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Forces driving late Pleistocene (ca. 77–12 ka) landscape evolution in the Cimarron River valley, southwestern Kansas

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ABSTRACT

This study presents stratigraphic, geomorphic, and paleoenvironmental ($\delta^{13}C$) data that provide insight into the late Pleistocene landscape evolution of the Cimarron River valley in the High Plains of southwestern Kansas. Two distinct valley fills (T-1 and T-2) were investigated. Three soils occur in the T-2 fill and five in the T-1 fill, all indicating periods of landscape stability or slow sedimentation. Of particular interest are two cumulic soils dating to ca. 48–28 and 13–12.5 ka. $\delta^{13}C$ values are consistent with regional paleoenvironmental proxy data that indicate the prevalence of warm, dry conditions at these times. The Cimarron River is interpreted to have responded to these climatic changes and to local base level control. Specifically, aggradation occurred during cool, wet periods and slow sedimentation with cumulic soil formation occurred under warmer, drier climates. Significant valley incision (~25 m) by ca. 28 ka likely resulted from a lowering of local base level caused by deep-seated dissolution of Permian evaporite deposits.

Introduction

Despite long-term interest in the late-Pleistocene stratigraphy and paleoenvironments of the High Plains (e.g., Antevs, 1935; Frye and Leonard, 1952; Haynes and Agogino, 1966; Holliday et al., 1994, 1996, 2011; Holliday, 1995, 2000), significant knowledge gaps still exist. Previous studies in the central High Plains (western Kansas, southwestern Nebraska and eastern Colorado) have typically focused on late-Quaternary eolian sands and loess (e.g., Forman et al., 1995; Maat and Johnson, 1996; Olson et al., 1997; Porter, 1997; Muhs et al., 1999, 2008; Olson and Porter, 2002; Johnson et al., 2007) whereas research on alluvial deposits has primarily been directed toward Holocene-aged fills (e.g., Johnson and Martin, 1987; Mandel, 1994; Bettis and Mandel, 2002). In contrast, with the exception of studies during the mid-twentieth century that focused on basic lithostratigraphic and paleontological relationships (e.g., Smith, 1940; Frye and Hibbard, 1941; Frye and Leonard, 1952; Gutentag, 1963), little work has been undertaken on the Pleistocene alluvium in this region. Our study expands upon this early work by assessing stratigraphic, geomorphic, and pedologic data from late-Pleistocene (Marine Oxygen Isotope Stage (MIS) 4–2, ca. 77–12 ka) alluvial fills stored in the Cimarron River valley, southwestern Kansas. Also, stable carbon isotope ($\delta^{13}C$) data from buried soils provide insight into late Pleistocene paleoenvironmental change and subsequent alluvial response. While $\delta^{13}C$ analyses have been successfully employed in Holocene alluvial environments (e.g., Nordt et al., 1994; Baker et al., 2000; Cordova et al., 2011; Mandel, 2013), here we report data from the late Pleistocene, a time interval for which such records are scarce in the Great Plains.

Study area

Physiography and geology

The study area is located in southwestern Kansas, which is part of the High Plains region of the Great Plains physiographic province (Fig. 1A; Fenneman, 1931). The High Plains covers ~450,000 km^2 and represents the remnant of an extensive alluvial apron formed by sediment shed from the Rocky Mountains. The High Plains is predominantly underlain by the Miocene and early Pliocene-aged sediments of the Ogallala Formation together with Quaternary alluvial and eolian deposits. Topographic relief is largely confined to stream valleys in which the Ogallala Formation crops out and Quaternary alluvium is stored.

Interfluves and alluvial terraces in the High Plains are mantled by at least three late-Quaternary loesses: the Loveland, Gilman Canyon, and Peoria (e.g., Muhs et al., 1999; Bettis et al., 2003). The Loveland loess is typically the oldest loess unit exposed and dates to the

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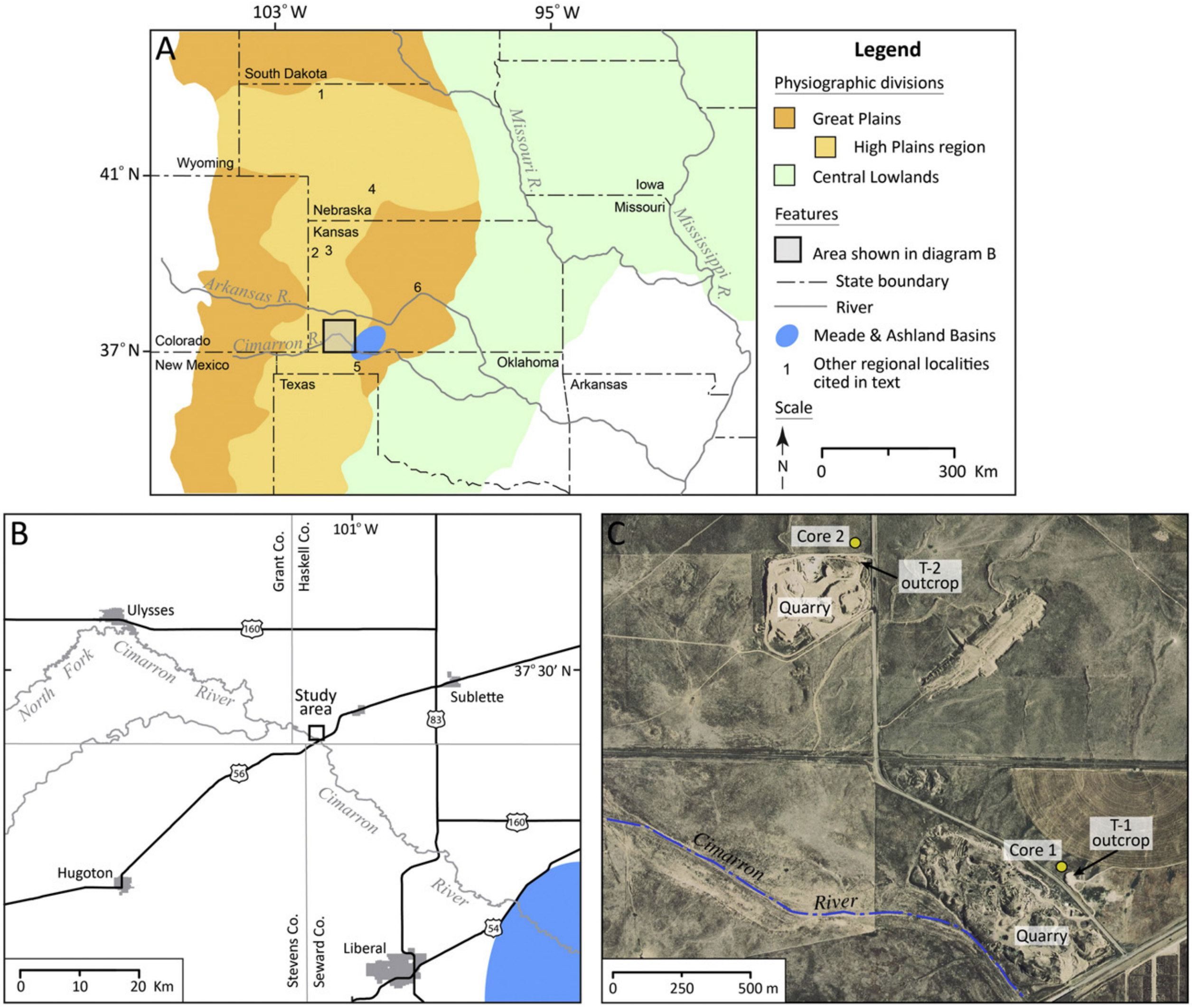

Figure 1. (A) Regional map showing the central High Plains and paleoenvironmental localities mentioned in text: 1) Cobb Basin, 2) Kanorado, 3) Willem Ranch, 4) Eustis Ash Pit, 5) Bull Creek, 6) Cheyenne Bottoms. (B) Study area location in the central High Plains, southwestern Kansas. (C) Aerial photograph of study area with quarry outcrop and core locations.

penultimate glaciation (MIS 6). The Gilman Canyon Formation overlies the Loveland Loess and is usually < 2 m thick (Bettis et al., 2003). The Gilman Canyon Formation typically consists of a dark noncalareous silt loam that has been modified by pedogenesis, commonly with two or more soils forming a pedocomplex (e.g., Reed and Dreeszen, 1965; Mandel and Bettis, 1995; Johnson et al., 2007). Radiocarbon ages from sites in the Great Plains indicate that the Gilman Canyon Formation dates between ca. 40,000 and 20,000 ^{14}C yr BP (Muhs et al., 1999; Johnson et al., 2007). The Peoria Loess mantles the Gilman Canyon Formation and is the thickest and most areally extensive loess deposit in the Great Plains (Bettis et al., 2003). Radiocarbon ages indicate that Peoria Loess deposition began ca. 30,000–20,000 ^{14}C yr BP and continued to accumulate until ca. 10,000 ^{14}C yr BP (e.g., Bettis et al., 2003; Muhs et al., 2008).

We investigated late-Quaternary alluvium stored in the Cimarron River valley of southwestern Kansas. Valley fills underlying two terraces (T-1 and T-2) were investigated in two sand and gravel quarries in the southwest corner of Haskell County (Fig. 1C). The T-1 and T-2 terrace treads stand 9 and 24 m above the modern river channel, respectively (Fig. 2A). The Cimarron River, a tributary of the Arkansas River, originates on the High Plains in northeastern New Mexico and drains ~49,000 km^2 (Fig. 1A). In the study area, the valley width is ~3 km, with ~45 m of elevation change between the uplands and the valley floor.

Climate and vegetation

The climate of southwestern Kansas is a semiarid steppe (Peel et al., 2007). Mean July and January temperatures for the period of record (AD 1918–2014) at Sublette, Kansas are 26°C and 0°C, respectively, and mean annual precipitation is 47.3 cm (High Plains Regional Climate Center, 2014). Approximately 75% of the precipitation falls during the months of April through September, largely as a result of frontal activity. The collision of Pacific and polar air masses with warm, moist maritime-tropical air from the Gulf of Mexico often produces intense rainfall of short duration along the zone of convergence. Periodic intensification of westerly (zonal) airflow, however, prevents moist Gulf air from penetrating into the High Plains and tends to cause drought in the region.

The natural vegetation of southwestern Kansas is short grass prairie (see Küchler, 1974). Uplands are dominated by C_4 plant communities, including blue gramma (*Bouteloua gracilis*) and buffalograss (*Buchloe dactyloides*). C_3 broadleaf deciduous trees,

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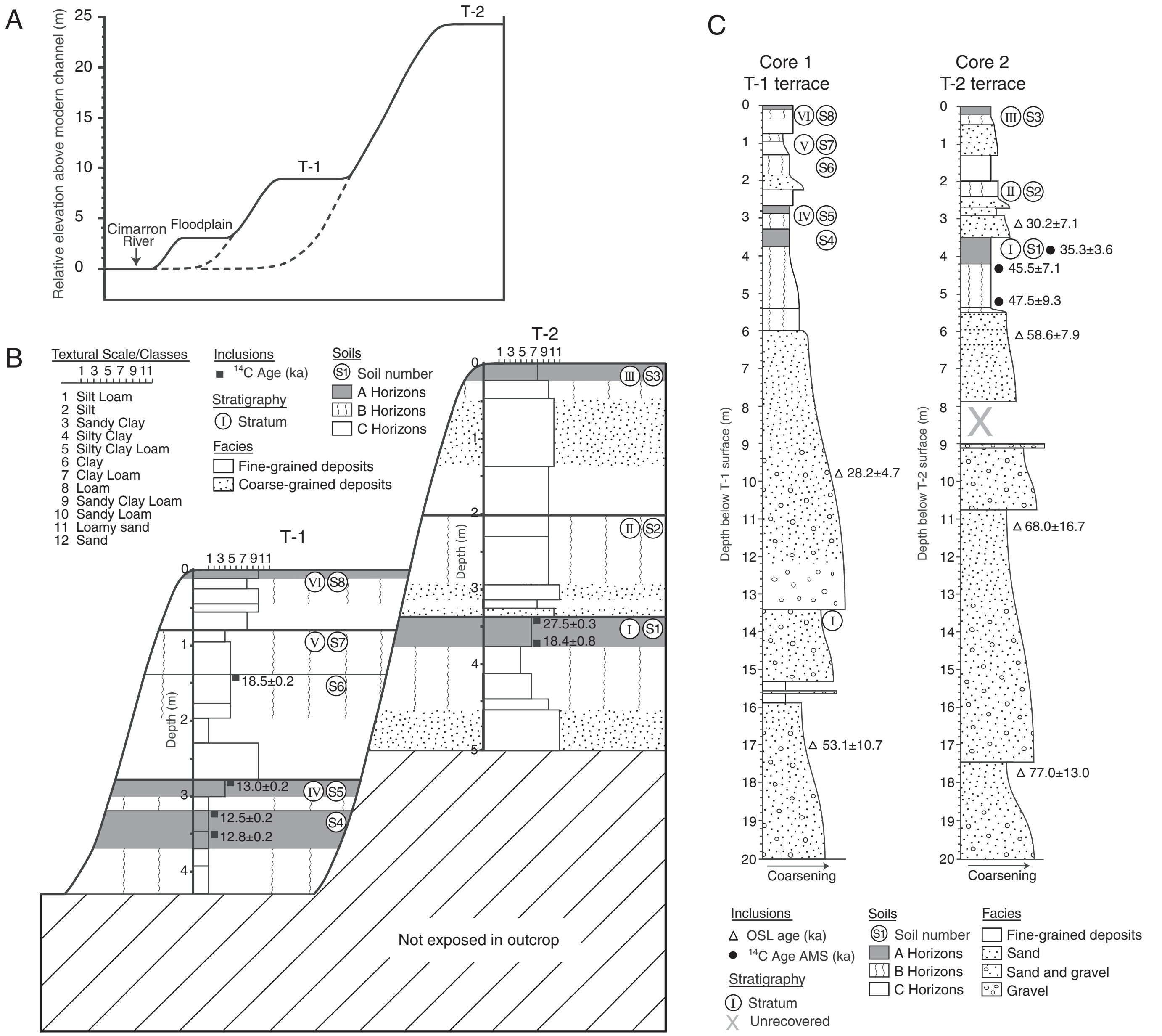

Figure 2. (A) Terrace stratigraphy and tread elevations relative to the stream channel. (B) Composite cross-section of alluvial fills exposed in quarries in the Cimarron River valley. Note that terrace fills are not to scale relative to each other. (C) Stratigraphy of soils and sediments from core. Note ^{14}C ages on figure are median calibrated ages reported in Table 1.

including cottonwood (*Populus sargentii*) and peachleaved willow (*Salix amygdaloides*) dominate riparian vegetation.

Methodology

Stratigraphic, sedimentological, and pedological descriptions were prepared for two sections exposed in outcrop (Figs. 2B and 3). Also, a 20-m-deep core was collected from each of the alluvial fills underlying the T-1 and T-2 terraces (Fig. 2C). Cores were taken ~25 m from the outcrop face using an Acker hollow-stem auger drill rig equipped with a split-spoon sampler. Soils in outcrop were described using standard terminology in Birkeland (1999) and Schoeneberger et al. (2012). Sedimentary features in C horizons were described where preserved. Soils were sampled by horizon for particle size (e.g., Schoeneberger et al., 2012). Where horizon thicknesses were > 50 cm, subsamples were collected so that the sample interval was no greater than 50 cm. Air-dried samples were ground to pass through a 2-mm sieve and 10 g of material was extracted for particle-size analysis using the pipette method (Soil Survey Staff, 1982). Calcium carbonate ($CaCO_3$) percent was determined by dry weight loss after HCl treatment. Organic matter content was determined with a Costech Elemental Analyzer during isotopic analysis.

Numerical-age control was provided by radiocarbon (^{14}C) dating of soil organic matter (SOM) from buried soils and by optically stimulated luminescence (OSL) dating of sediments. ^{14}C dating of SOM was selected because preferred materials, such as charcoal, are relatively scarce in buried soils on the High Plains. Ages determined on SOM provide an assessment of mean resident time for organic carbon that typically gives a minimum age for the onset of soil development. Although dating alluvial buried soils can be challenging due to potential inputs of carbon from multiple sources, ^{14}C dating of SOM has been successfully employed in alluvial settings from the Great Plains (e.g., Holliday, 1995; Mandel, 1994, 2008). ^{14}C samples collected from outcrop were submitted to the Illinois State Geological Survey (ISGS) and ^{14}C samples collected

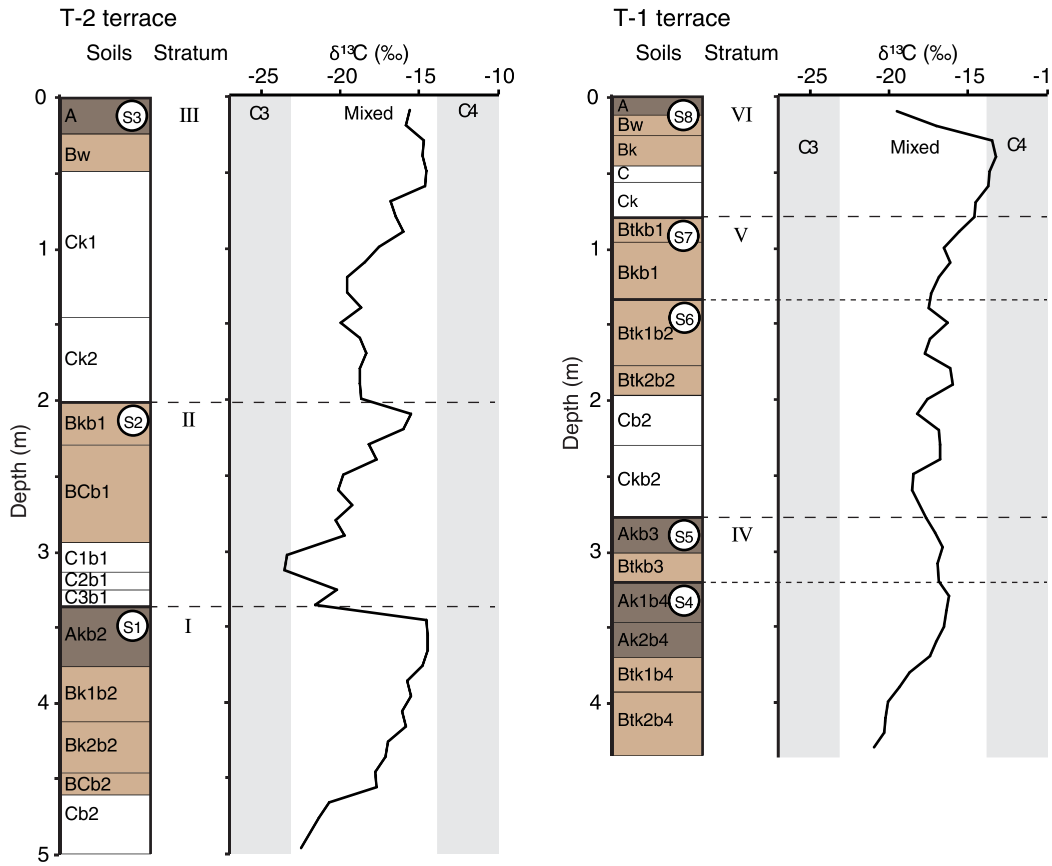

Figure 3. $\delta^{13}C$ depth plots from outcrop profiles.

from core were sent to the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) facility at the Woods Hole Oceanographic Institute. All ^{14}C samples underwent standard pretreatment, including the removal of roots and decalcification. ISGS sample dating was performed on the bulk organic fraction and ^{14}C activity was measured via the liquid scintillation technique. NOSAMS samples were combusted in a high-temperature muffle furnace and the resultant CO_2 reacted with a Fe catalyst to create graphite. After conversion to graphite, ^{14}C activity was measured via Accelerator Mass Spectrometry (AMS). All ^{14}C ages in the text are reported as median calibrated ages (ka) (Table 1) to facilitate comparisons with OSL ages.

OSL dating provides an age estimate of the time since sediment was last exposed to light (Huntley et al., 1985). Although fluvial environments can be challenging for OSL dating due to potential problems with solar resetting, termed partial bleaching, quartz small-aliquot OSL dating has been successfully applied to numerous fluvial settings (e.g., Wallinga, 2002; Rittenour, 2008). OSL samples were collected from the cores and analyzed at the Utah State University Luminescence Laboratory using the latest single-aliquot regenerative-dose (SAR) procedures for dating quartz sand (see Murray and Wintle, 2000). All samples were opened and processed under dim amber safelight conditions and were processed to isolate the fine-grained quartz sand fraction (using HCl to remove carbonates, chlorine bleach to remove organics, heavy liquid to remove accessory minerals (sodium polytungstate, 2.7 g/cm^3), and hydroflouric acid to etch the quartz and dissolve any remaining feldspars. Sample processing followed standard procedures outlined in Aitken (1998) and described in Rittenour et al. (2005). Dose rate calculations were based on ICP-MS determination of radio-elemental concentrations (Guerin et al., 2011), contribution from cosmic radiation (Prescott and Hutton, 1994), and includes attenuation by water (Aitken, 1998). OSL ages were calculated using the central age model (Galbraith and

Table 1
Radiocarbon ages.

Landform	Depth (cm)	Horizon	^{14}C age (yr BP)	Calibrated age range[1] (yr BP)	Median calibrated age (cal ka BP)	Lab no.	$\delta^{13}C$
T-1 (*Outcrop*)	190–200	Btk1b2	15,210 ± 90	18,700–18,250	18.5 ± 0.23	ISGS-6998	−6.4
	278–288	Akb3	11,110 ± 70	13,100–12,800	13.0 ± 0.15	ISGS-6999	−18.7
	320–330	Ak1b4	10,520 ± 70	12,670–12,370	12.5 ± 0.15	ISGS-7000	−16.5
	348–358	Ak2b4	10,930 ± 70	12,980–12,700	12.8 ± 0.14	ISGS-7001	−15.9
T-2 (*Outcrop*)	338–348	Akb2	23,330 ± 170	27,800–27,290	27.5 ± 0.26	ISGS-7010	−15.5
	367–377	Akb2	14,710 ± 130	19,240–17,570	18.4 ± 0.84	ISGS-7011	−13.4
T-2 (*Core*)	380	Akb2	31,300 ± 1650	38,950–31,720	35.3 ± 3.62	OS-111487	−17.3
	420	Bk1b2	42,400 ± 6600	52,620–38,380	45.5 ± 7.12	OS-111488	−14.3
	520	Bk2b2	45,100 ± 9300	56,730–38,210	47.5 ± 9.26	OS-111489	−13.9

[1] Calibration to calendar years (2 sigma) was performed with CALIB 7.0 using calibration dataset IntCal13 (Reimer et al., 2013). ^{14}C ages older than 40,000 yr were calibrated with CalPal using the calibration dataset CalPal2007_HULU.

Roberts, 2012) except USU-1588, which was calculated using the minimum age model (Galbraith and Roberts, 2012) due to evidence for partial bleaching.

Paleoenvironmental reconstructions were undertaken by analyzing stable carbon isotopes ($\delta^{13}C$) of SOM from buried soils, which reflect the relative contributions of C_3 and C_4 vegetation to below ground biomass production at the time of soil formation (e.g., Boutton, 1991). This technique has been successfully applied in many paleoenvironmental studies conducted in alluvial settings (e.g., Nordt et al., 1994, 2002; Fredlund and Tieszen, 1997; Mandel, 2008; Cordova et al., 2011). We estimate the relative contribution of C_4 plants to SOM with the following mass balance equation (Nordt et al., 1994)

$$\delta^{13}C_{SOM} = \delta^{13}C_{C4}(x) + \delta^{13}C_{C3}(x)(1-x) \quad (1)$$

where $\delta^{13}C_{SOM}$ is the measured $\delta^{13}C$ value, $\delta^{13}C_{C4} = -13‰$, $\delta^{13}C_{C3} = -27‰$ and x is the relative contribution of C_4 plants to SOM. Studies have shown a strong correlation between the proportion of C_4 vegetation and temperature (e.g., Boutton, 1996; Terri and Stowe, 1976; von Fischer et al., 2008), indicating that $\delta^{13}C$ values can function as an important proxy for paleoclimatic change. We reconstruct mean July temperatures based on $\delta^{13}C$ values from the A horizons of buried soils with the following equation (Nordt et al., 2007):

$$T(°C) = 0.685(\delta^{13}C) + 34.9 \quad (2)$$

$\delta^{13}C$ analysis was performed at the University of Kansas Keck Paleoenvironmental and Environmental Stable Isotope Laboratory (KPESIL). Bulk SOM samples were collected from outcrop at 10-cm intervals and prepared for $\delta^{13}C$ analysis following Boutton (1996). Samples were oven-dried, decarbonated with 10 ml of 0.5 M HCl, and then centrifuged and rinsed with deionized water until attainment of a neutral pH. Samples were dried at 60°C and then ground to powder with a mortar and pestle. Between 5 and 35 mg of decarbonated sample was weighed into tin capsules, combusted using a Costech Elemental Analyzer at 1060°C, and analyzed with a continuous-flow ThermoFinnigan MAT 253 mass spectrometer. Final $\delta^{13}C$ data were calibrated with National Institute Standards and are expressed relative to the Vienna Peedee Belemnite standard.

Results

Stratigraphy and chronology

The highest and oldest terrace (T-2) stands ~24 m above the modern river channel (Fig. 2A). The T-2 fill is over 20 m thick and three stratigraphic units were identified (Fig. 2B and C). The portion of the T-2 fill exposed in outcrop mostly consists of sandy loam, loamy sand, and sand in which three soils developed (Fig. 2B; see Table 3 for soil properties and Supplementary Fig. 1 for depth plots of organic carbon, $CaCO_3$ and clay content). Stratum I primarily consists of vertically stacked channel-fill sequences that fine upward and range in texture from fine sand to fine gravel. Horizontal bedding and trough cross-stratification are evident in outcrop. Sedimentological characteristics indicate deposition by a high-energy braided stream system. OSL ages of 77.0 ± 13.0, 68.0 ± 16.7 and 58.6 ± 7.9 ka at 17.7, 11.1 and 6.2 m depths, respectively, indicate that this depositional environment prevailed in the study area during late MIS 5 and MIS 4 (Fig. 2C; Table 2). The moderately developed cumulic soil (S1), formed in the top of stratum I, is 1.25 m thick (2 m thick in core) and has an Ak-Bk profile. The variability observed in horizon thickness between outcrop and core is likely due to localized truncation by the stream channel. SOM from the upper and lower 10 cm of the Ak horizon in outcrop yielded ages of 27.5 ± 0.26 ka and 18.4 ± 0.84 ka, respectively (Fig. 2B; Table 1). Three AMS ^{14}C ages from SOM were also obtained from the middle of the Ak horizon and the top and bottom of the Bk horizon in core. AMS samples yielded ages of 35.3 ± 3.62, 45.5 ± 7.12 and 47.5 ± 9.26 ^{14}C ka BP, respectively (Fig. 2C; Table 1). An OSL sample from the C horizon of soil 3 in core yielded an age of 58.6 ± 7.9 ka (Fig. 2C; Table 2).

Stratum II is 1.35 m thick and mostly consists of loamy sand and sand (Fig. 2B) in which a moderately developed soil (S2) formed. Soil S2 is 1.92 m thick with a 27 cm-thick Bk horizon. Primary bedding structures (cross stratification and gravel lenses) are preserved and gravel lenses consist of many stream-worn carbonate nodules that were likely derived from carbonate-rich soils developed in the Ogallala Formation. An OSL sample from the C horizon of soil 2 in core yielded an age of 30.2 ± 7.1 ka (Fig. 2C; Table 2).

Stratum III is ~2 m thick and lacks primary bedding structures. The sand fraction in the Ck1 and Ck2 horizons mostly consists of medium (500–355 μm) to fine (125–90 μm) and fine to very fine sand (125–63 μm), respectively. Based on these findings and the fact that eolian deposits typically blanket the upland landscape in the study area, we suspect that the massive, unstratified sediments on the T-2 surface, which represent the parent material for the surface soil (S3), are eolian. However, we cannot rule out an alluvial origin for these sediments. The modern surface soil (S3) is weakly developed with a 45 cm-thick A-Bw profile (Table 3).

The youngest terrace (T-1) stands ~9 m above the modern river channel (Fig. 2A). Fluvial sediments from core 1, collected from the alluvium underlying the T-1 terrace, produced OSL ages of 53.1 ± 10.7 ka at ~17 m depth and 28.2 ± 4.7 ka at ~10 m depth (Fig. 2C; Table 2). The

Table 2
Optically stimulated luminescence age information.

Core	Lab no.	Depth (m)	Num. of aliquots[1]	K[2] (%)	Rb[2] (ppm)	Th[2] (ppm)	U[2] (ppm)	Dose Rate[3] (Gy/ka)	Equivalent Dose, D_E[4] ± 2σ (Gy)	OD[5] (%)	OSL age ± 2σ (ka)
Core 1	USU-1581	9.9	20 (34)	1.68 ± 0.04	50.8 ± 2.0	2.6 ± 0.2	0.8 ± 0.1	1.82 ± 0.11	51.42 ± 6.42	25.7 ± 4.8	28.20 ± 4.68
	USU-1582	17.1	12 (49)	2.58 ± 0.06	77.1 ± 3.1	5.0 ± 0.5	1.1 ± 0.1	2.75 ± 0.17	145.7 ± 24.26	26.9 ± 6.4	53.07 ± 10.64
Core 2	USU-1588	3.2	15 (27)	2.16 ± 0.05	68.7 ± 2.7	4.8 ± 0.4	1.1 ± 0.1	2.76 ± 0.14	83.46 ± 17.73[6]	37.3 ± 7.5	30.19 ± 7.11
	USU-1589	6.2	15 (29)	2.22 ± 0.06	68.8 ± 2.8	3.8 ± 0.3	1.0 ± 0.1	2.43 ± 0.15	142.5 ± 11.40	12.1 ± 3.5	58.61 ± 7.92
	USU-1583	11.15	8 (47)	1.05 ± 0.03	36.7 ± 1.5	3.0 ± 0.3	0.9 ± 0.1	2.51 ± 0.16[7]	170.9 ± 37.49	29.7 ± 8.1	68.04 ± 16.69
	USU-1584	17.7	10 (44)	2.47 ± 0.06	78.7 ± 3.1	3.8 ± 0.3	1.0 ± 0.1	2.59 ± 0.16	199.0 ± 25.54	17.4 ± 5.2	76.97 ± 13.03

[1] Number of aliquots used in age calculation and number of aliquots analyzed in parentheses.
[2] Radioelemental concentrations determined by ALS Chemex using ICP-MS and ICP-AES techniques.
[3] Dose rate is derived from concentrations by conversion factors from Guerin et al. (2011), and includes cosmogenic contribution to total dose rate using latitude, longitude, elevation and depth following Prescott and Hutton (1994). Assumed 13.8% as average moisture content for all samples below 5 m depth, assumed 5% for USU-1588, and *in-situ* 6% for USU-1518.
[4] D_E calculated using the single-aliquot regenerative-dose procedure of Murray and Wintle (2000) on 1-mm small-aliquots of quartz sand (90–250μm). Weighted mean of D_E calculated using the Central Age Model (CAM) of Galbraith and Roberts (2012), except where noted otherwise.
[5] Overdispersion (OD) calculated as part of the CAM and represents variance in D_E data beyond measurement uncertainties, OD > 20% may indicate significant scatter due to depositional or post-depositional processes.
[6] D_E calculated using the Minimum Age Model of Galbraith and Roberts (2012).
[7] Dose rate calculated using chemistry data from USU-1584 and USU-1589.

Table 3
Soil descriptions from the T-2 outcrop profile.

Horizon	Depth (cm)	Color		Texture[1]	Structure[2]	Consistence[3]	Clay Films[4]	Boundary[5]	Pores[6]	Roots[6]	Notes
		Dry	Moist								
A	0–22	10YR 4/2	10YR 3/2	L	1 f sbk	vfr	---	c	2vf	1m 2f 3vf	Parts to 1 f gr; Common worm casts
Bw	22–45	10YR 5/2	10YR 4/1	SL	2 f sbk	fr	---	c	1f 2vf	1f 2vf	Parts to 2 f gr; Common worm casts
Ck1	45–137	10YR 6/4	10YR 5/4	LS	sg	lo	---	a	---	1f 2vf	Common carbonate nodules
Ck2	137–203	10YR 6/4	10YR 5/4	SL	1 f sbk	fr	---	a	2vf	1f 1vf	Parts to sg; Few carbonate nodules
Bkb	203–230	10YR 5/4	10YR 4/4	SL	2 m sbk	sh, vfr	---	g	1m 2f 3vf	2vf	Parts to 1 f sbk and 1 f gr; Common fine threads and common patchy distinct coats on ped faces
BCb	230–295	10YR 6/6	10YR 5/6	SL	2 f sbk	so, vfr	---	c	3f 2vf	1vf	Parts to 1 f gr and sg
C1b	295–316	10YR 6/6	10YR 5/6	S	sg	lo	---	a	---	---	Cross-stratification; gravel lenses
C2b	316–326	10YR 6/4	10YR 5/4	L	1 f sbk	fr	---	a	2vf	---	Parts to sg
C3b	326–338	10YR 6/6	10YR 5/6	LS	sg	lo	---	a	---	---	Thin gravel lenses
Akb2	338–377	10YR 5/3	10YR 4/3	CL	2 m pr	h, fi	---	g	3f 3vf	---	Parts to 1 f-m sbk; Few fine-medium carbonate threads
Bk1b2	377–414	10YR 6/3	10YR 5/3	SICL	2 m pr	h, fi	---	g	1m 3f 3vf	---	Parts to 2 f-m sbk; Few fine carbonate threads; Few wormcasts in burrows
Bk2b2	414–448	10YR 7/3	10YR 6/4	CL	2 f pr	h,fr	---	g	2f 3vf	---	Parts to 1 f sbk and 1 f gr; Common fine carbonate threads
BCb2	448–463	10YR 7/4	10YR 6/4	SL	1 m pr	so,vfr	---	c	3f 2vf	---	Parts to 1 f sbk and sg
Cb2	463–500+	10YR 6/4	10YR 5/4	S	sg	lo	---	---	---	---	

[1] Texture: C – Clay, CL – Clay Loam, SiCL – Silty Clay Loam, SiC – Silty Clay, L – Loam, SiL – Silty Loam, LS – Loamy Sand.
[2] Structure: 1 – weak, 2 – moderate, 3 – strong, m – massive, sg – single grain; f – fine, m – medium, c – coarse; abk – angular blocks, sbk – subangular blocks, pl – plates.
[3] Consistence: so – soft, sh – slightly hard, h – hard, vh – very hard; lo – loose, vfr – very friable, fr – friable, fi – firm; vfi – very firm.
[4] Clay Films: 1 – few, 2 – common, 3 – many; d – distinct, p – prominent; pf – ped faces, po – pores.
[5] Boundaries: a – abrupt, c – clear, g – gradual.
[6] Roots and pores: 1 – few, 2 – common, 3 – many; vf – very fine, f – fine, m – medium, c – coarse.

basal age is consistent with age estimates of the T-2 alluvial fill and is separated from the upper younger T-1 deposits by a coarse-grained unit interpreted to represent the base of channel incision into the T-2 alluvium at ~13.5 m. The T-1 alluvium is therefore inset into the T-2 fill. An alternative interpretation of the core 1 sequence is that the 53 ka sample was not fully reset prior to deposition and represents an age overestimate, although D_E distributions do not support this interpretations (see Supplementary Fig. 2).

The T-1 fill mostly consists of coarse-grained alluvium and three stratigraphic units were identified (Figs. 2B and 2C). The portion of the T-1 fill exposed in outcrop has sandy loam, loamy sand and sand textures in which five soils have developed (Fig. 2B, see Table 4 for soil properties and Supplementary Fig. 1 for depth plots of organic carbon, $CaCO_3$ and clay content).

Stratum IV is ~10 m thick and consists of sand and gravel. Two soils (S4 and S5) developed in the top of stratum IV. Soil S4 has cumulic properties with a 50 cm-thick, organic rich, Ak horizon and a 1.15 m-thick Btk horizon in core. SOM from the upper 10 cm of the Ak1 and Ak2 horizons yielded ages of 12.5 ± 0.15 and 12.8 ± 0.14 cal ka BP, respectively (Fig. 2B; Table 1). Soil S4 is similar to soil S5 in terms of its morphological expression (i.e., horizonation, structure, consistence, and carbonate morphology). SOM from the upper 10 cm of the Ak horizon yielded a radiocarbon age of 13.0 ± 0.15 cal ka BP (Fig. 2B; Table 1).

Stratum V consists of upward-fining alluvium (sandy loam to silt loam) in which two moderately expressed soils (S6 and S7) developed that were subsequently truncated by erosion. Soil S6 has a 59 cm-thick Btk horizon and SOM from the upper 10 cm of the horizon yielded an age of 18.5 ± 0.23 cal ka BP (Fig. 2B; Table 1). Soil S7 is moderately developed with Btk-Bk horizonation.

Stratum VI consists of loam and sandy loam deposits in which the modern surface soil (S8) developed. Soil S8 is weakly developed and has a 45 cm-thick A-Bw-Bk profile.

Stable carbon isotopes

$\delta^{13}C$ values of SOM from soil S1 in the T-2 fill increase from −17.8‰ at the bottom of the soil to −14.6‰ at the top, indicating the presence of a C_4-dominated environment (up to 89% C_4 biomass) during MIS 3 (ca. 48–28 ka) (Fig. 3). The increase in C_4 biomass is suggestive of climatic warming during this period; however, this trend may also reflect a change in the source of organic matter. Organic matter in late-Quaternary environments is derived from either pedogenic or detrital sources, which reflect the local vegetation growing directly in the soil and organic matter transported in alluvium from the wider drainage basin, respectively (e.g., Nordt et al., 1994, 2002). Given the cumulic nature of soil S1, the increase in $\delta^{13}C$ values may reflect an increase in the proportion of pedogenic organic matter contributed to the soil as sedimentation rates slowed during MIS 3. Temperature reconstructions indicate a mean July temperature of 25.0°C ($\delta^{13}C$ value of −14.6‰) at ca. 28 ka, which is similar to modern July temperatures in the study area. $\delta^{13}C$ values become significantly more negative (−23.5‰) in the C horizon of soil S2, suggesting the presence of a C_3-dominated community at this time (after ca. 28 ka). This shift likely reflects detrital contributions of organic matter from either older Pleistocene or Ogallala Formation alluvium in the drainage basin. It is also possible, however, that this signal represents local conditions (i.e., riparian vegetation) given that the Cimarron River headwaters on the High Plains and does not cross any major climatic or vegetation boundaries. $\delta^{13}C$ values progressively increase to −15.5‰ at the top of soil S2, indicating the presence of a C_4-dominated environment (up to 82% C_4 biomass). In soil S3, $\delta^{13}C$ values increase from −18.7 to −15.6‰ at the top of the soil. $\delta^{13}C$ values peak at −14.6‰, indicating the presence of up to 89% C_4 biomass.

The $\delta^{13}C$ values from the T-1 fill range from −21.0 to −13.2‰ (Fig. 3), indicating that vegetation communities varied from C_3-dominated to strongly C_4-dominated over the last ca. 13 ka. $\delta^{13}C$ values increase from −21.0‰ at the bottom of soil S4 to −16.3‰ near the top, indicating the presence of a C_4-dominated environment (up to 77% C_4 biomass) from ca. 13.0–12.5 ka. The increase in C_4 biomass is suggestive of climatic warming; however, as previously noted for soil S1, this trend may also reflect an increase in the contribution of local pedogenic organic matter during cumulic soil formation. Temperature reconstructions from soil S4 indicate a mean July temperature of 23.8°C at ca. 12.5 ka. $\delta^{13}C$ values decrease in soil S3 and yield a mean July temperature of 22.8°C, which is about 3.2°C cooler than modern July temperatures in the study area.

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Table 4
Soil descriptions from T-1 outcrop profile.

Horizon	Depth (cm)	Color		Texture[1]	Structure[2]	Consistence[3]	Clay Films[4]	Boundary[5]	Pores[6]	Roots[6]	Notes
		Dry	Moist								
A	0–7	10YR 4/2	10YR 3/2	SL	2 f gr	vfr	---	c	3f 3vf	3f 3vf	
Bw	7–25	10YR 4/2	10YR 3/2	L	2 f sbk	sh, vfr	---	c	1m 2f 3vf	1m 2f 2vf	Parts to 1 f gr; Common worm casts, few open burrows
Bk	25–45	10YR 6/4	10YR 5/4	SL	1 c pr	sh, vfr	---	g	1f 2vf	1m 2f 2vf	Parts to 1 f sbk and 2 f gr; Few medium soft carbonate masses; Common worm casts, few open burrows
C	45–55	10YR 6/4	10YR 5/4	SL	1 c pr	sh, vfr	---	c	2vf	1f 2vf	Few medium soft carbonate masses
Ck	55–80	10YR 7/4	10YR 5/4	L	1 c pr	sh, vfr	---	a	1f 3vf	2vf	Few thin beds, gravel lens at 80 cm; Few medium hard carbonate masses, common fine threads
Btkb1	80–94	10YR 6/4	10YR 5/4	SiC	2 c pr	h, fr	1 f pf, po	c	1f 2vf	1vf	Parts to 2 f-m abk; Patchy 10YR 4/4 clay films; Few fine hard carbonate masses, common distinct patchy coats on ped faces; Many wormcasts, Common open worm burrows
Bkb1	94–135	10YR 6/4	10YR 5/4	SiCL	2 c pr	h, fr	---	c	2f 3vf	1f	Parts to 1 m sbk; Common fine hard carbonate masses, few fine threads and common distinct patchy coats on ped faces; Many wormcasts, common open worm burrows
Btk1b2	135–168	10YR 6/4	10YR 5/4	SiCL	2 m pr	h, fr	2 d pf	g	3f 3vf	3vf	Parts to 2 f-m abk; Discontinuous 10YR 4/3 clay films; Few fine carbonate threads; Few faint 10YR 5/6 mottles along root paths; Many wormcasts, common open worm burrows
Btk2b2	168–194	10YR 7/4	10YR 6/4	SiCL	2 c pr	h, fr	1 f pf, po	c	2m 2f 2vf	2f	Parts to 2 f-m sbk; Patchy 10YR 4/3 clay films; Few fine hard carbonate masses, few fine threads; Many wormcasts, common open worm burrows
Cb2	194–230	10YR 7/4	10YR 6/4	SiL	1 m pr	h, vfr	---	a	2vf	1f	Few thin beds
Ckb2	230–278	10YR 7/4	10YR 6/4	SL	1 c pr	sh,,vfr	---	c	2f	1f	Common fine gravel beds, Few fine carbonate threads; Common 10YR 6/3 flood coats on ped faces
Akb3	278–300	10YR 5/2	10YR 4/2	SiCL	1 f pr	sh,fr	---	g	3f 3vf	1f	Parts to 2 m-c gr; Common fine carbonate threads and common distinct patchy coats on ped faces
Btkb3	300–320	10YR 6/3	10YR 5/3	SiL	2 m pr	sh,fr	1 f pf	g	1m 3f 3vf	1f	Parts to 2 f-m sbk; Discontinuous 10YR 4/2 clay films; Common fine carbonate threads and common distinct patchy coats on ped faces; Many wormcasts, common open worm burrows
Ak1b4	320–348	10YR 5/2	10YR 3/2	SiL	1 f-m pr	h, fr	---	g	2m 3f 3vf	1vf	50% 10YR 4/2 dry; Parts to 2 m-c gr; Common fine carbonate threads and many distinct patchy coats on ped faces; Many wormcasts, common open worm burrows
Ak2b4	348–369	10YR 4/2	10YR 3/2	SiL	1 f-m pr	h, fr	---	g	2m 3f 3vf	1vf	Parts to 2 m-c gr; Common fine carbonate threads and many distinct patchy coats on ped faces; Many wormcasts, common open worm burrows
Btk1b4	369–393	10YR 5/4	10YR 4/4	SiL	1 m pr	h, fr	2 d pf	g	1m 2f 2vf	1vf	Parts to 2 m sbk; Common fine and few medium carbonate threads; Continuous 10YR 4/2 clay films; Few wormcasts, few open worm burrows
Btk2b4	393–430+	10YR 6/4	10YR 5/4	SiL	1 m pr	h, fr	1 f pf	---	1m 2f 2vf	1vf	Parts to 1 f pr; Common fine and very fine carbonate threads; Patchy 10YR 5/2 clay films

[1] Texture: C – Clay, CL – Clay Loam, SiCL – Silty Clay Loam, SiC – Silty Clay, L – Loam, SiL – Silty Loam, LS – Loamy Sand.
[2] Structure: 1 – weak, 2 – moderate, 3 – strong, m – massive, sg – single grain; f – fine, m – medium, c – coarse; abk – angular blocks, sbk – subangular blocks, pl – plates.
[3] Consistence: so – soft, sh – slightly hard, h – hard, vh – very hard; lo – loose, vfr – very friable, fr – friable, fi – firm; vfi – very firm.
[4] Clay Films: 1 – few, 2 – common, 3 – many; d – distinct, p – prominent; pf – ped faces, po – pores.
[5] Boundaries: a – abrupt, c – clear, g – gradual.
[6] Roots and pores: 1 – few, 2 – common, 3 – many; vf – very fine, f – fine, m – medium, c – coarse.

Discussion

Chronology and correlation

Ages determined on SOM provide an assessment of mean resident time for organic carbon that, while not providing an absolute age, typically gives a minimum age for the onset of soil development. The number of stratigraphically inverted ^{14}C ages in this study (3 ages out of 9) highlights some of the problems associated with ^{14}C dating SOM (see Mayer et al., 2008). Inputs of old detrital carbon, reworked from older alluvial fills or surrounding uplands, can cause the measured ^{14}C of bulk SOM to be older than the actual age of the soil. Younger carbon, however, also can be introduced into older buried soil horizons via root growth, bioturbation or the leaching of mobile organic compounds.

With the exception of one ^{14}C age (14,710 ^{14}C yr BP), all ^{14}C ages from soil S1 in the T-2 fill are in stratigraphic order (Table 1). The anomalous age most likely reflects the input of younger carbon into the soil. Overall, the chronology indicates that soil S1 formed between ca. 48–28 ka (45,100 and 23,330 ^{14}C yr BP) and is coeval with the Gilman Canyon Formation pedocomplex. Published ^{14}C ages for the Gilman Canyon Formation range from ca. 44–24 ka (40,600–19,770 ^{14}C yr BP) from sites across Colorado, Nebraska and Kansas (e.g., Souders and Kuzila, 1990; Muhs et al., 1999; Johnson et al., 2007).

The stratigraphically inverted ages from soils S5 and S6 in the T-1 fill likely represent the input of detrital carbon either eroded from older fills (e.g., the T-2 fill) or, in the case of soil S5, older carbon reworked from soil S4 due to their close proximity. The two ^{14}C ages from soil S4 are in correct stratigraphic order and are considered accurate based on ^{14}C ages determined on similar alluvial soils dating to 12.9–11.5 ka that represent a major episode of quasi-stability and cumulic soil development throughout Kansas and Nebraska (see Mandel, 2008).

OSL ages from the T-1 and T-2 alluvial fills are in stratigraphic order with depth and are consistent with coherent ^{14}C ages (see comments above about adherent ^{14}C ages). Sample USU-1588 displayed positively skewed equivalent dose (D_E) with high over-dispersion (37%), indicative of partial bleaching. The age for this sample was calculated using a minimum age model. All other samples had normally distributed D_E values and over-dispersion <30%. The age of these sample was calculated using a central age model (weighted mean) of the D_E data.

Regional comparisons

MIS 3: ca. 57–28 ka

Stable oxygen isotope (δ^{18}O) records from Greenland ice cores indicate that MIS 3 experienced frequent and rapid climatic fluctuations known as Dansgaard–Oeschger (DO) events (Dansgaard et al., 1993). DO events are characterized by abrupt warming followed by a gradual return to cold, stadial conditions. During MIS 3, the Laurentide Ice Sheet retreated, resulting in meltwater discharge into the Gulf of Mexico (e.g., Hill et al., 2006; Tripsanas et al., 2007) and the deposition of glaciogenic loess in the Missouri and Mississippi River valleys (e.g., Forman and Pierson, 2002; Bettis et al., 2003). Paleoenvironmental change in the Central Plains during MIS 3, however, remains unclear due to the lack of high-resolution paleoenvironmental studies (see Voelker, 2002). Although the term "Central Plains" is informal, we use the term here to refer to the region extending from northern Nebraska, south through Kansas, into northern Oklahoma, as is often used in the literature. Regional paleoenvironmental studies support our δ^{13}C results, suggesting warm climatic conditions during late MIS 3. Regional paleoenvironmental records also point to increased aridity in the Central Plains at this time. This period of time corresponds to cumulic soil development and the formation of soil S1 in the T-2 fill in the Cimarron River valley. A review of the regional studies, used to support our findings, is provided below (specific localities shown on Fig. 1).

A late-Quaternary pollen record from Cheyenne Bottoms in central Kansas indicates the existence of a persistent shallow-water marshland at ca. 34 ka that began to rapidly dry around 30 ka, with deposition ceasing in the basin by ca. 28 ka (Fredlund, 1995). Jacobs et al. (2007) document a similar record from the Cobb Basin in the Nebraska Sand Hills. Age control indicates that a lake was present in the Cobb basin at ca. 45 ka but was completely dry by 27 ka. A progressive decrease in lake level during the latter stages of the basin's history suggests a decrease in precipitation. Diatom assemblages indicate slightly saline waters at this time, signifying evaporative conditions. Similarly, a summary of paleobotanical records from the Central Lowlands physiographic province suggests cooler temperatures and increased moisture at ca. 38 ka, with more arid conditions prevailing between 37–29 ka as indicated by a decline in aquatic and weedy taxa in wetland areas (Baker et al., 2009).

Warmer climatic conditions during late MIS 3, inferred in our study and other regional paleoenvironmental investigations, are consistent with climate models that suggest substantially warmer and drier conditions in the Great Plains during MIS 3 compared to the last glacial maximum (Van Meerbeeck et al. (2009). These simulations suggest enhanced westerlies during MIS 3, which may have reduced the influx of moist subtropical air masses, as well as enhanced seasonality with mainly warmer summers due to an increase in summer insolation.

Paleoenvironmental change in the mid-continent during MIS 3 also is indicated in mineralogical and sedimentological data from the Gulf of Mexico (GOM). Sionneau et al. (2013)) identified distinct mineralogical provinces that contributed sediment at different times to the GOM. Between 42.6–37.3 ka and 31.4–29.6 ka, clay mineralogy indicates a dominance of illite and chlorite clay minerals derived from the northeastern Mississippi watershed and the Great Lakes region, which suggests that moisture inflow from the GOM into the Central Plains was prevented at this time and that the region experienced generally arid conditions.

One potential mechanism for prevailing aridity in the Central Plains during late MIS 3 is meltwater discharge from the Laurentide Ice Sheet into the GOM. A δ^{18}O record from the GOM indicates five negative isotope excursions associated with increased meltwater discharge from ca. 44–28 ka (Hill et al., 2006) whereas sedimentological evidence from the northwest GOM indicates four meltwater events from ca. 53–32 ka (Tripsanas et al., 2007). Rittenour et al. (2007) also mapped extensive braidbelts in the lower Mississippi valley at this time (ca. 64–30 ka). Glacial meltwater discharge would have cooled sea surface temperatures in the GOM, resulting in reduced influx of subtropical air into the Central Plains during this period.

MIS 2: ca. 28–14 ka

Climatic models indicate a bifurcation of the jet stream in response to the presence of the Laurentide Ice Sheet during MIS 2 (e.g., Kutzbach, 1987; COHMAP, 1988; Bartlein et al., 1998). Paleowinds, originating from the northwest during this period, resulted in extensive loess deposition (Peoria Loess) across the Great Plains (e.g., Bettis et al., 2003; Muhs et al., 2008). Botanical and faunal studies from the Great Plains and adjacent Central Lowlands generally suggest colder climates during MIS 2, with greater effective moisture and less seasonality than today. Our chronology suggests that these climatic conditions were concurrent with the deposition of the bulk of the T-1 fill in the Cimarron River valley.

In central Kansas, pollen data from Cheyenne Bottoms indicate an expansion of pine and spruce at ca. 28 ka, suggesting a shift to colder climatic conditions (Fredlund, 1995). Similarly, across the Central Lowlands, evidence exists for a region-wide shift in vegetation from open parkland to spruce forest after ca. 29 ka (Baker et al., 2009). Wells and Stewart (1987) recorded spruce charcoal at multiple locations in south-central Nebraska and north-central Kansas that date to between ca. 21.5 ka and 17.6 ka. They also documented land-snail and vertebrate fauna that are typical of assemblages found in modern subalpine taiga regions. Rousseau and Kukla (1994) investigated mollusk assemblages at the Eustis Ash Pit in southwestern Nebraska that indicated relatively cool and moist conditions from ca. 24–16 ka. Inferences of colder climatic conditions from 24–14 ka in the Central Plains are also consistent with δ^{13}C data that indicate the dominance of C_3 plant communities during this period (e.g., Muhs et al., 1999; Johnson and Willey, 2000).

While there is general agreement that climatic conditions were colder from ca. 28–14 ka in the Central Plains, there is less certainty about moisture regimes. Although Wells and Stewart (1987) and Rousseau and Kukla (1994) suggest cold and moist conditions in north-central Kansas and southwest Nebraska, respectively, the pollen and sedimentological record from Cheyenne Bottoms suggests that climatic conditions were cold and dry during this period (Fredlund, 1995).

Early MIS 1: ca. 14–11 ka

The Pleistocene–Holocene transition was a period of considerable climatic variability on the Central Plains (e.g., Johnson and Willey, 2000; Mandel, 2008). Moreover, the stratigraphic records from the central and southern Great Plains during this time period show considerable temporal and spatial variability that reflect local paleoenvironmental conditions and probably complex response to the reorganization of mid-latitude climates following the collapse of the Laurentide Ice Sheet (Holliday et al., 2011). Warmer climatic conditions at ca. 12.5 ka inferred from the T-1 fill $\delta^{13}C$ record have, however, been documented by other regional paleoenvironmental studies. Temperature reconstructions for soil S4, which approach modern mean July temperatures, are consistent with a prominent warm excursion (stage IV) documented by Nordt et al. (2007). Also, temperature reconstructions from soil S5, which were 3.2°C cooler than modern temperatures, correlate well with a cool excursion (stage III; Younger Dryas) where Nordt et al. (2007) report temperatures that were ~3.5°C cooler across the Great Plains.

Regional paleoenvironmental studies have also recorded increasing aridity during early MIS 1. Cordova et al. (2011) suggest that a cool, wet environment prevailed at the Kanorado site in northwestern Kansas from ca. 14–12.9 ka and that conditions became warmer and more arid from ca. 12.9–12.2 ka, based on phytolith and $\delta^{13}C$ data. $\delta^{13}C$ analyses by Mandel (2008) at the Willem Ranch site in northwestern Kansas reveals a similar story of paleoenvironmental change. Bement et al. (2007) reported that C_3 plant communities dominated at Bull Creek in the Oklahoma Panhandle before ca. 12.9 ka followed by a shift to more mixed C_3 and C_4 communities at ca. 12.2 ka and C_4-dominated communities by ca. 11.2 ka. From ca. 12.9–12.7 ka grasslands begin to dominate and a drying of the environment is indicated by increases in high spine Asteraceae species. Further drying is indicated through ca. 11.2 ka by increases in Cheno-Am pollen.

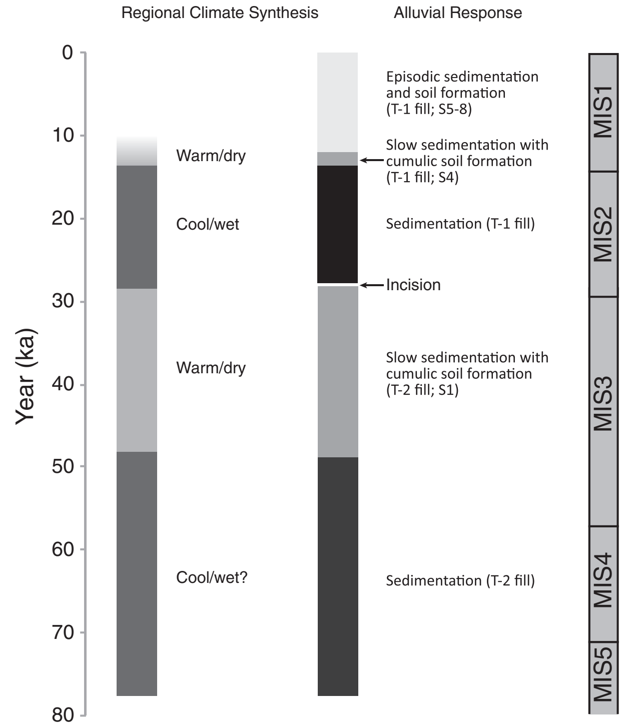

Figure 4. Generalized climate synthesis from regional paleoenvironmental studies and inferred alluvial response in the Cimarron River valley. Marine Isotope Stages (MIS) from Lisiecki and Raymo (2005).

Alluvial response to extrinsic and intrinsic forcing

Fluvial systems incise and aggrade based on the relationship between stream power (a function of discharge and slope) and sediment supply (Bull, 1991), both of which are climatically driven (e.g., Knox, 1983; Bull, 1991; Mandel, 1994; Blum and Törnqvist, 2000). Consideration of both extrinsic and intrinsic forcing mechanisms on stream power and sediment supply, therefore, allows for the establishment of multiple working hypotheses regarding the causes of incision and aggradation in the Cimarron River valley during the late Pleistocene.

Chronostratigraphic evidence (Fig. 2C) from core 2 indicates that the T-2 fill aggraded at least 12 m between ca. 77 and 58 ka followed by soil formation between ca. 48 and 28 ka, likely due to incision and terrace formation. Stratigraphic and sedimentological evidence indicate that the Cimarron River was an aggrading, braided stream between ca. 77 and 58 ka and paleoclimatic conditions were likely cool and wet. Between ca. 48 and 28 ka, $\delta^{13}C$ data and regional paleoenvironmental records indicate a shift to warmer and more arid climatic conditions, resulting in slower rates of sedimentation and the formation of a cumulic soil (S1) in the T-2 fill (Fig. 4).

One hypothesis for the slower rates of alluviation during late MIS 3 is that the channel began to incise in response to the prevailing dry climatic conditions. The relationship between incision and aridity on the Great Plains has been demonstrated in several studies and is often explained in terms of vegetation response and the subsequent alteration of runoff characteristics (e.g., Knox, 1983; Hall, 1990; Daniels and Knox, 2005). Under this scenario, less effective moisture would have resulted in decreased vegetation cover, which in turn decreased interception rates and increased surface runoff. In order for surface runoff to result in channel incision, sediment transport must have exceeded sediment supply to the channel. In semiarid regions, this is typically achieved via high-magnitude floods, produced by high-intensity precipitation from convective storms capable of producing significant in-channel sediment transport (e.g., Tucker et al., 2006). Although convective storms are small in terms of their areal extent, higher order valley segments in a semiarid basin generally experience more frequent floods, thereby favoring incision (Tucker et al., 2006). Overbank deposition, however, from high-magnitude floods typically favors rapid sedimentation instead of soil cumulization on floodplains. Under this scenario, therefore, incision must have been of sufficient magnitude that overbank flooding was minimal and so limited quantities of sediment were deposited on the T-2 surface during high-magnitude flooding. Another possibility, however, is that soil cumulization was facilitated through eolian deposition rather than alluviation on the T-2 surface after it was abandoned by incision. Eolian sedimentation is supported by the fact that this period coincides with the deposition of eolian sediments constituting the Gilman Canyon Formation across the Central Plains.

After ca. 28 ka, ~1.3 m of alluvium was deposited on the T-2 surface, burying soil S1. Our chronology, however, indicates that the bottom of the T-1 fill was also aggrading at this time. Stratigraphic and chronologic relationships indicate that the Cimarron River incised over 25 m before the beginning of T-1 aggradation. This estimate is based on the difference between the terrace treads (~15 m) and assumes that the river incised below the 28 ka OSL sample interval at 10-m depth before subsequently aggrading. If the axial stream deposited alluvium on the T-2 surface at ca. 28 ka, then the bulk of valley incision must have occurred shortly after and progressed rapidly over a span of a few thousand years. Alternatively, a tributary stream flowing across the T-2 surface may have deposited the alluvium. This inference is supported by the presence of tributary streams and paleochannels that incised through the T-2 fill in close proximity to the study site. If correct, then valley incision could have begun as early as ca. 48 ka, resulting in the formation of soil S1 in the T-2 fill as previously discussed.

Regardless of when channel incision was initiated, however, it is debatable whether climatic forcing alone could cause the magnitude of observed valley deepening. Considering alternative forcing mechanisms that could account for over 25 m of incision in the Cimarron River valley is therefore warranted.

One important consideration is the effect of local base-level fall on fluvial systems. Base-level lowering often results in channel incision and terrace generation through the creation and upstream migration of knickpoints (e.g., Schumm, 1977, 1993). Also, incision due to base-level lowering has been shown to generate significant entrenchment over relatively short periods. For example, the South Fork of the Big Nemaha River, in southeastern Nebraska, incised 8–12 m in response to downvalley channelization during the last century (Mandel and Bettis, 2001). In the Cimarron River valley, evidence exists for local base level control in the Meade and Ashland basins located ~50 km downstream of the study area (Fig. 1). These basins were formed by the coalescence of dissolution-subsidence depressions that became integrated during the Pleistocene (Frye and Schoff, 1942; Frye, 1950). In this area, deep-seated dissolution of Permian salt and gypsum deposits is favored by the presence of early Pleistocene faults that allow surface water access to deeply buried soluble beds (Frye and Hibbard, 1941; Frye, 1950). We hypothesize, therefore, that local base-level lowering, produced by the integration of drainages from formerly isolated dissolution-subsidence areas during the late Pleistocene, resulted in significant headward valley deepening in the Cimarron River valley and caused the abandonment of the T-2 surface shortly after ca. 28 ka. Gustavson (1986) tested a similar hypothesis of subsidence-controlled valley deepening in the Canadian River valley, located in the Texas Panhandle, and concluded that dissolution processes have been ongoing in the region since the late Pliocene.

Although a fall in local base level would tend to produce a response in one direction (i.e., incision), stratigraphic evidence indicates that the river re-aggraded at least 10 m to form the T-1 terrace, and possibly as much as 13.5 m, presuming the coarse-grained unit in core 1 represents the base of the T-1 fill. Whereas post-incisional aggradation typically favors a climatic driver, this mechanism assumes sediment bypass during knickpoint migration. As the knickpoint migrated upstream following base-level fall, however, it is feasible that the sediment generated by the wave of incision was subsequently stored in the Cimarron River valley, forming the bulk of the T-1 fill.

The Cimarron River deposited the bulk of the T-1 fill between ca. 28 and 13 ka. During this period regional paleoclimatic proxy data suggest a more mesic climate, with greater effective moisture than present in the Central Plains (e.g., Wells and Stewart, 1987; Rousseau and Kukla, 1994; Fig. 4). Furthermore, this period of aggradation coincides with the timing of extensive loess deposition (Peoria Loess) across the Great Plains.

Around 13 ka, sedimentation rates declined, resulting in the formation of a cumulic soil (S4) in the T-1 fill (Fig. 4). The period from 12.9 to 11.5 ka has been identified as a major period of quasi-stability, characterized by cumulic soil development in stream valleys throughout the Central Plains (Mandel, 2008). The $\delta^{13}C$ record, together with regional paleoenvironmental reconstructions, suggests that climatic conditions during this time were warm and dry in the study area. Effective moisture, however, must have been high enough to promote continued alluviation, albeit at a slower pace that favored cumulic soil formation (soil S4). Mandel (2008) hypothesized that zonal flow during this time may have restricted the northward penetration of moist air masses from the Gulf of Mexico into the Central Plains and that weak Pacific storms may have generated enough precipitation to promote slow alluviation and soil cumulization.

Summary and conclusions

The purpose of this study was to investigate stratigraphic, geomorphic, and sedimentological relationships from late-Pleistocene alluvial fills stored in the Cimarron River valley, southwestern Kansas. Stable carbon isotope ($\delta^{13}C$) analyses were also used to assess paleoenvironmental change and the subsequent alluvial response during this period. Two distinct valley fills (T-1 and T-2) were investigated from outcrop and core. Fluvial deposits consist primarily of vertically stacked, upward fining channel-fill sequences. Three soils were documented in the T-2 fill and five soils in the T-1 fill. These soils indicate periods of slow sedimentation and landscape stability in the Cimarron River valley. Of particular interest was the presence of two cumulic soils: one dating to ca. 48–28 ka (S1), the other dating to ca. 13–12.5 ka (S4). $\delta^{13}C$ values are consistent with regional paleoenvironmental proxy data and suggest that warm, dry conditions prevailed in the study area between ca. 48–28 ka. A return to cool, wet conditions occurred from ca. 28–14 ka, followed by a shift to warm, dry climates at ca. 13 ka.

The data collected in this study illustrate how the Cimarron River responded to changes in climate as well as to local base-level control. Our results indicate that 1) aggradation was facilitated by cool, wet climatic conditions (e.g., before ca. 48 ka and from ca. 28–13 ka); 2) slow sedimentation and cumulic soil formation, together with possible channel incision, occurred under warm, dry climatic conditions (e.g., from ca. 48–28 ka); and 3) significant valley deepening (~25 m) likely resulted from a lowering of local base level caused by deep-seated dissolution of Permian salt and gypsum deposits downstream of the study area. Overall, stratigraphic, paleopedologic, and paleoenvironmental data from the Cimarron River valley offer important insights into the late-Pleistocene geomorphic evolution of this fluvial system.

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