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Soil organic carbon storage as a key function of soils - A review of drivers and indicators at various scales

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A R T I C L E I N F O

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A B S T R A C T

The capacity of soils to store organic carbon represents a key function of soils that is not only decisive for climate regulation but also affects other soil functions. Recent efforts to assess the impact of land management on soil functionality proposed that an indicator- or proxy-based approach is a promising alternative to quantify soil functions compared to time- and cost-intensive measurements, particularly when larger regions are targeted. The objective of this review is to identify measurable biotic or abiotic properties that control soil organic carbon (SOC) storage at different spatial scales and could serve as indicators for an efficient quantification of SOC. These indicators should enable both an estimation of actual SOC storage as well as a prediction of the SOC storage potential, which is an important aspect in land use and management planning. There are many environmental conditions that affect SOC storage at different spatial scales. We provide a thorough overview of factors from micro-scales (particles to pedons) to the global scale and discuss their suitability as indicators for SOC storage: clay mineralogy, specific surface area, metal oxides, Ca and Mg cations, microorganisms, soil fauna, aggregation, texture, soil type, natural vegetation, land use and management, topography, parent material and climate. As a result, we propose a set of indicators that allow for time- and cost-efficient estimates of actual and potential SOC storage from the local to the regional and subcontinental scale. As a key element, the fine mineral fraction was identified to determine SOC stabilization in most soils. The quantification of SOC can be further refined by including climatic proxies, particularly elevation, as well as information on land use, soil management and vegetation characteristics. To enhance its indicative power towards land management effects, further "functional soil characteristics", particularly soil structural properties and changes in the soil microbial biomass pool should be included in this indicator system. The proposed system offers the potential to efficiently estimate the SOC storage capacity by means of simplified measures, such as soil fractionation procedures or infrared spectroscopic approaches.

1. Introduction

The multifunctionality of soils is increasingly recognized as a decisive aspect in global land management. After its introduction in the 1990s, the concept of soil functions was further developed to a systemic approach covering all aspects of soil-based environmental, social and economic services (Baveye et al., 2016; Blum, 1993). Although there is no agreed upon definition, the main ecological functions of soils, beyond technical and cultural aspects, are: biomass production, storage and filtration of water, storage and recycling of nutrients, habitat for biological activity and carbon storage. The latter can be regarded as a key function of soils, as it is not only decisive for climate regulation, but strongly affects all other functions as well. In recent years, the rising integration of soil functions in environmental policy making has

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created a need for more detailed and quantitative information on specific soil functions (Baveye et al., 2016; Lehmann and Stahr, 2010; Schulte et al., 2015). This is a prerequisite to assess the impact of land management changes on soil functionality, as it is intended by the "BonaRes" funding initiative in Germany (www.bonares.de). However, a direct quantification of soil functions requires various cost- and labour intensive analyses and, hence, is typically not feasible, especially when large spatial scales are targeted. An alternative approach could be to identify measureable biotic or abiotic properties that contain sufficient information to be used as indicators for the quantification of soil functions (Baveye et al., 2016; Rabot et al., 2017; Schulte et al., 2014; Vogel et al., 2018). Because of the soil's multifunctionality and complexity, an indicator system is needed that integrates the relevant soil processes and their interactions. Such indicators can be derived from "functional characteristics" as recently proposed by Vogel et al. (2018), which have to be measurable in a time- and cost-efficient way and should be reproducible on a relatively low level of expertise.

The main objective of this review is to identify a set of indicators that enables a quantification of soil organic carbon (SOC) storage at different spatial scales. This set of indicators should allow for not only a prediction of current SOC storage from local to regional scales but also for a spatial estimation of the SOC storage potential, which is an important aspect for land management decisions and for reponses to environmental disturbances. A quantification of current and potential SOC storage via a set of suitable indicators could be a promising alternative to precise and accurate determination of SOC storage via (expensive and time-consuming) field work and laboratory analyses. Moreover, for many purposes, an assessment of SOC storage and its role in ecosystem functioning requires less precision. For example, an indicator-based estimation of SOC is probably sufficient to help farmers decide which agricultural measures are appropriate to optimize the SOC level of a specific field. Similarly, landscape planners, administrative bodies or ecologists may wish to identify hotspots which are particularly valuable for conservation, without the need for precise measurements of SOC storage. To this end, the use of indicators of SOC storage is appropriate. Such indicators integrate one or several environmental and soil properties which are known to be related with SOC storage, allowing comparative assessment of soils at different locations with regard to expected SOC storage. The so-called "SCORPAN approach" of McBratney et al. (2003), building up on the factorial equation of Jenny (1941), aims to describe any soil property as a function of climate, organisms, topography, parent material, time, space and soil information itself. This approach provides a framework for an identification of factors controlling SOC storage that could serve as indicators. However, this is impeded by the fact that these factors are scale-dependent, indicating a "hierarchy of controls" (Manning et al., 2015). Important factors at the micro-scale (from particles to pedons), such as metal oxides or specific surface area (SSA), are less relevant at larger spatial scales, where SOC storage can be related to climate, topography, parent material, vegetation, land use and management (Manning et al., 2015; Moni et al., 2010). In this review, we present an overview of the importance of these factors for SOC storage with respect to scale. We further dicuss their suitability as indicators for SOC storage and provide an outlook on an indicator system to estimate actual and potential SOC storage.

2. Drivers and indicators of soil organic carbon storage

2.1. Climate

Climatic conditions, namely temperature and precipitation, are key drivers of SOC storage globally as well as at broad (sub-)regional scales, affecting both C input into the soil and SOC decomposition. Precipitation determines net primary productivity (NPP) in many (water-limited) terrestrial environments and thus the input of C into the soil. Furthermore, humid conditions favour the formation of SOC-stabilizing mineral surfaces by intensified weathering of the parent material (Chaplot et al., 2010; Doetterl et al., 2015) and often cause soil acidification leading to reduced decomposition of soil organic matter (SOM) (Meier and Leuschner, 2010). Temperature largely affects the microbial decomposition of SOM as its complex molecular attributes have a high intrinsic temperature sensitivity (Conant et al., 2011; Davidson and Janssens, 2006; von Lützow and Kögel-Knabner, 2009). Although this relationship is governed by multiple constraints, numerous studies have indicated a decrease of SOC with increasing temperatures (e.g., Jobbagy and Jackson, 2000; Koven et al., 2017; Sleutel et al., 2007; Smith et al., 2005;).

As a result of these combined influences, SOC stocks are generally highest under cool humid conditions and decrease under warmer and drier climates both at the global scale (Jobbagy and Jackson, 2000; Post et al., 1982) and at the (sub-)regional scale (Alvarez and Lavado, 1998; Badgery et al., 2013; Baritz et al., 2010; Burke et al., 1989; Callesen et al., 2003; de Brogniez et al., 2014; Gray et al., 2016; Hobley et al., 2015; Martin et al., 2011; Paul et al., 2002; Viscarra-Rossel et al., 2014). However, numerous studies have indicated that the relative importance of climate for SOC storage diminishes with increasing soil depth, where factors controlling the stabilization of SOM (see Section 2.5) become more important (Badgery et al., 2013; Gray et al., 2016; Hobley et al., 2015; Jobbagy and Jackson, 2000). At the regional scale, climate effects may be masked in topsoils by land use/management, particularly in cropland soils, where intensive management (fertilization, irrigation etc.) can counterbalance climate effects (Goidts et al., 2009; Tan et al., 2004; Wiesmeier et al., 2013). Moreover, climate variability at local scales is frequently small, so that climatic control of SOC storage is less relevant in contrast to other factors. In a regional study conducted in the conterminous USA by Homann et al. (2007), mean annual precipitation, evapotranspiration and clay content were positively correlated to SOC in the top 20 cm. The observed differences among regions were attributed to more localized differences in vegetation, clay mineralogy and other processes.

As climate constitutes a major controlling factor for SOC storage at regional to global scales, mean annual air temperature (MAT) has proven to be a suitable indicator that could serve as indicator for SOC storage due to its global availability in free climate databases. Closely related, vapor pressure deficit has also been shown to be closely associated with SOC storage (Allen et al., 2013). Although mean annual precipitation (MAP) has been shown to have a stronger association with SOC and its depth distribution in some studies at regional (Hobley et al., 2015) and global scales (Jobbagy and Jackson, 2000), other studies have reported stronger relationships between SOC with temperature than with precipitation (Allen et al., 2013; Wang et al., 2004). In a review of tropical plantation and successional forests, MAT was a better predictor of SOC than precipitation in a regression tree model, with cooler sites containing greater SOC (Marin-Spiotta and Sharma, 2013). As such, these relationships between climate drivers and SOC storage are not always easy to decipher.

To reconcile these conflicting reports, it has been suggested that the relative importance of either temperature or precipitation on SOC storage depends on the main limiting factors of SOC production and turnover in the study region (Hobley et al., 2015). In arid or semi-arid environments, water availability limits NPP, so that the system, and therefore SOC storage, is input limited (Hobley et al., 2016). In contrast, in regions with sufficient water-availability but colder temperatures, microbial activity is limited to a greater degree than NPP (von Lützow and Kögel-Knabner, 2009), so that the system and SOC storage is output limited. These patterns result in an accumulation of SOC in colder, moister areas, moderate SOC storage in warmer, moist regions and lower SOC storage in drier, hotter regions.

In this context, elevation above sea level may be a better indicator than climatic variables alone, as it integrates the effects of temperature and precipitation on NPP and decomposition and reflects erosional and depositional processes which influence spatial distribution of soil types as well as depth of soil. In regional studies, SOC stocks have been found to be highly correlated with elevation (de Brogniez et al., 2014; Leifeld et al., 2005; Neufeldt, 2005). Derived from digital elevation models with high accuracy and ease, elevation is therefore probably the most promising indicator for SOC storage at the regional scale. However, this may not hold true for continental climates in relatively flat areas, where elevation is not a good proxy for temperature and precipitation. A further exception to expected trends of SOC decreases with increasing elevation may occur when precipitation and temperature trends are decoupled across an elevation gradient, as may ocur in some tropical mountains, and topographic effects on soil moisture become more important (McGroddy and Silver, 2000).

2.2. Topography

Topographical features can play a crucial role for SOC storage as they control precipitation, water flow paths, water accumulation, and discharge and therefore contribute significantly to erosional processes. Sites with a high contributing area, low inclination and concave curvatures favour water accumulation, whereas steep slopes and convex curvatures lead to higher water discharge. Soil moisture controls not only NPP and therefore SOC input, but also microbial activity and SOC output. Both low and high soil moisture contents reduce microbial activity due to reduced substrate mobility/oxygen availability, favouring the accumulation of SOC. Thus, soil moisture was identified as a decisive factor for the spatial distribution of SOC in several regional studies (Mayes et al., 2014; Meersmans et al., 2008; Neufeldt, 2005; Tan et al., 2004; Vasques et al., 2010). Moreover, water flow paths are key determinants triggering soil erosion and redistribution, which in turn may have profound effects on SOC storage in dynamic landscapes (Doetterl et al., 2016; Seibert et al., 2007). Topographical features further influence the local microclimate which in turn may affect SOC storage at local scales. Differences in SOC depth profiles along a *catena* in tropical mountains of Puerto Rico were attributed to poor drainage and reducing conditions (Johnson et al., 2015; Silver et al., 1999). Similarly, variability in SOC storage along an elevational gradient in precipitation in Hawaii was better explained by the effect of redox and oxygen availability rather than climate, NPP or soil mineralogy (Schuur et al., 2001).

Despite these observations, there is no general agreement on the importance of topography-related factors for SOC storage, which seems to be partly explained by the various scales of observation. Analyses of different SOC data sets from global to subcontinental and regional scales revealed that topographical features, represented by e.g. topographic position, slope gradient and aspect, were of minor importance for SOC variability (Garten and Ashwood, 2002; Gray et al., 2016; Gray et al., 2009; Hobley et al., 2015; Schulp and Veldkamp, 2008; Tan et al., 2004; Viscarra-Rossel et al., 2014; Xiong et al., 2014). Local terrain attributes such as slope, aspect and curvature appear to influence SOC storage only at small spatial scales (< 100 m) and landscape position may become important at larger scales (> 100 m) (Hobley et al., 2015; Minasny et al., 2013). In fact, regional terrain attributes were the most important controlling factors for SOC variation within a relatively small study area in Panama, followed by local factors such as slope and curvature (Grimm et al., 2008). Furthermore, in a study on the SOC distribution in Laos, a significant proportion of SOC variability was explained by slope gradient (Chaplot et al., 2010). In a study on volcanic soils in Costa Rica, slope explained 27% of the variation in the magnitude and direction of differences in SOC between paired pastures and forests, even though at the regional scale, land use did not affect average SOC stocks (Powers and Veldkamp, 2005). The authors attributed these elevational differences to differences in soil weathering, with the fraction of non-crystalline clay content being a significant predictor of SOC. It therefore appears that topography is relevant at local (hillslope) scales where the soil properties do not vary a lot, but less relevant across larger landscapes, where SOC storage is averaged over a large variety of soil properties so that other factors become dominant.

In terms of topographical features which are indicative of the soil moisture-induced accumulation of SOC, the topographic wetness index (TWI) is a promising factor, which can easily be calculated from digital elevation models. Large TWI values are usually found at lower areas of a landscape with large contributing areas and indicate increased likelihood of saturated conditions and/or accumulation of eroded material, resulting in high SOC stocks (Sorensen et al., 2006). In a regional SOC inventory in Germany, TWI was identified as an important factor for the spatial variability of SOC in agricultural soils due to its indicative power for SOC accumulation in groundwater-affected soils (Wiesmeier et al., 2013). However, at a larger sub-continental scale, TWI was not important as a predictor of SOC stocks and content (Hobley et al., 2015). This suggests that TWI, similar to general topographic parameters, is relevant at smaller rather than larger scales.

2.3. Parent material

The role of parent material as a predictor of SOC storage has been examined in several studies from sub-regional to global scale. The potential influence of parent material is mainly attributed to its control on soil mineralogy, texture and fertility, which affect NPP and the stabilization of SOM (see Sections 2.5.2 and 2.5.6) (Gray et al., 2016; Herold et al., 2014). Moreover, parent material is stongly linked to soil type, which is associated with SOC storage (see Section 2.5.1). However, there is no consistency on the importance of parent material to SOC storage. A global analysis of the relationships between environmental factors and SOC based on 4382 soil profiles revealed only a weak correlation between parent material classes and SOC (Gray et al., 2009). However, the authors concluded that assignment of parent material into classes assuming specific silica contents may be a major simplification. In SOC inventories at the regional and catchment scale under different climatic conditions, parent material was also of minor importance for the spatial distribution of SOC (Grimm et al., 2008; Hobley et al., 2015; Wiesmeier et al., 2014a). However, in a study in Eastern Australia, classes of parent material became important for SOC storage in the subsoil (Hobley et al., 2015). Here again, it appears that the scale of investigation influences the relative importance of different drivers controlling SOC storage, as well as the depth of investigation.

This is confirmed by results from other studies, which found indications for a clear effect of parent material on SOC stocks (Vasques et al., 2010; Xiong et al., 2014). In a landscape study in France, differences in SOC between forests and crops were smaller in loess-derived soils than on other parent materials (Barre et al., 2017). In a semi-arid region in Turkey, parent material also mediated the effect of land use on SOC stocks within a soil order (Mayes et al., 2014). SOC stocks were greatest under grazing on alluvial and lacustrine soil sites, whereas cropland agriculture had greater SOC stocks on limestone terraces. The importance of parent material on SOC may be due to differences in silica (Si) contents (Badgery et al., 2013) based on geologic source and weathering stage. In two large subcontinental studies in Australia and Chile, Si was a key predictor for SOC storage, with increasing SOC stocks with increasing mafic character of the parent material and considerably lower stocks in soils developed from siliceous material (Doetterl et al., 2015; Gray et al., 2016). This may be due to a coarser soil texture and lower surface reactivity, associated with a lower fertility and C storage capacity. Doetterl et al. (2015) argued that climate may be of secondary importance for SOC storage from a long-term perspective, and that geochemical predictors, specifically Si concentrations, are more important. However, at larger superregional to global scales, studies clearly indicate the importance of climate for SOC storage in soils of widely varying parent material (Hobley et al., 2015; Jobbagy and Jackson, 2000), suggesting that the control of climate on production of SOC and its turnover cannot be neglected.

2.4. Organisms

2.4.1. Natural vegetation

At the (sub-)regional scale with more uniform climatic conditions, the level of SOC stocks is strongly affected by vegetation and land use (see Section 2.4.2). Among different climatic ranges of the world, significantly different SOC stocks and depth distributions were found between plant functional types (shrubland, grassland, forest) due to different C allocation patterns (Jobbagy and Jackson, 2000). Studies at the continental and regional scale in various climatic zones confirmed that vegetation type affects SOC storage by controlling both the input and decomposition of C. In a North American study that included a latitudinal gradient from tropical forests to arctic tundra, an estimate of the C input into the mineral pool was the best predictor of SOC stocks in topsoils (Frank et al., 2012). The analysis of factors controlling SOC stocks in Eastern Australia confirmed that vegetation (captured by vegetation cover) mainly affects SOC in the topsoil and that its influence declines with depth (Gray et al., 2016). Plant diversity and functional composition may also affect SOC storage at different spatial scales, particularly in natural grassland ecosystems under semi-arid conditions, but also in temperate environments (Soussana et al., 2004). At the (sub-)regional scale, SOC storage among different grassland types can differ significantly, e.g., at the Qinghai-Tibetan Plateau (Liu et al., 2016). However, different grassland types are probably coupled with different climatic conditions (particularly precipitation), which largely affect NPP. Vegetation-type effects on SOC often have been attributed to differences in plant litter chemistry, yet the molecular recalcitrance of plant inputs is strongly modulated by environmental factors, and there is little evidence that such intrinsic properties are important controls on SOC over timescales of decades to millennia (von Lützow et al., 2006). At the local scale, plant diversity has been found to be positively correlated with SOC accumulation, mainly due to higher root biomass production induced by specific plant functional traits (e.g., positive interactions between N-fixing legumes and grasses; enhanced N mineralization in species-rich grasslands without legumes) (Cong et al., 2014; Fornara and Tilman, 2008; Lange et al., 2014; Rees et al., 2005; Steinbeiss et al., 2008). For forests, there is no consensus on the effect of tree species on SOC storage. Under temperate climatic conditions, SOC stocks under coniferous species may be relatively high, particularly due to an accumulation of acidic litter in the organic layer (Jandl et al., 2007; Oostra et al., 2006; Schulp et al., 2008), yet other studies have failed to identify significant differences between SOC stocks from whole soil profiles among forest types (Vesterdal et al., 2008).

2.4.2. Land use and management

Land use change is assumed to be the most dynamic factor of SOC changes (Guo and Gifford, 2002; Poeplau et al., 2011; Post and Kwon, 2000; Viscarra-Rossel et al., 2014). In temperate regions, the storage of SOC increases in the order cropland < forest < grassland (Martin et al., 2011; Meersmans et al., 2008). However, depending on site and climatic conditions, forest SOC stocks can be higher than grassland stocks (Leifeld et al., 2005; Lettens et al., 2005). A large number of studies and meta-analyses found strong evidence for a large SOC decline of 30 to 80% when forests or grasslands were converted to cropland (Guo and Gifford, 2002; Poeplau et al., 2011; Wei et al., 2014). The loss of SOC in cultivated soils has been attributed to erosion, lower C inputs in croplands, a reduced stabilization of SOM due to deteriorated aggregation, and a subsequent mineralization promoted by increased soil temperature and aeration (Balesdent et al., 2000; Hamza and Anderson, 2005; Six et al., 1998). Contrasting with these losses, land use extensification, particularly the conversion of cropland to grasslands or forests, often leads SOC increases (Post and Kwon, 2000; Schulp and Veldkamp, 2008). Overall, land use information constitutes an important additional indicator for SOC storage from local to subcontinental scales.

SOC storage is also influenced by land management, though effects can be difficult to detect, especially deep in the soil profile. Agricultural management systems with crop rotations generally reveal higher amounts of SOC than monoculture systems (Jarecki and Lal, 2003). Increasing crop rotational diversity has been shown to play a major role in increasing SOC storage, driven by enhanced root C input, soil microbial diversity, and soil aggregate stability (Tiemann et al., 2015). Moreover, crop systems that are dominated by plants with distinctive and deep root systems promote SOC storage, as retention of root derived C in soil has been found to be 2.3 times higher than for aboveground residues (Kätterer et al., 2011). In particular, agricultural systems with perennial crops/forages and cover crops reveal distinctly higher SOC stocks (Jarecki and Lal, 2003; Paustian et al., 1997; Poeplau and Don, 2015). Crop rotational diversity, the use of organic amendments or perennial cropping systems may be indicative for SOC accumulation compared to conventional cropping systems. The indicative value of reduced tillage systems may be low, as there is evidence that reduced or no-tillage does not affect SOC storage but only leads to redistribution of SOC within the soil profile (Luo et al., 2010; Powlson et al., 2014).

In contrast to cropland management, less is known in terms of grassland management effects on SOC storage due to limited data availability and large spatial variability of SOC in grassland soils (Soussana et al., 2004). In global meta-analyses on grazing effects on SOC storage in grassland soils, no uniform trend was found (McSherry and Ritchie, 2013; Milchunas and Lauenroth, 1993; Pineiro et al., 2010; Smith et al., 2008). High grazing intensities caused a decrease of SOC in C3-dominated grasslands, but an increase in C4 and C4-C3 mixed grasslands (McSherry and Ritchie, 2013). In temperate grasslands, there are indications that the intensity of the grassland management, particularly in terms of number of cuttings and N fertilization, strongly affects SOC storage (Conant et al., 2001; Soussana et al., 2004). Specifically, an intermediate level of management intensity with regard to N fertilizer application and cutting times/stocking rates seems to be associated with higher SOC stocks compared to lower amounts in both extensively and intensively managed grasslands (Soussana et al., 2004; Ward et al., 2016). Grassland management effects are not restricted to the topsoil but also detected in the subsoil (Ward et al., 2016). In forest soils, management aspects such as thinning, extension of the rotation period, fertilization, liming, fire, storm and insect management, harvest management and input of harvest residues as well as stand age can affect SOC storage (Carroll et al., 2012; Jandl et al., 2007; Lal, 2005; Lorenz and Lal, 2014; Nabuurs et al., 2008). Due to the complexity of management practices on cropland, grassland and forest soils the suitability of management information as a indicator for SOC storage seems to be limited. A classification of management systems or the intensity level of management could be a promising strategy to refine an indicator system for SOC storage.

2.4.3. Soil biota

2.4.3.1. Microorganisms. Heterotrophic soil microorganisms (Bacteria, Eucarya (fungi) and Archaea) mediate turnover of SOM to cover their nutrient and energy needs for maintenance and growth. Their central role in the mineralization-immobilization-turnover of SOM was linked to the 'eye of the needle' or the 'microbial carbon pump' through which carbon and nutrient transformations are mediated (Jenkinson, 1977; Liang et al., 2017). The fact that microbial residues and exudates may contribute up to 80% of C in stable soil OM fractions (Gleixner et al., 2002; Kiem and Kögel-Knabner, 2002; Liang and Balser, 2011; Simpson et al., 2007; Sorensen, 1981, 1983) indicates, that the soil microbial biomass (SMB) plays a dual role in SOM turnover, balancing SOM mineralization and SOM stabilization processes.

Because most soil systems are resource (C, N, etc.) limited (exept peat/organic soils), the pool size of SMB is therefore usually positively related to OM input (NPP) and the SOM content in arable, grassland and forest ecosystems (Wardle, 1992). Strongest relationships between SMB-C and SOC have been demonstrated in soils with < 2.5% SOC

(Anderson and Domsch, 1980). As a living and active soil compartment with short turnover times of 1–5 years (Jenkinson and Ladd, 1981), the SMB pool changes more rapidly than total amounts of SOC. Therefore changes in the SMB-C pool and the ratio of SMB-C to SOC had been suggested as useful and sensitive indicators for monitoring changes in SOC e.g. due to changes in management (Powlson et al., 1987; Powlson and Jenkinson, 1981), also because small gradual changes in SOC are difficult to detect in the short term due to generally high background C levels. The ratio of SMB-C to SOC will increase for a time if the input of OM increased and decreases for a time if the input is decreased. Constancy of the SMB-C to SOC ratio is an indication that the soil ecosystem is in equilibrium (Anderson and Domsch, 1980). The establishment of a higher SMB-C to SOC ratio can further be discussed in a more efficient microbial C turnover.

A wide spectrum of SMB-C to SOC ratios ranging from 0.3 to over 7.0% was reportet in the literature (von Lützow et al., 2007). This wide range can be attributed to different soil management ecosystems (arable, grassland, forest) (Anderson and Domsch, 1980; Wardle, 1992) and environmental factors that influence C-input, especially macroclimate variables, soil texture and soil pH (Insam and Domsch, 1988; Wardle, 1992). On local and global spatial scales the SMB-C to SOC ratio is neutrally or negatively correlated with SOC and the soil C/N ratio (Wardle, 1992), indicating that microbial C limitation declines as SOC content increases and that in many soil ecosystems soil N rather than soil C may influence the immobilization of OM by the microbial biomass. The lower proportion in temperate forest soils than in most other ecosystems is probably a result of N limitation and low pH. How constraints on microbial physiology may influence SOC storage is demonstrated by increased C losses due to mineralization despite greater C inputs, when microbes are nutrient limited (Allison et al., 2010; Craine et al., 2007; Fontaine et al., 2004; Kirkby et al., 2014). This 'microbial nutrient mining' may produce predictions for ecosystem C storage that are opposite to those of basic stoichiometric decomposition theory. Therefore, to maintain a specific level of SOM and to predict SOC storage changes, maintenance requirements of the SMB and anthropogenic changes in nutrient availability for microbes have to be considered (Craine et al., 2007).

In conclusion, monitoring the SMB pool size can be used as an early indicator for changes in SOC contents e.g. due to changes in soil management. The SMB pool is easily measurable and a unique and non-composite fraction that can be equated to a model pool (Islam et al., 1997; McGill, 1996; Smith et al., 2002). The measurement of SMB is based on chloroform-fumigation methods that were developed by Jenkinson (1976) and Vance et al. (1987). It is a cheep and precise method to estimate the microbial biomass pool size (C, N, P, …). Moreover, simplified methods are now available e.g. the freeze-dried soil extraction method as proposed by Islam et al. (1997). Changes in the ratio of SMB-C to SOC indicate if a soil ecosystem is loosing or accumulating C. Correlations of SMB-C/SOC with soil chemical parameters (e.g. soil C/N content) may identify limiting factors of SOM turnover.

2.4.3.2. Soil fauna. Soil fauna and their interactions, both with the biotic and the abiotic environment, are involved in a number of processes that promote C stabilization or mineralization. SOM is incorporated into the soil profile by bioturbation, where it becomes available to the soil microbial community (Lavelle et al., 1997; Scheu and Parkinson, 1994) or is protected against mineralization by mixing with soil particles (Bossuyt et al., 2006; Lavelle et al., 1997; von Lützow et al., 2006). Soil fauna regulate microbial populations (Hedlund and Augustsson, 1995; Scheu and Parkinson, 1994; Scheu et al., 2002), inhibit or stimulate microbial activity by grazing (Hättenschwiler and Bretscher, 2001; Mikola and Setala, 1998; Scheu et al., 2002) and alter the microbial community composition by selective grazing or inoculum dispersal (i.e. passive transport of microorganisms) (Lavelle et al., 1997; Scheu and Parkinson, 1994; Scheu et al., 2002).

Furthermore, soil fauna affects SOC storage by affecting soil structure. Soil (macro)porosity is changed by the burrowing activity of earthworms and enchytraeids (Falco and Coviella, 2016; van Vliet et al., 1993). Earthworm feeding activities have been suggested to induce long-term stabilization of SOC as C is protected from microbial attack by the formation of microaggregates within macroaggregates in earthworm casts (Bossuyt et al., 2005; Lubbers et al., 2013; Six et al., 2004). While there is evidence that earthworms affect aggregate formation and aggregate size distribution (Al-Maliki and Scullion, 2013; Coq et al., 2007; Falco and Coviella, 2016; Fonte and Six, 2010; Hedde et al., 2013; Sanchez-de Leon et al., 2014), a recent meta-analysis comprising 21 studies did not find an overall effect of earthworms on SOC stocks (Lubbers et al., 2013). However, these authors noted that the time frame of many studies may have been too short to detect earthworm stimulation of C sequestration. Studies suggest that microbial activity and thus mineralization rates increase in the first days after casting and dcreases over time, indicating that the physical protection within earthworm casts may only be apparent in dried or aged casts (Kawaguchi et al., 2011; Marinissen and Dexter, 1990; Shipitalo and Protz, 1988). Aggregate stability within casts furthermore seems to depend on the soil texture (Angst et al., 2017; Shaw and Pawluk, 1986), the type of OM ingested (Flegel et al., 1998) and earthworm species or functional group (Flegel et al., 1998; Shaw and Pawluk, 1986).

The role of soil fauna other than earthworms on the physical protection of organic material by aggregate formation generally received less attention (Maass et al., 2015) although millipedes (Fujimaki et al., 2010), springtails (Siddiky et al., 2012a; Siddiky et al., 2012b), mites (Barois et al., 1998) and enchytraeid worms (Barois et al., 1998; Dawod and Fitzpatrick, 1993; Didden, 1990; Marinissen and Didden, 1997) were shown to facilitate soil aggregation through the production of fecal pellets. In fact, fecal pellets are important components of soil and may constitute a large amount of the soil in some ecosystems (Barois et al., 1998; Dawod and Fitzpatrick, 1993; Marinissen and Didden, 1997; Phillips and FitzPatrick, 1999). Within these fecal pellets, C may be accumulated (Schrader et al., 1997) and protected against water and microbial decomposition, e.g. by a silicon-rich coating or polysaccharides (Barois et al., 1998; Fujimaki et al., 2010). The potential of fecal pellets to protect C depends on additional factors such as soil texture, which can lead to different results: Enchytraeidae for instance were shown to decrease mineralization in loamy soils, whereas increased mineralization was found in sandy soils (Koutika et al., 2001).

Overall, the amount of comparative data for the impact of soil fauna on SOC storage is insufficient, especially as the faunal contribution to SOC dynamics depends on a number of additional factors. As soil faunal (functional) groups differ in their feeding strategies and microhabitat, they are involved in different processes within soils and can thus lead to the protection of C. This impact furthermore shows a temporal variability with stimulated OM mineralization in the short term and C stabilization in the longer term (Dignac et al., 2017). Site-specific properties such as soil texture, OM quality and distribution (Bossuyt et al., 2006; Coq et al., 2007; Fonte and Six, 2010), land use (Hedde et al., 2013; Pulleman et al., 2005; van Vliet et al., 1993) or agricultural management (Fonte et al., 2007; Pulleman et al., 2005) may largely influence the magnitude and direction of soil faunal effects. Due to these complex interactions, the suitability of soil fauna as an indicator for SOC storage is limited. Furthermore, the identification and quantification of soil fauna is labour-intensive and time-consuming, as repeated sampling is needed to account for the seasonality and climate-sensitivity of soil fauna.

2.5. Soil properties

2.5.1. Soil type

There is broad consensus that soil type is strongly associated with SOC storage at multiple scales and under different climatic conditions. During pedogenesis chemical weathering reactions lead to shifts in the composition of soil mineralogy which strongly influence the bulk reactive mineral surface area and C storage (Percival et al., 2000; Torn et al., 1997), and the resultant characteristics are categorically described by soil types. Generally, primary minerals weather to form secondary minerals including Fe-oxides (e.g. ferrihydrite and goethite), Al-oxyhydroxides (e.g. gibbsite), short range order minerals (e.g. allophane and imogolite) and phyllosilicates (e.g. kaolinite). In contrast to climate, vegetation and parent material, which may rather serve as indicators for SOC storage at larger scales, SOC stocks can be stratified according to soil type even at smaller (local to subregional) scales. In most soil classification systems, SOC is a major differentiation criterion (Micheli et al., 2014). Numerous studies reported a predominant influence of soil type on SOC stocks both in top- and subsoils (Chaplot et al., 2010; Grimm et al., 2008; Hobley et al., 2015; Schulp and Verburg, 2009; Tan et al., 2004; Vasques et al., 2010; Wiesmeier et al., 2014a; Xiong et al., 2014). Interestingly, evidence was found that land use effects on SOC storage, which are often assumed to be main drivers, are strongly differentiated within soil type (Mayes et al., 2014; Wiesmeier et al., 2015b).

However, soil type is not an independent controlling factor but integrates a number of climate, parent material and topography related properties which directly affect the potential of soils to store C, particularly soil moisture regime and soil texture (see Sections 2.2 and 2.5.2). For example, Gleysols often store large quantities of SOC, but this is not an intrinsic feature of the soil type Gleysol but the result of the topographical situation of Gleysols associated with high soil moisture contents (Mayes et al., 2014; Wiesmeier et al., 2013). However, if precise information on soil properties such as texture, soil moisture etc. is not available, soil type can be a suitable indicator for SOC storage integrating a wide range of decisive factors.

2.5.2. Soil aggregation

Physical protection of SOM by aggregates is considered to be an important mechanism for C stabilization. Besides abiotic formation of aggregates, soil biota and the material that they produce play an important role in formation and stabilization of soil structure and are strongly involved in the process of OM occlusion by aggregation (see Section 2.4.3) (Elliott, 1986; Oades, 1984, 1993; Six et al., 2002b). The protective capacity of aggregates is mainly related to a spatial separation of substrate and microorganisms as well as to a reduced microbial activity due to a reduced diffusion of oxygen into aggregates (Mikutta et al., 2006; Six et al., 2002b). Macroaggregates (soil particles > 250 μm) provide a minimal physical protection (Krull et al., 2003; Six et al., 2002b, 2004), while microaggregates play a decisive role for the long-term SOM stabilization (Totsche et al., 2017). However, because microaggregates are formed within macroaggregates (Angers et al., 1997; Oades, 1984; Pronk et al., 2012b; Six et al., 2002b), macroaggregation is the first step for long-term sequestration of fresh C input (Pronk et al., 2012a; Six et al., 2000). The physical protection of C provided by intra-aggregate occlusion can also be expressed from the perspective of the pore network, which regulates the availability of oxygen, water, and other resources for OM decomposition (Ananyeva et al., 2013; Kravchenko et al., 2015; Strong et al., 2004; Toosi et al., 2017). Small pores may also provide physical protection for OM by reducing accessibility to decomposers (Kravchenko et al., 2015; Negassa et al., 2015).

Soil cultivation influences aggregation and thus physically stabilized C pools. Macroaggregates are sensitive to land management and mechanical disruptive forces of cultivation because they are formed from plant input products and much less stable than microaggregates (Beare et al., 1994; Edwards and Bremner, 1967; Oades, 1984). Long-term tillage with frequent plowing accelerates the turnover rate of macroaggregates and limits the physical protection of labile SOM compounds associated with soil aggregates (Beare et al., 1994; Grandy and Robertson, 2006; Six et al., 1998, 1999). Similarly, reduced plant inputs due to harvesting of agricultural products reduces macroaggregate formation (Oades, 1984). Thus, macroaggregates may be good predictors of potential SOC responses to tillage or land use change because of their importance for storing more labile SOM (Angers and Giroux, 1996; Jastrow et al., 1996). The microaggregate fraction and microaggregates-within-macroaggregates were additionally proposed as diagnostic fractions for the SOC sequestration potential in sustainable agroecosystems (Kong et al., 2005). Different studies have shown that the C in that fraction is sensitive to management-induced changes in a wide range of soil types and under different climates (Denef et al., 2004; Kong et al., 2005; Six and Paustian, 2014). These findings suggested that microaggregates-within-macroaggregates may be indicative for changes in total SOC in response to changes in tillage and other management practices (Denef et al., 2004; Six et al., 2000). In a recent work, Segoli et al. (2013) successfully incorporated the explicit representation of the turnover of soil aggregates into the *AggModel* to decipher the influence on SOM dynamics. However, a major drawback is that the characterization of the aggregate size distribution is very sensitive to the soil sample pretreatment and to the methodology applied (e.g. type of sieving, its duration, etc.) (Rabot et al., 2017; von Lützow et al., 2007).

2.5.3. Silt and clay content

Among the major SOM stabilization mechanisms, interaction of OM with mineral surfaces is regarded as quantitatively most important in a wide range of soils (Sollins et al., 1996; von Lützow et al., 2006). This is indicated by a strong correlation of SOC stocks with clay contents observed in numerous studies at different spatial scales (Arrouays et al., 2006; Hassink et al., 1997; Kaiser and Guggenberger, 2000; Zinn et al., 2007). Therefore, soil texture is probably one of the most promising factors to be used as indicator for SOC storage over a wider range of scales. Soil texture data is available for many regions and there also global data sets with specific information on clay-size minerals (Hengl et al., 2017; Ito and Wagai, 2017).

Furthermore, on the basis of the stabilizing capacity of the fine mineral fraction (medium/fine silt and clay particles $< 20/< 50$ μm), simple regression models have been developed to estimate the (stable) SOC storage potential (Table 1). As the fine mineral fraction contains a large proportion of total SOC in most soils, particularly in cropland soils, it serves as a useful measure for potential SOC storage (Beare et al., 2014; Wiesmeier et al., 2014b). In his pioneer work, Hassink (1997) found a strong correlation between the content of fine soil particles < 20 μm and fine fraction SOC in a wide range of uncultivated and grassland topsoils of temperate and tropical regions and proposed that this correlation could be used to estimate the degree of saturation of stable SOC. Although Hassink's equation proved to be suitable to estimate the C sequestration potential of agricultural soils in temperate regions (Angers et al., 2011; Carter et al., 2003; Chan, 2001; Conant et al., 2003; Sparrow et al., 2006; Wiesmeier et al., 2014b; Zhao et al., 2006), more specific approaches have been developed for soils with different clay types, land uses and climatic regions. Feller and Beare (1997) delineated regression models based on the fine fraction < 20 μm for cultivated and non-cultivated soils and soils with different mineralogy in tropical environments. Further significant progress in this direction was achieved by Six et al. (2002a) who expanded the analysis of Hassink (1997) and developed specific regression models for cropland, grassland and forest soils with different clay types (1:1 vs. 2:1) and size delimitations of the fine fraction (< 20 μm vs. < 50 μm). Their results reveal that there are significant differences among land uses, clay types and fine fraction size classes. Generally higher intercepts (which represent the baseline C storage in particles at or below the size delimitation) of the regressions based on fine fractions < 50 μm compared to < 20 μm were explained by either additional C binding of organomineral associations into larger silt-sized aggregates or the presence of silt-sized particulate organic matter. The latter may also have caused the relatively high intercept in Hassink's equation, as POM was not separated from the fine fraction, e.g. by density fractionation (Wiesmeier et al., 2015a). It should be noted, however, that

Table 1
Overview on approaches to estimate the SOC storage capacity for soils under different climates, land uses and clay types based on linear regressions with the fine mineral fraction (silt and clay particles < 20/ < 50 µm).

Study	Climate	Land use	Clay type	Soil depth	n	Fine fraction	Linear regression	r
Hassink, 1997	Temperate and tropical	Grassland	Diverse	0–10 cm	33	< 20 µm	$4.09 + 0.37 \times ff_{<20}$	0.89
Feller and Beare, 1997	Tropical	Uncultivated	1:1	0–20 cm	25	< 20 µm	$-2.3 + 0.49 \times ff_{<20}$	0.91
		Cultivated			25		$3.2 + 0.29 \times ff_{<20}$	0.95
Six et al., 2002a	Diverse	Cropland	Diverse	–	146	< 20 µm	$4.38 + 0.26 \times ff_{<20}$	0.41
		Grassland					$2.21 + 0.42 \times ff_{<20}$	0.44
		Forest					$-2.51 + 0.63 \times ff_{<20}$	0.55
		Cropland	Diverse		66	< 50 µm	$7.18 + 0.20 \times ff_{<50}$	0.54
		Grassland					$16.33 + 0.32 \times ff_{<50}$	0.35
		Forest					$16.24 + 0.24 \times ff_{<50}$	0.35
		Diverse	1:1		47	< 20 µm	$1.22 + 0.30 \times ff_{<20}$	0.74
			2:1		93		$3.86 + 0.41 \times ff_{<20}$	0.39
		Diverse	1:1		10	< 50 µm	$5.50 + 0.26 \times ff_{<50}$	0.38
			2:1		52		$14.76 + 0.21 \times ff_{<50}$	0.07
Zhao et al., 2006	Temperate	Cropland	–	0–15 cm	12	< 20 µm	$-0.52 + 0.22 \times ff_{<20}$	0.88
Liang et al., 2009	Temperate	Grassland	2:1	0–30 cm	27	< 20 µm	$0.36 \times ff_{<20}$	0.51
Feng et al., 2013	Diverse	Diverse	1:1	–	101	< 20 µm	$1.04 + 0.23 \times ff_{<20}$	0.72
			2:1		241	< 20 µm	$1.68 + 0.32 \times ff_{<20}$	0.52
Beare et al., 2014	Temperate	Grassland	Allophanic	0–15 cm, 15–30 cm	29	< 20 µm	$1.35 \times ff_{<20}$	–
			Non-allophanic		184		$0.62 \times ff_{<20}$	–
Wiesmeier et al., 2015a	Semi-arid	Grassland	2:1	0–10 cm	14	< 20 µm	$-0.23 + 0.36 \times ff_{<20}$	0.98

these intercepts are mathematically derived and can be negative, so they should be interpreted with caution. In the study of Six et al. (2002a), further differences were found among clay types (lower C saturation in soils dominated by 1:1 clays compared to 2:1 clays, see Section 2.5.4) and land uses (higher slopes in forest soils compared to grassland and cropland). These results indicate that it is not just the quantity of the fine fraction, but also its quality which drives SOC retention in soils.

Feng et al. (2013) further developed the estimation of C stabilization by the fine fraction. They argued that predictions based on least-squares linear regressions are limited as data sets may contain high proportions of soils which are not saturated or over-saturated. They proposed that either specific surface area (SSA) and OC loading (see Section 2.5.4) should be incorporated into the models or that boundary line analysis should be used including thus only soils close to C saturation, e.g. by selecting only the upper 10th percentile of the data. The estimates of C stabilization in soils from 32 published studies based on boundary line analysis were substantially higher compared to linear regression approaches. Similar to the results of Six et al. (2002a), Feng et al. (2013) found general differences among land uses and clay types.

In addition to the 10th percentile approach proposed by Feng et al. (2013), quantile regression analysis was used to estimate the C storage capacity of grassland soils in New Zealand (Beare et al., 2014). In this study, a model that included extractable Al, pH and SSA predicted SOC stabilization capacity much better than models solely based on the fine fraction content. This could partly be related to the andic properties of soils in New Zealand (Matus et al., 2014). A comparison of allophanic and non-allophanic soils of New Zealand revealed that the best model for the prediction of the SOC storage capacity included SSA and extractable Al, although different coefficients between these two soil units were found (McNally et al., 2017). For soils in drier environments, Hassink (1997) found significantly lower OC contents in the fine fraction of natural grassland soils, probably due to a lower C input caused by a relatively low NPP in water-limited environments (Chan, 2001). Liang et al. (2009) and Wiesmeier et al. (2015a) used natural steppe relicts in Northern China to derive regression models for fine fraction SOC saturation similar to existing approaches. Lower intercepts were attributed to lower C inputs due to climate limitations, and potentially due to different clay mineralogy, indicating that where SOC storage is limited by other factors, soil texture may not be as important for SOC stabilization.

Overall, evidence for a global dependence of the SOC storage capacity on the fine fraction content was found, although this relationship strongly depends on climatic conditions, land use, clay type and size limits of the fine fraction. Certainly, there are limitations, particularly in soils where recalcitrance or other stabilizing agents such as Al/Fe oxides play an important role. Furthermore, one has to bear in mind that such approaches are not suitable to estimate the total SOC storage capacity of soils, as C associated with particles > 20/50 µm, e.g. in macro- and microaggregates are not accounted for (Baldock and Skjemstad, 2000). Nevertheless, it was demonstrated that the fine fraction content is a useful and widely available indicator for the SOC storage capacity in many soils.

2.5.4. Clay mineralogy and specific surface area

The widely reported association of SOC with soil mineralogy (Bosatta and Agren, 1997; Kaiser and Guggenberger, 2000; Kiem and Kögel-Knabner, 2002) is generally attributed to the large specific surface area (SSA) of fine soil minerals such as phyllosilicates (known as clay minerals), which promotes sorption of OC to the mineral matrix (Balabane and Plante, 2004; Tipping, 1981; von Lützow et al., 2006; Wagai and Mayer, 2007). Hence, there is generally a positive relationship between SSA and SOC storage, and soils dominated by clay minerals with high SSA are expected to adsorb more humic substances than soils dominated by minerals with low SSA (Tate and Theng, 1980). Clay mineralogy (clay type) affects the SSA available for OC sorption (Feng et al., 2013; Kahle et al., 2002; Mayer, 1994; Wiseman and Puttmann, 2005). In particular, expandable phyllosilicates, e.g., clay minerals such as smectites and vermiculites have a large impact on OC stabilization in soils (Beare et al., 2014; Ransom et al., 1998) and provide a high surface area up to 800 $m^2 g^{-1}$ (Carter et al., 1986), whereas SSA of non-expanding silicates (e.g., kaolinite, mica) ranges from 10 to 70 $m^2 g^{-1}$ (Dixon, 1977). Additionally, short range-order aluminosilicate clays, e.g., allophane and imogolite-type minerals have large surface areas (Beare et al., 2014; Matus et al., 2014; McNally et al., 2017; Torn et al., 1997), so that allophanic soils generally have a high SSA and large portions of SOC are associated with fine soil fractions. Illustrating the importance of allophane surfaces for C retention (Beare et al., 2014; McNally et al., 2017), Percival et al. (2000) found a positive relationship between SOC stocks and allophane, but not between SOC and total clay content in New Zealand grasslands. However, the covariance of these parameters (allophane and fine texture) makes it difficult to differentiate their importance as a driver of SOC storage, as high SSA of silt and clay fractions can be attributed to the presence of

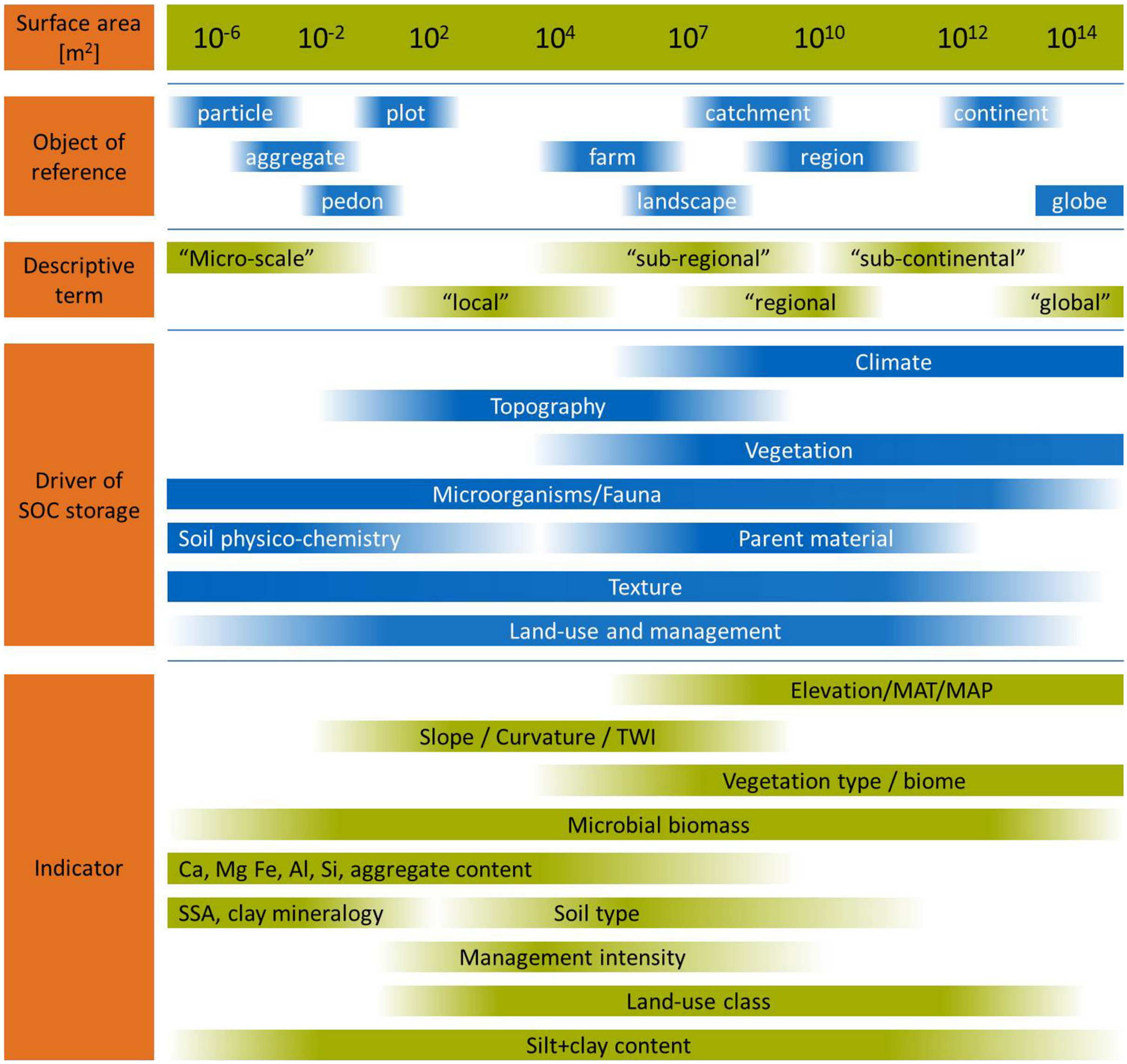

Fig. 1. Overview on the scale-dependent hierarchy of drivers and indicators for SOC storage (MAT = mean annual temperature, MAP = mean annual precipitation, TWI = topographic wetness index, SSA = specific surface area).

Table 2
Site-typical SOC contents for cropland soils of Bavaria as determined by soil texture class and elevation as proxy for climate conditions (Capriel, 2010, adapted).

Soil texture class	Elevation (m a.s.l.)	SOC (mg g^{-1})
Sand	< 350	0.70–1.37
	350–550	0.75–1.57
Silt	< 350	0.96–1.56
	350–550	1.17–1.88
	> 550	1.20–2.19
Clay	< 350	1.23–2.16
	350–550	1.31–2.52
	> 550	2.33–4.19
Loam	< 350	1.00–1.64
	350–550	1.16–2.32
	> 550	1.80–3.55

smectite and mixed-layer illite–smectite minerals (Wiseman and Puttmann, 2005).

There is no general agreement on the importance of clay mineral surfaces for SOC storage. It has been demonstrated that SOC stocks in kaolinitic soil profiles in the tropics are similar to those in temperate, less-weathered soils (Sanchez and Logan, 1992), despite faster decomposition rates associated with higher temperatures. (Wattel-Koekkoek et al., 2001) also showed that SOC contents of kaolinitic and smectitic soils did not differ to any significant extent in soils from seven different countries. Wiseman and Puttmann (2005) indicated that the two soil profiles with the highest SSA and percentage of clay and smectite concentration do not have significantly higher OC concentrations relative to the other soils. (Mayer and Xing, 2001), for instance, found that SSA did not correlate well with SOC in acid soils in the USA, and Hassink (1997) proposed that the surface area of clay minerals is not a good indicator for the potential of a soil to sorb SOC in tropical and temperate regions across the globe. Although many authors have described a positive correlation between SSA and SOC concentration, predicting SOC from SSA remains difficult. This may be because SOM is distributed in heterogenous patches within the soil microstructure (Steffens et al., 2017; Vogel et al., 2013), limiting the direct relationship between SSA and SOC concentration. Although recently developed regression models for estimating the SOC storage capacity included SSA (Beare et al., 2014; Feng et al., 2013; McNally et al., 2017; Zinn et al., 2007), most of these models were based on estimated SSA (Feng et al.,

2013) or calculated values from the water content of air-dried soil (Beare et al., 2014; McNally et al., 2017). An accurate determination of the SOC storage capacity based on SSA requires the accurate measurement of SSA for soil samples from a wide-range of diverse soils (Feng et al., 2013), limiting the suitability of SSA as a widely-adopted indicator for SOC storage. The suitability of mineralogy and SSA as indicators is further limited because the measurement of these parameters is labour-intensive and time-consuming. Although identification and quantification of phyllosilicates, as well as quantification of their surface properties (CEC, SSA) has been recommended to address their role in SOC turnover and stabilization (Barre et al., 2014), the widespread adoption of clay mineralogy and associated parameters as an indicator for SOM is unlikely due to laboratory constraints.

2.5.5. Oxides of aluminum (Al) and iron (Fe)

In addition to phyllosilicates, metallic sesquioxides (oxides, hydroxides, and oxyhydroxides of Al and Fe) significantly contribute to SSA and to OM stabilization in soils (Baldock and Skjemstad, 2000; Sollins et al., 1996). Additionally, Al- and Fe-oxides have a potentially greater affinity for SOM than other mineral surfaces such as clay minerals, and thus may be more important as drivers of SOC storage than other soil mineral components (Jones and Edwards, 1998; Kaiser et al., 2002; Kaiser and Zech, 1998). Fe-oxides and hydroxides have particularly high SSA (Kaiser and Guggenberger, 2003; Yong et al., 1992), resulting in positive relationships between SOC and Fe-oxide (Kaiser and Guggenberger, 2003; Kleber et al., 2005). As cations, Fe^{3+} and Al^{2+} can bind with organic ligands in acidic soils (e.g., carboxylic groups) to form organo–mineral complexes (Boudot et al., 1989; Higashi, 1983) and stabilize SOM via cation bridges (Oades, 1988; von Lützow et al., 2006) through a direct bond formation between OM and either Al^{3+} or Fe^{3+} at the surface of a short range-ordered mineral ((Duckworth et al., 2008; Mikutta et al., 2007). Whereas Fe^{3+} exists in a range of oxides with amphoteric surfaces on which organic anions are adsorbed (Oades and Tipping, 1986), the interaction of Al^{3+} with organic materials is more likely to be an electrostatic interaction (Oades, 1988). In addition to their direct role in OC sorption, Fe- and Al-oxides can help to stabilize aggregates (Rasmussen et al., 2005), resulting in occlusion of OM at the clay microstructure level ($< 20\,\mu m$) or the coating of phyllosilicates by Fe- and Al-oxides (Mayer et al., 2004), a phenomenon frequently described for kaolinite (Kanket et al., 2005; Singh and Gilkes, 1992) and supported by the widespread occurrence of kaolinite-oxides microaggregates in Oxisols (Balbino et al., 2004; Oades, 1993). Thus, soil mineralogy, in particular Fe and Al content, controls aggregate formation and stability through both mineral-mineral interactions and organo-mineral associations (Oades and Waters, 1991; Six et al., 2000).

Numerous studies have highlighted the importance of extractable Fe and Al in stabilizing SOC in top- and subsoils (Beare et al., 2014; Matus et al., 2014; McNally et al., 2017; Schneider et al., 2010; Wiseman and Puttmann, 2005), and Fe-oxides have been found to be the most relevant mineral phases for the formation of organo-mineral associations in the subsoil of acidic forest soils (Eusterhues et al., 2005). Here, it is likely that specific components of OC (O/N-alkyl C and microbially-derived sugars) are preferentially stabilized by association with Fe- and Al-minerals (Schoening et al., 2005; Spielvogel et al., 2008). For soils of volcanic origin (e.g., allophanic soils), it has been shown that extractable Al was much more important than Fe (McNally et al., 2017) or the mass proportion of fine particles or clay content (Beare et al., 2014; Curtin et al., 2016; Matus et al., 2006; Percival et al., 2000), likely a result of the dominance of Al in the fine mineral phase. As such, whether Al, Fe or clay content is more important for SOC sorption is likely a product of their presence, absence and relative proportions in a given soil.

The positive correlations of soil Fe and Al phases with SOC across a wide range of soils (Beare et al., 2014; Kaiser and Guggenberger, 2000; McNally et al., 2017; Tiessen et al., 1984) suggest that significant amounts of SOC are stabilized by interactions with Fe and Al (hydr-) oxides in soils (Bruun et al., 2010) and that Fe and Al could be suitable indicators for SOC storage. In a diverse landscape in South Africa, Fe oxides was a better predictor of SOC storage than clay content along topographic gradients and different lithologies, despite differences in mean annual precipitation (Khomo et al., 2017). However, time-consuming and costly soil fractionation procedures are required to isolate fine mineral fractions and determine their contents of Fe- and Al-oxides. Similar to clay mineralogy and SSA, sesquioxides thus represent only a moderately applicable indicator for SOC storage because of the high workload to extract them from soil samples.

2.5.6. Available Ca^{2+} and Mg^{2+} cations

The influence of the exchangeable cation concentration on the behavior of clay minerals, the biological availability of OC (Baldock and Nelson, 2000), and SOM stabilization is well documented (Clough and Skjemstad, 2000; Kaiser et al., 2012; O'Brien et al., 2015; Oades, 1988; Paul et al., 2003; von Lützow et al., 2006). Specifically, the presence of polyvalent cations in soil produces electrostatic bridging between soil mineral surfaces (clay minerals) and SOM (OM-cation-mineral: cation bridges), yielding an important mechanism of SOM stabilization (Muneer and Oades, 1989b; Stewart et al., 2008; von Lützow et al., 2006). Sorption of negatively charged organic groups through cation bridging to negatively charged clay minerals is possibly the most common mechanism that stabilizes OM against microbial decomposition (Oades, 1988; Sollins et al., 1996). In addition to the interaction with soil mineral surfaces via cation bridges, OM can be stabilized against microbial decomposition via chelation or cross-linking of OM with polyvalent cations (Kaiser et al., 2012; Masiello et al., 2004) or by promoting aggregation (Oades, 1988; Paul et al., 2003).

Although stabilization of OM due to the presence of Mg in arable soils of neutral pH has been documented (Kaiser et al., 2012), the presence of Ca, in particular, is more frequently attributed to enhanced OM stabilization. As a result, Ca has been found to be an important predictor of SOC accumulation across different landscapes and land-uses, e.g. under forests (Morris et al., 2007), pastures (Fornara et al., 2011), and cultivated soils (Kaiser et al., 2012). O'Brien et al. (2015) reported that exchangeable soil Ca^{2+} was the strongest predictor of SOC stocks in restored grasslands, indicated by a strong relationship ($R^2 = 0.74$) between Ca^{2+} and SOC, followed by clay content ($R^2 = 0.52$), highlighting the potentially large role Ca can play in stabilizing SOM. A recent analysis of 5500 soil proiles found that exchangeable calcium was a better predictor of SOC content than soil texture (Rasmussen et al., 2017). Oades (1988) reported higher OM contents and longer turnover times for OC of calcareous soils compared with adjacent non-calcareous, acidic soils, which may be partly due to Ca-humates. Other studies have also shown a decreased solubility of OC in the presence of Ca^{2+} (Muneer and Oades, 1989b), a reduced mineralization of native organic materials and organic substrates with the addition of Ca^{2+} (Muneer and Oades, 1989a, 1989b), or an increase in OM stabilization due to soil amendment with Ca (Krull et al., 2003). In contrast, Stewart et al. (2008) found no effect on the stabilization of added residue by increasing Ca^{2+} availability in agricultural soils. This is possibly due to other limitations on SOC storage in these soils. Here, the maxim that the limiting factor drives SOC storage and stabilization should also be applied.

Due to these inconsistencies, assessing SOC storage by Ca^{2+} cations is limited. Instead, the interaction of cations (Ca^{2+}, Mg^{2+}), clay content, and factors that influence biotic impacts on SOM dynamics (pH, P) with soil moisture and vegetation controls the accrual and retention of SOM (O'Brien et al., 2015). As such, the contents of the macronutrients Ca and Mg are not stand-alone indicators for SOC storage, but can be viewed in conjunction with other factors (temperature, water-availability, land-use) as an indicator for SOC storage.

3. Outlook: An indicator system to quantify current and potential SOC storage

In this review, we identified several core factors that control SOC storage at different spatial scales, which could be used to develop a system of indicators (Fig. 1). Such a system needs to consider the two most important aspects of SOC storage which are the C input and C stabilization within the soil. In the following, we outline a scheme of an indicator system that allows a time- and cost-efficient estimation of current and potential SOC storage from the local to the regional/sub-continental scale. This indicator system can be used in a wide range of soils in temperate (agro)ecosystems, but may be only partly suitable in tropical regions or for specific soils, e.g. paddy soils. As soil texture seems to be the most important factor controlling SOC stabilization, the silt and clay content represents a key indicator. Several fine fraction-based approaches were developed that allow an estimation of the SOC storage capacity of soils (Table 1). Since the precise determination of soil texture is laborious and time-consuming, an exact determination of the fine fraction content and its use as an accurate indicator may be limited. Nevertheless, soil texture can easily be estimated in the field with a relatively high precision using the “finger texturing” (Vos et al., 2016), enabling a rough estimation of SOC storage capacity if measured data is not available. Current developments in internet technology, with the launch of mobile apps to guide non-experts through soil textural classification (e.g. LandPKS app, landpotential.org) promise to further expand the everyday application of soil texture as an indicator of SOC storage. Lastly, the development of rapid, robust and low-cost estimation techniques for soil texture based on IR spectroscopy in the field or laboratory (Viscarra Rossel and Lark, 2009) will also expand the applicability of soil textural estimates as an indicator for the SOC storage capacity.

Besides the stabilizing capacity of the fine fraction, the potential SOC storage also depends on the C input level. As most promising indicators, climatic proxies, particularly elevation, as well as information on land use, land use management and/or vegetation characteristics were identified. This readily available information can be used to further specify the SOC storage capacity as it was done by Six et al. (2002a) and Feng et al. (2013) (Table 1). In this regard, regional approaches are needed to estimate the SOC storage capacity according to specific climatic conditions and land use/management/vegetation characteristics. A good example for such an approach, although not on a quantitative stock basis, is the classification of site-typical SOC contents in cropland soils of Bavaria (Germany) (Table 2) (Capriel, 2010). Using a large SOC database, site-typical SOC ranges were delineated for cropland soils according to texture classes, subdivided by elevation as a proxy for different climatic conditions in Bavaria. Such approaches can be adapted to regional environmental characteristics and expanded to cover other land uses, vegetation types, management aspects or topographic features.

In order to enhance the indicative power of such an indicator system towards impacts of land management, further “functional soil characteristics” should be included. Soil structural properties are most promising in this regard (Rabot et al., 2017; Vogel et al., 2018). Indeed, new insights emerged recently, recognizing a more important role to the physical protection of SOM (Ananyeva et al., 2013; Kravchenko et al., 2015; Strong et al., 2004; Toosi et al., 2017). The porosity, macroporosity, pore distances, pore connectivity and aggregate size classes (macro- and microaggregates, microaggregates-within-macroaggregates) were proposed as potential structural indicators (Rabot et al., 2017; Six and Paustian, 2014). Moreover, the SMB-C pool size and the SMB-C to SOC ratio are sensitive indicators that reflect cropping and management effects. Monitoring the SMB-C pool size and the SMB-C to SOC ratio can thus serve to predict SOC changes.

The sketched system of indicators including the fine fraction content, functional soil characteristics such as soil structural and microbial properties, climate proxies and land use/management/vegetation characteristics is a promising way to estimate the site-specific SOC storage potential time- and cost-efficiently. In order to estimate the fine fraction-based C sequestration potential, the actual amount of C in this fraction has to be estimated. Some authors used a fixed percentage of fine fraction C on total SOC of 77–85% (Angers et al., 1997; Di et al., 2017; McNally et al., 2017), but due to its large variability depending on land use, land use history and management, fine fraction C should be quantified. Several fractionation approaches were used to determine the actual C content of the fine fraction (Hassink, 1997; Wiesmeier et al., 2014b), but due to their complexity they are not practicable within the outlined indicator system. One solution would be a simplied fractionation approach in which the fine fraction is separated by sieving after shaking without time-consuming and expensive further separation of particulate organic matter (Trigalet et al., 2017). Such a simplified fractionation scheme is not as labour intensive as the currently used schemes and could thus be applied for soil monitoring at the regional scale. If simultaneously the VisNIR spectra are also collected, the resulting spectral library can be used to predict the fine fraction C content by scanning a bulk soil sample. Zimmermann et al. (2007) and Baldock et al. (2013) have already demonstrated the feasibility of predicting the C content in fractions using Mid Infrared spectroscopy, and Viscarra Rossel and Hicks (2015) have predicted the C content in three fractions using VisNIR spectroscopy on samples from an Australian database. Thus, IR spectroscopy-based approaches are a promising way for an efficient estimation of the current fine fraction C.

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