Late Cenozoic geomorphology of a bedrock-dominated landscape adjacent to the Laurentide glacial limit: ....

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Late Cenozoic geomorphology of a bedrock-dominated landscape adjacent to the Laurentide glacial limit: southeastern Nebraska, USA

by

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with 12 figures

Summary. The Rose Creek Escarpment in southeastern Nebraska, USA is a remnant Early Pleistocene or Late Pliocene river valley representing the northern terminus of a series of Cretaceous bedrock escarpments extending northward from central to west-central Kansas. Superficial (~ 1–4 m deep) and largely slope-accordant deformation of the Greenhorn Limestone (Upper Cretaceous) is common under slopes on the escarpment. Deformation features include simple involutions, folds, dipping strata, thrusts, detachment and decollement, boudinage-like bedding, brecciation, and faulting. It is likely that much of this deformation occurred because of ground ice growth and melting, solifluction, and slumping during the last glacial maximum, when the Laurentide ice margin was nearly 300 km to the north. The presence of bentonite layers facilitated rotational slumps and downslope movement of rock masses in general. The local stratigraphy of overlying loess deposits provides no strong evidence to refute a Late Pleistocene (Wisconsinan) origin for the deformation, although some sites showing deformation and brecciation lie downslope from surfaces that are also underlain by the Illinoian Loveland Loess. Considering the pre-Illinoian glacial history of the region and the age of the escarpment itself, it may indeed have been subject to multiple episodes of periglaciation after the onset of continental glaciation in the Late Pliocene.

Résumé. *Défunte géomorphologie cénozoïque d'un paysage Roche en place-Dominé à côté de la limite glaciaire de Laurentide : Le Nébraska Du sud-est, Etats-Unis.* – L'escarpement de Rose Creek dans le sud-est de Nebraska, E. U. est un restant du Pléistocène inférieur ou du pliocène supérieur d'une vallée fluviale qui signale le terminus nord d'une série d'escarpements de substratum rocheux crétacée qui s'étendent au nord du Kansas central jusqu'au Kansas central ouest. Une déformation superficielle (1–4 m de profondeur) et en grande partie concordant au versant du Castine Greenhorn (Crétacé supérieur) arrive souvent sous les pentes sur l'escarpement. Les caractéristiques de la déformation comptent des involutions simples, des plies, des strates plongeantes, des butées, des détachements de collements, des stratifications boudineuses, la formation des brèches et des failles. Il est probable que une grande partie de cette déformation est arrivée à cause de la croissance de la glace de la terre et de la fondante de la glace, solifluxion, et glissement synsédimentaire pendant à l'époque finale ou le glacier a été maximal, quand la rive du glacier Laurentides a été presque 300 km au nord. Le degré de présence des couches à bentonite a animé slumps en rotation et un mouvement de décente sur les pentes des masses de la roche en général. La stratigraphie locale des glissements de loess sur jacents ne

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fournissent pas d'évidence forte qui réfute une origine de l'époque Pléistocène supérieure (Wisconsinan) de déformation, malgré le fait que il y a des sites qui démontrent déformation et formation de brèches qui sont en pente décroissante de surface qui sont aussi sous-jacents du Loess Loveland (Illinoian). En pensant à l'histoire glacielle pre-Illinoian de la région et à l'age de l'escarpement elle-même, il a peut bien été sujet à époques de periglaciation après le commencement de glaciation continental dans le Pliocène supérieur.

1 *Introduction*

The meeting of three major landform regions around the Laurentide glacial limit in southeastern Nebraska (fig. 1) provides an exceptional opportunity to examine landscape evolution at a critical physiographic boundary. The Smoky Hills, which are dominated by shallow bedrock, abut till- and/or loess-covered terrains to the north (figs. 1, 2). This paper investigates recently-discovered superficial bedrock deformation at the northern end of the Smoky Hills (fig. 1) in the context of landscape evolution.

2 *Methods*

Local physiography was characterized through a focal analysis of relief in circular neighborhoods (r = 18 cells) on digital elevation models (DEMs) (fig. 2). Digital soil maps, outcrop maps, and other data were integrated in a GIS project to produce derivative surficial geologic maps. Digital photo panoramas were used to diagram outcrops.

3 *Geomorphology*

3.1 *Regional physical geography and geochronology*

Gently-westward-dipping (~2.0 m/km) Cretaceous strata cropping out across central to northern Kansas produce conspicuous east-facing escarpments with relief exceeding 100 m. These escarpments, generally referred to as the Smoky Hills (Adams 1903, Frye & Schoewe 1953, Wilson 1978, Steeples & Buchanan 1983, Chapman et al. 2001), disappear northward into Nebraska under Late Cenozoic sediments.

Eastern Nebraska experienced multiple glaciations during the interval 2.7 Ma–600 ka (Boellstorff 1978a, 1978b, Aber 1991, Sarna-Wojcicki & Davis 1991, Izett et al. 1992, Gansecki et al. 1998, Colgan 1999, Rovey & Kean 2001, Roy et al. 2004, Balco et al. 2005). During the Wisconsinan Stage, the ice sheet came within 300 km of the study area (fig. 1). Few periglacial features have yet been documented in Nebraska (Wayne, 1991, 2003, Wayne & Guthrie 1993), and overall there has been little documentation of ground ice effects on bedrock around the Laurentide ice margin compared to elsewhere in the northern latitudes (e. g., Krüger 1933, Paterson 1940, Bradshaw & Ingle Smith 1963, McCabe 1969, Williams 1980, Büdel 1982, French et al. 1986, Ballantyne & Harris 1994, Kerney et al. 1963, Murton 1996, Mason 2000, Mills 2000, Murton 1996, Murton & Lautridou 2003, Murton et al. 2003). The Peoria Loess, Gilman Canyon Formation, and Loveland Loess were deposited over eastern Nebraska during MIS 2, 3, and 6, respectively (Bettis et al. 2003).

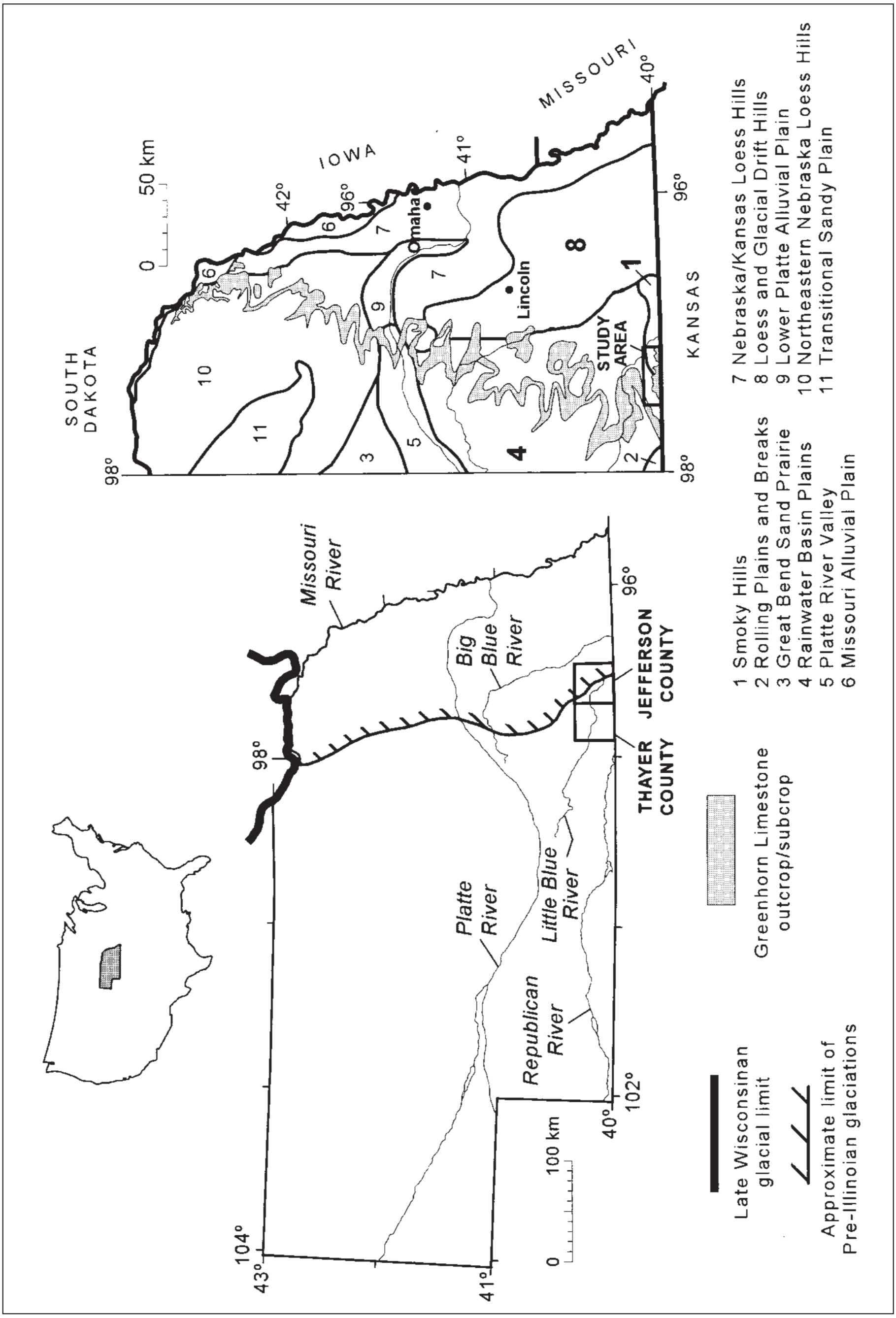

Fig. 1. Location map of study area. Geomorphic regions in eastern Nebraska are after CHAPMAN et al. (2001).

3.2 *Geology and physical geography of the study area*

The chief study area is the unglaciated "Rose Creek Escarpment" (RCE) of Joeckel et al. (2005a), the northern end of the Smoky Hills (figs. 1, 2). Maximum relief on the RCE (90 m) lies roughly *parallel* to bedrock dip, due to the incision of modern Rose Creek (figs. 2, 3). The strike-parallel escarpment edge to the east has about 75 m of relief, whereas the downdip (western) end of the RCE, has only 40 m of relief. There, the RCE is breached by a large, sediment-filled paleovalley extending east-northeastward from Kansas (fig. 2: ARP) and attributed by Fishel et al. (1949) to the eastward-draining ancestral Republican River. The modern course of that river now lies to the west (fig. 1). Overall, relief is much greater on the RCE (fig 2: area 1) than it is in adjacent areas (fig 2: areas 3 and 4). The RCE and an area of high relief, bedrock-dominated terrain across the Little Blue River (fig. 2: area 3a) delimit the partially-exhumed WSW-ENE-trending paleovalley, which is completely buried eastward of the glacial limit (fig. 2). The cross-valley remnant of high-relief bedrock terrain (fig. 2: 3a) is distinct from adjoining valley slopes (fig. 2: 3b, which are partially covered by regolith. A dissected terrace abutting the RCE (fig. 2: area 2) is underlain by coarse, poorly-sorted

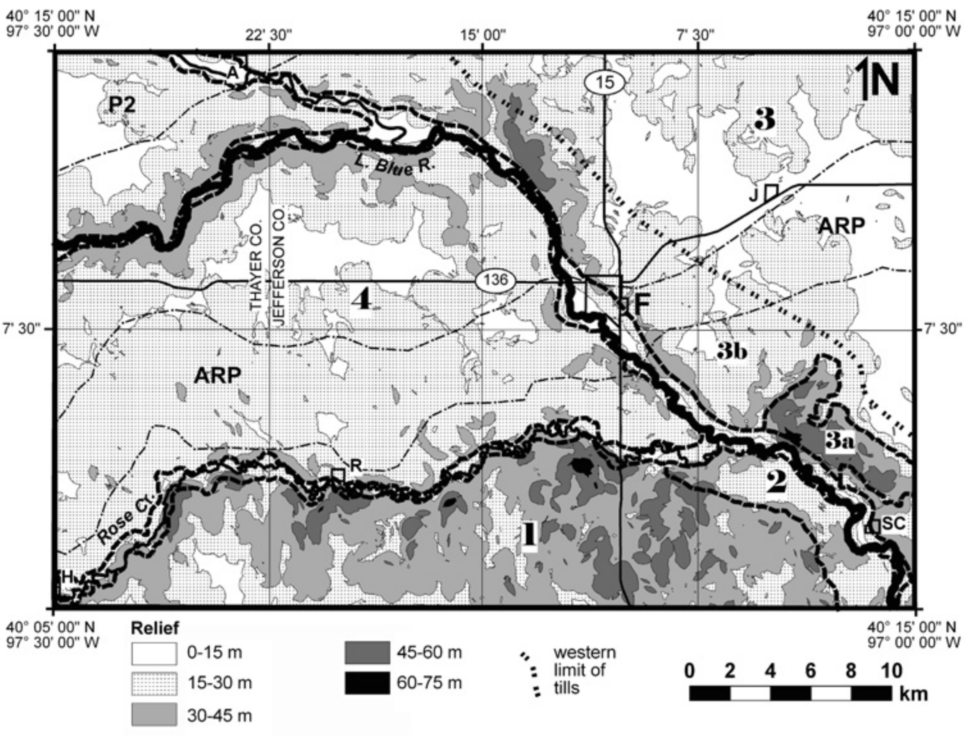

Fig. 2. Focal analysis of digital elevation model showing local relief across study area. Note putative ancestral Republican River palleovalley (ARP), and another paleovalley (P2). Large numerals indicate geomorphic subareas described in text. Towns/cities are: Alexandria (A), Fairbury (F), Hubbell (H). Jansen (J), Reynolds (R), and Steele City (SC).

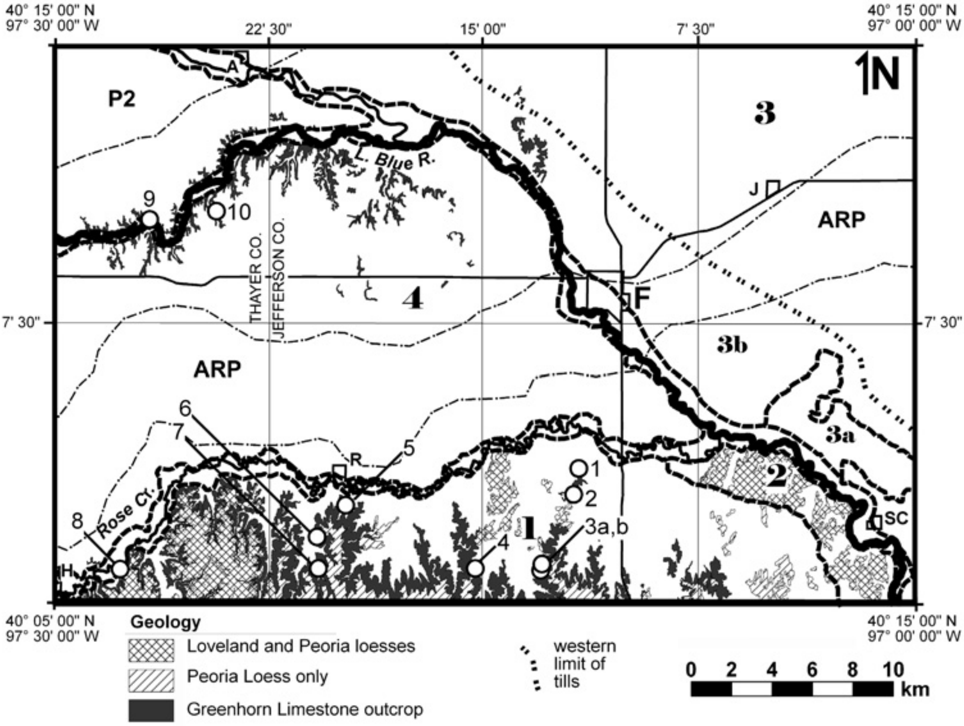

Fig. 3. Study sites in Greenhorn Limestone (see also fig. 2). Extent of Pleistocene loess is mapped only on Rose Creek Escarpment: area north of Rose Creek is covered by nearly continuous loess.

sands and gravels containing abundant cobbles and boulders from local bedrock and from glacial tills, and contrasting markedly with the Rocky Mountain-source sediments of the paleovalley fill (cf. Stanley & Wayne 1972).

Cretaceous bedrock under the RCE (fig. 4) area consists of: (1) the fluvial-estuarine sediments of the Dakota Formation (Brenner et al. 2000, Joeckel et al. 2004); (2) the marine Graneros Shale; and (3) the Greenhorn Limestone. Hill summits are capped by thin (0.6–3.2 m) Late Wisconsinan Peoria Loess, underlain by very thin (~ 0–1 m) Gilman Canyon Formation loess, which has well-developed soil structure. The Gilman Canyon Formation is absent where the Peoria Loess is very thin. Below the Gilman Canyon Formation, or the Peoria Loess where the latter is absent, is the Sangamon Geosol, a clay-enriched, reddish-brown paleosol that developed during late Illinoian and early Wisconsinan times (Bettis et al. 2003). Loveland Loess and older Pleistocene silts, at least partially of eolian origin, are widespread merely a few kilometers north of the RCE. On the RCE, however, Loveland Loess appears consistently only on the western dip-slope of the escarpment and atop the terrace bordering the eastern edge of the escarpment (figs. 2, 3). Late Pleistocene loess and Late Pleistocene to Holocene colluvium, consisting of reworked loess and eroded bedrock debris, irregularly mantles most slopes on the RCE. Paleogullies are filled with col-

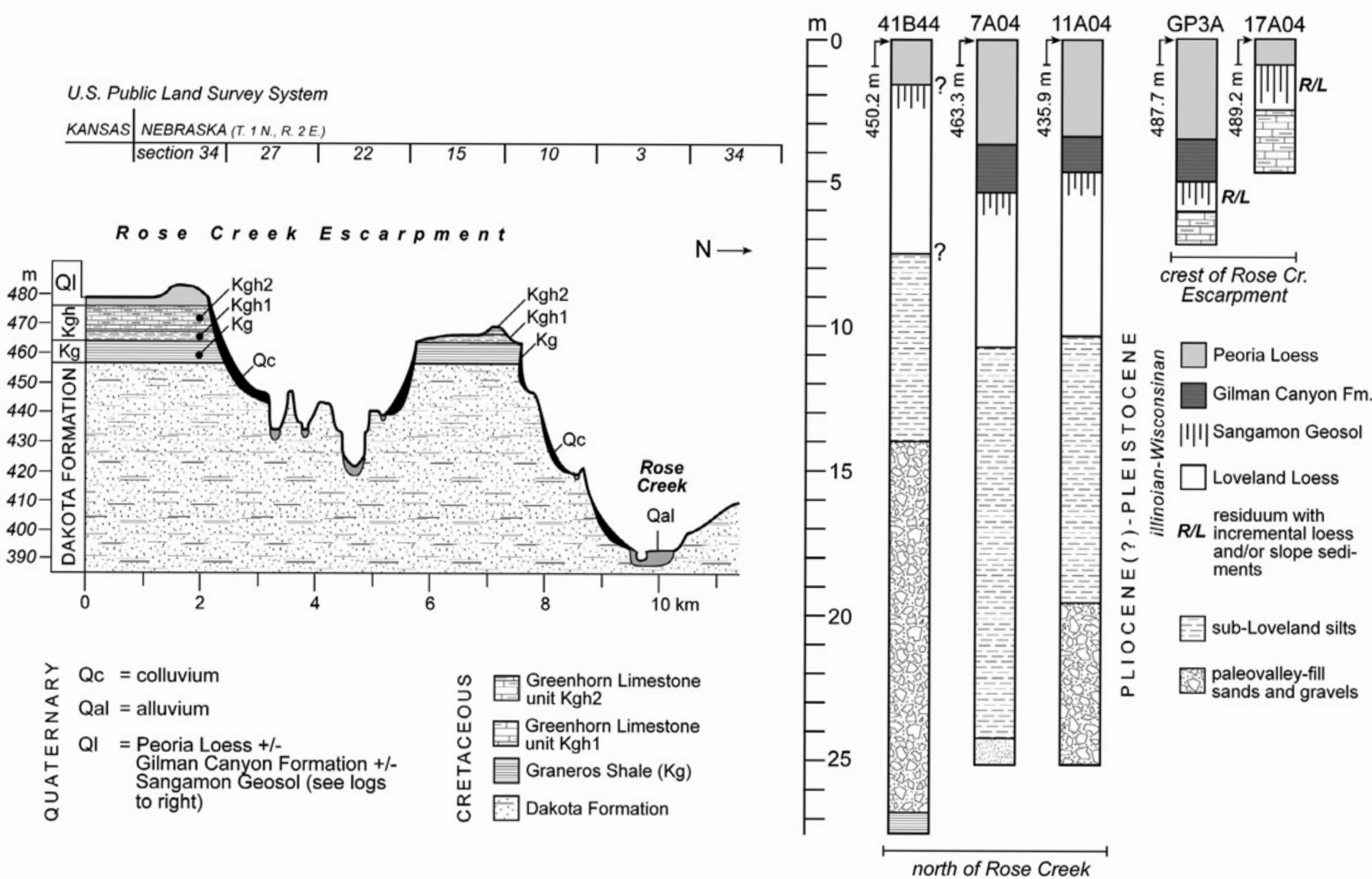

Fig. 4. A: North-south cross section of Rose Creek Escarpment. B: Core logs representative of stratigraphy through fill of putative ancestral Republican paleovalley and on Rose Creek Escarpment (see figs. 2, 3).

luvium (including transported, dark, soil material) and deoxidized loess and have no clear relationships with modern drainages, therefore they must predate the most recent episode of hillslope erosion. Locally, footslope colluvium is derived from eroded Illinoian and older loesses.

4 *Weathering response of bedrock*

4.1 *Comparison of weathering effects on different units*

Chemical weathering is locally prominent in the Dakota Formation and Graneros Shale (Joeckel et al. 2005b), but the Greenhorn Limestone shows, by far, the most striking response to physical weathering and slope processes. In this paper, it is subdivided informally into: (1) a lower unit (fig. 4: Kgh1) dominated by thinly- to thickly-laminated chalky shale to shaly chalk, roughly equivalent to the Lincoln and Hartland members in Kansas (Hattin 1975); (2) an upper unit (fig. 4: Kgh2), dominated by thinly- to medium-bedded chalky limestones, but also containing shaly chalk, and probably equivalent to the Jetmore and Pfeifer members in Kansas (Hattin 1975). Bentonite and bentonitic shale layers in the Graneros Shale and Greenhorn Limestone have plasticity indices of 45–63 %. A bentonite appears to separate Kgh1 and Kgh2 at localities 6 and 7 (fig. 2). The dominant weathering features of the Greenhorn Limestone are: (1) pronounced oxidative bleaching of shaly chalks (cf. Hattin 1975); (2) softening and brecciation by physical, chemical, and biological processes; (3) ferruginization by pyrite oxidation; and, most importantly; and (4) widespread, intense deformation of near-surface strata, the chief focus of this paper.

4.2 *Superficial deformation of the Greenhorn Limestone*

Deformation in the Greenhorn Limestone generally appears 1–4 m below the current land surface, and includes: (1) simple involutions, (2) downslope-trending folds on slopes, (3) dipping strata, (4) apparent thrusts of large masses of limestone downslope, (5) zones of detachment and decollement, (6) boudinage-like deformation, (7) brecciation, and (8) faulting. Land-surface expressions of these features are lacking.

Simple involutions are small (10–35 cm in width and as much as 20 cm deep), abrupt downwarpings of laminae in shaly shalk (fig. 5). They appear under nearly level or gently-sloping surfaces, in opposition to folds, which occur under slopes. Simple involutions were found much less frequently than folds, but this apparent rarity may merely be a bias generated by the location of road gradings on hillslopes.

Folds ($\leq$ 50 cm in height) are common in slopes on the chalky shales to shaly chalks of the lower Greenhorn Limestone (figs. 5–7). These folds are cuspate to elliptical, tight (interlimb angles of 30°–75°), and their axial planes are typically tilted downslope by 30° and 80° from the vertical. More lithified strata within folds are broken into trains of oriented fragments which, although they remain oriented relative to the form of the fold, also reflect the brittle behavior in those strata at the time of deformation (fig. 6). Less resistant shalier laminae or thin bentonite laminae within these strata do not exhibit such breakage and therefore underwent plastic deformation. In some cases, small folds clearly lie within larger masses, a meter or more in length, that were themselves folded or thrust downslope (fig. 6). At locality 4,

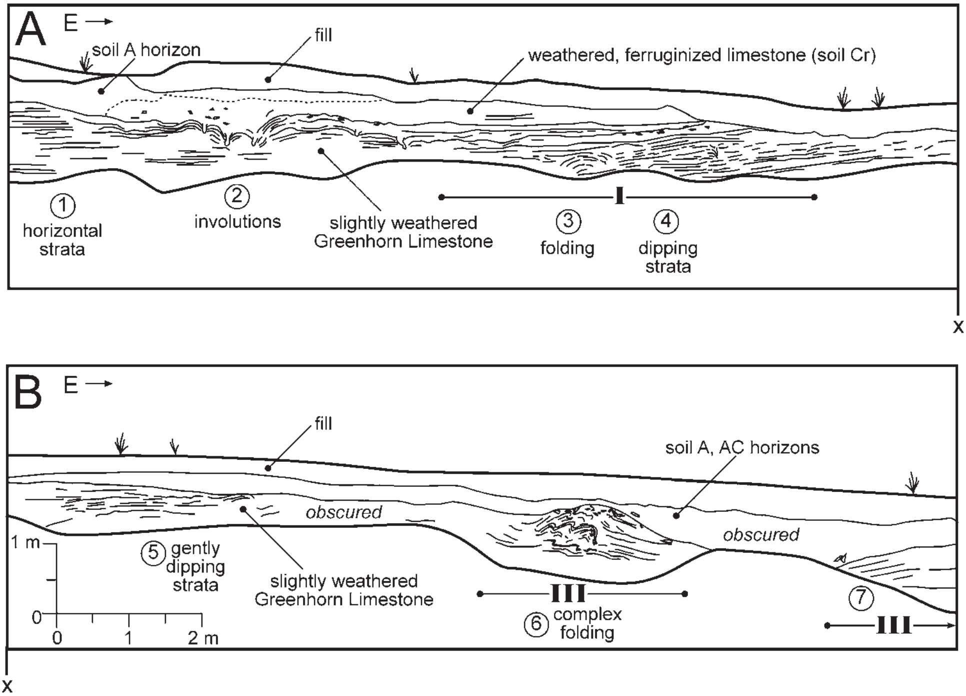

Fig. 5. Deformed shaly chalk of informal unit Kgh1 in cross-section of west-to-east backslope at locality 5. Bedding planes shown as lines. Two panels (A, B) articulate at "x". A: undeformed strata (1) transition into involutions (2) upslope, and dipping strata with folds (3, 4) downslope; note detachment and rotation of a large mass (I). B: Gently-dipping srata (5) have also been deformed by slippage. Complex folding (6; see also fig. 6) appears in translocated rock mass (II). Other dipping strata (7) were produced by basal detachment and rotation of a third mass (III). Soil horizons (A, AC, Cr) are indicated.

a downslope-trending series of asymmetrical, partially detached "roll-like" folds appears in association with asymmetrical, soil-filled cracks. These cracks widen upward and are almost all inclined downslope at angles of 35°–45° from the vertical, curving under the folds (figs. 6B, 7). Small faults, dipping downslope about 40°, appear in one of these folds (fig. 7).

Dipping strata (10° and 45°) are common in Kgh1 (figs. 8A, 9–11), but also appear in Kgh2. In several cases, dipping strata appear to be the results of the rotation of slump masses (figs. 9–11) along shear planes below the level of observation. Dipping strata within large blocks of Greenhorn Limestone that underwent downslope translation, are inclined downslope (fig. 12). Downslope-dipping strata can usually be associated with the apparent downslope "thrusting" or sliding of large masses exceeding 8 m in length and 2 m in thickness (fig. 12; cf. Chandler et al. 1976, Chandler & Johnson 1976). This phenomenon appears to be confined to the basal beds of Kgh2, and the best example may be locality 2, where bentonitic shales are deformed between two translocated masses of limestone (fig. 12).

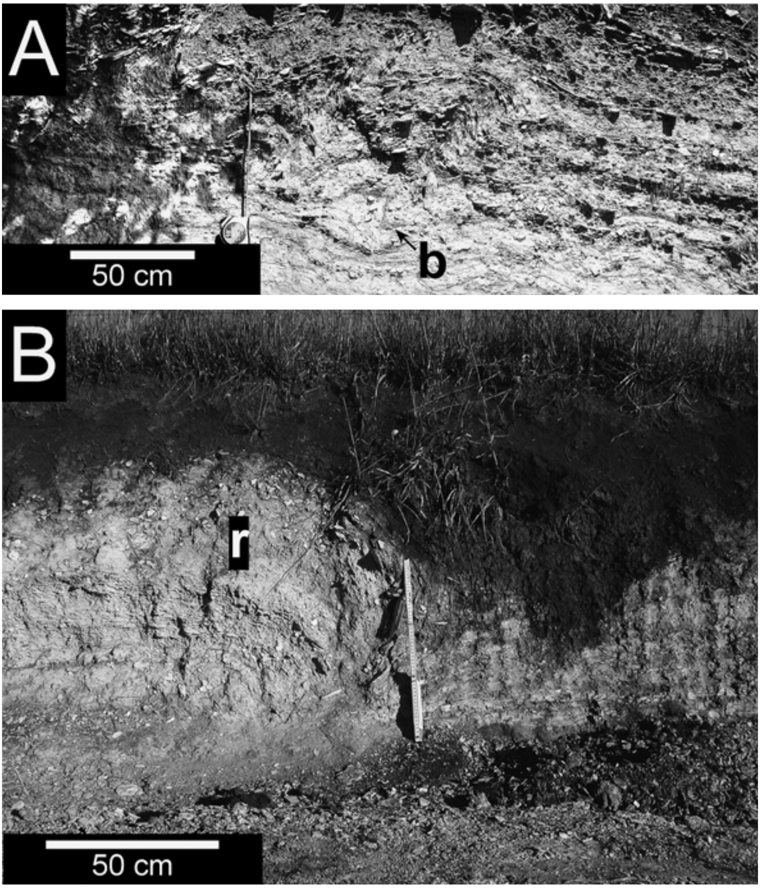

Fig. 6. Folding in informal unit Kgh1. A: Tight, folds with axial planes tilted downslope at locality 5 (see fig. 5); detachment occurred below thin bentonite (b). B: Roll-like fold (r) at locality 4 (see fig. 7).

Detachment and decollement are apparent under both folds and thrusts in association with bentonite seams within folds (figs. 9, 10). Boudinage-like deformation appears in association with the former plastic flow of surrounding or underlying bentonites, bentonitic shales, and chalky shales (fig. 12).

Brecciation of laminated shaly chalk in Kgh1 (fig. 8B) can be observed at a few localities. Chaotic brecciation at locality 7 (fig. 1) appears as a mass of randomly-oriented, fitted, angular fragments of weathered shaly chalk, each 1–7 cm thick and 2–20 cm in length (figs. 8B, 9, 10). Chaotically brecciated laminated shaly chalk at least 2 m thick underlies deformed strata downslope from a paleogully filled with deoxidized loess (*sensu* Ruhe 1969) and colluvium (figs. 9, 10). Soils developed in this fill at this locality have Bt horizons with clay coatings on ped faces.

5 *Discussion*

5.1 *Origin of superficial deformation of bedrock*

There is no evidence for major tectonic deformation in the study area, and the modern-slope-accordant nature of most of the deformation in the Greenhorn Limestone refutes both syndepositional and tectonic origins. Exceedingly rare major mass movements known in the area can be attributed to acceleration by engineering activities (see Joeckel et al. 2005b). Therefore, deformation must have occurred by mass movement after the development of existing slopes, but prior to the most recent episode of slope erosion. Such movement would require abundant water (e. g., Kerney et al. 1963), and therefore segregation of ice in the near-surface environment under periglacial conditions prior to the Holocene, followed by thawing and solifluction, is the most viable explanation (cf. Greene & Eden 1973, Fitzpatrick 1987, Hutchinson & Hight 1987, Wayne 1991, Ballantyne & Harris 1994). The lamination and thin bedding of Kgh1 sediments and their moderate to strong fissility facilitated deformation by providing myriad slip planes along which mass movement could occur. Bentonites must have also played an important role (cf. Kerr & Drew 1969, Ehlig 1998, Whisonant et al. 1998).

Observations made by other authors (e. g., Kerney et al. 1963, McCabe 1969, Williams 1980, Büdel 1982, French et al. 1986, Ballantyne & Harris 1994, Murton 1996, Murton & Lautridou 2003, Murton et al. 2000, 2001, 2003) support a hypothesis of periglaciation and solifluction. Brecciation and apparent solifluction effects on English chalks (e. g., Dines et al. 1940, Kerney et al. 1963, Murton 1996, Murton & Lautridou 2003) are noteworthy because of lithologic similarities. Also, slumping of the type indicated by small-scale faulting and block rotation

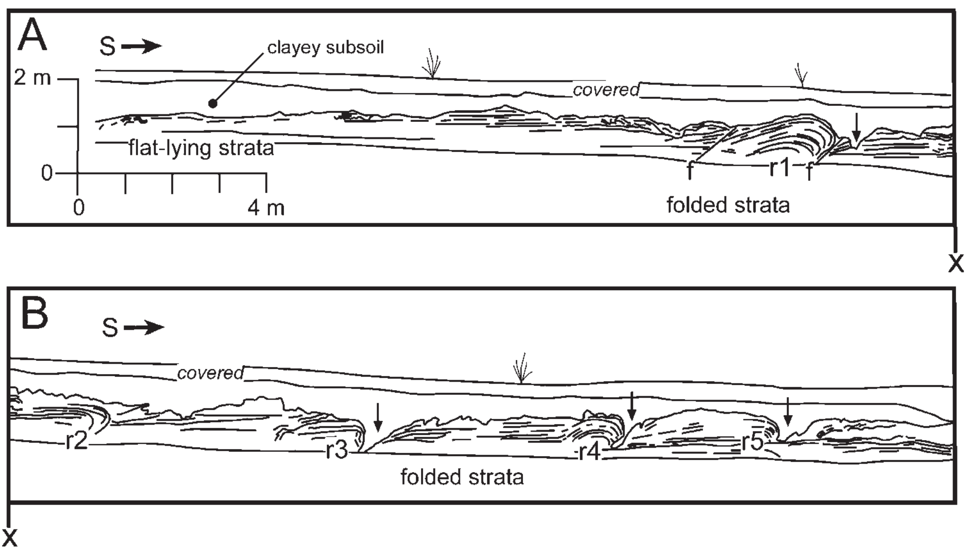

Fig. 7. Roll-like folds (r1-r5) trending downslope at locality 4 (see fig. 6B). Soil-filled cracks (arrows) appear to have formed under tension. Note small faults (f).

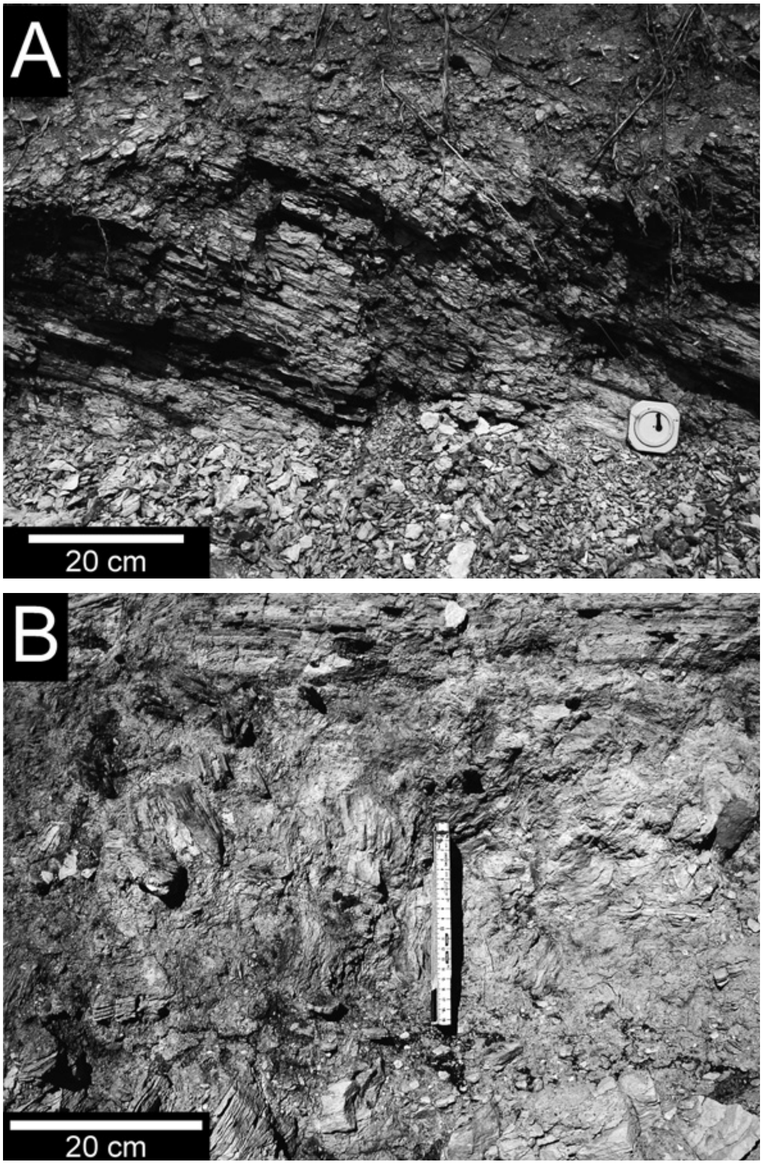

Fig. 8. A: Dipping (26°) strata in informal unit Kgh1, underneath thick bentonite at locality 7 (b* in figs. 9, 10). B: Brecciation in Kgh1 at locality 7 (see fig. 9).

occurs in ice-rich soils (e. g., Yershov 1990, Williams & Smith 1989). The scale of deformation features in the Greenhorn Limestone is similar to depths observed in both solifluction and ground-ice-weathering profiles elsewhere in the Northern Hemisphere (e. g., Krüger 1933, Bryan 1946, Kerney et al. 1963, French et al. 1986, Hutchinson & Hight 1987, Ballantyne & Harris 1994, Murton 1996).

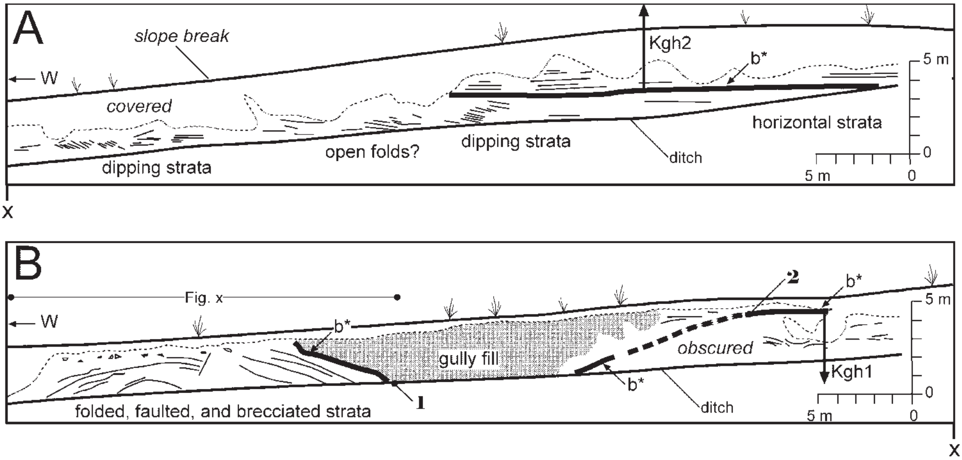

Fig. 9. Slope along north-flowing drainage at locality 7. Two panels (A, B) articulate at "x" and represent about 100 m of continuous exposure. A: Horizontal strata in informal unit Kgh2 above thick bentonite (b*), overlying dipping strata in Kgh1. B: Downslope dip (~ 12°) of same bentonite (b*). Strata above bentonite show slight drag folding. Bentonite (b*) dips to ditch level and remerges on opposite side of gully fill; vertical offset of bentonite from "1" to "2" is 3.65 m. Gully fill has well-developed surface soil with Bt horizon. Downslope (west) from gully fill, strata are highly deformed (see fig. 10).

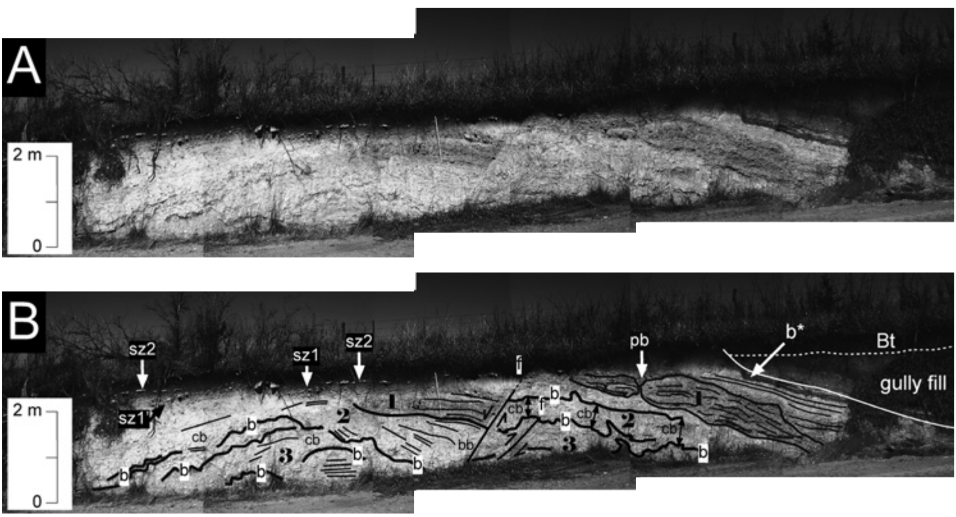

Fig. 10. Deformed Greenhorn Limestone and gully fill at locality 7 (see fig. 9). A: Photomontage of deformed strata. B: Deformed strata (thin lines), including distinct stratigraphic horizons (1, 2, 3) that have been folded and faulted (f). Thin (< 40 mm) bentonite seams ("b") are contorted, folded, and show boudinage-like separations; thick, marker bentonite seam (b*) dips upslope because of rotational slumping (see fig. 11). Chaotic brecciation (cb) is common. Stone zone sz2 consists of transported limestone channers, whereas sz1' consists of an *in-situ*, folded and weathered limestone layer; the two stone zones merge at sz1.

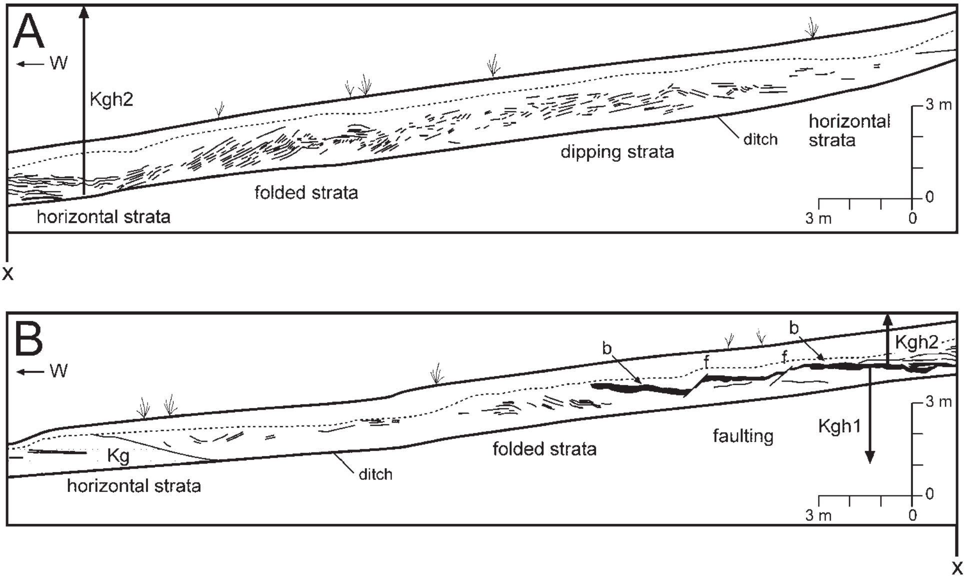

Fig. 11. Slope along north-flowing drainage dissecting RCE at locality 6. Two panels (A, B) articulate at "x" and represent about 90 m of continuous exposure. A: Horizontal strata (thin lines) appear upslope in informal unit Kgh2, but give way downslope to dipping strata, which are folded farther downslope, against horizontal strata at the base of Kgh2. B: Thick bentonite (b) has been broken by small faults (f) and also shows evidence for flow (changes in thickness and undulating geometry). Gentle folds appear farther downslope. At far left, Kgh1 appears to have slid over Graneros Shale (Kg).

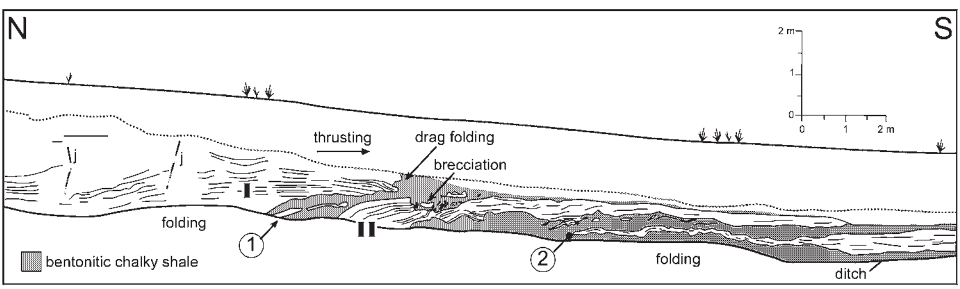

Fig. 12. Downslope "thrusting" by downslope sliding of one mass of informal unit Kgh2 (I) over another mass of Kgh2 (II) near the Kgh1-Kgh2 contact at locality 2. Observable are: subdued "drag folding" of strata, brecciation, and boudinage-like deformation of individual beds (1, 2). Strata dominated by thick bentonite and bentonitic shale (shaded) have undergone downslope plastic flow. Two joints (j) cross-cut deformation structures. At the footslope of the hill farther south (not shown), shaly chalks of informal unit Kgh1 show very open folds 50–80 cm in width.

5.2 *Chronology of landscape development*

The Early Pleistocene-Pliocene age assigned by Fishel et al. (1948) to the sediments filling the putative ancestral Republican paleovalley (fig. 2: ARP) is supported by the following observations: (1) pre-Illinoian tills overlie paleovalley deposits east of the Little Blue River; (2) the eastward extension of the ARP across that limit demonstrates a lack of ice-margin effect on the position of the ancestral Republican River; (3) pre-Loveland-Loess silts overlie the ARP west of the Little Blue River; and (4) the Late Pliocene-Early Pleistocene proboscidean *Stegomastodon* was found in silts above paleovalley sands and gravels north of Reynolds Nebraska (figs. 2, 3 "R"). This fossil was found in association with one of the Pearlette tephras, definitely making it older than 600 ka (Sarna-Wojcicki & Davis 1991, Izett et al. 1992, Gansecki et al. 1998). Fishel et al. (1948) described *Stegomastodon* from the ARP in northern Kansas.

High-relief slopes astride the Little Blue River match up as remnants of a formerly-continuous valley wall of the ancestral Republican River (fig. 2: areas 1 and 3a). The Little Blue River itself represents drainage diverted along an ice margin (cf. Reed & Dreeszen 1965, Wayne et al. 1991). Outwash was deposited on the terrace abutting the current RCE (fig. 2: area 2) only after the river breached the escarpment. The nearest unequivocally preglacial sediments to the study area are about 100 km west in Webster County, Nebraska, where the youngest mammal fossils date to ~11 Ma (Voorhies 1990), suggesting a regional biostratigraphic lacuna of about 8 million years. Therefore, a major episode of fluvial incision occurred well after 11 Ma and probably much later (Stanley & Wayne 1972, Aber 1997, Swinehart & Diffendal 1998). Slopes on the Rose Creek Escarpment must have undergone erosion of Loveland Loess in latest Illinoian-early Wisconsinan times. At locality 7, Bt horizons in post-deformation soils probably indicate pre-late-Holocene development (cf. Bettis 1992, Mandel & Bettis 2001). Solifluction on the Rose Creek Escarpment must have occurred at least as long ago as the Wisconsinan Stage. Almost all published examples of Quaternary periglacial features in North America (e. g., Sharp 1942, Schaefer 1949, Wayne 1967, 1991, 2003, Johnsson 1990, Wayne & Guthrie 1993, Walters 1994), in fact, have been attributed to Wisconsinan cold climates.

6 *Conclusions*

The Rose Creek Escarpment existed in some form in preglacial or early-glacial times. Superficial bedrock deformation there records periglaciation and solifluction, which likely occurred during the late Wisconsinan (the last glacial maximum in North America), although it is, as yet, impossible to be certain that some deformation did not take place earlier. If periglaciation and solifluction occurred during the late Wisconsinan, then it would have been taking place some 300 km south of the ice margin – far more southward than the periglacial features that have already been described from Nebraska (Wayne 1991). We suggest that relict periglacial effects on bedrock slopes in eastern Nebraska and nearby areas are insufficiently recognized (cf. Emerson 1970; Creemens et al. 2004). Therefore, northern European bedrock-escarpment slopes with complex histories (e. g., Chandler et al. 1976, Chandler & Johnson 1976) are analogs that should be applied heuristically to the North American Midcontinent.

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