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Pedogenic Isotopic Indicators of Climate and Carbon Cycling in Fish Lake Valley, Nevada

by

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1

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Abstract

Alluvial fans of four different ages were selected to study the isotope chemistry of pedogenic carboante. Rinds formed on the undersides of volcanic clasts were carefully removed and inner, indurated portions were analyzed for stable isotopic composition and radiocarbon content. Radiocarbon dates on carbonate and coexisting, occluded organic carbon indicate that rinds in these soils do not remain sealed over time and that inorganic and organic C are undergoing isotopic exchange. Younger dates higher in two profiles contrast with an older date below 120 cm in another profile, and may indicate that a greater degree of carbonate reprecipitation has occurred above about 50 cm depth.

The $\delta^{13}C$ values of carbonate and occluded organic C generally become heavier by about 2 o/oo (per mil) above about 50 cm depth. We expect carbonate to be heavier above about 50 cm due to a greater proportion of heavier, atmospheric CO_2 in upper than in lower profiles. However, the occluded organic C is also heavier above 50 cm, possibly indicating different rooting patterns for different plant species. $\delta^{13}C$ of inner rind carbonate is heavier in younger soils than in older soils by a maximum of 3 o/oo, suggesting that the proportion of C4 plants has increased by 10 to 20 % during the middle to late Holocene.

$\delta^{18}O$ of carbonate increases by about 4 o/oo toward the surface of all profiles, possibly resulting from greater evaporation above about 50 cm depth, isotopically heavier precipitation in middle to late Holocene storm events, change in effective wetting depth, or a combination of these effects. These data show that soil carbonate recrystallizes readily in the dynamic, open system characteristic of pedogenic environments.

Introduction

The isotopic composition of pedogenic carbonate has been used in recent years as an indicator of environmental conditions prevailing when carbonate precipitated. Amundson and others (1988a) showed that $\delta^{13}C$ of near-modern carbonate in the Mojave desert is related primarily to plant density and biological activity, which increase with increasing elevation, and to a lesser extent the isotopic composition of local plants. ($\delta \text{ o/oo} = [(R_x - R_{std})/R_{std}]1000$; C standard is Pee Dee Belemnite (PDB), Craig, 1957; O standard is Standard Mean Ocean Water (SMOW), Craig, 1961.) Carbonate $\delta^{18}O$ is related to the isotopic composition of precipitation, as influenced by storm tracks and temperature of condensation, and to evaporative effects (Amundson and others, 1989; Quade and others, 1989a). Organic matter $\delta^{13}C$ provides a direct measure of the plant and microbial C isotopic composition, and when combined with contemporaneous carbonate isotopic composition, provides an indication of soil respiration rate. Radiocarbon analysis of carbonate (particularly form limestone terranes) is subject to detrital carbonate contamination or reprecipitation of older carbonate (e.g. Williams and Polach, 1969; Callen and others, 1983; Gardner, 1984; Amundson and others, 1989), but may be useful in determining the amount of recently precipitated carbonate.

Much stable isotope research has focussed either on Holocene soils (e.g. Cerling, 1984; Amundson and Smith, 1987; Amundson and others, 1989; Pendall and Amundson, 1990) or on much older or undated soils (e.g. Cerling and others, 1977; Gardner, 1984); however, no systematic work has been done on isotopic composition of soils of increasing age, particularly ranging from the Holocene to the Pleistocene. The carbonate age in some cases has been assumed to be the age of the deposit in which the soil formed (e.g. Magaritz and others, 1979; Gardner, 1984; Quade and others, 1989b), but other studies use ^{14}C ages to assess the mean age of post-depositional, pedogenic carbonate (Magaritz and others, 1981; Amundson and others, 1989; Pendall and Amundson, 1990). Many studies have analyzed only the isotopic composition of soil carbonate, and use models to estimate the isotopic composition of soil CO_2 and organic matter which contribute to the carbonate isotope makeup (Cerling and others, 1977; Gardner, 1984; Rabenhorst and others, 1984). We hope to address some of these assumptions by measuring as many components of the C system as possible in order to better understand the systematics of inorganic and organic C movement through soils.

In this study, we examine the C and O isotopic composition of inner rind and bulk carbonate, and the C isotopic composition of occluded organic C in rinds and soil CO_2 in an age sequence of four soils. Soil CO_2 concentration and isotopic composition, measured in May, give an idea of soil respiration rate and its variability with profile depth (or stratigraphy) and the type of plants which are actively respiring at that time. Soil CO_2 concentration and isotopic composition are the most direct indicators of seasonal environmental conditions, especially in arid regions, because of plant and microbial sensitivity to moisture and warmth (Moitke, 1974). Soil carbonate, however, provides a better indication of long-term environmental conditions, as it accretes more slowly. Stable isotope analyses are most useful for environmental reconstruction if the components are also dated. We provide radiocarbon analyses for carbonate and coexisting, occluded oragnic C in three soils which have relative and absolute age controls ranging from <700 ka to 2 ka (see Harden and others, this volume). The apparent ages of carbonate and occluded organic C, measured by radiocarbon content, indicate the relative closure of the two fractions to re-equilibration, and may provide a way to assess organic C turnover and inorganic C residence time in these soils.

The soils in this study do not make up a chronosequence because lithology and elevation (and thus biota) are not held constant (morphology and chemistry of soils in this study discussed in Harden and others, this volume). We feel that the lithology does not directly affect the isotopic composition because no limestone exists in the drainage; however, lithology does influence the porosity of the soils and the relative induration of carbonate rinds. The older soils (< 700 ka) are developed on granitic alluvium and form K horizons but poor rinds because the clasts grusify, while soils less than about 200 (?) ka are developed mainly on volcanic alluvium, which maintains better, more indurated rinds. The elevation of the oldest profile is about 430 m higher than the other soils, and thus reflects slightly wetter and cooler conditions today and in the past.

The Trail Canyon area in the northwestern end of Fish Lake Valley (see Fig. 1 of Reheis and others, this volume) is a good place to study the dynamics and systematics of C and O stable isotopes, because all the carbonate here is pedogenic (derived from dust and thus dissolved as it washes down); the soils are well-drained and generally not completely indurated, although preferential flow may

be occurring; they are located at a modern ecologic boundary between woodland and shrubland; and there has been a long period of carbonate accumulation. We must be cautious in directly comparing our isotopic results with research done in southern Nevada (e.g. Amundson and others, 1988a and b, 1989; Quade and others, 1989a; Benson and Klieforth, 1989), because Trail Canyon is north of the Mojave desert province, as indicated by the absence of Larrea tridentata and Yucca Brevifolia (Spaulding, 1980). Thus, climatic conditions are different here, with the majority of precipitation falling from February to April (Rush and Katzer, 1973), lower minimum temperatures, and probably somewhat different storm tracks.

Methods

Clasts and bulk soil were collected from five soil profiles: TC1A and CH1A, located on surface Q1, <700 ka; TC4C, located on surface Q4, >50 ka; TC5B, located on surface Q5, 7-11 ka; and TC6C, located on surface Q6, 2-4 ka (locations shown on Fig. 1 of Harden and others, this volume; age control and soil descriptions also discussed). Soil gas was collected in May in separate vials for CO_2 content and stable isotope analysis at sites TC1A, TC6C, and CH1A. The clasts were cleaned and outer, powdery carbonate was removed by brushing with a stiff brush. Inner rinds, closest to the clast, were scraped off and ground to pass a 200 mesh sieve. Carbonate stable isotope ratios were measured on a Finnigan MAT 251 gas source mass spectrometer after reaction in 100% H_3PO_4 and CO_2 separation. Occluded, organic C and bulk soil organic C were prepared for stable isotope and radiocarbon analysis in the following manner: carbonate was removed by reacting splits of rind carbonate samples or bulk soil samples in 3 $\underline{N}$ HCl overnight; the non-reacted portion was rinsed and filtered onto Whatman glass fiber filters and dried at 105 °C for three hours. Organic samples were weighed and mixed with excess powdered cupric oxide, placed in molybdenum foil boats, and combusted at 850 °C for 3 hours; CO_2 was cryogenically purified and analyzed as in carbonate analysis. Splits of carbonate and organic C samples analyzed for stable isotope composition were separated for radiocarbon analysis by accelerator mass spectrometry (AMS) at Lawrence Livermore National Laboratory. There, carbonates were reacted in vacuo with ortho-phosphoric acid, and organic C was combusted with cupric oxide wire at 900 °C for 2 hours. CO_2 released was cryogenically purified and reduced to a graphite AMS target using methods described in Vogel and others (1984).

Results and Discussion

Radiocarbon Results

Apparent ages and percent modern C of inner rind carbonate and organic C in three profiles are shown in Table 1. It is clear that the pedogenic carbonate and organic C are much younger than the deposits in which the soils are forming. This shows that at Trail Canyon, the clast rinds undergo dissolution and reprecipitation of carbonate, and that organic C is not in isotopic equilibrium with the carbonate in which it was presumed to be "occluded." It is interesting that the deepest sample gives the oldest apparent age, and that the samples at about 50 cm depth have similar, younger apparent ages. This may indicate a change in effective wetting over time. Alternatively, it may simply show that the oldest soil contains more "dead" or older C. Amundson and others (1988b) obtained radiocarbon results from carbonate rinds that were generally too old, which they attributed to detrital limestone incorporation, whereas occluded organic C ages at Kyle Canyon were significantly younger than carbonate. Gile and Grossman (1979) obtained older radiocarbon dates for Pleistocene than for Holocene carbonate, but most of the Pleistocene samples appeared to "have undergone exchange with environmental C-14 after emplacement" (p. 197).

Table 1. AMS data from Lawrence Livermore Laboratory. Lab numbers and errors reported in Supplementary Table 1.

Surface Age	Profile Name	Depth (cm)	Apparent Age (BP)	
			Carbonate	Organic C
<700 ka	TC1A	120-140	11,410	3620
>50 ka	TC4C	37-63	2210	4220
7-11 ka	TC5B	40-50	2700	1640

Three models may be used to explain the results we obtained (Trumbore and others, 1989). First, a simple two-part mixing of old inorganic or organic C with modern C may be ocurring . This implies that the system remained closed for a certain length of time and has recently been perturbed in some way such that some old C is reprecipitating (in the case of carbonate) or is being replaced by young C (in the case of organic C), or that in sample preparation, we did not effectively remove younger rind material from older material. We feel that this simple, two-component system is not valid for these soils. Second, there may be a continual isotopic exchange over time, with a portion of the carbonate recrystallizing and a portion of the organic matter decomposing and being replaced by modern C. This exchange would not necessarily occur at similar

6

rates in organic and inorganic fractions. Third, there may be periods of carbonate accumulation or recrystallization and organic C turnover interspersed with periods of slower exchange. This could occur if pulses in dust deposition and wetting events are not evenly spaced through time (Chadwick and Davis, 1990). The radiocarbon results point out the difficulty of interpreting paleoclimatic signals in soils which have been exposed to changing climates over time.

Stable C Isotopic Compositon

Carbonate

Although the age of pedogenic carbonate in soils at Trail Canyon does not reflect deposit age, the stable isotopes may not be as profoundly influenced by modern C "contamination" as ^{14}C (Williams and Polach, 1971). $\delta^{13}C$ of inner rind carbonate shows trends with profile depth and with increasing age (Fig.1). Profile TC1A, on the oldest deposit, has the lightest $\delta^{13}C$ values; however, it is also higher in elevation than the other soils. Rather than assigning a correction factor for elevation which was not developed at this location, we present the raw data. Quade and others (1989a) measured a 4.6 o/oo decrease in $\delta^{13}C$ per 1000 m increase in elevation in Holocene soils in the Spring Mountains, which is about twice what we measure here. The average $\delta^{13}C$ of the other profiles appears to become heavier with decreasing deposit age. This suggests that there is an increasing proportion of C4 plants and/or decreasing soil respiration rates (more atmospheric CO_2) in younger soils. We must remember, however, that this carbonate represents a range of ages. The $\delta^{13}C$ also becomes heavier with decreasing depth in the profile. This may indicate increased diffusion of heavier atmospheric CO_2 into the upper profiles when soil moisture and respiration rates are low. Although soil CO_2 measured in May, 1989, indicated little atmospheric CO_2 (Fig. 3; discussed below), the growing season is likely to be quite brief, with atmospheric CO_2 making up a large fraction of annual average CO_2 (Terhune and Harden, in press). More rooting of C4 species in the upper profiles (discussed below) or different generations of carbonate at different depths may also explain the carbonate $\delta^{13}C$ decrease with depth.

Organic Carbon

The isotopic composition of organic C is determined by the plant species contributing to organic matter in the soil, with C3 (Calvin cycle of photosynthesis) plants averaging -27 o/oo and C4 (Hatch-Slack) plants averaging -12 o/oo (Hoefs, 1980). Decomposition

7

tends to fractionate C, volatilizing isotopically heavier components (carbohydrates, proteins) and leaving an increasing proportion of lighter components (e.g. lignin) (Benner and others, 1987). The large fraction of modern C in organic C in carbonate rinds suggests rapid decomposition of most of the organic matter in these soils, even deep in the profile (Table 1). The chemical nature of the organic C in rinds was not determined, but is assumed to be some combination of microbial and root exudates, and possibly dissolved organic C which has been incorporated into carbonate.

$\delta^{13}C$ of organic C in inner rinds exhibits nearly as much variability with profile depth as with deposit age (Fig. 2). Depth variability can be explained by variability in water holding capacity in these stratified soils, which leads to preferential rooting and microbial activity in finer-grained horizons. Different plant species are probably utilizing different zones in the profile, leading to isotopic differences if the plants use different photosynthetic pathways. If decomposition has not altered the isotopic composition drastically, $\delta^{13}C$ indicates that the proportion of C4 plants varies from about 45 to 55% in these soils. This is a significantly greater proportion than other researchers have found in southern Nevada (Amundson and others, 1988a, 1989; Quade and others, 1989a), and is surprising in light of cooler temperatures and spring precipitation.

We have not done any vegetation sampling at the Trail Canyon fans yet, so we cannot provide an accurate cross-check on the relative abundance of C3 and C4 plants. However, we have observed that Atriplex confertifolia, a C4 shrub (Dzurec and others, 1985), is common on all fans up to about 1900 m. It is possible that the organic C in rinds represents at least some plants which no longer grow in the area. Analysis of bulk soil organic C (assumed to have mean residence time of <1000 years; Fig. 2) shows that in younger soils, (TC5B, TC6C) modern organic matter is very close isotopically to rind organic C, but in TC4C, vegetation changes are apparent. In TC4C, a modern bulk organic C $\delta^{13}C$ value of -12.8 o/oo clearly indicates C4 plants at 65-100 cm depth, but shallow-rooted plants (-19.6 o/oo at 37-63 cm) are a mix of C3 and C4. This variability in the isotopic composition of soil organic C and thus CO_2 with depth complicates modeling of respiration rates and carbonate isotope composition.

8

A check on the overall equilibrium between carbonate and rind organic matter may be provided by determining the average difference between $\delta^{13}C$ of carbonate and organic C. We expect a 14 to 15 o/oo enrichment in carbonate relative to organic C (due to gas diffusion and phase changes (Deines and others, 1974; Cerling and others, 1989)); however, there is an average of 16 o/oo, and a maximum of 17.8 o/oo, fractionation in these soils. The overall greater than expected fractionation could be due to a higher contribution from atmospheric CO_2 to carbonate, although we would expect more of a depth trend if this were a large factor. Since the relatively high fractionation appears to be independent of depth in the profile, we attribute it to different rates of isotopic exchange for carbonate and organic C, i.e., these components are not truly contemporaneous.

Soil CO_2

The isotopic composition of soil CO_2 gas sampled in late May, 1990, is shown in Fig. 3. TC1A, 430 m higher in elevation than the other two sites, has the lightest $\delta^{13}C$, corresponding to higher respiration and/or fewer C4 plants, as expected. CH1A, on roughly the same age surface as TC1A but at a comparable elevation to the lower sites, has a CO_2 profile similar to that of TC6C down to 60 cm. This shows that the elevation (as it influences soil respiration rates) rather than soil age controls the soil CO_2 composition. TC6C has a layer of isotopically heavy air at about 80 cm, with a low (0.04 %) concentration of CO_2, suggesting that a gravelly lens has allowed lateral flux of atmospheric air ($\delta^{13}C = ^-7$ o/oo) into the profile at that depth. Except for this layer, $\delta^{13}C$ of CO_2 is 2 to 5 o/oo heavier than organic C, and concentrations range from 0.1 to 0.2 %, which indicates a high level of soil respiration and little downward mixing of atmospheric CO_2 at this time of year. Although the soil was not saturated at sampling time, recent rains had produced a bloom of annual vegetation (and presumably microbial growth) before the dessication of the summer.

In TC1A, the modern soil CO_2 is about 2 o/oo lighter than would be expected based on equilibrium with organic C and carbonate of inner rinds (we expect CO_2 to be about 4 o/oo heavier than organic C (Cerling and others, 1989) and about 10 o/oo lighter than carbonate at 25 °C (Deines and others, 1974)). In CH1A, soil CO_2 is a bit lighter than expected from equilibrium with rind organic C, but close to equilibrium with carbonate and bulk soil organic C, suggesting that at

9

shallow depths this profile may be in equilibrium with modern conditions. In TC6C, the only soil considered to be truly modern, soil CO_2 appears in equilibrium with rind organic C, bulk soil organic C, and carbonate. These trends are general and based on scanty CO_2 data, but it is encouraging that both carbonate and organic C in the TC6C profile appear to be in isotopic equilibrium with modern CO_2. If this is the case, it suggests that modern carbonate at Trail Canyon precipitates most actively after soil moisture has equilibrated with CO_2 produced in spring and early summer.

Stable O Isotopic Composition

Figure 4 shows trends in $\delta^{18}O$ of inner rind carbonate, which has been adjusted to reflect the isotopic composition of water (raw data minus 28.5 o/oo, the fractionation factor at 25 °C (O'Neil and others, 1969)); TC1A has been further corrected for elevation based on Milne and others (1987) by adding 1 o/oo. There is about a 4 o/oo increase in $\delta^{18}O$ with decreasing depth in profiles TC4C and TC5B which may be due to evaporative enrichment of isotopically heavier water remaining in the soil, or to isotopically lighter winter precipitation infiltrating more deeply into the profile. Other workers have noted similar trends with depth (e.g. Quade and others, 1989a). It is difficult to estimate the importance of evaporation without collecting precipitation and soil water, and without measuring soil temperatures to determine the correct fractionation factor. McKinley (personal communication, 1991) has measured precipitation in central-western Nevada for several years and notes lighter isotopic values in winter. Friedman (personal communication, 1991) states that snow is not subject to evaporation as it falls or as it melts, so that melt water from snow will be lighter than rain falling at similar elevations. The $\delta^{18}O$ of carbonate at all soil depths is heavier by at least 5 o/oo than would be expected from data collected by McKinley and by Kendall at sites north and east of Fish Lake Valley (Fig. 5), suggesting that evaporation is important in these soils. Precipitation actually falling at Fish Lake Valley is probably intermediate isotopically between data shown in Figure 5 and Benson and Klieforth's (1989) data shown in Figure 4.

These is also a trend of increasing $\delta^{18}O$ with decreasing deposit age, which indicates heavier precipitation or greater evaporation in more recent times (Fig. 4). Upper horizons of TC6C and TC5B are significantly heavier than in CH1A, at a similar elevation, suggesting that if carbonate on younger deposits is younger than on CH1A, there

has been more evaporation and/or warmer temperatures influencing the isotopic composition in the middle to late Holocene than during precipitation of CH1A carbonate (possibly late Pleistocene based on apparent age of TC1A carbonate at 120 cm depth (Table 1)). Additional radiocarbon and stable isotope analyses of carbonates at a variety of depths are planned to assess the possibility of a changing effective wetting depth with time.

Summary and Conclusions

Radiocarbon data shows that Pleistocene pedogenic carbonates in Trail Canyon soils are not well sealed from subsequent recrystallization. This implies that paleoenvironmental information obtained from other Pleistocene soil carbonates cannot be assumed to remain unaltered by post-depositional climatic change. Pleistocene carbonate rinds at Trail Canyon appear to average environmental signals (such as stable isotope ratios) over too long a peroid of time to be of direct use. Rinds formed in areas of extremely high dust flux or limestone may be less porous with a more compact micritic matrix than those observed and measured at Trail Canyon. These rinds may be better sealed and offer a greater chance for preserving climatic signals. Additionally, if better sampling methods can be developed to remove more discrete layers of material, more specific climatic information may be obtained.

Mixing models can be developed for Holocene soils, using apparent radiocarbon ages and modern dust and organic matter accumulation rates, to assess the turnover rates of organic C, and to assess the rates of carbonate isotope exchange. However, more radiocarbon data and rind accumulation rates are needed to assess trends with soil depth and with deposit age at Trail Canyon.

Stable isotope composition of inner rind carbonate was analyzed in hopes of determining relative respiration rates in soils (as a proxy for average soil CO_2), once the C3:C4 species ratio was obtained from occluded organic C in the rinds. However, because dating showed that the components are not truly co-existant, this was not done for all profiles. Carbonate and organic C in the youngest profile, TC6C (not analysed for radiocarbon), appear to be in isotopic equilibrium, and may be modeled to estimate the modern respiration rate. When measured values are entered into a model developed by Quade and others (1989a), using a porosity of 40 % at 20 °C, a respiration rate of approximately 1 mmole m^{-2} hr^{-1} is obtained, which is in line with values for arid ecosystems (Singh and

11

Gupta, 1977).   Ideally, a comparison of changing respiration rates with time would provide a sensitive indicator of available moisture; perhaps soils with better-sealed carbonates will be identified for additional work on this topic.

Stable O isotopic composition in carbonates is difficult to interpret due to unknown influences of evaporation and differential penetration of precipitation in different seasons.   Heavier $\delta^{18}O$ with decreasing profile depth may indicate greater evaporation above about 40 cm.   Heavier $\delta^{18}O$ with decreasing deposit age may indicate increased evaporation or storm tracks with heavier precipitation during the middle to late Holocene.   Additional dating in conjunction with stable isotope analyses may provide insights into differences between the glacial and interglacial water balance, as it is influenced by temperature and precipitation regime, at Fish Lake Valley.

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Carbonate
delta 13C, o/oo PDB

Organic C
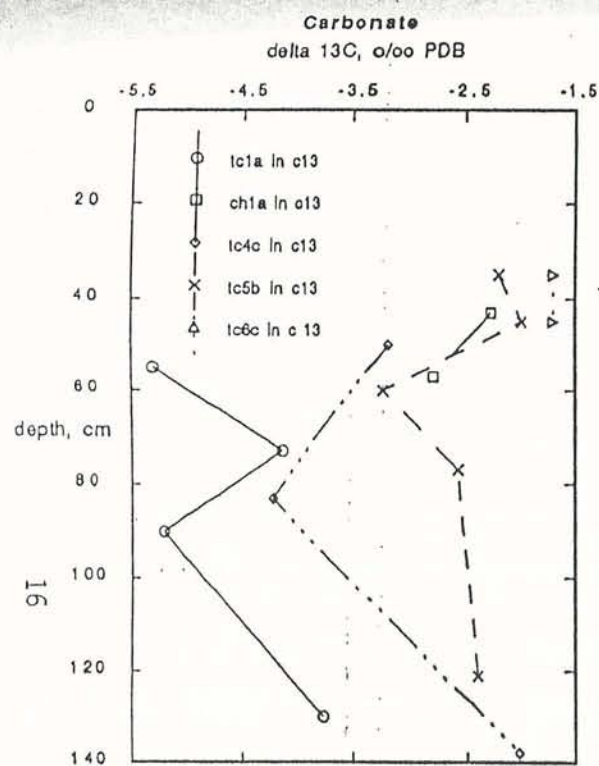delta 13C, o/oo PDB

Soil CO2
delta 13C o/oo PDB

Figure 1. C isotopic composition of inner rind carbonate in soil profiles at Trail Canyon, Fish Lake Valley. Profile TC1A is 430 m higher in elevation than the other soils. Radiocarbon analyses of three samples are shown in Table 1.
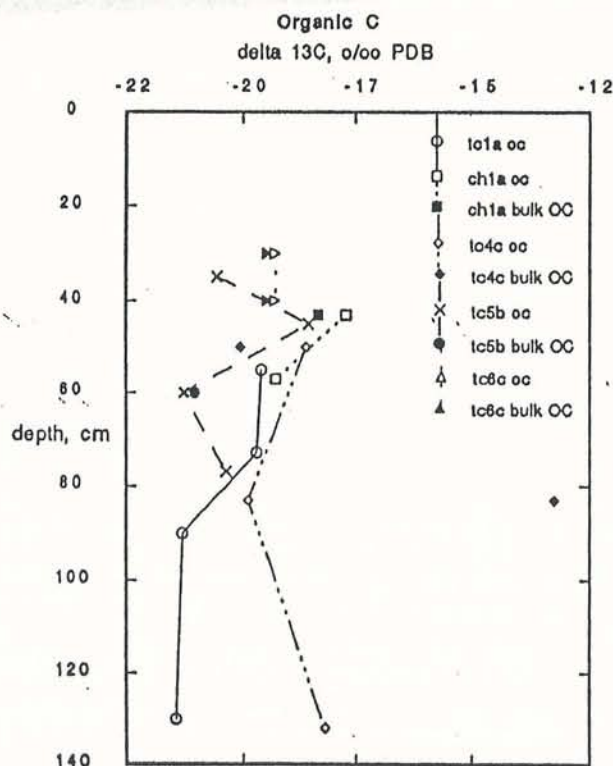
Figure 2. C isotopic composition of organic C "occluded" in inner rind carbonate and of bulk soil organic matter in soil profiles at Trail Canyon, Fish Lake Valley. Radiocarbon analyses of three samples are shown in Table 1.
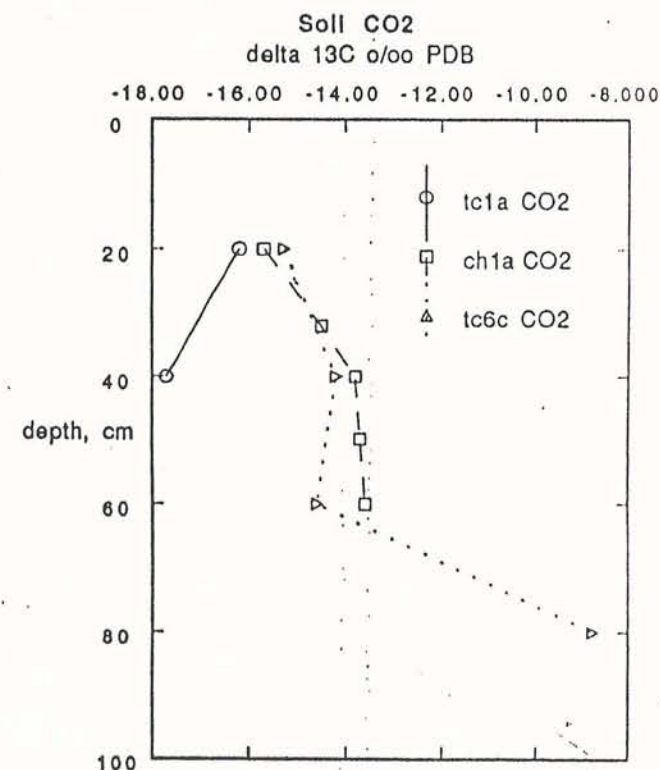
Figure 3. C isotopic composition of soil CO_2 collected from three soils in May, 1990. Atmospheric CO_2 has $\delta^{13}C$ of about -7.0 o/oo.

Carbonate and Precipitation

delta 18O, o/oo SMOW

Figure 4. O isotopic composition of inner rind carbonate, adjusted to show the isotopic composition of precipitation by adding 28.5 o/oo; and O isotopic composition of average precipitation from various storm tracks measured in southern Nevada (Benson and Klieforth, 1989). A = arctic, cP = continental polar, mP = maritime polar, mPk = maritime polar modified, mT = maritime tropical, cT = continental tropical. Precipitation values shown at 10 cm for comparative purposes only; no significance is given to soil profile depth.

O Isotopic Composition vs. Elevation
Central Great Basin, Nevada
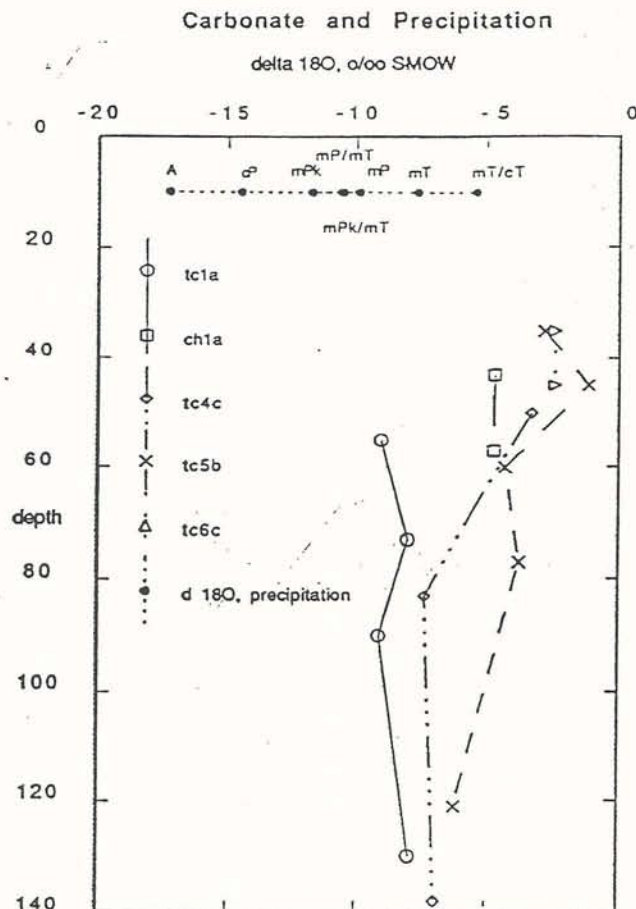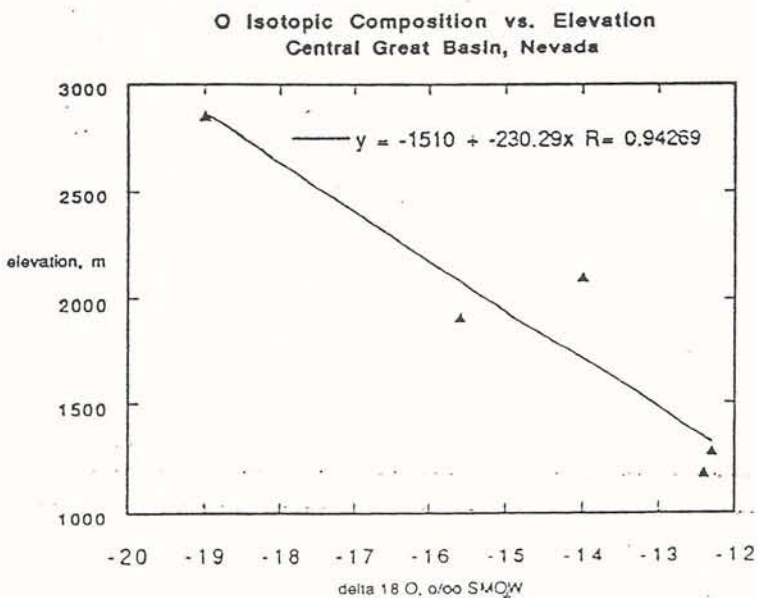
$y = -1510 + -230.29x$ R= 0.94269

Figure 5. Average O isotopic composition of precipitation and stream water measured at sites 100 to 200 km north and east of Fish Lake Valley, Nevada, at various elevations. Precipitation data courtesy of Pat McKinley; stream data courtesy of Carol Kendall.

17

CENTER FOR ACCELERATOR MASS SPECTROMETRY
Lawrence Livermore National Lab

USGS, Pendall Report Feb. 20, 1991

CAMS #	Identification			delta ^{13}C	D ^{14}C	+-	Age yrs. B.P.	+-
852	TC 1A	120-140	rinds-carbonate	-10	-758.3	5.0	11410	170
853	TC 1A	120-140	rind-organic	-25	-362.4	8.2	3620	110
854	TC 4C	37-63	rinds-carbonate	-10	-240.6	10.0	2210	110
855	TC 4C	37-63	rinds-organic	-25	-408.7	9.1	4220	130
856	TC 5B	40-50	rinds-carbonate	-10	-285.7	15.7	2700	180
857	TC 5B		40-50 organic	-25	-185.0	12.3	1640	130
858	TC 6.5A		65 cm charcoal	-25	-12.6	15.3	<150	
859	FLVTC-JH1		charcoal	-25	-381.9	7.6	3860	100

Notes:

1) Delta ^{13}C values are the assumed values according to Stuiver and Polach (Radiocarbon, v. 19, p.355, 1977) when given without decimal places. Values measured for the material itself are given with a single decimal place.

2) The quoted age is in radiocarbon years using the Libby half life of 5568 years and following the conventions of Stuiver and Polach (ibid.).