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Use of carbon isotope analysis to understand semi-arid erosion dynamics and long-term semi-arid land degradation†

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Many semi-arid areas worldwide are becoming degraded, in the form of C_4 grasslands being replaced by C_3 shrublands, which causes an increase in surface runoff and erosion, and altered nutrient cycling, which may affect global biogeochemical cycling. The prevention or control of vegetation transitions is hindered by a lack of understanding of their temporal and spatial dynamics, particularly in terms of interactions between biotic and abiotic processes. This research investigates (1) the effects of soil erosion on the $\delta^{13}C$ values of soil organic matter (SOM) throughout the soil profile and its implications for reconstructing vegetation change using carbon-isotope analysis and (2) the spatial properties of erosion over a grass-shrub transition to increase understanding of biotic-abiotic interactions by using $\delta^{13}C$ signals of eroded material as a sediment tracer. Results demonstrate that the soils over grass-shrub transitions are not in steady state. A complex interplay of factors determines the input of SOM to the surface horizon of the soil and its subsequent retention and turnover through the soil profile. A positive correlation between event runoff and $\delta^{13}C$ signatures of eroded sediment was found in all plots. This indicates that the $\delta^{13}C$ signatures of eroded sediment may provide a means of distinguishing between changes in erosion dynamics over runoff events of different magnitudes and over different vegetation types. The development of this technique using $\delta^{13}C$ signatures of eroded sediment provides a new means of furthering existing understanding of erosion dynamics over vegetation transitions. This is critical in terms of understanding biotic-abiotic feedbacks and the evolution of areas subject to vegetation change in semi-arid environments.

Semi-arid areas occupy approximately 17% of the global land area.[1] Many semi-arid areas have undergone desertification and land degradation since the intensification of agriculture. During the last 150 years in the south-western United States, for example, large areas of semi-arid grasslands have been replaced by arid shrublands.[2]. Replacement of grassland by shrubland induces a change in surface processes, notably increased runoff and erosion,[3] and altered nutrient cycling.[4–6] The replacement of grassland by shrubland is of major concern since it is thought that changes in nutrient cycling may affect biogeochemical cycles, thus affecting ecosystems and climate worldwide[7] through atmosphere and land-surface interactions. The enhanced soil erosion that occurs over shrublands is a major contributor to land degradation and the loss of essential nutrients such as carbon and nitrogen.[8] Understanding the past dynamics of vegetation change is thus important in terms of understanding ongoing vegetation change and associated degradation of the land.

Use of carbon-isotope analysis of sediment eroded at different stages over a grassland-to-shrubland transition provides the opportunity to understand the origin of eroded sediment, since the carbon isotope analysis of eroded sediment will discriminate between SOM originating from C_4 or C_3 vegetation which has changed over the last 150 years.[2] Carbon-isotope analysis of eroded sediment to trace the source of the sediment is not an entirely new concept. Bellanger *et al*.[21] employed carbon- and nitrogen-isotope analysis of eroded soil organic carbon to identify and monitor the sources of soil organic carbon in runoff waters in a tropical mountainous environment. Their research was based upon the assumption that constant organic carbon fluxes are generated during soil erosion and that the isotopic composition of displaced organic carbon does not vary with time.[21] Another example is that of Bird *et al*.[22] who used the $\delta^{13}C$ signature of particulate organic carbon in fluvial

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sediments to determine the photosynthetic pathway of vegetation at the origin of the eroded sediment, to infer changes in the extent of tropical forests on a basin-wide scale. These studies have demonstrated that, in their relative contexts, isotopic discrimination by C_3 and C_4 plants is recorded in eroded sediment, and this provides a sound justification for using carbon isotope analysis of eroded sediment to explore soil erosion over a grass-shrub transition.

The technique of reconstructing semi-arid vegetation change based on carbon-isotope analysis of soil organic matter (SOM) has been demonstrated previously, e.g.[9–12] The use of carbon-isotope analysis to reconstruct vegetation change is based upon the principle that C_4 grasses are enriched with ^{13}C relative to C_3 shrubs; therefore, C_4 grasses and C_3 shrubs have characteristic $\delta^{13}C$ values. However, it is well established that in most soils organic matter becomes ^{13}C-enriched with depth[12] by approximately 1 to 3‰[13] due to processes such as the systematic kinetic fractionation of the carbon isotopes through time,[14] although this can only account in-part for the increase in $\delta^{13}C$ of SOM with depth. Consequently, the carbon isotopic composition of SOM largely reflects the carbon isotopic composition of plant material from which it originated, making it an appropriate indicator of the photosynthetic pathway of past vegetation.[9] The isotopic discrepancy created by the vegetation change will persist for a length of time determined by the SOM turnover rate.[15] Implicit in the reconstruction of vegetation change and changing carbon stocks is the assumption that the soil is not undergoing erosion or deposition of sediment and that the changes are preserved in SOM throughout the soil profile (e.g.[12]). However, this assumption is compromised in semi-arid environments undergoing vegetation transitions since, typically, high levels of erosion result in the redistribution of sediment and organic material around the landscape.[3,16] There has been some previous recognition of the effects of soil erosion when reconstructing vegetation change; for example, Biedenbender *et al.*[17] suggested that evidence of a truncated soil profile from $\delta^{13}C$ analysis and ^{14}C dating of SOM may be due to the loss of SOM by erosion. However, no studies have yet adequately addressed the effects of soil erosion on reconstructing vegetation change using $\delta^{13}C$ analysis of SOM.

Research into erosion over grassland and shrubland has shown that sediment detached in grassland is redistributed locally, whereas in shrubland it is transported over much greater distances[3,16,18] because of a change in the cover and distribution of vegetation and increased flow connectivity in intershrub areas.[3] Hence, it is clear from previous research that soil erosion varies significantly over grassland and shrubland in terms of the amount of erosion that takes place, and the spatial variations of erosion processes. What remains poorly understood, however, is how vegetation and soil erosion change concurrently over a transition from grassland to shrubland and to what extent soil conditions over the grassland to shrubland transition diverge from the previously assumed 'non-eroding landscape'. Therefore, the objectives of this paper are twofold:

(1) to investigate the effects of soil erosion on the $\delta^{13}C$ values of SOM throughout the soil profile and its implications for reconstructing vegetation change using carbon-isotope analysis; and
(2) to investigate the spatial properties of erosion over a grass-shrub transition to increase understanding of biotic-abiotic interactions by using $\delta^{13}C$ signals of eroded material as a sediment tracer.

EXPERIMENTAL

Four study sites were set up over a grassland-to-shrubland ecotone at the Sevilleta National Wildlife Refuge in New Mexico, USA (34°19′N, 106°42′W), that are assumed to be representative of different stages of the grassland-to-shrubland transition. Thus, the experimental design of the field study employs the ergodic hypothesis where spatial dynamics are representative of changes through time (Fig. 1). Anecdotal accounts and historic photographs indicate that the ecotone is dynamic, since over the past 150 years desert shrubland has been encroaching northwards into the desert grasslands.[2] The 'native' grasslands are composed primarily of C_4 species (dominated by black grama, *Bouteloua eriopoda*) and the 'invading' shrublands mainly comprise C_3

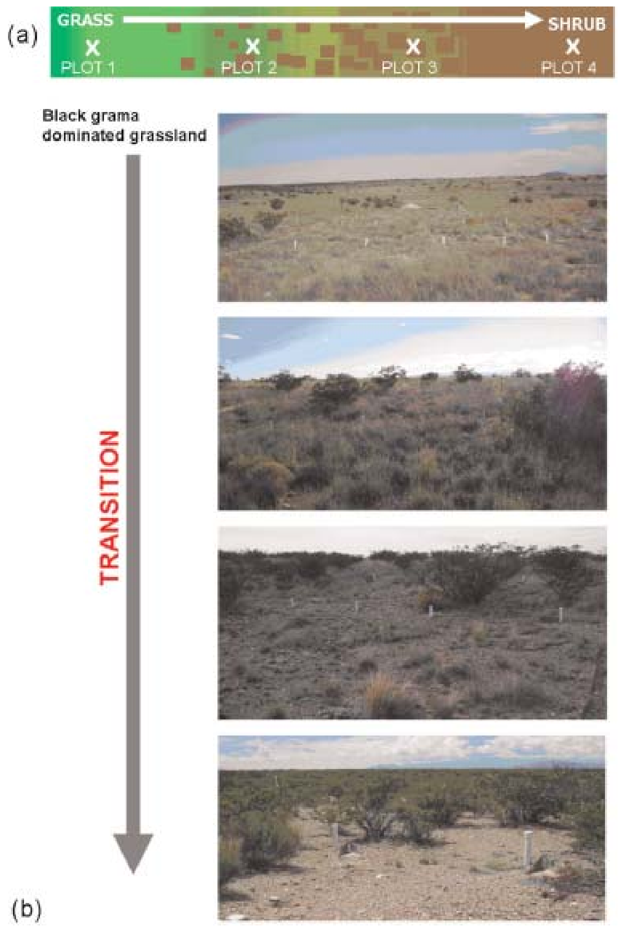

Figure 1. (a) Schematic of the experimental design, whereby four study plots have been set up over the grass-shrub ecotone, from the grass end-member, across two transitions plots, to the shrub end-member. (b) Photographs of the four plots located over the grass-shrub transition. This figure is available in colour online at www.interscience.wiley.com/journal/rcm

Table 1. Characteristics of rainfall-runoff events monitored over the grassland to shrubland transition

	Event date	Event rain (mm)	Event runoff (L)	Eroded sediment (kg)
Plot 1	07/09/2005	47.9	5275	4114
	04/08/2006	5.1	27	104
	15/08/2006	7.1	142	422
	29/08/2006	23.1	2820	2048
	07/09/2006	13.7	1426	863
Plot 2	07/09/2005	26.2	2294	6252
	11/08/2006	11.4	150	530
	15/08/2006	9.4	391	537
	29/08/2006	20.6	995	1261
	07/09/2006	15.0	1260	1120
Plot 3	11/08/2006	9.4	637	906
	29/08/2006	7.4	568	715
	07/09/2006	24.4	6550	6178
Plot 4	07/09/2005	14.1	1877	10823
	05/07/2006	17.0	1913	5836
	11/08/2006	7.1	450	810
	23/08/2006	9.9	1387	2595

species (dominated by creosotebush, *Larrea tridentata*). The study area is underlain by Quaternary, sedimentary geology. The soil in the study area is a sandy loam. The annual precipitation at the Sevilleta National Wildlife Refuge is 242 mm, of which an average of 140 mm falls during the summer monsoon period. Inter-annual variability in precipitation over the south-west USA is linked to shifts in the upper-air westerlies and the El Niño Southern Oscillation (ENSO).[19,20] The monsoonal precipitation tends to occur as convective rainfall events that are of high intensity.

At each of the four sites, three soil cores were taken down to the depth of the caliche layer (between 25 and 40 cm) for the analysis of SOM through the soil profile. Since the δ^{13}C variability of SOM is high in mixed C_4/C_3 ecosystems,[23] over all four plots, soil cores were taken in the centre of bare areas between vegetation in an attempt to reduce variability due to differences in sampling location. Soil was sampled from the top 2.5 cm of the soil profile, and then at 5 cm depth increments thereafter. At each site, a bounded 10×30 m runoff plot was constructed, and all sediment eroded from the plots during single runoff-generating rainfall events was collected at the plot outlet for analysis (Table 1).

Bulk soil samples from the soil cores and eroded sediment were sieved through a 2 mm screen and the sub-2-mm fraction was ground to a fine powder. As the soils are rich in carbonate, this inorganic carbon was removed from the soil using 75 mL of 2M HCl added to approximately 5 g of ground soil, left for 1 week, and then filtered through a glass-fibre filter paper and washed three times with 100 mL deionised water to remove the HCl. Samples were then air-dried prior to analysis. Pre-treated samples from the three soil cores from each of the four sites were analysed for δ^{13}C. The δ^{13}C values of plant and soil samples were analysed at IGER North Wyke using a NA 1500 elemental analyzer (Carlo Erba, Milan, Italy) and an automated continuous flow ANCA 20/20SL system (Europa, Crewe, UK). The natural abundance values were expressed as δ values, which represents the ratios of ^{13}C/^{12}C relative to the international VPDB and AIR standard, respectively. The δ^{13}C values (per mil) are defined as:

$$\delta^{13}C = [(\text{atom\%}\,^{13}C_{\text{sample}} - \text{atom\%}\,^{13}C_{\text{VPDB}})/\text{atom\%}\,^{13}C_{\text{VPDB}}] \times 1000$$

The analytical precision of the $\delta\ ^{13}$C measurements was <0.1. The δ^{13}C values for past C_4 and C_3 vegetation are generally based on contemporary measurements of δ^{13}C in foliage of the same vegetation types.[10,17] In the absence of detailed plant isotope measurements in this study, we used the means of the C_4 grasses (δ^{13}C value of −14.0) and C_3 shrubs (δ^{13}C value of −26.9) as determined by Boutton and co-workers[10] in a comparable ecosystem for the C_4 and the C_3 vegetation end-members in the equation below. The proportion of organic carbon derived from C_4 sources was estimated by:

$$\%C_4 = ((\delta_s - \delta_3)/(\delta_4 - \delta_3)) \times 100$$

where $\delta_s = \delta^{13}$C of the sample (‰); $\delta_3 = \delta^{13}$C value of C_3 plants (‰); and $\delta_4 = \delta^{13}$C value of C_4 plants (‰).

The above equation is applied with confidence to surface SOM since δ^{13}C values are not expected to change significantly during decomposition.[10]

Tamhane's T2 posthoc analysis of variance test was used to determine the variance between the mean δ^{13}C values of SOM between eroded sediment from the four plots and between the eroded sediment and in-situ surface soil at each plot. Tamhane's T2 test was used since it makes no assumption about the equality of variance.

RESULTS AND DISCUSSION

The effect of soil erosion on the δ^{13}C values of SOM throughout the soil profile

Results describing variation in δ^{13}C values and the percentage of soil organic carbon derived from C_4 plants with depth from each plot are presented in Table 2. Across all the plots, an increase in δ^{13}C occurs with depth (increase in C_4-derived organic matter), although this trend is most clear in plots 1, 2 and 4 (Table 2). Such a result might be

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Table 2. Descriptive statistics of δ^{13}C values of eroded sediment and at increasing depth throughout the soil profile at each plot

	Plot 1		Plot 2		Plot 3		Plot 4	
Depth (cm)	Mean δ^{13}C ± S.E.	% C_4	Mean δ^{13}C ± S.E.	% C_4	Mean δ^{13}C ± S.E.	% C_4	Mean δ^{13}C ± S.E.	% C_4
Eroded sediment	−21.0 ± 0.3♣[a]	46 ± 2	−21.1 ± 0.5♣[b]	45 ± 4	−21.4 ± 0.2♣[b]	43 ± 2	−25.1 ± 0.5◆[b]	14 ± 4
(Surface sediment) 0–2.5	−19.5 ± 1.2[a]	57 ± 9	−18.1 ± 0.2[a]	68 ± 2	−16.9 ± 0.2[a]	78 ± 2	−21.9 ± 0.5[a]	39 ± 4
2.5–7.5	−17.4 ± 0.9	74 ± 7	−17.6 ± 0.3	72 ± 2	−16.2 ± 0.4	83 ± 3	−19.1 ± 0.4	60 ± 4
7.5–12.5	−16.6 ± 0.6	80 ± 5	−17.2 ± 0.3	75 ± 2	−15.6 ± 0.4	88 ± 3	−18.3 ± 0.8	67 ± 6
12.5–17.5	−16.4 ± 0.4	81 ± 3	−17.1 ± 0.8	76 ± 6	−15.6 ± 0.5	88 ± 4	−18.6 ± 0.9	64 ± 7
17.5–22.5	−15.8 ± 0.7	86 ± 5	−16.6 ± 0.7	80 ± 5	−16.1 ± 0.4	84 ± 3	−17.9 ± 0.8	70 ± 6
22.5–27.5	−15.1 ± 0.2	91 ± 2	−16.9 ± 0.9	78 ± 7	−15.1 ± 0.3	91 ± 2	−16.4	81
27.5–32.5			−15.3 ± 0.4	90 ± 3	−16.0 ± 0.2	84 ± 2		
32.5–37.5			−15.3 ± 0.2	90 ± 2	−16.0 ± 0.1	84 ± 1		

♣◆Between-plot comparisons of mean δ^{13}C values of eroded sediment: mean δ^{13}C values of eroded sediment followed by the same symbol are not significantly different ($p < 0.05$).
[a] [b]Within-plot comparison of mean δ^{13}C value of eroded sediment and mean δ^{13}C value of surface sediment: δ^{13}C values of soil and eroded sediment followed by the same letter are not significantly different ($p = < 0.05$).

expected as increases in the δ^{13}C with depth have been observed in previous studies and have been primarily attributed to factors such as systematic kinetic fractionation[14] and the contribution of microbial biomass to the soil matrix.[12] The trends of increasing δ^{13}C of SOM at plot 1 are comparable with the trends observed over grasslands in other studies, e.g.[10,12] Results from plot 2, which is predominantly grass-covered with only a few creosotebush, suggest that throughout the soil profile, the majority of organic matter was C_4-derived. Plot 3, which has a greater shrub cover and only a sparse, fragmented grass-cover, had the highest δ^{13}C values of soil organic matter observed for the top 15 cm of the soil profile, and this changed only very slightly with depth. Results from the top 2.5 cm (i.e. the most recent organic matter inputs) of the soil profile at plot 3 suggest that 78 ± 2% of the soil organic carbon was derived from C_4 species, indicating that primary productivity is from C_4 species (i.e. black grama grass), which is in great contrast with the visual observations of primary productivity as the plot is partially covered by C_3 shrubs (Fig. 1). It is plausible that because shrubs do not occupy a great area within plot 3, the influence of C_3-derived organic matter has little effect on the overall SOM content of the soil because of the remnants of grass throughout the profile. The surface soils from plot 3 had higher δ^{13}C values than those from plot 1, implying that a greater percentage of SOM input to the soil at plot 3 is derived from C_4 plants than plot 1, which is counter-intuitive. The δ^{13}C signature of SOM in the surface horizon at plot 3 was similar to the δ^{13}C signature of SOM found at a depth of 10 cm at plot 1, which may indicate progressive erosion as shrubs have invaded the grassland causing erosion of the A-horizon, leaving older horizons (with a high δ^{13}C signature of SOM) exposed at the surface. Over plot 4, the shrub end-member, the lowest δ^{13}C values of SOM were measured, which is probably due to the increase in shrub cover producing more C_3-derived organic matter. Furthermore, the increase in shrub cover may impede the connectivity of flow, thereby decreasing the potential for erosion relative to plot 3 and consequently facilitating the incorporation of SOM into the soil profile.

The isotopic signatures and the estimated percentage of organic C derived from C_4 plants for the surface profiles over plots 1, 2, 3 and 4 do not match up with the visual observations and the distribution of C_3- and C_4-derived SOM throughout the soil profile from previous studies. Connin *et al.*[24] and Parsons *et al.*[25] observed that replacement of the grassland vegetation by shrubs can lead to a loss of surface soil up to a depth of 20–40 cm, encompassing a loss of both the A and the upper B soil horizons. Such rates of soil erosion are comparable with rates from studies at the Sevilleta National Wildlife Refuge[27] and similar arid environments[26] which indicates therefore that erosion may be responsible for the lack of comparability between profiles of δ^{13}C values from plots 1 to 4.

The spatial properties of erosion over a grass-shrub transition using δ^{13}C signals of eroded material as a sediment tracer

The mean δ^{13}C values of surface sediment and eroded sediment from each plot are presented in Table 2. The δ^{13}C (−25.1 ± 0.5) values of eroded sediment from plot 4 are significantly different from those of eroded sediment from all other plots ($p \leq 0.05$), with eroded sediments from plot 4 showing lower δ^{13}C characteristic of C_3-derived sediment (14 ± 4%). Within each plot, a means comparison of the δ^{13}C of surface sediment and eroded sediment shows that there are significant differences for all but plot 1 ($p \leq 0.05$). It seems likely that the plot 1 sediments are sourced from the younger surface soils, retained under the pristine grasslands. The mixed story that is shown for plots 2 and 3 is less easy to understand, although it might be assumed that such a signal represents the erosion of sediments from soils rich in both C_3 and C_4 vegetation and from soils rich in C_3 vegetation over plot 4. It is plausible that although plot 3 has a greater shrub cover than plot 2, the erosion of surface soil exposed older soil horizons that are enriched in δ^{13}C (due to δ^{13}C enrichment of SOM with depth). Thus, the input of C_3 SOM at plot 3 is outweighed by the enriched δ^{13}C signatures of the soil surface. Therefore, from the existing information it appears that more than one explanation of where sediments are sourced from and why they differ in isotopic signature from both the surrounding vegetation and the soils is viable.

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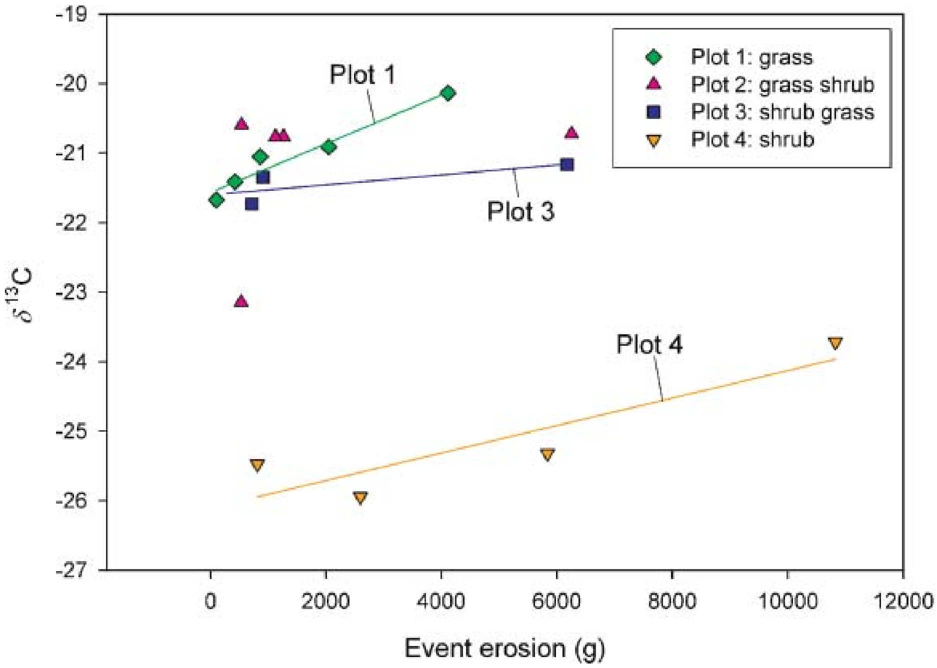

Figure 2. The relationship between event erosion and the $\delta^{13}C$ signatures of eroded material for each plot. This figure is available in colour online at www.interscience.wiley.com/journal/rcm

Figure 3. Concentration of runoff in inter-shrub areas leading to interill erosion. This figure is available in colour online at www.interscience.wiley.com/journal/rcm

Figure 2 shows the relationship between the $\delta^{13}C$ signature of eroded material and event erosion for all plots. There is a positive linear relationship on plots 1, 3 and 4 between the $\delta^{13}C$ signature of eroded material and the total amount of event erosion (plot 1: $R^2 = 0.95$; plot 3: $R^2 = 0.59$; plot 4: $R^2 = 0.81$). This positive relationship implies that with an increase in runoff and erosion, the eroded sediment has a higher $\delta^{13}C$ signature which is probably due to the increased incision into older soil horizons, in accordance with observations of increased runoff concentration in intershrub areas and increased rilling in shrublands, e.g.[16] (Fig. 3). Where grasses predominate, the increase in the $\delta^{13}C$ signature of eroded sediment could also be due to the increased detachment and entrainment of soil rich in C_4 SOM from grass patches. However, from these preliminary data, it is not possible to constrain exactly the processes leading to the observed variations in $\delta^{13}C$ signatures of eroded sediment over the grassland-to-shrubland transition.

CONCLUSIONS

Evidence in this paper has revealed the complexities of using $\delta^{13}C$ signatures of SOM through the soil profile for studying vegetation change and changes in carbon storage in an environment where changes in erosion are a key abiotic component of ecosystem change over a grassland-to-shrubland transition. It is likely that biotic-abiotic feedbacks over a grassland-to-shrubland transition do not follow a linear trajectory.[27] Therefore, at each stage of a grassland-to-shrubland transition, a complex interplay of factors is likely to determine the input of SOM to the surface horizon of the soil and its subsequent retention and turnover through the soil profile. As high rates of erosion have been observed in degrading semi-arid landscapes over recent years (and similar erosion is likely to have occurred over the last 150 years given the increases in erosion upon a transition to shrubland), it is probable that current soil surfaces under different vegetation types are of significantly different ages and that, therefore, profiles through depth also represent profiles through soils of different ages. Thus, reconstructions of semi-arid vegetation change that neglect the role of erosion and sediment deposition are fundamentally flawed.

An initial exploration using carbon isotope analysis to explore the spatial properties of soil erosion over a grass-shrub transition has revealed that there are distinct differences in the isotopic signatures of eroded sediment between plots 1 and 4 (representing the temporal transition of grass to shrub dominated landscapes). The positive correlation between event runoff and $\delta^{13}C$ signatures of

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eroded sediment that is consistent over plots 1, 3 and 4 indicates that the $\delta^{13}C$ signatures can be used to distinguish between changes in erosion dynamics over events of different magnitudes and over different vegetation types. The next step in refining the $\delta^{13}C$ signatures of eroded sediment with reference to their source is to undertake $\delta^{13}C$ analysis of particle-size fractions of soil and eroded sediment to derive more unique signatures.

This study has presented the problems associated with erosion in reconstructing vegetation change in semi-arid environments using $\delta^{13}C$ values of SOM. A key conclusion is that further work is required, perhaps using ^{14}C dating of SOM to determine the time course of vegetation change to ensure that comparisons of $\delta^{13}C$ signatures across the ecotone are well constrained through time. ^{14}C dating of SOM through the soil profile alongside $\delta^{13}C$ analysis of SOM will facilitate the past trajectory of vegetation change to be determined even when soils are eroded, although gaps in the historical record may persist due to prolonged periods of erosion. This study has also demonstrated that there are variations in $\delta^{13}C$ values of SOM in bulk eroded sediment over different vegetation types and under runoff events of different magnitudes, which provides a sound rationale for methodological refinements to be made in order to improve applications exploring the spatial dynamics of erosion using $\delta^{13}C$ signatures of surface soil and eroded sediment.

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