



Data synthesis of carbon distribution in particle size fractions of tropical soils: Implications for soil carbon storage potential in croplands



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ABSTRACT

Organic carbon saturation in soils refers to the theoretical maximum soil organic carbon (SOC) that can be associated with and stabilized on fine silt plus clay particles ($F < 20 \mu\text{m}$). We reviewed the literature dealing with SOC distribution between soil fractions to evaluate carbon saturation for tropical soils and estimate the C storage potential of cropland.

We collected 258 data points on SOC distribution between soil fractions in tropical soils from 84 sites in 27 countries. We used boundary line analysis to estimate the maximum stabilized SOC depending on soil group, clay type and land use. SOC storage potential was calculated as the SOC saturation deficit, the difference between the actual SOC content and the maximum stabilized SOC content.

We found that the maximum SOC in the fine fraction of tropical soils (53 g C kg^{-1} fine fraction) was lower than previous assessments of global SOC storage scale based mainly on temperate soils. The $F < 20 \mu\text{m}$ fractions were closer to SOC saturation in forest soils than in croplands. The cropland had a higher soil C storage potential, but changing agricultural management practices did not fill the deficit that is calculated using the whole dataset. The deficit was much lower when it was estimated with grassland or cropland data only: this provides a more realistic estimate for SOC storage potential for croplands.

The SOC content in the coarser fraction ($F > 50 \mu\text{m}$) did not depend on soil texture and significantly contributed to the total SOC, especially in sandy soils (41.3%). This is affected by changes in agricultural management practices. We concluded that, although the aim of increasing SOC stabilization originally arose from climate change mitigation strategies, it must now be more viewed as being more relevant to food security and local adaptation to climate change.

1. Introduction

Soil organic carbon (SOC) is the largest carbon sink in the terrestrial biosphere, amounting to about 1500 Pg for the top meter of soil (excluding Cryosols) (Jobbágy and Jackson, 2000). SOC plays a fundamental role in the fertility and productivity of terrestrial ecosystems, supporting important soil-derived ecosystem services such as soil quality, water filtration, erosion control, nutrient cycling, habitat and energy for soil organisms (Lal, 2016; Milne et al., 2015). Soil carbon storage has also been recognized as an efficient option to offset the rise in global atmospheric CO_2 concentration (Smith, 2016). Thus, storing carbon in soil is important for both food security and climate mitigation and adaptation.

Aboveground and belowground plant litter is the primary source of SOC. SOC accumulates in soils through the decomposition and

transformation of litter by soil organisms. The formation of organo-mineral complexes in the fine silt- and clay-size fractions ($F < 20 \mu\text{m}$) has been recognized as the most important process for SOC stabilization (Feller and Beare, 1997; Hassink, 1997; Six et al., 2002a), far more important than chemical recalcitrance of the organic matter (Dungait et al., 2012; Marschner et al., 2008). Hassink (1997) used least-squares linear regression of the relationship between SOC associated with the fine particle fraction, expressed in $\text{g C in } F < 20 \mu\text{m kg}^{-1}$ soil, and the relative mass of this fraction, expressed in g kg^{-1} soil, to estimate the capacity of the fine particles to stabilize carbon in uncultivated and grassland soils. The slope of this linear regression represented the theoretical capacity of the fine fraction of soil to accumulate and stabilize SOC. Further reviews studying the relationship between these variables highlighted the influence of the clay type (2:1 versus 1:1) and land use (i.e. forests, grasslands, croplands) on the slope of the

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regression (Feng et al., 2013; Six et al., 2002a). Feng et al. (2013) showed that the use of least-squares regression underestimated the maximum SOC that could be stabilized in the fine particle fraction and suggested the use of boundary line analysis to estimate the maximum stabilized SOC.

SOC saturation based on the maximum amount of SOC that can be stabilized on fine particles has been used in the literature to estimate the SOC storage potential, as the deficit between theoretical maximum stabilized SOC content for a given soil sample and the measured SOC content. Some studies (Angers et al., 2011; Wiesmeier et al., 2014) have estimated the SOC storage potential at national scale using the saturation limit as defined by Hassink (1997) or using fine particle content and other soil predictors like mineral surface area, exchangeable Al, Fe, and pH (Beare et al., 2014; McNally et al., 2017). These approaches address long-term, i.e. decadal, storage potential but do not take into account the short-term, i.e. annual, benefits of particulate organic matter. The size of this pool is highly variable and depends on land use and climate (Gregorich et al., 2006; Wang et al., 2016; Wiesmeier et al., 2014). Another way to estimate the SOC storage potential is to calculate the difference between measured SOC stocks or content in pristine conditions and measured SOC stocks in cultivated or human-impacted conditions in areas with the same soil and climate conditions (Akpa et al., 2016). This method also depends on the concept of SOC saturation as it assumes that soils in pristine conditions are higher in SOC than in anthropized areas due to higher C inputs, which is challenged by some studies dealing with pastures (e.g. Fujisaki et al., 2015; Stahl et al., 2016). A second assumption is the saturated state of soils in SOC under pristine vegetation. However, this latter assumption might overestimate the SOC storage potential, e.g. in the case of cultivated lands that remain cultivated. Stewart et al. (2007) introduced the concept of “effective stabilization capacity”, where the upper limit of SOC storage depends not only on physical and chemical properties of the soil but also on the levels of soil disturbance that vary depending on land use and within agroecosystems (e.g. tillage). Soils under native vegetation would, therefore, have a greater SOC stabilization capacity than cultivated soils.

SOC saturation and storage potential has not been investigated recently in tropical soils, despite its importance. The most recent meta-analysis of SOC distribution in the fine particle fraction was at global scale (Feng et al., 2013). However, tropical soils have specific chemical and mineralogical properties that could have a different effect on SOC saturation levels than those of temperate soils (Barthès et al., 2008; Feller and Beare, 1997). In recent years tropical soils have been subjected to land-use changes to increase production of food, timber, and fibers (Grace et al., 2014; Hansen et al., 2013). Large areas in the tropics have also been degraded and eroded (Gibbs and Salmon, 2015; Kiage, 2013), so we could expect low SOC stocks in these areas and thus a large potential for SOC storage.

To our knowledge the capacity of tropical cultivated soils to increase SOC content in the fine particle fraction to the SOC saturation level has not yet been evaluated. Is the SOC storage potential calculated using SOC saturation an achievable goal for increasing SOC to contribute to climate change mitigation and food security?

The general objective of this study was to review the literature to evaluate SOC storage potential in the tropics using the distribution of SOC in soil fractions. We aimed to i) define the level of SOC saturation in fine particle fractions depending on the soil type, clay type, and land use; ii) evaluate the potential for reducing the SOC saturation deficit by improving practices for annual crops; iii) assess the contribution of particulate organic matter to SOC content.

2. Materials and methods

2.1. Data collection

Data were collected by searching existing peer-reviewed literature

supplemented by searches for relevant grey literature. We searched literature published up to 2016 dealing with SOC distribution in particle-size fractions. The literature search was restricted to studies covering any area between the tropics or having a tropical climate according to the IPCC climate classification based on elevation, mean annual temperature and precipitation (IPCC, 2010, 2006). Google Scholar query «soil carbon “particle-size fractions” “20 μm” tropic*» produced 705 results. Only English language search terms were used but a few articles and PhD dissertations in French or Portuguese were also considered. We also checked papers that cited methodological papers dealing with this topic (Christensen, 1992; Gavinelli et al., 1995) and the papers cited in Feng et al. (2013) which is the most recent review dealing with this topic globally.

We selected studies where SOC content was measured in the fine ($F < 20 \mu\text{m}$) and/or coarse-size ($F > 50 \mu\text{m}$) fractions of the soil. SOC content associated to one of these fractions and expressed in g C in fraction kg^{-1} soil is calculated through Eq. (1).

$$\text{SOC}_F = F \times C_F \quad (1)$$

where SOC_F is the SOC associated to the considered fraction (g C in fraction kg^{-1} soil), F is the relative mass of the fraction (g fraction kg^{-1} soil), and C_F is the SOC content of the fraction (g C kg^{-1} fraction). Depending on the results presented in the collected studies, we could directly extract SOC_F values, or calculated them with Eq. (1). In addition some studies used separate analyses of the fractions $< 2 \mu\text{m}$ and $2\text{--}20 \mu\text{m}$. In this case, SOC_F of these fractions were summed to obtain the SOC associated to the $F < 20 \mu\text{m}$.

We included all studies using particle-size fractionation to separate organic matter depending on soil particle-size classes whatever the methods applied to disperse soil aggregates: these could use glass beads, ultrasonication, hexametaphosphate pre-treatment, Na resin, or combination of these techniques as described in Christensen (1992) and Gavinelli et al. (1995). In order to limit the dispersion in the dataset, we excluded studies using low energy dispersion fractionation methods aiming to generate aggregates. The recovery rate of soil fractionation was not always provided so we did not use a filter for this criterion. However, observations with a SOC recovery rate (sum of SOC across the fractions/bulk SOC content) higher than 1.2 were discarded. To produce the quantitative overview of the SOC content in the particle-size fractions, results from topsoil layers, i.e. up to 10 cm depth, were used. Some studies analyzed 0–5 and 5–10 cm layers separately, in this case we averaged the two values to obtain SOC content and particle-size mass for the 0–10 cm layer. The selection process gave 45 papers, 8 of which were included in the review by Feng et al. (2013), giving 258 observations in 84 sites and 27 countries in tropical areas. Most of the observations were located in South America and Sub-Saharan Africa. We found a few studies in Asia and Oceania (8 data points, see the map in Supplementary Fig. 1).

Soil type, clay type, and land use were collected from each paper (Supplementary Table 1). When not directly provided in the paper, we assigned soil to a WRB reference soil group (IUSS Working Group WRB, 2015) based on the soil classes in other classification systems, soil properties and/or any relevant information reported in each paper. Then, we further assembled these reference groups into four major groups according to the group sequences described in the Soil Atlas of Africa (Jones et al., 2013). These groups are based on the dominant factors or processes that most clearly control the formation of the soil. Group I included relatively homogeneous sandy and young soils with limited or poor profile development, such as Arenosol (24 data points in the collected dataset) and Cambisol (13 data points). Group II included soils with a clay-rich or argic subsoil horizon with a low base saturation, low activity clay (Acrisol, 47 data points), high base saturation, high activity clay (Luvisol, 8 data points) or high base saturation, low activity clay (Lixisol, 43 data points). Group III included soils where iron and/or aluminum chemistry plays a major role in their formation, mainly Ferralsols (98 data points) and few Nitisols (4 data points).

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