



Salinity and Sodicity Affect Organic Carbon Dynamics in Soil

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SOIL CARBON STOCKS IN SALINE AND SODIC SOILS

The classification of saline and sodic soils according to the U.S. Department of Agriculture, widely followed across the globe, is as follows: saline (nonsodic) soils are those which have exchangeable sodium percentage (ESP) < 15 and electrical conductivity in saturation paste (EC_e) > 4 dS m⁻¹; sodic (nonsaline) soils have ESP > 15 and EC_e < 4 dS m⁻¹; saline-sodic soils have ESP > 15 and EC_e > 4 dS m⁻¹; and nonsaline and nonsodic soils have ESP < 15 and EC_e < 4 dS m⁻¹ (for detailed discussion on salt-affected soils see Gupta and Abrol 1990; Rengasamy 2002; Qadir *et al.* 2007). Nonsaline soils having ESP > 6 and $EC_e \leq 4$ dS m⁻¹ and exhibiting undesirable physical characteristics associated with sodicity such as high clay dispersion, low infiltration and low hydraulic conductivity, are also classified as sodic soils in the Australian Soil Classification System (Isbell 1996).

Saline and sodic soils cover more than 800 million ha (Mha) worldwide (FAO 2008). In Australia, salinity affects about 17 Mha while sodicity affects approximately 340 Mha of land (NLWRA 2001); of this area, 3.7 Mha is used for agriculture and further 5.7 Mha is used for both agriculture and pasture production (Rengasamy 2010). In India, salinity affects 2.96 Mha, while sodicity affects another 3.77 Mha (Maji 2007; Singh *et al.* 2010). The area affected by sodicity in India is an underestimate, since soils having both ESP > 6 and < 15 and $EC_e \leq 4$ dS m⁻¹ are not accounted for, although in the presence of low EC_e , clay dispersion accompanied by low infiltration rate and low hydraulic conductivity limit the potential for crop productivity. Moreover, these soils in the semi-arid regions are also likely to have subsoil constraints (Dang *et al.* 2010), which further limit their production potential.

Soil organic matter (SOM) status is commonly used as an index of soil quality or soil health, and soil quality in turn is related to agricultural productivity; thus SOM is also used as an index of agricultural productivity (Dalal *et al.* 2003). The soil organic C (SOC) pool is not only important for the soil to perform its productivity and environmental functions, but is also of critical importance in its role in the global C cycle (Lal 2004a;

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Wong *et al.* 2010). However, it has been observed that sodic and saline soils have lost the soil C pool (Wang *et al.* 2010); and the magnitude of loss may range between 1 and 30 Mg C ha⁻¹, depending on the inherent pool and the severity of the salinity or sodicity-related soil degradation (Lal 2001).

In a long-term study on the dynamics of SOM and salinity and sodicity in a Vertisol under continuous cropping in Queensland, Dalal *et al.* (2003) reported a significant inverse relationship between SOC and ESP; the relationship was described by the following regression equation:

$$\text{ESP} = 3.57 - 1.44 \ln \text{ organic C (\%)}; R^2 = 0.64, P < 0.05.$$

Pathak and Rao (1998), Peinemann *et al.* (2005) and Wong *et al.* (2008b) also found decreasing SOC concentration with increasing sodicity (Fig. 1). As the SOC declines and ESP increases in the top soil layer, soil aggregation decreases (Dalal 1989), and bulk density increases (Wong *et al.* 2008b), resulting in decreased infiltration rate and reduced hydraulic conductivity and increased runoff (Rengasamy and Olsson 1991).

As mentioned above, both salinity and sodicity affect SOC stocks adversely. For example, Wong *et al.* (2010) reported that a scalded soil affected by salinity and sodicity had SOC stock of 19.8 Mg ha⁻¹ while the adjacent pasture site contained 42.1 Mg ha⁻¹ (relatively undisturbed pasture site, 35.2 – 53.5 Mg ha⁻¹); that is, the salt-affected site had <50% of the SOC stock. They also measured the SOC stocks in scalded-eroded site, which contained about 9.6 Mg/ha (7.7-11.4 Mg ha⁻¹), a decline in SOC stock of almost 80% compared to the adjacent pasture site (Fig. 1). Thus, the magnitude of SOC loss due to salinity and sodicity may vary across soil types, climate, vegetation and land use and soil management (Lal 2001).

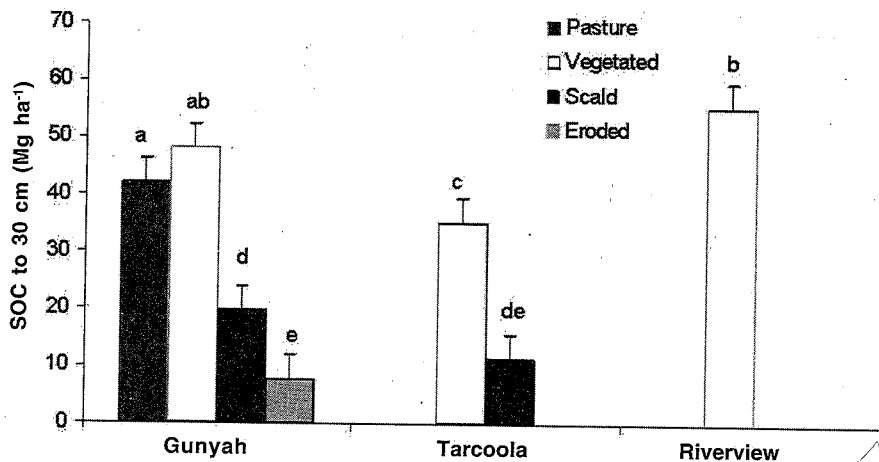


Fig. 1. SOC stocks in the top 30 cm depth from salt-affected sites (scald and scald and eroded) compared to adjacent pasture and vegetated sites, NSW, Australia. Scald sites on Gunyah and Tarcoola soil types were saline, while the latter site was also strongly sodic (ESP, 50-75 in the top 30 cm depth). Mean values (bar heights) having letters not in common differ significantly at $P < 0.05$. (Adapted from Wang *et al.* 2008b)

Soils in the arid and semi-arid tropical areas generally contain the largest pools of soil inorganic C (SIC) as carbonates, which could either be of lithogenic or pedogenic origin (for review see Lal 2002) and carbonate-C generally increases with depth. As a result, SIC stocks measured in the top 1 m depth of soil provide an underestimate of SIC stocks in these soils (Diaz-Hernandez 2010). Pedogenic SIC is formed through the dissolution of lithogenic or primary carbonates and from the precipitation of weathering products. The reaction of CO_2 with H_2O and Ca^{2+} and Mg^{2+} in the upper soil horizon, followed by the leaching of soluble products (bicarbonates) into the sub-soil and their subsequent re-precipitation results in the formation of pedogenic carbonates and in the sequestration of CO_2 (Sahrawat 2003). Thus, the leaching of bicarbonates through the soil profile, especially by irrigation management, could be a significant pathway leading to increased dissolved inorganic C (DIC) and thus temporarily increased SIC stocks.

Equally importantly, inorganic C can be converted to organic C by plants through photosynthesis, and in soils through the reaction of carbonate with added or soil decomposing organic matter. The transfer of C from inorganic to organic form provides a more conducive environment for C sequestration, soil conservation and environmental quality (Sahrawat *et al.* 2005; Bhattacharyya *et al.* 2009); the rate of C sequestration by this pathway (via SIC) ranges between 0.25 and 1.0 Mg C ha⁻¹ year⁻¹ (Wilding 1999). However, when phytoremediation is used to ameliorate sodic soils and bicarbonate is leached as a by-product of the overall reaction, the amelioration process could sequester SIC (Lal 2001; Sahrawat 2003). Thus, phytoremediation via application of plant materials could lead to the sequestration of both organic and inorganic C simultaneously (see Qadir *et al.* 2007 for review). However, there is a need to distinguish between the transfer of C between the two systems (without net C sequestration) and actual C sink for a better understanding and appreciation of the contribution by SIC and SOC to C sequestration in soils (Monger and Martinez-Rios 2001).

Bhattacharyya *et al.* (2009) reviewed the Indian literature on C stocks in Indian soils, covering 7 soil orders. It was found that Vertisols, Inceptisols and Alfisols have the major share of SOC stocks in the top 30 cm depth of soils. At similar soil depth, Aridisols contained low SOC, but high SIC. When Indian soils were broadly classified into 5 groups, then red soils (Alfisols and Ultisols), alluvial soils (Entisols and Inceptisols) and black soils (Vertisols) contained the largest stock of organic C, followed by arid soils (Aridisols) and brown forest soils (mostly Mollisols) in decreasing order. The SIC stock in the soil orders comprised 1 to 39% of the total C stocks in the soils (Bhattacharyya *et al.* 2009).

In a recent review on SOC dynamics in saline and sodic soils, Wong *et al.* (2010) reported that the effects of salinity and sodicity decreased the SOC status and stocks in the salt-affected soils. Also, the effects of salinity and sodicity on plant health and production adversely affected the SOC stocks in salt-affected areas. However, the presence of vegetation, which provides organic matter inputs to the soil tend to moderate the effects of salinity and sodicity on SOC stocks. Thus, it is not entirely surprising to note that the contents and stocks of SOC in salt-affected soil profiles without vegetation (bare) were consistently lower than those found in the salt-affected areas with vegetation (Wong

et al. 2008b). Evidently, the presence of vegetation and via its inputs of organic C ameliorates soil physical, chemical and biological properties, and is conducive to SOC accumulation in the profile compared to the site without vegetation (Rao and Pathak 1996; Tripathi and Singh 2005; Qadir *et al.* 2007; Wong *et al.* 2008b; Gill *et al.* 2009; Sharma *et al.* 2009; Ghosh *et al.* 2010).

Also, addition of organic material to sodic and saline soils increased the soil microbial biomass levels and respiration rates despite adverse soil environmental conditions. The results suggested that a dormant population of salt-tolerant soil microbial biomass is present in these soils, which has become adapted to such environmental conditions over time and multiplies rapidly when substrate is available (Wong *et al.* 2009).

It appears that salt-affected soils are generally low in organic matter; however, there is little detailed information on C stocks in saline and sodic soils. A number of studies based on the semi-arid areas may have included salt-affected soils but the results on the salt-affected soils cannot be separated from other soils in these studies (Knowles and Singh 2003; Bhattacharyya *et al.* 2009; Srinivasarao *et al.* 2009). It is evident that the salt-affected soils are relatively low in SOC compared to their non-saline counterparts in similar environments. However, such soils can be biologically revitalized and remediated by application of organic material over a longer-term time frame (Qadir *et al.* 2007; Wong *et al.* 2009).

It has been suggested that high salinity and alkalinity positively influence the CO₂ absorption capacity of soils, while high temperature and water content have a negative effect on the CO₂ absorption capacity of salt-affected soils (Xie *et al.* 2009). According to Xie *et al.* (2009), the process or mechanism involved in the absorption of CO₂ by alkali soils is inorganic and non-biological in nature. This research merits further investigation for overall global C budgeting and cycling. Precaution is required, however, in the interpretation of actual CO₂ sequestration into the SIC pool to differentiate between inorganic C transfer, sink, source and transport across the landscape (Monger and Martinez-Rios 2001) to avoid double-accounting.

With growing interest in C sequestration, the degraded soils including a large area under salt-affected soils in the arid and semi-arid regions globally are expected to play a central role in stabilizing the atmospheric concentration of CO₂ by employing sustainable agricultural practices (Izaurralde *et al.* 2001; Chmura *et al.* 2003; Lal 2002, 2004b; Srinivasarao *et al.* 2009).

CARBON TURNOVER PROCESSES IN SOIL AFFECTED BY SALINITY AND SODICITY

The rate of C sequestration and C stocks in soils depend on several soil and environmental factors. Among the soil factors, texture and mineralogy are more important. Among the environmental factors, moisture regime and temperature control the decomposition of organic matter added and the residence time of C in the soil. Equally importantly, the input and quality of organic matter added have an overwhelming effect on SOC turnover rates, and C storage in the soil profile (Torn *et al.* 1997; Six *et al.* 2002; Sariyildiz and Anderson 2003; Kiem and Koegel-Knabner 2003; Sahrawat 2004; Bhattacharyya *et al.* 2009).

Soil organic matter is dynamic and as a result, SOC concentrations and stocks increase or decline as a result of several soil and environmental factors including temperature, water regime, nutrient availability, management practices, and land use, which affect organic C inputs through plant biomass and organic inputs and organic C outputs through decomposition as well as C transport via leaching and erosion. Moreover, SOC is in a continual state of flux with simultaneous inputs and outputs of C through various organic C pools of various qualities (Paul 1984).

Since the SOC content depends upon the relative rates at which organic materials are added to the soil and lost from it through decomposition (and transport), it can mathematically be expressed as follows (Bartholomew and Kirkham 1960):

$$d \text{ SOC}/dt = A - k \text{ SOC} \quad \dots(1)$$

$$\text{or } \text{SOC}_t = \text{SOC}_0 \exp(-kt) + A/k [1 - \exp(-kt)] \quad \dots(2)$$

where SOC_0 and SOC_t are the SOC contents initially ($t=0$), and at a given time, t , respectively; A (mass of SOC per unit area) is the rate at which organic C is returned to the soil, and k (reciprocal of time) is the rate of loss of SOC. Since SOM is heterogeneous and its various components decompose at different rates, Eq. 2 is modified to:

$$\text{SOC}_t = \text{SOC}_1 \exp(-k_1 t) + A_1/k_1 [1 - \exp(-k_1 t)] + \dots + \text{SOC}_n \exp(-k_n t) + A_n/k_n [1 - \exp(-k_n t)] \quad \dots(3)$$

where SOC_1 , SOC_n refer to different components of organic matter, such as microbial biomass and light fraction (active C pool), heavy or clay-size C fraction (passive, inert SOC), and the remaining slow C pool (Jenkinson and Rayner 1977; Parton *et al.* 1987; Jenkinson 1990; Parton *et al.* 1996; Dalal and Chan 2001; Six 2002), and k_1 and k_n refer to the rate of loss of individual SOC components. SOC dynamic models such as Roth-C and Century consider these components as well as the factors affecting these components such as climate, organic C addition, soil matrix, cultural practices and some soil degradation processes such as soil erosion and leaching (Parton *et al.* 1996), but not salinity and sodicity.

When $t=0$, Eq. 3 is reduced to $\text{SOC}_t = \text{SOC}_1 + \dots + \text{SOC}_n = \text{SOC}_0$ and at time

$t = \alpha$, $\text{SOC}_t = A_1/k_1 + \dots + A_n/k_n = \text{SOC}_e$, which may be simplified to give:

$$\text{SOC}_e = A/k \quad (4)$$

where SOC_e is organic matter content at steady state after a long period of consistently similar soil management (Bartholomew and Kirkham 1960) or under native vegetation. The simplified form of Equations 2 and 3 is:

$$\text{SOC}_t = \text{SOC}_e + (\text{SOC}_0 - \text{SOC}_e) \exp(-kt) \quad (5)$$

where k is the overall rate of loss (decomposition) of various organic matter components.

The turnover period (or mean residence time) of SOC is then reciprocal of the k value.

As mentioned above, the turnover of C in SOM is determined using simple first-order kinetics; for specific purposes, C-labeled plant materials are used to determine the decomposition rates and turnover rates of C in various SOC pools (for details of methods see Jenkinson 1990; Paul *et al.* 2001).

Among the various factors which affect C turnover rates in soils, the most important include (i) primary production, which contributes to inputs of organic matter and (ii) soil microbial activity, which controls the overall rates of SOC decay and transformation that regulate SOC turnover. However, this is quite a simplistic concept of a complex process, which is controlled by several biotic and abiotic interactions and feedbacks. The most important factors remain temperature and rainfall as components of climate because they control both biomass production and the decomposition of SOM. Other factors that influence SOC turnover include production rate and quantity (added via below ground or above ground) and quality (including C/N ratio, lignin/N ratio) of organic inputs, soil type (especially pH, texture, structural stability and mineralogy), tillage and other management practices and the availability of plant nutrients influence decomposition rates. On the microbial side, the types, populations and activities of soil biota control both SOM decomposition and nutrient cycling, and eventually the release of nutrients via decomposition regulates plant productivity. Equally importantly, plant production and decomposition are influenced by temporal influences that integrate various other factors in these two processes.

Under a defined set of biotic and abiotic conditions, the turnover of different SOC pools depends on the quality and biochemical recalcitrance of the SOM and its accessibility to decomposers. Also, over time a relatively labile organic material may be physically protected by clay mineralogy or its incorporation into soil aggregates (Feller and Beare 1997; Baldock and Skjemstad 2000; Six *et al.* 2002; Kiem and Koegel-Knabner 2003; Sariyildiz and Anderson 2003; Sahrawat 2004).

Although, there is no systematic research on the SOC turnover rates in salt-affected soils or on comparative turnover rates of SOC in normal and salt-affected soils, it is known that high concentrations of salts or Na^+ adversely affect soil physical and biological properties, which in turn retard microbial activity and hence the decomposition of organic materials (Rao and Ghai 1985; Rao and Burns 1991; Batra *et al.* 1997; Qadir *et al.* 2007). Retarded decomposition of organic materials slows down the release of plant nutrients (Torn *et al.* 1997) and this adversely affects organic C production and turnover rates.

It has been universally found that SOC concentration decreases as the soil ESP increases (Fig. 2); the rate of decrease, however, depends on the soil type, soil mineralogy, the extent of salinity and other soil degradation processes such as soil fertility decline, soil structural decline and soil erosion (Lal 2001; Rengasamy 2010).

It has also been observed that in salt-affected soils, including sodic soils, as the soil pH increases SOC concentration decreases (Fig. 3) as a function of both reduced organic C inputs as well as SOC decomposition.

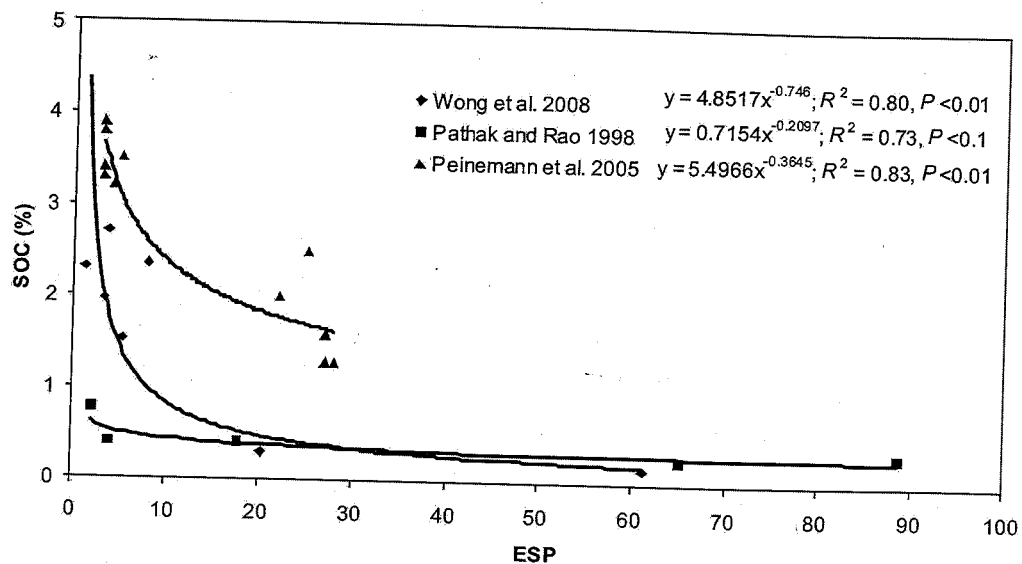


Fig. 2. SOC concentration declines with increasing soil sodicity. (Figure drawn from the data reported by Wong *et al.* (2008b) for 0-5 cm depth, Pathak and Rao (1998) for 0-15 cm depth, and Peinemann *et al.* (2005) for different depth intervals (up to 30 cm))

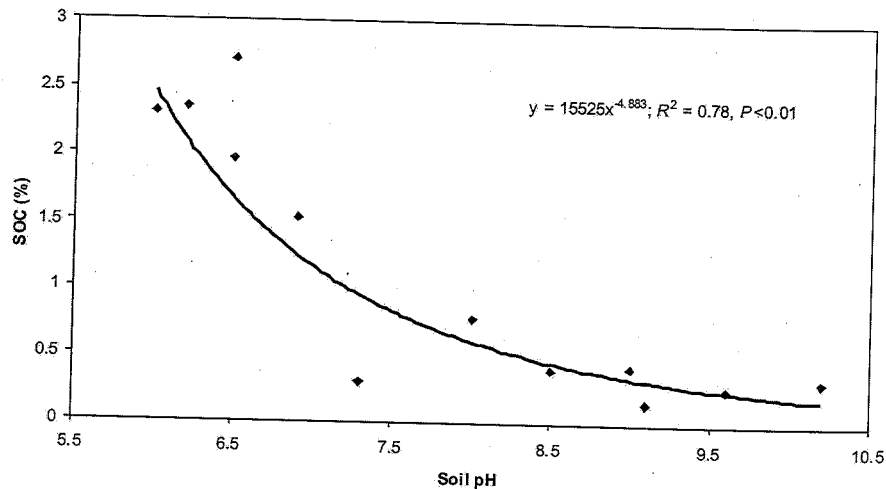


Fig. 3. SOC concentration as a function of soil pH. (Figure drawn from the data reported by Pathak and Rao, 1998, and Wong *et al.* 2008b)

SOIL ORGANIC CARBON DYNAMICS IN SALINE AND SODIC SOILS: CARBON FLUXES

Total Organic Carbon

Soil organic C stocks are a function of C inputs, which are dominated by litterfall, root exudates and fine root decomposition, and are, therefore, dependent on biomass

production, and outputs, which are dominated by microbial decomposition processes, leaching and erosion. As a result, decreases in biomass production due to soil degradation will directly influence the SOC levels. Those areas susceptible to salinity and sodicity are also the most susceptible to significant SOC losses as a proportion of total SOC. Salt-affected landscapes are usually found in areas of marginal agriculture or rendered marginal due to salinity and sodicity constraints, in association with soils of lower fertility and hence, lower biomass production, with a high susceptibility to erosion, which further accentuates the losses of C from the soil.

Following the onset of salinity and sodicity, native SOC can be rapidly lost since C inputs into the soil are decreased as salinity and sodicity cause plant health to decline through adverse soil physical and chemical conditions. Increased sodicity provides additional substrate from dispersed aggregates, which is easily decomposed by the microbial population (Jandl and Sollins 1997), and can also be easily lost by leaching. Thus, dispersion of aggregates due to sodicity, mainly with cores containing organic material (Tisdall and Oades 1982), increases the availability of organic C. As a result, SOC accessibility and degradability is increased for the microbial population under such conditions. In extreme cases of salinity and sodicity, the scalding of the soil surface occurs, resulting in very low SOC stocks. Scalded soils are highly susceptible to further losses caused by erosion, resulting in low levels of MBC in both scalded and eroded profiles, probably due to low levels of SOC.

Conversely, the potential to accumulate significant amounts of SOC in salt-affected landscapes is high, as SOC stocks are initially low in salt-scalded areas. It is suggested that SOC can also be rapidly accumulated during reclamation (Wong *et al.* 2008b), by the addition of organic material, replanting, or a combination of both, which can increase the standing biomass and hence, increase SOC stocks.

Organic Carbon Pools

The SOC can be partitioned into discrete pools according to its age or the time it takes to turnover (Jenkinson and Raynor 1977). Mean residence times are dependent on resistance to decay and the extent of protection against decomposition. The three main SOC pools are: i) the active pool, with a turnover time in the order of weeks (*i.e.* the MBC and particulate organic carbon, POC); ii) the slow pool with a turnover time in the order of decades (*i.e.* humus); and iii) the passive pool with a turnover time in the order of millennia (*i.e.* charcoal). The active pool is made up of readily oxidisable materials including the microbial biomass and its metabolites, and is largely controlled by climate and residue inputs (Schnurer *et al.* 1985). The slow and/or very slow pools contain macro- and microaggregates with chemically recalcitrant but moderately decomposable material, while the passive or recalcitrant pool includes recalcitrant and stable C formed from the turnover of microbial and slow SOC; this pool has organic compounds that are chemically resistant to, or protected from further microbial degradation and includes charcoal (Schimel *et al.* 1994; Lehmann *et al.* 2008). Most C found in detritus and microbes is oxidised and cycled rapidly. Some is transformed into a slow reservoir with a turnover time of the order of decades to centuries, most of which will eventually oxidise. The remainder is converted to the passive pool with turnover rates on a millennial timescale (Stallard 1998).

The SOC values were modeled using Roth-C for a 60-year old salt-affected site following scald formation for the total soil pool (soil), resistant plant material pool (RPM), humified organic matter pool (HUM) and inert organic matter pool (IOM) (Fig. 4). The limitation of this model is that it assumes all C losses are as a result of decomposition, and it does not take into account any C losses that occur as a result of erosion (Wong *et al.* 2008b).

Since sodicity causes disruption of soil aggregates and silt and clay dispersion, it exposes microbially-protected SOC in silt- and clay-size aggregates to microbial decomposition. Paired-site studies of relatively non-salt affected soils and adjacent sodic and saline-sodic soils showed that loss of organic C occurred primarily from the silt-size and clay-size fractions (Fig. 4). For example, although SOC concentration in a sodic soil was only 34% lower than that in the adjacent nonsaline-nonsodic soil, organic C concentrations in the silt-size and clay-size fractions were 70-72% lower in the sodic soil, while in the sand-size fractions it was unaffected. In fact, in a saline-sodic soil, particulate organic C concentration ($>50 \mu\text{m}$) was higher than in the adjacent nonsaline-nonsodic soil. This may be due to the fact that cellulolytic (cellulose hydrolyzing enzymes) activity is inhibited in soil of high pH and salinity (saline-sodic soil) (Horikoshi 1999), resulting in the

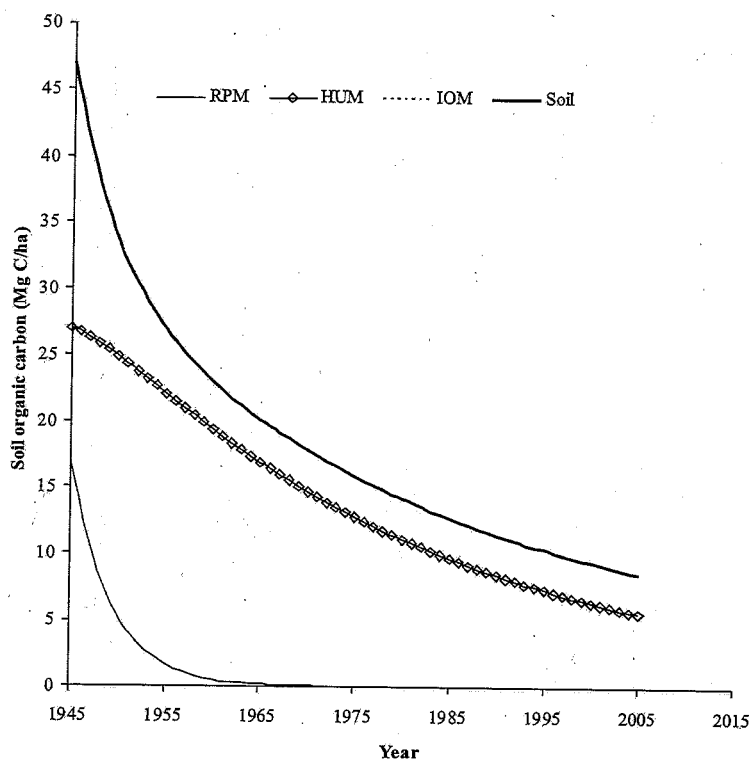


Fig. 4. Modeled SOC values for a salt-affected site following scald formation in 1945 in the total soil pool (soil), resistant plant material pool (RPM), humified organic matter pool (HUM) and inert organic matter pool (IOM), assuming no C loss due to erosion. (Redrawn from Wong *et al.* 2008b)

accumulation of particulate organic C (Peinemann *et al.* 2005), as was also found by Wichern *et al.* (2006) and Setia *et al.* (2010). Still organic C concentration of the silt-size and clay-size fraction was 75-78% lower in the saline-sodic soil than in the nonsaline-nonsodic soil (Fig. 5) (Peinemann *et al.* 2005). High pH and high concentrations of Na disperse aggregates, make organic C available to microorganisms and decrease the formation of organo-clay complexes and aggregates. Moreover, dissolved SOC is also subjected to the leaching and erosion and runoff losses.

Soil Microbial Biomass, Activity and Diversity

The active C pool is comprised of a living component, the MBC, and a non-living component, the dead biomass and its metabolic products, which comprise approximately 1-5 % of the total SOC (Sparling 1992). The active soil C pool is frequently used as an

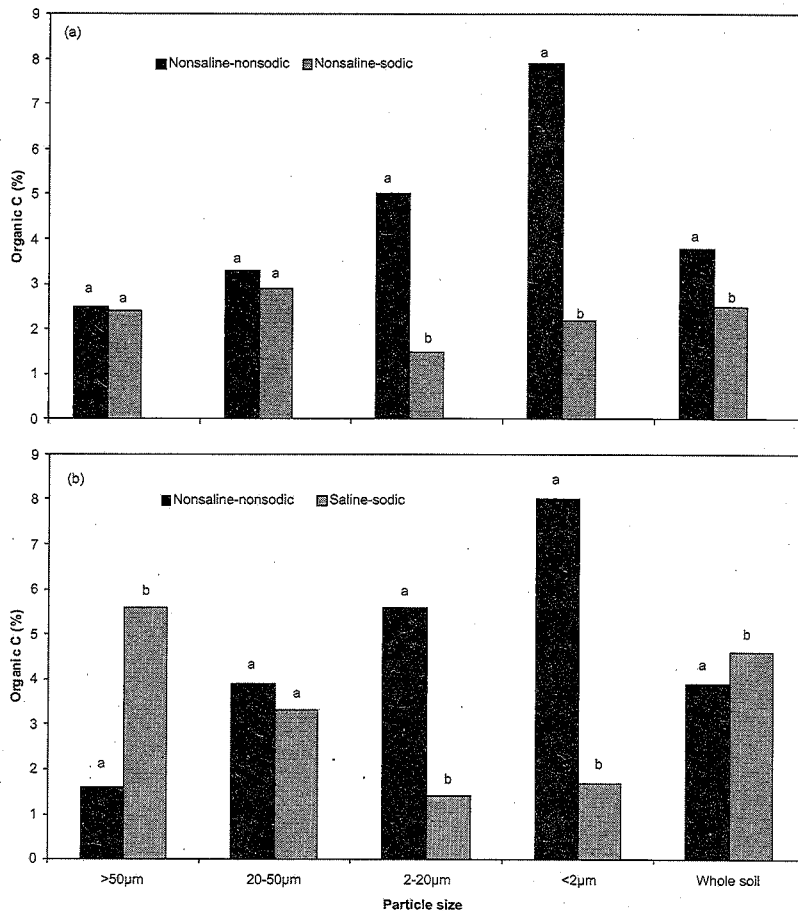


Fig. 5. Particle size distribution of organic C in (a) nonsaline-nonsodic and adjacent nonsaline-sodic soil, and (b) nonsaline-nonsodic and adjacent saline-sodic soil. For each figure, organic C concentrations within a particle size followed by letters not in common differ significantly at $P < 0.05$. (Figures drawn from the data reported by Peinemann *et al.* 2005)

early indicator of SOM dynamics, due to its faster turnover rate (Alvarez *et al.* 1998), as changes caused by management or environmental stresses can be detected earlier in this pool than in the SOM pool as a whole. Therefore, the effects on microbial activity caused by increasing salinity and sodicity should precede effects on the total C stock, and should be detected prior to the more obvious effects of declining vegetation health, changes in biomass C inputs and the total SOC pool.

A number of opposing processes affect SOC fluxes during salinisation and sodication. Dalal and Mayer (1986) have linked the loss of SOM to factors that affected its accessibility and stability to attack by the microbial population and enzymes. Macro- and microaggregates can contain, and physically protect a considerable portion of SOC (Conant *et al.* 2004). In sodic soils, the dispersion of aggregates on wetting can increase substrate accessibility and availability (Oades 1984). However, on drying, the bulk density of a soil increases and water holding capacity decreases which decreases the availability of SOM to the microbial biomass. Thus, the hardsetting soils of high bulk density restrict substrate availability to the microbial population due to the breakdown of soil structure on wetting and its subsequent formation of massive structure when dry, as substrate can be located in pores that are too small for the microbial population to access.

Sodicity has been shown to increase the rate of mineralization in previous studies. Chander *et al.* (1994) found that the rate of mineralisation of organic matter increased as sodicity increased, while the MBC decreased (Fig. 6). The smaller microbial population was partly due to decreased plant inputs from stresses placed on plants with increasing

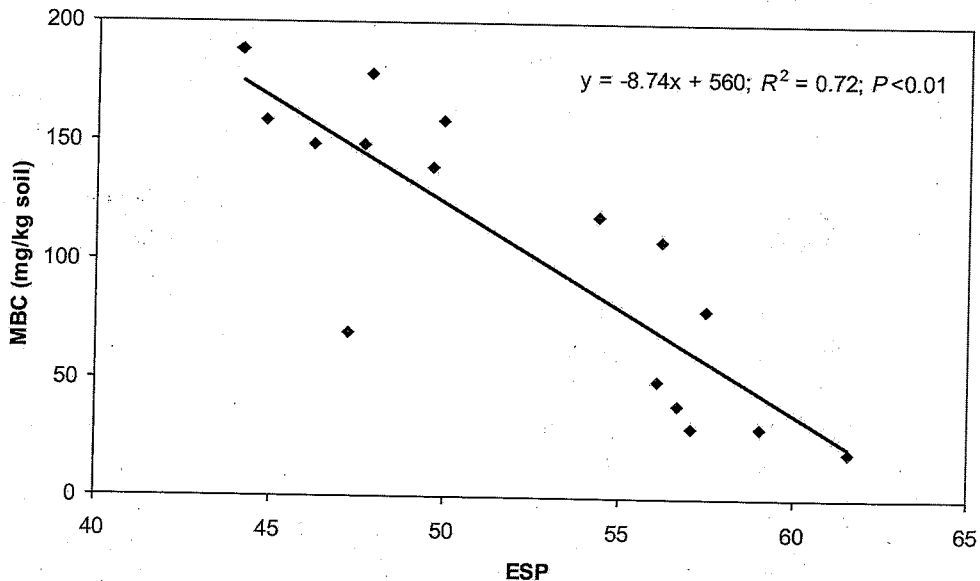


Fig. 6. Soil microbial biomass C (MBC) in sodic soil decreases with increasing sodicity (redrawn from Kaur *et al.* (2008) for organic-amended treatments after 14-year wheat-rice cropping for the reclamation of a sodic sandy loam soil). Essentially similar MBC and ESP relationship ($\text{MBC (mg/kg)} = -9.0 \text{ ESP} + 586$, $R^2 = 0.42$, $P < 0.01$) between 100 and 780 mg C/kg MBC and 1 and 52 ESP values was found by Rietz and Haynes (2003).

SOM. In saline soils, MBC is depressed due to osmotic stress, which means reclamation of these soils involves leaching of salts followed by the addition of organic amendments (Rao and Pathak 1996; Wong *et al.* 2008b; Wong *et al.* 2010).

Microbial community structure changes with increasing salinity and sodicity in soil, generally becoming bacterial-dominated with reduced fungal:bacterial ratio (Wichern *et al.* 2006), reduced MBC and yet maintaining high metabolic quotient (Yuan *et al.* 2007; Wong *et al.* 2008a). This appears to occur especially with increasing sodicity due to the accessibility to bacteria and exoenzymes of organic substrates in silt and clay-dispersed micropores in saline-sodic and sodic soils (Peinemann *et al.* 2005). Hollister *et al.* (2010) examined the structure and diversity of microbial communities in hypersaline and sodic environments using community quantitative PCR, 16s rRNA gene cloning, and Sanger sequencing. Bacteria were numerically dominant relative to Archea (Dendooven *et al.* 2010; Hollister *et al.* 2010), although increasing salinity reduces the total microbial population (Bastida *et al.* 2006).

Litter and Root Decomposition

It is generally accepted that increasing levels of organic matter content in soil improve soil structure. Whilst only present when plants are growing, plant roots, mycorrhizal hyphae and fungal hyphae at larger scales enmesh macroaggregates, inhibiting slaking and dispersion. At smaller scales, mucilages and colloidal organo-minerals are the primary binding agents in microaggregates (Nelson and Oades 1998). High bulk density is a common symptom of sodic soils, which restricts the access by microorganisms for decomposition of roots and root exudates. The presence of roots promotes aggregate stability through the *in situ* production of polysaccharides and fungal hyphae (Tisdall and Oades 1982).

Whilst the addition of gypsum ameliorates soil chemical properties, tillage or deep ripping of clay layers is required to improve soil physical and hydraulic properties, and soil aeration, and organic amendments for aggregate stabilisation. The formation of soil aggregates can physically protect soil C, such that any process which disrupts these aggregates will most likely increase C mineralisation. Under conventional tillage treatments, aggregates are frequently disrupted, resulting in fewer stable macroaggregates and the mineralisation of previously protected organic matter (Paustian *et al.* 2000). This is due to the more labile nature of the organic matter associated with macroaggregates, which is, therefore, more readily mineralisable compared to that associated with microaggregates (Waters and Oades 1991), unless aggregates are dispersed by sodicity.

The addition of organic amendments such as farmyard manure, composts, organic byproducts and plant residue enhances microbial biomass, microbial activity and microbial biomass diversity in saline soils (Tejada *et al.* 2006; Khan *et al.* 2008; Wong *et al.* 2009; Hüllugalle *et al.* 2010; Setia *et al.* 2010), and accompanied by gypsum application in sodic soils (Kaur *et al.* 2008). These effects may largely be due to increased soil aggregation, formation and stabilization of Ca-organo-clay complexes, and displacement of Na from the exchange sites, and to an extent on increased partial pressure of CO₂ from microbial decomposition of organic substrates. Soil with non-SEMCTITIC clay mineralogy enhances

added organic residue decomposition while smectitic-clay soil reduces organic residue decomposition (Nelson *et al.* 1997). Thus, SOC retention is improved by reducing ESP before the addition of organic amendments, irrespective of clay mineralogy (Nelson *et al.* 1997).

The presence of plant roots has been shown to increase the partial pressure of CO_2 (P_{CO_2}) in aerobic (Mishra and Sharma 2003) and waterlogged soils (Boivin *et al.* 2002), and to produce root exudates for microbial activity and soil aggregation. Because salt-affected soils commonly occur in alkaline conditions, the increase in P_{CO_2} has been shown to decrease soil pH. In Australia, many sodic soils are generally alkaline (exceptions being highly saline-sodic but acidic subsoils in certain regions) and contain CaCO_3 in the profile, usually in the subsoil, but this remains relatively insoluble due to high pH conditions. Decreasing soil pH by increasing P_{CO_2} and organic matter in soils prevents CaCO_3 precipitation and enhance its solubility, facilitating the reclamation of sodic areas (Chorom and Rengasamy 1997).

PRACTICES TO ENHANCE CARBON INPUTS TO SALINE AND SODIC SOILS

The need to produce more food for the growing population especially in the developing countries makes it imperative to increase the use of salt-affected soils to meet the increasing needs for food, feed, fiber, and bioenergy production. To make most efficient use of saline and sodic soil resources, it is important to ameliorate and make them more productive. Minhas and Sharma (2003) have provided detailed discussion on the management of saline and sodic soils in India; only brief discussion is given below.

Historically, sodic soils have been ameliorated by using an amendment (mostly gypsum or other chemicals) that is a readily available source of Ca^{2+} to replace excess Na^+ on the cation exchange complex. The displaced Na^+ is leached out of the root zone through the application of excess water. In the case of saline soils, the salts are washed out of the root zone with excess water by providing unimpeded flow of water through the soil profile; there is no need to add an amendment containing Ca^{2+} (for review see Gupta and Abrol 1990; Qadir *et al.* 2007). Gill *et al.* (2008) suggested deep application of organic amendments, which were found to be more efficacious than application of gypsum, especially for the amelioration of dense sodic subsoil, possibly by increasing the infiltration rate, hydraulic conductivity, and aggregation. However, farm-scale application of organic amendments to the subsoil may not be economical. There is also the possibility of inversion of highly sodic subsoil and bringing it closer to the soil surface. In addition to use of chemical amendments such as gypsum, the amelioration of sodic and saline soils by using growing plants, vegetation or external inputs of organic matter termed phytoremediation, has also been in vogue for a long time for amendment of these soils (Qadir *et al.* 2007).

The amelioration of salt-affected soils by phytoremediation is based on the principle that phytoremediation of calcareous sodic and saline soils in the arid regions assists in increasing the dissolution rate of natural calcite (and or dolomite) through the production of CO_2 and its reaction with CaCO_3 results in enhanced dissolution of calcite. The partial

pressure of CO_2 or P_{CO_2} is much higher in the anaerobic submerged than aerobic upland soils (Ponnamperuma 1972), and hence the amelioration of sodic and saline soils is much faster under submerged conditions— an enhanced amount of CO_2 is generated under these conditions and because of the flooded condition, CO_2 escape is hampered and it has longer residence time, and the flooded condition facilitates drainage (Qadir *et al.* 2007). In non-calcareous soils, increase in CO_2 concentration results in the production of H^+ and a corresponding decrease in soil pH, although the decrease in pH in calcareous soils may be smaller as the soil pH is buffered by the increased dissolution of CaCO_3 .

However, the efficacy of application of amendments especially plant materials in enhancing SOC status and amelioration of these soils depends on the plant species and their ability to grow and produce biomass (Tripathi and Singh 2005; Qadir *et al.* 2002). In general, it has been found that the amelioration of sodic and saline soils through the use of plants in the form of vegetation and crop residues is a slow process and this process can be enhanced by the application of amendments such as gypsum to reclaim the sodic soils. However, phytoremediation is advantageous in that in addition to supplying organic matter, it provides source of plant nutrients, which are released during their mineralization in the soil. Moreover, plant roots produce root exudates and mucilages, resulting in increased microbial activity and microbial products in and around rhizosphere for aggregate formation and stabilization. Growing roots also provide channels for enhanced infiltration and hydraulic conductivity for rapid leaching of excess salts.

In a long-term experiment on a sodic soil, Garg (1998) monitored the changes in organic C under four tree species and reported that *shisham* (*Dalbergia sissoo*) and mesquite (*Prosopis juliflora*) were more efficient than acacia (*Acacia nilotica*) and *arjuna* (*Terminalia arjuna*) species in terms of decreasing Na^+ levels and increasing SOC accumulation. The rate of increase of SOC was low for the first 2 to 4 years, exponential between 4 to 6 years, and reached a plateau at a low rate for 6 to 8 years. In a related study, Bhojvaid and Timmer (1998) found that the establishment of mesquite on a sodic soil increased SOC of the top 1.2 m soil from 11.8 to 13.3 Mg C ha^{-1} in 5 years, 34.2 Mg C ha^{-1} in 7 years and 54.3 Mg C ha^{-1} in 30 years. The average annual rate of increase of SOC was 1.4 Mg C ha^{-1} over the 30-year period. Based on the estimates made from field studies on sodic soils, Kaur *et al.* (2002) suggested that various land-use systems consisting of a number of grasses and trees can sequester organic C in the range of 0.2 to 0.8 Mg C ha^{-1} per year.

Thus, phytoremediation of saline and sodic soils occurs through plant root activity to increase the dissolution of CaCO_3 , enhanced Ca^{2+} concentration in the soil solution and on cation exchange sites to replace Na^+ , primarily through the increased partial pressure of CO_2 within the root zone. In addition, leguminous species also release H^+ , which further enhances dissolution of CaCO_3 , besides providing N for crop growth. Moreover, organic substrates and resulting microbial products in rhizosphere assist in the formation and stabilization of aggregates and protection of SOC, thereby improving physical condition of soil for leaching of salts below the root zone. Organic amendments provide some of these benefits while main benefit of gypsum application is to provide Ca^{2+} to replace Na^+ from soil solution and cation exchange complex. A combination of these practices is

the most appropriate option for remediation of saline and sodic soils and increasing C stock by these soils since these degraded soils have a high potential for C sequestration.

PERSPECTIVES

Salinity and sodicity affect a large area of the world, in excess of 800 Mha; of which significant proportion of the land is utilized for agricultural production, including intensive cropping. Compared to nonsaline and nonsodic soils, SOC levels in these soils are larger, due primarily to reduced biomass production, hence reduced organic C input, but also to enhanced or similar organic C decomposition rates. The latter is accentuated or maintained by increased dispersion of aggregates due to sodicity and hence exposure of protected organic C within aggregates, and its dissolution and availability to the bacterial-dominated microbial population. Furthermore, as this source of SOC is depleted, even MBC is utilized, which further reduces microbial biomass. Also, the dissolved organic C is subjected to leaching and runoff losses.

Limited work from paired-site studies show SOC losses in excess of 60%; from a relatively salt-unaaffected site containing about 50 Mg C ha⁻¹, to <20 Mg C ha⁻¹ in the top 30 cm depth in the salt-affected site (Wong *et al.* 2008b). Conversely, limited data show that effective revegetation of sodicity-affected land can restore the SOC stock to close to its original level over a 40-50 year period. Even under intensive cropping, including the application of amendments and optimum nutrient management, SOC increases of almost 50% in the top 15 cm depth have been measured after a 25-year period (Benbi and Brar 2009). Remediation of even 10% area of salt-affected lands, achieving an estimated SOC sequestration of 0.2 Mg C ha⁻¹ year⁻¹ over a 50-year period (approximately 50% of the potential C sequestration rate), may lead to 0.8 Pg C sequestered in SOC in these soils. Therefore, the potential for salt-affected soils to sequester SOC is large and significant. It is expected that large proportion of C sequestration will occur or result in the formation and stabilization of soil aggregates such as SOM-Ca²⁺-clay aggregates, and as protected SOC against rapid microbial decomposition. However, research is required to validate this SOC sequestration mechanism after restoration of salt-affected soils, since, besides SOC benefits other benefits occur in improved physical and chemical characteristics of the soil.

It has been suggested that SIC stock may increase in saline-sodic and sodic (alkaline) soils since they provide an environment for CO₂ sink as HCO₃⁻ and CO₃²⁻ in dissolved form as well as precipitated as calcite as atmospheric CO₂ concentration increases. However, limited information is available to estimate SIC sequestration and inorganic C dynamics by these processes in these soils. Moreover, all three major phenomena of inorganic C transport, inorganic C transfer such as inorganic C-inorganic C and inorganic C-organic C, and net sink in soil have not yet been quantified. Also, these soils, especially alkaline sodic soils provide an unfavourable environment for plant production, and hence even with the remotely net SIC sequestration it is a perverse outcome for agricultural productivity and the environment.

Limited information available on nitrous oxide (N₂O) emission and methane (CH₄) fluxes from saline and sodic environments show similar or lower N₂O and CH₄ emissions,

especially the latter decreases as salinity increases (Allen *et al.* 2006), and Archae population decreases as sodicity increases (Dendooven *et al.* 2010). Organic amendments, N fertilization and flooding will restore or even increase N₂O and CH₄ emissions. Therefore, besides estimating the changes in SOC stocks, N₂O and CH₄ fluxes should also be included in the assessment of overall improvement in the greenhouse gas mitigation.

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