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# Paleoclimatic interpretations of buried paleosols within the pre-Illinoian till sequence in northern Missouri, USA

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ABSTRACT

Northern Missouri preserves shallow buried paleosols ranging in age from Early to Late Pleistocene that are developed in six pre-Illinoian tills beneath Illinoian (MIS 6) and Wisconsinan (MIS 4-2) loess. The morphology of these paleosols changes with age, reflecting a changing climate during the Pleistocene, and cosmogenic-nuclide burial dates of the respective sola provide age control on the timing of these changes.

The depth to secondary calcium carbonate nodules within the weathering profiles increases with younger age, indicating a transition to moister conditions during the Pleistocene, and these nodules are absent entirely within the modern soils. After approximately 0.4 Ma, the sola became distinctly redder, even as the time available for pedogenesis became shorter, culminating in the bright red Sangamon Geosol (MIS 5). This trend is consistent with increasing interglacial temperatures and/or precipitation. Finally, erosion rates determined from cosmogenic-nuclide concentrations within the sola also increase systematically with younger age. This increase may be due to some combination of changing climate, more-frequent glaciations and the deposition of a thick cover of unconsolidated glacial sediment above the stable residuum-dominated preglacial landscape.



## 1. Introduction

A sequence of buried paleosols ranging in age from Early to Late Pleistocene is preserved in northern Missouri within a sequence of pre-Illinoian (pre-MIS 6) tills and overlying loess (Figs. 1, 2). The pedology and geochemistry of these paleosols represent a potential record of paleoclimate and paleoenvironment during Pleistocene interglacial periods. In this paper, we compile and summarize data from multiple paleosols at 34 stratigraphic sections in northern Missouri (Table 1) in order to i) describe these paleosols; ii) highlight temporal variations in soil morphology, stratigraphy, and geochemistry during the Pleistocene; and iii) relate these variations to likely paleoclimatic or paleoenvironmental causes.

We focus on three striking examples of systematic temporal changes. First, pedogenic calcium carbonate (colloquially called "caliche"), which is not present in the modern soils, occurs more commonly and at shallower depths in older paleosols. As the presence and depth of secondary carbonate is related to mean annual precipitation (MAP), we propose that this records an increase in interglacial MAP during the Pleistocene. Second, the morphology of paleosols changes with age. Late Middle Pleistocene soils, although representing shorter durations of pedogenesis, have redder sola than early Pleistocene soils, culminating in the bright red Sangamon Geosol (MIS 5). Although the reasons for the distinctive red color of the Sangamon and the (somewhat older) Yarmouth Geosol have been discussed for many years, our observations place these observations in a temporal context and thus help to assess the competing hypotheses of warmer interglacial temperatures, longer weathering durations, and burial diagenesis as the dominant factor for reddening of the younger paleosols. Finally, we use cosmogenic-nuclide measurements, already carried out on these paleosols in previous research for the purposes of determining their age, to also estimate soil surface erosion rates prior to burial. Younger paleosols display higher apparent erosion rates, which may reflect either climate change or an increased frequency of landscape disturbance due to glaciation and glacial–interglacial climate change. Overall, we conclude that the Missouri paleosol sequence records a systematic increase in interglacial temperature, moisture availability and rates of erosion and landscape development during the Pleistocene.

## 2. Stratigraphy

Six Pleistocene tills in northern Missouri are grouped into three formal formations and three informal members (Figs. 1, 2; Rovey and Tandarich, 2006; Rovey and Balco, 2011). These tills display consistent clay loam textures, but have distinct and laterally consistent differences in lithology, thus in most cases allowing identification by lithologic characteristics alone. Four mature weathering profiles capped by argillic B horizons divide this till sequence into five major glacial sedimentary sequences corresponding to each of the named lithostratigraphic units

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**Fig. 1.** Location and physiography of study area in central Missouri. Numbered triangles correspond to site numbers in Table 1. The dashed line in the main panel is the southern boundary of known glacial deposits.

(Fig. 2). Informal names have been applied to some of these paleosols (Guccione and Tandarich, 1993; Rovey, 1997), but here we designate paleosols older than the Yarmouth Geosol (a paleosol buried by Illinoian deposits; Hallberg et al., 1980) simply by the lithostratigraphic name of the unit on which the soil formed; e.g., the Atlanta paleosol is the paleosol developed on the Atlanta-Formation till.

The two oldest tills (Atlanta and Moberly) have a reversed magnetic remanence, which restricts their age to the Matuyama Chron or Early Pleistocence. Cosmogenic-nuclide burial ages for these tills are $2.42 \pm 0.14$ Ma and $1.31 \pm 0.09$ Ma, respectively (Balco and Rovey, 2010). The younger three tills of the McCredie Formation are magnetically normal and underlie the 0.16 Ma Loveland Silt (Foreman and Pierson, 2002; Mason et al., 2006). The oldest of the three normal-polarity tills, the Fulton till, has a cosmogenic-nuclide burial age of $0.80 \pm 0.06$ Ma (Balco and Rovey, 2010). Given its normal magnetic remanence, it is most likely that the Fulton was emplaced during MIS 18 near 0.76 Ma.

The two youngest tills (Columbia and Macon) have similar lithologies, indicating that they share similar ice-accumulation and source areas, and are closer in age than the older three tills. Burial ages for the Columbia and Macon are $0.22 \pm 0.15$ Ma and $0.21 \pm 0.17$ Ma, respectively (Balco and Rovey, 2010). An argillic horizon in the Columbia paleosol suggests that deposition of the two tills was separated by at least one relatively long interglacial period. These observations, along with the fact that the Macon till underlies MIS 6 loess, suggest that the Columbia and Macon tills were emplaced during two of marine oxygen isotope stages 12 (0.48–0.42 Ma), 10 (0.37–0.34 Ma), or 8 (0.3–0.24 Ma).

The till sequence is overlain by several loess units, the oldest of which is the Illinoian (MIS 6) Loveland Silt. The Loveland began accumulating at ~0.16 Ma (Foreman and Pierson, 2002; Mason et al., 2006) and is distinguished from younger loess by pinkish hues, higher kaolinite contents, and the Sangamon Geosol at the top. The Loveland is most commonly preserved on flat uplands near the Missouri River (Guccione, 1983). However, it occurs in a variety of landscape positions and directly overlies various tills ranging in age from that of the Macon to the Moberly (Figs. 2, 3). Therefore, much of the present topographic relief developed prior to 0.16 Ma.

The Loveland Silt is buried by Wisconsinan (MIS 4-2) loess. The Early Wisconsinan (MIS 4) Roxana Silt may be present locally as a thin layer with a faintly developed A horizon of the Farmdale Geosol (Guccione, 1983; Rovey, 1997). Nevertheless, the Roxana is rarely preserved or at least distinguishable from the overlying Peoria Loess. The Peoria is the

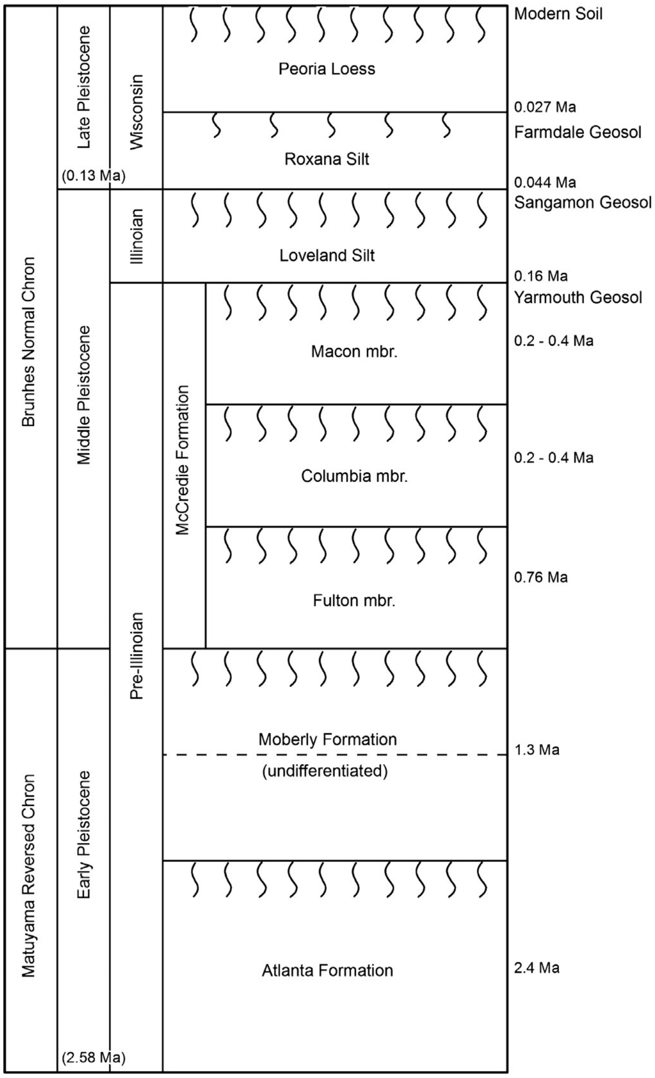


**Fig. 2.** Stratigraphic sequence of tills, loess and paleosols in northern Missouri. The McCredie, Moberly and Atlanta Formations are mostly till. Ages of the tills are from Balco and Rovey (2010); see text for error limits. Ages for the loess are from the compilation in Mason et al. (2006).

most common and thickest loess that, like the underlying Loveland Silt, drapes over a dissected landscape predating loess deposition.

## 3. Background

### 3.1. Soil development rates

In general, rates of soil development within freshly exposed materials are initially rapid and decrease exponentially with time after the onset of pedogenesis (Bockheim, 1980; Birkeland, 1999; Schaetzl and Anderson, 2005). For soils with formation times up to approximately 0.1 Ma, quantitative measures of soil development are commonly accurately approximated by simple exponentially decreasing rate models, but it is not clear whether this is appropriate for longer durations, mainly because there are fewer soils older than this that are accurately dated. In particular, B-horizon thickness and clay content approach constant steady-state values after 0.1–0.2 Myr of soil formation in climates of the midwestern U.S. (Grimley et al., 2003).

Rubification or reddening is caused by accumulation of hematite and other iron oxides produced by weathering of iron-bearing minerals within the solum. Iron oxide content is primarily correlated with both surface temperature and soil formation time, and secondarily by clay content, which provides suitable surfaces for iron oxide coatings (Schaetzl and Anderson, 2005; Birkeland, 1999; Muhs et al., 2001). Rubification can proceed within a developing solum past 0.1 Myr and possibly beyond 1 Myr (Harden and Taylor, 1983; Markewich et al., 1989).

The Sangamon and Yarmouth Geosols are distinguished visually from younger (and older) soils by their bright red coloration, although this aspect is better documented for the Sangamon due to its broader geographic extent and better preservation. Many early workers recognized that the redness of the Sangamon is geographically and climatically anomalous compared to the local modern soils (e.g. Thorp et al., 1951; Ruhe, 1974, and references therein). However, the redness of the Sangamon could be due to a warmer climate, longer development, or both (Simonson, 1954; Ruhe, 1974; Boardman, 1985; Grimley et al., 2003). More recently Thompson and Soukup (1990) proposed burial diagenesis for the red coloration of a Sangamon profile in western Iowa, based on the relative position of argillans and iron oxide coatings observed in thin sections. Under this hypothesis the Sangamon's red color could reflect a warmer post-burial climate instead of conditions during formation.

**Table 1**
Locations and names of surface exposures and cores used in this study.

| Number | Location | Latitude | Longitude |
|---|---|---|---|
| 1 | Maitland Quarry | 40°13.63′ | 95° 01.71′ |
| 2 | Breit Quarry | 39°54.08′ | 94°49.89′ |
| 3 | Cameron Quarry | 39°47.38′ | 94°17.81′ |
| 4 | Jeffries Quarry | 40° 33.49′ | 93° 50.11′ |
| 5 | Blue Mound Quarry | 39° 37.15′ | 93° 34.30′ |
| 6 | Mercer Quarry | 40° 30.61′ | 93° 29.12′ |
| 7 | AECI Pit | 39° 32.50′ | 92° 40.17′ |
| 8 | SMS92C | 40° 02.02′ | 92° 29.16′ |
| 9 | SMS92B | 39° 53.92′ | 92° 28.29′ |
| 10 | SMS92A | 39° 48.15′ | 92° 28.45′ |
| 11 | Stadium Dr. Cut | 38° 56.07′ | 92° 17.56′ |
| 12 | FU02 | 38° 49.30′ | 92° 00.00′ |
| 13 | FU03 | 38° 38.68′ | 91° 59.20′ |
| 14 | KT17 | 38° 56.19′ | 91° 52.63′ |
| 15 | Harrison Pit | 38° 57.53′ | 91° 52.42′ |
| 16 | Sieger Pit | 39° 14.88′ | 91° 48.36′ |
| 17 | PF2 | 38° 54.24′ | 91° 44.58′ |
| 18 | RV17 | 38° 48.94′ | 91° 43.05′ |
| 19 | Riedell Pit | 39° 26.65′ | 91° 42.08′ |
| 20 | WB19 | 38° 57.84′ | 91° 41.22′ |
| 21 | Readsville Pit | 38° 45.66′ | 91° 38.41′ |
| 22 | Johnson & Deeker Pits | 38° 52.20′ | 91° 27.10′ |
| 23 | NF06 | 38° 54.20′ | 91° 26.61′ |
| 24 | Musgrove Pit | 38° 51.80′ | 91° 26.10′ |
| 25 | JB12 | 38° 51.81′ | 91° 19.13′ |
| 26 | JB10 | 38° 51.08′ | 91° 17.58′ |
| 27 | Pendelton Pit | 38° 47.28′ | 91° 15.11′ |
| 28 | HP01 | 38° 58.71′ | 91° 13.65′ |
| 29 | Polston Pit | 38° 45.79′ | 91° 10.45′ |
| 30 | WNE04 | 38° 52.73′ | 91° 02.59′ |
| 31 | Warrenton Pit | 38° 43.08′ | 91° 01.67′ |
| 32 | WL3 | 38° 51.78′ | 90° 59.40′ |
| 33 | WL2 | – | – |
| 34 | WL4 | – | – |

### 3.2. Secondary calcium carbonate

The depth to pedogenic calcium carbonate is widely recognized as a proxy for MAP (Busacca, 1989; Birkeland et al., 1991; Birkeland, 1999; Retallack, 1997, 2005). In this paper we generally avoid the term "calcic horizon" for these accumulations, because some of the horizons considered here contain less than 15% secondary calcium carbonate.

Pedogenic calcium carbonate ($CaCO_3$) forms within or below a solum within dry to seasonally dry climates. The carbonate begins to precipitate at a depth where the pore water is held immobile under tension

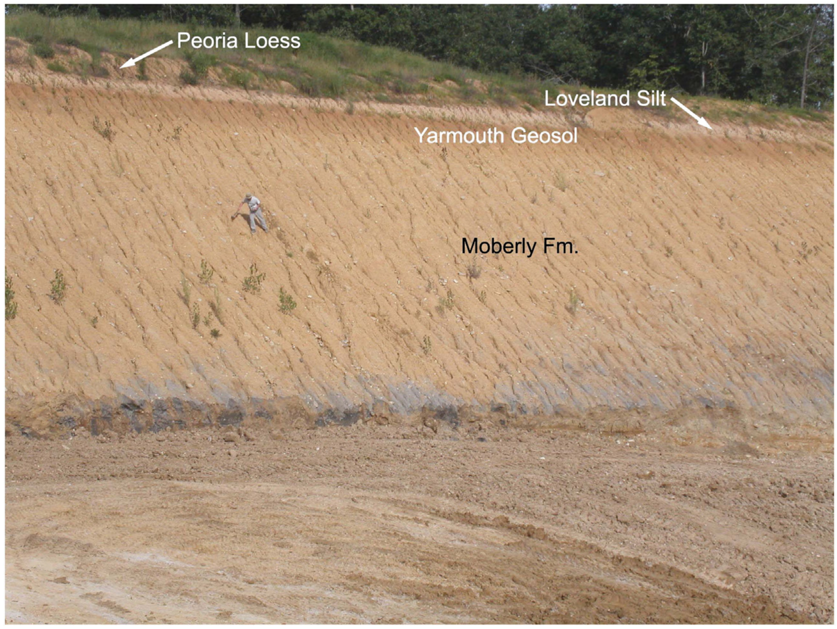


**Fig. 3.** Yarmouth Geosol buried by Loveland Silt at the Musgrove Pit (Table 1, Site 24).

for extended durations. There, the water reaches supersaturation with respect to $CaCO_3$ due to evapotranspiration and/or an increase in pH caused by hydrolysis of noncarbonate minerals. Thus, the top of the pedogenic carbonate corresponds to the maximum depth of significant water flux, which increases with MAP (Arkley, 1963; Birkeland, 1999; Retallack, 2005).

Jenny and Leonard (1934) first developed a relationship between MAP and the depth to pedogenic $CaCO_3$ with 104 observations along an east-west transect from western Missouri to eastern Colorado. To our knowledge, the only modern soils in Missouri with pedogenic carbonate are present in the extreme northwest portion of the state, which was the eastern margin of their transect. The present study area lies between this bounday, where MAP is near 90 cm year$^{-1}$, and east-central Missouri, where MAP is near 100 cm year$^{-1}$. Thus, 90 cm year$^{-1}$ is the approximate local MAP threshold for pedogenic calcium carbonate accumulation, and it is close to the highest values recorded for calcic horizons in the North American data bases of Royer (1999) and Retallack (2005).

Jenny and Leonard's relationship between MAP and "depth to carbonate zone," as recalculated by Royer (1999), is:

$$P = 2.324D + 420.2 \quad \left(r^2 = 0.64, \sigma = 109\right) \tag{1}$$

where P is MAP (mm), D is depth to pedogenic carbonate (cm), and σ is the standard deviation of the residuals. This equation is valid for MAP below approximately 900 mm; above that value it systematically overestimates MAP.

Retallack (2005) compiled a larger database with 807 soils from different regions, including the original Jenny and Leonard values. To improve the fit to data above MAP = 900 mm, he used the quadratic equation:

$$P = 137.24 + 6.45D–0.013D^2 \left(r^2 = 0.52, \sigma = 147\right) \tag{2}$$

where variables are the same as in Eq. (1). This function better represents the observations that i) correlation between D (depth to pedogenic carbonate) and P (MAP) breaks down at high MAP, and ii) pedogenic carbonate is rarely observed at MAP exceeding ~900 mm. Thus, values of D exceeding ~2 m show only that MAP was likely greater than 900 mm.

### 3.3. Leaching depth

We use the term "leaching depth" to denote the depth to which primary carbonates have been dissolved by infiltrating groundwater from the bulk matrix of an originally calcareous material, in this case till. Within mature weathering profiles the leaching depth may be related to the depth of pedogenic carbonate, because water that precipitates secondary $CaCO_3$ is supersaturated and will not dissolve additional primary $CaCO_3$ below that depth (Birkeland et al., 1991). Moreover, if infiltrating water regained capacity for leaching of calcareous parent material beneath an accumulation of pedogenic carbonate at some later time, that water would be even more aggressive in dissolving and removing any pedogenic carbonate higher in the profile. Therefore, leaching and carbonate precipitation depths should be similar in most cases, but because rates of calcite dissolution and precipitation may vary with respect to changes in seasonal saturation and MAP, these depths may not be identical at all times. Jenny and Leonard, though, apparently considered them to be equivalent and used the leaching depth in cases where pedogenic carbonates were not observed.

The rate and depth of leaching in a soil increase with MAP (Egli and Fitze, 2001). Nevertheless, in contrast to the well-documented association between pedogenic-carbonate depth and MAP, the leaching depth in paleosols of similar texture is commonly interpreted as primarily a function of soil age and secondarily, if at all, a function of climate. For relatively young soils, this is certainly the case (e.g., latest Pleistocene soils; Flint, 1949; Mickelson and Evenson, 1975). However, geochemical modeling indicates that leaching rates, like rates of other pedogenic processes, decrease exponentially over time (McFadden and Tinsley, 1985). Such a decrease is evident when comparing measured leaching depths in tills of various age to the time of exposure (Flint, 1949; Hallberg, 1980a, 1980b; Kemmis et al., 1981; Birkeland, 1999). This implies that the rate of leaching would eventually decrease to a value approaching the erosion rate of the overlying soil, in which case the leaching depth, like the depth to secondary carbonate, would correlate with MAP.

### 3.4. Surface erosion rates inferred from cosmogenic-nuclide measurements

In previous work (Balco and Rovey, 2010), we applied the technique of cosmogenic-nuclide burial dating to the Early and Middle Pleistocene paleosols in the Missouri glacial sequence to determine the age of the tills. This technique relies on the fact that the cosmic-ray-produced radionuclides $^{10}$Be and $^{26}$Al are produced at a fixed ratio in quartz grains exposed to the cosmic-ray flux at the Earth's surface. If these quartz grains are then buried by a thick enough till (several meters) to effectively halt the cosmic-ray flux, then $^{10}$Be and $^{26}$Al production stops and the concentrations of these nuclides decrease due to radioactive decay. As $^{26}$Al has a shorter half-life than $^{10}$Be, the $^{26}$Al/$^{10}$Be ratio decreases over time. Thus, measuring $^{10}$Be and $^{26}$Al concentrations in a buried paleosol enables one to determine the age of an overlying till (Balco and Rovey, 2008, 2010).

Besides providing age control for the paleosols in the Missouri glacial sequence, these cosmogenic-nuclide measurements are also useful to the present study for two reasons. First, the difference in age between successive tills provides information on the time available for development of the paleosol formed on the lower till. For example, 1.1 Ma was available for development of the Atlanta paleosol, and 0.5 Ma for the development of the Moberly paleosol. Second, the absolute concentrations (in contrast to the ratios used to determine burial ages) of cosmic-ray-produced nuclides in paleosol quartz provide information on the surface erosion rate prior to paleosol burial. The process of erosion can be thought of as advection of fresh rock or sediment, that has not been exposed to the surface cosmic-ray flux, towards the surface through a relatively thin zone (~2 m, depending on material density) where the majority of cosmogenic-nuclide production takes place. The faster the erosion rate, the less time a particular soil particle spends in the production zone before being lost to erosion, and the lower its cosmogenic-nuclide concentration when it reaches the surface. Thus, the $^{10}$Be or $^{26}$Al concentration in soil quartz is inversely proportional to the erosion rate and, given several assumptions that are described below and in the supplemental material, provides a quantitative estimate of the erosion rate of the soil surface (Brown et al., 1995; Bierman and Steig, 1996; Granger et al., 1996).

## 4. Data collection

The majority of publications dealing with the pre-Illinoian till sequence in Missouri describe paleosols very briefly and only as necessary to establish stratigraphic boundaries (e.g. Rovey and Kean, 1996). More detail on the paleosols is given for the various type sections and other important sites in Rovey (1997) and Rovey and Tandarich (2006), but these descriptions do not include pedogenic carbonate except where it constitutes a prominent calcic horizon. The data compiled in this paper are drawn exclusively from field and laboratory notes collected by the first author since the early 1990s, and include many new examples provided by cores from the Missouri Division of Geology and Land Survey's STATEMAP program. The base of B horizons, leaching depths, and depths to the top of pedogenic $CaCO_3$ were recorded, usually to either the nearest 0.5 or 1.0 ft (0.15, 0.30 m) and less commonly at 0.25 ft (0.08 m). These values are converted to the nearest 0.1 m in Table 2; thus, the accuracy is slightly less than implied in some cases. Leaching depths were determined initially by matrix reactions with dilute HCl, and most of these depths were corroborated with grain counts of the coarse-sand fraction. Depletion of carbonate grains occurring at the same elevation as the loss of effervescence shows that profiles have not recalcified after burial. The redox state of each paleosol was recorded in general terms as “gleyed” (hues 2.5–5 Y, value >5 and chroma <3) or “oxidized” (hues of ~10 YR or redder, value >4, chroma >2). More precise hues were also recorded by comparison with Munsell charts at most surface exposures and for some core samples.

Clay content and clay–mineral alteration were also measured for a series of closely spaced samples spanning stratigraphic contacts to confirm the base of the B horizon and depth of pedogenic alteration; many of these measurements are summarized and/or tabulated in Rovey (2012). The base of the B horizon is taken here as the depth at which the clay content begins a systematic upward increase. This depth usually corresponds with visual estimates based on soil structure, prominent cutans, and redox features, although in some instances a B subhorizon or BC horizon might be delineated beneath this boundary, based on coloration, and/or weak soil structure. If pedogenic carbonate is concentrated beneath the base of clay enrichment, we consider this to be a Ck horizon, not a Bk.

Pedogenic $CaCO_3$ usually is present as discrete nodules (Stage II; Gile et al., 1966), although at three sites it locally reaches Stage III, an indurated crust. The size of the nodules was not consistently measured, but most are less than 10 cm in diameter. The range in depth at a given site typically is less than one meter within the pre-Yarmouth paleosols, but may exceed two meters within the Yarmouth.

To estimate pre-burial erosion rates, we use $^{10}$Be measurements on paleosol quartz from 12 sites, 10 from the Missouri till sequence and one from a correlative paleosol in Iowa, from Balco and Rovey (2010). We omitted a poorly developed weathering horizon between tills of the Alburnett Formation in Iowa that was burial-dated by Balco and Rovey (2010) (their site 11), because it shows negligible soil development and most likely represents only a short period of surface exposure. We also include $^{10}$Be measurements from one additional site in Missouri (the Columbia paleosol in the PF2 core; see Table S1), where cosmogenic-nuclide data were insufficient for burial dating, but the age of the overlying till is known from other correlative sites.

Here we describe the procedure for estimating the paleo-erosion rate from measured $^{10}$Be concentrations (although both $^{10}$Be and $^{26}$Al concentrations were measured, they provide redundant information for this purpose and we focus only on $^{10}$Be here). First, we reconstruct the $^{10}$Be concentrations attributable to cosmic-ray exposure during paleosol formation by correcting measured $^{10}$Be concentrations for i) radioactive decay; ii) post-burial nuclide production by deeply penetrating muons; and iii) inherited $^{10}$Be present in the till parent material at the time soil formation began. In addition, because some of the paleosols were truncated by subglacial erosion during emplacement of overlying tills, we use observations of preserved soil horizon thicknesses to estimate the truncation depth and therefore the depth of the samples below the paleosol surface immediately prior to burial. These steps are described in detail in the supplementary information, and yield a range of permissible $^{10}$Be concentrations in the paleosol at the time it was buried by the overlying till.

Second, given reconstructed $^{10}$Be concentrations at one or more depths in each paleosol prior to burial, we can use this information to estimate the paleosol surface erosion rate. Although the basic principle that the $^{10}$Be concentration in quartz near the soil surface is inversely proportional to the erosion rate is true no matter what the process of soil formation, the actual form of this relationship depends on the rate and extent of vertical mixing in the soil (e.g., Lal and Chen, 2005). In theory, given enough samples, the form of the $^{10}$Be concentration–depth relationship should enable quantitative reconstruction of both the erosion rate and the depth and degree of mixing. In this case, however, we need a consistent means of estimating pre-burial erosion rates from $^{10}$Be concentrations in a set of paleosols where: i) in most cases, the number of samples is inadequate to characterize the $^{10}$Be concentration–depth profile in detail; ii) complex variations in concentration–depth profiles indicate significant vertical soil mixing; and iii) some of these variations are most likely the result of subglacial deformation. Given these uncertainties, we did not attempt to fit a deterministic mixing model to $^{10}$Be concentrations in each paleosol. Instead, we assume that the sample with the highest $^{10}$Be concentration observed in each profile was, at the time of burial, most likely part of a surface mixed layer. We then convert the permissible range of reconstructed $^{10}$Be concentrations for this sample to a range of permissible

**Table 2**
Measured properties of paleosols at sites in Table 1.

| Unit | Location | B Thickness (m) | Hue | Leaching depth (m) | Pedogenic carbonate depth (m) | Comments |
|---|---|---|---|---|---|---|
| Macon (Yarmouth and Yarmouth–Sangamon geosol) | | | | | | |
| | Breit Quarry | 1.9 | 5 YR | 0.8 | 1.1 | Buried by Peoria Loess |
| | Cameron Q. | 1.1 | 5 YR | 0.3 | 0.3 | Buried by Peoria Loess |
| | Mercer Quarry | 1.2 | 7.5 YR | 1.8 | 2.1 | Buried by Peoria Loess |
| | SMS92a | 1.5 (g) | | 0.9 | – | Buried by Loveland Silt |
| | SMS92b | 2.7 (g) | | 2.1 | 1.8 | Buried by Loveland Silt |
| | SMS92c | 1.2 (g) | | 3 | – | Buried by Loveland |
| | Sieger Pit | 1.5 (g) | | 1.2 | – | Buried by Loveland |
| | PF2 | 1.8 | 2.5 YR | 1.5 | – | Buried by Loveland |
| | Riedell Pit | 1.2 | 5 YR | 0.9 | – | Buried by Peoria |
| Columbia | | | | | | |
| | SMS92a | – | | – | >0.2 | Nodules are beneath |
| | SMS92b | – | | – | >1.4 | truncated solumn |
| | PF2 | 0.6 | 7.5 YR | 0.0 | 0.5 | Bg/Bt profile |
| | Sieger Pit | 0.3 (g) | 7.5YR | 1.3 | 1.2 | A/Bg/BC profile |
| | Riedell Pit | 0.6 | 10 YR | 1.5 | 0.6 | |
| Fulton | | | | | | |
| | Mercer Quarry | 0.6 | 10 YR | 0.0 | 0.3 | |
| | AECI Pit | 0.6 | 10 YR | 0.9 | – | |
| | SMS92a | 2.1 (g) | | 0.9 | −1.7 | |
| | SMS92b | 0.3 | 10 YR | 0.0 | −0.3 | |
| | KT17 | 1.4 (g) | | 2.2 | – | |
| | Harrison Pit | 0.2 | 10 YR | 0.3 | – | |
| | RV17 | 0.8 | 10 YR | 0.0 | −0.3 | |
| | WL3 | 0.2 | 10 YR | 0.3 | <−0.2 | |
| Moberly | | | | | | |
| | Blue Mound Q. | 1.5 | 7.5–10 YR | 0.0 | −0.8 | |
| | Mercer Quarry | 0.6 (g) | | 0.0 | – | |
| | AECI Pit | 0.6 (g) | | 0.0 | – | |
| | FU02 | 1.8 (g) | | 0.0 | −0.3 | |
| | Harrison Pit | 0.6 | 10 YR | 0.0 | <−0.6 | Nodules to top of truncated B |
| | Deeker Pit | 0.2 | 10 YR | 0.0 | <−0.2 | Nodules to top of truncated B |
| | RV17 | 2.1 (g) | | 0.0 | – | |
| | NF06 | 1.1 (g) | | 0.0 | −0.2 | |
| | Pendelton Pit | 0.3 | 10 YR | <0.0 | – | Incomplete leaching |
| | WL2 | 0.2 | 10 YR | 0.2 | 0.8 | |
| Atlanta | | | | | | |
| | Stadium Rd. Cut | 0.9 | 7.5–10 YR | <0.0 | <−.9 | Incomplete leaching, Gypsum |
| | Readsville Pit | 0.5 | 10 YR | 0.0 | – | |
| | Johnson Pit | 1.5 (g) | | 0.0 | −0.5 | |
| | Musgrove Pit | 2.1 (g) | | 0.0 | −0.9 | A horizon, Gypsum |
| | Pendleton Pit | 1.2 | 7.5–10 YR | <0.0 | – | Incomplete leaching |
| | Warrenton Pit | 2.1 (g) | | 0.0 | −0.3 | |
| | Additional Yarmouth–Sangamon profiles in older tills | | | | | |
| Columbia | | | | | | |
| | WNE04 | 0.9 (g) | | 0.9 | – | |
| | Harrison Pit | 0.9 | 7.5 YR | 0.5 | 2.0 | Peoria(?) |
| | AECI | 1.5 | 10 YR | 3.6 | −0.3 | Peoria |
| | JB10 | 2.7 (g) | | 0.5 | – | |
| | WL2 | 1.2 (g) | | 1.5 | 0.8 | |
| Fulton | | | | | | |
| | FU02 | 2.4 (g) | | 2.1 | – | |
| | FU03 | 1.8 (g) | | 1.8 | – | |
| | WB19 | >1.2 | | 2.1 | – | |
| | NF06 | 2.1 (g) | | 1.4 | – | |
| | HP01 | 0.8 | | 0.5 | – | |
| | WL4 | 1.1 | 7.5 YR | 1.8 | – | |
| | Pendleton Pit | 1.4 | 5 YR | 0.6 | – | Peoria |
| Moberly | | | | | | |
| | Maitland Q. | 2.1 | 7.5 YR | | 2.0 | Peoria |
| | Jeffries Q. | 1.5 | 7.5 YR | | – | Peoria |
| | JB12 | 1.4 (g) | | 3.0 | 1.5 | |
| | Musgrove Pit | 1.4 | 7.5 YR | 1.7 | 0.6 | Loveland |
| | Johnson Pit | 0.8 | 7.5 YR | 1.0 | 1.2 | Loveland |
| Atlanta | | | | | | |
| | Polston Pit | 2.1 | 2.5 YR | 0.0 | – | Peoria |

Notes: Depths are given relative to the base of the B horizon. A positive value means that this depth is below the base of the B horizon; a negative value means that it is above. (g) denotes a gleyed profile. Hues are given for aerated profiles only; dashes indicate that pedogenic carbonate was not observed.

erosion rates given a range of possible mixing models (see additional discussion in the supplemental information). In this study, we argue that this highly simplified procedure is adequate because the most important observation we present is that we observed nearly two orders of magnitude variation in reconstructed near-surface $^{10}$Be concentrations in our study paleosols. Thus, our goal in this calculation is mainly to place this observation in a framework of erosion rates, which are geomorphically meaningful, rather than nuclide concentrations, which are less so. More precise estimates of paleo-erosion rates are not required to support our conclusions.

## 5. Results

### 5.1. B horizons and carbonate data

Depths to the top of secondary pedogenic carbonate, leaching depths of primary carbonate and the preserved thickness and hues of respective B horizons are listed by stratigraphic unit in Table 2 and summarized in Fig. 4. This compilation includes instances in which i) portions of a B horizon were found atop the four oldest tills (Columbia to Atlanta), and ii) successive weathering zones (excluding oxidation) had not merged. We excluded rare instances in which a gap or discontinuity is present within the till sequence, meaning that the next-youngest till, which normally would overlie a given paleosol, was missing; in this case we could not verify that the paleosol was buried at the same time and thus records the same paleoenvironmental conditions as correlative paleosols (however, all results are included for the youngest Macon paleosol, regardless of the overlying loess unit). Likewise, we excluded cases in which the solum was clearly pervasively sheared and deformed during burial or overthickened by accretionary deposition as indicated by lithologic discontinuities and/or erratic fluctuations in weathering parameters (e.g. Rovey, 1997). At some locations, continued excavation over several years exposed minor variation within a given paleosol and from previously reported thicknesses. In these cases we use measurements from the thickest measured solum.

The number of sites at which we observed each paleosol is mainly the result of the regional stratigraphy. For example, we made relatively few observations (6) of the Atlanta till, which lies at the bottom of the section and generally is only observed in deep excavations. In addition, it is likely that it is poorly preserved due to its long exposure, ~1.1 Myr. A location bias also lowers totals for the youngest two paleosols (Macon and Columbia). Many of these sites cluster along the I-70 corridor, close to the Missouri River and the approximate southern limit of glaciation. All of the tills have been found within a few kilometers of this boundary, but near the river, the youngest tills are preserved only along the midportion of stable flat interfluves due to headward erosion by tributary streams.

All of the pre-Yarmouth paleosols were truncated to some extent by proglacial and/or subglacial erosion, even though portions of A horizons are preserved at several sites. The maximum preserved thickness of the B horizon is ~2 m for the Atlanta, Moberly and Fulton paleosols. The Yarmouth maximum is somewhat thicker, but these profiles were likely overthickened by loess additions, as discussed later. If we assume that the depth of truncation of B horizons by till emplacement is random and not systematic, this consistency supports the idea that B horizons approached a quasi-steady thickness over times shorter than the interval between most of the pre-Illinoian glaciations. Therefore, pedogenic carbonate and leaching depths are listed relative to the base of the B horizon, since the former land surface is not preserved. A negative value for the pedogenic-carbonate depth means that the top of the nodules is **above** the base of the B horizon; a positive value means it is below.

The predominant redox state of each solum given in Table 2 is a general indicator of paleolandscape position. Gleying commonly occurs in two different positions: low positions that collect runoff and flat uplands where runoff is impeded. Based on the paleotopography observed in the larger surface exposures, the latter (stable, poorly drained uplands) is usually the case for the gleyed profiles, and these tend to be somewhat thicker than the well-drained aerated profiles that developed on slopes.

Gleyed instances occur in all paleosols, which shows that the Pleistocene climate in northern Missouri was persistently moist enough to produce gleyed B horizons, albeit with prominent oxidized mottling, implying that water saturation was only seasonal. Given the distribution of modern soils with gleyed horizons spanning humid to semiarid regions, this attribute provides little constraint on paleoclimate. Some of the profiles also preserve vertic properties such as curving joints with slickensides (Rovey, 1997). These develop during seasonal wetting following prolonged dry periods with extreme moisture deficits (e.g. Schaetzl and Anderson, 2005).

Other soil properties summarized in Table 3 and Fig. 4 indicate significant temporal changes in the environment of soil formation. For the oldest two paleosols (Atlanta and Moberly) the leaching depth

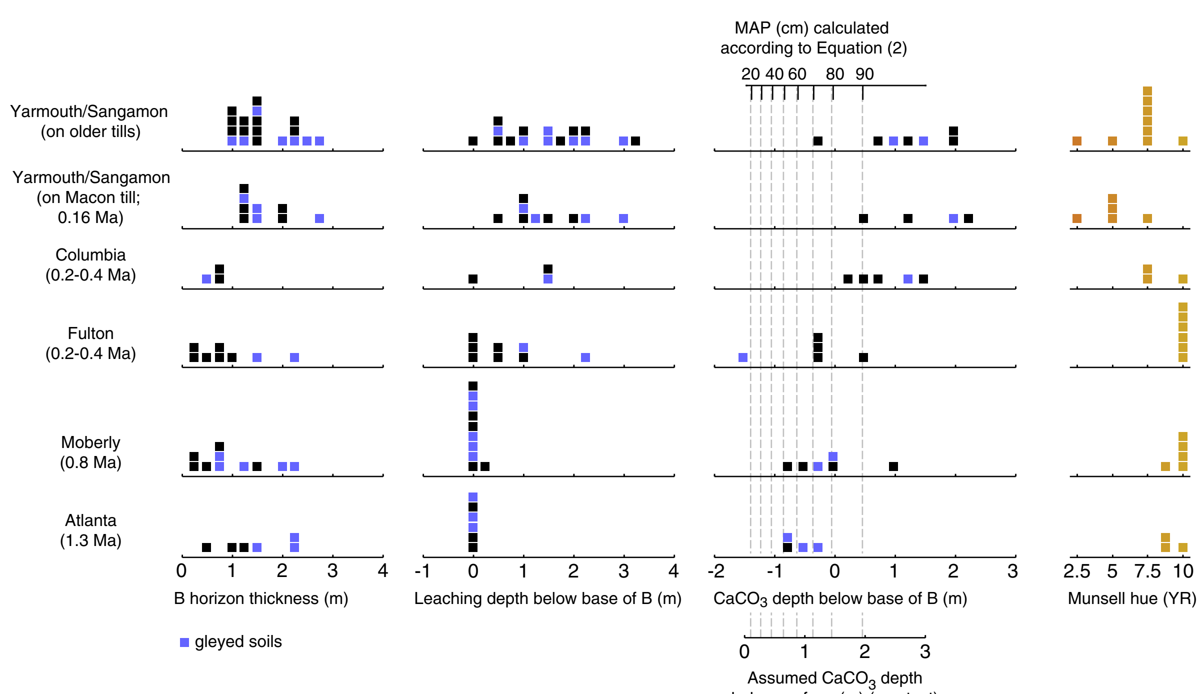


**Fig. 4.** Temporal variation in soil properties in paleosols within the Missouri glacial sequence. Source data for this figure appear in Table 2. Each symbol represents a single observation; histogram bins in all axes except Munsell color are 0.25 m. Depth of carbonate precipitation is equated with mean annual precipitation (MAP) according to Eq. (2) and the assumption of a uniform 1.5-m B horizon thickness (see text).

**Table 3**
Representative properties and burial ages for Missouri paleosols.

| Unit | Burial date (Ma) | Exposure duration (Myr) | Typical hues of profiles | Max. thickness B Hor. (m) | Median leaching depth beneath B (m) | Median nodule depth beneath B (m) |
|---|---|---|---|---|---|---|
| *Wisc. Loess* | | | | | | |
| Loveland (Loess) (Sangamon) | 0.04 | <0.1 | 5–7.5 YR | 1.2 (7) | [a] | Not present |
| Macon (Yarmouth) | 0.16 | <0.22 | 5 YR | 2.7 (7) | +1.2 (9) | +1.4 (4) |
| Columbia | 0.2–0.4 | <0.2 | 7.5 YR | 0.6 (3) | +1.3 (3) | +0.9 (4) |
| Fulton | 0.2–0.4 | 0.4–0.6 | 10 YR | 2.1 (8) | +0.3 (8) | −0.3 (5) |
| Moberly | 0.8 | 0.5 | 10 YR | 2.1 (10) | 0.0 (10) | −0.3 (5) |
| Atlanta | 1.3 | 1.1 | 7.5–10 YR | 2.1 (6) | 0.0 (6) | −0.7 (4) |
| Residuum | 2.4 | | | Not measured | | |
| Yarmouth in older tills | 0.16 | <0.35 | 7.5 YR | 2.7 (16) | +1.4 (16) | +1.2 (7) |
| Total Yarmouth | 0.16 | <0.35 | 5–7.5 YR | 2.7 (23) | +1.4 (25) | +1.2 (11) |

*Notes*: Median values for soil properties are calculated from data in Table 2. The exposure duration is the difference between the burial ages of successive units. Burial dates for the Columbia and older paleosols are from Balco and Rovey (2010); those for the Yarmouth and Sangamon Geosols are from the compilation in Mason et al. (2006). The exposure duration for the Macon (Yarmouth) paleosol is the difference between its burial age and a maximum depositional age of 0.38 Ma (see text). Limiting values in Table 2 are used in calculating medians whenever these maximum or minimum values introduce no ambiguity.
Properties of the Sangamon Geosol are not listed in Table 2; these are taken directly from field and lab notes.
Numbers in parentheses are the number of observations.
[a] The Loveland is leached to its base in all cases.

coincides with the base of the B horizon, and the top of pedogenic carbonate is consistently above that datum. In the overlying paleosols both of these depths are inversely correlated with paleosol age, which implies that younger soils formed during wetter conditions, most likely due to increased precipitation.

This trend coincides with other observations. First, authigenic gypsum within the (oldest) Atlanta paleosol indicates a dry climate; sulfates are absent from the younger paleosols. Second, pedogenic $CaCO_3$ is more common within the pre-Yarmouth profiles. Nineteen of 29 (66%) pre-Yarmouth profiles have secondary nodules within or beneath the B Horizon. Those that lack nodules within this group in most cases have thinner preserved B horizons, indicating that they may have been present higher in the profile before truncation. In contrast, just ten of 26 (38%) of the total Yarmouth and Yarmouth–Sangamon profiles preserve pedogenic $CaCO_3$. Because the Yarmouth is buried by loess (not till), subglacial truncation cannot account for the lower percentage. In some cases the Yarmouth is within several meters of the modern soil, and secondary carbonate may have dissolved during recent weathering in a moister climate. Nevertheless, the presence/absence of secondary $CaCO_3$ nodules in Yarmouth paleosols does not correlate well with their present burial depth, so modern leaching does not fully account for their relative scarcity compared to older sola.

## 5.2. Color

Hues of the oxidized B horizons are shown in Tables 2, 3 and Figs. 4, 5. The hues of the parent materials (oxidized C horizons) are approximately 10 YR within each till, except the Atlanta. The Atlanta commonly has a slight pinkish tinge imparting redder hues closer to 7.5 YR at some exposures.

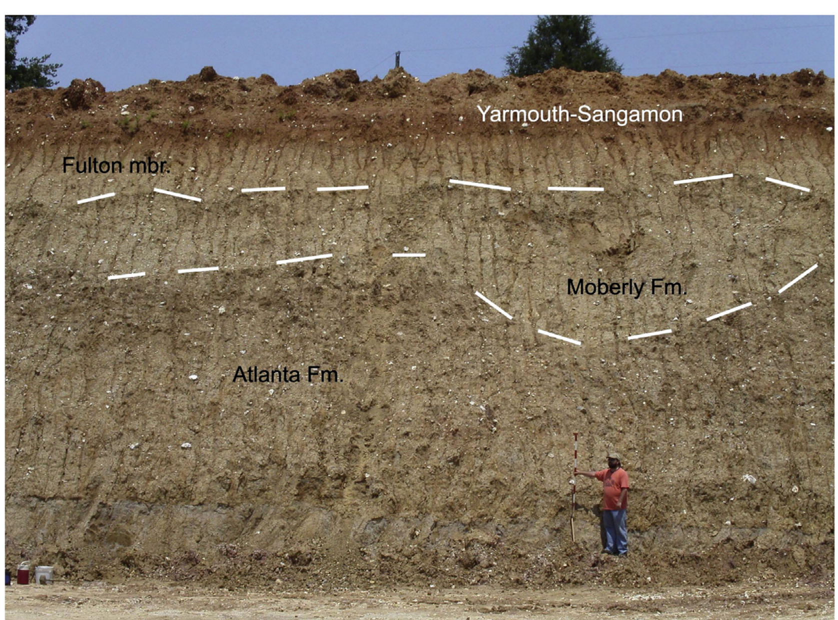


**Fig. 5.** Stratigraphic section at the Pendleton Pit (Table 1, Site 27) showing the contrast in coloration between the Yarmouth and older paleosols.

The oxidized Yarmouth and Yarmouth–Sangamon profiles (hues of about 5–7.5 YR) are consistently redder than their older counterparts (Figs. 4, 5). For example, oxidized hues within the three oldest paleosols (Atlanta, Moberly and Fulton) are approximately 10 YR in nearly all cases. Hues of the Atlanta paleosol sometimes approach 7.5 YR, but given the pinkish parent material, this does not indicate substantial rubification. The Columbia paleosol, immediately below the Yarmouth, however, seems to mark a transition to the bright red hues of the younger paleosols (Fig. 6). Two of the three oxidized profiles atop the Columbia till attain hues of 7.5 YR within Bt and BC horizons below thin Bg horizons. The redder hues may indicate warmer interglacial temperatures, or alternatively, the color may have been enhanced by the release of reduced iron from the overlying Bg horizon, which supplied additional iron coatings to the subjacent horizons upon oxidation. Additional examples of the Columbia paleosol would be useful in assessing these possibilities.

### 5.3. $^{10}$Be concentrations and inferred erosion rates

Fig. 7 shows reconstructed sample depths and $^{10}$Be concentrations in the study paleosols.

Older paleosols had significantly higher $^{10}$Be concentrations at the time of burial than younger ones, implying that surface erosion rates were lower in early Pleistocene interglaciations than in later ones. Paleosols developed on Paleozoic bedrock prior to Pleistocene glaciation that were buried at 2.4 Ma (although these paleosols are not the main subject of this paper we show their reconstructed $^{10}$Be concentrations in Fig. 7) had $^{10}$Be concentrations at the time of burial corresponding to surface erosion rates of 0.5–1.5 m Myr$^{-1}$. A similar paleosol developed on Paleozoic bedrock that was buried at 1.3 Ma implies erosion rates of 2–3 m Myr$^{-1}$. Near-surface samples from the Atlanta paleosol that were buried at 1.3 Ma imply erosion rates of 8–12 m Myr$^{-1}$, and those from the Moberly paleosol that were buried at 0.8 Ma imply erosion rates of 9–17 m Myr$^{-1}$. Near-surface samples from Fulton and Columbia paleosols imply erosion rates of 12–20 m Myr$^{-1}$ and 16–40 m Myr$^{-1}$, respectively (Fig. 7, Table 4).

As discussed above and in the supplementary data, we made numerous assumptions in constructing these erosion rate estimates and they are highly uncertain for any particular till. However, both directly measured and reconstructed $^{10}$Be concentrations in near-surface paleosol samples show a systematic order-of-magnitude decrease in $^{10}$Be concentration from older to younger paleosols, and it is difficult to construct an explanation for this observation that does not require a similar order-of-magnitude variation in erosion rates. For example, if we systematically underestimated the thickness of younger paleosols that was truncated by subglacial erosion, observed $^{10}$Be concentrations would systematically underestimate true surface concentrations, which in turn would lead to an overestimate of erosion rates. However, to explain an order-of-magnitude variation in erosion rates, this hypothesis would require that we systematically failed to recognize 2.5 m of truncation in all younger paleosols. This, in turn, is not compatible with the preservation of thick B horizons and, in some cases, portions of A horizons, in some younger paleosols. In addition, it would not be compatible with the consistency in the maximum preserved thickness of B horizons in paleosols of different ages. A more likely potential scenario would propose that younger paleosols were not exposed for a long enough time for near-surface $^{10}$Be concentrations to reach production–erosion equilibrium, in which case the $^{10}$Be concentration would reflect the exposure time of the soil rather than a steady erosion rate. Reaching production–erosion equilibrium requires enough time for approximately 1.5 m of erosion to occur; at an erosion rate of 10 m/Myr this is 0.15 Ma. For the Atlanta, Moberly, and Fulton paleosols, burial ages of overlying tills imply that they developed for 1.1, 0.5, and >0.4 Ma, respectively. Thus, it is highly unlikely that near-surface $^{10}$Be concentrations in these paleosols did not reach equilibrium with a steady erosion rate. It is possible, however, that emplacement of the Columbia till and the Macon till occurred during consecutive glacial–interglacial cycles, in which case the time available for development of the Columbia paleosol would be less than 0.1 Ma, and $^{10}$Be concentrations might not have reached production–erosion equilibrium. Thus, our steady-erosion assumption might result in an overestimation of the erosion rate for the Columbia paleosol, but not for older paleosols.

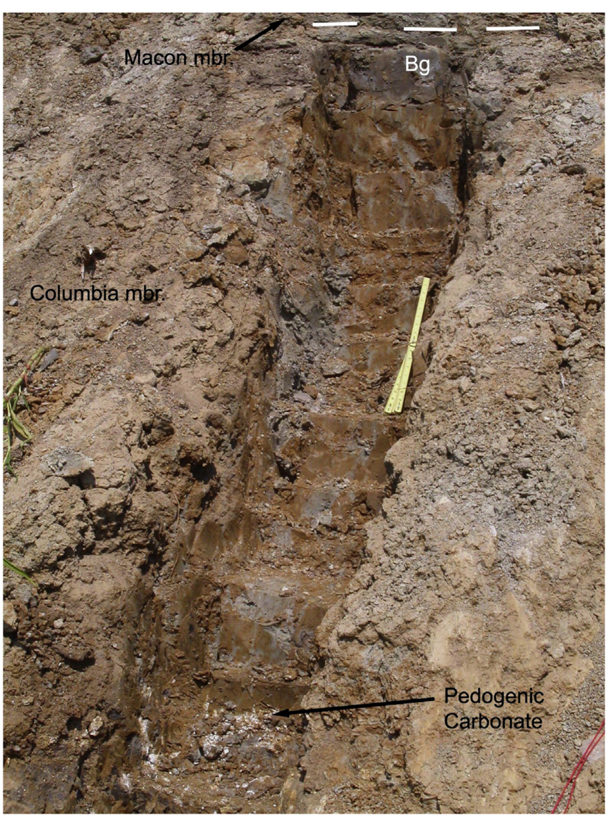


**Fig. 6.** Columbia paleosol at the Sieger Pit (Table 1, Site 16). This exposure shows a Bg horizon above a reddish brown BC horizon with reduced mottles. Nearby sections in this exposure revealed an organic-rich A horizon above the Bg.

## 6. Discussion

### 6.1. Exposure and development times

Based on the time between successive glaciations, most of these paleosols (except possibly the Columbia) formed over more than one glacial–interglacial cycle. Therefore, the soil characteristics probably reflect climate conditions averaged over some extended time prior to burial. Thus, these observations do not provide a high-resolution record of Pleistocene climate, but clearly show several general trends (Tables 3, 4, Fig. 4).

The exposure time between deposition of successive tills (the maximum possible duration of pedogenesis) generally decreases with younger age. However, all of these durations, again with the possible exception of the Columbia, are longer than the available time for Sangamon pedogenesis (<0.1 Myr). A maximum 2.1 m of B horizon is preserved atop each of the three oldest tills, despite a range in nominal exposure times from 1.1 to ~0.5 Myr. This consistency would be unlikely unless these profiles had reached a quasi-equilibrium between formation and erosion.

The maximum B horizon thickness for the Yarmouth (2.7 m; Table 2) is thicker than observed for older paleosols. This difference most likely reflects gradual burial beneath loess instead of rapid and erosive burial by till. The upper boundary of the (aerated) Yarmouth is

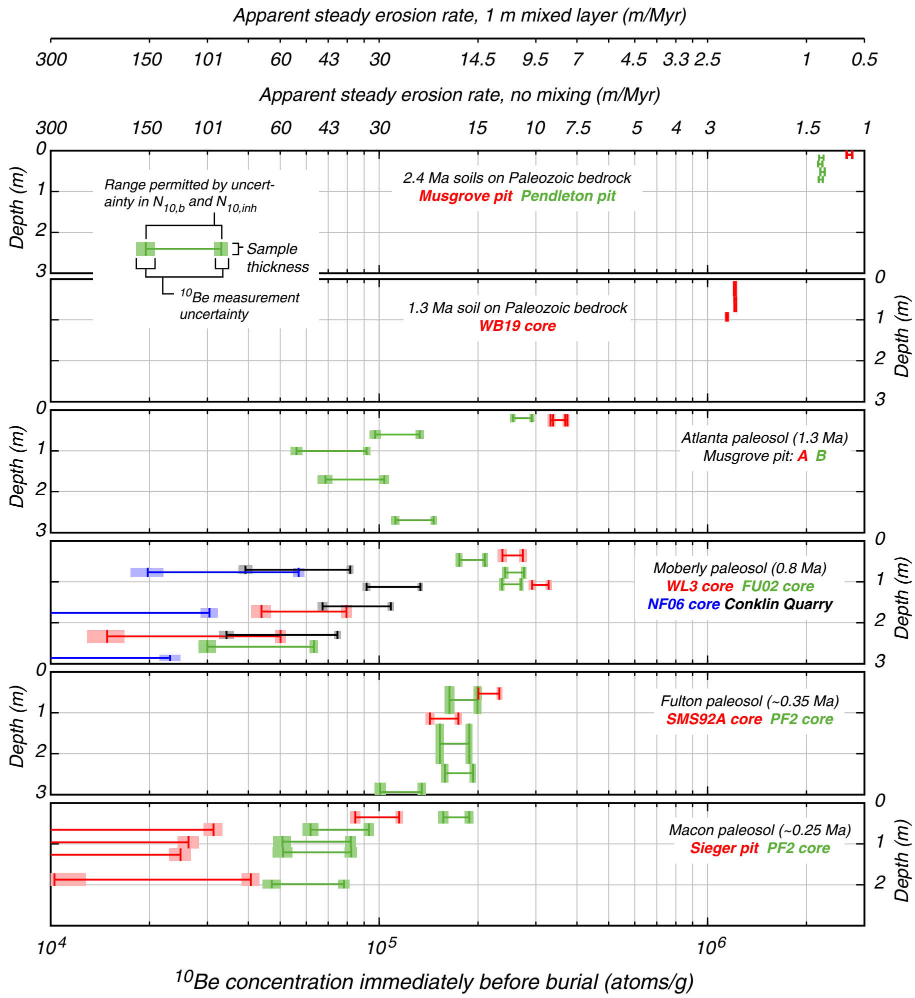


**Fig. 7.** Reconstructed $^{10}$Be concentrations at time of burial for Missouri paleosols. Color-coding distinguishes results from different sites where each paleosol is exposed. Scales at top show implied erosion rates corresponding to nuclide concentrations given the assumptions of steady erosion with complete mixing of a 1-m surface layer (uppermost scale bar) and steady erosion with no mixing (scale bar below); these are calculated for the mean elevation and latitude of the sample sites (see text and supplemental information). Because we represent shielding depths prior to burial as well as the inherited $^{10}$Be concentration in tills as a range of possible values, this implies a range of permissible reconstructed $^{10}$Be concentrations. Thus, each observation is plotted as a horizontal "barbell" with two elements. The height and width of the thick portions of each bar represent the sample thickness and the 1-sigma measurement uncertainty in the $^{10}$Be concentration, respectively. The length of the thinner line connecting the ends of the "barbell" represents the range of possible reconstructed $^{10}$Be concentrations permitted by the range of shielding depth and inherited $^{10}$Be concentration. The purpose of representing the reconstructed $^{10}$Be concentrations in this way is twofold. First, it highlights that the uncertainty in shielding depth and inherited $^{10}$Be concentrations is common to all samples in a particular paleosol, so would act to shift the entire depth-nuclide concentration array together. In contrast, measurement uncertainties in individual concentrations could affect the shape of the depth-concentration profile. Second, it highlights that for paleosols with low nuclide concentrations overall, measurement uncertainty makes a negligible contribution to the uncertainty in reconstructed $^{10}$Be concentrations relative to the uncertainty contributed by imprecise knowledge of the shielding depth and inherited $^{10}$Be concentration. Details of the calculation appear in the supplemental information.

nearly always diffuse with maximum reddening 20–40 cm below the top. Our observations are consistent with the interpretation that the Yarmouth A horizon generally grew upward into the initial loess cover while converting the former A horizon (in pre-loess sediment) into a new B horizon. Thus, the preserved B horizon as listed here likely approximates the thickness of the entire Yarmouth solum. We use this finding later to estimate a representative solum thickness for the older paleosols.

The Yarmouth Geosol is more mature than the overlying Sangamon profile, which formed over <0.1 Myr (Willman and Frye, 1970; Hall, 1999; Hall and Anderson, 2000; Grimley et al., 2003; Table 3). If the formation time for the Yarmouth paleosol, where it is buried by Loveland Silt, was at least as long as that for the Sangamon profile, this would imply that the Macon till is at least 0.25 Ma (e.g., MIS 8), and would also be consistent with a MIS 10 age.

In many cases the Yarmouth and Yarmouth–Sangamon are developed in tills older than the Macon. Conceivably, these profiles could have developed over longer durations, if the Macon had not been deposited in that area. However, this is unlikely at most locations, given that the younger tills occur in stable landscape positions throughout the study area. It seems more likely that the Yarmouth profile within the older tills represents downward migration into successively older tills as the younger ones were truncated by post-Macon erosion. Given the consistency in thickness between the Yarmouth and Yarmouth–Sangamon profile atop various tills, this again supports the idea that, in general, B horizons closely approached an equilibrium thickness within less than two complete glacial–interglacial cycles during the Middle Pleistocene.

### 6.2. Depths to carbonate

We use medians (instead of means) as the best central measure of leaching and pedogenic-carbonate depths (Table 4), because some of the observations in Table 2 are limiting values. This would occur, for example, when secondary nodules were present at the very top of a truncated profile. The median leaching and pedogenic-carbonate depths within each paleosol are strongly correlated, though not identical. Within the three oldest profiles (Atlanta, Moberly, Fulton), the top of

**Table 4**
Estimated erosion rates and mean annual precipitation (MAP) for Missouri paleosols.

| Unit | Burial date (Ma) | Apparent erosion rates (m $Myr^{-1}$) | Estimated median nodule depth (cm) | Estimated MAP cm $year^{-1}$ |
|---|---|---|---|---|
| *Wisc. Loess* | | | | |
| Loveland (Loess) (Sangamon) | 0.04 | | | |
| Macon (Yarmouth) | 0.16 | | 270 (11) | 91 |
| Columbia | 0.2–0.4 | 16–40 (2) | 240 (4) | 94 |
| | | | 180[a] (1) | 88 |
| Fulton | 0.2–0.4 | 12–20 (3) | 120 (5) | 72 |
| Moberly | 0.8 | 9–17 (3) | 120 (5) | 72 |
| Atlanta | 1.3 | 8–12 (2) | 80 (4) | 57 |
| Residuum | 2.4 | 0.5–1.5 (2) | | |

*Notes*: Calculation of apparent erosion rates from cosmogenic-nuclide measurements is described in detail in the supplementry material. The estimated median pedogenic-carbonate (nodule) depths are calculated from the relative depths in Table 3 by assuming a representative solum thickness of 150 cm. The MAP is calculated from the estimated median pedogenic-carbonate depths using Eq. (2). Numbers in parentheses are the number of measurements.

[a] This is the individual value for the Sieger Pit which preserves portions of an A horizon, estimated at 30 cm.

the pedogenic carbonate tends to be slightly above the leaching depth, which coincides closely with the base of the B horizon. In the two younger paleosols (Columbia and Macon/Yarmouth) these depths nearly coincide and are more than a meter below the base of the B horizon.

An important question in interpreting these results is the timescale on which the depth of pedogenic carbonate nodules records the climate during soil formation. In semiarid climates with high rates of dustfall (supplying $Ca^{+2}$), Stage II nodules can form in less than ~0.02 Myr (Birkeland et al., 1991). Geochemical modeling indicates that these nodules can reach an equilibrium depth with respect to MAP in even shorter times, ≤~0.01 Myr (McFadden and Tinsley, 1985). Dustfall decreases from west to east within the central United States toward this study area, so formation and equilibrium times may be longer here. However, these estimates imply that the depths of the uppermost pedogenic carbonate could record the beginning of colder and drier glacial periods shortly before burial, rather than average interglacial conditions. This seems likely for the three oldest paleosols where the top of the pedogenic carbonate is slightly above the base of the leached matrix. On the other hand, the overall consistency between leaching depths, which presumably reflects a longer integration timescale, and pedogenic-carbonate depths indicates that these observations record paleoclimate during soil formation on the time scale of one or more interglacial periods prior to paleosol burial.

The Sangamon Geosol at these sections does not preserve secondary $CaCO_3$ within its solum or directly beneath it within the Loveland Silt, which is leached to its base. However, elsewhere and just to the northwest the Sangamon commonly preserves calcic horizons (Mason et al., 2006; Mandel and Bettis, 1995). Possibly, the uppermost pedogenic carbonate within the underlying Yarmouth profile reflects accumulation during Sangamon formation, thus accounting for the greater range in depth of carbonate nodules compared to the older profiles. However, this would mean that nodules formed at depths >6 m below the top of the Sangamon at some of the sections, an unlikely event in light of the exhaustive compilations of Royer (1999) and Retallack (2005). More likely, the greater range in depth indicates a greater seasonal variation in precipitation, compared to the older profiles (Retallack, 2005).

We use Eq. (2) to equate median pedogenic-carbonate depths (Table 3) with MAP at the time of soil formation during the Early and Middle Pleistocene (Table 4; Fig. 4). Because Eq. (2) uses the land surface as the datum for pedogenic-carbonate depth, we also need to relate the base of the B horizon (the reference depth in Tables 2, 3) to the ground surface. For this we use the median thickness of the Yarmouth profile (1.5 m) as a representative value for the entire solum of each paleosol.

The (oldest) Atlanta paleosol implies the lowest recorded MAP, near 60 cm $year^{-1}$, at 1.3 Ma. Greater pedogenic-carbonate depths in the Moberly and Fulton paleosols imply MAP near 70 cm $year^{-1}$ at 0.8 and ca. 0.4 Ma, respectively. Even greater depths in the younger Columbia, Yarmouth, and Sangamon soils imply MAP exceeding 90 cm $year^{-1}$ after ca. 0.4 Ma. This value is at the upper limit of applicability of Eq. (2), and essentially equal to the threshold value near 90 cm $year^{-1}$ for pedogenic $CaCO_3$ accumulation along the eastern margin of Jenny and Leonard's transect. This is consistent with the relative scarcity of secondary $CaCO_3$ within the Yarmouth profile.

Additional factors influence the formation and depth of pedogenic calcium carbonate including vegetation, slope, dustfall and seasonality of precipitation. These factors account for the relatively low $r^2$ values in Eqs. (1) and (2); an estimate of MAP based on a single observation would have error limits exceeding 10 cm. In addition, here we have created the potential for systematic error by assuming a single solum thickness in converting the relative depths of the carbonate nodules to an absolute depth beneath the ground surface. This issue is especially relevant for the Columbia paleosol, which may be less mature (and therefore thinner) than the others. We can address this issue for the Columbia paleosol at the Sieger Pit, where a portion of an A horizon is preserved above the B, and the position of the former ground surface is relatively well constrained and indicates an original ~30 cm thickness for the A horizon. With these contraints, the pedogenic-carbonate depth at this site implies MAP of 88 cm $year^{-1}$, which is indistinguishable from the value based on the median depth (see discussion in Section 3.2). To summarize, although there are significant uncertainties in estimating the absolute value of MAP from our observations of pedogenic-carbonate depths, it is difficult to explain the observed significant and systematic variation in these depths in the Missouri paleosol sequence by anything other than a corresponding systematic change in MAP during soil formation.

### 6.3. Color

The origin of the bright red colors within the Sangamon and Yarmouth Geosols is a long-standing issue. Here we cannot address differences between these paleosols and their modern counterparts (see reviews in Curry and Baker, 2000, and Grimley et al., 2003), but we can compare them to older paleosols that developed over much longer periods of time. This shows that younger paleosols in the Missouri sequence, although they formed during shorter time periods than older paleosols, have redder sola. Therefore, we propose that the red colors of the Sangamon and Yarmouth reflect warmer interglacial temperatures beginning sometime after about 0.4 Ma. Higher precipitation could also enhance hydrolysis and accelerate accumulation of iron oxide coatings within the solum, although higher MAP would generally correlate with higher temperatures. Many proxy records of global climate also indicate warmer interglacials beginning at the same approximate time. Starting with MIS 11 (~0.42 Ma), $^{18}O$ interglacial excursions have larger amplitudes and higher reconstructed temperatures through the Sangamon interglacial (MIS 5) (e.g. Lisiecki and Raymo, 2005; Cronin, 2009). Refinement of the age of the Columbia and Macon tills would help in a more rigorous assessment of the consistency between the oceanic record and the till/paleosol record of the midcontinent.

Burial diagenesis expressed as iron oxide crystallization on ped surfaces has also been proposed for the red colors within the Sangamon Geosol (Thompson and Soukup, 1990). But, if this mechanism were responsible for the red coloration of the Sangamon and Yarmouth, why would it not have affected the older paleosols as well, which often are buried just a few meters below the Yarmouth? Moreover, the interior of peds within both the Sangamon and Yarmouth is commonly redder than the latest coatings (Mason et al., 2006; observations this study). Therefore, post burial diagenetic coatings cannot be a sufficient explanation for the distinctive red color in all cases.

Dehydration of goethite to hematite also has been proposed as a mechanism for producing reddish paleosols upon burial (Retallack, 1991), although not specifically for these paleosols. However, increased moisture content (and saturation) upon burial would not promote dehydration at the shallow depths here, which are generally less than 30 m. Moreover, this mechanism would produce redder colors within the older more deeply buried paleosols, not the opposite sequence preserved here.

Burial reduction is another possibility that is nominally consistent with the observed decrease in redness with soil age. Under this hypothesis, the Sangamon and Yarmouth retain their original color, while the older more deeply buried paleosols have been reduced to more-drab hues. Burial reduction, however, is not fully consistent with these paleosols, as there is no color trend among the three oldest and most deeply buried tills. Moreover, gleyed B horizons with no visible organic matter are present immediately below thick oxidized C horizons of overlying tills. In these cases the reduction was controlled by near-surface conditions within respective O or A horizons, which reflects the fact that pervasive gleying is difficult to accomplish within tills. Pervasive and uniform reduction requires not just water saturation, but also extreme $O_2$ depletion and a high dissolved organic-matter content to serve as the electron donor for reduction of $Fe^{+3}$ (e.g. Hallberg et al., 1978). As an example, reduction halos are present around large wood fragments within the Moberly till, but these extend just a few centimeters. These halos may have formed after the Moberly was buried, but if the reduction extends such a short distance around discrete organic materials, how could post-burial reduction have extended uniformly across entire horizons? Instead the field evidence indicates that pervasive reduction is limited to B horizons that were immediately beneath organic-rich surface horizons which provided $O_2$-deficient water and dissolved organic matter to the immediately subjacent horizon prior to burial. To summarize, burial reduction is unlikely to account for the differences in color within the pre-Illinoian paleosol sequence.

### 6.4. Erosion rates

Apparent erosion rates inferred from $^{10}$Be concentrations are higher for younger paleosols than for older paleosols. Preglacial (pre-Pleistocene to Early Pleistocene) erosion rates of 0.5–1.5 m $Myr^{-1}$ inferred from paleosols developed on bedrock residuum are low relative to recent erosion rates observed worldwide (Portenga and Bierman, 2011) and consistent with long periods of slow weathering of a stable, low-relief cratonic landscape, which is the classical understanding of preglacial topography, e.g. Anderson (1988). These rates are similar to but slightly lower than Pliocene through Pleistocene erosion rates inferred from burial dating of cave sediments elsewhere in unglaciated central North America (Granger et al., 2001). Low preglacial erosion rates here are also consistent with low erosion rates in the Mississippi drainage basin prior to 2.0 Ma inferred from Gulf of Mexico cores (Peizhen et al., 2001).

Apparent erosion rates for paleosols developed on till, rather than the preglacial landscape, are uniformly higher. Apparent erosion rates inferred from $^{10}$Be concentrations are correlated with MAP inferred from pedogenic-carbonate depths, suggesting that erosion rates during interglacials were influenced by climate, although the magnitude of variation in apparent erosion rates (a factor of 10) is disproportionate to the magnitude of the variation in MAP (a factor of 2). It appears more likely that higher erosion rates observed in paleosols formed on till may reflect the replacement of a stable landscape covered with bedrock residuum and dominated by chemical weathering with a thick layer of unweathered and unconsolidated glacial sediment. Thus, higher apparent erosion rates in younger paleosols may simply reflect more frequent resurfacing of the landscape and/or higher surface elevations relative to local drainages. The interval between glaciations was shorter for younger paleosols, so the idea that erosion rates are most likely highest directly after glaciations and decrease exponentially thereafter (e.g., Schumm and Rea, 1995), could potentially also result in a correlation between shorter intervals between glaciations and higher apparent erosion rates.

## 7. Summary

A sequence of buried paleosols within pre-Illinoian tills and younger loess formations in Missouri, USA provides information on paleoclimate during soil formation. Cosmogenic nuclide burial dates of these paleosols allow this paleoclimate information to be placed in a temporal context and provide estimates of soil erosion rates prior to burial.

Primarily, observations of leaching and carbonate accumulation depth show that modern MAP (approximately 100 cm $year^{-1}$ in the study area) is significantly higher than that during most of the Pleistocene. Many of the paleosols preserve pedogenic calcium carbonate, but it is absent in modern soils at all the sites investigated here. The depth of the uppermost pedogenic carbonate nodules within a given solum is greater, indicating higher MAP, for younger paleosols. Based on reconstructed depths of these nodules beneath the former ground surface, we estimate an increase in MAP of 60% between Early and Middle Pleistocene paleosols, although much of this increase occurs as a large difference in MAP inferred for paleosols older and younger than 0.4 Ma. Again, sola in paleosols younger than 0.4 Ma are distinctly redder than in older paleosols, even though the duration of pedogenesis is shorter for the younger soils due to more frequent glaciations. This relationship implies that higher interglacial temperatures and/or precipitation are the principal cause of the increased rubification.

Finally, reconstructed cosmogenic-nuclide concentrations are orders of magnitude higher in early Pleistocene than in late Pleistocene soils, which imply correspondingly higher erosion rates in the late Pleistocene relative to the early Pleistocene. Apparent erosion rates are near 1 m $Myr^{-1}$ for paleosols formed on preglacial residuum buried by Early Pleistocene tills and greater than 16 m $Myr^{-1}$ for late Middle Pleistocene paleosols. This variation in erosion rates is correlated with variation in MAP, but also may be related to other geomorphic factors such as more frequent landscape disturbance and resurfacing with fresh glacial sediment at higher elevations above base level.

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## Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.palaeo.2014.10.018.

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