Soil Organic Matter and Soil Function – Review of the Literature and Underlying Data

Effects of soil organic matter on functional soil properties

Project in the Grains Research and Development Corporation and Department of the Environment

"Improved management of soil organic matter for sustainable Cropping "

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Summary

A review has been undertaken into how soil organic matter affects a range of soil properties that are important for the productive capacity of the soils. The potential effect of varying the amount of soil organic matter in soil on a range of individual soil properties was investigated using a literature search of published information largely from Australia, but also including relevant information from overseas. The soil properties considered included aggregate stability, bulk density, water holding capacity, soil erodibility, soil thermal properties, soil colour, soil strength, compaction characteristics, friability, nutrient cycling, cation exchange capacity, soil acidity and buffering capacity, capacity to form ligands and complexes, salinity and the interaction of soil organic matter with soil biology.

Overall this review concentrated on the soil physical properties and results from some of the field studies on soil organic matter and soil properties and so is intended to compliment the earlier review by Krull et al (2004) and another on the role of soil carbon in nutrient cycling (Macdonald and Baldock 2010).

Some conclusions on the effects of soil organic matter on several soil properties were made. Soil organic matter had clear effects on water holding capacity, cation exchange capacity, aggregate stability and buffering capacity to acidification. Soil organic matter also had a definite effect on the compaction and strength characteristics of soils which in combination with friability can determine how the soil responds to traffic and tillage. Soil organic matter was an important factor in providing a nutrient supply and in nutrient cycling, especially of nitrogen, but also of significant proportions of phosphorus and sulphur and other micronutrients. The relative importance of soil organic matter to provide certain functions in soils varied with texture, with soil organic matter generally being more critical in soils with lower clay contents.

Different soil organic fractions had different properties and varied in their importance for different soil properties. While data was generally lacking, it appears that the humus fraction is more important for the functions requiring chemical activity such as cation exchange capacity and pH buffering capacity and the particulate organic carbon is more important for aggregate stability of the larger aggregates, the rapid recycling of nutrients and as a food and energy source for the soil microbial population.

The level of soil organic matter and the levels of certain factions can influence the biological activity and microorganisms in the soil. The development of new methodologies such as the measurement of microbial biomass, substrate analysis, phospholipid analysis, DNA analysis and enzyme evaluation has given improved insight into many of the biological processes in soils. This has shown that land management and even different plants or crops can change the biological populations. However, a more complete understanding of the overall effect some of these changes can have on overall productivity and nutrient cycling requires a higher level of understanding then is currently available.

There were some values for the levels of soil organic carbon required to maintain the soil properties in a functional form and the value of 2% soil organic carbon occurred a number of times as a useful guide. One important conclusion is that by increasing soil organic matter it is possible to improve several soil properties simultaneously and so have a cumulative effect on productivity.

A final part of the review was to consider the capacity of the dry land agricultural systems to achieve levels of soil organic matter that could improve soil properties. This was done using soil carbon data

from across Australia from a range of sources including the National Soil Carbon Project. In general there was a percentage of soils that had soil organic carbon levels sufficiently low as to affect soil properties. These also seemed to occur in areas where available land management practices could increase the levels of soil organic carbon above the measured levels and so there is a potential to improve productivity in these areas by increasing levels of soil organic carbon. A large proportion of the soils also had higher levels of soil organic carbon which although not ideal for soil properties, could be more difficult to increase under standard land management systems. A small percentage of soils had high levels and probably indicate what is possible under the more specialised land management systems.

One conclusion was that for many soils, increases in soil organic matter have the capacity to strongly influence only the surface soils, perhaps only the top 10 cms and the top 20 cms at most. This limits the capacity of soil organic matter to influence soil productivity. Even so the top 10 to 20 cm is a critical zone for the soil. It is the interface where seeds are sown, germinate and emerge. It is where a large proportion of plant materials are added to the soil for decomposition and the recycling of nutrients and where the rainfall either enters the soil or runs off. So the potential to improve soil condition in the top 10 to 20 cm is still critical for plant productivity.

A deficiency in the studies on the effects of soil organic matter on soil properties and productivity is the need to consider the impact of soil organic matter across a wide range of seasons and soil types. The means to address this deficiency is perhaps a program of applying biophysical models using the known effects of soil organic matter on the soil properties.

1. Introduction – Setting the Scene

1.1 Outline and Rationale

Organic matter has long been seen as an important component of soils. Many soil scientists and geologists consider that it is the influence of organic matter and biological activity that distinguishes soil from the underlying rock and parent material (Soil Survey Staff 2010). What is soil organic matter and its relationship to soil organic carbon is described in Box 1.

The influence of the soil organic matter on soils can be summarised as:

> *"Organic matter increases the water retaining power of soils, decreases water-runoff losses, improves aeration…. and produces a better soil structure or tilth……by promoting granulation … the damage done by water and wind erosion is greatly reduced.*
> *Organic matter improves the soil chemically by serving as a storehouse or supply of plant – nutrient elements. ….Most of the soil's nitrogen exists in organic form…. The organic colloidal materials have a much greater base-exchange capacity per unit weight than mineral colloidal materials, and hence the may act as buffers in the soil. These colloidal substances have a strong ability to adsorb and hold on to constituents of fertilisers and nutrients released from soil minerals.*
> *Organic matter improves the soil for the growth of microorganisms which are ….the agents whereby the plant-food elements of the soil are kept in circulation. It is the source of food and energy for the majority of soil microorganisms."*
> (Millar and Turk 1943, pp 223- 225)
>
> *"Soil organic matter ….is the natural nitrogen reservoir; furnishes large portions of soil phosphorus and sulphur; it protects against erosion; it supplies substances for desirable aggregate formation; and it loosens up the soil to provide aeration and water movement."*
> (Donahue et al. 1983, Chapter 5)

While these quotations and many more like it give a qualitative assessment of the effects of organic matter on soils they do not provide a quantitative, analysable assessment of the effects of organic matter on soil properties. They do not provide sufficient information to enable an objective quantitative estimate of how soil organic matter can influence a range of functional soil properties that are important for productivity and the sustainability of productive systems.

Much of the recent interest in soil organic matter has been as a vehicle to increase soil carbon levels in order to trade carbon or to gain benefits in market based instruments for carbon. However, it can be argued that the potential effects of soil organic matter on the productive capacity of soils are also of practical and economic importance and of significant interest to many in the agricultural community. It is the objective of this review, to concentrate on the capacity of soil organic matter to be an agent to maintain and improve soil condition and soil health and so the productive capacity of the soil.

This review examines the evidence in published information for the effects of soil organic matter on a range of soil properties that control the productive capacity of soils and the sustainability of productive systems. The intention is to develop some quantitative relationships between soil organic matter and some of the important functional soil properties. It examines the potential for setting

Box 1 - Basic Definitions

Soil Organic Matter

Soil organic matter in its broadest sense, encompasses all of the organic materials found in soils irrespective of its origin or state of decomposition. Included are living organic matter (plants, microbial biomass and faunal biomass), dissolved organic matter, particulate organic matter, humus and inert or highly carbonised organic matter (charcoal and charred organic materials). The functional definition of soil organic matter excludes organic materials larger than 2 mm in size.
(Baldock and Skjemstad 1999)

Soil Organic Carbon

Soil organic matter is made up of significant quantities of C, H, O, N, P and S. For practical reasons, most analytical methods used to determine the levels of soil organic matter actually determine the content of soil organic carbon in the soil. Conversion factors can be applied to the level of soil organic carbon to provide an estimate of the level of soil organic matter based on the content of carbon in the soil organic matter. The general conversion factor is 1.72, so the level of soil organic matter is ≈ 1.72 x the soil organic carbon. However this conversion factor does vary depending on the origin and nature of the soil organic matter from 1.72 to 2.0. The general convention now is to report results as soil organic carbon rather than as soil organic matter.
(Baldock and Skjemstad 1999)

Inorganic Soil Carbon

Significant amounts of inorganic carbon can occur in soils especially in more arid areas and in association with more mafic parent materials (limestones, basalts). Calcium carbonate as concretions, nodules or as diffuse carbonate can be very common in some soils. Carbon can also occur as dolomite or magnesium carbonate. Carbonates can be formed in the soil (pedogenic) or have a lithogenic origin (be derived from the parent material). The inorganic carbon is not included in the soil organic carbon content and measures are required to ensure it is not included in any determination of the soil organic carbon levels. Inorganic carbon does not contribute to the soil organic matter (Drees and Hallmark 2002).

various critical levels of soil organic matter for a range of important soil properties as well as the soil, climate and land management conditions to maintain these critical levels of soil organic matter. In some cases setting critical levels may be difficult because of differences in soil types and the criteria for setting the critical values. A previous review has been undertaken by Krull et al. (2004) which concentrated more on the scientific aspects of soil organic matter. It is hoped this review will complement and update the review by Krull et al. to some extent and hopes to address some of the issues more directly related to specific land management.

Loveland and Webb (2003) undertook a review of critical values of organic matter for agricultural soils. They concluded that although it is widely believed a major threshold is 2% soil organic carbon (3.4% soil organic matter), the quantitative evidence for this threshold is slight. The data suggested that more research is required on the nature of soil organic matter and its influence on the properties of a range of soil types under different land uses.

It is the objective of this review to review how soil organic matter influences a range of functional soil properties across a number of important soil types. An important principle is that while each the effect of soil organic matter on each individual soil property will be evaluated, the overall impact of organic matter on the productive capacity of the soil will include the cumulative effects of all the soil properties.

Box 2 - Outline of the Report

The report takes several steps as outlined below:

1. Introduction – setting the scene
2. Identifying the effect of organic matter on individual functional soil properties
3. Identifying the capacity of soils to achieve and maintain critical soil organic carbon levels under a range of environmental and land management options.

1.2 Historical Note

1.2.1 Crop-ley rotations

The importance of organic matter to maintain soil condition and soil productive capacity in cereal growing areas and pasture lands of Australia has long been recognised. Hamblin and Kyneur (1993) in their review of trends in wheat yields and soil fertility in Australia, identified nutrient exhaustion as a major factor in the initial decline in yields with the commencement of cereal growing in many districts in the pre-1930's. The practices of fallowing and the introduction of superphosphate in crop–ley rotations in the 1930's to 1960's reversed the decline in yields. The crop-ley system utilises legume based pastures for forage and for nitrogen nutrition and improves soil physical condition. Since the 1970's, better wheat varieties, especially the semi-dwarf varieties, and the introduction of better nitrogen nutrition, herbicides and alternative crops to improve rotations has continued to improve yields.

Several published papers confirmed the importance of the use of superphosphate and legumes in the ley phase for increasing soil nitrogen and organic matter levels and improving soil structure. Clarke et al. (1967) showed that growing ryegrass can improve aggregation of the surface soil of a red-brown earth that had been under a wheat-fallow rotation for 80 years. After 80 years of wheat –fallow soil carbon content had fallen to 0.84% in the top 7.5 cm which is considered low (Baldock and Skjemstad 1999; Hazelton and Murphy 2007). An important result was that the grass was necessary to improve the soil structure, and that the clover (subterranean clover) was not effective in improving aggregation. This was attributed to the low carbon: nitrogen ratio which meant the clover roots were rapidly decomposed, as well as the advantages of the fibrous root system of the grasses in promoting aggregation. The improved aggregation led to increased infiltration in the surface soil.

In a detailed study of crop-ley rotations on red-brown earths, Tisdall and Oades (1980a) demonstrated that the inclusion of a ryegrass pasture phase in a wheat rotation increased the soil carbon and nitrogen levels in the soil. The inclusion of a pasture phase also increased the proportion of soil allocated to water stable aggregates larger than 2.0 mm. Their conclusion was that the stability of aggregates larger than 250 μm requires actively growing root systems and fungal hyphae. This conclusion was enhanced by their subsequent paper (Tisdall and Oades 1980b). A note of clarification from these papers was that the results did not apply to soils where binding agents such as calcium carbonate, hydrous oxides of aluminium and iron were responsible for the stability of soil aggregates. In soils high in clay, aggregates are bound by the electrochemical effects of the clays and the interacting exchangeable cations. The plant roots, fungal hyphae and soil organic matter will have a lesser, but still important role in the stability of soil aggregates.

A study of the effect of poor structure on wheat germination was made by McIntyre (1955) and Millington (1959). They showed that a soil under long term wheat-fallow had some problems with wheat germination and emergence because of low aggregate stability and poor aeration in a year following heavy rain after sowing which restricted the ultimate wheat yield. The wheat yields were 2.61 t/ha for the unaffected crop but 0.61 t/ha for the crop affected by the germination problem. The soil carbon levels were estimated at less than 1.0 to 1.2% for the 0 – 10 cm depth. In other years when there was no heavy rain after sowing the effect was not present. It was also a red-brown earth soil.

1.2.2 Pasture systems

The effect of increased pasture growth on soil organic matter and soil nitrogen levels is shown in the pasture areas. Barrow (1969) on a sandy soil on the coastal plain in Western Australia showed large increases in soil carbon, soil nitrogen and cation exchange capacity with improved pastures and superphosphate. Soil carbon increased from 0.77 to 2.03%, total N from 0.038 to 0.122% and cation exchange capacity from 4.09 to 8.28 cmole(+)/kg.

Williams and Donald (1957) working on a podzolic soil with loamy sand topsoil and yellow clay subsoil in the southern tablelands of NSW showed improved pasture with subterranean clover and superphosphate increased total nitrogen, soil carbon and cation exchange capacity in comparison to a soil with native pasture. They made a rough estimate that for each 1% increase in soil organic matter the cation exchange capacity increased by 2.21 cmole(+)/kg. On the basis of this calculation they also estimated that soil organic matter has a cation exchange capacity of 220 cmole(+)/kg. The conclusion is that in these sandy soils soil organic matter is a very important component of the total cation exchange capacity. The data also indicated that water holding capacity of these soils increased by about 3% after the introduction of improved pastures and associated increase in soil organic matter. One important conclusion was that soil organic matter could not increase without sufficient nutrients to maintain a standard ratio of C:N:S:P within the organic matter. This ratio was approximately 100:10:1:1 but can vary.

In South Australia, Russell (1960) recorded the effects of adding phosphate to a series of pasture plots. Phosphate encouraged the growth of clovers to different degrees. The growth of the clovers had similar effects to those reported in the publications above. An additional effect was the decrease in bulk density from 1.43 to 1.20 t/m^3. This occurred where there was the strongest increase in total nitrogen and soil organic matter. Soil organic carbon increased from less than 1.0% to 2.5% where both phosphate and lime were added. This soil was also loamy sand.

1.2.4 Soil acidification

An important effect recorded by Williams and Donald (1957), Russell (1960) and Barrow (1969) was that the increase in soil organic matter and total nitrogen associated with the improved pasture was accompanied by increased soil acidity and a lowering of pH. The fact all these soils are quite sandy and have relatively low buffering capacity is one reason for this. While increasing soil organic matter can lower soil pH, the strong acidification recorded by Williams (1980) is not due solely to the accumulation of soil organic matter. The strong acidification was largely a result of nitrate leaching and the removal of exchangeable cations such as Ca and Mg in agricultural produce. In fact, the presence of soil organic matter can buffer against such strong acidification (Fenton and Helyar 2007; Aitken et al 1990). The strong acidification effect of soils under pasture improvement and increased nitrogen was documented by Williams (1980) who noted that a soil with a natural pH of about 6.0 in the topsoil could fall to close to 5.0 after 40 to 50 years of pasture based on subterranean clovers. Soil acidification can affect the soil to up to 30 cm depth.

1.2.5 Conclusion

This short historical note is presented mainly to set the scene and put the importance of soil organic matter for agricultural productivity into context. A few points are relevant:

1. Soil organic matter provides a substantial part of the physico-chemical activity in soils needed to for productivity, especially in soils with lower amounts of clay.
2. Much of the early work concentrated on red-brown earths with the lighter textured soils or other lighter textured surface soils. It is necessary to consider how soil organic matter influences a wider range of soil types.
3. Soil organic matter has the capacity to have substantial effects on soil properties that impact on the productivity of soils.
4. The management of soil organic matter is complex and it is only one factor in the overall maintenance of the productive capacity of soils.
5. Soil organic matter can improve a wide range of soil properties.

1.3 Nature of Soil

Soil is made of several components, including the solid mineral component of the soil, the organic component and the porous component. How the porous component is divided between water and air will vary with the moisture conditions of the soil. Each of the components can be further subdivided. How the soil is constructed into these different components will have a large effect on the functional soil properties and the stability of those functional soil properties. The amount and type of organic matter can have an effect on how the soil is constructed between the different components. In turn, how the solid mineral component of the soil is constructed can affect how the soil organic matter will influence the soil. Generally the more clay, the less will be the influence of the soil organic matter.

The key components of soils are then:

- The primary particles which are clays (< 0.002 mm), silt (0.002 to 0.02 mm), fine sand (0.02 to 0.2 mm), coarse sand (0.2 to 2.0 mm) and gravel (2 to 6.0 mm). The relative proportions of these different sized particles determine the physical and chemical characteristics of the soil.
- The porosity is determined by how the different primary particles are held together to form aggregates and soil structural units. Porosity is often divided into: macro-porosity for aeration and rapid water flow (pores > 0.050 mm), micro to meso-porosity for water storage for plants (0.0002 to 0.05 mm) and residual porosity (< 0.0002 mm). A healthy, functional soil requires a good balance between all pore sizes which requires soils to have a good stable structure and aggregate stability.
- The physico-chemical activity which enables the soil to interact with soil nutrients and toxic compounds in order to process, store or mineralise them depending on what is required to ensure that plants and the soil microbial and fauna populations have the nutrients they need and do not encounter adverse chemical agents. The primary sources of physio-chemical activity in soils are clay particles and soil organic matter. Silt size particles can provide smaller amounts of physico-chemical activity. The nature of the organic matter determines the degree of physico-chemical activity provided by the organic matter.

1.4 Functions of Soil – Soil Health

Soil is one of the nation’s most valuable assets, and it is frequently it is the characteristics and fertility of this resource which determines a region’s wealth. Soil is the main resource upon which agricultural production depends. It is the foundation for many roads and buildings and is often used as construction material. Soils are used as a filter or sink for effluent and waste materials. More recently soil has been seen as a possible sink for carbon to assist in reducing the level of carbon dioxide in the atmosphere and is also seen as playing a major part in the cycle of other greenhouse gases such as nitrous oxide and methane. Soil health and soil condition is important and the management of soil organic matter is an important part of managing soil health and maintaining soil condition. The pressure on the management of soil organic matter is increasing as costs of inputs for agriculture increase and the capacity and ability to overcome soils in poor condition by adding more fertiliser, adding one more cultivation, adding one more irrigation or adding another input are diminished.

Soil health is defined as:

> “Soil health is the condition of the soil in relation to its inherent (or potential) capability, to sustain biological productivity, maintain environmental quality, and promote plant and animal health. A healthy soil is productive, sustainable and profitable.” (MacEwan 2007).
> http://vro.dpi.vic.gov.au/dpi/vro/vrosite.nsf/pages/soilhealth_what_is

From a practical viewpoint it is useful to consider that soils provide several functions across the landscape for growing crops and pastures, and for the community.

Some of the functions provided by soils include:

- Base and medium for plant growth (supplies water, air and nutrients)
- Filtering and storage of water
- Control of water flow
- Pool of nutrients
- Resists erosion
- Foundation and base for infrastructure
- Reduce availability and movement of toxic materials
- Populated by soil biota essential for decomposing and recycling nutrients

From a purely agricultural viewpoint, a healthy soil:

- Supplies nutrients, water and oxygen for healthy plant growth
- Allows water to infiltrate freely
- Resists erosion
- Stores water
- Readily exchanges gases with the atmosphere
- Retains nutrients
- Acts as an environmental buffer in the landscape
- Resists disease
- Contains a large and diverse population of soil biota
- Is not acidifying or salinising
- Has a range of pore spaces to house organisms, nutrients and water” http://vro.dpi.vic.gov.au/dpi/vro/vrosite.nsf/pages/soilhealth_what_is

For agriculture the most important is the soil’s function as a base and medium for plant growth. However, storage of water, control of water flow and pool of nutrients are clearly important functions for the supply of water and nutrients to plants, control of erosion and the long term sustainability of agricultural production. The important question is however, what are the precise functions that soil organic matter can maintain and improve in soils?

1.5 The Nature of Soil Organic Matter

1.5.1What is soil organic matter?

The ultimate source of organic matter for most soils is through the fixation of carbon dioxide from the atmosphere through photosynthetic reactions by plants. There is also a very small input from autotrophic bacteria (Tate 1987). However, in some instances there may also be some input from industrial and mining products derived from petroleum or coal. At the broad scale these sources of soil carbon are insignificant.

Soil organic matter is derived from organic materials that are added to the soil and the majority of soil organic matter derives from the breakdown of residues remaining after plants have died. These residues can take the form of root residues located in the soil matrix or leaves, stems and stubble existing as litter on the soil surface. Animals also provide a proportion of the soil organic matter to varying degrees depending on management and the ecosystem.

Soil organic matter includes plant debris, root exudates and animal materials; their degradation products; and products synthesised by soil microorganisms and other soil biota.

Included in soil organic matter is the finer root material of living plants, fungal hyphae and bacteria, as well as small soil fauna.

For cereal or grass straw and stubble the amount of carbon that remains in the soil as soil organic matter after 12 months was estimated by Stevenson (1982, pp12) at between 14 to 47%, while the amount of root material that remained was estimated at 63 to 74% . While only a limited set of information, this does give an indication of how much carbon in plant material is returned to the soil as soil organic matter.

Additions of large amounts of organic materials as composts or as biochar can increase the levels of soil organic matter in soils (Gibson et al. 2002; Quilty and Cattle 2011).

1.5.2 The Measurement of Soil Organic Matter

In the laboratory, soil organic carbon is usually measured to estimate the amount of soil organic matter. This is because it is the most reliable and easy method to estimate soil organic matter. However to convert the soil organic carbon measurement to soil organic matter it is necessary to convert the soil organic carbon reading based on the assumed percentage of carbon in the organic matter. This can vary and the conversion can range from 1.72 to 2.0 depending on the source of the materials for the soil organic matter. In practice the value of 1.72 is used (Nelson and Sommers 1996; Baldock and Skjemstad 1999). For more precise scientific work only the values of soil organic carbon are used.

Soil carbon is measured by oxidation using wet or dry methods (Baldock and Skjemstad 1999). The wet method uses a chemical agent such as dichromate but this method can vary depending on whether samples are heated. In the original method (Walkley and Black 1934) the samples were not heated and there is incomplete oxidation of the soil organic carbon. With heating this method (Heanes 1984) generally does oxidise the soil organic carbon completely although some of the coarser carbonised carbon may resist oxidation (Baldock and Skjemstad 1999).

In dry oxidation the soil is heated in oxygen and all carbon, organic and inorganic, is converted to CO_2. The amount of CO_2 is analysed usually spectroscopically. The result must be corrected for inorganic carbon such as carbonates or these must be removed by acid treatment before the sample is analysed (Baldock and Skjemstad 1999). The dry oxidation method is currently the preferred method for most samples, although where there are large amount of carbonate present, the wet oxidation method may still have some advantages (Schmidt et al. 2012).

A review of how effective the different methods of measuring soil organic matter and soil carbon are at detecting different organic materials was undertaken by Conyers et al (2011). The general conclusion was that the dry oxidation method and Heanes method (1984) were the most reliable across a wide range of materials.

These methods measure total soil carbon which is used to estimate soil organic matter. None of them measure the different components of soil organic carbon or soil organic matter which is more difficult and costly.

1.5.3 Soil Organic Matter Fractions

As a result of the way soil organic matter is formed, there are several different types of soil organic matter present in soils. Over time several approaches have been taken to defining the types of soil organic matter. In general, three distinct types have been defined (adapted from Stevenson 1982; and Baldock and Skjemstad(1999) and Six and Jastrow 2002):

- Decomposing plant residues, fungal hyphae, fine plant roots and associated biomass which usually is unstable and has a short turn over time (< 1 – 2 years). This is sometimes referred to as the labile fraction.
- An active component that includes partially stabilised organic materials and microbial metabolites – usually referred to as the humus fraction (turn over time 5 to 25 years)
- Resistant fraction which is strongly stabilised either chemically or physically. This is also may include part of the humus faction (turnover time 250 to 2500 years).

Six and Jastrow (2002) explain that this division is largely conceptual and it has been difficult to define them in functionally meaningful pools or specifically measurable fractions. Huang (2004) also provides a summary of a range of methods of dividing soil organic matter into fractions. In Australia the soil carbon fractions are defined generally as in Table 1.1 with the original three classes evident but a more comprehensive set of definitions is given. The factions are defined experimentally in Table 1.2. When measured in the laboratory and reported on a mass basis in soil test results, soil organic matter consists of all these fractions. The most important fractions are the particulate organic matter (POC) and humus. The common percentages of the different fractions are shown in Table 1.1.

The labile fraction of the soil organic matter has been considered to be of considerable practical importance, as Tisdall and Oades (1980a) identified, it is primarily responsible for the development of large soil aggregates (> 250μm). Therefore there have been attempts to measure the labile fraction of soil organic matter using potassium permanganate as an extractant (Blair et al. 1995). However there is some doubt on exactly which components of soil organic matter are oxidised by the potassium permanganate and therefore its use as a universal measure of labile soil organic matter is uncertain (Skjemstad et al. 2006). It is probably more suited for use with calibration to local soils, practices and land uses. At present the physical measure of particulate organic carbon (POC) is generally preferred as an estimate of the labile soil organic matter fraction.

For functionality it is potentially useful to consider the effects of the different fractions on key features as described in Table 1.1

1.5.4 Accumulation and formation of soil organic matter - a simple description

In the initial stages, organic material or litter, largely plant material, is composed of complex carbohydrates (celluloses, polysaccharides); plant proteins (high in nitrogen), cuticular waxes and lipids and lignin (can be difficult to decompose). The stability of the components within initial organic material is from least to most stable: simple sugars < amino acids < proteins < cellulose < hemicellulose < fats, starches and waxes < lignin and tannins (Quideau 2002). These compounds undergo decomposition releasing carbon dioxide.

The decomposition process often forms small particulate pieces of partially decomposed organic materials. While the carbohydrates and some of the proteins can be rapidly broken down, the lignins are typically viewed as being the more resistant. The lignin in combination with other partially decomposed organic compounds often form the highly complex organic materials often referred to as humus, a complex amorphous organic material. The source material for humus includes incompletely decomposed materials from the organic litter as well as the products of microbial activity (see Figure 1.1). The products derived from lignin degradation together with those synthesized by micro-organisms undergo polymerization to form humus compounds. The components in humus have been shown to have been derived from lignin decomposition and from synthesis by microorganisms (Stevenson 1982, pp206). There are several environments in which humus has formed in the absence of lignin rich plant material so the synthesis of humic substances has occurred without the input of lignin (Stevenson 1982, pp 209). Other organic compounds may also combine with amino compounds to form nitrogen rich humus. Some humus compounds are complex molecules that in the form of the original bio-molecules are synthesised by plants and microorganisms.

Humus is often strongly associated with individual soil particles and can coat individual particles or exist in pores between the particles, and form complexes with clays. Humus can develop colloidal properties and so have some physio-chemical activity being able to absorb water and have cation exchange capability. Humus usually has an approximate composition in regard to the ratio of C:N:P:S of about 100:10:1:1 (Rice 2002; Williams and Donald 1957), although this can vary depending on the source of the organic matter. This ratio can vary depending on the source and history of the soil organic matter (Kirkby et al. 2011).

Humus is composed of a range of organic materials including:

- polymers with various combinations of carboxylic, aromatic and phenolic substituents – the balance and size of these determines if the molecules will be in the humic acid group, the fulvic acid group or the humin group. The humin group is highly insoluble or so tightly bound to clay particles of hydroxides that it is very resistant to decomposition (Allison 1973, Stevenson 1982). These three groups can vary in their physical and chemical properties.
- Fats, waxes and resins including lipids, long chain fatty acids (C20 to C34), paraffins, polycyclic hydrocarbons and some porphyrins.
- Amino acids, which often react with the humic substances
- The carbohydrates including sugars, celluloses, polysaccharides and hemicelluloses which are largely unstable and also react with the humic materials or are decomposed by microorganisms.

1.5.5 Conclusion

The fractions of soil organic matter vary in the physico-chemical properties and in their effects and interactions with the soil mineral fraction. Therefore each of the fractions can be expected to have a different effect on the processes and soil properties that affect the productivity of a soil. To date there is only limited information on the relative importance of each of the fractions for each of the soil properties and soil processes that determine the prodctivity and sustainability of soil. It is one of the aims of this review to identify that information and reach some conclusions about how the different soil fractions affect the soil properties and soil processes that determine the productivity of the soil.

Table 1.1 Fractions of soil organic matter based on Baldock and Skemstad (1999) and use for soil carbon models (Six and Jastrow 2002).

		Composition	**Amount in soil**	**Fractions for models**
Non-living organic matter	Dissolved organic matter		< 0.1%	Labile soil carbon Active pool Decomposable plant materials (low C:N ratio, low lignin) Resistant plant material (high C:N ratio, high lignin)
	Particulate organic matter	Litter	5 to 20%	
		Macro-organic material		
		Light fraction		
	Humus	Non-humic biomolecules	65 to 80%	Slow soil carbon
		Humic substances		
	Inert organic matter	Charcoal/char	1 to 5%	Passive soil carbon Inert organic materials
Living organic matter	Phytomass	Plant roots, litter	1%	Labile soil carbon Active pool Decomposable plant materials (low C:N ratio, low lignin) Resistant plant material (high C:N ratio, high lignin)
	Microbial biomass	Bacteria	2 to 5%	
		Fungal		
	Faunal biomass		<1%	

Note that results from the National Soil Carbon Research Program (SCaRP) indicate values for the fractions as below (Jeff Baldock pers comm.):

Particulate Organic Matter - 10 to 60%
Humus - 20 to 80%
Inert Organic Matter - 0 to 60%

Fractions for description

- Dissolved organic matter - < 0.45 mm organic materials in solution
- Particulate organic matter (POC) – organic fragments with a recognisable cell structure
- Litter organic materials located at the surface and devoid of mineral particles
- Macro-organic matter fragments of organic material >50µm
- Light fraction – organic materials separated from soils by flotation
- Humus – Amorphous organic materials
- Non-humic biomolecules – organic molecules that can be placed into discrete classes of biopolymers
    - polysaccharides
    - proteins
    - waxes
    - lignin
- Humic substances – organic molecules that are biopolymers without discrete structures
    - humic acid
    - fulvic acid
    - humin
- Inert organic matter – highly carbonised organic materials including charcoal and charred plant materials

Fractions for carbon modelling

- Active soil carbon pool – turnover time 1.5 years
- Slow soil carbon pool – turnover time 25 years
- Passive soil carbon pool – turnover time 1000 years

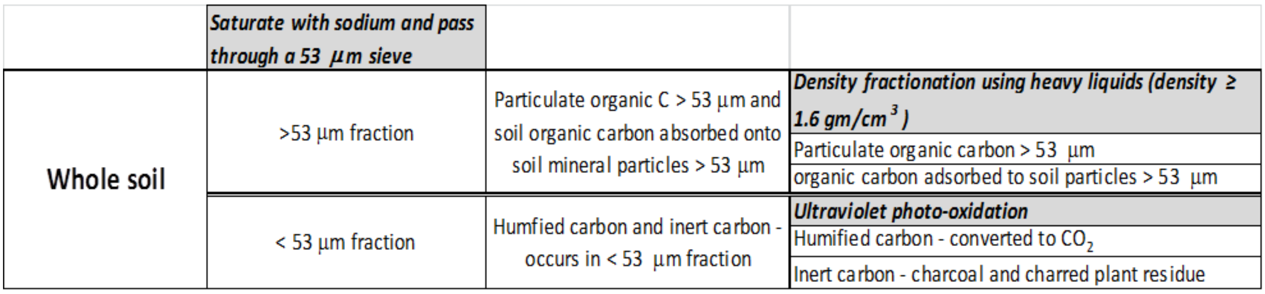
Table 1.2. Method for determining fractions of soil organic matter in the laboratory (Baldock and Skjemstad 1999; Skjemstad et al. 1996).

Step 1

Soil is passed through a 2 mm sieve. Material including organic material such as gravel, stubble and large roots is removed from the soil sample. The gravel is weighed and the organic material can also be weighed. The soil passing through the 2 mm sieve is then used for further analysis.

Step 2

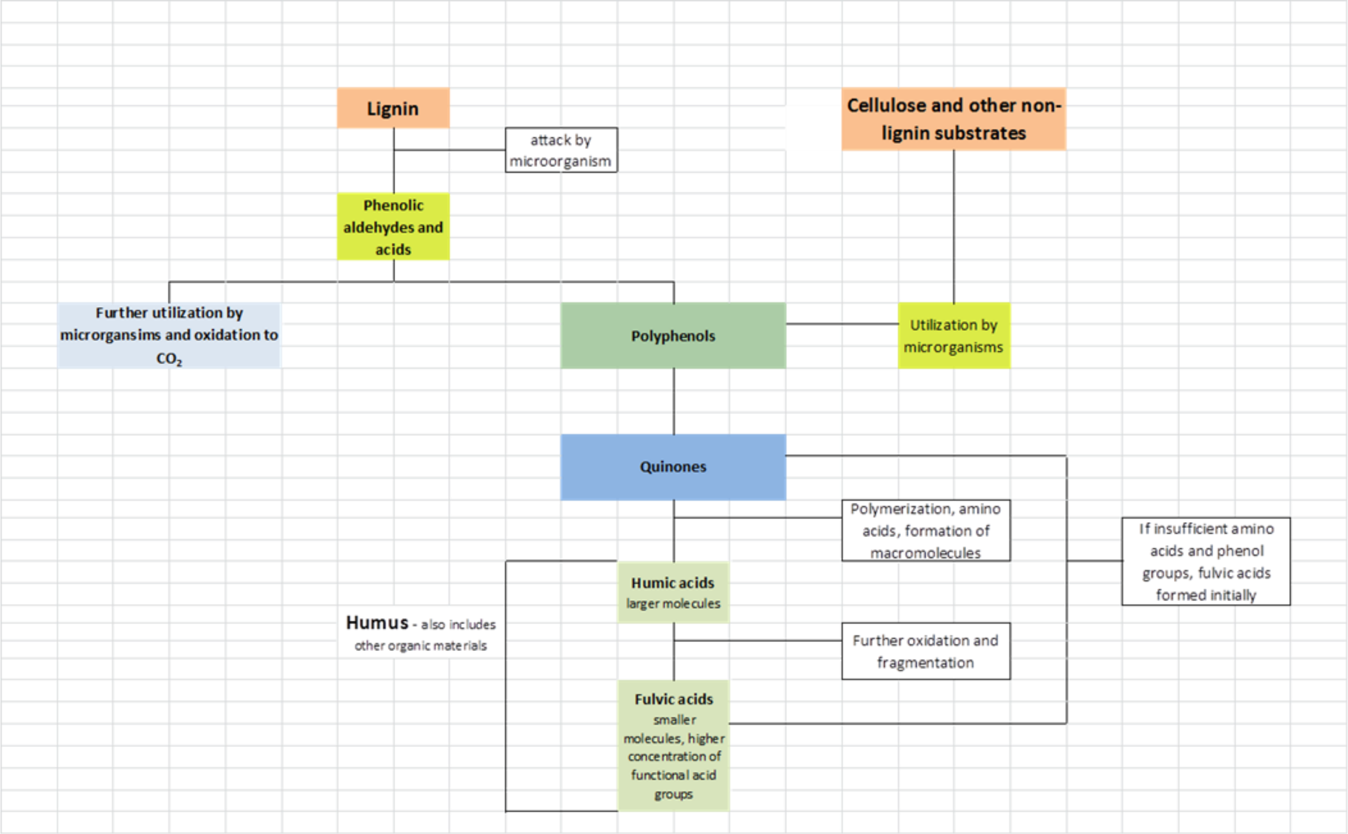
The < 2 mm material is further processed as below.



	Saturate with sodium and pass through a 53 µm sieve		
Whole soil	>53 µm fraction	Particulate organic C > 53 µm and soil organic carbon absorbed onto soil mineral particles > 53 µm	***Density fractionation using heavy liquids (density ≥ 1.6 gm/cm^3)***
			Particulate organic carbon > 53 µm
			organic carbon adsorbed to soil particles > 53 µm
	< 53 µm fraction	Humfied carbon and inert carbon - occurs in < 53 µm fraction	***Ultraviolet photo-oxidation***
			Humified carbon - converted to CO_2
			Inert carbon - charcoal and charred plant residue

Figure 1.1 Scheme showing the development of soil organic matter in soils (Based on information in Stevenson 1982).

a. Technical view



b. Simplified view.

- Living organisms - < 5%
- Fresh organic matter, leaves, roots, exudates, animals bodies etc. - < 10%

⬇

- Active fraction, decomposing organic material – 33 to 50%

⬇

- Stabilised organic matter – 33 to 55%
- Charcoal, resistant organic material - <2% to 20% depending on soil history.

1.6 The Functions of Soil Organic Matter in Soils

The functions of soil organic matter can be summarised as below (after Baldock and Skjemstad 1999)

- Physical Functions
    - Improves the structural stability of soils at size scales ranging from the < 250 μm scale of microaggregates to large aggregates (> 250 μm).
    - Influences the hydrological properties of soil including the water holding capacity, infiltration properties and hydraulic conductivities of subsoil layers.
    - Improves the friability and tilth of the soil, lower soil strength
    - Improves aeration
    - Alters thermal properties
- Chemical Functions
    - Contributes to the cation exchange capacity of the soil.
    - Enhances the ability of the soil to buffer against changes in pH.
    - Complexes cations and anions which can:
        - reduce the availability of toxic agents such as Al^{3+} in the soil solution.
        - maintain the availability of nutrients such as phosphorus in a potentially available form and prevent them from becoming unavailable.
        - Promote the binding of organic matter to soil minerals.
        - Interacts with herbicides and pesticides
- Biological functions
    - Provides a source of energy and food for microorganisms that are essential to biological processes in the soil.
    - Serves as large reservoir of nutrients (especially nitrogen but also phosphorus and sulphur, and the micronutrients) which are released to the plant – available pool by decomposition processes – essential to the recycling of nutrients.
    - Can contribute to the biological control of disease organisms in the soil.

1.6.1 The Influence of the Soil Organic Fractions on the Functions of Soil Organic Matter

As described in Section 1.5, soil organic matter is not a single uniform material but is very diverse in its chemical and physical properties which will influence its capacity to do certain functions. An estimate of the capacity for the different soil carbon fractions to undertake some of the functions of soil organic matter is shown in Table 1.3. The humic and fulvic acids would seem to have the highest chemical activity, but the particulate organic carbon has the highest capacity to maintain the stability of larger soil aggregates. Overall, it would seem that a balanced amount of the soil carbon fractions is desirable for all functions. Krull et al. (2004) also identified the importance of the different fractions for different functions (see Figure 1.2, reproduced from Krull et al. 2004). An important aspect identifed by Krull et al was that the importance of soil organic matter to carry out certain functions in the soil will vary with soil type. For example the need for soil organic matter to provide cation exchange capacity is most important in sandy soils. On the other hand the need for soil organic matter to provide a food and energy source for the microbial population is needed in all soils, regardless of clay content or texture.

Figure 1.2. The optimum expression of each soil organic matter function requires different proportions of the soil organic carbon pools (soluble, particulate, humic and inert). The degree to which soil organic matter can influence a particular function may vary with soil type (texture, clay content, sesquioxide content, sodicity (From Krull et al 2004).

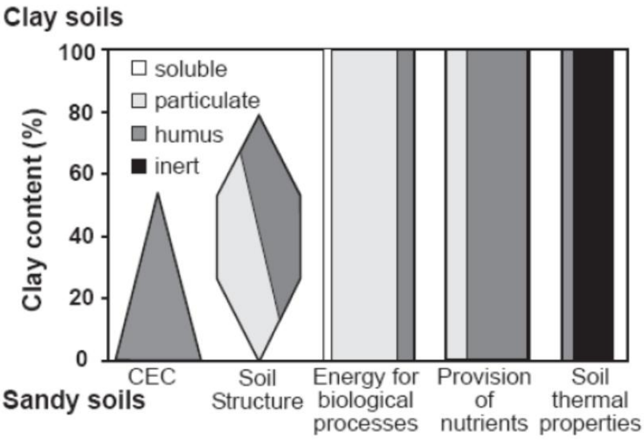


Table 1.3. Fractions of soil organic matter and estimated potential of the fractions to contribute to soil functions. This is an initial estimate to be used to understand the variation in results of how soil organic matter relates to some functional soil properties. The humus is divided into humin (H)(largely insoluble organic material), humic acid (HA)(very large polymers – molecules) and fulvic acid (FA)(smaller polymers – molecules) fractions based on Stevenson (1982) and Slattery et al. (1998). The cation exchange capacity of the organic materials is highly dependent on pH, generally increasing from low values in acid conditions to higher levels as the pH increases.

Fraction		**Function**				
		Cation exchange capacity	**Aggregate stability-structure development**	**Water holding capacity**	**Complexing metal polyvalent cations (eg Al^{3+}), anions (P)**	**Provide food and energy source for microbial population**
Particulate organic carbon (POC)		Low to very low	high for aggregates > 250 µm	low	very low	very high
Humus	**H**	low	high for aggregates < 250µm	high	Low	moderate to high
	HA	High* (560 – 770 cmole(+)/kg)		high	high	moderate to high
	FA	very high* (640 - 1420 cmole(+)/kg)	low to moderate for aggregates < 250 µ	moderate	very high	moderate to high
Resistant organic carbon /Char/Charcoal		low	low	Low to moderate	moderate	low

- Soil organic matter has a CEC of 100 to 500 cmole(+)/kg. This because the CEC for organic matter includes all the fractions of soil organic matter

1.7 Current Interpretive Levels of Soil Organic Carbon

The levels of soil carbon that are considered good or adequate have been estimated for various locations and soil types throughout Australia (Table 1.4). However, it is usually not entirely clear if these levels are necessarily related to the condition and functionality of the soils or whether the defined levels relate more to what are the "expected " levels of soil carbon given the soils and climate for a particular area. Given the range of values, it would appear that the latter is the case. An important implication of the critical levels of soil carbon given by Isbell for the Australian Soil Classification is that as the levels of soil organic carbon increase into the range of above 4 to 6% depending on texture, the soils begin to change from mineral dominated soil materials, to soil materials dominated by organic materials and as the soil organic carbon increases beyond 13 to 17%, the soil materials are predominately organic materials.

It is clear that it is possible there may be a different scale for interpreting soil organic carbon levels as they affect soil function for each soil type and for each soil property.

As mentioned by Loveland and Webb (2003) a universal value of 2% soil organic carbon is considered to be a critical value that accounts for a large change in the functionality of soil properties. In a review of data from dryland sites in Alberta Canada, Janzen et al. (1992), as reported in Krull et al. (2004), showed that dry matter yields decrease after soil organic matter levels fall below 2%. Although Loveland and Webb admitted they could not find strong evidence to support the 2% value as being critical, it may provide a useful guideline.

Some of these critical values can be tested against the published data on the effects of soil organic matter on functional soil properties.

1.8 Some Intended Guidelines for the Report

a. The report is focussed on the effect of soil organic matter on soil condition rather than on the capacity of the soil to store carbon. Therefore the emphasis is on the soil organic carbon content or soil organic carbon per cent (g/100g) and not on the soil carbon store or soil carbon stocks. In general the soil carbon content for the 0 to 10 cm layer is taken as a standard, but obviously there will be a range of soil depths in the published data.

   The procedure will be where possible to convert all soil organic matter and soil organic carbon values (g/100g) to a standard value based on:

   - Soil organic matter = 1.72*soil carbon (see Section 1.5.2)
   - Soil carbon values converted to the values obtained using the dry oxidation method in a LECO furnace (Baldock and Skjemstad 1999; CSIRO 2013a) using the standard conversions in Australian Greenhouse Office Technical Report (Skjemstad 2000). The LECO furnace is a commercially available instrument which uses high temperature oxidative combustion to convert carbon in the soil to CO_2 under standard conditions. The amount of CO_2 emitted is measured by non-dispersive infrared radiation.

Because soil organic carbon is the commonly measured soil property and values discussed will be soil organic carbon rather than soil organic matter. However, it is the soil organic matter that affects the soil properties. The review will attempt to clarify exactly which soil quantity is being discussed but it may require the reader to be aware that the quantity referred to will vary depending on the particular data sets or issue under discussion.

When estimating the amounts of nutrients available in units such as kg/ha, there are some advantages in using the soil carbon store in t/ha as the basic unit to assess soil condition or the potential impact of soil organic matter on productivity.

b. Develop relationships between soil organic carbon and functional soil properties for different surface soil types and some subsoil types based on soil texture (clay and sand content), sodicity and amount of sesquioxides (Ferrosols). It was recognised in the original GRDC Report (Krull et al. 2004) that the amount of clay and sand content would affect the relationships between soil carbon and functional soil properties.
c. Use published pedotransfer functions to enhance and fill in gaps of relationships between soil organic carbon and functional soil properties.
d. Discussion and description will include sections on both soil properties and soil processes.

1.8.1 Different surface soil types

The productivity of soils is influenced by the overall physico-chemical activity of the soil which determines the ability of the soil to:

- interact with and store soil nutrients,
- buffer against changes in soil pH,
- provide energy sources for microorganisms,
- provide an environment for biodiversity to develop (ensure nutrient cycling and disease control)
- interact with potentially toxic chemicals and compounds in the soil environment
- develop soil structure
- provide water holding capacity
- protection against erosion

In general the sum of the physico-chemical activity of a soil is provided by the content and type of clay and by the soil organic matter. As can be seen from Table 1.5, clay minerals such as the smectites, which are commonly derived from basic rocks such as basalts, have a high cation exchange capacity and swelling capacity and so have a high physico-chemical activity. Other clays such as kaolinite which are derived from highly weathered and infertile shales have low physico-chemical fertility, even though the clay contents may appear adequate.

The consequence of this is that to establish effective relationships between soil organic matter and soil properties it is necessary to identify different surface soil types based largely on soil texture. Soils low in clay are much more dependent on soil organic matter to provide physico-chemical activity and so tend to have several functional soil properties that are more affected by changes in soil organic matter levels. This was recognised by Oades (1993) in rating the importance of biology and abiotic processes in developing soil structure. Oades identified that in sands (< 15% clay) biotic processes dominate the development of soil structure whereas in clays (> 35% clay), biotic processes are minimal with shrink-swell processes and abiotic influences being the most important. For loams (clay content of 15 to 35%) biotic and abiotic processes are important.

The effect of clay content and texture on the importance of soil organic matter is shown in Figure 1.2 reproduced from Krull et al (2004). The relative effects of soil organic matter and some of the fractions on the key functions of soil can be seen in the diagram.

Murphy et al. (2013) in a field classification of surface soils identified the importance of sodicity levels for more clayey soils and organic matter for soils of lower clay content. Water holding capacity is also dependent on soil texture and the effects of soil organic matter are strongest in more sandy soils (Hudson 1994; Minasny et al. 1999; Rawls et al. 2003). Other soil chemical properties are influenced by soil organic matter with the importance of soil organic matter being modified by soil texture. Aitken et al. (1990) recognised the effect of clay and soil organic matter on the buffering capacity to acidification.

In this review, the general approach is to consider relationships between soil organic matter and the functional soil properties for sands (< 15% clay), loams (15% to 35 % clay) and clays > 35% clay). For some specific soil properties, soil textures are considered in a higher resolution because the specific effects require this or because the special properties of the clay minerals (e.g. self-mulching surface soils) require individual consideration.

Table 1.4. Interpretive values or critical values for soil organic carbon from various sources (adapted from Baldock and Skjemstad 1999). These are generally not based on the effects on the functions of the soil but rather on the expected values for a given climate and soil type.

Soil Organic Carbon Rating / Status	Soil organic carbon g/100g			
Peverill et al. 1991 - Victoria				
	Low rainfall		High rainfall	
	Crops	Pastures	Crops	Pastures
Low	<0.90	< 1.74	<1.45	< 2.90
Normal	0.90 – 1.45	1.74 – 2.62	1.45 – 2.90	2.90 – 5.81
High	> 1.45	> 2.62	> 2.90	> 5.81
Hughes et al. 1996 – South Australia				
	Sand	Sandy loam	Loam	Clay loam/clay
Low	<0.50	< 0.70	< 0.90	< 1.20
Moderate	0.50 – 1.00	0.70 – 1.40	0.90 – 1.80	1.20 – 2.00
High	> 1.00	> 1.40	> 1.80	> 2.00
Purdie 1998 – Western Australia				
	A horizons	A2, B horizons		
Low	<1.00	< 0.01		
Medium	1.00 – 2.00	0.01 – 0.05		
High	> 2.00	> 0.05		
Hazelton and Murphy (2007) – New South Wales				
Extremely low	< 0.40			
Very low	0.40 – 0.60			
Low	0.60 – 1.00			
Moderate	1.00 – 1.80			
High	1.80 – 3.00			
Very High	>3.00			
Isbell 2002 – Australian Soil Classification				
	10% clay	20% clay	30% clay	50% clay
Humose soil	>4.3	>4.7	>5.0	>5.7
Organic soil	>13	>14	>15	>17

Table 1.5. Cation exchange properties of common soil components (after McKenzie et al. 2004).

Soil Component	Cation exchange capacity cmol (+)/kg
Coarse sand	Negligible
Fine sand	Negligible
Silt	Negligible in most soils ≈ 1 to 2 in some soils
Kaolinite clay	3 to 20
Illite clay	10 to 40
Smectite clay	80 to 150
Iron and aluminium oxides	4
Organic matter (Individual fractions of the organic matter may be lower or higher)	150 to 500

Summary of Key Points – 1. Introduction – Setting the Scene

- Managing soil organic matter levels is generally considered an effective way to maintain and improve the productive capacity of soils.
- Soil organic matter has physical, chemical and biological functions in soils. As the amount of soil organic matter is reduced, the capacity of soils to undertake these functions can be impaired or diminished.
- Soil organic carbon is only a fraction of soil organic matter and is usually considered to be about 58% of the soil organic matter, but this can vary from about 50 to 58% depending on the source of the organic matter. Nitrogen, phosphorus, sulphur and other elements make up the remainder of soil organic matter. By convention the level of soil organic carbon is measured to estimate the levels of soil organic matter in soils.
- The soil organic carbon concentration (g/100g) in soils is the most direct measure of the soil condition. For evaluating the effect of soil organic matter on soil function, the soil organic carbon concentration or soil carbon content (g/100g) is the most suitable measurement.
- Soil organic matter is composed of the following:
    - living organisms such as fine root and their exudates, microbial populations (2 to 5%)
    - decomposing organic material such as plant debris, dead microorganisms, animal materials (5 to 20%);
    - humus which includes large complex organic molecules and some smaller complex molecules often attached to soil particles (65 to 80%); and
    - inert or resistant organic materials such as charcoal or char (1 to 5%).
- Historically the introduction of a pasture phase into land management systems increased soil organic matter and improved soil properties including water holding capacity, aggregate stability and cation exchange capacity and reduced soil erosion. However the introduction of legume based pastures also enhanced the process of soil acidification.
- There are some guidelines for interpreting values of soil organic carbon that are in current use across Australia, but these are used for a wide range of purposes and do not necessarily relate to the influence of soil organic matter or soil organic matter on the functions of soils.
- The relative importance of functions attributed to soil organic matter will vary with soil type. The effects of soil organic matter on soil chemical activity such as cation exchange capacity and buffering capacity are most important for sandy, low clay content soils. The capacity of soil organic matter to provide aggregate stability is more important in loam and some clay soils.
- Different fractions of the soil organic matter are dominant for different functions of soil organic matter. The fibrous, particulate soil organic matter tends to be more important for soil physical properties and the humus, more molecular soil organic matter has a larger influence on soil chemical properties.

2. Effects of Soil Organic Matter on Soil Physical Properties

2.1 Ideal Soil – Non Limiting Water Range

2.1.1 General

The physical properties of the soil can affect plant growth and productivity but several key factors such as structural stability and erodibility also affect the sustainability of the soil for production. To summarise the effects of soil structure on plant productivity it is useful to examine the non-limiting water range as defined by Letey (1985 and 1991). Soil organic matter also influences the aggregate stability of soils, bulk density and other critical soil physical properties. These will be discussed below, including the potential for different soil organic matter fractions to affect these properties and whether there are any guidelines for the levels of soil organic matter which will affect these properties.

2.1.2 Ideal Soil – Non Limiting Water Range

A convenient method to summarise the physical fertility of soils or the impact of soil physical properties on productivity is the non-limiting water range. This was developed by Letey (1985 and 1991) and recognises that effective root growth in soils is bounded by high mechanical resistance as the soil dries and low air porosity as the water content increases. In soils with “good” soil structure there is a wide range of moisture contents between these two boundary conditions. In soils with “poor” soil structure, there is only a narrow range of moisture contents between these two boundary conditions. In some cases as in dense clods of some sodic soils and some compacted soils, there may be no moisture content where there are suitable conditions for root growth. This can be evident in the field when roots clearly grow on the outside of the clods or peds but do not penetrate inside. Micro-variation occurs within soils when plant roots can grow into existing pores created by previous roots, despite conditions within clods or soil peds.

2.1.3 Boundary conditions

The mechanical resistance to root growth varies. A summary of the values that limit root growth is provided by Hazelton and Murphy (2007). Anderson et al. (1980) provide useful guidelines and they define the specific dimensions of a penetrometer for measurement of mechanical resistance. They suggest that once the penetrometer resistance reaches 2.4MPa root growth is largely restricted to existing pores, planes of weakness or cracks between soil structural units or peds. However as indicated by Hazelton and Murphy (2007), root growth can be severely restricted well before that value of resistance is reached. Cass (1999) provides further guidelines for interpreting mechanical resistance and recommends that mechanical resistance be measured at field capacity. Broad interpretations are:

- < 1 MPa at field capacity and drier – optimal conditions for root growth. Physical fertility optimum.
- 1 to 2 MPa at field capacity – seedling emergence retarded and may fail if cracks not present, root growth may be restricted as mechanical resistance likely to reach 3MPa as soils dry towards permanent wilting point. Soil physical fertility moderate.
- 2 to 3 MPa at field capacity – root growth impeded and restricted to old root channels and cracks. Water and nutrient uptake restricted. Physical fertility poor.

The growth of roots requires oxygen and soil air. Baver (1940) suggested that a minimum air porosity of 10% was required for root growth, and recommendations about the required air porosity require for root growth have not changed significantly since then, although the optimum has since been suggested as 15% (Dexter and Zoebisch 2002) and a specific requirement of at least 5% oxygen in the soil air.

2.2 Soil Structural Stability

Individual soil particles, be they clay, silt or sand, usually do not remain as individual particles in soils. They bond together by a range of mechanisms to form soil structural units or aggregates. The pore space created by these aggregates and the pore space between the aggregates is what is meant by soil structure (see Figure 2.1). Aggregate stability is the capacity of these aggregates to remain intact when exposed to stress imposed by wetting under tension, wetting under rainfall or tillage. Maintaining the aggregate structure is important for plant productivity as the aggregate structure can influence soil strength and mechanical resistance to emergence and root growth, aeration, surface crusting, erosion, infiltration, water holding capacity and bulk density. Although the direct relationship between aggregate stability and these functional soil properties is not always well established, aggregate stability is a general indicator of the physical fertility of the soil.

Unlike pH or soil strength or air porosity, aggregate is not a direct measure of a soil property that has direct impact on productivity. However it is a fundamental indicator of the physical health of the soil.

2.2.1 Aggregate hierarchy

Most soils have a hierarchy of aggregate structure. While several schemes exist for describing the aggregate hierarchy the simplest and perhaps most practical is that proposed by Tisdall and Oades (1982) and expanded by Cambardella (2002). In this scheme there are three levels of aggregates or soil structural units.

- Microaggregates 2 – 20 μm diameter formed by the flocculation of clays and some silt particles. Flocculation of the negatively charged clay particles is enhanced by the presence of exchangeable cations, especially calcium and greatly enhanced by the presence of trivalent aluminium cations. Organic matter may also enhance some flocculation and this organic matter present is primarily of microbial origin.
- Microaggregates 20 – 250 μm diameter formed around organic debris and the stability is maintained by decomposition products from the organic debris. It is thought that these aggregates are very stable and the organic materials bonding the aggregate together are somewhat resistant to decomposition.
- Macroaggregates > 250 μm diameter which consist of microaggregates, primary particles and particulate organic matter (POC) held together by a network of fine plant roots and mycorrhizae. The surface of the roots and mycorrhizae are covered with organic materials. The stability of these macroaggregates is related to the amount of POC and the presence and growth of the plant roots and mycorrhizae. Hence its stability is very much determined by soil management.

From this description it is apparent how important the different fractions of soil organic matter can be to defining the stability of soil aggregates.

A summary of the major bonding agents for soil aggregates is given in Table 2.1.

2.2.2 Soil organic matter and aggregate stability

The amount and type of organic matter in soil has a large influence on the aggregate stability. Krull et al. (2004) have done a major review of how soil organic matter influence aggregate stability. A few key points can be drawn from that review.

- The processes of aggregate stability can be different depending on whether actively growing plants are present. Aggregate stability increases more rapidly under actively growing plants, especially grasses with fine root systems than under the simple addition of organic materials to the soil (see Tisdall and Oades 1982, 1980a, 1980b; Lynch 1984; Ekwue 1990; Angers and Carter 1996; Jastrow 1996; Degens 1997: Piccolo et al 1997). Actively growing plants stabilise aggregates by binding soil particles with actively growing fine roots and with fungal hyphae that are dependent on the growth of the plant roots (Carter et al. 1994). Exudates from the roots and the hyphae provide mono and polysaccharides that enhance the binding of the aggregates. The limitation of this binding mechanism is that it is relatively unstable once plant growth ceases, although the stability is dependent on the nature of the plant materials. For example ryegrass residues remained effective for 4 to 6 months. Other plant materials with lower C:N ratios are more readily decomposed and can be expected to be more transient in their effects on aggregate stability. Simply adding organic materials requires a different set of processes and is much more reliant on the microbial populations becoming active and producing the biological binding agents such as microbial poly saccharides and humic materials, which takes more time.
- The processes and mechanisms of aggregate stability are complex and aggregate stability is not directly related to the level of soil carbon in the soil. As Jastrow (1996) and Angers and Carter (1996) show aggregate stability can increase much more rapidly than the level of soil carbon, especially in the presence of actively growing plants. Given the mechanisms that develop stable macroaggregates this is not unexpected, as the roots systems, fungal hyphae and bacterial populations can stabilise large soil aggregates with relatively small increases in soil carbon. Despite this general observation there are a few notes of caution in applying this conclusion.
    - The growth of roots and fungal hyphae and the addition of their exudates to the soil increase soil carbon levels. There is then a general trend for aggregate stability to increase with increasing levels of soil carbon. Some authors have suggested general critical values for aggregate stability including:
        - Kay and Angers (1999) – aggregate stability declines rapidly as SOC decreases from 1.5 to 1.2% . A level of 2.0% SOC is required to maintain aggregate stability.
        - Haynes (2000) – A maximum of mean weight diameter for aggregate stability occurs about 3.2 to 4.0% SOC.
        - Carter (1992) – 2.5% SOC is the minimum levels to maintain aggregate stability, 4.5% SOC gives a maximum aggregate stability
        - Chaney and Swift (1984) – indicated that aggregates stability continues to increase with increases in soil organic matter, does not reach a limit.
        - The different results and thresholds reflect the different soil types and probably the different methods used for measuring aggregate stability. However it is apparent that even though different soil organic matter

        fractions are important for aggregate stability, it is possible to define some general guidelines for overall SOC levels that are required to maintain minimum levels of aggregate stability.

    - The overall aggregate stability in the soil is very dependent on stable micro-aggregates. Having stable micro-aggregates is important to the development of stable macro-aggregates. In turn the formation of macro-aggregates is thought to play a major role in the development of the stable microaggregates as the macroaggregates provide an environment for the accumulation of the polysaccharides and humic materials to stabilise the microaggregates (Oades 1984). As the roots and hyphae decompose, the fragments become the centre of a macroaggregate and decomposition products interact with clay which encrusts the organic fragment to the extent that decomposition is retarded. The final product is a microaggregate that is bound by a matrix of humic materials. The microagregates are much more stable than the macroaggregates as the humic substances that form the binding agents are more chemically stable and resistant to decomposition but also are often physically protected from decomposition to some extent. This process is most effective in loam and clay soils (Oades 1993) which have significant amounts of clay particles.
- The different fractions of organic matter are responsible for the stability of different aggregate sizes. As a general rule the stability of macroaggregates (> 250μm) is associated with the living soil organic matter (phytomass including finer plant roots and fungal hyphae), particulate organic matter (POC) and the dissolved organic matter. The stability of the microaggregates (< 250μm to 20 μm) is associated with the humic fraction (See Table 2.1). The stability of even smaller microaggregates < 2μm to 20 μm is influenced by the humic fraction but is probably more determined by the relative percentages of the Ca, Mg, Na and K cations on the clay exchange sites and the nature of the clay minerals in the clay particles. While this division is not always so simple, it is a useful guideline for developing recommendations for land management practices to manage aggregate stability (Oades 1993).
- Adding organic materials to soils is an effective way to increase soil organic matter levels and aggregate stability. Results are complicated however and much depends on the nature of the organic materials added, the effects of the organic materials on plant growth and the nutrient levels of the organic materials and the soil. The reviews conducted by NSW Agriculture (Gibson et al. 2002) and Quilty and Cattle (2011) summarise some of the characteristics of organic waste materials that can be added to soils.
- Aggregate stability is not a soil property that can be readily assigned to soil functions as can be done for soil properties such as pH or infiltration or bulk density. Aggregate stability is important for its impact or effects on other functional soil properties. It is necessary to relate measures of aggregate stability to other functional soil properties such as bulk density, soil strength and water holding capacity.

2.2.3 Use of soil aggregate stability tests for dispersion and sodicity

Several aggregate dispersion tests have been advocated for testing aggregate stability and the stability of soil structure to wetting. Emerson (1967) proposed a classification scheme for soil structure stability based on the coherence of aggregates in distilled water. The scheme identified 8 classes. A full interpretation of these 8 classes is presented in Emerson (1991), as well as a slight modification to the procedure and definitions for classes 7 and 8 (see Table 2.2).

Interpretation of the different classes is as follows:

Classes 1 and 2
Soil materials in which there are unstable clays, usually as a consequence of substantial levels of exchangeable sodium being present on the exchange sites of the clays. Where surface soils show this instability an ameliorant such as gypsum is usually required to stabilise the soil.
Classes 3a and 3b
Indicate weakly structured soils that will show dispersion after raindrop impact and may form a surface crust. Soil structure is strongly influenced by soil organic matter levels. Emerson (1991) shows the effect of soil organic carbon on the bulk density of 5 mm cubes (Figure 2.2) for the surface soil of a red-brown earth. The occurrence of dispersion reduces structural soil porosity and increases bulk density. Emerson concluded that at a level of SOC of 1.2% no true soil aggregates are left and all structural porosity is lost.
Class 4
Carbonate or gypsum is present. This tends to stabilise soil aggregates to dispersion and slaking.
Class 5
Neutral to acid soils which are strongly aggregated by organic matter. There are generally no problems with crusting in these soils. Decline in levels of organic matter with time may move these soils into Class 3.
Class 6
Acid soils with high amounts of kaolin clay and often high amounts of iron oxides.
Class 7
Largely cemented soils with silicon, iron or manganese
Class 8
Very stable soils with very high levels of organic matter.

A caution provided by Emerson (1991) is that compaction by traffic can result in reduced porosity that cannot be predicted by aggregate stability tests, the presence of sodium or calcium ions and to some extent organic matter.

Subsequent aggregate tests such as Loveday and Pyle (Loveday and Pyle 1973) and the ASWAT test (Field et al. 1997) have concentrated on the use of aggregate tests to detect clay dispersion and the overall presence of sodicity in soils.

Figure 2.1. Soil structure. The arrangement of primary particles into aggregates to form porosity (Adapted from Geeves et al. 2007a).

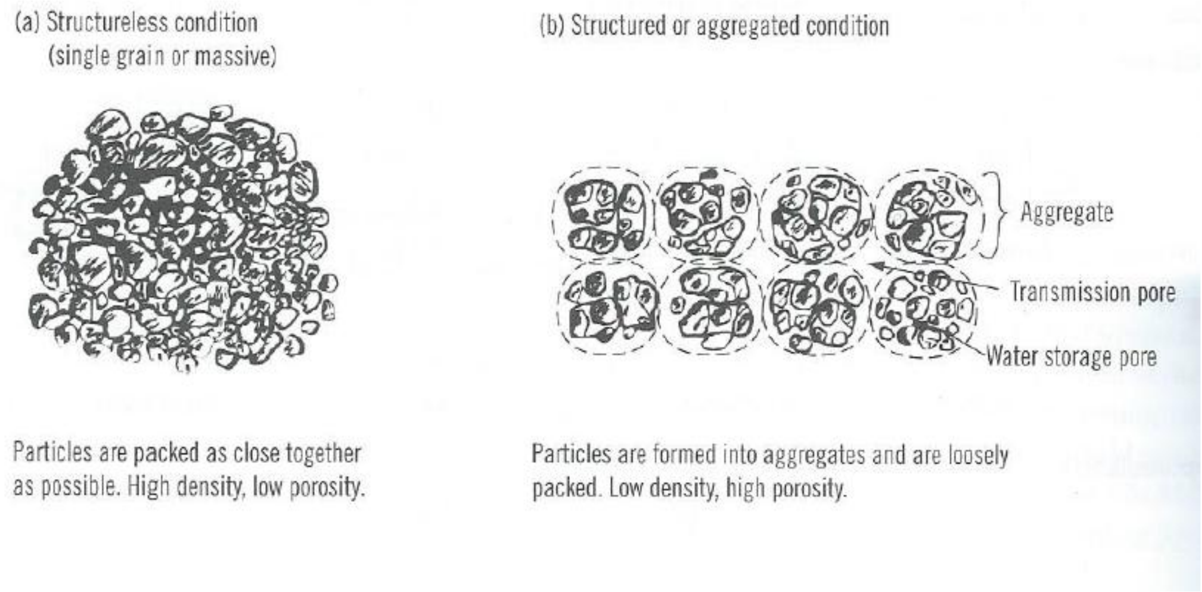


Table 2.1. Summary of bonding agents for aggregates in soils (Adapted from Cambardella 2002)

Aggregating Agent	**Aggregating process**	**Size of aggregates**
Humic substances	Strong bonds with mineral components of the soil	microaggregates
Polysaccharides	Gelatinous gluing agents	Micro and macroaggregates
Plant roots	Exude polysaccharides Enmesh soil microaggregates Form organo-mineral associations	Macroaggregate formation and short term binding
Fungal hyphae	Exude polysaccharides Enmesh soil microaggregates	Macroaggregate formation and short term binding
Earthworms	Mix organic matter and clay colloids Mix decaying organic material (particulate organic matter) and bulk soil	Macroaggregate formation

Figure 2.2. The bulk density of A horizons of red-brown earths near Tatura in relation to their organic carbon levels. The graph shows the bulk density of 5 mm cubes as they are remoulded and the relation to soil organic matter, showing effect of dispersion and aggregate stability. The graph also shows the effect of soil organic carbon on the bulk density of 75 mm cores. (This is Figure 2 from Emerson 1991).

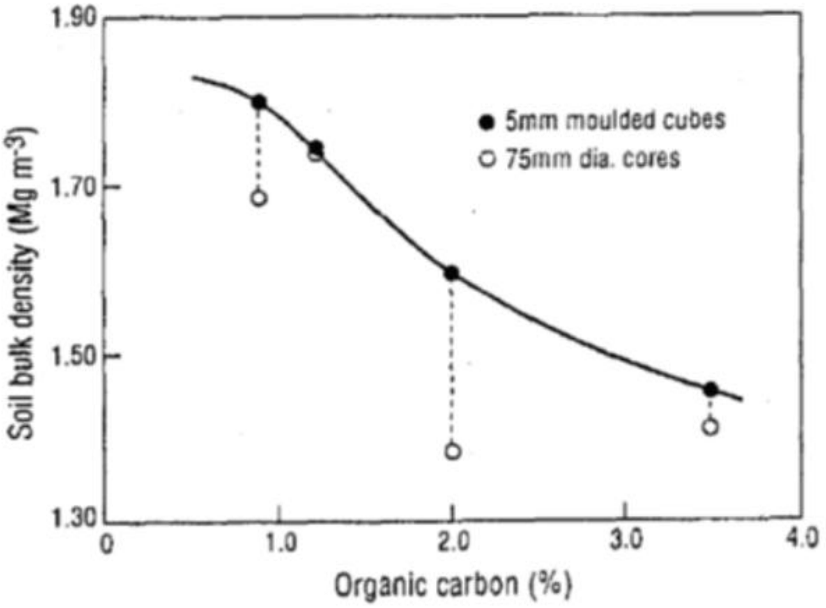


Table 2.2. Description of the Emerson Aggregate Test (Emerson 1967).

Step1

Air dry aggregate between 0.20 to 0.5 g or about 5mm in diameter is immersed in 100 ml distilled water. Stability is assessed at 24 hours.

Class 1

Aggregate slakes and shows severe dispersion with the majority of the aggregate breaking down into a cloud of dispersed clay. The cloud of dispersed clay covers most of the dish or beaker used for the test.

Class 2

Aggregate slakes and shows some clay dispersion of the aggregate. The cloud of dispersed clay covers part of the dish or beaker used for the test.

Step 2 – if no dispersion from Step1

A 5 mm cube of soil is wetted to field capacity and then physically worked and deformed. The aggregate is then immersed in 100ml of distilled water. Stability assessed after 24 hours.

Class 3a

Aggregate slakes and shows severe dispersion with the majority of the aggregate breaking down into a cloud of dispersed clay. The cloud of dispersed clay covers most of the dish or beaker used for the test.

Class 3b

Aggregate slakes and shows some clay dispersion of the aggregate. The cloud of dispersed clay covers part of the dish or beaker used for the test.

Class 4 to 6

Aggregate slakes but no dispersion.

Class 8

Aggregates do not slake

Step 3 - aggregates not showing any dispersion after physical working and immersion in distilled water
Is gypsum or lime present?

Class 4

Gypsum or lime present

Step 4 –Gypsum or lime not present
Soil is added to a 1:5 soil:water suspension and vigorously shaken for 10 minutes.

Class 5

Clay disperses

Class 6

Clay remains flocculated – usually with stabilisation from iron and aluminium oxides (Krasnozems, Ferrosols).

Class 7

Materials cemented by silica, iron or manganese – pan like materials.

2.3 Bulk Density – Soil Porosity

2.3.1 Definitions

The bulk density is the mass of dry soil per bulk volume of the soil. The mass of dry soil is determined by drying the soil to constant weight at 105^0C. The soil volume is determined by the volume of the total soil, usually by measuring the volume of a cylinder of soil sampled by a ring or by excavation. The term bulk density is used because the weight used in the measurement includes only the weight of the soil solids and does not include the weight of water or air within the soil. By using the bulk density, calculations can be made to determine the porosity of the soil. Low values mean a porous soil and high values a soil with low porosity.

2.3.2 Relationship to soil organic matter

Bulk density has a strong relationship with organic matter. Generally, the higher the level of organic matter, the lower the bulk density. Higher aggregate stability associated with higher levels of soil organic matter increases soil porosity which results in a lower bulk density. However bulk density is also affected by other soil properties such as soil texture, clay mineral type, sodicity and exchangeable cations, and the presence of iron and aluminium oxides. Land use history can also affect bulk density through cultivation, the time since cultivation and the amount of rain since the cultivation and compaction by stock or machinery. Despite these complications several relationships between bulk density and organic matter have been published. These were reviewed by Valzano et al. (2005) who identified the most useful ones for Australian soils. Tranter et al. (2007) also reviewed pedotransfer functions to predict bulk density and developed several for Australian soils. Shaykewich and Zwearich (1968) also developed some informative relationships between soil organic matter and bulk density for Canadian soils as shown in Figure 2.3. Unfortunately they used different particle size classes to those conventionally used in Australia.

Based on the reviews by Valzano et al. (2005) and Tranter et al. (2007) further investigation into the relationships between bulk density and soil organic matter will be based on the following relationships:

- Manrique and Jones (1991) A: Bulk density = 1.51 – 0.113 * soil organic carbon
- Manrique and Jones (1991) B: bulk density = 1.66 – 0.318 * $\sqrt{}$ (soil organic carbon)
- Merry in Valzano et al. (2005): Bulk density = 1.608 – 0.0872 * soil organic carbon
- Tranter et al. (2007):
    - ρ_b (soil bulk density) = $\rho_m + \Delta\rho$
        - Where ρ_b is the soil bulk density
        - ρ_m is the soil bulk density based on particle size and the mineral fraction
        - $\Delta\rho$ is the structure component associated with soil organic matter, tillage and sodicity
    - $\rho_m = 1.35 + 0.0045 * \text{sand } \% + 6*10^{-5} (44.7 - \text{sand } \%)^2 + 0.060*\log_e(\text{depth in cm})$
        - Sand is 20 to 2000μm
    - $\Delta\rho = - 0.217 - 0.114 * \log_e(\text{soil organic carbon}) - \log_e(\text{soil depth in cm})$
- Shaykewich and Zwarich (1968):
    - BD = 1.7756 – 0.0016vfs – 0.0017Si – 0.0047C – 0.0707 Soil OM – 0.0008*Soil OM*C
        - vfs is very fine sand, 0.1 to 0.05 mm; Si is silt 0.05 to 0.002 mm (US standard); C is clay, < 0.002 mm

Several relationships are used because it is possible different relationships are more effective for different ranges of soil organic matter and for different soil textures. For example Tranter et al. (2007) cautioned that their model fits best between the bulk densities of 1.2 and 1.6 t/m^3, and that it overestimates for bulk densities less than 1.2 t/m^3 and underestimates for bulk densities > 1.6 t/m^3. The results for the Tranter PTF are shown in Figure 2.4.

2.3.3 Changes in bulk density with rainfall

Tilled soil settles under the influence of rainfall and so bulk density changes with rainfall. Onstad et al. (1984) developed a mathematical description of how the bulk density changes with rainfall and the rate of change and the degree of settlement and final bulk density of the soil depends on the aggregate stability of the soil. This effect was quantified by Andales et al. (2000) with the equation:

$$BD_t = BD_{settled} + (BD_{till} - BD_{settled}) *\exp(-R*KE_{rain})$$

Where:

BD_t is the bulk density at time t after a given amount of rainfall kinetic energy,

$BD_{settled}$ is the bulk density after the soil is fully settled,

BD_{till} is the bulk density after the tillage or sowing operation,

R is the rate of change in bulk density with the kinetic energy of the rainfall, where R is given by the equation, $R = 5.0 * (1 - A)$, and A is given by the equation $A = 0.025 *SOC$,

KE_{rain} is the sum of the kinetic energy of the rainfall since the last tillage in J/m^2, and is given by the equation – $KE_{rain} = 0.00217 * Rain$, where Rain is given in mm.

The relationship is developed for the soils of Iowa in the Mid-West USA, which tend to be silt loams and loams derived from glacial parent materials. This does limit the applicability of the specific equation for Ausrtralian soils, but the principles remains valid.

2.3.4 Discussions of results

As a general rule the higher the levels of soil organic matter in a soil the higher the aggregate stability and the lower the bulk density. There has however much uncertainty and variation about this relationship because of the large number of factors that influence the bulk density of a soil. The following conclusions can be made.

1. Texture or particle size distribution is a principle determinant of the potential bulk density values of a soil. Prediction equations or pedotransfer functions that do not take texture into account will give incorrect predictions and usually appears largely for the loams and clay soils.
2. The derived pedotransfer function used (Tranter et al 2007) appears to predict bulk densities that are possibly too low for loam and clay soils. These predictions need to be compared to further published data on bulk densities.
3. When assessing the levels of soil organic matter expected in agricultural soils the predicted bulk densities are lower than those that are considered strongly limiting to plant productivity. This would imply that changing bulk density by increasing soil organic matter is not likely to improve plant productivity solely on the basis of reduced bulk density levels. Two cautions apply to this conclusion:
    a. Higher and limiting bulk densities than those predicted using texture and soil organic matter alone can occur because of compaction by machinery or stock. In these cases limiting values of bulk density can be reached even with the levels of soil organic matter present. Soil sodicity is also a severe factor and because of

dispersion can result in higher bulk densities than expected on texture and soil organic matter alone.

b. Many other physical properties not directly accounted for by a measurement of bulk density may still affect plant productivity. These include water holding capacity, soil strength, friability or tilth and susceptibility to erosion.

Based on the predicted values of bulk density there is a strong indication that bulk density will become too high at levels less than about 0.5% organic carbon which is 0.86% organic matter. Of course bulk density can still become limiting at higher levels of soil carbon when there is excessive compaction or high levels of sodicity.

Figure 2.3. Predicted bulk density for the pedotransfer function of Shaykewich and Zwarich (1968). Bulk density is affected by changes in clay and soil organic matter contents.

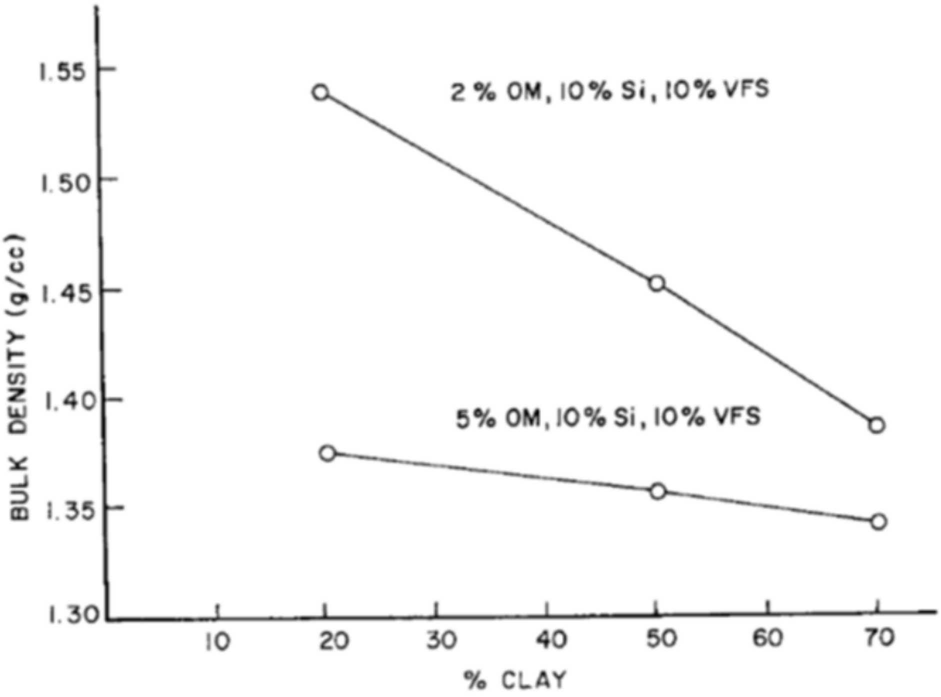
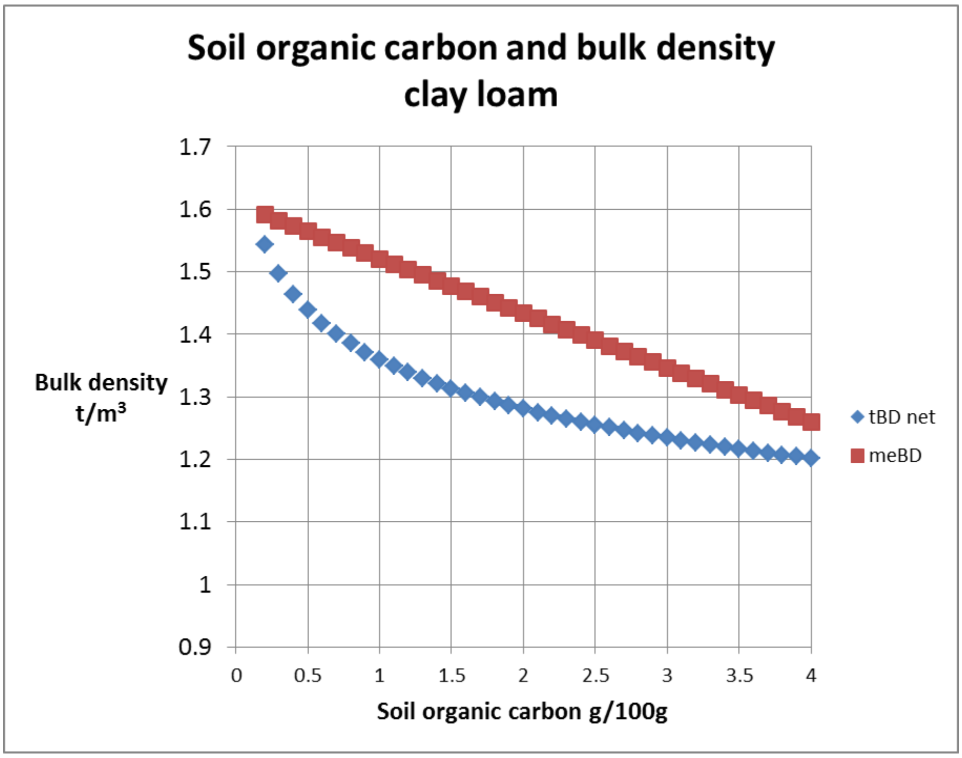


Figure 2.4 a - c. Predicted bulk density using two pedotransfer functions of Tranter et al. (2007) and Merry in Valzano et al. (2005). Bulk density is predicted for three soil textures.

a.



b.

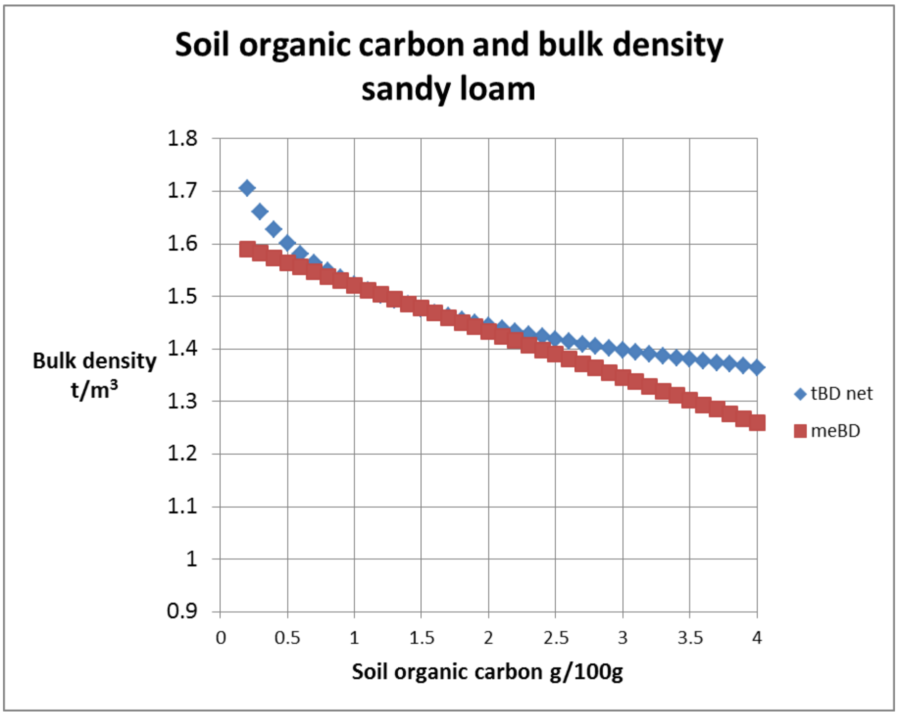
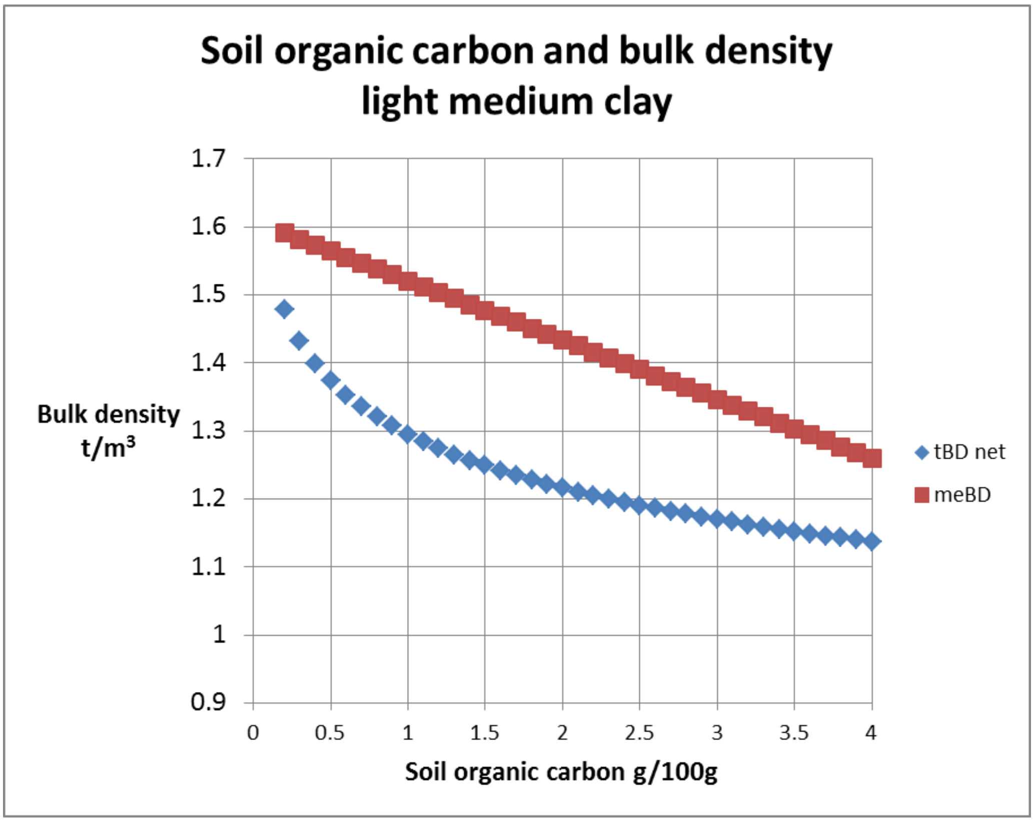


Figure 2.4 a - c (cont.). Predicted bulk density using two pedotransfer functions of Tranter et al. (2007) and Merry in Valzano et al. (2005). Bulk density is predicted for three soil textures.

c



2.4 Soil Hydraulic Properties

2.4.1 Basic definitions

Water holding capacity is an important part of the soil physical fertility as it determines the capacity of the soil to store and supply water for plant growth. Because of the links to soil porosity, there is a strong interaction between soil structure, aggregate stability, bulk density and water holding capacity (Krull et al 2004 p47, Emerson and McGarry 2003; Hudson 1994). Therefore discussions and evaluations about water holding capacity cannot be made independent of these other soil properties.

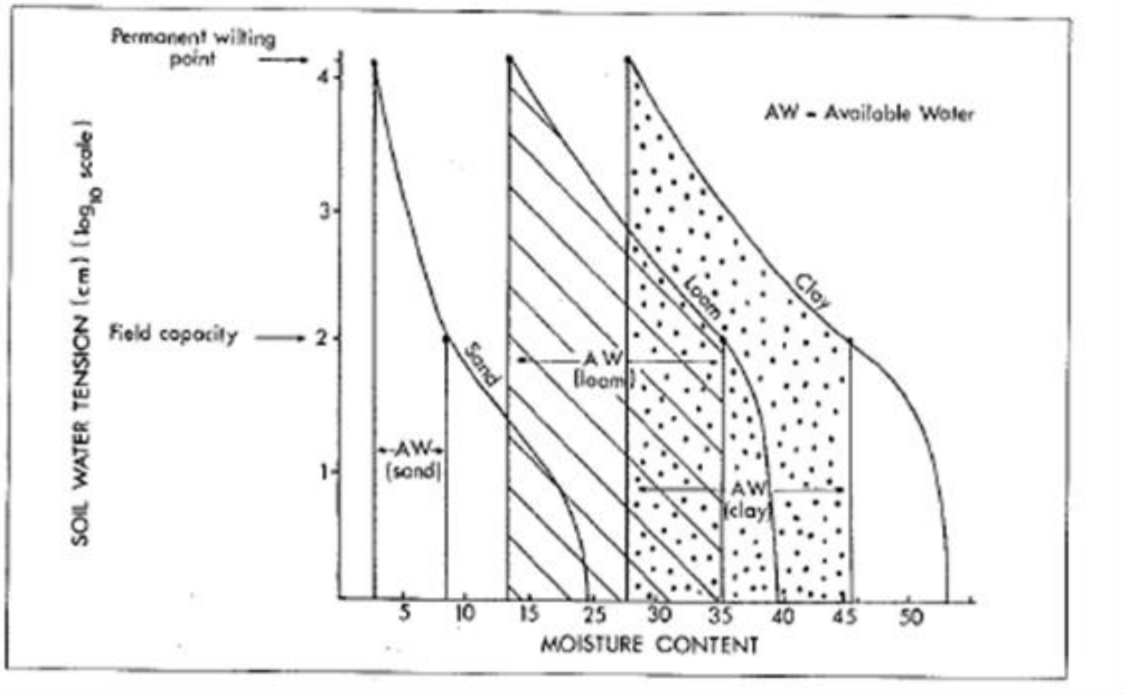
The most important aspect of water holding capacity is generally considered to be the plant available water (PAW) defined as the difference between the water in the soil held at field capacity (FC), and the water held at permanent wilting point (PWP). These two important components of water holding capacity are defined as:

1. Field capacity (FC) - If all soil pores are filled with water the soil is said to be saturated. There is no air left in the soil. After the drainage has stopped, the large soil pores are filled with both air and water while the smaller pores are still full of water. At this stage, the soil is said to be at field capacity. At field capacity, the water and air contents of the soil are sometimes considered to be ideal for crop growth, but this depends on the pore size distribution of the soil. The field capacity is the amount of water held in the soil after the gravitational water has drained away and after the rate of downward movement of water has materially decreased, which usually takes place within 2 to 3 days in pervious soils of uniform structure and texture (Veihmeyer and Hendrickson 1931). It is also referred to as the drained upper limit which is the highest, field-measured moisture content of a soil after thorough wetting and draining until the drainage rate becomes negligible. Field capacity corresponds to a matric tension in the range of 10 to 33 kPa (Gerakis and Ritchie 2002). For this report field capacity will be taken as 10 kPa.
2. Permanent wilting point (PWP) – Minimum amount of water held in the soil that plants can avoid wilting. If moisture decreases to this or any lower point a plant wilts and can no longer recover when placed in a saturated atmosphere for 12 hours. The physical definition of the wilting point is generally given as the water content at −1500 kPa of tension. It corresponds to the moisture content and tension at which plants can no longer extract moisture from the soil (Gerakis and Ritchie 2002). At permanent wilting point, water is only held in very small pores or on the surface of soil particles.

Both FC and PWP vary widely between soils and are determined by factors such as clay and clay mineral type, silt, fine sand, coarse sand, organic matter, cation characteristics and sesquioxide minerals. Soil organic matter has some potential to affect these basic water holding properties.

The water holding capacity of the soil can be completely defined by the moisture characteristic which shows how much water is held at a range of tensions from zero or saturation to oven dry or 10^{-6} kPa. Examples of moisture characteristics are shown in Figure 2.5 for sand, loam and clay soils. The effect of soil texture on the moisture characteristic is apparent. The moisture characteristic shows the pore size distribution of the soil as the tension required to drain a soil pore in the soil depends on its size. Clays have smaller pores and hold more water at higher tensions.

Figure 2.5. Relationship between soil water potential and water retention for different soil textures. The values shown are pF values and a value of 2 is 10kPa (field capacity) and 4.2 is 1500 kPa (permanent wilting point). (Source Geeves et al. 2007a).



2.4.2 Moisture content units

To clarify it is essential to be clear on how the moisture contents at FC, PWP and the different moisture tensions are defined. Moisture contents can be defines as below:

- Gravimetric moisture content –
    - gms of water / gm of soil or moisture % gravimetric
- Volumetric moisture content
    - cm^3 of water cm^3 of soil, equivalent to
    - mm of water /mm of soil, equivalent to 1/100 * mm of water / 10 cm of soil
- Volumetric moisture content = gravimetric moisture content * bulk density

Generally the preferable units are volumetric moisture content.

2.4.3 Pedotransfer Functions to predict the effects of soil organic matter on water holding capacity of soils

The soil hydraulic properties include the water contents at field capacity (FC), permanent wilting point and saturation (PWP), the available water (FC - PWP) and the hydraulic conductivities and infiltration are defined by the moisture characteristic. A large number of publications have established relationships between the soil water retention curve and water holding properties and basic soil properties such as soil texture but also soil organic carbon content. Concentration is given to those that established pedotransfer functions as these ensure a large data set has been utilised. The intended approach is to utilise these relationships to establish some general relationships between soil organic carbon and the hydraulic properties for the different soil texture classes and then test these against published relationships. The outcome is to establish some general relationships between soil organic carbon levels and the hydraulic properties with some confidence.

The summary of results for the different pedotransfer functions is presented in Figure 2.6 and Appendix 2. The advantage of using more general pedotransfer functions to estimate the effects of soil organic matter on soil water holding properties is that these give a more complete indication of the effects of soil organic matter rather than single one-off studies of limited data sets. The limited one-off data sets are useful to check the outcomes from the pedotransfer functions.

Several of the pedotransfer functions rely on the use of mathematical expressions to describe the soil water retention curve. The two that are most commonly used for Australian soils are the Campbell equation(Campbell 1974) and the van Genuchten equation (van Genuchten 1980) and Cresswell and Paydar (1996) have shown that these are applicable to Australian soils.

a. Kay et al 1997

The soil water retention curve or moisture characteristic can be approximated by various mathematical functions. One of the simplest is the Campbell equation:

$$\Psi = \Psi_e [\theta/\theta_s]^{-b}$$

Where Ψ is matric potential, Ψ_e is air entry potential, θ is the volumetric water content, θ_s is volumetric water content at saturation, b is a fitted constant.

This can be expressed as:

$\theta_v = a (\Psi m)^b$

Where θ_v is the volumetric moisture content, “a” is a constant to account for Ψ_e and θ_s.

Kay et al. 1997 in Krull et al (2004) has used this expression as a pedotransfer function to predict the moisture characteristics for a range of Canadian soils. It can also be used to estimate the impact of changes in soil carbon on the water holding properties of soils. Using Kay et al. established for the set of Canadian soils with a range of clay and sand contents that the values for the constants “a” and “b” in Equation X.2 are given by:

$$a = \exp[-4.1518 + 0.6851*\ln(clay\%) + 0.4025*\ln(soc\%) + 0.2731*\ln(BD)]$$

$$b = -0.5456 + 0.1127 * \ln(clay\%) +0.0233* \ln(SOC\%) + 0.1013 * \ln(BD)$$

Baldock (unpublished data) in Krull et al (2004) analysed some red-brown earth surface soils and showed significant increases in water holding capacity and PAW, especially for soil with less than 20% clay (Figure 23 in Krull et al 2004).

b. Hudson 1994

Hudson supplied a short review as well as analysing the effect of soil organic matter on water holding for soils with specific textures. This was to ensure that the effects of soil organic matter (SOM) were not confounded by soil texture effects. The soils used were from cropping lands in Midwest USA in Wisconsin, Minnesota and Kansas. The results clearly showed that the water held at FC increased more rapidly with increase with soil organic matter than increases in PWP (3.2% increase in FC per 1% increase in SOM for sand v 0.92 % increase in PWP with 1% increase in SOM). For silt loam the changes were 4.5 % increase in 1% increase in SOM for FC and 0.8 % increase in PWP for 1% increase in PWP. Hudson also showed that soil organic matter occupied 5 to 25% of the soil volume, even though the soil organic matter by weight was apparently less (1 to 6%).

c. Neuroman Calculations (Minasny and McBratney 2002; Minansy et al. 1999)

Another alternative for an equation to estimate the soil water retention curve is the van Genuchten equation (1980):

$\theta v = \theta r + (\theta s - \theta r) / [1 - (\alpha\Psi)^n)^m]$

Where θr is the residual volumetric water content, θs is the water content at saturation, m, n and α are empirical fitting coefficients and $m = 1 - 1/n$.

Minasny and McBratney (2002) have developed a pedotransfer function calculator to estimate soil hydraulic properties from basic soil properties including soil texture and bulk density, using the analysis published in Geoderma (Minasny et al. 1999). The pedotransfer function (PTF) uses Australian wide soil data and is based on analysis using the van Genuchten equation for the soil moisture characteristic.

Soil organic matter is not a direct input into the Neuroman calculator. However bulk density is a major input. The potential effect of organic matter on the water holding properties is estimated by using the bulk density pedotransfer function developed by Tranter et al. (2007), and the subsequent effect of the organic matter on the bulk density is used as a surrogate to predict the impact of soil organic matter on the hydraulic properties in the Neuroman calculator. The use of the Tranter PTF to predict the impact of soil organic matter on bulk density has been discussed in Section 2.3.

The pedotransfer function indicated a consistent increase in water holding properties associated with increasing organic matter although there was a tendency for the water holding properties to peak at about 3 to 4 % organic matter rather than continuing to increase. This could be an artefact of using the bulk density as a surrogate of the water holding capacity.

d. Williams et al. (1992)

Williams et al developed 6 functions to predict the water holding properties of soil from general soil properties. They utilised the Campbell equation to predict the moisture characteristic and used two Australian data sets to derive pedotransfer equations. Williams et al (1992) used the Prebble and Forest data sets of Australian soils to develop the pedotransfer function (PTF). Their Function 6 used soil organic matter and so is used to predict the impact of soil organic matter on water holding properties of soils.

Paydar and Cresswell (1996) confirmed the applicability of this (PTF) for Australian soils as well as providing as estimate of the uncertainty of using this PTF. Generally clays formed a large part of this data set. The equation used was:

$\ln \theta = A + B \ln(h)$

where for Function 6, A is given by:

A = 2.124 +0.1257* [ln (clay*clay)]-0.0051(fine sand) +0.0848(texture group) +0.00238*(organic matter*organic matter)-0.00398*silt

Where for Function 6 B is given by:

B = -0.292-0.00128*fine sand-0.0189*ln(organic matter)+0.0282*ln(texture group)+ 0.0627*ln(clay) - 0.00002*clay*clay.

e. Rawls et al. (1992)

Rawls et al developed as series of linear regression equations based on over 2500 soil horizons covering a wide range of soil textures and soil organic matter levels. The data was largely based on soils in the USA. The correlation coefficient of the relationships was generally > 0.80. This PTF was used to estimate the effect of soil organic matter on water retention.

The equations used by Rawls are:

At 10kPa:

θ = 0.4118 – 0.0030 * sand +0.0023*clay + 0.0317*organic matter

At 1500kPa:

θ = 0.0260 + 0.0050*clay + 0.0158*organic matter

f. Rawls et al. (2003)

Further analysis of the same data set above was undertaken and published in Rawls et al. (2003). Rawls et al. initially review studies on the effects of soil organic matter on soil water holding properties and they note that the studies are contradictory with about half the studies reporting that soil organic matter has no effect on soil water holding properties. In summary they concluded that the studies showing that soil organic matter had minimal effect on water holding properties usually involved small sample sets and often had soils with low levels and a small range of soil organic matter. Where studies had larger data sets from a wide range of regions, textures and soil organic matter levels, they tended to find that soil organic matter was an important predictor of water retention at a range of soil water potentials.

Rawls et al. went on to undertake an analysis of a large data set of 12, 000 samples from the National Soil Characterisation database from the Soil Survey Staff of the USA. Most of the soils were from Mollisols and Alfisols but included some Vertosols. Textures were mainly sandy loams, loams, silty loams and silty clay loams. Clays were less represented. Regression tree type analysis was undertaken on the data set for the soil water potentials of 33 kPa and 1500 kPa (see Figures 2.7 and 2.8).

The regression tree for the 33kPa soil water potential showed that the first node was based on texture with the sands, loamy sands and sandy loams being separated from the more clayey textures. The interesting result for the sands, loamy sands and sandy loams was that the next division was based on the soil organic carbon level being greater than or less than 2.1%. Where the soil organic carbon levels is greater than 2.1%, soil organic carbon continues to feature in predicting the moisture content at 33 kPa with critical levels at 4.8% and 7.7%. Where the soil organic carbon is less than 2.1% texture is used to predict the moisture at 33kPa with the division into sand and loamy sand and sandy loam. However the sandy loam soils are divided depending on whether the soil organic carbon levels are greater or less than 1.1%. The conclusion then is that based purely on the data set, soil organic carbon has a large effect on the moisture content at 33 kPa for the soils with sand, loamy sand and sandy loam textures.

For the loams to clay textures, texture has a primary effect on the amount of moisture held at 33 kPa. The regression tree does show that soil organic carbon has a modifying effect on the amount of water held at 33 kPa for each of the individual texture groups of loams, sandy, clay loams, sandy clays, silt loams, clay loams and silty clay loams. The regression tree does appear to have an effect on the water held in the clays, but as mentioned the number of clay samples was limited.

The regression tree for the water held at the soil water potential of 1500 kPa showed that the amount of clay dominated the prediction of the amount of water held at this water potential. Soil organic carbon only played a role where the clay content was < 14% (see Figure 2.8).

This study is important because it provides strong evidence for the effect of soil organic carbon and so soil organic matter on water holding properties of soil.

Several PTF's are proposed in this paper, but these are complex and have not been processed at this stage.

g. <u>Krull et al. (2004)</u>

Krull et al. (2004) showed that an increase of 1 gm of soil organic carbon /100 g of soil increased soil water holding capacity by the following amounts:

- Soils less than 10 % clay – 20 to 30% increase
- Soils 10 to 15 % clay – 10 to 25 % increase
- Soils 15 to 20 % clay – 10 to 18% increase
- Soils > 20% clay – about 10% increase or less.

h. <u>Publications with limited data sets.</u>

Several publications report on relationships between water holding properties and soil organic matter. Several of these such as Bauer and Black (1992) in Krull et al (2004), report on complex and uncertain relationships between water holding and soil organic matter and some suggest soil organic matter has no effect on water holding properties. However, as discussed by Hudson (1994), often individual studies have a limited range of textures or factors such as soil compaction confound the relationships between soil organic matter and water holding properties. Also adding organic materials is not necessarily a guaranteed method to increase water holding using soil organic matter (see P52, Krull), as it does not necessary result in increases in the effective soil organic matter fractions. This highlights the importance of knowing about the fractions of the soil organic matter to predict the effects on soil properties.

Other pedotransfer functions not used include Vereecken et al (1989) and Wostern (1999).

Several authors have reported no effects of soil organic matter on water holding capacity. Lal (1979) and Danalatos et al. (1994) reported no effects and Thomasson and Carter (1989) and Loveland and Webb (2003) reported that soil organic matter only accounted for a small part of the variation in the water holding capacity of soils. The effects of soil organic matter on soil water holding capacity can be complex including the following:

- Soil organic matter can reduce bulk density which can affect the water holding capacity of the soil. The reduced bulk density may tend to decrease the overall volumetric moisture content of the soil as the volumetric moisture content is calculated as the gravimetric moisture content x bulk density. The effect of soil organic matter on water holding capacity may need to take account of the effect on bulk density.
- The composition of the soil organic can be important in determining the water holding capacity. If the soil organic matter has a large proportion of particulate organic matter or

litter as can happen as the soil organic matter levels increase, this organic fraction may not contribute greatly to the water holding capacity of the soil.

- In soils with lower sand content, the clay and silt contents tend to dominate the water holding capacities. This can be seen especially in Tables 2.3 and 2.4 where the sandy soils the water holding capacity tends to increase by about 3 mm/10 cm of soil for every increase of 1 gm of soil carbon per 100 gm of soil. In the clay soil the increase is less than 0.5 to 1 mm /100gm of soil.
- The water holding capacity of the soil is only one factor in determining how much water is available to the plant, crop or pasture. The other factors are:
    - The infiltration rate at the soil surface and the capacity of the soil to allow rainfall to infiltrate and redistribute through the soil.
    - The suitability of the soil for root growth so that plants are able to grow with few limitations and explore and exploit the maximum volume of soil to utilise water and nutrients stored in the soil.
    - The management of surface cover has a large effect on the hydraulic properties of the soil and on the infiltration and amount of water that will infiltrate into a soil. While this related to erosion control, maintaining surface cover also greatly increases the amount of water that infiltrates into the soil and so increases the amount of water available to plants (Bot and Benites 2005).

2.4.4 Summary of results for organic matter and water holding

The results of the calculations of the effects of soil organic matter on water retention for the pedotransfer functions (PTF's) are presented in Tables 2.3, 2.4, Figure 2.6 and Appendix 2. The levels of soil organic carbon presented are from 0.7% to 3.0 % as this is considered to be general operational level of soil organic carbon in most agricultural soils under dryland agriculture. The major features of the results are outlined below. Note that discussion is based on soil organic carbon rather than soil organic matter because soil organic carbon is the basis for the measurements and calculations.

- The PTF's consistently indicate that the higher levels of soil organic carbon increase the water holding capacity of the soil at field capacity or the drained upper limit, and the plant available water. Some, but not all, indicate that increasing soil organic carbon will also increase the water retained at permanent wilting point or the lower limit. This supports the conclusion in Rawls et al (2003) of the effects of soil organic carbon on water retention.
- The amount of the increase in water holding capacity shows some variation with texture, but this is not consistent across the PTF's. Generally, it appears the loamy sands and sandy loams have a greater potential to increase water holding potential with increases in soil organic carbon, but this is not constantly shown by all PTF's.
- The Neuroman PTF appears to underestimate the increase in water holding capacity of the soil with increasing soil organic carbon. The mechanism of the increased water holding capacity in the Neuroman PTF is based solely on increased porosity associated with a reduction in bulk density. As discussed in Rawls et al (2003), this can underestimate the change in water holding capacity as it takes no account of the capacity of the organic matter itself to retain water. Depending solely on changes in bulk density may under estimate the effects on soil water water-holding capacity.

- The Williams 6 PTF appears to overestimate the water holding capacity of the coarser texture soils, especially the loamy sands. This is maybe because of a lack of these soils in the data set used to develop the PTF. This is despite the conclusion of Cresswell and Payder (1996) that this PTF is suitable for Australian soils. Of course this probably reflects the lack of such coarse textured soil in both data sets.
- Based on a summary of average values, the rate of increase in plant available water with increased soil organic carbon appears to be is relatively consistent across textures. Based on the values in Tables 2.4 and 2.5 the plant available water increases by about 2 to 3 mm/10cm of soil for each 1.0% increase in soil organic carbon over the range of 0.7% to 3.0% soil carbon. Above 3.0% soil carbon the increase will be less as the effect of soil organic carbon on water retention diminishes. Further work is required to confirm this, but this appears to be the order of magnitude of the effect of soil organic matter on water retention.
- It can be hypothesised on the basis of these preliminary results, that perhaps 0.5 mm/10 cm increase is a result of improved porosity and the remainder to the moisture holding properties of soil organic matter itself. This is a conclusion from the Neuroman PTF, which is based largely on changes in bulk density. Further investigation is required to confirm this.
- These levels are generally consistent with the tree regression of Rawls et al (2003), but further evaluation is required.
- In terms of yield effects, assuming a water use efficiency of 10kg/ha/mm of rainfall utilised, an extra 2 to 2.5 mm of rainfall stored can result in 20 to 25kg /ha of yield. However, this may occur several times during the year, or may occur at the critical flowering or grain-fill stage of growth. Therefore yield effects of the extra moisture storage are not that easy to predict.

2.4.5 Effect of Soil Organic Matter on Profile Available Water

Increases in soil organic matter are likely to be concentrated in the surface soils and usually in the top 10 cms and unlikely to extend beyond 20 cms. This limits the capacity of increases in soil organic matter to increase total available water in the soil profile. In Table 2.6 the typical effect of increasing soil organic matter on the profile available water shows that the overall effect on total profile available water is marginal because the increases in soil organic matter are largely confined to the surface soils. The increases in available water is about 5 mm, which may be important at sowing in some years or at anthesis or grain-fill in some critical years, but overall the effect is likely to be marginal. To fully test the impact of the increased available water, a full modelling program is required.

2.4.6 Potential Effects of Soil Organic Carbon on Infiltration

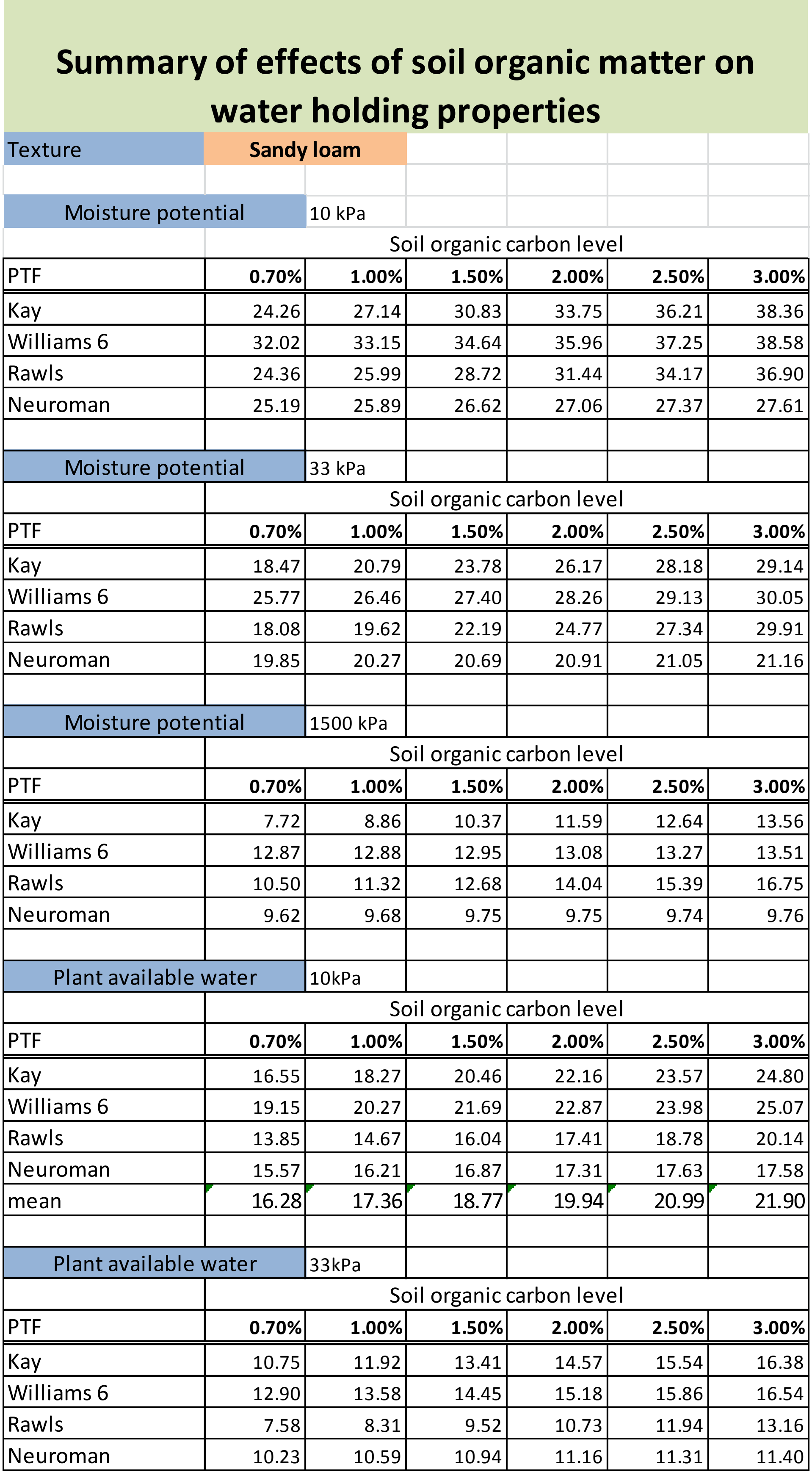
Soil organic matter, by providing aggregate stability to wetting and raindrop impact has the capability of maintaining the infiltration capacity of the soil and preventing surface crusting. The amount of runoff is then minimised, especially under higher intensity rainfall. Reducing runoff has the dual advantages of minimising water erosion and runoff, and ensuring the rainfall enters the soil to be stored for crop and pasture growth. In a study on the Ferrosols of eastern Australia, Bell et al. (1999) demonstrated a clear link between runoff under storms of different intensity and the soil organic matter levels. They measured total carbon (TOC) using LECO but also the “active carbon fraction" (C1) which was measured using 33mM $KmnO_4$. On analysis the total carbon was related to the active carbon fraction by the equation:

$$C1\ (g/kg) = 0.1071 * TOC(g/kg) – 0.5589 \qquad (r^2 = 0.9626)$$

The level of C1 was shown to be highly correlated to the percentage weight of aggregates < 0.125 mm (P125), which was in turn related to the final infiltration rate of the soil under rainfall simulation under a rainfall of different intensities (FIR). The critical outcome was that Bell et al (1999) were able to relate FIR to the C1 value. Using the rainfall intensity records for eastern Australia and using the 30 minute rainfall intensity as a standard with return periods of 1 year and 10 years, they were able to make recommendations about the levels of soil carbon required to minimise runoff from intense rainfall. For northern Tasmania they concluded that levels of 0.79 to 1.32 g/kg of C1 were sufficient to prevent most runoff. Storm intensities were only 19mm/h for the 1 year return period and 40 mm/h for the 10 year return period. This contrasts with Innisfail where the corresponding rainfall intensities are 113 and 139 mm/h and the required levels of C1 are 3.19 and 3.85 g/kg. Using the equation above, these values convert to TOC values of 1.26 to 1.75 g/100g for northern Tasmania and 3.50 to 4.12 g/100g for Innisfail.

It is possible to apply the same principles to locations such as Cowra and Gunnedah in NSW, where the corresponding 1 year and 10 year 30 minute rainfall intensities are 28 and 52 mm/h and 30 and 60 mm/h respectively (www.bom.gov.au). Based on Table 3 in Bell et al. (1999) and the above relationship between C1 and TOC, this corresponds to TOC values of about 1.10 and 1.78 g/100g for Cowra and 1.16 and 2.01 g/100g for Gunnedah with the higher value being that required to prevent runoff for a 30 minute rainfall intensity with a 1 in 10 year return period. The interesting outcome is that the general value of 2.00g/100g of TOC being a general guideline for a limiting value for a soil function appears again. The limitation of this conclusion is that it applies to Ferrosols and not necessarily to other soil types.

Table 2.3 a. Results of pedotransfer function calculations for 2 textures.



Summary of effects of soil organic matter on water holding properties

Texture	**Sandy loam**

Moisture potential	10 kPa				
	Soil organic carbon level					

PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	24.26	27.14	30.83	33.75	36.21	38.36
Williams 6	32.02	33.15	34.64	35.96	37.25	38.58
Rawls	24.36	25.99	28.72	31.44	34.17	36.90
Neuroman	25.19	25.89	26.62	27.06	27.37	27.61

Moisture potential	33 kPa
	Soil organic carbon level

PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	18.47	20.79	23.78	26.17	28.18	29.14
Williams 6	25.77	26.46	27.40	28.26	29.13	30.05
Rawls	18.08	19.62	22.19	24.77	27.34	29.91
Neuroman	19.85	20.27	20.69	20.91	21.05	21.16

Moisture potential	1500 kPa
	Soil organic carbon level

PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	7.72	8.86	10.37	11.59	12.64	13.56
Williams 6	12.87	12.88	12.95	13.08	13.27	13.51
Rawls	10.50	11.32	12.68	14.04	15.39	16.75
Neuroman	9.62	9.68	9.75	9.75	9.74	9.76

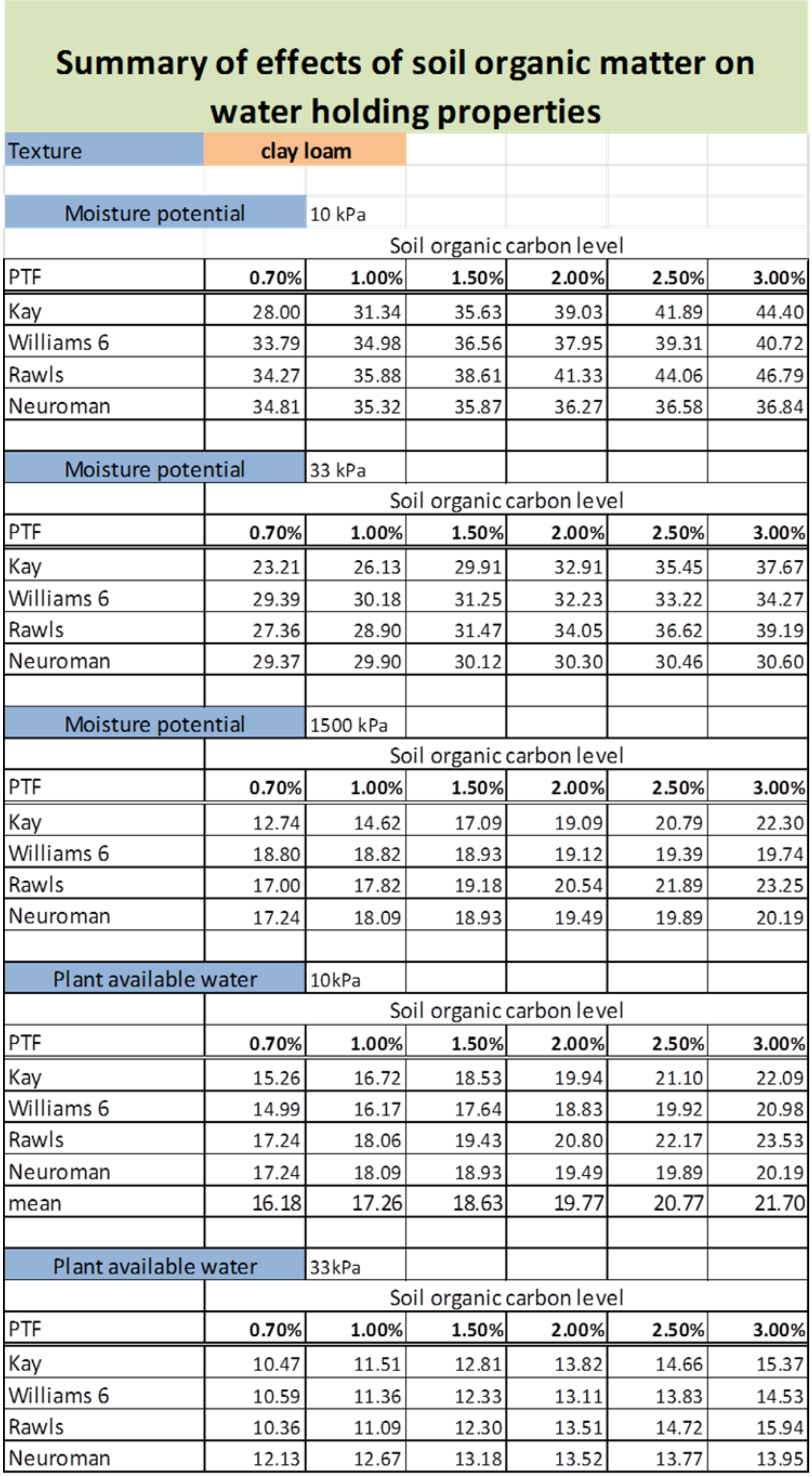
Plant available water	10kPa
	Soil organic carbon level

PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	16.55	18.27	20.46	22.16	23.57	24.80
Williams 6	19.15	20.27	21.69	22.87	23.98	25.07
Rawls	13.85	14.67	16.04	17.41	18.78	20.14
Neuroman	15.57	16.21	16.87	17.31	17.63	17.58
mean	16.28	17.36	18.77	19.94	20.99	21.90

Plant available water	33kPa
	Soil organic carbon level

PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	10.75	11.92	13.41	14.57	15.54	16.38
Williams 6	12.90	13.58	14.45	15.18	15.86	16.54
Rawls	7.58	8.31	9.52	10.73	11.94	13.16
Neuroman	10.23	10.59	10.94	11.16	11.31	11.40

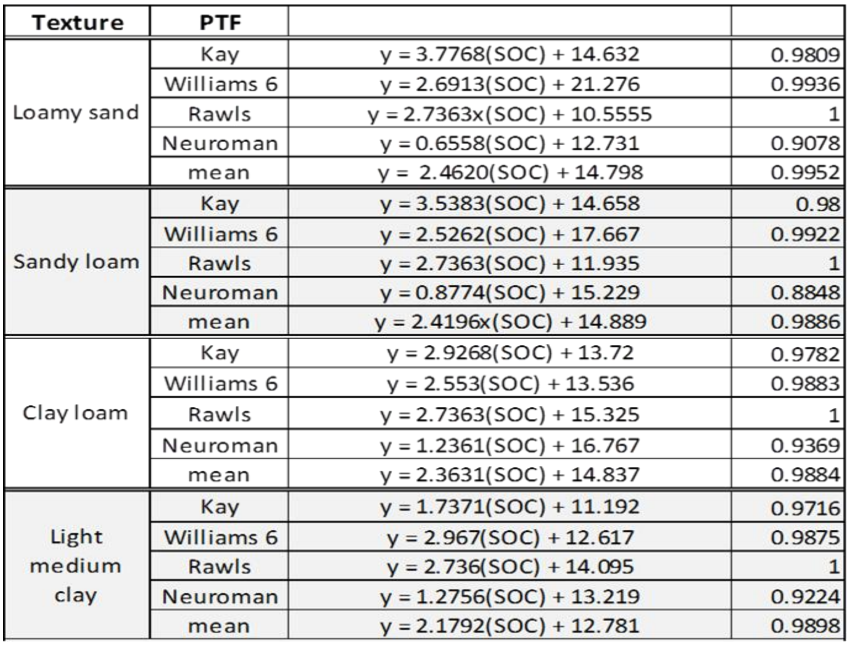
Table 2.3b. Results of pedotransfer function calculations for 2 textures. See Appendix 2 for further textures.



Summary of effects of soil organic matter on water holding properties

Texture	clay loam					
Moisture potential	10 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	28.00	31.34	35.63	39.03	41.89	44.40
Williams 6	33.79	34.98	36.56	37.95	39.31	40.72
Rawls	34.27	35.88	38.61	41.33	44.06	46.79
Neuroman	34.81	35.32	35.87	36.27	36.58	36.84
Moisture potential	33 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	23.21	26.13	29.91	32.91	35.45	37.67
Williams 6	29.39	30.18	31.25	32.23	33.22	34.27
Rawls	27.36	28.90	31.47	34.05	36.62	39.19
Neuroman	29.37	29.90	30.12	30.30	30.46	30.60
Moisture potential	1500 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	12.74	14.62	17.09	19.09	20.79	22.30
Williams 6	18.80	18.82	18.93	19.12	19.39	19.74
Rawls	17.00	17.82	19.18	20.54	21.89	23.25
Neuroman	17.24	18.09	18.93	19.49	19.89	20.19
Plant available water	10kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	15.26	16.72	18.53	19.94	21.10	22.09
Williams 6	14.99	16.17	17.64	18.83	19.92	20.98
Rawls	17.24	18.06	19.43	20.80	22.17	23.53
Neuroman	17.24	18.09	18.93	19.49	19.89	20.19
mean	16.18	17.26	18.63	19.77	20.77	21.70
Plant available water	33kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	10.47	11.51	12.81	13.82	14.66	15.37
Williams 6	10.59	11.36	12.33	13.11	13.83	14.53
Rawls	10.36	11.09	12.30	13.51	14.72	15.94
Neuroman	12.13	12.67	13.18	13.52	13.77	13.95

Table 2.4. Relationships for PTFS of different textures. These data show the relationship between plant available water from 10kPa to 1500kPa for each of the pedotransfer relationships.



Texture	PTF		
Loamy sand	Kay	y = 3.7768(SOC) + 14.632	0.9809
	Williams 6	y = 2.6913(SOC) + 21.276	0.9936
	Rawls	y = 2.7363x(SOC) + 10.5555	1
	Neuroman	y = 0.6558(SOC) + 12.731	0.9078
	mean	y = 2.4620(SOC) + 14.798	0.9952
Sandy loam	Kay	y = 3.5383(SOC) + 14.658	0.98
	Williams 6	y = 2.5262(SOC) + 17.667	0.9922
	Rawls	y = 2.7363(SOC) + 11.935	1
	Neuroman	y = 0.8774(SOC) + 15.229	0.8848
	mean	y = 2.4196x(SOC) + 14.889	0.9886
Clay loam	Kay	y = 2.9268(SOC) + 13.72	0.9782
	Williams 6	y = 2.553(SOC) + 13.536	0.9883
	Rawls	y = 2.7363(SOC) + 15.325	1
	Neuroman	y = 1.2361(SOC) + 16.767	0.9369
	mean	y = 2.3631(SOC) + 14.837	0.9884
Light medium clay	Kay	y = 1.7371(SOC) + 11.192	0.9716
	Williams 6	y = 2.967(SOC) + 12.617	0.9875
	Rawls	y = 2.736(SOC) + 14.095	1
	Neuroman	y = 1.2756(SOC) + 13.219	0.9224
	mean	y = 2.1792(SOC) + 12.781	0.9898

Table 2.5. Relationships for pedotransfer functions (PTF) of different textures. These data show the relationship between plant available water from 10kPa to 1500kPa for the Kay pedotransfer relationship based on constant bulk density.

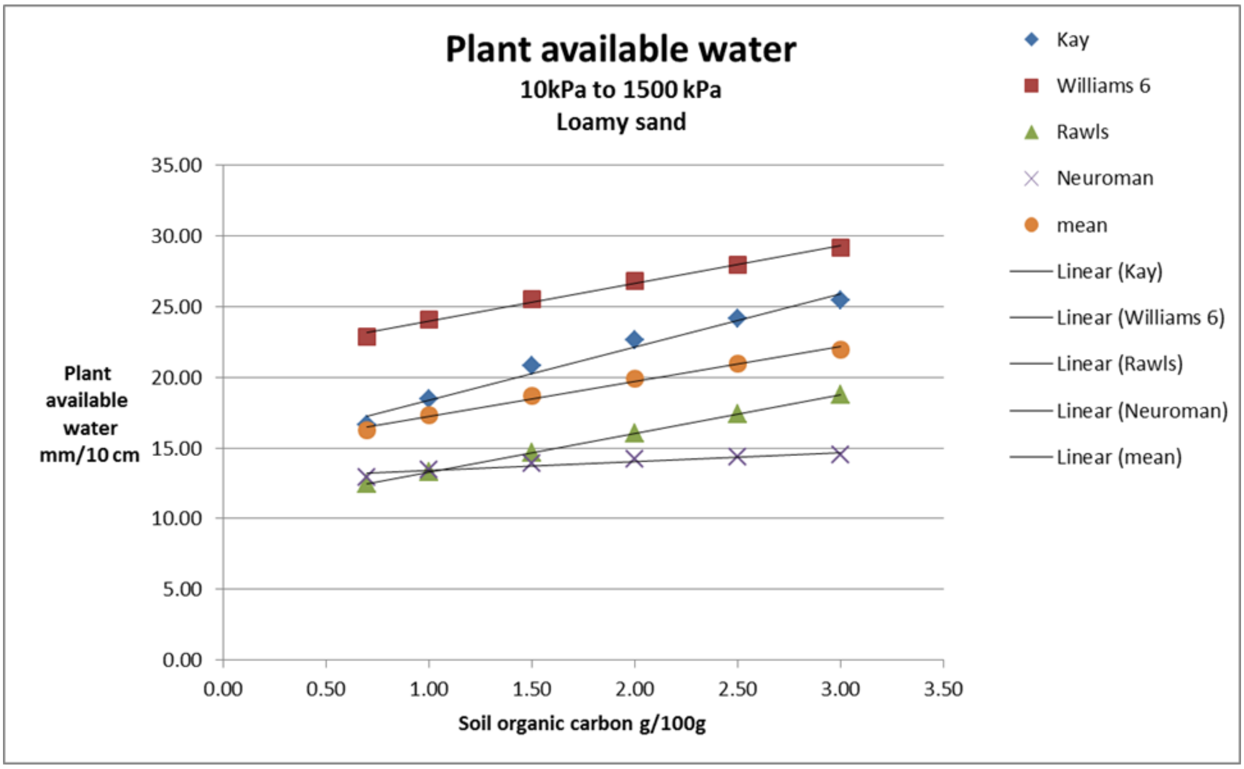
Texture	**PTF**	**Using constant bulk density**	Change in PAW mm/10 cm/1% increase in soil organic carbon
Loamy sand	Kay	y = 3.3542(SOC) + 14.906	3.35 mm/10 cm
Sandy loam	Kay	y = 2.9258(SOC) + 14.878	2.93 mm /10 cm
Clay loam	Kay	y = 1.811(SOC) + 13.49	1.81 mm/10cm
Light medium clay	Kay	y = 0.4182(SOC) + 11.456	0.42 mm/10cm

Table 2.6. Changes in profile available water associated with changes in soil organic matter. Profile from Wellington NSW, Red Chromosol under wheat crop (data from Geeves et al. 1995).

Layer	texture	colour	pH	Total soil organic carbon (g/100g)	Water holding mm/layer Based on Kay in Table 2.5			
					Original	TOC 1.5	TOC 2.0	TOC2.5
0 – 10cm	sandy loam	light reddish brown	5.07	0.94	17.63	19.27	20.73	22.19
10 – 22 cm	light medium clay	reddish brown	5.81	0.65	14.07	14.07	14.07	14.07
22 – 51 cm	light medium clay	brown	-	-	33.76	33.76	33.76	33.76
51 – 76	light medium clay	yellowish brown	6.7	0.23	28.90	28.90	28.90	28.90
Total available water (mm)					94.36	96.00	97.46	98.92

Figure 2.6 a – d. Results of pedotransfer function calculations for different textures

a.



b.

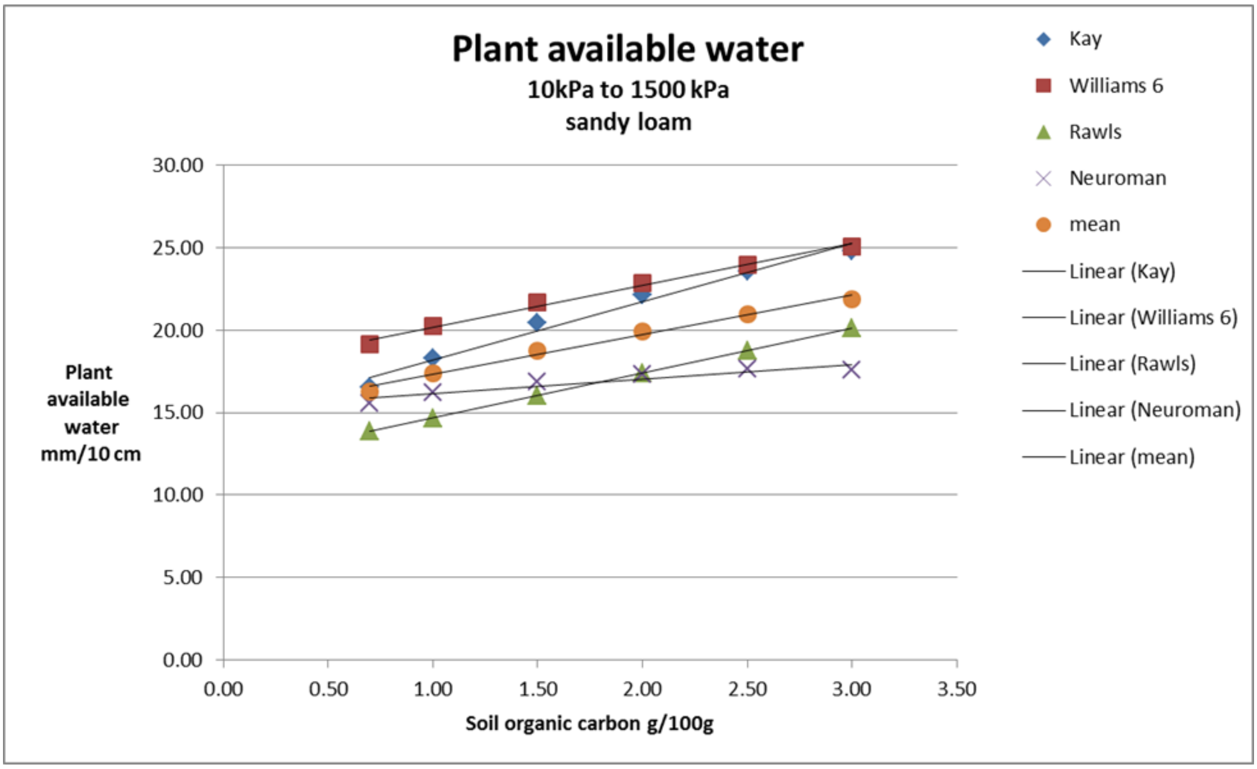
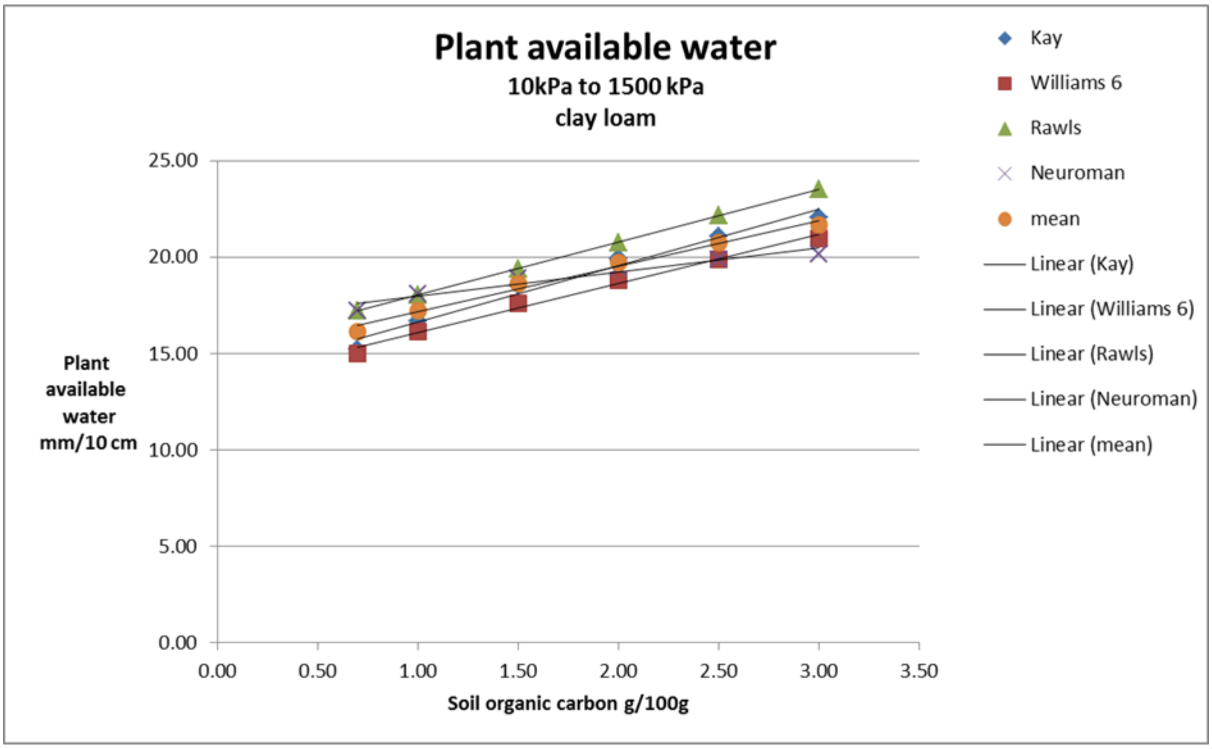


Figure 2.6 a – d. Results of PTFs for different textures

c.



d.

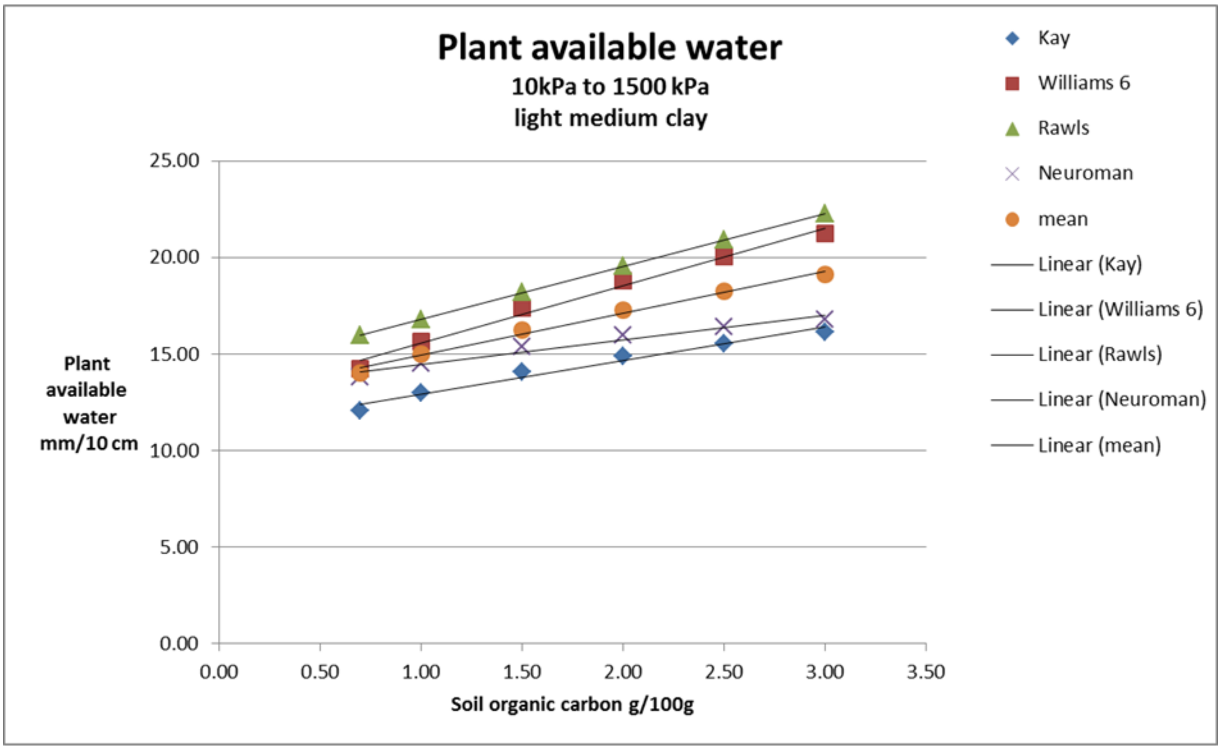


Figure 2.7. Regression tree for Soil Water Potential at -33kPa. (Rawls et al. 2003) . Textural class and soil organic carbon (C_{org}) used as predictors. The group or number node is shown in brackets , the average volumetric water content in the group is shown below the node number, the standard deviation within the group is shown in parentheses, and the bottom number is the count of samples in the group.
(With permission from Geoderma)

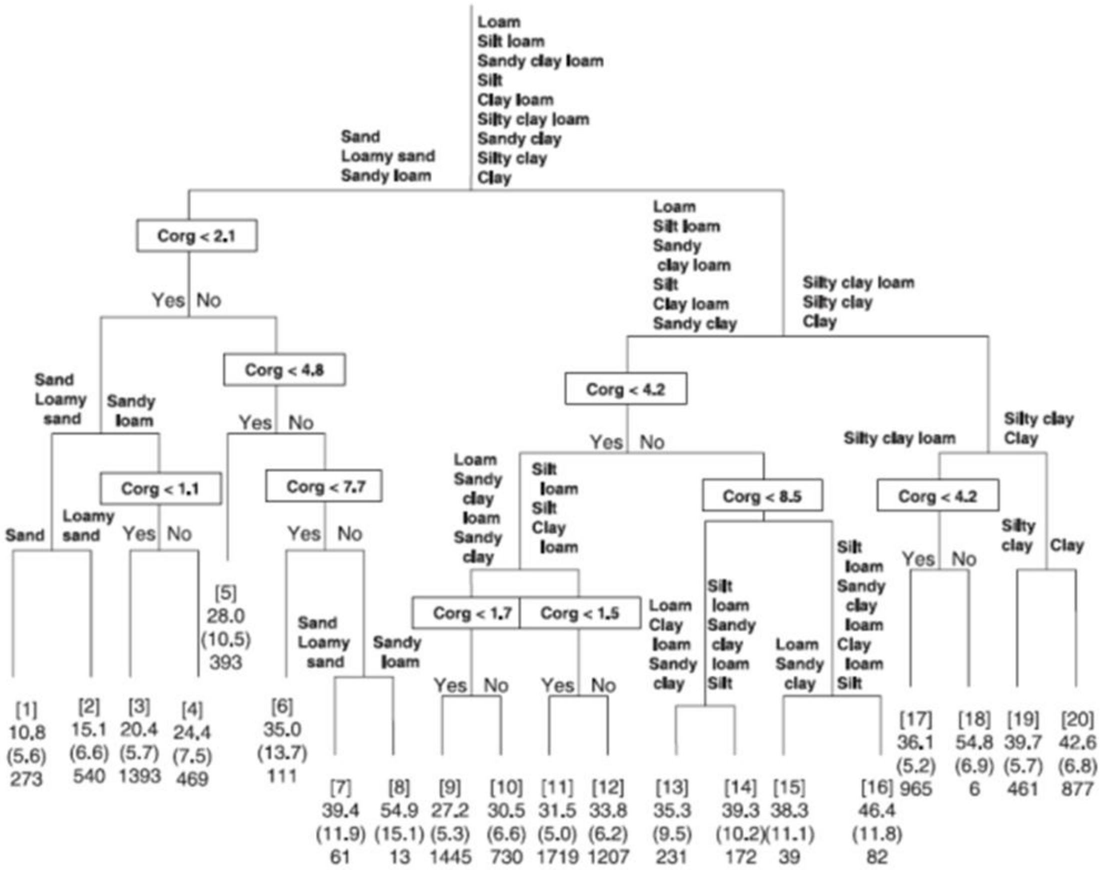
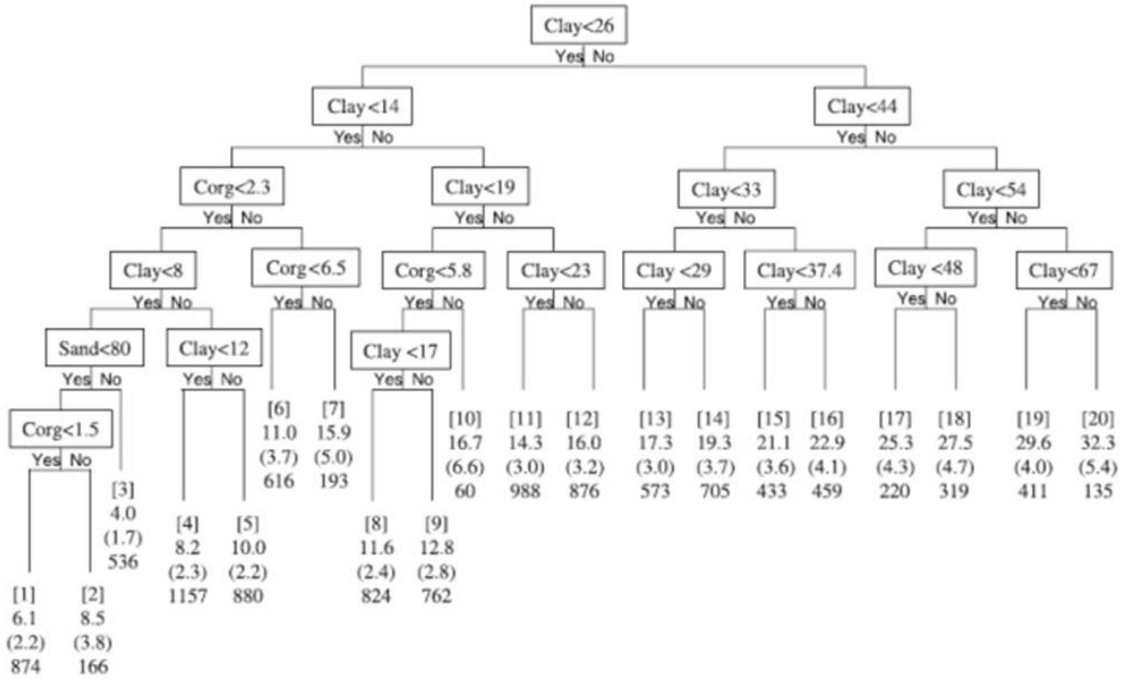


Figure 2.8. Regression tree for Soil Water Potential at -1500kPa. (Rawls et al. 2003) . Textural class and soil organic carbon (Corg) used as predictors. The group or number node is shown in brackets , the average volumetric water content in the group is shown below the node number, the standard deviation within the group is shown in parentheses, and the bottom number is the count of samples in the group.
(With permission from Geoderma)



2.5 Soil Thermal Properties

Soil organic matter can alter the thermal properties of soil because it is usually dark and black. The practical overall effect needs to be evaluated. By potentially increasing the dark colour of the soil, the albedo of the soil is reduced and so increases the potential for the soil to absorb more heat. In the USDA erosion prediction model WEPP (Flanagan and Nearing 1995), the suggested model is that the soil albedo is given by the equation:

$$\text{Soil Albedo} = 0.6/\exp(0.4 * \text{soil organic matter})$$

This results in values of about 0.6 for soils with very little organic matter and of 0.08 for soils with 5% organic matter. This equation has been developed for soils in the United States and it is unlikely that it would account for the Black Vertosols in northern NSW and southern Queensland. Based on this equation a level of soil organic matter of 2.5 to 3 % would be required to bring the soil albedo down significantly. This corresponds to soil organic carbon levels of 1.5 to 2%.

Nevertheless, it does give an indication of the potential for soil organic matter to reduce the albedo of soil surfaces and increase the amount of heart absorbed. This may be important for germination and early growth for some crops or be a disadvantage if it enhances evaporation from the soil in hot dry weather.

The actual thermal heat properties of soil responsible for the storage and flow of heat through the soil are also affected to some degree by soil organic matter (Abu-hamdeh and Reeder 2000 in Krull et al 2004). The heat storage and flow properties are summarised in Table 2.6. Soil organic matter has substantially different physical properties to the other soil constituents so as the proportion of soil organic matter increases; there is the potential to change the thermal properties of the soil. Soil organic matter can also affect the bulk density which will affect the heat capacity and conductivity of the soil. Generally increasing the soil organic matter reduces the thermal conductivity of the soil and so the amount of heat that flows through the soil (Abu-hamdeh and Reeder 2000).

An important conclusion supported by Krull et al. (2004) was that moist soils have a higher heat capacity (see Table 2.7) and so require a lot more heat to raise the temperature, irrespective of the colour of the soil. Wet soil can also transfer heat more rapidly. Generally the soil moisture content has a larger effect on the soil thermal properties than soil organic matter.

Table 2.7. Thermal properties of soil constituents. Marshal and Holmes (1979)

	Specific heat capacity	Density	Volumetric heat capacity	Thermal conductivity
	kJ/kg/^{0}C	Kg/m^3	kJ/m^3/^{0}C	W/m/^{0}C
Air 20 ^{o}C	1	1.2	1.2	0.025
Water	4.2	1.0 x 10^3	4.2 x 10^3	0.6
Quartz	0.8	2.7 x 10^3	2 x 10^3	8.8
Clay minerals	0.8	2.7 x 10^3	2 x 10^3	2.9
Soil organic matter	2.5	1.1x 10^3	2.7 x 10^3	0.25

2.6. Soil Erodibility to Wind and Water

Erosion, both wind and water is a major problem for soils under agricultural production. The management of groundcover is critical in the control of erosion and the adoption of land management practices to maintain ground cover levels is an important part of erosion control measures. Under cropping, tillage is often an agronomic necessity and when tillage leaves soil in a loose tilled condition, the soil is at its most vulnerable to wind and water erosion.

2.6.1 Soil erodibility to water erosion

Estimation of soil erodibility (K)

Soil erodibility is only one factor in determining the potential for water erosion. The overall potential for water erosion of a field or paddock is determined by the rainfall erosivity of the area, the length and degree of slope of the paddock, land management factors (the ground or vegetation cover and whether the soil is in a loose tilled condition) and the existence of any erosion control works affecting water flow from runoff (Rosewell and Loch 2002; Geeves et al. 2007b). The soil erodibility or K factor is probably a minor factor compared to some of the other ones in determining the overall erosion rate occurring at a site. However it can have an effect on the final erosion rate from a given storm as it can vary by a rate of up to 6 times with the value of K varying from 0.010 to 0.060. A relationship has been developed to estimate K in the USA using particle size, soil organic matter, structure and soil permeability. The soils used were largely high in silt and fine sands and associated with Mollisols and some Alfisols. Rosewell and Loch (2002) suggest that this relationship is not necessarily readily applicable in Australian soils. Australian soils have the complications of self-mulching, highly aggregated Ferrosols, dispersion and a larger number of coarse textures. Experimental data has shown the K values for the Red Ferrosols are 0.013 but can increase to 0.03 when degraded. The K values for the self-mulching Vertosols are 0.045, which is higher than predicted by the US relationship. Rosewell and Loch (2002) also note that some coarse textured granite soils have high K values because of low cohesion between soil particles.

Rosewell (1993) has developed a table of K values for different textures assuming a soil organic matter content of 2%. If the soil organic matter levels fall below 2%, the soil erodibility will increase.

It can be difficult to separate the effect of the management factors and the soil erodibility factor on the overall erosion rate. As an example, a major storm occurred at Cowra in 1991 in which 81 mm of rain fell in 45 minutes (Hairsine et al. 1993, reported in Edwards and Zierholz 2007). On the plot which had been under traditional till for 10 years and was in a loose tilled condition, the erosion rate was 360t/ha. The erosion rate on the adjacent direct drill plot which remained in a seedbed condition was 65 t/ha. The differences in the erosion rate can be attributed to the cover effect, the condition of the soil being tilled or untilled and the soil erodibility.

A study by Loch and Foley (1994) indicated that maintaining stable aggregates > 125 μm can reduce erosion rates. This supports the importance of aggregate stability in reducing soil erodibility.

Potential effects of soil organic matter on water erosion

Soil organic matter can have a range of effects on water through its effects on aggregate stability, bulk density, soil porosity and soil infiltration. A conclusion can be made that increasing soil organic matter improves aggregate stability and tends to reduce water erosion. However it has been difficult to separate this effect from the general management factors associated with increased vegetation cover and reduced exposure of the soil in a loose tilled condition to rainfall. Certainly bare fallow soils have much higher erosion rates (50 to 100 t/ha/yr on average) compared to 1 to 10 t/ha/yr for better managed soils (Edwards and Zierholz 2007; Rosewell 1993).

2.6.2 Soil erodibility to wind erosion

As for water erosion the potential for wind erosion is determined by a series of factors, including the local wind power or wind erosion climate factor, the field length or fetch, the soil ridge roughness factor, the quantity of vegetation cover and the soil erodibility to wind (Geeves et al. 2007b). The soil erodibility to wind is critical as wind erosion will be minimal for many soils with low soil erodibility to wind erosion. The soil erodibility to wind erosion can have a large effect on wind erosion.

In general soils are most susceptible to wind erosion when the hold less 10% clay which includes textures of sands, loamy sands and clayey sands (Leys et al 2002). However there are exceptions and if dry aggregates are less than 0.85 mm or 850 μm, they can also become susceptible to wind erosion even if they are made up of clay or silt. Aggregate stability is important as if aggregates become less than 850 μm they can become susceptible to wind erosion. No mention was made of the part of soil organic matter in this aggregation.

As for water erosion, land management practices that maintain ground cover or vegetative cover are the best way to control wind erosion (Geeves et al. 2007b).

2.7 Soil Colour

Soil colour is affected by many factors other than soil organic matter. However, it is assumed that for the surface soils, organic matter is often a major factor. The surface soils are the most highly weathered part of the soil and the parent material affects and other pedogenic effects such as drainage, nodule formation, segregation of minerals and weathering of primary minerals has largely been completed. Soil organic matter is a major colouring agent in soil formation in the surface soils. In some cases such in the case of Red Ferrosols or krasnozems and euchrozems and the Black Vertosols, the pedogenic processes remain a major colouring agent in the soils. For a large range of surface soils in the Chromosol, Kandosol and Dermosols groups, the accumulation of soil organic matter is a major pedologic process. The surface soils of these groups have been called the “fragile surface soils” by Murphy et al. (2013) because of the importance of soil organic matter in these soils.

Soil organic matter which tends to be dark or black can have an effect on soil colours but the effect varies with soil type. Peverill et al (1999) as described in Krull et al. (2004) has identified that dark brown colours often indicates soils in better condition because of higher organic matter levels. Schulze et al (1993) found a strong relationship between soil colour and the levels of soil organic matter. This is presented by Krull et al. (2004) and showed that:

- Munsell values between 3 and 4 are associated with organic matter levels of 1.0 to 2.0%
- Munsell values between 2 and 3 are associated with organic matter levels of 2.0 to 3.0%
- Munsell values less than 2 and associated with soil organic matter levels > 3.0%

This study by Schulze was for US soils in the glacial soils of Indiana and Illinois and so might not be directly applicable to Australian soils. There is a considerable volume of data in several data sets in Australia which could be analysed in a similar way and this would be potentially useful exercise.

Soil colour can be an indicator of other soil properties such as the degree of leaching and nutrient levels, sodicity, levels of sesquioxides and salinity.

In the Australian Soil Classification (Isbell 2002), one of the properties to define the melanic and melacic horizons is a black moist colour defined as having a value of 3 or less and a chroma of 2 or less. A melanic horizon must have the additional properties of being a least 20 cm thick and be well structured. These horizons are intended to account for soils with significant amounts of organic matter but are not highly organic such as the humose and organic horizons which are dominated by organic matter.

2.8 Soil Strength

Soil strength is a very complex soil property affected by a large amount of temporal and spatial variability. Soil strength not only varies with moisture content, but also with bulk density. At the critical time of emergence in cropping soils, bulk density also varies with time (Onstad et al. 1984). They showed that the bulk density of a tilled soil varies with rainfall after tillage. A soil with low aggregate stability can settle rapidly with low amounts of rainfall and increase in bulk density or form a surface crust. This means that the relationship of soil strength with moisture content also varies with time.

Soil strength becomes limiting based on penetrometer measurements when the resistance exceeds 2 to 3MPa as discussed in Section 2.1.3. Several references have discussed how improved cropping practices can reduce soil strength. While increased soil organic carbon may be part of the reason, reduced compaction and bulk density are also involved in reducing soil strength.

2.8.1 Pedotransfer functions for soil strength and soil organic matter

Comprehensive pedotransfer functions relating soil strength to soil organic matter are not common. One PTF developed for soil strength is that proposed by de Silva and Kay (1997) in Kay et al. (1997), which includes moisture content, bulk density and soil organic matter. This equation shows the complex nature of the relationship between soil strength and soil organic matter. The PTF is:

$$SR = c\,\theta^{d}\,BD^{e}$$

SR is the soil resistance to penetration, θ is the moisture content, and BD is the bulk density. The coefficients c, d and e are functions of clay content and the soil organic carbon.

$$\log_e c = -3.67 + 0.765\,(SOC\%) - 0.145\,(clay\,\%)$$
$$d = -0.481 + 0.208\,(SOC\%) - 0.124\,(clay\%)$$
$$e = 3.85 + 0.0963\,(clay\%)$$

A full application of this PTF has not yet been applied. It was developed on a Canadian set of soils including sandy loams and clay loams that were derived from calcareous parent materials and had illite clay minerals.

Applying this PTF, Kay et al. (1997) concluded that soil organic matter had a large effect on the soil resistance to penetration. In general increasing the soil organic carbon by 1g/100g halved the soil resistance to penetration, especially at moisture contents drier than field capacity. In many cases this reduced the resistance to a level below 2 MPa which permitted root growth to recommence. However, Kay et al. do caution on applying these results to other soils in other regions.

2.8.2 Soil strength data

When soil strength is measured in the field, it is dependent on the land management history and actions such as tillage, compaction events and moisture conditions will affect the actual soil strength measured. It can be difficult to find direct relationships with soil organic matter under these conditions. For example in Chan and Mead (1988), the soil strength measured by a shear vane was 37.8 kPa under pasture even though the bulk density was less (1.24t/m^3), the soil moisture was higher (0.25%) and the soil organic carbon was 2.5%, compared to 22.3 kPa under traditional tillage which had a bulk density of 1.48 t/m^3, a moisture content of 0.19% and soil organic carbon content of 0.9%. This demonstrates the impact of a tillage operation on soil strength. Any relationships developed between soil strength and soil organic matter have to be the basis of a long term equilibrium or as a general trend and be aware of the practical implications in the field that land management actions can affect soil strength outside of the general relationships.

2.8.3 Conclusion

Soil strength has complex relationships with soil moisture, bulk density and soil organic matter. The most effective way to identify the influence of soil organic matter on soil strength is to examine how soil organic matter affects the soil strength soil moisture relationships.

2.9 Compaction and Engineering Characteristics of Soils

2.9.1 The Engineering Soil Properties

Increasing levels of soil organic matter change the compaction and engineering characteristics of soils which can be important for when soils are subjected to compaction pressures either from stock or machinery and affect their response to tillage.
Soil organic matter affects the engineering properties of soils including the Atterberg Limits and the Proctor Maximum Compaction Density or Maximum Density.

The Atterberg Limits (Atterberg 1911) are widely used to classify cohesive soils for engineering properties (Hicks 2007). They can be related to soil mechanical behaviour and so to soil behaviour under traffic and tillage. These limits are generally described as the soil behaviour changes from a saturated condition to drier condition. When the soil is wetter than the liquid limit, the soil is in a liquid state and flows under its own weight. In this condition it is not suitable for traffic or tillage. As the soil dries past the liquid limit it begins to behave in a plastic state where the soil can be remoulded and deformed but does not crumble. Again this is not the optimum condition for traffic or tillage. Once the soil is slightly drier than the plastic limit the soil begins to crumble on deformation and the definition of the plastic limit is:

> “The plastic limit is the moisture content at which the soil begins to break apart and crumble when rolled by hand into threads 3 mm in diameter (Hicks 2007) .”

When the soil just drier than the plastic limit is the optimum condition for tillage and is suitable for traffic as well. The recommended moisture content for tillage is ≈ 0.7 to 0.9 plastic limit when the soil is in a crumbly condition and not too dry (Keller and Dexter 2012; Ojeniyi and Dexter 1979).

The Proctor Maximum Density or Maximum Density (as defined in United States Department of the Interior 1960 and Standards Association of Australia 1993 - 2003) is a measure of the maximum density that can be achieved for a given soil under standard compaction conditions and the moisture content at which that compaction occurs. The density and moisture content will vary with texture, clay type and soil organic matter levels.

2.9.1 The Importance of the Engineering Soil Properties for Soil Management

The engineering soil properties can be of relevance to agricultural for tillage operations, trafficability and for identifying soils susceptible to excessive compaction under stock and machinery. Kirkby (2002) makes the point that:

"The Atterberg Limits are often quoted for soils as if they are a constant. For many applications this maybe reasonable but when organic matter content or exchangeable cation of a soil varies as a result of agricultural management, this assumption is no longer reasonable."

The Atterberg Limits are important for agriculture because they define the moisture contents when the soil is at the optimum for cultivation, friability, workability and trafficability and most susceptible to compaction damage. Dexter (1988) defines "soil workability" as the difference between the plastic limit, the optimum for tillage, and the field capacity. The closer the field capacity is to the plastic limit, the more likely the soil is to be at desirable moisture contents for tillage after rainfall.

The Atterberg Limits describe the effect of varying water content on the consistency of a fine-grained soil. They define the water contents at which the soil passes from a solid state to a plastic state (the plastic limit) to a liquid state (liquid limit) (Kirkby 2002; Hicks 2007). As discussed above the moisture content at each of these limits can have practical implications for the response of the soil to tillage and compaction in the field under cropping and grazing.

An example of how soil organic matter can affect the Proctor Maximum Density of soils is presented in Hamblin (1987). The maximum density is significantly less as the soil organic matter increases. Thomas et al (1996) undertook a more detailed study of how soil organic matter affects the density of soils after compaction and showed a very strong relationship between the Proctor Maximum Density and the soil organic matter content over the range of 0.5 to 3% soil organic carbon. In unpublished data Murphy (1994) showed that the bulk density under a range of compaction pressures was about 0.10 t/m^3 less, when a sandy loam soil that had a soil organic carbon level 0.80% compared to a soil with 0.54%.

It is clear that soil organic matter can significantly affect the engineering properties of soils. Most likely this is a consequence, at least in part, of the influence on aggregate formation and stability.

Pedotransfer functions relating liquid and plastic limits to clay content and soil organic matter have been developed by Keller and Dexter (2012). They developed this PTF using soils from a wide range of source including Australia. The liquid limits increased with increasing soil organic matter for a given soil texture. Significantly, Keller and Dexter (2012) suggest that any influence of soil organic matter on the plastic and liquid limits of soils is likely to be associated with the humified soil organic matter which are operating at the molecular levels and intimately at the soil particle scale and not the soil organic matter that is a more fibrous or particulate form.

2.10 Friability – Response to Tillage

Soil friability is a key soil physical property to provide information about how soils behave for cropping operations including tillage and trafficability. It enables predictions to be made about the ease of producing favourable seedbeds for germination and emergence, the energy required for tillage, and the suitability of soils for producing suitable seedbeds with minimum disturbance machinery (Munkholm 2011). Friability is defined as:

> ".. the tendency of a mass of unconfined soil to break down and crumble under applied stress into a particular size range of smaller fragments" (Utomo and Dexter 1981).

The optimal seedbed to enhance germination and emergence of seedlings is produced by a bed of soil aggregates in the size range of 0.5 to 8 mm (Braunack and Dexter 1989). Aggregates larger than 5 to 8 mm are not desirable because of the risk of rapid drying of the germinating seed. Where aggregates smaller than 0.5 mm occur there is the possibility of surface sealing and crust formation. Good friability of a surface soil is even more crucial for good crop establishment under no-till and seeding systems with minimal disturbance.

In the review of friability measurements and the practical implications of friability, Munkholm (2011) clearly identifies that for many soils there is a strong relationship between soil organic matter and friability, but also identifies that there is no universal relationship between soil organic matter and friability. The reason for this is clear given the nature of self-mulching surface soils which are common on many of the fertile Vertosols derived from basaltic parent materials. These strongly self- mulching surface soils have high friability which has no relationship to soil organic matter as it is a consequence of the smectite clays and the high levels of exchangeable calcium. As Oades (1993) has identified, the mechanisms for maintaining favourable soil structure are different for sandy soils, loams and clays. For soils without high amounts of clay and high amounts of expandable clays, soil organic matter can have a strong effect on friability. For sandy soils, soil organic matter is crucial and adequate inputs of soil organic matter are required to maintain and improve soil friability. This was demonstrated by Macks et al. (1996) in measurements of soil friability on a set of soils from the wheat belt of south eastern Australia. The soils with low soil carbon in the 0 – 50 mm layer (0.50 to 0.70%) had friability values of < 0.15 while the soils with higher soil carbon (1.00 to 6.7%) had friability values of 0.29 to 0.80). Values > 0.25 are considered very friable. The relationships between friability and soil organic carbon because they are indirect, indicate that there is an effect of different soil organic matter fractions on the levels of friability and aggregate strength.

Summary of Key Points - 2. General Soil Physical Properties

- Soil organic matter has general impact on a broad range of soil physical properties. Key to this is its effect on aggregate stability. Soil organic matter determines the aggregate stability of soil with lower levels of clay (loamy sand to loam) and is still a factor in aggregate stability in clayey soils. As a general rule, the particulate organic matter fraction (POC) is the key factor for aggregate stability of larger aggregates (>250μm) and humus for smaller aggregates (<250μm). Because of the importance of different fractions not all measures of aggregate stability can be directly relate to soil organic matter levels. Aggregate stability declines rapidly below 1.2 to 1.5%, and at least 2% soil organic carbon is generally considered necessary for optimum aggregate stability.
- While plant available water is largely determined by the particle size distribution of the soil, it can be significantly modified by soil organic matter. Largely through increasing field capacity the soil organic matter can increase the plant available water by about 1.5 to 3 mm/10 cm depth for each 1% increase in soil organic carbon for the lighter textured soils and about 0.5 mm/10 cm depth for the heavier textured soils.
- Maintaining soil organic matter levels is a major factor in controlling water and wind erosion.
- Soil organic matter can have beneficial effects on the soil strength, soil compaction characteristics and soil friability. All these are especially important for cropping and tillage operations. By maintaining soil organic matter levels the soils have lower strength, are less susceptible to compaction and have better tillage characteristics. All these features can result in more timelines of cropping operations, reduced costs, better seedbeds and ultimately better productivity.

3. Effects of Soil Organic Matter on Soil Chemical Properties

3.1 Nutrient Cycling – Nutrient Sources

3.1.1General nutrient cycling

Soil organic matter is central to nutrient cycling and nutrient supply in most soils. Although industrial fertilisers have been a major source of nutrients for agricultural soils, the soil organic matter, including the microbial component, still remains a major store and pathway for plants to obtain the nutrients needed for growth. The general stores and pathways for nutrients in soils are summarised in Figure 3.1. Soil organic matter and the microbial population associated with soil organic matter plays a major part in the cycling of the nutrients in soils and in ensuring that the nutrients are in a form available to plants. In some cases as for phosphorus, the microbial population may be important in enhancing the dissolution of the crystalline fraction of the nutrients.

A major review of the role and importance of soil organic matter in nutrient cycling has been undertaken by Macdonald and Baldock (2010). In this review some of the key conclusions are:

- In agricultural systems the natural nutrient cycling processes are disrupted by the removal of nutrients in agricultural products and harvests.
- The biochemical recalcitrance and bioavailability of the organic substrate, and the functional potential of the soil microbial community determine the decomposition rate of soil organic materials and the supply of nutrients
- The decomposition and flow of nutrients in soil has to follow the principles of stoichiometry and all energy and nutrient flows need to maintain a balance. This requires dividing soil organic matter into manageable pools and keeping balances between C:N:P:S and the trace elements.

Nutrient cycling from organic matter is of increasing interest giving the increasing cost of artificial fertilisers.

3.1.2 Amount of nutrients held in soil organic matter

The amount of nutrients stored in soil organic matter can be estimated using the general assumption that soil organic matter has a constant ratio of nutrients of C:N:P:S. The constant ratio has been suggested by Williams and Donald (1957) 155 : 10 : 0.68 : 1.4, Rice (2002) 100 : 10: 1 : 1, Stevenson (1982) 140 : 10 : 1.3 : 1.3 and Himes (1998). With the need for the formation of soil organic matter to have a relatively constant nutrient ratio an adequate supply of nutrients is essential for the increase of soil organic matter. Spain et al. (1993) have suggested that a lack of P has limited the amount of soil organic matter in some Australian soils, especially as it must be supplied from the parent material. Similarly other authors have suggested that ultimately P can set limits to the amount of soil organic matter that can be accumulated in soils (Walker and Adams 1958; McGill and Cole 1981).

In a major review, Kirkby et al. (2011) identified that it is the humus fraction of the soil which has a very constant nutrient ratio. The light fraction or a fraction that approximates to the particulate carbon fraction is more variable and depends on the land management history and the nature and source of the organic materials that are being are being inputted into the soil carbon system. However, Kirkby et al did identify that while there was generally a strong link between C and N, and also C and S, the link between C and P was much more variable. The results for Kirkby et al. indicated that the humus fraction has a nutrient ratio closer to that proposed by Rice (2002). If the nutrient ratio proposed by Rice is assumed the amount of nutrient stored in the soil organic matter is indicated in Table 3.1.

While the figures in Table 3.1 indicate that soil organic matter holds a large store of nutrients it is not easily accessible as much of it is held in that faction of the soil organic matter that is more resistant to decomposition. It estimated that 2 to 4% of soil organic matter can be decomposed to provide nutrients on an annual basis (Rice 2002). Decomposition of the soil organic matter also can reduce the soil organic matter levels which has other potentially deleterious effects on soil properties.

Comparing the normal amounts of nutrients required to grow crops with levels of nutrients held in the soil organic matter (Table3.1) suggests that the soil organic matter has a valuable store of nutrients. Wheat requires about 100 kg of N to produce a 3 t/ha crop with about 10 to 11% protein and about 10 kg of P. Based on the levels in Table 3.1 there is sufficient in the nutrient store for several years of crops. Of course the nutrients in the soil organic matter are not in a form readily available to the plant, nor would it be advisable to keep running down the levels of soil of organic matter to utilise the entire nutrient reserves held in the soil organic matter. The availability of the nutrients very much depends upon which fraction of the organic matter the nutrients are held (see Table 1.1 and McDonald and Baldock 2010).

Soil organic matter holds 90 to 95 % of the nitrogen held in soils and the N nutrient cycle is intimately tied in with the soil organic matter and the soil microbial population. Sulfur is also closely tied to soil organic matter it is estimated 90 % of the S is tied up with the soil organic matter (Rice 2002). Phosphorus is very different. While a large proportion of the P can be tied up with organic matter, up to 80% (Stevenson 1982), organic P is more commonly about 40 % (Rice 2002). The reminder occurs as crystalline form in a range of P minerals such as the apatite group (Kirkby et al. 2011). Kirkby et al. found a weak relationship between soil C and acid extractable P suggesting that there are several sources of inorganic P.

Some of the important conclusions to be drawn from Figure 3.1 are the following:

- Potentially there are three major stores from which nutrients can move into the soil solution to become available to plants. These are soil organic matter, the crystalline or soil mineral store, and the adsorption surfaces of the clay minerals and soil organic matter.
- The movement of the nutrients from these stores is dependent on:
    - The action of the soil microbial population to decompose soil organic matter or dissolve crystalline materials.
    - The chemical environment of the soil solution to have the appropriate pH, redox conditions or electrochemical concentrations to dissolve the crystalline minerals or cause desorption of the nutrients from the adsorption surfaces.
- Soil organic matter can provide some adsorption surfaces for some nutrients, especially for the negative anions such as S and P.
- Some of the nutrient can move between the different stores which can be an advantage if the nutrients are more available from some stores.

- The mineralization of nutrients from soil organic matter and release of nutrients in a form available for plants requires the soil microbial population to decompose soil organic matter
- Conversely, the formation of soil organic matter requires nutrients to be immobilized in the soil organic matter.
- The addition of fertilisers as shown in Figure 3.1 can add nutrients to the soil solution but also to the different nutrient pools in the soil.

While soil organic matter does provide a large store of nutrients, it is not readily accessible. Nevertheless it does provide a nutrients store and the base for the soil microbial population which is essential for nutrient cycling in the soil.

The capacity of soil organic matter to provide nutrients is dependent on the ease with which it can be decomposed and on the nutrients present within the soil organic matter.

3.1.3 Nitrogen

Of the total nitrogen in the soil, 90 to 95 % is held in organic form as soil organic matter including soil organisms. The remainder occurs as ammonia, nitrates and nitrites. Most plants can absorb NH_4^+ and NO_3^- rapidly, but plants can also absorb urea [$CO(NH_2)_2$] and amino acids, but to gain access to the nitrogen in soil organic matter, the organic matter has to be mineralized to release the N in a form that the plants can uptake (Allison 1973). Even though a measurement of total nitrogen might indicate there is abundant nitrogen present in the soil, it will not be immediately available to the crop. Plant N requirements and the release of N from soil organic matter reserves vary with seasonal conditions (Strong and Mason 1999). The principle pathways of the nitrogen cycle are shown in Strong and Mason (1999). The main features of this pathway are:

- Plants require the N to be in a mineral form in order to take up the N. This usually means it has to be in the form of NH_4^+, or NO_3^- .
- The mineral form of N can be sourced from:
    - Mineralization of soil organic matter – requires activity of the soil microbial population
    - Additions of fertilizer as NH_4^+, or NO_3^- or urea.
- The NO_3^- is not stable, it can be:
    - Leached out of the soil if soil is too permeable and the rainfall exceeds the evapotranspiration
    - Converted to N_2O and then to N_2 by denitrification in anaerobic conditions or near anaerobic conditions.
- NH_4^+ can be converted to NO_3^- in a further mineralization process.
- Nitrogen can be fixed from the atmosphere in which N_2 is converted to NH_4^+ usually by rhizobia in the roots of legumes. This N then becomes part of the biomass of the legumes which eventually becomes part of the soil organic matter when the legume biomass decomposes in the soil.

The levels of the mineral forms of N are subject to high spatial and temporal variability and are not always a reliable indicator of the N needs of crops (Strong and Mason 1999). Total N and total C, largely determined by soil organic matter, do provide the main store of N for plant growth.

Nitrogen Cycling

Nitrogen is a dynamic nutrient and it is being continuously cycled between the atmosphere, the soil and living organisms. Within the soil it can exist as the unreactive gases N_2, N_20, the ions NH_4^+, NO_3^- or in organic forms as amino acids, lignin and a wide range of other compounds in living organisms, residues and soil organic matter. There are a number of pools of N in soils (Herridge 2011):

- Plants – present in proteins, transport and storage compounds, structural compounds (eg lignin) and genetic material
- Residues - straw, shoots, leaves, old root systems
- Dung and urine from animals
- Soil organic matter - active fractions, humus, char or charcoal
- Microbial biomass
- Mineral N – ammonium (NH^+,) nitrate (NO_3^-), nitrite (NO_2^-). This is the critical N for plants.

N transformations between these pools could not happen without the organisms that live in the soil and soil biology. These major N transformation processes include (Herridge 2011):

- Urea hydrolysis which involves the conversion of urea to ammonia by the enzyme urease:
    - $CO(NH_2)_2 + 2\ H_2O \rightarrow 2NH_3 + CO_2 + H_20$
- Humification which is the decomposition of plant and animal residues to relatively stable organic matter which is usually in the form of humic and fulvic acids or humus.
    - The humification process will be influenced by the quality of the organic residues. The residue quality can be defined by two factors as defined by Praveen – Kumar et al. (2003) who used the C/N and the Lignin/N ratios to define the residue quality (see Table 3.2).
- Mineralisation
    - Ammonification which is the conversion of organic materials in the soil to ammonia and ammonium by soil organisms
    - Nitrification – the conversion of ammonium to nitrate and nitrite. This is inhibited in acid soils.
- Immobilisation where mineral N (ammonium and nitrate) is incorporated into microbial biomass and occurs in the humification or breakdown of plant residues with a high C:N ratio. Generally C:N ratios greater than 30 will immobilise N (Herridge 2011).
- Denitrification in which nitrate and nitrite is converted to N_2O and then to N_2 by soil bacteria in anaerobic or near anaerobic conditions. This can be a serious loss of N from the soil N pool but it also has important implications for greenhouse gas emissions (Dalal 2003a, b) as N_2O is a potent greenhouse gas. This process is most likely to occur when soils are at or near saturation or soils are compacted.

It is possible to estimate the amount of N available for a crop from the following (Herridge 2011):

- N mineralisation of organic matter or humus

This depends on the amount of soil organic matter or humus in the soil. Tilled soils and fallowing result in higher mineralisation rates than untilled soils with untilled soils having about 80% the rate of tilled soils. It is the humus fraction which is considered the most important for providing N after mineralisation of soil organic matter. Soil moisture is also required for mineralisation to occur so in dry times mineralisation is reduced. The supply of N by the mineralisation of soil organic matter or humus ranges from 10 to 32 kgN/ha for in-crop mineralisation under no-till and 13 to 40kgN/ha for in-crop mineralisation under cultivation. Higher amounts of N can be released under fallow, but this depletes larger amounts of humus from the soil.

- N mineralisation of fresh crop residues

For estimating the amount of N available from the fresh crop residue the assumption is made that 65 to 70% of the fresh crop residue is mineralised. How much N is released depends on the C:N ratio of the crop residue. Breakdown of a residue such as wheat stubble may immobilise N because of the high C:N ratio, but mineralisation of a residue such as chickpeas, Faba beans or canola will release N because of their lower C:N ratio (Herridge 2011). Adequate moisture and warm temperatures are required for the mineralisation of the crop residues. Substantial amounts of residue in the form of stubble from a wheat crop with a grain yield of 3t/ha is likely to immobilise 21kgN/ha while a 1.5 t/ha canola crop can release at least 16kgN/ha.

- Unused nitrate from previous crop or land use

The unused nitrate can be estimated from soil measurements of nitrate and crop yields, although the amount of nitrate leached is always difficult to estimate.

- N mineralisation of urine or dung.

Ladd and Russell (1993) estimated that 20 to 40% and up to 54% of the N in urine and dung can be lost as a result NH_3 volatilization. Much of the N is present as urea in the urine and the urease is present in the dung to break down the urea into NH_3 which makes it difficult to manage and preserve the N in the animal waste and the use of urease inhibitors has been one option to reduce the emission of ammonia and nitrous oxide (Dalal et al. 2003a).

Fixation of Nitrogen by Legumes

While the legumes are the primary beneficiaries of their capacity to convert N_2 to mineral N as they can grow in the soil with no inputs of N fertiliser, the legumes do produce N-rich plant residues after the crop is harvested or after the growth cycle of the annual legume pasture species. The mineral N released by these residues can be taken up by subsequent crops or grass pasture species, generally enhancing the soil organic levels matter of the soil. The input of N into the soil is a critical step in the increase of soil organic matter levels.

The amount of N fixed by the legumes has been reviewed by Unkovich et al. (2010) and Herridge (2011) and the amounts include:

- 105 kgN/ha/yr for field peas
- 180 kgN/ha/yr for soybeans
- 130 kgN/ha/yr for lupins
- 70 kgN/ha/yr for chickpeas
- 120 kgN/ha/yr for subterranean clover
- 80 kgN/ha/yr for annual medic and
- 180 kgN/ha/yr for lucerne.

The amount of N in the shoot and root dry matter that would be added to the soil is estimated at (Unkovich et al. 2010 and Herridge 2011):

- Subterranean clover – shoot 88kgN/ha; root 62kgN/ha
- Annual medics – shoot 78 kgN/ha; root 31kgN/ha
- Annual clovers – shoot 167 kgN/ha; root 67kgN/ha
- Lucerne – shoot 149 kgN/ha; root 149kgN/ha

- Chickpeas shoot 85 kgN/ha, root 85kgN/ha
- Faba beans – shoot 122 kgN/ha; root 50kgN/ha
- Lupins – shoot 125 kgN/ha; root 51 kgN/ha

The figures above indicate the potential for the residues from the legume plants to add N to the soil organic matter. The amount of N fixed is dependent on yields as well as on the efficiency of fixing N. Soybeans has a high potential to fix N because it is a high water demanding crop and high yield potential compared to crops such as field peas, lupins and Faba beans which not only have lower yield potential but are also usually grown under moisture limiting conditions. In soils with high nitrate levels the rate of N-fixation will also be suppressed (Herridge 2011). Other factors to ensure adequate N fixation are prevention of soil acidity and P deficiency. Management of P nutrition and soil pH are essential to manage N fixation in soils. The N-fixing legumes have extra nutrient requirements for calcium, boron and molybdenum. The addition of lime in soils with acidity is essential as some rhizobia are sensitive to acidity and Al toxicity. The other requirement is to avoid soils with high levels of salinity and sodicity.

Conclusion for Nitrogen

The soil organic matter and soil biology are a key part of the management of N nutrition of the soil. Managing the soil organic matter and the soil biology are necessary to maintain adequate levels of soil N. The maintenance and increases in soil N in most Australian soils require the inputs of N from N fixation by legumes and the subsequent mineralisation of the soil organic matter and humus to supply plants with at least some of their N requirements. While fertilisers are also an important source of N, the increasing cost of N fertilisers is likely to increase the reliance on the legume and soil organic matter sources of N.

3.1.4 Phosphorus

Phosphorus is a complex nutrient. Soil organic matter is only one source of the nutrient and soil organic matter and soil organisms interact in a complex way with some of the other mineral and adsorption surface sources of the nutrient. The estimate is that about 40% of P is in an organic form in the soil organic matter (Rice 2002), although this can vary from 25 to 80% (Probert 1993). Plants usually take P as the orthophosphate anions ($H_2PO_4^-$ and HPO_4^{2-}) which are present in the soil solution (Stevenson 1982). The plants usually take up these anions in the rhizosphere and as they are taken up the P is replenished by diffusion from the soil solution (see Figure 3.1). Any P compounds need to be mineralized from the soil organic matter or from the mineral phase to form the orthophosphates which requires the action of extracellular phosphatases (Moody and Bolland 1999). In the rhizosphere the chemical environment can be very different because of the presence of enzymes, differences in pH, partial pressure of carbon dioxide, and cation and anion activities. This can affect the availability of P.

Soil organic matter can be a source of P when it is mineralized. Soil organic matter might also influence how much P is fixed by Al and Fe oxides and calcium carbonate as the organic matter can prevent this from occurring. The effect on the amount of P fixation depends on the nature of the organic compounds (Moody and Bolland 1999).

Finer root systems and the extent of mycorrhizae hyphae can increase the access of plants to sources of P (Bolan et al. 1984). A more detailed discussion of the role of mycorrhizae and P is given in Watt et al. (2006).

The relationship of soil organic matter to P is not as direct as it is for N. However, higher levels of soil organic matter are likely to be an advantage in providing a reserve supply of P but also supplying a source of carbohydrate for microbial populations that can increase access to P sources.

Pools of Phosphorus in Soils

As for N, a similar type of mineralization and immobilisation process which moves the P between the soluble and organic pools exists for phosphorus. However instead of the dentrification process which removes the N to the atmosphere, there is a crystallization or fixation process that removes the P to the insoluble mineral pool in the soil to form "fixed P" (Moody and Boland 1999). It is the soluble P which is generally considered to be available to the plant. The P held in the soil organic matter or humus, organic residues and microbial biomass is not immediately available to the plant and mineralisation of these organic materials is required for the P to become available to the plants. Soils differ in the capacity to move P into the "fixed P" pool. Those soils high in iron and aluminium sesquioxides (krasnozems or Ferrosols) have the greatest capacity to fix P. As will be discussed in Section 3.4.2 soil organic matter has the capacity to reduce the amount of P which is fixed, as it can absorb some of the P into the organic fraction keeping it in a form which is at least available in the long term.

Interaction between Nitrogen and Phosphorus

The critical role of legumes in the nitrogen nutrition of soils is not possible unless the legumes have an adequate supply of P (Herridge 2011). The overall amount of N fixed into the crop or plant is dependent on the dry matter yield and the effectiveness of the plant in fixing N and this requires a healthy plant adequately supplied with nutrient. While the legume can obtain the N by fixation, it must have a source of P and other nutrients.

Importance of Micronutrients

The recycling of nutrients is also dependent on certain micronutrients being present or can be suppressed by other cations or anions. Molybdenum (which is at the centre of the nitrogenase enzyme) and boron (Bolanos et al. 2002) are essential for N-fixation and high levels of P can induce zinc deficiency in plants (Zhu et al. 2001). High levels of Ca^{2+} can inhibit the availability of soluble P. Overall nutrition requires the micronutrients to be present in sufficient quantities.

3.1.5 Sulfur

Most of the sulfur in soils is an organic form and can be mineralised into the inorganic form by soil microorganisms. About 1 to 5% of the sulfur is usually in the inorganic forms as soluble sulfate (Stevenson 1982). The exception to this is soils that have high amounts of sulfur rich minerals such as the sulfides as occurs in acid sulphate soils that occur most often in coastal plains.

A major source of S results from the decomposition of plant materials, especially of crops such as canola that have a high S content (Singh et al. 2004). In Australia S was generally not considered deficient for many soils as it was added as part of superphosphate for many years (Lewis 1999). The widespread growing of canola has indicated more widespread deficiency of S in Australian soils with the use of high analysis P fertilisers instead of superphosphate and the high S needs of canola. Gypsum is often now applied with canola crops.

While S can be supplied to soils by sea spray in coastal areas in inland areas a major source of S is plant residues and animal excreta. Organic accumulation of S is a major source for plant growth

(Lewis 1999). The activity of soil microorganisms is a major determining factor in the availability of S.

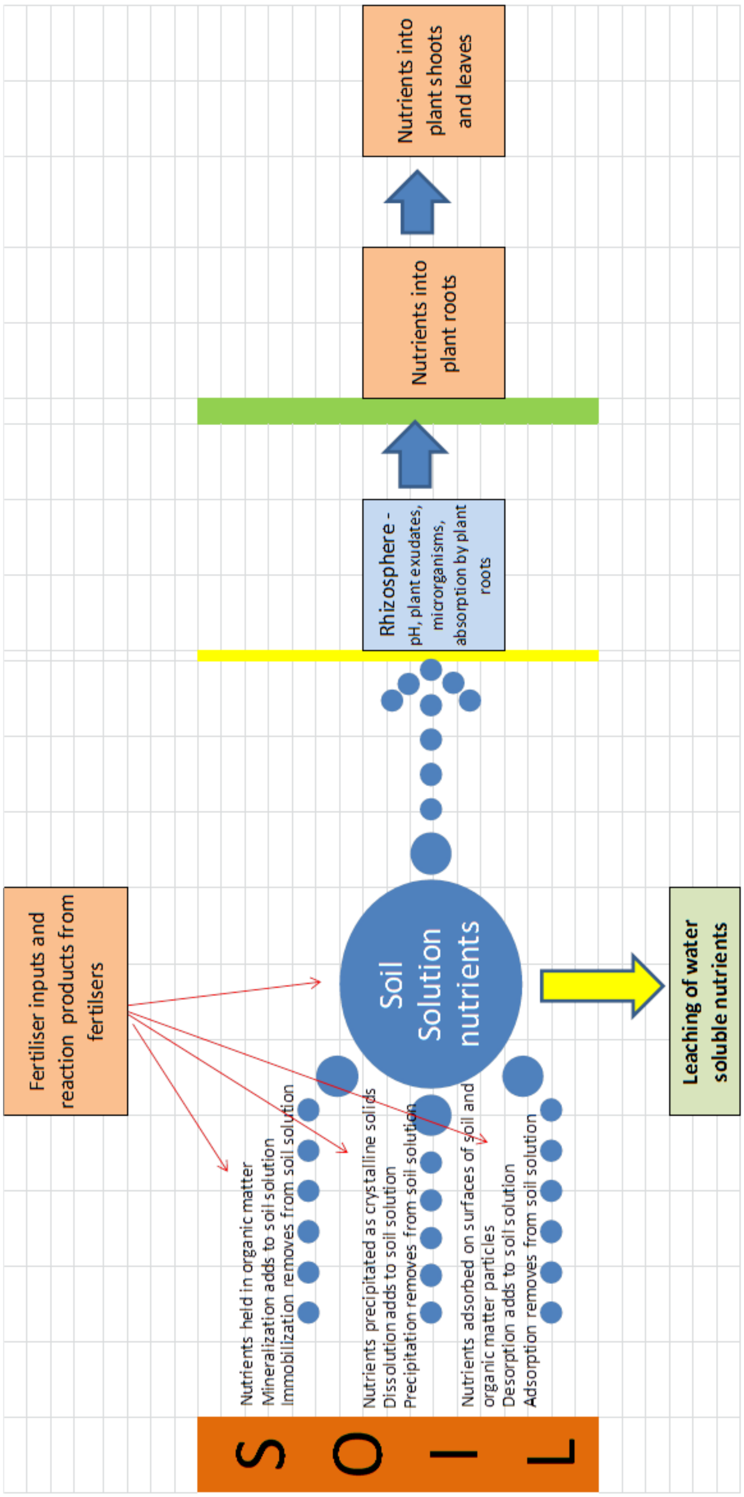
Table 3.1. Estimated amounts of nutrients present in soil organic matter. However, these nutrients are only available to plants if the soil organic matter is mineralised and much of the nutrient is held in the more resistant part of the soil organic matter, the humus. The clay soils appear to hold less nutrients because of lower bulk densities, but are more likely to have higher levels of soil organic matter. These estimates are based on soil organic matter having a nutrient ratio of C:N:P:S of 100:10:1:1 (see Rice 2002).

Texture	Mass of soil 0 - 10 cm	soil organic carbon	soil organic matter	N kg/ha	P kg/ha	S kg /ha
sandy loam	1650 t/ha	0.5	0.86	1419	142	142
		1.0	1.72	2838	284	284
		1.5	2.58	4257	426	426
		2.0	3.44	5676	568	568
		3.0	5.16	8514	851	851
loam	1400 t/ha	0.5	0.86	1204	120.4	120.4
		1	1.72	2408	240.8	240.8
		1.5	2.58	3612	361.2	361.2
		2	3.44	4816	481.6	481.6
		3	5.16	7224	722.4	722.4
clay	1300 t/ha	0.5	0.86	1118	111.8	111.8
		1	1.72	2236	223.6	223.6
		1.5	2.58	3354	335.4	335.4
		2	3.44	4472	447.2	447.2
		3	5.16	6708	670.8	670.8

Table 3.2. General categories of soil organic matter quality for decomposition based on quality index. (Adapted from Krull et al 2004 and Preveen – Kumar et al. 2003)

Organic matter quality	C/N ratio	Lignin/N ratio
Highly decomposable	< 18	< 5
Moderate	18 to 27	5 to 7
Slow	28 - 60	7.5 to 15
Least decomposable	➢ 60	➢ 15

Figure 3.1 Model for nutrient cycling in soils



3.2 Cation Exchange Capacity and Soil Organic Matter

3.2.1 Cation Exchange Capacity

Cation exchange capacity is the capacity of the soil to hold exchangeable cations. While Ca^{2+}, Mg^{2+}, K^+ and Na^+ comprise the bulk of the cations, other cations can include NH^{4+}, and micronutrients such as Cu^{2+}, Co^{2+} and Zn^{2+}. In acid soils (pH < 5.5) the sum of Ca^{2+}, Mg^{2+}, K^+ and Na^+ is often less than the exchange capacity of the soil, the remainder being filled by Al^{3+}, H^+ and by Mn (Rengasamy and Churchman 1999).

The negative charge that accounts for the cation exchange capacity of soils (CEC) has two major sources:

- The clay or soil minerals which vary in the net negative charge which arises from the crystalline structure of clay or soil minerals. The typical values for the different clay and soil minerals are shown in Table 1.5. The clay minerals derived from basic rocks such as the basalts (smectite) have higher cation exchange than those derived from the more acidic rocks or highly weathered shales and sandstones (illite and kaolinite).
- Humus in soil organic matter – this has the complication that the net charge on the organic matter can be dependent on the soil pH. This occurs because the net charge on the organic compounds responsible for the cation exchange capacity can be pH dependent and the ionic strength of the soil solution (Rengasamy and Churchman 1999). To counter this the range of pH for many soils experience in the field is relatively narrow, usually covering less than 2 to 3 pH units at most and often less than 2 pH units.

Relationships to predict the CEC of soils usually have the clay content as the major component but can have soil organic matter as a secondary component to improve the prediction. However it is clear from Table 1.5, that clay type can have a dominant effect on CEC and therefore any PTF's to predict CEC using clay content need to be for a specific region or soil type where there is reasonable confidence that the mineral type of the clay is constant.

By convention cation exchange capacity is the CEC at pH 7.0. The sum of the exchangeable cations Ca^{2+}, Mg^{2+}, K^+ and Na^+ is effectively the CEC at the field pH and is the effective CEC or ECEC (Rengasamy and Churchman 1999).

3.2.2 Potential of soil organic matter to provide cation exchange capacity

Soil organic matter includes a wide range of compounds. Included are a group of organic compounds known as humic and fulvic acids as well as derivatives of amino acids and phenolic acids which have the general form RCH_3 – COOH and R-CH_2-COOH (Stevenson 1982). The importance of these compounds is that as the pH of soils increases from 5.0 to 7.0 these hydroxyl groups become disassociated or ionised to form negatively charged R^*COO^- sites which are available for cation exchange. The precise pH when this occurs varies depending on the exact nature of the different compounds. The carboxyl groups in humus ionise or dissociate mainly in the acid part of the pH scale or above pH 6.0 in the case of the phenolic hydroxyls. This has important practical implications as it means that much of the increase in the net charge on humus is likely to occur below pH 7.0. When the pH falls below 5.0 many of the organic acids which make up the soil organic matter are not disassociated and so do not display a negative charge.

Generally the greater the degree of decomposition or humification of the organic matter the higher the CEC of the organic matter (Stevenson 1982). A general rule is that 25 to 90% of the CEC of a soil is attributed to soil organic matter but this varies greatly depending on the parent material, derivation of the soil organic matter and land management history of the soil. In sandy forest soils soil organic matter contributes much more to CEC than in Vertosols derived from basalt. Stevenson (1982) outlines several laboratory methods for determining the relative contribution of clays and soil organic matter to CEC (pp327). Helling et al. (1964) show a strong relationship between pH and the CEC contributed to CEC by soil organic matter. The contribution at pH 3.5 is 28% but at pH 7.0 the contribution is 40%. Helling et al. showed large contributions of soil organic matter to CEC for forest and grassland soils.

Hallsworth and Wilkinson (1958) developed some regression equations to predict CEC from clay content and soil organic matter. The coefficient for clay content varies considerably depending on soil type as would be expected as this depends on the clay minerals, and the coefficient for the organic matter also varies depending on the local environment and land use. This means that the CEC of the soil organic matter varies depending on the pH of the soils, but also on the nature of the soil organic matter. The “Chernozem” soil organic matter was estimated to have a CEC of 297 cmol(+)/kg while the soil organic matter from the acid soils from the same areas was estimated to have a CEC of 134 cmol(+)/kg.

The PTF suggested by Hallsworth and Wilkinson (1958) for “miscellaneous acid soils” from Australia is:

$$\text{CEC} = 5.13 + 0.23 * \text{clay } \% + 2.27 * \text{soil organic carbon } \%$$

This is applied to some of the soils to test the effect of organic matter on CEC. For the chernozem soils the coefficients for the clay and soil organic carbon are 0.82 and 5.12 respectively so they consider the clays have higher CEC and the soil organic matter has higher CEC as well.

Chan et al. (1992) developed a relationship of:

$$\text{CEC} = 0.123 + 2.97 * \text{soil organic carbon } \%$$

For a Red Kandosol under a cropping system. This gave a cation exchange capacity of the soil organic matter of 172 cmole(+)/kg. They concluded that the soil organic matter contributed a large part of the soil cation exchange capacity.

Stevenson (1982) explains that the potential sites for cation exchange in soil organic matter are substantially more than is usually measured as CEC. Many sites become unavailable because of associations with polyvalent cations.

A further relationship including soil organic carbon was developed by Tranter et al. (2009) for soils on the north coast of NSW in the Woodburn area. The relationship involving soil organic carbon and cation exchange capacity was:

$$\text{CEC} = 3.7 - 0.28\ \text{pH} + 2.6 \log_{10}(\text{SOC}) + 0.28\ (\text{clay } \%)$$

Where SOC was the soil organic carbon measured using the Walkley Black method. Unfortunately they made no mention of the soil type or geology of the soils in the study.

Krull et al. (2004) have done a major review of the potential for soil organic matter to provide cation exchange capacity and the effects of pH and ionic strength on this capacity. The major conclusions from this review were the following.

- The CEC is made up of a permanent charge component and a variable charge component. The soil organic matter provides a major part of the variable charge component of the CEC.
- The carboxyl groups in the soil organic matter are a major source of the negative charge that contributes to CEC (Oades et al 1989).
- Soil organic matter contributes as much as 70% to the ECEC in Australian Oxisols (krasnozems) (Moody 1994).
- Emerson and McGarry (2003) in examining soil organic matter under trees identified an uncharged component of the soil organic matter in a sodic Hydrosol that did not contribute to the CEC. They also found the charged part of the soil organic matter that contributed to the CEC which they hypothesised was derived from the lignified portion of feeder roots.
- Parfitt et al. (1995) found that most of the CEC attributed top soil organic matter was from carboxyl functional groups. They also suggested that a critical limit of 2% of soil organic carbon was necessary for soil organic matter to have an appreciable effect on CEC. A further observation was that the presence of soil organic matter reduced the CEC attributed to smectite which suggests that the soil organic matter has some interaction with smectite to reduce the availability of exchange sites. Tan and Dowling (1984) made a similar observation on a Houston Black Vertisol.
- Lopes and Cox (1977) observed that at pH < 5.0 there is no relationship between CEC and soil organic matter. However at pH above 5.5 there was a markedly increasing relationship between CEC and soil organic matter. This perhaps confirms that the dissociation or ionisation of carboxyl groups in soil organic matter does not occur until pH is above 5.5.
- The suggestion to use surface area as a predictor for CEC (Curtain and Smilie 1976) is consistent with the suggestion of Wiklander (1969) that much of the negatively charged sites on the soil organic matter that are available for CEC, are attached to the surfaces of soil particles.
- Slattery et al. (1998) in a comprehensive study accounted for the cation exchange capacity associated with the different fractions of the soil organic matter and the different buffering capacity provided by the fractions in the acidification of a cropping soil in north eastern Victoria. They found that the humic and fulvic acid provided significant cation exchange as well significant buffering capacity against acidification.

3.2.3 Conclusion

It is clear from the published information that soil organic matter has a large effect on the cation exchange capacity of soils. However it is complex and is dependent on the texture of the soil and the clay minerals of the soil. The pH of the soil also affects the contribution that soil organic matter can make to the soil. This makes the widespread development of soil acidity of considerable concern apart from the known problems with acidity. If the soil pH falls below 5.5, it appears that the contribution of soil organic matter to CEC falls to negligible levels

3.3 Soil Acidity and Soil Organic Matter

3.3.1 Soil Acidity

The optimum range of pH for plant growth is from 6.0 to 8.0. A major problem for many agricultural areas is increasing acidity associated with pasture improvement and the removal of agricultural products from the landscape (Williams 1980; Slattery et al. 1999; Upjohn et al. 2005; Fenton and Helyar 2007). Soil acidification has resulted in soil pH_{water} levels falling below 5.5 across a wide area and at these pH levels plant growth and yields decline. In many soils when the pH_{water} falls below 5.5, Al and Mn toxicity occurs as these cations come into solution from the minerals in the soil. Al and Mn in solution inhibit the growth of plant roots, especially the growth of finer plant roots. Whether Al^{3+} and Mn ions appear in solution at low pH depends on the soil type. For example , when the soil pH_{water} falls below 5.5 Al and Mn will not appear in the soil solution in some soils such as those that are extremely weathered and consist predominantly of siliceous sand like the mallee sands of western Victoria and parts of Western Australia (Slattery et al 1999) . The highly developed podzolic soils and the krasnozems with abundant soil minerals have large stores of Al in their crystalline structures and can release large amounts of Al (Fenton and Helyar 2007).

The measurement of soil pH is usually done on a 1:5 calcium chloride solution. A general correction to convert pH_{water} to pH_{CaCl} is (Slattery et al 1999):

$pH_{CaCl} = pH_{water} – 0.84$

For a range of practical and theoretical reasons discussed in Slattery et al. (1999), pH_{CaCl} is the preferred measurement method for soil pH especially for pH values around those concerned with soil acidity. This means that Al and Mn begin to appear in the soil solution and begin to affect plant roots at pH_{CaCl} 4.8.

3.3.2Acidification Process

The acidification of agricultural soils is driven by the following processes (see Fenton and Helyar 2007):

- Nitrate leaching - legumes by fixing N and adding N to the soil to form nitrates can be acidifying. N fertilizers can also be acidifying. Use of N by plants can reduce nitrate leaching.
- Removal of cations such as Ca^{2+} and Mg^{2+} in agricultural products. Hay is especially acidifying because of the removal of large amounts of biomass.
- Build-up of organic matter can make the soil more acidic with the formation of organic / humic acids (Conyers et al 2012).

The acidification does not affect all soils but is generally concentrated in soils with sandy to medium textured surface soils which includes areas with the following soils where annual average rainfall is more than 500 mm per annum (Fenton and Helyar 2007):

- Kurosols – strongly weathered podzolic soils, soloths
- Red and Yellow Chromosols – red and yellow podzolic soils, red-brown earths, non-calcic brown soils
- Red Kandosols – red earths
- Sodosols – Solodic soils
- Ferrosols - krasonzems
- Sandy Orthic Tenosols – earthy sands

- Less fertile Dermosols - red and yellow podzolic soils, red-brown earths, non-calcic brown soils

The expected acidification rates of different farming systems are shown in Table 3.3. These give a general indication of the rates of acidification that are occurring in the soils in eastern Australia.

3.3.3 Buffering Capacity of soils against acidification

The supply of N and nutrients to the plants and the leaching of nitrate from the soil is an overall acidification process with protons being added to the soil. If this is not balanced by the return of the plant biomass to the soil, the net result is the acidification of the soil. The capacity of the soil to absorb this acidification process is its buffering capacity. The buffering capacity of the soil is an important aspect of soil health as it can give some stability to soil pH and prevent reduction in soil pH (Krull et al 2004). The buffering capacity is defined as the resistance of the soil to changes in pH when acid or base are added. From a more practical viewpoint the buffering capacity can be defined as the amount of lime required to affect a unit increase in pH (Slattery et al 1999). The buffering capacity is related to clay content and type, soil organic matter content, CEC and initial soil pH.

The buffering capacity of the soil is driven by similar factors as the soil CEC. That is:

- The clay or soil minerals which vary in the net negative charge which arises from the crystalline structure of clay or soil minerals.
- Humus in soil organic matter – the functional groups especially the hydroxyl, phenolic, acidic, amine and amide (Krull et al 2004).

The buffering capacity of soil organic matter may exceed that of clay on a per gram basis, but there may be less of it in soil.

Recorded buffering capacities for 27 field trials in Queensland were in the range 1.2 to 7.1 t of lime /pH unit (Aitken et al. 1998 reported in Slattery et al. 1999). These values equate to pH increases of 0.14 to 0.82 pH units per tonne of lime. These values can be applied in reverse, and using the Table 3.3. An annual pasture on the southern slopes with more than 500 mm rainfall will take 1.2/0.2 yrs to fall 1 pH unit which is 6 years for the soil with the low buffering capacity, based on the acidification rate of 200kg/ha/yr from Table 3.3. For the soil with the high buffering capacity the fall of 1 pH unit will take 7.1/0.2 which is 35.5 years.

A feature of many of the soils subject to acidification is that they are highly weathered (Fenton and Helyar 2007).

Slattery et al. (1998) investigated acidification on a cropping soil in northern Victoria and looked at the relative importance of the humus fraction in providing buffering capacity against acidification. They found that the fulvic acid was the most effective followed by humic acid and then the water soluble organic acids. A difficulty was that the transfer of protons to the organic matter tended to result in the movement of the acidification into the subsoil. The organic acids, especially fulvic acid, were leached into the subsoil to depths of 40 cm or more. The organic acids also tended to remove some of the Al^{3+} from solution preventing it having a toxic effect on plant roots.

3.3.4 Pedotransfer function for buffering capacity

The potential for soil components to buffer against changes in pH is summarised in Table 3.4. It is clear that soil organic matter is one of the more effective components in the soil to buffer against changes in pH. To predict the buffering capacity of the soil it is necessary to know the clay content, the type of clay minerals including the content of Fe and Al oxides and hydroxides, and the amount of soil organic matter. In more acidic and highly weathered soils where clays are dominated by kaolinite and illite and Fe and Al oxides and hydroxides, the soil organic matter becomes more important to provide buffering capacity. This is what Aitken et al. (1990) found this for a range of 40 more acidic, highly weathered soils in Queensland. The developed a pedotransfer function to predict the buffering capacity for these kinds of soils.

$BCpH = \rho * (0.955 * SOC\% + 0.011 * clay\ \%)$
where:

- BC_{pH} is the buffering capacity to pH change expressed as tonnes of lime / unit of pH change
- ρ is the bulk density of the soil for 0 – 15 cm in tonne/m^3
- SOC% is the soil carbon % for 0 – 15 cm
- Clay % is the clay % for 15 cm.

The PTF is not applicable for soil that has clays with higher CEC's. The ECEC's in the data set for the PTF ranged from 0.2 to 14 cmol(+)/kg and clays 1 to 77%. This indicates that the clay minerals present are largely kaolinites and Fe and Al oxides and hydroxides.

Helyar et al. (1990) summarised the relative effects of buffering capacity of soil to changes in soil pH in a single table for the soils of NSW based on results from the soil testing service of NSW Agriculture (see Table 3.4). It is clear that soil organic matter has a large impact on the buffering capacity. The buffering capacity of the soil organic matter arises from the hydroxyl, carboxyl and phenolic functional groups on the organic compounds. The actual buffering capacity of soil organic matter varies from 4 to 10 kmol H^+/ha/10 cm/pH unit (Helyar and Porter 1989) but an average value of 4.2 is assumed as being representative (Helling et al 1964). The buffering capacity of clays for each 1 per cent clay can be estimated at 0.6 for kaolinite, 2.5 for illite, and 5.6 for montmorillonite kmol H^+/ha/10 cm/pH unit. An average of 2.0 is the assumed average for soils in the areas of acid soils in NSW (Helyar et al 1990). On this basis Helyar et al (1990) propose the following PTF to predict he buffering capacity of soils to changes in pH:

BC_{pH} (kmol H^+/ha/10cm/unit of pH) = 4.2 * SOM + 2.0 * Clay %
Or
BC_{pH} (tonnes of lime/ha/10cm/unit of pH) = 0.05*[4.2 * SOM + 2.0 * Clay %]

Conversion based on the assumption that 2 moles of H^+ are consumed by 1 mole of calcium carbonate. 1 mole of calcium carbonate is 100.09 gms, therefore 1 kmole of H^+ is equivalent to 0.5 (100.09) kmole of calcium carbonate which is approximately 50 kg.

This is a slightly different PTF to Aitken et al. (1990) and gives slightly different estimates especially for sandy soils. This indicates the variation in buffering capacities of the soil organic fractions and the clay minerals fractions of the soils from the Queensland data set (Aitken et al 1990) and from the

NSW data set (Helyar et al 1990) (Figure 3.2). This is perhaps not surprising as Helyar et al. has given the expected large range in the buffering capacities for both clay fraction and the soil organic matter fractions. Any PTF for BC_{pH} and soil organic matter should show a band rather than single line. A good knowledge of the types of clays expected and the characteristics of the soil organic matter is required if PTF's are to be used to predict the buffering capacity to pH changes.

Regardless of the results, the importance of soil organic matter to the buffering capacity of soils to soil acidification is established.

An important effect of soil organic matter in acid soils with soil $pH_{water} < 5.5$ is to absorb some of the Al^{3+} ions from the soil solution reducing the potential toxic effects of these cations. Hargrove et al. (1981) reported in Thomas (2002) reported an 85% reduction in the amount of Al in the soil solution at pH_{water} 4.0 and a 98% reduction at pH_{water} 5.0 by increasing soil organic matter levels to 10%. Aitken (1992) observed significant amounts of Al bound to soil organic matter and it actually contributed to the pH buffering capacity.

3.3.5 Acidifying effects of soil organic matter

A recent study by Conyers et al. (2012) indicated that ultimately the accumulation of soil organic matter has an overall acidifying effect on the soil. However the review by Krull et al. (2004) suggests that much depends on the nature of the organic matter added to the soil and the management history to which the soil is subjected. While soil organic matter is potentially acidifying, it also has the capacity to buffer against acidification. Some soil organic materials and therefore plant materials are more acidifying than others. There does appear be a need to develop guidelines on the management of soil organic matter around this issue, as while there are potential benefits of higher levels of soil organic matter, there is also some caution needed.

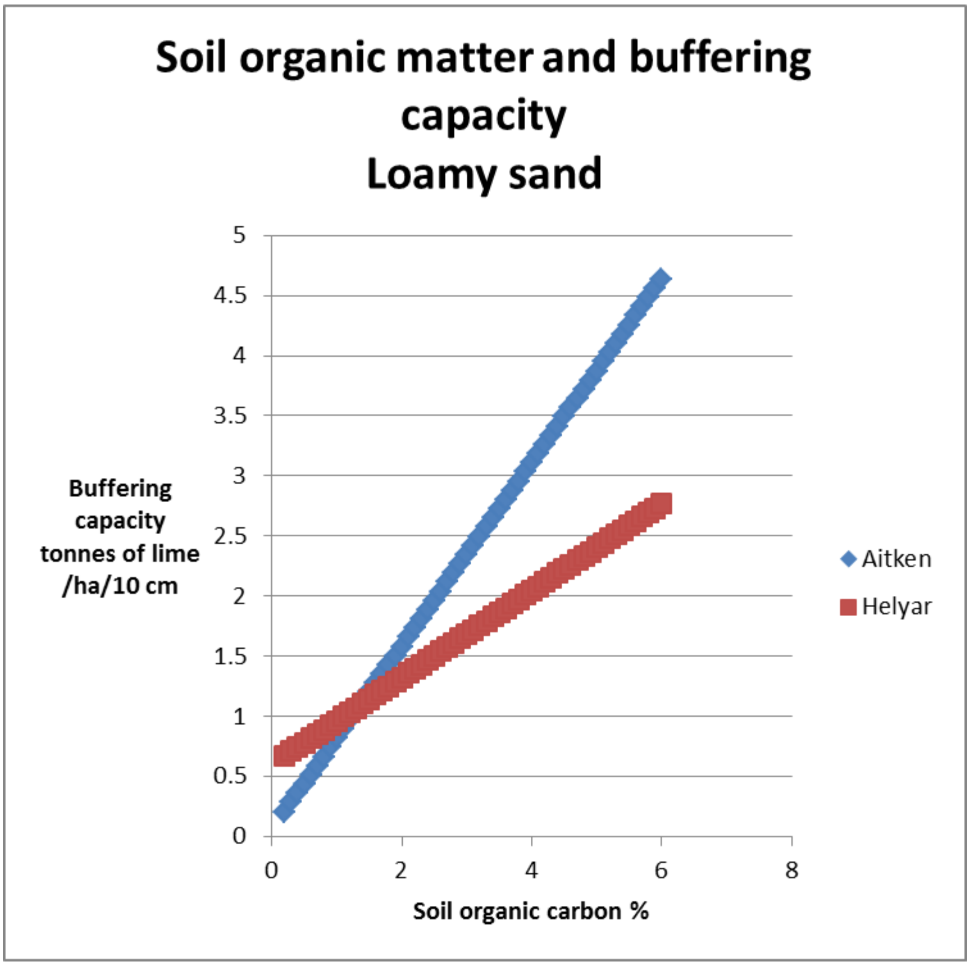
Table 3.3. The amount of lime needed to neutralise the acidification caused by the farming systems in New South Wales. Assumptions are sufficient P added to maintain 20% clover in pastures, crops are 60 to 80% of maximum yield, and soil pH in $CaCl_2$ is > 5.1. (Adapted from Table 13.8 in Fenton and Helyar 2007).

Farming System	**Lime requirement (kg/ha/yr)**
Plains, less than 500 mm annual average rainfall	
Crop or crop/pasture	75
Coast and Tablelands	
Perennial pasture	100
Southern Tablelands	
Perennial pasture	150
25% crop, 75% perennial pasture	175
Annual pasture	200
25% crop, 75% annual pasture	250
Southern Slopes more than 500 mm annual average rainfall	
Perennial pasture	150
Annual pasture	200
50% crop, 50% annual pasture	250
crop	300
Irrigation or more than 1000 mm annual average rainfall	
Less than 300 kg acidifying nitrogenous fertiliser /ha/yr	450
More than 300 kg acidifying nitrogenous fertiliser /ha/yr	1000

Table 3.4. Soil Buffering capacity to pH in relation to texture and soil organic matter for soils in NSW. (Adapted from Helyar et al. 1990).

		BC_{pH} kmol(H^+)/10 cm/ha			
Soil texture	Clay %	Soil organic matter (%)(g/100g)			
		0	1	2	4
Clay	70 (40 – 100)	140	144	158	157
Silty clay	50 (40 – 60)	100	104	108	117
Sandy clay	45 (35 – 55)	90	94	108	117
Silty clay loam	34 (27 – 40)	68	72	76	85
Clay loam	34 (27 – 40)	68	72	76	85
Sandy clay loam	28 (20 – 35)	56	60	64	73
Silt	6 (0 -12)	12	16	20	29
Silt loam	14 (0 – 27)	28	32	36	45
Loam	17 (7 – 27)	34	38	42	51
Sandy loam	10 (0 – 20)	20	24	28	37
Loamy sand	8 (0 – 15)	16	20	24	33
Sand	5 (0 - 10)	10	14	18	27

Figure 3.2. Effect of clay and soil organic matter on buffering capacity to changes in soil pH for pedotransfer functions from two data sets of Australian soils (Aitken et al. 1990 and Helyar et al. 1990). Note soil organic carbon % is equivalent to g/100g.



3.4 Capacity of Soil Organic Matter to form Complexes and Ligands with Cations and Anions

3.4.1 Chemical Activity of Soil Organic Matter

Soil organic matter is chemically active undertaking simple ion exchange with the simple cations Ca^{2+}, Mg^{2+}. N^{+} and K^{+} by reaction with the COOH groups. It can also undertake exchange with multivalent cations such as Cu, Zn, Mn and Co as well as Cd, Pb and Hg. However because of their more complex electron structures these cations can form metal complexes with the molecules in the soil organic matter which are more stable than the simple ion exchange mechanisms (Stevenson 1982). This chemical activity of soil organic matter can also be important for the complexing of pesticides or organic pollutants. The adsorption of Atrazine onto soil organic matter is an example (Stevenson 1982).

As mentioned in Section 3.3, soil organic matter can also form stable bonds with Al and remove it from the soil solution.

In a major review of the interaction of soil organic materials and the soil mineral fraction, Huang (2004) identified that organic compounds can have an influence on the reactions between Al and Fe oxides and the soil solution and on the formation, transformation and surface properties of the oxides of Al and Fe. When Al^{3+} is in solution are three possible pathways for precipitation including the formation of ordered gibbsite, the formation of poorly ordered Al hydroxides or the formation of Al organic acid complexes. The amount of Al in each pathway depends on the nature and concentration of soluble inorganic and organic ions and clay minerals and humus in the soil at the time. In general the presence of organic compounds retard crystallization. A further important effect is that the organic compounds can affect the surface properties of Al transformation products which have a large influence on the charge characteristics and physical properties of the Al compounds. The surface of the minerals is the region of interactions with the soil solution, organic and inorganic particles, plant roots, microorganisms and other soil biota.

A similar effect is suggested for the formation of Fe oxides and hydroxides in soils (Huang 2004). In fact the suggestion is that no hematite will form in soils if the soil organic matter is high. Unfortunately, no suggested values of soil organic matter content are given.

3.4.2 Phosphorus and soil organic matter

Phosphorus can exist in a series of different pools in the soil but only one of these is readily available to plants, which is the solution pool in which the water soluble orthophosphates occur. The pools include (Moody and Bolland 1999; Stevenson 1982):

- The soluble pool which includes the orthophosphates of $H_2PO_4^-$ and HPO_4^{2-} . These forms of P can be readily taken up by the plant roots from the rhizosphere - very high availability
- The soluble organic compounds such as nucleic acids and sugar phosphates which can also be taken up by the plant roots but to a lesser extent - high availability
- Soil organic matter as residues – phospholipids, inositols, insoluble soil organic compounds – decomposition required to mineralise the P in this pool – moderate availability
- Microbial biomass – decomposition required to mineralise the P in this pool - moderate availability
- P adsorbed onto soil particles including clays and Fe and Al minerals and calcium carbonate – low availability

- Crystalline P fixed into Ca-P minerals of extremely high stability – very low availability

One of the potential advantages of soil organic matter is that it can delay the movement of P into the crystalline forms of P which have very low availability (Alvarez et al. 2004 reported in Krull et al 2004). Toreau et al. (1988) also reported that soil organic matter resulted in a reduction in the sorption of P. However other studies have produced different results with Saunders (1965) showing increased sorption with more soil organic matter and Moody and Stanley (1979) reporting no effect in basaltic soils. In a major review, Simpson et al. (2011) point out that the soil organic matter is a major reserve of P for plant growth and suggest that soil organic matter is one pathway for transferring the reserve of N and P built up during the pasture phases for use in a cropping phase. They also emphasised that in soil with high P sorption, managing to store P in plant residues is a method to slow strong P sorption reactions. McLaughlin et al. (2011) has identified soil organic matter as an important source of P for Australian agriculture.

The capacity of soil organic materials to decrease P sorption was shown by Schefe et al. (2007) using lignite and compost. The P sorption of di-ammonium phosphate (DAP) was reduced from 265 mg P/kg in the untreated soil, to 234 mg P/kg for the soil treated with lignite, and 244 mg P/kg for soil treated with compost and an effective addition of soil organic carbon of 2.5%. The soil was an acid Yellow Dermosol from Rutherglen Victoria.

The capacity of organic acids to prevent the fixation of phosphate by iron and aluminium was demonstrated by Swenson et al. (1949), using the additions of manure to soils.

There is then some evidence that soil organic matter can enhance the availability of P in soils by preventing the sorption of P into less available forms.

3.5 Soil Organic Matter and Effects of Salinity

3.5.1 Soil salinity

Salinity is severe problem for the maintenance of soil organic matter. As shown by Wong et al. (2008), a saline site can often start having very low levels of soil organic matter because the high salinity virtually shuts down the input of biomass and organic material into the soil so that the soil organic matter runs down (see Figure 3.3). Saline sites are often devoid of vegetative and surface cover so can lose soil and soil organic matter in erosion as well.

However, soil organic matter can assist soils with salinity in a number of ways.

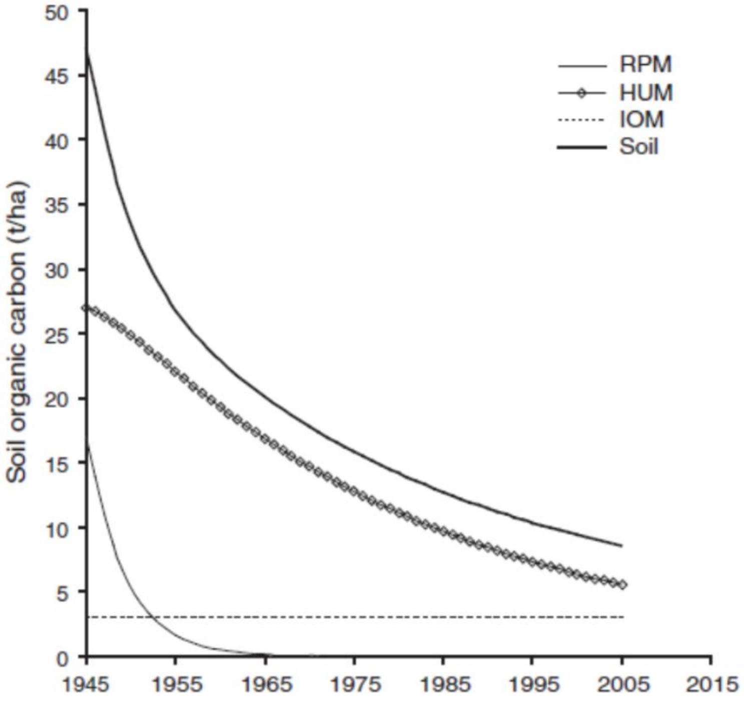
- Providing surface cover can prevent erosion and increase surface infiltration. This will not however rehabilitate a soil which has a severe underlying salinity problem. If the electrical conductivity levels and /or the pH values of the surface soil remain high, surface cover alone will not rehabilitate a site, especially if the source of the salinity is continuing to operate.
- The soil organic matter in the soil can complex and absorb some of the salts and anions, inactivating them from the soil solution and preventing them from interacting with or harming any plants. Clearly there is a limit to which the soil organic matter can carry out this function but this will be related to the cation exchange capacity of the soil organic matter.
- The soil organic matter can increase the water holding capacity of the soil and so reduce the impact of salinity on the plants. The effect of this can be seen in the values used to convert

electrical conductivity measured in 1:5 soil:water extract (EC1:5) to electrical conductivity saturation extract (ECsat) (Slavich and Petterson 1993). The saturation water content is the critical value and many pedotransfer functions tend not to give this because there is much more interest in the values for field capacity and permanent wilting point which give plant available water and it is difficult to predict. The conversion factor “f” for EC1:5 to ECsat is given by:

- f = 2.46 + 3.03 /saturation water content.
- The effect of soil organic matter on the saturation water content can be estimated from pedotransfer functions (see Section 2.4). However a series of relationships developed by Shaw (1991) to predict the saturation percentage for estimating the electrical conductivity of the saturation extract from the 1:5 electrical conductivity, suggests that the amount of water held at saturation is dominated by clay content and cation exchange capacity, with soil organic matter playing a minor role.
- The strong relationship between bulk density and soil organic matter described in Section 2.3 indicates that soil organic matter can influence the amount of water held in the soil at saturation. The higher the soil organic matter levels the more water held in the soil at saturation and so soil organic matter has the capacity to attenuate to some extent the effects of salinity for plant growth and soils. The actual electrolyte experienced by plants may be reduced in soils that are higher in soil organic matter.

The conclusion is that while soil organic matter can partially attenuate the effects of soil salinity, soil organic matter will not be able to overcome or rehabilitate major salinity problems unless the primary source of the salinity problem is solved.

Figure 3.3. Fate of soil organic carbon under high salinity levels. Taken from Figure 5 of Wong et al. (2008). The measured soil carbon stocks were 10t/ha to 30 cm and the initial stocks about 45 to 50 t/ha based on measurements from surrounding pastures. Modelled soil organic carbon values for a scald assuming scald formation in 1945. Shown are total soil pool (soil), resistant plant material pool (RPM), humified organic matter pool (HUM) and the inert organic matter pool (IOM). This model assumes all carbon losses are a result of decomposition and does take account of any carbon losses associated with erosion.



Summary of Key Points – 3. General Soil Chemical Properties

- The nutrient cycling from soil organic matter is a major source of nutrients for plants. Soil organic matter consists of a balance of nutrients C:N:P:S approximately in the 100:10:1:1 although this can vary a little depending on the origin of the organic matter and the history of decomposition. To accumulate soil organic matter in soils requires the addition of nutrients as well as a carbon source. The availability of nutrients from soil organic matter depends on the fraction of soil organic matter in which it is stored and the activity of the soil microorganisms. The newly added plant material and particulate organic matter are the most available, the humus next and the inert soil organic carbon the least. Other key outcomes.
    - The nutrients stored in the soil organic matter are in the order of 3000 to 5000 kg/ha of N, 300 to 600 kg/ha of P and 300 to 600 kg of S. This is not readily available of course. This store is in a high state of flux as the soil organic matter is being decomposed and increased with about 2 to 4% being tuned over annually. A single wheat crop requires about 100 kg N and 10 kg of P.
    - About 90 to 95% of the N in the soil is held in the soil organic matter store
    - About 40% of the potentially available P in the soil is stored in the soil organic matter.
    - About 90 to 95% of the S is in the soil organic matter.
    - The movement of the nutrients between the soil organic matter store and the soil solution where it is available to plants is the function of soil microorganisms
    - Soil organic matter keeps the P in a form that it can become available to plants
- Cation exchange capacity is the capacity of the soil to hold cations including Ca, Mg, Na. K and micronutrients such as Cu and Zn.
    - Cation exchange capacity of the soil is provided by the clay particles and by soil organic matter. In clay rich soils most of the cation exchange capacity is provided by the clay. In sandy soils, much of the cation exchange capacity is provided by soil organic matter.
    - The cation exchange capacity of the soil organic matter fraction is pH dependent and is highly variable between the different soil organic matter fractions. Most of the cation exchange capacity is provided by the fulvic and humic acids of the humus fraction. The particulate organic carbon and inert carbon tend to have very low cation exchange capacities.
    - Various studies confirm that soil organic matter provides the cation exchange capacity of soils with lower clay contents. The effectiveness of the soil organic matter to provide the cation exchange capacity depends on the nature of the soil organic matter and the amount of humus and the pH of the soil. When the soil falls below 5.0, many soil organic materials will have very low cation exchange capacity.
    - A critical limit of at least 2% soil organic carbon is recommended when soil organic matter provides the major part of the soil's cation exchange capacity.
- Soil acidification is a problem across a wide range of agricultural soils in south eastern Australia. It is caused by nitrate leaching, removal of cations in agricultural produce and the in-situ accumulation of soil organic matter.
    - The in-situ accumulation of soil organic matter tends to make the soil more acidic as the amount of humus with the humic acids and fulvic acids accumulate in the soils

  - The soil organic matter has a buffering effect against the general acidifying process despite being slightly acidifying itself. There are a number of pedotransfer functions available to predict this effect and they tend to be regionally specific to account for the specific properties of local soils. The buffering capacity of the soil organic matter varies depending on the nature of the soil organic matter.
  - While soil organic matter can provide some buffering capacity against acidification, this only delays the onset of acidification and does not solve the fundamental problem.
  - To provide a credible buffering capacity a level of about 2 % soil organic carbon is probably required.
- Soil organic matter has components that are chemically active that can combine with cations and anions, removing them from the soil solution to form complexes and ligands. Species include Cd, Pb, Hg, P, and a range of biologically active pesticides and herbicides.
  - Soil organic matter can interact with Al and Fe, preventing these forming stable crystalline forms. This can be important for soil structure stability.
  - Soil organic matter can form complexes with P, preventing the P being removed from the soil solution into solid crystalline forms that are no longer available to the plants. Hence it can prolong the availability of P.
- Salinity, when severe, can virtually shut down plant growth and the input of biomass material into the soil causing the rundown of soil organic matter in the soil
  - Soil organic matter can attenuate the effects of salinity when it is not too severe by:
    - Complexing the salts in the soil and removing them from the soil solution
    - Increasing the water holding capacity of the soil and reducing the salt concentration in the soil solution.
    - These beneficial results are likely to require high levels of soil organic carbon, probably more than 2%.

4. Soil Organic Matter and Soil Biology

4.1 Introduction

Soil biology is fundamental to the processes in soils that develop and maintain soil structure, is responsible for the continued cycling of nutrients to ensure their availability for plants and can suppress the occurrence of disease in plants. It is the size, diversity and activity of the micro and macro-organisms that are fundamental to these functions. An understanding of organisms in soil and soil biology is highly relevant to maintaining or increasing yields with low fertiliser input and reducing losses from soil-borne diseases in Australian cereal and pasture production (Martin 1993). Watt et al. (2006) undertook a major review of the effects of soil biology on productivity and identified the key mechanisms of how soil biological processes could affect crop productivity. Microbial biomass was a common measure of biological function in soils and Dalal (1998) has given a detailed review of the functions of the soil microbial biomass in soils, especially of its role in the recycling and supply of N, P and S to plants, but also of the difficulties of obtaining meaningful measurements of microbial biomass and the interpretation of those measurements to make predictions about soil productivity and soil management. More recently a wider range of biological measurements have been developed. These were described by Insam (2001) and Mondini et al. (2004). Insam (2001) describes the use of enzyme analysis including inflorescence, physiological profiling testing the capacity of a soil extract to degrade different substrates (eg BIOLOG), the use of biomarkers such as phospholipid fatty acids (PLFA), the use of DNA extraction in soils and improved measurement of soil microbial biomass. These new measurements soil biological populations and functions have been applied in a range of programs to increase the understanding of how soil biology can be managed to improve crop productivity and sustainability.

Soil is a habitat for large populations of microorganisms and although the microbial biomass is only a small amount of the soil organic matter < 5%, it is a most important and dynamic fraction of the soil organic matter. Much of the microbial population is associated with the surface of soil colloids as this provides an environment that is enhanced with ions, water, nutrients and organic matter (Huang 2004). The rhizosphere surrounding roots would be similar. Bacteria also produce extracellular polysaccharides that help bind the cells to the surfaces. The amount of soil organic matter can influence how easy and effective it is for microorganisms to bind to the surface of colloids and soil particles (Huang 2004).

By adhering to soil colloids microorganisms can obtain access to nutrients or have a sink which removes metabolites. The colloids may also adsorb toxic substances such as antibiotics and pesticides which might otherwise inhibit the growth of the microorganisms. With time, the microorganisms can become enveloped by the colloids which can restrict the access of oxygen but might also protect them from desiccation as the soil dries or from predators and from decomposition by other microorganisms. With pore sizes < 6μm bacterial predators have difficulty gaining access (Huang 2004).

One important aspect that influences soil biological functionality is the capacity of mineral colloids and humic substances to sorb enzymes released by the soil microbial populations and maintain the activity of these enzymes for substantial periods of time (Huang 2004). In some cases the activity of the enzymes is enhanced when bound to the mineral colloids or the humic substances (Gianfreda and Bollag 1994).

Degens et al (2000) reported in Krull et al. (2004) introduced the concept that falling levels of soil carbon reduce the biodiversity and functional diversity of soil microbial communities which supports the results of Yan et al. (2000).

The biology of the soil environment is very complex and may be difficult to understand and manage. However recent advances in methodologies and land management practices offer opportunities for maintaining productivity with diminishing resources and for improving productivity through soil biology.

4.2 The Rhizosphere

Plant roots are considered as an integral part of soil biology and therefore the rhizosphere is important in understanding the function of soil biology in maintaining soil productivity (Foster 1993; Watt et al. 2006). The rhizosphere is linked to the root:soil interface and is the special zone near the plant root surface that has very different physical, chemical and biological features to the bulk of the soil. In simplest terms it can be considered to extend to about 1 mm out from the surface of the root (Foster 1993). However, Watt et al. (2006) suggest that the rhizosphere is very complex and varies widely depending on the plant species, the age of the plant and even within a single plant will vary with the type of root (whether tap root, seminal root or nodal root) and position along a root. Roots of different age and type have different suites of exudates and this can result in different suites of organisms. Fungi associated with a root might extend the rhizosphere some distance from the root.

The rhizosphere is usually slightly compacted after the physical action of the root growing through the soil. The chemistry is strongly influenced by several effects (Foster 1993; Watt et al. 2006):

- Depletion of soluble nutrients and cations such as N, P, S, Ca and Mg as they are absorbed by the plants and by the microorganisms growing in the zone.
- Balance reactions such as the plant adding protons to balance the uptake of the positively charged nutrient ions and cations, which can significantly reduce the pH. In other cases the exudates my include hydroxyl ions
- Depletion of oxygen and increases in carbon dioxide.
- Depletion of water as the plant absorbs water especially in times of high evaporative demand.
- Enhancement of organic compounds from the root surface including:
    - Mucilages – such as insoluble poly saccharides which act as lubricants to assist the movement of roots as they grow.
    - Exudates - metabolite products including sugars (sucrose), carboxylic anions, amino acids, proteins, phenolics and enzymes.
    - Lysates – breakdown products from old cells, especially from older roots and wheat roots after anthesis.
    - Cell walls.
    - Signal compounds that can inhibit or stimulate the growth of different microorganisms.
- The rhizosphere has a different microbial population to the bulk soil (Singh et al 2007). Often the bacteria are within 0.03 mm of the root surface (Watt et al 2006).

The enhancement of rhizosphere with organic compounds makes it a very high energy environment and it supports a rich microbial population. The microbial population within this zone has the potential to have the following effects:

- Can affect the growth and longevity of root hairs and so the uptake on nutrient and water
- Control root morphology
- Influence the uptake of trace elements especially Mb and Mn
- Affect the development of symbiotic relationships such as Rhizobia and Mycorrhiza
- Influence the infection of roots with pathogens (Raaijmakers et al 2009)
- Detoxify soils and determine the extent that roots are affected by toxins such as Al^{3+} ions.

The rhizosphere can persist for years depending on whether a plant is an annual or a perennial. Even when a root dies, it will continue to support a succession of organisms and products that can have important nutrition, disease or other effects on the roots of plants growing in the future. It can have significance in rotation systems (Watt et al. 2006). Such rhizospheres may be especially important in soil environments such as the deeper subsoils of with high sodicity where the same biopore channels maybe reused by the roots of successive plants and crops (Watt et al. 2006).

A further important aspect of the rhizosphere is that up to 20% of the net photosynthate produced by many plants can be passed through the root system into the rhizosphere. Watt et al. (2006) suggest this is 10%. While this may be a seen as a useful product for soil organic matter and the soil ecosystem and is also a potential loss of productivity (Martin 1993).

4.3 Biological Processes with the Potential to Contribute to Crop Productivity

Biological processes that can contribute to crop productivity include nutrient cycling and nitrogen fixation; aggregate stability and soil structural development; and disease suppression (Table 4.1) (Watt et al. 2006). Processes that relate to soil biological processes and have been used specifically to improve productivity in land management practices include symbiotic nitrogen fixation and the use of crop rotations to control plant diseases and inhibitory organisms. These processes operate in the rhizospheres around plant roots. The improvement of soil structure using plant roots and the associated microorganisms has been widely used to improve soil structural stability (Oades 1993).

The management of soil biological processes is an area of investigation with a large potential to increase productivity. The specific effects of the biological processes on productivity are discussed below.

4.3.1 Nutrient Cycling

Soil organic matter is a major source of nutrients for plants in soils and when soil organic matter is decomposed or mineralised these nutrients can become available to plants. The fixing of N from the atmosphere is a major source of N for most soils and this is undertaken by microorganisms in the soil or in symbiotic relationships with legumes and some other plant species. The microbial population has to undertake a range of functions to ensure that adequate mineralization of the available nutrient pools takes place.

(i) Mineralisation of soil organic matter

The soil organic matter can hold a large amount of the available nutrients in the soil. However much of that nutrient is not available to plants while it is held in an organic form, either because the organic form is insoluble, such as many organic P compounds, or because it is held in the organic material itself. The process of mineralisation is required to convert the nutrients into a form in which they are available to plants (see Figure 3.1). The soil biology or soil microorganisms are primarily responsible for this process. The implication of this is that the availability of nutrients from soil organic matter is the consequence of the balance between the formation of soil organic matter, which absorbs or immobilises the nutrients, and mineralisation which releases the nutrients (Janzen 2006). As Probert (1993) has described the traditional pasture-crop rotation system fitted well into this balance, with the build of soil organic matter in the pasture phase and the subsequent mineralisation of soil organic matter and release of nutrients in the cropping phase. Under current prospective systems that require a long term increase or maintenance of soil organic matter, the mineralisation of soil organic matter to release nutrients requires more subtle management systems. The rates of immobilisation of nutrients and mineralisation vary through the year and this offers some options for the management of the nutrients available for plants. The mineralisation of nutrients by decomposition of crop and plant residues or by decomposing of existing soil organic matter may have different implications for the maintenance and/or increase in levels of organic matter in soils (see Table 4.1). If the nutrients can be sourced from the decomposition of the input plant residues rather than existing soil organic matter, this reduces the pressure on existing soil organic matter.

The rate of decomposition of organic materials and the amount of nutrient that becomes available to plants depends on the nature of the organic material being decomposed and the microbial populations in the soil. As discussed in Section 3.1.3 the amount of N made available to plants depends on the C:N ratio of the organic material being decomposed. If the organic material is low in N, the soil microorganisms will take up more N than they release to the soil solution.

The mineralisation of soil organic matter can be inhibited as the soil organic matter is stabilised in the soil. The soil organic matter can be stabilised by the following: (Six et al. 2002)

- Biochemical stabilisation – resistance of the organic compounds to breakdown or attack by soil organisms as occurs with lignin and polyphenols.
- Stabilization by association with silt and clay particles – silt and clay particle provide some protection against the access of microorganisms to the organic materials.
- Physical protection within aggregate structures – the inside of aggregates are protected from soil microbial organisms to some degree, which cannot gain access or there is a lack of oxygen.

The soil organic matter, especially the humus fraction has a relatively long residence time and usually averages 36 to 63 years (Six et al 2002), depending on climate, soils and land use. An important mechanism which has been proposed for the formation of aggregates which has implications for the mineralisation of soil organic matter and the recycling of nutrients is that the macro-aggregates are often composed of more stable micro-aggregates which have a higher concentration of soil organic matter (Six et al 2002). The soil microbial population may be prevented from having access to any soil organic matter within micro-aggregates that form into larger macro-aggregates, and oxygen levels may be low in the micro-aggregates inside the macro-aggregates. Given this potential anaerobic environment, nitrous oxide emissions may occur (Six et al. 2002).

More recently Schmidt et al. (2011) have argued that the microbial population of soil can adapt to any quality or composition of soil organic material so that even organic materials high in lignin and aromatics can be decomposed rapidly if the appropriate population of microorganisms is present. Traditionally, organic materials high in lignin and aromatics were considered to be resistant and to have a high residence time. The implication is that the microbial processes such as nutrient cycling in soils can be managed by manipulating the soil microbial populations. The land management actions to do this however were not well defined and there may be practical difficulties in managing the soil microbial populations but the possibilities are there.

The mineralisation of crop and pasture residues is a major importance in the supply of nutrients, especially N, for crop productivity (Probert 1993; Angus et al. 2006; Herridge 2011). The rates of mineralisation depend on the substrate materials and on the biological populations breaking them down. The amount of soil moisture is also a critical factor. Angus et al (2006) noted that for every extra 1% volume of water in the soil there is an additional 0.054 kg/ha/day of N. This becomes relevant when stubble retained treatments have soils with more moisture than stubble burnt treatments. In a series of 20 pasture management systems on the south west slopes of NSW, Angus et al. (2006) estimated that the overall mean mineralisation rate was 0.56 kg N/ha/mm of rainfall. This involved the mineralisation of residues from annual pastures, lucerne, perennial grasses and from combinations of lucerne and perennial grass. This extended for up to 4 years following the pasture phase. Differences in N mineralisation rates between cropping treatments were recorded but these were attributed to differences in soil moisture contents and yield differences under different tillage and stubble treatments rather than differences in soil biology or the microbial populations. An interesting finding was that pastures containing perennial grasses had high N mineralisation rates than those with annual grasses and this was considered to possibly be the result of moving subsoil N to the surface by the perennial grasses.

The possibility of different microbial populations resulting in different rates of mobilisation and mineralisation of N has been considered by Ryan et al. (2006). They showed increased accumulation of N in the 0 to 0.05 m layer of soil for a short time under brassica crops (canola) compared to cereals. One possible cause for this was a change in the microbial populations but other possible causes such was the low C:N ratio of the canola leaf material (8 – 12), which can fall and create litter at this time.

Stubble management was shown to affect the mass and activity of microorganisms by Hoyle and Murphy (2006). Stubble burnt treatment demonstrated lower N mineralisation rates and CO_2 evolution than retained stubble treatments. However, despite a detailed investigation using substrate analysis (CLPP) a difference in the microbial populations could not be statistically demonstrated. Biochemical enzyme analysis did demonstrate that there were differences in activity between stubble retained and stubble burnt treatments, with the stubble retained treatments having higher cellulase and β - glucosidase activity. These two enzymes can increase the rate of breakdown of cellulose and release of N from organic materials. It would seem the stubble burnt treatment have less capacity to breakdown plant residues. The possibility of preconditioning microbial populations for specific types or quality of substrate in order to increase the rates of decomposition and N supply was demonstrated by Cookson et al. (1998) for wheat straw.

(ii) Nitrogen Fixation

Nitrogen fixation by microorganisms is one of the most important functions in terrestrial ecosystems and is essential for the nitrogen cycle. The nitrogen fixing organisms operate both in in symbiosis (Rhizobia) with the plants and there are also some free living organisms that fix nitrogen from the atmosphere. One of the problems with fixing nitrogen from the atmosphere is that to convert N_2 to NH_4^+ or mineral nitrogen is highly energy intensive.

Rhizobia and legumes

The major nitrogen fixers that operate in symbiosis in plants are the Rhizobia. The amounts of N fixed and the amount of N that can be added by other soil sources and crops by this process are discussed in Section 3.1.3.

Nitrogen Fixation by Free Living Bacteria

A continuous supply of N is required to maintain agricultural production. A potential valuable source of N is that supplied by free living N-fixers. N fixation by free living bacteria does occur and it has been estimated at the rate of 0 to 15 kg N/ha/yr (Peoples 2002), although occasional rates up to 80 N/kg/ha/yr have been suggested. The bacteria involved include *Azotobacter , Clostridium, Chromotorium and Chlorobium* (Burris 2002). The rates of N fixation by free living bacteria are much lower than those achieved by bacteria in symbiotic relationships with plants, especially the *Rhizobia* in legumes (commonly 50 to 200 kg N.ha/yr). N fixation is highly energy demanding as the free energy demand to convert N_2 to NH_4^+ is very high. If the N fixing organism obtains an alternative source of N they tend to turn off the N fixing process. There are also conflicting demands in the N fixing process. An oxygenated environment is required to supply the ATP to provide sufficient energy for the process, but the nitrogenase enzyme for the process is denatured by exposure to oxygen (Peoples 2002). It takes 16 ATP molecules to convert 1 N_2 to $2NH_4^+$ (Burris 2002). It is estimated that a free living bacteria would need to decompose 50 to 200 kg of soil organic matter to fix 1 kg of N (Stevenson 1969). A symbiotic relationship with a plant maximise the chances of meeting these conflicting demands although several free living organisms also do, but the production rates of N fixation are low.

Given the high energy requirements for nitrogen fixation, the rhizosphere is a favourable environment for N-fixing organisms. Rovira (1963) inoculated the rhizosphere of wheat, maize and tomato with free living N-fixing bacteria and showed some positive responses to these when the plants were grown in low nutrient sand.

Gupta et al. (2006) in a review of soil environments Australia wide concluded that there is some potential for free-living N fixers to contribute N to crops utilising stubble and plant residue from the previous crop, especially if those residues are wheat stubble or similar with a wide C:N ratio. However in some seasons at some locations there is only a narrow window of adequate moisture and available moisture for these microbial populations to operate.

The conclusion from this discussion is that the opportunities to utilise free living N fixing organisms are limited but can be a useful addition in some circumstances.

(iii) Phosphorus

Microorganisms are responsible for a large proportion of P used by plants either by the mineralisation of plant and animal residues or by the mobilisation of mineral sources of P by agents such as mycorrhiza. Organic compounds might also be responsible for preventing P from being precipitated into mineral forms that become unavailable to plants (See Section 3.4.2). About 40% of P is considered to be in held in organic form in the soil and hence potentially available if the organic material is mineralised. A small amount of P occurs as water soluble form already available to the plant (2 to 100 ppm). The arbuscular mycorrhiza are often considered to enhance the capacity of plant roots to utilise P sources in soils (Bolan 1991). This is not universally accepted as it has been difficult to gain convincing evidence in the field (Ryan et al. 2002, 2005; Ryan and Graham 2002; Ryan and Angus 2003; Watt et al. 2006), but the evidence is present that the mycorrhizal fungi can source P and supply it to the infected host roots under some conditions (Yao et al 2001; Bucking and Shachar –Hill 2005). However it is also clear that the conditions and processes under which this occurs require more investigation and understanding, and that conditions in the field are not well understood.

(iv) Sulfur

A large proportion of the S available to plants is associated with the plant and animal residues or with the soil organic matter. The S becomes available to plants when the soil microorganisms mineralise the plant materials or the soil organic matter. Plant materials differ in the capacity to supply S. Canola is a good source of S and when microorganisms decompose canola residue, significant levels of SO_4^{2-} are released into the soil solution (Singh et al 2004). A proportion of the S from the canola was immobilised as it was taken up in the biomass of the microorganisms. The occurrence of S in different pools is an example of how the soil microorganisms are responsible for transferring nutrients from one pool to another within the soil.

4.3.2 Soil Structure

Soil microorganisms, in combination with plant roots, use the processes that occur in the rhizosphere to enhance the aggregate stability of soils (Oades 1993; Tisdall and Oades 1982; Preston et al. 2001). This is discussed more fully in Section 2.2.1 but it is a role for soil biology, especially for soils that are dominated by silt and sand (Oades 1984).

When plants grow into soils, especially into subsurface and subsoils which may have a high soil strength that can restrict root penetration, the volume of soil that roots can access is minimised. The roots can be constrained to grow in existing cracks or old biopores formed by previous plant roots. These spaces become niches for root growth for successive generations of roots (Watt et al. 2006). These niches can be in the unploughed subsurface soil or in the subsoil. The biopores are a microenvironment of their own with old root material and a microbial population living on the old root systems. The organisms in these biopores or cracks can be important to stabilise these structural units but can also be important as they can mineralise the old roots to release nutrients. They may also be pathogenic. The soil surrounding the macropore can contain more microbial biomass per unit mass than the bulk soil and is able to utilise a wider range of carbon substrates and to a greater extent than the bacterial population in the corresponding bulk soil (Pierret et al. 1999).

The distribution of different pore sizes and soil texture of the soil determines the activity of the different microorganisms as the moisture content changes in the soil. The porosity will affect the aeration and the availability of moisture (Moyano et al. 2013).

4.3.4 Disease Suppression

A major part of soil biology is the occurrence of plant diseases and the suppression of diseases is a major part of the management of soil biology. As outlined in Table 4.1, there are a number of strategies for suppressing diseases using soil biological processes:

- Avoiding the supply of a host to the pathogen for a specified period to reduce the level of inoculum
- Grow a population of bacteria that outcompetes the pathogen, usually a fungi.
- Grow microorganisms that produce antibiotic exudates that suppress the pathogenic microorganisms.

A general approach to biological disease control is reviewed by Wilhelm (1973) who suggests the basic strategies for suppression of disease include crop diversity, decoy hosts which initiate germination of resisting structures which may deplete pathogens of energy reserves, and serve to protect the major host, and symbiotic fungi which occupies root cortical tissue of most plants. In addition to the traditional mycorrhizal function, this may afford protection against invasion by pathogens.

The soil borne pathogen Rhizoctonia solani AG-8 causes major yield losses worldwide With the retention of crop residues the soil can become suppressive to this disease (Barnett et al. 2006). This suppression effect was demonstrated to be a result of 3 groups of bacteria including Exiguobacterium acetylicum, Pantoea agglomerans and Microbacteria. The P. agglomerans and E. acetylicum microorganisms associated with the roots promoted the growth of infected wheat plants and soil associated Microbacteria which reduced the amount of root infected with R. solani. The objective is to understand how land management practices can promote the spread and growth of the disease suppressive microorganisms.

4.4 Methodologies for Measuring Soil Microbial Populations and Function

Recent years have seen the development of a wide range of new methods for measuring soil biology and for characterising the microbial populations in soils. Insam (2001) provides a summary of many of the new methods. A generalised summary of the methods is presented in Table 4.2 and this broadly groups the methods based on their complexity and difficulty of interpretation. Some of them are highly specialised and require well equipped laboratories and specially trained technicians.

One of the objectives of soil science is to develop soil biology tests that are the equivalent of some of the established tests in conventional soil testing such as pH, EC and Exchangeable Sodium Percentage. As Dalal (1998) and Murphy et al. (2011) suggest, microbial biomass is a difficult test to interpret based on current knowledge, although Gonzalles-Quinones et al (2011) do provide some guidelines on the interpretation of soil microbial biomass data.

The modern techniques of soil enzyme assays, DNA analysis, fatty acid analysis such as PLFA and the substrate analysis such as BIOLOG and CLPP are being applied to Australian soils and especially the cropping soils. Some of the examples and references are given in Table 4.2, but many of the studies referred to in this Section use these modern methods to characterise the soil biology and the microbial populations in the trials and fields being tested. Many of these techniques have been used in a special issue of Soil Research (Vol. 44 in 2006) on Soil Biology in Australian Farming Systems.

4.5 Microorganisms in Soils

Although often assumed as general knowledge it is potentially useful to have a general summary of the microorganisms that occur in soils.

The microorganisms in soils include the following (Allison 1973):

- Bacteria – 0.5 to 5 μm – mainly aerobic that decompose soil organic matter, includes some free living N fixing bacteria and anaerobic bacteria – other important bacteria:
    - *Nitrosomonas* that oxidises ammonia to nitrite.
    - *Nitrobacter* that converts nitrite to nitrate.
    - *Rhizobia* that form symbiotic relationships with legumes to fix N.
- Actinomycetes – fall between bacteria and fungi - form mycelium with 0.5 to 1.2 μm diameter. Largely aerobic and decompose soil organic matter, especially the fragments that make up the particulate soil organic matter (POC). Often produce antibiotics including streptomycin. Can attack many of the more resistant organic compounds but not necessarily lignin.
- Fungi – widely diverse group of organisms. Form mycelium with 5 μm diameter. Can form large mass in soil but maybe a single organism. More acid tolerant than bacteria and Actinomycetes. Fungi have lower N demands than bacteria. Can readily attack cellulose, but only a few can decompose lignin. Some of the fungi have formed symbiotic relationships with plants which seem to be for mutual benefit (myccorhiza).
- Algae – 2 - 5 μm in diameter. All autotrophic and photosynthesise. Include the blue-green algae (*Cyanophyceae*) which can also fix N from the atmosphere. Largely of minor importance in agricultural soils but can be of importance in rangelands and in initiating soil formation on bare rock surfaces.
- The Mycorrhizas including Ectomycorrhizas, Ericaceous mycorrhizas and Vesicular-arbuscular mycorrhizas are symbiotic fungi that colonise plant roots and are capable enhancing nutrient uptake of some plant species, especially P (Bowen 1993).
    - They can increase the volume of soil that plants can access nutrients and possibly moisture
    - Access insoluble sources of nutrients, especially P
    - Reduce pathogen effects on plant roots
    - The effects are often highly species and soil specific
    - Most effective in plants having “stunted” root morphologies
    - Effects very much concentrated in the native and Pinus species rather than in cereals and commercial legumes. Only limited commercial importance to date. Impacts of high fertiliser levels on mycorrhizas a potential problem.
    - More research and development required.

Commercial tests are now available to estimate the relative activity of some of these different organisms, but the meaningful interpretation of the results for production is still difficult.

4.6 Land Management and Soil Microbial Populations and Function

Land management can affect the level of soil organic matter, the soil biology and the soil microbial population. Numerous studies have shown that land management practices such as tillage, stubble burning and over grazing can affect the level of soil organic matter, but fewer studies have looked directly at the effects on the soil biology and the soil microbial populations.

The soil microbial population is a very sensitive indicator of soil condition and effects of land management. Bending et al. (2000) found that the substrate utilisation and metabolic diversity of the microbial population was much more sensitive to changes in soil management than the levels of soil carbon and the sizes of the different soil carbon pools. In a detailed study Nelson and Mele (2006) investigated the effects of crop residues, crop rotation and liming on the soil microbial population. Using substrate and DNA analysis they were able to show that the diversity of the soil microbial population was increased by lime especially. Where soils were amended with N-rich crop residues including lucerne, lupin and pea residues, the biodiversity was also increased. Each crop residue, which also included wheat, canola and a control of no residue, had a unique microbial community. The wheat residue had the lowest biodiversity and was approximately the same as the control. However the largest overall effect on the microbial population was the addition of lime. In the un-limed soils, Mn, Cu and Na had an effect on the microbial populations, suggesting a toxic or adverse effect on the microorganisms in the lower pH environment. The strong effect of the lime was perhaps expected as the soil had pH 4.14 in $Ca\,Cl_2$ and there was a significant level of soluble Al^{3+} without the addition of lime.

Stubble retention of wheat was found to increase the enzyme activity and N mineralisation (Hoyle and Murphy 2006). This was attributed to increased soil biological activity and was at least partly attributed to the fact that there was a higher levels of soil moisture under the stubble retained soil compared to the stubble burnt soil. There was a higher levels of microbial biomass in the stubble retained soil, but there was not a significant difference in the total organic carbon.

Carter and Mele (1992) showed that direct drill cropping systems increased the microbial biomass under direct drill compared to conventional tillage systems.

A useful comparison of the effects of land management on soil biology and the microbial populations is shown by Bell et al. (2006). They included not only standard tillage treatments in their investigation but also some native/undisturbed pasture sites. The biological populations of the native and undisturbed sites are quite different to the agricultural sites. The total carbon is higher, the microbial biomass is higher, the enzyme activity is higher, and total DNA is higher. All expected but the data confirmed the soil biology of the native undisturbed sites is very different to the sites under cropping. Within the cropping sites, there were no consistent trends in the soil biological measurements to give definite conclusions about treatment effects between stubble retention and stubble burning and tillage and no tillage. An example of their results is shown in Table 4.3. This table gives an indication of the parameters used in the analysis of microbial communities and of the comparison of cropped soils to woodland soils. One of the results from the study of Bell et al. was that they could not find a consistent impact of industrial fertilisers on the microbial populations (see their Table 6).

Cookson et al (2006) found that any effects of land management practices on the microbial populations are moderated by the size of the C and N pools which in turn are affected by the amount of silt and clay in the soil. The amount of clay and silt can control the amount of water and organic matter available to the soil ecosystem.

A new approach to soil monitoring is to investigate two genes for two broad land uses. Hayden et al (2010) examined the abundance of two critical genes for N cycling across regional Victoria. The genes were *amoA* (responsible for metabolising ammonia) and nifA (has a role in the N_2 to ammonia process). The distribution of the 2 genes was investigated at a regional scale. The *amoA* gene was strongly influenced by land use and was most common in neutral to alkaline soils and especially the Calcarosols in western Victoria. The *nifA* gene is more affected by land-use, and there is was no significant interaction between land-use and geomorphic zones. Variables identified as drivers for *amoA* included pH, Olsen P, microbial biomass carbon, nitrate and total nitrogen while for *nifH* the main drivers were microbial biomass carbon, electrical conductivity, microbial biomass nitrogen, total nitrogen and total potassium.

Land use can significantly affect both the amount of soil organic matter and its chemical and physical nature (Murphy et al. 2011). In turn the amount and nature of soil organic matter can affect the microbial biomass and character of the microbial population. Murphy et al (2011) show clearly that while the size, function and structure of the soil microbial community are related to total soil carbon, they are more strongly related to the labile pools of soil organic matter. In the same study there were significant differences in the biological activity between the land management systems based on potentially mineralisable nitrogen (PMN), β-glucosidase (decomposition of cellulose), and acid phosphatase. The land management systems with a pasture phase or under pasture had a higher biological activity of enzymes. The continuous wheat had the lowest activity. Total soil carbon was a significant general indicator of the amount of biological activity, with the higher soil carbon levels having higher activity. However, the soil carbon fractions were a stronger indicator, with the higher activity being associated with higher levels of the labile soil carbon fraction. One striking feature was that the C:N ratio of the labile soil organic matter fraction was significantly higher than for the general soil organic matter and had a value > 20. A general conclusion was that microbial processes were inextricably linked to the quantity and quality of available soil organic matter resources and this is a consequence of soil organic matter being a primary source of energy for most of the soil microbial population.

These examples give an indication of the change in direction of soil biological and soil microbiological studies in more recent years.

4.7 Critical Levels of Soil organic Carbon for Soil Microbial Populations and Function

The question arises if there is a threshold level of soil organic matter that is required to maintain all the functions of the microbial population (Yan et al. 2000). Is it possible that if the if the soil organic matter levels fall below a certain threshold that some functions are lost and the access to some nutrient stores is lost because they are no longer being mineralized? A “species redundancy” view has been expressed by Lawton and Brown (1994) which maintains that several species usually perform a similar task in an ecosystem and that given the dynamic nature of species, especially microbial populations (Yan et al. 2000), that other species will perform the task. However, an alternative view is that all species are unique and that while similar, are not identical in their function or range and that if sufficient species are lost beyond a threshold the ecosystem function maybe compromised (Ehrlich and Wilson 1991). In a study of biodiversity in Vertosols near Narrabri, Yan et al. (2000) found that to maintain the maximum biodiversity of the microbial population it was necessary to maintain a level of soil organic carbon at 2.0%. The study did not explain what the levels of biodiversity meant for decomposition of wheat straw or fixation of N or other function of agricultural interest, but was an indication that if the soil carbon levels fell below 2.0%, there was a loss of microbial biodiversity.

The links between soil biodiversity and soil functions such as disease control, nutrient availability and water use efficiency are described by Brussaard et al. (2007). They emphasise that the soil microbial population can influence these soil functions and have developed a scheme for assessing how the different soil populations can affect these soil functions.

The amount of soil organic carbon can affect the levels of biological activity. Chan et al. (1992) found a direct relationship between the amount of activity and the level of soil organic carbon in a tillage trial on a Red Kandosol in the south eastern wheat belt of Australia. Singh et al. (2007) found that the growth of plants and the provision of a source of carbon for energy is a major requirement for the maintenance of the microbial population in grassland soils and that using DNA techniques, demonstrated that the microbial population within the rhizosphere was different to that outside the rhizosphere. The microbial population outside the rhizosphere had more spore forming bacteria.

Table 4.1 Land management practices and agricultural practices that exploit soil biological processes to improve productivity (Adapted from Watt et al. 2006).

Agronomic Issue	**Underlying soil biological processes**	**Time Scale**
Nutrient cycling	• Decomposition of crop and pasture residues • Decomposition of existing soil organic matter	Weeks, months Weeks, months, years. decades for more resistant fractions
Nitrogen fixation	•Infection of root hairs by rhizobia •Nodule development and function •Decomposition of nitrogen containing tissue	Hours Days to weeks Hours to years
Crop rotation and crop sequences	•Non hosting of common diseases – time gap •Inoculum decline	Hours to months Months to years
Disease suppression	•Bacteria increase on successive generations of dead roots •Bacteria suppressive to fungi produce antifungal exudates in the plant rhizosphere	Years Hours to days
Development of soil structure	•Development of aggregate stability – root and microbial exudates, hyphae •Stabilisation of soil macropores – root and microbial exudates, hyphae	Weeks, months Weeks, months

Table 4.2. Methods for measuring soil biology and characterising soil microbial populations

General class of measurements	**Output and measurement**	**References** Insam (2001) as a general reference for the methods
Soil organic carbon measurements	Total organic carbon Soil carbon fractions g/100g	Baldock and Skjemstad (1999) CSIRO (2013a)
Microbial biomass Respiration studies Carbon dioxide evolution	g/100g μg/day	Gonzalez-Quinones et al. (2011) Dalal (1998) Used in Hoyle and Murphy (2006)
Growth of microorganisms on different substrates Community level physiological profiles (CLPP) BIOLOG	Usually requires a statistical analysis such as principal components or a biodiversity measure such as the Shannon Diversity Index (Stephan et al. 2000)	Degens et al (2001) Nicholson and Hirsch (1998) Used in Hoyle and Murphy (2006) and Nelson and Mele (2006)
Phospholipid fatty acid extraction PLFA Fatty acid methyl ester FAME	Usually requires a statistical analysis such as principal components or a biodiversity measure such as the Shannon Diversity Index	Zelles et al. 1992 Zelles and Bai (1993) Mele and Crowley (2008) Ibekwe and Kennedy (1999) Used in Hoyle and Murphy (2006)
DNA analysis	Usually requires a statistical analysis such as principal components or a biodiversity measure such as the Shannon Diversity Index	Yeates and Gilling (1998) McCaig et al. (1999) Used in Barnet et al (2006)
Enzyme fluorescent analysis Biochemical enzyme assays	Measurement of enzyme activity	Marx et al. 2001, Mondini et al. 2004; Burns et al. 2013 Fontvieille et al. (1991) Used in Hoyle and Murphy (2006) Used in Bell et al (2006)

Table 4.3. Impact of cropping on soil biological properties in the central Darling Downs of Southern Queensland. (Adapted from Bell et al. 2006). The parameters EL – FAME and PL-FAME refer to the fatty acid analysis of the microbial communities including the ester-linked fatty acid methyl esters (EL-FAMEs) and the phospholipid-linked fatty aid methyl esters (PL-FAMEs).

Biological parameter		**Jandowae**	
0 – 0.05 m	**Units**	**Cropped**	**Open woodland**
Microbial biomass	mg biomass/g soil	0.25 ± 0.04	0.76 ± 0.06
Microbial activity	mg fluorescein/gsoil.45 min	4.0 ± 0.6	9.3 ± 1.1
Total free living nematodes	(/g soil)	7.3 ± 1.9	9.9 ± 1.5
Total DNA	μg/g soil	4.3 ± 0.3	17.3 ± 4.0
EL - FAMEs	μg/g soil	33.4	55.9
PL - FAMEs	μg/g soil	8.0	2.1
Total Organic Carbon	mg/g	7.0 ± 0.1	17.9 ± 2.1
Total N	mg/g	0.76 ± 0.00	1.51 ± 0.13

Summary of Key Points – 4. Soil Organic Matter and Soil Biology

- Soil organic matter provides the food and energy source for the soil microbial population and the soil fauna. In turn the soil microbial population and soil fauna are responsible for the recycling of nutrients from decomposing plant materials and soil organic matter. More specifically the soil microbial population is responsible for N fixation either in symbiotic relationships in plants such as legumes or as free living nitrogen fixers.
- The soil organisms in the soil have a function of maintaining or improving aggregate stability either through providing mucilage and polysaccharides that can act as binding agents or physically binding aggregates as hyphae.
- In the dynamic soil environment soil organisms can also have an important function of disease suppression of certain plant pathogens.
- Recent developments in methodologies to measure soil biology have enabled a wide range of measurements to be made on soils including microbial biomass, substrate analysis, phospholipid analysis, DNA analysis and enzyme evaluation. These new methods have enabled new insights and a better understanding of soil biology.
- Land management has been shown to clearly affect soil biological populations using the new soil biological methods. Land management factors that affect soil biological populations include the addition of lime, stubble management, tillage, use of perennial grasses, different crop species and preservation of native vegetation.
- It is early yet to establish critical levels of soil organic carbon based on the new soil biological methodologies, but early indications show a level of about 2% may be required for optimum levels of biological activity.

5. Capacity of Soils to Achieve and Maintain Critical Soil Organic Carbon Levels under a Range of Environmental and Land Management Options

5.1 Introduction

Soil organic matter has some capacity to improve soil properties but increasing soil organic matter levels requires a suitable soil and climate and input of biomass and nutrients. Soils tend to reach a long term equilibrium level of soil organic matter related to the local soil (texture, nutrient levels, depth, structure),climate and land management practices (biomass and nutrient inputs, balance of decomposition pressures - tillage, stubble management and grazing pressures) (Gray et al. 2012). It is probably not possible to build soil organic matter beyond some biophysical boundary set by the soil, climate and landform position. Of course it may be possible by introducing soil organic matter from outside such as compost or biochar, but the limit is difficult to exceed by in-situ increase of soil organic matter using plant photosynthesis.

Some land management practices have lower inputs of biomass and nutrients and a higher decomposition pressure on the exiting soil organic matter. It will be very difficult to build up soil organic matter under these land management practices.

With potential limits on the increases in soil organic matter it becomes highly relevant how these limits relate to any critical values of soil organic matter or soil organic carbon for particular soil functional properties.

In this Section the aim is to identify what some of the expected levels of soil organic matter exist for particular regions under various land management practices. Discussion will be based on soil organic carbon as most of the published data is based on soil organic carbon in the published data sets.

5.2 Changes in Soil Carbon / Soil Organic Matter in the 0 – 10 cm Layer

Recent studies in soil organic carbon indicate that the major changes in soil organic carbon associated with changes in land management take place in the top 10 cm. It can be further argued that for changes in soil organic carbon and soil organic matter to have a significant impact on soil properties they have to be substantial. The largest changes in soil organic carbon take place in the top 10 cm as can be seen in the 3 graphs in Figure 5.1. The data in these graphs are from the dry land cropping belt and slopes of NSW, and is possible changes in soil carbon can occur deeper in the soil, especially in higher rainfall areas on the coast or on the tablelands. However, it is useful to concentrate initially on the 0 – 10 cm layer initially to investigate two objectives:

1. The capacity of land management systems to maintain soil organic carbon levels that can improve soil functional properties
2. The capacity of land management systems for a first approximation to investigate the potential for land management to achieve to improvements in soil condition by increasing soil organic carbon.

As has been seen throughout this review, the relationships between soil organic matter are based on measurements of soil organic carbon expressed as soil organic content of the soil in g/100g or soil carbon per cent. This is the standard output from most routine soil analyses.

A number of data sets are reviewed as typical examples of what levels of soil organic carbon can be expected in Australian soils. This is only a selection of data sets, but together they indicate the expected levels of soil carbon.

1. *Baldock and Skjemstad (1999), Spain et a. (1993)*

The soil organic carbon levels of the topsoils or A horizons of the major great soil groups are shown as below:

- Red-brown earths (Red Chromosols) – 1 to 1.5%
- Red earths (Red Kandosols) – 1 to 1.5%
- Black earths (Black Vertosols) – 2.0 to 2.5%
- Grey clays (Grey Vertosols) – 0.7 to 1.0%
- Krasnozems (Red Ferrosols) – 6.0%

The samples were grouped on the basis of great soil group alone and land uses within each great soil group were variable. These values do give an indication of the levels of soil organic carbon to expect but they do probably underestimate the levels that can be achieved under good cropping and pasture systems in the red Chromosols and Red Kandosols, especially in higher rainfall areas.

The organic concentrations in the surface soils of the main great soil groups from a wider range of data are given by Spain et al. (1993) (Table 5.1). These are indicative only of the of the soil carbon levels to be expected as they come from surface soils ranging from 5 to 15 cms deep. They do show however the likely levels of soil carbon to be expected for the main great soil groups which are used for agricultural production.

2. Geeves et al (1995)

A comprehensive sampling program was undertaken of the wheat belt in south eastern Australia in 1991 and reported in Geeves et al. (1995). A total of 78 sites were selected non-randomly to represent the full range of land management systems used on the major soil types in the region. Descriptions were made of each site of surface soil and vegetation features and a profile description. Measurements were made of hydraulic conductivity, soil moisture characteristic, bulk density, particle sizes, soil strength, aggregate stability, total carbon, soluble and exchangeable cations, soil pH and soil minerals. Information on paddock history was also collected. All sampling and measurements were made within a 1 m radius of the soil core taken for the profile description. A number of woodland or native vegetation sites were selected for benchmark comparisons.

Soils were crushed and passed through a 2 mm sieve. Soil carbon was determined using a LECO CR_12 furnace. The soil carbon data is summarised in Table 5.2

The pH values were all less than 7 except one which was 7.47 so the values shown can be expected to show soil organic carbon.

Most of the soils examined were in the Red Chromosol, Red Kandosol and Red Dermosol groups. There were occasional Sodosols. The soils would comply with the old red-brown earth and red earth groups and many were also in the non-calcic brown soils in the slopes. There were occasional euchrozem like soils associated with more basic parent materials.

The surface soils were largely lightly textured sandy loams to fine sandy clay loams with weak structure. The exception would be the better structured clay loams of the Red Dermosol or euchrozem like soils.

The soil organic carbon levels shown in Table 5.2 are indicative of the values that can be expected across the wheat belt in south eastern Australia. Several key features are evident from this table.

- The original levels of soil carbon in the woodland soils have been greatly reduced. This has been reported by many authors.
- Overall across the cropping and pasture sites the soil carbon levels are largely in the range of 1 to 2%.
- Some outstanding performers where the soil carbon levels exceed 2%. This has happened in both cropping and pasture land management systems. These results need to be confirmed by better sampling systems to ensure they are just not a result of local spatial variability, but they are an interesting result.
- Overall pasture has slightly higher soil carbon levels than cropping. However, adopting a pasture land management system is no guarantee of increasing soil carbon. Some of the lowest soil carbon levels are associated with pasture land management. In the report the category of "heavy grazing" is used for some of the pasture sites and is accompanied by comments such as "poor stand of clover and lucerne, mainly saffron thistle and barley grass", "volunteer pasture" and "lucerne, cut for hay". It appears for this category a feature of the pasture management suggested that there were aspects that may impact on the soil condition. This has important implications for general recommendations about using pastures as a means to increase soil carbon levels. The pastures must be well managed and grazing well managed.
- The cropping practices considered conservation based appeared to have a very slight advantage but overall it would be difficult to separate them from the more conventional practices in their overall effect on soil carbon. There is however a large range under both cropping systems.

Based on this data it would seem possible to readily maintain soil carbon levels in the range of 1.2 to 1.5% soil organic carbon. To increase soil organic carbon levels above 1.5% and then above 2% is likely to require the application of innovative and specialised land management systems. The better structured clay loams of the Red Dermosol or euchrozem like soils can be expected to have higher soil carbon levels because of high clay contents (Carter et al. 2003).

3. Central West Catchment of NSW – Soil Carbon Benchmark Report

In 2011 all the soil carbon from a range of sources was summarised for the Central West Catchment in NSW. The Central West Catchment includes the catchments of the Macquarie and Castlereagh rivers in central west NSW. The catchment was divided into a series based on soil type and climate and the soil carbon information for these zones summarised. Various programs of soil carbon measurements and soil mapping were active in the area over the last few years and the opportunity taken to compile this information into one document. Information was also utilised from the NSW SALIS database. The report is summarised in Murphy et al. (2010). Some output from the report is summarised in Table 5.3. While the data generally confirms the results from the Geeves et al. (1995) data it also confirms the greater capacity of the basalt type soils (Red Dermosols, and Vertosols) to have higher levels of soil organic carbon. It also shows the general trend for soil organic carbon as that shown in the SCaRP data report (Figure 5.1) (CSIRO 2013b).

4. Northern tablelands of NSW (Wilson et al. 2011)

In the northern tablelands of NSW soil carbon measurement were made on a series of geologies and soil types. The values for the soil organic carbon contents for 0 – 10 cm for the different soil types and geologies were:

- Basalt – Black and Red Ferrosols
  - Cultivation 2.43%
  - Improved pasture 2.87%
  - Unimproved pasture 2.69%
  - Woodland 5.06%
- Granite – Grey and Yellow Chromosols
  - Cultivation 1.28%
  - Improved pasture 1.43%
  - Unimproved pasture 1.52%
  - Woodland 1.92%
- Metasediments – Red and Brown Chromosols
  - Cultivation 1.98%
  - Improved pasture 2.48%
  - Unimproved pasture 3.00%
  - Woodland 4.55%

The data from this paper largely confirms the values from the previous data sets but also reinforces the strong effect of soil type on the potential to store soil organic carbon. The better soils with higher clay contents and more fertility can store higher levels of soil organic carbon.

5. Soil Health in Broadacre Crops Project Queensland (Lawrence et al 2008).

A summary of the soil carbon data collected from landholders' paddocks for 0 – 10 cm in a soil health project in the eastern Darling Downs shows that the cropped soils have soil organic carbon levels of 1.63 ± 0.38 %. This includes one outlier value of 2.80% which is very unusual. If this is excluded the soil carbon level is 1.53 ± 0.14%. The comparison soil carbon values done along fence lines and intended to indicate a soil in better condition is 2.11 ± 0.56 %. The soil type is a Vertosol with higher clay levels. The results again are consistent with the general levels of the other data.

6. Pastures – Chan et al. (2011)

In a study of the potential for pasture improvement to sequester soil carbon, Chan et al. (2011) showed that the soil organic carbon levels in the 0 – 10 cm layer are in the range of 2 to 4%. While their Figure 3 demonstrates that the major changes in soil carbon occur in the top 10 cm of soil, closer examination shows that changes in soil carbon of the order of 0.5% soil carbon can take place in the 10 to 20 cm layer and slightly less in the 20 to 30 cm layer. This may have implications for the small improvement of some functional soil properties such as water holding properties and cation exchange capacity.

7. NSW MER Soil Carbon Data

The NSW Office of Environment and Heritage undertook a soil monitoring program in 2008/2009 (see Table 5.4). Soil organic carbon was measured as part of this program (DECC 2009). More extensive measurements of the samples (carbon fractions) have been undertaken in conjunction with CSIRO as part of the National Soil Carbon Program with CSIRO Adelaide (SCaRP). Some of these results can be reported here (Table 5.4). The measurements of soil carbon in Table 5.4 show the following.

- The coastal areas tend to have higher soil carbon levels in the plains, slopes and tablelands areas. The rainfall is higher in these areas.
- The western areas have the lowest soil carbon levels and this corresponds to the lowest rainfall areas.
- The inland slope areas tend to have higher soil carbon levels than the plains.
- The inland slopes and inland plains areas have soil carbon levels in the range of 1.0 to 2.0 g/100g. This is the range where changes in soil carbon levels are likely to result in changes in the basic functional soil properties. This implies that in the inland slopes and plains there is a potential for increases in soil carbon and so soil organic matter levels to improve soil functional properties.

8. National Soil Carbon Research Program Data Set

A preliminary summary of the soil carbon data from the National Soil Carbon Research Program is presented in Table 5.5. This data was collected over 2010 to 2012 throughout Australia (CSIRO 2013b). A detailed analysis of this data is currently being undertaken and it will be published in a special issue of Soil Research. This very preliminary summary of the data has been undertaken to obtain some indication of the expected soil carbon contents across a range of soil types, environments and land uses throughout Australia. A few brief conclusions can be made about the expected soil carbon contents across Australia and therefore the likely effects on soil functional properties.

- The higher levels of soil organic carbon content (> 3%) are associated with regions having higher rainfall areas and cooler temperatures in Tasmania especially but also such areas as the Southern Tablelands of NSW and the Strzelecki Ranges and Otways of Victoria.
- Generally Queensland has relatively lower levels of soil organic carbon, which is consistent with the higher temperatures resulting in higher decomposition rates, even though the rainfall and higher temperatures can result in high biomass productivity.
- As a general rule the major cropping areas of the slopes and plains are on the lower end of the soil carbon levels often having median soil organic carbon levels < 2% and some areas such as the drier Mallee areas being < 1%.
- The range of soil organic carbon values (20th to 80th percentile) within each region indicates there is scope for improving soil carbon levels in most regions. The development of land management practices to increase soil carbon levels would appear to be a potentially viable objective. The range of change in soil organic carbon levels for many agricultural areas of the slopes and plains appears to be in the range where there are gains to be made in improving the functionality of soils.
- As a general rule, pasture tends to have higher soil carbon levels than cropping land uses.
- There apparent high values in the Western Australian data are associated with irrigated dairy pastures.

This data is a very preliminary analysis and only gives an indication of the kinds of soil carbon concentration of soil carbon contents that can be expected in different regions.

9. Conclusion

There is sufficient data presented to indicate what levels of soil carbon are to be expected. Sufficient information is also presented to confirm the possibility for land management systems to bring about increases in soil organic carbon and for soil organic matter to improve and maintain functional soil properties. What is not presented is a comprehensive modelling program to predict what levels of soil carbon will occur for different combinations of soil type, climate and specific land management practices. This would be beyond the scope of the current review.

5.3 The Importance of Soil Depth on Improving Soil Condition

The above data is based on the 0 – 10 cm layer, and the reason for concentrating on that layer is the amount of data supporting the view that the major changes in soil organic carbon occur in the 0 – 10 cm layer. The 0 – 10 cm layer is an important soil layer as it interacts with atmosphere with rainfall and air exchange and it is the layer where crops are planted and germination and emergence takes place. It is the layer that is subjected to the pressures of water and wind erosion, and compaction by machinery and stock. It is important that is in condition and has a strong resilience. Improving the condition of the surface 0 - 10 cm layer has potential benefits such as:

- Better infiltration
- Reduced erosion
- Better soil tilth for cultivation and unimpeded germination and emergence
- Better water holding capacity for the 0 – 10 cm layer
- Higher cation exchange capacity and chemical activity for the 0 – 10 cm layer
- Better resistance to soil compaction resulting in better aeration and root growth

However, these benefits are restricted to the surface 10 cm. If some of these benefits such as higher cation exchange capacity and higher water holding capacity could be extended by increasing soil organic matter levels in deeper layers in the soil, the benefits of soil organic matter to soils could be enhanced. In a recent paper, Read et al. (2012) noted that soil carbon increases in revegetating scalds took place down to 30cm. The data in Chan et al (2010) give some indications of changes in soil organic carbon of 0.5% in the 10 to 20 cm layer and slightly less in the 20 to 30 layer. It is a potentially useful avenue for future investigation because if soil carbon levels can be increased deeper than 10 cm, then the potential benefits of soil organic carbon and soil organic matter will be significantly enhanced. The data from the current National Soil Carbon Project is a source of useful information for this.

5.4 Implied potential for improving soils by increasing levels of soil organic carbon

A comparison of the general levels of soil carbon occurring on the landscape under agriculture in the dryland cropping areas and the critical values of soil organic carbon that can improve soil functional properties would seem to imply some potential exists to improve soil condition and soil productive potential by increasing levels of soil organic carbon. From Table 5.2 and Figure 5.2, it is possible to identify three sets of soils in relation to the potential to improve soils by increasing levels of soil organic carbon:

- ❖ At the low end are the soils with the minimum and low levels of soil organic carbon. These are sufficiently low that the soil properties are being affected adversely by the low soil carbon and there is a genuine potential to increase the soil carbon levels by applying suitable land management practices. It would be a strong possibility to achieve an increase in soil carbon of 0.75 to 1.0%.
- ❖ The soils mid-range or median levels of soil carbon are not being severely affected by low soil carbon, but soil properties could be improved by higher levels of soil carbon. To

increase soil carbon levels probably requires large changes in land management practices to ensure soil carbon levels increase. Some of these changes may not be economically or agronomically easy to achieve. Realistic changes in soil carbon are likely to be of the order of 0.25 to 0.75%.

- Soils with currently high levels of soil carbon will make it difficult to maintain increases in soil carbon. No large continued increases can probably be expected. However, there is the need to maintain the high levels of soil carbon which needs continued inputs management.

The potential to increase soil carbon levels may be much higher in higher rainfall areas on the coast and in the tablelands.

Throughout the review it appears a level of soil organic carbon of 2.00% is a sound level to ensure functioning soil properties and the biological populations in the soil are in good condition. It does seem to be useful guide as a target to ensure the soil is in good functioning condition. Based on much of the measured data this can be a difficult level of soil carbon to achieve under dryland agriculture on the slopes, but Chan et al. (2011) suggest it can be often attained under pastures in the tablelands.

Table 5.1. Measured levels of soil carbon in great soil groups from across Australia based on Spain et al. (1993).

Great Soil Group	**Australian Soil Classification**	**Range of soil carbon contents in the top 5 to 15 cms of soil g/100g**
Red-brown earths	Red Chromosols, Red Dermosols	0.5 to 1.5
Red earths	Red Kandosols	0.6 to 1.36
Black earths	Self –mulching Black Vertosols	1.6 to 3.5
Red Podzolic Soils	Red Chromosols, Red Dermosols, Red Kurosols	1.6 to 4.1
Non-calcic Brown Soils	Red Chromosols, Red Dermosols	1.0 to 2.2
Siliceous sands	Arenic Tenosols	0.3 to 1.5
Earthy sands	Tenosols	0.3 to 0.8
Solodic soils group	Sodosols	0.8 to 1.8
Grey, brown and red cracking clays	Epipedal and Massive Vertosols	1.0 to 2.4
Euchrozems	Red Dermosols, Red Ferrosols	1.8 to 3.0
Solonised brown soils	Calcarosols	0.7 to 1.5

Table 5.2. Summary of the total soil carbon content data for the 0 – 10 cm layer from Geeves et al. (1995). Based on a survey from the wheat belt of south eastern Australia. Data are in g/100g.

Data set	**mean**	**sd**	**min**	**max**	**median**	**Perc_5**	**Perc_95**	**n**
All	1.75	1.17	0.48	6.29	1.47	0.68	4.74	64
Cropping	1.39	0.59	0.48	3.62	1.32	0.54	2.76	38
Pasture	1.62	0.66	0.64	3.31	1.42	0.64	3.31	19
Woodland	4.11	1.89	1.74	6.29	4.36	-	-	7
Pasture – well managed	1.86	0.66	1.11	3.31	1.65	-	-	12
Pasture - heavily grazed	1.20	0.40	0.64	1.87	1.32	-	-	7
Cropping - conservation	1.45	0.65	0.82	3.62	1.26	0.82	3.62	19
Cropping - conventional	1.32	0.54	0.48	2.71	1.38	0.48	2.71	19

- sd is standard deviation
- max is maximum
- min is minimum
- Perc_5 is the 55 percentile
- Perc_95 is the 95% percentile
- N is the number of sites in the data set.

Table 5.3. Summary of some of the soil organic carbon data from the Central West Catchment Soil Carbon Benchmark report. This data reports on the Basalt derived soils and the fragile red soils which include the Red Chromosols and Red Kandosols. Soil carbon data is soil carbon in g/100 g for the 0 – 10 cm layer.

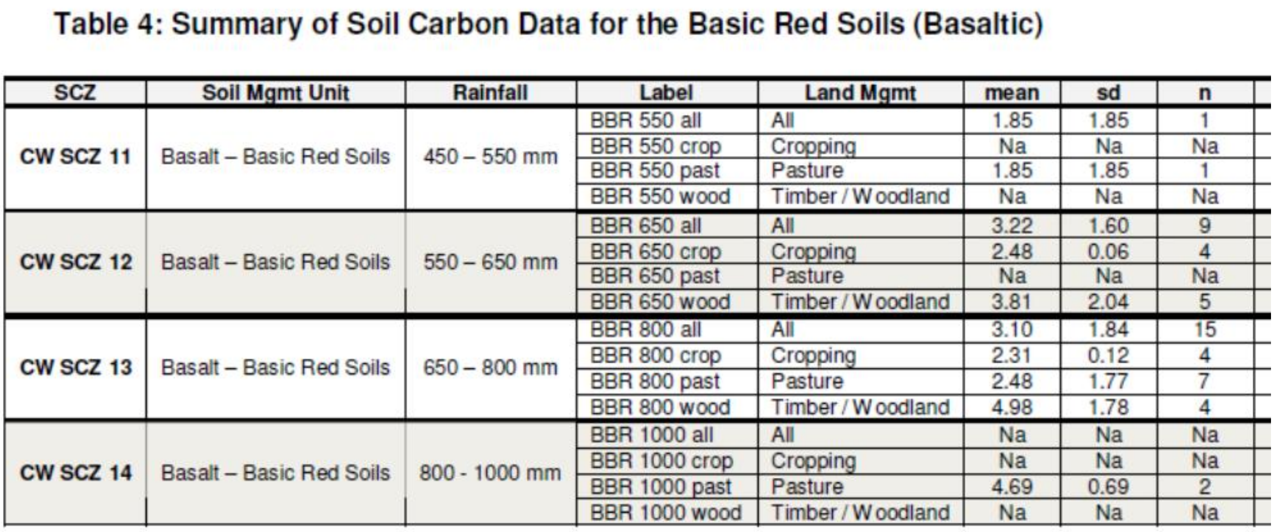


Table 4: Summary of Soil Carbon Data for the Basic Red Soils (Basaltic)

SCZ	Soil Mgmt Unit	Rainfall	Label	Land Mgmt	mean	sd	n
CW SCZ 11	Basalt – Basic Red Soils	450 – 550 mm	BBR 550 all	All	1.85	1.85	1
			BBR 550 crop	Cropping	Na	Na	Na
			BBR 550 past	Pasture	1.85	1.85	1
			BBR 550 wood	Timber / Woodland	Na	Na	Na
CW SCZ 12	Basalt – Basic Red Soils	550 – 650 mm	BBR 650 all	All	3.22	1.60	9
			BBR 650 crop	Cropping	2.48	0.06	4
			BBR 650 past	Pasture	Na	Na	Na
			BBR 650 wood	Timber / Woodland	3.81	2.04	5
CW SCZ 13	Basalt – Basic Red Soils	650 – 800 mm	BBR 800 all	All	3.10	1.84	15
			BBR 800 crop	Cropping	2.31	0.12	4
			BBR 800 past	Pasture	2.48	1.77	7
			BBR 800 wood	Timber / Woodland	4.98	1.78	4
CW SCZ 14	Basalt – Basic Red Soils	800 - 1000 mm	BBR 1000 all	All	Na	Na	Na
			BBR 1000 crop	Cropping	Na	Na	Na
			BBR 1000 past	Pasture	4.69	0.69	2
			BBR 1000 wood	Timber / Woodland	Na	Na	Na

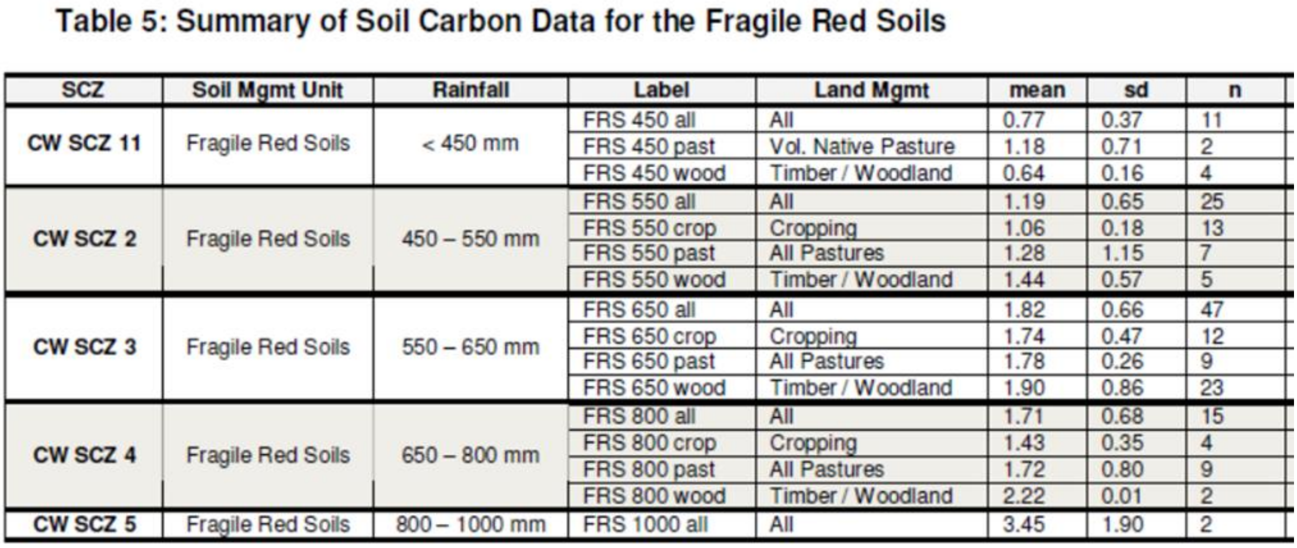


Table 5: Summary of Soil Carbon Data for the Fragile Red Soils

SCZ	Soil Mgmt Unit	Rainfall	Label	Land Mgmt	mean	sd	n
CW SCZ 11	Fragile Red Soils	< 450 mm	FRS 450 all	All	0.77	0.37	11
			FRS 450 past	Vol. Native Pasture	1.18	0.71	2
			FRS 450 wood	Timber / Woodland	0.64	0.16	4
CW SCZ 2	Fragile Red Soils	450 – 550 mm	FRS 550 all	All	1.19	0.65	25
			FRS 550 crop	Cropping	1.06	0.18	13
			FRS 550 past	All Pastures	1.28	1.15	7
			FRS 550 wood	Timber / Woodland	1.44	0.57	5
CW SCZ 3	Fragile Red Soils	550 – 650 mm	FRS 650 all	All	1.82	0.66	47
			FRS 650 crop	Cropping	1.74	0.47	12
			FRS 650 past	All Pastures	1.78	0.26	9
			FRS 650 wood	Timber / Woodland	1.90	0.86	23
CW SCZ 4	Fragile Red Soils	650 – 800 mm	FRS 800 all	All	1.71	0.68	15
			FRS 800 crop	Cropping	1.43	0.35	4
			FRS 800 past	All Pastures	1.72	0.80	9
			FRS 800 wood	Timber / Woodland	2.22	0.01	2
CW SCZ 5	Fragile Red Soils	800 – 1000 mm	FRS 1000 all	All	3.45	1.90	2

Table 5.4. Estimated soil carbon levels for NSW soils under agricultural land use. Based on the data from the soil monitoring project of the NSW Office of Environment and Heritage with assistance from CSIRO Adelaide (see DECCW 2009 for methodology and Murphy et al. in prep. for analysis)

Region	Soil organic carbon levels 0 – 10 cm (g/100g)					
	median	mean	standard deviation	maximum	minimum	n
Border Rivers Gwydir						
Northern plains	1.33	1.32	0.39	2.03	0.80	13
Northern slopes	1.52	1.94	0.56	2.78	0.98	11
Central West						
Central plains	0.94	0.94	0.10	1.09	0.79	8
Central slopes	1.51	1.61	0.44	2.36	0.81	30
Central tablelands	1.00	1.13	0.29	1.46	0.92	3
Hunter Central Rivers						
Central slopes	2.73	3.02	2.38	7.37	1.80	14
Coastal plains	2.48	2.67	1.03	4.81	1.42	10
Coastal slopes	3.09	3.18	1.07	5.16	1.23	21
Hawkesbury Nepean						
Coastal slopes	3.96	3.64	1.70	7.20	1.20	16
Central tablelands	2.04	2.05	0.65	3.17	0.99	20
Lachlan						
Central slopes	1.45	1.38	0.36	1.89	0.77	8
Central plains	1.38	1.38	0.16	1.57	1.10	9
Lower Murray Darling						
Western plains	1.04	1.37	0.80	3.49	0.46	14
Western ranges	0.53	0.53	0.04	0.56	0.51	2
Murray – Murrumbidgee						
Alpine subalpine areas	3.23	3.00	1.00	3.83	1.69	4
Southern plains	1.64	1.69	0.71	3.99	0.93	25
Southern slopes	1.87	2.16	1.21	5.99	1.18	20
Southern tablelands	3.26	3.14	0.95	4.80	1.77	15
Namoi						
Northern plains	1.18	1.43	0.89	4.56	0.48	38
Northern slopes	2.15	2.42	1.01	4.89	1.03	35
Northern Rivers						
Coastal plains	2.71	2.84	1.05	5.53	1.34	16
Coastal slopes	3.57	3.72	2.80	9.76	0.74	16
Northern tablelands	2.41	2.71	0.97	5.87	1.60	18
Southern Rivers						
Coastal slopes	3.21	3.05	0.69	4.02	2.18	8
Southern tablelands	2.80	3.33	1.66	7.85	1.04	34
Western						
Western plains	0.68	0.86	0.34	1.47	0.56	11
Western ranges	0.61	0.75	0.37	1.40	0.38	10

Table 5.5a. Estimated soil carbon levels for Australian soils under agricultural land use. Based on the data from the National Soil Carbon Project (SCaRP) of the CSIRO and the Department of Climate Change and Energy Efficiency (CSIRO 2013b). Note that this is a very preliminary summary of the available data and is indicative only of the overall results.

				Soil organic carbon g/100g 0 to10 cm				
Comb	**State**	**Region**	**Broad Land Use**	**median**	**percentile 20%**	**percentile 80%**	**No of sites**	**Main soils**
1	NSW	Central plains	Cropping	0.968	0.737	1.208	49	Chromosols/Dermosols and Vertosols
2	NSW	Central plains	Pasture	1.058	0.750	1.517	69	Chromosols/Dermosols and Vertosols
3	NSW	Central slopes	Crop - pasture rotation	1.353	1.095	1.699	51	Chromosols/Dermosols
4	NSW	Central slopes	Pasture	1.724	1.406	2.231	67	Chromosols/Dermosols
5	NSW	Central slopes	Pasture cropping	1.773	1.408	2.570	23	Chromosols/Dermosols
6	NSW	Central tablelands	Pasture	1.859	1.001	2.499	219	Chromosols/Dermosols , Ferrosols
7	NSW	Murray Slopes	Crop - pasture rotation	1.765	1.422	2.100	27	Chromosols, Kurosols
8	NSW	Murray Slopes	Pasture	2.644	1.883	3.486	73	Chromosols, Kurosols
9	NSW	Northern plains	Irrigated cropping	1.197	0.717	1.630	50	Vertosols
10	NSW	Northern plains	Pasture	1.211	0.878	2.276	25	Vertosols
11	NSW	Northern slopes	Crop - pasture rotation_org	1.739	1.311	2.745	16	Vertosols, Dermosols
12	NSW	Northern slopes	Cropping	1.700	1.072	1.953	11	Vertosols
13	NSW	Northern slopes	Pasture	1.954	1.483	2.492	100	Chromosols
14	NSW	Northern tablelands	Pasture	1.916	1.503	2.651	115	Chromosols, Kurosols, Kandosols
15	NSW	Riverina	Crop - pasture rotation	1.467	1.288	1.781	39	Chromosols, Sodosols
16	NSW	Riverina	Cropping	1.329	1.061	1.618	41	Chromosols, Sodosols, Vertosols
17	NSW	Riverina	Pasture	1.669	1.361	1.852	32	Chromosols, Sodosols
18	NSW	Riverina	Pasture - irrigated	1.802	1.106	2.020	7	Vertosols
19	NSW	Southern tablelands	Pasture	4.292	2.423	6.381	22	Chromosols, Dermosols, Kurosols
20	QLD	Balonne	Cropping	1.040	0.884	1.168	19	Vertosols, Dermosols, Sodosols
21	QLD	Biloela/Fitzroy	Cropping	1.494	1.136	1.897	77	Vertosols, Dermosols
22	QLD	Border Rivers	Cropping	0.912	0.792	1.137	19	Vertosols, Sodosols
23	QLD	Bunderberg	Cropping	1.856	1.376	2.032	12	Ferrosols
24	QLD	Goodger - Dalby	Cropping	2.016	1.800	2.299	30	Ferrosols
25	QLD	Hermitage	Cropping	2.073	1.938	2.303	93	Vertosols
26	QLD	Ingham	Cropping	1.141	1.001	1.207	18	Kandosols
27	QLD	Lower Condamine	Cropping	1.361	1.042	1.760	30	Vertosols, Sodosols
28	QLD	Mackay	Cropping	1.221	1.107	1.328	34	Chromosols
29	QLD	Rangelands	Pasture	0.709	0.448	1.264	190	Vertosols, Chromosols, Calcarosols, Sodosols, Kandosols
30	QLD	South Burnett	Cropping	1.701	1.590	2.025	22	Ferrosols
31	QLD	Sugarcane	Cropping	1.798	0.842	1.978	19	Ferrosols, Kandosols, Chromosols, Hydrosols
32	QLD	Upper Condamine	Cropping	1.260	0.889	1.785	30	Vertosols, Chromosols, Sodosols

Table 5.5b. Estimated soil carbon levels for Australian soils under agricultural land use. Based on the data from the National Soil Carbon Project (SCaRP) of the CSIRO and the Department of Climate Change and Energy Efficiency (CSIRO 2013b). Note that this is a very preliminary summary of the available data and is indicative only of the overall results.

				Soil organic carbon g/100g 0 to10 cm				
Comb	**State**	**Region**	**Broad Land Use**	**median**	**percentile 20%**	**percentile 80%**	**No of sites**	**Main soils**
33	SA	Eyre Pen	Crop - pasture rotation	1.246	0.907	1.574	29	Chromosols
34	SA	Eyre Pen	Cropping	1.333	1.020	1.805	39	Chromosols
35	SA	Mid North	Crop - pasture rotation	1.427	1.102	1.835	77	Chromosols
36	SA	Mid North	Cropping	1.499	1.182	1.812	48	Chromosols
37	SA	Mid North	Pasture	1.329	1.075	1.829	7	Chromosols
38	TAS	Brighton	Cropping	2.504	2.169	2.918	5	Chromosols, Dermosols
39	TAS	Burnie	Cropping	4.491	3.804	5.259	11	Ferrosols
40	TAS	Burnie	Pasture	8.412	6.655	9.167	7	Ferrosols, Dermosols
41	TAS	Central Highlands	Cropping	2.520	1.985	3.227	7	Chromosols, Dermosols
42	TAS	Central Highlands	Pasture	4.337	2.392	5.186	15	Chromosols, Dermosols, Vertosols
43	TAS	Clarence	Cropping	2.566	2.014	3.456	12	Chromosols, Dermosols, Vertosols
44	TAS	Clarence	Pasture	3.557	2.607	4.200	7	Chromosols, Dermosols, Vertosols
45	TAS	Deloraine	Cropping	4.930	3.941	6.317	7	Ferrosols, Dermosols
46	TAS	Deloraine	Pasture	9.536	7.237	11.445	4	Ferrosols, Dermosols
47	TAS	Derwent Valley	Pasture	3.102	2.497	3.758	6	Chromosols, Dermosols
48	TAS	Devonport	Cropping	4.579	3.631	5.484	42	Ferrosols
49	TAS	Devonport	Pasture	6.163	5.438	7.647	16	Ferrosols
50	TAS	East Coast	Cropping	4.327	2.487	4.618	7	Vertosols, Dermosols
51	TAS	East Coast	Pasture	5.220	4.375	5.832	11	Dermosols, Chromosols, Vertosols
52	TAS	Far North East	Cropping	4.522	4.222	5.672	14	Ferrosols
53	TAS	Far North East	Pasture	7.263	6.638	9.082	12	Ferrosols, Dermosols
54	TAS	Northern Midlands	Cropping	2.765	2.130	3.675	43	Dermosols, Chromosols, Vertosols
55	TAS	Northern Midlands	Pasture	3.896	2.552	6.079	22	Dermosols, Chromosols, Vertosols
56	TAS	Sorell	Pasture	6.274	3.453	7.455	7	Dermosols, Chromosols, Vertosols
57	TAS	Southern Midlands	Cropping	2.961	2.311	3.429	19	Dermosols, Chromosols, Vertosols
58	TAS	Southern Midlands	Pasture	4.784	2.876	6.567	20	Dermosols, Chromosols, Vertosols
59	TAS	Southport	Pasture	5.835	5.209	5.912	3	Dermosols

Table 5.5c. Estimated soil carbon levels for Australian soils under agricultural land use. Based on the data from the National Soil Carbon Project (SCaRP) of the CSIRO and the Department of Climate Change and Energy Efficiency (CSIRO 2013b). Note that this is a very preliminary summary of the available data and is indicative only of the overall results.

				Soil organic carbon g/100g 0 to10 cm				
Comb	**State**	**Region**	**Broad Land Use**	**median**	**percentile 20%**	**percentile 80%**	**No of sites**	**Main soils**
64	VIC	Eastern Plains	Pasture	3.308	2.533	4.401	27	Chromosols, Sodosols
65	VIC	North Eastern Plains	Pasture	2.242	2.001	2.580	12	Chrmosols, Dermosols
66	VIC	Gippsland	Crop - pasture rotation	2.666	2.255	3.125	25	Sodosols
67	VIC	Gippsland	Pasture	3.333	2.548	5.100	154	Dermosols, Sodosols, Chromosols, Hydrosols, Podosols, Tenosols
68	VIC	Gippsland	Pasture - irrigated	2.827	2.050	5.283	6	Kandosols
69	VIC	Mallee	Crop - pasture rotation	1.085	0.793	1.110	5	Calacarosols, Vertosols
70	VIC	Mallee	Cropping	0.358	0.246	0.706	183	Tenosols, Calcarosols
71	VIC	Northern Wimmera	Crop - pasture rotation	1.037	0.963	1.145	19	Vertosols
72	VIC	Northern Wimmera	Cropping	0.953	0.787	1.147	113	Vertosols, Sodosols, Chromosols
73	VIC	Nth/Sth Slopes	Crop - pasture rotation	2.245	1.651	3.192	28	Chromosols, Sodosols
74	VIC	Nth/Sth Slopes	Pasture	2.951	2.593	3.743	34	Chromosols, Sodosols
75	VIC	Otways	Pasture	5.566	4.174	8.044	43	Dermosols, Chromosols, Kurosols
76	VIC	Sth Wimmera	Crop - pasture rotation	1.414	1.092	1.608	9	Sodosols
77	VIC	Sth Wimmera	Cropping	1.439	1.048	1.795	49	Vertosols, Sodosols
78	VIC	Sth Wimmera	Pasture	1.509	1.093	1.980	23	Sodosols, Chrmosols
79	VIC	Strzelecki Ranges	Pasture	5.171	4.468	6.216	56	Dermosols, Chromosols, Kurosols
84	VIC	Volcanic Plains	Crop - pasture rotation	2.823	2.317	3.498	30	Chromosols, Sodosols, Vertosols
85	VIC	Volcanic Plains	Cropping	2.174	1.882	2.628	17	Chromosols, Sodosols, Vertosols
86	VIC	Volcanic Plains	Pasture	3.856	3.062	5.110	113	Chromosols, Sodosols, Vertosols

Table 5.5d. Estimated soil carbon levels for Australian soils under agricultural land use. Based on the data from the National Soil Carbon Project (SCaRP) of the CSIRO and the Department of Climate Change and Energy Efficiency (CSIRO 2013b). Note that this is a very preliminary summary of the available data and is indicative only of the overall results.

				Soil organic carbon g/100g 0 to10 cm				
Comb	**State**	**Region**	**Broad Land Use**	**median**	**percentile 20%**	**percentile 80%**	**No of sites**	**Main soils**
87	WA	Albany Sand Plain	Crop - pasture rotation	2.179	1.366	2.975	101	Tenosols, Podosol, Sodosol, Chromosol
88	WA	Albany Sand Plain	Cropping	1.591	1.086	2.259	50	Tenosols, Podosol, Sodosol, Chromosol
89	WA	Albany Sand Plain	Pasture	3.767	2.591	5.872	109	Tenosols, Podosol, Sodosol, Chromosol
90	WA	Avon Basin	Crop - pasture rotation	0.782	0.577	1.026	117	Tenosols
91	WA	Avon Basin	Cropping	0.722	0.565	0.999	34	Tenosols
92	WA	Avon Basin	Pasture	1.038	0.604	1.188	10	Tenosols
93	WA	Esperance Sand Plain	Pasture	2.005	1.583	2.702	102	Tenosols, Sodosols, Chrmosols, Kandosols
94	WA	Geographe	Pasture	4.254	3.247	5.452	200	Tenosols, Chromosols, Kandosols
95	WA	Kojonup	Crop - pasture rotation	3.795	2.883	4.270	27	Sodosols, Kurosols, Chromosols
96	WA	Kojonup	Pasture	2.770	2.231	3.740	26	Sodosols, Kurosols, Chromosols
97	WA	Mingenew	Crop - pasture rotation	0.703	0.550	0.933	106	Kandosols, Dermosols, Tenosols, Chromosols
98	WA	Mingenew	Cropping	0.725	0.633	0.911	73	Kandosols, Dermosols, Tenosols, Chromosols
99	WA	Young River	Crop - pasture rotation	1.498	0.911	1.873	47	Sodosols, Tenosols, Dermosols, Chromosols
100	WA	Young River	Pasture	1.267	1.190	1.420	3	Chrmosols, Tenosols

Figure 5.1. Distribution of soil organic carbon with depth for a set of data including Figure 29 from the CSIRO SCaRP Program. All the data sets show that the major changes in soil organic carbon occur in the top 10 cm.

A. Chan et al (2011)

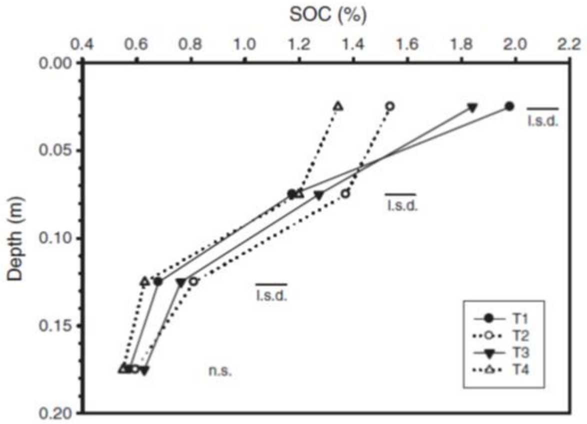
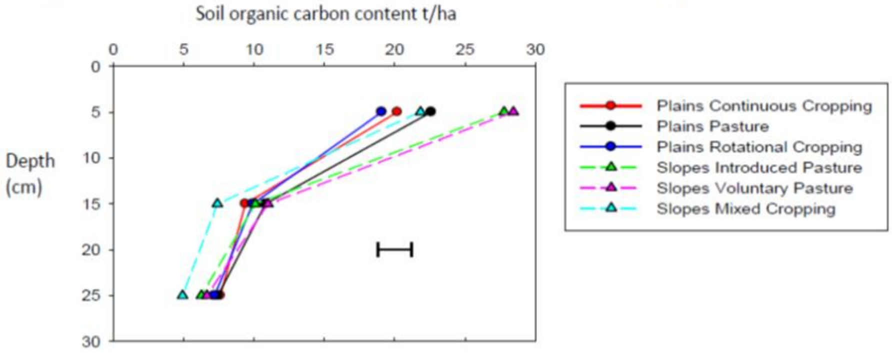


Fig. 2. Soil organic carbon (SOC) concentration distribution with depth under different tillage and stubble management practices (2004) in the SATWAGL long-term field experiment. Horizontal bars are l.s.d. ($P = 0.05$).

B (SCaRP)(CSIRO 2013b)

Figure 29. Comparison of soil organic carbon stocks by major land-use groups.



C. Murphy et al. (2002)

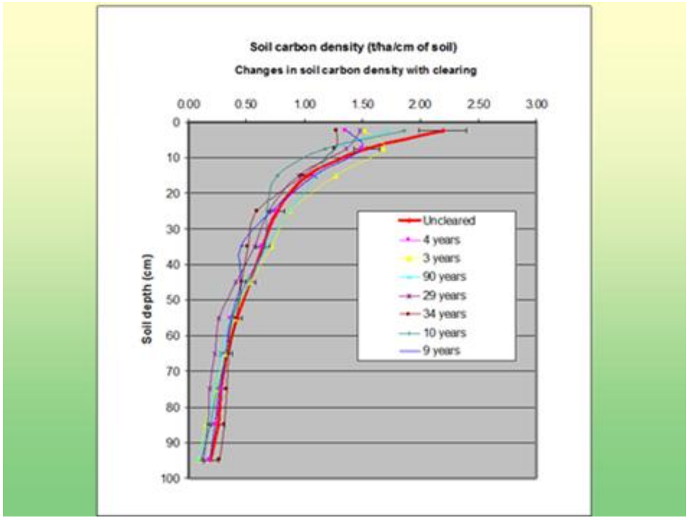
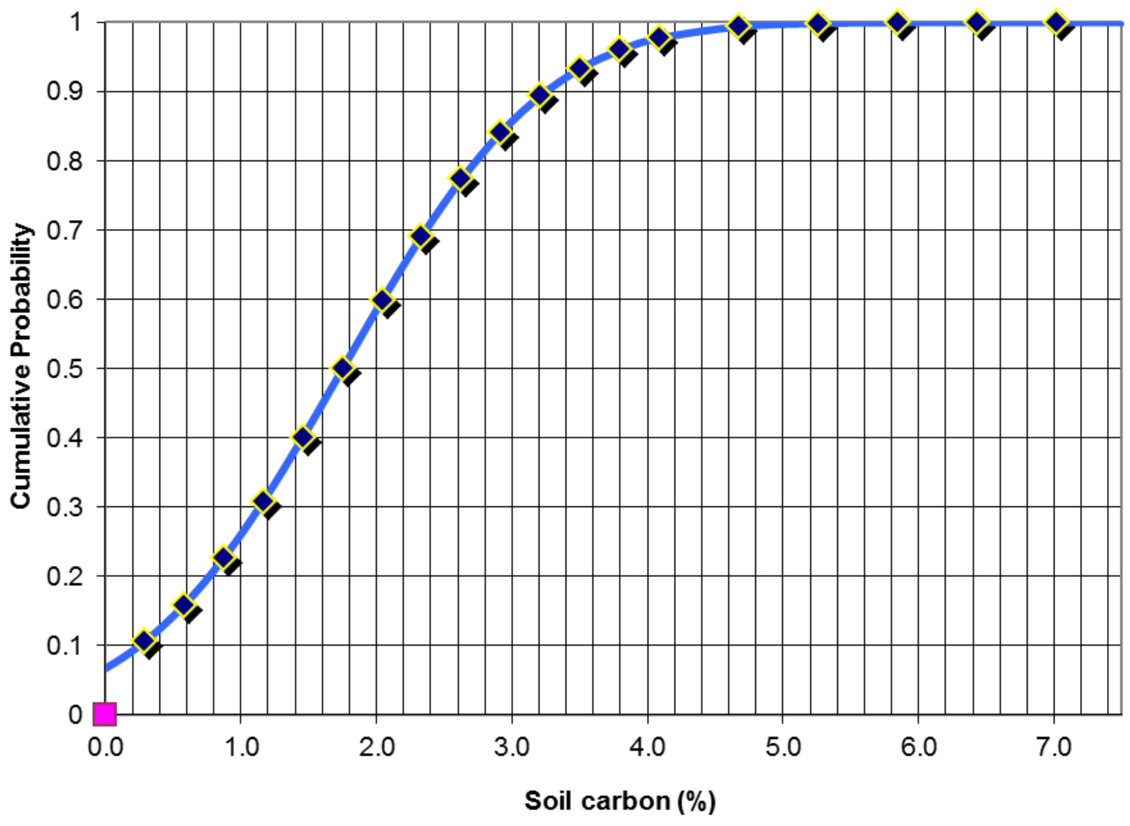


Figure 5.2. Distribution of soil organic carbon 0 – 10 cm in the sout eastern wheat bealt using data from Geeves et al (1995). This is a cumulative distribution using mean 1.75% and standard deviation 1.17%. It indicates that about 25% of soils have soil carbon < 1% and terefore can significantly improve their soil properties by incrasing soil carbon levels.



6. Conclusions- Impact of Soil Organic Matter on Productivity

6.1 Discussion

Some conclusions about the impact of soil organic matter on productivity can be derived from the review. These are outlined below.

1. Soil organic matter can significantly change a wide range of physical, chemical and biological soil properties and affect their functionality (NRCS 2003).
2. The depth or volume of soil that can be influenced by changes in soil organic matter is limited. For soils in many regions it is likely that the changes in soil organic matter will only be a maximum in the top 10 cm of soil and diminish with depth, often minimising by 20 cm. Changes in soil carbon may extend deeper in some circumstances such as those associated with perennial pasture or under higher rainfall. More investigation is required to confirm these conclusions.
3. The soil organic matter fractions can individually influence soil properties more effectively than is indicated by total soil organic matter. This is the case for the effects of particulate soil organic matter effects on soil physical properties and the effects of humus on the soil chemical properties such as cation exchange and nutrient cycling. However it is the case that total soil organic carbon remains a reasonable first approximation to the functionality of soil organic matter on soil properties. It is hypothesised that more information on the soil carbon fractions would be informative on the functionality of soil organic matter when total soil organic carbon levels are in the range of about 0.7% to 2.0 %. Below 0.7% it is likely there is insufficient soil organic matter to maintain functions, and at levels higher than 2.0%, it is likely that there will be sufficient soil organic matter to maintain most functions.
4. Soil organic matter is not an agent to solve all soil limitations and problems. In the case of soil acidification, soil organic matter can increase the soil buffering capacity, but if the acidification pressure remains on the soil, the soil will still acidify with time and the soil organic matter will merely delay the process by a number of years. While this may be economically and environmentally important, it does not resolve the underlying limitation or soil degradation process.
5. In nutrient cycling there is potential conflict in the management of soil organic matter. One objective may be to increase soil organic and soil carbon levels but the second objective is to ensure adequate N, P and S nutrition for crops and pastures (see Janzen 2006). The utilisation of the nutrient pool in soil organic matter requires the mineralisation of the soil organic matter which by definition required the break down and reduction of the soil organic matter pool in the soil. While this does not necessarily mean that both objectives cannot be met by well balanced and strategic management, it is a fundamental conflict requiring the development of specialised management strategies. The key would appear to be in the development of management strategies that ensure rapid nutrient cycling and take full advantage of the capacity of legumes to fix N. This immediately brings the system under a potential acidification pressure which needs to be managed. Hence the need for finely balanced management systems.
6. In the past it has been generally recognised that the use of industrially produced fertilisers could be used to replace the nutrients lost when soil organic matter levels were falling and

so the effects of lower soil organic matter have been less pronounced (Loveland and Webb 2003). However, with the increased cost of fertilisers and the potential environmental constraints on industrially processed N, the relative importance of maintaining nutrient levels through recycling from soil organic matter and plant and animal materials in the soil is increasing.

7. There is a clear link between the build-up of soil organic matter, the fixation of N by legumes and the need for adequate P nutrition. This has been recognised before by Spain et al. (1993), Chan et al. (2010) and Herridge (2011).
8. Some effects of soil organic matter are often likely to be “event based” rather than continuous and unchanging. This especially applies to soil structure where a high intensity storm may cause erosion damage on a poorly structured soil or cause surface crusting that inhibit germination. The effects may be seasonally dependent where rainfall distribution in one season is such that the extra water storage associated with better levels of soil organic matter is not required by the crop. In another season, the extra stored water may prevent severe moisture stress at anthesis.
9. An important effect of soil organic matter is potentially the maintenance of soil in good structural condition that enables the timeliness of agronomic operations such as sowing, spraying for weeds, adding fertiliser and even harvesting.
10. The impact of soil organic matter on productivity is complex because soil organic matter affects a range of soil properties not a single property. It is tempting to refer to the quote from Herridge (2011), as it seems to relate very well to the overall effects of soil organic matter on productivity:

    > “Major leaps in the productivity of agricultural systems rarely arise from the interventions related to single factors, but rather from synergistic interactions among many interventions working together…”
    > (Watt et al. 2006, quoted in Herridge 2011).

    Soil organic matter can influence a number of soil properties and therefore potentially a wide range of functions in soils. By influencing a range of functions the effects of soil organic matter can be substantial.

An example of evaluating the effect of soil organic matter is in predicting how increasing soil organic matter and improving soil structure might affect the amount of water available to a crop. It is possible to consider this in 3 stages:

- Infiltration at the surface – as the rain falls the water has to penetrate into the soil and move through the soil. If the surface seals, the water can be lost as runoff.
- The water has to be stored in the soil. Soil organic matter will increase the amount of water stored in the surface layers.
- The plants have to remove the water from the soil. Most of the water used by a crop will not be stored in the top 10 cm, but the plant will tend to use moisture from deeper in the profile. The plants need to have abundant root growth that accesses a large volume of soil. A successful germination and emergence is essential for this to occur. However some of the water stored in the top 10 cm will flow into the deeper soil and at germination and emergence the water in the top 10cm is potentially very valuable in providing the germinating seed and emerging seedling.

Based on this simple example it is clear that to fully evaluate the impact of soil organic matter on productivity some kind of modelling exercise is required. In this way the effects of changes in soil properties associated with soil organic matter can be evaluated across a wide range of soil types and environmental conditions (van Rees pers. comm.). A set of modelling exercises should be undertaken to fully evaluate the potential impacts of soil organic matter on productivity and on environmental issue such as wind and water erosion.

6.2 An estimate of the economic value of soil carbon

While the review has shown the capacity of soil organic matter to influence a range of functional soil physical, chemical and biological properties and to play an important role in nutrient cycling, there is still a need to quantify the potential economic impact of these effects on soil properties and productivity. Studies to quantify the effects of soil organic matter on yield and productivity and the economic impacts are difficult to undertake because of the complexity of the effects of soil organic matter. However, given the effects quantified in this review it would seem likely that some economic impacts would be detected.

An estimate of the economic value of soil organic matter was undertaken by Ringrose Voase et al. (1997). In an investigation which included yield data from crops and pastures as well as soil data from 80 paddocks , the costs, returns, gross margins and soil properties were compared over 3 years (1992 – 1995). These results are highly localised for a specific soil type and set of enterprises, however the results indicated there was potential for an economic benefit of having higher levels of soil organic matter in soils. The report concludes:

> "Organic carbon content is closely linked to organic matter content which has a range of beneficial effects including improved nutrient cycling, soil structure and soil structure stability. Improved soil structure enhances water entry and reduces runoff and hence erosion. Improved structural stability helps prevention of hardsetting and crusting." pp26

Managing for soil organic matter remains a sound basis for maintaining soil in a good condition for optimising productivity and for maintaining the productive capacity of the soil in the long term. It is soil organic matter that provides much of the physio-chemical activity required by the soil to carry out the functions required for crop and pasture production. Soil organic matter provides a source for the cycling of nutrients in the soil as well as providing a source of food for the microorganisms responsible for the recycling of nutrients. A soil with adequate soil organic matter is always likely to be more productive and have the capacity to sustain productivity than a soil that is deficient in soil organic matter. The most practical way to enhance soil quality is to promote the better management of soil carbon and soil organic matter.

7. Future Issues for Soil Organic Matter and Soil Function

This review has indicated that the following areas of knowledge about the impact of soil organic matter on soil properties and soil productivity require further detailed investigation.

1. The effect of soil organic matter on water holding capacity of soils and infiltration (through aggregate stability) is established but the subsequent impact of this on productivity is event based. Therefore it is difficult to make a single prediction about how changes in water holding capacity and infiltration resulting from changes in soil organic matter will impact on yields. A series of modelling exercises is required to examine how the change in water holding capacity and infiltration will affect yields across a range of seasons, locations and soil types. Included in this study would be any effects of surface crusting to prevent germination and the problems of tillage and sowing that might occur in structurally degraded soil with low soil organic matter levels.
2. The effects of soil organic matter are recorded for a scattered set of soil types and conditions. It is clear that the effect of soil organic matter on soil properties will vary depending soil type depending on clay content, clay type, sesquioxide content, silt content and other soil properties. A more detailed and thorough analysis of soil organic matter effects on the properties of different soil types is required.
3. Soil organic matter is a very heterogeneous material and is made up of different fractions with differing chemical and physical properties. These different fractions will affect soil properties in different ways. How these fractions will affect the soil properties and the critical amounts required to have significant effects requires further investigation. It would generally appear that the particulate organic matter or the coarser fractions largely influence soil physical properties while the finer fractions (humus) have a larger effect on the soil chemical properties but this hypothesis needs to be confirmed. Preliminary analysis indicates that total soil carbon is a first approximation for the amount of most fractions. However in the range of total soil organic carbon levels of 0.7% to 2.0%, the effectiveness of the soil organic matter on different soil properties may depend very much on the relative proportions of the different fractions of soil organic matter. This requires further investigation
4. The fundamental conflict between increasing soil organic carbon and using the mineralisation of soil organic matter as a source of nutrients requires further investigation and resolution. The development of land management strategies to optimise both the increase in soil organic carbon levels and the recycling of nutrients from soil organic matter needs to be a priority. The modelling of nutrient cycling across seasons, locations and soil types would be a recommendation.
5. The importance of phosphorus for increasing soil organic carbon and soil organic matter levels became apparent from the review. The optimal use of P from fertilisers, reserves in soils and recycling is an important aspect of land management.
6. The potential for increasing acidity associated with increasing soil organic matter needs to be managed.

7. Despite its obvious importance and the critical functions it performs in soils, soil biology remains somewhat of an uncertain area of soil science for developing land management strategies to obtain specific objectives to manage soil organic matter. Given the availability of a new range of tests that can provide information on soil biology there is a need to develop a series of guidelines for the interpretation of soil biology tests.
8. The influence of soil organic matter on soil strength and the compaction characteristics needs to be investigated in more detail, at least initially using published soil information.
9. Some quantification of the impact of soil organic matter on wind and water erosion would be a valuable exercise.

Ultimately an economic objective is to determine some economic value for soil organic matter. It might be a difficult task but if at least items 1, 4 and 7 were able to provide some data, this could be done.

8. References

Abu-hamdeh, HN and Reeder, R. (2000). Soil thermal conductivity. Effects of density, moisture, salt concentration and organic matter. Soil Science Society America Journal. 64, 1285 – 1290.

Aitken, RL. (1992). Relationships between extractable Al, selected soil properties, pH buffering capacity and lime requreiemnt in some acidic Queensland soils. Australian Journal of Soil research 30, 119 – 30.

Aitken, RL, Moody, PW and McKinley PG. (1990). Lime requirement of acidic Queensland soils. I relationships between soil properties and pH buffer capacity. Australian Journal of Soil Research 28, 695 – 701.

Aitken, RL, Moody, PW and Dickson, T (1998). Field amelioration of acidic soils in south east Queensland. I Effect of amendments on soil properties. Australian Journal of Agricultural Research. 49, 627 -37

Allison, FE. (1973) Soil organic matter and its role in crop production. Developments in Soil Science 3. Elsevier Scientific Publishing Company, Amsterdam.

Alvarez, R, Evans, LA, Milham, PJ and Wilson, MA. (2004). Effects of humic material on the precipitation of calcium phosphate. Geoderma 118, 245 – 260.

Andales, AA, Batchelor, WD, Anderson, CE, Farnham, DE and Whigham, DK. (2000). Incorporating tillage effects into a soybean model. Agricultural Systems 66, 69 – 98.

Anderson, G, Pidgeon, JD, Spencer, HB and Parks, R. (1980). A new hand held recording penetrometer for soil studies, Journal of Soil Science. 31, 279 – 96

Angers, DA and Carter, MR.(1996). Aggregation and organic matter storage in cool, humid agricultural soils. In MR Carter and BA Stewart (eds). “Structure and Organic Matter in Agricultural Soils”. (CRC Press, Boca Raton). pp193 – 211.

Angus, JF, Bolger, TP, Kirkegaard, JA and Peoples, MB. (2006). Nitrogen mineralisation in relation to previous crops and pastures. Australian Journal of Soil Research 44, 355 – 365.

Atterberg, A . (1911). Uber die physikalische Bodenuntersuchung und uber die Plastizitat der Tone. Udi. Internationale Mitteilungen fur Bodenkunde 1, 10 – 43.

Baldock, JA and Skjemstad, JO. (1999). Soil organic carbon /Soil organic matter. In Peverill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - an interpretation manual. CSIRO Publishing Collingwood Australia.

Barnett, SJ, Roget, DK and Ryder, MH (2006). Suppression of Rhizoctonia solani AG-8 induced disease on wheat byn the interaction between Pantoea, Exiguobacterium and Microbacterium. Australian Journal of Soil Research 44, 331 – 342.

Barrow, NJ. (1969). Accumulation of organic matter under pastures. Australian Journal of Experimental Agriculture and Animal Husbandry 9, 437 -444.

Baver, LD. (1940). Soil Physics. John Wiley and Sons, New York.

Bauer, A and Black, AL. (1992). Organic carbon effects on available water capacity of three soil textural groups. Soil Society of America Journal. 56, 248 – 254.

Bell, MJ, Moody, PW, Yo, SA and Connolly, RD. (1999). Using active fractions of soil carbon as indicators of the sustainability of Ferrosol farming systems. Australian Journal Soil Research 37, 279 – 287.

Bell, M, Seymour, N, Stirling, GR, Stirling, AM, Van Zwieten, Vancov, T, Sutton, G and Moody, P. (2006). Impacts of management on soil biota in Vertosols supporting the broadacre grains industry in northern Australia. Australian Journal of Soil research 44, 433 – 451.

Bending, GD, Putland, C and Rayns, F. (2000). Changes in microbial community metabolism and labile organic matter fractions as early indicators of the impact of management on soil biological quality. Biology and Fertility of Soils 31, 78 – 84.

Blair, GJ, Lefroy, RDB & Lisle, L. 1995. Soil carbon fractions based on their degree of oxidation, and the development of a carbon management index for agricultural systems. Australian Journal of Agricultural Research, 46, 1459-1466.

Bolan, NS, (1991). A critical review on the role of mycorrhizal fungi in the uptake of phosphorus by plants. Plant and Soil. 134, 189 – 207.

Bolan, NS, Robson, AD, Barrow, NJ and Aylmore, LAG. (1984). Specific activity of phosphorus in mycorrhizal and non-mycorrhizal plants in relation to the availability of phosphorus to plants. Soil Biology and Biochemistry 16, 299 – 304.

Bolanos, L, Redondo,- Nieto, M, El-Hamdaoui, A and Bonilli, A. (2002). Interaction of Boron and Calcium in the Rhizobium-Legume N2-Fixing Symbiosis. In Goldbach, H.E.; Rerkasem, B.; Wimmer, M.A.; Brown, P.H.; Thellier, M.; Bell, R.W. (Eds.). Boron in Plant and Animal Nutrition. (Springer, US).

Bot, A and Benites, J. (2005). Chapter 5. Creating drought resistant soil. In “The Importance of Soil Organic Matter. Key to Drought resistant Soil and Sustained Food and Production”. FAO Soil Bulletin No 80. Food and Agriculture organisation of the United Nations, Rome.

Bowen, GD. (1993). Microbial determinants of plant nutrient uptake. In Soils: An Australian Viewpoint. CSIRO Publications, Collingwood, Victoria, Australia

Braunack, MV and Dexter, AR. (1989). Soil aggregation in the seedbed: a review. II. Effect of aggregate sizes on plant growth. Soil and Tillage Research 14, 281 – 298.

Brussaard, L, deRuiter, PC abd Broan, GG. (2007). Soil biodiversity for agricultural sustainability. Agriculture, Ecosystems and Environment 121, 233 – 44.

Bucking, H and Shachar-Hill, Y. (2005). Phosphate uptake, transport and transfer by the arbuscular mycorrhizal fungus *Glomus intraradices*is stimulated by increased carbohydrate availability. New Phytologist 165, 899 – 912.

Burns, RG, De Forest, JL, Marxsen, J, Sinsabaugh, RL, Stromberger, ME, Wallenstein, MD, Weintraub, MN and Zoppin, A. (2013). Soil enzymes in a changing environment: Current knowledge and future directions. Soil Biology and Biochemistry. 58, 216 – 234.

Burris, RH. (2002). Biological nitrogen fixation. In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker. New York.

Cambardella, CA. (2002). Aggregation and organic matter. In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker. New York.

Campbell, GS. (1974). A simple model for determining unsaturated conductivity from moisture retention data. Soil Science 117, 311-314.

Carter, MR. (1992). Influence of reduced tillage systems on organic matter, microbial biomass, macroaggregate distribution and structural stability of the surface soil in a humid climate. Soil and Tillage Research 23, 361 – 372.

Carter, MR and Mele, PM. (1992). Changes in microbial biomass and structural stability at the surface of a duplex soil under direct drilling and stubble retention in north-eastern Victoria Australian Journal of Soil Research 30, 493 - 503

Carter, MR, Angers, DA and Kunelius, HT. (1994). Soil structural form and stability and organic matter under cool-season perennial grasses. Soil Science Society of America Journal. 58, 1194 – 99.

Carter, MR, Angers, DA, Gregorich, EG and Bolinder, MA. (2003). Characterizing organic matter retention for surface soils in eastern Canada using density and particle size fractions. Canadian Journal of Soil Science. 83, 11 – 23.

Cass, A. (1999). Interpretation of some soil physical indicators for assessing soil physical fertility. In Peverill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - an interpretation manual. CSIRO Publishing Collingwood Australia.

Chan, KY and Mead, A. (1988). Physical properties of a sandy loam under different tillage practices. Australian Journal of Soil Research. 26, 549 – 59.

Chan, KY, Roberts, WP, Heenan, DP. (1992). Organic carbon and associated soil properties of a red earth after 10 years of rotation under different stubble and tillage practices. Australian Journal of Soil Research 30, 71 – 83.

Chan, KY, Oates, A, Li, GD, Conyers, MK, Prangnell, RJ, Polie, G, Liu, DL, Barchia, IM, (2010). Soil carbon stocks under different pastures and pasture management in the higher rainfall areas of south-eastern Australia. Australian Journal of Soil Research. 48, 7–15.

Chan, KY, Conyers, MK, Li, GD, Helyar, KR, Poile, G, Oates, A and Barchia, IM. (2011). Soil carbon dynamics under different cropping and pasture management in temperate Australia. Results of three long term experiments. Soil Research. 49, 320- 328.

Chaney, K and Swift, RS. (1984). The influence of organic matter on aggregate stability in some British soils. Journal of Soil Science 35, 223 - 230

Clarke, AL, Greenland, DJ and Quirk, JP. (1967).Changes in some physical properties of the surface of an impoverished red-brown earth. Australian Journal of Soil Research 5, 59 -68.

Conyers, MK, Poile, GJ, Oates, AA, Waters, D and Chan, Y. (2011). Comparison of three carbon determination methods on naturally occurring substrates and the implication for the quantification of ‘soil carbon’. Australian Journal of Soil Research 49, 27 – 33.

Conyers, M, Newton, P, Condon, J, Poil, G, Mele, P and Ash, G. (2012). Three long-term trials end with a quasi-equilibrium between C, N and pH and implication for C sequestration. Soil Research 50, 527 – 35.

Cookson, WR, Marschner, Clark, IM, Milton, N, Smirk, MN, Murphy, DV, Osman, M, Stockdale, EA and Hirsch, PR. (2006). The influence of season agriculture management and soil properties on gross nitrogen transformations and bacterial community structure. Australian Journal of Soil Research 44, 453 – 465.

Cresswell, HP and Paydar, Z. (1996) Water retention in Australian soils. I. Description and prediction using parametric function. Australian Journal of Soil Research. 34, 195 – 212.

CSIRO. (2013a). Soil Carbon Research Program (SCaRP). Project 1. Field and laboratory Methodologies. A collaborative project supported by the Climate Change Reduction Program of the Australian Department of Agriculture, Fisheries and Forestry and the Grains Research and Development Corporation involving staff and contributions from CSIRO, University of Western Australia, Department of Agriculture and Food of Western Australia, Victorian Department of Primary Industries, Murray Catchment Management Authority, Department of Environment and Natural Resources of South Australia, Queensland Government, University of New England, New South Wales Department of Primary Industries, The University of Tasmania, and The Tasmanian Institute of Agricultural Research.

CSIRO. (2013b). Soil Carbon Research Program (SCaRP). A summary of program results. A collaborative project supported by the Climate Change Reduction Program of the Australian Department of Agriculture, Fisheries and Forestry and the Grains Research and Development

Corporation involving staff and contributions from CSIRO, University of Western Australia, Department of Agriculture and Food of Western Australia, Victorian Department of Primary Industries, Murray Catchment Management Authority, Department of Environment and Natural Resources of South Australia, Queensland Government, University of New England, New South Wales Department of Primary Industries, The University of Tasmania, and The Tasmanian Institute of Agricultural Research.

Curtain, D and Smillie, GW, (1976). Estimation of the components cation exchange capacity from measurements of specific surface area and organic matter. Soil Science Society of America Journal 40, 461-2.

da Silva, A. P. and Kay, B. D. (1997). Estimating the least limiting water range of soils from properties and management. Soil Science Society of America Journal 61, 877–883

Dalal, RC. (1998). Soil microbial biomass—what do the numbers really mean? Australian Journal of Experimental Agriculture. 38, 649–65

Dalal R, Wang W, Robertson GP, Parton WJ, Myer CM, Raison RJ. (2003a) Emission sources of N2O from Australian agricultural and forest lands and mitigation options. Technical report no 35, National carbon accounting system, Australian Greenhouse Office, Australian Government, Canberra.

Dalal, RC, Wang W, Robertson GP, Parton WJ. (2003b) Nitrous oxide emission from Australian agricultural lands and mitigation options: a review. Australian Journal of Soil Research 41:165–195.

Danalatos, NG, Kosmas, CS, Driessen, PM and Yassoglou, N. (1994). Estimation of the draining soil moisture characteristics from standard data as recorded in soil surveys. Geoderma 64, 155 – 65.

DECCW (2009), Protocols for Soil Condition and Land Capability Monitoring, prepared by G. Bowman, G.A. Chapman, B.R. Wilson, G. Bowman, B.W. Murphy, B.R. Jenkins, T. Koen, J.M. Gray, H.B. Milford and D.T. Morand, Department of Environment, Climate Change and Water NSW, Sydney.

Degens, BP. (1997). Macro-aggregation of soils by biological bonding and binding mechanisms and the factors that affect these: A review. Australian Journal of Soil Research 35, 431 – 460.

Degens, BP, Schipper, LA Sparling, GP and Vojvodic – Vukovic, M. (2000). Decreases in organic C reserves in soil can reduce catabolic diversity of soil microbial communities. Soil Biology and Biochemistry. 32, 189 – 96.

Degens, BP, Schipper, LA, Sparling, GP and Duncan, LC. (2001). Is the microbial community in a soil with reduced catabolic diversity less resistant to stress or disturbance? Soil Biology and Biochemistry. 33, 1143 – 1153.

Dexter, AR (1988). Advances in characterization of soil structure. Soil and Tillage Research. 11, 199 – 238.

Dexter, Anthony R and Zoebisch, Michael, A. (2002). Degradation, critical limits of soil properties and irreversible. In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker, New York.

Donahue, Rl, Miller, RW, Schickluna, JC. (1983). Soils: An introduction to Soils and Plant Growth. Prentice-Hall International. New Jersey, USA.

Drees, LR and Hallmark, CT. (2002). Inorganic carbon analysis. . In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker. New York.

Edwards, K and Zierholz, C. (2007) Soil Formation and erosion rates. In Peter EV Charman and Brian W Murphy (eds). Soils-Their Properties and Management. 3rd edition. Oxford University Press, Melbourne.

Ehlich, PR and Wilson, EO. (1991). Biodiversity studies: science and policy. Science 253, 758 – 762.

Ekwue, EI. (1990). Organic matter effects on soil strength properties. Soil and Tillage Research 16, 289 -97.

Emerson, WW. (1967). A classification of soil aggregates based on their coherence in water. Australian Journal of Soil Research. 5, 47 – 57.

Emerson, WW. (1991). Structure decline in soil, assessment and prevention. Australian Journal of Soil Research. 29, 905 – 922.

Emerson, WW and McGarry, D (2003). Organic carbon and soil porosity. Australian Journal of Soil Research. 41, 107 -18.

Fenton, G and Helyar, K. (2007) Soil acidification. In Peter EV Charman and Brian W Murphy (eds). Soils-Their Properties and Management. 3rd edition. Oxford University Press, Melbourne.

Field, DJ, McKenzie, DC and Koppi, AJ. (1997) Development of an improved Vertosol stability test for SOILpak. Australian Journal of Soil Research. 35, 843 – 52.

Flanagan, D.C., and M.A. Nearing (eds.). 1995. USDA-Water Erosion Prediction Project (WEPP) Hillslope Profile and Watershed Model Documentation. NSERL Report No. 10, National Soil Erosion Research Laboratory, USDA-Agricultural Research Service, West Lafayette, Indiana.

Foster, RC. (1993). The plant root environment. In Soils: An Australian Viewpoint. CSIRO Publications, Collingwood, Victoria, Australia.

Fontvieille, DA, Outaguerouine, A and Thevenot, DR. (1991). Fluorescein diacetate hydrolysis as a measure of microbial activity in aquatic systems: application to activated sludges. Environmental Technology 13, 531 – 540.

Geeves, GW, Cresswell, HP, Murphy, BW, Gessler, PE, Chartres, CJ, Little, IP and Bowman, GM. (1995). The physical, chemical and morphological properties in the wheat-belt of south eastern New South Wales and north eastern Victoria. CSIRO Division of Soils, Canberra and NSW Department of Conservation and land Management.

Geeves, GW, Craze, B and Hamilton, GH. (2007a). Soil physical properties. . In Peter EV Charman and Brian W Murphy (eds). Soils-Their Properties and Management. 3rd edition. Oxford University Press, Melbourne.

Geeves, GW, Leys, JF and Mctainsh, GH. (2007b). Soil erodibility to water and wind. In Peter EV Charman and Brian W Murphy (eds). Soils-Their Properties and Management. 3rd edition. Oxford University Press, Melbourne.

Gerakis, A and Ritchie, JT. (2002). Plant available water. . In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker. New York.

Gianfreda, L and Bollag, JM. (1994). Effect of soils on the behaviour of immobilized enzymes. Soil Science Society of America Proceedings 58, 1672 – 1681.

Gibson. TS, Chan, KY, Sharma, G and Shearman, R. (2002). Soil Carbon Sequestration Utilising Recycled Organics - A review of the scientific literature. Project 00/01R-3.2.6A. The Organic Waste Recycling Unit, NSW Agriculture. Report prepared for Resource NSW. http://www.environment.nsw.gov.au/resources/warr/SPD_ORG_0208SoilCarbonSeq.pdf.

Gonzalez-Quinones V, Stockdale,EA, Banning, NC, Hoyle, FC, Sawada, Y. Wherrett, AD, Jones DL and Murphy, DV. (2011). Soil microbial biomass – interpretation and consideration for soil monitoring. Soil Research 49. 287 – 304.

Gray, J, Bishop, T, Smith, P, Robinson, N and Brough, D. (2012). A pragmatic quantitative model for soil carbon distribution in eastern Australia. In Minansy, Malone and McBratney (eds). Digital Soil Assessments and Beyond. Taylor and Francis Group, London. ISBN 978 – 0 415-62155-7.

Gupta, VVSR, Roper, MM and Roget, DK. (2006). Potential for non-symbiotic N2-fixation in different agroecological zones of southern Australia. Australian Journal of Soil Research 343 – 54.

Hairsine, P, Murphy, B, Packer, I and Rosewell, C. (1993). Profile of erosion from a major storm in the South-east cropping zone. Australian Journal of Soil and Water Conservation 6 (4), 50 – 55.

Hallsworth, EG and Wilkinson, GK. (1958). The contribution of clay and organic matter to the cation exchange capacity of soils. Journal of Agricultural Science 51, 1 – 3.

Hamblin, A. (1987). The effect of tillage on soil physical conditions. In PS Cornish and JE Pratley (eds). Tillage – new directions in Australian Agriculture. Inkata Press, Melbourne.

Hamblin Ann and Kyneur, Geoff. (1993). Trends in wheat yields and soil fertility in Australia. Department of Primary Industries and Energy. Bureau of Resource Sciences. Australian Government Publishing Service, Canberra, Australia.

Hargrove, WL and Thomas, GW. (1981). Effect of organic matter on exchangeable aluminium and plant growth in acid soils. In Chemistry in the soil environment. American Society of Agronomy and Soil Science Society of America. Madison WI.

Hayden, HL, Drake, J, Imhof, M, Oxley, PA, Norng, Sorn, Mele, PM. (2010). The abundance of nitrogen cycle genes (amoA and nifA)depends on land use and soil type in south eastern Australia. Soil Biology and Biochemistry 42, 1774 1783.

Haynes, RJ. (2000). Interactions between soil organic matter status, cropping history, method of quantification and sample pretreatment and their effects on measured aggregate stability. Biology and Fertility of Soils. 30, 270 – 275.

Hazelton, Pam and Murphy, Brian. (2007) Interpreting Soil test – What do all the numbers Mean? CSIRO Publishing, Collingwood, Australia.

Heanes, DL. (1984). Determination of total organic –C in soils by an improved chromic acid digestion and spectrophotomic procedure. Communication in Soil Science and Plant Analysis 15, 1191 – 1213.

Helling, KR, Chesters, G and Corey, RB. (1964). Contribution of soil organic matter and clay to soil cation exchange capacity as affected by pH on the saturation solution. Soil Science Society Proceedings. 28, 517 – 20.

Helyar, KR and Porter, WM. (1989). Soil acidification, its measurement and the process involved. In AD Robson (ed) “Soil Acidity and Plant growth”. pp 61 – 102. Academic Press, Sydney.

Helyar, KR, Cregan, PD and Godyn, DL. (1990). Soil acidity in New South Wales – current pH values and estimates of acidification rates. Australian Journal of Soil Research. 28, 523 – 37.

Herridge, David. (2011). Managing Legumes and Fertiliser N for Northern Grains Cropping. Grains Research and Development Corporation. DRDC Project UNE00014. ISBN 978-1-921779-33-6. GRDC, Kingston ACT, Australia.

Hicks, RW. (2007). Soil engineering properties. In Peter EV Charman and Brian W Murphy (eds). Soils-Their Properties and Management. 3rd edition. Oxford University Press, Melbourne.

Himes, FL. (1998). Nitrogen, sulphur and phosphorus and the sequestration of carbon. In R Lal et al. (eds) Soil Processes and the Carbon Cycle. CRC Press, Boca Raton, FL. Pp315 – 319.

Hoyle, FC and Murphy, DV. (2006). Seasonal changes in microbial function and diversity associated with stubble retention versus burning. Australian Journal of Soil Research 44, 407 – 423.

Huang, PM. (2004). Soil mineral – organic matter – microorganism interactions: fundamentals and impacts. Advances in Agronomy. 82, 391 – 472.

Hudson, BD. (1994). Soil organic matter and available water capacity. Journal of Soil and Water Conservation. 49, 1898 – 194.

Hughes, B, Jacka, J, Lewis, D and Prance, T. (1996). Guidelines for interpreting soil results. South Australia Soil and Plant Analysis Service. Primary Industries, South Australia.

Ibekwe, AM and Kennedy, AC. (1999). Fatty acid methyl ester (FAME) profiles as a tool to investigate community structure of two agricultural soils. Plant and Soil 206, 151 – 161.

Insam, H. (2001). Developments in soil microbiology since the mid 1960s. Geoderma. 100, 389 – 402.

Isbell, RF. (2002). The Australian Soil Classification. CSIRO Publishing, Collingwood Victoria, Australia.

Jastrow, JD. (1996). Soil aggregate formation and the accrual of particulate and mineral associated organic matter. Soil Biology and Biochemistry 28, 665 – 676.

Janzen, HH, Larney, FJ and Olson, BM. (1992). Soil quality factors of problem soils in Alberta. Proceedings of the Alberta Soil Science Workshop, 17 – 28.

Janzen, HH. (2006). The soil carbon dilemma: Shall we hoard it or use it? Soil Biology & Biochemistry 38 (2006) 419–424.

Kay, BD and Angers, DA. (1999). Soil Structure. In ME Sumner (ed) ”Handbook of Soil Science”. (CRC Press. Boca Raton USA).

Kay, BD, da Silva, AP and Baldock, JA. (1997). Sensitivity of soil structure to changes in organic carbon content using pedotransfer functions. Canadian Journal of Soil Research. 77, 655 – 667.

Keller, Thomas and Dexter, Anthon.y (2012). Plastic limits of agricultural soils as functions of soil texture and organic matter. Soil Research. 50, 7 – 17.

Kirkby (2002). Liquid and plastic limits. In Neil McKenzie, Kep Coughlan and Hamish Cresswell (eds). Soil Physical Measurement and Interpretation for Land Evaluation. CSIRO Publishing, Collingwood Australia.

Kirkby, CA, Kirkegaard, Richardson, AE, Wade, LJ, Blanchard, C and Batten, G. (2011). Stable soil organic matter: A comparison of C:N:P:S ratios in Australian and other world soils. Geoderma, 163, 197-208.

Krull, ES, Skjemstad, JO and Baldock, JA. (2004). Functions of Soil organic matter and the effect on soil properties. CSIRO Land and Water, PMB2, Glen Osmond SA 5064. GRDC Project No CSO 00029. Residue, Soil Organic Carbon and Crop Performance.

Ladd, JN and Russell, JS. (1993). Chapter 37 Soil Nitrogen. In “Soils An Australian Viewpoint”. Division of Soils, CSIRO. pp589 – 607. (CSIRO:Melbourne).

Lal, R. (1979). Physical properties and moisture retention characteristics of some Nigerian soils. Geoderma. 21, 209 – 223.

Lawrence, D, Bell, M, Gentry, J, Schwenke and Muir, S. (2008). Soil health in Broadacre Crops. Queensland and northern NSW. Project Sustainable soil health management workshops for northern broadacre cropping industries. Queensland Department of Primary industries and NSW Department of Primary Industries.

Lawton, JH and Brown, VK. (1994). Redundancy in ecosystems. In ED Schulze and HA Mooney (eds). “Biodiversity and Ecosystem Function”. (Springer, Berlin). pp255 – 270.

Letey, J. (1985). Relationships between soil physical properties and crop production. Advances in Soil Science. 1, 277-94.

Letey, J. (1991). The study of soil structure: science or art. Australian Journal of Soil Research 29, 699 -707.

Lewis, DC. (1999) Sulfur. In Peverrill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - an interpretation manual. CSIRO Publishing Collingwood Australia.

Leys, J, Semple, W, Raupach, M, Findlater, P and Hamilton, GJ. (2002). Measurement of size distributions of dry soil aggregates. In Neil McKenzie, Kep Coughlan and Hamish Cresswell (eds). Soil Physical Measurement and Interpretation for Land Evaluation. CSIRO Publishing, Collingwood Australia.

Loch, RJ and Foley, JL. (1994). Measurement of aggregate breakdown under rain. Comparison with tests of water stability and relationship with field measurement of infiltration. Australian Journal of Soil Research 32 (4). 701 – 720.

Lopes, AS and Cox, FR. (1977). A study of the fertility status of surface soils under “cerrado” vegetation in Brazil. Soil Science Society of America Journal. 41, 742 -7.

Loveday, J and Pyle, J. (1973). The Emerson dispersion test and its relationship to hydraulic conductivity. Technical Paper 15. CSIRO Division of Soils, CSIRO, Canberra, Australia.

Loveland, P and Webb, J. (2003). Is there a critical level of organic matter in the agricultural soils of temperate regions: a review. Soil and Tillage Research 70, 1 – 18.

Lynch, JM. (1984). Interactions between biological processes, cultivation and soil structure. Plant and Soil 76, 307 – 318.

Macdonald, LM and Baldock, JA. (2010). Manipulating soil carbon and nutrients: advancing understanding of soil nutrient cycling using approaches based on ecological stoichiometry. CSIRO Land & Water. National Research Flagships, Sustainable Agriculture. CSIRO. Private Mail Bag 2,Glen Osmond, SA 5064, Australia and Grains Research and Development Corporation.

McCaig, AE, Glover, LA and Prosser, JI. (1999). Molecular Analysis of Bacterial Community Structure and Diversity in Unimproved and Improved Upland Grass Pastures. Applied Environmental Microbiology. April; 65(4): 1721–1730.

McCarty, G.W., J.M. Bremmer and H.S. Chai. (1989) Effects of N-(n-butyl) thiophosphoric triamide on hydrolysis of urea by plant, microbial and soil urease. Biology and Fertility of Soils 8, 123-127.

McGill, WB and Cole, CV. (1981). Comparative aspects of cycling organic C, N, S and P through soil organic matter. Geoderma 26, 267 – 286.

McIntyre, DS. (1955) Effect of soil structure on wheat germination in a red-brown earth. Australian Journal of Agricultural Research. 6, 797 – 803.

McKenzie N, Jacquier, D, Isbell, R and Brown, K. (2004). Australian Soils and Landscapes – An illustrated compendium. CSIRO Publishing. Collingwood Victoria Australia.

McLaughlin, MJ, McBeath, TM Smernik, R, Stacey, SP, Ajiboye, B and Guppy, C. (2011). The chemical nature of P accumulation in agricultural soils – implications for fertiliser management and design: an Australian perspective. Plant Soil 349, 69 – 87.

MacEwan R.J. (2007). Soil Health for Victoria’s Agriculture- Context, Terminology and Concepts. MIS 07898 Final Report. Department of Primary Industries Research Victoria. Bendigo. March / 2007. State of Victoria. http://www.dpi.vic.gov.au/vro/

Macks, SP, Murphy, BW, Cresswell, HP and Koen, TB. (1996). Soil friability in relation to management history and suitability for direct drilling. Australian Journal of Soil Research. 34, 343 – 360.

Manrique, LA and Jones, CA. (1991). Bulk density of soils in relation to physical and chemical properties. Soil Science Society of America Journal. 55, 476 – 81.

Marshal, TJ and Holmes, JW. (1979). Soil Physics. Cambridge University Press. London, Melbourne.

Martin, JK. (1993). Biology of the rhizosphere. In Soils: An Australian Viewpoint. CSIRO Publications, Collingwood, Victoria, Australia.

Marx, MC, Wood, M and Jarvis, SC. (2001). A microplate fluorometric assay for the study of enzyme diversity in soils. Soil Biology and Biochemistry 33, 1633 – 1640.

Mele, PM and Crowley, DR. (2008). Biological properties of soils: Application of self-organising maps to assess soil quality (A review). Agriculture, Ecosystems and Environment 126, 139 – 152.

Millar, CE and Turk, LM. (1943). Fundamentals of Soil Science. John Wiley and Sons Inc., New York.

Millington, RJ (1959). Establishment of wheat in relation to apparent density of the surface soil. Australian Journal of Agricultural Research 10, 487 – 94.

Minasny, B, McBratney, AB and Bristow, KL. (1999). Comparison of different approaches to the development of pedotransfer functions for water-retention curves. Geoderma 93, 225 – 53.

Minasny, B. and McBratney, A. (2002). Neuroman Evaluator. Neural Networks Pedotransfer Functions. University of Sydney Centre for Precision Agriculture.

Mondini, C, Fornasier, F, Sinicco, T. (2004). Enzymatic activity as a parameter for the characterization of the composting process Soil Biology and Biochemistry. 36, 1587-1594.

Moody, PW and Stanley, J. (1979). Factors affecting phosphorus retention by basaltic soils on the Atherton Tableland, Queensland. Journal of the Australian Institute of Agricultural Science 45, 201-2.

Moody, PW. (1994). Chemical fertility of krasnozems – a review. Australian Journal of Soil Research 32, 1015 – 41.

Moody, PW and Bolland, MDA. (1999). Phosphorus. In Peverrill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - an interpretation manual. CSIRO Publishing Collingwood Australia.

Moyano, F., Manzoni, S., and Chenu, C. (2013). Responses of soil heterotrophic respiration to moisture availability: An exploration of processes and models (review). Soil Biology and Biochemistry 59, 72–85,

Munkholm, LJ. (2011). Soil friability: A review of the concept, assessment and effects of soil properties and management. Geoderma. 167 – 168, 236 – 246.

Murphy, BW. (1994). Mechanisms that cause changes in soil structure - Emphasising the wheatbelt soils of south eastern Australia. PhD Thesis, University of New South Wales, Sydney, Australia.

Murphy, B., Rawson, A., Ravenscroft, L., Rankin, M. & Millard, R. (2002). Paired site sampling for soil carbon estimation – New South Wales. Australian Greenhouse Office, Technical Report No 34, Canberra, Australia.

Murphy. BW, Young, J, Rawson, A and Marchant, A. (2010). Soil Carbon Benchmark Matrix for the Central West Catchment Management Authority. Central West Catchment Authority and NSW Department of Environment, Climate Change and Water.

Murphy, DV, Cookson, WR, Brainbridge, M, Marschner, P, Jones, DL, Stockdale, EA and Abbott, LK. (2011). Relationships between soil organic matter and the soil microbial biomass (size, functional diversity and community structure) in crop and pasture systems in a semi-arid environment). Soil Research 49, 582 – 594.

Murphy, BW, Crawford, MH, Duncan, DA, McKenzie, DC and Koen, TB. (2013). The use of visual soil assessment schemes to evaluate surface structure in a soil monitoring program. Soil Use and Management 127, 3 – 12.

Nelson DW and Sommers LE. (1996) Total Carbon, Organic Carbon and Organic Matter. In Methods of Soil Analysis Part 3 Chemical Methods p1001-2 SSSA Madison. USA.

Nelson, DR and Mele, PM. (2006). The impact of crop residue amendments and lime on microbial community structure and nitrogen –fixing bacteria in the wheat rhizosphere. Australian Journal of Soil Research. 44, 319 – 329.

Nicholson, PS and Hirsch, PR. (1998). The effects of pesticides on the diversity of culturable bacteria. Journal of Applied Microbiology 84, 551 – 558.

NRCS. (2003). Managing soil organic matter. The key to air and water quality. Technical Note No 5. Natural Resources Conservation Service. United States Department of Agriculture. soils.usda.gov/sqi.

Oades, JM. (1984). Soil organic matter and structural stability: mechanisms and implications for management. Plant and Soil. 76, 319 -337.

Oades, JM. (1993). The role of soil biology in the formation, stabilization and degradation of soil structure. Geoderma. 56, 377-400.

Oades, JM, Gillman, GP and Uehara, G. (1989). Interactions of soil organic matter and variable charge clays. In DC Coleman, JM Oades, And G. Uehara, (eds). Dynamics of soil organic matter in tropical ecosystems. pp 69 -95. University of Hawaii Press, Honolulu.

Ojeniyi, So and Dexter, AR. (1979). Soil factors affecting the macrostructures produced by tillage. Transactions of the American Society of Agricultural Engineers. 22, 339 – 343.

Onstad, CA, Wolfe, ML, Larson, CL and Slack, DC. (1984). Tilled soil subsidence during repeated wetting. Transactions of the American Society of Agricultural Engineering 27, 733 -36.

Parfitt, RL, Giltrap, DJ and Whittin, JS. (1995). Contribution of organic matter and clay minerals to the cation exchange capacity of soils. Communications in Soil Science and Plant Analysis 26, 1343 -55.

Paydar, Z and Cresswell HP. (1996) Water retention in Australian soils. II. Prediction using particle size bulk density and other soil properties. Australian Journal of Soil Research, 34, 679 – 93.

Peoples, MB. (2002). Biological fixation, contributions to agriculture. In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker. New York.

Peverill,KI , Margetts, SW, Brown, AJ, Grenhill, NB and Monro, GR. (1991). The Analytical and Interpretive Service. Ag-Plus, Melbourne.

Peverill, K. I., Sparrow, L. A., and Reuter, D. J. (1999). 'Soil Analysis. An Interpretation Manual.' (CSIRO Publishing: Collingwood.)

Piccolo, A, Pietramellara, G and Mbagwu, JSC. (1997). Use of humic substances as soil conditioners to increase aggregate stability. Geoderma 75, 267 – 277.

Pierret, A, Moran, CJ and Pankhurst, CE. (1999). Differentiation of soil properties related to the spatial association of wheat roots and soil mcaropores. Plant and Soil. 211, 51 – 58.

Praveen – Kumar, T, Jagadish, C, Panwar, J, and Kathju, S. (2003). A rapid method for assessment of plant residue quality. Journal of Plant Nutrition and Soil Science. 166, 662.

Preston, S, Wirth, S, Ritz, K, Griffith, BS and Young, IM. (2001). The role played by microorganisms in the biogenesis of soil cracks: importance of substrate quantity and quality. Soil Biology and Biochemistry 33, 1851 – 1858.

Probert, ME. (1993). Organic phosphorus and sulfur. In Soils: An Australian Viewpoint. CSIRO Publications, Collingwood, Victoria, Australia.

Purdie, BR. (1998). Understanding and interpreting soil chemical and physical data. In: "Soil Guide. A Handbook for Understanding and Managing Agricultural Soils." Bulletin 4343. Agriculture Western Australia.

Quidea, Sylvie A. (2002). Organic matter accumulation. In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker, New York.

Quilty, JR and Cattle, SR. (2011). Use and understanding of organic amendments in Australian agriculture. Soil Research 49, 1 – 26.

Raaijmakers, JM, Paulitz, TC, Steinberg, C, Alabouvette, C, Moënne-Loccoz, Y. (2009). The rhizosphere : a playground and battlefield for soilborne pathogens and beneficial microorganisms. Plant and Soil 321, 341 – 361.

Rawls, WJ, Ahuja, LR and Brakensiek, DL. (1992). Estimating soil hydraulic properties from soils data. In M. Th. Van Genuchten and FJ Leij (eds). Indirect methods for estimating hydraulic properties of unsaturated soils. Proceedings of International Workshop on Indirect Methods for Estimating the Hydraulic Properties of Unsaturated Soils. Riverside, California, October 11 – 13, 1989. US salinity Laboratory, Agricultural research Service, US Department of Agriculture, Riverside, California.

Rawls, WJ, Pachepsky, YA, Ritchie, JC, Sobecki, TM and Bloodworth, H. (2003). Effect of organic carbon on soil water retention. Geoderma. 116, 61 – 76.

Read, ZJ, Murphy, B and Greene, RSB. (2012). Soil carbon sequestration potential of revegetated scalded soils following water ponding. Proceedings of the 5th Joint Australian and New Zealand Soil Science Conference, Hobart, December 2012. Australian Society of Soil Science Incorporated.

Rengasamy, P. and Churchman, GJ. (1999). Cation exchange capacity, exchangeable cations and sodicity. In Peverrill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - an interpretation manual. CSIRO Publishing Collingwood Australia.

Rice, Charles W. (2002). Organic matter and nutrient dynamics. In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker, New York.

Ringrose-Voase, AJ, Geeves, GW, Merry, RH and Wood, JT. (1997). Soil indicators of changing land quality and capital value. CSIRO (Australia). Land and Water Technical Report 17/97.

Rosewell, CJ. (1993). SOILOSS – A program to assist in the selection of management practice to reduce erosion. Technical Handbook No 11. 2nd Edition. Department of Conservation and Land Management , Sydney.

Rosewell, CJ and Loch, RJ. (2002). Estimation of the RUSLE soil erodibility factor. In Neil McKenzie, Kep Coughlan and Hamish Cresswell (eds). Soil Physical Measurement and Interpretation for Land Evaluation. CSIRO Publishing, Collingwood Australia.

Rovira, AD. (1963). Microbial inoculation of plants. I. Establishment of free-living nitrogen-fixing bacteria in the rhizosphere and their effects on maize, tomato and wheat. Plant and Soil 19, 304-14.

Russell, JS. (1960) Soil fertility changes in the long term experimental plots at Kybybolite, South Australia. I. Changes in pH, total nitrogen, organic carbon and bulk density. Australian Journal of Agricultural Research 11, 902 – 926.

Ryan, MH and Graham, JH. (2002). Is there a role for arbuscular mycorrhizal fungi in production agriculture? Plant and Soil 244, 255 – 239.

Ryan, MH, Norton, RM, Kirkegaard, JA, McCormick, KM, Knights, SE and Angus, JF. (2002). Increasing mycorrhizal colonisation does not improve growth and nutrition of wheat on Vertosols in south eastern Australia. Australian Journal of Agricultural Research 53, 1173 – 1181.

Ryan, MH and Angus, JF. (2003). Arbuscular mycorrhizae in wheat and field pea crops on low P soil: increased Zn uptake but no increase in P-uptake or yield. Plant and Soil. 250, 225 – 239.

Ryan, MH, van Herwaarden, AF, Angus, JF and Kirkegaard, JA. (2005). Colonisation by arbuscular mycorrhizal fungi is associated with reductions in biomass of wheat in low-P soil under field conditions. Plant and Soil 250, 275 – 286.

Ryan, MH, Kirkegaard, JA and Angus, JF. (2006). *Brassica* crops stimulate soil mineral N accumulation. Australian Journal of Soil Research 44, 367 – 377.

Saunders, WMH. (1965). Phosphate retention by New Zealand soils and its relationship to free sesquioxides. organic matter and other soil properties. New Zealand Journal of Agricultural Research. 8, 30 – 57.

Schefe, CR, Patti, AF, Clune,TS, and Jackson, WR. (2007). Soil amendments modify phosphate sorption in an acid soil: the importance of P source (KH2 PO4, TSP, DAP. Australian Journal of Soil Research 45, 246 – 54.

Schmidt, MW, Torn, MS, Abiven, S, Dittmar, T, Guggenberger, G, Janssens, IA, Kleber, M, Kogel-Knabner, I, Lehmann, J, Manning, DAC, Nannipieri, P, Rasse, DP, Weiner, S and Trumbore. (2011). Persistence of soil organic matter as an ecosystem property. Nature, 478, 49 – 56.

Schmidt, A, Smernik, RJ, McBeath, TM. (2012). Measuring organic carbon in Calcarosols: understanding the pitfalls and complications. Soil Research. 50, (5), 397 – 405.

Schulze, D. G., Nagel, L. L., van Scoyoc, G. E., Henderson, T. L., Baumgardner, M. F., and Stott, D. E. (1993). Significance of organic matter in determining soil color. In 'Soil color. Vol. SSSA Special Publication number 31.' (Eds J. M. Bigham and E. J. Ciolkosz.) pp. 71-90. (Soil Science Society of America: Madison, WI.).

Shaw, RJ. (1991). Soil salinity – electrical conductivity and chloride. In Peverrill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - An Interpretation Manual. CSIRO Publishing Collingwood Australia.

Shaykewich, CF and Zwarich, MA. (1968). Relationships between soil physical constants and soil physical components of some Manitoba soils. Canadian Journal of Soil Science. 48, 199 – 204.

Shepherd, TG, Saggar, S, Newman, RH, Ross, CW & Dando, JL. (2001). Tillage-induced changes to soil structure and organic carbon fractions in New Zealand soils. Australian Journal of Soil Research, 39, 465-489.

Sherrod, LA, Peterson, GA, Westfall, DG & Ahuja, LR. (2005). Soil Organic Carbon Pools After 12 Years in No-Till Dryland Agroecosystems. Soil Science Society of America Journal, 69, 1600-1608.

Simpson, RJ, Oberson, A, Culvenor, RA, Ryan, MH, Veneklaas, EJ, Lambers, H, Lynch, JP, Ryan, PR, Delhaize, E, Smith, FA, Harvey, PR and Richardson, AE. (2011). Strategies and agronomic interventions to improve the phosphorus-use efficiency of farming systems. (2011). Plant and Soil. 349, 89 – 120.

Singh, B, Rengel, Z and Bowden, JW. (2004). Canola residues decomposition: the effect of particle size on microbial respiration and cycling of sulphur in a sandy soil. SuperSoil 2004: 3rd Australian New Zealand Soils Conference, 5 – 9 December 2004, University of Sydney, Australia. Published on CDROM. Website www.regional.org.au/au/asssi/.

Singh, BK, Munro, DP, Potts, JM and Miillard, P. (2007). Influence of grass species and soil type on rhizosphere microbial community structure in grassland soils. Applied Soil Biology. 36, 147 - 55.

Six, J, Feller, C, Denef, K, Ogle, SM, de Moraes, JC, Albrecht, A. (2002). Soil organic matter, biota and aggregation in temperate and tropical soils – Effects of no-tillage. Agronomie 22 (2002) 755–775.

Six, J and Jastrow, JD. (2002). Organic matter turnover. . In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker, New York.

Skjemstad, JO, Clarke, P, Taylor, JA, Oades, JM and McClure, SG. (1996). The chemistry and nature of protected carbon in soil. Australian Journal of Soil Research 34, 251 -71.

Skjemstad, JO, Spouncer, LR and Beech, A . (2000). Carbon conversion factors for historical soil carbon data. National Carbon Accounting System Technical Report No. 15. Australian Greenhouse Office. Canberra ACT, Australia.

Skjemstad, JO, Swift, RS and McGowan, JA. (2006). Comparison of the particulate organic and permanganate oxidation methods for estimating labile soil organic carbon. Australian Journal of Soil Research 44, 255 – 63.

Slattery, WJ, Edwards, DG, Bell, LC, Coventry, DR and Helyar, KR. (1998). Soil acidification and the carbon cycle in a cropping soil of north eastern Victoria. Australian Journal of Soil Research. 36, 273 – 290.

Slattery WJ, Conyers, MK and Aitken, RL. (1999). Soil pH, Aluminium, manganese and lime requirement. In Peverrill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - an interpretation manual. CSIRO Publishing Collingwood Australia.

Slavich, PG and Petersson, GH (1993). Estimating the electrical conductivity of saturated paste extracts from 1:5 soil:water suspensions and texture. Australian Journal of Soil Research. 31, 73 – 81.

Soil Survey Staff. (2010). Keys to Soil Taxonomy. (Eleventh Edition). US Department of Agriculture. Natural Resources Conservation Service. US Government, Washington DC.

Spain, AV, Isbell, RF and Probert, ME. (1993). Soil organic matter. In Soils: An Australian Viewpoint. CSIRO Publications, Collingwood, Victoria, Australia.

Sparling, GP. (1992). Ratio of microbial biomass carbon to soil organic carbon as a sensitive indicator of changes in soil organic matter. Australian Journal of Soil Research, 30, 195-207.

Standards Association of Australia. (1993 – 2003) Methods of testing soils for engineering purposes. Australian Standards AS 1289 5.1.1. Soil compaction and density tests. Determination of minimum and maximum dry density of a cohesionless soil. The Standards Association of Australia, North Sydney.

Stephan, A, Meyer, Ah and Schmid, B. (2000). Plant diversity affects culturable soil bacteria in experimental grassland communities. Journal of Ecology. 88, 988 – 998.

Stevenson, IL. (1969). Biochemistry of Soil. In Firman E. Bear (ed). Chemistry of the Soil. American Chemical Society Monograph Series. Van Nostrand Reinhold Company, New York.

Stevenson FJ. (1982). Humus Chemistry – Genesis, Composition and Reactions. John Wiley and Sons, New York.

Stewart, C, Paustian, K, Conant, R, Plante, A & Six, J. (2007). Soil carbon saturation: concept, evidence and evaluation. Biogeochemistry, 86, 19-31.

Strong, WM and Mason, MG. (1999). Nitrogen. In Peverrill, KI, Sparrow, LA and Reuter, DJ (eds). Soil Analysis - an interpretation manual. CSIRO Publishing Collingwood Australia.

Swenson, RM, Cole, VC and Sieling, DH. (1949). Fixation of phosphorus by iron and aluminium and replacement by organic and inorganic ions. Soil Science 67, 1 – 22.

Tan, KH and Dowling, PS. (1984). Effect of organic matter on CEC due to permanent and variable charges in selected temperate region soils. Geoderma, 32, 89 – 101.

Tate, Robert L. (1987). Soil Organic Matter – Biological and Ecological Effects. John Wiley and Sons, New York.

Thomas, GW, Hazler, GR and Blevins, RL. (1996). The effects of organic matter and tillage on maximum compaction of soils using the proctor test. Soil Science 1 61, 502 – 508.

Thomas, GW. (2002). pH. In Rattan Lal (ed). Encyclopedia of Soil Science. Marcel Dekker, New York.

Thomasson, AJ and Carter, AD. (1989). Current and future uses of the UK soil water retention dataset. In "Indirect methods for estimating the hydraulic properties of unsaturated soils". Proceedings of an International Workshop. MT van Genuchten, FJ Leij and LJ Lund (eds). pp 355 – 358. USDA and Department of Soil and Environmental Science: Riverside.

Tisdall, JM and Oades, JM. (1980a). The effect of crop rotation on aggregation in a red-brown earth. Australian Journal of Soil Research 18, 423-433.

Tisdall, JM and Oades, JM. (1980b). The management of ryegrass to stabilise aggregates of a red-brown earth. Australian Journal of Soil Research 18, 415-422.

Tisdall, JM and Oades, JM. (1982). Organic matter and water stable aggregates in soil. Journal of Soil Science. 33, 141 – 63.

Toreu, BN, Thomas, FG and Gillman, GP. (1988). Phosphate characteristics of the soils of the north Queensland coastal region. Australian Journal of Soil Research. 26, 465 – 77.

Tranter, G, Minasny, B, McBratney, AB, Murphy, B, McKenzie, NJ, Grundy M & Brough, D. (2007) Building and testing conceptual and empirical models for predicting soil bulk density. Soil Use and Management 23, 437 - 43.

United States Department of Interior. (1960). Design of Small Dams. USDI Bureau of Reclamation. Government Printing Office. Washington DC.

Unkovich, MJ, Baldock, J and Peoples, MB. (2010). Prospects and problems of simple linear models for estimating symbiotic N2 fixation by crop and pasture legumes. Plant and Soil 329, 75 – 89.

Upjohn, Brett, Fenton, Greg and Conyers, Mark. (2005). Soil Acidity and Liming. Agfact AC19, 3rd Edition. NSW Department of Primary Industries. http://www.dpi.nsw.gov.au/__data/assets/pdf_file/0007/167209/soil-acidity-liming.pdf

Utomo, WH and Dexter, AR. (1981). Soil friability. Journal of Soil Science. 32, 203 – 213.

Valzano F, Murphy BW, Koen T. (2005). The impact of tillage on changes in soil carbon density with special emphasis on Australian conditions. Technical Report No 43, National carbon Accounting System, Australian Greenhouse Office, Canberra.

van Genuchten, MTh. (1980). A closed form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Science Society America Journal. 44, 892 -8.

Veimeyer, FJ and Hendrickson, AH. (1928). Soil moisture at permanent wilting point. Plant Physiology. 3(3): 355–357.

Veimeyer, FJ and Hendrickson, AH. (1931). The moisture equivalent as a measure of the field capacity of soils. Soil Science. 32, 181 – 194.

Vereecken, H., Maes, J., Feyen, J., Darius, P. (1989). Estimating the soil moisture retention characteristic from texture, bulk density and carbon content, Soil Science, 148, 389–403.

Walkley, A and Black, IA. (1934). An examination of the Degtjareff method for determining soil organic matter and a proposed modification of the chrmic acid titration method. Soil Science 37, 29 – 38.

Walker, TW and Adams, AFR (1958). Studies in soil organic matter.I. Influence of parent material on accumulation of C, N, S and organic Pin grassland soils. Soil Science 85, 307 – 318.

Watt, M., Kirkegaard, JA, Passioura, JB. (2006). Rhizosphere biology and crop productivity—a review Australian Journal of Soil Research 44, 299 – 317.

West, T and Six, J. (2007). Considering the influence of sequestration duration and carbon saturation on estimates of soil carbon capacity. Climatic Change, 80, 25-41.

Wiklander, Lambert. (1969). Cation and anion exchange phenomena. . In Firman E. Bear (ed). Chemistry of the Soil. American Chemical Society Monograph Series. Van Nostrand Reinhold Company, New York.

Wilhelm, S. (1973). Principles of biological control of soil-borne plant diseases. *Soil Biology and Biochemistry, 5*(6), 729-737. http://dx.doi.org/10.1016/0038-0717(73)90017-5

Williams, CH. (1980). Soil acidification under clover pasture. Australian Journal of Experimental Agriculture and Animal Husbandry 20, 561 -567.

Williams, CH and Donald, CM. (1957). Changes in organic matter and pH in a podzolic soil as influenced by subterranean clover and superphosphate. Australian Journal of Agricultural Research 8, 179-189.

Williams, J, Ross, P and Bristow, K. (1992). Prediction of the Campbell water retention function from texture, structure and organic matter. In M. Th. Van Genuchten and FJ Leij (eds). Indirect methods for estimating hydraulic properties of unsaturated soils. Proceedings of International Workshop on Indirect Methods for Estimating the Hydraulic Properties of Unsaturated Soils. Riverside, California, October 11 – 13, 1989. US salinity Laboratory, Agricultural research Service, US Department of Agriculture, Riverside, California.

Wilson, BR, Koen, TB, Barnes, P, Ghosh, S and King, D. (2011). Soil carbon and related properties along a soil type and land use intensity gradient, New South Wales, Australia. Soil Use and Management. 27, 437 – 447.

Wong, VNL, Murphy, BW, Koen, TB, Greene, RSB and Dalal, RC. (2008). Soil organic carbon stocks in saline and sodic landscapes. Australian Journal of Soil Research 378 – 389.

Wösten, J.H.M., Lilly, A., Nemes, A., Le Bas, C. (1999). Development and use of a database of hydraulic properties of European soils. Geoderma. 90, 169-185.

Yan,F, McBratney, AB and Copeland, L. (2000). Functional substrate biodiversity of cultivated and uncultivated A horizons of Vertisols in NW New South Wales. Geoderma. 96, 321 – 343.

Yao, Q, Li, X, Feng, G and Christie, P. (2001). Mobilization of sparingly soluble inorganic phosphates by the external mycelium of an abuscular mycorrhizal fungus. Plant and Soil. 279 – 285.

Yeates, C and Gillings, M. (1998). Rapid purification of DNA from soil for molecular biodiversity analysis. Letters in Applied Microbiology 27, 49 – 53.

Zelles, L., Bai, Q.Y., Beck, T., Beese, F, (1992). Signature fatty acids in phospholipids and lipopolysaccharides as indicators of microbial biomass and community structure in agricultural soils. Soil Biology and Biochemistry. 24, 317–323.

Zelles, L and Bai, QY. (1993). Fractionation of fatty acids derived from soil lipids by solid phase extraction and their quantitative analysis by GC-MS. Soil Biology and Biochemistry 25, 495 – 507.

Zhu YG, Smith SE, Smith FA. (2001). Zinc-phosphorus interactions in two cultivars of spring wheat (Triticum aestivum L.) Differing in P uptake efficiency. Annals of Botany. 88, 941-945.

Appendix 1

Table Summarising Effects of Soil Organic Matter and its Fractions on Soil Properties.

Caution-

A table such as this can be an over simplification. While the table can give a general indication of the effects of soil organic matter on soil properties, it is necessary to consider the soils, landscape, climate and crops/pastures being grown to fully evaluate the effects of soil organic matter on soil properties.

Potential effect of soil organic matter parameters on soil functional properties _Part 1

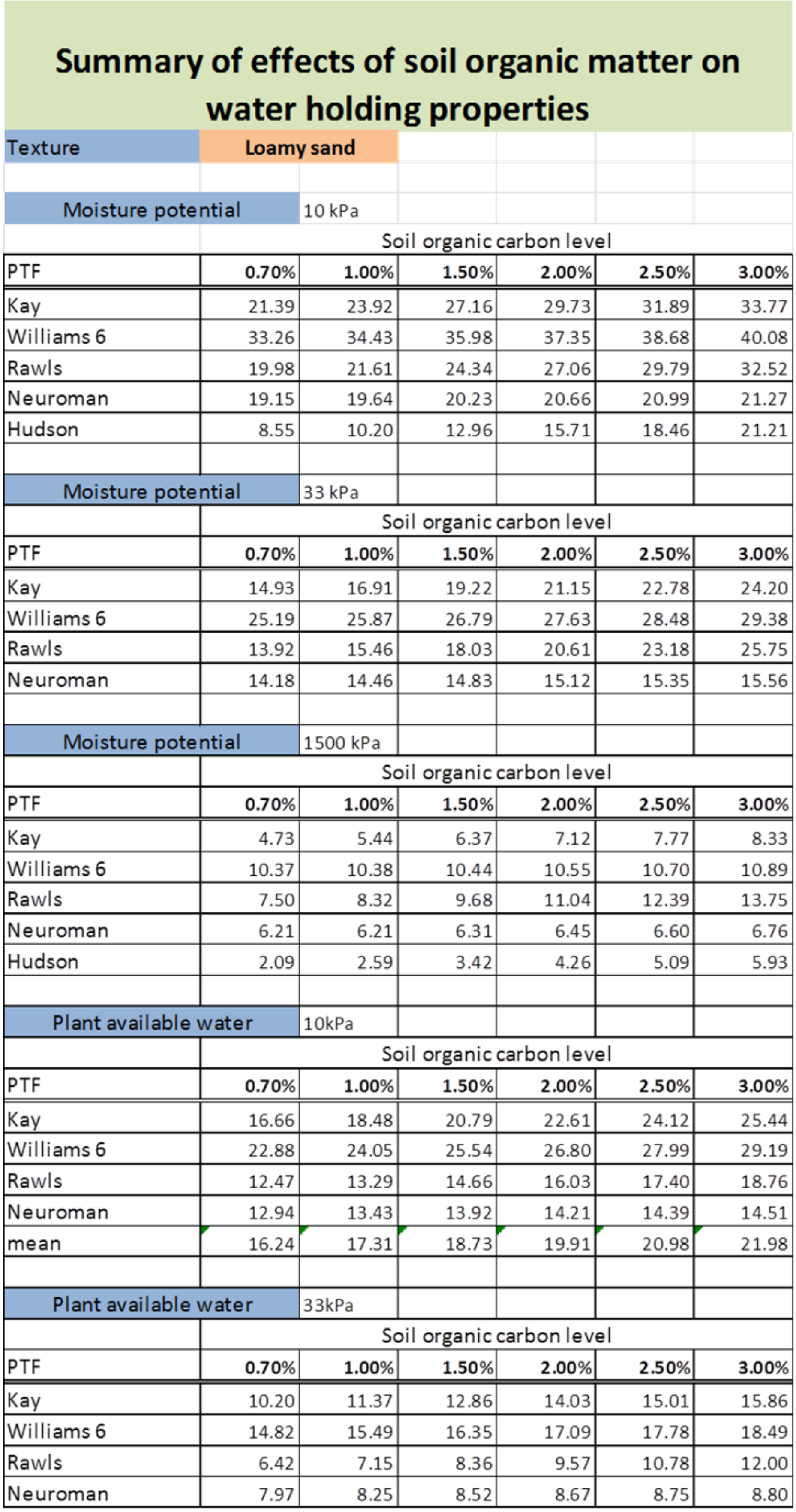
Soil functional property	**Soil organic carbon parameter**				
	Total soil organic carbon (TOC)	**C/N ratio of TOC**	**Particulate soil organic carbon (POC)**	**Humus soil organic carbon (HOC)**	**Resistant soil organic carbon (ROC)**
Plant available water					
Field capacity (10 kPa)	Minor for loams and clays Minor to moderate for sandy loams and sands	Not applicable	Minor for loams and clays Minor to moderate for sandy loams and sands	Minor	Minor for loams and clays Minor to moderate for sandy loams and sands r
Wilting point **1500 (kPa)**	Minor	Not applicable	None	Minor	Minor
Plant available water **(10 kPa – 1500 kPa)**	Minor for loams and clays Minor to moderate for sandy loams and sands	Not applicable	Minor for loams and clays Minor to moderate for sandy loams and sands	Minor	Minor for loams and clays Minor to moderate for sandy loams and sands
Note – Plant available water is often considered part of soil structure but because it is such a critical function of soils it is considered separately.					
Soil structure – Aggregate stability, texture, clay mineralogy (swelling potential), ESP and cations					
Infiltration – crusting and surface sealing	Moderate to high for loam and sandy soils, minor for clayey soils	Not applicable	Moderate to high for loam and sandy soils, minor for clayey soils	Minor to moderate for all soils	None
Tilth - response to tillage – friability – draught – seedbed conditions	Moderate to high for loam and sandy soils, minor for clayey soils	Not applicable	Moderate to high for loam and sandy soils, minor for clayey soils	Minor to moderate for all soils	None
Compaction characteristics – engineering characteristics - Atterberg Limits	Moderate to high for loam and sandy soils, minor for clayey soils	Not applicable	Moderate to high for loam and sandy soils, minor for clayey soils	Minor to moderate for all soils	None

Potential effect of soil organic matter parameters on soil functional properties _Part 2

Soil functional property	Soil organic carbon parameter				
	Total soil organic carbon (TOC)	**C/N ratio of TOC**	**Particulate soil organic carbon (POC)**	**Humus soil organic carbon (HOC)**	**Resistant soil organic carbon (ROC)**
Nutrient Cycling - Mineralisation of nutrients					
N cycling	High for all soils	High for all soils	High for all soils	High for all soils	Minor for all soils
P cycling	High for all soils	High for all soils	High for all soils	High for all soils	Minor for all soils
S cycling	High for all soils	High for all soils	High for all soils	High for all soils	Minor for all soils
K cycling	High for all soils	High for all soils	High for all soils	High for all soils	Minor for all soils
Trace elements	High for all soils	High for all soils	High for all soils	High for all soils	Minor for all soils
Source of electric charge in soils – chemical activity					
Cation exchange capacity	Moderate to high for loamy and sandy soils	Not applicable	None to low for all soils	Moderate to high for loamy and sandy soils	None to low for all soils Some biochars can have moderate effect
Buffering capacity against acidification	Moderate to high for loamy and sandy soils	Not applicable	None to low for all soils	Moderate to high for loamy and sandy soils	None to low for all soils Some biochars can have moderate effect
Complexing of anions (P)	Moderate to high for loamy and sandy soils	Not applicable	None to low for all soils	Moderate to high for loamy and sandy soils	None to low for all soils Some biochars can have moderate effect
Complexing of metal ions (Al, heavy metals)	Moderate to high for loamy and sandy soils	Not applicable	None to low for all soils	Moderate to high for loamy and sandy soils	None to low for all soils Some biochars can have moderate effect
Biological Activity – microbial activity					
Food source	Moderate to high for all soils	Moderate to high for all soils	Moderate to high for all soils	Moderate to high for all soils	Low for all soils
Capacity to change functional groups	Moderate to high for all soils	Moderate to high for all soils	Moderate to high for all soils	Moderate to high for all soils	Low for all soils

Appendix 2

Results of pedotransfer calculations for 4 textures.



Summary of effects of soil organic matter on water holding properties

Texture	Loamy sand

Moisture potential	10 kPa				

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	21.39	23.92	27.16	29.73	31.89	33.77
Williams 6	33.26	34.43	35.98	37.35	38.68	40.08
Rawls	19.98	21.61	24.34	27.06	29.79	32.52
Neuroman	19.15	19.64	20.23	20.66	20.99	21.27
Hudson	8.55	10.20	12.96	15.71	18.46	21.21

Moisture potential	33 kPa				

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	14.93	16.91	19.22	21.15	22.78	24.20
Williams 6	25.19	25.87	26.79	27.63	28.48	29.38
Rawls	13.92	15.46	18.03	20.61	23.18	25.75
Neuroman	14.18	14.46	14.83	15.12	15.35	15.56

Moisture potential	1500 kPa				

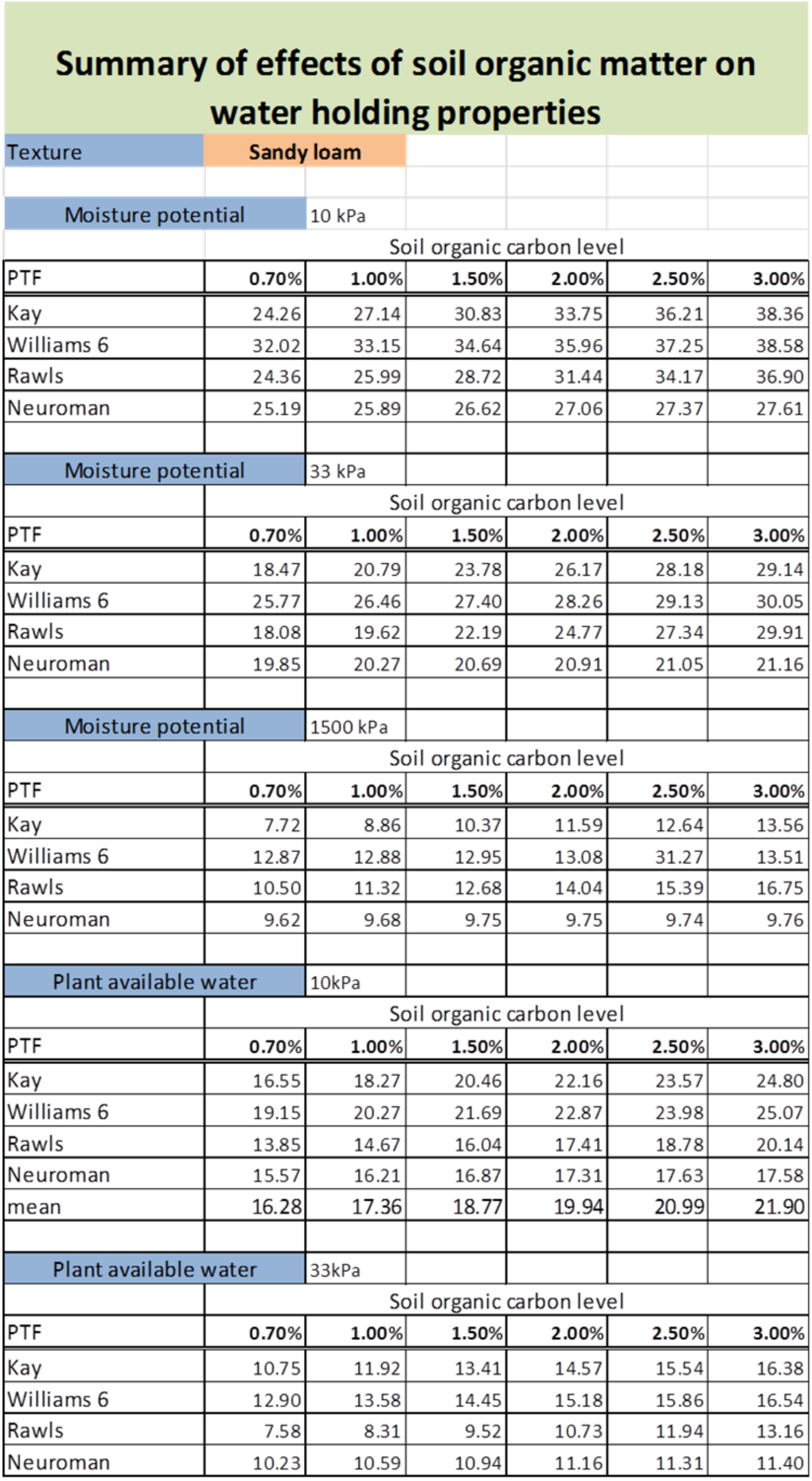
	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	4.73	5.44	6.37	7.12	7.77	8.33
Williams 6	10.37	10.38	10.44	10.55	10.70	10.89
Rawls	7.50	8.32	9.68	11.04	12.39	13.75
Neuroman	6.21	6.21	6.31	6.45	6.60	6.76
Hudson	2.09	2.59	3.42	4.26	5.09	5.93

Plant available water	10kPa				

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	16.66	18.48	20.79	22.61	24.12	25.44
Williams 6	22.88	24.05	25.54	26.80	27.99	29.19
Rawls	12.47	13.29	14.66	16.03	17.40	18.76
Neuroman	12.94	13.43	13.92	14.21	14.39	14.51
mean	16.24	17.31	18.73	19.91	20.98	21.98

Plant available water	33kPa				

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	10.20	11.37	12.86	14.03	15.01	15.86
Williams 6	14.82	15.49	16.35	17.09	17.78	18.49
Rawls	6.42	7.15	8.36	9.57	10.78	12.00
Neuroman	7.97	8.25	8.52	8.67	8.75	8.80



Summary of effects of soil organic matter on water holding properties

Texture	Sandy loam

Moisture potential	10 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	24.26	27.14	30.83	33.75	36.21	38.36
Williams 6	32.02	33.15	34.64	35.96	37.25	38.58
Rawls	24.36	25.99	28.72	31.44	34.17	36.90
Neuroman	25.19	25.89	26.62	27.06	27.37	27.61

Moisture potential	33 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	18.47	20.79	23.78	26.17	28.18	29.14
Williams 6	25.77	26.46	27.40	28.26	29.13	30.05
Rawls	18.08	19.62	22.19	24.77	27.34	29.91
Neuroman	19.85	20.27	20.69	20.91	21.05	21.16

Moisture potential	1500 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	7.72	8.86	10.37	11.59	12.64	13.56
Williams 6	12.87	12.88	12.95	13.08	31.27	13.51
Rawls	10.50	11.32	12.68	14.04	15.39	16.75
Neuroman	9.62	9.68	9.75	9.75	9.74	9.76

Plant available water	10kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	16.55	18.27	20.46	22.16	23.57	24.80
Williams 6	19.15	20.27	21.69	22.87	23.98	25.07
Rawls	13.85	14.67	16.04	17.41	18.78	20.14
Neuroman	15.57	16.21	16.87	17.31	17.63	17.58
mean	16.28	17.36	18.77	19.94	20.99	21.90

Plant available water	33kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	10.75	11.92	13.41	14.57	15.54	16.38
Williams 6	12.90	13.58	14.45	15.18	15.86	16.54
Rawls	7.58	8.31	9.52	10.73	11.94	13.16
Neuroman	10.23	10.59	10.94	11.16	11.31	11.40

Summary of effects of soil organic matter on water holding properties

Texture	light medium clay

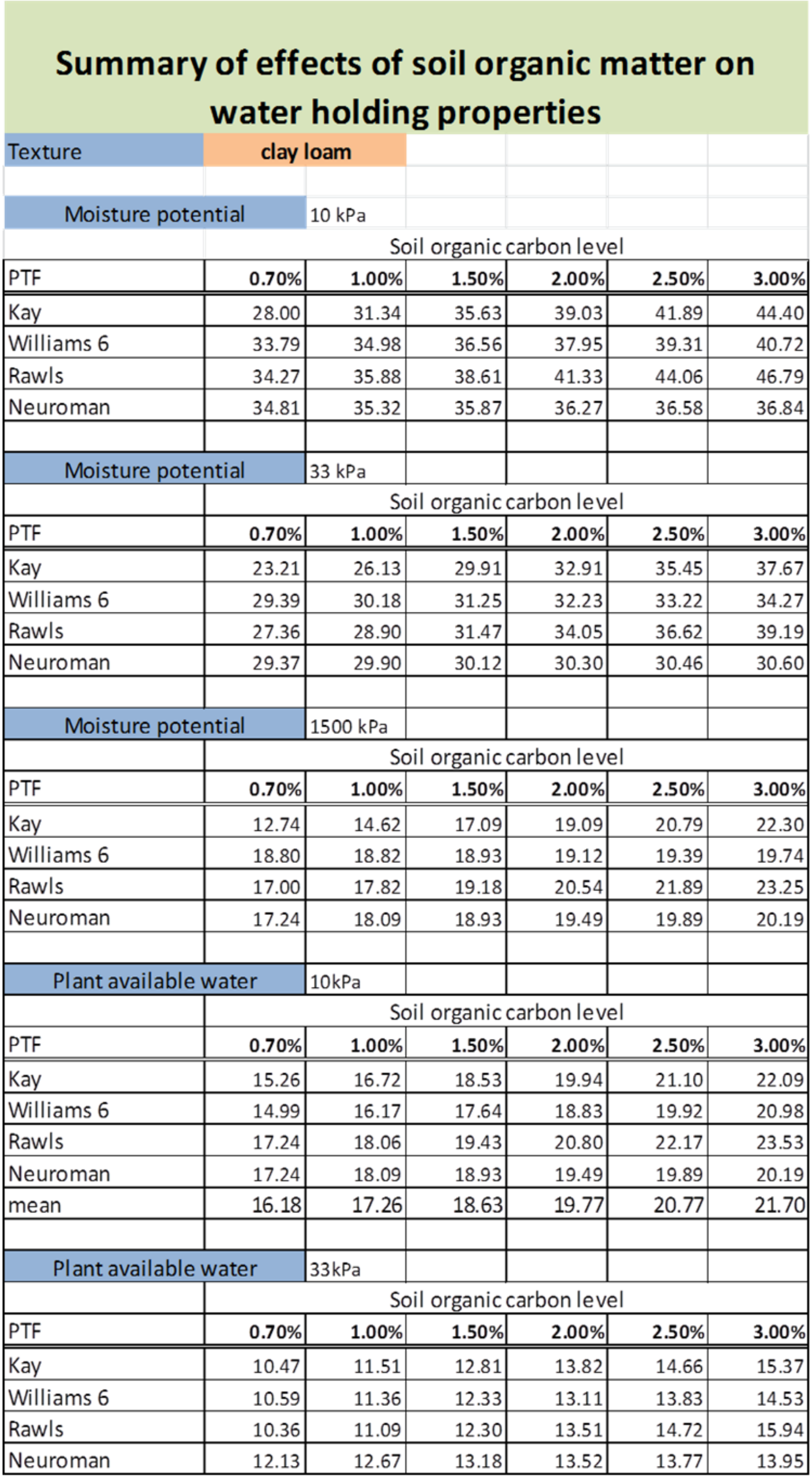
Moisture potential	10 kPa				
	Soil organic carbon level				
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	31.04	34.75	39.52	43.30	46.48	49.26
Williams 6	40.26	41.68	43.56	45.22	46.84	48.52
Rawls	42.52	44.15	46.88	49.60	52.33	55.06
Neuroman	38.76	39.32	39.91	40.32	40.64	40.91

Moisture potential	33 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	27.61	31.08	35.58	39.15	42.18	44.82
Williams 6	36.27	37.25	38.57	39.78	41.00	42.30
Rawls	36.80	38.34	40.91	43.49	46.06	48.63
Neuroman	34.93	35.25	35.56	35.76	35.93	36.07

Moisture potential	1500 kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	18.98	21.77	25.43	28.39	30.91	33.14
Williams 6	25.98	26.00	26.15	26.42	26.79	27.28
Rawls	26.50	27.32	28.68	30.04	31.39	32.75
Neuroman	13.82	14.53	15.38	15.98	16.44	16.81

Plant available water	10kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	12.06	12.98	14.09	14.91	15.57	16.13
Williams 6	14.28	15.68	17.41	18.80	20.05	21.24
Rawls	16.01	16.83	18.20	19.57	20.94	22.30
Neuroman	13.82	14.53	15.38	15.98	16.44	16.81
mean	14.04	15.01	16.27	17.32	18.25	19.12

Plant available water	33kPa					
	Soil organic carbon level					
PTF	**0.70%**	**1.00%**	**1.50%**	**2.00%**	**2.50%**	**3.00%**
Kay	8.63	9.32	10.15	10.77	11.27	11.68
Williams 6	10.29	11.24	12.42	13.36	14.21	15.02
Rawls	10.30	11.25	12.24	13.45	14.66	15.88
Neuroman	9.99	10.46	11.03	11.42	11.73	11.97



Summary of effects of soil organic matter on water holding properties

Texture	clay loam

Moisture potential	10 kPa				
	Soil organic carbon level					

PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	28.00	31.34	35.63	39.03	41.89	44.40
Williams 6	33.79	34.98	36.56	37.95	39.31	40.72
Rawls	34.27	35.88	38.61	41.33	44.06	46.79
Neuroman	34.81	35.32	35.87	36.27	36.58	36.84

Moisture potential	33 kPa

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	23.21	26.13	29.91	32.91	35.45	37.67
Williams 6	29.39	30.18	31.25	32.23	33.22	34.27
Rawls	27.36	28.90	31.47	34.05	36.62	39.19
Neuroman	29.37	29.90	30.12	30.30	30.46	30.60

Moisture potential	1500 kPa

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	12.74	14.62	17.09	19.09	20.79	22.30
Williams 6	18.80	18.82	18.93	19.12	19.39	19.74
Rawls	17.00	17.82	19.18	20.54	21.89	23.25
Neuroman	17.24	18.09	18.93	19.49	19.89	20.19

Plant available water	10kPa

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	15.26	16.72	18.53	19.94	21.10	22.09
Williams 6	14.99	16.17	17.64	18.83	19.92	20.98
Rawls	17.24	18.06	19.43	20.80	22.17	23.53
Neuroman	17.24	18.09	18.93	19.49	19.89	20.19
mean	16.18	17.26	18.63	19.77	20.77	21.70

Plant available water	33kPa

	Soil organic carbon level					
PTF	0.70%	1.00%	1.50%	2.00%	2.50%	3.00%
Kay	10.47	11.51	12.81	13.82	14.66	15.37
Williams 6	10.59	11.36	12.33	13.11	13.83	14.53
Rawls	10.36	11.09	12.30	13.51	14.72	15.94
Neuroman	12.13	12.67	13.18	13.52	13.77	13.95