the last glaciation can provide new insight on low-relief landscapes like the one in northeastern Iowa that Ruhe studied. A final section of the paper highlights two quite different avenues of research that could yield valuable new insights on the evolution of landscapes outside the ice margins in the Upper Mississippi valley, when they were 'up in the refrigerator.'

2. Geographic and temporal focus

2.1. Study area

This paper is focused largely on the area outside the limits of the Laurentide Ice Sheet during Oxygen Isotope Stage (OIS) 2, east of the Des Moines Lobe of the ice sheet, west of the Green Bay and Lake Michigan lobes, and north of about N 41°30′ latitude (Fig. 1). This area of interest is entirely drained by the Upper Mississippi River and its tributaries. It includes the Wisconsin Driftless Area, where no evidence exists of late Cenozoic glaciation (Mickelson et al., 1983), along with surrounding landscapes that were affected by glaciation prior to the last glacial cycle. Illinoian (OIS 6) glaciation covered parts of the study area in northern Illinois and possibly western Wisconsin and southeastern Minnesota, but much of the study area was affected only by pre-Illinoian glaciation (Goebel et al., 1983; Lineback et al., 1983; Hallberg et al., 1991; Syverson and Colgan, 2011). Multiple pre-Ilinoian glaciations from around 2.4 Ma into the middle Pleistocene extended far south of the study area (Roy et al., 2004; Balco and Rovey, 2010), but fluvial dissection has removed almost all glacial landforms from areas of Illinoian or older glaciation.

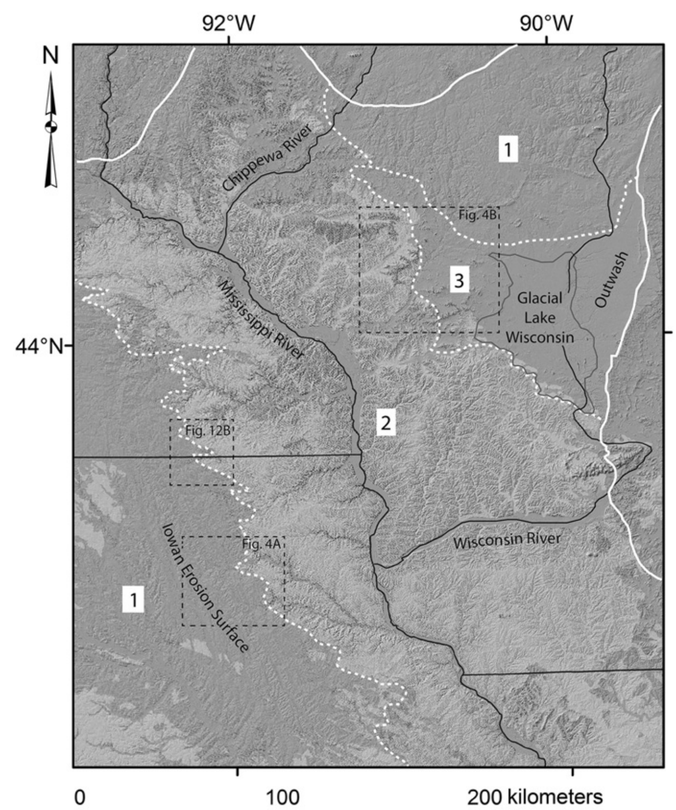

Fig. 3. Contrasting landscapes of the central part of the study area, portrayed with shaded relief image from 30-m digital elevation model (USGS National Elevation Dataset). Light gray shade indicates areas of loess thicker than about 1 m, interpreted from soil survey data (STATSGO, Natural Resource Conservation Service, USDA). Solid white lines indicate outermost ice margins of the last glaciation. Major landscape types separated by dashed white lines and labeled by numbers: 1. Low-relief landscapes underlain by pre-Illinoian glacial sediment or shallow bedrock, mantled with thin (<1) loess or loam sediment with patches of eolian sand (including Iowan Erosion Surface); 2. Relatively high-relief, highly dissected bedrock-controlled landscapes with no evidence of glaciation or with patchy remnants of pre-Illinoian glacial sediment, ridgetops mantled by thick loess and slopes partly covered with colluvium; 3. Central Sand Plain of Wisconsin, very low-relief landscapes mantled with eolian sand, with scattered narrow sandstone ridges (portions of sand plain occupied by bed of Glacial Lake Wisconsin and late Wisconsin outwash are labeled). Boxes indicate areas portrayed in Figs. 4A, B, and 11B.

For the purposes of this paper, it is more useful to subdivide the study area by topography and loess cover rather than by glacial history. Three general types of landscapes can be identified within the study area (Figs. 3 and 4). The first (labeled 1 in Fig. 3) is made up of low-relief landscapes where pre-Wisconsin glacial sediment is still fairly extensive, though with significant areas of shallow bedrock. Glacial deposits and bedrock are generally covered with thin (<1 m) loess, or less commonly eolian sand or loam surface mantles. Fig. 4A illustrates the typically abrupt transition between the first and second types of landscapes. The second type (2 in Fig. 3) includes areas with greater relief (up to about 160 m km^{-2}), steep slopes, and deeply incised valleys, in which glacial sediment is either entirely absent or preserved only in small patches. This type of landscape has moderately to very thick loess cover on summits and colluvium-mantled hillslopes. Much of the Driftless Area fits within this second landscape type, along with bordering regions with very similar topography that were affected by pre-Illinoian glaciation.

The third type of landscape is the Central Sand Plain of Wisconsin (3 in Figs. 3; 4B). The lowest and flattest part of this mostly low-relief plain is the former bed of Glacial Lake Wisconsin (Clayton and Attig, 1989). To the west of the lake bed, the plain is underlain by shallow sandstone bedrock. This bedrock landscape and the lake bed are covered with a sandy mantle that is probably largely eolian, as discussed below. A westward-sloping late Wisconsin outwash plain separates the bed of Glacial Lake Wisconsin from the moraine that marks the east side of the study area (Figs. 3, 4B). Isolated narrow ridges and pinnacles of sandstone rise above all parts of the sand plain. The Central Sand Plain is generally considered to form the northernmost part of the Driftless Area, although the limit of glaciation in that area is poorly defined.

Extensive fill terraces line major river valleys in all parts of the study area, and even occur along many smaller streams. Within the study area at present, mean annual temperatures range from 6 to 10 °C, and mean annual precipitation ranges from 800 to 980 mm. The natural vegetation of the study area during the Holocene included forest, mainly to the east, and tallgrass prairie, mainly to the west.

2.2. Time period of interest

Most of the physical evidence of landscape response to periglacial environments in the study area is from the time spanning the Last Glacial Maximum (LGM) and the subsequent deglaciation, between about 30 and 11.5 cal ka, or OIS 2 and latest OIS 3 in terms of global oxygen isotope stages. Much of the literature relevant to this paper refers to roughly the same time period as the late Wisconsin (or late Wisconsinan) substage in chronostratigraphic terms, and I will use that term here as well, where a more specific time reference is not possible. Reference to chronology in this paper is entirely in terms of approximate age ranges, using 'cal ka' to emphasize that all are either based on calibrated ^{14}C ages or result from other dating methods thought to give directly comparable results. When I cite age ranges originally given in ^{14}C years, I have estimated equivalent ranges in cal ka using the IntCal13 calibration curves as implemented in Calib 7.1 (Reimer et al., 2013).

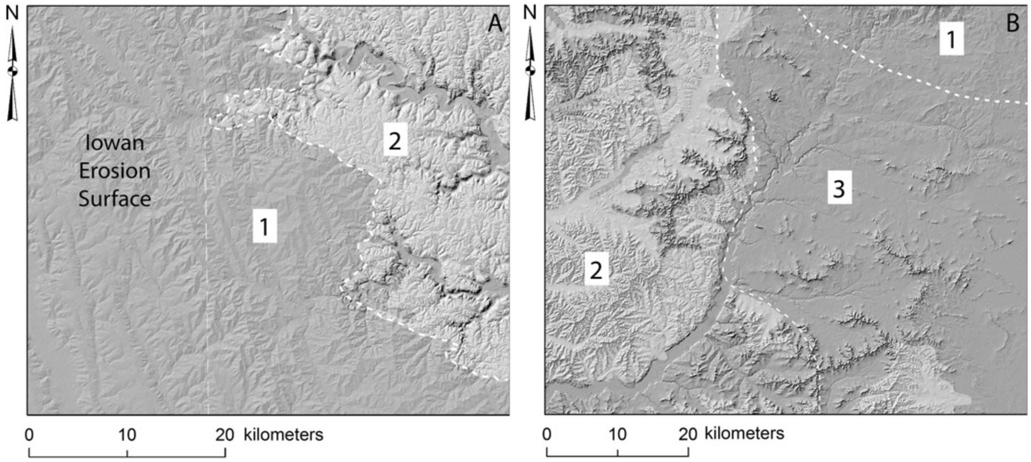

Fig. 4. Examples of topographic change across boundaries between the contrasting landscapes shown in Fig. 3. Source of shaded relief image same as in Fig. 3, and light gray shades indicate loess thicker than about 1 m. (A) Eastward transition from low-relief landscape with thin loess over glacial sediment or bedrock (Iowan Erosion Surface) to higher-relief dissected landscape mantled by thick loess and colluvium. (B) Eastward transition from higher-relief thick loess- and colluvium-mantled landscape to Central Sand Plain, with area of low-relief thin loess-mantled glacial sediment and bedrock to the northeast.

3. Environment of the study area during the Last Glacial Maximum and the late-glacial period

3.1. Proximity to the ice sheet and paleoecological evidence

During the LGM, lobes of the Laurentide Ice Sheet bordered the study area to the west, north, and east, though the available data suggest differences in timing of maximum extent and retreat of various lobes. Overall though, the ice sheet bordered the study area from before 25 cal ka to as late as 16.5 cal ka and was nearby until after 14 cal ka. An early advance of the Des Moines Lobe west of the study area occurred between about 44 and 30 cal ka (Bettis, 1997). That lobe then retreated and readvanced to reach its maximum extent, not far west of the study area, at about 16.5–17 cal ka. It then retreated somewhat, but was still in close proximity until about 14 cal ka (Bettis et al., 1996). Optically stimulated luminescence (OSL) dating at two sites suggests the Green Bay Lobe reached its maximum limit bordering the east side of the study area around 26 cal ka (Attig et al., 2011; Carson et al., 2012). Ullman et al. (2015) concluded from terrestrial cosmogenic nuclide (TCN) dating that retreat of the Chippewa and Green Bay lobes began at about 23 cal ka, but Attig et al. (2011) and Carson et al. (2012) placed the start of the Green Bay Lobe retreat somewhat later at about 18.5 and 21 cal ka, respectively. Retreat was interrupted by readvances, including a well-dated final readvance east of the study area around 13.5 cal ka (Leavitt and Kahn, 1992; Hooyer, 2007). The Lake Michigan Lobe reached its Marengo moraine bordering the study area around 29–30 cal ka; this lobe as a whole retreated after about 23 cal ka (Curry and Petras, 2011).

Paleoecological evidence from the time when the ice sheet was at its local maximum limit remains sparse but provides a relatively coherent picture of environments similar to the modern low Arctic tundra or tundra–forest transition. The Conklin quarry site (Fig. 1) at the southern edge of the study area in Iowa, has yielded a rich variety of data from plant macrofossils, pollen, bryophytes, insects, molluscs, and small mammal remains dated to 20–22 cal ka, indicating the presence of tundra and boreal forest taxa (Baker et al., 1986). The environment at that time is interpreted as a *Picea–Larix* krummholtz with tundra openings. Tundra plant macrofossils occur in sediments of small ice-walled lakes in stagnant ice of the Lake Michigan Lobe near the southern edge of the study area and yield ages between 21.7 and 16.2 cal ka (Curry et al., 2010). Tundra species dominated a sparse assemblage of plant macrofossils at the Jore site (Fig. 1) in southeastern Minnesota, with an age of 22.6 cal ka (Baker et al., 1999). Small mammals with tundra affinities were present in full-glacial material from the Elkader site in northeastern Iowa and the Moscow Fissure site in the Wisconsin Driftless Area (Fig. 1), along with an arctic/subarctic insect assemblage at the former site and mammals with forest and grassland affinities at the latter site (Foley, 1984; Woodman, 1982). Pollen and plant macrofossils at the Wolf Creek site, northwest of the study area, record a tundra environment from about 24 to 17 cal ka (Birks, 1976) in an area north of the outer ice sheet limits shown in Fig. 1, but presumed to be exposed by local glacial retreat. Tundra plants were present as late as 14.2 cal ka east of the study area at Valders quarry in eastern Wisconsin (Maher et al., 1998), but forest was established before about 13 cal ka at sites in and around the study area in south-central and eastern Wisconsin, northern Illinois, and northeastern Iowa (Chumbley et al., 1990; Baker et al., 1992; Leavitt and Kahn, 1992; Gonzales and Grimm, 2009).

3.2. Ice-wedge casts and other geomorphic evidence for permafrost

James Knox appreciated the importance of paleoecological data but always emphasized geomorphic evidence on the environment of the last glaciation, especially ice-wedge casts and hillslope deposits. In this, and in his overall emphasis on the importance of periglacial processes in shaping the landscape of the study area, he was probably strongly influenced by his interactions with European geomorphologists. Knox had a particularly long-standing friendship with Leszek Starkel, a Polish fluvial geomorphologist. Starkel's emphasis on late Pleistocene periglacial environments as a major factor in the long-term dynamics of fluvial systems (e.g., Starkel, 2003; Starkel et al., 2007), most likely influenced the views of Knox on the importance of past periglacial conditions in the Midwestern U.S., and increased his appreciation of the substantial body of research on relict periglacial landforms and sediments in Poland and more broadly across central and western Europe (see French (2003) for more historical background on the importance of research on periglacial processes in Poland and the significance of the journal *Biuletyn Peryglacialny* for the broader development of this field).

Ice-wedge casts – vertical wedges of sediment interpreted as having replaced the ice wedges that are common in regions of continuous permafrost (Harry and Gozdzik, 1988; French, 2007, pp. 310–315) – have been documented at numerous sites in the study area and surrounding regions (Black, 1965; Wayne, 1967, 1991; Ruhe, 1969; Clayton and Bailey, 1970; Péwé, 1973; Johnson, 1990; Bettis and Kemmis, 1992; Mason et al., 1994a; Walters, 1994; Clayton et al., 2001). If correctly identified, these features provide unambiguous evidence for the past presence of permafrost, although the exact implication in terms of paleotemperature is debated (Murton and Kolstrup, 2003; French, 2007, p. 310). Walters (1994) made a particularly detailed and compelling case for interpreting sediment-filled wedges in northeastern Iowa as ice-wedge casts, showing that they form polygonal networks and that beds of host sediment are often either upturned or downturned adjacent to the wedges. Other ice-wedge casts identified in the region (e.g., Fig. 5A, B) are morphologically quite similar to those described by Walters (1994) and to ice-wedge casts from Europe described by Harry and Gozdzik (1988). Some wedges in the study area are filled with gravel-free fine sand and have a vertical fabric in the sediment fill, suggesting that they are relict sand wedges (Péwé, 1959; Murton et al., 2000), formed when contraction cracks in permafrost filled with eolian sand. Much smaller sand-filled wedges apparently formed during aggradation of late Wisconsin fluvial sediment in southeastern Minnesota (Fig. 5C), resembling similar features reported from full glacial fluvial sediments in the Netherlands (Van Huissteden et al., 2000); given their shallow depth these may represent contraction cracking of seasonally frozen sediments rather than permafrost. Polygonal patterns visible on air photos are widely distributed across the study area and adjacent regions; whereas excavation does not always reveal distinct wedge structures associated with these polygonal patterns, it is likely that they represent former ice-wedge polygons in many cases (Konen, 1995; Lusch et al., 2009).

Near the study area in Wisconsin and to the south of it in Illinois, ice-wedge casts occur in sediments deposited during the last glaciation (Johnson, 1990; Clayton et al., 2001), conclusively demonstrating the occurrence of permafrost at that time. Other ice-wedge casts within the study area formed in sediments predating the last glaciation or of unknown age, but in most cases it seems likely that the ice-wedge casts themselves formed in the last glaciation. Older features would often have been removed by erosion, and soil development in wedge-filling sediment is relatively weak and similar to Holocene soil development in sandy sediments deposited during the last glaciation. In some

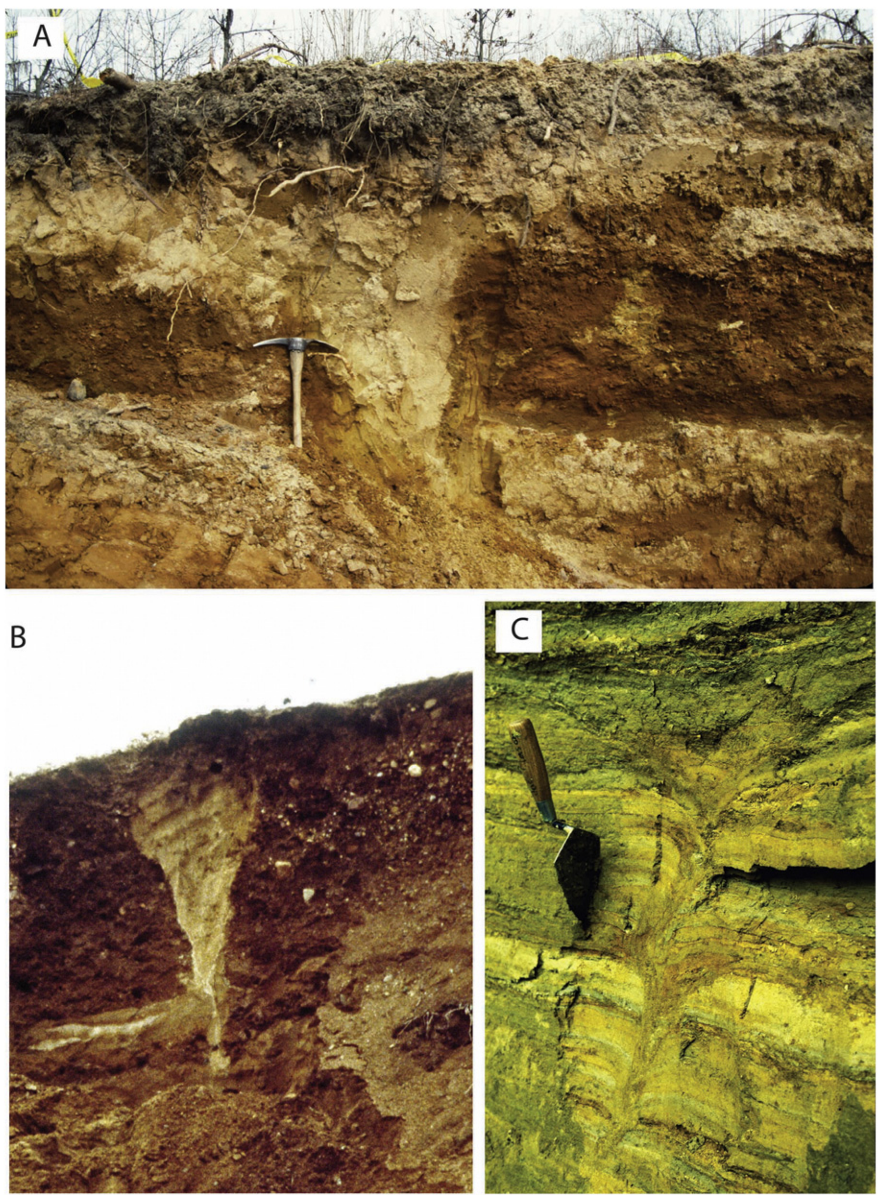

Fig. 5. Relict frost wedges in the study area, at locations labeled in Fig. 1. (A) Ice-wedge cast near Sparta, Wisconsin (photo by James C. Knox, 1994). Host material is well-developed soil with reddish brown Bt horizon formed in fluvial sand and gravel of unknown age. Mattock is about 60 cm long. (B) Ice-wedge cast at Hampton Town Hall site, Minnesota (photo by J. Mason, 1991). Host material is well-developed soil with clay-rich reddish brown Bt horizon formed in pre-Wisconsin glaciofluvial sediment. Modern surface soil is truncated by 15 cm or more within and outside of wedge. Vertical dimension of wedge is about 1 m. (C) Small wedges, possibly produced by deep seasonal frost rather than forming in permafrost, in late Wisconsin fluvial silt and fine sand in Root River valley, Lehman site, Minnesota (Mason and Knox, 1997; photo by J. Mason, 1993). Note two vertically stacked wedge forms, above and below trowel. Trowel blade is about 12 cm long.

cases this weak soil development within wedges is in strong contrast to strongly expressed soils in the material hosting the wedges, probably representing long-term pedogenesis before the last glaciation (Mason et al., 1994a; the soils within wedge fills in Fig. 5A and B are also good examples of this pattern).

Overall, the spatial distribution and likely age of ice-wedge casts and relict sand wedges indicates that the entire study area was affected by widespread and probably continuous permafrost during the coldest part of the last glaciation, consistent with the continental-scale map of French and Millar (2014). Besides ice-wedge casts, however, few other landforms that are unambiguously linked to permafrost have been identified in and around the study area. Clayton and Attig (1987) interpreted shallow trenches that parallel shorelines of Glacial Lake Wisconsin to have formed through burial of lake ice by aggrading glacial outwash fans. The buried ice was initially preserved within permafrost but melted to form the trenches during permafrost degradation. Low mounds in northern Illinois that were interpreted by Flemal (1976; Flemal et al., 1973) as relict pingos are in fact ice-walled lake plains (Curry et al., 2010). Other interpretations of Driftless Area landforms as reflecting specific aspects of the periglacial environment have been proposed but are not widely accepted (Iannicelli, 2010). Many aspects of the glacial geomorphology of Wisconsin may reflect the presence of permafrost at the ice margin and extending some distance under the ice sheet (Clayton et al., 2001), but more detailed discussion of the specific processes and landforms involved is beyond the scope of this paper.

4. Loess, hillslope, and fluvial stratigraphy in relatively high-relief, dissected landscapes

4.1. Stratigraphy and sedimentology of loess and colluvium

Wherever bedrock-controlled slopes predominate in parts of the study area with relatively high relief (type 2 landscapes), they typically are mantled by sediment that is a mixture of weathered local bedrock, loess, and occasionally other materials. The nature of this sediment mantle varies with local bedrock stratigraphy and local relief. For example, on the quartzite slopes of the Baraboo Hills and isolated monadnocks in central Wisconsin, large boulders form talus slopes, and lower gradient boulder streams are also present. The boulder streams,

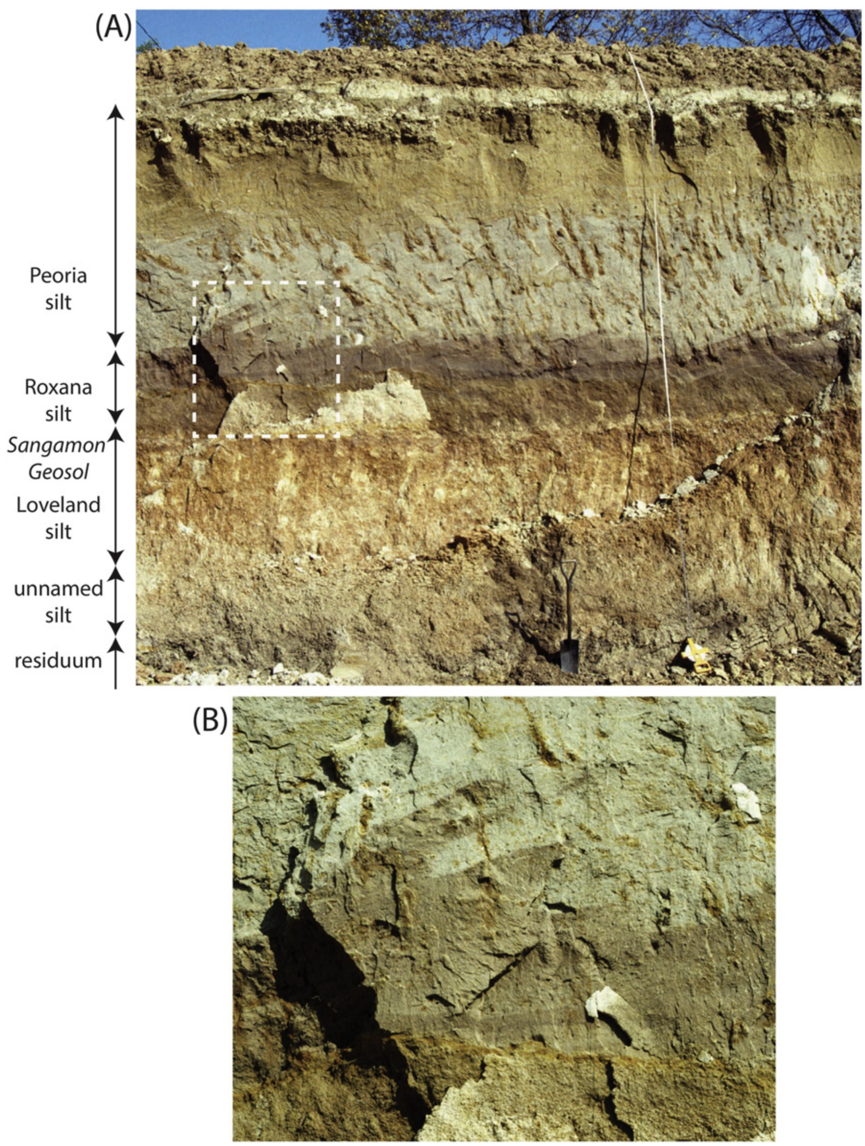

Fig. 6. (A) Loess section on a broad ridgetop at the Kieler site (Fig. 1), southwestern Wisconsin Driftless Area (Jacobs et al., 1997). Downslope is to right and out of the exposed face. Spade at base of exposure is about 1 m long. Note slabs of Roxana silt thrust over basal Peoria silt, producing structures similar to those in some soils affected by solifluction today (see text). (B) Closeup of structures near base of Peoria silt, area within dashed line in (A). Photos by James C. Knox, 1994.

in particular, have been attributed to past periglacial environments (Smith, 1949; Clayton et al., 2001).

Here I focus on the more extensive type 2 landscapes within about 80 km of the Mississippi River in which broad ridgetops underlain by resistant dolomite are separated from deeply incised valleys by steep slopes underlain mainly by sandstone (Fig. 4A). These landscapes include parts of the western Driftless Area where James Knox carried out much of his research, southeastern Minnesota, and northeastern Iowa. In this setting, loess caps the wide ridgetops (Fig. 6), with several middle to late Pleistocene loess units preserved at some well-studied sites (Leigh and Knox, 1993, 1994; Jacobs and Knox, 1994; Leigh, 1994; Mason et al., 1994b; Jacobs et al., 1997). These include late Pleistocene Peoria silt (or Peoria loess, deposited in OIS 2, 25–12 ka), Roxana silt (mostly OIS 3, 55–25 ka), Loveland silt (or Loveland loess, OIS 6) and rarely one or two older loesses (Fig. 6). Peoria silt is the only unit preserved on many summits, where it can rest directly on weakly weathered bedrock or pre-Illinoian glacial sediment (Fig. 7). The Mississippi River valley was the major source of Roxana silt and an important source of Peoria silt (Leigh, 1994; Leigh and Knox, 1994), but the most volumetrically important source of Peoria silt in the region was probably silt from Des Moines Lobe glacial sediments and eroding older tills west of the Mississippi (Mason et al., 1994b). The maximum local thickness of these two loess units decreases away from their major sources, but there are many local anomalies in these sedimentary trends; for example, broad ridgetops where Peoria silt is thinner than expected based on distance from sources (Leigh and Knox, 1994; Mason et al., 1994b). Locally, preservation of Roxana silt and older loess seems related in part to the width and slope of the ridgetops, but the longest known sequence is on a narrow spur ridge (Jacobs and Knox, 1994). By far the most consistent pattern is that all loess units present on ridgetops decrease substantially in thickness or pinch out entirely from ridge summits to the convex shoulder slopes below (Fig. 7).

The simplest explanation for these spatial patterns of loess preservation is differential erosion, during and after the accumulation of each unit (Leigh and Knox, 1994). Slopewash and gully erosion are highly effective in eroding sparsely vegetated thick loess, especially under intense rainfall. Intense convective rainstorms are probably much more likely under warm interglacial conditions than during glacial periods in the study area, but modeling of the LGM climate suggests that occasional strong cyclonic storms produced heavy rainfall in the ice marginal region of North America (Bromwich et al., 2005). If so, relatively high rates of slopewash erosion might have been possible at that time in the study area, especially if the rain fell on a thawing active layer above permafrost (Bettis et al., 2008).

At least two exposures have revealed clear evidence that another erosional process, solifluction, was active on gently sloping ridgetops early in the accumulation of Peoria silt, probably just before the LGM (one is shown in Fig. 6 and described by Jacobs et al., 1997; the other was a site near Caledonia, Minnesota). In these exposures, slabs detached from the underlying Roxana silt rise up over basal Peoria silt, very similar to the movement of material downslope and toward the ground surface observed on some modern slopes undergoing solifluction over permafrost and attributed to shear within the lower part of the mobile active layer (Egginton and French, 1985). The structures shown in Fig. 6 provide direct evidence for activation of a hillslope process characteristic of most modern periglacial environments (Lewkowicz, 1988) around the time of the LGM in the study area. Solifluction can occur, sometimes at high rates, in cold environments with deep seasonal frost rather than permafrost (Matsuoka, 2001). On the dry, well-drained ridgetops in the study area, however, permafrost would most likely have played an important role in activating solifluction, by impeding drainage and allowing abundant segregated ice formation and *two-sided* freezing (Lewkowicz, 1988). Furthermore, local variation in active layer hydrology and rates of solifluction (e.g., Egginton and French, 1985) – because of variations in vegetation, snow cover, or aspect – might explain some of the otherwise anomalous local variations in loess unit thickness on wide ridgetops. Finally, wind erosion through direct entrainment of coarse loess (Sweeney and Mason, 2013) should not be discounted as an explanation for unusually thin loess on some high, wind-exposed summits.

Below the gently sloping ridgetops, slopes steepen, with some segments exceeding 40°. These steeper slope segments are mantled by poorly sorted sediment, with abundant gravel- to boulder-sized clasts of dolomite in a silt-rich matrix (Fig. 8). Outcrops of dolomite and sandstone protrude through this mantle. Gentler footslopes are underlain by massive silt that closely resembles loess but has a distinctly higher sand content and often contains scattered large clasts. The silt matrix in the rocky colluvium on upper slopes and the massive silt on footslopes are largely reworked but unweathered Peoria silt, based on grain-size modes, color, and carbonate mineral content (Mason, 1995). On some slopes, rocky, dolomite-rich colluvium extends below the massive footslope silt (Fig. 8A, C), while on others the rocky colluvium gives way downslope to a thin sandy mantle that covers sandstone bedrock and extends downslope over the massive footslope silt (Fig. 8B).

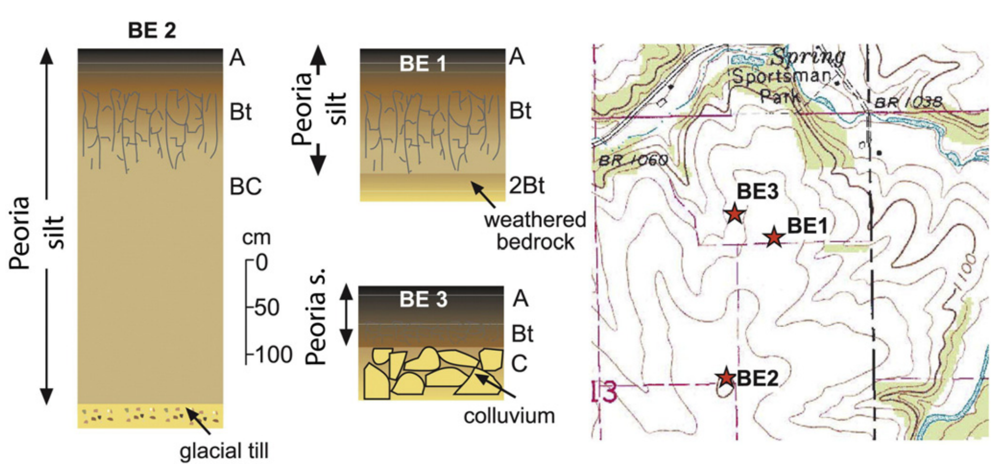

Fig. 7. Loess stratigraphy and downslope decrease in loess thickness at BE site, southeastern Minnesota (Fig. 1, Mason et al., 1994b). On broad ridgetop (BE-2), only Peoria silt is present above pre-Illinoian glacial sediment that displays minimal soil development. Much thinner loess overlies bedrock on slopes (BE-1, -3). Absence of older loess units here is typical of many ridgetop sites in high-relief, dissected part of study area (Mason et al., 1994b; Leigh and Knox, 1994). Contour interval of topographic map is 20 ft (6.1 m).

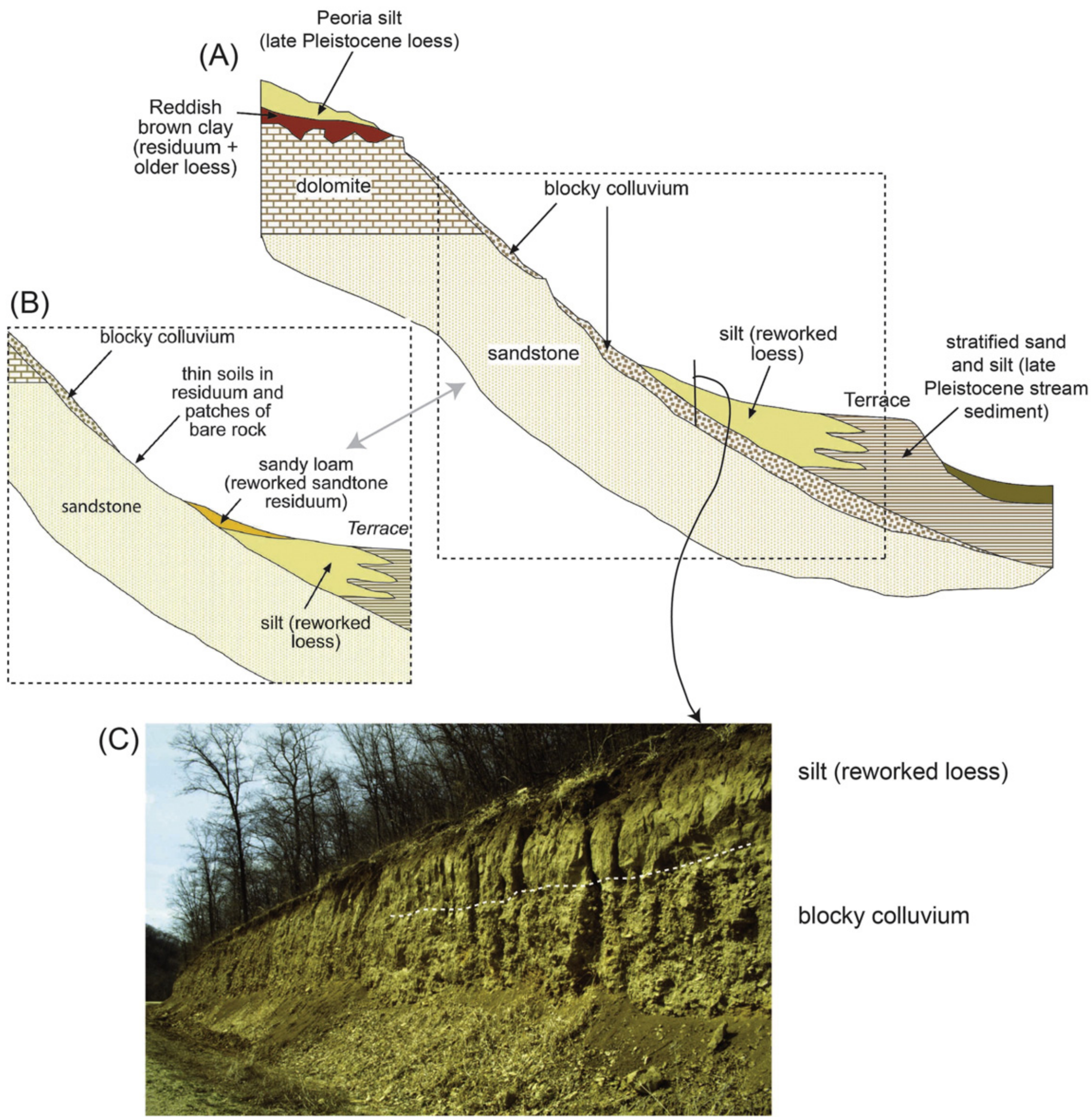

Fig. 8. Hillslope stratigraphy on steep bedrock-controlled slopes in a relatively high-relief study area landscape in southeastern Minnesota (Mason, 1995; Mason and Knox, 1997); broadly similar to stratigraphy in similar landscapes of the Driftless Area and northeastern Iowa. (A) Schematic view of stratigraphy. (B) Alternate stratigraphic pattern observed on some slopes in the same area. (C) Photo of footslope silts overlying rocky colluvium, Whitewater valley, Minnesota (slope position marked in (A), photo by J. Mason, 1990).

Downslope movement of the entire sediment mantle is demonstrated by dolomite debris far downslope over sandstone bedrock (Fig. 8A, C) and by a preferential orientation of the long axes of clasts parallel to the slope gradient, which is particularly strong for scattered clasts in the footslope silts (Mason, 1995). The rocky, dolomite-rich sediment on upper slopes likely moved through slow creep rather than rapid slides or flows; loess could have been mixed into dolomite debris as it moved downhill. Large landslides that occurred on steep slopes in southeastern Minnesota after an extreme rainfall event in 2007 stripped upper slopes down to bedrock while depositing fans of clast-rich sediment on footslopes and valley floors. No evidence of analogous events, such as clast-rich fans or lenses embedded within the massive footslope silts, were observed in extensive roadcut and gully outcrops that were studied throughout the area around these recent landslides.

The massive footslope silts originated in part through reworking of loess initially deposited on steeper slopes above, which could have occurred through solifluction as suggested earlier for ridgetop loess. Possibly, however, rapid flows triggered by saturation during seasonal thawing (Lewkowicz, 1988) or heavy rainfall were also involved in remobilizing loess and depositing it on footslopes. The footslope silts might also contain slopewash deposits, with any diagnostic sedimentary structures subsequently destroyed by bioturbation; and they also must include a component of primary loess deposited directly on footslopes.

Mason and Knox (1997) reported radiocarbon ages from sites in southeastern Minnesota and the Driftless Area demonstrating that the footslope silts and rocky, dolomite-rich colluvium accumulated between about 22 and 14 cal ka. They also noted that the footslope silts grade laterally into late Pleistocene fluvial sediment underlying a well-defined terrace, and at a few sites the rocky colluvium also interfingers with late Pleistocene fluvial sediment. Thus, much of the mantle of sediment on these hillslopes accumulated in the late Pleistocene, at a time when a variety of independent evidence indicates the presence of permafrost and tundra vegetation in the study area. The processes inferred to have produced this mantle do not necessarily require permafrost or periglacial conditions. Presence of permafrost, however, would likely have accelerated downslope movement through frost creep of the rocky colluvium and favored development of a saturated active layer in loess upper slopes, fostering downhill movement through solifluction, rapid flows, and or slopewash. More rapid frost weathering of rock outcrops in a periglacial environment and high rates of loess influx would have supplied abundant material to continually generate a slope mantle, which moved downhill rapidly enough to replace whatever previously covered the bedrock surface on these slopes. A shift toward much slower rates of movement by creep and slopewash in the Holocene would have allowed long-term persistence of much of the late Pleistocene slope mantle, although it has locally been removed by gully erosion.

4.2. Glacial–interglacial contrast in sediment yield

Knox (1989) made an initial effort to move beyond qualitative assessments of the effects of glacial–interglacial climate change on sediment yields with a study of a low-order drainage basin in the southwestern Driftless Area. In this small basin with local relief of about 30 m, loess covers upper slopes, footslopes are mantled with thick loess-derived silts, and the rocky colluvium mantling midslopes in Fig. 8 is largely absent. Using intensive core sampling, Knox mapped the thickness of Peoria loess and loess-derived footslope silts from drainage divide to valley floor. Truncation of loess thickness relative to the maximum thickness on the drainage divide was used to estimate the total volume of loess removed from summits and upper slopes. Knox concluded that the amount of loess eroded from uplands implied a sediment yield of 112 t km^{-2} y^{-1} if averaged over the past 20,000 years. About 70% of the eroded loess was at least temporarily stored in footslope silts, and based on radiocarbon dating at one locality, Knox estimated those silts accumulated before about 12,000 ^{14}C y BP (about 13.9 cal ka). Assuming on that basis that most erosion of the loess took place between 20,000 and 12,000 years ago, he then calculated a higher sediment yield of 280 t km^{-2} y^{-1} that he attributed to accelerated erosion under periglacial conditions. The estimated sediment yields during the Holocene would then be considerably <112 t km^{-2} y^{-1} (today these estimates would more appropriately be made using calibrated radiocarbon ages, but the relative magnitudes would be fairly similar).

The contrasting late Pleistocene and Holocene rates of erosion estimated by Knox (1989) clearly suggest that slope processes during full-glacial environments play a disproportionate role in long-term erosion and landscape evolution of the study area. Note, however, that estimated full-glacial sediment yields on the order of 250–300 t km^{-2} y^{-1} are actually much lower than modern sediment yields from tectonically active mountain ranges and even some semiarid landscapes of lower relief (Walling and Webb, 1996). Furthermore, the estimate of full-glacial sediment yield is based on erosion of loess, and rates of erosion could have been much lower on slopes where other materials such as glacial till or bedrock were exposed. Thus, even where tectonically stable landscapes like the Driftless Area are episodically subjected to periglacial conditions, they may not reach peaks of sediment yield comparable to those observed today in areas of active uplift and rapidly eroding drylands.

To summarize, loess and hillslope sediment stratigraphy in the highly dissected, bedrock-controlled type 2 landscapes of the study area is consistent with the hypothesis of accelerated erosion and increased supply of sediment to the fluvial system in the coldest part of the last glaciation, at a time when independent evidence indicates widespread permafrost and tundra vegetation. Preservation of a largely late Pleistocene slope mantle on steep slopes is difficult to explain without relatively low rates of erosion during the Holocene. Estimates of sediment yield by Knox (1989) provided more quantitative support for this hypothesis, though they are limited to erosion of loess in a single small basin. Full-glacial conditions are likely to have a disproportionate influence on long-term rates of erosion; but even under those conditions, the results of Knox (1989) suggested that landscapes of the study area do not reach sediment yields comparable to areas of active tectonics. All of these conclusions needed further testing, e.g., through dating of the slope mantle by a variety of methods at more localities across the study area. More detailed observations of hillslope sediments at many sites could also reveal more about specific processes that produced them and whether they are processes that would be particularly effective in a periglacial environment.

4.3. Fluvial response to hillslope erosion

Accelerated hillslope erosion would have increased sediment supply to streams, although as the study by Knox (1989) demonstrated, the effects would have been moderated by storage of eroded sediment on lower slopes and low-order valley floors. Assessing the impact of increased nonglacial sediment yield on the fluvial system is complicated, however, because it occurred at a time when enormous quantities of sediment were fed from the ice sheet into major rivers of the region, including the Mississippi and important tributaries (e.g., the Wisconsin and Chippewa rivers, Fig. 3). Those streams aggraded substantially in response and later incised episodically during deglaciation (Knox, 1996; Bettis et al., 2008), in the process changing base level for basins in the Driftless Area and adjacent landscapes of southeastern Minnesota and northeastern Iowa that were outside the ice limits.

Most stream valleys in those basins outside the glacial limits contain alluvial fills recording late Wisconsin aggradation (Fig. 9). Near the Mississippi valley these fills can be dominated by slackwater sediment of the Mississippi River itself (Bettis and Hallberg, 1985; Flock, 1983; Knox, 1996), but farther upstream they are made up of locally derived and often coarse sediment (Fig. 9B), likely to be a response to increased

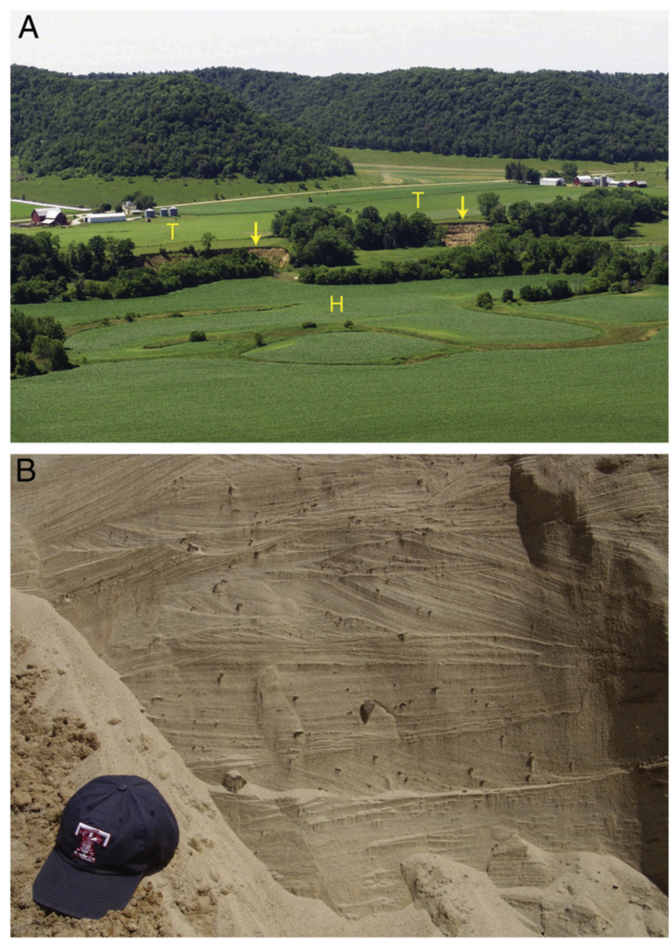

Fig. 9. (A) Late Wisconsin fill terrace at Jore site, southeastern Minnesota (Fig. 1). Terrace surface labeled 'T,' with arrows indicating scarps exposing 15 m of fluvial sand and silt above modern river level. Plant macrofossils representing a full-glacial Arctic plant assemblage were collected just above river level (Baker et al., 1999) on scarp to left. (B) Typical exposure of channel sands in fill below late Wisconsin terrace, lower Root River valley near Lehman site, southeastern Minnesota (Fig. 1). Pebble lithology suggests derivation mainly from local Cambrian and Ordovician bedrock, but some pebbles are glacial erratics from pre-Illinoian sediments in Root River basin. Photos by J. Mason, 2007 and 2009.

sediment input from hillslopes rather than base level effects (Knox and Maher, 1974; Knox, 1983; Hallberg and Bettis, 1985; Bettis and Autin, 1997; Mason and Knox, 1997; Bettis et al., 2008). The late Wisconsin fills in basins outside the glacial limit often resemble glacial outwash despite nonglacigenic origin (Bettis et al., 2008), and to my knowledge, distinctive sedimentological characteristics attributable to an origin through periglacial hillslope erosion have not been identified (see also Vandenberghe, 2011). More detailed studies of how hillslope deposits interfinger with or grade into late Wisconsin alluvial fills (Hallberg and Bettis, 1985; Mason, 1995; Mason and Knox, 1997) could be highly informative.

5. Low relief landscapes: the Iowan Erosion Surface and the influence of R.V. Ruhe

5.1. Characteristics and contrasting interpretations of the Iowan Erosion Surface

As noted above, the study area contains landscapes very different from the highly dissected, relatively high-relief areas discussed in the previous section. One of the most extensive is a region east of the Des Moines Lobe in northeastern Iowa and southeastern Minnesota usually referred to as the Iowan Erosion Surface (IES; Figs. 3 and 4B), an example of the type 1 landscapes described above. The IES and varied interpretations of its evolution are rarely mentioned in the broader geomorphological literature, but it has often had an important role in informal discussions of geomorphic response to periglacial conditions among geomorphologists in the Midwest. Knox (1996) emphasized the IES in discussing the influence of accelerated late Wisconsin erosion on the fluvial system in the Upper Mississippi basin.

The topography and stratigraphy of the IES have been described in great detail, though with very different interpretations, by several geologists over the past 125 years (Alden and Leighton, 1915; McGee, 1891; Ruhe et al., 1968; Ruhe, 1969). To summarize their observations in modern terms, this landscape is dominated by long and usually gentle slopes, underlain by pre-Illinoian till and other glacial sediment (Fig. 10). The drainage network is fully integrated, and no identifiable glacial landforms are preserved. A thin (<1 m) mantle of loess or loam-textured sediment covers much of the central and northern IES, with loess thickening to 3–4 m in some southern parts of the IES. A stone-line, or thin layer of gravel and sand, often occurs at a shallow depth, at the contact between loess or loam sediment and till. The IES is bordered to the east and south by landscapes of denser drainage networks, steeper slopes, and greater relief, which have considerably thicker loess cover over pre-Illinoian tills and/or weathered bedrock (Fig. 4A). The transition from the thin loess of the IES to these adjacent areas of thick loess is remarkably abrupt, occurring within a few hundred meters in some localities (Mason et al., 1999).

A striking feature of the IES is the occurrence of long, narrow isolated hills of thick loess, oriented northwest–southeast (Fig. 10A and B). McGee (1891, pp. 403–404) described these distinctive "canoe-shaped loess ridges" and adopted the Native American term *paha* for them.

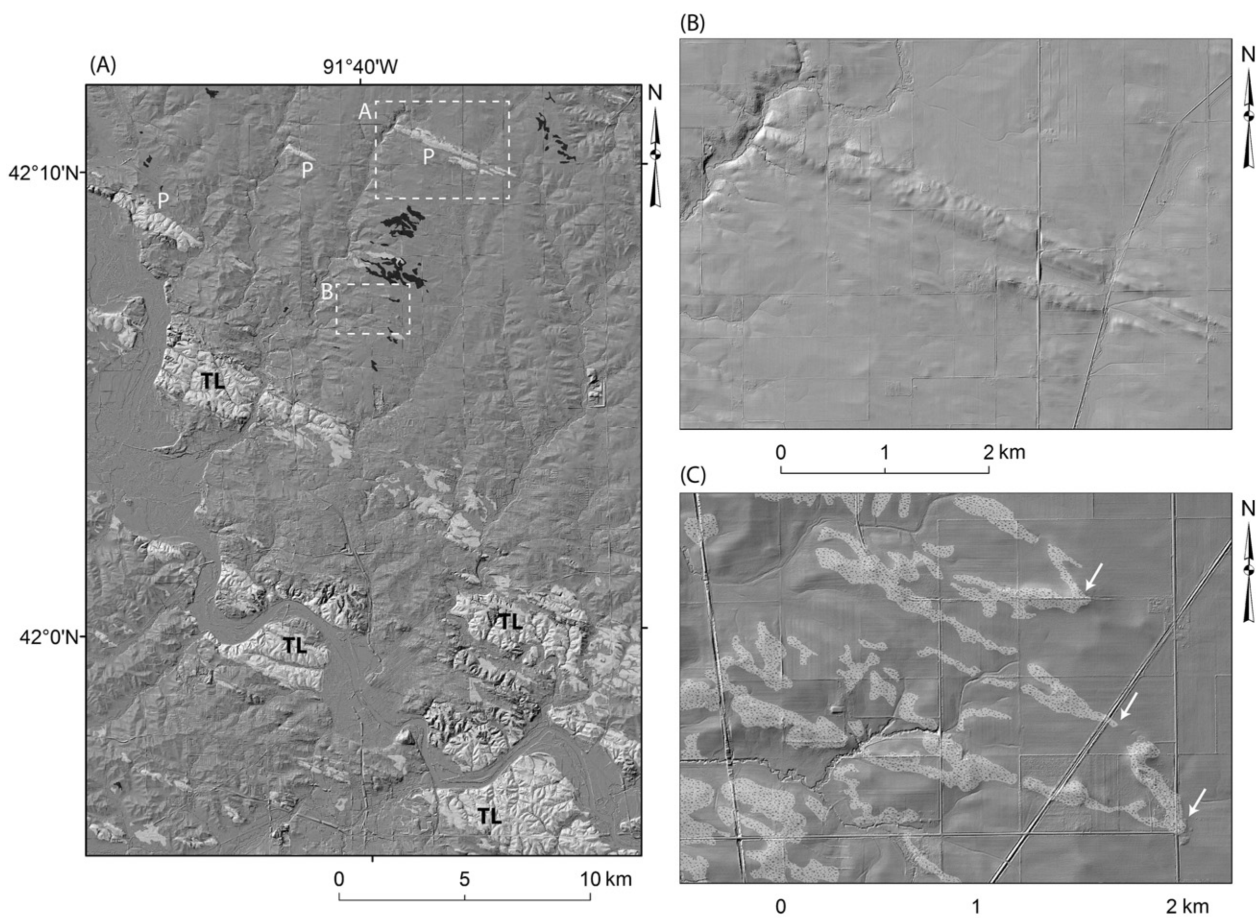

Fig. 10. Topography and eolian sediment distribution on IES, Linn County, Iowa. (A) Overview, with several typical paha (P) and larger areas of thick loess (TL) labeled. Light gray tone indicates thick loess interpreted from soil survey data as in Fig. 3. Dark tone indicates patches of Donnan soils, interpreted as relict pre-Wisconsin soils, from county soil survey (SSURGO data, Natural Resource Conservation Service, USDA). Areas shown in (B) and (C) marked with dashed lines. (B) Well-developed paha. Northwest–southeast aligned ridge is underlain by thick loess, with thin loess on surrounding landscape (Alburnett Paha of Ruhe et al., 1968). (C) Low dunes. Light tones with dark stipple indicated eolian sands interpreted from county soil survey. Three distinct parabolic dunes marked with arrows. Shaded relief image used in all panels produced from 1-m LiDAR topographic data and distributed by Iowa Department of Natural Resources.

Larger and more irregularly shaped patches of thick loess also occur, often with northwest–southeast oriented boundaries (Fig. 10A). In contrast to the surrounding landscape, Peoria loess can exceed 10 m in thickness within the paha and larger thick loess patches (Ruhe et al., 1968; Hallberg et al., 1978).

The name now applied to the IES reflects a sequence of two contrasting interpretations of the region. Alden and Leighton (1915) made an extended argument for interpreting the IES as the area affected by the 'Iowan glaciation,' thought to be an early phase of the last glacial cycle. They noted that the IES did not retain glacial landforms like those in the area of the younger Wisconsin glaciation to the west, but it was not deeply dissected like areas thought to have been affected by earlier Kansan (now pre-Illinoian) glaciation. They also described typical weathering profiles on the IES and noted that they suggest less weathering than the reddish-brown 'feretto' profiles found on Kansan tills (prominent interglacial paleosols preserved under Wisconsin loess).

In the 1960s, Robert Ruhe, then investigating soil-landscape relationships for the Soil Conservation Service, returned to the problem of the 'Iowan glaciation,' using extensive shallow subsurface data and a more sophisticated understanding of soils (Ruhe et al., 1968). He used new evidence from drilling into paha, which in some cases revealed strongly expressed paleosols with thick, clay-rich Bt horizons, formed in the tills beneath the thick loess. This evidence of advanced pedogenesis is largely absent in the same tills in the surrounding landscape. Instead, on the northern and central IES relatively weak soils with Bw or weakly expressed Bt horizons have developed through thin loess or loam sediment into the uppermost till. Ruhe and coworkers also used drilling to trace the till at the surface of the southernmost IES under thick loess to the south, where it was capped by a well-developed paleosol, as in many paha. Based on this soil-geomorphic evidence, Ruhe concluded that the tills across the IES were, in fact, pre-Illinoian (then referred to as Kansan and Nebraskan), consistent with the soil development observed in buried profiles within the paha.

This conclusion has been supported by later work on the IES (Hallberg et al., 1978). The surrounding landscape is an extensive erosion surface much younger than the glaciation in which the tills were deposited, which removed evidence for advanced pedogenesis that might be expected at the surface of the tills given their true age. Ruhe used radiocarbon ages from the base of the loess in the paha and under thinner loess on the surrounding landscape to argue that much of the erosion surface formed around 30–21 cal ka, though also suggesting that formation continued into the Holocene in some areas (Ruhe, 1969; Ruhe et al., 1968). Important though limited exceptions to the general soil geomorphic pattern emphasized by Ruhe do occur and will be reconsidered below. First, paha exist without a paleosol between thick late Pleistocene loess and underlying till (e.g., the Alburnett paha of Ruhe et al., 1968, illustrated in Fig. 10B). Second, patches of unusually thick, clay-rich surface soils occur on the IES with morphology similar to strongly expressed buried paleosols within the paha, but with thin or no loess cover; these are often mapped as the Donnan series in county soil surveys (Fig. 10A).

5.2. Ruhe's conceptual model of landscape evolution and alternatives to it

The development of an extensive erosion surface around the peak of the last glaciation is clearly relevant to the broader topic of this paper; however, it is essential to recognize the conceptual framework in which Ruhe worked and to consider how interpretations of the IES might change if we move beyond the constraints of that framework. Throughout his career, probably starting with an early project on soils and landscape evolution in central Africa (Ruhe, 1956), he was strongly influenced by Lester King's theory of landscape evolution. In a comprehensive statement of his conceptual model, *The Morphology of the Earth* (King, 1962), King counterposed it to the ideas of William Morris Davis, claiming it to be applicable across most climatic environments (King, 1962, pp. 152–154). King envisioned landscapes evolving through the formation of pediments (or larger 'pediplains') that grow through slope retreat and are shaped largely by flowing water, with mass wasting effective only on the retreating scarps (King, 1962, pp. 135–152). Following incision in response to tectonic uplift, younger pediments grow at the expense of older surfaces, ultimately producing a stepped landscape descending from old, high surfaces to younger ones. King's key examples of stepped landscapes evolving through this process involved retreat of major escarpments in southern Africa, South America, and Australia over the Cenozoic or longer.

Ruhe applied King's concept of stepped landscape development at the spatial scale of small study areas in Iowa and on the temporal scale of the middle Pleistocene to Holocene (e.g., Ruhe, 1967; Ruhe et al., 1968). His methods, which involved rigorous subsurface investigation and voluminous data collection, were quite different from the broad-brush interpretations of King. Nevertheless, he apparently believed as strongly as King did in the general applicability of King's model (e.g., Ruhe et al., 1967, p. 69), and he interpreted Midwestern landscapes beyond the late Pleistocene glacial limits as stepped sequences of erosion surfaces, formed through pedimentation. Ruhe did incorporate new elements necessary in the Midwest, such as the burial of some older surfaces by younger loess (e.g., Ruhe et al., 1967, p. 85). Ruhe's published work displays little interest in the details of geomorphic processes, though he worked in an era when geomorphology was revolutionized by quantitative process studies. To my knowledge, Ruhe did not discuss the underlying causes of cyclic erosion. While he recognized the likelihood of periglacial conditions during glacials and described ice-wedge casts on the IES and involutions in loess (Ruhe, 1969, pp. 177–179), he did not acknowledge potential effects of permafrost on the dominant geomorphic processes or the mode of landscape evolution. Instead, he clearly saw the pediments he identified in Iowa as having formed according to the general, nonclimatic model of King. Ruhe would most likely not have approved of identifying these surfaces as cryopediments (French and Harry, 1992) because of the implication that they were formed by a distinct set of cold climate processes in a periglacial environment, although at least some landforms described as cryopediments are interpreted as having formed largely through slopewash and stream erosion (Vandenberghe and Czudek, 2008), as Ruhe envisioned for the IES. Ruhe cited the stone lines of the IES as specific evidence on the role of 'running water' (Ruhe et al., 1968, p. 34).

In the late 1950s, Ruhe and coworkers identified a stepped sequence of several geomorphic surfaces, higher and progressively older toward the drainage divides, in a dissected landscape of southwestern Iowa (Fig. 1), underlain mainly by loess and pre-Illinoian tills (Ruhe et al., 1967). Each surface was marked by characteristic soils, with greater morphological expression and weathering on older surfaces. Later, working on the IES, Ruhe correlated the buried surface underlying loess within the paha with the oldest, or Yarmouth-Sangamon, geomorphic surface in southwestern Iowa, capped by the deep, weathered Yarmouth-Sangamon paleosol (Ruhe et al., 1967, pp.77–80; 1968, pp. 25–26). The much more extensive areas around the paha were then considered to form a single relatively young erosion surface, the IES in a strict sense.

That is, Ruhe interpreted much of the IES landscape as made up of relatively young pediment-like surfaces that extended headward from major streams in a fairly short period of time during the last glaciation, removing deep, well-developed soils (Ruhe et al., 1968, pp. 25–26), a view that clearly implies substantial rates of erosion. Even higher rates are required if it is assumed that Peoria loess as thick as is now observed in the paha was also deposited across the rest of the IES and then removed during erosion surface formation. Ruhe's discussion of loess on the IES – as I read it – implies an initial increment of Peoria loess deposition before erosion surface development, subsequently eroded everywhere but in the paha. As loess continued to be deposited, it was removed from the growing erosion surface, except for a final increment that blanketed the whole landscape. Ruhe did briefly suggest a role for eolian processes in erosion surface development, stating that loess was deflated from the IES and carried downwind from it and noting that eolian sand was deposited with loess in the paha (Ruhe et al., 1968, pp. 29–34). The pedimentation process that he favored clearly would have to have primarily occurred through slopewash and stream flow, however.

How much of this interpretation follows directly from the field evidence that Ruhe and others collected, and how much was determined more by Ruhe's particular model of landscape evolution? The stepped pediment-like surfaces portrayed so unambiguously by Ruhe in his research reports are not particularly evident in shaded relief maps of the IES (Fig. 10), or in southwestern Iowa for that matter. Interpretation of the paha as erosional remnants left on major drainage divides by pediment expansion is irreconcilable with their actual topographic setting, often extending across divides and perpendicular to them (e.g., Fig. 10A and B). One key observation, however, was and still is compelling: the large contrast between well-developed soils on pre-Illinoian tills in local areas, and much weaker soil development on the same tills over much of the IES is difficult to explain without a corresponding contrast in erosional history. In southwestern Iowa, Ruhe et al. (1967, pp. 104–149) found that 'Yarmouth-Sangamon' soils similar to those within the paha display much greater depletion of weatherable primary minerals and clay mineral alteration than soils formed entirely since the last glaciation, strengthening the argument that residence time is a key factor differentiating these soils. Later research has documented this contrast between Holocene soils and those formed in earlier interglacials in other parts of the Midwest (Follmer, 1983; Jacobs, 1998; Hall and Anderson, 2000; Grimley et al., 2003). Even if other variables such as the climate or vegetation conditions of past interglacials also contribute to this differentiation of soil morphology (as suggested by Ruhe, 1965), the implication is that the soils across much of the IES are too young to have experienced those past interglacial conditions.

This key observation in itself provides little information on the timing, duration, or episodicity of the erosion that produced the distinctive soil-geomorphic patterns of the IES. Ruhe's conceptual model of sequentially developing, discrete erosion surfaces inherently favors the idea of discrete episodes of erosion and pedimentation. If we assume that much of the landscape of the IES was covered by thick clay-rich soils comparable to the 'Yarmouth-Sangamon paleosols', the implication is that a long period of quite slow erosion occurred, allowing deep soil development and possibly spanning multiple glacial cycles. In this scenario, the episode of much more rapid surface lowering at the peak of the last glaciation (OIS 2) was anomalous, relative to the longer term history of this landscape. The loess stratigraphy of the Great Plains and Midwest does raise the possibility that conditions of climate and vegetation during OIS 2 differed from those of earlier glacials, favoring substantially more dust production and loess accumulation (Bettis et al., 2003; Roberts et al., 2003; Mason et al., 2007). If so, then perhaps those unique conditions of OIS 2 also favored substantially higher rates of hillslope erosion in regions like the IES that were close to the ice margin.

Importantly, there is no compelling reason to assume that accelerated erosion during OIS 2 occurred by pedimentation spreading from streams toward drainage divides as envisioned by Ruhe. Broad-scale changes in climate and vegetation could instead trigger accelerated erosion at more or less the same time across much of the landscape, with local factors reducing erosion rates in small areas such as the paha. This would certainly be consistent with the concept of accelerated erosion across much of the entire study area in the periglacial environment of the last glaciation, as often advocated by Knox.

An alternative interpretation of the IES could place it within a broader view of landscape evolution in parts of the Midwestern U.S. that were last glaciated in pre-Illinoian time. These landscapes have lost all recognizable glacial topography through long-term development of stream networks and associated hillslopes; they have also experienced episodic loess deposition. Mason et al. (2007) described the stratigraphy and local distribution of thick loess overlying pre-Illinoian glacial sediments in a landscape of this type in eastern Nebraska (Fig. 1). Middle to late Pleistocene loess units generally are thickest on broad ridgetops, and thin as they drape down hillslopes, eventually pinching out on steep slopes where tills are exposed at the surface (Fig. 11A). The implication is that the current drainage network was present in at least partial form when the loess began accumulating, and erosion truncated – or in local areas completely removed – the loess on hillslopes during or after each episode of loess deposition (Fig. 11B). On ridgetops, well-developed paleosols are preserved within the loess sequence and at the surface of the underlying glacial deposits, and some of the loess units are pedogenically altered to varying degrees throughout their thickness (Fig. 11A). On adjacent steep slopes that lack loess cover, surface soil development in glacial tills is relatively weak. In areas of pre-Illinoian glaciation farther from major loess sources, such as south-central Iowa or north-central Missouri, loess units older than Peoria loess may not be clearly recognizable but may be preserved within thick buried soil complexes similar to those described by Ruhe as 'Yarmouth-Sangamon paleosols' (Woida and Thompson, 1993).

Importantly, all of these observations can be explained without invoking a discrete and anomalously severe episode of hillslope erosion during the last glaciation, and certainly without the pedimentation model that Ruhe employed. Instead, generally higher rates of erosion on steeper slopes can be seen as a persistent feature of landscape evolution after pre-Illinoian glaciation. That in turn can explain truncation or complete absence of loess and weak soil development there. On gentler slopes, and especially on summits, persistently slower erosion has allowed net accumulation of loess, manifested as upbuilding of thick soil profiles in areas farther from loess sources. The accumulation of loess itself probably favors development of deeper and more strongly expressed soils, relative to dense, slowly permeable pre-Illinoian tills.

In effect, the conceptual model just outlined suggests a divergence between the most steeply sloping parts of the upland landscape (net erosion and limited soil development) and more gently sloping areas (net loess accumulation and deeper soils). Could the IES be a case of this kind of divergence as well, but with the area of net erosion extending over much more of the landscape and with net loess accumulation and advanced soil development confined to small patches such as the paha? This concept seems plausible in part simply because of the geographic setting of the IES. The northern IES, in particular, was 'up in the refrigerator' during the last glaciation, within a major reentrant in the ice margin, and may have been in a similar position relative to the ice sheet in earlier glaciations; thus, it could have repeatedly experienced especially severe periglacial conditions and high rates of erosion in that setting. To fully explain why the low-relief IES became a landscape dominated by net erosion, however, and to explain why net upbuilding by loess accumulation occurred in more rugged landscapes nearby, we need to consider the role of eolian processes active under periglacial conditions.

5.3. Importance of eolian processes

Mason et al. (1999) proposed that the low-relief landscape of the IES acted as a surface of eolian sand transport under conditions of sparse vegetation at the peak of the last glaciation. That is, eolian sand migrated fairly freely across that landscape in the absence of major topographic barriers such as incised valleys or escarpments. Patches of eolian sand are common on the IES, especially in Iowa, and are readily identifiable in soil survey maps. Newly available LiDAR-based digital elevation models have revealed many more details of the dunes containing these sands, usually parabolic in form (Fig. 10C; Loope et al., 2013). Loope et al. (2013) reported preliminary results of optically stimulated luminescence (OSL) dating indicating that these dunes were active between 20 and 13 cal ka. Assuming that much of the eolian sand on the IES records late Wisconsin activity, it clearly indicates a sparsely vegetated landscape at that time.

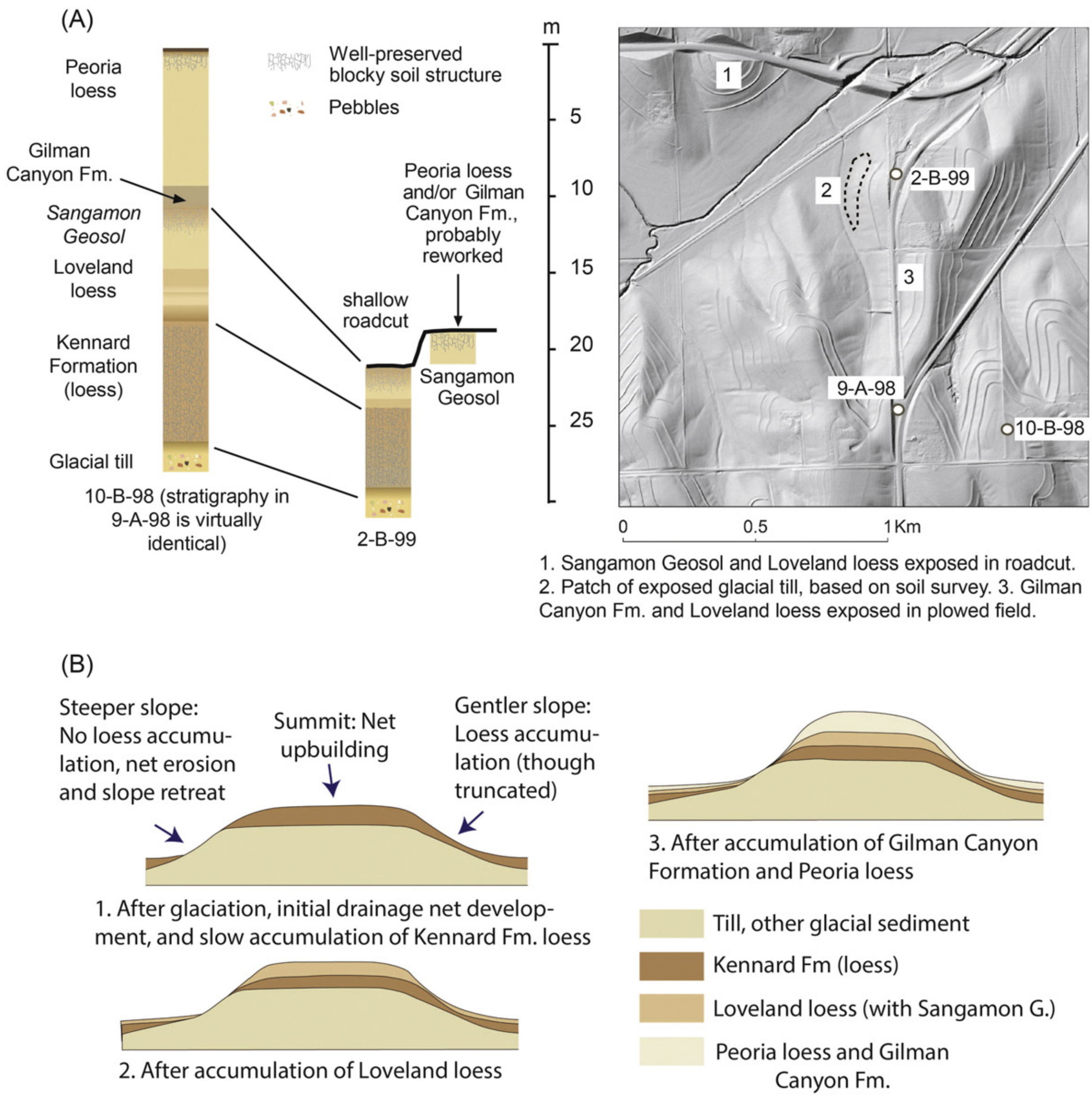

Fig. 11. Quaternary stratigraphy in a landscape of pre-Illinoian glaciation, southeastern Nebraska, and model of landscape based on observations there (Mason et al., 2007). (A) Small part of the study area of Mason et al. (2007), showing observations from drill holes and exposures, typical of those in many other localities. Gilman Canyon Formation is loess, stratigraphically equivalent to Roxana silt in study area. Kennard Formation is slowly accumulated middle Pleistocene loess. See Mason et al. (2007) for details. (B) Conceptual model of landscape evolution, starting after initial fluvial dissection of glaciated landscape and slow accumulation of oldest loess unit (Kennard Formation). Landscape diverges between large areas of net upbuilding through loess accumulation and steep slopes that experience net erosion and retreat.

Widespread eolian sand transport also provides an explanation for the distinctive form and location of the paha and for the abrupt increase in loess thickness at the eastern and southern margins of the IES. Mason et al. (1999), using the IES as a case study, proposed a conceptual model in which loess can move across extensive surfaces of eolian sand transport without accumulating because active saltation favors near-complete reentrainment of the loess after its initial deposition. In this model, loess may be transported long distances from its original sources through repeated reentrainment until it reaches an area that is shielded by a barrier to sand migration, such as an incised stream valley (Fig. 12A). A recent process study indicated that coarse loess may be unstable and prone to deflation in sparsely vegetated landscapes even in the absence of abrasion by saltating sand (Sweeney and Mason, 2013), so the model of Mason et al. (1999) is not the only viable explanation for abrupt local changes in loess thickness. Nonetheless, many details of loess distribution on and around the IES are consistent with that model. Loess carried by northwesterly winds from glacial sources associated with the Des Moines Lobe as well as areas of wind erosion on the IES itself would be transported southeastward in multiple steps until it crossed barriers to saltating sand transport. The abrupt transition from the IES to thick loess-mantled landscapes coincides closely with incised, steep-sided stream valleys (Fig. 12B). The northwest- to southeast-oriented paha can then be interpreted as outlying areas of thick loess that accumulated downwind of local obstacles to sand transport, such as short valley segments with steep bedrock-controlled slopes (Fig. 10B). This is a far more satisfying explanation of the morphology of the paha than Ruhe's view that they simply represent the most stable primary divides, unreached by late Wisconsin pedimentation. The final, thin mantle of loess that covers much of the IES likely accumulated as denser vegetation reduced or ended sand movement. While Mason et al. (1999) focused on the last glaciation, the IES could also have been a surface of eolian sand transport during earlier glaciations as well, favoring long-term net erosion of much of this landscape rather than upbuilding through loess accumulation.

The eolian processes just described should be seen as complementing the hillslope erosion emphasized by Ruhe rather than replacing

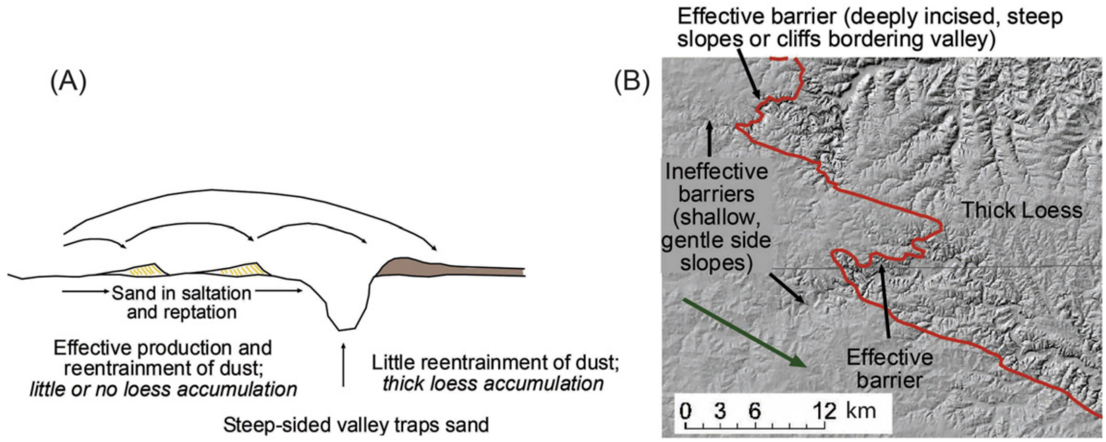

Fig. 12. Conceptual model of topographic effects on distribution of thick loess (Mason et al., 1999). (A) Schematic view of thick loess accumulation downwind of stream valley barrier to eolian sand; area upwind (to left) of valley is a surface of dust and sand transport. (B) Application of model to the IES (area shown in Fig. 3). Green arrow indicates predominant wind direction inferred from loess trends and dunes in nearby areas. Thick loess accumulated to east where deep dissection limited sand transport, but not to west where sand could move freely over shallow valleys. Thick loess extends farther west on south side of larger valleys, where it was shielded from sand moved by northwesterly winds.

it in explaining the distinctive landscape of the IES. Even very widespread eolian activity may not have led to much lowering of the till surface by deflation – because of the cohesion of the dense, clay-rich till – and its gravel content. Furthermore, streams of the IES aggraded with coarse-textured fill during OIS 2; because this aggradation occurred far upstream from these streams' rising base level where they joined the Mississippi, it was likely a response to higher rates of hillslope erosion (Bettis and Kemmis, 1992; Bettis and Autin, 1997). Solifluction over permafrost could have played an important role in this erosion, given the direct evidence for permafrost provided by ice-wedge casts on the IES. The widespread stonelines make a case for slopewash as a key process, however, possibly involving runoff from rapid snowmelt on frozen soils (Vandenberghe and Czudek, 2008) as well as heavy rain from the occasional mid-latitude cyclones mentioned above (Bromwich et al., 2005; Bettis et al., 2008). Extensive eolian sand activity would have enhanced erosion of the glacial sediments underlying the IES and soils formed in them by hillslope processes, by limiting loess accumulation and inhibiting vegetation growth, thereby keeping the glacial sediments exposed to erosion.

The potentially crucial role of eolian processes is highlighted by ridgetop stratigraphy in the highly dissected type 2 landscapes just to the east of the IES. There, late Wisconsin loess accumulation evidently exceeded the rate of hillslope erosion, causing net upbuilding of relatively narrow, rolling summits and, in some cases, burying and preserving underlying paleosols and older loess units (Jacobs and Knox, 1994; Leigh and Knox, 1994; Jacobs et al., 1997). Greater preservation of highly erodible loess and underlying paleosols in this dissected landscape than on the IES, in the same periglacial environment, seems anomalous. Yet it can be explained by considering the combined effect of eolian sand transport and hillslope erosion in limiting loess accumulation on the IES. In contrast, while hillslope erosion was clearly active in the higher-relief type 2 landscapes to the east, ridgetops there were largely unaffected by eolian sand transport because they were separated from sand sources by deep, steep-sided, narrow valleys.

Certain observations discussed above imply that the areas of net loess accumulation and net surface lowering on the IES and in dissected landscapes to the east shifted somewhat during the last glaciation. Localities where thick loess overlies a till surface marked by minimal soil development, within some of the paha and parts of the dissected landscape east of the IES, suggest net loess accumulation after an initial phase of rapid erosion. The patches of unusually well-developed soils with minimal loess cover on the IES could represent local shifts in the other direction, from stability to erosion and removal of whatever loess had accumulated earlier.

5.4. Beyond the IES: other low-relief landscapes of the study area

Beyond the IES, it is increasingly clear that eolian processes played an important role in the evolution of other low-relief landscapes of the study area during the last glaciation. One of the best cases for this view is provided by the Central Sand Plain of Wisconsin (Figs. 3 and 4B). The Central Sand Plain lacks loess cover, although at least thin loess covers surrounding areas. Jacobs et al. (2011) proposed that loess was carried into the sand plain from sources to the west but did not accumulate there because of widespread eolian sand activity; instead the loess was reentrained and carried farther downwind. In other words, the sand plain acted as a surface of dust transport similar to the model proposed for the IES by Mason et al. (1999). Dunes have been recognized on parts of the sand plain for many years, and Rawling et al. (2008) used OSL dating to demonstrate that large dunes on the bed of Glacial Lake Wisconsin developed in the late Wisconsin. Schaetzl (2012) argued that these large dunes occur in limited areas and questioned the more widespread eolian sand activity proposed by Jacobs et al. (2011). Loope et al. (2013) noted, however, that newly available LiDAR-based topographic data demonstrate the occurrence of smaller, low-relief dunes over much of the sand plain (Fig. 13A). A thin sand mantle, rather than loess, covers late Wisconsin glacial deposits over a large area east of the Central Sand Plain, and LiDAR data reveal patches of low-relief dunes in that area as well (Fig. 13B). Thus, widespread eolian sand transport may have extended eastward from the Central Sand Plain as the ice sheet retreated.

Despite this convincing evidence for widespread full- and late-glacial eolian activity, the Central Sand Plain outside of the Lake Wisconsin basin is characterized by an integrated drainage network, with slopes descending to stream channels and few if any closed basins. In other words, this is a landscape clearly shaped by fluvial and hillslope processes; however, recurrent eolian activity under periglacial conditions can help explain its surficial stratigraphy. As with the IES, widespread eolian sand activity can explain the absence of loess from the Central Sand Plain, whether from the last glaciation or older; and the absence of loess cover exposed the underlying weathered sandstone to erosion by eolian and hillslope processes.

The northernmost part of the study area is a broad zone across central Wisconsin in which generally thin (<1 m) loess covers pre-Wisconsin glacial sediment or weathered bedrock (Scull and Schaetzl, 2011; Stanley and Schaetzl, 2011). Thicker loess occurs locally on bedrock uplands near the Chippewa River valley (Schaetzl et al., 2014). Presumably this landscape, like the IES, experienced hillslope erosion in a periglacial environment (Stanley and Schaetzl, 2011), given its

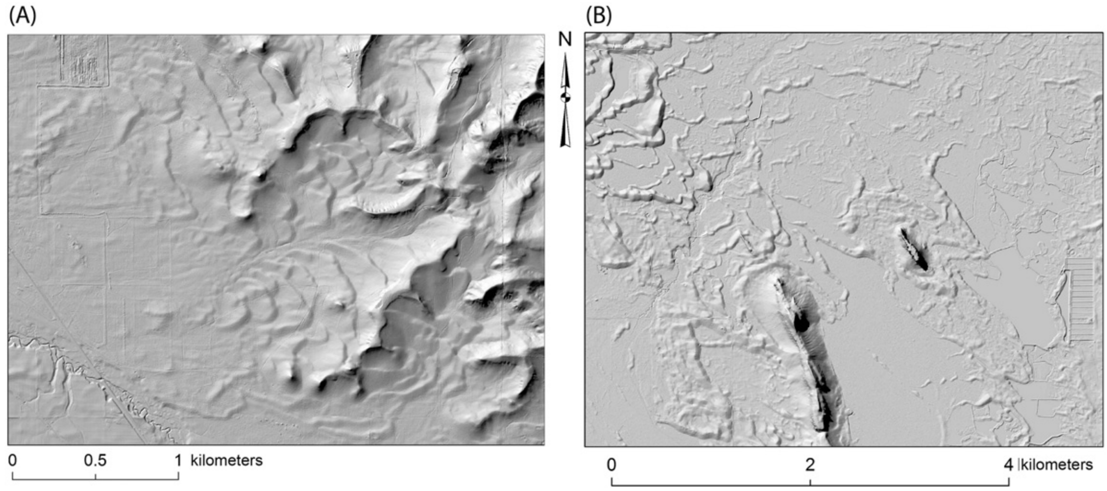

Fig. 13. Dunes on the Central Sand Plain of Wisconsin, in shaded relief images from LiDAR data (obtained by Monroe and Adams counties, distributed through WisconsinView portal, http://www.wisconsinview.org/). (A) Parabolic and transverse ridge dunes on slopes of narrow bedrock ridges, near Sparta, Wisconsin (Fig. 1). (B) Parabolic dunes with a range of sizes on bed of Glacial Lake Wisconsin and outwash plain to the east (Fig. 3). Large dunes on west side of image were apparent on older topographic maps, but small dunes to the east were not.

location deep in a re-entrant in the ice margin (Fig. 1). Interaction with eolian processes is also likely here; for example, Stanley and Schaetzl (2011) found a mantle of eolian sand or loam west of the incised Black River, with thin loess to the east of the river valley. This abrupt transition was interpreted as the result of the river valley acting as a topographic barrier, limiting the eastward extent of an active surface of sand transport and allowing loess accumulation farther downwind (eastward).

6. Directions for future research

As reviewed here, abundant evidence exists for widespread permafrost, eolian activity, and accelerated hillslope erosion in the study area, around the peak of the last glaciation. It should also be clear from this review, however, that many of the sediments and landforms providing this evidence remain poorly dated, and quantitative estimates of the rates of erosion and sediment accumulation are generally lacking. Thus, a key direction for future research should be the much more widespread application of modern geochronological techniques. Radiocarbon dating has provided a great deal of valuable data on the chronology of loess and colluvium accumulation but depends on site-specific circumstances that allow preservation of charcoal, wood (generally below the water table), or gastropod shells (in unleached calcareous sediments). Progress has been made in the application of optically stimulated luminescence (OSL) dating to late Wisconsin dune sands in or near the study area (Booth et al., 2005; Rawling et al., 2008; Miao et al., 2010; Loope et al., 2013; Hanson et al., 2014), but application to loess has been much more limited (Schaetzl et al., 2014). Given the importance of loess stratigraphy as evidence for contrasting rates of erosion across local landscapes in the study area, developing more robust chronology for the loess sequence should be a high priority. Loess of the last glaciation can be dated with quartz OSL, especially where it is coarse in source-proximal settings like the eastern (downwind) edge of the IES. The emerging technique of post-IR infrared-stimulated luminescence dating (Buylaert et al., 2009; Thiel et al., 2011) may provide better chronology for older loess units.

Methods based on cosmogenic nuclides (Gosse and Phillips, 2001; Granger et al., 2013) have had surprising limited application within the study area, mainly focused on the goal of dating glaciation (Colgan et al., 2002; Ullman et al., 2015). One early study used meteoric ^{10}Be accumulation to estimate the exposure age (i.e., the residence time at the land surface) of a now-buried profile of the Sangamon Geosol within the study area in northern Illinois. The estimated exposure age of about 100,000 years, incorporating a correction for inherited ^{10}Be, is consistent with soil formation starting at the end of OIS 6 (the Illinoian glaciation) and continuing until burial around 55 ka (Curry and Pavich, 1996).

Despite the limited application of these methods so far, the potential for new insights from a variety of cosmogenic nuclide methods seems high, although the nature of the study area also suggests some potential complications and limitations. Late Quaternary fluvial sediments of the study area are often rich in quartz sand and should be well-suited for analysis of in situ-produced ^{10}Be and ^{26}Al to produce basin-averaged estimates of rates of denudation (Brown et al., 1995; Bierman and Steig, 1996; Granger et al., 1996; Von Blanckenburg, 2005). To focus strictly on the hypotheses of accelerated hillslope erosion under periglacial conditions, this approach could be applied to well-dated late Wisconsin and Holocene sediments in basins that were entirely outside the ice margins of the last glaciation (e.g., Knox, 1983, 1996, 2000; Bettis and Autin, 1997; Mason and Knox, 1997; Bettis et al., 2008). Besides the potential contrast between Holocene and full-glacial samples, it would be of interest to test whether samples from a range of depths in thick late Wisconsin fluvial sediments record an increase in apparent rate of denudation as accelerated erosion progressively removed material with long surface exposure under the previous regime of slower erosion. The entire range of full-glacial to interglacial denudation rates may be too low to register as a significant change with this method (Schaller and Ehlers, 2006), but that in itself would be an important finding. Recycling of older fluvial sediment is a complication in some settings (Hidy et al., 2014) but may not be significant in the study area because of the limited volume of sediment storage possible in the narrow bedrock valleys, relative to the upland areas producing fresh sediment.

Another potential application of cosmogenic nuclide analysis follows from the interpretations of soil-geomorphic evidence on the IES that were discussed above. If past assumptions about how soil morphology reflects soil age are correct, the well-developed paleosols within the paha should have had a relatively long residence time – reflecting low rates of erosion in their local geomorphic setting – before burial by loess during the last glaciation. This long residence time should have resulted in substantial accumulation of in situ-produced ^{10}Be and ^{26}Al (loss through radioactive decay after burial by late Wisconsin loess would be limited and can be accounted for if the overlying loess is dated with OSL and/or radiocarbon). In contrast, if the much more weakly developed soils across the rest of the IES landscape truly represent a fresh start for pedogenesis after accelerated late Wisconsin erosion, this should also be reflected by lower nuclide accumulation corresponding to surface exposure only during the Holocene. A somewhat similar comparison might be possible in the Driftless Area and other high-relief parts of the study area, between paleosols buried by late Wisconsin loess and surface soils on adjacent steep hillslopes. The contrast in nuclide accumulation in these cases should be detectable as it would be of at least the same order of magnitude as the contrast between surfaces exposed on moraines formed in the last glaciation and those formed in OIS 6 (Gosse and Phillips, 2001; Balco, 2011). Inheritance would be an issue here as well, but could probably be addressed through analysis of ^{10}Be and ^{26}Al depth profiles (Anderson et al., 1996). An alternative approach would be to compare accumulation of meteoric ^{10}Be between the buried paleosols and nearby soils thought to have much shorter residence times. Many of the soils involved are relatively clay rich and would, therefore, have optimum retention of ^{10}Be; in addition, substantial data now exist on meteoric cosmogenic nuclide accumulation in soils worldwide for comparison (Willenbring and von Blanckenburg, 2010).

I would also like to emphasize another, quite different avenue toward better understanding of how the landscapes of the study area responded to past periglacial conditions. As mentioned earlier, a large body of published research exists on relict periglacial features, hillslope deposits, and eolian sediments and landforms attributed to periglacial conditions in Europe. The literature on hillslope and eolian deposits dating to the last glaciation in the study area and through the broader periglacial zone south of the Laurentide Ice Sheet is depauperate by comparison, particularly in its description of sedimentological and stratigraphic details. This disparity cannot be quickly remedied, given the current funding environment and the relatively few researchers working today on the nonglacial Quaternary history of the study area. Nonetheless, even limited efforts to compare the sedimentology and stratigraphy of periglacial hillslope deposits across the study area with those reported from Europe could yield important insights on common patterns and contrasts, which might in turn be related to specific environmental conditions and processes. Examples of specific topics studied in detail in various parts of Europe and worth pursuing in the study area include (i) the sedimentology and micromorphology of periglacial slope deposits and their three-dimensional stratigraphy as revealed by geophysical methods (van Steijn et al., 1995; Harris, 1998; Bertran and Texier, 1999; Gerber et al., 2007); (ii) cryoturbation structures, including those associated with ice-wedge casts and slope deposits (Vliet-Lanoë, 1988, 1991; Vandenberghe, 1992); (iii) the occurrence of ice-wedge casts or other frost fissures within the loess sequence (Jary, 2009); (iv) the sedimentology of periglacial eolian sands (Ruegg, 1983; Schwan, 1986, 1987; Koster, 1988; Vandenberghe, 1991; Kasse, 2002); and (v) the sedimentary record of fluvio-eolian interactions in the periglacial environment (Schwan, 1987; Van Huissteden et al., 2000; Kasse et al., 2007).

Only limited comparisons have been made between presumed periglacial features and sediments in the study area and those of modern permafrost environments, with the notable exception of ice-wedge casts and relict sand wedges. Other opportunities to use modern analogues remain largely unexplored. For example, it is worth considering whether footslope silts in high-relief parts of the study area originated as the organic-rich silts found in permafrost regions, referred to as yedoma (Kanevskiy et al., 2011; Strauss et al., 2012; Schirrmeister et al., 2013). The footslope silts of the study area could be yedoma that has lost an originally high content of organic matter through oxidation in the Holocene, or alternatively, they may record accumulation of silt with little organic content in an environment that was different in some important way from those in which yedoma occurs today. Exposures of the footslope silts in the study area are sufficient that their internal features and distribution in the landscape could be studied in detail in an effort to resolve this question. Emphasis on the need for more detailed observations of this kind would no doubt meet with the approval of James Knox, who always saw careful field observation as the most essential tool of geomorphology.

7. Conclusions

The concept of accelerated erosion during glacial periods in regions like the study area is important in part because of its possible relevance to the overall contribution of low-elevation continental landscapes to global sediment flux (Willenbring et al., 2013; Larsen et al., 2014b). Cosmogenic nuclide-based estimates of erosion rates in the study area using widely distributed late Pleistocene fluvial sands could allow a better estimate of that contribution. Even full-glacial rates of erosion may turn out to be quite low by comparison with mountains experiencing tectonic uplift or even many low elevation drylands, as suggested by the study of Knox (1989). However, the scope of geomorphology is not limited to areas that contribute significantly to global sediment flux or that display record-breaking rates of erosion and soil production (Larsen et al., 2014a), interesting as those landscapes may be. In fact, landscapes like the study area are of geomorphological interest because of, not in spite of, the persistence of sediments and landforms developed in periglacial environments quite different than the present. This is the direct consequence of very low Holocene rates of erosion (prior to widespread human impacts in the last two centuries), and it is in stark contrast to landscapes where high rates of erosion make it plausible to assume that hillslopes and soils on them are in some form of dynamic equilibrium.

When rare modern episodes of dramatic erosion affect the study area, the processes involved are strongly influenced by the legacy of the last glacial environment, as often emphasized by James Knox. For example, as mentioned earlier, an extreme rainfall event in 2007 triggered large landslides and extensive gully erosion in a relatively high-relief part of southeastern Minnesota and an adjacent portion of the Wisconsin Driftless Area. Much of the total mass of sediment mobilized by this event was colluvium or fluvial sediment deposited in the periglacial environment of the last glaciation. Landslides stripped colluvium from steep slopes, and gullies deeply incised late Wisconsin fill terraces. The stability of the hillslopes under all but the most extreme rainfall is largely a function of the high frictional strength of the rocky colluvium (Mason, 1995). The geomorphic response to minor Holocene climate fluctuations (Knox, 1985, 1993) and greatly expanded human impacts in the Driftless Area in the nineteenth century (Knox, 1987) were also conditioned by sediments and landforms that are the legacy of the last glacial period. For example, the rocky colluvial mantle on steep slopes may have protected them from more severe erosion in response to grazing or logging; intense gully erosion is more often observed in loess or sandy to silty footslope deposits. The dense tills with only a thin loess mantle on the IES and the eolian sand mantle of Wisconsin's Central Sand Plain have strongly influenced historical land use in both regions.

Given the importance of this late Pleistocene legacy in the Holocene and modern landscape, there is clearly a need to better understand the processes that were active in the periglacial environment and how they shaped the local distribution and sedimentology of hillslope and fluvial deposits. We have some understanding of the distribution of permafrost and tundra-like vegetation across the study area during the last glaciation, based on ice-wedge casts and paleoecological data, although more well-studied localities and better chronology would certainly be helpful. Sediments mantling hillslopes in higher-relief areas are known to be of late Wisconsin age, based on stratigraphic relations and radiocarbon dating in a few localities. The arguments for widespread, substantial hillslope erosion made by Ruhe and others, on the basis of soil-geomorphic evidence and loess stratigraphy, are still largely valid; and new tools such as LiDAR elevation data and OSL dating are revealing a great deal about the importance of eolian processes in the periglacial environment.

Quantitative estimates of erosion rates using cosmogenic nuclides, which have revolutionized geomorphological research in other regions, are certainly needed in the study area. A need also exists for more time-consuming and less immediately rewarding field-based work on the sedimentology and stratigraphy of late Wisconsin hillslope, fluvial, and eolian sediments and landforms, in the tradition of James Knox and his good friend Robert Ruhe. More thorough study of this evidence, at many more localities, will allow more effective comparison with modern permafrost environments and with the well-studied geomorphic record of past periglacial environments in Europe. The landscape of the Upper Mississippi River basin still bears a strong imprint of the last time it was 'up in the refrigerator,' but a great deal is still to be learned about how that imprint developed.

Acknowledgments

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