

WHAT PROCESSES CAN CONTRIBUTE TO ENHANCING SOIL CARBON STORAGE AT GREATER SOIL DEPTHS?

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Abstract

Under some circumstances, a one-off use of full inversion tillage for pasture renewal can lead to an increase in overall soil carbon (C) stocks by burying and slowing the decomposition of C rich topsoil and exposing C poor subsoil to large fresh C inputs (pasture roots, litter, dung). However, it is not yet known what processes might be responsible for the slower decomposition of buried topsoil carbon. Here, we investigated the following factors that could potentially contribute to slower soil carbon turnover at depth in the soil:

- 1) Soil temperature;
- 2) Soil moisture;
- 3) Maximum protective capacity (for soil carbon) of the mineral soil;
- 4) Oxygen concentration / redox potential;
- 5) Biological synergy (priming) whereby the decomposition rate of poorly decomposable material is positively correlated with the overall activity of the decomposer community.

Deeper soil layers experience lesser temperature fluctuations than upper layers, which can reduce decomposition rates at depth. However, detailed simulations for NZ soil showed this effect not to be quantitatively important with decomposition rates changing by only a few percent. Similarly, upper soil layers undergo more intense wetting and drying cycles, whereas deeper soil layers are likely to experience more stable moisture conditions. In dryer regions, soils at depth can be permanently dry which could greatly reduce decomposition rates. Carbon storage in topsoils could, in principle, also be limited by soils' maximum protective capacity, which could be overcome if undersaturated soil is brought up from the sub-soil. However, there is no tangible evidence that such maximum storage capacity actually limits carbon storage in any soils, and hence, there is no evidence that overcoming maximum storage limitations could enhance overall carbon storage. Oxygen limitations could potentially play a role in regions with very high rainfall or in soils with restricted drainage where soils remain water-logged for extended periods of time. However, the role of oxygen limitation is likely to be restricted to particular sites and conditions where prolonged water logging can occur.

Finally, we investigated the potential role of biological synergy (priming) which essentially assumes that the specific organic carbon decomposition rate can be enhanced if there is greater overall decomposer activity in the soil. This provides a plausible mechanism that could lead to greater persistence of soil carbon at depth irrespective of specific soil and environmental conditions. If further experimental work can further substantiate the role of this mechanism

under typical field conditions, it could support the role of full-inversion tillage as a climate change mitigation option in grasslands.

Introduction

With ongoing climate change, there is a desire to decrease net greenhouse gas emissions, and one way to achieve that is by sequestering more atmospheric carbon (C) in soils (Paustian et al., 2016; Smith, 2016). One way to achieve extra carbon storage could be by burial of carbon deeper within the soil through full-inversion tillage (FIT) during pasture renewal (Calvelo Pereira et al., 2018, 2020; Beare et al., 2020; Lawrence-Smith et al., 2020). Deep tillage (> 60 cm depth) increased soil carbon stocks on long-term by on average 43% in sandy and loamy croplands of Northern Germany (Alcantara et al. 2016). Buried topsoil carbon was preserved while a new topsoil accumulated organic carbon over more than 40 years.

Full-inversion tillage causes both the transfer of low C subsoil to the surface where they can then receive fresh carbon inputs from roots, litter and dung and the burial of C-rich topsoil deep into the subsoil where it can potentially increase its permanence, which together could increase soil C storage. In New Zealand, Calvelo Pereira et al. (2018) reported a short-term increase in C stocks following full-inversion tillage of an extra 13.9 tC ha⁻¹ (0-30 cm) over the first 4 years of the study. A review by Lawrence-Smith et al. (2020) suggests an increase of 36 MtC could be achieved by the introduction of FIT during pasture renewal in New Zealand. Also, carbon burial with deep soil flipping of NZ grasslands resulted in an extraordinary carbon sequestration of more than 160 tC ha⁻¹ (Schiedung et al., 2019). Longer-term monitoring of changes in soil carbon storage is needed to verify benefits of FIT for increasing soil C stocks. On farm research trials have commenced to test this potential (Beare et al., 2020; Calvelo Pereira et al., 2020).

Here, we have explored the theoretical potential of five possible mechanisms (temperature, soil moisture, maximum protective capacity, oxygen limitation and the role of biological synergy), to explain soil C sequestration following FIT mostly by running specific simulations targeted to address each specific process. Through that, we aimed to contribute to a better understanding of possible mechanisms that could enhance carbon storage at depth.

Results and Discussion

Temperature variability

In soils, temperatures generally fluctuate more nearer the soil surface, whereas temperatures tend to be more stable deeper within the soil (Fig. 1a). Decomposition is generally held to be a strongly non-linear function of temperature (e.g. Kirschbaum, 2000; Fig. 1b). This implies that

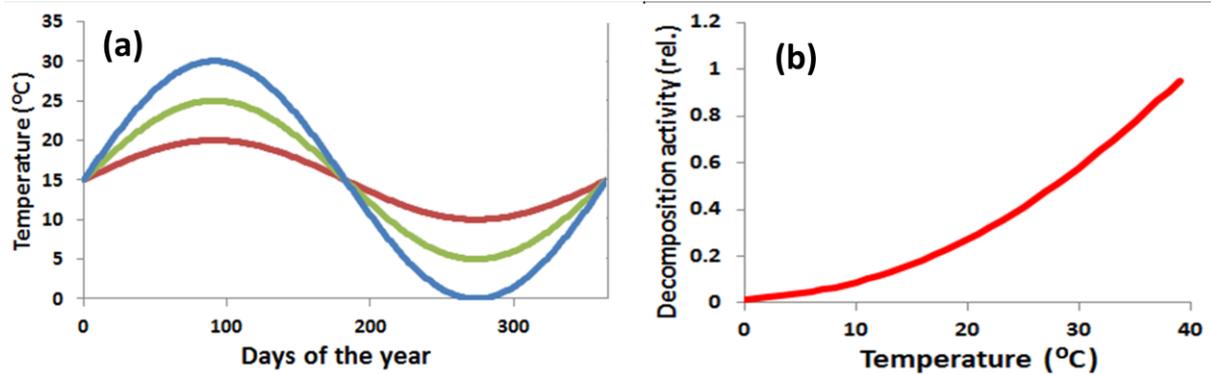


Figure 1: The interaction between soil depth and decomposition activity, showing (a) a notional seasonal temperature cycle at three notional depths in the soil and (b) the response of decomposition activity to temperature (after Kirschbaum, 2000).

decomposition could be stimulated by temperature variations, with more activity gained when temperatures exceed averages than would be lost when temperatures are below average because the temperature response curve is steeper at higher than lower temperature. For a soil that is always moist, that would suggest that decomposition activity would be higher in the topsoil along with its greater temperature variations.

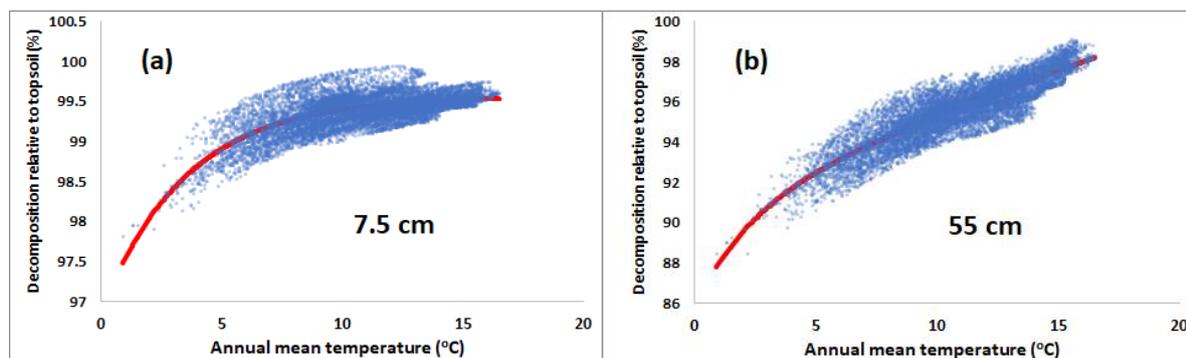


Figure 2: Reduction in annual decomposition activity at depths of (a) 7.5 cm or (b) 55 cm in the soil. This graph considers only temperature effects, with data at lower depths expressed as a percentage of expected decomposition activity at the top 2.5 cm. Simulations were conducted for 10,491 specific locations from New Zealand.

The quantitative significance of that temperature effect was explored by running simulations for actual observed annual weather sequences from New Zealand (Fig. 2). The simulations confirmed the expected patterns: compared to the top 2.5 cm of the soil, annual decomposition activity was reduced by 1-2% at 7.5 cm depth (Fig. 2a) and by 2-10% at 55 cm depth (Fig. 2b), with the extent of reduction decreasing with increasing temperature. The calculations suggest, however, that the temperature effect on its own may be insufficient to explain the large observed increases in carbon storage at depth that have been observed in some studies.

Temperature and Moisture Interactions

Bioclimatic effects become considerably more complicated once temperature-moisture interactions are factored in as well. When soils dry out, decomposer activity can be greatly curtailed (Paul et al., 2003). The question is whether drying has a greater effect on topsoils or soil layers deeper in the profile. We investigated that by first considering modelled soil water contents for an intensively studied dairy farm (Scott Farm) in New Zealand's Waikato region

(Kirschbaum et al., 2015) where we used the CenW model to simulate soil water contents to a depth of 70 cm (Fig. 3).

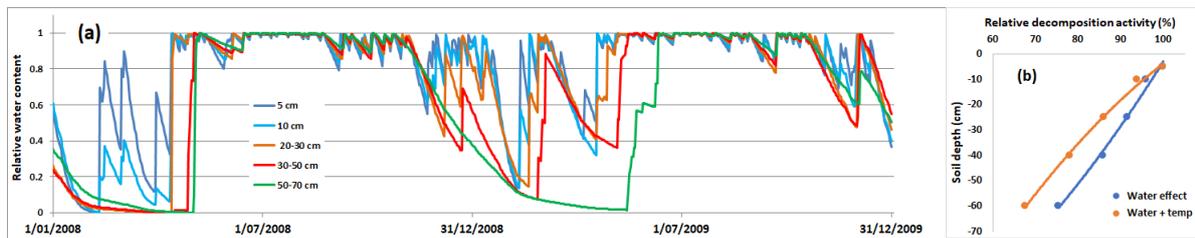


Figure 3: Soil water contents and consequent limitations to decomposition activity in the soil, showing (a) Modelled soil water contents at five different soil depths over two years for a dairy farm in the Waikato region (after Kirschbaum et al., 2015) and (b) relative decomposition activity as a function of soil depth. This site received 1267 mm yr^{-1} rainfall.

The experimental period started with an intensive drought period during which all layers of the soil dried out (the first 3-4 months in Fig. 3). There were a few smaller rainfall events, but they rewetted only the upper soil layers while the deep soil remained dry until heavier rainfall events rewetted the whole profile in late autumn. The whole soil then remained wet throughout winter and the following spring, but over the summer months, the previous year's drought conditions were repeated, with the deepest soils layers continuing to dry out while upper soil layers were occasionally rewetted by smaller rainfall events that were not heavy enough to rewet the whole profile until late autumn (Fig. 3a).

When that pattern of soil water was combined with a simple function to describe the dependence of decomposition rate on soil moisture, it showed total decomposition activity to decrease with soil depth (Fig. 3b), with decomposition activity in the deepest layer being only about 75% of that in the topsoil. The drying effect occurred mostly over the summer months. If the soil had remained wet at that time of the year, it could have allowed the fastest decomposition activity. So, since the drying effect reduced decomposition activity predominantly over the summer months, it led to an even greater effect of the combined temperature and soil drying effect in making the subsoil less conducive to decomposition activity than the topsoil, with the deepest soil only allowing about 2/3 of the decomposition activity of the topsoil (Fig. 3b).

The Scott-Farm site (Fig. 3) was reasonably wet, receiving annual rainfall of 1267 mm yr^{-1} over the experimental period. To further investigate how the pattern observed in Figure 3 might shift with annual rainfall, we again utilised weather data sets from New Zealand, including rainfall and solar radiation as inputs to calculating soil water balances. For this work, we used 10% of the available data sets.

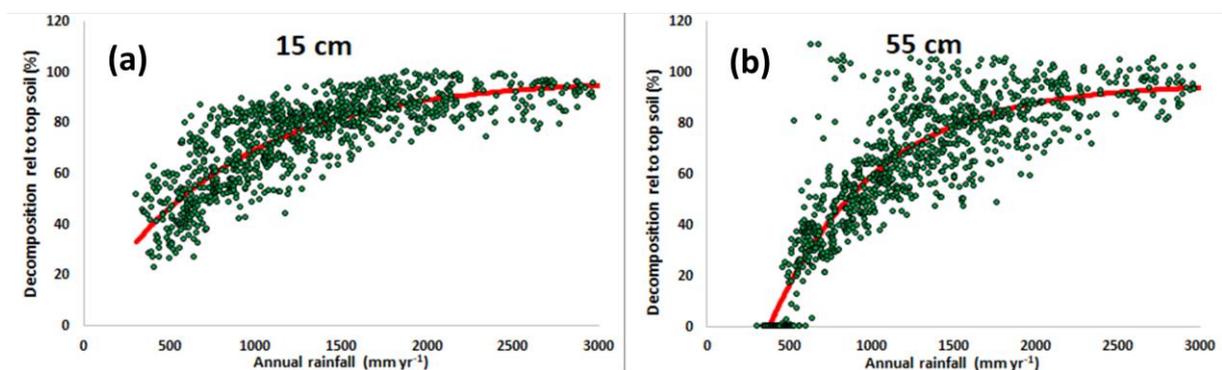


Figure 4: Total annual decomposition activity at (a) 15 cm and (b) 55 cm relative to that calculated for the upper soil layer, expressed here as a function of received annual rainfall.

The simulations showed that for sites receiving annual rainfall of more than about 1500 mm yr⁻¹, decomposition rates become independent of soil depth. There was generally enough rain for soils to remain wet at all depths (Fig. 4). For drier soils, however, the effect of soil depth on decomposition rates became increasingly prominent, with soils receiving only 500 mm yr⁻¹, decomposition activity at 15 cm depth was calculated to be only about 50% of that in the topsoil (Fig. 4a). The pattern became even more pronounced at 55 cm depth, with many soils at that depth and with annual rainfall of about 500 mm yr⁻¹ never getting moistened and remaining in a permanent dry state (Fig. 4b).

Where such dry subsoils can be found requires further experimental data and field observations. Preferential flow paths causing direct rainfall infiltration into the subsoil were not accounted for in the simulated soil moisture gradients (e.g. Hardie et al 2013). The calculations indicated that in drier regions, carbon at depth could well be protected through soils remaining dry over prolonged periods, and under very dry conditions, that could even amount to permanently dry conditions. Any buried carbon could then well be protected from decomposition through the dry conditions.

Maximum protective capacity

In principle, if the topsoil has reached a maximum protective capacity then any newly added C may simply decompose without being stabilised. Soil inversion could be beneficial by storing the protected carbon at depth where it remains protected while newly added carbon can be protected by bonding with the undersaturated mineral soil brought up from depth. Provided that matrix protection provides a very strong protection that renders C resistant over centuries, that mechanism could, in principle, provide persistent long-term storage.

However, the role of that mechanism would be restricted to soils with saturated topsoils. It could also play a role only if the whole notion of maximum protective capacity is indeed valid at all. At present, there appears to be no tangible evidence in support of the existence of such limits. In previous work, Kirschbaum et al. (2018) studied the distribution of soil carbon contents in New Zealand's national soils database and tried to identify a pattern in their distribution that would have indicated a controlling role for maximum protective capacity. However, the observations did not follow a pattern that would have indicated a role for maximum storage capacity. Instead, the distribution of soil carbon contents followed a pattern that would have been expected if none of the soils had reached a maximum storage limit (Kirschbaum et al., 2018). We therefore concluded that there is no tangible evidence to support the overcoming of a maximum storage capacity as a factor that could contribute to the usefulness of FIT.

Oxygen Limitations

Organic carbon decomposition is the oxidation of reduced carbon compounds. High rates therefore require adequate availability of oxygen although some decomposition can also occur under anaerobic conditions with CH₄ (instead of CO₂) as the end product. Ongoing decomposition at high rates does require oxygen, however, or carbon continues to accumulate as it does in peat bogs, for example (e.g. Kuhry and Vitt, 1996). The inversion of soils through deep inversion tillage could potentially take soil organic carbon from an aerobic zone near the soil surface and deposits it deeper within the soil where oxygen access may be more restrictive, especially in soils with restricted drainage or a high ground water table, or if they receive frequent rainfall that may lead to water saturated conditions in deeper soil horizons for part of the year.

This was further explored here by considering the fate of soil organic carbon placed at different positions within the soil with different ease of access to oxygen. The different locations were characterised here by their relative oxygen diffusion resistances from the atmosphere with a normal concentration of 21%. It was further assumed that the decomposability of organic carbon decreased in proportion to the amount of carbon remaining and that the draw-down in oxygen concentration was proportional to the actual decomposition rate. Hence, the oxygen draw-down decreased over time as the reducing quality and quantity of any remaining soil carbon reduced its specific and absolute decomposability and its consequent oxygen requirement.

The simulations showed decreasing organic carbon decomposition rates over time (Fig. 5b) as some carbon decomposed and less carbon remained and its quality reduced. These reducing decomposition rates consequently affected the oxygen draw-down (Fig. 5a), which trended upwards over time as the reducing organic carbon decomposition rate lowered the oxygen demand. From the assumption of easy access to oxygen under the default assumption (relative diffusion resistance = 1), substantial increases in diffusion resistances would be required to make any substantial difference to the patterns of soil carbon loss over time.

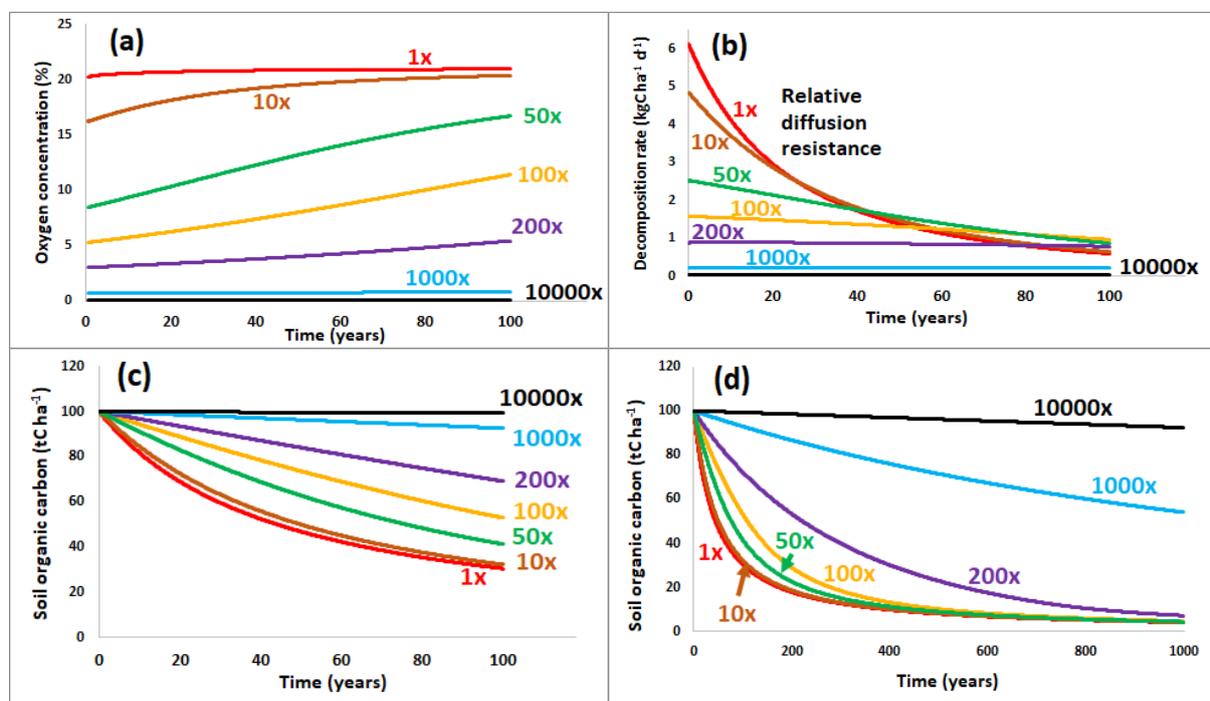


Figure 5: The interaction between oxygen concentration and organic carbon decomposition. Simulations are shown here for conditions characterised by different oxygen diffusion resistance relative to that placed near the soil surfaces as shown by the numbers in the figure. The figure shows (a) oxygen concentration over time and with different diffusion resistance, (b) resultant decomposition rates and remaining organic carbon over the next (c) 100 or (d) 1000 years.

A ten-fold increase in diffusion resistances was modelled to make only a very minor differences to the amount of carbon remaining over time (Fig. 5c) even though there was some difference in the initial decomposition rates (Fig. 5b). Diffusion resistances needed to increase more substantially (50 to 100-fold) to make more tangible differences to the patterns of carbon loss (Fig. 5c), and even increases in diffusion resistance of those magnitudes would have little effect at times scales beyond 100 years (Fig. 5d).

Diffusion resistance would have to increase by 1000 to 10,000 times to tangibly reduce patterns of organic carbon loss over time frames beyond 100 years (Fig. 5d). Such increases in diffusion resistance could be found in soils with permanent water coverage, such as in peat bogs, but are unlikely without the presence of water as a sealant. Under such conditions, oxygen limitation could be possible and act as a mechanism of protection of soil carbon if it is placed deeper in the soil where a permanent water seal restricts access to oxygen.

Biological Synergy

Soil organic carbon models such as CENTURY (Parton et al., 1987) and RothC (Jenkinson and Rayner, 1977) mostly consider the decomposition of organic carbon in specified pools to follow simple first-order decay kinetics in dependence on the size of those pools and biophysical factors such as temperature, soil moisture, soil texture and, for some pools, a measure of the quality of organic carbon. This has worked very well for most applications that these models have been tested against, but it is not certain whether they also correctly describe the patterns under the unique condition of carbon buried deep within the soil profile.

Don et al. (2013), for example, showed that soil decomposition rates can be more rapid when organic carbon becomes more concentrated. More generally, Kuzyakov et al. (2010) described a range of observations, collectively termed ‘priming effects’ or ‘biological synergy’, that showed that the decomposition of soil organic carbon can be affected by the presence of additional (easily available) carbon sources. Subsoil carbon stability has been attributed to lack of such fresh C inputs (Fontaine et al. 2007) In essence, the concept of biological synergies postulates that fresh labile organic carbon in the soil generates a micro-environment replete with microbes and extracellular enzymes that are primarily generated to process the fresh available carbon sources, but these cells and their enzymes would also be able to degrade older and less readily available soil organic carbon. So, reduced carbon influx, as would occur after soils have been relocated to depth through FIT, would lead to the loss of those agents and thus slow decomposition rates.

This is simulated here through a simple conceptual model of soil organic carbon and its decomposition (Fig. 6). It distinguishes three types of soil organic carbon (active, slow and resistant), with the slow and resistant pools typically comprising the majority of soil organic carbon (as illustrated through the size of the boxes). In contrast, the flux of organic carbon in the soil predominantly flows through the active pool, with lower flux rates through the slow pool and very low rates through the resistant pool (as illustrated by the thickness of the arrows).

Decomposition rates of these pools are determined by the size of respective pools, illustrated by the red arrows, assumed to follow simple first order decay dynamics. At steady state, these loss rates are matched by carbon influx of the same magnitude into each pool so that pool sizes do not change once equilibrium has been reached. This model is essentially a simplified version of the well-established CENTURY model (Parton et al., 1987) and its CenW derivative (Kirschbaum & Paul, 2002).

In addition, one can assume a degree of biological synergy between these pathways, illustrated by the green arrows in Fig. 6. It assumes that the decomposition of slow and resistant organic carbon depends not only on external factors and the size of these two reservoirs but is also enhanced by the decomposition rate of the active pool, which is the primary process determining the activity of the decomposer community.

Figure 7 illustrates the fate of organic carbon after a soil has been inverted at year 1. It is assumed here that before inversion, all pools (active, slow, resistant) in the topsoil were at steady state with inputs equal to outputs, and all pools in the subsoil were zero. After inversion, the subsoil was assumed to receive no further carbon inputs, leading to gradual loss of the pools whereas the new topsoil received the same input of fresh carbon as the topsoil did before inversion, thus allowing a gradual build-up of new respective pools. Without biological synergy, gains would match losses perfectly, with no overall change in total SOC, but biological synergy created an asymmetry.

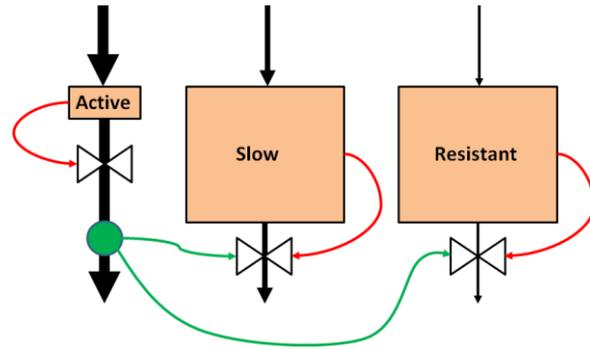


Figure 6: Conceptual 3-pool model of soil organic carbon. Red arrows represent the standard formulation of organic carbon models, illustrating the first-order dependence of organic carbon degradation in dependence on their own pool sizes. Green arrows illustrate the additional biological synergy illustrated here.

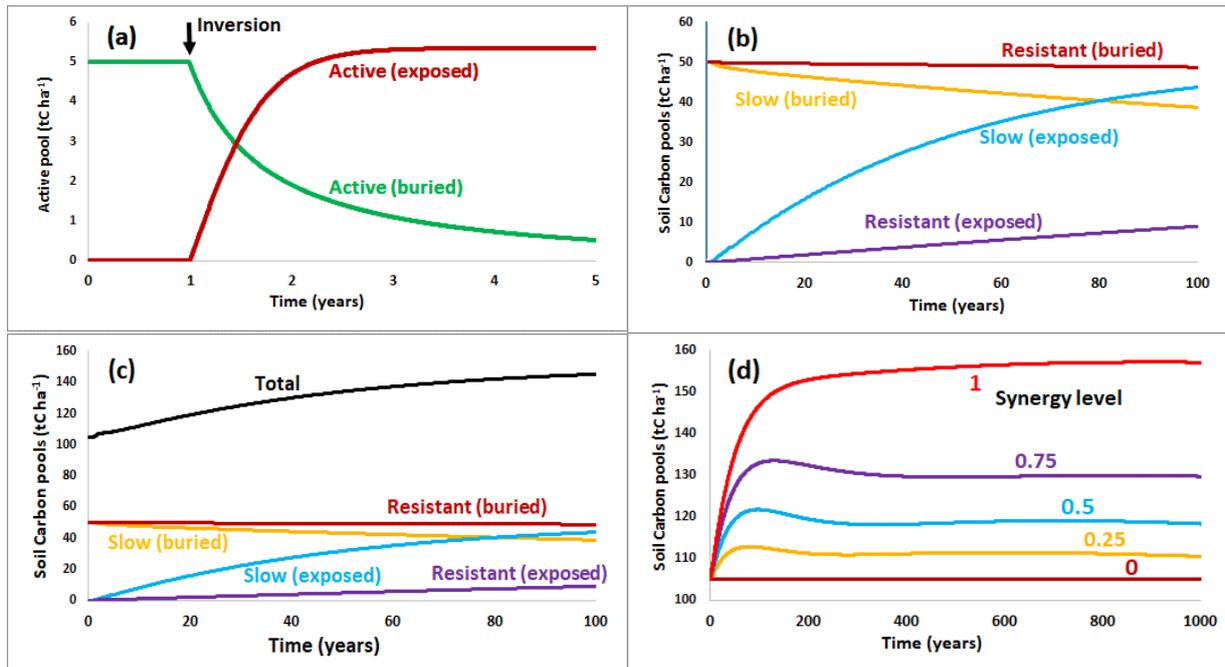


Figure 7: Illustrating the functioning of a conceptual 3-pool soil-organic-carbon model with biological synergy. The model run started with all pools in steady state. At year one, the soil was inverted so that the inverted soil received no further inputs of fresh organic carbon, while the newly exposed soil was assumed to have zero initial pool sizes but received new carbon inputs from that time onwards. Shown here are (a) active SOM in the exposed and buried soil layers shown over 5 years, (b) slow and resistant SOM in the same layers over 100 years, (c) the same pools as in (b) plus total SOC and (d) the total amount of SOC over 1000 years with different levels of biological synergy.

For the active pools, the asymmetry was only slight but apparent even for those pools (Fig. 7a). While a new active pool had essentially established itself within 2 years, the loss of the buried active pool started with a rapid initial loss, but after the first couple of years, its ongoing rate of loss slowed, and after about 5 years, about 5% of the initial mass of the active pool still remained. Changes in the sizes of the slow and resistant pools were quantitatively much more important for total carbon balances (Fig. 7b, c). The formation of a new resistant pool in the newly exposed topsoil was clearly faster than the loss of the buried resistant pool. However, its decomposition rate was so slow even with maximum stimulation through biology synergy that even after 100 years, the new resistant pool had not yet built up to even 10 tC ha⁻¹, making its asymmetric changes quantitatively not very important over that initial period over the first 10 years, although its changed dominated the slower soil carbon changes at the time scale beyond the first century (Fig. 7d).

Instead, over the first 100 years, the most important contribution came from asymmetric changes in the slow pool (Fig. 7b). Its initial size was large, and once biological synergy had been lost as an agent to stimulate its decomposition slowed considerably. The slow pool of the exposed soil increased in size significantly over the time frame of decades to almost reach its steady size within less than 100 years.

Summing carbon in all organic carbon pools then yields an overall increase in site carbon from 105 to about 140 tC ha⁻¹ (Fig. 7c) as buried carbon was protected from decomposition through the cessation of biological synergy while the newly exposed subsoil built sizeable pools through the ongoing input of fresh carbon. These changes became even greater over the longer time frame of centuries and were effectively not reversed (Fig. 7d). The extent of these overall carbon gains was essentially proportional to the extent of the biological synergy (Fig. 7d). If

decomposition rates would be only partly stimulated through biological synergy then the overall carbon gains through soil inversion would also only partly be realised.

Overall, these stimulations clearly illustrate how resistant carbon in the soil can, in principle, be protected and remain in the soil almost indefinitely once the decomposition engine provided through biological synergy is lost as it would through burial of carbon at depth without ongoing contact with fresh carbon inputs.

General Discussion and Conclusions

The most obvious benefit lies in changes in temperature variability with depth with the consequence of slowing decomposition rate under less variable temperatures (Fig. 1). However, detailed simulations showed that, while the effect is likely to be real and slow down decomposition at depth, it only makes a very small difference (Fig. 2). On its own the temperature effect is unlikely to provide a sufficient rationale for carbon gains through FIT.

Water limitations could potentially play a more important role, especially in drier environments, with conditions at depth potentially remaining permanently dry so that they could severely restrict decomposition activity (Figs. 3-4). This could potentially play an important role, provided soils in question form no major cracks during dry periods as cracks could allow by-passing of upper soils and rewetting on lower parts of the profile.

In contrast to soils in dry regions, in wet regions, it may be water-logged conditions, either permanently or for parts of the year, that may restrict access to oxygen and lower decomposition rates (Fig. 5). However, it seems likely that access to oxygen would have to be restricted quite severely before it could have a significant effect on decomposition rates, essentially requiring a layer of stagnant water to reduce oxygen diffusion rates to such an extent for oxygen to become limiting.

Finally, we investigated the possible role of biological synergy or ‘priming’, which we consider to be the enhancement of the decomposition of recalcitrant organic carbon by increased overall soil biological activity through the availability of fresh labile organic carbon (Fig. 6). The potential effectiveness has long been demonstrated through a range of experiments under manipulative experiments (e.g. Kuzyakov et al., 2000; Don et al., 2013), while its role under actual field conditions is much harder to demonstrate. However, a simple extrapolation from these observations under artificial to field condition would make it likely that similar processes play an important stimulatory role under field conditions as well.

Our simulations suggested that this process could well explain the persistence of organic carbon for long periods of time if the synergistic supply of fresh organic carbon can be cut off or is reduced (Fig. 7). In particular, it would not primarily slow down the initial rates of decomposition but become progressively more important as less and less organic carbon remains, especially more labile organic carbon.

In terms of overall carbon storage in the soil, this could lead to a substantial increase in total soil carbon storage. In the simulations shown here and with 100% biological synergy, the extra carbon storage over the first few decades after soil inversion corresponded to an average increase of about 1% per annum, which then gradually diminished as the soil moved towards a new steady-state condition, but reassuringly, the gain would not be reversed.

If biological synergy can be substantiated as a general mechanism to enhance soil carbon storage at depth, it could constitute a very valuable mechanism for extra carbon storage irrespective of the peculiarities or conditions of specific soils. It could then support the wider use of full-inversion tillage as a useful climate change mitigation option on grasslands.

Acknowledgments

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