



# Decomposition of distinct organic matter pools is regulated by moisture status in structured wetland soils



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## ABSTRACT

Peatlands are garnering much attention for their greenhouse gas feedback potential in a warming climate. As of yet, the coupled biogeochemical and hydrological processes that control the amount and timing of soil organic matter (SOM) mineralization and, ultimately, whether peatlands will be sinks or sources of atmospheric CO<sub>2</sub> are not fully understood. Soil structure is a key feature of soils that mediates the coupling between biogeochemical and hydrological processes. However, we know very little about how soil structure responds when soils are exposed to wetting–drying cycles outside their normal range. In order to better understand how high elevation peatlands will respond to increasingly dry years, we incubated soils from high elevation meadows in the Sierra Nevada at 5 different water potentials and measured the CO<sub>2</sub> flux for over one year. We found that the cumulative carbon mineralization had a U-shaped pattern, with the greatest mineralization at the wettest (–0.1 bar) and driest (–4 bar) water potentials, across all hydrologic regions of the meadow. We propose a conceptual model that reproduces a similar pattern by incorporating the concept of dual porosity medium, with two distinct pore-size populations representing inter- and intra-aggregate porosity. Availability of water and oxygen to the two pore-size populations depends on the soil's equilibrium water potential. The model and the data suggest that the decomposition rates of intra-aggregate SOM may increase due to prolonged drought events that lead to accelerated release of C from previously untapped pool.

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## 1. Introduction

The accumulation of carbon in wetland soils, including most peatlands and high-elevation meadows in the Sierra Nevada and other mountain ranges, is primarily driven by low rates of decomposition as the soils stay near saturation for the bulk of the growing season (Clymo, 1965; Malmer, 1986). The low rate of decomposition coupled to a high degree of above and belowground productivity results in an ecosystem sink of carbon dioxide from the atmosphere (Arnold et al., 2014).

The key soil environmental conditions that regulate the rate of soil OM decomposition include (a) soil water content and potential, (b) air-filled porosity, and (c) temperature (Linn and Doran, 1984; Baldock and Skjemstad, 2000; Schmidt et al., 2011). Soil water content indirectly regulates decomposition (Berhe, 2012; Berhe

et al., 2012) by controlling the diffusion of microbial substrate on the dry end as well as by limiting oxygen (through air-filled porosity) to microorganisms on the wet end (Kaiser et al., 2015; Orchard and Cook, 1983). The degree of cell hydration and fluid exchange between microbial cells and soil is governed by water potential gradients (Stark and Firestone, 1995). Moreover, there is also an interdependency between temperature and water effects on OM decomposition, where maximum rates of soil respiration are dependent upon temperature (Wildung et al., 1975).

Each of the above parameters that determine the physical environment of the soil are inherently functions of the architecture of the soil matrix (Kay, 1998; Baldock and Skjemstad, 2000; Rawls et al., 2003). This refers to the distribution and arrangement of soil pores in addition to soil particles and aggregates. In many soils, the pore-sizes exhibit unimodal probabilistic distribution with a distinct modal pore-size. While others, including macroporous and aggregated soils as well as most organic soils, exhibit bimodal or multimodal distributions (populations) of soil pore sizes. While the mechanisms that form multimodality are not fully understood, it has been observed that this property is prevalent in soils with high

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SOM and high degree of aggregation (Ghezzehei, 2012) and gets accentuated by wetting and drying cycles (Horn and Smucker, 2005). The large and inter-connected macropores (inter-aggregate pores) in these soils belong to a different pore-size population from the finer intra-aggregate pores. The change in soil structure due to compaction and shrinkage (Arnold and Ghezzehei, in revision) adds another layer of complication to the pore size distributions of multimodal soils. However, there is no general prediction that could be made for how compaction impacts pore size distribution across soil taxa as the different pore-size populations respond differently to the deforming stresses (Kutílek et al., 2006).

In multimodal soils, water and air content tend to vary spatially even under equilibrium water potential conditions. Thus, different portions of a given soil could experience markedly different rates of mineralization. Although from the surface the relative contributions of the different components are not discernible. As a result, the response of such soils to changing hydrologic and/or temperature regimes may appear contradictory and difficult to interpret. For example, Baldock and Skjemstad (2000) found that soils with a high degree of larger pore sizes exhibited increased rates of carbon mineralization as compared to soils with a smaller range of pores (at equivalent values of air-filled porosity). In another study, Torbert and Wood, (1992) found that the degree of water-filled pore space in addition to the size and structure of pore space heavily influence microbial activity in the soil. Studies in the laboratory have also shown that a lowering of the water table can stimulate carbon dioxide flux from peat microcosms (Moore and Dalva, 1993; Funk et al., 1994; Blodau et al., 2004). However field studies have shown mixed results with some reporting no significant change in soil respiration (Freeman et al., 1996), and yet others reporting importance of microtopographical controls (Strack and Waddington, 2007), and even significant carbon losses (Oechel et al., 1998). In one study, field based measurements of ecosystem respiration showed little to no response to drying whereas laboratory incubated peat cores showed a decrease in respiration with drying at the surface (Lafleur et al., 2005). Drying in an Alaskan fen induced plant stress, which turned out to be more responsible for reducing the carbon sink in their ecosystem through a reduction of gross primary productivity rather than an increase in ecosystem respiration (Chivers et al., 2009). In one study, loss of porosity was correlated with reduction of OM mineralization (Franzluebbers, 1999).

These disagreements clearly warrant an explanation based on a systematic study that evaluates how different moisture regimes influence decomposition in soils with multimodal pore-size distributions. The objective of this study is to provide a mechanistic understanding of how SOM decomposition varies across a wide range of soil water content in highly organic soils with bi-modal pore-size distribution. We conducted a long-term soil incubation experiment utilizing soil samples from high elevation meadows in the Sierra Nevada mountain range. Gas fluxes from samples that were kept at five different water potentials (0.1–4.5 bar) were collected and analyzed for over one year. In order to aid in synthesizing the various effects across the range, we introduce a simple modeling framework that incorporates water and air content distributions in a bimodal soil (Section 2). Projections of this model adequately explain the trends observed in our experiments as well as reported elsewhere.

## 2. Theory

### 2.1. Environmental controls on mineralization

Here we develop a simplified conceptual/mathematical model that integrates the interactions and feedbacks between soil

structure, structural dynamics, and hydrologic conditions in controlling soil organic matter decomposition. The model does not attempt to resolve temporal dynamics of organic matter mineralization or spatial distribution of concentrations and rates of decomposition. Rather, the overall goal is to mathematically represent how complexity of soil structure, through its influence on physical environmental conditions, controls the relative degree of mineralization over a wide range of moisture regimes.

Consider a macroscopically homogeneous soil volume that is at constant and homogeneous water potential. The instantaneous rate of mineralization of soil carbon in this volume can be described using a “one-pool” model (Jenny, 1980)

$$\frac{dC}{dt} = kC \quad (1)$$

where  $C$  (C-mass/soil-mass) is the quantity of mineralizable  $C$  and  $k$  (1/time) is the rate constant. Assuming that the soil remains under fairly constant environmental conditions (i.e., the rate constant and the pool of mineralizable carbon remain unchanged), Eq. (1) can be solved to provide an exponential decay of the soil  $C$  content

$$C = C_0 e^{-kt} \quad (2)$$

where  $C_0$  is the initial mineralizable  $C$  pool. Alternatively, the quantity of the mineralized  $C$  over the course of a given period of mineralization under stable conditions can be described as

$$C_t = C_0 (1 - e^{-kt}) \quad (3)$$

Note that  $C_0$  represents only the portion of the total SOM that is accessible for mineralization under the equilibrium physical conditions.

We postulate that the initial mineralizable pool ( $C_0$ ) and rate constant ( $k$ ) of any given soil that is incubated under equilibrium conditions are dependent on physical conditions. This assumption is justified by the numerous studies that documented the dependence of fitted  $C_0$  and  $k$  on physical conditions such as temperature and water-filled porosity (Linn and Doran, 1984; Kechavarzi et al., 2010a). Under isothermal conditions (which covers most laboratory experiment) the total mineralized  $C$  over a period of  $t = T$  can be related to water-filled porosity and air-filled porosity as

$$C_t(T) = C_0 (1 - e^{-kT}) = S f_W(\theta) f_A(a) \quad (4)$$

where  $\theta$  is volumetric water content,  $a = \phi - \theta$  is volumetric air content, and  $\phi$  is porosity. The functions  $f_W$  and  $f_A$  describe the role of water content and air contents as necessary variables that control access to nutrients and oxygen. Specifically, the second term of Eq. (4) implies that mineralization is dependent on sufficient substrate diffusion (which is positively correlated with water content) as well as efficient gas-exchange (which increases with air-filled porosity). The proportionality constant  $S$  is a factor that accounts for total SOM, SOM chemistry, and other soil biogeochemical characteristics that are not directly dependent on soil water content status. The factor  $S$ , varies depending on the quality and quantity of the SOM. Without quantitative understanding of the value of  $S$ , we cannot use the model for quantitative prediction. Nevertheless, it gives mechanistic explanation of how water-content and air-content influence mineralization.

For the purpose of illustrating these concepts, we define the functions  $f_W$  and  $f_A$  as linearly dependent on water-filled and air-filled fractions of the total porosity, respectively, as

$$f_W(\theta) = \frac{\theta}{\phi} \quad \text{and} \quad f_A(a) = \frac{a}{\phi} \quad (5)$$

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