RECONSTRUCTING CLIMATE ON THE GREAT PLAINS FROM BURIED SOILS: A QUANTITATIVE APPROACH

By

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M.A., University of Kansas, 2008

Submitted to the graduate degree program in Geography and the Graduate Faculty of the University of Kansas in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

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Date approved: April 23, 2013

ABSTRACT

The Great Plains, U.S.A. lack quantitative paleoclimatic data for the late Quaternary largely because two common sources of paleoclimatic data, tree ring and pollen records, are rare in the region. Sequences of buried soils, however, are commonly preserved in eolian and alluvial sediments on the Great Plains and have the potential to enhance the region's paleoclimate record. This research presents a study of buried soils preserved in the Caddo Canyons of central Oklahoma to highlight opportunities and considerations for using buried soils to reconstruct past climate. Results indicate that sequences of buried soils dating to the mid- and late-Holocene are commonly preserved in the canyons. Canyon geomorphology dictates the nature of the fill contained in the canyons, and the effect of geomorphology and microclimate on soil formation must be carefully considered when interpreting the buried soil and stable carbon isotope record from the canyons.

In an effort to capitalize on the rich paleoenvironmental record that buried soils can provide, this study presents the Buried Soil Reconstruction Model (BuSCR), a method for reconstructing paleoclimate based on properties of buried soils. The model was developed based on a study of modern analogue soils and climate on the Great Plains. BuSCR reconstructs mean annual precipitation (MAP), moisture index (I_m), and mean annual temperature (MAT) with statistically significant results ($r^2 = 0.4$, $p < .0001$) and low mean average errors. While error increases on the edges of the Great Plains climate envelope, application of BuSCR to a series of buried soils across the Great Plains, including soils from the Caddo Canyons, shows that it corroborates both paleoenvironmental reconstructions using other proxies (e.g. dune activation histories) and model-simulated hindcasts. In particular, BuSCR reconstructions corroborate model simulations of a -25% MAP anomaly during the Medieval Warm Period and a drastic reduction in MAP and I_m across the Great Plains during the Altithermal. These results indicate that the BuSCR model, with further testing and if applied widely to buried soils across the Great Plains, could provide quantitative reconstructions of past climate that fill a current hole in the North American paleoclimate database.

ACKNOWLEDGEMENTS

The work presented here would not have been possible without the help of many others. First, the members of my dissertation committee, Dr. Johan Feddema, Dr. Rolfe Mandel, Dr. Terry Slocum, Dr. Sharon Billings, Dr. Jack Hofman, and Dr. Joy Ward, provided valuable critiques, asked tough questions, and ushered me through to completion. I thank them for their time and attention to this project. My advisors, Dr. Johan Feddema and Dr. Rolfe Mandel, in particular, spent numerous hours working with me on my research. Their perspectives were invaluable and greatly influenced the direction of the project. I thank them for their dedication to me and my dissertation research. Also, I am indebted to the KU Geography Department and numerous graduate students in Geography and other departments across campus for their assistance and support in completing the Ph.D.

Numerous landowners in Caddo and Blaine Counties, Oklahoma provided access to their land to conduct reconnaissance surveys of the canyons. Their hospitality was much appreciated. In particular, Dean Smith assisted me in contacting land owners and getting oriented to the area. Jimmy Wells and Mike Propps allowed me repeated access to their property to describe and sample the cut-bank exposures in Farra Canyon and Armstrong Creek. Aaron Benson, Celeste McCoy, Carmen Sigler, and Russell Zung assisted me in the field. Also, Dr. Rolfe Mandel accompanied me to Oklahoma and helped describe and sample Farra 2. The study of the canyons would not have been possible without the assistance of the above-mentioned land-owners, friends, and colleagues.

The NSF IGERT C-CHANGE program at KU provided generous financial support during my graduate career to cover the cost or research, tuition, and living expenses. I am grateful for the financial support as well as the personal and professional opportunities the C-CHANGE program provided. In particular, I want to thank Dr. Joane Nagel and members of the C-CHANGE steering committee for giving me the chance to be part of a program that shaped who I am as a scientist, my research interests, and my future career path.

Additional research funding was provided by Dr. Rolfe Mandel, a Student Research Grant from the Geological Society of America, grants from the Osage Chapter of the Association of Women Geoscientists, and the KU Department of Geography Kollmorgen Scholarship. The laboratory analyses conducted for this project would not have been possible without generous financial support from the aforementioned sources. Much of the laboratory analysis was completed at the Kansas Geological Survey Geoarchaeology and Paleoenvironmental Research Laboratory, where Laura Murphy and Bridget Sanderson provided valuable assistance and training needed to run numerous soil characterization analyses. Dr. Hong Wang, Illinois State Geological Survey, conducted the radiocarbon analyses and Greg Kane, KU Peck Paleoenvironmental and Environmental Stable Isotope, provided access to the laboratory and training for the stable carbon isotope analyses.

Lastly, I want to thank my friends and family who supported and encouraged me throughout this process. In particular, Ferdouz Cochran, Diana Franklin, Trish Jackson, Laura Murphy, and Ariana Wall Postlethwait were always there to listen, celebrate, and urge me on to completion. Also, my parents, Steve and Sonya Bonner, provide endless encouragement and support. I am grateful they were able to share this journey with me. Most importantly, my husband, Russell Zung, and my son, Mather – you two are my rock. I am so grateful to come home to you every day and could never have done this without remembering I was doing it for us. I love you.

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INTRODUCTION

The purpose of my dissertation research is to quantify modern soil-climate relationships on the Great Plains in order to develop a model for reconstructing past climates based on buried soils. Chapter 1 presents a study describing the buried soil record preserved in two canyons in central Oklahoma. The Holocene paleoenvironmental history of the region was reconstructed based on soil stratigraphy and stable carbon isotope analysis of soil organic matter from the buried soils and alluvial deposits in the canyons. This analysis provided an on-the-ground study of the opportunities and challenges in reconstructing past climate based on buried soil characteristics, and the findings greatly informed development of the model presented in Chapter 2. Chapter 2 reviews the data mining and statistical methods employed to develop the Buried Soil Climate Reconstruction (BuSCR) model, and presents the results of testing the model on 35 modern soils across the Great Plains. In Chapter 3, the BuSCR model was applied to 17 buried soils across the Great Plains, and the results were compared to current understanding of the paleoenvironmental history of the region. This served as an excellent test for the predictability of the model and its applicability in the buried soil environment.

Over the last 100 years, soil scientists have proposed numerous models for explaining and attempting to predict soil pedogenic pathways (see Schaetzl and Anderson 2005; Minasny et al. 2008 for reviews). Few realize that observations of soil-climate relationships, first in Russia and then in North America, prompted Hans Jenny to propose one of the first and, probably most influential models of soil formation - the fundamental equation of soil forming factors. Further, it can be argued that publication of Jenny's influential book, *Factors of Soil Formation: A system* *of quantitative pedology*, spearheaded the sub-field of quantitative pedology (Minasny et al. 2008). The BuSCR model was developed with Hans Jenny's state factors approach as a framework, and Jenny's work was a great resource and inspiration in development of the model. To conclude the introduction to my dissertation research, I've provided a brief history of the early efforts to quantify soil-climate relationships that led to the birth of quantitative pedology.

By 1900, soil science was developing as a discipline separate from agriculture and geology. Nearing the turn of the century, pochvovedenie (translated as pedology or soil science) developed and flourished in Russia, under the leadership of V.V. Dokuchaev (Tandarich and Sprecher 1994). Around the same time, the Division of Soils (later the Bureau of Soils) in the U.S. Department of Agriculture, with Milton Whitney at its helm, began surveying American soils (Cline 1977). Approaches to soil classification and mapping in the two countries differed considerably. Influenced by field observations of changes in soil character along the steep climate gradient of the Russian Steppe, the Russian soil scheme classified and conceptualized soils in terms of environmental factors of soil formation, with special emphasis on climate. In contrast, impacted by observations of soils in mountainous regions of California, U.S. soil scientists mapped and classified soils based on geographic province, stressing the importance of geology. Regardless of these contrasts, scientists worldwide were increasingly aware that soils differed from the rock in which they formed and were much more than simply a medium for plant growth.

Amid growing interest in the study of soils as independent natural bodies, initiation of the journals *Soil Science* and *Pedology* in 1916 connected soil scientists from around the world. The First International Congress of Soil Science met in Washington, D.C. in 1927. The congress brought together scientists from 34 countries for a ten-day meeting with the intent of discussing and agreeing on "uniform methods of procedure in the handling of soil problems of like character in all the countries represented, with the aim of eventually effecting a correlation of the soils of the whole world" (Lipman 1928). The meeting also integrated American soil scientists into the international disciplinary fold for the first time (De'Sigmond 1935).

The highlight of the congress was a transcontinental excursion that facilitated discussion and debate on issues related to soil genesis, classification, mapping, and soils as a resource for agriculture and industry. C.F. Marbut (Bureau of Chemistry and Soils, Department of Agriculture) proposed the transcontinental excursion and led the 235-person party, of which approximately half were foreign delegates and half were American (McCall 1928), on a one-month trip that crossed the U.S. and Canada.

In 1917, Marbut had read the German translation of a text by Russian soil scientist K.D. Glinka, *Die Typen der Bodenbildung* (Tandarich and Sprecher 1994). This was his first exposure to the Russian soil scheme with its emphasis on climate. Marbut translated the text into English in 1927, making it widely accessible to American soil scientists for the first time. Stimulated by exposure to the Russian scheme, Marbut used the transcontinental excursion as an opportunity to take the "old-fashioned American soil classification" system and "throw it overboard," focusing on the soil-climate relationship instead (Jenny 1989). While these ideas were welcomed and familiar to the European scientists in attendance, he met resistance from the Americans. Hans Jenny was a young post-graduate who had trained at the Swiss Federal Institute of Technology under George Wiegner when he attended the congress and the transcontinental field trip. Jenny remembered the resistance Marbut encountered on the field trip when he emphasized the influence of climate on soils (Jenny 1989):

> "Marbut was in a difficult position. On one side was what he thought about the soil, and on the other side was what the government thought about the soil. He was confronted with politics. There was this big soil survey scheme and these soil survey people were not going to change their thinking right away... the older people who were entrenched and had administrative positions were not keen about it [Marbut's emphasis on the Russian scheme]. So there was this conflict which appeared on the trip quite often between Marbut and [A.G.] McCall. McCall was in charge of all of the soils work [by U.S. Department of Agriculture], and I guess he was Marbut's boss. He was very conservative, and we had the feeling that he opposed Marbut."

Unbeknownst to the Americans on the excursion (Jenny 1989), an American soil scientist by the name of E.W. Hilgard had already outlined independently a set of ideas very similar to the Russian concept of soil formation (Hilgard 1892, 1906; Jenny 1961; Cline 1977). His 1906 text, *Soils: their formation, composition, and relations to climate and plant growth in humid and arid regions*, presented numerous examples of the influence of climate on soils, as well as a soil profile concept. While it appears that his soil profile concept was used by the Bureau of Soils, Milton Whitney largely discouraged use of Hilgard's publications by soil surveyors (Jenny 1961). Hilgard criticized Whitney's theories of soil genesis, which held that geology, clay constituents, and soil moisture largely determined soil formation and fertility (Cline 1977), and this feud may have been motive for Whitney to keep Hilgard's publications from those working in the Bureau. Marbut's introduction of the climatic theory of soil formation on the transcontinental excursion, therefore, was novel to the Americans and is to this day largely attributed to him, Glinka and Dokuchaev.

While McCall and Marbut disagreed on the influence of climate on soils, the soils featured on the excursion across the expansive North American continent left a significant impression on Hans Jenny (Jenny 1989):

> "I was impressed when we went from Washington south and saw the southern red soils and a few weeks later the black soils in Canada. There must be a relationship, I thought, but in Europe the landscape is all cut up… The red soils of the South and the black soils of Canada were showcases of the climatic theory of soil formation… I was intrigued by the 1,000 mile-long smooth transition of the Great Plains…There the rolling plains, I fancied, must harbor the secret of mathematical soil functions. At times, I could hardly sleep thinking about it."

On the excursion, Robert Bradfield offered Jenny use of his lab at the University of Missouri while he was on sabbatical for one year, which Jenny eagerly accepted. He arrived in Missouri soon after and immediately began to research the relationship between climate and soil nitrogen, initially, and later organic matter and clay content. Jenny (1989) spent long hours traveling the Midwest conducting field work, "longing to meet this idol of a climatic soil profile [a Russian Chernozem - a soil rich in humus with calcium carbonate accumulation in the subsoil]." He eventually amassed a substantial collection of soil samples that he analyzed in the laboratory and derived a three-dimensional nitrogen-climate surface relating soil nitrogen to moisture and temperature. Four articles published between 1929-1935 in the journal *Soil Science* testify to Jenny's focus on defining the soil-climate relationship mathematically (Jenny 1929, 1930, 1931, 1935). Jenny presented maps, scatter plots, and mathematical equations in all four articles to make a case for quantifying soil properties in terms of climatic factors. His 1935 *Soil Science* article, "The clay content of the soil as related to climatic factors, particularly temperature," included one of the earliest presentations of a version of his later famous state-factor equation of soil formation.

Hans Jenny's research contrasted greatly with the research of others in soil science at the time. For example, in the 1930 volume of *Soil Science* an article titled, "A holder for soil sample bags," with an accompanying illustration was published opposite Jenny's article titled, "The nitrogen content of the soil as related to the precipitation-evaporation ratio," where he presented a differential equation to describe the logarithmic relationship between nitrogen content and a humidity factor. While this is an extreme example of the chasm, articles in *Soil Science* and *Soil Science Society of America Proceedings* published between 1930-1935, were largely devoid of maps or scatterplots, and mathematical functions were almost completely absent. Most questions addressed by the discipline at that time were descriptive in nature, recording the occurrence of soils and describing their properties, with no effort to explain the mechanisms that created said soil. By 1945, scatterplots were common, as researchers frequently published studies on the "effect" of a treatment or factor on soil properties or plant growth. However, very little mathematics, with the exception of Pearson's correlation coefficients, was used to study these effects until at least the mid-1950s.

In 1936, Hans Jenny accepted a position at the University of California, Berkeley teaching pedology and colloid science (Amundson et al. 1994), where he stayed for the remainder of his career. Jenny's published research during the late 1930s centered on colloids and cation exchange in soils. In 1941, however, Jenny published *Factors of Soil Formation. A system of quantitative pedology*, which introduced a grand theory of soil formation. This theory described soil as an open, physical system defined by the five soil forming factors: climate, organisms, topography (relief), parent material, and time. In *Factors,* Jenny presented the fundamental equation of soil forming factors:

$$S = f(cl, o, r, p, t, ...)$$

Where S is the soil, cl is climate, o is organisms, r is relief or topography, p is parent material, and t is time. With publication of *Factors of Soil Formation*, Jenny tentatively introduced quantitative pedology to soil science. Of specific importance was the following assertion from Jenny's book (1941): "The fundamental equation of soil formation is of little value unless it is solved... It is the purpose of the present book to assemble known correlations between soil properties and soil-forming factors and, as far as is possible, to express them as quantitative relationships or functions." To solve the equation, Jenny proposed using soil functions where one factor varies while the other four remain constant. For example, in climofunctions climate varies but relief, parent material, biology, and time are constant across the sample sites (Jenny 1941, 1946).

Jenny's now famous equation has been called "elegant" and his book "eloquent" and a "classic in the field" (Wysocki et al. 2000; Brady and Weil 2005; Brevik and Hartemink 2010). Further, it is considered today to be potentially the most influential theory in the history of soil science, called the "granddaddy" of theories of soil formation (Schaetzl and Anderson 2005). It is the basis for most models in environmental science (Hoosbeek et al. 2000), and most research in pedology begins with an explanation of the soil forming factors (Birkeland 1999).

According to Crocker (1932), within ten years, *Factors of Soil Formation* "had a tremendous effect in clarifying and directing American pedological thinking ... [as it was] well illustrated by a comparison of ... recent ... soil survey literature ... with that of ten or fifteen years ago." As early as 1943, ecologists, soil scientists, and geomorphologists published articles supporting Jenny's theory of soil formation, and specifically the influence of climate on soil properties (e.g., Bryan and Albritton 1943; Barshad 1946; Aandahl 1948; Muckenhirn et al. 1949). In later years, however, numerous scientists challenged Jenny's approach (e.g., Crocker 1952; Stevens and Walker 1970; Dijkerman 1974; Huggett 1975; Chesworth 1976), giving rise to several competing theories of soil genesis. Process-oriented models developed by Simonson (1959) and Johnson and Watson-Stegner (1987), mechanistic models introduced by Kline (1973) and Huggett (1975), and energy budget models such as Runge's (1973), have greatly impacted the field of soil science; yet the ideas introduced by Jenny in *Factors* remains a cornerstone of both our understanding of soil formation and the way we conduct soil science today.

Textbooks and journal articles today frequently refer to Jenny's equation as "unsolvable" (see Brevik and Hartemink 2010; Schaetzl and Anderson 2005; Wysocki et al. 2000). Nevertheless, the use of chrono-, litho-, climo-, bio-, and toposequences over the last sixty-plus years have provided quantitative answers to numerous questions about the relationship of soil properties to environmental factors. Also, while some soil scientists have criticized Jenny's quantitative approach, his students Rod Arkley and Jennifer Harden, developed the leaching index and profile development index, respectively, two frequently cited quantitative methods for relating soil properties to the factors of soil formation (Arkley 1961; Harden 1982).

In recent years, soil scientists have once again become interested in quantifying soil formation, and some have suggested that Jenny's equation is now solvable (Bryant et al. 1994; Hoosbeek et al. 2000; Scull et al. 2003). Many of the challenges posed to the theory in the mid-20th century can now be overcome with the use of computers, detailed global soil databases, and new statistical techniques. In the face of global climate change, solving Jenny's equation, especially with reference to climate as a soil-forming factor, is crucial. Questions regarding global carbon cycling, soil fertility, and food production in light of global warming and exponential population growth may be answered by returning to efforts to solve Jenny's "elegant" equation.

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GEOMORPHOLOGY, SOIL STRATIGRAPHY, AND HOLOCENE PALEOENVIRONMENTS OF THE CADDO CANYONS, CENTRAL OKLAHOMA, U.S.A.

1. Introduction

The Caddo Canyons in the Southern Rolling Plains of Oklahoma hold an archive of late-Quaternary environmental change. The numerous canyons, cut during the Pleistocene, contain alluvial fills with multiple buried soils that date to the late Pleistocene and mid-late Holocene (Albritton 1966; Ferring 1982; Hall and Lintz 1984; Hofman 1988, 1990), as well as buried tree stumps and other fossil plant and animal remains. The soil-forming periods represented by the buried soils appear to be regionally synchronous. This synchronicity suggests that alluvial soil formation in the canyons responded to region-wide environmental factors, rather than local conditions, and that the regional late-Quaternary climate may be reconstructed based on properties of the buried soils (Ferring 1990; Hall 1990; Holliday 1997). Alluvial soil stratigraphy and stable carbon isotope analysis of soil organic matter preserved in buried paleosols are commonly used to reconstruct paleoenvironments in the North American Great Plains (e.g., Humphrey and Ferring 1994; Fredlund and Tieszen 1997; Nordt 2001; Cordova et al. 2011; Murphy and Mandel 2012) and are well-suited for developing a late-Quaternary paleoenvironmental history for the Southern Rolling Plains based on the record preserved in the canyons.

The Caddo Canyons have been most intensively studied by archaeologists, inspired by the 1961 discovery of a Clovis mammoth kill site in Domebo Canyon (Albritton 1966; Leonhardy 1966). Because many archaeological sites have been recorded in the canyons in the years since the discovery of the Domebo site, these geomorphic settings have captured the attention of archaeologists (e.g., Ferring 1982; Lintz and Hall 1983; Hofman 1988, 1990). While studies of the canyons have primarily focused on the archaeological record, a few also have documented soil-stratigraphic sequences (e.g., Nials 1977; Pheasant 1982; Lintz and Hall 1983). However, these studies have examined only a single canyon, and none have focused on clues the soil stratigraphy provides about the environmental history of the region. Also, no studies have analyzed stable carbon isotopes of pedogenic organic matter preserved in the canyons' Holocene valley fills.

The Holocene paleoclimatic history of the Southern Rolling Plains of North America remains unclear due to poor preservation of traditional paleoclimatic indicators (e.g., pollen, plant macro-fossils) and a paucity of paleoenvironmental studies in the region (Hall 1990; Hall and Valastro 1995; Baker et al. 2000; Holliday 2000). Paleoenvironmental reconstructions elucidate the potential effects of climate change on the region, such as the effect of prolonged droughts, providing context for policymakers developing climate change mitigation strategies (Birks and Birks 1980; Bradley 1985; National Research Council 2006). The best examples of a prolonged drought in the historical record are the Dust Bowls on the Central and Southern Plains, U.S. during the 1930s and 1950s. While historians often cite modification of the landscape by humans as the cause of the Dust Bowl (Worster 2004), paleoenvironmental records indicate periods of prolonged aridity that activated dunes and dust storms were common on the Great Plains during the late Holocene (see Holliday 1995; Madole 1995; Arbogast 1996; Hanson et al. 2010; Halfen et al. 2012). A long-term environmental history of this region would aid policy makers in developing land use and energy policies that sustain landscape stability in the face of climate change on the Southern Rolling Plains. Determining the feasibility of and strategies for avoiding a future Dust Bowl in this agriculturally productive, yet drought prone region is of crucial importance.

Previous studies suggest that coeval sedimentary deposits and soils, as well as archaeological deposits and plant macrofossils, are preserved in the canyons of Central Oklahoma. However, prior to my study, no multi-canyon investigation had been conducted that correlated the soil stratigraphy among and between the canyons to confirm this supposition. Also, no study had used the soil stratigraphy and stable carbon isotope record to reconstruct landscape evolution in the canyons and paleoenvironmental history of the region, respectively. The objectives of this study were to: 1) detail the geomorphology and soil stratigraphy of two previously undescribed canyons, Farra Canyon and Armstrong Creek, 2) determine if patterns of sediment flux and preservation are consistent between the canyons and other streams in the region, 3) develop an alluvial chronology for the canyons, and 4) reconstruct late-Quaternary bioclimates based on the soil stratigraphic record and results of stable carbon isotopes analysis of organic matter from buried soils.

2. Study Area

2.1 Physiography and Geomorphology

The Southern Rolling Plains, also known as the Osage Plains, is the southward extension of the Central Lowland physiographic province that was never glaciated (Fenneman 1938). This area, comprised primarily of western Oklahoma and central Texas, is bordered by the Great Plains to the west and south, the Coastal Plain to the southeast, and the Ozark Plateaus and Ouachita Province to the east. The canyons are located within the Western Sandstone Hills geomorphic province of west-central Oklahoma (Curtis et al. 2008; Fig. 1). The province is characterized by rolling hills and broad plains composed of Permian-age sandstone. Tributaries to the Washita and Canadian rivers have deeply incised the sandstone to form steep, north-south trending canyons. The v-shaped canyons become more entrenched toward the mouth of the streams. There are many canyons, and each canyon drains only a small catchment area, generally less than 10 km^2 (Fig. 1, inset).

Over at least the last 13,000 years, large volumes of sediment have been deposited and stored in the canyons. The thickness of valley fills increases downstream, ranging from less than one meter thick at the headwaters of the canyons to about 30 m thick at the mouth of the canyons. Two to four alluvial terraces occur in the canyons. The oldest terraces only occur as remnants high in valley landscapes, and geomorphic surfaces comprising the valley floors of the canyons range from historic alluvial terraces to late-Pleistocene surfaces and Permian sandstone (Norris 1951; Retallick 1966)

2.2 Climate and Vegetation

Central Oklahoma has a dry, subhumid continental climate with fairly mild winters and long, hot summers. Based on meteorological records for Caddo County, Oklahoma, the mean annual precipitation is 760 mm, with most occurring in late spring and early fall (Oklahoma Climatological Survey 2004). The mean annual temperature is 16.1˚ C, and the mean January and July temperature is 3.0˚ C,and 28.5˚ C, respectively. Flash flooding is common in the canyons, and severe droughts have afflicted the region roughly every 10 to 20 years for the period of record (Oklahoma Climatological Survey 2011).

Two plant communities occur in the study area: tallgrass prairie and post oak-blackjack forest, also known as the Cross Timbers (Hoagland 2008). Dominant tallgrass prairie species include big bluestem (*Andropogon gerardii*), little bluestem (*Schizachyrium scoparium*), switchgrass (*Panicum virgatum*), and Indiangrass (*Sorghastrum nutans*). The Cross Timbers is a mosaic of woodland and grassland species, including post oak (*Quercus stellata*), blackjack oak (*Quercus marilandica*), red cedar (*Juniperus virginiana*), buckbursh (*Ceanothus cuneatus*), eastern redbud (*Cercis canadensis*), big bluestem (*Andropogon gerardii*), and Indiangrass (*Sorghastrum nutans*) (Hoagland 2008). The canyon bottoms are sheltered from the wind and hot sun, and springs provide a year-round water supply. Hence, a moist microclimate prevails inside the canyons (Vining 1964). Average daily temperatures, air movement, and evaporation are lower and relative humidity is higher in the canyons compared to the uplands (Rice 1960). This microenvironment is more conducive to the growth of mesic plant species than the surrounding upland (Vining 1964), and supports relict stands of sugar maple (*Acer saccharum*) and a variety of mosses, liverworts and ferns, approximately 250 km west of where they commonly occur (Little 1939, Hoagland 2008).

3. Methods

3.1 Site Selection

In order to address the research goals, the soil stratigraphy and geomorphology of two canyons in central Oklahoma, Farra Canyon and Armstrong Creek, were studied and compared to four other canyons in the region: Domebo Canyon (Albritton 1966), Cedar Creek (Nials 1977), Carnegie Canyon (Lintz and Hall 1983), and Delaware Canyon (Ferring 1986). Archaeological and paleoenvironmental investigations in Farra Canyon and Armstrong Creek have revealed the presence of buried soils (Hofman 1988; Banks et al. 1994; Jack Hofman 2009; Don Wyckoff 2010), and multiple buried soils were observed during reconnaissance trips to these two canyons in the summer of 2010.

3.2 Soil stratigraphic descriptions and sample collection

We characterized the soil stratigraphy of the Caddo Canyons by examining stream-bank exposures at two locations in Farra Canyon and one location in Armstrong Creek and performing laboratory analyses on samples collected from these exposures. We described the cut-bank exposures following procedures outlined by Scheoneberger et al. (2002) and Birkeland (1999). Bulk samples of sediment and soil from cut-bank exposures were collected at 10 cm increments within all buried soils and at 20-30 cm increments within deposits above and below buried soils. Samples were placed in ziplocked bags and returned to the laboratory for analysis.

3.3 Laboratory analysis

Samples were collected for determination of particle-size distribution, soil organic matter (SOM) content, radiocarbon ages, and stable carbon isotope values. Samples were analyzed at the Geoarchaeology and Paleoenvironmental Research Laboratory at the Kansas Geological Survey. A modified pipette method (Burt 2004) was employed to determine the particle-size distribution of soils and sediments. The sand fraction was further analyzed to determine the proportion by weight of very fine, fine, medium, coarse, and very coarse-sized sands in each sample. This was accomplished by passing the dried sand fractions through a series of sieves agitated on a Ro-tap shaker for two minutes and weighing the amount of sample caught on each sieve. Soil organic matter content was determined using the loss-on-ignition technique following the procedure of Konen et al. (2002). Results of these laboratory analyses were used to confirm the presence of buried soils and unconformities, to differentiate eolian from alluvial deposits, and to detect nuances in canyon alluvial sequences.

Subsamples of all soil samples and a representative proportion of the sediment samples from all cut-bank exposures were collected for stable carbon isotope ($\delta^{13}C$) analysis. $\delta^{13}C$ of soil organic carbon reflects the proportion of C_3/C_4 plants that have contributed organic matter to the soil (Schlesinger 1997; Nordt 2001). C_3 plants include trees, shrubs, forbs, and cool-season grasses, and C_4 plants are warm-season grasses. Previous studies have indicated that shifts in C3/C4 plant communities are strongly related to temperature on the Great Plains (Nordt et al. 2008; Von Fischer et al. 2008). $\delta^{13}C$ values of bulk SOM in the samples were determined at the University of Kansas Keck Paleoenvironmental & Environmental Stable Isotope Laboratory using a dual-inlet Finnigan MAT 253 stable isotope ratio mass spectrometer. All samples were pre-treated to remove carbonates and pulverized to insure homogenous samples. The findings from the stable carbon isotope analysis were used in concert with the soil stratigraphy to reconstruct late-Quaternary landscape evolution and paleoenvironments of the Caddo Canyon region

3.4 Numerical chronology

To identify synchronous periods of soil formation between canyons and reconstruct the regional late-Quaternary climate, we developed a reliable numerical chronology of canyon environmental change during the Holocene and terminal Pleistocene. Radiocarbon ages of buried soils were determined on SOM. Radiocarbon dating of SOM provides a mean age of organic carbon in the soil. Radiocarbon ages can vary depending on factors including soil texture, soil biology community composition, and the types of plant residue incorporated into the soil through time. Although radiocarbon dating of SOM can be problematic (Birkeland 1999; Holliday 2004: 174-183), with proper care in sampling and interpretation, SOM can provide accurate age control, especially in drier environments (Holliday et al. 2006). Reliable dating of SOM has been demonstrated in many studies (e.g., Holliday et al. 1994; Rawling III et al. 2003; Mayer and Mahan 2004; Mandel 2008).

In this study, large (1 to 2 kg) samples were collected from buried soils and processed at the Illinois State Geological Survey Isotope Geochemistry Laboratory. The samples underwent standard pretreatment to remove rootlets and calcium carbonate. Radiocarbon ages were determined for the total decalcified soil carbon using liquid scintillation counting or accelerator mass spectrometry (AMS), thereby providing mean residence times for organic carbon in the samples (see Campbell et al. 1967). Although mean residence time does not provide the absolute age of a buried soil, it does give a minimum age for the period of soil development, and it provides a limiting age on the overlying material (Haas et al. 1986; Birkeland 1999: 137). Where possible, in-situ charcoal contained in sediment above and below buried soils was dated in order to build the numerical chronology and define the soil-forming period for each buried soil. AMS was used to date four charcoal samples and two soil organic matter samples.

4. Results

4.1 Geomorphology of the Caddo Canyons

We discovered through reconnaissance of four canyons in the Caddo Canyon region, Farra Canyon, Armstrong Creek, Cedar Canyon, and Dead Woman Creek, that the geomorphology of the sandstone canyons is remarkably similar. The geomorphology of these four canyons also resembles what has been observed in other canyons in the region (see Retallick 1966; Lintz and Hall 1983; Ferring 1986). All of the canyons have formed through headward erosion of the soft sandstone bedrock that dominates the region, and the streams in the canyons are tributaries to high-order streams such as Deer Creek and the Washita, and Canadian rivers. They continue to lengthen through headward erosion (Fig. 2). The canyons tend to be v-shaped and relatively short, ranging from 3 to 10 km in length. The canyons are parallel to each other, and are oriented in a southwest-northeast direction, draining into the high-order streams at their mouths.

Thick packages of alluvial fill are preserved in the canyons, especially beneath the T-1 and T-2 terraces in the middle and lower reaches of the streams. Proximity to the headwaters versus mouth of the canyon, in part, dictates the locations of terraces and patterns of alluvial fill preservation. Near the headwaters, where modern erosional episodes have weakened the sandstone bedrock, only recently deposited alluvium occurs beneath T-0 and T-1 surfaces. Sequences of stratified alluvium, evidence for episodic deposition of sediment through flooding or slopewash, are common in fills in the upper reaches of the canyons. Weakly developed soils have formed at the surface of these fills. In the middle reaches of the canyons, T-1 and T-2 fills are 3 to 20 m thick and have stratified flood deposits and multiple buried soils. However, in the middle reaches, soil stratigraphic sequences vary among sections within the same canyon as well as between canyons. Thick packages of alluvium also occur beneath T-1 terraces in the lower reaches of the canyons, although buried soils were not commonly observed in these alluvial fills. T-2 surfaces were not found in either the upper or lower reaches of the four canyons.

While multiple terraces occur in all of the canyons, contiguous terraces running the length of the canyons do not exist. In other words, all T-0, T-1, and T-2 surfaces remain only as remnants due to the highly variable bedrock geometry of the canyons and extreme erodibility of the sandstone. Where the canyons become quite narrow due to the form and structure of the bedrock, there are no terraces and hence no alluvial fills preserved (Fig. 3). At locations where the canyons are narrow and deep, the channels are incised in bedrock. Hence, these locations also lack alluvial fills. In contrast, where the canyons are wide and fairly shallow, alluvial fills often occur beneath multiple terraces (Fig. 4). The canyons typically are narrow in the upper and middle reaches and widen toward the mouth. However, the width of a canyon can be extremely variable along its length.

Meanders in the canyons also dictate presence or absence of remnant terraces. The thickest packages of alluvial fill, often exceeding 10 m, tend to occur beneath remnant T-2 terraces tucked up against the canyon wall along meander bends where the canyon widens. However, preservation of T-2 terraces varies among the four canyons and is fairly rare. Remnants of the T-2 terrace generally occur in wide portions of the canyons, commonly at meander bends in the middle reaches.

4.2 Soil stratigraphy and radiocarbon chronology of Farra Canyon

Farra Canyon is 3.75 km long and drains into Deer Creek to the south. The thickness of the valley fill generally increases downstream, but varies greatly, ranging from less than one meter thick at the headwaters to 20 m or more thick in the middle and lower reaches of the canyon. One to three remnant alluvial terraces frequently occur in Farra Canyon. Geomorphic surfaces comprising the valley floor of Farra Canyon ranged from historic alluvial terraces to late-Pleistocene surfaces and sandstone bedrock. In order to determine the nature of the fill and the paleoenvironmental record preserved beneath remnant terraces in Farra Canyon, we described the soil stratigraphy from two cut-bank exposures in the canyon, hereafter referred to as Farra 1 and Farra 2.

At Farra 1, nearly 6 m of fill are exposed beneath the T-1 terrace (Fig. 5). The upper 3 m of fill is characterized by 1-2 m-thick packages of massive sands separated by buried soils, while fluvial sands interbedded with clay-rich flood drapes dominate the lower 3 m. The fill is primarily composed of fine and very fine sands, ranging in texture from sand in C horizons to sandy loams within the A and B horizons of buried soils, and clay-rich flood drapes (Fig. 6). Sandy deposits that mantle the T-1 terrace have a greater proportion of very fine sands compared to the underlying fill, and bedded sands composing the C4b3 horizon have the greatest proportion of medium-sized grains.

Three buried soils were recorded at Farra 1 (soils 2, 3, and 4 in Fig. 6). Structure, color, and peaks in organic matter content were indicators of buried soils in all cut-banks. With the exception of soil 3, the surface and buried soils are all weakly developed at Farra 1. The surface soil has an A-AC-C profile, and the first and third buried soils have A-C horizonation. In contrast, the second buried soil (soil 3) displays A-Btk-C horizonation. Also, biogenic features, including multiple krotovina as large as 7 cm in diameter, are common in the A horizon of soil 3. The biogenic features and horizonation of soil 3 indicate that it probably formed during a longer period of landscape stability compared to the other buried soils at Farra 1. We also observed a sandy, organic-rich deposit at the base of the cut-bank at Farra 1. The top of this deposit was marked by a sharp erosional contact with the overlying bedded sands (Fig. 7). While the deposit was dark in color and heavily bioturbated, we did not observe soil structure or horizonation indicating pedogenesis. Hence, we classified this deposit as Cb4, and believed it to be equivalent to the lower member of the Domebo formation (Qdl) described by Albritton (1966) in Domebo Canyon.

Radiocarbon dating of bulk soil and charcoal collected from the cut-bank at Farra 1 was conducted during Fall 2010 (see Fig. 6). SOM from the shallowest buried soil (soil 2) and from soil 3 beneath it, yielded radiocarbon ages of 610 $\pm$ 15 B.P. and 375 $\pm$ 15 B.P., respectively. The inverted ages may be due to contamination by younger carbon, as a significant amount of mixing from bioturbation was observed in the A horizon of soil 3. Differences in soil texture or the nature of plant residue incorporated into the soil can cause younger carbon to be preserved in one soil but not another, which also could explain the inverted radiocarbon ages of soil 2 and soil 3. SOM from soil 4 yielded a radiocarbon age of 1795 $\pm$ 15 yr. B.P. In-situ charcoal collected from the top of the organic-rich deposit approximately 5 m below the surface yielded a radiocarbon age of 9620 $\pm$ 35 B.P. The age of soil 4 compared to the age of the charcoal in the organic-rich deposit confirm a likely erosional episode during the early to mid-Holocene prior to deposition of the bedded sands in which Soil 4 formed. The radiocarbon age of the charcoal from the organic-rich deposit near the bottom of the section is consistent with ages of the Qdl and other basal deposits in the region assumed to be equivalent to the Qdl (see Albritton 1966; Hofman 1988, 1990; Banks et al. 1994).

Farra 2, a cut-bank exposure of fill beneath a T-2 terrace in Farra Canyon, exhibited both similarities and differences compared to Farra 1 (Fig. 8). Most notably, the T-2 fill is over 14 m thick at Farra 2, compared to 6 m thick at Farra 1. Like Farra 1, the fill at Farra 2 consists primarily of fine and very fine sands, with the upper portions (top 8 m) dominated by massive sands separated by buried soils, and the lower portions (bottom 6 m) dominated by fluvial sands interbedded with clay-rich flood drapes. The top 2 m of fill are sandier and have a smaller proportion of very fine sand compared to Farra 1. The size of sand grains is fairly uniform and similar to Farra 1 through the remainder of the section (approximately 80% sand comprised of 60-70% fine sands and 20-30% very fine sands), except where buried soils or flood drapes occur.

Three buried soils were recorded and described at Farra 2 (soils 2, 3, and 4 in Fig. 9). The surface soil (soil 1) and shallowest buried soil (soil 2) are nearly identical in depth and character to soil 1 and soil 2 at Farra 1. Both soils are weakly developed, displaying A-AC-C profiles dominated by very fine sands. Few, wavy lamellae occur near the base of the sandy subsoil of soil 2. A second buried soil was recorded at Farra 2 (soil 3) at almost the same depth as the second buried soil at Farra 1. However, there was no A horizon in soil 3 at Farra 2, indicating that an erosional episode trunctated the soil. Soil 3 at Farra 2 consists of a 2 m-thick Bw horizon with medium prismatic parting to subangular blocky structure that formed in a 4 m thick deposit of sand. A 1 m-thick Ckb2/Bkb2 horizon is at the base of soil 3 at Farra 2, and calcium carbonate from this horizon overprinted the A horizon of the third buried soil (soil 4). The Ckb2/Bkb2 horizon was described as part of soil 3, though radiocarbon ages indicate it could be a horizon of another soil (Fig. 9). The presence of the Bw horizon and the thick, carbonate-rich subsoil of soil 3 at Farra 2, in which common, distinct threads and films of carbonate were observed, differs significantly from the thin, Btk horizon of soil 3 at Farra 1 with few, faint carbonate threads. The third buried soil at Farra 2 (soil 4) is at the top of a series of flood deposits, similar to soil 4 at Farra 1. Also, an organic-rich deposit similar to the deposit observed 5 m below the surface at Farra 1 occurs at the base of the cut-bank at Farra 2.

At Farra 2, bulk soil samples from the uppermost horizon of every buried soil, the Bkb2/Ckb2 horizon, and sediments from the base of the section, as well as charcoal from a flood-drape and the lowermost organic-rich deposit, were collected and radiocarbon dated. SOM from the A horizon of soil 2 at yielded a conventional radiocarbon age of 950 $\pm$ 70 yr. B.P. This age confirmed what we suspected based on observations of the soil stratigraphy at Farra 1 and Farra 2: the most recent periods of soil formation (modern soil and the first buried soil) are coeval at Farra 1 and 2. The SOM from the upper-most horizon of soil 3, the Bw1b2 horizon, returned a conventional radiocarbon age of 5250 $\pm$ 100 yr. B.P., indicating that soil 3 is much older at Farra 2 compared to soil 3 at Farra 1. The radiocarbon ages are supported by the substantially different properties of the second buried soil at Farra 1 compared to Farra 2 and corroborate the presence of an erosional unconformity at the top of soil 3 at Farra 2.

We used AMS rather than conventional techniques to date SOM from bulk samples of the Ckb2 and Akb3 horizons at Farra 2. While we prefer conventional dating of bulk SOM to determine the mean age of carbon in a soil, AMS dating was used for these samples because there was not a sufficient amount of carbon to date them without assuming large errors. The Bkb2/Ckb2 horizon of soil 3 returned an AMS radiocarbon date of 10,075 $\pm$ 35 yr. B.P., and the Akb3 horizon was dated to 10,420 $\pm$ 35 yr. B.P. While these dates are in sequential order considering their location in the section, they are much older than expected considering two ages determined on charcoal and one conventional age on bulk sediment from below soils 3 and 4 (Fig. 9). It is likely that the presence of old detrital carbon in samples from the Bkb2/Ckb2 and Akb3 horizons explains the anomalous ages. Reliable radiocarbon ages determined on soil organic matter above and below soil 3, however, bracket its age between 5250-8800 ^{14}C yr. B.P.

Charcoal collected from a flood-drape approximately 10 m below the surface and from the lowermost organic-rich deposit 12 m below the surface yielded radiocarbon ages of 8800 $\pm$ 25 yr. B.P. and 9530 $\pm$ 35 yr. B.P., respectively. Organic matter from a bulk sediment sample collected near the bottom of the cut-bank yielded an age of 9460 $\pm$ 160 yr. B.P. These ages indicate that the organic-rich sediment at the base of Farra 2 and Farra 1 are coeval and equivalent to the Qdl.

4.3 Soil stratigraphy and radiocarbon chronology of Armstrong Creek

Armstrong Creek runs along a 6 km course and drains to the north into Deer Creek. This canyon contains little to no fill near the headwaters, especially along a narrow 1-2 km stretch of canyon in the upper reaches. Toward the mouth of the canyon, the amount of fill in the valley increases to as much as 8 m thick, but no fill greater than 10 m thick was recorded in Armstrong Creek. Typically, one or two alluvial terraces occur in Armstrong Creek. Like Farra Canyon, the geometry of the canyon dictates the nature of the fill preserved in its bottom, and geomorphic surfaces found in the canyon range from historic alluvial terraces to late-Pleistocene surfaces and sandstone bedrock. The soil stratigraphy exposed in one cut-bank in Armstrong Creek was described and compared to the soil stratigraphy recorded in Farra Canyon.

At Armstrong Creek, a 7 m-thick section of alluvial fill is exposed beneath a T-1 terrace (Fig. 10). The fill consists primarily of fine and very fine sands, with a greater proportion of very fine sands in the top 2 m compared to lower portions of the section. The top half of the fill consists of the surface soil and a pedocomplex of buried soils formed in a sandy deposit. The lower 4 m of fill consists of fluvial sands interbedded with clay-rich flood drapes, similar to the fill at the base of both cut-banks examined in Farra Canyon. Textures of the alluvial deposits range from clay loams or loams in the flood drapes to loamy sands in the coarser fluvial deposits. An organic-rich, sandy deposit similar to those observed in Farra Canyon is approximately 6 m below the surface.

A surface soil and three thin but well-developed buried soils were recorded in the cut-bank exposure in Armstrong Creek (Fig. 11). The surface soil (soil 1) has a well-expressed A-AB-Btk profile. Carbonate nodules measuring up to 0.3 cm, as well as carbonate threads and films were described in the Btk horizon. Soil textures ranged from sandy loams in the A and AB horizons to sandy clay loam in the Btk horizon. Three buried soils form a pedocomplex just below the surface soil. The first buried soil (soil 2) has an Ab-Bwb-Bkb profile. Carbonates from the Bkb horizon of soil 2 have overprinted on soil 3, welding these two soils together and creating paled colors in the Bkb1 horizon. Darkening of the Akb2 horizon distinguishes soil 3 from soil 2, and laboratory analyses confirmed higher organic matter contents typical of a buried A horizon. The second buried soil (soil 3) has an Akb-Bkb pofile, and like the soil above it, carbonates from soil 3 have overprinted on the soil below it, the third buried soil (soil 4). Soil 4 has a weakly expressed Akb-Cb profile developed in sandy alluvium. There is no evidence of soil formation in the clay-rich flood-drapes or sandy alluvial deposits present in the lower half of the section.

SOM from the A horizons of the three buried soils was radiocarbon dated using conventional techniques. SOM from the Ab1 horizon of soil 2 dated to 2170 $\pm$ 100 yr. B.P. (Fig. 11). Samples from the buried A horizons of soil 3 and 4 returned mid-Holocene radiocarbon dates of 4110 $\pm$ 130 yr. B.P. and 4750 $\pm$ 200 yr. B.P., respectively. In-situ charcoal recovered from the organic-rich sediment approximately 6 m below the surface was radiocarbon dated using AMS techniques. The charcoal dated to 9380 $\pm$ 25 yr. B.P. The age of the charcoal and the stratigraphic position and characteristics of the organic-rich deposit in Armstrong Creek indicates that it is both age and geomorphically equivalent to the organic-rich deposit recorded and dated in Farra Canyon.

4.4 Stable Carbon Isotope Compositions for Farra Canyon and Armstrong Creek

The stable carbon isotope ratios ($\delta^{13}C$) determined on SOM exhibited notable consistencies as well as significant differences between the three cut-banks. The surface soils at all sites exhibited the most depleted $\delta^{13}C$ values at the top, becoming enriched in ^{13}C with depth through the soil. Additionally, the $\delta^{13}C$ values of surface soils are clearly dictated by modern vegetation at the site today. The surface soils at both Armstrong Canyon and Farra 1, where cedar trees with extensive root systems are growing today (Fig. 5 and 10) have depleted $\delta^{13}C$ values (-25.2‰ and -25.5‰, respectively). The surface soil at Farra 2, in contrast, was enriched in ^{13}C, as indicated by $\delta^{13}C$ values of -18.2‰ in the A horizon. The vegetation at Farra 2 was more of a savanna, with grasses growing in the area where the surface soil was sampled, and cedar trees nearby (Fig. 8).

Analysis of the $\delta^{13}C$ of sandy alluvium contained in both canyons exhibited some consistencies among the three sites. The stable carbon isotope values of the organic matter in the sands generally centered around -22‰, ranging from -19.4‰ at a depth of 2.75 m at Farra 1 to -24.1‰ at a depth of 7 m at Farra 2. Relatively constant $\delta^{13}C$ values occur through thick deposits of sandy alluvium in the lower 3 m of fill at Armstrong Creek. The $\delta^{13}C$ values also remain fairly constant at depths of 1-2 m and 4-8 m below the surface and below the Qdl at Farra 2.

The $\delta^{13}C$ values from buried soils in the sections showed dramatic variability, both within individual cut-banks and when comparing coeval soils in different canyons. $\delta^{13}C$ values of buried A horizons ranged from -16.4‰ in soil 4 at Farra 1 to -25.6‰ in soil 4 at Armstrong Creek. While the surface soil at Farra 1 exhibited a depleted $\delta^{13}C$ value, the first and second buried soils were enriched in C^{13}, with $\delta^{13}C$ values of -20.4‰ and -20.3‰, respectively. The third buried soil at Fara 1 (soil 4) that formed at the top of a flood drape was even more enriched, and had the least negative $\delta^{13}C$ value in the study (-16.4‰). A stable carbon isotope value of -17‰ was determined on organic matter from the Ab1 horizon of the first buried soil at Farra 2, which is the same value determined for organic matter from the surface soil. However, the $\delta^{13}C$ values for the second buried soil (soil 3) became steadily more depleted deeper in the profile. $\delta^{13}C$ values for the Bw1b1 horizon of soil 3 averaged -20.5‰, while the Bw2b2 horizon had $\delta^{13}C$ values averaging -22.6‰. $\delta^{13}C$ value for the Akb3 horizon of soil 4 at Farra 2 was -23.2‰. The first buried soil at Armstrong Creek had a $\delta^{13}C$ value similar to those for the sands near the bottom of the section, with -21.0‰ in the Ab1 horizon. Soils 3 and 4, which are welded together by overprinting of carbonates and date to the mid-Holocene, exhibited the most depleted $\delta^{13}C$

values in the Armstrong Creek cut-bank, with the exception of the A horizon of the surface soil. $\delta^{13}C$ values for these two soils ranged from -22.1‰ in the Akb2 horizon to 25.6‰ near the top of the C1b3 horizon.

Three pairs of approximately coeval soils were identified at the three sites based upon radiocarbon ages determined on SOM. Comparison of the $\delta^{13}C$ values of these soils is useful for reconstructing the paleoenvironment of the Caddo Canyon region. Soil 2 at Farra 1 and soil 2 at Farra 2 were radiocarbon dated to 610 $\pm$ 15 yr. B.P. and 950 $\pm$ 70 yr. B.P., respectively, indicating they both formed in the last 1000 years. The $\delta^{13}C$ value for the A horizon of soil 2 at Farra 1 is -20.4‰ and the A horizon in soil 2 at Farra 2 has a $\delta^{13}C$ value of -16.7‰. Soil 4 at Farra 1 and soil 2 at Armstrong Creek both formed during an earlier period of the late Holocene, with radiocarbon ages of 1795 $\pm$ 15 yr. B.P. and 2170 $\pm$ 100 yr. B.P., respectively. The $\delta^{13}C$ values of organic matter from the A horizons of these soils were -16.4‰ and -21.6‰, respectively. Lastly, soil 3 at Farra 2 and soil 4 at Armstrong Creek formed during the same part of the mid-Holocene, with SOM from these two soils radiocarbon dated to 5250 $\pm$ 100 yr. B.P. and 4750 $\pm$ 200 yr. B.P., respectively. $\delta^{13}C$ values from the A horizons of these two soils were -19.8‰ and -25.6‰, respectively. Like the surface soils, organic matter from the A horizons of buried soils that are coeval yielded $\delta^{13}C$ values more than 5 ‰ different from each other.

5. Discussion

5.1 Geomorphic and bedrock controls on canyon alluvial fill

Canyon geomorphology and geology control many of the characteristics of the late-Quaternary alluvial fills in the Caddo Canyons. Specifically, the orientation of the canyons with respect to nearby high-order streams, the highly variable geometry of the canyons' bedrock bottoms, and the exceedingly erodible nature of the sandstone bedrock affect the lithology and stratigraphy of alluvial fills. The lowermost fill in the three cut-banks that were described in detail, as well as in other cut-banks that were observed, consist of flood-drapes interbedded with sandy deposits overlying the Qdl. Since the canyons contain short, spring-fed, first order streams flowing directly into high-order streams at their mouth, changes in base level of the high-order streams directly controls flooding in the canyons. The energy level of flooding in the large stream determines the nature and volume of sediment, and the distance upstream that alluvium accumulates in the canyons. Backwater flooding rapidly fills canyons with sediment during the first phase of aggradation, depositing mostly clay-rich sediments on the floors of the canyons. Sandy alluvium derived from sandstone bedrock is transported down the canyons during larger flood events, becoming interbedded with the clay-rich backwater flood drapes. The Qdl at the base of most canyons in the region, as surmised by Leonhardy (1966) and Retallick (1966) in the original investigation at Domebo Canyon, is the product of backwater embayment that was common in the canyons during the Pleistocene-Holocene transition. The Qdl remains at the base of most canyons today due to its clay-rich consistency that is resistant to erosion.

After a sufficient volume of sediment has accumulated in a canyon bottom, the geomorphology of the canyon is no longer affected by streamflow in the trunk stream. Further, once fill accumulates in the canyon to a sufficient height, flooding cannot overtop the fill surface. Entrenchment transforms the floodplain into a terrace, and soil formation occurs on the elevated surface. After creation of the terrace, only major floods or eolian deposits overtop the terrace fills and bury surface soils. This process explains the presence of pedogenically-unmodified flood deposits in the lower half of all sections and the presence of the buried soils typically closer to the modern surface.

As previously noted, the highly variable geometry of the canyons determines where alluvial fills are preserved. For example, an 11 m-thick alluvial fill that aggraded between 10,000 and 5000 yr. B.P. is preserved at Farra 2 tucked up against the canyon wall in a wide portion of a meander bend. In contrast, only 5 m of fill occurs beneath the T-1 terrace at Farra 1 in a much narrower portion of Farra Canyon. The radiocarbon age determined on in-situ charcoal from the Qdl at Farra 1 compared to the age of the third buried soil above it (Fig. 6), as well as the erosional unconformity at the top of the Qdl (Fig. 7), indicate that an erosional event removed sediment overlying the Qdl during the mid- to late-Holocene, and flood deposits observed there today accumulated since 2000 ^{14}C yr. B.P. Hence, while a mid- to late-Holocene erosional event removed sediment from Farra 1 and likely affected much of the valley, the fill at Farra 2 was protected from erosion due to its position in the canyon. Consequently, the fill at Farra 2 contains a rich record of individual mid-Holocene soil forming events while all buried soils observed at Farra 1 formed during the late Holocene.

During the late Holocene, landscape stability and concomitant soil formation appear to have dominated the region. This conclusion is based upon the common occurrence of buried soils dated to the late Holocene in canyon fills (Ferring 1986; Lintz and Hall 1983). However, due to the abundance of highly erodible sandy alluvium in the canyons, preservation of alluvial fills and associated buried soils is not guaranteed. Hence, in some canyons or portions of canyons the absence of buried soils dating to the mid-Holocene may be related to removal of mid-Holocene alluvium, not because soil formation did not occur during that period. In fact, the presence of buried soils dating to the mid-Holocene at both Farra 2 and Armstrong Creek provide evidence that landscape stability and soil formation occurred in the region during the mid-Holocene. In summary, buried soils contained in the fills of the Caddo Canyons may be accurately interpreted as representing periods of regional landscape stability. However, the absence of buried soils in some portions of the canyons should not be interpreted as widespread landscape instability. As this study illustrates, examination of the soil stratigraphy from multiple cut-banks in a single canyon as well as in other nearby canyons helps elucidate the story of late-Quaternary landscape evolution in the region.

5.2 Paleoenvironmental reconstruction of the Caddo Canyon region

Using the soil stratigraphic and stable isotope record from Farra Canyon and Armstrong Creek, in addition to information gleaned from previous studies conducted in the canyons, we reconstructed a paleoenvironmental history of the Caddo Canyon region (Fig. 12). It appears that the Qdl is ubiquitous in the canyon bottoms of the region and aggraded in a backwater embayment during the Pleistocene-Holocene transition. The radiocarbon chronology presented here in combination with previous work (e.g., Albritton 1966; Hofman 1988) indicates that the backwater embayment engulfed the canyon bottoms between 11,200 to 8800 ^{14}C yr. B.P., with most of the aggradation occurring between 10,000 and 9000 ^{14}C yr. B.P. The Qdl formation was first described at a Clovis mammoth kill site in Domebo Canyon, and mammoth remains have been recovered from it in other canyons (Jack Hofman 2009). δ^{13}C values from Qdl range between -20‰ to -22‰, indicating that a mixed C3/C4 community dominated this region during the terminal Pleistocene. Additional study of the Qdl formation and numerous buried tree stumps recovered from it (Leonhardy 1966; Jack Hofman 2009) could provide important clues to the environmental history of the region at the end of the Pleistocene. Episodic high energy and low energy flooding was common during the early Holocene, as evidenced by clay-rich flood drapes interfingered with fluvial sands overlying the Qdl in Armstrong Creek, Farra Canyon, Cedar Creek, and Domebo Canyon (see Albritton 1966; Nials 1977).

A few episodes of soil formation appear to be ubiquitous in the Caddo Canyon region, although preservation of buried soils in the late-Quaternary stratigraphic record from these periods is inconsistent for the reasons described above. Multiple buried soils dating to the mid-Holocene are preserved in both Farra Canyon and Armstrong Creek. While soils dating to this period have been recorded at nearby Mustang Creek in central Oklahoma (Beale and McKay 2009), no previous reports have documented mid-Holocene soil formation in the Caddo Canyon region. In Carnegie Canyon, 63 buried tree stumps, 50 of which were eastern red cedar, were recovered and radiocarbon dated to between 2000-4000 yr. B.P. (Hall and Lintz 1984). While no buried tree stumps of this age were recovered during our investigations, their presence in another canyon along with mid-Holocene-age buried soils in Armstrong Creek and Farra Canyon indicate that canyon fills may contain a rich, mid-Holocene paleoenvironmental record. We observed carbonate accumulations in the subsoil of all mid-Holocene age buried soils. Subsoil carbonate accumulation is not common to soils forming in this region today, but is observed in soils under more arid moisture regimes. Accordingly, the characteristics of the mid-Holocene soils indicate the climate was likely more arid in the mid-Holocene in central Oklahoma compared to today (Fig. 12). $\delta^{13}C$ values from the A horizons of these soils were -19.8‰ and -25.6‰. Hence, we cannot draw conclusions regarding the climate from these values, as they likely reflect differences in local vegetation or microclimate effects rather than regional climate.

During the late Holocene, episodic flooding and/or eolian activity was punctuated by periods of landscape stability, when soils formed in the alluvial and/or eolian deposits (Fig. 12). Episodes of landscape instability re-mobilized sediment repeatedly during this period, causing surface soils to be buried and thereby creating a sequence of late-Holocene buried soils. The weakly expressed profiles of the late-Holocene soils indicate that episodes of landscape stability were brief. Numerous studies in the canyons (see Nials 1977; Ferring 1986; Hall 1990) have documented late-Holocene buried soils, including the Caddo Paleosol named for the area. We observed buried late-Holocene soils in all three cut-banks examined for this study. The $\delta^{13}C$ values from the A horizons of these soils ranged from approximately -16 to -21‰. While this large range makes it difficult to draw significant conclusions, these values suggest that the vegetative community was mixed and temperatures were possibly warmer than today or the mid-Holocene. Other studies in Oklahoma have interpreted the late Holocene as more moist than today (Albert 1981), and the soil stratigraphy preserved in the Caddo Canyons certainly supports that interpretation.

5.3 Considerations for using $\delta^{13}C$ for paleoenvironmental reconstructions

The results of the $\delta^{13}C$ analysis highlight the potential impact microclimate and geomorphic processes can have on vegetation, and hence, on ^{13}C / ^{12}C ratios in SOM. Factors such as elevation, aspect, and the effect of slope on soil moisture can create microclimates at sites like the canyons and dramatically influence vegetation. The vegetation, in turn, affects the $\delta^{13}C$ value of SOM. In these environments, the $\delta^{13}C$ values reflect the effects of microclimate instead of regional climate on vegetation. Geomorphic processes, specifically disturbance due to fluvial erosion and slope processes, also can affect vegetation communities and create a mosaic of grass, woody shrubs, and trees. Hence, at geomorphically active sites such as the canyons, a savanna results in highly variable $\delta^{13}C$ values across the landscape. Additionally, while the ratio of C3 to C4 plants is the dominant driver of $\delta^{13}C$ values of SOM, differences in microbial communities and associated transformations also occur across the landscape and affect $\delta^{13}C$ values. Preferential decomposition of certain plant materials over others causes increases in $\delta^{13}C$ through time, further challenging interpretations of climate based on $\delta^{13}C$ values of SOM alone.

The $\delta^{13}C$ values determined on SOM from surface soils at the three cut-banks, as well as $\delta^{13}C$ values for pairs of coeval buried soils, illustrates the potential impact that microclimate and geomorphic processes have on $\delta^{13}C$, making these values a problematic proxy for settings such as the Caddo canyons. While vegetation directly affects the $\delta^{13}C$ values of SOM, it does not appear to significantly affect soil formation in this region, at least over short periods (< 1000 yr.). It is crucial, therefore, to interpret $\delta^{13}C$ results in consideration of microclimate and vegetation, and to use multiple paleoenvironmental proxies for an accurate picture of the paleoenvironment.

6. Conclusion

Geomorphic controls dictate the nature of the alluvial fills and the paleoenvironmental record contained within the fills in the canyons, as well as vegetative patterns in and around the canyons. The soil stratigraphy of Farra Canyon and Armstrong Creek indicate that the Caddo canyons have undergone numerous episodes of cutting and filling controlled by flooding in the trunk streams. Patterns of fill are not consistent among canyons, and historical human manipulation of the landscape (e.g. modern dam building), further complicates matters. Still, results from the two canyons investigated here indicate that mid-Holocene and late-Holocene buried soils, as well as the Qdl, are commonly preserved in canyon fills. Further, morphologic consistency of coeval buried soils among the canyons indicate that analysis of $\delta^{13}C$ values of SOM from the buried soils as well as application of paleoclimate models to the buried soils (e.g. Zung and Feddema 2013, in preparation) may be used to reconstruct late-Quaternary climate in the region.

This study presents a high-resolution $\delta^{13}C$ record from the canyons of central Oklahoma, the first such record. We have demonstrated that it is crucial to differentiate geomorphic and vegetation signals from climate-controlled properties when using soil stratigraphy and $\delta^{13}C$ to reconstruct past environments. This study highlights the potential pitfalls in attributing $\delta^{13}C$ values to regional climate in locations where microclimates and geomorphology strongly affect distribution of vegetation. Despite these complexities, our research indicates that downcutting and erosion in the canyons were accompanied by dry conditions in the uplands during the mid-Holocene, while more moist and warm conditions, indicated by aggrading fluvial sediments separated by several short periods of soil formation, characterize the late Holocene in central Oklahoma. A wealth of additional paleoenvironmental information on this region of the Southern Plains could be obtained from the fill contained in the canyons, especially during the Pleistocene-Holocene transition, mid-Holocene and late Holocene, but care must be taken with interpreting the record.

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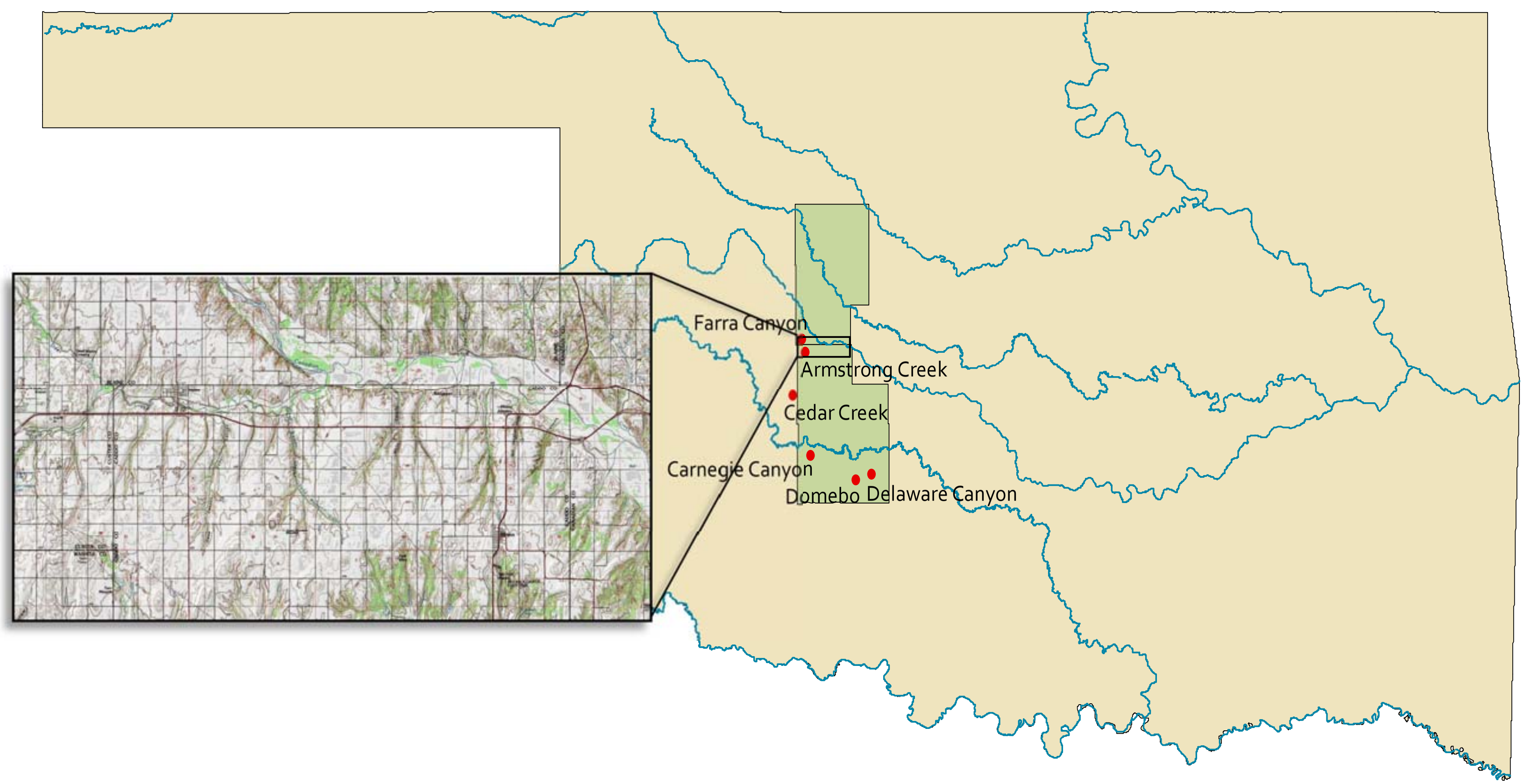

Fig. 1 – Study area. Topographic map shows area where Armstrong Creek and Farra Canyon drain into Deer Creek.

Fig. 2 - Headward erosion in Cedar Canyon,Caddo County, Oklahoma.

Fig. 3 - Sandstone canyon wall exposed along a narrow section of Farra Canyon. At this location and others like it, no alluvial fill is preserved.

Fig. 4 - T-1 and T-0 terraces along a wide portion of Armstrong Creek.

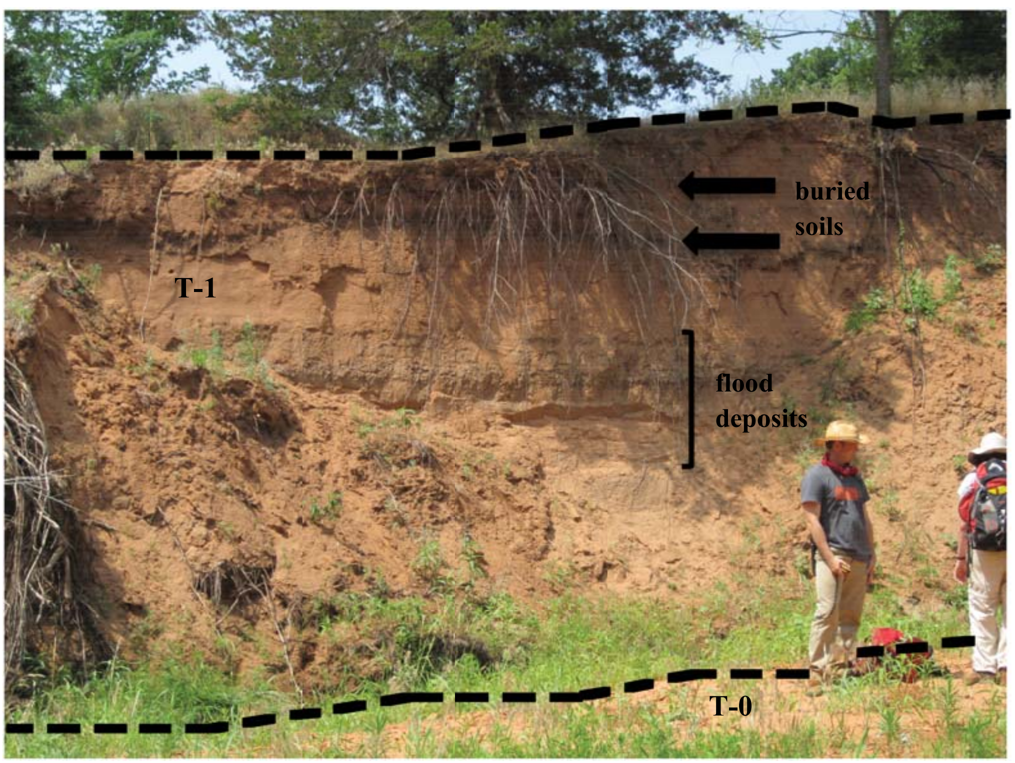

Fig. 5 - The older terrace (T-1) and modern floodplain (T-0) surfaces at Farra 1 in Farra Canyon. Note the buried soils and flood deposits preserved in the fill.

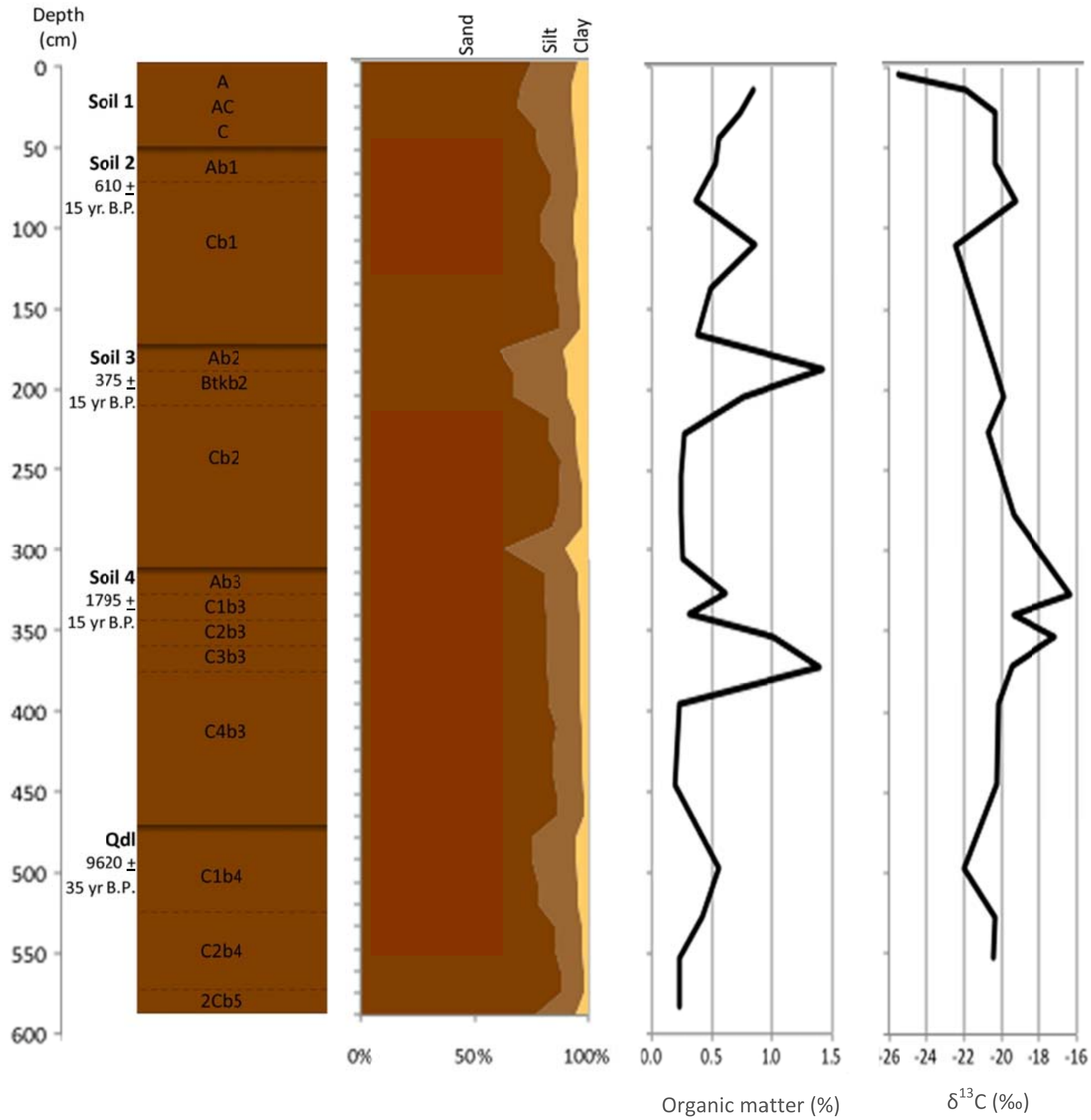

Fig. 6 – Farra 1 soil stratigraphy, particle size distribution, organic matter content, and stable carbon isotopic ratio of soil organic matter. Radiocarbon ages for material dated from the section also shown.

Fig. 7 – Erosional contact between the Qdl and bedded sands at Farra 1.

Fig. 8 – The cut-bank beneath the T-2 terrace at Farra 2.

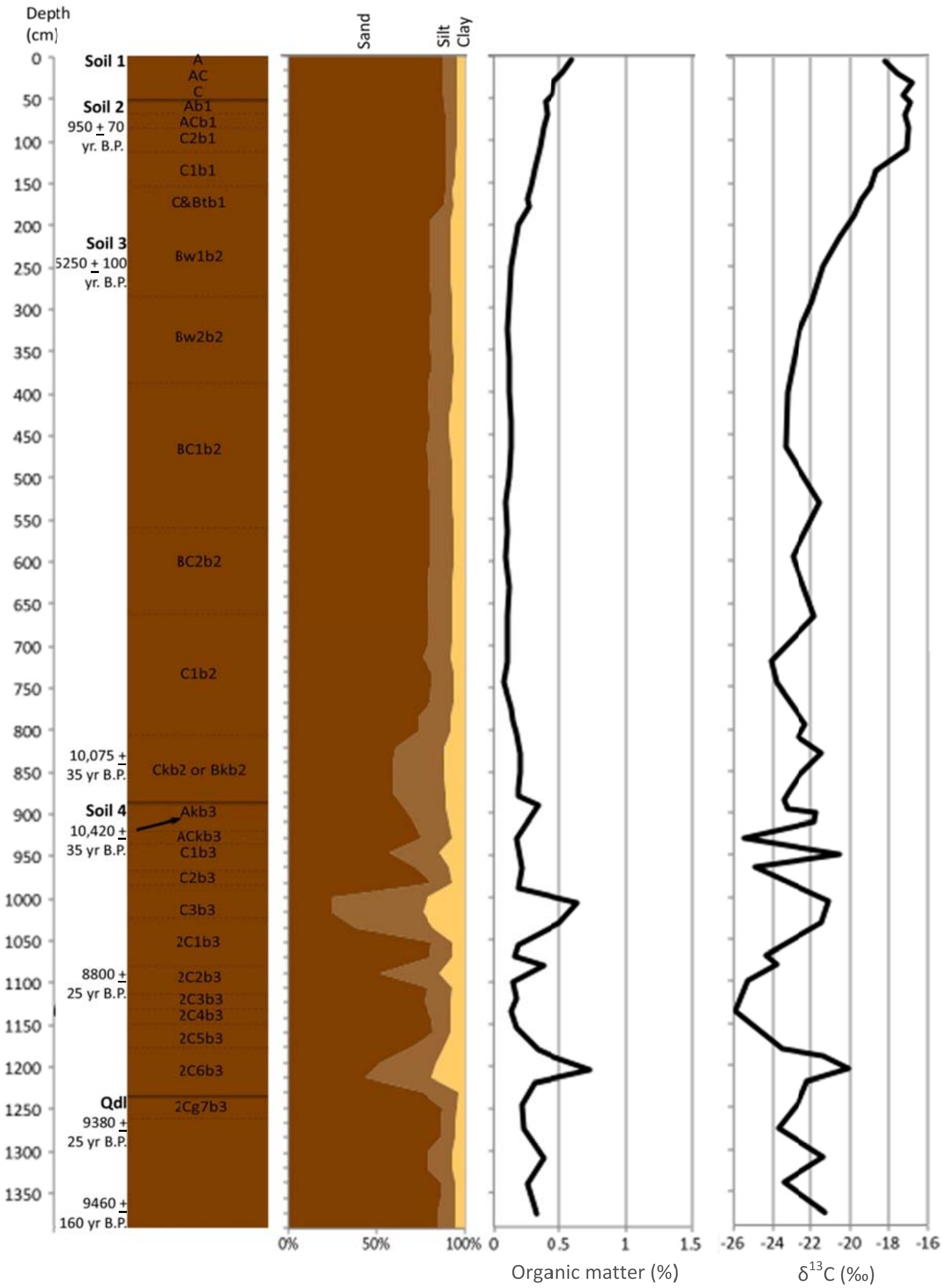

Fig. 9 – Farra 2 soil stratigraphy, particle size distribution, organic matter content and stable carbon isotopic ratio in soil organic matter. Radiocarbon ages for material dated from the section also shown.

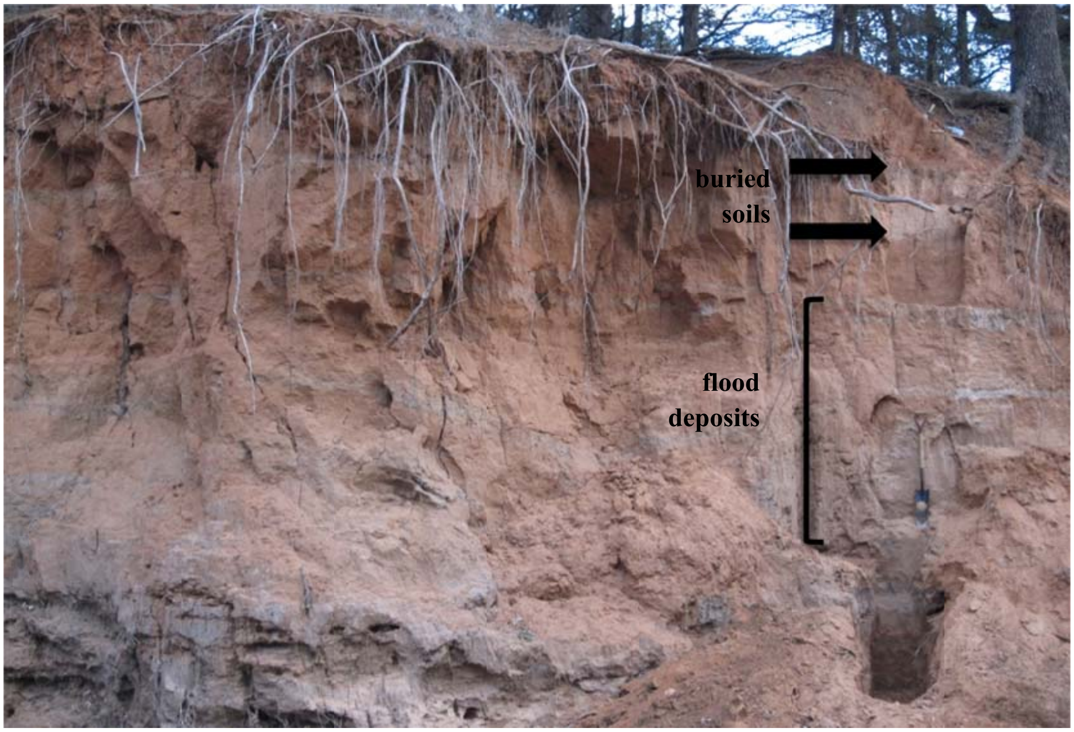

Fig. 10 – The cut-bank beneath the T-1 terrace at Armstrong Creek.

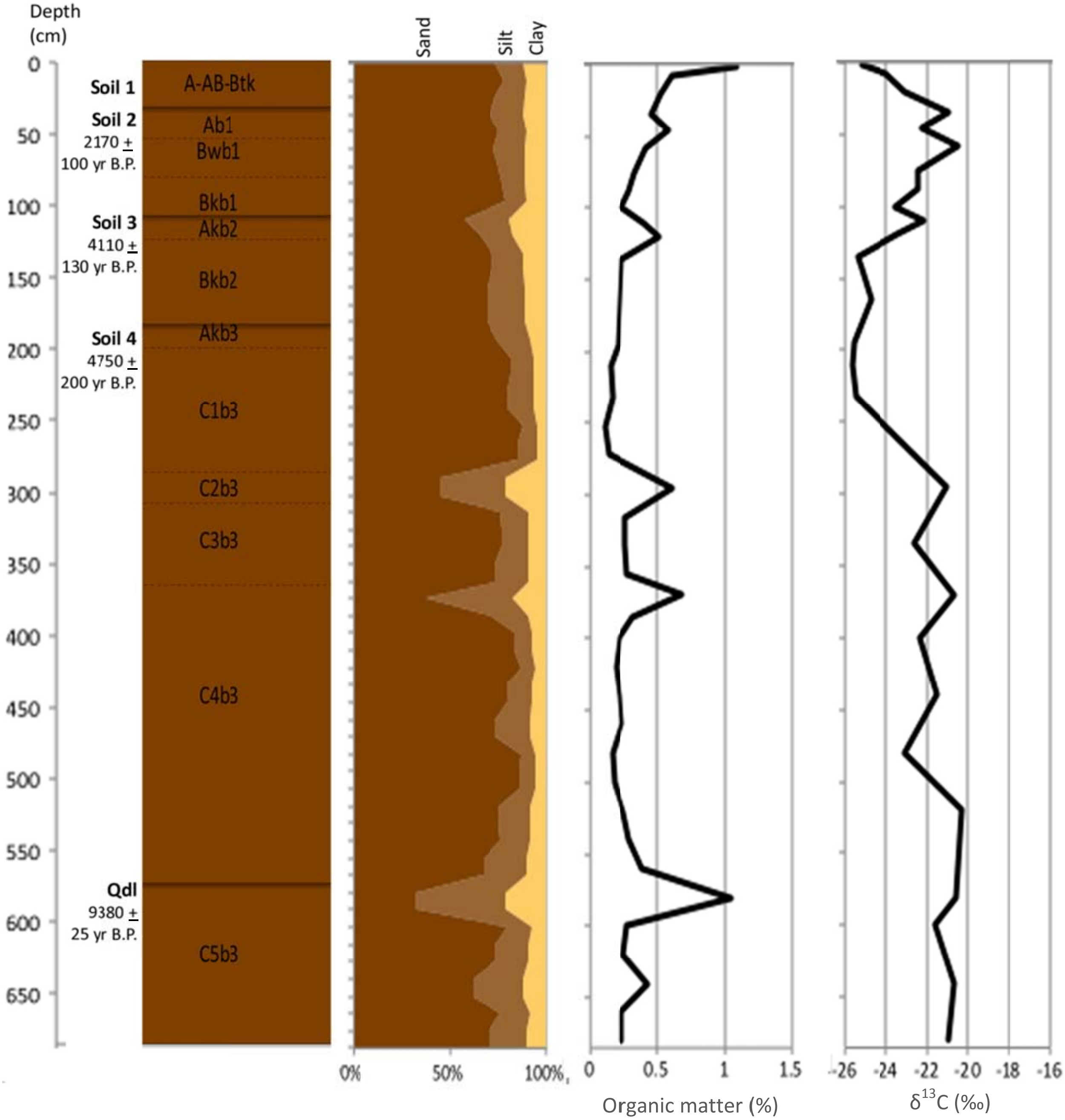

Fig. 11 – Armstrong Creek soil stratigraphy, particle size distribution, organic matter content and stable carbon isotopic ratio in soil organic matter. Radiocarbon ages for material dated from the section also shown.

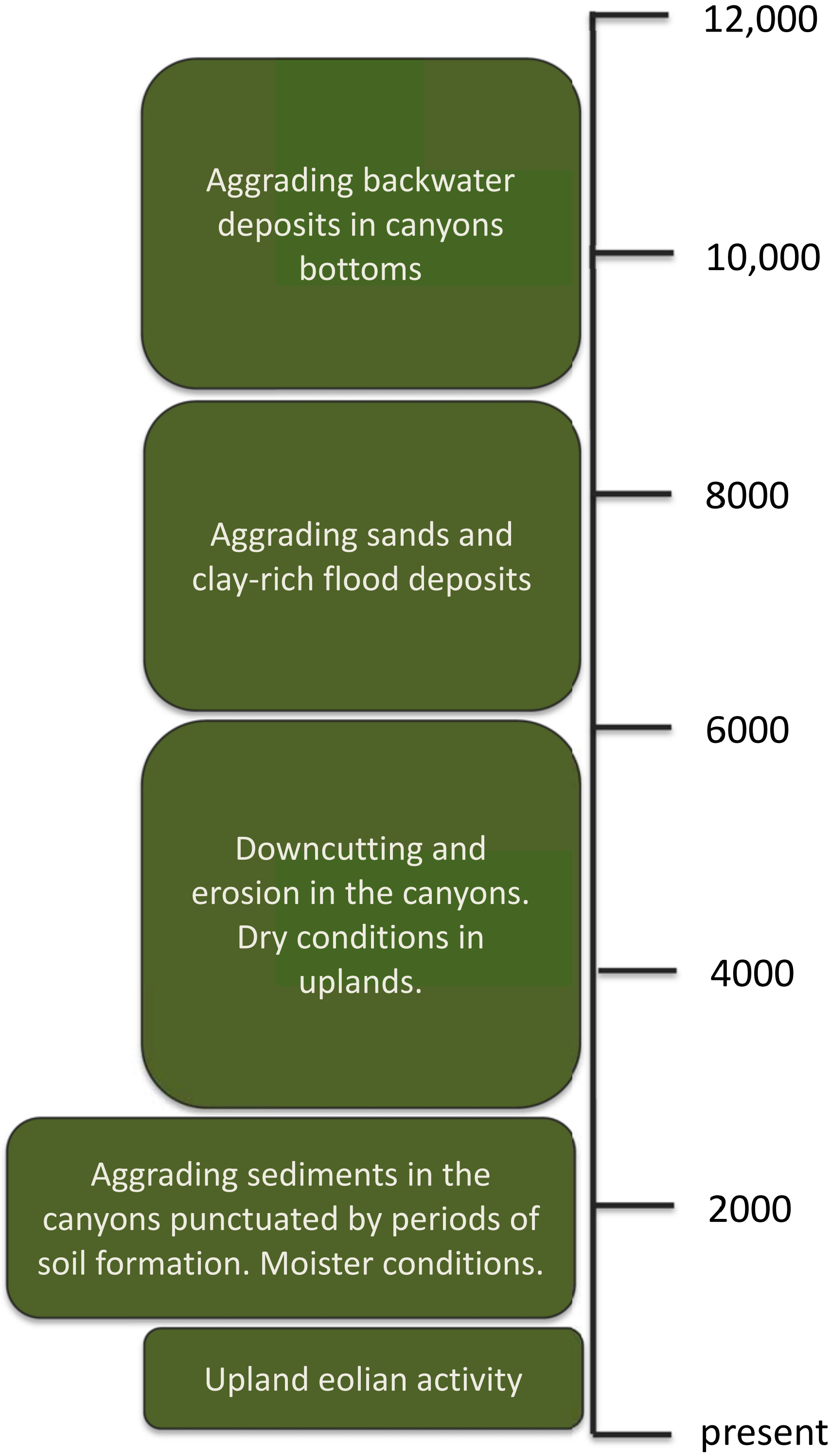

Fig. 12 – Reconstructed landscape evolution of the Caddo Canyon region from 12,000 ^{14}C yr. B.P. to present.

THE BuSCR MODEL: A METHOD FOR RECONSTRUCTING PAST CLIMATE OF THE GREAT PLAINS, U.S.A. BASED ON BURIED SOIL PROPERTIES

1. Introduction

The Great Plains, U.S.A. lack quantitative paleoclimatic data for the late Quaternary (14,000 cal. yr. B.P. to present), largely because two common sources of paleoclimatic data, tree ring and pollen records, are rare in the region (Hall and Valastro Jr. 1995; Baker, Fredlund et al. 2000; Holliday 2000). However, sequences of buried soils are commonly preserved in eolian and alluvial sediments on the Great Plains and have the potential to enhance the region's paleoclimate record (e.g. Nordt 1994; Holliday 1995; Mandel 2008). Buried soils have long been used to reconstruct past environmental conditions in the region qualitatively, where researchers describe past conditions as "warm" or "wet", for example, based on buried soil properties (e.g. Bryan and Albritton 1943; Judson 1953; Wendorf et al. 1955; Holliday 1995; Mandel 2008), but there have been few attempts to quantify such reconstructions.

Buried soils formed during periods of past landscape stability, and paleoenvironmental reconstructions using buried soils are based on the premise that bioclimatic factors exert great influence on soil genesis, with soil properties reflecting this influence. A quantitative paleoclimate proxy based on properties of buried soils would provide a new source of paleoclimate data in the region, one that has experienced significant change even during just the last 150 years. A more complete understanding of the Great Plains climate history over long timescales (thousands of years) also provides context for projected future climate change. Additionally, an enhanced paleoclimate data set for the Great Plains would help answer questions regarding patterns of spatial and temporal climate variability in the region. To help achieve that end, this study presents a numerical model to reconstruct past climates from buried soils.

Development of a paleoclimatic proxy based on buried soil properties depends on calibration to modern analogues. Calibration involves analysis of modern surface soils and historical climate records to determine the soil properties that are climate-dependent and to what extent. Calibration and its validity for developing paleoclimatic proxies rest on the principle of uniformitarianism. Uniformitarianism assumes that processes and relationships observed in the natural world today also operated in the past and throughout the period of interest. In the case of buried soils on the Great Plains, this means that modern soil-forming processes operated throughout the late Quaternary. If we assume this to be the case, then examination of soil-climate relationships in the region today will point us toward climate-dependent buried soil properties that would be useful as paleoclimatic proxies. Quaternary scientists commonly use statistical methods to develop models for hindcasting climate based on paleoclimate proxies (Bradley 1985). Typically, mechanistic models based on known natural and physical laws exhibit less error compared to statistical models when forecasting future trajectories of a natural system. When hindcasting, however, initial conditions of the system are unknown, and so statistical models provide strong results in paleoclimate models.

Soils form when a set of extrinsic factors alters a stable land surface over time. The dominant *in situ* factors of soil formation include climate, biology, topography, parent material, and time; together these determine the pedogenic pathway of a soil (Jenny 1941; Arnold 1965; Schaetzl and Anderson 2005). 'Pedogenic pathway' describes the set of soil-forming processes that lead to a given soil morphology, producing predictable soil properties (Arnold 1965). For example, melanization, a dominant pathway in semi-arid, mid-continental locations where grasses populate the terrain, produces dark A horizons and granular soil structure due to high soil organic matter content. In arid regions, calcification occurs as water evaporates from soils under severe moisture deficits, and calcium carbonate and salts precipitate out of solution and accumulate in the soil. Where a single soil-forming factor dominates pedogenesis, it may be possible to predict a soil's pedogenic pathway considering this factor alone and quantitatively link soil properties to that extrinsic forcing. For example, if climate is the dominant soil-forming factor and the other four factors of soil formation can be held relatively constant, soil properties can be quantitatively linked to climate (Jenny 1941, 1946, 1980).

Over the last century, scientists have repeatedly established that climate, in particular moisture balance, explains the geographic distribution of soils (Hilgard 1892; Jenny and Leonard 1934; Marbut 1935; Blumenstock and Thornthwaite 1941; Arkley 1967; Mather 1978). The water balance determines the amount of excessive precipitation available for leaching and the atmospheric demand for water or amount that evaporates from the soil. Leaching mobilizes clays and organic acids, creating E horizons and clay-enriched subsoils typical of Ultisols, Alfisols, and Spodosols in humid regions. High evaporative demand leads to precipitation of carbonates and salts in soils like Aridisols forming under arid climates. Rather than attributing soil properties to a single climate factor, studies over the last century point to the balance between precipitation and potential evapotranspiration as the key factor dictating soil properties, such as clay and carbonate depth functions (e.g. Jenny 1935; Arkley 1963; Rasmussen et al. 2005)

Several more recent studies have successfully quantitatively linked soil properties to climate. Specifically, calcium carbonate accumulations in the subsoil (McFadden and Tinsley 1985), organic carbon accumulation in A horizons (e.g. Rasmussen et al. 2005; Dai and Huang 2006; Gray et al. 2009; Scull 2010; Suuster et al. 2011), and clay translocation through the solum of surface soils (Rasmussen et al. 2005; Gray et al. 2009; Scull 2010) have been quantitatively linked to modern climate. Depth functions of carbonate content display a strong relationship with regional climates represented by water balance calculations (Arkley 1963; McFadden and Tinsley 1985). Mean depth to the zone of calcium carbonate accumulation has been reconstructed using water balance calculations to estimate the volume of water passing through depth increments in a soil (Arkley 1963). Further, compartment models that consider soil texture, porosity, and saturated hydraulic conductivity have found precipitation to be the most influential factor in determining mean depth of carbonate accumulation in a soil (Arkley 1963; McFadden and Tinsley 1985). Parton et al. (1994) created a model that predicted soil organic carbon (SOC) content with R^2 values of 0.75 to 0.93 by estimating turnover rates for three pools of SOC based on temperature, precipitation, evapotranspiration rates, chemical composition of litter, and clay content of the soil. Gray et al. (2009) used multiple regression to demonstrate that annual precipitation and silica content of the parent material predict clay content of the B horizon with moderate effectiveness on the global scale.

Climate-dependent soil properties must be carefully selected to study modern soil and climate relationships and this selection should include consideration of the potential for diagenesis (physical, chemical or biological change) in the buried environment. Soil properties preserved in soils buried for short durations (less than 10,000 years) and resistant to post-burial diagenesis are the most appropriate proxies. Post-burial diagenesis is the physical and chemical alteration of soil properties that can occur after burial (Birkeland 1999: 342). Examples of potential diagenetic changes in buried soils include compaction, cementation, mixing, transformations due to microbial activity, and soil welding; however, the potential for diagenesis is less for young buried soils (less than 10,000 years old) isolated from the effects of the external environment. Bt horizon properties and clay mineralogy in buried soils are quite resistant to post-burial alteration and, hence, are commonly used proxies for reconstructing paleoenvironments from buried soils. In contrast, soil pH and nitrogen content are closely linked to climate in modern surface soils (Jenny 1929, 1930; Gray et al. 2009), but are not persistent in the buried soil environment. Only soil characteristics that are persistent in buried soils and reflect climate conditions during formation of the soil are suitable for paleoenvironmental reconstructions as these properties are likely to reflect conditions during the period of soil formation, rather than modern state factors.

In this paper, we present a quantitative paleoclimate model for reconstructing past climates on the Great Plains, U.S. based on the properties of buried soils, hereafter called the BuSCR (**Bu**ried **S**oil **C**limate **R**econstruction) model. We searched soil profiles in the National Cooperative Soil Survey (NCSS) Characterization Database (http://ssldata.nrcs.usda.gov/) to identify appropriate modern soils to consider in relation to modern climate and water balance variables. Based on our study of modern soil and climate relationships, we calibrated a multiple regression model for application to buried soils. We statistically validated the BuSCR model by testing it on a series of modern soils not included in the regression analysis. The paper concludes with the results of the statistical validation and discussion of considerations for applying the model to buried soils.

2. Methodology

Development of the BuSCR model proceeded through the following multi-step process: 1) conducting a pilot study in Kansas, the results of which guided variable selection for the model, 2) choosing soils across the Great Plains that are modern analogues to soils commonly preserved in buried environments for the calibration dataset, 3) spatially joining modern soil and climate data, 4) analyzing the dataset for outliers and testing the data for the assumptions of linear regression, 5) conducting a multiple linear regression analysis to create the BuSCR model, and 6) statistically validating the model to determine errors. In the following section, we provide details on the methods employed for each of these steps.

2.1 Soil-climate pilot study

We developed the BuSCR model based on an analysis of modern soil-climate relationships across the Great Plains, beginning with a pilot study of soil-climate relationships across Kansas. Analysis of modern soil-climate relationships will indicate buried soil properties that may be useful paleoclimate proxies. We conducted the pilot study to determine the soil properties with the best potential for serving as proxies for climate on the Great Plains, U.S. Data from 31 soil pedons surrounding five long-term weather stations in Kansas were correlated with climate parameters. The five stations were selected to represent the east-west moisture gradient and dominant geology across Kansas, as well as in consideration of data availability and quality. Subsoil carbonate content (%), clay content (%), A horizon organic carbon content (%), and cation exchange capacity (CEC) of the A horizon and the subsoil (cmolc/kg) were extracted from the National Cooperative Soil Survey (NCSS) Characterization Database (National Cooperative Soil Survey 2013) for soils within 35 km of the five long-term weather stations. Only Alfisols, Entisols, Inceptisols, and Mollisols were included in the analysis because these represent the majority of soil orders on the Great Plains that form in environments where buried soils are commonly preserved.

To relate soil conditions to climate, we elected to use annual temperature and precipitation data, as well as a number of water balance variables. We used a daily water balance methodology

(Feddema et al. 2013) to assess thermal and moisture conditions, including changes in soil water storage, at five weather stations along an east-west transect across Kansas (Fig. 1). Daily records of minimum temperature (C), maximum temperature (C), and precipitation (mm) measured between 1900-2009 were used as input to the model at each location. The model partitions overland runoff from precipitation using the Soil Conservation Service (Soil Conservation Service 1972, Mather 1978) overland runoff model. From infiltrated precipitation and temperature, estimates of potential evapotranspiration (PE), snow accumulation and snowmelt (as implemented in Willmott et al. 1985), soil moisture conditions, actual evapotranspiration (AE), moisture surplus and deficit conditions were estimated on a daily basis. PE was estimated using the Thornthwaite equation (Thornthwaite 1948, Thornthwaite and Mather 1955, Feddema 2005, Feddema et al. 2013). Key variables for assessing soil-climate relationships across Kansas were aggregated by season and by year, including mean annual precipitation (MAP), mean annual temperature (MAT), annual average PE, annual average AE, AE/PE ratio, annual moisture surplus, annual moisture deficit, and annual moisture index (I_m). See Willmott and Feddema (1992) for the formulation of the moisture index used throughout this study. Seasonal precipitation, temperature, moisture surplus and deficit, and moisture indices were included in the analysis as well because seasonal moisture balances influence pedogenesis (Amundson et al. 1997; Breecker et al. 2009). For example, because moisture availability during the growing season limits plant productivity in semi-arid regions, spring and summer I_m may be more influential on A horizon organic carbon content than MAP or the annual I_m in this region. We conducted a correlation analysis to identify the soil and climate variables with the strongest relationships. Highly correlated soil and climate variables identified in the pilot study (see results) were included in the multivariate regression analysis used to develop the model.

2.2 Developing the BuSCR Model

2.2.1 Soil Data Selection

For this study, we defined the Great Plains by geographic and climatic parameters, specifically parts of the Great Plains or Central Lowland physiographic provinces that receive less than 1000 mm MAP and have an annual I_m less than or equal to 0.2. Within the geographic region defined by these criteria, we selected all Alfisols, Entisols, Inceptisols, and Mollisols from the NCSS database. Most soils with laboratory characterization data in the NCSS database did not include information on soil parent material. We linked parent material information to each soil for all soil series extracted from the NCSS database by referencing Natural Resource Conservation Service Official Soil Descriptions (Soil Survey Staff 2013).

Development of the BuSCR model is based on the Jenny (1941) state factor approach, where:

$$S = f(cl, o, r, p, t, \ldots)$$

Here, the five factors of soil formation, climate (cl), organisms (o), relief (r), parent material (p), and time (t), determine a soil's properties (S). In order to quantitatively relate soil properties to climate, we needed to ensure that most of the variation in the soils' properties was attributable to climate and not the other four state factors. To accomplish this, we filtered the soil data set based on slope, parent material, and taxonomic classification at the order and great group level to ensure that only relatively young soils that formed in flat to gently sloping eolian and alluvial deposits were included in the analysis. Biota was not considered in data set development since the influence of biology and climate are difficult to separate in semi-arid environments. Instead, we accounted for biotic influences on soil formation by including soil variables in the regression analysis that are representative of biotic production.

Specific filter criteria were applied as follows. First, we controlled for relief by determining the elevation and slope for each soil using a U.S. Geological Survey 10-meter resolution digital elevation model (Gesch 2007) in ArcGIS. Slope processes dominate pedogenesis on slopes greater than 12 percent (Schaetzl and Anderson 2005: 475) and high elevation sites (above 1400 meters) could exhibit climate regimes uncharacteristic of the Great Plains. Therefore, we removed from the data set soils on slopes greater than 12 percent or above 1400 meters elevation. Next, to control for parent material, we only included soils that formed in eolian or alluvial deposits based on Official Soil Descriptions for each soil series. Also, we applied a standardization considering soil saturated hydraulic conductivity (K_{sat}). K_{sat} impacts depth of the wetting front and is largely determined by the particle size distribution of the parent material in which a soil forms. The K_{sat} standardization attempts to control for differences in water flux among soils with different parent materials. Application of the K_{sat} standardization greatly improved correlations between carbonate and clay depth functions and climate in the pilot study (see Results). To control for time, soil taxonomic classifications were used; excluding Entisols, Inceptisols, Mollisols, and Alfisols classified as paleo- at the great group level ensured ancient soils were not included in the analysis.

2.2.2 Soil-climate database development

We queried from the NCSS Soil Characterization Database soil data by horizon and filtered the data based on the above mentioned selection criteria. Horizon-level data was used to calculate variables for the regression analysis. Specifically, we obtained horizon depths and profile designations, organic carbon content of the A horizon, CaCO3 content of B and C horizons, fine and total clay content of all horizons, and total sand content for all horizons for each soil pedon. Using these data, organic carbon content in the A horizon ($SOC_{A\ hzn}$), thickness of the A horizon (A_{thick}), depth to the top of subsoil pedogenic carbonates ($CaCO_{3\ dep}$), and the eluviation index (I_e) were calculated.

Because soils can contain multiple A horizons with variable thicknesses, we calculated a depth-weighted average organic carbon content of the A horizon. $SOC_{A\ hzn}$ is closely tied to temperature (Parton et al. 1987, Kirschbaum 1995), moisture index (Rasmussen et al. 2005), and net primary productivity (Parton et al. 1987, Rasmussen et al. 2005). Inclusion of $SOC_{A\ hzn}$ in the analysis should, therefore, provide a bioclimatic signal on the Great Plains. A_{thick} was also recorded for each soil. This variable was included as a potential additional control for time.

$CaCO_{3\ dep}$ and the I_e (Cremeens and Mokma 1986) were determined using the aforementioned data from NCSS and standardizing these values using an estimated K_{sat}. $CaCO_{3\ dep}$ was determined for all soils containing pedogenic carbonates. We defined pedogenic carbonates as carbonates contained in Bk horizons or in B or Bw horizons where the ratio of carbonates in the C horizon(s) to carbonates in the B horizon(s) is less than or equal to one. I_e was calculated for soils with illuviated clays. The I_e estimates the degree of clay eluviation in a soil by comparing fine and total clay content in surface horizons to fine and total clay contents in the subsoil:

$$Ie = \frac{Fine\ clay(B_t,\ B\ or\ B_w)\ /\ Total\ clay(B_t,\ B\ or\ B_w)}{Fine\ clay\ (A\ \&\ E)\ /\ Total\ clay(A\ \&\ E)}$$

A soil showed evidence of clay illuviation if it contained a B, Bt, or Bw horizon and had an I_e greater than 0. $CaCO_{3\ dep}$ and I_e were standardized by the estimated K_{sat} (cm/day) of the soil. We employed a pedotransfer function to empirically estimate K_{sat} for each soil (Cosby et al. 1984, Tietje and Hennings 1996). Specifically, mean clay and sand contents for the upper 150 cm of each soil were determined, and K_{sat} was calculated using these values and the following equation (Cosby et al. 1984):

$$K_{sat} = 60.96 * 10^{(-0.6+0.0126s-0.0064c)}$$

where s = sand content (%), c = clay content (%). K_{sat} for all soils in the database averaged 65.89 cm/day. We divided the K_{sat} of each individual soil by the average of the database to find a normalized K_{sat} with reference to the database. $CaCO_{3dep}$ and I_e were standardized by dividing by the normalized K_{sat} determined for each soil.

2.2.3 Development of the Illuviation/Calcification Index

In order to represent the full range of moisture conditions and precipitation regimes across the Great Plains, the clay and carbonate data were combined into a single index called the Illuviation/Calcification index (I_{ic}). Soils were identified as illuviated soils if clay illuviation was evident, and calcified soils if they had pedogenic carbonates in the subsoil. We calculated I_{ic} by converting K_{sat} standardized I_e and $CaCO_{3\ dep}$ to unitless numbers. I_e (standardized for K_{sat}) ranged from 0.4 to 9.0. These values were converted to unitless numbers between 0-1 by calculating:

$$\textit{Illuviation index} = I_e\ (K_{sat}\ \textit{standardized}) / 10$$

Standardized depths to top of the carbonate accumulations ranged from 5.2 to 545.0. Unitless values ranging from 0 to 1 were calculated for the carbonate data using:

$$\textit{Calcification index} = 1 - [(550 - CaCO_{3\ dep}\ (K_{sat}\ \textit{standardized})) / 550]$$

All soils in the dataset had either pedogenic carbonates or illuviated clay in the subsoil. Those categorized as calcified contributed a unitless negative value to the index variable, and those designated as illuviated contributed a positive unitless value to the index variable. Therefore, I_{ic}

provides an estimate of amount of clay illuviation or pedogenic carbonate accumulation in the soil with values ranging from -1 to 1.

2.2.4 Analysis of Outliers

Outliers in the data set were examined individually. Based on this review, we established three additional parameters to control for time and parent material effects. First, potentially polygenetic soils were identified and removed from the dataset. We defined polygenetic soils as soils with a Btk horizon. The presence of a Btk horizon indicates that the period of soil formation probably included two disparate climatic regimes - one more humid and another more arid. Consequently, the properties of the soil do not reflect a single climate regime, and cannot be attributed to one climate condition. Second, soils with subsoil carbonates inherited from the parent material were removed. Carbonate-rich parent materials are quite common across the Great Plains, and subsoil carbonates inherited from the parent material differ genetically from pedogenic carbonates. Hence, carbonates in the susbsoil could not be assumed to be pedogenic. We defined pedogenic carbonates as subsoil carbonates in Bk horizons or in other B horizons when the ratio of carbonates in the C horizon to carbonates in the B horizon was less than one. Only soils with pedogenic carbonates according to the above definition were selected for model development. Third, we removed soils with very rapid, rapid, and moderately rapid permeability (equates to K_{sat} > 120 cm/day, Pacific Northwest Soil Survey Region 2006) from the data set because clays and carbonates do not readily accumulate in soils with rapid saturated hydraulic conductivities. The final data set included 140 relatively young soils across the Great Plains that formed in settings similar to those in which buried soils formed during the Quaternary.

2.2.5 Controls for Oversampling

Long-term annual climatologies (1950-1999 means) from the University of Delaware Center for Climatic Research's Climate Data Archive were used to characterize the Great Plains climate for the statistical analysis. The Willmott & Matsuura monthly and annual climatologies (Willmott and Matsuura 2001a, 2001b) provide precipitation, temperature, and moisture index estimates for 0.5-degree grid cells globally. Climate variables were chosen for the final analysis based on the results of the pilot study and included mean annual temperature (MAT), mean annual precipitation (MAP), and annual moisture index (I_m). We spatially joined the climate data to the modern soil data using ArcGIS.

Review of the spatially joined soil and climate data confirmed a priori knowledge that soil scientists commonly oversample locations with easy access at long-term research sites, and/or where soils representative of a specific soil order or soil-geomorphic relationship are known to occur. Including groups of similar soils in the analysis could bias it due to overrepresentation of certain soil-climate relationships and consequential underrepresentation of others. To correct for this local oversampling bias, we calculated median values for spatially clustered groups of soils. Specifically, we found the median $SOC_{A\ hzn}$, $A_{thick,}$ and I_{ic} for soils located in the same 0.5-degree grid cell that had the same parent material. We used the single median value to represent the group of similar soils, rather than the individual values of all the soils, in the statistical analysis.

2.2.6 Statistical Analysis and Model Validation

To develop the final BuSCR model, we performed a multivariate regression analysis. The first step of the analysis involved testing for violations of the assumptions of multiple regression. Specifically, we examined histograms, scatterplots, and Q-Q plots to determine if the data satisfied the assumptions of normality, linearity, and homoscedasticity. The relationships between all soil and climate variables were relatively linear, but the data distributions of all soil variables exhibited positive skews, violating the assumptions of both normality and homoscedasticity. To correct for non-normality and satisfy the assumption of homoscedasticity, we applied transformations to the data.

Stepwise multiple linear regression was used to develop the BuSCR model. We randomly selected, using a random number generator in Microsoft Excel, three-quarters of the sample (n=105) to run the stepwise multiple regression and generate the regression functions. The remaining cases (n=35) were used to statistically validate the model. We stepped multiple linear regression analyses forward and backward for explaining the variance in MAP (mm), MAT (°C), and I_m using all combinations of transformed and non-transformed soil variables (A_{thick}, I_{ic}, and $SOC_{A\ hzn}$). The function that explained the greatest amount of variation for each climate variable based on the adjusted R^2 and F statistic values was selected and used to estimate the amount of error in each model.

In order to statistically validate the models and estimate error, we calculated MAP, MAT, and I_m for the 35 soils not included in the regression analysis using the best model for each respective variable. We then compared the estimated climate to observed climate for each case and calculated the average difference in predicted and observed values known as the mean absolute error (MAE), the revised index of agreement (Willmott et al. 2012), and root mean square error (RMSE) for each model. Systematic and unsystematic RMSE was also calculated in order to partition and deduce potential sources of error in the model (Willmott et al. 1985). Comparison of these error estimations, adjusted R^2, and F statistics for the models determined selection of the final BuSCR model.

3. **Results**

3.1 Pilot Study

A statistical analysis of soil properties and climate across Kansas indicated that several soil properties are significantly correlated with climate in the central Great Plains. Table 1 shows correlation coefficients for soil property-climate characteristic pairs, as well as their significance levels. We selected soil properties for the pilot study that had been quantitatively linked to climate in previous research and that represent climate-driven pedogenic processes common on the Great Plains. Calcification, organic matter accumulation, mineral weathering, and clay illuviation, also known as lessivage, were the particular foci for the pilot study.

Carbonates in the subsoil are significantly negatively correlated with MAT, MAP, PE, AE, annual I_m, and summer I_m (Sig < .0001). The highest correlated pairs were: thickness of the Bk horizon with MAT (-0.681) and PE (-0.681); subsoil carbonate content with AE (-0.732); and maximum carbonate content in the subsoil with PE (-0.701). Note that PE and MAT are closely related because the Thornthwaite equation uses daily average temperature to estimate PE. These results indicate that the subsoil carbonate content of soils in the central Great Plains have a strong, significant negative correlation with annual moisture availability, where carbonate content increases when moisture availability decreases.

$SOC_{A\ hzn}$ exhibits only a few moderately significant correlations with climate across Kansas, with $SOC_{A\ hzn}$ most strongly related to summertime climate. Summer moisture surplus showed the strongest correlation (0.62, Sig < .001). Summer I_m (0.572), summer AE/PE (0.567), the annual range in I_m (-0.577), and annual moisture deficit (-0.589) were also moderately significantly correlated to $SOC_{A\ hzn}$. The correlation analysis shows that accumulation of organic carbon in the A horizon of soils in Kansas is significantly related to moisture availability, especially during the summer growing season. With increased moisture, $SOC_{A\ hzn}$ increases.

Soil CEC was also moderately significantly related to climate in Kansas, with subsoil CEC (CEC_{sub}) exhibiting stronger relationships to climate than A horizon CEC ($CEC_{A\ hzn}$). $CEC_{A\ hzn}$ was most strongly tied to summer deficit (-0.639, Sig < .05). CEC_{sub} was significantly (Sig < .001) related to annual moisture deficit (-0.702), annual range in I_m (-0.669), AE/PE (-0.655), summer deficit (-0.663), and summer AE/PE (-0.711). CEC_{sub} was significantly negatively related to available water, with lower subsoil CECs occurring where moisture deficits, measured in terms of I_m, AE/PE, and deficit, were higher.

Subsoil clay content was highly significantly related to moisture as represented by the climate parameters MAP, AE, surplus, and I_m. The maximum clay content was more significantly related to climate than mean subsoil clay content, and produced the highest correlation coefficients observed in the pilot study. Specifically, maximum subsoil clay content was highly significantly (Sig < .0001) and positively correlated with MAP (0.780), AE (0.747), surplus (0.771), I_m (0.775), AE/PE (0.757), and summer I_m (0.761). These results indicate that clay content in the subsoil, specifically the maximum clay content found in the subsoil, is strongly significantly related to the amount of available water annually, with subsoil clay content increasing with moisture surplus.

After reviewing correlations between soil characteristics and climate across Kansas, we chose to focus development of the BuSCR model around three primary soil properties and three annual climate parameters: carbonate in the subsoil, clay content in the subsoil, organic carbon content in the A horizon, MAT, MAP, and I_m. These soil properties were as highly or more highly correlated with MAT, MAP, and I_m as compared with correlations against deficit, surplus or seasonal moisture balance estimates. In fact, the highest correlations observed, with accompanying high significance levels, were for maximum clay content in the subsoil correlated with annual MAP and I_m. Subsoil carbonate content exhibited strong, significant correlations with temperature, as illustrated by highly significant correlation coefficients with MAT, PE, and AE. While organic carbon content in the A horizon exhibited the weakest correlations, we included it in the analysis in order to potentially account for variation within the data not observed at the scale considered in the pilot study. These three soil properties serve as indicators of the occurrence and relative magnitude of the pedogenic processes calcification, lessivage, and organic carbon accumulation, respectively.

These variables were also selected based on data availability and access: particle size distribution, carbonate content, and organic carbon in the A horizon are commonly measured values recorded in the NCSS database. Additionally, these three soil properties are generally resistant to post-burial alteration at timescales less than 10,000 years (Birkeland 1999). The Willmott and Matruura database includes data for all three climate variables, and MAP and MAT are also commonly available or easily calculated for any long-term weather station. In addition, I_m is readily calculated from mean monthly temperature and precipitation statistics using the Thronthwaite (1948) methodology. Therefore, we felt confident we could generate a sizable sample of soils with the data we needed and an accompanying climate data set by focusing on these three soil properties and three climate parameters.

The results of the correlation analysis not only helped us identify the soil properties most significantly related to climate, but also highlighted potential confounding factors. Specifically, the correlation analysis called attention to the impact of parent material on subsoil texture and movement of water through the soil. The following two examples illustrate this issue. While organic carbon generally displayed a significant positive correlation with surplus and a negative relationship with deficit, soils surrounding the weather station at Larned were exceptional. These soils had organic carbon contents in the A horizon lower than would be expected considering Larned's annual moisture balance (Fig. 2). As can be seen in Fig. 1, Larned is located in a unique region geologically, in the sand hills and dunes of the Arkansas River Lowlands. Soils here form in very sandy parent material and, as a result, have lower water holding capacities. Lower water holding capacities decrease plant available water and NPP, contributing to low organic carbon contents in soil A horizons. Additionally, sandy soils offer SOC little physical protection from microbial attack, increasing rates of SOC loss from A horizons. In contrast, soils in the area of Ottawa dominantly form in clay-rich, Cretaceous-age limestones and shales. Consequently, soils in this region are commonly fine-textured. The correlation analysis shows that subsoil clay content was positively correlated with moisture surplus, but soils surrounding the Ottawa weather station had higher clay contents than expected considering climate (Fig. 3). This finding indicates again the strong influence parent material exerts on soil properties. Our observation of parent material as a potential confounding factor to quantifying soil-climate relationships on the Great Plains led to development of the K_{sat} standardization outlined in the methods section.

3.2 Finalized Soil-Climate Dataset

An initial query of soils from the NCSS database considering slope, elevation, taxonomic classification, climate, and physiographic region identified 666 soils appropriate for the analysis. Lack of data needed to calculate the soil variables for the analysis resulted in deletion of 417 soils, leaving 249 soils. After calibrating for parent material effects, eliminating polygenetic soils from the dataset, and combining clusters of soils to control for oversampling, the final data set consisted of 140 cases. Of these, 105 were used to develop the BuSCR model and 35 were left out of the multiple regression analysis and used only to statistically validate the model (Fig. 4).

The data set spatially covers the Great Plains region adequately and also represents a range of soil-forming environments, climates, and soil types (Fig. 4). Most of the soils in the finalized data set formed in alluvium (n = 73) or loess (n = 59), and a few formed in other types of eolian deposits (n = 8). Mean annual temperatures ranged from 4.8 °C in North Dakota to 17.7 °C in the panhandle of Texas, and MAP ran from 315 mm in central Montana to 984 mm in eastern Kansas. Central Montana also had the driest I_m at -0.5, and a location in western Missouri and another in eastern Kansas were wettest with I_m values of 0.2. K_{sat} varied from less than 7 cm/day to 116 cm/day. $SOC_{A\ hzn}$ ranged from 0.4 % to over 5.0 %. Fourty-two soils showed signs of clay illuviation, and we calculated an illuviation index for these soils. We designated 98 soils as calcified due to presence of measurable pedogenic carbonate accumulations in the subsoil and calculated the calcification index for these soils.

As in the preliminary study in Kansas, soil properties representing the pedogenic processes of calcification and lessivage were most highly correlated with climate on the Great Plains, while organic carbon accumulation exhibited lower correlations with climate. $CaCO_{3\ dep}$ and I_e, both standardized for soil K_{sat}, were significantly correlated with MAP and I_m exhibiting positive relationships to increased moisture (Fig. 5). $CaCO_{3\ dep}$ was most strongly correlated with MAP (r = 0.542, $p = 9.88 \times 10^{-10}$). In comparison, I_e exhibited high correlation coefficients and significance levels for MAP (r = 0.642, $p = 3.27 \times 10^{-8}$) and I_m (r = 0.628, $p = 7.76 \times 10^{-8}$). Both $CaCO_{3\ dep}$ and I_e were weakly related to MAT (r = 0.226, $p = 0.017$ and r = 0.126, $p = 0.337$, respectively). $SOC_{A\ hzn}$ did not generally exhibit strong correlations with MAP or I_m. However, $SOC_{A\ hzn}$ displayed a strong negative relationship with MAT across the Great Plains (r = -0.478, $p = 2.28 \times 10^{-9}$; Fig. 6). This correlation with MAT is a marked improvement compared to observations from the pilot study, which had a much lower variation in temperature. This indicates that variation in temperature and $SOC_{A\ hzn}$ at the spatial scale of the Great Plains is higher than between site variation. In contrast, site-to-site variation was great enough at the spatial scale of Kansas to blur any temperature/organic carbon relationship that in fact occurs at coarser scales. The I_{ic} was more significantly correlated with climate than either of the soil variables (I_e $CaCO_{3\ dep}$) alone, illustrating the importance of combining these variables into a single index to represent the full spectrum of moisture and soil conditions across the Great Plains. I_{ic} was most highly correlated with MAP (r = 0.592, $p = 1.28 \times 10^{-14}$), and also strongly related to patterns in I_m across the region, with r = 0.535 *($p = 9.86 \times 10^{-12}$).*

None of the soil properties considered in this study displayed normal distributions. Both results of the Shapiro-Wilk test for normality and visual displays of the data indicated that all three soil variables were significantly positively skewed (Fig. 7). In order to correct for skewness, transformations were applied to all three variables. Re-running the Shapiro-Wilk test on the transformed data indicated that a log transformation was most suitable for handling the slight positive skews of A_{thick} and I_{ic}. We used a square root transformation to correct for the more severe positive skew observed for data on $SOC_{A\ hzn}$.

3.3 Multiple Regression Analysis or Model Development

Both the transformed and non-transformed data sets were stepped through the multiple regression analysis to develop the BuSCR model. Table 2 provides the results of the multivariate regression analysis. A model using I_{ic}, log transformed A_{thick}, and $SOC_{A\ hzn}$ accounted for 40% of the variation in I_m. This model generated the highest F-statistic and lowest p-value of any models for any climate variable. The best model for predicting MAP uses I_{ic}, log of A_{thick}, and square root of $SOC_{A\ hzn}$. The model accounts for 38% of the variability in MAP and also was highly statistically significant, with a F-statistic and p-value just slightly lower than the model for I_m.

The stepwise multiple regression analysis showed that soil properties considered here account for far less variance in MAT than in MAP and I_m. The most effective model for predicting MAT uses I_{ic}, log of A_{thick}, and square root of $SOC_{A\ hzn}$ to account for 30% of MAT variance ($F = 16.09$, $p = 1.27 \times 10\text{-}8$).

3.4 Model Validation Statistics

We applied the three regression equations to all 35 soils not used to create the models and compared the BuSCR predicted values to observed climate values. Scatterplots of observed versus predicted values show that all three models underestimate the magnitude of the dispersion in the data (Fig. 8). We calculated mean predicted and observed values, standard deviation of the observed and predicted data sets, and several statistical estimates of model error useful for comparing the three data sets (Table 3). While differences in predicted and observed means ranged from only 2% (MAT) to 10% (I_m), all observed data sets exhibited greater variability, as indicated by standard deviations almost double the variability compared to predicted values. The mean predicted values were slightly higher than mean observed values for all three models, these differences varied systematically with climate. For example, the MAP regression model predicted an average MAP of 587.16 mm (s.d. = 87.40 mm) for the 35 soils compared to a mean observed MAP of 572.22 (s.d. = 155.96). We estimated the model's mean average error (MAE) as 97.49 mm; however MAP was underpredicted by an average of -272.51 for soils with MAP greater than 800 mm, and the model over-predicted rainfall amounts by 136.94 in regions with less than 400 mm MAP. Similarly, the model predicted a mean I_m of -0.17 (s.d. = 0.10), compared to a mean observed I_m of -0.19 (s.d. = 0.19). MAE for the I_m model was 0.12, but the model underestimated locations with a I_m greater than 0.1 by -0.24, and over-predicted I_m by 0.17 in locations where I_m is less than -0.4. The MAT model predicted an average MAT of 10.10 °C (s.d. = 1.8 °C) compared to an average observed MAT of 9.92 °C (s.d. = 3.17 °C) with a MAE of 1.82, but in cold locations the model over-predicted MAT by 2.68 °C, while in warm regions temperatures were under-predicted by -4.04 °C.

The revised index of agreement (d_r) corroborates the MAE and standard deviations - all three models underestimate climate variability observed across the Great Plains. The revised index of agreement compares model estimates to pair-wise matched observations and provides a unitless estimate of model performance ranging from -1 to 1, where $d_r = 0$ indicates a perfect model fit (Willmott et al. 2012). The d_r was 0.64, 0.64 and 0.62 for the MAT, MAP, and I_m models, respectively. While observed and predicted values most closely agreed with the I_m model, it performed only slightly better than the other two. All three underestimate the deviation of observed climate from mean observed climate by about 50%.

Examination of RMSE and its components indicates that there may be room for improvement in the models. Comparison of the systematic and unsystematic components of root mean square error confirms that systematic error exists in all three models, and that this error accounts for a much greater proportion of total error compared to unsystematic error. $RMSE_s$ accounts for 68%, 69%, and 67% of the total error in the MAT, MAP, and I_m models, respectively. $RMSE_u$ estimates the amount of random error in the model, while $RMSE_s$ estimates the proportion of systematic error, which may be corrected through improvements to research design, data collection, or sampling procedures. These error estimates are confirmed by the systematic patterns of error apparent from the scatterplot of observed versus predicted values (Fig. 9) and MAE variation with climate. In short, the models accurately predict the location of mean climate conditions based on soil properties, but only moderately estimate climates that deviate severely from the mean.

4. Discussion

4.1 Model Effectiveness and Potential Sources of Model Error

Results of this study show that there is great potential for using buried soils as proxies for paleoclimate. The BuSCR model uses four soil properties to reconstruct MAP, MAT, and I_m at a single location with highly statistically significant results. Additionally, the four soil properties needed to apply the model are easily obtained from soil profile descriptions and/or standard laboratory analyses. Further, statistical validation of the model showed that while model errors increase at the edges of the Great Plains climate regime, differences in predicted compared to observed values were exceptionally low across most of the region (see Fig. 8, Table 3).
While there is great promise for this methodology, our tests of the model also suggest that there is room for significant improvement. Specifically, most of the observed model error is systematic in all cases, suggesting that the climate variability was underpredicted for the three simulated climate variables. Our analysis suggests that much of this error has to do with the selection of soil samples for this study, and is also related to other factors that influence soil properties across the study region.

We have identified several potential sources of the systematic error in the BuSCR model. First, sampling of soils across the Great Plains and analyses in NRCS laboratories has depended upon the historical efforts of NRCS personnel and soil surveys across the region. Therefore, soils included in the database were sampled using a variety of strategies, frequently opportunistically rather than systematically or randomly, and for a multitude of different purposes and projects. The map of soils considered in this study (Fig. 4) illustrates the spatial bias present in the NRCS database. The majority of the soils included in this analysis were sampled from the central

Plains. Inadequate representation of soils that formed under more extreme (wet/dry, hot/cold) climatic regimes could cause systematic errors, such as those observed in the BuSCR model. While the nature of the database and its purpose is not ideal for the purpose of statistical modeling, we preferred the detailed profile, characterization, and location data it provided over a spatially interpolated soil map for developing the BuSCR model and determining its potential effectiveness for reconstructing past climate.

Second, at the ends of the climate spectrum, where the largest errors in the models occur in a predictable fashion, vegetation may alter pedogenic pathways. For example, in cold regions, podsolization commonly occurs under evergreen vegetation. In hot semi-arid regions, low net primary productivity would affect organic matter content in soils. The model does not account for differences in vegetation across the region. Therefore, variability in soil properties due to systematic changes in vegetation not modeled by BuSCR may be a source of error.

Third, underestimation of the climate variability of the Great Plains may simply be due to inclusion of soil properties not correctly calibrated to the climate range of the region. Including other soil properties that account for additional variation in climate may improve the model and decrease the systematic error. Splitting the model into regional versions, one for application to soils that formed in more humid regimes and another for more arid regions could also decrease systematic error.

4.2 Sub-regional versus regional relationships

The strength of correlations of soil properties to climate decreased when considering the entire Great Plains region compared to the sub-region of Kansas. This likely results because the Great Plains region includes soils that formed under a greater variety of soil-forming conditions than in Kansas. We did not control for vegetation or parent material mineralogy, both of which would vary more across the Great Plains than across Kansas alone. Not controlling for these two factors of soil formation across the Great Plains may have blurred the climate signal and its relationship to soil properties. Further, even with great care taken to control for other differences in soil-forming factors, as outlined here in the methods, these efforts depend on the accuracy of the data used to categorize soil-forming factors. NRCS Official Soil Descriptions (OSDs) were used to determine soil parent material and deduce soil age. In some cases, the OSDs listed multiple possible parent material sources. Consequently, we used a generalized parent material designation when the OSD listed multiple parent material sources. Further, use of OSDs to designate a soil's parent material and probable age depends on correlation of the soil to the appropriate soil series. Correlation of a soil classified in the field to a soil series remains a subjective undertaking, occasionally forcing soil scientists to identify the soil series that best fits field observations, even if the fit is not ideal. Therefore, there may be errors in our soil age and parent material estimates due to poor sources of data on these soil-forming factors.

Soils in the data set we examined for the entire Great Plains formed under a much larger range of climates than are present in Kansas. Specifically, soils in the Great Plains data set formed under a wider temperature range (12.9 °C) compared to the soils sampled in Kansas (3.4 °C). This difference undermines our assumption that soil formation is controlled by the three primary pathways considered in this analysis (organic matter accumulation, lessivage, and calcification). Consequently, crucial pedogenic pathways not apparent in the pilot study, such as oxidation, may nonetheless drive soil formation in some parts of the Great Plains not represented in the pilot study. No variables included in the modeling account for pedogenic pathways uncommon throughout the Plains.

4.3 Considerations for applying the model

There are several constraints to applying the model considering problems with preservation of soil properties in a buried context and the parameters of the data used to create it. Soils found in buried contexts sometimes have undergone post-burial alteration, and this may prevent application of the model. The BuSCR model cannot be utilized on truncated buried soils missing an A horizon. Measures of organic carbon content in the A horizon and depth from the surface to carbonate accumulations are both required inputs for the model, and these cannot be accurately estimated on soils with upper portions of the soil profile missing. Also, buried soils which exhibit evidence of overprinting of clays or carbonates from the overlying material cannot be used in the model. If overprinting has occurred, estimates of depth to carbonates and the I/E index would be incorrect.

We developed the model using a sample data set that satisfied a variety of criteria. Hence, this model should be applied to buried soils that satisfy similar criteria. Specifically, the model should not be applied to soils that exhibit the following: 1) evidence of polygenesis, such as the presence of a Btk horizon or a Bk over a Bt or Btk horizon, 2) a rapid K_{sat} (i.e. $K_{sat} > 120$ cm/day), 3) Bg, Bss, Bn, By, or Bz, horizons or with these overlying Bt, Bw, B, or Bk horizons, and/or 4) no evidence of pedogenic clays or carbonates. The presence of a Bt, Bw or B horizon were indicators of the presence of pedogenic clays, and pedogenic carbonates were identified when the ratio of carbonates in the C horizon compared to the B horizon was less than or equal to one. If Bg, Bss or other horizons are observed below Bk or Bt horizons, the model may be applied only considering the upper Bt or Bk horizon when estimating model parameters. In addition to considerations regarding the specifics of the buried soil's properties, it must be remembered that the BuSCR model was developed for use on soils in the Great Plains that formed under climates with I_m ranges from 0.2 to -0.5. In short, validity of the reconstructed climate depends upon appropriate application of the model. Therefore, careful consideration of the dominant pedogenic pathway and the probable pedogenic history of the buried soil should be considered to determine if it is an appropriate soil for the BuSCR model.

5. Conclusion

In this paper, we present the BuSCR model, a multiple regression model which uses A horizon organic carbon content, depth to carbonate accumulations, and calculation of the I/E index to reconstruct annual I_m, MAP, and MAT for the Great Plains, U.S. The model is highly statistically significant ($p < .0001$) and estimates 40% of the variability in I_m, 38% of the variability in MAP, and 30% of the variability in MAT using these three soil properties. We tested the model by reconstructing climate based on properties of 35 soils sampled from across the Great Plains. The results of the test indicate that the model estimates mean conditions quite accurately, with MAE = 0.12, 97.5 mm, and 1.8 °C for annual I_m, MAP, and MAT, respectively. However, it severely under-predicts deviations from the mean or conditions on the arid and humid extremes of the Great Plains climate envelope. Comparison of $RMSE_u$ and $RMSE_s$ indicates that approximately two-thirds of the error in the model is systematic, suggesting there is room for improvement. Improved sampling criteria for selecting soil profiles to develop and calibrate the model and including variables in the model that account for variation in vegetation or age across soils could significantly improve the model. We created this model for application to buried soils across the Great Plains, in order to improve our understanding of past climate conditions in the region where few other paleoclimate records have been recorded. The results of our analysis indicate that while there may be room for improvement in the BuSCR model, there is also great promise for applying it. Testing the model through application to buried soils across the region and comparing the results to other paleoclimate proxies is a natural next step for assessing its validity.

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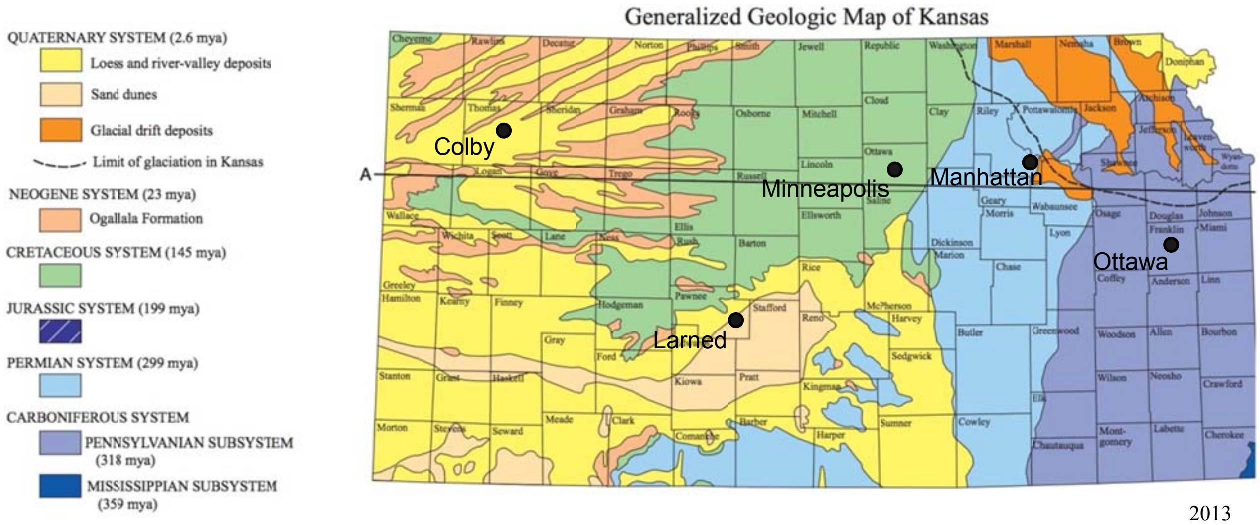

Fig. 1 - Locations of long-term weather stations used for the pilot study. All soils in the pilot study were located within 35 km of a long-term weather station.

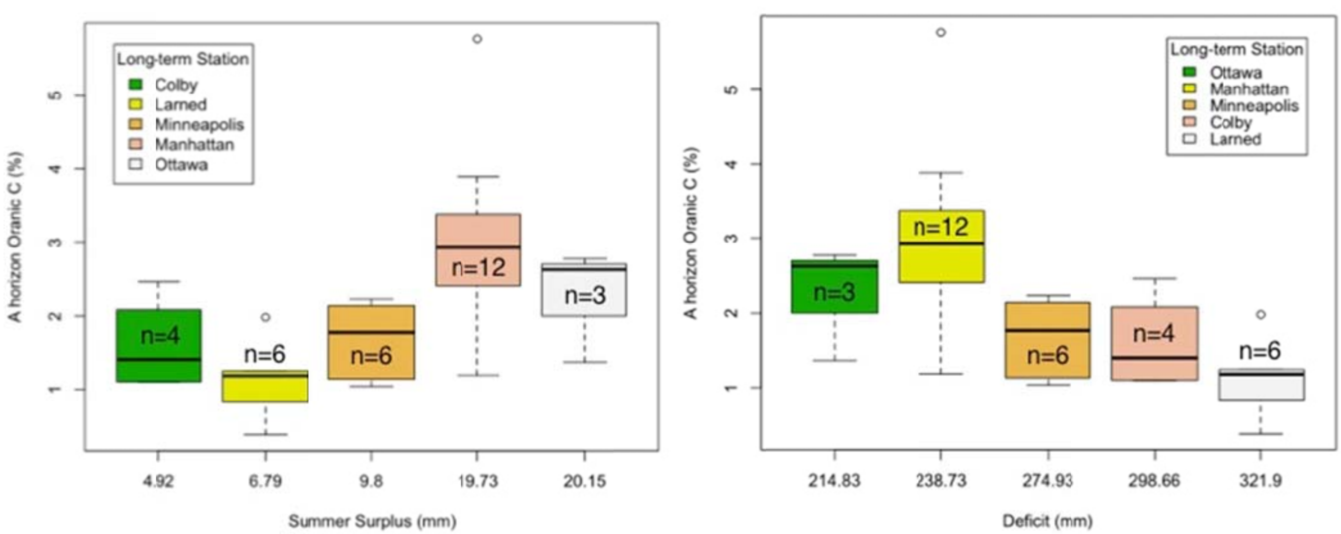

Fig. 2 - Boxplots showing A horizon organic carbon content in relation to summer surplus (left) and annual deficit (right). Notice soils surrounding the Larned weather station exhibit lower than expected organic carbon contents due to the soils' sandy parent material.

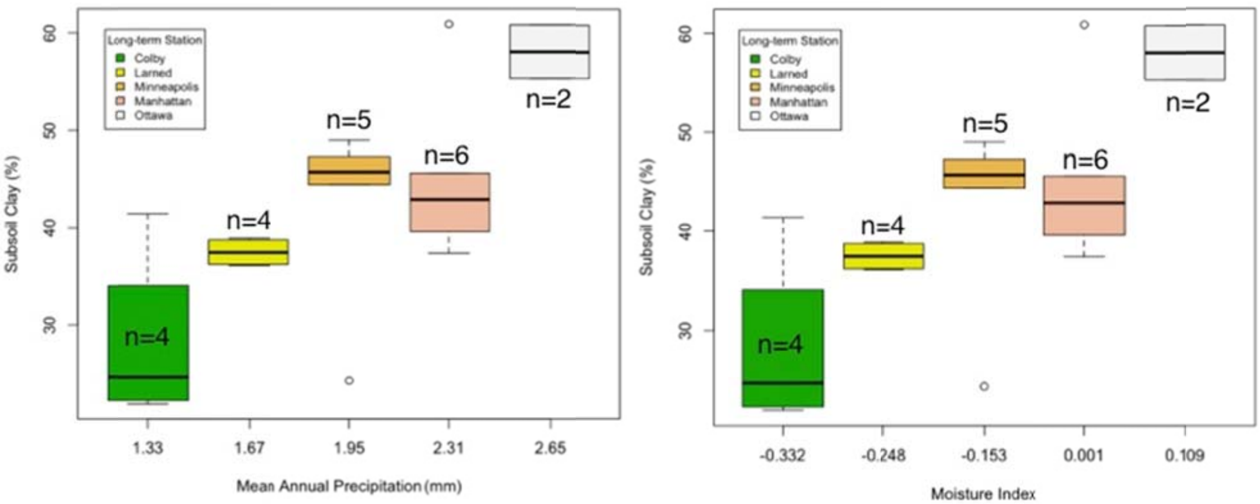

Fig. 3 - Boxplots showing subsoil clay content in relation to MAP (left) and I_m (right). Notice soils surrounding the Ottawa weather station exhibit higher clay content than expected due to clays inherited from the fine-textured parent material.

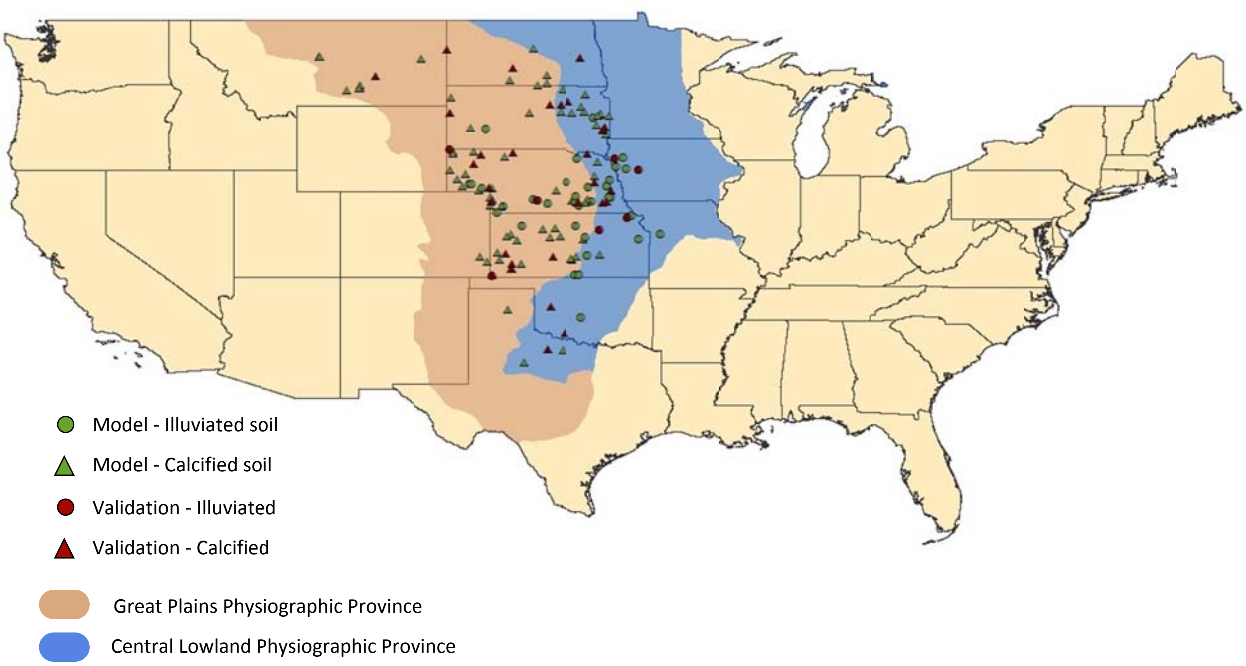

Fig. 4 - Map showing locations of modern soils used in development and statistical validation of the BuSCR model.

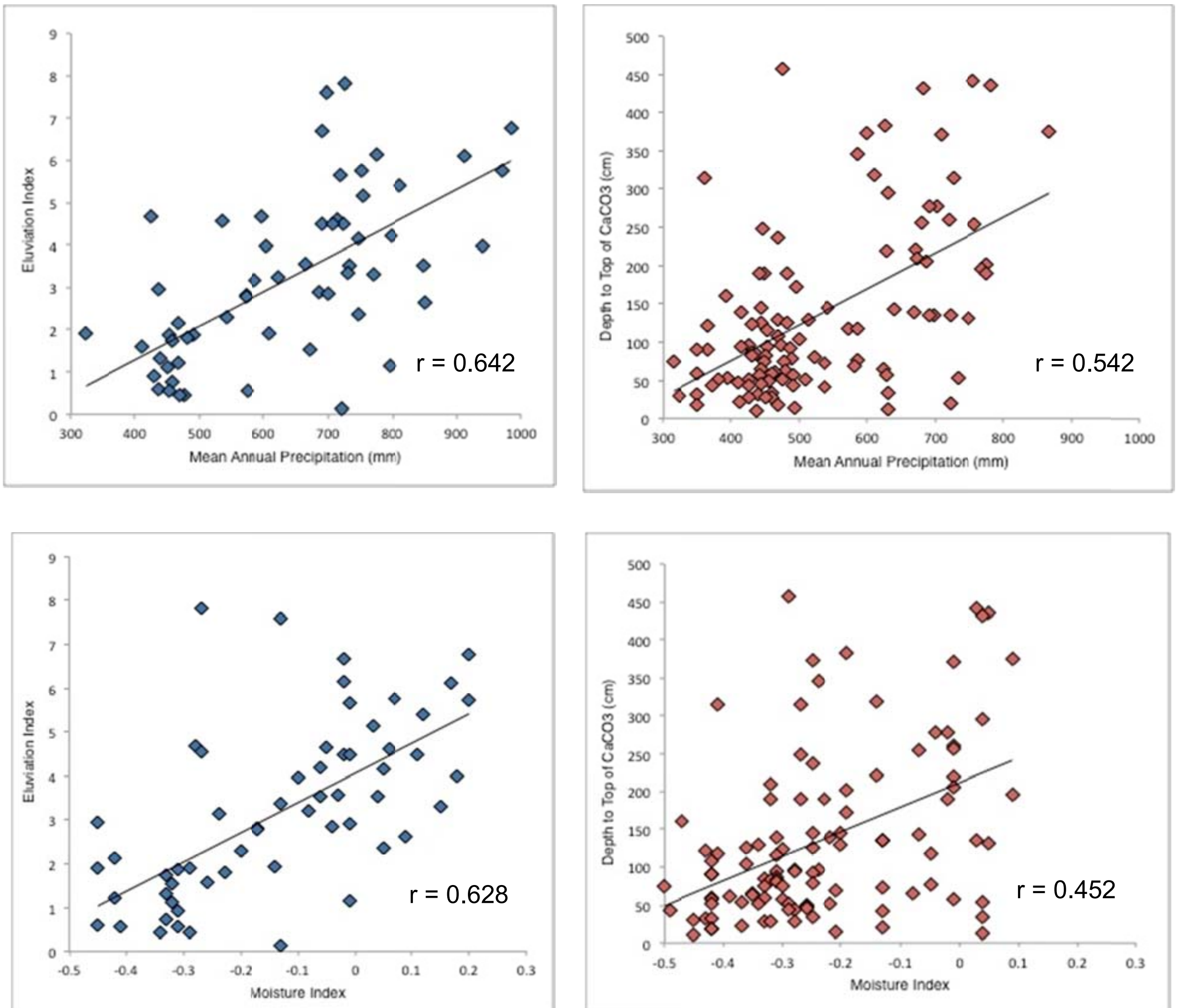

Fig. 5 - Scatterplots showing the relationship between illuviation and calcification and moisture. Plots on the left in blue show I_e and MAP (top) and I_m (bottom). Plots on right in red show depth to the top of pedogenic carbonate accumulations and MAP (top) and I_m (bottom).

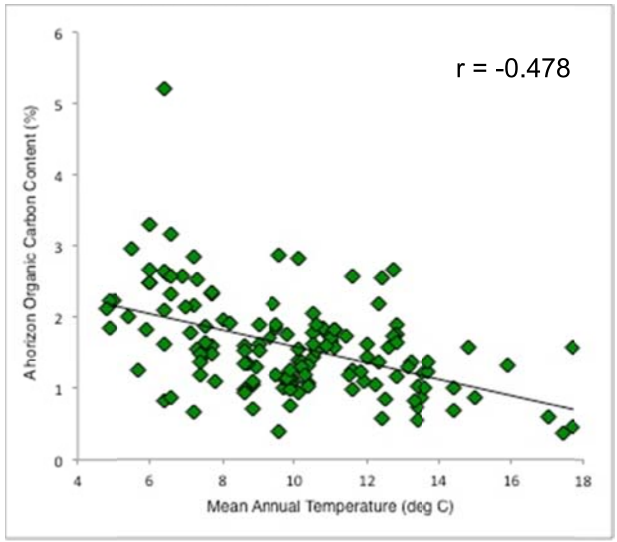

Fig. 6 - Scatterplot showing negative relationship between organic carbon content in the A horizon and MAT.

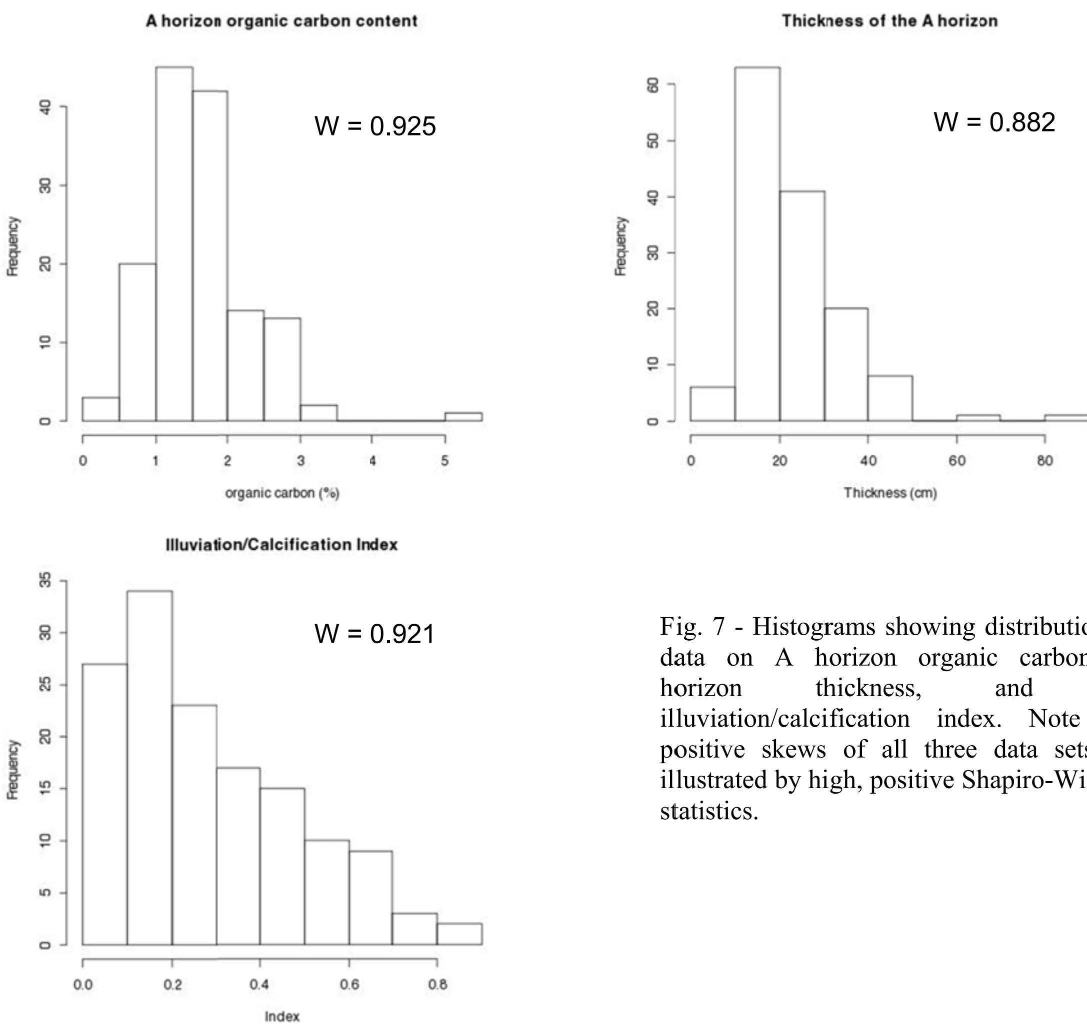

Fig. 7 - Histograms showing distribution of data on A horizon organic carbon, A horizon thickness, and the illuviation/calcification index. Note the positive skews of all three data sets, as illustrated by high, positive Shapiro-Wilk W statistics.

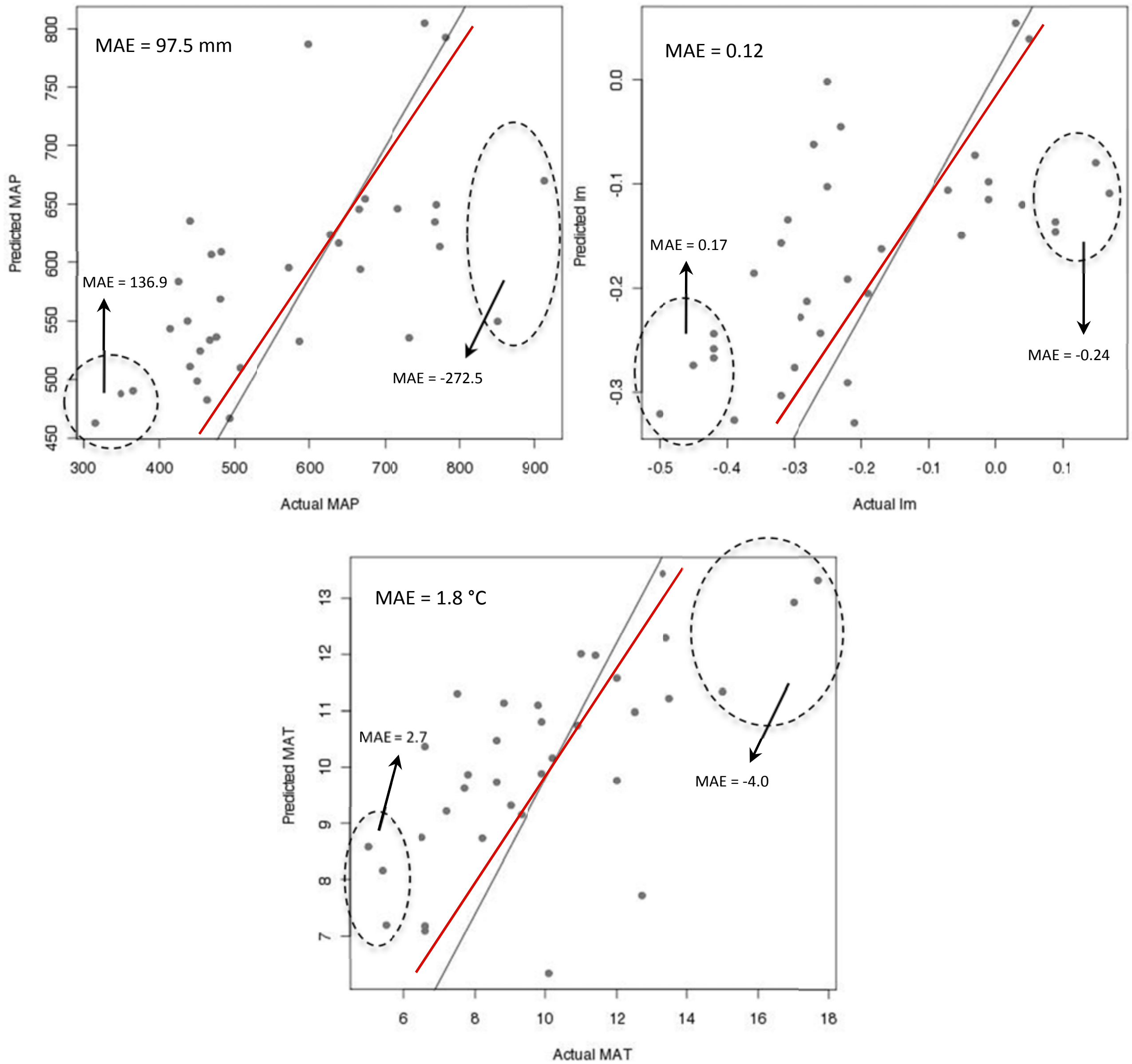

Fig. 8 - Predicted compared to observed values for the three regression models. MAE values for each model are shown. MAE calculated for cases with the highest and lowest climate values in the data set (circled) are also provided to illustrate the increase in error at the edges of the climate envelope. Red lines show perfect match of predicted and observed values and gray lines are best fit.

Table 1 – Correlation coefficients and significance levels for soil properties and and climate characteristics across Kansas.

	MAT	MAP	PE	AE	D	S	Im	Im range	AE/PE	Summer D	Summer S	Summer Im	Summer AE/PE
Thickness of Bk horizon	-0.681***	-0.595**	-0.681***	-0.640***	0.371*	-0.538*	-0.562*	0.256	-0.516*	0.159	-0.511*	-0.58**	-0.376*
Subsoil % CaCO3	-0.664***	-0.685***	-0.685***	-0.732***	0.506*	-0.630**	-0.657***	0.386*	-0.638**	0.299	-0.610**	-0.666***	-0.522*
Max % CaCO3 in subsoil	-0.690***	-0.650***	-0.701***	-0.698***	0.442*	-0.592**	-0.618**	0.321	-0.585**	0.227	-0.569**	-0.633**	-0.453*
A horizon % organic C	0.087	0.514*	0.152	0.476*	-0.589**	0.537*	0.547*	-0.577**	0.558*	-0.548*	0.62**	0.572**	0.567**
A horizon CEC	-0.034	0.407	-0.002	0.324	-0.586*	0.461*	0.454*	-0.626*	0.470*	-0.639*	0.537*	0.466*	0.552*
Subsoil CEC	0.24	0.594*	0.264	0.553*	-0.702**	0.609*	0.615*	-0.669**	0.655**	-0.663**	0.626*	0.599*	0.711**
Subsoil % clay	0.571*	0.728**	0.558*	0.709**	-0.602*	0.707**	0.709**	-0.502*	0.688**	-0.424	0.625*	0.683**	0.612*
Max % clay in subsoil	0.528*	0.78***	0.531*	0.747***	-0.695**	0.771***	0.775***	-0.606*	0.757***	-0.526*	0.728**	0.761***	0.694**

Highest r (shaded) | * - Sig < .05 | ** - Sig < .001 | *** - Sig < .0001

Table 2 - Results of the stepwise multiple regression analysis. The equation that accounts for the most variation in each climate variable (MAT, MAP, I_m) is shown.

Regression Equation	df	F-statistic	p-value	Adj. R^2
MAT = 4.945(index) - 0.369log(A horizon thickness) - 5.512√(A horizon OC) + 16.484	101	16.09	1.27×10^{-8}	0.303
MAP = 411.19(index) + 66.76log(A horizon thickness) - 11.37√(A horizon OC) + 266.96	101	22.42	3.32×10^{-11}	0.382
I_m = 0.389(index) + 0.099log(A horizon thickness) - 0.057(A horizon OC) - 0.684	101	24.14	7.38×10^{-12}	0.400

Table 3 - Comparison of predicted values for the three models and observed values of the validation data set, including various statistical measures of model performance.

Model	p		o		MAE	RMSE	$RMSE_u$	$RMSE_s$	d_r
	mean	s.d.	mean	s.d.					
MAT (°C)	10.1	1.8	9.9	3.2	1.8	2.3	1.3	1.9	0.636
MAP (mm)	587.2	87.4	572.2	155.9	97.5	121.0	67.1	100.8	0.639
I_m	-0.17	0.10	-0.19	0.19	0.12	0.14	0.08	0.12	0.618

TESTING THE BuSCR MODEL ON HOLOCENE-AGE BURIED SOILS ACROSS THE GREAT PLAINS, U.S.A.

(in preparation as Ashley B. Zung, Rolfe D. Mandel, and Johannes J. Feddema. 2013 The BuSCR Model, Part 2 – Testing the model on Holocene-age buried soils across the Great Plains, *Quaternary Research*)

1. Introduction

Buried soils, which are commonly preserved in the stratigraphic record of the Great Plains, have long been used to reconstruct past environmental conditions in the region (Bryan and Albritton 1943; Judson 1953; Wendorf et al. 1955; Holliday 1995; Mandel 2008). Such reconstructions are based on the premise that bioclimatic factors exert great influence on soil genesis, and soil properties reflect this influence. Because two of the most commonly utilized terrestrial paleoclimate proxies, pollen and tree rings, are rare on the Great Plains, alluvial and eolian soil stratigraphy have emerged as effective techniques for reconstructing environmental change in the region.

Soil stratigraphic investigations in combination with stable carbon isotope ($\delta^{13}C$) analysis of soil organic matter (SOM) have been widely utilized on the Great Plains to discern shifts in moisture and temperature regimes through the late Quaternary (Nordt 1994; Holliday 1995; Baker et al. 2000; Forman et al. 2001; Bement et al. 2007; Mandel 2008; Nordt et al. 2008; Cordova et al. 2011). Studies of valley fills exposed at numerous localities across the region have demonstrated that sediment accumulation over the last 14,000 years has been punctuated by regionally synchronous soil-forming events (Hall 1990; Holliday 1995; Mandel 2008).

Specifically, alluvial soil stratigraphy provides evidence for a regional drying trend from the terminal Pleistocene into the early Holocene in the Central and Southern Plains (Holliday 1995; Mandel 2008), and an episode of aridity around 1000 B.P. in the Southern Rolling Plains (Hall 1990). The geomorphic and soil stratigraphic record of Holocene eolian activity on the Great Plains indicates widespread, sustained aridity between 10,000 and 5000 cal. yr. B.P. and numerous discrete drought events over the last 2000 years (Forman et al. 2001; Hanson et al. 2010; Halfen et al. 2012). The record of dune activity and loess deposition indicates that these arid periods probably had a greater than 25% growing season precipitation deficit compared to today (Forman et al. 2001).

Although reconstructions based upon soil stratigraphy and $\delta^{13}C$ values of SOM have provided important insight into shifts in late-Quaternary moisture regimes on the Great Plains, these methods alone do not provide quantitative estimates of temperature and precipitation. Nor do they differentiate the impact of temperature from shifts in precipitation or elucidate the mechanisms of climate change. Nordt et al. (2007) attempted to resolve this by developing a transfer function for mean July temperatures (MJT) based on the $\delta^{13}C$ composition of SOM from buried soil A horizons. Other than this model, however, no quantitative method for reconstructing past climate, such as those for pollen assemblages and tree-ring growth patterns (e.g. Briffa et al. 2001; Jackson and Williams 2004; Minckley et al. 2008) exists for buried soils.

In order to more fully utilize the buried soils record of the Great Plains to reconstruct past climate, we developed the BuSCR (**Bu**ried **S**oil **C**limate **R**econstruction) model (Zung and Feddema 2013, in preparation). The BuSCR model reconstructs mean annual precipitation (MAP), mean annual temperature (MAT), and annual moisture index (I_m) based upon morphological, physical, and chemical characteristics of the buried soil. We calibrated the model based on modern soil-climate relationships at 105 sites across the Great Plains and statistically validated the model by applying it to 35 modern surface soils and comparing the results to modern climate records. Results of the statistical validation were quite promising, as predicted values exhibited highly statistically significant p-values when correlated with observed values and very low mean average errors from observed values (Zung and Feddema 2013). While the model exhibited high precision when tested on modern surface soils, we are uncertain of its success in reconstructing past climates based on buried soil properties.

The purpose of this study is to test the BuSCR model by applying it to 17 buried soils preserved in a variety of depositional settings across the Great Plains. The results of BuSCR are reviewed with respect to our current understanding of climate conditions during four late-Quaternary climate periods in the region (Table 1). Also, we compared the results of the BuSCR modeled MAT and I_m to MJT reconstructed using the Nordt et al. (2007) model at 6 of the 17 sites where data on $\delta^{13}C$ of SOM was available. These comparisons allowed us to determine the accuracy of the BuSCR model with regard to the currently accepted understanding of late-Quaternary climate change on the Great Plains and reconstructions based on a widely accepted paleoenvironmental proxies.

2. Methodology

2.1 Data on buried soils across the Great Plains

To find soils for our test, we reviewed scientific publications that contained information about buried soils that seemed appropriate for the BuSCR model. Buried soils must meet a series of criteria to apply the BuSCR model. Specifically, the soil must exhibit evidence of pedogenic clay or carbonate accumulation and have an A horizon. The presence of a Bt or Bw horizon were indicators of the presence of pedogenic clays, and pedogenic carbonates were identified when the ratio of carbonates in the C horizon compared to the B horizon was < 1. If Bg, Bss or other horizons were observed below Bk or Bt horizons, the model was applied but only considered the upper Bt or Bk horizon. Also, soils that showed evidence of polygenesis, such as the presence of a Btk horizon or a Bk over a Bt or Btk horizon were not used in our test.

In addition to meeting specific morphological criteria, data availability was crucial for application of the model. In particular, implementation of BuSCR requires estimating the buried soil's saturated hydraulic conductivity (K_{sat}), calculating the eluviation index (I_e), and approximating soil organic carbon (SOC) content by weight. We estimated K_{sat} using a pedotransfer function based on the total sand and clay content of the subsoil (Cosby et al. 1984; Tietje and Hennings 1996). K_{sat} estimations were used to normalize depth to carbonates and I_e, as needed to calculate the illuviation/calcification index (Zung and Feddema 2013). Determining the I_e required estimates of fine clay and total clay content for all horizons. Application of a conversion factor of 1.724 (NRCS 2009) to SOM allowed estimation of SOC content when a laboratory measurement of SOC was unavailable.

Published detailed soil descriptions were typically necessary to determine soil horizonation and other soil profile characteristics as input to the model (e.g. depth to carbonates and thickness of the A horizon). These extensive data requirements for applying the model limited the sample size for the test. However, while some of data were unavailable from published sources, metrics used in the model are commonly measured in standard laboratory analyses for soil characterization. $\delta^{13}C$ values of SOM in the A horizon of buried soils also was collected, when available. These data were collected in order to reconstruct MJT using the Nordt et al. (2007) forest-grassland ecotone and grassland pedotransfer functions.

A final condition for soil selection was reliable chronologic control. This was done in order to assign buried soils to climate periods with relative confidence. Radiocarbon ages determined on materials such as SOM, charcoal, or wood were used to provide chronologies for buried soils. The properties of buried soils used in the BuSCR model reflect the dominant climate during the period of soil formation, as determined by the radiocarbon age of the soil. Reliable estimated soils ages ensured that the period of soil formation was known and that buried soils in the study were assigned to appropriate complementary climate periods. Where uncertainty existed with regards to soil age and the estimated age crossed two climate periods, we assigned the soil to both climate periods and analyzed the results accordingly. Climate period delineations used in this study are based on a literature review of widely accepted climate periods, with specific attention to the Great Plains (Table 1).

2.2 Modern Climate Data

To relate past estimates of climate to present day climatology, we downloaded long-term annual and monthly climatologies (1950-1999 means) from the University of Delaware Center for Climatic Research's Climate Data Archive and used these to characterize the Great Plains modern climate. The Willmott & Matsuura monthly and annual climatologies (Willmott and Matsuura 2001a, 2001b) provide precipitation, temperature, and moisture index estimates for 0.5 degree grid cells globally. Modern climate data were used to calculate anomalies of the BuSCR estimates from present conditions. We also collected long-term climatologies for MJT in order to calculate anomalies from present-day July temperature for the reconstructions based on $\delta^{13}C$ values. We spatially joined the climate data to the buried soil data using ArcGIS in order to calculate the anomalies.

2.3 Climate reconstructions and model comparison

We reconstructed MAP, MAT, and I_m for 17 buried soils using the BuSCR model (Fig. 1; Zung and Feddema 2013). Where $\delta^{13}C$ data were available, MJT was modeled for the period of soil formation using the forest-grassland ecotone and grassland model (Nordt et al. 2007). Climate anomalies from present conditions were calculated for each site by comparing the reconstructed MAP, MAT, I_m and MJT to the Willmott and Matsuura (2001) climatologies at each location. We compared MAP, MAT, and I_m reconstructed values and anomalies to multi-proxy qualitative reconstructions of climate from the literature and also compared MAT and I_m results to the Nordt et al. model reconstructions.

3. Results

Soils used to test the BuSCR model formed during the Holocene at locations across the Great Plains in a variety of depositional settings (Fig. 1; Table 2). Soil 1 in southwest South Dakota formed in the most northern location, and soil 17 in central Oklahoma formed in the southernmost location. Soil 1 also formed in the westernmost location, while soils 8 and 9 at the Coffey archaeological site in eastern Kansas formed in the easternmost location. Three soils formed in loess (soils 2, 5, and 6), and the remaining 14 soils formed in alluvium. Ten soils exhibited evidence of pedogenic carbonate accumulations, and seven soils had properties that indicated clay illuviation. Soil radiocarbon ages ranged from ca. 7200 to 950 yr. B.P.

3.1 BuSCR reconstructions of MAP

The BuSCR model reconstructed MAP as drier at all locations for all time periods compared to the modern climate (Table 3; Fig. 3). The magnitude of the anomalies was greatest for locations in the eastern Great Plains and least for locations in the western Great Plains, regardless of climate period. Soil 10 exhibited the highest anomaly, with MAP hindcasted as 454 mm below present day MAP values during the Medieval Warm Period. Soil 1, which formed during the Neoglacial in southwestern South Dakota, was reconstructed as having the least anomalous MAP versus today at 99 mm less. BuSCR reconstructed MAP and MAP anomalies were highly correlated with longitude ($r = 0.899$), and not correlated with estimated soil age ($r = -0.161$). This indicates that semi-arid conditions, similar to the western Great Plains today, were widespread during the soil forming periods represented by soils in the data set. These results are also an artifact of smaller modern MAP in western compared to eastern Kansas, and hence smaller anomalies in the west. Patterns of MAP percent change, however, were consistent with the magnitude of total anomalies.

Soils at the same site that formed during the same climate period showed notable consistencies, indicating internal consistency in the model. For example, soils 5 and 6 are a pair of stacked buried soils described at Old Waunetta Roadcut that both formed during the Neoglacial or Altithermal Period (Table 2). MAP reconstructed for these soils were 381 and 378 mm, respectively. Soils 13 and 14, another pair of stacked buried soils from a site in the panhandle of Texas (Table 2), also showed remarkable consistency, with MAP modeled as 399 and 406 mm. Soils 8 and 9, however, showed less consistency. The BuSCR model estimated MAP during the Altithermal as 539 mm based on the properties of soil 8 and 453 based on the properties of soil 9. It is important to note, however, uncertainty in the age of soil 8 may mean that these soils are not coeval.

3.2 BuSCR reconstructions of Im

Like MAP, reconstructed I_m was consistent, showing values lower than modern I_m for all locations during all time periods (Table 3; Fig. 2). I_m anomalies were also greater to the east compared to the west, similar to MAP. The greatest I_m anomaly (-0.48) was modeled for soil 9 that formed during the Altithermal in eastern Kansas, and the lowest I_m anomaly (-0.10) was determined for soil 15, which formed during the Neoglacial or Medieval Warm Period in the Texas Panhandle. These results indicate that the moisture balance on the Great Plains during periods of the Holocene represented by the 17 soils was typically drier than today.

I_m anomalies displayed a consistent spatial pattern from east to west, as demonstrated by a high correlation with longitude (r = 0.829), but exhibited weaker correlations with age (r = -0.397) than the reconstructed MAP values. Soils that formed during the Altithermal and Early Holocene exhibited the greatest mean I_m anomalies (-0.315 and -0.293, respectively), and soils representing the more recent Neoglacial and Medieval Warm Periods exhibited much lower anomalies, averaging -0.215 and -0.217, respectively. Within site consistencies were also observed for reconstructed I_m, and, like MAP, I_m modeled for both soils 5 and 6 and soils 13 and 14 were incredibly close, while soils 8 and 9 exhibited the greatest within site variation (Table 3).

It should be noted that while I_m and MAP anomalies are strongly correlated (r = 0.836), they are not directly related, as evident by the soils with highest and lowest anomalies for the two climate parameters. This is due to the fact that I_m anomalies depend not only on the magnitude of change in precipitation, but also on change in potential evapotranspiration. Hence, changes in I_m are best understood in consideration of changes in both temperature and precipitation at the respective location.

3.3 BuSCR reconstructions of MAT

Reconstructions of MAT showed significantly greater variation across locations compared to MAP and I_m, although a strong spatial correlation was still observed (Fig. 4). MAT reconstructed by the BuSCR model was strongly correlated with latitude (r = 0.719); the model hindcasted positive temperature anomalies for locations in the Southern Plains and negative temperature anomalies in the northern portion of the Great Plains. Soil 17, which formed during the Medieval Warm Period on the Southern Plains, displayed the most negative MAT anomaly at -5 °C. The BuSCR model reconstructed the highest positive temperature anomaly (3.8 °C) for soil 4 that formed during the Neoglacial in central Nebraska. On average, the climate of locations north of 40 degrees were 2.3 °C warmer than present, regardless of climate period, while soils south of 40 degrees were 1.2 °C cooler, on average.

MAT was more strongly correlated to age than the other two climate parameters (r = 0.406), indicating that temperatures were generally warmer during earlier periods of the Holocene than later periods. Specifically, early Holocene temperature anomalies averaged 1.7 °C and Altithermal MAT was 1.1°C. Neoglacial MAT anomalies averaged 0.9 °C, and the BuSCR model reconstructed the most negative temperature anomalies for the Medieval Warm Period, -1.4 °C on average. MAT reconstructions exhibited within site consistency similar to MAP and I_m reconstructions.

3.4 July mean temperature reconstructions from δ13C values

Table 4 provides MJT reconstructions based on $\delta^{13}C$ values of SOM from six buried soils in the data set. We applied both the forest-grassland ecotone model and the grassland model from Nordt et al. (2007) to reconstruct MJT. MAT reconstructed by BuSCR exhibited a weak correlation with MJT reconstructed from $\delta^{13}C$ values, (Fig. 5; r = -0.057, p value = 0.916). In contrast, I_m reconstructed by BuSCR was strongly positively correlated with MJT (Fig. 5; r = 0.603, p value = 0.210). These results indicate that the direct pedotransfer function the Nordt model uses to calculate MJT from $\delta^{13}C$ values, and therefore the $\delta^{13}C$ values themselves, are most highly correlated with I_m for the six soils considered here.

Interestingly, the MAT anomaly reconstructed with the BuSCR model exhibited a statistically significant correlation with the MJT anomaly ($r = 0.799$, p value = 0.059), while the I_m anomaly was not significantly correlated with the MJT anomaly (Fig. 6; $r = 0.169$, p value = 0.745). MJT was strongly correlated with latitude, with the exception of the buried soil at Armstrong Creek (soil 16; Table 4). The BuSCR modeled MAT did not exhibit a strong correlation with latitude.

4. Discussion

4.1 Late-Quaternary Paleoclimate on the Great Plains, U.S.A.

The results of the BuSCR climate reconstructions generally agree with findings from other paleoenvironmental studies of the Great Plains. Specifically, I_m anomalies indicate that the early Holocene I_m on the Great Plains was on average -0.293 lower than today and Altithermal I_m was -0.315 lower. These results indicate that persistent aridity dominated the Great Plains during the first half of the Holocene, supporting conclusions from other late-Quaternary paleoenvirnomental reconstructions on the Great Plains (e.g. Nordt et al. 2008; Mandel 2008; Forman et al. 2001; Bement et al. 2007; Cordova et al. 2011).

Multiple lines of evidence also suggest the occurrence of discrete mega-drought events during the last 2000 years, and specifically during the Medieval Warm Period (Hall 1990; Sridhar et al. 2006; Halfen et al. 2012). Both the BuSCR I_m and MAP reconstructions support this assertion, with negative I_m and MAP anomalies reconstructed for all soils that formed during the Medieval Warm Period. Further, negative MAP anomalies during the Medieval Warm Period far exceed those of the Neoglacial, -206.4 mm on average compared to -151.7 mm, respectively, indicating a greater than 25% decrease in MAP during the Medieval Warm Period compared to earlier and later periods of the Neoglacial. A shift in Great Plains spring-summer circulation from moist southerly flow to dry southwesterly flow (Sridhar et al. 2006; Feng et al. 2008) would explain the significant decrease in MAP reconstructed by the BuSCR model during the Medieval Warm Period.

The BuSCR model greatly improves the quality of climate reconstructions based on buried soils by quantifying I_m and also separating MAP from MAT. Frequently, paleoenvironmental reconstructions based on soil stratigraphy identify shifts in climate as 'more arid' or 'moister.' Such conclusions typically assume that more arid conditions, for example, require both a shift to lower MAP as well as higher MAT. BuSCR reconstructions for the Altithermal show that this is not the case. Typically, the Altithermal is interpreted as being both warmer and drier than today. Eolian soil stratigraphy and dune activity has frequently been used to deduce this (see Forman 2001). However, our data show that the Altithermal was dramatically drier but not significantly hotter. Discriminating between shifts in MAP and MAT provide direction for hypothesizing as to the mechanisms of past climate change, and as a result, better prepare us to project the results of changing mechanisms for future climate.

BuSCR model results for MAT and I_m matched MJT reconstructions rather well. Correlations improve when soil 16, a potential outlier is removed from the data set. In fact, the correlation between BuSCR MAT and MJT jumps to 0.692 when soil 16 is removed from the test. The $\delta^{13}C$ value for soil 16 at Armstrong Creek may be problematic due to microclimate and geomorphic affects on vegetation at that location (see Zung and Mandel 2013, in preparation). This potential problem highlights the importance of understanding site-specific impacts on vegetation, and consequently $\delta^{13}C$ values of SOM in buried soils, when applying the Nordt et al. (2007) model.

The strong correlation of MJT and I_m indicates a potential problem with attributing $\delta^{13}C$ values to temperature alone. Nordt et al. (2008) acknowledge this potential problem, noting that the curvilinear relationship of $\delta^{13}C$ values and latitude in the Central and Southern Plains is probably due to the increased importance of moisture availability to determining vegetation in lower latitudes of the mid-continent. Clearly, interpretation of $\delta^{13}C$ values of SOM and MJT temperature reconstructions that result from application of the Nordt et al. (2007) model requires an understanding of the geomorphic and biotic context of each site, as well as the potential impact of moisture balance on vegetation.

5. Conclusion

We applied the BuSCR model to 17 buried soils located across the Great Plains, from South Dakota to the Texas Panhandle. The BuSCR reconstructions, especially for I_m and MAP, are in good agreement with previous late-Quaternary paleoclimatic reconstructions on the Great Plains. This indicates that we can apply the BuSCR model confidently to Great Plains buried soils and reconstruct past climates. We also compared the results of the BuSCR model to reconstructed MJT based on $\delta^{13}C$ values determined on SOM, a paleoenvironmental technique widely utilized for reconstructing past climate on the Great Plains. Strong correlations of the MAT anomalies and I_m modeled by BuSCR with MJT reconstructions and anomalies further indicate the reliability of the BuSCR model. While these findings indicate the applicability of the BuSCR model, site and soil specific issues must be considered when selecting soils to use for climate reconstructions.

The findings of this study also highlight the importance of differentiating the effects of temperature on moisture balance from those of precipitation. Differentiating the drivers that led to shifts in the moisture balance in the past and estimating quantitatively the amount of change that occurred provides valuable insight to scientists regarding drivers of mid-continental climate in North America over long time scales. This knowledge equips climate modelers to better project the impact of future climate change on semi-arid regions like the Great Plains, U.S.

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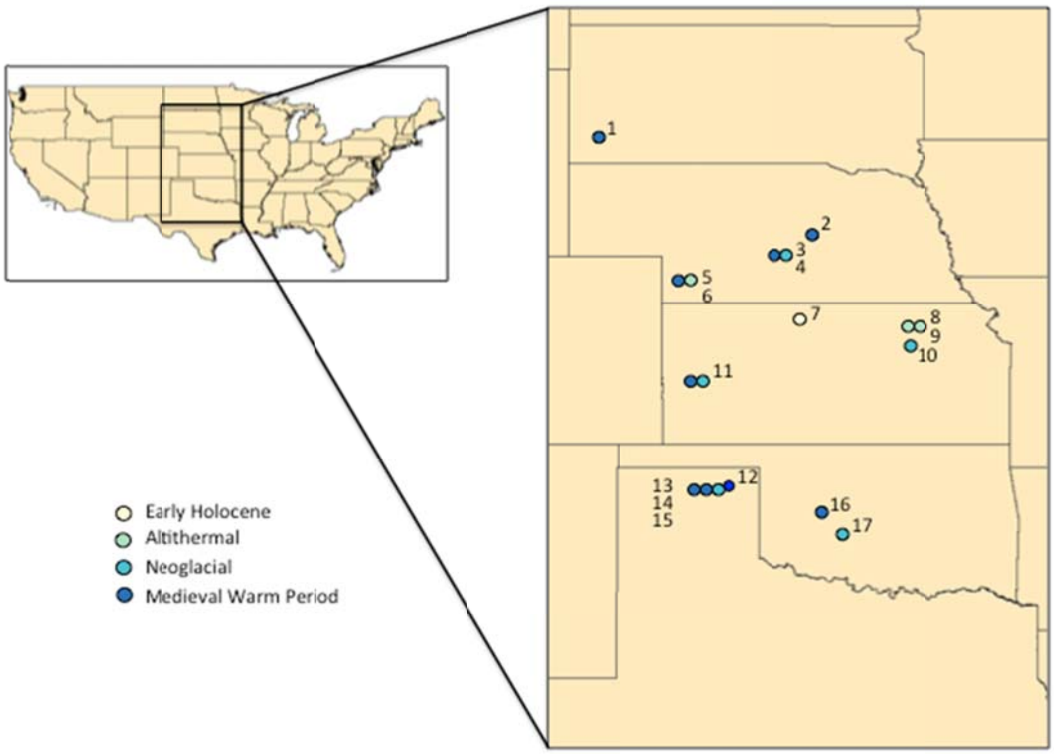


Fig. 1 – Locations and climate periods assigned to the 17 buried soils used to test the BuSCR model. Multiple soils at the same site are represented by individual circles placed in a horizontal line.

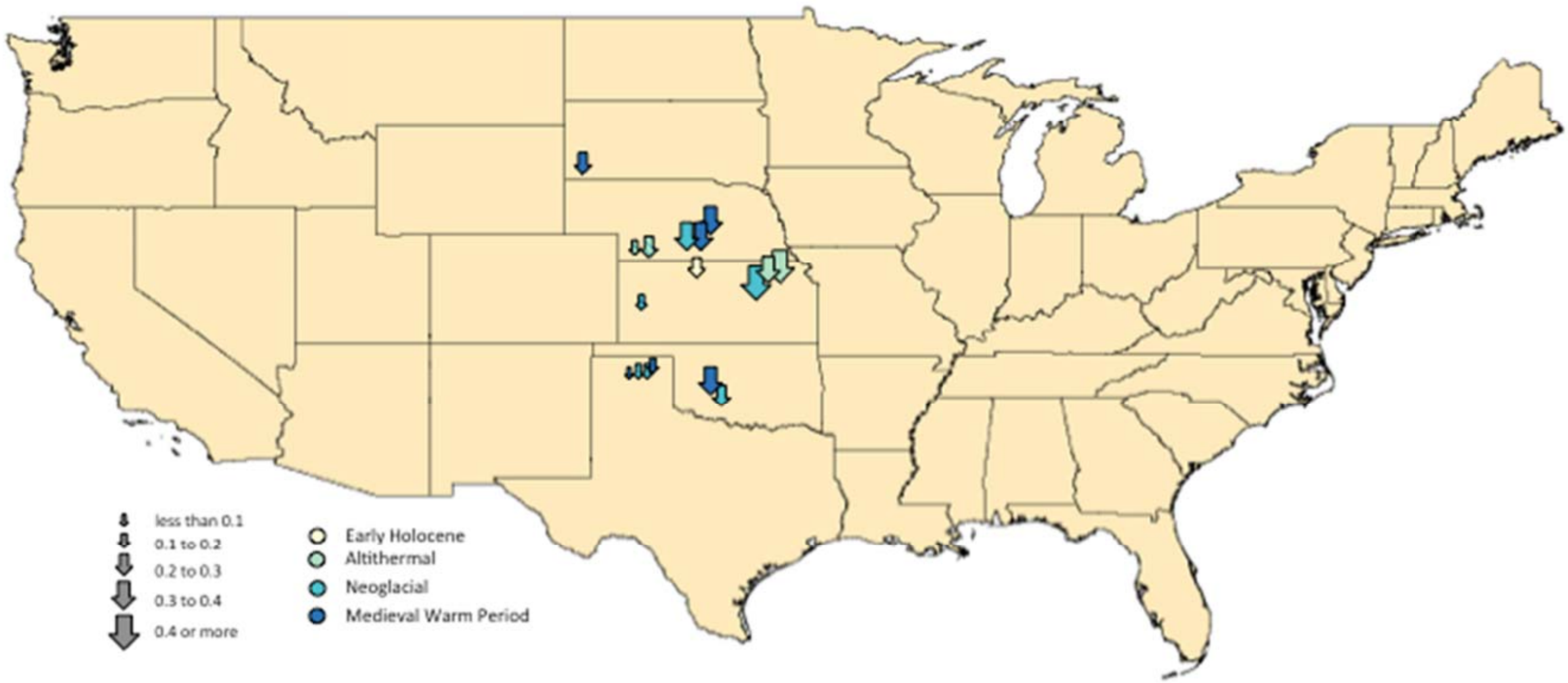


Fig. 2 – Results of the BuSCR model reconstructions for I_m compared to modern climate records. Size of arrow represents magnitude of the anomaly and color identifies the climate period. Striped markers are soils assigned to two climate periods, and stripe colors represent the respective climate periods.

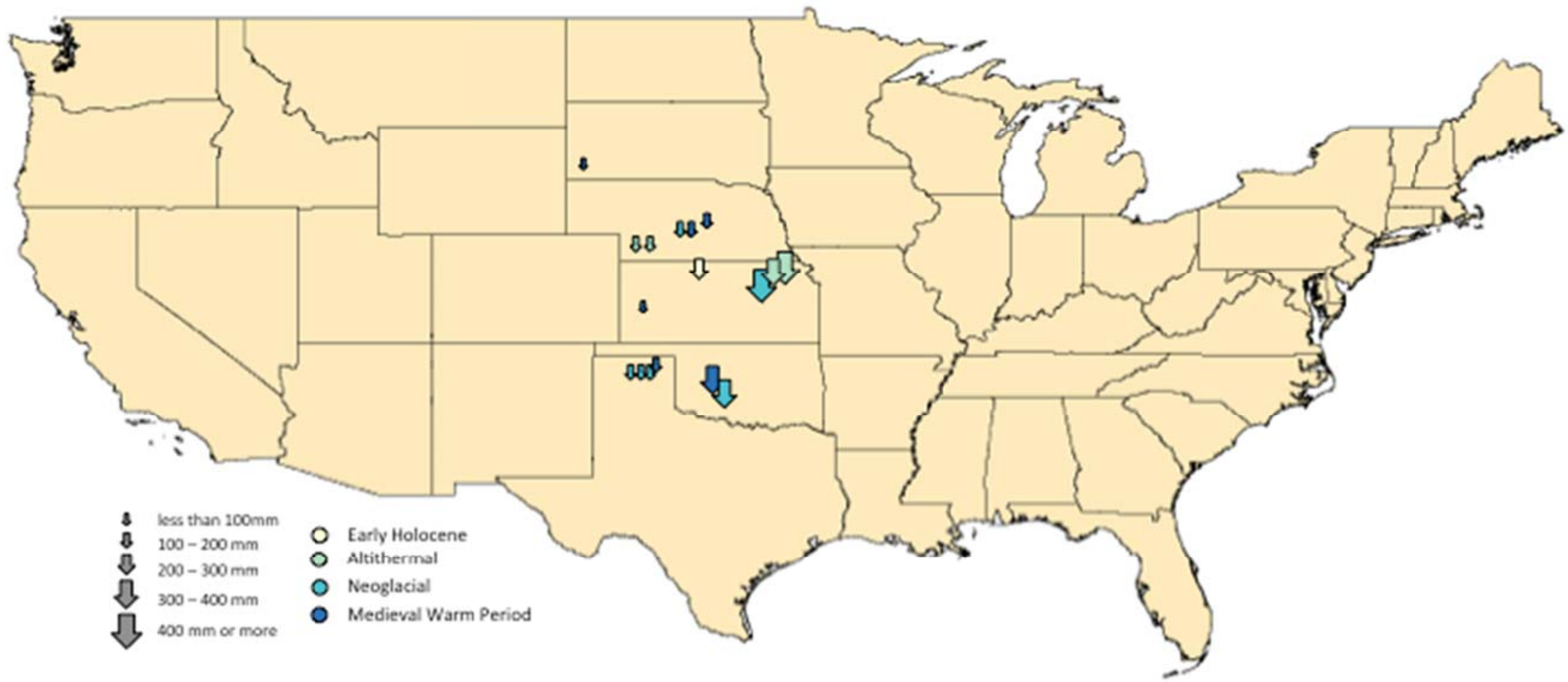


Fig. 3 – Results of the BuSCR model reconstructions for MAP compared to modern climate records. Size of arrow represents magnitude of the anomaly and color identifies the climate period. Striped markers are soils assigned to two climate periods, and stripe colors represent the respective climate periods.

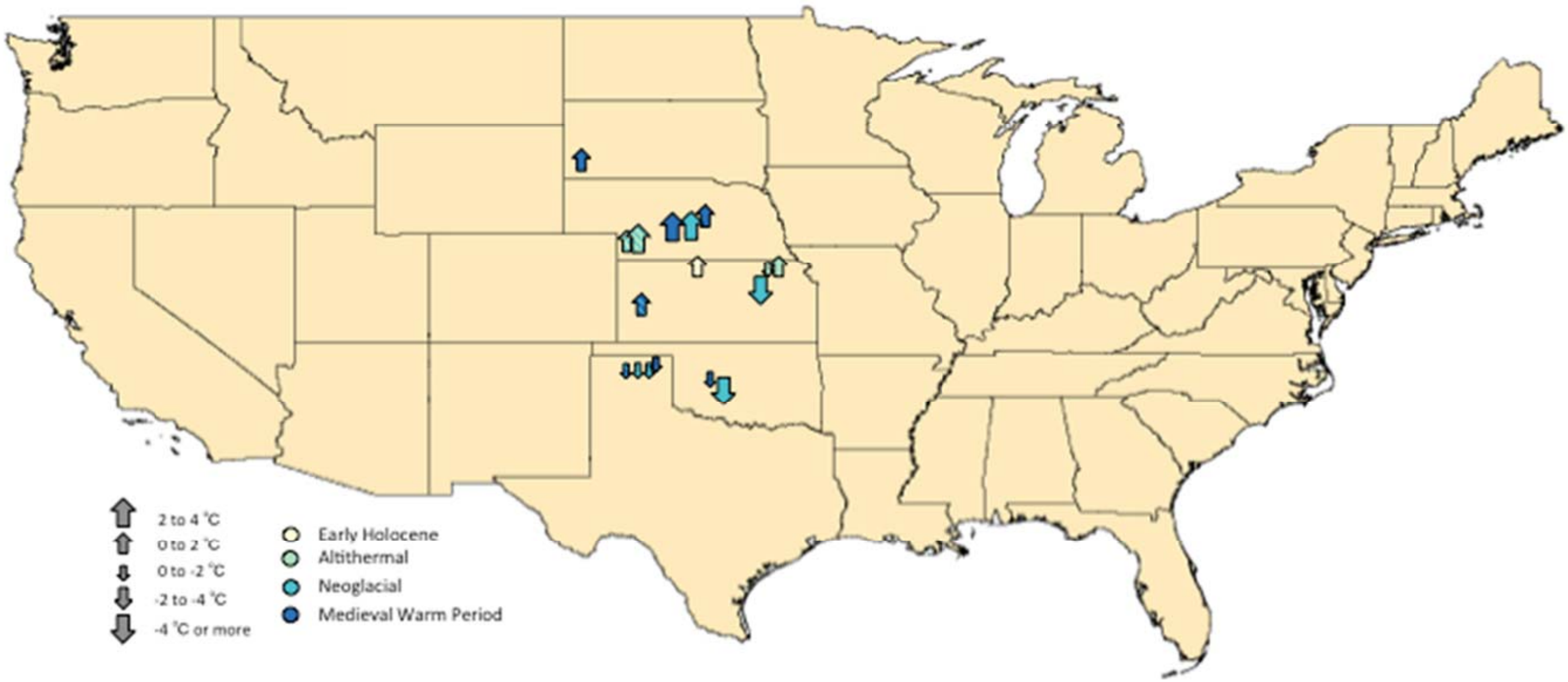
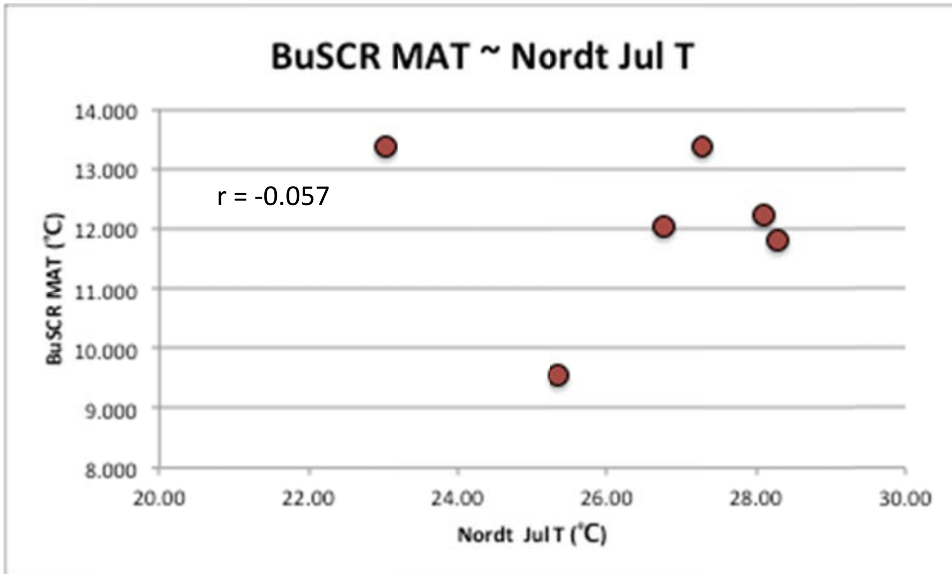


Fig. 4 – Results of the BuSCR model reconstructions for MAT compared to modern climate records. Size of arrow represents magnitude of the anomaly and color identifies the climate period. Striped markers are soils assigned to two climate periods, and stripe colors represent the respective climate periods.



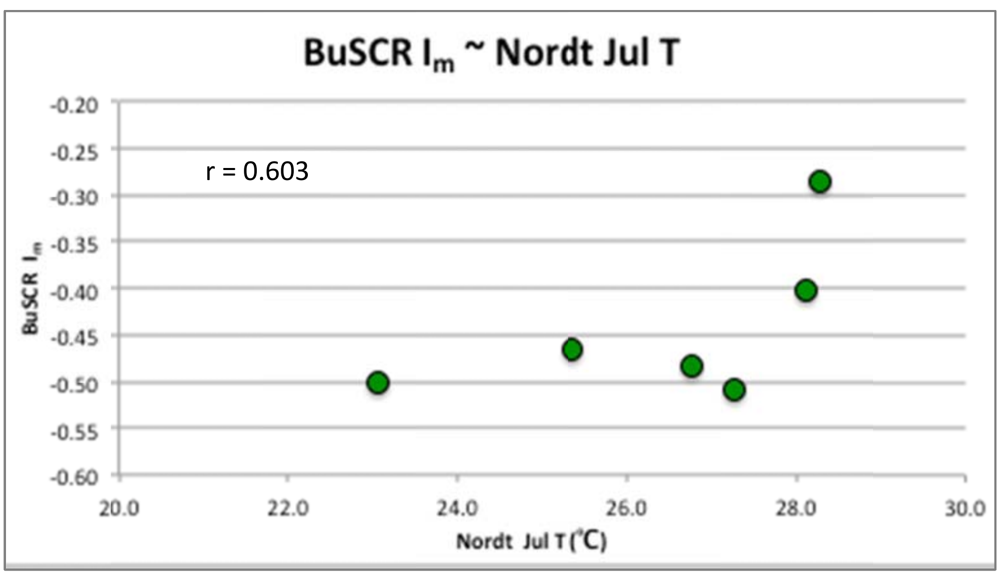
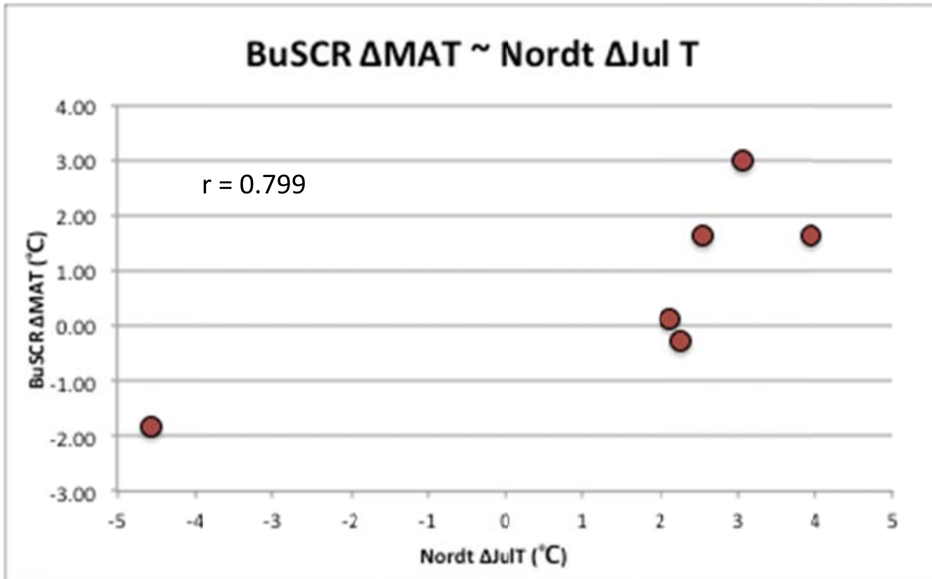


Fig. 5 – Scatterplots showing reconstructed MAT and I_m from the BuSCR model in comparison to estimates of MJT from the Nordt et al. (2007) forest-grassland ecotone model.



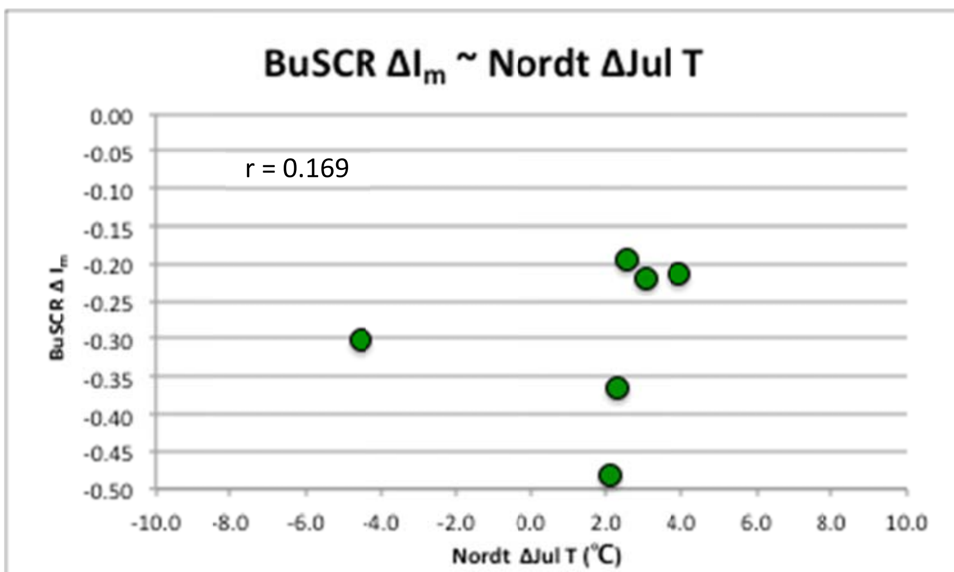


Fig. 6 – Scatterplots showing reconstructed MAT and I_m anomalies from the BuSCR model in comparison to MJT anomalies from the Nordt et al. (2007) forest-grassland ecotone model.

Table 1 – Major climate periods of the Great Plains, U.S.A. Soils used in this analysis were assigned to a climate period based on estimated age of the soil.

	TIME (cal. yr. B.P.)	TIME (^{14}C yr. B.P.)	CLIMATE CONDITIONS	SOURCE
Early Holocene	11,500 - 8000	10,000 - 7000	Post-glacial warming and drying	Bond et al. 2001; Forman et al. 2001; Williams et al. 2010
Altithermal	8000 - 4500	7000 - 4000	Extremely dry and warm; Holocene thermal maximum	Valero-Garas et al. 1997; Bond et al. 2001; Forman et al. 2001; Nordt et al. 2008
Neoglacial (Late Holocene)	4500 - present	4000 - present	Increasing moisture and decreased temperatures	Holliday 1995; Bond et al. 2001; Forman et al. 2001
Meidevel Warm Period	1200 - 800	1200 - 800	Sharp return to more arid and warm conditions	Sridhor et al. 2006; Bonan 2008; McGann 2008; Nordt et al. 2008

Table 2 – Soils used to test the BuSCR model.

Soil Number	Location Description	State, U.S.A.	Latitude (°N)	Longitude (°W)	Soil Horizonation	Soil Age (^{14}C yr. B.P.)	Climate Period	Buried SOC $\delta^{13}C$	Source
1	Soil 4, Highland Creek Terrace	SD	43.55	103.39	A-Bk-C[†]	3870 + 50	Neoglacial	-18	Fredlund 1996, Fredlund and Tieszen 1997
2	Soil 1, Mirdan Canal Section	NE	41.45	98.83	Ab-Bwb-BCb-Cb	3010 + 90	Neoglacial	na	Mason and Kuzila 2000
3	Soil 2, Pressy Park	NE	41.03	99.39	Ab-ACb-Cgb	1190 + 60	Medieval Warm Period	na	D. May 1992
4	Soil 3, Pressy Park	NE	41.03	99.39	Ab-ACb-Cb	3000 to 1200	Neoglacial	na	D. May 1992
5	Soil 2, Old Wauneta Roadcut[§]	NE	40.5	101.42	A-Bk	6200 to 2430	Neoglacial/Altithermal	-15.8	Feggestad et al. 2004, Jacobs and Mason 2004, Miao et al. 2006
6	Soil 3, Old Wauneta Roadcut[§]	NE	40.5	101.42	A-AB-Bk-C	6200 to 2430	Neoglacial/Altithermal	-15	Feggestad et al. 2004, Jacobs and Mason 2004, Miao et al. 2006
7	Soil 3, Wyrill Section 2, Deer Creek	KS	39.67	99.12	Ab-Bk-BCk	7160 + 70	Early Holocene	na	Mandel (in press)
8	Soil 2, Profile 1, Coffey Site	KS	39.5	96.5	Ab-Abb-Btb-Btgssb	est. 7000 to 5000[a]	Altithermal	-13.4	Mandel et al. 2010
9	Soil 2, Profile 2, Coffey Site	KS	39.5	96.5	Ab-ABb-Btb	7210 to 5620	Altithermal	-13.7	Mandel et al. 2010
10	Soil 2, Core KR-4	KS	39.08	96.75	Ab-ABkb-Bkb	1200 + 70	Medieval Warm Period	na	Mandel (in press)
11	Soil 2. Sand Creek. Wichita Co.^	KS	38.31	101.16	Ab-Bkb-BCkb	1360 + 70	Neoglacial/Medieval Warm	na	Mandel (in press)
12	Soil 3, BeeJay Site	TX	36.03	100.82	Ab-Bk	2710 + 15	Neoglacial	na	Mandel (2013)
13	Soil 2, Cutbank #3, Dugout Creek	TX	36.03	100.86	Ab-Bw	< 1150 + 15	Medieval Warm Period	na	Mandel (2013)
14	Soil 3, Cutbank #3, Dugout Creek	TX	36.03	100.86	Ab-Bw	< 1150 + 15	Medieval Warm Period	na	Mandel (2013)
15	Soil 2, Dugout Creek Site^	TX	36.03	100.86	Ab-Abb-Bwb	1280 + 15	Neoglacial/Medieval Warm	na	Mandel (2013)
16	Soil 2, Armstrong Creek, Caddo Co.	OK	35.52	98.62	Ab-Bwb-Bkb	2170 + 100	Neoglacial	-21.6	Zung & Mandel (in prep)
17	Soil 3, Delaware Canyon, Caddo Co.	OK	35.04	98.17	A-Bk[‡]	940 + 100	Medieval Warm Period	na	Ferring 1986, Holliday 1992 (p. 14)

[§] - Soil assigned to multiple climate periods due to uncertainty of the age

^ - Soil assigned to multiple climate periods because age straddles a break between periods

[†] - Carbonates in the Bk horizon of soil 4 may not be pedogenic. See text for additional details.

[‡] - K_{sat} estimated for soil __ exceeds 120 cm/day threshold. See text for additional details.

[a] - Based upon a stratigraphically equivalent soil

Table 3 – Climate reconstructions from the BuSCR model, modern climate values, and anomalies of past climate from modern climate for the 17 soils considered in this study.

Soil number	Latitude (°N)	Longitude (°W)	Horizonation	Climate Period	Reconstructed Climate			Modern Climate			Climate Anomaly		
					I_m	MAP	MAT	I_m	MAP	MAT	I_m	MAP	MAT
1	43.55	103.39	A-Bk-C	Neoglacial	-0.46	351	9.5	-0.25	450	7.9	-0.21	-99	1.6
2	41.45	98.83	Ab-Bwb-BCb-Cb	Neoglacial	-0.41	425	10.9	-0.08	624	9.6	-0.33	-199	1.3
3	41.03	99.39	Ab-ACb-Cgb	Medieval Warm Period	-0.45	405	11.9	-0.15	572	9.5	-0.30	-167	2.4
4	41.03	99.39	Ab-ACb-Cb	Neoglacial	-0.47	411	13.3	-0.15	572	9.5	-0.32	-161	3.8
5	40.50	101.42	A-Bk	Neoglacial/Altithermal	-0.48	381	12.0	-0.29	491	10.4	-0.19	-110	1.6
6	40.50	101.42	A-AB-Bk-C	Neoglacial/Altithermal	-0.51	378	13.4	-0.29	491	10.4	-0.22	-113	3.0
7	39.67	99.11	Ab-Bk-BCk	Early Holocene	-0.48	403	13.7	-0.19	616	12.0	-0.29	-212	1.7
8	39.50	96.50	Ab-Abb-Btb-Btgssb	Altithermal	-0.29	539	11.8	0.08	860	12.1	-0.37	-320	-0.3
9	39.50	96.50	Ab-ABb-Btb	Altithermal	-0.40	453	12.2	0.08	860	12.1	-0.48	-407	0.1
10	39.08	96.75	Ab-ABkb-Bkb	Medieval Warm Period	-0.38	397	8.6	0.09	851	12.7	-0.47	-454	-4.1
11	38.31	101.16	Ab-Bkb-BCkb	Neoglacial/Medieval Warm	-0.48	389	12.5	-0.36	481	12.0	-0.12	-92	0.5
12	36.03	100.82	Ab-Bk	Neoglacial	-0.52	365	13.4	-0.36	524	13.9	-0.16	-158	-0.5
13	36.03	100.86	Ab-Bw	Medieval Warm Period	-0.49	399	13.2	-0.36	524	13.9	-0.13	-125	-0.7
14	36.03	100.86	Ab-Bw	Medieval Warm Period	-0.47	406	12.6	-0.36	524	13.9	-0.11	-118	-1.3
15	36.03	100.86	Ab-Abb-Bwb	Neoglacial/Medieval Warm	-0.46	402	12.1	-0.36	524	13.9	-0.10	-122	-1.8
16	35.52	98.62	Ab-Bwb-Bkb	Neoglacial	-0.50	386	13.4	-0.20	697	15.2	-0.30	-311	-1.8
17	35.04	98.17	A-Bk	Medieval Warm Period	-0.43	408	11.1	-0.15	776	16.1	-0.28	-368	-5.0

Table 4 - Climate reconstructions of MAT and I_m from the BuSCR model and MJT from the Nordt et al. (2007) forest-grassland ecotone and grassland models, modern climate, and anomalies of past climate from modern climate for the six soils with $\delta^{13}C$ data.

					Reconstructed Climate				Modern Climate			Climate Anomaly		
					Ecotone	Grassland								
Soil number	Latitude (°N)	Longitude (°W)	Climate Period	$\delta^{13}C$	MJT	MJT	I_m	MAT	MJT	I_m	MAT	MJT	I_m	MAT
1	43.55	103.39	Neoglacial	-18.0	25.3	22.6	-0.46	9.5	21.4	-0.25	7.9	3.9	-0.21	1.6
5	40.5	101.42	Neoglacial/Altithermal	-15.8	26.8	24.1	-0.48	12.0	24.2	-0.29	10.4	2.6	-0.19	1.6
6	40.5	101.42	Neoglacial/Altithermal	-15.0	27.3	24.6	-0.51	13.4	24.2	-0.29	10.4	3.1	-0.22	3.0
8	39.5	96.5	Altithermal	-13.4	28.3	25.7	-0.29	11.8	26.0	0.08	12.1	2.3	-0.37	-0.3
9	39.5	96.5	Altithermal	-13.7	28.1	25.5	-0.40	12.2	26.0	0.08	12.1	2.1	-0.48	0.1
16	35.52	98.62	Neoglacial	-21.6	23.0	20.1	-0.50	13.4	27.6	-0.20	15.2	-4.6	-0.30	-1.8

CONCLUSION

Resilience, as defined first by Holling (1973) and later by Walker et al. (2004), is the capacity of a complex system to undergo change or disturbance while still retaining the same function, structure, and identity. Landscape resilience requires adaptability to external drivers while still enabling the system to function and serve a defined purpose (Folke et al. 2010). In other words, resiliency estimates the ability of a system to sustain shocks and still function for a defined purpose. Shocks to landscape resilience could include major geomorphic disturbances as occur with natural disasters or human land use or climate change. Climate change challenges landscape resilience because availability of key resources (e.g. water, cultivable soil, forage for grazing animals, habitable dwellings) can be affected due to dramatic biotic and geomorphic shifts that result from climatic shifts. For this discussion, I define the function and purpose of the semi-arid landscape in terms of the people who live there and use its resources. Specifically, a semi-arid landscape must produce food, either through livestock, crops or both, to sustain the people living there and be comfortably habitable.

Understanding resilience to climate change in semi-arid landscapes, like the Great Plains of North America or the African Savanna is especially tricky because these systems display bistability. Bistable ecosystems have more than one stable state and frequently a threshold or potential barrier between the two states (D'Odorico and Bhattachan 2012). When a shock occurs in semi-arid regions that exhibit bistability, the landscape may move to its alternate stable state, which does not support food production (D'Odorico and Bhattachan 2012). Further, because the alternate state is also stable, this landscape regime shift is highly irreversible (D'Odorico and Bhattachan 2012). The dominance of blowing dust and unarable soil on the Great Plains during the Dust Bowl is an example of a shift to an alternate stable state. The potential for shifting into an alternate stable state is a key concern considering the paleoenvironmental record indicates that some of the most agriculturally productive regions of the world, including the Great Plains of North America, display bistability.

Paleoenvironmental and paleoclimatological reconstructions elucidate patterns of environmental change and stable states during our most recent geologic past, furnishing a perspective on landscape resiliency beyond the scale of a human lifetime. Since most science is conducted on time scales far shorter than thousands of years, paleonvironmental studies supply scientists and policy makers with a long-term perspective on climate variability that can, in some cases, redefine what is considered "stable" for a landscape. Further, through quantitative paleoclimate modeling, such as the BuSCR model, we move beyond qualitative descriptions of past climatic conditions to understand the mechanisms of landscape change and the climatic thresholds that, when surpassed, cause landscape instability or a shift to an alternate stable state to occur. Numerical estimates of the precipitation, temperature, and moisture balance associated with landscape instability, e.g. widespread flooding or dune activation, facilitates scenario building, whereby landscape impacts associated with precipitation and temperature benchmarks are forecasted for a region under various climate change projections. In other words, quantitative paleoclimate reconstructions aid in defining the threshold associated with the divide between two alternate stable states in a bistable system

Paleoenvironmental reconstructions around the world indicate that over at least the last 10,000 years semi-arid regions repeatedly experienced long periods of drought, some lasting as long as 4000 years (Gasse 2000; Forman et al. 2001; Booth et al. 2005; Kiage and Liu 2006). During these times, widespread dune activation, decreasing lake levels, and shifts in vegetation are typically recorded. MAP during the Medieval Warm Period was hindcasted as 25-30% below present day in the central and southern Plains of North America (Feng et al. 2008), while growing season deficits are estimated as >25% (Forman et al. 2001). Models of future climate change project precipitation anomalies for the Great Plains similar to conditions during the Medieval Warm Period and Dust Bowl (Karl et al. 2009). Widespread dune activation and dust storms have been recorded for both of these periods. The fact that we may soon cross a threshold over which landscape stability cannot be maintained on the U.S. Great Plains begs the question: Can humans engineer landscape stability in semi-arid regions around the world through their land use practices?

U.S. government response to the Dust Bowl of the 1930s indicated that we believed we could. In reaction to the widely held belief that agricultural practices that failed to protect soils from erosion were at least in part responsible for the Dust Bowl, the Department of the Interior established the Soil Conservation Service (SCS) in 1933. The primary goal of this organization was to educate farmers and ranchers on soil conservation techniques such as contour plowing, carrying capacity, and crop rotation. The SCS later became the modern-day Natural Resources Conservation Service (NRCS) that continues today to educate farmers and facilitate soil conservation programs. In recent years, the NRCS has introduced the Conservation Reserve Program (CRP), which pays farmers to remove marginal lands from crop rotation in order to allow soil nutrient stores to rebuild, and promotes no till farming. The NRCS has become an invaluable partner in educating farmers and ranchers regarding the latest science on soil conservation, and preserving the nation's soil resource.

While CRP has been popular with farmers and the public, and cited by the USDA as successful in decreasing soil erosion and protecting sensitive environments, economic interests threaten the program. Specifically, the need to decrease U.S. government expenditures and increased demand for biofuels means millions of acres that have been set aside to protect soil resources, may move back into crop rotation in the coming years. If programs like CRP are phased out, the acreage of semi-arid land at risk of disturbance will increase and the ability of the Great Plains to sustain a major shock, such as prolonged drought due to future climate change, will decrease in proportion.

Despite the apparent success of conservation programs such as CRP, recent research on maintaining landscape stability in semi-arid regions calls into question long-held truths about carrying capacity in semi-arid grazed lands. Carrying capacity is defined as the maximum number of animals able to graze a plot of land while maintaining vegetative quality and landscape stability. Conventional wisdom has associated overgrazing or exceeding the carrying capacity with desertification. Flying in the face of conventional wisdom, Allan Savory presented research in a recent TED Talk supporting holistic management of grasslands. He suggests that large, dense, moving herds of grazing animals prevents desertification of grasslands and savannas and maintains landscape stability. He explains that the addition of nitrogen through livestock excrement and trampling of above-ground biomass accelerates decomposition, and creates a positive feedback promoting grassland health. He further critiques the assumption that grazing practices by groups such as pastoralists of eastern Africa created desertification, instead proposing that these highly mobile societies with large moving herds execute an optimum management strategy for maintaining landscape stability in semi-arid regions.

The Karamojong of northeastern Uganda are an excellent example of a people whose traditional livelihood exemplifies the methods Savory proposes for maintaining landscape resilience yet, U.S. foreign aid policy, diplomatic relationships, and geopolitics create significant barriers to continuing this way of life. The Karamojong are a herding people who typically move their family herds of 25 cattle and 50 goats as dense packs across range 550 to 2000 km^2 in area in the Eastern Africa (Levine 2010). Research indicates that pastoral or agro-pastoral livelihoods are most resilient in this semi-arid region (Levine 2010). State boundaries between Uganda and Kenya as well as intertribal conflict, however, greatly limit movement of the Karamojong, and lack of formalized methods for protecting communal land tenure rights create additional barriers to open range.

Posing additional challenges, USAID efforts in Uganda are aimed at supporting the farmer. For example, Feed the Future initiatives in Uganda focus on promoting production of three cash crops – maize, coffee, and beans, completely ignoring pastoralism and northeastern Uganda altogether. This strategy supports the majority of Ugandans, who are farmers, but highlights the problem with appropriating foreign aid budgets by country rather than by region and in consideration of various lifeways. Championing the farmer supports the views of the President and First Lady of Uganda, key U.S. allies in central Africa. In fact, the First Lady and Minister of Karamoja Affairs, Janet Museveni, wrote in a 2011 letter to the European Union of the many "dangers of pastoralism," saying it was a "social ill" and that the government should help the Karamojong "become civilized and settle down" (Vidal 2011). Therefore, while the Karamojong pastoral livelihood may actually promote landscape resilience in eastern Uganda and provide a sustainable foodway for the region, foreign aid operations, geopolitics, outright bigotry, and systems of power significantly challenge the Karamojong in retaining this lifeway and culture.

In summary, paleoclimatic reconstructions provide a more wholistic long-term view of landscape resiliency and the thresholds that exist, especially in semi-arid regions. Information gleaned from these reconstructions indicate that when thresholds are crossed, landscapes change to an unarable stable state. Also, indigenous knowledge provides a long-term perspective, yet there are significant cultural and geopolitical challenges to implementing “indigenuity” as a mitigation strategy for climate change. Economic pressures driving land use in semi-arid regions of the U.S. and abroad further challenge implementation of consevation practices that support landscape resilience under the pressure of future climate change. Critical examination of land-use policies in the U.S., as well as our aid efforts abroad, in the context of long-term landscape stability may be required to ensure resilient semi-arid landscapes in the face of climate change.

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