

Understanding and fostering soil carbon sequestration

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E-CHAPTER FROM THIS BOOK



Mechanisms of soil organic carbon sequestration and implications for management

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1 Introduction

1.1 Introduction and definitions

A major potential for increasing carbon sequestration in mineral soils is in agricultural systems under cropland use (Amelung et al., 2020). Understanding organic carbon (OC) sequestration in (mineral) soils requires considering the pathways and the associated different types of organic matter (OM) input. As pointed out in Box 1, OC sequestration refers to ‘the process of transferring CO₂ from the atmosphere into the soil of a land unit, through plants, plant residues and other organic solids which are stored or retained in the unit as part of the soil organic matter (humus)’ (Chenu et al., 2019; Olson et al., 2014). If we accept this definition, all processes are relevant that lead to a storage or retention of OC in soils. A number of mechanisms have been described that lead to the retention of OC in soils. As the OC that enters the soil is in dynamic equilibrium, all the different OM pools that are retained in a soil must be considered. SOC sequestration implies raising soil organic carbon (SOC) levels, where they are currently undersaturated, and to maintain maximal OC levels in well-managed soil systems (Lehmann et al., 2020).

Olson et al. (2014) pointed out that it is essential to strictly differentiate between the application of any of OM to soils from sources external or outside a land unit (e.g. amendments like manure, compost, biochar) and OC sequestration *sensu strictu*. Sequestration of OC in soils as defined here (Box 1) requires that atmospheric CO₂ is fixed through photosynthesis and stored in the soil. No atmospheric CO₂ is converted and stored as a result of amendment transfer and it does not add to reducing atmospheric CO₂ levels. Therefore, we will not consider the application of organic amendments in this chapter. As organic amendments may in specific cases influence SOC sequestration through their impact on plant growth and soil microbial functioning, their management is discussed in Chapter 9 of this book.

Box 1: Definitions

'We generally use the comprehensive term stabilization of OM for processes or mechanisms that lead to prolonged turnover times in soil. Stabilization is defined as protection of OM from mineralization. Stability is the integrated effect of recalcitrance, organo-mineral interactions and accessibility. Generally, stabilized OM is older than unstabilized OM, a fact that is demonstrated in longer turnover and mean residence times. Turnover times of stabilized OM cannot be given as absolute values because they are dependent on environmental conditions and soil properties (e.g. climate, soil mineralogy)' (von Lützow et al., 2006).

OC sequestration is defined as 'the process of transferring CO₂ from the atmosphere into the soil of a land unit, through plants, plant residues and other organic solids which are stored or retained in the unit as part of the soil organic matter (humus)'. Retention time of sequestered OC in the soil (terrestrial pool) can range from short-term (not immediately released back to atmosphere) to long-term (millennia) storage. Sequestration can therefore be quantified for a given duration (Chenu et al., 2019; Olson et al., 2014).

'OC storage is broader as it is defined as the increase in SOC stocks over time in the soils of a given land unit, not necessarily associated with a net removal of CO₂ from the atmosphere' (Chenu et al., 2019).

'The OC storage potential of a given soil unit may be defined as the maximum gain in SOC stock attainable under a given climate and a given timeline (e.g. time required to attain a new equilibrium or a given time period, such as 20 years for IPCC)' (Chenu et al., 2019).

'Similarly, the OC sequestration potential of a given soil would be the maximum gain in SOC allowing a net removal of CO₂ from the atmosphere under a given climate and for a specified timeline' (Chenu et al., 2019).

1.2 Organic matter input to soils

Organic C enters the soil mainly as:

- aboveground litter or crop residues,
- belowground litter or crop residues, and
- rhizodeposition.

Both above and belowground litter or crop residues are mainly composed of OC bound in large polymers (celluloses, hemicelluloses, lignin, cutin, suberin) in leaves, stems, twigs and other woody debris, or roots, with only a small contribution of low-molecular weight organic components (Kögel-Knabner, 2017). They are either deposited on the mineral soil surface, or in different soil depths as root litter. Incorporation of OC from aboveground litter occurs via bioturbation or leaching of soluble components. In contrast, rhizodeposition consists mainly of low molecular weight compounds released from roots into the surrounding soil at various depths (Fig. 1).

Long-term OC storage in soils occurs primarily when OC derived from plant biomass is stabilized in soils as soil OM. Plant biomass makes up the majority of OC input also to agricultural soils. But we have to take into

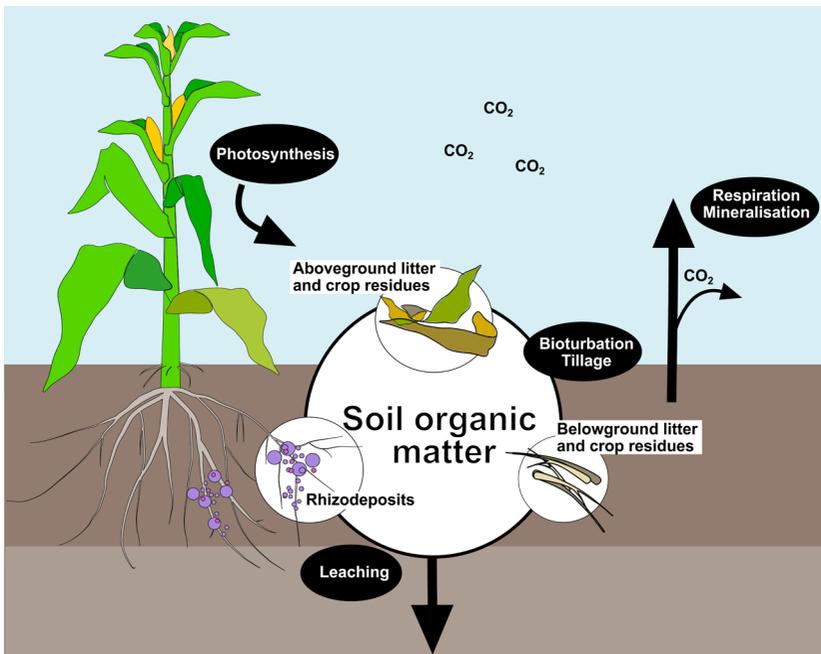


Figure 1 Aboveground and belowground OM input to soils.

account that OM is also added to cultivated soils through fertilization and waste disposal (e.g. liquid manure, compost, sludge, animal excreta, biochar, biogas digestate), which contribute significant amounts (Jacobs et al., 2020). Soils are often also contaminated with organic constituents from the petroleum and coal chemistry/industry, as well as from coal combustion, e.g. tar oil, coal dusts, black carbon, specifically in industrial-urban areas (Kiem and Kögel-Knabner, 2003; Schmidt and Noack, 2000), as well as plastics (Rillig et al., 2021). Geogenic C such as kerogen or black shale can also be inherited from the parent material (Fox et al., 2020 and references therein). This short listing demonstrates the large diversity of OM input to soils. Lehmann et al. (2020) suggest that the molecular diversity of the organic compounds rather than the material properties of individual compounds controls decomposition in soils. As pointed out above, the amendments help to increase the OC content and stocks of a soil, but may not help sequestering OC in soils. At the same time, it is important to return organic residue materials to soils, rather than burning them or using them otherwise, e.g. for energy production or production of chemicals.

A considerable portion of biomass OM reaches the soil as subsurface input, i.e. as root litter and rhizodeposition. About 30%, 50%, and 75% of the root biomass are found in the topmost 10 cm, 20 cm and 40 cm of the soil, respectively; however, the maximum rooting depth, and thus also the range influenced by root residues and rhizodeposition, is much larger. Root debris/residues and exudates also reach greater depths (down to several meters in some cases), and together with displaced/relocated OM from the topsoil, contribute significantly to SOC formation in subsoils (below the uppermost (A) horizon). In grassland and steppe soils, root-C generally contributes in greater proportion to the total carbon input than in forest ecosystems under comparable climatic conditions. In forests of cool temperate climates, the contribution of root litter to the OM input, accounts for between 20% and 50% depending on the tree species and the life form (evergreen or deciduous). Under wheat cultivation, the subsurface input represents around 25% of the total OC input, under grassland around 40%. The predominantly low-molecular, usually N-rich OM that is secreted by plant roots (rhizodeposition) contributes significantly to the OC input to the soil. Most of the root exudates are rapidly metabolized by microorganisms, and due to this constant OC source, there is a high (microbial) population density in the rhizosphere. The proportion of belowground input flux (much less documented than inputs from leaves and stems) is considered to contribute substantially to soil OM through dead roots and rhizodeposition, but the exact proportions and quantities remain a matter of debate. Quantitative estimates are complicated by the fact that a major proportion of the root exudates is transferred to the soil via symbiotic fungi (mycorrhiza) (Dijkstra et al., 2021; Zhou et al., 2020).

1.3 Resynthesis of organic compounds in soil: Formation of microbial necromass and exudates

When the OM is deposited on or in the soil and is broken down by the soil decomposer community, it is rapidly mineralized with the liberation of CO_2 , while the majority of the mineralized N is incorporated into the microbial biomass.

As described in Fig. 2, a portion of the plant OC is used by the soil (micro-) organisms for build-up of their biomass (anabolism). Microbial biomass is the mass of living microorganisms in soil. He et al. (2020) estimated the global stocks of living microbial biomass C to be 12.6 (6.6–16.4) Pg C for fungal biomass and 4.3 (0.5–10.3) Pg C for bacterial biomass in topsoil (0–30 cm), but most of the microbes are dormant or barely active. Microbial products are generally composed of the same type of (macro)molecules as plants, except for lignin, which is exclusively found in plant tissue. Microorganisms also introduce new monomers that build up the cell-wall biopolymers composed

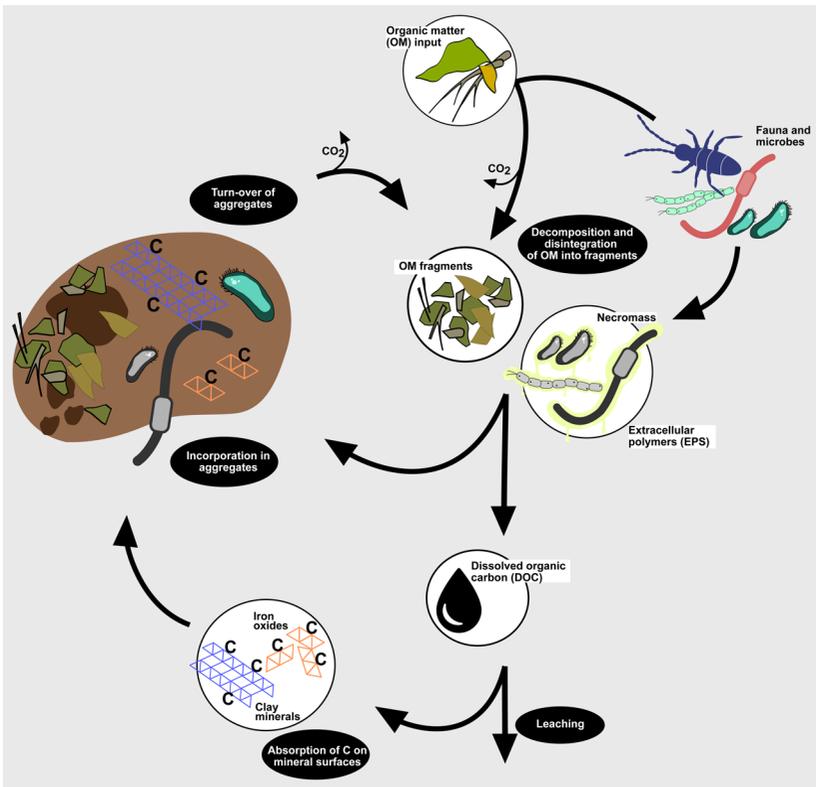


Figure 2 Flow of OC through different binding forms in soils.

of amino sugars (Kögel-Knabner, 2002). This means that the C/N ratio of decomposing plant residues decreases during the course of decomposition. The microbial necromass, which is further decomposed in soils has a narrow C:N ratio, ranging often between 4 and 8. The OC assimilated by microbes and forming their biomass is thus recycled several times through the microbial biomass (Gleixner et al., 2002; Buckeridge et al., 2020). This fact is obvious in the isotopic enrichment of SOM with ^{13}C as the microbial fractionation preferentially transfers the lighter isotope (^{12}C) into the product (CO_2) and leaves the substrate (decomposing plant tissues or individual organic compounds) enriched in ^{13}C . The proportion of OC that is used for building the microbial biomass and its exudates (so-called extracellular polymeric substances, EPS) is the C use efficiency (CUE) of the soil microbial biomass. It is often considered to be between 0.4 and 0.5, but varies in a large range (Manzoni et al., 2018; Qiao et al., 2019).

The decomposition and release of organic residues introduced to soils is to a large extent through the activity of microorganisms. The direct contribution of soil fauna to the release of CO_2 from soils is comparably small, but the magnitude of direct and indirect animal effects on SOM dynamics remains poorly quantified. The ^{13}C signature of soil macrofauna reveals the importance of microorganisms, in contrast to dead plant material, as a major food resource for the soil faunal community. Direct feeding on living plant material as well as on mycorrhizal fungi is likely rare among soil invertebrates. Plant C is taken up predominantly by saprotrophic microorganisms and channeled to higher trophic levels of the soil food web (Potapov et al., 2019). The activity of soil microorganisms is regulated by soil fauna, as they affect many aspects of OM turnover, particularly by regulating the activity and functional composition of soil microorganisms and their physical-chemical connectivity with soil OM (Erktan et al., 2020; Filser et al., 2016; Thakur and Geisen, 2019).

For large molecules comprising the major part of the plant residues (found in soils as particulate organic matter, POM) extracellular enzymes are necessary to break them down for microbial uptake and metabolism, implying that degraders and their enzymes must come into contact with substrate. The degradation reactions of organic compounds (catabolic reactions) are mainly hydrolytic or oxidative depolymerization reactions, essential for the uptake of the released low molecular weight degradation products in microbial cells. Some compounds, which are not taken up, are transported with the soil solution and/or are adsorbed onto other organic or mineral compounds. These small molecules resulting from enzymatic degradation thus contribute to the formation of organo-mineral associations (Kleber et al., 2015; Kögel-Knabner et al., 2008). In contrast to earlier views on soil OM formation, this implies that soil OM is also derived from easily decomposable organic molecules (Cotrufo et al., 2013).

The mineralization of SOC is thus a two-stage process: first, non-bioavailable forms are converted to bioavailable forms, which, only then, undergo the second process, i.e. mineralization after uptake in the cell. It is also considered that other unknown nonenzymatic processes contribute to the first step (Brookes et al., 2017). Such depolymerization reactions are not necessary for the exudates released from plant roots in the rhizosphere, which are low molecular weight compounds.

The capacity of a soil to effectively retain OC is described by the CUE, i.e. the ratio of OC remaining in a system to OC entering that system. Microbial CUE is the fraction of OC taken up allocated to biosynthesis, forming new biomass and biological products, including exudates (Manzoni et al., 2018). The efficiency of long-term OC storage in soils depends on how much OC enters protected compartments or is removed from the system by abiotic transport in vertical and/or horizontal direction (e.g. leaching or erosion).

1.4 Protective mechanisms leading to storage of organic carbon in soils

Plant and microbial residues and their transformation products can be protected from further microbial degradation through various mechanisms. These stabilization processes have an effect on the mixture of fresh and transformed plant residues resulting from decomposition processes, as well as on newly formed microbial residues. They lead to a significant deceleration of the initial rapid mineralization rate. As synthesized recently by Angst et al. (2021), OM stabilized in aggregates and mineral-associated OM is with about equal contributions of plant and microbial origin.

The action of decomposers on their organic substrates takes place in soil pores. As most bacteria are in the size range between 0.5 and 2 μm (Liang et al., 2019), the soil structure and its heterogeneity controls accessibility and biodegradation at the scale of the microbial habitat (Juarez et al., 2013). Large tracts of the soil-pore network are practically devoid of decomposer organisms, and the distribution and forms of OM are equally patchy. Physical separation between decomposer and substrate is therefore an important stabilization mechanism (Lehmann et al., 2020). The structural arrangement of the soil particles creates microsites with limited O_2 diffusion, enzymatic and microbial access (Chevallier et al., 2010; Keiluweit et al., 2017; Mbé et al., 2021; Zimmerman et al., 2004). Different microdomains were identified within an aggregate associated with either occluded POM or dissolved organic matter (DOM), and different spatial and functional properties (Steffens et al., 2017). Domains with POM are densely packed, having low porosity, and are considered to store and protect OC, whereas other domains are porous and consist of mineral grains covered with thin OM films, serving as sites of microbial

growth and OM turnover (Steffens et al., 2017). Organic compounds can be stabilized and protected from microbial degradation through interactions with minerals. This mainly involves interactions with particles in the fine fraction and leads to the formation of 'organo-mineral associations, the OM being present in the form of mineral-associated OM. If isolated from soil they are often small organo-mineral microaggregates, rather than primary particles with sorbed OM' (Chenu and Plante, 2006; Totsche et al., 2018), the differentiation between microaggregates and mineral-associated OM being impossible at the smallest scales. Depending on the pedogenetic environment controlled mainly by parent material and climate, different pedogenic minerals form (Kögel-Knabner and Amelung, 2021), which are highly reactive and provide a large specific surface area for interaction with different organic compounds. In addition, dissolution of metals, and the synthesis of organo-metallic phases and their association with clays may also add to protection of OM from decomposition (Basile-Doelsch et al., 2015; Wagai et al., 2020). The contribution of these different processes in combination with protection mechanisms, such as the associations with mineral surfaces, and the physical isolation of decomposers and substrates, strongly depend on soil properties and site conditions, leading to a wide range of OC turnover rates in different soil types.

1.4.1 Particulate organic matter

Plant residues in or transported into the mineral soil are an easily available carbon and nutrient source and are thus immediately attacked and degraded by soil fauna and microorganisms. They thereby act as nuclei for the formation of soil aggregates (Bucka et al., 2019; Witzgall et al., 2021) and at the same time are stabilized against degradation. One way of stabilizing otherwise readily decomposable substrates in the soil can result from the spatial separation of the substrate from the decomposers, which thus prevents degradation (see above). They can be isolated from the mineral soil as POM, existing in different stages of decomposition. Although all organic compounds of plant or animal origin entering the soil are prone to decomposition, the more recalcitrant compounds are relatively enriched in older/stabilized POM fractions while the polysaccharides are decomposed (Baldock et al., 1992; Leifeld and Kögel-Knabner, 2005; Mueller and Koegel-Knabner, 2009). Bacteria and fungi attack the plant particles already comminuted by soil fauna and remain attached to their surfaces, together with fine mineral particles (Chenu and Cosentino, 2011; Tisdall and Oades, 1982). The proliferating microbes excrete EPS during growth and metabolism (Costa et al., 2018; Flemming and Wingender, 2010). Those patches of microbes and their EPS are sticky and adhere as gluing agents to surrounding minerals leading to soil structure formation by cross-linking of particles (Chenu and Jaunet, 1992; Gaillard et al., 1999; Rabbi et al.,

2020; Young and Crawford, 2004). This leads to the formation of aggregated organo-mineral compounds inside the aggregates and as the pores are filled with growing organisms and their exudates, degradation ceases, and the partly degraded POM particles are inaccessible to other microorganisms (McCarthy et al., 2008). This is supported by results from (Steffens et al., 2017). Different microdomains were identified within an aggregate associated with either occluded POM or DOM, and different spatial and functional properties. Domains with POM are densely packed, having low porosity, and are considered to store and protect OC, whereas other domains are porous and consist of mineral grains covered with thin OM films, serving as sites of microbial growth and OM turnover. Smaller POM particles have a larger specific surface area compared to larger ones, which provides a higher potential contact area with mineral particles and colonizing microbes. With increasing degree of decomposition, the POM isolated has a more narrow C:N ratio, due to the accumulation of N-rich microbes and their necromass and exudates, acting as glue between the surface of the POM particles and fine-grained soil minerals. This induces particle assembly around the POM particles and the formation of aggregates (Witzgall et al., 2021). Within time frame of decades and turnover of the aggregates, the particulate plant residues become available again for microbial decay and thus get more and more degraded. The amount of OC stored in a soil as POM is the net result of stabilizing and destabilizing processes. Release of POM from physical occlusion can occur through processes such as tillage, bioturbation, or freeze-thaw and wetting-drying cycles (Bailey et al., 2019).

1.4.2 Interactions of organic matter with mineral surfaces

The predominating minerals forming associations with OM in soil environments include metal oxides, hydroxides, and oxyhydroxides, phyllosilicates, and short-range ordered aluminosilicates. A detailed assessment of the mineral surfaces found in soils and the mechanisms of interaction is given by Kleber et al. (2015, 2021).

Phyllosilicates provide permanently charged surfaces in external and interlayer siloxane surfaces with significant isomorphous substitution of lattice cations, resulting in a pH-independent and permanent net negative charge of the siloxane surfaces. Pedogenic Fe and Al can be present in monomeric form by chelating with organic ligands or in polymeric form as polynuclear complexes and as secondary minerals. The latter includes Fe and Al oxides, hydroxides, and oxyhydroxides as well as short-range-order aluminosilicates (allophane, imogolite, and proto-imogolite) that have high sorptive capacity for OM due to their small size (down to several nanometers) and high surface reactivity via surface hydroxyl groups. Since phyllosilicates possess variable-charge surfaces at crystallite edges, which behave similarly to the ones found

on metal oxides, inner- and outersphere adsorption as well as H-bonding of organic ligands may also contribute to OM adsorption to phyllosilicates (Kleber et al., 2015). The extent of permanent negative charge of siloxane surfaces varies with the amount of isomorphic substitution, phyllosilicates thus differ considerably in their ability to bind OM. Protective effects of phyllosilicate clays and pedogenic metal phases on OM are likely to occur concurrently, and possibly synergistically (Wagai et al., 2020).

The organic compounds associated with minerals range from low-molecular weight compounds to large polymeric molecules or cell wall residues. They originate from both plant and microbial origin. Data compiled by Angst et al. (2021) show that plant-derived compounds account for ~50% of the SOM in aggregates and mineral-associated OM in cropland soils, and even more in forest soils. They found that soil types less conducive to the proliferation of microbes, such as Podzols or Alisols, may contain relatively more plant compounds than soil types with favorable conditions for microbial growth, such as Chernozems and Luvisols. Grasslands, in particular, with a continuous input of plant material, low C:N ratios and ratios of lignin to N in plants, and a higher soil pH, seem to favor the accumulation of microbial residues in stabilized SOM as compared to forest or agricultural soils. Thus, if the clay fraction is isolated from the soil and its composition is investigated, mainly alkyl compounds and polysaccharides are found, however, only low concentrations of aromatic compounds, such as strongly transformed lignin fragments. The composition of mineral-associated material may vary not only in different soil types but also with soil depth (Rumpel et al., 2015). This is most probably associated with the microbial processing and exchange of DOM with increasing depth (Roth et al., 2019). Bound OM is also rich in carboxyl and N-containing groups, consistent with the low C:N ratio and a microbial origin of its components.

Low-molecular weight compounds include alcohols, sugars, amino acids, and amines, as well as simple aromatic compounds such as benzene, phenols, etc. that are bound via sorption processes. EPS are released by microbes and strongly associate with mineral surfaces. They are highly variable, hydrated polymers composed of polysaccharides, proteins, nucleic acids, and lipids (Costa et al., 2018). EPS are responsible for the cohesion of microorganisms and adhesion of (groups of) microorganisms to surfaces. It is important to understand that many microorganisms can form supracellular structures comprised of surface-associated microbial cells embedded in hydrated EPS that facilitates adhesion and survival (Cai et al., 2019; Flemming and Wuertz, 2019). Foster (1988) already described small groups of cells, sometimes coated with clay particles, and embedded in partially degraded OM or surrounded by EPS. A planktonic lifestyle (i.e. free swimming) is not possible in the unsaturated zone and most bacterial cells are restricted in their motility and found as small colonies attached to surfaces, protecting themselves from

changing hydration conditions using a matrix of EPS (Kleber et al., 2015). This implies that microbial residues are often already attached to mineral surfaces in soils when building necromass. Buckeridge et al. (2020) describe that necromass added to soil strongly binds to mineral surfaces. According to the estimates by Liang et al. (2019) up to around half of the total SOC is microbial necromass. How much of this is associated with mineral surfaces is presently not known.

The binding between minerals and organic compounds can take place through various mechanisms and depends on the one hand on the type of mineral and its surface charge, and on the other, on the type and charge of the functional groups in the OM. Therefore, the pH value and the base saturation are of great importance for the development of organo-mineral compounds (Rasmussen et al., 2018). The binding modes that contribute to the stabilization of organic ligands at mineral surfaces include specific and nonspecific adsorption, H-bond formation, van der Waals interactions, and hydrophobic interactions (see Kleber et al. (2015) for a detailed description). Due to the different mineral composition of soils and the associated large variability of properties several binding mechanisms operate simultaneously. The decomposability of OM can also be strongly reduced by complex formation with metal cations. In this case, the bonding of metal cations (Ca, Al, Fe, heavy metals) onto the OM inhibits attack by enzymes. Soluble complexes of Fe and Al with organic ligands can be precipitated especially in acidic, OM-rich environments such as volcanic and podzolic soils (Wagai et al., 2020). These secondary mineral phases such as allophanes, Fe and Al oxyhydroxides form in the presence of OM via coprecipitation. The size of these mineral phases ranges from one to a few dozen nanometers (Eusterhues et al., 2008; Kleber et al., 2015; Levard et al., 2012; Tamrat et al., 2019; Torn et al., 1997), as the recrystallization ('aging') of such pedogenic Al and Fe oxides can be delayed or prevented if adsorbed or coprecipitated organic compounds block the growth positions (crystallization nuclei) of the crystals or form very stable complexes with the Al and Fe ions.

Evidence of the high stability of organo-mineral compounds is reflected by the fact that the temperature of thermal decomposition of OM increases after bonding onto the mineral phase, and at the same time, the decomposition rates of these bound organic substances are clearly lower compared to free OM. The radiocarbon data compiled by Kleber et al. (2015) show that turnover times of OC stored in association with minerals are on average four times longer than those of OC in free or occluded OM, and range from decades to millennia. Mineral-associated OM is in a state of dynamic equilibrium with DOM, resulting in a constant exchange between the soil solution and mineral surfaces. Thus, a significant fraction of the mineral-associated OM is released into the soil solution and becomes available for microbial uptake and mineralization. This

process has been suggested as regulatory gate for microbial activity in soils (Kemmitt et al., 2008).

The formation of organo-mineral associations depends on the presence, but also the accessibility of mineral surfaces for interaction with OM. Interaction of mineral surfaces with DOM is constrained by soil texture and the associated soil-pore system, as preferential flow paths restrict the direct interaction of DOM with mineral surfaces. Changes to the soil environment that affect the solubility of minerals or surface charges of minerals including increased pH or in the reductive dissolution of Fe-bearing minerals can lead to a release of sorbed OM (Kleber et al., 2015). Turbation processes and specifically bioturbation may enhance the exposure of mineral surfaces for interaction with OM. Earthworms induce the formation of organo-mineral associations and at the same time create closer association of partly degraded OM and iron oxides (Barthod et al., 2020; Vidal et al., 2019).

1.4.3 Isolation of particulate organic matter and organo-mineral associations

Physical fractionation, based on a separation resulting from differences in density or particle size, is used to separate plant residues from organo-mineral compounds (Poeplau et al., 2018). It is generally performed following pretreatment consisting of agitation or ultrasound to destroy aggregates. The obtained fractions are then separated using sieves, sedimentation, and/or density fractionation (Christensen, 2001; Oades, 1988). During density fractionation, the sample is suspended in a liquid with a density between 1.6 and 2.4 g cm⁻³ (organic solvent mixtures or inorganic salts), so that the light, floating organic material can be separated. For particle size fractionation, the material in the >63 µm fraction (sand fraction) is separated, which can be further fractionated by sieving. The organo-mineral compounds <63 µm are separated using sedimentation in water, and the <2 µm fraction can be further separated via centrifugation. Fractionation according to the particle size and density are often combined. Such combined particle size and density fractionation approaches are used to isolate free, and occluded POM as well as mineral-associated OM. The POM is found in the light or coarse fraction, occluded POM after dispersion of aggregates. If the aggregation is selectively destroyed, the occluded POM, mainly plant residues at an advanced state of decomposition, can be isolated. It is most often not possible to isolate the 'pure' POM particles, as some mineral material is strongly attached to the free, but especially to the occluded POM fraction. This leads to the fact that the OC concentration of occluded POM ranges between 200 mg g⁻¹ and 300 mg g⁻¹ fraction. The heavy and fine fractions mainly contain mineral-associated OM. Recently, simple fractionation protocols for agricultural topsoils have been proposed

and tested for routine analysis of large sample sets to obtain indicators for soil management and OC input strategies (Just et al., 2021; van Wesemael et al., 2019).

1.5 Organic carbon sequestration in soil compartments with high and specific input: detritosphere and rhizosphere

The detritosphere and the rhizosphere are soil microenvironments characterized by a high input of OM. The detritosphere is defined as a zone of recognizable plant and animal detritus undergoing decay and the soil in their vicinity. It represents a small volume of the soil, as it is limited to the soil adjacent to the plant residues, where most of the microbial activity and biogeochemical processes take place. The detritosphere constitutes an area a few millimeters thick around decomposing residues, and most of the OC transferred from the residues to the soil via diffusive or advective transport is found in this zone (Gaillard et al., 1999; Marschner and Rengel, 2012; Védère et al., 2020). In the first stages of decomposition, most of the residue OC mineralization takes place on the residues themselves (Védère et al., 2020). The incorporation of plant residues into soil leads to the sequestration of OM and the formation of mineral-associated OM in aggregates within weeks of decomposition (Bucka et al., 2019). Detritus can be located on top of the soil or within the mineral soil. These fresh litter surfaces serve as hotspots of microbial activity driving the formation of organo-mineral associations together with comprising a nucleus for aggregate formation. Both processes induced by plant detritus, i.e. the occlusion of POM in soil aggregates and the association of OM with mineral surfaces, strongly rely on the spatial proximity of particulate litter and its surfaces, microbial residues, and fine-sized mineral particles (Witzgall et al., 2021). Detritus is distributed in the mineral soil via bioturbation, peloturbation in Vertisols or cryoturbation in permafrost soils. In cropland soils, the incorporation of detritus is supported by ploughing operations. Most often this is limited to the topsoil (0–30 cm), but deep ploughing may introduce topsoil and detritus to larger depths (Alcántara et al., 2017, 2016). Ploughing converts subsoil material with low OC contents into topsoil with higher OC concentrations and renders a larger soil volume and its associated mineral surfaces available for storing OM (Wiesmeier et al., 2014b). Thus, extending the detritosphere into larger soil depths may enhance OC sequestration. This is in contrast to the concept of no or reduced tillage which relies on bioturbation for incorporation of plant residues into the mineral soil. Here the OC increase in the uppermost topsoil is most often counterbalanced by a release of stored OC in deeper soil layers, leading to a low to nonsignificant effect on SOC stocks over the soil profile (Chenu et al., 2019). However, it is important to note that ploughing also leads to tillage erosion (Chapter 5 of this book), the disruption of aggregates and

release of CO₂ to the atmosphere, until a new vegetation is established. These processes must be carefully balanced for a given soil and agricultural system to enhance SOC sequestration or preventing SOC loss.

Plant roots affect OC dynamics in soils by providing OC primarily in the forms of root litter and rhizodeposition (Fig. 1), thus promoting OC sequestration (Dijkstra et al., 2021; Rasse et al., 2005). Root depth distribution is thus an important plant trait that influences OC sequestration, specifically in the subsoil (Jobbagy and Jackson, 2000; Poirier et al., 2018; Thorup-Kristensen et al., 2020). However, roots can also promote SOC mineralization, exposing previously protected OC to microbial decomposition and leading to a loss of SOC. The role of plant roots contributing to SOC stabilization has been often studied in isolation from their role in contributing to destabilization. As pointed out by Dijkstra et al. (2021), there are results from studies on SOC stabilization by plant roots and their rhizodeposition suggesting that roots stabilize or increase SOC, and reduce SOC loss, whereas, other studies show that plant roots or their exudates accelerate SOC mineralization (Keiluweit et al., 2015; Li et al., 2021), often referred to as rhizosphere priming. The contribution of roots to OC sequestration has to be differentiated for root litter and rhizodeposition (Kögel-Knabner, 2000, 2017). The OC inputs from these sources may be retained preferentially in different SOC pools and by different mechanisms (Villarino et al., 2021). Root necromass in soils is a major OC input to the soil (Rasse et al., 2005; Sokol and Bradford, 2019), creating new detritosphere regions with all implications for OC sequestration resulting from the formation and storage of POM year after year. Rhizodeposition is easily taken up by microorganisms and its OC remaining in the soil ends up as microbial necromass and microbial EPS, finally contributing to the mineral-associated OM (Villarino et al., 2021), after internal microbial recycling. Increasing OC allocation to roots and rhizodeposition may thus be an important tool for increasing SOC storage.

2 Potential of different soils for carbon sequestration

The efficiency of OC sequestration in soils is largely affected by site conditions. The build-up and decomposition of SOM occur as nonlinear processes and the turnover of SOM depends on the amount of SOM (Chenu et al., 2019; Poeplau et al., 2011). This relationship is site- and soil-specific: SOM turnover is generally higher in coarse-textured soils and under warmer climates than in fine-textured soils and at lower temperatures. Therefore, the SOC storage potential clearly differs among different soils and depends on several site characteristics such as parent material, soil type, texture, mineralogy, and climate.

At the same time, it is relevant to identify soils with high SOC stocks and to maintain or even improve their SOC sequestering capacity through increasing OC inputs and reducing OC loss from these soils. Similarly, a high

OC sequestration potential is identified for soils that are depleted in OM, due to use as grazing land or cropland (Sanderman et al., 2017b).

2.1 Identifying soils with high organic carbon sequestration potential

In general, SOC accumulation depends on pedogenic soil differentiation in terms of the formation of reactive mineral surfaces, soil aggregation, and translocation processes (Kögel-Knabner and Amelung, 2021). Soil texture plays a decisive role as fine-textured soils with high silt and clay contents can have considerably higher SOC stocks than coarse-textured soils (Arrouays et al., 2006; Hassink, 1997; Zinn et al., 2007). Moreover, baseline SOC stocks are important, as the efficiency of OC sequestration is highest in soils with a relatively low SOC level and a corresponding high SOC storage potential (Chenu et al., 2019).

Soil type/soil group is important for understanding and quantifying the sequestration of OM (Keyvanshokouhi et al., 2019; Mayer et al., 2019; Soucémariadin et al., 2018; Wiesmeier et al., 2015). The distribution, content, and stocks of OM differ between major soil groups as differentiated according to the World Reference Base for Soil Resources (WRB) (IUSS Working Group WRB, 2015). As already described by Batjes (1996) large proportions of SOC are stored below 1 m, which requires consideration in relation to pedogenic OM formation and translocation mechanisms. Yost and Hartemink (2020) report that many investigations do not report soil depth and do not consider the subsoil. As estimations of the areas covered by the Reference Soil Groups are still vague, our understanding on the OC state of soil groups and their changes has to rely on very rough estimates of soil changes (FAO, 2015).

Whereas some mechanisms of OM accumulation occur in many soils, some are specific for a soil group (Kögel-Knabner and Amelung, 2021). While sorptive preservation mechanisms occur more or less in all soil orders, though largely in those with larger portions of reactive clay minerals and oxides, physical protection mechanisms dominate in the steppe soils (Chernozems, Phaeozems, Kastanozems) and are relevant also in Nitisols, Ferralsols, Cambisols, and Luvisols. Some mechanisms are related to specific pedogenic processes of OC accumulation and translocation. This refers to Cryosols, Vertisols, Luvisols, Podzols, but also some other soil groups have specific translocation mechanisms into the subsoil (Kögel-Knabner and Amelung, 2021). These pedogenic OM formation mechanisms and their effects with respect to climate change are only considered for some soil groups (e.g. for Cryosols) in the current discussions on the role of soils in the global C cycle. In addition, we are lacking information on the concentration and stocks of OM in many soils, and specifically for depths below 30 cm. There is increasing evidence that OC storage in subsoils is controlled by soil group or associated parent material,

while land use type and soil management often only affects topsoil OC stocks (Chen et al., 2019; Mayer et al., 2019; Poeplau et al., 2020). Moreover, most studies on OM stabilisation mechanisms have been carried out in temperate regions with acid soils, while tropical and arid region with alkaline soils received less attention. The specific processes occurring in these soils and in particular the interaction between organic and inorganic carbon compounds are less known (Chapters 7 and 23 of this book).

Building and maintaining high SOM levels in soils has to rely on high or increasing plant inputs (Amelung et al., 2020). High plant inputs foster microbial activity, and in turn accumulation of microbially derived OC through the accumulation of microbial-derived compounds, including necromass. This process complex implies that SOC sequestration can be improved via increasing productivity in soils (Sanderman et al., 2017a), which is in turn specifically promising in soils with presently low yields. It is therefore essential to provide soil information containing localized information on soil group, degradation status, crop yield gap, and the associated OC sequestration potentials (Amelung et al., 2020). As summarized by Kögel-Knabner and Amelung (2021) only for about 52% of the total area, soil group-specific OC stock data estimates are available. From these considerations, it becomes evident that there may also be a large global potential to sequester OC in soil groups that cover large areas, even if their capacity to store OC is limited and they are only partially suitable for plant production. Arenosols and Leptosols cover 8.7% and 11.1% of the total soil area, but their SOC stocks are not known (Kögel-Knabner and Amelung, 2021) (see, e.g. review by Yost and Hartemink (2020) for sandy soils). It also becomes obvious that soil-specific land-use (e.g. cropland versus afforestation) and management practices for OC sequestration need to be evaluated and adjusted. Degraded soils with low initial SOC stocks could be consequently targeted to foster C sequestration (Lal, 2018). Positive feedbacks of measures to enhance SOC on soil fertility and thus yields and OC inputs in degraded soils could in turn contribute to SOC accrual. Moreover, fine-textured subsoils with a high protective capacity and relatively low SOC contents could efficiently store additional SOC (Balesdent et al., 2018; Chenu et al., 2019).

2.2 Organic carbon saturation concept

In the course of global efforts to foster C sequestration in soils, information on the potential of soils to sequester additional C is required with regard to a prioritization of land management practices to increase SOC stocks (Minasny et al., 2017). There is growing evidence that soils have a finite capacity to store C as their potential to stabilize SOM against microbial mineralization may be limited by the amount of fine mineral particles (Baldock and Skjemstad, 2000; Chung et al., 2008; Six et al., 2002; Stewart et al., 2007). The OC saturation concept

is based on the assumption that there is an upper limit of stable OC storage via interaction of SOM with mineral surfaces as quantitatively most important stabilization mechanism in most soils (Arrouays et al., 2006; Oades, 1988; Sollins et al., 1996; von Lütow et al., 2006). The capacity of soils to preserve SOC may be determined by the amount of the fine mineral fraction (<20/<53 μm , medium/fine silt and clay) as proxy for mineral surfaces. In his pioneer work, Hassink (1997) proposed an equation to estimate the maximum OC storage capacity related to the fine fraction <20 μm based on a range of uncultivated and grassland soils in temperate and tropical regions. According to this equation, the OC sequestration potential (corresponding to the OC saturation deficit) can be calculated as the difference between the maximum OC storage capacity and the actual amount of OC in the fine fraction, which can be determined by fractionation. The OC saturation concept was further expanded to other land uses and climatic regions and specific equations were proposed for cropland, grassland, and forest soils, different clay types (1:1 vs. 2:1 clays) and size definitions of the fine fraction (<20 μm vs. <53 μm) (Beare et al., 2014; Feller and Beare, 1997; Feng et al., 2013; Six et al., 2002; Wiesmeier et al., 2015; Zhao et al., 2006). Further refinements of the OC saturation approach were proposed with regard to the incorporation of additional parameters such as specific surface area, extractable Al and pH as well as boundary line or quantile regression analysis of the data (Beare et al., 2014; Feng et al., 2013; McNally et al., 2017).

The OC saturation concept was frequently used to estimate the OC sequestration potential of agricultural soils (Angers et al., 2011; Chan, 2001; Chen et al., 2019; Six et al., 2002; Sparrow et al., 2006; Wiesmeier et al., 2014a; Zhao et al., 2006). However, the suitability of this concept was discussed controversially. A critical aspect of the OC saturation concept is the fact that it is not suitable to estimate the total OC sequestration potential of soils as it only refers to mineral-protected OC and does not include POM, which may potentially accumulate indefinitely (Cotrufo et al., 2019) and OC in the sand fraction (Barré et al., 2017; Chen et al., 2019). Thus, a considerable proportion of total SOC may be not taken into account, although most agriculturally used soils contain relatively large proportions (often >80% of total SOC) of OC in the fine fraction (Beare et al., 2014; Chen et al., 2019; Wiesmeier et al., 2014a). Moreover, the OC saturation concept refers to the biophysically determined SOC storage capacity independent from management aspects and economic or policy constraints and may thus not be suitable to estimate the 'effective' OC sequestration potential of soils that can actually be achieved by means of improved management (Castellano et al., 2015). Therefore, alternative 'data- or model-driven' approaches were proposed that provide references for maximum SOC storage derived from highest SOC stocks that can be achieved in specific regions/soils under given land use and management practices (Barré et al., 2017; Chenu et al., 2019). In order to assess the applicability of the OC

saturation approach for an estimation of the SOC sequestration potential, field studies on managed soils with varying degrees of OC saturation are needed.

3 Soil management options for organic carbon sequestration

3.1 Increasing organic carbon input vs. decreasing organic carbon output

The storage of SOC depends on a variety of pedogenic, biological, topographic, and climatic properties depending on the spatial scale considered (Wiesmeier et al., 2019; Chapter 5 of this book). Under long-term constant environmental conditions, SOC stocks approach a dynamic equilibrium of OC input (e.g. harvest residues, organic fertilizers, roots, rhizodeposition) and OC output (by mineralization or erosion). In order to foster OC sequestration in soils, the soil management can be changed in a way to either increase OC inputs to the soil by improved land use/management practices or to decrease OC outputs by increasing the retention time of OM through addition of stable OM (e.g. biochar) or increasing the interaction with soil minerals. What is crucial in this regard is to avoid or at least reduce trade-offs with other soil functions/ ecosystem services, e.g. reduced biomass production/food security or emission of greenhouse gases such as N₂O (Guenet et al., 2021) that potentially offsets SOC gains (Chapter 6 of this book).

3.1.1 Management practices to increase organic carbon input

As the type of land use/management largely controls OC input in soils, land use changes or improved management of agricultural soils proved to be successful strategies to increase OC inputs and SOC sequestration in mineral soils.

As there are substantial differences of SOC stocks among main land use types, a land use change from intensively managed land use systems such as croplands to more extensively managed land uses such as grasslands or forests normally results in a long-term increase of SOC stocks (Guo and Gifford, 2002; Poeplau et al., 2011; Post and Kwon, 2000; Schulp and Veldkamp, 2008). The conversion of cropland to permanent grassland is an efficient way to sequester C in the long-term (Freibauer et al., 2004; Post and Kwon, 2000; Smith, 2004; Soussana et al., 2004). Converting cropland to forest normally results in lower OC sequestration than its conversion to grassland, particularly in temperate environments, and may lead to a relative shift from stable to labile SOC, particularly when coniferous species are used (Paul et al., 2002; Poeplau and Don, 2013; Post and Kwon, 2000; Smith, 2004). For afforestation of grassland, no globally consistent effect on SOC stocks was observed, while in temperate environments SOC tended to decrease (Guo and Gifford, 2002; Paul et al.,

2002; Poeplau et al., 2011; Post and Kwon, 2000). While OC sequestration by land use extensification was mainly attributed to increased OC inputs, other factors that increase OC retention time such as reduced tillage/soil disturbance probably also play a role. In fact, evidence was found that total OC inputs do not differ between cropland and grassland soils in temperate regions and that OM quality (more root-derived OC inputs in grassland soils) rather than quantity drives differences in SOC stocks between land use types (Jacobs et al., 2020).

For cropland soils, there are different management practices that increase OC inputs and proved to sequester OC in the long-term (Chapter 16 of this book). In this regard, the design of crop rotations is decisive to maximise OC inputs. Generally, increasing the diversity of cropping systems has been shown to be beneficial for the build-up of SOC (Jarecki and Lal, 2003; McDaniel et al., 2014; Tiemann and Grandy, 2015). In particular, the integration of legumes, ley grassland, cover crops, undersown crops, deep-rooting crops, and perennial crops into cropping systems leads to sequestration of C, mainly by increased belowground OC inputs (Bolinder et al., 2012; Ledo et al., 2020; Martin et al., 2020; Paustian et al., 2019; Poeplau and Don, 2015; West and Post, 2002). An increased retention of crop residues is another strategy to enhance the OC input into agricultural soils and their SOC stocks. In particular, the retention of aboveground harvest residues such as stubble, beet leaves, and straw or their return via organic fertilizers (slurry, manure, or digestate) is an efficient way to increase OC inputs and thus SOC stocks (Lal, 2005; Liu et al., 2014; Mann et al., 2002; Powlson et al., 2008). In this regard, it should be mentioned that the application of external organic fertilizers does not result in OC sequestration if they return to land anyway, as in this case, OC sources are only relocated and locally concentrated. Organic fertilizers would only directly contribute to soil carbon sequestration, if the alternative were burning as in the case of straw in many region of the world. They would indirectly contribute to OC sequestration, when crop yields and thus OC inputs increased because of improved chemical and physical soil properties (Paustian et al., 2019; Sykes et al., 2019).

Agroforestry systems that integrate woody species in agricultural land provide diverse ecosystem services and increase SOC stocks mainly due to increased above and belowground OC inputs (Chapter 20 of this book). Several global meta-analyses provided evidence for OC sequestration by agroforestry systems, particularly in tropical and subtropical regions, but also in temperate environments (Cardinael et al., 2018; Chatterjee et al., 2018; De Stefano and Jacobson, 2018; Feliciano et al., 2018; Hübner et al., 2021; Kim et al., 2016; Mayer et al., 2022; Shi et al., 2018). The introduction of trees in agricultural systems increases the overall OC input by pruning residues, litterfall, root turnover, and rhizodeposition within the sphere of influence of trees (Cardinael et al., 2018; De Stefano and Jacobson, 2018). Indications were found that root-derived OC inputs are deposited even in deep subsoils as tree

roots may grow deeper in agroforestry system compared to forest ecosystems (Cardinael et al., 2015; Germon et al., 2016). Moreover, SOC increases can be attributed to OC inputs of understorey vegetation within tree rows of agroforestry systems as well as increased yields on adjacent agricultural land due to improved microclimate conditions (Cardinael et al., 2018; Lorenz and Lal, 2014). Although increased OC inputs may be the most important factor for OC sequestration in agroforestry systems, other processes leading to better microbial functioning due to enhanced soil quality (Guillot et al., 2021), and/or an increased retention of OM through reduced soil disturbance and erosion and reduced decomposition of recalcitrant tree litter may be relevant.

For grassland ecosystems, several improved grassland management practices are discussed in order to increase SOC stocks mainly via increased root-derived OC inputs (Chapter 18 of this book). Most importantly, improved grazing management such as optimized grazing intensity or holistic grazing systems may increase SOC stocks (Conant et al., 2017), although effects of grazing on SOC are highly context-specific and grazing management has to be locally adapted in order to contribute to C sequestration (McSherry and Ritchie, 2013). Particularly in semiarid regions, reduced grazing intensity or spatial/temporal grazing exclusion leads to increased above and belowground OC inputs and increased SOC stocks (Hobley et al., 2018; Wang et al., 2018). In temperate regions, optimized grassland management intensity in terms of the frequency of cutting and fertilization may affect SOC storage, but evidence for that is scarce (Soussana et al., 2004; Ward et al., 2016). Moreover, plant diversity in temperate grasslands seems to be positively correlated with SOC storage, mainly due to increased rhizosphere OC input (Lange et al., 2015; Yang et al., 2019). Potential positive effects of management practices such as sward renewal and integration of deep rooting and/or leguminous species on belowground OC inputs and SOC storage have to be further studied (Whitehead, 2020; Whitehead et al., 2018). Moreover, these positive effects must be balanced with trade-offs in form of N₂O and/or CH₄ emissions from livestock to evaluate the contribution SOC sequestration in grazing systems can make to GHG emission mitigation (Chapters 6 and 18 of this book).

3.1.2 Approaches to decrease organic carbon loss

Besides strategies to increase OC inputs into agricultural soils, there are several management practices that can contribute to OC sequestration by reducing the mineralization of OM and increasing its retention time in mineral soils. This can generally be achieved by increasing the spatial extend of the detritosphere-mineral soil interface through incorporation of OM deeper in the mineral soil by inversion tillage, or by increasing the available mineral surface for interaction with OM by amending the soil with fine-sized minerals. Another approach is the addition of stable OM, such as biochar.

In general, a deepening of cropland topsoils by tillage may increase OC inputs in deeper parts of the mineral soil and promote the formation of organo-mineral associations, as it was indicated by relatively high proportions of mineral-associated OM in cropland topsoils compared to other land uses (Stemmer et al., 1999; Wiesmeier et al., 2014a). As a result, SOC stocks may be increased in the long term, as it was observed in temperate cropland soils (Don et al., 2009; Nieder and Richter, 1986; Wiesmeier et al., 2012). Moreover, deep inversion tillage in soils with compacted subsoil layers (e.g. hard or plough pan) or for pasture renewal may be an effective measure to foster C sequestration (Alcántara et al., 2016, 2017; Lawrence-Smith et al., 2021; Schiedung et al., 2019). The burial of OC-rich topsoils (usually below 60 cm depth) results in reduced decomposition of OM, while fresh OC inputs in exposed subsoils depleted in C may be effectively stabilized. Reduced tillage or no-till systems are practiced around the world to control erosion and improve soil structure, water holding capacity, and soil macrofauna by increased soil cover and decreased soil disturbance. Moreover, it is commonly assumed that under reduced tillage/no-till systems soil aggregation and thus the physical protection of OM is improved, while conventional tillage leads to a disruption of (macro) aggregates and increased mineralization of OM (Balesdent et al., 2000; Beare et al., 1994; Six et al., 1998). Although some studies reported increased SOC stocks in topsoils as a result of reduced tillage/no till (West and Post, 2002), there is growing evidence from several studies and meta-analyses that when subsoil layers are considered, reduced tillage/no-till systems do not necessarily contribute to a net OC sequestration but rather induce a vertical redistribution of C in the soil profile (Haddaway et al., 2017; Luo et al., 2010; Meurer et al., 2018; Powlson et al., 2014). In reduced tillage/no-till systems, SOC stocks are often higher in the upmost topsoil (usually in 0-10 cm) compared to conventional tillage systems, as input of OC is concentrated in this depth. However, in the lower topsoil SOC stocks are lower as the incorporation of OM in this depth is reduced (except for root-derived OC inputs).

Amending soils with minerals to increase their protective capacity is a promising strategy to enhance the retention of OC inputs, particularly in sandy soils with a naturally low SOC storage potential. In Australia, sandy topsoils are often amended with clay to improve their chemical and physical properties, reduce soil erosion and nutrient leaching, and increase SOC stocks (Cann, 2000; Churchman et al., 2014; Churchman et al., 2020; Hall et al., 2010; Schapel et al., 2018). Besides the incorporation of external clay-rich material into agricultural topsoils, mixing subsoil clay into sandy topsoils ('delving') is a common practice (Betti et al., 2015; Churchman et al., 2020). The amendment of soils with crushed silicate rocks such as basalt is discussed as option to foster the sequestration of inorganic C, as enhanced weathering of silicate rocks in soils liberates Ca and/or Mg that delivers stable carbonates that either remain in the

soil or are leached to the oceans (Beerling et al., 2020, 2018). Besides silicate rocks, also silicate by-products from industrial processes such as steel slag can be used (Reddy et al., 2019). Despite potential risks with regard to the release of metals and persistent organic compounds, further benefits in terms of crop production can be expected due to soil pH increases and nutrient additions that in turn may increase organic OC sequestration (Lehmann and Possinger, 2020). However, evidence for efficient organic and/or inorganic C sequestration by mineral amendments is still poor and additional GHG emissions for mining/excavation, grinding, transport, and application of minerals have to be taken into account (Beerling et al., 2020; Chenu et al., 2019).

The application of thermally transformed OM (biochar and hydrochar) to agricultural soils is a strategy to contribute to OC sequestration by adding OM with a high stability against microbial degradation and thus low turnover (Lehmann et al., 2006). Thermal conversion of organic material at high temperatures may occur at high temperatures (>450°C) through pyrolysis or through heating at lower temperatures (around 200°C) in the presence of water. The latter procedure transforms substrates with high residual water into lignite-like products by hydrothermal carbonization (HTC) under high pressure. The stability of thermally transformed products and thus their potential contribution to SOC sequestration largely depends on biomass parent material and production conditions, particularly the pyrolysis temperature (Spokas, 2010). Generally, biochars produced by pyrolysis at temperatures >450°C are characterized by an overall high stability (Crombie and Mašek, 2015), whereas materials produced at lower temperatures such as HTC coals have a much lower stability against decomposition (Naisse et al., 2015).

3.2 Organic carbon sequestration in soils due to historic management techniques

On the other hand, specific soil management techniques have been developed already in historic times that produce significant OC accumulation, while at the same time strongly modifying the soils, such as in the Terra Preta, paddy soils, or plaggen soils (Kögel-Knabner and Amelung, 2021). The Terra Preta or Amazonian Dark Earths are anthropic soils formed on land surfaces enriched by inputs resulting from pre-Columbian settlers. These inputs resulted in soils characterized by a dark thick A horizon, enriched in OM, and elevated contents of plant-available nutrients (P and Ca), a less acidic pH value and a higher cation exchange capacity than the adjacent tropical soils. Terra Preta is one of the rare but prominent examples, how the addition and alteration of OM to (oxidic) soils may change their properties so dramatically that they turn into fertile soils for centuries. Similarly, the plaggen management established in medieval times in NW Europe resulted in the improvement of sandy soils with respect to OC

contents and stocks. The addition of plaggen material to the topsoils induced a volume increase as well as higher OC concentrations in the Ap horizons, resulting in significantly higher SOC stocks and fertility in the Plaggic Anthrosols compared to surrounding soils. The increase in yields, in turn resulted in higher in-situ OM input, most probably also contributing to high SOC stocks (Urbanski et al., 2022). Paddy soils occurring widely in Asia are also highly modified by specific paddy management operations, especially artificial submergence and drainage, ploughing and puddling (= ploughing and levelling the surface layer of a submerged soil), organic manuring (animal manure, rice straw, and other crop residues, often fermented with sediments taken from the river or channel), liming, and fertilization. Paddy soil management has a clear effect on the accumulation of OM and the OM storage in paddy soils exceeded OM storage in corresponding non-paddy soils (Kalbitz et al., 2013; Wissing et al., 2011). The large accumulation of OM observed in paddy soils is considered to be due to high input of plant residues coupled to alternating redox conditions (Chen et al., 2021; Kögel-Knabner et al., 2010; Kölbl et al., 2014).

These are examples for historic soil management that resulted in higher OC contents and stocks in combination with higher fertility, increased yields and thus in turn a further increase in OC sequestration. The amount of OM input materials available is a major constraint for increasing OC sequestration in soils, specifically when taking into account the higher amounts of OM input necessary to comply for the increased turnover of OM due to increasing temperatures (Bruni et al., 2021; Riggers et al., 2021; Wiesmeier et al., 2016). If OM input materials are available, it seems worthwhile to consider such 'extreme' management options to increase OC sequestration in soils. This calls for judicious management of organic waste materials, which should be considered as raw materials for transformation into soil amendment instead of allowing their destruction and rapid release of CO₂ to the atmosphere.

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