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Soil Carbon Saturation: What Do We Really Know?

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ABSTRACT

Managing soils to increase organic carbon storage presents a potential opportunity to mitigate and adapt to global change challenges, while providing numerous co-benefits and ecosystem services. However, soils differ widely in their potential for carbon sequestration, and knowledge of biophysical limits to carbon accumulation may aid in informing priority regions. Consequently, there is great interest in assessing whether soils exhibit a maximum capacity for storing organic carbon, particularly within organo–mineral associations given the finite nature of reactive minerals in a soil. While the concept of soil carbon saturation has existed for over 25 years, recent studies have argued for and against its importance. Here, we summarize the conceptual understanding of soil carbon saturation at both micro- and macro-scales, define key terminology, and address common concerns and misconceptions. We review methods used to quantify soil carbon saturation, highlighting the theory and potential caveats of each approach. Critically, we explore the utility of the principles of soil carbon saturation for informing carbon accumulation, vulnerability to loss, and representations in process-based models. We highlight key knowledge gaps and propose next steps for furthering our mechanistic understanding of soil carbon saturation and its implications for soil management.

1 | Introduction

Soil organic matter provides a myriad of ecosystem services (Smith et al. 2015), yet it has been severely degraded by anthropogenic land use, resulting in large organic carbon (C) losses in much of the world (Sanderman et al. 2017). Targeted management strategies that maintain or enhance soil organic carbon (SOC) stocks are paramount to restoring these ecosystem services, with increasing interest across a broad range of stakeholders (Derrien et al. 2023). To inform management decisions under a changing climate, a predictive understanding of the underlying controls driving SOC accrual and potential limits to SOC storage is critical.

Fundamentally, SOC storage results from a balance between C inputs and outputs, where underlying transformations are driven by a suite of environmental and management factors (West and Six 2007). Separating SOC into functionally distinct fractions—namely, mineral-associated organic carbon (MAOC) and particulate organic carbon (POC)—can provide further insights because of their different formation pathways, turnover times, and response to disturbance (Lavallee et al. 2020). MAOC is typically characterized as more persistent with, on average, slower turnover times than POC (Cotrufo and Lavallee 2022; Heckman et al. 2022). However, MAOC is not a uniform pool. Indeed, MAOC can be complex and heterogeneous, both in terms of its spatial distribution (Schweizer 2022; Vogel et al. 2014) and temporal dynamics (Neurath et al. 2021; Schweizer et al. 2024). The persistence of MAOC, and surface coverage of mineral particles, can depend on the type of mineral, solution chemistry, and bond strengths (Lutfalla et al. 2019; Mayer et al. 2023; Schrumpf et al. 2021). Despite these complexities, managing soils to increase MAOC may contribute to more durable C sequestration in soils. However, we recognize that MAOC and POC must ultimately be managed together (Angst et al. 2023), given their inherent linkages and distinct functionality (e.g., stoichiometry and nutrient availability) in ecosystems (Rocci et al. 2024; Villarino et al. 2023).

Another aspect of MAOC that is central to our understanding of soil C cycling and storage is its potential to saturate based on the finite nature of reactive minerals in a soil. This concept of 'soil C saturation' was first proposed over 25 years ago (Hassink 1997) and has seen much interest since then (Feng et al. 2013; Georgiou et al. 2022; Poeplau et al. 2024; Six et al. 2002). Although the concept has also been applied to describe apparent plateaus in total SOC with increasing C inputs (Stewart et al. 2007; West and Six 2007), we emphasize that MAOC and POC fractions differ in the controls governing their formation and potential limits (Castellano et al. 2015; Cotrufo et al. 2019). Given the mechanistic underpinnings of the soil C saturation concept, we strictly focus on MAOC here. MAOC saturation corresponds to a maximum capacity defined solely by chemical and physical properties of the soil mineral matrix. In contrast, effective capacities can depend on climate, management, and microbial responses (Craig et al. 2021; Stewart et al. 2007). Given interest in natural climate solutions to store C on land (Bossio et al. 2020; Griscom et al. 2017) and potential feedbacks between soil C and climate (Bradford et al. 2016), the concept of MAOC saturation is an important constraint on estimates of soil C storage potential. As such, attention is needed to define precise terminology and avoid muddling concepts.

Several approaches have been used to investigate the maximum capacity of MAOC across ecosystems—namely, by relating MAOC content to clay and silt content, SOC, and C inputs. However, each of these methods provides distinct information and has its own set of caveats. Furthermore, the use of different fractionation methods to separate MAOC may also lead to biases (Leuthold et al. 2024; Poeplau et al. 2018). These factors have contributed to a perceived debate in recent literature about the validity and utility of the MAOC saturation concept (Begill et al. 2023; Cotrufo et al. 2019; Georgiou et al. 2022; Six et al. 2024).

Here, we define key terminology and review the most common approaches and best practices for quantifying MAOC saturation. We explore the concept of C saturation at both the macro-scale of soil fractionation measurements as well as at the micro-scale of heterogeneous mineral interactions. We summarize evidence from field and laboratory experiments on the utility of the C saturation concept and note the importance of constraining the capacity of MAOC in process-based models that explicitly represent organo–mineral associations. Finally, we highlight key knowledge gaps and discuss next steps to further probe our mechanistic understanding of C saturation at both micro- and macro-scales.

2 | Conceptual Understanding of Soil Carbon Saturation

2.1 | Macro-Scale Understanding and Terminology

Given the finite nature of reactive minerals in the soil, it has been hypothesized that the physicochemical stabilization of carbon by association with mineral surfaces (i.e., MAOC) should also be finite (Hassink 1997; Stewart et al. 2007). This forms the basis for the concept of carbon saturation. Some studies have sought to apply this concept more broadly to total SOC [including data-driven approaches for total soil carbon potential, as noted in Barré et al. (2017)], but here, we strictly focus on the saturation of MAOC and note that POC can reach much higher concentrations (as in organic soils; Mirabito and Chambers 2023) with no theoretical basis for saturation. In this section, we distinguish and define consistent terminology that is central to the concept of MAOC saturation and shed light on common misconceptions.

First and foremost, we define the *theoretical mineral capacity* as the absolute maximum MAOC independent from soil climatic conditions and management (Table 1; Figures 1 and 2). The *theoretical mineral capacity* is unique to a given soil based solely on its mineral content and composition—that is, the 'size of the bucket' given the analogy of filling a bucket with water. If the *theoretical mineral capacity* is reached, by definition, no further increase in MAOC is possible. We note that this conceptual limit does not presume a monolayer. Indeed, spatial heterogeneity and organo–organic interactions on mineral surfaces are well-established (reviewed in Section 2.2). However, the existence of a *theoretical mineral capacity* does assume that subsequent layering will be finite, likely due to weaker organo–organic interactions (with potential implications on vulnerability; Section 4.2), such that the amount of MAOC will eventually reach an utmost maximum capacity defined solely by the mineral content and composition (given operational definitions of MAOC; Section 3.2). This limit remains conceptual and is difficult to quantify in practice (Section 3.1) because observational datasets span a finite range of environmental conditions, and it is unknown whether the dataset includes soils that are currently

TABLE 1 | Key terminology, definitions, environmental controls, and quantification approaches.

Terminology	Definition	Expected behavior	Environmental controls	Quantification	Examples
Theoretical mineral capacity[a]	Theoretical maximum potential for MAOC storage independent from climate and management under not limiting C input conditions	No increase in MAOC with increasing C input rate under any conditions	Clay + silt, mineralogy	___ *Approximated* as the maximum observed capacity (below) assuming some sites globally have sufficient conditions to reach the theoretical mineral capacity	___ Ahrens et al. (2020) and Georgiou et al. (2022)
Maximum observed capacity[a]	Maximum potential for MAOC storage observed under existing climate and management globally	Negligible increase in MAOC with increasing C input rate under existing conditions	Clay + silt, mineralogy[b]	Maximum MAOC as a function of clay + silt (BL) across all soils within a given mineral category	Beare et al. (2014), Feng et al. (2013) and Georgiou et al. (2022)
Effective capacity	Apparent limit in MAOC storage under a particular set of management and climate conditions	Negligible increase in MAOC with increasing C input rate, but potential to increase MAOC under different conditions that change formation efficiency and decrease C outputs	Management, nutrient availability, vegetation and microbial traits, soil moisture, temperature, pH, clay + silt, mineralogy	Manipulation experiments with different levels of increased C inputs; Maximum MAOC as a function of clay + silt (LR or BL) for a given management or climate	Feng et al. (2014), Hassink (1997), Matus (2021), Six et al. (2002), Stewart et al. (2007) and Viscarra Rossel et al. (2024)
Steady-state	Balance of MAOC inputs and outputs under any environmental conditions	No change in MAOC over time for constant C input rate	Time, management, nutrient availability, vegetation and microbial traits, soil moisture, temperature, pH, clay + silt, mineralogy	Manipulation experiments that change C inputs	Stewart et al. (2007) and West and Six (2007)

Abbreviations: BL, boundary line; LR, linear regression.

[a] *MAOC saturation* corresponds to the state in which the *maximum observed capacity* is reached, given practical limitations in quantifying the *theoretical mineral capacity* (see Section 4.1 for further details on how the *maximum observed capacity* has been used to calculate MAOC saturation and deficits).

[b] Could include other factors responsible for *effective capacities*, if the *maximum observed capacity* quantified globally has not reached the *theoretical mineral capacity* for a given mineral category.

at their *theoretical mineral capacity*. As such, we define the *maximum observed capacity* as the physicochemical stabilization limit for a given mineral content and composition under the range of current environmental conditions globally [similar to 'attainable maxima' by Ingram and Fernandes (2001)]. However, we note that minerals may not be the most limiting factor for the *maximum observed capacity* (Karunaratne et al. 2024; Poeplau et al. 2024), and further work should explore the existence and proximity of the *theoretical mineral capacity* to the *maximum observed capacity* (as illustrated with dashed lines in Figure 2).

Depending on the particular range of soil, climate, and management conditions, microbial constraints may also limit MAOC formation and persistence under increased C inputs. Although this behavior can resemble carbon saturation, MAOC may in fact be much lower than the *theoretical mineral capacity* or even the *maximum observed capacity*. For example, ecological constraints on microbial biomass (e.g., via competition or predation) could reduce C flow through microbes and thereby reduce MAOC formation as soil C inputs increase. This was nicely illustrated by Craig et al. (2021) using a meta-analysis and soil C model with and without density-dependent microbial growth and mortality. By driving a disconnect between C inputs entering the soil and those available to form MAOC, microbial constraints can result in apparent saturation behavior. However, this behavior is context-dependent, and more MAOC may be attainable for the same C inputs under different conditions (e.g., climate, litter quality, soil pH, and management) that reduce C outputs (Figure 2). Following Stewart et al. (2007), we define the *effective capacity* as an apparent saturation limit (i.e., plateau) in MAOC with increasing C inputs, under a particular set of management and climate conditions. Both natural and managed systems may exhibit *effective capacities*. We highlight that this phenomenon has been widely observed in field and laboratory experiments (reviewed further in Section 3.1), but its context-dependency is critical when comparing estimates from different studies. Furthermore, the utility of the capacities defined above is still an evolving area of research and ultimately depends on the application (see Section 4).

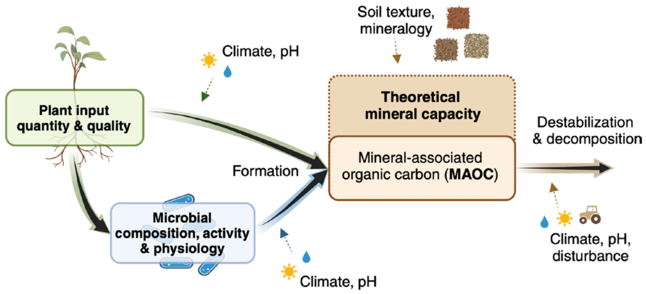

FIGURE 1 | Environmental and management factors controlling carbon inputs and outputs to mineral-associated organic carbon (MAOC) in soil. MAOC formation depends on the quantity and chemical composition of inputs from plant and microbial sources, as well as the physical and chemical properties of the mineral matrix (which define the *theoretical mineral capacity*). Both formation and destabilization are further influenced by soil environmental conditions (e.g., soil moisture, temperature, and pH) and disturbance.

Finally, and perhaps most importantly, we note that C saturation—or any of the above capacities—should not be concluded simply because a soil has received C inputs for a long time and MAOC is unchanging (West and Six 2007). To constitute a capacity, MAOC should be unchanging with further increases in C inputs, not only with time (Figure 2). The latter merely implies a *steady-state* where C inputs are equal to C outputs over time. There can be many *steady-states* for a given soil, depending on the environmental and management conditions (Karunaratne

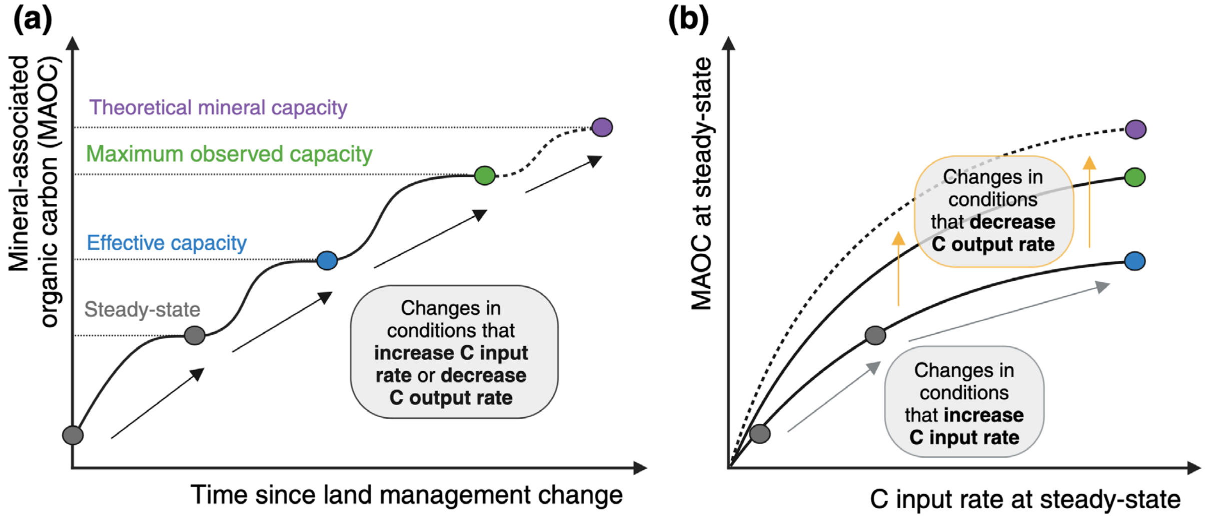

FIGURE 2 | Key terminology and relationships for the concept of soil carbon saturation—namely, *steady-state*, *effective capacity*, *maximum observed capacity*, and *theoretical mineral capacity*; see Table 1 for definitions. Soil carbon saturation corresponds to the state in which the *maximum observed capacity* is reached, given limitations in quantifying the *theoretical mineral capacity*. (a) Mineral-associated organic carbon (MAOC) as a function of time since land management, starting from a soil with low MAOC and following changes in conditions that increase the C input rate, decrease the C output rate, or both. (b) MAOC at steady-state as a function of C input rate at steady-state. Steady-states corresponding to the defined terminology are depicted using matching colors as in panel (a). Schematics adapted from Stewart et al. (2007) and West and Six (2007). Dashed lines signify that the proximity of the *theoretical mineral capacity* to the *maximum observed capacity* is unknown.

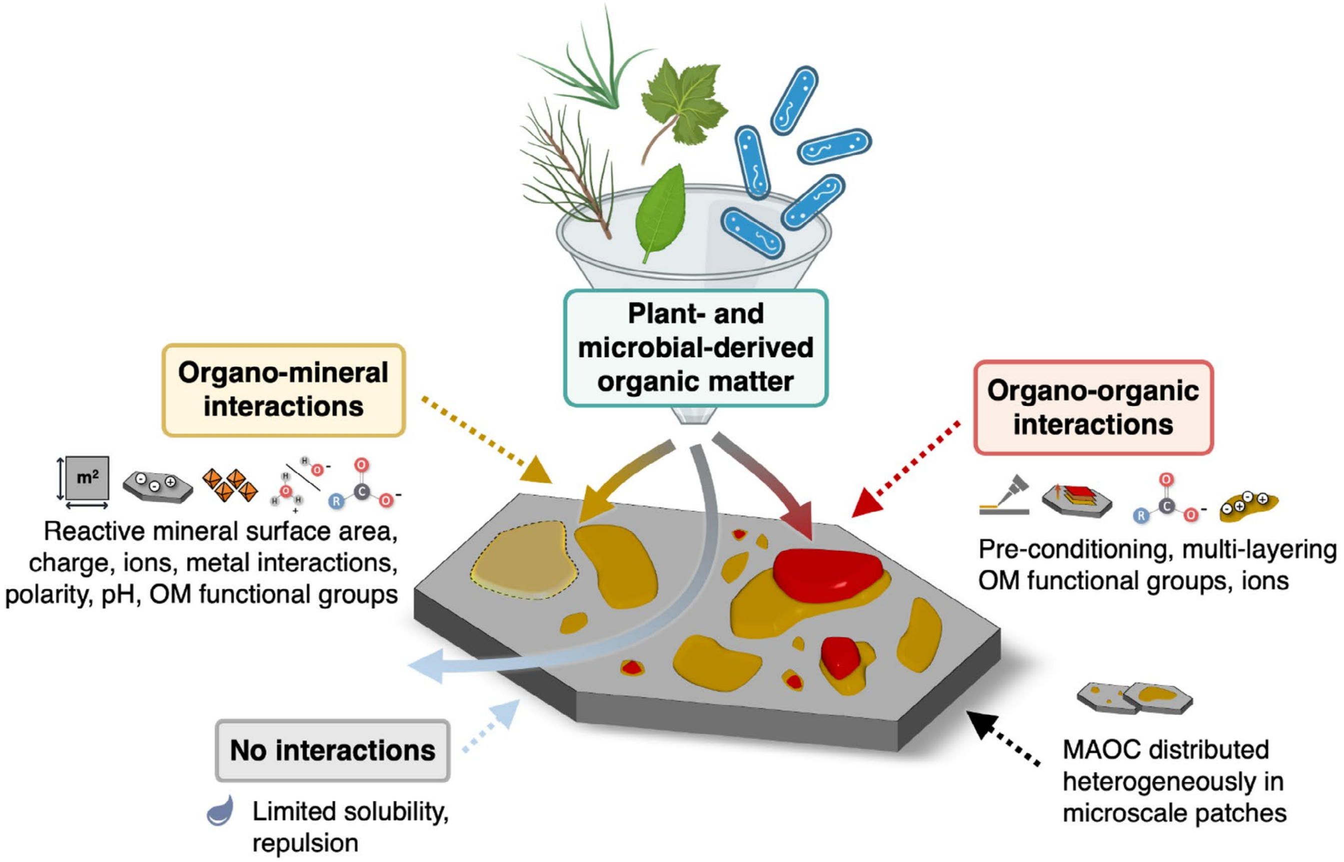

FIGURE 3 | Pathways and controlling factors of mineral-associated organic carbon (MAOC) formation through organo–mineral and organo–organic interactions. Many mineral and organic matter (OM) properties influence MAOC formation, spatial distribution, and estimates of C loading.

et al. 2024; Kirschbaum et al. 2020). We also note that just because a soil is at a *steady state*, including any of the defined capacities, does not imply that its MAOC is old. It only implies that MAOC concentrations are unchanging, but not that a given molecule within MAOC is 'stable' from decomposition. Indeed, sorption can be dynamic and reversible (Kleber et al. 2021; Yang et al. 2021), especially depending on the nature of organo–mineral and organo–organic interactions (Section 2.2).

2.2 | Surface Interactions and Microscale Heterogeneity

Analyses of surface interactions and direct imaging point to an array of factors driving organo–mineral and organo–organic interactions (Figure 3). Organo–mineral interactions typically depend on the mineral specific surface area (SSA), charge density, crystal structure, ionic solution composition, metal-to-carbon ratio, and pH (Kaiser and Guggenberger 2003; Kleber et al. 2021; Nierop et al. 2002; Saidy et al. 2013; Willemsen et al. 2019). Pre-weathered microsites at mineral surfaces have been shown to provide preferential binding sites (van der Kellen et al. 2022), motivating further study of the heterogeneous spatial distribution of mineral surface reactivity. Reactive mineral surface availability and properties (e.g., SSA) and reactive metal ions are central to explaining MAOC capacities, yet data limitations still hinder their exploration at regional- to global-scales beyond a rough distribution of clay-sized minerals (Ito and Wagai 2017). As a consequence, mineral surface properties are only coarsely captured in current estimates of *maximum observed capacities* globally (Section 3.1), and future data synthesis studies may allow for refined capacities that better resolve microscale complexity.

In addition to mineral surface properties, the formation of MAOC critically depends on the properties of organic matter (OM) compounds—for example, functional group composition, charge, polarity/solubility, and ion interactions (Curti et al. 2021; Kalinichev and Kirkpatrick 2007; Marschner and Kalbitz 2003; Newcomb et al. 2017). Consequently, certain OM compounds can be selectively retained and shape MAOC composition (Avneri-Katz et al. 2017; Liang et al. 2019; Mikutta et al. 2019). Moreover, the composition and amount of pre-existing OM can influence subsequent sorption (Gao et al. 2018; Willemsen et al. 2022). OM compounds with both hydrophilic and hydrophobic regions may introduce polar functionality, which can induce the accretion of additional OM compounds according to the zonal model (Kleber et al. 2007; Underwood et al. 2024). C loadings several times higher than a theorized monolayer-equivalent coverage (i.e., $>1\,\mathrm{mg\,C\,m^2}$) have been observed (Kahle et al. 2002; Kögel-Knabner et al. 2008; Mayer 1994; Mayer and Xing 2001) indicating the importance of organo–organic interactions, especially in soils with lower clay contents (Schweizer et al. 2021). The extent of organo–organic interactions may thus partially decouple MAOC accumulation from the factors involved in organo–mineral interactions (Figure 3) and contribute to preferential binding sites and patchy surface coverage. The co-precipitation of OM with metal ions (Mikutta et al. 2011; Tamrat et al. 2019) may further decouple MAOC accumulation. As a secondary effect beyond organo–mineral and organo–organic interactions, the aggregation and occlusion of OM in microscale and nanoscale structures can also contribute to measured MAOC, requiring careful operational isolation (Section 3.2).

Indeed, microspectroscopic and microspectrometric techniques reveal a patchy distribution of MAOC on mineral surfaces (Schweizer 2022; Solomon et al. 2012) and, when combined with isotopic tracers, indicate that fresh OM can be attracted to pre-existing MAOC patches (Keiluweit et al. 2012; Vogel et al. 2014). Surface-sensitive analytical approaches, such as atomic force microscopy and X-ray photoelectron spectroscopy, provide direct evidence for a thicker accrual of MAOC through organo–organic interactions (Gazze et al. 2018; Huang et al. 2020), as previously suggested (Wagai et al. 2009a). Moreover, the coverage of OM across mineral surfaces has been shown to increase over the time of soil formation, from patchy to more connected OM coatings, but a proportion of the mineral surface often appears to remain free of OM (Schweizer et al. 2018). This patchiness has also been reported using gas sorption approaches (Kaiser and Guggenberger 2003; Mayer and Xing 2001; Wagai et al. 2009a). Co-precipitation of OM with metal ions was shown to contribute to higher C loadings than adsorption and is related to a different spatial arrangement at the nanoscale as indicated by scanning transmission electron microscopy with electron energy loss spectroscopy (Chen et al. 2014; Possinger et al. 2021). Although it may be tempting to assume that the prevalence of patchy MAOC distributions implies that soils are still below their respective *theoretical mineral capacities*, we emphasize that there are other inherent reasons for OM patchiness and full surface coverage may not always be possible. In particular, future studies should continue to explore the spatial distribution of mineral reactivity across soils as well as steric interactions with sorbed OM and their role in the observed patchiness. Targeted studies of high OM soils may also provide further insights (see Section 5). Furthermore, studies examining the formation of organo–organic interactions will be paramount to understanding the role of MAOC layering in the context of C saturation and MAOC formation in soils near their *maximum observed capacity*.

3 | Quantifying Mineral-Associated Carbon Limits

3.1 | Approaches for Quantifying Saturation

Soil C saturation is generally conceptualized as a non-responsiveness of the soil C—and in particular MAOC—content to increasing C inputs (Six et al. 2002; Stewart et al. 2007). However, quantifying soil C inputs can be challenging and uncertain (Price et al. 2012), especially those entering the MAOC fraction given their physical–chemical nature (i.e., structural vs. soluble) (Cotrufo et al. 2022) and the microbial processes driving their transformations (Craig et al. 2021; Sokol et al. 2022), representing a pressing gap for future work. Thus, several alternative approaches have been proposed over the last two decades to quantify MAOC saturation. We provide an overview of the most common approaches used to quantify the *maximum observed capacity* or context-dependent *effective capacities*. Namely, we review the theory and evidence underpinning the relationships between MAOC versus clay + silt, MAOC versus SOC, and MAOC versus C inputs and discuss the significance and caveats of each.

3.1.1 | MAOC and Clay + Silt Linear Regression

The most widely used approach for quantifying the maximum capacity of MAOC considers the relationship between MAOC and clay and silt particles (Figure 4a). While the importance of clay + silt for the physicochemical protection of organic C has long been documented (Mayer 1994; Theng 1979), seminal work by Hassink (1997) framed this relationship in the context of a MAOC capacity. Hassink (1997) found that MAOC was negligibly different between (grazed) grassland and arable soils and concluded that grasslands had reached their MAOC capacity given their higher SOC and presumed higher C inputs; however, C inputs were not quantified or directly manipulated. The MAOC capacity across temperate and tropical grasslands was then estimated using a linear regression between MAOC and clay + silt. This regression—often termed the Hassink equation: MAOC capacity = 0.37 (% clay + silt) + 4.09—has been supported by some studies (Carter et al. 2003; Guillaume et al. 2022) and challenged by others who have reported 'over-saturated' soils when using this equation (Angers et al. 2011; Feng et al. 2013). This lack of agreement is, at least in part, because the dataset was of limited size and did not represent soils from enough pedoclimatic contexts to obtain a general threshold—that is, it estimated an *effective capacity* based on a particular set of grasslands and not the *maximum observed capacity* (Figure 4a). Furthermore, since the Hassink equation is derived from a linear regression, it does not represent an upper limit but rather the average MAOC for a given clay + silt content under natural vegetation. Thus, we emphasize that, by definition, even the original dataset contained soils above the linear regression that would be considered 'over-saturated' in the other studies noted above.

Increasing evidence over the last two decades suggests that alternative analyses—namely, using a quantile regression or boundary line—and knowledge of the soil mineralogy are needed to quantify the *maximum observed capacity* for a given clay + silt content (Section 3.1.2). Despite these recent advances, the Hassink equation continues to be widely applied for assessing C deficits at field- to national-scales (Chen et al. 2018; Just et al. 2023; McNally et al. 2017; Soinne et al. 2024; Wiesmeier et al. 2014, 2015). We argue that it is important to consider the foundation for each approach when selecting an *effective* versus *maximum observed capacity* in future studies.

3.1.2 | MAOC and Clay + Silt Relationship: Recent Advances

A notable advance in quantifying the *maximum observed capacity* of MAOC is the boundary line (BL) analysis proposed by Feng et al. (2013) and subsequently used by others (Beare et al. 2014; Cai et al. 2016; Fujisaki et al. 2018). This approach was inspired by literature on maximal plant productivity (Webb 1972) but

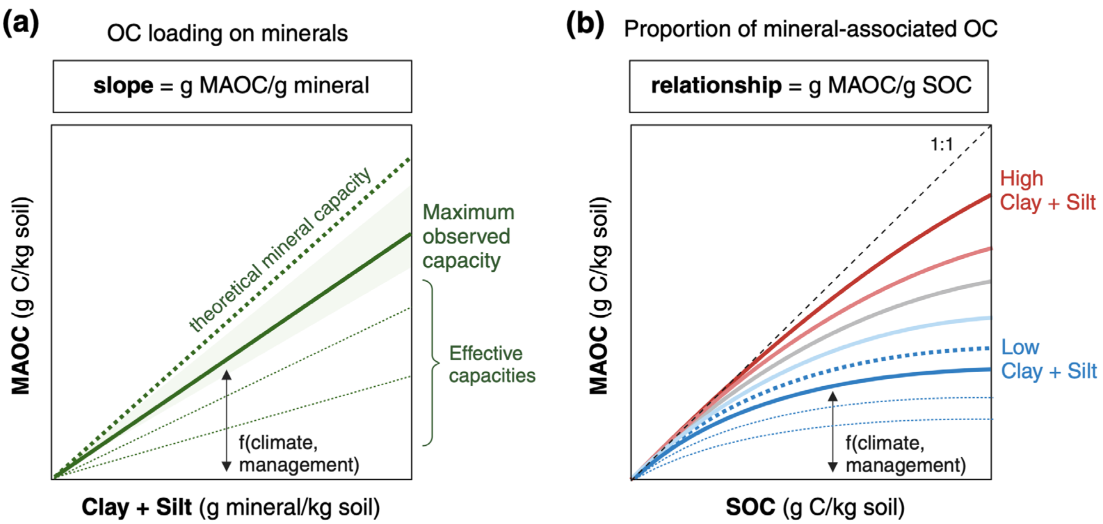

FIGURE 4 | Conceptual schematics of approaches used to quantify the *maximum observed capacity* across soils. Namely, relationships between mineral-associated organic carbon (MAOC) and either (a) clay + silt content or (b) total soil organic carbon (SOC) are often explored.

can be derived for organo–mineral associations using a saturating sorption isotherm with a *theoretical mineral capacity* (e.g., Langmuir or Sips) (Georgiou et al. 2022). Although studies have proposed alternative isotherms that warrant further investigation (e.g., linear or Freundlich; Section 4.3) (Kirschbaum et al. 2020), sorption experiments and C input manipulations do suggest saturating behavior in many contexts (Section 3.1.4). With sufficient conditions for some soils to have reached MAOC saturation within a dataset, the BL is expected to closely approach the *theoretical mineral capacity* for each mineral type (Figure 4a). In practice, however, the BL analysis constrains the *maximum observed capacity* for the range of environmental conditions encompassed in the dataset. Moreover, most studies conservatively estimate the BL as the 95th or 90th percentile of MAOC as a function of clay + silt (Beare et al. 2014; Feng et al. 2013; Georgiou et al. 2022) to capture the upper limit yet mitigate measurement uncertainties (Section 3.2). As such, the *maximum observed capacity* likely underestimates the *theoretical mineral capacity*, and the proximity of these two quantities to each other remains an open question.

In essence, the relationship between MAOC and clay + silt describes the C loading on mineral particles (Figure 4a), and BL slopes represent a revised approach to quantify the maximum. Leveraging an expanded global dataset of MAOC measurements (Figure 5; see Supporting Information), we find that the *maximum observed capacity* is 87 mg C g^{-1} clay + silt (calculated from the slope of the 95th percentile; 90th to 99th percentile: 75–120 mg C g^{-1} cla + silt) in soils dominated by 2:1 phyllosilicates, and 49 (41–54) mg C g^{-1} clay + silt in soils dominated by 1:1 phyllosilicates. These values are consistent with studies that used subsets of this dataset (Feng et al. 2013; Six et al. 2024), and hold across soil depths (Georgiou et al. 2022) and fractionation methods (see Figure S1). It is possible that targeted sampling in carbon-rich surface soils and sandy soils may result in revised BL estimates—for example, C loading tends to be higher in sandy soils (Poeplau et al. 2024), though this appears to partly depend on the fractionation method (Figure S2; see Section 3.2 and Supporting Information). High C loading in the fine fraction of sandy soils could also be related to mineral composition, which can in certain cases be dominated by pedogenic metal (hydr)oxides instead of phyllosilicates (Eusterhues et al. 2005). However, we emphasize that slightly increasing or decreasing BL estimates does not in itself falsify the concept of C saturation. Rather, it highlights the well-established importance of other environmental factors in driving observed MAOC (Table 1) and the need for spanning a wide range in these factors to quantify the *maximum observed capacity*.

Soil minerals vary markedly in their specific surface area (SSA) and charge (Kleber et al. 2021; Kögel-Knabner et al. 2008) and hence in their capacities to bind organic matter. Consequently, soils should be characterized based on their mineralogy when quantifying the *maximum observed capacity* (Feng et al. 2013)—for example, a differentiation between soils dominated by 1:1 versus 2:1 phyllosilicates. Similarly, soils with greater quantities of allophanic and amorphous clays may warrant a separate category, though data are more limited. For instance, Beare et al. (2014) demonstrated that the MAOC stabilization capacity of allophanic soils (reaching BL slopes of 153 ± 17 mg C g^{-1} clay + silt) was 33%–91% higher than in non-allophanic soils. However, the SSA of allophane is very high (e.g., 700–900 m^2 g^{-1} clay + silt; Kleber et al. 2021), such that the C loading may in fact be lower than the maximum observed on other minerals (i.e., <1 mg C m^{-2}; Section 2.2). Metal oxides can also play an important role in binding organic matter depending on the soil pH (Doetterl et al. 2015; Rasmussen et al. 2018; Salonen et al. 2023). Thus, finer-scale mineralogical properties (e.g., SSA, surface heterogeneity) warrant study to advance mechanistic understanding and predictive models of the *theoretical mineral capacity*; but in the meantime, statistical models based on an upper envelope (i.e., BL) are best suited to constrain the *maximum observed capacity* for each mineral type.

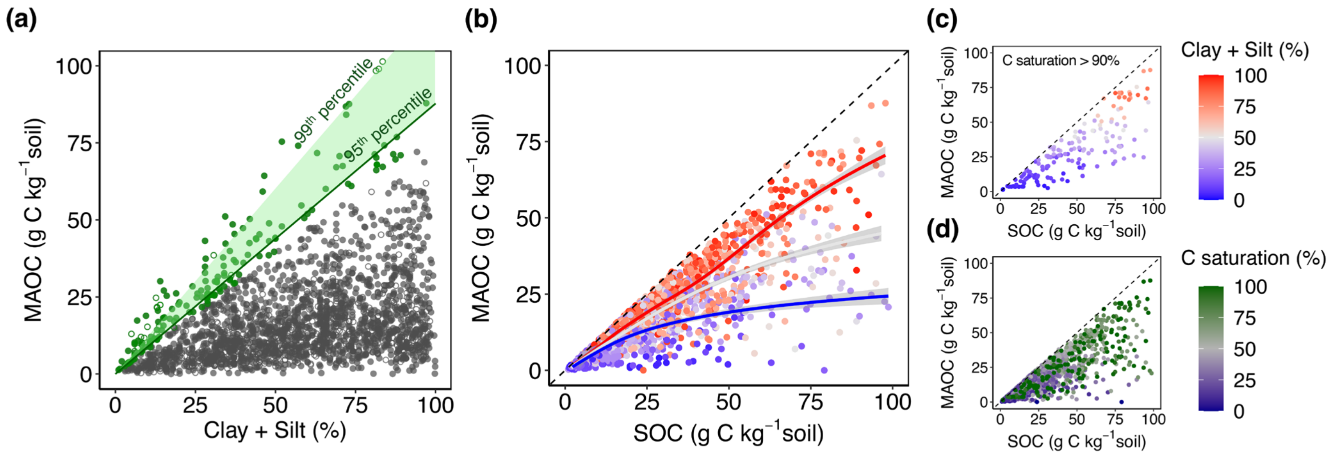

FIGURE 5 | Quantifying the *maximum observed capacity* globally from measurements of mineral-associated organic carbon (MAOC) and soil organic carbon (SOC) in soils dominated by 2:1 phyllosilicate (high-activity) minerals ($n = 2252$). (a) The *maximum observed capacity* (green line) is derived using the boundary line analysis (see Section 3.1.2 and Supporting Information), and observations with > 90% C saturation are depicted in green (solid and open circles denote natural and agricultural ecosystems, respectively). C saturation (as a %) is calculated by ratio of the MAOC and the corresponding *maximum observed capacity* for each clay + silt content. (b) The relationship between MAOC and SOC depends on the clay + silt content, where best fit lines correspond to subsets of < 33% clay + silt (blue), > 66% clay + silt (red), and intermediate (gray). Solid circles are colored by clay + silt content across (b) all soils and (c) only observations with > 90% C saturation. (d) MAOC and SOC with all observations colored by %C saturation.

Although most studies and theory suggest that the upper envelope should be a linear function of clay + silt, nonlinear functions have been proposed to constrain the *maximum observed capacity*; but importantly, not the *theoretical mineral capacity*. A nonlinear function implies that the *observed* C loading (i.e., mg C g^{-1} clay + silt) is itself a function of clay and silt content. There may be mechanisms for this—for instance, pore space constraints in clay-rich soils could limit the degree of organo–organic layering (and hence, C loading) on mineral particles. Similarly, aggregation and occlusion may spatially separate potentially available reactive surface area from organic matter (Lehndorff et al. 2021). It is also likely that C inputs are not sufficient in many clay-rich soils to show saturation (Poeplau et al. 2024)—in which case, it is important not to mistake a lack of observed saturation under current conditions as an indication of a lower *maximum observed capacity* or the non-existence of a *theoretical mineral capacity*. For instance, Viscarra Rossel et al. (2024) used frontier lines to define nonlinear envelopes of MAOC stocks and reported 'maximum attainable potentials' (or *effective capacities*) for a particular range of conditions—namely, more arid, input-limited environments in Australia. Although attainable potentials may be relevant for management decisions in certain contexts (Section 4.1), we emphasize the importance of quantifying the *maximum observed capacity* globally and understanding the environmental factors that drive a deviation from the maximum. That is, in addition to being a scientific curiosity, the *maximum observed capacity* represents a key quantity in many process-rich SOC models that is important for predicting MAOC (Section 4.3).

3.1.3 | MAOC and SOC Relationship

An alternative approach that some studies have proposed for detecting, and even quantifying, C saturation is the relationship between MAOC and SOC (Figure 4b). This approach was first derived by Stewart, Plante, et al. (2008) using a two-pool SOC model with explicit C saturation and applied to 8 agroecosystem experiments. They found that MAOC approached a plateau as a function of SOC, whereas POC increased linearly, and concluded that MAOC may be saturated in the soils with high SOC. It is important to note that they focused on individual sites that mainly varied in land management. Subsequent studies have applied this method more broadly across regional to continental scales (Begill et al. 2023; Cotrufo et al. 2019; Fernández-Catinot et al. 2023; Matus 2021), even though this relationship can be more complex when applied across a range of soils and climates.

Indeed, this approach ultimately depicts the MAOC/SOC ratio (Figure 4b), which can be influenced by many environmental factors including (but not limited to) C saturation—for instance, climate and land use can influence the distribution of MAOC and POC, and hence, MAOC/SOC (Georgiou et al. 2024; Hansen et al. 2024; Sokol et al. 2022). Nevertheless, recent studies have used the curvature, or lack thereof, in the relationship between MAOC and SOC as evidence for or against the concept of C saturation (Begill et al. 2023; Cotrufo et al. 2019; Fernández-Catinot et al. 2023). Notably, MAOC/SOC can be quite stable in temperate agricultural soils across a wide range of SOC contents, where MAOC comprises on average ~74% of SOC (Begill et al. 2023). Yet, MAOC values above 45–50 g C kg^{-1} soil are relatively limited in some datasets (Cotrufo et al. 2019), though we note that this alone does not imply C saturation nor do values above this presumed limit disprove the concept. We highlight that a single universal relationship or plateau in MAOC versus SOC is not expected across sites (Figure 4b) but rather a continuum of *effective* or *maximum observed capacities* that depend on a suite of environmental factors.

Most importantly, we show that the relationship between MAOC and SOC can depend on clay + silt content (Figure 5b; also in Georgiou et al. (2022) and Begill et al. (2023)). Climate, land-use, and management also influence this relationship, resulting in a large degree of variability in MAOC for a given SOC globally. Moreover, we find that soils along the whole spectrum of MAOC/SOC ratios may be at or near C saturation (Figure 5c,d)—calculated as the ratio of MAOC to the *maximum observed capacity* (Figure 5a). The subset of points with > 90% C saturation (Figure 5c) exhibit a large degree of variability but a notable lack of curvature (or plateau) in the relationship between MAOC and SOC. Thus, we emphasize that this relationship on its own cannot be used to support or refute C saturation across multiple locations with different environmental conditions (Six et al. 2024). However, it may be useful for agroecosystem experiments at individual sites where C inputs are difficult to quantify or in conjunction with the relationship between MAOC and clay + silt (Stewart, Plante, et al. 2008). Furthermore, the MAOC/SOC ratio may provide insights on nutrient availability and vulnerabilities across land-use types, given differences in MAOC and POC stoichiometric composition and responses to global change factors (Lugato et al. 2021; Rocci et al. 2024).

3.1.4 | MAOC and C Inputs Relationship

Another way to test if, or when, a soil's capacity to store MAOC may be reached, is to amend a soil with different amounts of C inputs over a longer time period (e.g., years to decades) and to follow the change in MAOC content resulting from the C additions. We stress that a single C input rate is not sufficient because it does not allow for the distinction between simply achieving a new steady-state as a function of time versus a maximum capacity as a function of C inputs (Figure 2). Therefore, multiple additions at different C input rates are needed to detect whether the response of MAOC will plateau with increasing C inputs.

Input manipulation experiments can be conducted as either field trials or laboratory incubations. To interpret their results, however, we note a few inherent challenges and considerations. C inputs are often added to the soil surface and, even if they are mixed into the soil, may not be immediately able to sorb to a mineral without some degree of microbial processing (Angst et al. 2021). As such, the efficiency of MAOC formation can be very low; for example, less than 10% of added aboveground C inputs may be retained as MAOC (Villarino et al. 2021). In addition to the point-of-entry, C input quality can affect MAOC formation efficiency (Lavallee et al. 2020; Sokol et al. 2019). For instance, nutrient-rich organic matter can have a higher affinity for mineral surfaces, leading to preferential sorption (Spohn 2024). Although MAOC formation efficiency should not ultimately impact a soil's capacity to store MAOC, it can alter the level of C inputs needed to achieve this capacity (Castellano et al. 2015). As such, very high doses of C inputs may be needed to approach a plateau in MAOC. This is especially true for cropland soils, which are often C depleted (averaging only ~30% MAOC saturation; Georgiou et al. 2022). Linear increases in SOC or MAOC contents are thus expected in many agricultural soils, even under high C inputs, making it difficult to quantify, or even detect, C saturation in this way.

Consequently, it is not surprising that studies have found conflicting evidence of saturating behavior in many long-term field trials. For example, leveraging 7 long-term (10–32 years) manure amendment experiments, Feng et al. (2014) showed that it was difficult to distinguish linear and saturating relationships between MAOC and C inputs at most of the sites, concluding that the C input rates were insufficient to achieve C saturation in the intensively cultivated soils studied. Similarly, Hao et al. (2003) reported linear increases in SOC over 25 years of manure amendments but did not fractionate soils to quantify changes in MAOC and POC. In contrast, Gulde et al. (2008) found saturating behavior in the MAOC and microaggregate fractions—and linear increases in POC—across several levels of increased manure input rates at a long-term (> 30 years) agricultural experiment. Targeted experiments in soils with already high MAOC concentrations close to the *maximum observed capacity* (as estimated in Section 3.1.2 and Figure 5) may be necessary to further test the C saturation concept (see Section 5). Future studies may also target soils with similar mineral properties but large variations in initial MAOC contents (see Section 4.1). Moreover, given the slower turnover times of MAOC, the experiment duration is also important. West and Six (2007) estimate that over two decades may be needed for soils to equilibrate to a change in C input rates. For long-term experiments that extend over multiple decades, it is critical to characterize, and control for, other chemical, biological, or physical changes that may occur over the same time period—for example, changes in pH (Wang and Kuzyakov 2024), microbial community composition (Bradley et al. 2014), or even weathering of geochemically young soils (Khedim et al. 2021).

Finally, innovative short-term studies may also have merit for testing theory and underlying mechanisms. At the most simplified level, laboratory sorption experiments have been conducted with specific mineral types and multiple doses (i.e., concentrations) of DOC. Such experiments are used to derive sorption isotherms—most often based on a saturating, Langmuir relationship—and constrain the maximum amount of sorbed C for each mineral type (Jagadamma et al. 2014; Mayes et al. 2012). However, these experiments are often conducted under sterile conditions and typically underestimate the capacity of MAOC by up to an order of magnitude compared to observations from field experiments (Abramoff et al. 2021; Georgiou et al. 2022). While such sorption experiments can be used to inform select model parameters (see Section 4.3), the mismatch in MAOC contents highlights the importance of biotic processes. Microbial necromass and residues contribute substantially to MAOC formation and thus laboratory experiments inoculated with microbes are essential to achieve MAOC values that are better aligned with field studies (Kallenbach et al. 2016).

3.2 | Fractionation Methods and Potential Biases

Any potential biases in fractionation methods may ultimately affect quantified MAOC capacities and conclusions on MAOC saturation. MAOC is typically isolated via the fractionation of soils by size (as the fine fraction < 20–63 μm) or density (as the heavy fraction typically > 1.6–1.85 g cm^{-3}) or both (Christensen 2001; Elliott and Cambardella 1991; Poeplau et al. 2018; von Lützow et al. 2008). Before MAOC can be isolated, however, the soil must be dispersed to carefully break up all aggregates. Aggregates contain both POC and MAOC, and their partial dispersion can

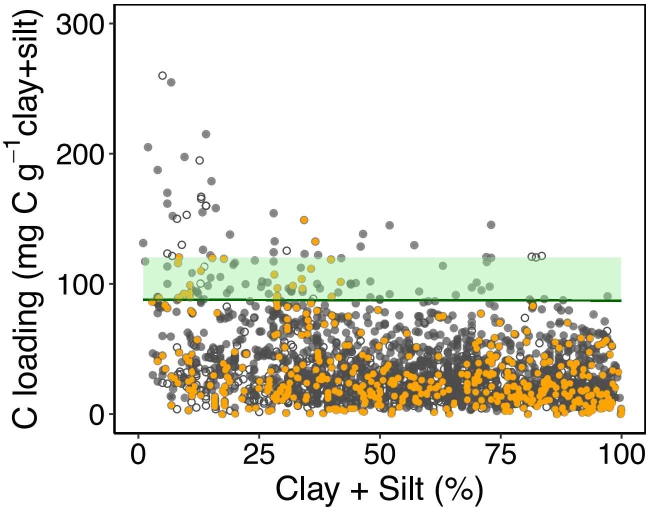

FIGURE 6 | Carbon loading in the MAOC fraction across clay + silt content in soils dominated by 2:1 phyllosilicate minerals. Size fractions are shown with gray symbols (n = 1751; solid and open circles denote natural and agricultural ecosystems, respectively) and density fractions are shown with orange symbols (n = 501). The *maximum observed capacity* is derived using the boundary line analysis, as the 95th percentile (green line) with shaded uncertainty spanning the 99th percentile (as shown in Figure 5a).

bias both size and density separation methods, though in different ways.

The dispersion of aggregates is typically achieved via shaking or sonication, while striving to avoid breaking POC into fine fragments that may lead to the overestimation of MAOC, especially when separated by size (Amelung and Zech 1999; Cerli et al. 2012; Golchin et al. 1994; Poeplau and Don 2014). Given the high C content of POC (~20%–30%), fine POC contamination could bias estimates of C loading (i.e., mg C g^{-1} clay + silt fraction) towards higher values. Methodological biases are expected to affect the quantification of MAOC proportionally more in soils with higher POC/MAOC ratios and/or smaller amounts of fine fraction—for example, in soils with high sand (low clay + silt) content—resulting in a perceived oversaturation of these soils. Indeed, estimates of C loading in sandy soils are higher from size separations, while those from density separations appear to correspond more closely to the BL-derived *maximum observed capacity* (Figure 6; Supporting Information). Additionally, it is possible for density separations following insufficient disruption of soil aggregates to overestimate MAOC due to the inclusion of occluded POC (Golchin et al. 1994; Wagai et al. 2009b). Thus, future studies that use both fractionation methods on a systematic set of target samples are needed to constrain methodological biases and verify maximum C loading estimates. In soils where methodological biases may be more pronounced, we strongly encourage the inspection of the MAOC fraction (e.g., by C:N ratios, isotopic composition, or infrared spectra) (Chenu et al. 2015; Leuthold et al. 2024; Sollins et al. 1999) and recommend a combination of density and size fractionation when possible (Mirabito and Chambers 2023).

Finally, we note that each of the procedures currently used for aggregate dispersion and the subsequent fractionation by size or density has some drawbacks (Kaiser and Berhe 2014; Plaza et al. 2019). As a result, there is not a perfect match between the fractionated MAOC and its conceptual definition (Leuthold et al. 2023). Furthermore, fractionated MAOC includes both mineral-organic and organo–organic interactions (Figure 3; Possinger et al. 2020), which are represented as distinct pools in select process-rich models and may differ in their relative stability and saturation dynamics (Section 4). Although using more complex fractionation schemes—either combining one or more size and density separation steps (e.g., Moni et al. 2012) or separating free MAOC from MAOC occluded in different aggregate fractions (e.g., Stewart, Paustian, et al. 2008; Fulton-Smith and Cotrufo 2019)—could result in more consistently defined MAOC, the low OC recoveries and throughput associated with complex fractionation schemes make these procedures difficult to implement (Poeplau et al. 2018), especially in large-scale analyses.

4 | Utility of the Carbon Saturation Concept

4.1 | Informing the Potential for Carbon Accrual

How and when MAOC saturates can have major implications for SOC accrual, including efforts to manage it directly in soils. Of particular interest is whether MAOC saturation impacts the efficiency with which C inputs are transformed into MAOC. This expectation is based on adsorption–desorption kinetics, as modeled and proposed by Hassink and Whitmore (1997), where the rate of new MAOC formation depends on the fraction of the capacity already occupied by MAOC. That is, for a given increase in C inputs and all else being equal, a soil further from its capacity may display greater MAOC accrual than one closer to its capacity. As a soil approaches MAOC saturation, new C inputs may be more likely to remain as POC (also see Section 3.1.4) (Castellano et al. 2015; Gulde et al. 2008) or, if transformed into microbial products, may be less likely to form MAOC, resulting in higher respiration and less SOC formed per unit C added (Stewart, Paustian, et al. 2008).

Thus far, only a handful of experiments have investigated C accrual in relation to MAOC saturation, and even fewer measured changes in MAOC in addition to SOC and respiration, with contrasting results. For instance, in a 2.5-year incubation experiment using soil depth as a proxy for saturation status, litter additions to C-horizon soils (lower saturation) resulted in higher SOC and lower CO_2 efflux than A-horizon soils (higher saturation) (Stewart, Paustian, et al. 2008). Yet, the opposite was observed in an incubation experiment of soils with differing initial SOC, with higher SOC formation efficiency from added litter occurring in soils with higher initial SOC contents (Wu et al. 2024). We note, however, that the saturation status of the soils in Wu et al. (2024) spanned a relatively narrow range—from 14.5% to 27.4% by dividing MAOC (estimated as SOC minus POC) by the *maximum observed capacity* from the BL approach (Section 3.1.2) or 25.9% to 52.2% from the Hassink equation (Section 3.1.1). Furthermore, both experiments likely experienced confounding factors that affect MAOC formation efficiency, including microbial community composition (Craig et al. 2021; Kallenbach et al. 2016) and soil pH (Malik et al. 2018). Indeed, the effect of MAOC saturation status on MAOC accrual can be difficult to discern in practice because multiple mechanisms can contribute to nonlinearities and thresholds in the MAOC response to C inputs. For instance, soils could accumulate MAOC faster once an initial degraded state is surpassed, due to changes in microbial community composition and carbon-use efficiency (Craig et al. 2018, 2021), organo–organic layering (Schweizer 2022), or aggregation. While the complex MAOC dynamics emerging from these interacting processes may obscure the relationship between MAOC saturation status and MAOC accrual, they do not necessarily invalidate its existence. Experiments that parse apart these interacting influences are difficult, yet necessary for probing the link between C accrual and MAOC saturation.

Finally, we note that studies differ in their use of MAOC deficit or saturation status—calculated by subtraction or ratio of MAOC to the *maximum observed capacity*, respectively. These metrics can serve different functions. MAOC saturation status (as a percentage) may be more appropriate for informing the efficiency of further MAOC formation, since it relates to the curvature of the saturating relationship, regardless of the absolute limit (Georgiou et al. 2022; Hassink and Whitmore 1997). In contrast, MAOC deficit (as a concentration or stock) may be useful for estimating absolute MAOC accrual potentials (Angers et al. 2011; Chen et al. 2018) and may even be calculated based on *effective capacities* instead of the *maximum observed capacity*. Indeed, achieving the *maximum observed capacity* may not be realistic everywhere, since climatic constraints are difficult to overcome. For agricultural regions in temperate climates, MAOC concentrations in natural ecosystems may serve as an attainable upper limit (i.e., *effective capacity*) to strive for under management (Georgiou et al. 2022), unless nutrients or water limit productivity in the natural ecosystems (Six et al. 2002). For arid climates, managed ecosystems can reach higher *effective capacities* than natural ecosystems—as demonstrated for Australia (Viscarra Rossel et al. 2024). Furthermore, studies continue to calculate both MAOC deficit and saturation status using the Hassink equation (Heinemann et al. 2024; Soinne et al. 2024; Wu et al. 2024), though it can greatly underestimate observed MAOC (Section 3.1.2). These decisions influence conclusions on the utility, or even existence, of MAOC saturation, and thus, we urge future studies to carefully consider the appropriate metric and capacity for their application (Table 1).

4.2 | Proximity to Saturation May Affect Vulnerability

In addition to influencing C accrual rates, the MAOC saturation status of a soil may also affect its vulnerability to lose C under disturbances caused by changes in land use or climate. However, there are contrasting hypotheses and very limited evidence to date for whether soils closer to or further from MAOC saturation are more vulnerable to desorption and subsequent decomposition under disturbance.

For instance, greater loss of MAOC could occur from soils closer to MAOC saturation due to more loosely held or exchangeable MAOC and organo–organic associations (Kögel-Knabner et al. 2008). In this context, it is especially important to consider the fractionation method and sonication energy, which can impact the distribution of exchangeable MAOC and fine POC between measured fractions (Section 3.2). Results from an 11-year fallow experiment on a sloped site with spatial variation in MAOC saturation status indicated that SOC loss was higher in soils with higher initial SOC (Meyer et al. 2017), even when accounting for potential regression to the mean effects (Lark et al. 2006; Slessarev et al. 2023). Although a fraction of SOC losses was driven by POC, Meyer et al. (2017) concluded that the majority of losses were attributable to MAOC and were positively correlated to MAOC saturation status (calculated using the Hassink equation). We note, however, that this study used high sonication energies, potentially leading to fine POC contamination in MAOC that could also explain this result (Six et al. 2024). Nevertheless, taken together with the smaller C gains near MAOC saturation (Section 4.1), the larger C losses near MAOC saturation suggest a potential hysteresis in MAOC as a function of C inputs that requires further study.

In contrast, based on adsorption–desorption kinetics alone (e.g., represented by a saturating Langmuir relationship) and no hysteretic behavior, smaller C losses could instead occur near MAOC saturation—where the relationship between MAOC and C inputs is expected to plateau. That is, soils further from MAOC saturation could be relatively more vulnerable to MAOC loss, for example, in response to warming temperatures. This is because, mathematically, MAOC depends more strongly on the temperature-sensitive ratio of adsorption to desorption constants in the limit of low C concentrations, as illustrated by Georgiou et al. (2022) using a global data synthesis. However, MAOC is a heterogeneous and dynamic pool that is not exclusively formed by surface adsorption (Kleber et al. 2021), and the nature of organo–mineral or organo–organic associations will likely play an important role in the MAOC response. Further work is needed to constrain the vulnerability of MAOC (and SOC) to warming across soils that differ in their MAOC saturation status, while importantly considering the distribution between MAOC and POC fractions when investigating SOC responses (Rocci et al. 2021).

Ultimately, it remains uncertain whether the degree of MAOC saturation has a positive or negative effect on MAOC vulnerability, and how context-dependent this effect may be, given the limited evidence thus far. It is also possible that different types of perturbations (e.g., changes in temperature, dry/wetting, or C inputs) may lead to contrasting conclusions on the vulnerability of MAOC (or SOC) as a function of MAOC saturation. Thus, it is critically important for future experimental studies to probe and elucidate the underlying mechanisms of C loss for a range of disturbances and for synthesis studies to account for the type of disturbance when drawing conclusions on the utility of MAOC saturation for informing a soil's vulnerability to lose carbon.

4.3 | Representations in Process-Based Models

Advancing predictive understanding of SOC dynamics in response to climate and land-use change relies on process-based models, especially at regional to global scales. Because environmental changes often drive large alterations in C inputs to soil, the representation of MAOC saturation can play an important role in resulting model predictions. Furthermore, the distribution of C among MAOC and other soil pools mediates emergent

SOC responses. However, the biotic and abiotic processes driving the formation pathways and capacities of MAOC are either absent or vary widely across models (Sulman et al. 2018).

Critically, most SOC models with first-order kinetics (as used in all Earth System Models) lack the representation of a *theoretical mineral capacity* and also lack explicit microbial mechanisms that could constrain SOC responses to C inputs (Wieder et al. 2015). As a result, such models predict strong linear increases in all SOC pools in response to increased plant productivity or C input rates (Craig et al. 2021; Georgiou et al. 2017; Six et al. 2002), a prediction that is often not borne out in ecosystem experiments (Sulman et al. 2018; Terrer et al. 2021). Although modifications to first-order SOC models that allow for MAOC saturation have been explored in past studies (Segoli et al. 2013; Stewart et al. 2007), recent efforts aim to explicitly represent microbe-mediated MAOC formation pathways and mineral capacities within more process-rich SOC models (Sulman et al. 2018).

At present, process-rich SOC models vary in their assumptions about the saturation (or lack thereof) of mineral surfaces, the composition and structure of MAOC pools, and the microbial processes that drive *effective capacities*. Models that represent saturating sorption to minerals typically employ formulations based on a Langmuir isotherm (Abramoff et al. 2022; Ahrens et al. 2020; Zhang et al. 2021) or less commonly a Freundlich isotherm (Grant et al. 1993). Both equations are used to describe the relationship between solute concentrations and MAOC and differ in that the Langmuir represents an explicit *theoretical mineral capacity* whereas the Freundlich is an empirical power-law that allows MAOC to accumulate in perpetuity at an increasingly diminished rate. Both approaches typically result in attenuated responses of MAOC to C inputs when compared with first-order or process-rich models that lack MAOC saturation.

Process-rich SOC models also vary in how saturation is represented and parameterized across pools. For instance, the MEND model assumes that only a small kinetic fraction of MAOC can saturate (Wang, Post, et al. 2013), MILLENNIAL applies a capacity to sorbed DOC but not microbial residues (Abramoff et al. 2022), and COMISSION and MEMS have a capacity for both sorbed DOC and microbial residues (i.e., exchangeable and stable MAOC, respectively, in MEMS) (Ahrens et al. 2020; Zhang et al. 2021). Predictions can be highly sensitive to these assumptions as well as the parameters—for example, using a capacity derived from DOC sorption experiments versus the boundary line (BL) approach (Section 3.1) (Ahrens et al. 2015, 2020). Although the BL-derived *maximum observed capacity* is typically used as the *theoretical mineral capacity* in models (Abramoff et al. 2022; Ahrens et al. 2020), future work should explore environmental factors that drive potential differences between these two capacities (Table 1).

Notably, in some process-rich SOC models, biotic controls can enforce a limit on MAOC or SOC, even in models that do not represent an explicit capacity of minerals. In particular, the representation of density-dependent constraints on microbial growth and turnover can modulate the simulated relationship between C inputs and SOC accrual (Georgiou et al. 2017). Such biotic controls are likely to be an essential ingredient in modeling apparent saturation behavior (i.e., *effective capacities*), as shown by Craig et al. (2021) for MAOC and even POC. Models that simulate a strong microbial priming effect may also exhibit strongly diminished or even absent SOC responses to increased C inputs (Wang, Chen, et al. 2013).

Given prevailing uncertainties in both process understanding and model formulations, tighter coupling is needed between empirical and modeling studies to probe the range of potential responses—that is, from complete insensitivity to saturating behavior to strong linear increases in MAOC. For example, model comparisons could be used for hypothesis testing and to identify process representations that have large effects on simulated patterns of MAOC saturation. These uncertain processes could then be targeted with experiments, and data from long-term input gradients could be used for model validation. Such integrated efforts offer a unique opportunity to enhance understanding of the mechanisms and implications of MAOC saturation.

5 | Conclusions and Future Directions

Soil C saturation is not only an interesting theoretical concept but also stands to inform process-rich SOC models and targeted climate mitigation and soil health initiatives. Therefore, the concept of C saturation—especially as it pertains to MAOC—has seen much renewed interest and debate in recent years. However, findings on the utility, or even existence, of MAOC saturation can be greatly influenced by the selected capacity (Table 1) and method of quantification. Careful consideration is thus needed depending on the application. In particular, we encourage the use of the *maximum observed capacity* (derived globally from the clay + silt boundary line for each mineral category) in estimating MAOC saturation status but also note that it may be difficult to reach this capacity in many soils due to other inherent limitations. Future work should continue to explore additional mineral categories (e.g., non-crystalline minerals) (Beare et al. 2014), as well as reactive metals, which play important roles in stabilizing organic matter in soils (King et al. 2023; Rasmussen et al. 2018; von Fromm et al. 2021). Many geochemical variables are often not measured or reported together with MAOC but have the potential to complement the clay + silt content and broad mineral categories to better characterize mineral surface area and reactivity. Furthermore, studies may explore mechanisms underlying potential nonlinearities in the *maximum observed capacity* as a function of clay + silt, as well as the biotic factors driving *effective capacities*.

Recent studies have used microspectroscopic and microspectrometric techniques to advance understanding of the spatial distribution and dynamic nature of MAOC. The observed patchy surface coverage and layering of OC on minerals illustrate the complexity of MAOC and highlight that MAOC saturation is not a monolayer. Further work is needed to better characterize the concept of MAOC saturation at the micro-scale and reconcile findings with field-scale observations. For instance, microspectroscopic characterization of soils near their *maximum observed capacity* could offer valuable insights, especially regarding their proximity to a *theoretical mineral capacity* and the underlying mechanisms of stabilization. Furthermore, in-depth characterization of soils following targeted C accrual and vulnerability experiments—especially with isotopic tracers—would help to elucidate where new MAOC formation or loss occurs, as well as the stability and reversibility of organo–mineral associations.

The MAOC saturation status of a soil may ultimately impact the potential for further C accrual and the vulnerability of C loss. Although some studies have found less efficient MAOC accrual in soils closer to saturation, as expected from theory, evidence is still scarce and sometimes contradictory. Targeted experiments are needed to assess the formation of new MAOC under increased C input rates, especially in soils near their *maximum observed capacity* or across soils that vary in MAOC saturation status. Incubation or in situ manipulation experiments could also be used to probe the vulnerability of MAOC loss across these targeted soils. In particular, future experiments should explore several global change and land-use factors—including dry/wet cycles, warming, and changes in C inputs—that may result in contrasting MAOC responses. Capturing these responses in process-rich SOC models is critical for predictions of soil C cycling. The landscape of soil management is continuously evolving, and a predictive understanding of soil C storage and function in response to climate change and across pedoclimatic contexts is paramount for informed decision-making.

Author Contributions

Katerina Georgiou: conceptualization, data curation, formal analysis, methodology, visualization, writing – original draft, writing – review and editing. **Denis Angers:** conceptualization, methodology, writing – review and editing. **Ryan E. Champiny:** methodology, visualization, writing – review and editing. **M. Francesca Cotrufo:** conceptualization, methodology, writing – original draft, writing – review and editing. **Matthew E. Craig:** methodology, visualization, writing – original draft, writing – review and editing. **Sebastian Doetterl:** conceptualization, writing – review and editing. **A. Stuart Grandy:** conceptualization, visualization, writing – review and editing. **Jocelyn M. Lavallee:** conceptualization, data curation, methodology, writing – original draft, writing – review and editing. **Yang Lin:** methodology, writing – original draft, writing – review and editing. **Emanuele Lugato:** methodology, visualization, writing – review and editing. **Christopher Poeplau:** data curation, methodology, visualization, writing – original draft, writing – review and editing. **Katherine S. Rocci:** data curation, writing – original draft, writing – review and editing. **Steffen A. Schweizer:** conceptualization, visualization, writing – original draft, writing – review and editing. **Johan Six:** conceptualization, methodology, visualization, writing – review and editing. **William R. Wieder:** visualization, writing – review and editing.

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Conflicts of Interest

There are no conflicts of interest associated with this manuscript. M.F.C. discloses to be one of the cofounders of Cquester Analytics LLC, a service facility that provides analyses of soil carbon fractions.

Data Availability Statement

The data that support this review article are freely available from the original papers and open-access repositories listed in the Data Sources and Supporting Information. The final combined dataset is also archived on Zenodo (https://doi.org/10.5281/zenodo.15171280).

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Data Sources

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Supporting Information

Additional supporting information can be found online in the Supporting Information section.