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The stable carbon isotope composition of soil organic carbon and pedogenic carbonates along a bioclimatic gradient in the Palouse region, Washington State, USA

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Abstract

Isotopic signatures of soil components are commonly used to infer past ecologic and climatic shifts in the soil record. The theory behind the fractionation of isotopes that occurs during ecosystem processes is well understood; however, few isotopic studies have explored ecosystem relationships in modern soils. We discuss relationships of stable carbon isotopic signatures in plant tissue, soil organic carbon (SOC), laboratory-respired CO_2, and modern carbonates at 10 sites (seven containing pedogenic carbonates) along a C_3-dominated climatic gradient (mean annual precipitation (MAP) ranging from 200 to 1000 mm) in the Palouse region of eastern Washington state.

A horizon soil organic carbon (SOC) $\delta^{13}C$ values varied from $-24.3‰$ to $-25.9‰$ PDB. Values in the arid portion of the gradient (200 to approximately 500 mm MAP) generally decreased and linear regression of SOM ^{13}C vs. MAP was significant ($r^2 = 0.71$, $p = 0.02$). Trends in plant-^{13}C of two grass species (*Agropyron spicatum* and *Festuca idahoensis*) found throughout this portion of the gradient were similar to that of SOC. Mean pedogenic carbonate $\delta^{13}C$ values varied from $-4.1‰$ to $-10.8‰$ PDB. Linear regression was significant for carbonate ^{13}C vs. MAP ($r^2 = 0.79$, $p = 0.007$), estimated above-ground productivity ($r^2 = 0.88$, $p = 0.002$) and soil carbon content ($r^2 = 0.83$, $p = 0.004$). Carbonate $\delta^{13}C$ values at the most arid site exhibited higher variability than other sites (presumably due to greater spatial variation in plant respiration vs. atmospheric diffusion). Our data suggest that carbon isotopic relationships among ecosystem components may prove useful in determining ecosystem level properties in modern systems, and potentially in ancient systems as well.
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1. Introduction

The fractionation of carbon isotopes that occurs during ecosystem processes makes stable isotopes an ideal tool in understanding terrestrial ecosystem pathways. Much of the carbon isotope work has focused

on C_3 vs. C_4 plant dynamics because of the substantial difference in ^{13}C discrimination in these photosynthetic pathways. However, other ecological factors that affect carbon isotope values in the soil system may be more easily observed in the absence of C_3-C_4 vegetation shifts.

The fractionation of carbon isotopes due to C_3 and C_4 photosynthetic mechanisms has been well documented (O'Leary, 1988; Farquhar et al., 1989). C_3 plants have a mean $\delta^{13}C$ value of $-27‰$ PDB whereas C_4 plants have a mean value of $-12‰$ PDB. The relative difference between C_3 and C_4 organic carbon isotope signatures is maintained as plant tissue is incorporated into soil organic matter, respired as carbon dioxide, and in arid and semi-arid regions, converted into pedogenic carbonates (via the dissolution of respired CO_2). These characteristic isotope values have been used to track C_3 vs. C_4 plant dynamics through measurements of SOC (Dzurec et al., 1985; McPherson et al., 1993; Mariotti and Peterschmitt, 1994) and pedogenic carbonates (Quade et al., 1989; Amundson et al., 1989a,b; Kelly et al., 1991; Wang et al., 1993).

Variation in plant carbon isotope values can generally be attributed to two primary causes: variation in the isotopic signature of assimilated CO_2 and photosynthetic ^{13}C discrimination due to changes in the ratio of internal versus external CO_2 partial pressures (Farquhar et al., 1989). Factors that have been shown to affect photosynthetic ^{13}C discrimination include: drought stress (Farquhar and Richards, 1984; Ehleringer and Cooper, 1988; Read et al., 1991, 1992; Ehleringer et al., 1993), plant mineral nutrient (Sparks and Ehleringer, 1997), and irradiance levels (Zimmerman and Ehleringer, 1990; Berry et al., 1997; Buchmann et al., 1997). Differences in assimilated CO_2 arise from altitude (Korner et al., 1991), CO_2 gradients caused by dense overstory foliage limiting atmospheric diffusion of soil-respired CO_2 (Von Fischer and Tieszen, 1995; Berry et al., 1997), and over longer timescales, changes in atmospheric CO_2 $\delta^{13}C$.

Cerling (1984) provided much of the theoretical groundwork for the isotopic relationships found in pedogenic carbonates in arid and semiarid ecosystems. The source of carbon for pedogenic carbonates is primarily respired CO_2 from below-ground plant and microbial respiration. However, mixing of atmospheric CO_2 into the soil may alter the isotopic signature of respired CO_2. Cerling et al. (1991) make the distinction between "soil CO_2" and "soil-respired CO_2", the former being the gas occupying the soil pore space and the latter the flux of gas through the soil space. This distinction is important in that the isotopic composition of these entities is not the same, differing by a fractionation factor associated with the diffusion of CO_2 out of the soil.

The amount of atmospheric mixing and the isotopic signature of resulting soil CO_2 are dependent upon the rate and isotopic signature of respired CO_2, the depth of carbonate formation, and the isotopic signature and pCO_2 of atmospheric CO_2. Other factors, such as temperature of CO_2 dissolution and carbonate precipitation, may affect carbonate-C isotopic values to a lesser extent.

The carbon isotopic composition of pedogenic carbonates has been shown to reflect changes in atmospheric pCO_2 levels over geologic time scales (Cerling, 1992). Soil pCO_2 can vary by $1-2$ orders of magnitude depending upon soil respiration rates and other soil conditions (Amundson and Davidson, 1990). Although it has been suggested that pedogenic carbonate-^{13}C values (or more accurately $\Delta^{13}C$ (carbonate $-$ SOC) values) may also reflect changes in soil respiration rates (and related ecosystem parameters such as soil organic carbon content and productivity), few studies have demonstrated this relationship. The overall objective of this research is (1) to assess the isotopic relationships among ecosystem components (plant tissue, SOC, respired CO_2 and pedogenic carbonates) across a C_3-dominated bioclimatic sequence (MAP ranging from 200 to 1000 mm) in the Palouse region of eastern Washington state, and (2) to assess the potential of using the ä^{13}C values of soil carbonates as proxies for estimating long-term productivity in modern ecosystems (and potentially paleoecosystems) of the northwestern United States.

2. Methods

2.1. Site description and field methods

The study site is located in the Palouse loess region of southeastern Washington State (Fig. 1). The Palouse region is an area of long-term loess
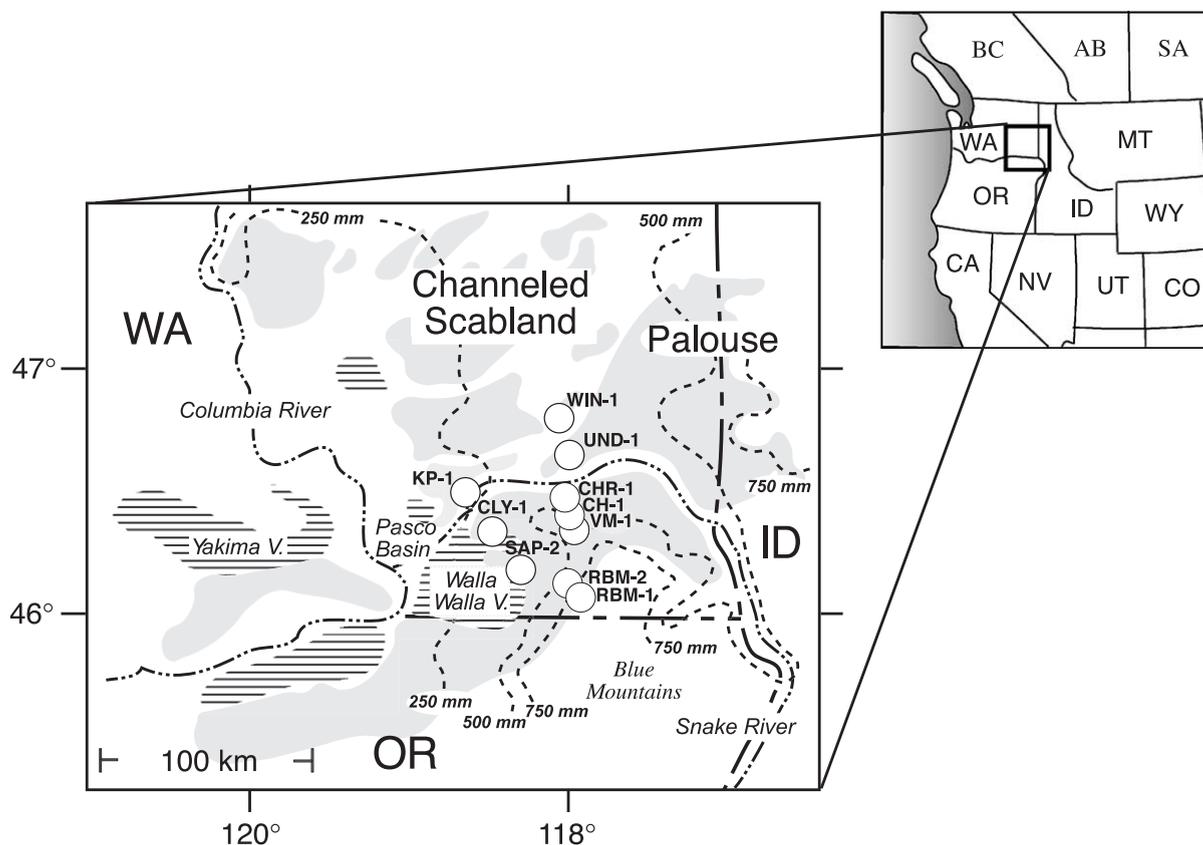
Fig. 1. Palouse study site location map. Shading indicates areas of loess deposition, horizontal lines indicates areas of slackwater deposits.

deposition where numerous soil profiles have been formed and subsequently buried under the episodically aggrading landscape (Busacca, 1989; McDonald and Busacca, 1992). The regional climatic gradient extends from the rainshadow of the Cascade Range in the west to the Blue Mountains in the east. Mean annual precipitation values range from 200 mm to upwards of 1000 mm.

Dominant vegetation along the gradient includes: shrub (*Artemisia tridenta*), perennial bunchgrass (*Festuca idahoensis*, *Agropyron spicatum*), and coniferous communities (*Pinus ponderosa*, *Pseudotsuga menzeisii*, *Abies grandis*, *Abies lasiocarpa*, and *Picea engelmannii*). Above-ground productivity values for each site were estimated from Daubenmire (1969) based on location, community composition, and percent vegetative cover (Table 1). Soils at each site were all derived from loess parent material and sites were chosen to minimize topographic and other variations (a true bioclimatic sequence in the sense of Jenny (1941)).

Bulked soils were sampled by genetic soil horizon (Soil Survey Staff, 1982) to a minimum depth of 1.0 m. Soil samples were mixed well, air dried, run through a 2-mm sieve, and bagged to await further analysis. Samples for carbonate ^{13}C analysis (a minimum of three from each site) were collected from the uppermost zone of carbonate accumulation. Plant samples were composited from the above-ground portion of several plants of selected species. To avoid isotopic variation associated with microhabitat differences, plants were collected from within a 25-m radius of the soil site.

2.2. Laboratory methods

All plant samples were ground in a Wiley mill to pass through a number 60-mesh screen. Soil organic

Table 1
Palouse study site characteristics

Site	MAP[a] (mm)	Elevation (m)	MAT[b] (°C)	Dominant[c] vegetation	Productivity[d] (g^{-2} $year^{-1}$)	Depth to carbonate zone (m)
KP-1	220	410	12	1, 2	60	0.54
CLY-1	305	385	12	1, 2	75	1.13
UND-1	330	450	11	2	100	1.10
WIN-1	430	520	11	2, 3	140	1.23
CHR-1	485	758	11	2, 3	190	1.71
SAP-2	495	400	11	2, 3	200	2.41
CH-1	580	945	10	3, 2, 4	380	1.58
VM-1	790	1090	8	3, 2, 4	480	–[e]
RBM-2	900	990	7	4	900	–
RBM-1	1000	1220	6	5	1150	–

[a] Mean annual precipitation.

[b] Mean annual temperature.

[c] Dominant plant species: (1) *Artemisia tridentate*, (2) *Agropyron spicatum*, (3) *Festuca idahoensis*, (4) *Pinus ponderosa*, (5) *Pseudotsuga menzeisii*.

[d] Above-ground productivity estimates from Daubenmire (1969).

[e] No carbonates.

matter samples (from the A horizon of each site) were acidified with 0.1 M HCl to remove carbonates and ball milled for 3 min prior to analysis. Organic carbon (plant organic carbon and soil organic matter) ^{13}C samples (three subsamples per plant or soil sample) were combusted and purified in a Carlo-Erba prep system (Fisons NA 1500 NC) and analyzed using a VG Isochrom isotope ratio mass spectrometer in continuous flow mode. Precision of standards was 0.03‰ and average standard deviation of several replicated samples ($n=8$) was 0.18‰ PDB. Soil organic carbon was measured using a Leco CHN analyzer.

Carbonate samples were baked under vacuum at 400 °C, reacted with 100% phosphoric acid and cryogenically purified (adapted from Bowen, 1966). The resulting CO_2 was analyzed on a VG Isochrom isotope ratio mass spectrometer in dual inlet mode. Standard deviation for replicated standards prepared on-line ($n=10$) was 0.19‰. All isotopic measurements are presented in the delta (δ) notation:

$$\delta_a = [R_a/R_{std}]*1000$$

where R_a is the $^{13}C/^{12}C$ ratio in the sample and R_{std} is the $^{13}C/^{12}C$ ratio in the standard. All values are expressed as parts per mil (‰) and are referenced to the PDB standard.

Carbon isotope composition of respired CO_2 was assessed in the laboratory by incubating triplicate soil

samples from the A horizon of each site in sealed jars fitted with Leurlock valves. Approximately 50 g soil was brought to field moisture capacity, placed in mason jars, purged with CO_2-free air and incubated for 1 week at 25 °C. Air samples (duplicate subsamples from the head-space of each jar) were analyzed using a VG Micro-gas injector connected to a VG Isochrom isotope ratio mass spectrometer in continuous flow mode.

3. Results and discussion

3.1. Organic carbon

Soil organic carbon $\delta^{13}C$ values (from the A horizon of each site) varied from -24.1‰ to -25.9‰ PDB. Values in the arid portion of the gradient (200 to approximately 500 mm MAP) generally decreased and a linear regression of SOC ^{13}C vs. MAP was significant ($r^2=0.71$, $p=0.02$). SOC ^{13}C values at sites greater than 500 mm MAP appeared to remain relatively constant (Fig. 2).

Two grass species, *Agropyron spicatum* (bluebunch wheatgrass) and *Festuca idahoensis* (Idaho fescue), were present throughout the arid portion of the gradient (200–450 mm MAP). Although each species contribution to the SOC pool certainly varied over the precipitation gradient, carbon isotope values
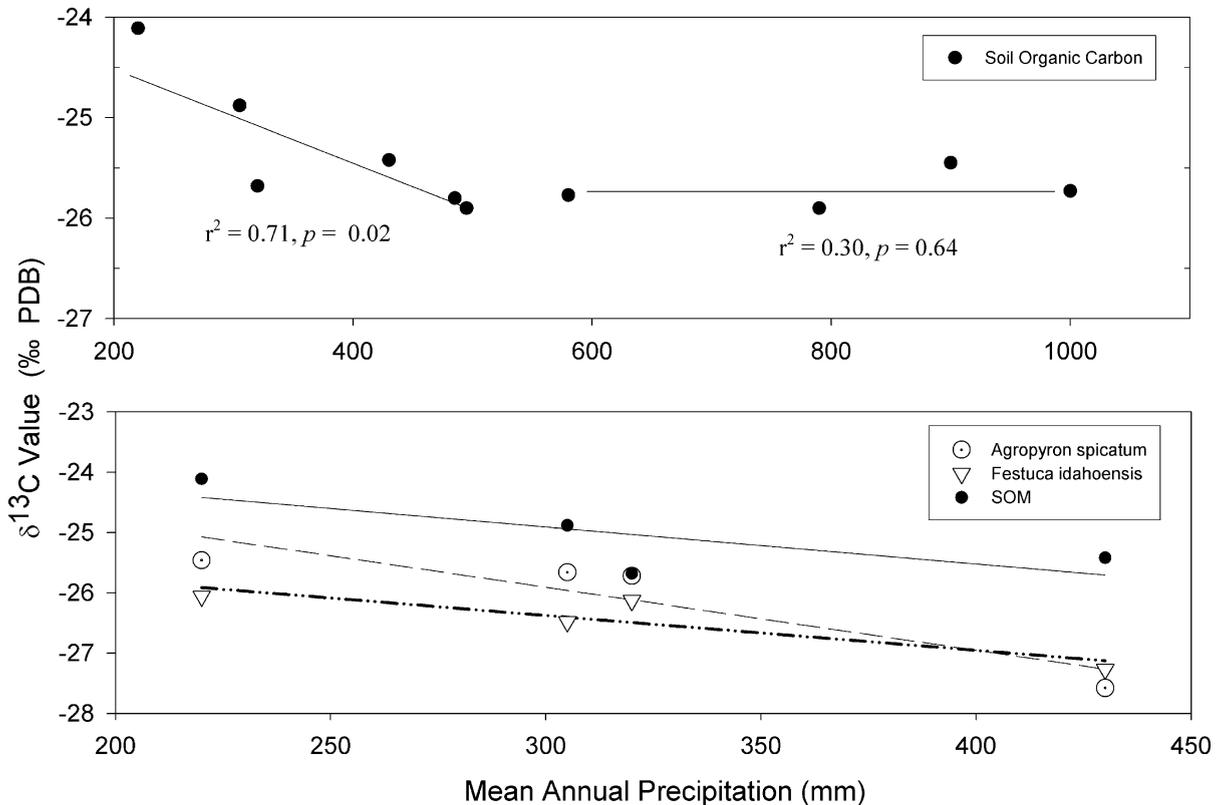
Fig. 2. Soil organic matter and plant tissue $\delta^{13}C$ values vs. mean annual precipitation.

of plant tissues for both species show trends similar to SOC isotope values (Fig. 2).

Decompositional processes and turnover rate of carbon are the primary factors affecting SOC ^{13}C. SOC ^{13}C changes of approximately 1–2‰ generally occur with depth in the soil (Nadelhoffer and Fry, 1988; Wedin et al., 1995). Although these changes are suspected to be from decompositional processes, the exact nature of this enrichment is not known. All of our sites showed 1–2‰ enrichment in SOC-^{13}C (the A horizon being the most depleted) with depth to 1 m (Stevenson, 1997).

The turnover rate of carbon, and incorporation of new carbon into SOC at each site, may impact SOC ^{13}C values (particularly if older inputs of carbon were isotopically dissimilar). The small number of samples makes it difficult to ascertain the factors responsible for the trend in SOC ^{13}C. The similarity in trends between plant ^{13}C and SOC ^{13}C values, however, suggests that there is little differential modification of

SOC isotopic signature due to varying turnover time across the arid portion of the gradient.

The trend in $\delta^{13}C$ plant and SOC values is evident from 200 to approximately 500 mm MAP; above 500 mm, vegetation appears to be insensitive to changes in precipitation. Drought stress would be the most likely explanation for this trend. Drought stress decreases stomatal conductance, increasing the ratio of external to internal CO_2 partial pressures. As a result, $\delta^{13}C$ differences of 1–3‰ have been noted in laboratory and field experiments involving a single species grown at various moisture levels (Farquhar and Richards, 1984; Read et al., 1991, 1992). Regional climatic records (Phillips, 1965, 1970) indicate that calculated values of annual actual evapotranspiration (AET) generally exceed MAP in the range of 450–550 mm (the value for each site is dependent upon species water-use, vegetative cover and soil characteristics as well as precipitation amount).

Although water-use may vary greatly from species to species, it is reasonable to assume that drought-stress ^{13}C discrimination occurs primarily when AET exceeds precipitation over the period of active plant growth. When precipitation exceeds AET, even though species water-use may differ, sufficient moisture is present so that little ^{13}C discrimination occurs.

Apart from C_3 vs. C_4 shifts, the translation of plant discriminatory factors to SOC has only recently been intensively investigated. Although not conclusive, our data suggest that annual calculated AET and MAP values correspond with plant and SOC $\delta^{13}C$ values and indicate drought-induced ^{13}C discrimination on a community level.

3.2. Respired carbon dioxide

The intermediary between SOC and pedogenic carbonates is CO_2 produced from microbial and root respiration. Temporal and spatial variability in soil respiration have been shown to be high (Singh and Gupta, 1977), and intensive measurement is needed to adequately determine the production rate and seasonal flux of CO_2 through soil. For the purpose of this study we performed a laboratory incubation experiment to determine if differential fractionation (due to differences in substrate composition) occurs between SOC and microbially respired CO_2 in soils from across the precipitation gradient.

Various components of plant tissue (e.g., lignin vs. cellulose) have different isotopic signatures (Benner et al., 1987). Because both microbial communities and litter quality may vary from site to site, it is possible that differential fractionation from SOC to respired CO_2 may occur over the precipitation gradient. Results from the respiration experiment (Fig. 3) show that isotopic composition of respired CO_2 ranged from $+0.6$ to -1.0‰ PDB above and below the $\delta^{13}C$ SOC $\delta^{13}C$ values. The mean $\Delta^{13}C$ ($CO_2 -$ SOC) value across all sites was -0.14‰. The lack of a significant relationship between $\Delta^{13}C$ ($CO_2 -$ SOC) and MAP indicates there is no trend in differential fractionation across the precipitation gradient.

3.3. Pedogenic carbonates

Mean carbonate $\delta^{13}C$ values from the top $1-2$ mm of the Bk horizon varied from -4.1‰ to -10.8‰

PDB along the gradient (Fig. 4). Linear regression of mean carbonate $\delta^{13}C$ values vs. carbonate ^{13}C was significant for MAP ($r^2 = 0.79$, $p = 0.007$), estimated productivity ($r^2 = 0.88$, $p = 0.002$) and soil organic carbon content ($r^2 = 0.83$, $p = 0.004$). Values at the most arid site (KP-1) exhibited much higher variation in $\delta^{13}C$ values than other sites (individual samples ranged from -3.65‰ to -8.17‰ PDB) and somewhat lessened the overall fit of carbonate $\delta^{13}C$ vs. MAP relationship.

We believe the explanation for the high variation at the most arid site is twofold. First, carbonate values near the surface are inherently variable as the balance between atmospheric and soil CO_2 is temporally and spatially more variable (Cerling, 1984). Several of the values from the KP-1 site were more depleted than the others and may represent the influence of atmospheric mixing; however, more intensive testing across several arid sites would be needed as verification. Secondly, at the most arid sites (unlike the other sites), there was much higher variation between samples taken from the very upper part of the Bk horizon and bulked samples of the Bk horizon as a whole (data not shown), indicating a possible disequilibrium between present-day carbonate formation and that of the formation of the bulk of the Bk horizon.

The $\Delta^{13}C$ (carbonate – SOC) value (the difference in $\delta^{13}C$ signatures between carbonate and SOC) provides information on the summed fractionations that occur between the carbon isotope composition of soil organic matter and carbonates, and is independent of $C_3 - C_4$ shifts. Trends in MAP ($r^2 = 0.59$, $p = 0.04$), estimated productivity ($r^2 = 0.81$, $p = 0.006$) and soil organic carbon content ($r^2 = 0.58$, $p = 0.05$) were significantly related with $\Delta^{13}C$ (carbonate – SOC) values; however, largely because of the arid site data point, the r^2 values were not as high as the $\delta^{13}C$ relationships.

The primary determinants of the isotopic composition of pedogenic carbonate-carbon are the isotopic composition of soil CO_2, and the amount and isotopic signature of diffused atmospheric CO_2, and temperature (Cerling, 1984). Theoretically, if soil respiration is the sole carbon source for carbonate precipitation, the difference between SOC and carbonate isotopic signatures is the sum of fractionations that occur during this process. This
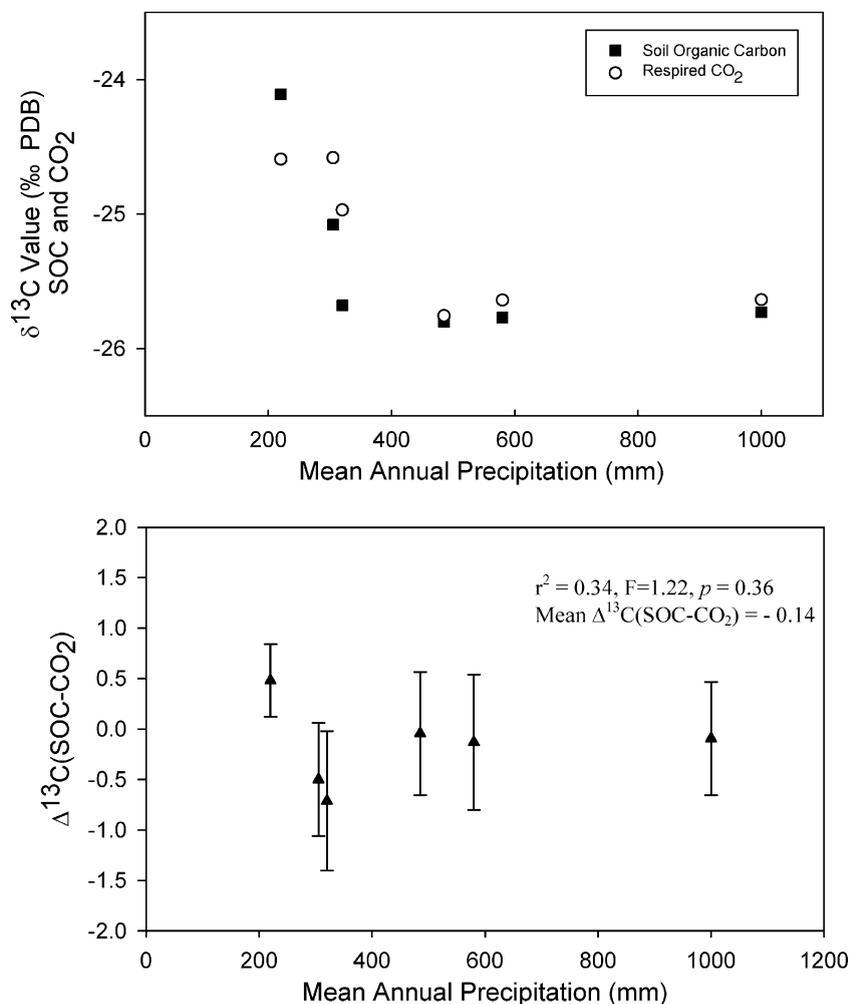
Fig. 3. Soil organic matter $\delta^{13}C$ values, mean respired CO_2 $\delta^{13}C$, and mean $\Delta^{13}C$ (SOC – CO_2) values vs. mean annual precipitation (error bars represent standard deviation, $n = 3$).

has been calculated at $14-16‰$ (Cerling, 1984, 1992; Cerling et al., 1989) and is dependent upon the magnitude of respired CO_2 flux and the temperatures of dissolution of CO_2 and precipitation of carbonate. Because atmospheric CO_2 is more enriched in ^{13}C than soil CO_2, increasing additions of atmospheric CO_2 will increase $\Delta^{13}C$ (carbonate – SOC) values.

Increased mixing of atmospheric CO_2 into the soil as mean annual precipitation (and soil respiration) decrease would explain the trends evident in Figs. 4 and 5. As MAP decreases, depth of carbonate precipitation also decreases, and in general, so does

productivity and soil respiration. Both decreased depth of carbonate formation and a lower soil respiration rate increase atmospheric $CO_2 - ^{13}C$ input in pedogenic carbonate formation.

In the Palouse system, C_3 plants predominate (Daubenmire, 1969) and there are only small (and relatively constant) changes in SOC $\delta^{13}C$ over the precipitation gradient. Since SOC isotopic signatures remain relatively constant, the $\delta^{13}C$ of the carbonates also correlates strongly with productivity. The carbonate $\delta^{13}C$ vs. productivity relationship (as opposed to $\Delta^{13}C$ (carbonate – SOC) vs. productivity relationship) will only hold when SOC isotopic values
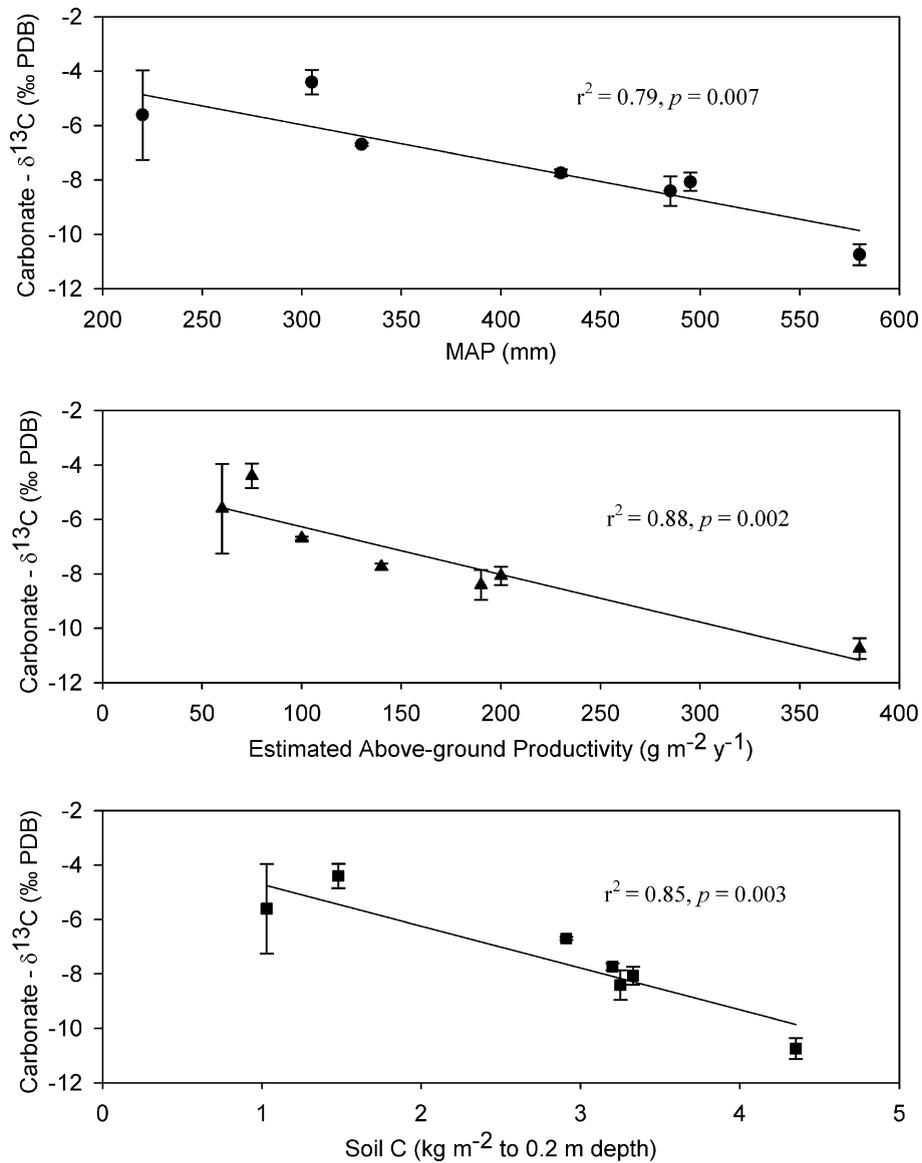
Fig. 4. Carbonate δ^{13}C values vs. mean annual precipitation, above-ground productivity, and soil organic carbon at seven sites along a precipitation gradient in the Palouse region. Error bars represent the standard deviation of samples from each site ($n=5$ for the most arid site, $n=3$ for all other sites).

remain relatively constant (i.e. in the absence of C_3 vs. C_4 shifts).

Amundson et al. (1989b) attributed changes in observed patterns of carbonate-δ^{13}C isotope values to plant density (and thus soil respiration) in Nevada, and Cerling (1992) has shown how paleosol carbonate can be used as an indicator of pCO$_2$ in paleo-atmospheres.

Cerling (1984) has suggested that atmospheric CO$_2$ mixing in most arid soils occurs in the top 50 cm and may reach as far as 100 cm below the soil surface, but that soil freezing may facilitate mixing even deeper. Most other carbonate isotope studies have been performed in warm desert environments where freezing of the soil does not occur frequently. Our data suggest
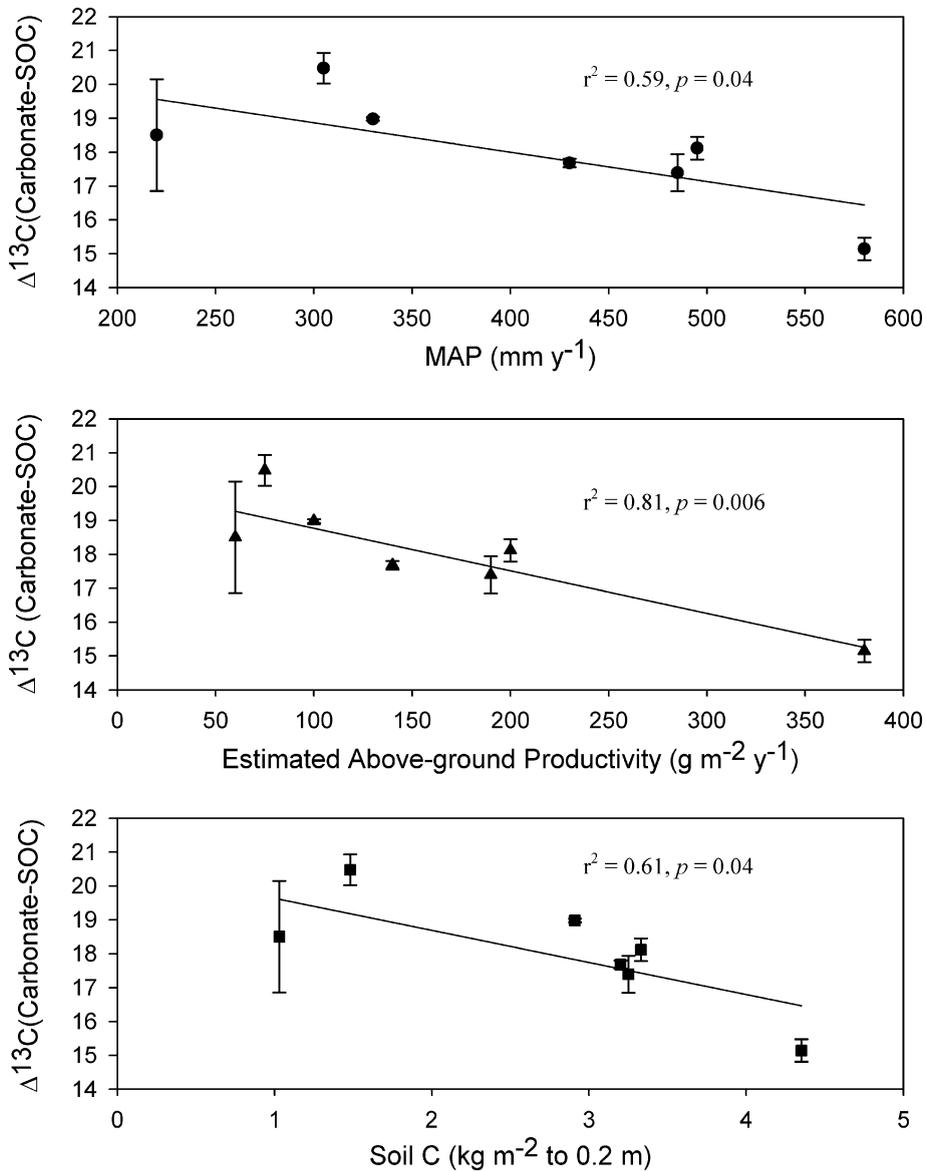
Fig. 5. $\Delta^{13}C$ (carbonate – SOC) values vs. mean annual precipitation, above-ground productivity, and soil organic carbon at seven sites along a precipitation gradient in the Palouse region. Error bars represent the standard deviation of samples from each site ($n = 5$ for the most arid site, $n = 3$ for all other sites).

that atmospheric mixing may occur to a much greater extent in temperate regions and reach to a depth of 250 cm below the soil surface.

The depth to which atmospheric mixing occurs is a function of both soil porosity and soil CO_2 flux. In temperate regions, freezing of the soil tends to produce frost heave, resulting in macropores between frozen soil aggregates. These macropores greatly facilitate diffusion of atmospheric CO_2 into the soil at a time when soil respiration is at its minimum.

Although the high correlation of carbonate-carbon isotopic signature with productivity is useful, the driving variable behind the isotopic signatures is the rate of soil respiration. In order to ascertain approx-

imate respiration values, the Cerling isotopic diffusion model (Cerling, 1984) was used to back-calculate respiration rates needed to account for changes in carbonate-carbon isotope values over the climatic gradient. Calculated values ranged from 1.0 to 10.0 mmol C m^{-2} h^{-1} and were similar to rates obtained by Wildung et al. (1975) of 0.63 to 4.20 mmol C m^{-2} h^{-1} measured over a 2-year period in a sagebrush–bunchgrass community in south-central Washington state.

4. Summary and conclusions

Jenny's (1941) state factor approach asserts that soils are a function of the environment under which they formed. Conversely, soil features are a result of various ecosystem processes. Our data suggest that in the absence of large C_3 to C_4 plant-induced shifts in isotopic composition of soil components, smaller shifts associated with process-level ecosystem properties may be discerned.

On a C_3-dominated climatic gradient in southeastern Washington state, we found that SOC values showed a significant relationship with MAP over the range 200 to approximately 500 mm and were similar to trends in two plant species present throughout this range. The similarity in trends (possibly due to drought-induced ^{13}C discrimination in plants) suggests that even small changes in plant community isotope values are transmitted to SOC.

Trends in pedogenic carbonate $\delta^{13}C$ values were significantly related with MAP, total soil carbon, and estimated productivity values. Carbonate $\delta^{13}C$ was highly variable at the most arid site (KP-1) possibly reflecting the inherent variability in the balance between soil respiration and atmospheric diffusion. Our data suggest that pedogenic carbonate ^{13}C signatures may serve as a useful proxy for long-term productivity (and amounts of precipitation required to sustain such productivity) at sites where carbonate is present.

On time scales of centuries to millennia, various methods have been developed to reconstruct ecosystem structure (i.e., pollen records, etc.); however, few measures of ecosystem function are available. The isotopic relationships between ecosystem components presented here may prove useful in defining these measures.

Paleosols are increasingly being studied to determine how past climatic shifts have affected terrestrial ecosystems, and conversely, how projected climate change may affect present and future terrestrial ecosystems. The utility of the isotopic characterization of soil components recovered from paleosols can never be fully exploited until further research is directed toward assessing the specific controls on the range and variation of these data in modern ecosystems.

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