**Parent Material Influence on Soil Response to Vegetation Change, Southeastern
Minnesota, U.S.A.**

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ABSTRACT

Soil morphology changes dramatically across the former transition from forest to grassland in the
Midwestern U.S.A. That vegetation boundary shifted as a result of Holocene climatic change and
fire suppression following Euroamerican settlement, but the timescale of soil response to those
vegetation changes and the factors that influence it are poorly known. On steep colluvial slopes
of southeastern Minnesota, Mollisols with thick, dark A horizons typically associated with
grassland are found today under deciduous forest. Soils with much thinner and/or lighter-colored
A horizons occur immediately up- and downslope of the forest-covered Mollisols. Most of the
soils with thick A horizons are not in topographic settings found to favor organic matter
accumulation in other landscapes. Principal components analysis highlights important axes of
textural and mineralogical variation among horizons of these soils, often related to parent
material properties. Soils with and without thick A horizons are separated along one principal
component heavily loaded by high pH, reflecting the presence of dolomite fragments in the
hillslope sediment that soils with thick, dark A horizons formed in. Stable C isotope analysis
reveals that the Mollisols with thick, dark A horizons under forest had C input from vegetation

with abundant C₄ grasses in the past. Public land survey data also indicate replacement of grassland or savanna by forest since 1854 on colluvial slopes where forest-covered Mollisols occur today. We propose that the soils with thick, dark A horizons under forest today reflect a lagged response to vegetation change, from grassland or savanna to closed forest. Their thick A horizons and some C₄-derived organic matter may be preserved because dolomite weathering releases abundant Ca⁺⁺ and Mg⁺⁺ that favor aggregation and organic matter stabilization. Soils with thin A horizons just up- or downslope may also have formed partly under grasses, but have responded more quickly and/or more substantially to vegetation change.

KEYWORDS: Soil response to vegetation change, C isotopes, Mollisols, Alfisols, parent material

1. Introduction

Major changes in soil morphology associated with the midlatitude forest to grassland transition were recognized early in the history of pedology, most notably in the work of Dokuchaev (1883). There is also a history of hypotheses on relict soil characteristics related to geologically recent shifts in the location of the forest-grassland boundary (Bronger, 1991; Miedema et al., 1999; Neustruev, 1927). In the Midwestern USA prior to the nineteenth century expansion of agriculture, there was a generally westward transition from deciduous forest to tallgrass prairie, with an intermediate zone of savanna or open woodlands (Fig. 1a). The complex spatial pattern of this vegetation boundary was recorded in detail by nineteenth century public land surveys and has been attributed to effects of both climate and wildfire, with discussion of the relative role of each factor continuing up to the present (Curtis, 1959; Grimm, 1984; Nelson and Hu, 2008; Shuman et al., 2009). Across this forest-prairie ecotone, there is a corresponding

48 transition from Alfisols to Mollisols (generally from Luvisols to Chernozems or Phaeozems in
49 the WRB classification). Studies in Minnesota, Iowa, and Illinois have documented marked
50 changes in soil morphology over a distance of a few tens of kilometers or less, with thinner A
51 horizons and more distinct textural contrasts between A or E and B horizons on the forest side of
52 the boundary (Bailey et al., 1964; Mason and Nater, 1994; Severson and Arneman, 1973; White
53 and Riecken, 1955).

54 The fact that these morphological changes occur at, or close to, the nineteenth century
55 vegetation boundary is especially noteworthy because paleoecological research has demonstrated
56 that the forest-prairie boundary has shifted substantially over the past 10,000 yr (Webb et al.,
57 1983; Williams et al., 2009). In fact, even where forest expanded into prairie within the last
58 millennium, Alfisols sometimes dominate the landscape recently occupied by forest (Grimm,
59 1983; Umbanhowar, 2004). Thus, it appears that under the right circumstances, soil morphology
60 can change rapidly in response to the transition from prairie to forest.

61 On the other hand, some Mollisols near the forest-prairie boundary in Minnesota are forest-
62 covered today, or were at the time of nineteenth-century land surveys (Almendinger, 1990; Buell
63 and Cantlon, 1951; Grimm, 1983). One obvious explanation for these anomalous soils is that
64 they have not had enough time to respond to recent advance of forest into prairie (e.g. Geis et al.,
65 1970). This raises the question of whether response time to vegetation change could be longer for
66 soils in certain geomorphic settings or parent materials. In many landscapes, A horizon
67 thickness and whole-profile organic carbon (OC) content are greater on gentle and concave
68 lower slopes than on steep and convex upper slopes. This pattern is attributed to downslope
69 translocation of OC-rich soil, greater plant productivity and OC addition to soils, and/or
70 conditions that reduce decomposition rates on lower slopes that are often depositional settings

(Aandahl, 1948; Berhe et al., 2008, 2012; Doetterl et al., 2012; Gessler et al., 2000; Schimel et al., 1985; Yoo et al., 2006). Thus, some geomorphic settings might allow longer-term persistence of thick OC-rich A horizons after forest invasion of prairie, or could even allow development of such horizons entirely under forest. Alternatively, soils with thick, dark A horizons may form entirely under forest in carbonate-rich parent materials, because continual release of abundant Ca^{++} and Mg^{++} from weathering carbonates favors OC accumulation. This has generally been the interpretation of *rendzina* soils on chalk and other carbonate rocks (Kubiena, 1953, p. 179; Schreier and Lavkulich, 1985; Ugolini and Tedrow, 1963), and has been invoked as at least a partial explanation for other examples of thick OC-rich A horizon formation under forest (Anderson et al., 1975a; Anderson et al., 1975b; Schaetzl, 1991). The mechanism responsible probably involves Ca^{++} and Mg^{++} acting as “bridges” that allow OC sorption on mineral surfaces, an important mechanism of OC stabilization (Oades, 1988; Mikutta et al., 2007; von Lützow et al., 2006, 2008).

The questions raised by soil patterns near the forest-prairie border have clear implications for the broader issue of soil organic carbon (OC) storage and how it is affected by natural or anthropogenic environmental change. Where forest replaced prairie in the Midwest, rapid transformation of Mollisols with thick, OC-rich A horizons into Alfisols could have quickly produced an overall reduction in OC storage at the landscape scale. The rate and spatial pattern of this response might have been substantially affected by topographic setting and parent material, however.

This paper reports results from a study of soils on steep bedrock-controlled hillslopes in southeastern Minnesota, U.S.A. (Fig. 1b), which include Mollisols with thick, OC-rich A horizons under mature deciduous forest. These A horizons may be relict features, formed under

prairie that has now been replaced by forest. This interpretation is supported by the presence of similar A horizons under small patches of prairie on similar parent materials nearby, and by reconstructions of regional vegetation history. Across the Midwestern U.S.A., prairie spread eastward in the early or middle Holocene, followed by westward readvance of forest in the late Holocene (Baker et al., 1992; Webb et al., 1983). Plant macrofossil, pollen, and OC isotopic data from stream sediments near the study area indicate expansion of prairie between about 6500 and 3500 cal yr B.P., with a subsequent shift back to greater forest cover (Baker et al., 2002; Chumbley et al., 1990). It is unlikely that either forest or prairie ever completely covered the study area, however, because its relatively rugged topography created local settings with widely varying microclimates and susceptibility to wildfire. In parts of Minnesota just north of our study area, there is evidence that some of the expansion of deciduous forest took place within the past millennium or even the last several hundred years (Grimm, 1983; Shuman et al., 2009; Umbanhowar, 2004). By the nineteenth century, the landscape of the study area was a mosaic of forest, prairie, and savanna (Marschner, 1974).

While these reconstructions of past vegetation change clearly suggest that the soils with thick, dark A horizons under forest are relicts of former prairie, this interpretation is complicated by other observations. These Mollisols are not distributed across the entire catena on the hillslopes in our study area, but instead occur on specific slope segments. Alfisols or Inceptisols with much thinner A horizons occur both uphill and downhill from the Mollisols, on the same slopes, and on ridgetops. Many of the Alfisols have distinct E horizons and strong textural contrast between E and underlying Bt horizons. It seems unlikely that prairie would occupy only certain segments of these slopes. In addition, the thickest and darkest A horizons under forest are generally close to dolomite outcrops, with abundant dolomite fragments in part or all of the

profile. This observation raises the possibility that a carbonate-rich parent material can entirely explain the genesis of those distinctive A horizons, as described above for *rendzinas* and similar soils in other regions, and they did not develop under grassland vegetation at all. In addition, the soils are formed in hillslope sediments made up of various mixtures of Late Pleistocene loess and bedrock-derived materials. The relative proportion of loess could potentially have affected OC accumulation and turnover, given the distinctive silt-rich texture, high moisture retention capacity, and smectitic clay mineralogy of the loess (Jacobs et al., 2011; Jacobs et al., 2012).

We studied whether distinctive parent material or topographic setting could have favored the occurrence of thick, dark A horizons under forest, and then used stable C isotope analysis of soil organic matter to detect any past influence of prairie vegetation. Warm-season grasses that are abundant in the tallgrass prairie of southeastern Minnesota (Minnesota DNR, 2005), including remnant patches within the study area, all use the C₄ photosynthetic pathway (Sage et al., 2011; Teeri and Stowe, 1976). In contrast, trees and forest understory species in this setting exclusively use the C₃ pathway. The two pathways involve different degrees of discrimination against the heavy isotope ¹³C (O'Leary, 1988) and as a result, organic matter added to soils from C₄ and C₃ plants has quite different isotopic composition (Boutton, 1996; Cerling et al., 1989). Thus, where prairie was replaced by forest, its past presence can potentially be detected in soil OC (e.g. Wang et al., 1993), especially in deeper horizons where OC turnover time is hundreds to thousands of years (Wang et al., 1996). We also obtained data on stable N isotope ratios, which can help interpret depth trends of C isotopes as effects of vegetation change or other processes.

2. Materials and Methods

2.1 Study area and prior research on hillslope stratigraphy

The study area is a bedrock-controlled, highly dissected landscape in Houston County, southeastern Minnesota, U.S.A. (Fig. 1b). This region was glaciated in the Pliocene or Early to Middle Pleistocene (Hobbs and Goebel, 1982), but glacial sediments have largely been removed by subsequent erosion and the landscape has been dissected by streams, producing deep valleys with local relief of 100 to 150 m km⁻². Valley-margin slopes often have segments steeper than 35°. The climate is humid and continental, with a mean annual temperature of 6.7° C and mean annual precipitation of 889 mm yr⁻¹ (1971-2000 means for Caledonia, Minnesota).

We focused on the genesis of Mollisols and other soils on steep valley-margin slopes. The soils on gently sloping ridgetops and valley floors of the study area have distinctly different parent materials and present-day land use (largely agricultural fields) and are beyond the scope of this paper. Throughout the study area, steep slopes similar to the ones we sampled are covered almost entirely by natural vegetation, predominantly deciduous forest but also including patches of prairie on south-facing slopes.

Soils on these steep slopes formed in a mixture of Late Pleistocene loess and hillslope sediments derived from local Paleozoic bedrock (Fig. 2). The Oneota Dolomite (Ordovician) forms a resistant caprock that weathers into angular boulders, which are distributed downslope, often mantling the uppermost part of the underlying Jordan Sandstone (Cambrian). Figure 2 also illustrates typical surficial material stratigraphy in the study area, based on extensive prior research (Lively et al., 1987; Mason, 1995; Mason and Knox, 1997; Mason et al., 1994). Late Pleistocene Peoria Loess was deposited across the study area, and is preserved in thicknesses up

to 4 m on wide ridgetops (Mason et al., 1994). Unweathered Peoria Loess is predominantly fine to medium silt with 10-20% clay and contains some detrital dolomite and calcite (Mason, 1992; Mason and Nater, 1994). Peoria Loess thins greatly toward steep valley-side slopes we sampled. At the upper end of those slopes surface soils formed in thin loess and the underlying reddish brown, clay-rich material, probably a mixture of bedrock residuum and weathered older loess (Frolking et al., 1983; Jacobs et al., 1997).

Farther downslope, the only loess preserved is mixed with bedrock-derived material in a predictable sequence of hillslope deposits (Fig. 2), summarized here using three *lithofacies* defined by Mason (1995). Steep slopes above and just below the dolomite-sandstone contact are mantled by a clast-supported diamicton (*Dc*) with a silt loam to sandy loam matrix, from 0.5 to about 2 m thick. Almost all clasts are angular fragments of Oneota Dolomite derived from a short distance upslope. Downslope from the area mantled by the *Dc* lithofacies, there is a somewhat gentler slope segment with scattered sandstone outcrops and a relatively thin mantle of massive sand (*Sm*) containing only scattered dolomite fragments. As the gradient decreases across the upper boundary of a prominent concave footslope, the sandy sediment of *Sm* overlies massive silt-rich deposits (*Fm*), which resemble Peoria Loess but contain substantially more sand and occasional rock fragments.

All three lithofacies accumulated primarily in the Late Pleistocene when a periglacial climate limited vegetation cover and produced widespread permafrost (Mason and Knox, 1997). At that time, Peoria Loess was rapidly eroded from slopes and narrow ridgetops and reworked onto footslopes to form the *Fm* lithofacies. The *Dc* lithofacies includes abundant debris produced by accelerated frost-weathering of dolomite bedrock, mixed with loess and bedrock-derived sand and moved downslope by frost creep. The *Sm* lithofacies is weathered sandstone moved

downslope by slopewash and bioturbation. Some downslope movement of soil by bioturbation, frost creep, and rainsplash has undoubtedly continued through the Holocene, but at low rates.

The general distribution of Mollisols and other soils on steep bedrock slopes of our study area is evident in the Houston County soil survey (Lueth, 1984). Mollisols with thick, dark A horizons and weakly developed Bw horizons are mapped on steep upper slopes (Fig. 3). These include Brodale soils (Entic Hapludolls), often under small patches of prairie on south- or southwest-facing slopes, and more extensive Lacrescent soils (Typic Hapludolls) under deciduous forest on all slope aspects. These soils formed mainly in the Dc lithofacies, based on typical profile descriptions. Soils mapped up- and downslope from and these Mollisols are mostly Alfisols (Typic and Glossic Hapludalfs), with patches of Entisols (Udipsamments or Quartzipsamments) in the sandy Sm lithofacies.

2.2 Site Selection and Field Sampling

This paper is based on 29 soil profiles, 26 of which were studied at four hillslope sites, on slopes with northeast (CAR), northwest (RN, RH), and southwest (SF) aspects (Fig. 1b, Fig. 4). Three other soils were sampled on ridgetops mantled with thick loess (Fig. 1b), to compare pedogenesis in that setting with the hillslope soils that were the primary focus. Field descriptions and all laboratory data are provided in supplemental online tables S1-S3.

We selected the hillslope sites because a) reconnaissance suggested that they represent downslope sequences of soils similar to those that were mapped over many of the slopes in the study area by Lueth (1984), b) three sites include forest-covered Mollisols and the fourth represents Mollisols under a prairie patch on a southwest-facing slope for comparison, and c) at

each site there were fresh exposures of soils along hundreds of meters of trails cut into the hillsides for logging or all-terrain vehicle use. These exposures often reached weathered bedrock, and while they were not aligned along the direction of steepest descent, they provided an outstanding opportunity to examine soil variation from the shoulder slope down to the upper footslope (from A to A' in Fig. 2). Based on reconnaissance of the exposures, we selected profiles for detailed study (nine profiles at CAR, six each at SF and RH, and five at RN, see Fig. 4). At all sites, sampled profiles were distributed across the elevation range of soils with relatively thick, dark A horizons at each site, and at the sites where exposures allowed this (CAR, RN, RH), we also sampled soils with thinner A horizons lower and/or higher on the slope. Within the band of thicker A horizons, profiles were selected to represent the overall range of variability of A horizon thickness and B horizon morphology that we observed in the continuous exposures. At the CAR site we also supplemented exposures with a profile sampled to bedrock at 1.2 m depth on the lower slope using a 2.5-cm diameter core (CAR-9), and a hand-dug pit on the opposite side of the ridge from the other profiles (CAR-12).

None of the sampled soils at these four hillslope sites have been cultivated. The SF site is located on a small patch of prairie, similar to many others on steep southwest- or south-facing slopes in the surrounding region (Curtis, 1959; Minnesota DNR, 2005). Vegetation at the SF site is dominated by warm-season C₄ grasses, including little bluestem (*Schizachyrium scoparium*), big bluestem (*Andropogon gerardii*), Indiangrass (*Sorghastrum nutans*), and sideoats grama (*Bouteloua curtipendula*). Scattered small trees, mostly eastern red cedar (*Juniperus virginiana*) are also present. At the CAR site, CAR-12 is in a similar but much smaller patch of prairie grasses, while the rest of that site and all of the RN and RH sites are covered with deciduous forest, including trees with diameters > 40 cm. Larger trees are predominantly red oak (*Quercus*

rubra) and white oak (*Q. alba*), with smaller numbers of bur oak (*Q. macrocarpa*). Smaller trees and saplings include the same three oak species, American elm (*Ulmus americana*), shagbark hickory (*Carya ovata*), sugar maple (*Acer saccharum*), basswood (*Tilia americana*), and black cherry (*Prunus serotina*). Groundcover plants at all sites are those common in oak forests of southern Minnesota (Minnesota DNR, 2005), with few invasive species, suggesting that these sites do not have unusual soil conditions or disturbance history.

The three ridgetop soils formed in loess ≥ 1 m thick, and were described and sampled from 2.5-cm diameter cores at two sites (CAR-South and VR) and from a pit at one (AL). All three soils are in remnant patches of mature deciduous forest. The CAR-S profile is on the loess-mantled ridgetop closest to the CAR hillslope site.

Freshly exposed faces were created by digging back from the existing trail cuts, and soil morphology was then described in the field using standard methods (Schoeneberger et al., 2002). Content of rock fragments > 2 mm was estimated visually during profile description, based on previous quantitative measurements of Dc lithofacies rock fragment content (Mason, 1995). Samples for most laboratory analyses were collected from the entire thickness of horizons < 25 cm thick; thicker horizons were subdivided for sampling. Small samples for C isotope analysis or radiocarbon dating were collected from selected profiles at the CAR site, from freshly exposed faces.

2.3. Laboratory Methods

Bulk samples were air-dried, ground, and sieved at 2 mm for all laboratory procedures. Laboratory analyses were carried out on subsets of 28 soils (OC and pH), 24 soils (particle size

analysis), 18 soils (clay mineralogy) and 14 soils (C and N isotopic analysis); the subsets used are listed by site in Table 1 and were intended to represent the range of observed morphological variation. We measured pH in 1:1 soil:water suspensions. Loss on ignition (LOI) at 360°C for 2 hr was used to estimate OC content (Konen et al., 2002, MLRA 108 equation), in part because this method does not require pretreatment for carbonate removal. We measured clay (< 2 µm), silt (2-63 µm), and sand (> 63 µm) by the pipette-sieve method (Gee and Bauder, 1986). Dark-colored A horizon samples were pretreated with 30% H₂O₂ at 70-90°C. Carbonate minerals were not removed because detrital dolomite grains form a substantial fraction of the sand and silt in some samples and secondary carbonate cementation is rare and quite weak.

Samples used for analysis of clay mineralogy were dispersed with 0.5 mg L⁻¹ Na-metaphosphate, and clay (< 2 µm) was separated by settling. Clay mineralogy was determined by x-ray diffraction of clay smears on glass slides using Cu-kα radiation, with a scan speed of 2° per minute and a step of 0.02°, in a Rigaku Miniflex diffractometer. Clay samples were saturated with Mg²⁺ followed by ethylene glycol solvation (Moore and Reynolds, 1989). Height above background for the 001 peak of each major clay mineral was determined using JADE software (Materials Data Inc.), and clay mineral percentages were estimated from relative peak heights following a method of the Illinois State Geological Survey (Method I of Hughes and Warren, 1989). All clay associated with a 1.4 nm peak was counted as vermiculite, though it may also include hydroxy-interlayered smectite and/or hydroxy-interlayered vermiculite.

Soil used for OC and N isotope analysis was subsampled in the lab, from either larger samples collected specifically for this purpose or the dried and crushed large bulk samples used for other lab analyses. In either case, roots or other macroscopic organics were removed from subsamples under a dissecting microscope, and the remaining soil was weighed into silver

capsules. Only five of the 55 samples used for OC isotope analysis contained carbonates detectable with 10% HCl in the fine earth fraction (all from the SF site), but many others came from horizons containing large dolomite coarse fragments, though with a noneffervescent matrix. Therefore, all samples were treated to remove carbonates by exposing them to HCl fumes in a desiccator for 8 hr (Harris et al., 2001). After carbonate removal, the samples were submitted to the UC-Davis Stable Isotope Facility (SIF), for analysis of C and N isotopes. Analysis at that laboratory used a PDZ Europa ANCA-GSL elemental analyzer interfaced with a PDZ Europa 20-20 isotope ratio mass spectrometer, correcting the measured values using standards interspersed within each sample batch. Results are reported as $\delta^{13}\text{C}$ (the standardized ratio of ^{13}C to ^{12}C , using the V-PDB standard), $\delta^{15}\text{N}$ (the standardized ratio of ^{15}N to ^{14}N , using the Air standard), and mass of C and N in each sample, from which we calculated C:N ratios. The UC-Davis SIF reports that long-term standard deviations of $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ measurements are 0.2 ‰ and 0.3 ‰, respectively.

Because of concerns about complete carbonate removal using 8 hr acid fumigation, and variable effects of acid treatments in general (Brodie et al., 2011; Ramnarine et al., 2011), we removed carbonates from 33 replicate subsamples by immersing them in 2M H_2SO_4 for 48 hr, then rinsing repeatedly to $\text{pH} > 5$ and drying, before isotopic analysis by the same methods at UC-Davis SIF. For the 33 pairs of replicate samples treated by both carbonate removal methods, $\delta^{13}\text{C}$ values were highly correlated ($R^2 = 0.98$, $\delta^{13}\text{C}$ (fumigated) = $0.98 \delta^{13}\text{C}$ (acid immersed) – 0.75). The small negative intercept is in the opposite direction of that expected for incomplete carbonate dissolution with fumigation, and may reflect preferential loss of ^{13}C in OC that dissolves during acid immersion. Overall, though, the small differences between results for most samples indicate that observed patterns of $\delta^{13}\text{C}$ variation within and between profiles are not

substantially affected by the method of carbonate removal. The C:N ratios calculated from UC-Davis SIF measurements are all in the range expected for soil organic matter, with no anomalously high values suggesting residual carbonates.

Samples of bulk soil, collected for radiocarbon dating from three depths in one profile (CAR-7), were subsampled under the microscope to avoid roots or other macroscopic organics, then submitted to NOSAMS (Woods Hole Oceanographic Institute) for accelerator mass spectrometer dating (carbonates were removed at NOSAMS using their standard methods).

2.4. GIS methods

Relationships between topography and soil morphology at the study sites and in the surrounding region were characterized using a digital elevation model with 3-m cell size, produced by the Minnesota Department of Natural Resources from LiDAR data. Slope gradient, profile curvature (land surface curvature along the line of steepest descent), and plan curvature (equivalent to curvature of contours) were calculated at the original 3-m resolution using standard tools in ArcGIS v. 10.1. We then smoothed these data, using the average values of cells within a 4-cell (12 m) radius, to remove small topographic irregularities present in the high resolution of the LiDAR data, in particular the trails that we sampled along (the smoothed grid still had a 3 m cell size).

Point locations of the sampled soil profiles were then assigned terrain attributes (slope and plan and profile curvature) from the smoothed surfaces. Because the relatively small number of sampling sites could make it difficult to detect significant contrasts in topographic setting between the forest-covered Mollisols and other soils, we also applied the same methods to large

numbers of randomly located points in polygons of Mollisols and adjacent Alfisols on colluvial slopes in the Houston County soil survey (Lueth, 1984, available as digital SSURGO data). One set of 31,700 points was sampled within polygons of Lacrescent and Brodale soils, Mollisols on steep upper slopes similar to those we targeted for sampling (Fig. 3). Another 31,701 points were sampled in polygons of Alfisols (Council, Elbaville, and Seaton valley phase) typically mapped immediately downslope from Lacrescent and Brodale soils, on footslopes mantled by the Sm or Fm lithofacies (Fig. 3).

To study vegetation change within the last 160 yr, we used a GIS dataset of vegetation at intersections of the grid laid out during the original 1854 public land survey of the study area, interpreted from surveyors' field notes (Almendinger, 1997). Using digital soil survey data, we selected 232 section and quarter section corners (the survey grid intersections) that fell within polygons of Brodale and Lacrescent soils, excluding a few that were in locations where error in mapping is highly likely (e.g. where polygons of those soils were drawn as if they extended onto valley floors). The vegetation types recorded on these Mollisols in 1854 were then tabulated separately for corners on slopes with south (90° to 270° azimuths) and north (0° to 90° and 270° to 0° azimuths) aspects. Virtually all slopes with Lacrescent soils on north aspects are forest-covered today, so we were especially interested in whether the land survey notes record prairie or savanna on those slope aspects in 1854.

2.5. Statistical analysis at the soil horizons level

As in most soil genesis research, we initially focused on relationships between whole-profile morphology, parent materials, and topographic settings. However, particularly important

relationships for our research question, such as those between OC accumulation and Ca^{++} abundance, texture, or mineralogy should be evident at the level of individual horizons. Analyzing those relationships at the horizon level is especially important because most of the soils we sampled have heterogeneous parent materials, resulting in substantial variation among horizons in texture, mineralogy, and rock fragment content. Thus, we used principal component analysis (PCA) and linear models to examine relationships between parent material-related properties and OC content in individual horizons, and then considered which of those horizon-level relationships can help explain variations in whole-profile morphology across the hillslopes we studied.

The content of OC decreases greatly with depth in all the sampled profiles; however, the key issue of interest is the degree to which the OC content of a horizon at a particular depths is anomalously high or low, as compared to other horizons at that depth in the whole set of soils sampled at all sites. To determine that, *standard %OC profile* representing the average of all individual soil OC profiles was calculated by 1-cm depth increments, assigning each increment the average %OC for all individual horizons that include that depth increment. We then calculated the *differential OC content* of each horizon:

$$\text{Diff OC} = \text{OC}_{i, z1-z2} - \text{OC}_{\text{ave}, z1-z2}$$

where $\text{OC}_{i, z1-z2}$ is measured %OC between depths $z1$ and $z2$ (upper and lower horizon boundaries) for a particular sample, and $\text{OC}_{\text{av}, z1-z2}$ is the average %OC for that depth interval in the standard %OC profile. In essence, Diff OC of an individual horizon is the deviation of its OC content from the standard profile.

Principal components analysis (PCA) was used to explore the major axes of variation in soil horizon properties, in terms of Diff OC as well as potentially parent material-related properties

of individual horizons. This analysis used 111 individual A to C horizon samples from a subset of 18 soil profiles (Table 1) for which clay mineralogy data were available, and the input variables were pH, Diff OC, particle size fractions (sand, silt, clay), and clay mineral fractions (smectite, illite, vermiculite, kaolinite+chlorite). The raw percentages of particle size and clay mineral fractions were transformed using centered log ratios to avoid the inflated correlations associated with compositional data that sum to 100% (Kucera and Malmgren, 1998). The PCA was implemented using the PRINCOMP module of R (R DevelopmentTeam, 2008); the correlation matrix was used, equivalent to standardizing the original variables, which differ in units and variance. Based on the results of PCA, we used R to test linear models in which potentially parent material-related properties were used as predictors of Diff OC for the same 111 individual horizons, using the Akaike Information Criterion (AIC) (Akaike, 1973) to evaluate the information gained by adding predictor variables.

3. Results

3.1. Soil morphology in relation to hillslope stratigraphy

The 29 soils that we studied varied most strikingly in A horizon thickness and color. Using those characteristics, the soils were placed in two groups, one of which was subdivided based on dolomite rock fragment content (Fig. 5). Group 1 includes 17 Mollisols at the four hillslope sites (e.g. RN-4 and CAR-7, Fig. 5), all of which have A and AB horizons with moist Munsell value and chroma of 3/3 or darker totaling >25 cm in thickness (the basic requirements for mollic epipedons). Of these, 10 are under forest (at RN, RH, and CAR sites), and seven under prairie

(all soils at SF site and CAR-12) (Fig. 4, Table 1). Most Group 1 soils (12 of 17) have Bw horizons with well-expressed pedogenic structure but little or no field evidence of clay illuviation. Three others, all under forest, have Bt horizons identified in the field by clay coatings on peds or in pores and possible small increases in clay from A to B horizons. One Group 1 soil (SF-4) has no B horizon and the B horizon was not exposed enough to characterize in CAR-12.

Most of the Group 1 soils (15 of 17) clearly formed in the Dc lithofacies (Fig. 2) and have visually estimated B horizon rock fragment contents >20% (virtually all dolomite), although rock fragment content decreases to < 5% in upper A horizons of some of these soils. The exceptions are a soil formed in 27 cm of stone-free silt over dolomite bedrock (SF-4), and one formed mainly in weathered sandstone but with some dolomite fragments (SF-6). Despite the high content of dolomite fragments throughout much of the profile in Group 1 soils, the fine earth is not effervescent with 10% HCl above a depth of 50-70 cm in Group 1 profiles from CAR, RH, and RN sites. Strong to violent effervescence of the fine earth was observed below that depth. At the south-facing SF site, all but the upper A horizon generally effervesced with 10% HCl, and most dolomite fragments in lower A and B horizons have well-developed pedogenic carbonate pendants on their undersides. No redoximorphic features or low-chroma colors indicating past or present soil wetness were observed in any Group 1 soil.

Group 2 is composed of nine soils at the hillslope sites (e.g. CAR-1 and RH-5, Fig. 5) and the three thick loess soils. In this group, A or AB horizons with value and chroma of 3/3 or darker have total thickness < 25 cm (often much < 25 cm). All but two of the Group 2 soils are Alfisols with Bt horizons identified in the field by texture and clay coatings and argillic horizons confirmed by laboratory particle size data; most also have distinct E horizons. The exceptions (CAR-2 and RN-2) are Inceptisols with Bw horizons. Group 2 soils can be divided into two

subgroups, based on parent material. Group 2A soils include the three formed in ridgetop loess and four formed in a thin mantle of the sandy Sm lithofacies that extends downhill over the silt-rich Fm lithofacies on lower footslopes (Fig. 2). Upper horizons of the latter four soils have high sand content, but Bt horizons are more silt-rich. Group 2A soil profiles are non-effervescent in all horizons. In contrast, the five soils in Group 2B (CAR-3, RN-2, RN-4, RN-5, and RH-5) contain relatively abundant dolomite fragments in lower horizons, where the fine earth matrix is also effervescent. The parent material of three of these soils is clearly the Dc lithofacies. The Bt horizon clay content and color in the other two (RN-4 and -5) suggest they formed in a mixture of the Dc lithofacies the reddish brown clays that underlie loess on ridgetops, possibly resulting from cryoturbation. One Group 2B soil, RH-5, forms an especially interesting contrast with the Group 1 soil RH-4, only 5 m away along the same roadcut face (Fig. 5).

3.2. Topographic setting

Field observations and topographic maps (e.g. Fig. 3, 4) suggest that the band of Mollisols on colluvial hillslopes (including our Group 1 soils) closely corresponds to steep upper slopes with convex or planar profiles, while the soils with thinner A horizons just downhill are on gentler concave slopes. Our quantitative analysis supports that view as a broad generalization, but indicates substantial variation in local topographic settings of these soils (Fig. 6). The 31,701-point sample from soil survey polygons of Mollisols similar to our Group 1 soils displays a wide range of curvatures, with a slightly negative median value of -0.03, indicating a slightly convex slope profile (Fig. 6a) This sample includes slopes ranging from 2° to 42° (median of 26°), possibly exaggerating the actual range for these soils because it is clear that some of the

originally hand-drawn soil survey polygons were erroneously extended onto ridgetops or footslopes. The soil survey-based sample of Alfisols typically mapped just downslope of those Mollisols displays a distinctly more concave (positive) median profile curvature of 0.22 (Fig. 6a), with slopes ranging from $< 1^{\circ}$ to 26° (median of 9°). The profile curvature distributions for these two large samples are significantly different ($p < 0.001$), based on a two-sample Kolmogorov-Smirnov test (*k.test* in R). The distributions of profile curvature for the 17 Group 1 soils (Fig. 6b) and the large sample of similar Mollisols are not significantly different ($p = 0.19$), but the Group 1 distribution is significantly different from the one for the large sample of Alfisols on lower slopes ($p = 0.001$). The median profile curvature for Group 1, -0.37, is more convex than for the large Mollisol sample, however, and the Group 1 soils are on relatively steep slopes (range of 18° to 33° , median of 31°).

The nine Group 2 soils that we sampled on hillslopes have range of convex to concave profile curvatures (Fig. 6b), with a slightly convex median of -0.09, and their distribution is not significantly different from those of Group 1, the large Mollisol sample, or the large Alfisol sample (not surprising given the small sample size). Our Group 2 soils are on relatively steep slopes, compared to the large Alfisol sample, however (range of 17° to 31° , median of 26°). Median plan curvature is positive, indicating topographic settings where surface water flow lines would diverge, for Group 1 (0.10), Group 2 (0.16), and both large samples (0.11 for Mollisols, 0.02 for Alfisols); none of the distributions of plan curvature are significantly different from any others ($p > 0.05$ in all cases).

The most important result of this topographic analysis is that despite considerable local variability, our Group 1 soils and similar Mollisols on colluvial slopes in the study area are *not* typically on concave slopes of relatively low gradient, in areas of convergent flow (concave plan

curvature). That is, they are not in parts of the landscapes found to have relatively high accumulation of soil OC in modeling and observational studies where topographic effects were significant (Aandahl, 1948; Berhe et al., 2008, 2012; Gessler et al., 2000; Schimel et al., 1985; Yoo et al., 2006). Only the large sample of Alfisols on footslopes tends to be in those topographic settings, yet those soils display much thinner A horizons and probably contain much less OC than the Mollisols higher on the slopes. Our sample of Group 2 soils, though mainly Alfisols, is from steeper and less concave slopes than the larger Alfisol sample, and in some cases intermixed with Group 1 soils in very similar topographic settings (e.g. RN and RH sites, Fig. 4).

3.3. OC content, texture, mineralogy, and pH

As expected from contrasts in A horizon thickness and color, soils in Group 1 tend to have higher OC content at a given depth than those in Group 2 above a depth of 70 cm, as shown by plots of average OC by depth for each group and across all the soils (Fig. 7). Individual samples display considerable scatter and overlap between the groups, however. Group 2A and Group 2B have very similar profiles when averaged separately (not shown). The Diff OC variable is essentially the distance to the right or left of the overall average profile shown in Figure 7, so soils in Group 1 generally have higher Diff OC values than those in Group 2.

The texture, pH, and clay mineralogy of the soils are all strongly related to parent materials, though overprinted to varying degrees by pedogenesis, particularly for the Group 2 soils (Fig. 8). The three Group 1 soils and the Group 2B soil (RH-5) shown in Figure 8 all have fine earth textures within the broad range observed in the matrix of the Dc lithofacies that they formed in.

This matrix is a mixture of sand derived mainly from local bedrock and silt and clay from Peoria Loess (Mason, 1992), with the proportions often varying considerably across hillslopes and with depth in individual profiles. Comparison with the lower horizons of the AL profile, formed entirely in Peoria Loess, indicates that the clay mineralogy of Group 1 and Group 2B soils is largely inherited from the loess added to the Dc matrix, though minor amounts of clay may come from weathered bedrock. In Group 2A soils formed in the Sm and Fm lithofacies (e.g. CAR-1, Fig. 5, 8) much of the clay is probably loess-derived, but in some cases relatively high kaolinite content may reflect some clay input from the sandstone bedrock. The upper part of the AL and CAR-1 profiles (Group 2A) displays distinct depletion of smectite, probably through a combination of eluviation and conversion to hydroxy-interlayered forms which are counted as vermiculite; otherwise, there is little evidence of pedogenic alteration of the clay mineralogy inherited from parent materials.

Presence or absence of carbonate minerals is clearly the predominant influence on pH in these soils (Fig. 8). As noted above, Group 1 soils typically contain coarse fragments of dolomite throughout the profile, and they are present in lower horizons of Group 2B soils. Across all sampled soils, horizons that effervesce in 10% HCl and contain dolomite fragments have pH values between 7.0 and 8.2, consistent with pH controlled by equilibrium with carbonates. Non-effervescent horizons containing coarse dolomite fragments have pH values of 6.5 to 7.5. Horizons from Group 2 soils that lack dolomite fragments and are non-effervescent all have pH < 7.0, and in many cases have pH between 4.8 and 6.0 (e.g. CAR-1 and AL profiles in Fig. 8).

Statistical analyses highlight the range of variation of texture and mineralogy *within* groups, that is, the contrast in A horizon morphology between Groups 1 and 2 is only one dimension of the variability of these soils. Nonetheless, these analyses clearly identify pH as a variable

associated with high Diff OC, and therefore also point to parent material dolomite content as a key explanation for the morphological contrasts between Groups 1 and 2. The 111 samples used in these analyses come from nine Group 1 soil profiles (53 horizons), five Group 2A profiles (33 horizons), and four Group 2B profiles (35 horizons) (Table 1), using A to BC or C horizons from each profile.

The results of PCA can be interpreted using biplots of individual horizon sample scores on the first three principal components (PC1, PC2, and PC3), which together account for 86.5% of the variance in the input data (Fig. 9, Table 2). Of these, PC1 and PC2 (65.8% of the variance) are most heavily loaded by particle size and clay mineral fractions, with smaller but substantial loading by Diff OC as well (Fig. 9a). A biplot of horizon sample scores on PC1 and PC2 (Fig. 9b) reveals the large textural and mineralogical variation *within* morphological groups, more than distinctions between them. Group 1 samples mostly cluster near the center of the plot, reflecting intermediate textures and a mixture of clay minerals. Several Group 1 samples plot farther to the right ($PC1 > 2$), however. These are mostly samples from deeper BC and C horizons with relatively low Diff OC and higher sand and kaolinite content than other Group 1 samples, reflecting greater contribution of sandstone bedrock. Group 2A horizons occur in three distinct locations on the PC1-PC2 biplot (Fig. 9b). Several, from a loess-derived ridgetop soil, plot far to the upper left ($PC1-$, $PC2+$), to a large extent because of high silt and clay content, and some cases also a relatively high proportion of illite. Another cluster of Group 2A horizons falls in the lower right quadrant ($PC1+$, $PC2-$); these formed in the sandy Sm lithofacies and have relatively high sand, kaolinite, and vermiculite contents. Still other Group 2A samples overlap those from Group 1 near the center of the plot. Group 2B horizons are widely distributed on the PC1-PC2 plot, overlapping areas occupied by samples from both Group 1 and Group 2A.

The morphological groups are much more clearly separated on a biplot of PC2 and PC3 (Fig. 9d), though mainly along PC3. While PC3 accounts for only 20.7% of total variance, it is the first component strongly loaded by pH and it is also loaded in the same negative direction by Diff OC (Fig. 9c). Almost all Group 1 horizons fall in the lower half of the PC2-PC3 biplot (Fig. 9d), with most in the lower right quadrant (PC2+, PC3-), mainly reflecting their high pH and Diff OC; deeper Group 1 horizons (BC and C) with higher illite tend to be toward the right (high PC2) side of the Group 1 cluster. Group 2A horizons are almost all in the upper half of the PC2-PC3 biplot, largely because of lower pH and/or Diff OC, along with high vermiculite or kaolinite proportions in some samples. Acidic Group 2A A and E horizons plot especially high on PC3. Group 2B horizon samples plot in areas occupied by both Group 1 and Group 2A samples. Dolomite-rich B and C horizons from Group 2B plot close to many Group 1 horizons.

To summarize, PCA reveals a great deal of interesting textural and mineralogical variation within morphological groups, much of which may ultimately be related to heterogeneous parent materials. Despite their striking morphological contrast, Groups 1 and 2A clearly separate only along one of the major axes of variation identified by PCA. That axis, PC3, is most strongly loaded by pH and Diff OC, clearly suggesting that pH (or dolomite content) was an important factor in the OC accumulation that formed the thick A horizons of Group 1. Diff OC is loaded almost as strongly on PC1, however, in the same direction as clay, silt, and smectite, suggesting an association between texture and OC accumulation for individual horizons, if not for morphological groups.

With that background, we tested linear models for prediction of Diff OC in the same set of 111 horizon samples, from either pH or a textural variable alone, and from combinations of pH and a textural variable (Table 3). Either % sand or % clay alone is a significant, though quite

weak, predictor of Diff OC; pH alone explains almost none of the variance in Diff OC. Combining % sand, or especially % clay, with pH yields substantially better predictive equation (adjusted $R^2 = 0.56$ for % clay and pH), although it still only explains about half of the variance in Diff OC. Adding a clay mineral variable (proportion of smectite, vermiculite, illite, or kaolinite+chlorite) does not improve the model, based on AIC (two examples shown in Table 3). The residuals from the model using % clay and pH as predictors indicate substantial underestimation of six highest Diff OC values, and overestimation of the four lowest, but the other 101 residuals are normally distributed.

These results can be explained by considering the varying relations between texture, pH, and Diff OC with depth in the horizon samples used for this analysis. Diff OC (not just total OC) is greatest in the upper horizons of Group 1 soils, which also tend to contain more clay and less sand than deeper horizons; pH is relatively high throughout many of these soils. Thus, *within* Group 1, sand or especially clay content may be a better predictor of Diff OC than pH. On the other hand, pH differentiates Group 1 from Group 2A or 2B, for upper horizons with similar clay or sand content but contrasting Diff OC. Clay mineralogy does not appear to be significantly related to anomalously high or low OC accumulation in the soils we sampled.

3.4. Stable isotope analysis and radiocarbon dating

Results discussed here are based on OC and N isotopic data from 13 vertical profiles in Group 1 soils (Table 1). One of these was profile CAR-12 which was sampled only for isotopic analysis, and four were profiles that were essentially replicates of CAR-7, distributed at 5 m intervals over 20 m away from that profile along the same vertical face, and sampled only for

isotopic analysis. Five Group 2 soils were also sampled for isotopic analysis, three from Group 2A and two from 2B (Table 1).

All C isotopic results are reported as $\delta^{13}\text{C}$ VPDB, the ratio of ^{13}C to ^{12}C normalized by the ratio in the standard Vienna Pee Dee Belemnite, in per mil (‰) units; N isotopic data are reported as $\delta^{15}\text{N}$, the ratio of ^{15}N to ^{14}N normalized by the isotopic ratio of atmospheric N. Radiocarbon dating of OC from CAR-7 indicates that the mean residence time of OC increases substantially with depth, given a modern age at 5 cm and 895 ± 30 ^{14}C yr BP at 45 cm (Table 4). This trend of increasing age with depth is widely observed and expected for most soils because of low rates of OC addition and decomposition in deeper horizons (Scharpenseel, 1971; Wang et al., 1996). Assuming OC mean residence time increases with depth in this fashion in all of the soils we studied, deeper samples are dominated by OC added earlier in time, and depth trends of $\delta^{13}\text{C}$ can potentially be interpreted as the result of vegetation change.

The most straightforward results were from the three Group 2A soils, all currently under forest (Fig. 10a). Near the ground surface in these soils, $\delta^{13}\text{C}$ is -28‰ to -26‰, close to the mean value of -28‰ for OC from C_3 plants when freshly added to the soil (Boutton, 1996). All three Group 2A profiles display trends of increasing (less negative) $\delta^{13}\text{C}$ with depth, with a total increase of 2-4‰. A downward increase of this magnitude is common and is generally attributed to a combination of more advanced organic matter decomposition in deeper horizons, greater input of less ^{13}C -depleted root biomass relative to leaf litter with increasing depth, and the influence of fossil fuel combustion on isotopic composition of atmospheric CO_2 (the Suess effect). The isotopic shift during decomposition is likely the largest factor, although the specific mechanisms involved are still a matter of discussion (Bostrom et al., 2007; Ehleringer et al., 2000; Nadelhoffer and Fry, 1988; Wedin et al., 1995; Wynn et al., 2005; Wynn et al., 2006). No

past input from C₄ plants is necessary to explain this modest shift in $\delta^{13}\text{C}$ with depth. The value of $\delta^{15}\text{N}$ increases with depth and C:N decreases with depth in Group 2A profiles (Fig. 10b-c). Both trends are commonly reported; the C:N decrease is attributed to decomposition, while the increase in $\delta^{15}\text{N}$ results from decomposition effects, fractionation during transfer of N from mycorrhizal fungi to host plants, and N loss through nitrification and denitrification (Hobbie and Ouimette, 2009; Melillo et al., 1989; Nadelhoffer and Fry, 1988). In these profiles $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ are positively correlated ($R^2=0.80$), as expected if both are strongly influenced by greater decomposition with depth. The slope of this relationship is 1.81 (Fig. 11a).

Interpretation of isotopic profiles from the four Group 1 soils that are under prairie grass-dominated vegetation today (three from the SF site plus CAR-12) is also fairly straightforward (Fig. 10a-c). These soils have OC with $\delta^{13}\text{C}$ values between -16‰ and -13‰, indicating organic matter inputs largely dominated by C₄ plants (mean value of -13‰; Boutton, 1996). These soils do not consistently display depth trends of $\delta^{13}\text{C}$ or C:N that would indicate decomposition effects; however, $\delta^{15}\text{N}$ does increase consistently with depth in three of the four profiles and $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ are weakly positively correlated ($R^2=0.42$), with a slope of 1.12 (Fig. 11a). A stronger depth trend might have been apparent if we had sampled these soils closer to the ground surface (the uppermost samples we analyzed were composited from the thick uppermost A horizons). The weakness of depth trends may also reflect the large belowground input of fresh carbon from roots in the prairie ecosystem. Importantly, the $\delta^{15}\text{N}$ and C:N values for horizons of these soils under prairie overlap with those from similar depths in forest-covered Group 2A (Fig. 10b-c).

Isotopic profiles of OC are distinctly different in the profiles of Group 1 soils that are under forest today and from Group 2B soils, also under forest (Fig. 10d-f). All of these soils display similar steep trends of increasing $\delta^{13}\text{C}$ with depth (Fig. 10d), with a shift of 4 to 7‰ between

samples at 5-10 cm depth and those from > 30 cm. Near-surface $\delta^{13}\text{C}$ is in the range expected for largely C_3 -derived OC, consistent with the modern forest cover on these soils. Values of $\delta^{13}\text{C}$ at depth are as high as -19 to -17.5‰, which are difficult to explain without substantial input from C_4 vegetation at some time in the past. The shift with depth is much larger than that attributed to effects of decomposition in prior research (Bostrom et al., 2007; Ehleringer et al., 2000; Nadelhoffer and Fry, 1988; Wedin et al., 1995; Wynn et al., 2005). Furthermore, if we were to try to explain this large shift through decomposition alone, we would expect to find values of $\delta^{15}\text{N}$ and C:N at depth also recording an extraordinary degree of decomposition. Instead, the $\delta^{15}\text{N}$ and C:N profiles in these 11 soils are generally similar to those of all the others that were analyzed (compare Fig. 10e-f with Fig. 10b-c), suggesting that large differences in the degree of decomposition in deeper horizons cannot explain contrasts in C isotope profiles. Combining Group 1 soils under forest and Group 2B soils, $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ are positively correlated ($R^2=0.81$), but with a much lower slope (0.66) than for the other soils (Fig. 11a). That is, $\delta^{15}\text{N}$ displays a smaller increase with depth in these soils, for a given increase in $\delta^{13}\text{C}$. This is expected if $\delta^{15}\text{N}$ increases largely through effects of decomposition, while the increase in $\delta^{13}\text{C}$ reflects the shift due to decomposition superimposed on a larger shift caused by vegetation change.

In summary, Group 2A soils under forest and Group 1 soils under prairie yielded C isotope profiles that can be explained as the result of decomposition, combined with the Suess effect and greater input from roots at depth, but *without* vegetation change. In contrast, the Group 1 soils and Group 2B soils all have considerably larger increases of $\delta^{13}\text{C}$ with depth, which appear to require vegetation change for adequate explanation, especially when profiles of $\delta^{13}\text{C}$ are compared to those of $\delta^{15}\text{N}$ and C:N. The implied vegetation change is from a mixture of C_4

grasses and C₃ plants, to vegetation with near-complete C₃ dominance, most likely forest. All of these soils that preserve evidence of vegetation change contain coarse dolomite samples throughout the profile or in deeper horizons, although it should be emphasized that the dolomite does not contribute to the isotope shift because the samples were treated to remove carbonates.

The four replicate profiles sampled at 5-m intervals adjacent to CAR-7 (a Group 1 soil under forest) provide one additional insight, although we only collected two samples from each of these profiles (Fig. 11b). All four display a shift toward less negative $\delta^{13}\text{C}$, as does CAR-7, but there is considerable local variability in $\delta^{13}\text{C}$ at a given depth and in the slope of the depth trend. Thus, although the four replicate profiles are very similar in morphology to the CAR-7 profile studied in detail, they could differ significantly in the degree of decomposition at a given depth, or even in past vegetation history.

3.5. Vegetation change since the public land survey

Public land survey records from 1854 indicate considerable areas of vegetation more open than closed forest on steep colluvial slopes, on both north and south aspects but especially on the latter (Table 5). Specific vegetation structure cannot be accurately inferred from many of the terms used by land surveyors for open vegetation (e.g. “dry ridge” and “scattering oaks”), but probably ranged from small areas of treeless prairie to more extensive savanna and open woodland. This interpretation is supported by the typical absence of witness trees (nearby trees recorded at corners to allow the survey markers to be relocated) where vegetation was described as prairie or “dry ridge”, and the predominance of bur oak (*Quercus macrocarpa*) when witness trees are recorded in vegetation described as “scattering oaks”, “oak barrens,” etc. In the

nineteenth century, fire-tolerant bur oak was commonly observed in open stands with grass understories (Curtis, 1959).

What is particularly notable about the land survey records is that relatively open vegetation is recorded on 52% of north-facing colluvial slopes, virtually all covered by closed forest today. While no corners are close enough to our study sites to infer vegetation change there, these results do suggest that soils on forest-covered north-facing slopes at the CAR, RN, and RH sites could have had more open vegetation with abundant grasses as recently as 160 years ago. While there are large trees that may be older than 160 yr at all three sites, they are scattered enough that they could have once formed part of a savanna or open woodland.

4. Discussion

We found no evidence that topographic setting can explain the occurrence of thick, dark A horizons under forest in our study area. Quantitative analysis indicates that the soils of Group 1 are not generally in topographic settings found to favor OC accumulation in previous research (Aandahl, 1948; Berhe et al., 2008, 2012; Gessler et al., 2000; Schimel et al., 1985; Yoo et al., 2006). In fact, lower concave slopes that might favor OC accumulation in other landscapes are actually occupied by Alfisols and other soils with thin A horizons in the study area. Some of our Group 2 soils are in topographic settings very similar to Group 1 soils, also suggesting that topographic effects are not important in explaining the morphological contrast between groups. Low erosion rates under closed forest or dense grassland may help explain why topographic effects on soil OC are not more evident on the slopes we studied, as opposed to landscapes affected by long-term anthropogenic erosion (Doetterl et al., 2012).

The relative importance of parent material and vegetation history is a more complex issue. Without the stable C isotopic data and the land survey records, the rest of our results could lead us to interpret thick, dark OC-rich A horizons as primarily the result of parent materials rich in dolomite fragments. Such horizons occur under forest as well as prairie, but only in the dolomite-rich Dc lithofacies, consistent with parent material control. Specifically, without additional information, we could interpret thick OC-rich A horizons mainly as an effect of abundant Ca^{++} and Mg^{++} released by weathering dolomite, which may favor development of organic matter-mineral cation bridges and more generally favors especially stable aggregation, both of which act to stabilize OC (Oades, 1988, von Lützow et al., 2006). That in turn could lead to slower OC turnover and accumulation of a larger pool in the A horizon, regardless of vegetation structure. Higher clay and silt content also appear to be associated with greater OC accumulation in individual horizons at a given depth in these soils, possibly also by allowing more stable aggregation. Nonetheless, the presence or absence of dolomite fragments is the parent material effect most clearly distinguishing the groups of soils we defined by A horizon morphology.

The stable C isotopic data, however, provide clear evidence that Group 1 soils with thick, dark, OC-rich A horizons that are under forest cover today contain substantial C_4 -derived OC preserved in deeper horizons. That is, they must have formed partly under grassland or savanna with relatively abundant C_4 grasses. Interestingly, the two Group 2B soils we sampled for isotopic analysis, Alfisols with thinner A horizons formed in dolomite-rich parent material, also preserve isotopic evidence of past OC input from C_4 grasses. The land survey data also demonstrate that many forest-covered slopes where thick, dark A horizons occur had more open vegetation, most likely including a grass understory, as recently as the mid-19th century.

Replacement of open vegetation by forest could be the result of wildfire suppression after Euroamerican settlement of the study area, beginning in the mid-19th century; this is a common explanation for similar vegetation change in other parts of the Midwest (Curtis, 1959). On the other hand, there is good evidence for a somewhat longer-term Late Holocene trend of climatically driven forest expansion in nearby areas, as discussed above (Grimm, 1983; Shuman et al., 2009; Umbanhowar, 2004).

Taking all lines of evidence into account, the thick, dark-colored A horizons of Group 1 soils occur almost entirely in the dolomite-rich Dc lithofacies, under prairie at present or under forest that has probably replaced prairie or savanna with abundant C₄ grasses. Two hypotheses can explain this pattern. First, prairie or savanna could have occurred on north as well as south slope aspects in the past, but was limited to the steep upper slopes mantled by the Dc lithofacies. This would account for restriction of Group 1 soils to that setting today. Alternatively, prairie or savanna may have covered much more of the landscape in the past, but older C with the isotopic signature of C₄ plants is better preserved today in Group 1 soils formed in the Dc lithofacies than in other soils on the same hillslopes.

The first hypothesis cannot be ruled out without direct evidence for the past presence of prairie above or below the slope segments now occupied by Group 1 soils at our study sites. There is good reason to doubt this explanation, however. At the local scale where macroclimatic gradients are minimal, boundaries between forest and prairie in the humid Midwestern U.S. generally appear to have been related to spatial variation in fire frequency, which was influenced by topographic or hydrological features that limited the spread of fire and allowed establishment of trees (Grimm, 1984). At our study sites, however, there is no apparent reason why fires would consistently burn only a particular segment of the slope. Especially low soil moisture on certain

slope segments might conceivably have favored prairie grasses over trees, either directly or through its effect on fire frequency. Given soil textures, however, it is likely the lowest moisture values are actually reached on sandy upper footslopes (Sm lithofacies), not on the steep slopes above mantled by the Dc lithofacies with its siltier matrix (Fig. 2).

The second hypothesis is more likely to be valid. Preferential preservation of isotopic evidence for past C₄ vegetation in the Group 1 soils, can in fact be explained by the same effect of abundant dolomite-derived Ca⁺⁺ and Mg⁺⁺ on OC stability that was outlined above. In this case, however, the development of thick OC-rich A horizons in Group 1 soils is not attributed entirely to that effect of parent material lithology. Instead, we suggest that prairie or savanna was present across much of the hillslope at each of the three presently forest-covered study sites, and relatively thick A horizons may have developed in the full range of parent materials under the influence of that vegetation. As closed forest occupied these sites, slower cycling of OC in horizons formed from the dolomite-rich Dc lithofacies led to more prolonged preservation of both thick, dark A horizons and the isotopic signature of prairie grasses.

The second hypothesis suggests that there could be soils that are intermediate in A horizon morphology and preservation of C₄-derived OC, because they contain only a few dolomite fragments or they have sandy textures that limit OC accumulation. Soils of Group 2B, are in some cases intermediate in A horizon morphology, with A horizons less than 25 cm thick but still thicker than those in Group 2A. The two contrasting profiles only 5 m apart at the RH site (RH-4, RH-5) offer intriguing evidence in this regard (Fig. 5). Group 1 soil RH-4, with a thick, dark A horizon and a distinctly OC-rich profile, has abundant dolomite fragments throughout the profile and relatively high silt content. Group 2B soil RH-5, with a dark but thin A and a distinct EB horizon has relatively high sand content and few dolomite fragments in its upper horizons,

possibly because they have weathered rapidly in the sandy matrix. These two soils have similar $\delta^{13}\text{C}$ values in deeper horizons indicating past importance of C_4 grasses (Fig. 10d). Thus, even if RH-5 has lost more OC from upper horizons than RH-4, after the transition to closed forest, faster turnover in RH-5 has apparently not had as much effect on OC in dolomite-rich deeper horizons.

5. Conclusion

On hillslopes of the study area, soils with thick, dark A horizons under closed forest today preserve isotopic evidence that they formed under prairie or savanna in the past. Yet these soils are almost exclusively formed in colluvial parent material that contains coarse dolomite fragments, which is likely to have created soil chemical conditions favorable for OC accumulation and preservation. Other soils nearby are likely to have also sustained prairie or savanna in the past, but preserve no evidence of it today.

We propose that parent material controls the response of soils to vegetation change in this landscape. Where it is dolomite-rich, morphological response is slower and/or more limited, and long-term preservation of OC derived from past vegetation in deep horizons is more likely. The key evidence needed to more conclusively test this interpretation, and more generally the effect of abundant dolomite fragments on OC turnover, would be extensive radiocarbon dating of organic matter in soils formed in parent materials ranging from dolomite-rich to dolomite-free.

Even without such definitive evidence, the results of this study strongly suggest a need to consider *interactions* of vegetation history and parent material in interpreting soils near dynamic vegetation boundaries. More broadly, not only parent material but also topographic setting and

soil hydrology could influence the timescale of soil response to vegetation change in other landscapes, resulting in a soil landscape pattern that has only partially adjusted to the most recent changes in vegetation.

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Figure Captions

Figure 1. a. Vegetation of Minnesota and Wisconsin at the time of 19th century public land surveys (Marschner, 1974; Finley, 1976). White box is study area. b. Topography of the study area (based on US Geological Survey 30-m DEM) with study site locations. Sites in italics are individual ridgetop profiles in thick loess, at other sites, multiple soils were sampled on hillslopes.

Figure 2. Schematic view of typical hillslope stratigraphy in the study area, showing landscape position of Dc and Sm lithofacies and other depositional units (Mason, 1995; Mason and Knox, 1997). A and A' mark upper and lower limits of soil sampling at study sites.

Figure 3. Topography typical of the study area. Darkest shading indicates Mollisols with thick, dark-colored A horizons on steep slopes (Brodale and Lacrescent soils). Intermediate shading indicates Alfisols typically mapped on footslopes just below those Mollisols (Council, Elbaville, Seaton valleys soils). Alfisols also predominate on ridgetops above the Mollisols on steep slopes. Two study sites are indicated (CAR, SF). Topography from LiDAR-based 3-m DEM, Minnesota Department of Natural Resources. Soils from Natural Resource Conservation Service SSURGO data.

Figure 4. Topographic setting of soils sampled at the four hillslope study sites, identified by number, with symbols indicating which group they fall in. Topography from LiDAR-based 3-m DEM, Minnesota Department of Natural Resources; contour interval is 5 m.

Figure 5. Photos of soils representing groups and subgroups identified in this study (see Fig. 4 for locations and topographic settings). Soil horizons are labeled; subhorizons are not distinguished for clarity although they were described and sampled separately. Soils RH-4 and RH-5 are separated by about 5 m.

Figure 6. Terrain attributes for hillslope soil locations. a. Profile curvature at randomly located points in Mollisols on steep slopes (Brodale, Lacrescent soils, 31,701 points) and Alfisols on adjacent lower slopes (Council, Elbaville, Seaton valleys, 31,701 points). b. Profile curvature of Group 1 soils (17 profiles) and Group 2 soils on hillslopes (9 profiles). c. Plan curvature at randomly located points in Mollisols and Alfisols (same points as in a). d. Plan curvature of Group 1 and 2 soils.

Figure 7. Organic carbon content versus depth, for individual horizons of Group 1 and Group 2 soils. Lines indicate overall average across all sampled soils, and separate averages for Groups 1 and 2.

Figure 8. Texture, clay mineralogy, pH, and OC content with depth, for six soils representative of Group 1 and Groups 2A and 2B. Includes the four soils illustrated in Fig. 5 plus a ridgetop soil in thick loess (AL) and a Group 1 soil under prairie (SF-2). Dashed gray line to right of OC panel indicates coarse dolomite fragments occupying greater than about 5% of soil volume; solid gray line indicates effervescence of fine earth in 10% HCl.

1027 **Figure 9.** Results of principal components analysis. Panels at left show loadings of original
1028 variables on pairs of components. Panels at right plot individual soil horizons by their scores on
1029 pairs of components, with symbols differentiating samples by group and horizon type.

1030 **Figure 10.** Stable C and N isotopic data and C:N ratios for all samples analyzed, except the four
1031 replicate profiles of CAR-7 (see Fig. 11). C:N ratios are based on determinations of total C and
1032 N at UC-Davis SIF, on the same samples used for isotopic analysis.

1033 **Figure 11.** a. Relationship between $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ in the same horizons, with results of linear
1034 regression analysis, by group. b. C isotope composition with depth for CAR-7 and four replicate
1035 profiles (-R1, -R2, etc.).

Table 1. Soil profiles used for each type of analysis, by site and morphological group (see Results section for group definitions). Profiles at hillslope sites (CAR, RN, RH, SF) are identified by site-specific numbers; see Fig. 3 for locations of these profiles within each site.

Analysis	CAR (NE-facing, forest)	RN (NW-facing, forest)	RH (NW-facing, forest)	SF (SW-facing, prairie)	Ridgetop loess sites (forest)
Topographic analysis	Group 1: 4, 5, 6, 7, 12 Group 2A: 1, 2, 9 Group 2B: 3	Group 1: 3 Group 2A: 1 Group 2B: 2, 4, 5	Group 1: 1, 2, 3, 4, 6 Group 2B: 5	Group 1: 1, 2, 3, 4, 5, 6	
OC, pH	Group 1: 4, 5, 6, 7 Group 2A: 1, 2, 9 Group 2B: 3	Group 1: 3 Group 2A: 1 Group 2B: 2, 4, 5	Group 1: 1, 2, 3, 4, 6 Group 2B: 5	Group 1: 1, 2, 3, 4, 5, 6	Group 2A: Albert, CAR-South, VR
Particle size	Group 1: 4, 5, 6, 7 Group 2A: 1, 2, 9 Group 2B: 3	Group 1: 3 Group 2A: 1 Group 2B: 2, 4, 5	Group 1: 1, 2, 3, 4, 6 Group 2B: 5	Group 1: 2, 3, 6	Group 2A: Albert, CAR-South
Clay mineral., PCA, linear modeling	Group 1: 4, 5, 6, 7 Group 2A: 1, 2, 9 Group 2B: 3	Group 2A: 1 Group 2B: 2, 5	Group 1: 2, 4 Group 2B: 5	Group 1: 2, 3, 6	Group 2A: Albert, VR (VR for clay mineralogy only)
C and N isotopes	Group 1: 5, 7 (+ 3 replicates), 12 Group 2A: 1, 2 Group 2B: 3	Group 1: 3 Group 2A: 1	Group 1: 2, 4 Group 2B: 5	Group 1: 1, 2, 5	

Table 2. Results of principal components analysis

Variables: Diff OC, pH, % sand, % silt, % clay, % smectite, % vermiculite, % illite, % kaolinite + chlorite (variables in italics were logratio transformed before analysis)

Importance of components

	Comp. 1	Comp. 2	Comp. 3
Standard Deviation	1.870	1.539	1.383
Proportion of Variance	0.389	0.263	0.212
Cumulative Proportion	0.389	0.652	0.865

Loadings

	Comp. 1	Comp. 2	Comp. 3
Diff OC	-0.289	0.242	-0.372
pH	0.144	0.224	-0.589
sand	0.459	-0.228	-0.257
silt	-0.305	0.300	0.384
clay	-0.486	0.106	
smectite	-0.353	-0.261	-0.404
vermiculite	-0.109	-0.558	0.229
illite	0.164	0.593	
kaolinite+chlorite	0.439		0.275

Table 3. Linear models predicting Diff OC. For all models, the coefficients and intercept are all significantly different from zero ($p < 0.001$), except for the coefficients for % smectite and % vermiculite in the last two equations, which are not significantly different from zero. AIC is Akaike Information Criterion; n = 111.

Model	Overall F-statistic (p)	Adj. R^2	AIC
Diff OC = $0.781 - 0.018\text{Sand}$	32.18 (< 0.001)	0.23	257
Diff OC = $-0.951 + 0.052\text{Clay}$	51.41 (< 0.001)	0.31	242
Diff OC = $-1.794 - 0.025\text{Sand} + 0.426\text{pH}$	53.73 (< 0.001)	0.49	211
Diff OC = $-3.850 + 0.064\text{Clay} + 0.398\text{pH}$	70.82 (< 0.001)	0.56	194
Diff OC = $-3.861 + 0.062\text{Clay} + 0.391\text{pH} - 0.003\text{Smect}$	46.96 (< 0.001)	0.56	196
Diff OC = $-3.411 + 0.063\text{Clay} + 0.365\text{pH} - 0.010\text{Vermic}$	47.62 (< 0.001)	0.56	195

Table 4. Radiocarbon ages of bulk soil organic matter, CAR-7 soil profile

Depth (cm)	Lab Number	Age (^{14}C yr BP)	$\delta^{13}\text{C}$
0-5	OS-85327	[110% modern]	25.05
20-25	OS-85328	415 $\pm$ 25	19.14
42-47	OS-85342	895 $\pm$ 30	17.41

Table 5. Vegetation recorded near section and quarter section corners, located on Mollisols likely to have formed in the DC lithofacies, Houston County, Minnesota, public land survey of 1854 (calculated from dataset of Almendinger, 1990).

Vegetation Description	Corners with NW, N, NE Aspects	Corners with SW, S, SE Aspect
Forest	88 (50%)	64 (37%)
Scattering timber/oaks	60 (34%)	58 (33%)
Brush	1 (<1%)	1 (<1%)
Oak openings	5 (3%)	9 (5%)
Grove	3 (2%)	0
Oak barrens	0	2 (1%)
Prairie	7 (4%)	12 (7%)
Dry land/Dry ridge	11 (6%)	29 (17%)
	175 (100%)	175 (100%)

Figure 1a

Figure 1b

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11