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Isotopic analysis of respired CO_2 during decomposition of separated soil organic matter pools

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Abstract

A detailed understanding of the processes that contribute to the $\delta^{13}C$ value of respired CO_2 is necessary to make links between the isotopic signature of CO_2 efflux from the soil surface and various sources within the soil profile. We used density fractionation to divide soils from two forested sites that are a part of an ongoing detrital manipulation experiment (the Detrital Input and Removal Treatments, or DIRT project) into two soil organic matter pools, each of which contributes differently to total soil CO_2 efflux. In both sites, distinct biological pools resulted from density fractionation; however, our results do not always support the concept that the light fraction is readily decomposable whereas the heavy fraction is recalcitrant. In a laboratory incubation following density fractionation we found that cumulative respiration over the course of the incubation period was greater from the light fraction than from the heavy fraction for the deciduous site, while the opposite was true for the coniferous site.

Use of stable isotopes yielded insight as to the nature of the density fractions, with the heavy fraction solids from both forests isotopically enriched relative to those of the light fraction. The isotopic signature of respired CO_2, however, was more complicated. During incubation of the fractions there was an initial isotopic depletion of the respired CO_2 compared to the substrate for both soil fractions from both forests. Over time for both fractions of both soils the respired $\delta^{13}C$ reflected more closely the initial substrate value; however, the transition from depleted to enriched respiration relative to substrate occurs at a different stage of decomposition depending on site and substrate recalcitrance. We found a relationship between cumulative respiration during the incubation period and the duration of the transition from isotopically depleted to enriched respiration in the coniferous site but not the deciduous site. Our results suggest that a shift in microbial community or to dead microbial biomass as a substrate could be responsible for the transition in the isotopic signature of respired CO_2 during decomposition. It is likely that a combination of organic matter quality and isotopic discrimination by microbes, in addition to differences in microbial community composition, contribute to the isotopic signature of different organic matter fractions. It is apparent that respired $\delta^{13}CO_2$ cannot be assumed to be a direct representation of the substrate $\delta^{13}C$. Detailed knowledge of the soil characteristics at a particular site is necessary to interpret relationships between the isotopic values of a substrate and respired CO_2.

Keywords: Alfisol; Andisol; Coniferous forest; Deciduous forest; $\delta^{13}C$; Density separation; Isotopic fractionation; Organic matter; Soil respiration

1. Introduction

Soil organic matter (SOM) is a complex mixture of material from various sources that exists along a continuum of decomposition and stabilization in the soil profile. For simplicity and for modeling purposes, soil organic matter often is divided into several pools with different turnover times and recalcitrance. Each pool

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contributes to total efflux in different proportions depending on availability as a substrate for microbial decomposition (Parton et al., 1987; Coleman and Jenkinson, 1996; Trumbore, 1997). Many approaches have been taken to physically or chemically separate these pools in the laboratory (e.g., Strickland and Sollins, 1987; Six et al., 2000, 2001; Swanston et al., 2004). Analyses of the carbon isotopic composition of SOM pools have yielded insight into turnover rates and microbial processing (e.g., Balesdent and Mariotti, 1987; Buchmann et al., 1998; Six et al., 2001) and some progress has been made to use $\delta^{13}C$ values of respired CO_2 to identify source pools for CO_2 efflux from the soil surface (Ehleringer et al., 2000). Often, these studies are in ecosystems where a shift between C_3 and C_4 vegetation has significantly altered the isotopic signatures of the C inputs to SOM pools. Consequent shifts of C isotopic inputs and accompanying changes in the isotopic signatures of SOM pools in these systems can help identify the sources of respiratory CO_2 (e.g., Rochette and Flanagan, 1997). There is greater difficulty determining the source of respired CO_2, however, when only small variations in isotopic signatures of inputs exist, or when there is little difference in the isotopic composition between inputs and SOM pools. For these systems in particular, we need a more precise understanding of processes that control the C isotopic signature during respiration to make links between the isotopic composition of respired CO_2 and its source in the soil.

The differences between the $\delta^{13}C$ value of vegetation biomass, SOM, and respired CO_2 have already been used to gain insight into biological processes that mediate C transfers among ecosystem pools (e.g., Nadelhoffer and Fry, 1988; Šantrůčková et al., 2000b; Niklaus et al., 2001). Plant litter generally has lower $\delta^{13}C$ values than bulk soil and serves as continuous inputs into SOM in the form of both above and below ground sources (Accoe et al., 2003; Bird et al., 2003). Individual molecular components of these inputs have highly variable isotopic signatures; for example, lignin is depleted in ^{13}C content by 2–6‰ compared to the bulk plant material and by 4–7‰ relative to cellulose (Benner et al., 1987), and wood cellulose is 2‰ enriched compared to leaf cellulose (Gleixner et al., 1993). Invertebrates have been shown to excrete frass with lower $\delta^{13}C$ values compared to food (Šantrůčková et al., 2000a). Recent studies suggest that soil microorganisms may alter the isotopic composition of SOM during decomposition through mechanisms such as metabolic discrimination (Schmidt and Gleixner, 1998, Šantrůčková et al., 2000b), selective consumption of substrates (Macko and Estep, 1984), or preferential use of intramolecular position within substrates (Schweizer et al., 1999; Hobbie and Werner, 2004). In general, processes that control the isotopic signature of CO_2 during decomposition and efflux from the soil back to the atmosphere are not well understood; indeed, whether isotopic fractionation during decomposition even occurs is currently under debate (Lin and Ehleringer, 1997; Henn and Chapela, 2000; Šantrůčková et al., 2000b; Fernandez et al., 2003; Klumpp et al., 2005).

As decomposition of fresh plant litter progresses and the decomposition products become incorporated into the soil profile, $\delta^{13}C$ content has been observed to increase (Buchmann et al., 1998). Multiple theories have been proposed to explain the observed trend of $\delta^{13}C$ with depth (Ehleringer et al., 2000), including changes in the atmospheric $\delta^{13}CO_2$ value since the Industrial Revolution, preferential feeding by microbes on isotopically light material, and metabolic fractionation during decomposition, among others. Boutton (1991) suggested that deeper SOM is older, and thus presumably more resistant to further decomposition than is surficial SOM. However, it is not clear that the observed pattern of increasing $\delta^{13}C$ value with depth necessarily means that more labile SOM is less ^{13}C enriched. In addition to incomplete understanding of processes that contribute to isotopic fractionation, we also know little about differences in the degree of fractionation during decomposition from SOM pools of different ecosystems (Ehleringer et al., 2000).

Numerous authors have used density fractions of SOM to represent different pools of SOM that might turnover at different rates (cf., Strickland and Sollins, 1987; Trumbore, 1997; Six et al., 2001; Baisden et al., 2002; Swanston et al., 2002). Light fraction material (LF, $< 1.6\,g\,cm^{-3}$) is composed of partially decomposed litter debris, charcoal, and humus. Heavy fraction material (HF, $> 1.6\,g\,cm^{-3}$) consists of mineral clays and organic material in close chemical association with mineral surfaces. Heavy fraction material typically has a lower C:N than light fraction material and is thought to contain soil organic C that is more processed and stabilized (i.e., resistant to further decay). Heavy fraction SOM is generally found to be ^{13}C-enriched compared to the light fraction (Ehleringer et al., 2000; Six et al., 2001; Fernandez et al., 2003). The purpose of this study was two-fold: (1) to follow the dynamics of respired $\delta^{13}CO_2$ during SOM decomposition and (2) to determine whether, in a soil incubation system where root contributions to soil respiration are removed from heterotrophic decomposition of SOM pools, respired $\delta^{13}CO_2$ is a reflection of substrate $\delta^{13}C$ in two very different forests. We expected greater cumulative respiration to occur during incubation of light fraction material than from heavy fraction material and that the isotopic signature of CO_2 respired during decomposition of the light and heavy fractions would be distinct from each other and reflective of the isotopic signature of the source material of different recalcitrance for both forest soils.

2. Materials and methods

Soil from the 0–5 cm layer mineral A horizon was collected from two long-term experimental field sites at the H. J. Andrews Experimental Forest in the Cascade range of western Oregon in June 2002 (coniferous site) and at the Allegheny College Bousson Experimental Forest in western