**Stratigraphy, morphology, and geochemistry of late Quaternary buried soils
on the High Plains of southwestern Kansas, USA.**

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Abstract

This study investigated two stratigraphic sequences that record the complex interplay of sedimentation and pedogenesis over the past ca. 84 ka on the High Plains of southwestern Kansas. Up to eight eolian sand and loess units with associated soils were identified in two cores collected from an upland setting. Soil morphological and geochemical data were used to quantitatively assess and compare soil development between multiple buried soils. Chronostratigraphic relationships indicate that 1) loess and eolian sands were episodically deposited during late Marine Isotope Stage (MIS) 5 (ca. 84–70 ka), 2) the Sangamon soil at this site formed between MIS 3/4 (ca. 70–52 ka), which is consistent with age estimates for renewed Sangamon pedogenesis in the Mississippi River valley, and 3) soil development in the Gilman Canyon Formation began at ca. 44 ka and continued until at least 29.2 ka. Multiple lines of evidence, including grain-size distributions, structure, clay content, and chemical weathering index data, indicate that the morphology and chemistry of the Ak horizon developed in the Gilman Canyon

Formation is a product of cumulization by slow loess additions. Similar evidence suggests that cumulization processes are also responsible for the morphology and chemistry of other buried soils in the stratigraphic sequence. Overall, loess inputs during pedogenesis complicate the quantification of weathering processes in these soils.

Key words

Great Plains, loess, paleosols, cumulization, Gilman Canyon Formation, Sangamon soil

1. Introduction

Many investigations of Quaternary loess-paleosol sequences, particularly in Europe and China, have utilized chemical weathering indices to investigate weathering intensity and paleoclimatic change (e.g., Yang et al., 2004; Jeong et al., 2008; Buggle et al., 2011; Schatz et al., 2015). In the Midwestern USA, the application of weathering indices has mostly been directed to glacial loess and associated intercalated paleosols (e.g., Muhs et al., 2001; Grimley et al., 2003). In the central Great Plains of North America, however, where loess deposits are non-glacial (e.g., Aleinikoff et al., 2008), weathering indices have typically been employed as an indicator of sediment provenance (e.g., Muhs et al., 1999, 2008; Jacobs and Mason, 2007). In addition to chemical weathering, pedogenic processes also result in the formation and accumulation of secondary minerals (e.g., clay, Fe oxides, and CaCO_3),

which are commonly reflected in the morphological properties of the soil. Relatively few sediment-paleosol studies, however, have attempted to quantitatively compare morphological development with the intensity of chemical weathering. Together, such data allow for a more complete assessment of 1) the relative contributions of pedogenic and sedimentary processes, 2) the relative morphological differences among sequences of buried soils, and 3) regional differences and comparisons.

In this paper we investigate two sediment-paleosol sequences that record soil formation and the episodic deposition of loess and eolian sands during the late Quaternary on the High Plains of southwestern Kansas. A 98 m-deep core (HP1A) was collected from an upland setting approximately 25 km south of the Arkansas River (Fig. 1) as part of the High Plains Ogallala Drilling Program at the Kansas Geological Survey. Here, we present the stratigraphy, geochronology, soil morphology, and geochemistry of the upper 14 m of this core. The stratigraphy, numerical ages, and soil morphology from the upper 20 m of a similar core (CMC) collected from an upland setting just north of the Cimarron River valley (Fig. 1) are also presented. The HP1A and CMC cores provide ca. 77 and 84 ka records, respectively, of eolian sedimentation and soil formation on the High Plains surface. By correlating eolian units with regionally extensive loess units and quantifying the morphological development and weathering intensity of buried soils, this study provides a better understanding of late-Quaternary landscape evolution as well as the complex interplay between eolian sedimentation and pedogenesis in the central Great Plains of North America.

2. Background

2.1. Study area

The study area is located in southwestern Kansas and is part of the High Plains region of the Great Plains physiographic province (Fig. 1; Fenneman, 1931). The High Plains region represents the remnant of a vast alluvial plain (Ogallala Formation) formed by sediments deposited by streams discharging from the Rocky Mountains during the late Miocene-early Pliocene epochs. The High Plains surface is relatively flat and featureless and is mantled by less than 5 m of loess in southwestern Kansas (Welch and Hale, 1987; Bettis et al., 2003); however, river valleys, dune fields, and numerous small playa basins provide some topographic relief (5–15 m).

Southwestern Kansas is dominated by short grass prairie and has a semiarid continental climate. Mean annual temperature and precipitation are 12.5°C and 490 mm, respectively, for the period of record (AD 1947–2015) at Garden City, Kansas (High Plains Regional Climate Center, 2015). Most precipitation falls during the early summer due to frontal activity where cold, dry polar air masses collide with warm, moist air masses from the Gulf of Mexico. Also, periodic intensification of the North American Monsoon brings peak summer rainfall to southwestern Kansas (e.g., Higgins et al., 1997). Droughts are common in the region due to the development of

upper level anticyclones, which inhibit frontal activity and are driven primarily by sea surface temperature anomalies in the tropical Pacific Ocean (e.g., Cook et al., 2007; Hoerling et al., 2014).

2.2. Late-Quaternary Stratigraphy

The late-Quaternary stratigraphic framework for the Great Plains includes at least three superimposed loess units: the Loveland Loess, Gilman Canyon Formation, and Peoria Loess (e.g., Feng et al., 1994; Bettis et al., 2003). These loess units are regional in extent and thus provide a means to stratigraphically correlate more localized units.

The Loveland Loess typically is the oldest loess unit in the Great Plains and usually is only a few meters thick (Bettis et al., 2003). Studies suggest that the Loveland Loess was deposited during the penultimate glaciation (Marine Isotope Stage (MIS) 6) (e.g., Forman et al., 1992; Maat and Johnson, 1996; Forman and Pierson, 2002), although there is evidence of intermittent deposition throughout MIS 5 in some locations (e.g., Rodbell et al., 1997; Markewich et al., 2011). The Loveland Loess is commonly recognized by the well developed, rubified soil formed in its upper part. This soil, referred to as the Sangamon Geosol, has been documented in the Great Plains (e.g., Frye and Leonard, 1952; Schaetzl, 1986; Feng et al., 1994) and throughout the Midwestern USA (e.g., Willman and Frye, 1970; Ruhe, 1974; Follmer, 1978; Jacobs and Knox, 1994; Grimley et al., 2003). The Sangamon Geosol typically

is 1–2 m thick with silt loam or silty clay loam textures, 7.5YR to 5YR hues, thick Bt horizons, and well-developed prismatic or angular blocky structure. Based on studies in the Mississippi River valley, Sangamon pedogenesis occurred between ca. 130–90 ka (most of MIS 5) with renewed pedogenesis from ca. 74–54 ka (MIS 4) (Markewich et al., 2011).

The Gilman Canyon Formation (GCF) mantles the Sangamon Geosol and is usually < 2 m thick on the Great Plains (Bettis et al., 2003). The GCF is the stratigraphic equivalent of the Pisgah Formation in western Iowa (Bettis, 1990) and the Roxana Silt of the Mississippi River valley (Follmer, 1983; Leigh and Knox, 1993). The GCF is typically noncalcareous with dark brown to grayish brown coloration and silt loam textures. The GCF has often been modified by pedogenesis, commonly with two or more soils (e.g., Reed and Dreeszen, 1965; Mandel and Bettis, 1995; Johnson et al., 2007). In Kansas and Nebraska, the GCF has yielded calibrated ¹⁴C ages ranging from ca. 45 to 24 ka (Johnson et al., 2007; Muhs et al., 2008).

The Peoria Loess overlies the GCF and is the thickest and most areally extensive loess deposit in the Great Plains (Bettis et al., 2003). This loess unit typically is calcareous, massive, light yellowish tan to brown in color, and has silt loam textures. Peoria Loess accumulated in the Great Plains between ca. 25–11 ka (Bettis et al., 2003; Mason et al., 2007). The Brady Geosol, developed in the Peoria Loess but only distinguishable from the modern soil when buried by the Holocene Bignell Loess, typically is a dark gray to grayish brown soil with a well-developed B horizon

(Johnson and Willey, 2000). Calibrated¹⁴C ages indicate that the Brady Geosol formed from ca. 13.8 to 8.8 ka (Johnson and Willey, 2000; Mason et al., 2003) and was buried by the Bignell Loess beginning ca. 9.9–8.8 ka, with continued loess deposition spanning much of the Holocene in some locations (Mason et al., 2003).

3. Methods

Continuous 6.5-cm diameter cores were collected with an Acker hollow-stem auger and wireline, split-spoon sampler. Soils and sediments were described in detail using standard terminology outlined in Birkeland(1999). Buried soils were numbered consecutively from the top of the core downward, with the number following the suffix “b” (e.g., Akb1, Btkb2, Btkb3, etc.), following Holliday (2004, p. 339). Soils were also, however, placed in a stratigraphic framework by numbering them consecutively, beginning with 1, the modern surface soil, at the top of the core (e.g., soil 1, soil 2, etc.). Soilmorphological properties (color paling and lightening, texture, structure, dry consistence, clay films, and secondary carbonate)were used to calculate normalized horizon index (HI) values, which enable a quantitative comparison of the degree of soil development (Harden, 1982). Also, Profile Development Index (PDI) values were determined by summing the HI values multiplied by the respective horizon thicknessesfor each soil profile.

Soils were sampled by horizon for particle size. Where horizon thicknesses were > 50 cm, subsamples were collected so that the sample interval was no greater than

50 cm. Particle-size analyses were performed using the pipette method (Soil Survey Staff, 1982). Bulk geochemical samples (~10 g each) were collected at 10–20 cm intervals from the HP1A core and were ground and analyzed by ALS Minerals using four-acid digestion and Inductively-Coupled Plasma Mass-Spectrometry (ICP-MS). Analyses included measurements of the following major elements: Al, Ca, Fe, K, Mg, Mn, Na, and Ti.

Chemical weathering during pedogenesis results in changes to the elemental composition of the original parent material through the selective removal of mobile elements from the soil relative to the more immobile elements (e.g., Middelburg et al., 1988; Buggle et al., 2011). Geochemically based weathering indices, which utilize the ratio of select mobile to immobile elements, were therefore used to assess the intensity of chemical weathering. Such indices have been widely applied over a range of soil parent materials and time scales (see reviews by Sheldon and Tabor, 2009; Buggle et al., 2011). However, determining the most appropriate index can be problematic. The most common mobile elements used in weathering indices are Ca, K, Mg, and Na. Ca and Mg may not reflect the true magnitude of weathering in a soil, however, owing to the formation of secondary pedogenic carbonates, which may mask primary silicate weathering (Buggle et al., 2011). This fact is particularly relevant for the soils investigated in this study. In terms of immobile elements, Al and Ti are generally used.

For this study we selected the classic “Chemical Index of Alteration” (CIA; Nesbitt and Young, 1982) and the “Chemical Proxy of Alteration” (CPA; Buggle et al., 2011). The CIA provides a measure of feldspar weathering by relating molar proportions of Ca, K and Na to Al:

$$\text{CIA} = [\text{Al}_2\text{O}_3 / (\text{Al}_2\text{O}_3 + \text{CaO} + \text{K}_2\text{O} + \text{Na}_2\text{O})] \times 100 \quad (1)$$

One of the complications with this index, however, is the determination of silicate Ca (CaO). Given the presence of secondary carbonates in our buried soils, we corrected CaO for apatite using P_2O_5 values, following McLennan (1993). Where corrected CaO values were greater than Na_2O , CaO was assumed to equal Na_2O , as Ca typically is lost more rapidly than Na during weathering. Percentages of other elemental oxides were rescaled after CaO correction.

Given the uncertainties in determining CaO, we also employed the CPA index, which simply relates molar proportions of Na to Al and also provides a measure of feldspar, particularly plagioclase, weathering (Buggle et al., 2011):

$$\text{CPA} = [\text{Al}_2\text{O}_3 / (\text{Al}_2\text{O}_3 + \text{Na}_2\text{O})] \times 100 \quad (2)$$

The CPA is advantageous for this study because it overcomes complications associated with the presence of secondary carbonates in the buried soils. In order to accurately determine elemental oxide content, CaO was excluded

for the CPA analysis and the percentages of other elemental oxides were rescaled. Relating Al to Na is also favored because these elements form a homogenous mineralogical weathering system where both the immobile (Al) and mobile (Na) elements are hosted by the same mineral (e.g., albite; $\text{NaAlSi}_3\text{O}_8$), which minimizes the effects of variations in parent material composition (Buggle et al., 2011).

In order to compare normalized HI values calculated from morphological properties with chemical weathering index values, CIA and CPA data were also normalized by horizon with respect to the parent material (using mean values for each horizon; normalized values are denoted as ΔCIA and ΔCPA). ΔCIA and ΔCPA values were also multiplied by horizon thickness and summed (denoted as ΔCIA_t and ΔCPA_t) to assess overall profile weathering and facilitate comparison with PDI values.

Numerical age control was provided by optically stimulated luminescence (OSL) dating, which provides an age estimate of the time since sediment was last exposed to light (e.g., Wintle, 2008). OSL samples were collected from core and analyzed at the Utah State University Luminescence Laboratory using the latest single-aliquot regenerative-dose (SAR) procedures for dating quartz sand (see Murray and Wintle, 2000). All samples were opened and processed under dim amber safelight conditions and were processed with HCl and chlorine bleach to remove carbonates and organics, respectively. Heavy liquid (sodium polytungstate, 2.7 g/cm^3) separation was undertaken to isolate the fine-grained quartz sand fraction and

hydrofluoric acid was used to etch the quartz and dissolve any remaining feldspars. Sample processing followed standard procedures outlined in Aitken (1998) and described in Rittenour et al. (2005). OSL ages were calculated using the central age model (Galbraith and Roberts, 2012).

In addition to OSL ages, we also radiocarbon-dated two soil organic matter samples from the CMC core. ^{14}C samples were sent to the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) facility at the Woods Hole Oceanographic Institute. Samples underwent standard pretreatment, including the removal of roots and decalcification, and were combusted in a high-temperature muffle furnace. The resultant CO_2 was reacted with a Fe catalyst to create graphite and ^{14}C activity was measured via Accelerator Mass Spectrometry. All ^{14}C ages in the text are reported as median calibrated ages (ka) to facilitate comparisons with OSL ages.

4. Results and discussion

4.1. Stratigraphy, chronology, and correlation

The HP1A core contains six eolian units and five buried soils whereas the CMC core contains eight eolian units and seven buried soils (Fig. 2). Our numerical age results allow us to chronostratigraphically correlate these local units with regionally extensive loess units. OSL ages from the bottom of the cores (ca. 76–84 ka) indicate

that units 4 and 5 from HP1A and units 5–8 from CMC are coeval with the late Loveland Loess (Fig. 2; Table 1). Markewich et al. (2011) suggest that although much of the Loveland Loess was deposited during MIS 6 in the Mississippi River valley, loess accumulation continued intermittently through late MIS 5 (up to ca. 74 ka). Although the Loveland Loess has been interpreted as essentially a single depositional unit in the Missouri River valley (e.g., Mason et al., 2007), multiple carbonate-rich soils have been identified in this loess deposit in the Great Plains (e.g., Feng et al., 1994), indicating multiple periods of loess deposition and soil formation. The OSL ages reported in this study, together with the presence of multiple buried soils, indicate episodic eolian deposition during late MIS 5 on the High Plains of southwestern Kansas.

Investigations from the Mississippi River valley indicate that the latest period of Sangamon soil development occurred from ca. 74–54 ka (Markewich et al., 2011). Based on our chronostratigraphic results, it appears that soil 3 most likely represents this latest period of Sangamon pedogenesis in the study area. Two OSL ages from the HP1A core constrain the depositional age of unit 3 to between ca. 70.3 and 54.0 ka (Fig. 2; Table 1) and suggest that sediment deposition and pedogenesis were concurrent during MIS 4 and early MIS 3. The basal OSL date from unit 2 in the CMC core suggests that soil 3 was buried at ca. 52 ka. It should be noted that whereas Markewich et al. (2011) refer to the soil that formed from ca. 74–54 ka as the “Sangamon paleosol”, other authors working in Kansas have informally termed this soil as a “reddish soil” or “reddish pedocomplex” distinct from the Sangamon

soil (e.g., Feng et al., 1994; Karlstrom et al., 2008). The basal age of unit 3 from the HP1A core is similar to the basal age of this “reddish soil” (ca. 69 ka) at the Barton Landfill in central Kansas (Feng et al., 1994).

OSL and ^{14}C ages from unit 2 in the HP1A and CMC cores range from 52.2 to 29.2 ka, indicating that this unit chronostratigraphically correlates to the GCF (Fig. 2; Table 1). In Kansas and Nebraska, the GCF has yielded calibrated ^{14}C ages ranging from about 45 to 24 ka (Johnson et al., 2007; Muhs et al., 2008), although the basal age of the unit is poorly constrained. Johnson et al. (2007) infer a basal age of 55–50 ka for the GCF. The correlative Roxana Silthas yielded basal OSL ages of ca. 50–64 ka in the Mississippi River valley (Forman and Pierson, 2002). The ca. 44 ka OSL age from the HP1A core provides a limiting age for the onset of soil development in this unit. Layzell et al., (2015) present calibrated ^{14}C ages of 48–28 ka for a similar soil developed in alluvium in the Cimarron River valley.

The surface soil (soil 1) is developed in Peoria Loess. No age control was obtained for unit 1; however, ages from the top of the GCF provide maximum-limiting ages for the interval of Peoria Loess deposition. The youngest reported age for the GCF is 23.9 ± 1.3 ka ($19,770 \pm 590$ ^{14}C yr. B.P.) from a locality in south central Nebraska (Souders and Kuzila, 1990) whereas the youngest age for the GCF in Kansas is 24.7 ± 0.8 ka ($20,510 \pm 310$ ^{14}C yr. B.P.) (Mandel, 2002).

The numerical chronology presented herein, together with regional data, allows us to also assess the approximate duration of pedogenesis for soils 1–5 in the HP1A core. Although uncertainties in the OSL ages (15–18% for all samples) preclude a precise assessment, taking these uncertainties into account for bounding C horizon ages provides the maximum possible duration for a given soil-forming interval. For example, soil 4 has bounding ages of 76.1 ± 13.5 and 70.3 ± 10.6 ka, which indicates that pedogenesis could have occurred over a period of up to 30 ka (89.6–59.7 ka). Similarly, bounding C horizon ages for soil 5 (76.8 ± 13.1 and 76.1 ± 13.5 ka) indicate that pedogenesis could have occurred over a period of up to 27 ka. For soil 3 we use bounding ages of 70.3 ± 10.6 and 52.2 ± 9.6 ka from the HP1A and CMC cores, respectively, which gives a maximum period of soil formation of 38 ka. The youngest reported age for the GCF of 23.9 ± 1.3 ka and the 44.3 ± 7.8 ka OSL age from the HP1A core were used for soil 2 and provide a maximum duration for soil formation of 33 ka. No age control was obtained for soil 1, but given that the Brady Geosol began to form in the Peoria Loess at ca. 13.8 ka (e.g., Johnson and Willey, 2000), we assume that the maximum amount of time for the formation of the surface soil is ~14 k years.

4.2. Soil morphology

Soils 3–6 and 3–8 in the HP1A and CMC cores, respectively, are interpreted to have formed in eolian deposits that are coeval with late phases of Loveland Loess aggradation. The majority of these soils are developed in sandy loam or loamy

sandsediments; however,soil 5in HP1A developed in loess with a silty loam texture (Tables 2 and 3). The relatively coarse-grained deposits are interpreted as eolian sands derived either from the Arkansas or Cimarron River valleys or from local reworking of the alluvial deposits of the Ogallala Formation.

In the HP1A core, soils 3–6are well developed andall havemorphological features that are typical of Sangamon pedogenesis, including 7.5 YR hues, 0.7–1.1 m-thick Btk horizons, angular blocky structure with well expressed clay films, and 18 to 32% clay in the Btk horizons (Table 2; Fig. 3). These soils lack obviousA horizons, which have either been eroded before burial or oxidized after burial. Soils 3 and 6 have the highest HI values (0.61 and 0.78)(Table 4). Soil 3, however, has the highest PDI value (69) because of its greater thickness. Although soil 5 has similar morphological properties to the other soils of Loveland age, it has lower HI (0.35) and PDI values (31) due to the lack of morphological change relative to the parent material (loess).

In the CMC core, soils 3, 5, and 7 are the most developed soils of Loveland age and display morphological features similar to soils 3–6 in the HP1A core, including 7.5 YR hues, thick Btk horizons, angular blocky structure with distinct clay films, and up to 40% clay in the Btk horizons (Table 3; Fig. 4).Soils 3, 5, and 7, however, have thicker Btk horizons (1.4–3 m) than soils 3–6 in HP1A and therefore higher PDI values (98, 78, and 73) (Table 4). Soil 4 has similar morphological properties to soils 3, 5, and 7 but has a thinner Btk horizon and therefore a lower PDI value (21). Soils

3 and 5 have the highest HI values (0.48 and 0.52). Soils 6 and 8 are less developed with 0.4–0.55 m-thick Bk horizons, weak to moderate subangular blocky structure, no clay films, and only 11–17% clay (Table 4; Fig. 4).

In both cores, soil 2 is interpreted as representing a soil developed in the GCF. At the type section in southwestern Nebraska, the GCF has multiple soils due to varying rates of loess accumulation (Johnson et al., 2007). In southwestern Kansas, however, we find one soil, suggesting a more distal location relative to the loess source. Soil 2 in both cores is relatively well developed with a ~1.8 m-thick profile, Ak-Bk-Ck horizonation, and blocky structure (Tables 2 and 3). The darker matrix colors in the Ak horizons compared to the Bk horizons are probably a product of organic matter additions during pedogenesis, which is commonly found in GCF soils (see Mandel and Bettis, 1995; Johnson et al., 2007). In the HP1A core, texture fines upward from sandy loam to loam through the C horizons and from clay loam to silty clay loam in the soil solum (Table 2). We interpret the upward-fining sequence through the C horizons as evidence for increased loess deposition relative to more locally derived eolian sands as indicated by the upward increase in silt (Fig. 3). Clay contents are relatively high in the Ak horizons (27–36%) of both soils and are similar to the overlying unmodified Peoria Loess (Figs. 3 and 4). We therefore interpret the high clay content in the Ak horizons as resulting from continual inputs of fine dust during pedogenesis. In this sense, the Ak horizons are products of cumulation (see Birkeland, 1999 p. 165; Schaetzl and Thompson, 2015 p. 436) and therefore are more stratigraphic artifacts due to increased dust influx than products of

pedogenesis (see Jacobs and Mason, 2005, 2007). An influx of fine-grained material, particularly clay, would also explain the well-developed blocky structure evident in the Ak horizons (see Mason and Nater, 1994; Birkeland, 1999). Furthermore, if dust inputs were rich in CaCO_3 then carbonate accumulation in the Ak horizon of soil 2 may also be related to cumulation processes. It is also possible, however, that secondary carbonate accumulated after the soil was buried. The longer a soil remains buried, the more likely it will be affected by secondary enrichment from above (Birkeland, 1999 p. 27–28). Hence, through time, as water moved from overlying carbonate-rich sediments into soil 2, calcium carbonate precipitated in the Ak horizon.

The Ak and Bk horizons of soil 2 have HI values between 0.29 and 0.39 and the soil has relatively high PDI values of 63 and 80 (Table 4). It should be noted that the HI values for the Ak horizons might be artificially high given that some morphological characteristics used to construct the indices (i.e. texture and structure) are very likely related to sedimentary rather than pedogenic processes. Also, the high PDI values likely result from ongoing loess accumulation, which would increase the overall thickness of the soil profile.

The surface soils in both cores are moderately developed silty clay loams with A-Bk-Ck horizonation formed in the upper 1.1 to 1.2 m of Peoria Loess (Tables 2 and 3). This soil has relatively low PDI values (19 and 25), indicating relatively weak development. Clay contents (23 to 34%) are similar to soil 2 (Figs. 3 and 4). Beneath

the solum the loess is a calcareous, massive, light yellowish brown colored silty clay loam. The loess has higher clay contents than one would typically expect for Peoria Loess, which probably indicates a location distal to loess sources and therefore reflects the deposition of fine far-traveled dust (see Mason and Jacobs, 1998; Jacobs and Mason, 2007).

4.3. Chemical weathering indices

Chemical weathering indices were calculated for the HP1A core (Fig. 5). The CIA and CPA (equations 1 and 2) display similar depth trends throughout the stratigraphic sequence; however, the CIA displays more variability in most of the parent materials (e.g., unit 4) compared to the CPA and we attribute this to uncertainty in CaO correction. Parent material variability is minimized with the CPA not only due to the exclusion of CaO but also because index elements are hosted by the same mineral group. We therefore assume that the CPA provides a more reliable assessment of chemical weathering in this setting.

The highest weathering index values are in soils 5, 2, and 1, suggesting more intense weathering during these periods of soil formation. Soil parent material index values, however, differ between the eolian units, highlighting the importance of normalizing index values respective to the parent material. For example, the C horizons of units 1, 2, and 5 (loess) have much higher index values than units 3, 4, and 6 (eolian sands), which indicates that the loess parent materials already contained highly weathered

material at the onset of pedogenesis. Also, loess parent materials appear to be more homogenous in terms of their elemental composition than the eolian sands. Both indices show erratic shifts in soil 3, which likely reflect variability in original sediment composition. Both indices also show increased values through the C horizons of unit 2, which supports the previous inference from grain-size data of increased loess deposition at the expense of eolian sands through the unit.

Normalizing chemical weathering index values for the parent materials (ΔCIA and ΔCPA) indicates that soils 4 and 6 are actually the most chemically weathered (Table 5). Soil 2 has relatively high ΔCIA values but low ΔCPA values. However, if soil profile thicknesses are taken into account (ΔCIA_t and ΔCPA_t), then soil 2 appears to be the most chemically weathered (Table 5). The surface soil (soil 1) has the lowest chemical index values. The low ΔCPA value in the A horizon may reflect either recent deposition of unaltered loess (note that the chemical index values for the A horizon are similar to the Peoria Loess) or land use practices where the addition of fertilizer and other soil amendments may have altered base element concentrations.

4.4. Comparison of morphological and chemical weathering indices

Figure 6 compares soil morphological and chemical weathering index data for the HP1A core. Relationships between the morphological indices and the CPA (i.e., HI vs ΔCPA and PDI vs ΔCPA_t) are statistically significant at the 0.01 α level and have higher coefficients of determination (R^2), providing further evidence that the CPA

more appropriately represents the magnitude of weathering in this setting. Comparing HI and ΔCPA values indicates that soils 3, 4, and 6, which formed in eolian sands of Loveland age, are the most developed (Tables 4 and 5). Comparing PDI and ΔCPA_t values, however, which take into account profile thickness, indicates that soils 2, 3, and 6 are the most developed.

In soil 2, both the HI and the ΔCPA values are higher for the Ak horizon than the Bk horizon. This result implies that not only are additions of fine-grained material to the A horizon contributing to its morphological development as previously discussed but also may be imparting a higher chemical weathering signature. As previously noted, chemical index data show that loess typically has higher index values than eolian sand. Therefore, if rates of loess deposition continually increased at the expense of eolian sands during GCF time, then designating the Ck1 horizon, which contains up to 30% sand, as the parent material for soil 2 likely resulted in an overestimation of both morphological development and weathering intensity. Alternatively, loess incorporated into the Ak horizon of soil 2 could have a different source and therefore have a different chemical composition to that of the Ck1 horizon. Furthermore, if profile thicknesses are a product of loess inputs rather than pedogenesis then PDI and ΔCPA_t values will also provide an overestimation of overall soil development.

The multiple lines of evidence for ongoing loess accumulation during the formation of soil 2 raises the question as to whether this process was also operating during

other periods of soil formation. In the HP1A core, soils 3 and 6 show a relatively large increase in silt from the C to the Btk horizons (increases of 34 and 20 %, respectively) (Fig. 3). Similarly in the CMC core, soils 3 and 5 show increases of 50 and 37%, respectively (Fig. 4). Furthermore, examination of the silt-sized fraction (Figs. 3 and 4) indicates that the dominant fraction is coarse silt (50–20 μ m) in most soils, particularly in the HP1A core. Also, its relatively uniform distribution in many of the soil profiles suggests a sedimentological rather than pedogenic origin. Loess contributions during pedogenesis would also help explain some of the variability observed in the chemical weathering indices (e.g., soil 3) (Fig. 5).

Considering the effect of time on soil development also may shed some light on potential loess inputs. Schaetzl et al. (1994) recommend that chronofunctions employ mathematical models that best fit the data (i.e., have the highest coefficient of determination) but also be based on a theoretical understanding of pedogenesis at a given site. For this study, the PDI and ΔCPA_t data best fit an exponential function and a power function, respectively (Fig. 7). These functions, however, suggest that rates of soil development increase over time, which is contrary to steady state models of pedogenesis as well as a myriad of chronofunctions that typically show the opposite trend (see Schaetzl and Thompson, 2015). Although this trend can be explained by changing climatic conditions (i.e., that the most developed soils formed under warmer, wetter climates) it can also be explained by significant loess additions that may have 1) contributed to morphological development and imparted a higher chemical signature, 2) created thicker soil profiles due to cumulation

processes, or 3) resulted in an overestimation of soil development in soils for which the sandy loam C horizons were designated as the parent material. The effect of loess inputs on soil development cannot be quantified, however, without information on the composition of loess inputs and the deposition rate.

5. Summary and conclusions

This study investigated two late-Quaternary stratigraphic sequences that provide a complex pedosedimentary record from the High Plains of southwestern Kansas. Chronostratigraphic relationships indicate that 1) loess and eolian sands were episodically deposited between ca. 84–70 ka, 2) the Sangamon soil formed between ca. 70–52 ka, which is consistent with age estimates for late Sangamon pedogenesis in the Mississippi River valley, and 3) the Gilman Canyon Formation pedocomplex began forming at ca. 44 ka and continued until at least 29.2 ka, which is consistent with other regional chronologies.

Of the different weathering indices used in this study, the Chemical Proxy of Alteration (CPA), which simply relates Na to Al, provided the best metric for assessing chemical weathering intensity. Geochemical results highlight the importance of accounting for soil parent material composition when utilizing weathering indices as loess displayed higher index values compared to eolian sands. Quantitative comparisons between morphological development and chemical weathering intensity indicate statistically significant relationships with high

coefficients of determination, implying that postdepositional pedogenesis has appreciably affected both soil morphology and chemistry. However, it also appears that ongoing loess additions during soil formation significantly affected soil morphological and chemical properties. Multiple lines of evidence, including grain-size distributions, soil morphology and chemical weathering index values, suggest that the Ak horizon of the Gilman Canyon Formation is a product of cumulation processes due to increased loess influx. Also, cumulation likely played a role in the morphological and chemical development of other soils in the stratigraphic sequence. This finding limits the usefulness of soil development indices and chronofunctions that are based on soil profile thicknesses and illustrate the pedosedimentary complexity associated with soils in the central Great Plains. Buried soils in this setting and other settings that may have had significant loess inputs should therefore be interpreted in light of the interactions between sedimentary and pedogenic processes.

Acknowledgements

This paper benefited from constructive comments by an anonymous reviewer as well as comments from Missy Eppes who reviewed an earlier version of the manuscript. This research was funded from multiple sources including the Geological Society of America, the Society for Sedimentary Geology (SEPM), the National Science Foundation (NSF EAR-1023285), and the US Geological Survey (STATEMAP program). We thank Bridget Sanderson and Kevin Fenyak for assisting with particle-size analyses.

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Figure titles

Figure 1. A) Regional map showing study area location on the High Plains of southwestern Kansas. B) Lidar image showing location of the HP1A and CMC cores.

703

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708

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716

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719

720 Figure 6. Plots of morphological indices versus chemical weathering indices. HI =
721 Horizon Index, PDI = Profile Development Index, ΔCIA and ΔCPA = chemical index
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724

725 Figure 7. Plot showing changes in profile index values over time. See section 4.1 for
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Figure captions

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Table 1. Age information

OSL ages¹

Core	Lab no.	Depth (m)	No. of aliquots ²	Dose Rate (Gy/ka)	De ³ ± 2σ (Gy)	OD ⁴ (%)	OSL age ± 2σ (ka)
HP1A	USU-1110	3.9	29 (46)	3.39 ± 0.16	150.08 ± 21.79	37.2 ± 5.4	44.3 ± 7.8
	USU-1111	6.9	20 (46)	3.47 ± 0.19	187.57 ± 24.24	27.5 ± 4.8	54.0 ± 8.9
	USU-1112	7.5	29 (55)	2.94 ± 0.15	206.90 ± 23.13	26.4 ± 4.4	70.3 ± 10.6
	USU-1113	8.2	30 (54)	3.06 ± 0.14	231.38 ± 25.89	27.5 ± 4.4	75.7 ± 11.3
	USU-1114	9.3	24 (65)	3.16 ± 0.16	240.11 ± 35.20	33.5 ± 5.5	76.1 ± 13.5
	USU-1115	11.9	26 (59)	2.78 ± 0.14	213.76 ± 29.55	31.9 ± 5.4	76.8 ± 13.1
CMC	USU-1518	4.4	8 (55)	2.79 ± 0.14	145.36 ± 22.24	17.2 ± 6.5	52.2 ± 9.6
	USU-1519	11.7	15 (45)	2.66 ± 0.17	200.83 ± 39.44	36.1 ± 7.3	75.5 ± 16.9
	USU-1569	18.65	13 (44)	2.78 ± 0.17	233.27 ± 42.69 ⁵	33.6 ± 7.2	84.0 ± 17.9

Radiocarbon ages

Core	Lab no.	Depth (m)	¹⁴ C age (yr B.P.)	Calibrated age range ⁶ (yr B.P.)	Median calibrated age (ka)	δ ¹³ C
CMC	OS-112461	2.8	24,900±840	27,600-30,760	29.2±1.6	-17.4
	OS-112462	3.2	30,300±1700	31,070-38,170	34.6±3.6	-15.9

¹ OSL age analysis using the single-aliquot regenerative-dose procedure of Murray and Wintle (2000) on 1-mm small-aliquots of quartz sand.

² Number of aliquots used in age calculation and number of aliquots analyzed in parentheses.

³ Equivalent dose (De) calculated using the Central Age Model of Galbraith et al. (1999), except where noted otherwise.

⁴ Overdispersion (OD) represents variance in De data beyond measurement uncertainties, OD >20% may indicate significant scatter due to depositional or post-depositional processes.

⁵ De calculated using arithmetic mean.

⁶ Calibration to calendar years (2 sigma) was performed with CALIB 7.0 using calibration dataset IntCal13 (Reimer et al. 2013).

Table 2. Soil descriptions from HP1A core

Soil #	Horizon	Depth (m)	Matrix Color	Texture ^a	Structure ^b	Consistence ^c	Clay Films ^d	Boundary ^e	Notes
1	A	0-0.4	10YR 4/2	SiL	2 f sbk	so	---	c	Parts to 1 f gr
	Bk1	0.4-0.75	10YR 5/2	SiCL	2 m sbk	h	---	g	Few carbonate nodules, common fine threads
	Ck	0.75-2.0	10YR 6/4	SiCL	m	h	---	g	Very few carbonate nodules, few fine threads
2	Akb1	2.0-3.0	10YR 4/4	SiCL	3 m abk	h	---	g	Very few carbonate nodules, few fine threads
	Bkb1	3.0-3.8	10YR 5/4	CL	3 m abk	h	---	g	Common carbonate nodules, few fine threads
	Ck1b1	3.8-4.3	10YR 6/4	L	m	sh	---	g	Common carbonate nodules
	Ck2b1	4.3-5.1	10YR 6/6	L	m	so	---	c	Parts to sg; Few carbonate nodules
	Ck3b1	5.1-6.08	7.5YR 6/6	SL	sg	lo	---	a	Very few carbonate nodules
3	Btkb2	6.08-7.2	7.5YR 5/4	L	3 m abk	h	2 d po	g	Common carbonate nodules
	Cb2	7.2-7.96	7.5 YR 6/6	SL	sg	lo	---	a	
4	Btkb3	7.96-8.7	7.5YR 6/4	SL	3 m abk	h	2 d pf, po	c	Common carbonate nodules
	Ckb3	8.7-10.51	7.5YR 6/6	SL	sg	lo	---	a	Very few carbonate nodules
5	Btkb4	10.51-11.4	7.5YR 5/4	SiCL	2 m abk	h	2 p po	c	Few carbonate nodules
	Ckb4	11.4-12.0	7.5YR 6/4	SiL	m	sh	---	a	Very few carbonate nodules
6	Btkb5	12.0-12.8	7.5YR 7/3	L	3 m abk	vh	2 d pf, po	c	Many coalescing carbonate nodules
	Cb5	12.8-14.0	7.5YR 6/6	SL	sg	lo	---	---	

^a Texture: SiL–Silt Loam, SiCL–Silty Clay Loam, CL–Clay Loam, L–Loam, SL–Sandy Loam

^b Structure: 1–weak, 2–moderate, 3–strong; f–fine, m–medium, c–coarse; abk–angular blocks, sbk–subangular blocks, gr–granular, m–massive, sg–single grain

^c Consistence: lo–loose, so–soft, sh–slightly hard, h–hard, vh–very hard

^d Clay Films: 1–few, 2–common, 3–many; d–distinct, p–prominent; pf–ped faces, po–pores

^e Boundaries: a–abrupt, c–clear, g–gradual

Table 3. Soil descriptions from the CMCcore

Soil #	Horizon	Depth (m)	Matrix Color	Texture ^a	Structure ^b	Consistence ^c	Clay Films ^d	Boundary ^e	Notes
1	A1	0-0.2	10YR 4/2	L	2 m gr	so	---	c	
	A2	0.2-0.6	10YR 4/3	SiCL	1 m sbk	so	---	c	Parts to 2 m gr
	Bk	0.6-1.1	10YR 5/3	SiCL	2 m sbk	h	---	g	Very few carbonate nodules, few fine threads
	Ck	1.1-1.95	10YR 6/4	SiCL	m	h	---	c	Few carbonate nodules, few fine threads
2	Akb1	1.95-3.2	10YR 4/3	SiCL	2 m abk	sh	---	g	Few carbonate nodules, common fine threads
	Bkb1	3.2-3.8	10YR6/4	L	3 m sbk	h	---	g	Common carbonate nodules and threads
	Ckb1	3.8-4.6	7.5YR 6/4	SL	sg	lo	---	a	Very few carbonate nodules
3	Btk1b2	4.6-5.4	7.5YR 5/4	CL	2 m abk	h	2 d pf	g	Many coalescing carbonate nodules
	Btk2b2	5.4-6.9	7.5YR 6/4	SL	2 m abk	h	2 d pf	g	Common carbonate nodules
	Cb2	6.9-7.29	7.5YR 5/6	SL	sg	lo	---	a	
4	Btkb3	7.29-7.8	7.5YR 6/3	CL	2 m abk	h	2 d pf	g	Many coalescing carbonate nodules
	Ckb3	7.8-8.5	7.5YR 5/6	L	m	sh	---	a	Few carbonate nodules
5	Btk1b4	8.5-9.4	7.5 YR 6/4	SiCL	3 m abk	vh	3 d pf	g	Many coalescing carbonate nodules
	Btk2b4	9.4-9.9	7.5YR 6/4	SiL	2 m abk	sh	2 d pf	g	Common carbonate nodules
	BCb4	9.9-10.28	7.5YR 6/4	L	1 f sbk	sh	---	g	Parts to 2 m gr
	Cb4	10.28-12.45	7.5YR 5/6	SL	sg	lo	---	c	
6	Bkb5	12.45-12.9	7.5YR 6/4	L	2 m sbk	sh	---	g	Common fine carbonate threads
	Cb5	12.9-13.42	7.5YR 6/6	LS	sg	lo	---	a	
7	Btk1b6	13.42-13.8	7.5YR 6/6	L	2 f sbk	sh	1 d pf	g	Few carbonate nodules, common fine threads
	Btk2b6	13.8-14.6	7.5YR 6/6	SL	2 m abk	h	1 d pf	g	Few carbonate nodules, few fine threads
	Btk3b6	14.6-16.4	7.5YR 6/4	L	2 f sbk	h	1 d pf	c	Few carbonate nodules, common fine threads
	Cb6	16.4-16.7	7.5YR 6/4	SL	m	so	---	a	Parts to sg
8	Bkb7	16.7-17.1	7.5YR 6/3	SL	1 f sbk	sh	---	c	Few carbonate nodules and fine threads
	C1b7	17.1-17.9	7.5YR 6/3	LS	m	so	---	c	Parts to sg
	C2b7	17.1-19.59	7.5YR 5/6	LS	sg	lo	---	a	

^aTexture: SiL–Silt Loam, SiCL–Silty Clay Loam, CL–Clay Loam, L–Loam, SL–Sandy Loam

^bStructure: 1–weak, 2–moderate, 3–strong; f–fine, m–medium, c–coarse; abk–angular blocks, sbk–subangular blocks, gr–granular, m–massive, sg–single grain

^cConsistence: lo–loose, so–soft, sh–slightly hard, h–hard, vh–very hard

^dClay Films: 1–few, 2–common, 3–many; d–distinct, p–prominent; pf–ped faces, po–pores

^eBoundaries: a–abrupt, c–clear, g–gradual

Table 4. Calculated morphological index values

Core	Soil No.	Horizon	HI	PDI
HP1A	1	A	0.23	19
		Bk1	0.29	
	2	Ak1b1	0.39	63
		Bkb2	0.29	
	3	Btkb2	0.61	69
	4	Btkb3	0.55	42
	5	Btkb4	0.35	31
CMC	1	Btkb5	0.78	62
	2	A1	0.24	25
		A2	0.19	
		Bk	0.26	
	3	Akb1	0.33	80
		Bkb1	0.33	
	4	Btk1b2	0.48	98
		Btk2b2	0.39	
	5	Btkb3	0.38	21
		Btk1b4	0.52	
		Btk2b4	0.45	
	6	BCb4	0.21	78
		Bkb5	0.25	
		Btk1b6	0.21	
	7	Btk2b6	0.25	73
		Btk3b6	0.25	
	8	Bkb7	0.1	4

HI = Horizon Index; PDI = Profile Development Index.

Table 5. Calculated chemical index values from the HP1A core

Core	Soil No.	Horizon	ΔCIA	ΔCPA	ΔCIA_t	ΔCPA_t
HP1A	1	A	0.68	0.0	95	37
		Bk1	1.95	1.10		
	2	Ak1b1	4.39	1.50	662	245
		Bkb2	2.79	1.19		
	3	Btkb2	1.76	1.96	197	219
	4	Btkb3	3.32	2.34	252	178
	5	Btkb4	1.49	1.26	134	113
	6	Btkb5	4.14	2.61	248	209

ΔCIA and ΔCPA = chemical index values by horizon normalized for soil parent material;
 ΔCIA_t and ΔCPA_t = chemical index values by soil profile normalized for soil parent material and profile thickness. Note that the Ck1 horizon was utilized as the parent material for soil 2.

Figure 1

Figure 2

Figure 3

Figure 5

Figure 6