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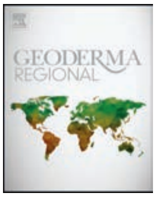

Kinetic parameters of soil organic matter decomposition in soils under forest in Hungary

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ABSTRACT

To obtain better knowledge on how soil organic matter (SOM) decomposition is regulated by different soil properties, 13 soil samples amended with maize residues were incubated for 163 days. Carbon mineralization kinetics was modelled by fitting five different kinetic models, from which the first-order two pools model was selected for further analysis. The results showed that residue addition increased the rate of C mineralization, particularly in sand samples, where four times as much C was released as CO_2 in amended samples as in control samples. Residue addition also affected the kinetic parameters of the two C pools modelled. The mean residence time (MRT) averaged 31.9 days for the fast C pool and 65.1 days for the slow C pool phase in the amended soils, whereas in the control soils MRT_1 and MRT_2 averaged 52.3 and 94.2 days, respectively. The kinetics of C decomposition was found to be influenced to the greatest extent by soil organic carbon (SOC) content and soil texture. The texture primarily affected the size of the fast carbon pool (C_1): in both the amended and control samples C_1 was increased by the clay content ($r = 0.574$ and $r = 0.554$, respectively) and decreased by the sand content ($r = -0.771$ and $r = -0.583$, respectively). Our results confirm the fact that texture has a significant role in soil organic matter mineralization, however other parameters also play a crucial role in the decomposition due to the complexity of the processes.

1. Introduction

The significant role of SOM in the global carbon cycle is well known. The global soil carbon pool (2500 Gt) is 3.3 times the size of the atmospheric pool and 4.5 times the size of the biotic pool (Lal, 2004). Promoting the amount of SOM sequestered in terrestrial ecosystems is important as it may help slow the rise in atmospheric CO_2 in coming decades (Janzen, 2006). Therefore, a better understanding of the properties and dynamics of SOM and the identification of the factors that regulate soil respiration in natural and managed ecosystems is critical in predicting ecosystem responses to global change (Ahn et al., 2009).

Soil organic matter consists of a variety of materials, ranging in age from a few hours for root exudates through weeks and months for plant residues to thousands of years for resistant organic compounds (Kuzyakov, 2006; Schmidt et al., 2011; Trumbore, 2000). Models that aim to describe the cycles of these compounds cannot characterize every carbon pool with different turnovers, but differentiate one, two or three pools of SOM. The mathematical description of nutrient release patterns and the fitting of kinetic equations to mineralization curves make it possible to characterize SOM pools and calculate the fraction of potentially mineralizable C and its mineralization rate (Bernal et al., 1998; Saviozzi et al., 1993).

There are many potential approaches for modelling the C and N decomposition kinetics of soils. The original approximation was the first-order logarithmic form used by Stanford and Smith (1972). According to this exponential model, there is only one pool of potentially mineralizable C, decomposing at a rate proportional to its initial concentration. Since then many other models have been postulated. Some authors modified the simple first-order model and used approaches that take into account several organic pools with different decomposition rates. For example, Murayama (1984) described a first-order model with two compartments (two exponential equations), one for SOM with fast turnover and the other for SOM which decomposes more slowly. Others (Jones, 1984; Beauchamp et al., 1986) used an additional parameter in the simple first-order model to differentiate the fast initial flush caused by easily decomposable substrates at the beginning of the incubation period. Although first-order approaches are more common, other models with zero-order (Addiscott, 1983; Tabatabai and Al-Khafaji, 1980) and mixed-order (first-order plus zero-order) equations

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(Bonde and Rosswall, 1987; Seyfried and Rao, 1988) have also been used to study SOM decomposition kinetics.

SOM decomposition depends on many biotic and abiotic factors and may be altered by climate change (IPCC, 2013). Among the abiotic factors, climatic conditions, soil moisture, chemical recalcitrance and the physical accessibility of C compounds are significant (Manzoni et al., 2012). Soil texture is also an important parameter influencing SOM decomposition through factors such as particle surface area and porosity, which affect water-holding capacity, cation exchange capacity and many other factors (Procter et al., 2015). Soils with higher silt and clay content generally sequester more C than sandier soils, because decomposition is slower in the finer textured soils where lower O_2 inhibits aerobic microbial processes and the chemical and physical stabilization of SOM by soil minerals is more effective (Barré et al., 2014; Hassink, 1997; Wattel-Koekkoek et al., 2003). Since the specific surface area of the fine particle-size fractions is several orders of magnitude higher than that of the sand fraction, and since minerals belonging to these groups have reactive surfaces, not just clay minerals but also Fe-oxides play an important role in SOM binding (Eusterhues et al., 2005; Gu et al., 1994; Kaiser and Guggenberger, 2003). In incubation experiments the major influencing factors are soil pH and redox conditions, clay content, the size and quality of soil C pools and microbial activity (Ahn et al., 2009; Delaune et al., 1981; Riffaldi, 1996). In addition to soil parameters, the organic materials added to soils also have a great influence on the amount and rate of organic matter decay (Kassam et al., 2017; Singh and Gupta, 1977).

Soil incubation is a direct approach to quantifying the mineralizable soil C (Ahn et al., 2009), so many experiments have used this method to model SOM decomposition by fitting kinetic equations to mineralization curves. However, most incubation experiments test one or two soil samples, so it is not clear how soil type, texture and other soil parameters influence SOM decomposition kinetics. Therefore, it was aimed to provide more complete insights into (1) the description of the dynamics and kinetics of C mineralization in 13 different soil samples with widely differing properties; (2) the effectiveness of some commonly used decay models for describing C mineralization; and (3) the quantification of the relationship between soil properties and mineralization parameters derived from the best model.

2. Materials and methods

2.1. Soil samples

Thirteen topsoil samples were collected from seven sites in Hungary (Fig. 1, Table 1). The samples were mainly forest soils with high clay, silt and organic matter content, but three forest soils developed on sand were also collected (Tables 1 and 2). Soil samples were taken from the upper 0—20 cm horizon with one exception, where the bedrock was near the surface (sample code BAT in Table 1). Undisturbed soil samples were taken in order to determine the water-holding capacity (WHC). The soils were air-dried, homogenized, passed through a 2-mm sieve and stored at room temperature. Soil texture, total and dissolved organic and nitrogen content, iron content and the cation exchange capacity were measured before the incubation experiment.

The soil texture was determined by the pipette method (Gee and Bauder, 1986). The total organic carbon (TOC) content was analyzed using an NDIR-chemiluminescent analyzer (Apollo 9000, Tekmar Dohrmann). The total N content was determined by the Kjeldahl method (Conklin, 2014). The water extractable organic carbon (WEOC) and the total soluble nitrogen (TSN) were measured after extraction with ultrapure water (> 18 MΩ cm) at a 1:10 soil:solution ratio for 2 h and filtration through a 0.45 μm membrane nylon filter (Millipore), in a TOC/TN analyzer (TOC-L, Shimadzu). Iron was determined in acid ammonium oxalate extracts (Schwertmann, 1973) and in dithionite–citrate–bicarbonate extracts (Holmgren, 1967) and measured using an atomic absorption spectrophotometer (210VGP, Buck Scientific). Cation exchange capacity (CEC) was determined according to the method of Gillman (1979). In this method 2 g of air-dried soil was saturated with 20 ml of 0.1 M $BaCl_2$ solution and equilibrated with 20 ml of 0.2 mM $BaCl_2$ solution. Then reacted with 10 ml of 5 mM $MgSO_4$ to replace Ba with Mg. The electrical conductivity of the resultant reactant suspension was adjusted to that of the 1.5 mM $MgSO_4$ ionic strength reference solution. The loss of Mg from the reactant $MgSO_4$ solution was determined by analysis of the equilibrium solution and is equivalent to that adsorbed and hence to CEC. The Mg concentration was measured by an atomic absorption spectrophotometer (210VGP, Buck Scientific). The use of unbuffered solutions throughout ensures that pH approaches the natural soil pH.

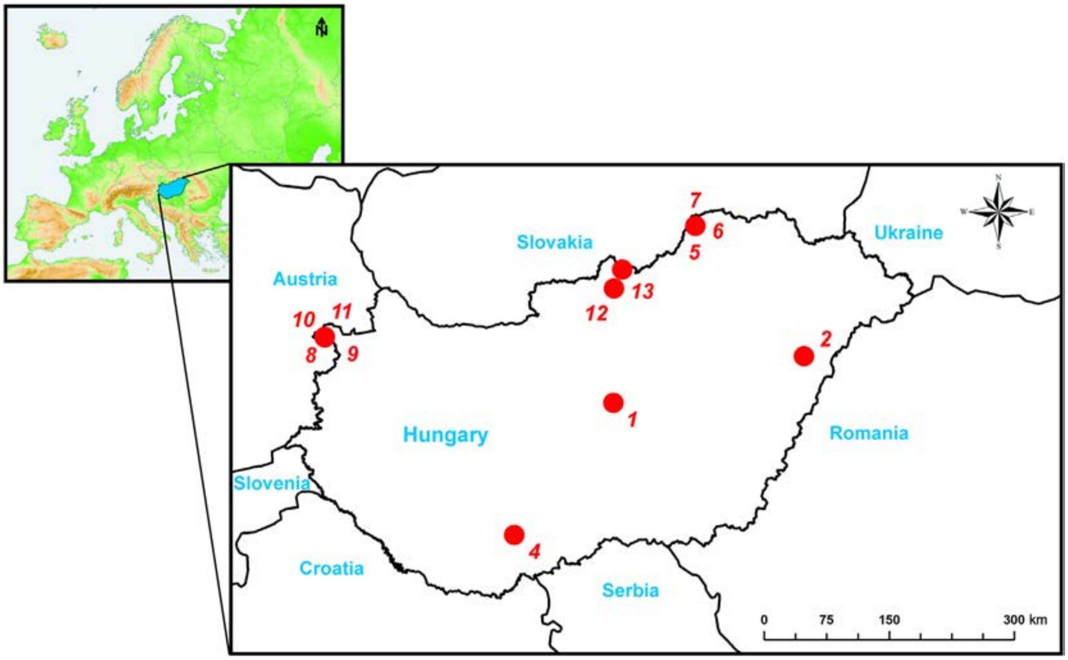

Fig. 1. Location of the sampling sites. Numbers indicate the numbers of the samples. See Tables 1 and 2 for details.