Carbon isotope ratios of soil organic matter and their use in assessing community composition changes in Curlew Valley, Utah

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Summary. Stable carbon isotope ratios of roots and soil organic matter were measured in Curlew Valley, Utah to determine if changes in the relative dominance of two shrub species had occurred in this salt-desert community. Measurements were made on soil cores along transects stretching from monospecific stands of Ceratoides lanata, a C3 shrub, to monospecific stands of Atriplex confertifolia, a C4 shrub. δ13C values of roots and soil organic matter under Ceratoides appeared to be in equilibrium with the current plant community. By contrast, δ13C values of roots and soils under Atriplex portions of the transects were more negative than would be expected for a C4-dominated community. These results indicate that a change in relative C3/C4 dominance has occurred, and suggest that the C4 shrub Atriplex confertifolia is increasing in importance in this salt-desert community.

The salt-desert shrub ecosystem type occupies 17 million hectares in the western United States (West 1983), principally within the physiographic confines of the Great Basin. This ecosystem is dominated by shrubs, with Atriplex confertifolia (Torr. et Frem.) S. Wats. (shadscale) being a principal dominant. Ceratoides lanata (Pursh) J.T. Howell (winterfat) also occurs thoughout the Intermountain West and achieves local dominance in many areas. Vegetational characteristics of the salt-desert shrub ecosystem are described in detail by Cronquist et al. (1972).

Atriplex confertifolia and Ceratoides lanata are members of the Chenopodiaceae family which contains both C3 and C4 species. Ceratoides lanata has the C3 pathway of photosynthesis, whereas Atriplex confertifolia is a C4 species (Welkie and Caldwell 1970). Comprehensive studies on the ecophysiology of these two shrubs have revealed few significant differences in productivity, water use efficiency, and utilization of the annual moisture resource (Caldwell et al. 1977). These studies demonstrated that the amount of carbon fixed per unit of ground area was approximately the same in both winterfat and shadscale communities. However, long term studies indicate that communities dominated by shadscale and winterfat are not compositionally stable (Stewart et al. 1940; Hutchings and Stewart 1953; Holmgren and Hutchings 1972; Norton 1978). While the instability observed in many of these studies was caused by grazing, changes in the relative abundances of shadscale and winterfat have been observed in ungrazed areas as well (Holmgren and Hutchings 1972).

The natural difference in the stable carbon isotope composition of C3 and C4 plants provides an opportunity for assessing the “long-term” compositional stability of the salt-desert shrub ecosystem. During photosynthesis, C4 plants discriminate less against 13CO2 than C3 plants (Vogel 1980; O'Leary 1981). The result of this fractionation of stable carbon isotopes during photosynthesis is a characteristic carbon isotope ratio (expressed as δ13C) in the plant tissue which serves as an effective marker for the occurrence of C3 and C4 photosynthesis. The δ13C values of C3 plants are between -23 and -34‰, with an average of -26‰, whereas C4 plants range from -9 to -17‰, and average -12‰ (Smith and Epstein 1971). Thus, the average difference in δ13C between C3 and C4 plants is approximately 14‰.

This difference in stable carbon isotope ratios between photosynthetic types can be used to quantify the proportion of C3 and C4 species contributing to a mixture of plant material (Ludlow et al. 1976; Ode et al. 1980). The δ13C value of soil organic matter should also reflect the relative importance of C3 and C4 plants in the community. Theoretically, the δ13C value of the soil organic matter should be identical to the existing vegetation at a site if: 1) the existing vegetation has remained unchanged for a period of time equal to that of the oldest carbon in the soil profile, 2) the δ13C value of atmospheric CO2 has remained constant through time, 3) no fractionation of carbon isotopes has taken place in the soil as a result of either decomposition processes or differential preservation of plant biochemicals. If assumption 2 and 3 hold true and there is a difference in the δ13C value between the soil and the vegetation, then in all likelihood there has been some degree of compositional change in the C3/C4 biomass of the community.

The assumption that the δ13C value of atmospheric CO2 has remained constant over time is difficult to verify directly. However, some circumstantial evidence is available. For example, carbon isotope ratios of coal derived from terrestrial plants are similar to those of modern day C3 plants.
(Galimov 1976; Deines 1980). C$_3$ and C$_4$ grass cuticles from the early Pliocene (~15 mya) were found to have $\delta^{13}$C values of $-24.6\%_o$ and $-13.7\%_o$, respectively (Nambudiri et al. 1978). Similarly, the $\delta^{13}$C value of a specimen of *Atriplex confertifolia* more than 14,000 years old (~13.4\%) is close to present day values (~14.8\%) (Troughton et al. 1974b). Since plant $\delta^{13}$C values depend on the $\delta^{13}$C values of the source CO$_2$ available for fixation during photosynthesis (Smith et al. 1976), these data suggest that the $\delta^{13}$C value of atmospheric CO$_2$ has not changed by more than 1 or 2\%o since the Mesozoic or earlier.

Fractionation of carbon isotopes during decomposition of plant material in the soil has received little attention thus far. $\delta^{13}$C values of organic matter from mineral soils tend to become enriched in carbon-13 by 1–3\%o with age and increasing depth in the soil profile, suggesting a slight fractionation during decomposition (Troughton et al. 1974a; Stout et al. 1975; Goh et al. 1976, 1977; O'Brien and Stout 1978; Stout and Rafter 1978; Schleser and Pohl- ing 1980; Schleser and Bertram 1981; O'Brien et al. 1981; Stout et al. 1981). This fractionation is probably a result of microbial metabolism. It has been repeatedly demonstrated that microorganisms utilize $\delta^{12}$C in preference to $\delta^{13}$C, leading to slight $\delta^{13}$C enrichment of the residual organic substrate (Rosenfeld and Silverman 1959; Kaplan and Rittenberg 1964). In addition, inorganic decomposition of marine sediments has also been shown to cause $\delta^{13}$C enrichment in organic matter (Sackett and Thompson 1963). Thus, metabolic and inorganic processes occurring in the soil may be responsible for the small fractionations noted in soil organic matter.

By contrast, there appears to be little fractionation in soils where decomposition is restricted, such as peats and New Zealand *Agathis* forest soils (Stout et al. 1975; Stout et al. 1981). Based on available evidence, there seems to be a pattern of little or no isotope fractionation in soils where decomposition is restricted and undecomposed plant material accumulates, and small but significant $\delta^{13}$C enrichment of organic matter in mineral soils where decomposition is relatively rapid (Troughton et al. 1974a; Stout and Rafter 1978). However, the magnitude of this soil related fractionation is not large enough to mask gross changes in C$_3$ and C$_4$ vegetation types, provided the change in vegetation has not occurred so long ago to have been obscured by the build-up of organic material from the current community.

Differential decomposition of plant biochemical fractions in the soil could alter the isotopic composition of the parent material, since different biochemical fractions such as starch, cellulose, lipids, etc. can have different $\delta^{13}$C values (Park and Epstein 1960; Smith and Benedict 1974). However, these studies show that most biochemical fractions differ from whole plant values by only 1–3\%o in isotopic composition. The major exception is plant lipid which is consistently 5–10\%o more depleted in $\delta^{13}$C than whole-plant material (Park and Epstein 1960; Smith and Epstein 1971; Smith and Benedict 1974). Differential preservation of plant lipids is unlikely in the majority of normal, well-drained soils, since most lipids decompose within one year under aerobic conditions (Waksman and Stevens 1929; Lukoshko 1965).

In summary, available evidence suggests that the carbon isotopic composition of atmospheric CO$_2$ has not changed appreciably in recent time, and that carbon isotope fractionation occurring during decomposition in the soil is small. Therefore, given the time limits imposed by the turnover rates of soil organic carbon, it should be possible to determine whether or not a change in the relative proportions of C$_3$ and C$_4$ plants has occurred by measuring the carbon isotope composition of the current plant community and soil organic matter. The turnover time of organic carbon in the soil of interest will ultimately determine the length of time within which a change in community composition can be detected. While radiocarbon ages of modern soil organic matter are frequently on the order of several thousand years, studies on soil respiration rates and decomposition of labelled plant material indicate mean ages of a few hundred years for soil organic matter (O'Brien et al. 1981). Since it is impossible to reconcile these differences at the present time, it is not clear how long a change in the C$_3$—C$_4$ composition of a plant community will remain detectable in the isotopic composition of soil organic carbon.

Few studies have attempted to use stable carbon isotope ratios of soil organic matter to assess plant community changes. Hendy et al. (1972) studied the $\delta^{13}$C and $\delta^{15}$N content of organic matter in an Australian soil and found that the $\delta^{13}$C values increased significantly between 14,000 and 9,000 years ago. They interpreted these data as indicating a change in vegetation from either a temperate forest (C$_3$) or temperate grassland (C$_3$) to a subtropical grassland (C$_4$). This change to a C$_4$ dominated ecosystem was found to correspond with a worldwide warming trend 11,000 years ago. Krishnamurthy et al. (1982) measured the isotopic composition of organic matter in paleosols from Kashmir, India and found isotopic evidence for a change from C$_4$ vegetation existing under arid conditions ~31,000 years ago to C$_3$ dominated vegetation existing under relatively more humid conditions ~19,000 years ago.

$\delta^{13}$C values of soil organic matter from preliminary cores taken in Curlew Valley, Utah revealed that significant differences existed between the isotopic composition of soil organic matter and the current community vegetation, suggesting that either *Ceratoides* or *Atriplex* was increasing in relative importance. Since previous ecological studies have indicated that this plant community type may be in a state of change (Stewart et al. 1940; Hutchings and Stewart 1953; Holmgren and Hutchings 1972; Norton 1978), the present study was initiated in an attempt to assess the long-term stability of the *Atriplex-Ceratoides* salt desert shrub community through the use of stable carbon isotope ratios of soil organic matter.

**Methods**

**Study area**

The study site is located in Curlew Valley, Utah (41°52' N, 113°5 W) at an elevation of 1,350 m. This broad, lacustrine valley extends northward from the northernmost extent of the Great Salt Lake and was largely inundated by the waters of Lake Bonneville during the last glacial period. Valley soils are generally homogeneous and vary in texture from silt loam to sandy loam. Soil profiles are of the Thiemol Series, a fine silty mixed mesic family of Xer- olic Calcisols (Skujins and West 1973). The soils are highly calcareous in nature with carbonates comprising 10–25% of the soil by weight. Chemical and physical prop-