Two case studies exploring opportunities and constraints for soil organic carbon sequestration following land use change

By

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Declaration of originality

This thesis contains no material which has been accepted for the award of any other degree or diploma in any university. To the best of the author's knowledge the work contains no material previously published or written by another person, except where due reference is made in the text.

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Abstract

The flux of carbon (C) between the terrestrial biosphere and the atmosphere occurs naturally as part of the carbon cycle (CC). However, anthropogenic activity can disrupt the natural equilibrium of the CC. Although fossil fuel burning is the largest cause of human induced C emissions to the atmosphere, land use and land use change (LULUC) also results in emissions. Of particular concern is soil disturbance and erosion caused by LULUC activities including deforestation, tillage and overgrazing. These activities can lead to a net loss of C from the terrestrial biosphere leading to increased levels of greenhouse gas in the atmosphere. This outcome is a major factor widely understood to adversely affect the rate of global warming, climate change and extreme weather events.

There has been a great deal of research undertaken to show that LULUC activities can also lead to the sequestration and long term storage of C in soil. LULUC practices that result in increases in soil organic matter (SOM), humification and aggregation improve soil quality, water quality, enhance food security and increase biodiversity. The activities also result in the sequestration of atmospheric CO_2 as soil organic carbon (SOC) and therefore may contribute to the reduction or stabilisation of atmospheric CO_2 levels and mitigate the adverse effects of climate change. This thesis describes the effects of two different LULUC activities on the C sequestration potential of soil.

In this research two case studies have been used to explore three principal aims. Firstly to identify the C sequestration potential of agricultural soil following LUC; then to determine whether there is a steady rate of SOC sequestration over time following LUC, and finally to identify biogeochemical factors that contribute to the C sequestration potential of soil after LUC.

The first case study was undertaken on the Southern Tablelands of New South Wales (NSW), a temperate region with approximately 670 mm mean average rainfall (MAR). The study involved comparing the soil C stock (SCS) to 30 cm depth, in agricultural land (AL) principally used for grazing, with the SCS from 20 biodiverse environmental plantings (BEP) established between one and 19 years prior to sampling. BEPs, otherwise referred to as mixed species environmental plantings (MSEP), windbreaks or shelterbelts, can be established using either a direct seeding method or from tube-stock. The sites included in this study were all directly seeded in either linear or block configurations using a mix of native eucalypt and acacia trees and understory shrubs. At each site three paired 10 m transects were established on a tree line (TL), the inter-row (IR) and in the adjacent AL. The AGB was determined for each TL transect. Three soil cores to 30 cm depth were taken from each transect and separated into 0-5, 5-10, 10-20 and 20-30 cm increments. The soil samples were analysed for SOC, total nitrogen (TN), total phosphorus (TP), and by using mid infrared spectroscopy (MIRS) analysis soil fraction changes

were also determined. The measured AGB and SCS results were compared with FullCAM modelled data. The overall aim of this case study was to determine whether BEPs sown into AL led to SOC sequestration, and whether the rate of increase could be predicted based on tree age and biomass.

The results of the BEP case study show that the average soil CS in the AL based on an equivalent soil mass (ESM) was 41.1 Mg C ha^{-1} (41.1 tonnes C ha^{-1}). BEPs aged 11-15 years had the highest SCS of 49.3 Mg C ha^{-1}. The oldest BEP sites aged between 16 and 19 years had a lower SCS of 45.7 Mg C ha^{-1}. Overall, the average SCS rate of increase was 0.5 Mg C ha^{-1} yr^{-1}. However, the results show significant variation between sites and no trend suggesting there is a temporal increase in SCS. The average AGB for all sites was found to be 31.4 Mg C ha^{-1} with the range between 0.2 Mg C ha^{-1} in a two year old site, and 97.0 Mg C ha^{-1} in a site aged 16 years. The average annual rate of AGB change was found to be 2.4 Mg C ha^{-1} yr^{-1}. The results show there is no relationship between AGB and SCS. Therefore the prediction of changes to SCS based on AGB increase alone is not possible. The results also show that site factors and nutrient availability have an influence on the SOC sequestration potential of the BEPs and the sustainability of the SCS.

The second case study was undertaken in the semi-arid rangelands of the Central West of NSW. The MAR of the region is approximately 440 mm. The study examined whether waterponding, an effective environmental restoration activity used to rehabilitate scalded soil, would result in increases in SCS. The research involved comparing the SCS of three scalded sites with 12 waterpond sites established between 1 and 27 years prior to sampling. Nine soil cores to 30 cm depth were taken at each scald site. Three waterponds were sampled at each of the 12 waterpond sites. Nine soil cores to 30 cm depth were taken from three positions (wall, mid and top) within in each waterpond. Each soil core was separated into 0-5, 5-10, 10-20 and 20-30 cm increments. Every sample was analysed for pH, electrical conductivity (EC), SOC, TN, and total C. The results show that waterponds have a significantly lower EC than scalded soil. This change occurs because soluble salts are leached from the saline scald soil profile after waterponding. The results also show that the scalded soil has an average SCS to 30cm depth of 18.7 Mg C ha^{-1}. Waterponds aged five years had a SCS of 26.1 Mg C ha^{-1}. This represents a rate of increase of 1.5 Mg C $ha^{-1}yr^{-1}$. The 10 year old and 25-27 year old waterponds have a slightly lower SCS of 25.3 and 24.9 Mg C ha^{-1}. The probable reason for the apparent decline in SCS in the older water ponds is associated with the lower bulk density found in these older waterponds and the increased presence of SOM.

A number of edaphic factors were found to influence the potential of the scalds to sequester SOC. To make such assessments within research resource constraints three different subsets of the soil samples were compiled. Initially, a sub-set of 144 samples including one scald, and waterponds aged 1 year, 5 years and 25 years since establishment were analysed for aggregate

stability, TP, total sulfur (TS), and available P (AP). The results from this subset were used to identify temporal changes to soil stability and to ascertain whether these soil nutrient concentrations were affected by waterponding or influenced the SOC sequestration potential of the soils. The aggregate stability results show there is a temporal change in the dispersability of the soils following waterponding. Scalded soils are stable because of the high concentration of soluble salts. After the salts are leached from the profile the soils become more susceptible to dispersion. The results also indicate that total N, P and S is not limiting, but AP is limiting ongoing sequestration of C in the waterponds.

A smaller subset of samples made up of one scald core and a paired 27 year old waterpond core were analysed for cation exchange capacity, x-ray diffraction and x-ray fluorescence to identify any mineralogical differences that may be attributable to waterponding. Calcite was found to be leached from the scald profile, hematite increased due to iron oxide formation, gypsum, kaolinite, illite and vermiculite was higher in the waterponded soil than the scalded soil whereas smectite was lower in the waterponds. These differences are most likely due to weathering processes and aeolian deposition.

Across the scalded soil surface there are occasional hummocks of vegetation. Cesium-137 analysis was undertaken to determine whether the hummocks were recent deposits or relics of the original soil profile. For this analysis composite samples were made up from the three scald sites, two hummock sites and waterponds aged between 25 and 27 years since establishment. The results from this small study indicate the hummocks are most likely to be recent deposits of aeolian material.

The experimental results from this project show that LULUC can lead to an increase in SCS. The higher SCS found in the BEPs may not be sustainable because the oldest sites show a lower SCS than the younger sites. The results suggest reasons for the decline are most likely due to site factors and lower levels of nutrient availability. The SCS sequestered in the scalded soil rehabilitated by waterponding however, is likely to have been sequestered permanently so long as the rehabilitated landscape is managed sustainably.

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List of acronyms and abbreviations

AGB	above ground biomass
AL	agricultural land
ANOVA	analysis of variance
AP	available phosphorus
ASWAT	aggregate stability in water
BD	Bulk density (Mg m^{-3})
BEP	biodiverse environmental planting
C	carbon
CC	carbon cycle
CEC	cation exchange capacity
CFI	carbon farming initiative
CO_2	carbon dioxide
CO_2-e	carbon dioxide equivalent
C:Avail P	carbon (SOC) to available phosphorus ratio
C:N:P:S	the ratio of SOC to TN, TP and TS in soil, expressed as 1000:80:20:14
C:N	carbon (SOC) to nitrogen (total nitrogen) ratio
C:P	carbon (SOC) to phosphorus (total phosphorus) ratio
C:S	carbon (SOC) to sulphur (total sulphur) ratio
CS	carbon stock
CT	conventional tillage
DCHTC	Dumas catalysed high temperature combustion
EC	electrical conductivity
ECEC	effective cation exchange capacity
ERF	emissions reduction fund
ESM	equivalent soil mass
ESP	exchangeable sodium percentage
FullCAM	full carbon accounting model
GA	Greening Australia
GHG	greenhouse gas
HOC	humic organic carbon
HS	humic substances
ICP-AES	Inductively coupled plasma- atomic emission spectrometry
IR	inter-row
LUC	land use change
LULUCF	land use, land use change and forestry
Mg ha^{-1}	mega grams (tonnes) per hectare
MIRS	mid infrared spectroscopy
MSEP	mixed species environmental planting
NCA	national carbon accounting
NPP	net primary productivity
NT	no-till
OM	organic matter
POC	particulate organic carbon

POM	particulate organic matter
PS	particle size
PSA	particle size analysis
ROC	resistant organic carbon
SAT	state and transition
SCS	soil carbon stock (Mg C ha^{-1})
SEM	standard error of the mean
SOC	soil organic carbon
SOM	soil organic matter
TL	tree line
TN	total nitrogen
TP	total phosphorus
TS	total sulfur
TSC	total soil carbon
UNFCCC	United Nations Framework Convention on Climate Change
XRD	X-ray diffraction

Glossary and Terms

Aggregate stability in water	A test to determine the stability of soil aggregates used as an indicator of soil resilience (Field *et al*. 1997).
A-horizon	The top, mineral layer of the soil profile.
Alluvial material	Particles of gravel, sand, silt and clay deposited by water movement.
Available phosphorus	Inorganic orthophosphate ion ($H_2PO_4^-$, HPO_4^{2-} and PO_4^{3-}) in the soil solution.
Biodiverse environmental planting	A form of agroforestry where seed from native trees and understory species have been directly drilled into agricultural land. Also referred to in the literature as windbreak, mixed species environmental planting (MSEP), environmental planting or shelterbelt (Paul *et al*. 2013).
Carbon to nitrogen ratio	The mass of soil C to the mass of soil TN.
Carbon to phosphorus ratio	The mass of soil C to the mass of soil TP.
Carbon to sulphur ratio	The mass of soil C to the mass of soil TS.
Chronosequence	Study sites of similar soil and vegetation attributes but of different ages.
Carbon dioxide equivalent (CO_2-e)	Measure used to report emissions from greenhouse gases based on their global warming potential (IPCC, 2007).
Equivalent soil mass	A measure of SCS that takes into account changes to soil bulk density, volume and SOC (Ellert and Bettany, 1995).
Hard scald	Compacted clay or cemented sand forming a crusty scald on the soil surface (Beadle, 1948a).
Heterogeneity	The random nature, patchiness or variability in spatial and temporal patterns of specific soil attributes.
Heterotroph	Animals, fungi and most bacteria that feed on OM produced by other organisms.
Native soil	Natural, undisturbed soil representative of pre-disturbance soil conditions.
Soft scald	Soil surface scalds with a spongy surface and fine cracks (Beadle, 1948a).
Soil organic carbon	The C component of soil organic matter.
Soil carbon sequestration	The capture and storage of atmospheric CO_2 in soil.
Soil carbon stock	The quantity of C in soil expressed as Mg C ha^{-1}.
Soil organic matter	The organic component of soil that is made up of living and dead plant and animal residues at various stages of decomposition.
Soil bulk density	The weight of soil in given volume recorded as Mg m^{-3}.
Soluble salt	Soluble salt in soil includes cations Na^+, K^+, Ca^{2+} and Mg^{2+} and anions Cl^-, SO_4^{2-}, CO_3^{2-}.
Total nitrogen	a measure of the total N concentration and includes all inorganic and organic pools of soil N.
Total phosphorus	The quantity of both organic and inorganic P in soil.

Total sulphur	The quantity of both organic and inorganic forms of S in soil.
Total soil carbon	The concentration of both organic and inorganic C in soil
Variability	Changes to specific soil attributes over space and/or time.
Waterponding	A mechanical method used to rehabilitate scalded soil (Thompson, 2008).

Chapter 1: Introduction

1.1 Background

There is compelling evidence to prove that the climate of the earth is being changed as a consequence of increased anthropogenic induced atmospheric greenhouse gas (GHG) emissions, and most notably by carbon dioxide (CO_2) emissions. Fossil fuel burning, land clearing and agriculture have been proven to be the dominant causes of CO_2 emission increase during the post industrialisation period. Since the mid 1700's atmospheric CO_2 concentrations have increased by approximately 27 %, from a pre-industrial concentration of 280 ppm to 385 ppm in 2010. During the 1990's 22 to 25 Gt of CO_2 were emitted annually, but this increased to between 25.3 and 27.5 Gt CO_2 per year during the period 2000-2005 (IPCC, 2007) and by 2011, global CO_2 emissions were estimated to be 34 Gt CO_2 yr^{-1} (Oliver *et al*. 2012). The increase in atmospheric CO_2 concentration has contributed to a rise in the global mean surface air temperature, higher evaporation rates, and changes in regional rainfall patterns which may contribute to the occurrence and frequency of extreme weather events (IPCC, 2007). The consequences of GHG driven climate change include sea level rise, changes to temperature and rainfall patterns and greater frequency and extremes in adverse weather events such as storms, droughts and floods. Extreme climatic events are predicted to worsen unless action is taken to slow GHG emissions (Hartman *et al*. 2013). Such events can lead to the loss of prime agricultural land due to soil erosion. The erosion is initiated by loss of ground cover resulting from land clearing, overgrazing and drought and in extreme cases can result in desertification.

When compared with fossil fuel burning, land use, land use change and forestry (LULUCF) is a secondary contributor of CO_2 emissions. Nevertheless LULUCF is an important source contributing approximately 25 % of total CO_2 global emissions. For example, globally, during the 1990's land use change (LUC) was estimated to contribute 5.9 Gt of CO_2 into the atmosphere annually (IPCC, 2007), and in the period between 2002 and 2011 LUC average CO_2 emissions were 0.9 Gt C yr^{-1} (Hartman *et al.* 2013).

By comparison, Australia's total GHG annual emissions in the year to December 2013, were estimated to be 0.55 Gt CO_2-e (DOE, 2015). These estimates include emissions by the energy, industrial, agricultural and waste sectors, but excludes emissions caused by the LULUCF sector. The estimated emissions represent a 25.5% increase when compared with the 1990 emission estimate of 0.43 Gt CO_2-e. Australian emissions from the LULUCF sector 2013 were estimated to be -0.00043 Gt of CO_2 equivalent (CO_2-e) with the negative trend attributed to a decrease in emissions from the *forest land converted to cropland and grassland* sector (DOE, 2015). CO_2-e refers to the global warming potential (GWP) of greenhouse gases, including methane (CH_4) and nitrous oxide (N_2O), hydrochlorofluorocarbons (HCFC), perfluorocarbons (PFC) and sulfur hexafluoride (SF6) (IPCC, 2007). For example, CO_2 has a reference GWP of 1 for 20, 100 and

500 years, whereas the GWP of CH_4 is 56 over 20 years, and for N_2O is 280 over 20 years (UNFCCC, 2015).

In 2013 afforestation and reforestation undertaken in Australia were estimated to remove 0.03 Gt CO_2-e (DOE, 2015a). Emissions from the Australian agriculture sector in the June quarter of 2015 were estimated to be 0.081 Gt CO_2-e. This represents a 0.7% lower emission rate than the 1990 baseline agricultural sector emission rate of 0.089 Gt CO_2-e (DOE, 2015b).

In addition to emitting CO_2 into the atmosphere, LULUCF can contribute to soil degradation which is manifested as wind and water erosion, soil compaction and desertification (Chappell *et al*. 2015). In Australia key LULUCF activities that contribute to CO_2 emissions and land degradation include development of marginal rangelands for agriculture, clearing of forests for farmland, and the conversion of grasslands and pastures for cropping. Strategies that can be applied to restore or mitigate against the adverse effects of LULUCF activities are constantly being developed and implemented. In the context of the current research, such activities include the sequestration of C into the soil. Soil C sequestration refers to an increase in soil organic carbon (SOC) and thus soil C stock (SCS) that can occur naturally or as a result of anthropogenic intervention. In terms of SCS to mitigate the effects of climate change strategies, are being developed for economical and sustainable ways to permanently sequester atmospheric C in soil. SOC sequestration is related to increases in soil organic matter (SOM) as well as offsetting GHG emissions into the atmosphere; it thus has many environmental and ecological co-benefits.

1.1.1 Australian SOC stocks

Prior to European settlement the Australian landscape included an extensive matrix of native forest, woodland, shrub-land and grassland. European settlement resulted in an extensive range of LUC activities involving clearing or thinning of land cover to make way for urban settlement, agriculture and mining (Barson, *et al*. 2000). An overview of the relative proportion of land covered by vegetation on the Australian continent prior to and after European settlement, as detailed by Barson *et al*. (2000) is summarised in Table 1-1. The distribution and proportion of land use, including vegetation cover across Australia during the period 1997-2008 is shown at Figure 1-1.

Table 1-1. Extent of land cover vegetation in Australia pre and post European settlement (After Barson *et al*. 2000).

Vegetation type	Pre-European land cover	200 years post European settlement (1980s)
Forest	9%	5%
Woodlands	21%	14%
Open woodlands	21%	25%
Shrublands	40 %	37% (Note: ~ 3 % cleared and ~ 20% thinned)
Grassland	7%	16%
Unvegetated	< 1%	Not recorded

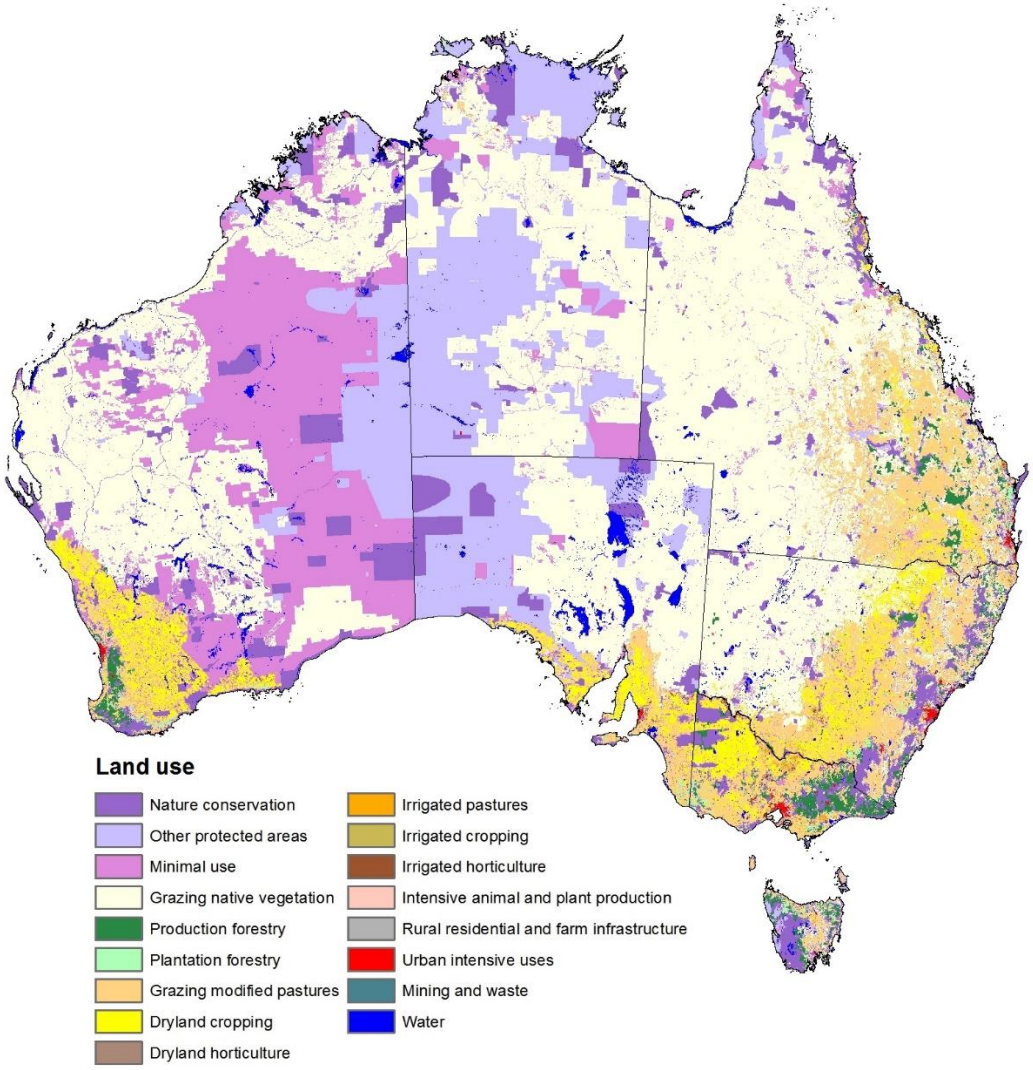


Figure 1-1. Australian Land Use (1997-2008). (Source: State of the Environment (SOE) Report, 2011).

More recently it has been estimated that approximately 456 million hectares or 59 percent of the Australian continent is used for agriculture (ABARE-BRS 2010). Of the agricultural land, 56% is estimated to be used for grazing much of which is in the arid and semi-arid areas of Australia; 37% for conservation, 4% for cropping and 2% for forestry (Mewett *et al*, 2013).

There is scant detail regarding the loss of SOC across Australian biomes. Chan *et al*. (2008) estimated that the loss of SOC from Australian cropping soils compared with pre-clearing levels ranges between 10-60% and further calculates a historical loss of SOC to be roughly 0.65 Gt C. Chan *et al*. (2008) further suggest that 0.65 Gt C is therefore the theoretical SOC sequestration potential for Australia soil which is slightly more than the Australian total sector emissions during 2013 of 0.55 Gt CO_2-e.

In another study Viscarra Rossel *et al*. (2014) estimated the total SOC stock across Australia in the 0-30 cm soil layer to be 24.97 Gt C in 2010 and the total stock on agricultural land to be 12.8 Gt C ha^{-1}. The distribution of estimated SOC stocks across the Australia continent is shown in Figure 1-2. The map shows that soils with the highest SOC stocks are found along the eastern sea board, Tasmania and the far south west corner of the continent. Lowest SOC stocks in the range of 6-10 Mg C ha^{-1} are located in the central arid region of the continent. Viscarra Rossel *et al*. (2014) also made estimates of SOC stocks to 30 cm for various agricultural land uses. The authors estimated that soils used for grazing land had the lowest average SOC stock of 24 Mg C ha^{-1} whereas soils used for grazing with improved pastures had 46 Mg C ha^{-1}. The authors also estimated soils used for cropping had 36 Mg C ha^{-} whereas soils used for irrigated cropping contained approximately 44 Mg C ha^{-}. Soils under forestry activity had 57 Mg C ha^{-1} and nature conservation 83 Mg C ha.

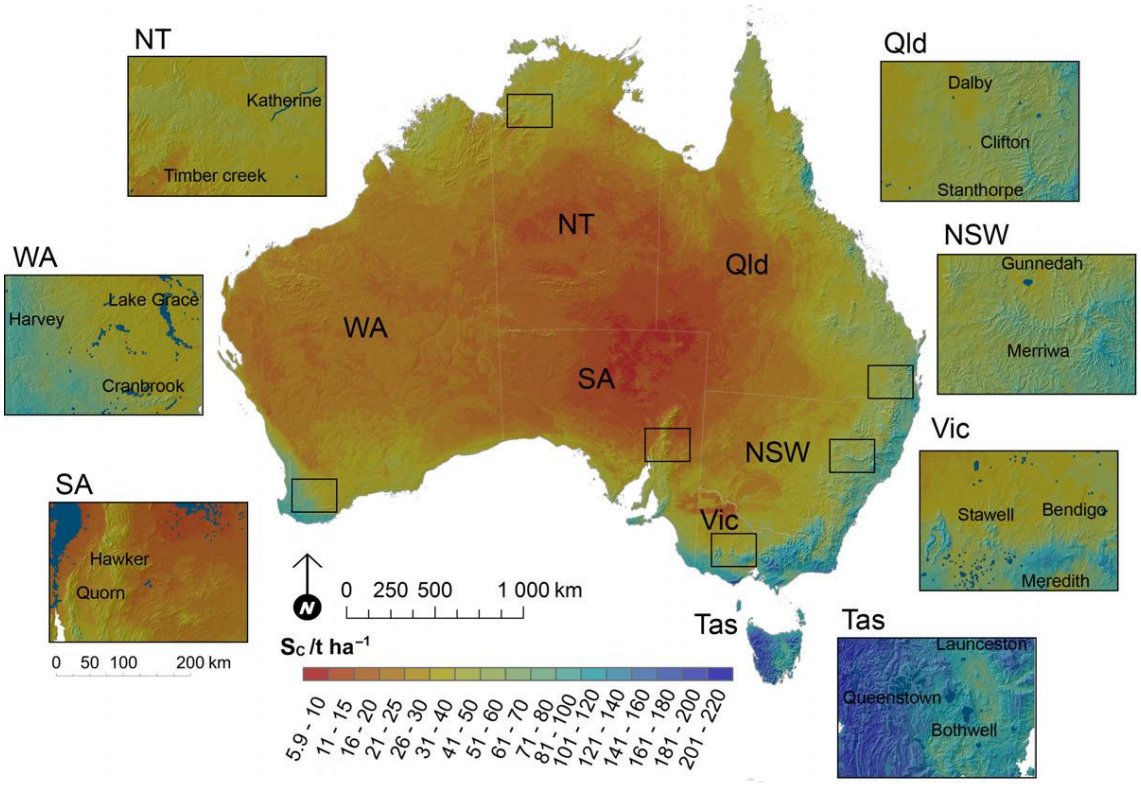


Figure 1-2. Australian SOC stock estimates for 2010 with insets showing estimates for each state or territory. Source Viscarra Rossel *et al*. (2014).

Viscarra Rossel *et al*. (2014) also estimated the area of land used for each sector. They estimated the total area of land used for agricultural activities including grazing on native vegetation to be 4.7 M km^2. Of that 3.7 M km^2 is used for grazing of native pastures, 2.9 had minimal use, 0.03 M km^2 for cropping and 0.71 M km^2 for improved grazing.

1.1.2 Historical context

In the early years of Australian agriculture there was little if any cognisance that land clearing, the introduction of cloven hoofed animals, soil tillage, stubble burning and other European based agricultural practices would have a degrading impact on Australian soils and landscapes. High stocking rates and heavy grazing, exacerbated by the introduction of rabbits resulted in the deterioration of vegetation cover across the landscape and increased soil erosion, and as a result contributed to a reduction in the natural resilience of the landscape against drought. Noble and Tongway (1986) suggest that in the first 20 years following pastoral settlement in Australia, up to 70 % of the landscape became degraded due to vegetation removal and soil erosion following the introduction of grazing into the arid rangelands. Within another 60 years large areas of the remaining 30% of the landscape also became degraded as the area of agriculture expanded and native vegetation was cleared. Beadle (1948a) explains that soil erosion in western New South Wales (NSW) can be attributed to overgrazing, notably by sheep and rabbits. In the higher rainfall areas of NSW, he attributes the erosion to land clearing for pasture improvement together with overgrazing. These activities contribute to the degradation of the grass sward to ground level thereby exposing the soil surface to wind and water erosion. Finally, in the Mulga country, Beadle attributes the removal or ring barking of timber as the cause of tree death in adjacent trees. He calls such tree death as the "isolation factor" (Beadle, 1948a:76), and he also notes that drought, old age and parasitic infections exacerbate the extent of tree death but most particularly in trees suffering "isolation".

These problems are also experienced globally. Land degradation arising from the removal of perennial native vegetation and subsequent loss of SOM through grazing or cropping is limiting the productivity of the agricultural landscape worldwide (Janzen *et al.* 1998a). Soil degradation arising from erosion leads to loss of SOC (Lal, 1998), which also reduces soil productivity. Thus, the loss of SOM and SOC leads to the loss of essential plant nutrients, reduces water infiltration and plant available water and thereby results in the loss of productive land (Lal, 2010a,b) and ultimately the ability to sustainably produce sufficient food for the world's population

There are many examples explaining how SOM is lost from the landscape, one being the use of conventional tillage methods which disrupt soil aggregates making organic material that was previously stable within aggregates susceptible to decomposition. Albrecht (1938) eloquently described the effect of modern agricultural practices on SOM

> '…the aeration of the soil by cultivation, what the pioneers did in effect was to fan the former simmering fires of acidification and preservation into a blaze of bacterial oxidation and more complete combustion.'

Albrecht was also concerned that the loss of SOM occurred at a faster rate than it could be accumulated under certain agricultural systems, but notably cropping systems. In this

circumstance and other land clearing activities, when coupled with poor management, overgrazing, and adverse climatic events, some landscapes can become degraded below a critical threshold. These fragile landscapes have lost functionality and their ability to retain essential resources. Thus their capacity to recover without some form of intervention has been severely retarded, even lost. However, other landscapes with functionality above a critical threshold can be self-sustaining (Tongway and Ludwig 2007).

A contemporary example of natural land degrading events has occurred during the first two decades of the 21st century where climatic features have impacted significantly across the Australian soil landscape. The so-called Millennium drought (Van Dijk *et al*. 2013) contributed to depleted vegetation cover, tree deaths, and widespread dust storms resulting in severe wind erosion (Leys *et al*. 2011; McTainsh *et al*. 2005; Semple *et al*. 2010) and desertification (Lal, 2010b). Such degrading events were often followed with severe flooding across large areas of the continent which exacerbated soil erosion and concomitant loss of SOM by wind and water (Heberger, 2011). Such levels of soil and SOM loss are not sustainable and lead to loss of productivity and ecosystem services. However, by increasing SOM, through LUC activities, multiple benefits can also be derived. For example, recent improvements to land management such as conservation farming, LUC and management programs promoted and supported by organisations including Landcare, Local Land Services (and its predecessors) and Greening Australia (GA) amongst others, have been aimed at developing landscape resilience. Such programs have contributed to restoration of nutrient cycling processes, soil water retention, biodiversity increase, prevention of erosion and deposition and have led to productivity improvements (SOE, 2011). The improved land management programs and strategies have also contributed to increases in SOM and SOC.

1.1.3 Land use change to sequester soil organic carbon

Soil can be both a source and sink of atmospheric CO_2, but the SOC stocks are fundamentally determined by the balance between carbon inputs and losses with the balance being driven by the processes of CO_2 uptake via photosynthesis and CO2 losses by decomposition, respiring plants animals and microbes. Other factors that influence the potential of soil to sequester carbon include microbial activity, climate, edaphic characteristics and land use. A known way to sequester atmospheric CO_2 and reduce emissions in soil is to adopt changes to soil management practices. Consequently, there is ongoing interest in the potential of the terrestrial biosphere to help mitigate against the effects of human induced CO_2 emissions and in particular the potential of the soil to act as a sink for atmospheric CO_2. However, the quantity of atmospheric CO_2 that actually can be sequestered by soil is constrained to the quantity previously lost through disturbance (Lal, 2004b; Mackey *et al.* 2013; Stockman *et al*. 2013) such as tillage or LUC including conversion from forest to agriculture or from pasture, or native grass to crop. This is because before disturbances, the quantity of SOC is in a state of equilibrium with the

environment; simply fluctuating with climatic and other environmental factors. After disturbance SOC is lost from the soil through mineralisation and decomposition, leading to a new, often lower, equilibrium (Batjes, 1998; Powlson *et al*. 2011).

Soils that have previously lost SOC provide the best opportunity to effectively sequester CO_2 (Cowie *et al*. 2013; Govers *et al*. 2013). Rehabilitation of C depleted landscapes can be achieved through the adoption of contemporary farming techniques and LUC activities aimed at increasing SOM. The farming methods most likely to lead to improvements in degraded landscapes include conservation tillage (e.g., no till, reduced till), pasture cropping, conservation grazing techniques (e.g., rotational grazing, high intensity short duration (HISD) grazing or holistic management), conversion of cropping to pasture land, retention of stubble and reforestation. These methods can result in the retention and SOC sequestration, which in turn can improve soil structure, improve infiltration, retain soil moisture, be a food and habitat source for soil biota, improve soil nutrient status and provide resilience against erosion. Many of the LUC methods have been developed to cope with the unique environmental conditions found in Australia and many can be adopted at either or both the macro and micro scale (Chan *et al*. 2010; Luo *et al*. 2010). However, the evidence that contemporary farming methods actually lead to an increase in SOC especially in the short term in Australian soil is limited (Chan *et al*. 2010; Cowie *et al*. 2013; Hoyle *et al*. 2013; Eyles *et al*. 2015).

The addition of organic soil amendments such as manure, composts and char has also been proposed as providing resources that enhance the potential to sequester SOC. However, Cowie *et al*. (2013) found no significant difference in SOC between cropped sites treated with organic amendments compared with those that received chemical fertiliser. Quilty and Cattle (2011) also showed that mineralisation of C can be enhanced following application of organic amendments because it can stimulate biological activity which in turn, may result in a loss of SOC.

1.1.4 Australian position on soil carbon sequestration

Since the commencement of the research reported in this thesis Australia's political emphasis on climate change has ebbed and flowed largely as a result of the policies of different Governments. However, SOC sequestration has consistently remained a part of the political debate because it is seen as a plausible way for Australian farmers to contribute to mitigating the effects of rising atmospheric CO_2 levels. The management and monitoring of SOC was stated to be a matter of national and international importance (SOE, 2011:270). Maintaining SOC is important for soil health, nutrient cycling, productivity, soil structure and water storage, but more recently has emerged as a strategy to capture and store atmospheric CO_2 in soil previously depleted of C, and hence is seen as a way of mitigating increased atmospheric CO_2 levels. Underpinning Australia's position on SOC sequestration is a commitment to the Kyoto Protocol (1998) of the United Nations Framework Convention on Climate Change (UNFCCC).

Signatories to these international agreements are bound to reduce GHG emissions and meet GHG inventory reporting obligations. As a signatory to the Kyoto Protocol, Australia agreed to reduce GHG emissions during the 2008-2012 commitment period to 108 percent of the emissions in 1990. Most recently, in 2015 the 21st Conference of the Parties (COP 21) was held in Paris. The Paris meeting achieved universal agreement and legally binding commitments from nations to further reduce GHG emissions in an endeavour to combat climate change and aim to retain average global temperature increases to below 2°C, and aspire to keep increases below 1.5°C. At this meeting Australia set an emissions reduction target of 26 to 28 % below 2005 levels by 2030 (DPMC, 2015).

Australia's total GHG emissions was 0.53 Gt CO_2-e during 2014 (National Inventory by Economic Sector, 2014). The agriculture, forestry, fishing sector GHG emission contribution during 2014 was 0.083 Gt CO_2-e, or 16% of the total emissions (Figure 1-3).

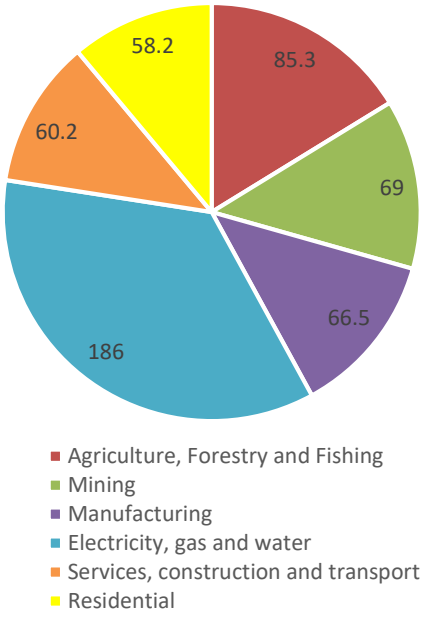


Figure 1-3 Australia's direct Greenhouse Gas Emissions (Mt CO_2-e) by Economic Sector 2014 (Source: Department of Environment Report: National Inventory by Economic Sector, 2014).

Concerns about the increase in GHG emissions and global warming have generated a considerable body of research both in Australian and internationally, that has explored the causes of emissions, and ways to both reduce atmospheric CO_2 levels and alleviate the effects of increased CO_2 emissions (Guo and Gifford, 2002; Sanderman *et al*. 2010; Janzen, 2015). The enormity and gravity of the situation has been the impetus for international and Australian political action aimed at both reducing CO_2 emissions and slowing the rate of climate change.

In Australia, the actions have included legislation to establish emission reduction targets and an emissions trading scheme. Garnaut (2011), who investigated scientific, political and economic elements of climate change action concluded with the premise that science had provided

evidence beyond reasonable doubt that anthropogenic activity was causing global warming. In 2011, the Federal Government implemented legislation establishing the Carbon Farming Initiative (CFI) (*The Carbon Credits (Carbon Farming Initiative) Act, 2011)*. The CFI is designed to help Australia meet Kyoto Protocol target obligations. The methods enable landholders to voluntarily submit projects to earn C credits through activities that lead to C sequestration or avoid C emissions.

More recently, the Federal Government has proposed changes to the CFI that include the establishment of an Emissions Reductions Fund (ERF) and an expansion of the range of activities that may become eligible greenhouse gas offset activities. The Government has stated that reforestation, C sink plantings and SOC sequestration will be a key land sector abatement activities that can reduce GHG emissions and therefore will be eligible for inclusion in the ERF; a key component of the Direct Action Plan. The ERF is designed to help Australia meet emissions reduction targets and involves the purchase, by the Government by auction, of Australian Carbon Credit Units from businesses and landholders who adopt practices aimed at emission reduction. Further, Australia signed a voluntary pledge to increase SOC by 0.4% annually. The global declaration, referred to as "4/1000 soils for food security and climate", was an aspirational initiative introduced by the French government during the Paris climate talks held in 2015 to promote mechanisms such as improved land management, conservation agriculture and forestry as ways to sequester SOC (MAAF, n.d).

1.2 Research approach

This research project has examined two LULUCF methods and their potential to sequester SOC. The methods include:

(i) Establishment of biodiverse environmental plantings (BEPs) of woody perennials on land previously used for agriculture; and

(ii) Rehabilitation of scalded clay pan soil by waterponding.

The Domestic Offsets Integrity Committee (recently replaced by the Emissions Reduction Assurance Committee) approved both of these type of activities on the CFI Positive Activity List which means that they are activities that can earn carbon credits for storing C or reducing GHG emissions. A CFI Methodology has also been developed for native mixed species environmental plantings (MSEP) (*Carbon Credits (Carbon Farming Initiative) (Reforestation by Environmental or Mallee Plantings-FullCAM) Methodology Determination 2014)*. The methodology provides a mechanism for landholders wanting to establish BEPs by seeding or planting native trees and thus increasing above ground biomass (AGB) to earn C credits when they meet prescribed eligibility, monitoring, reporting and auditing criteria. At present there is no equivalent methodology specifically for rehabilitation of scalded soil areas, although it is possible landholders may, in the future be able to claim C credits by using the CFI methodology

applicable for sequestering C in soils in grazing systems (*Carbon Farming (Carbon Farming Initiative) (Sequestering Carbon in Soils in Grazing Systems) Determination, 2014)* .

The two LULUCF methodologies mentioned form the basis of two experimental case studies to examine the capacity of each to sequester SOC against a background of LUC as follows:

1. **Biodiverse environmental plantings (BEP)** established on agricultural soil across the temperate region of the Southern Tablelands of NSW. BEPs are commonly established to provide wind shelter and shade for livestock, pastures and crops, wildlife corridors, or to a lesser extent as source of firewood. In addition BEPs produce large quantities of AGB and have potential to sequester C in the AGB. The potential to sequester SOC is yet to be investigated.
2. **Waterponds** constructed on clay pan scalded soils across the semi-arid rangelands of the Central West region of NSW. Due to overgrazing these soils have lost their A-horizons and have a very low concentration and SCS. Waterponds are established to facilitate revegetation of scalded soil on clay pans. Once restored soil structure improves and a range of plant species grow, many of which are suitable for grazing. Soil changes following waterponding have the potential to increase SCS.

It is anticipated that the results of this research will provide empirical data that gives a better understanding of how the two LUC activities result in the SOC sequestration. In both case studies a chronosequence of study sites has been used thus allowing an estimation of a temporal rate of change. In particular the research will:

- Contribute data that will improve modelled predictions of SOC sequestration in BEPs, and hence also for the CFI Methodology (*Carbon Farming (Carbon Farming Initiative) (Sequestering Carbon in Soils in Grazing Systems) Determination, 2014)*; and
- Provide empirical evidence that waterponding is an effective method of sequestering SOC and thereby contribute data towards the development of a CFI methodology appropriate to waterponding.

1.3 Hypothesis

The key assumption of this research has been that both of the LULUCF activities studied will lead to sequestration of atmospheric CO_2 in soil. Leading on from this assumption is the expectation that once sequestered, the concentration and SCS will remain relatively constant over time. Additionally, the factors contributing to the C sequestration potential of the soil have been assumed to be consistent between the different soil types studied, their landscape and climatic environments. Thus, the hypothesis of this research project is that:

- The two LUC methods studied will result in sequestration of atmospheric CO_2 in soil.

1.3.1 Aims and objectives

The overall aim of this project has been to identify the C sequestration potential of agricultural soil that has undergone LUC either through the establishment of BEPs or following the construction of waterponds on scalded soil.

Both case studies had the following common principal aims:

- To identify the C sequestration potential of agricultural soil following each LUC activity; and
- To identify biogeochemical factors that contribute to the C sequestration potential of soil after each of two LUC methods have been implemented.

Secondary aims specific to individual case studies include:

- An assessment as to whether AGB in BEPs leads to a commensurate increase in SOC, and whether biomass measurements can be used to predict SOC;
- A comparison of traditional analysis methods for soil C with mid infrared spectroscopy (MIRS) predictions, and the use of MIRS to predict changes to soil C fractions following establishment of BEPs;
- A comparison of the measured AGB and soil C in BEPs with modelled data using the Full Carbon Accounting Model (FullCAM); and
- To gain an understanding of clay mineralogical changes associated with SOC sequestration following waterponding.

1.4 Thesis Structure

The two case studies will be addressed sequentially by this research and are therefore presented in individual chapters. The results from both case studies will then be considered together in a synthesis chapter to compare the potential of each LUC activity to sequester atmospheric CO_2 as SOC. An outline of this thesis is presented in Table 1-2.

Table 1-2. Thesis structure.

Chapter	Description
1. Introduction	Introduction to thesis; background, aims and objectives
2. Literature review	Review of literature on topics including the causes of global CO_2 emissions, LUC effects on soil C, SOC sequestration, the role of organic matter in SOC sequestration, a discussion on the chronosequence approach and SOC and geomorphic processes.
3. Methods	Detail of the field and laboratory methods common to both case studies.
4. Case study 1 – Biodiverse Environmental Plantings (BEP) and SOC sequestration	Detail of the aim and objectives, site details and sampling methods. Provides a literature review of BEPs and their potential to sequester SOC, the Australian reforestation policy and C farming methodology for environmental plantings, the use of FullCAM to estimate SCS in BEPs, SOC fractions, and the use of MIRS for SOC analysis. Finally, the results, discussion and conclusion of the study into the AGB and SOC sequestration potential of BEPs are presented.
5. Case study 2 – C sequestration potential of scalded soil rehabilitated by waterponding	Literature review of scalding and the evolution of the waterponding method; provide details of the study sites, methods of analysis undertaken unique to Case Study 2, and present the results and discussion of the findings relevant to SCS, and soil mineralogical changes following waterponding.
6. Synthesis	Synthesis of results related to SCS change in the context of state-and-transition and ecological response frameworks for the two case studies to draw out the findings in terms of the stated aims and hypothesis of this research project.
7. Conclusion	Includes a summary of the key findings from the two case studies in relation to the stated aims and objectives, an overview of limitations of the research and recommendations for further study.

Chapter 2: Literature Review

2.1 Introduction

Carbon is present in four actively cycling, interlinked pools including the atmosphere, biosphere, oceans and the soil and the carbon cycle (CC) regulates the natural flux between these pools. Atmospheric CO_2 is produced naturally by respiring plants, animals, fungi and aerobic organisms, during consumption of organic compounds, from the decay of organic matter and during volcanic outgassing. These natural sources of CO_2 are counterbalanced by the three natural sinks of CO_2; the oceans (and sediments), the atmosphere and the terrestrial biosphere. Naturally occurring CO_2 is sequestered in the oceans by the burial of marine sediments and in the terrestrial biosphere through the rock weathering cycle. Plants also sequester C by converting atmospheric C into plant biomass through the process of photosynthesis. These sources and sinks of CO_2 exist in balance, regulating the exchange of C produced by biomass inputs and losses, through the CC process.

The Intergovernmental Panel on Climate Change (IPCC) presents evidence to indicate that the CC balance has been disrupted due to anthropogenic activities, and those activities have contributed to significant increases of CO_2 emissions (IPCC, 2007). Further, the evidence shows that increased CO_2 emissions have affected the atmospheric CO_2 levels. There is evidence from climate observations that climate change and global warming is occurring as a result of the increases to GHG concentrations and positive radiative forcing. There are two principal causes of the rise in global CO_2 emissions; fossil fuel combustion and LUC which includes deforestation.

There is considerable global effort being made towards lowering the concentration of atmospheric GHGs. Indeed the charter of the UNFCCC is to implement strategies to constrain climate change, global temperature rise and to promote adoption of global emission reduction targets (Anjani *et al*. 2013). Reducing the quantity of fossil fuel emissions through clean energy options, greater use of renewable energy and adoption of energy saving technologies, will help to alleviate energy demand, reduce C emissions and eventually lead to lower atmospheric CO_2 concentrations. One way that is promoted as being able to reduce atmospheric CO_2 is to sequester C in soil and vegetation. SOC sequestration is particularly possible where SCS have previously been largely depleted. The depletion arises, most typically through land degradation, erosion, desertification and LUC activities that are used to convert land to make way for agriculture (lal, 2004a,b). However, LUC can also be undertaken for the purpose of sequestering atmospheric CO_2 by both biomass and the soil.

This literature review will first explain the two principal anthropogenic causes of CO_2 emissions that contribute to increased atmospheric CO_2 concentration, and why this is of concern. Then the potential of the terrestrial biosphere pool, particularly the role of soil to sequester

atmospheric CO_2, and LUC activities that can be applied to enhance SOC sequestration, will be examined. This review will also:

i. Consider the role LUC has on C emissions and opportunities for SOC sequestration;
ii. Identify the pools of C that are relevant to LUC;
iii. Discuss the forms of C;
iv. Review the literature pertaining to factors leading to soil organic C (SOC) loss;
v. Review the literature pertaining to SOC sequestration;
vi. Consider the C sequestration potential of soil;
vii. Discuss the concept of SOC equilibrium;
viii. Examine the effects of soil management on SOM and SOC;
ix. Discuss the use of chronosequence studies;
x. Consider SOC storage with regards to geomorphic processes; and
xi. Summarise the processes.

2.2 Anthropogenic causes of increased atmospheric CO_2 concentrations

Concerns surrounding the increase in GHG in earth's atmosphere as a result of anthropogenic activities include the potential for global warming and climate change. Climate change can be influenced either by internal processes or by external forcing processes. Internal processes are caused by a number of natural forces including sea surface temperature, El-Niño/La-Niña conditions, condensation of water vapour, ocean fluxes, and ice sheet changes. External forcing processes can occur due to solar variability, volcanic eruption and anthropogenic GHG emissions. It is recognised that the two processes are linked. The IPCC provide evidence to show it is extremely unlikely that internal variability alone can explain the global warming trend observed over the past 50 years. They also provide evidence that anthropogenic activity contributes to climate change and has very likely caused warming of the global climate system (IPCC, 2007; Hartman *et al*. 2013).

In the Fifth Assessment Report of the IPCC, Working Group 1 provided detailed atmospheric and surface observations (Hartman *et al*. 2013). In their report they provided evidence of changes to the composition of the atmosphere. They reported that the concentration of atmospheric CO_2 was 40% greater in 2011 than in 1750 (390.5 ppm and 278 ppm respectively. CH_4 was 150% greater in 2011 than in 1750 (1803 ppb and 722 ppb respectively) and N_2O was 20% greater in 2011 than in 1750 (324.2 ppb and 270 ppb respectively) (Hartman *et al*. 2013).

The increased reliance on fossil fuel combustion and LUC has contributed to the increase in anthropogenically induced CO_2 emissions. Fossil fuel combustion and LULUCF are widely accepted as being extremely likely to have contributed to global warming and climate change (Gifford *et al.* 1992; IPCC, 2007; Lal, 2009; Raupach and Fraser, 2011). The IPCC (2007)

estimates that fossil fuel combustion has contributed approximately 2/3^{rds} of anthropogenic CO_2 emissions and LUC has contributed approximately 1/3rd. Of these emissions, approximately 45% remains in the atmosphere, 30% has been taken up by the oceans, and 25% taken up by the terrestrial biosphere.

Evidence of the effects of climate change due to increases in atmospheric CO_2 include an increase in global mean surface temperature (Figure 2-1), changes in extremes of temperature, increases in arctic temperatures with associated temperature increases in the oceans and sea level rise, increased incidence and duration of droughts in the tropics and sub tropics, and an increased incidence in heavy precipitation events, even where overall rainfall has reduced (IPCC, 2007; Hartmann *et al*. 2013).

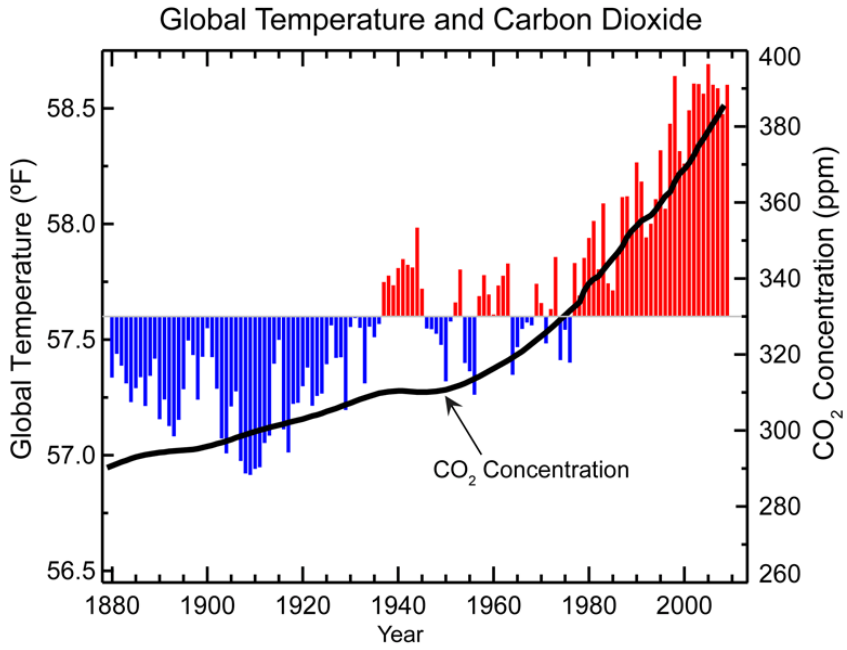


Figure 2-1. The relationship between the atmospheric CO2 concentration and increases in temperature. (Source: National Oceanic and Atmospheric Administration: National Centres for Environmental Information available from: https://www.ncdc.noaa.gov/indicators/)

2.2.1 Fossil fuel combustion as a source of increased atmospheric CO_2

Globally, fossil fuel combustion is the single largest source of CO_2 emissions. Fossil C including coal, oil and gas is naturally stored in the geosphere. When undisturbed this C pool is dormant and plays no role as part of the natural CC. However, when disturbed particularly through anthropogenic activity, this pool becomes a source of atmospheric CO_2 emissions. The activities include mining and burning of fossil fuels to meet global energy requirements.

The rate of anthropogenic CO_2 emissions from fossil fuel burning has increased over the past 150 years since the beginning of the industrial revolution and during that time there has also been a commensurate rise in atmospheric CO_2 concentrations. The emissions are cycled into the atmosphere and disrupt the natural equilibrium of the CC (Schimel *et al*. 2000a; Janzen, 2004).

The IPCC (2007) has demonstrated that the rapid increase in fossil fuel CO_2 emissions over the past 150 years has caused the elevated atmospheric CO_2 level. Evidence from polar ice core records quantifies the change in atmospheric CO_2 concentration that has occurred since the industrial revolution (IPCC, 2007). Pre-industrial levels of CO_2 have been found to be approximately 280 ppm (Batjes, 1992). More recently, direct measurements recorded by the National Oceanic and Atmospheric Administration (NOAA, 2013) reported that earth's atmosphere contained on average approximately 394 ppm CO_2 in 2012. This is equivalent to a mean annual increase in atmospheric CO_2 of approximately 2.7 ppm $year^{-1}$. By comparison, the IPCC have estimated that there was an increase of only 20 ppm atmospheric CO_2 in the 8 000 years prior to industrialisation (IPCC, 2007). Most recently the global monthly mean atmospheric CO_2 concentration exceeded 400 ppm in March 2015, a 70% increase over pre-industrial levels. (NOAA, 2015).

2.2.2 Land use change as a source of increased atmospheric CO_2

The terrestrial biosphere has a role as both an emitter and sink for atmospheric CO_2. Emissions of CO_2 from the biosphere occur naturally from fire and through heterotrophic respiration, which involves the release of CO_2 by soil organisms during decomposition of organic matter. Natural variations in atmospheric CO_2 levels and rates of GHG emissions occur due to seasonal, spatial and temporal factors (Batjes, 1992; Baldock *et al*. 2012). Rates of emission depend on climatic conditions particularly rainfall and temperature (Post *et al.* 1982). Anthropogenic activities result in large changes to the terrestrial biosphere. These activities have a major effect on the natural pools and fluxes of C and thus the CC. Anthropogenically induced emissions of CO_2 occur through LULUC activities including deforestation, burning of biomass and cultivation. Changes to the natural biophysical land cover (e.g. native forests, grasslands, tundra) through deforestation or reforestation also change the albedo, and the natural water and C fluxes of the planet. Changes to land use results in a net release of C to the atmosphere and are also important in terms of accounting for C credits and debits under C trading schemes.

A summary of literature on anthropogenic LULUC activities that influence the C pools and fluxes and the mechanisms involved are summarised in Table 2-1. The activities are broadly divided into conversion of forest, cropland, grassland and degraded land and the management change. Studies into changes show variability to SCS following LUC depending on edaphic characteristics, prior land use, new land use and climate (Post and Kwon, 2000). For example conversion of degraded land usually results in soil C increase whereas conversion of productive land can result in soil C decrease (Lal, 2003, Powlson *et al*. 2011; Govers *et al*. 2013). The conversion of forest to cropland normally leads to decreases in SOC, whereas conversion of forest to grassland can lead to SOC increases of up to 50% (Guo and Gifford 2002; Murty *et al*. 2002). Where cropland is converted to forest there are usually gains in SOC, but this may take

up to 30 years (Paul *et al*. 2002a). Grassland converted to forest may result in very little change or even loss of SOC (Guo and Gifford 2002; Paul *et al*. 2002a; Cunningham *et al*. 2012).

Table 2-1. Summary of literature identifying influences on C pools and C fluxes following land-use change.

Activity	**Outcome (L = Loss, G = Gain, NC = minimal effect)**	**Mechanisms**
Forest conversion		
Forest remaining forest	NC	In undisturbed forests systems SOC is in equilibrium. However, changes to SCS can arise from above and below ground biomass growth. Losses occur to biomass and SOC following harvesting, burning, dead wood and litter decomposition especially when converting old growth to plantation forests (Harmon *et al*. 1990; IPCC, 2003).
Conversion of existing forest to cropland	L	Normally results in SCS decrease to 60cm depth (Guo and Gifford, 2002). Losses due to forest biomass harvesting, and native soil cultivation (Houghton, 1999; Houghton, 2010; Sanderman *et al*. 2010). Losses to soil C following conversion (Murty *et al*. 2002).
Conversion of existing forest to grassland/pasture	G	Highly variable temporal and spatial effect on SOC with some research identifying losses, others finding gains (Post and Kwon, 2000). SCS increase, but rainfall dependent (Guo and Gifford, 2002). Increases to SOC in approximately 50 % of studies. Some studies showed large increases attributed to low baseline content of SOC. Management practices affect SOC sequestration potential (Murty *et al*. 2002).
Cropland conversion		
Cropland remaining cropland	L/NC	Cropping sites have on average up to 35 % less SOC than pasture sites in the 0-10 cm layer, depending on soil type and clay content (Cotching *et al*. 2013). SOC loss following long-term (30-50 years of cultivation) (Post and Kwon, 2000). Potential for SOC loss or gain depends on previous management history and current management practices (Chan *et al*. 2011). Cultivation disrupts aggregates and results in loss of C from the particulate organic matter (POM) fraction (Cambardella and Elliott 1993). Cultivation results in SOC loss especially in top 0.1 m (Luo *et al*. 2010).
Land converted to cropland	L	Normally results in CO_2 emissions from biomass and soil. Globally 93 Pg C emitted between 1850- 2005 (Houghton, 2013).

		Change from forests and grasslands normally result in net loss of C from both biomass and soil (IPCC, 2003). Compared with adjoining native land, cultivation of land leads to SOC loss (Luo *et al*. 2011).
Annual crops (cereals, vegetables)	L	No long term storage of C in the biomass (Lemus and Lal, 2005). Soil C loss up to 51% over 50 years. New low equilibrium achieved (Luo *et al*. 2011).
Perennial woody crops (orchards, vineyards and agroforestry)	G	Long term C storage in biomass depends on pruning, harvesting, growth rates.
BEPs	G/L	Increase in biomass. Variability in SOC sequestration rates (Paul *et al*., 2002a; Cunningham *et al*. 2012; Prior *et al*. 2015)
Cropland converted to grassland	G	Results in increased SCS. Most increase occurs in top 20 cm (Guo and Gifford, 2002; Powlson *et al*. 2011). Chan *et al*. (2011) found pasture phase is needed to increase SOC in cropland.
Cropland converted to forest	G	Increased SCS and biomass C (Guo and Gifford, 2002); Decrease initially, with potential for increase after 30 years (Paul *et al.* 2002). SOC sequestration is more likely following conversion from cropland to forest than from pasture. (Laganiere *et al*. 2010).
Grassland conversion		
Grassland remaining grassland	NC	Normally in equilibrium with environmental factors including climate, biomass, topography and parent material (Dalal and Mayer, 1986). SOC variation influenced by management (harvesting, rangeland degradation, grazing, fire, pasture management) (Chan *et al*. 2010).
Conversion of grassland to cropland	L	SCS decrease following cultivation although the quantity lost is rainfall dependent (Guo and Gifford, 2002). Losses of 10 to 30 Mg C ha^{-1} to 30 cm depth following clearing of pasture or native vegetation for cropping (Valzano *et al*. 2005). Poorly managed or overgrazed pastures may lose only a small quantity of C when converted to cropland, relative to well manage pastures (Valzano *et al*. 2005).
Conversion of grassland to forest	L / NC /G	Tree type, climate and previous land use will determine the extent to which SOC can be sequestered in soil following afforestation (Lugo and Brown (1993). Tree type and rainfall influence whether there is an increase or decrease in SCS following conversion of grassland to forest (Guo and Gifford, 2002). Decrease in first 10 years, possible increase after 30 years (Paul *et al*, 2002a; Paul *et al*. 2004; Polglase *et al*. 2000).

		SOC increases in the top 0-5 cm of the soil profile are possible. However, losses down to 25 cm depth result in a net loss of SOC from the profile following afforestation (Vesterdal *et al*. 2002). No change over a 30 year time frame following reforestation with mixed species environmental plantings (MSEP) (Hoogmoed *et al*. 2012; Cunningham *et al*. 2012).
Conversion of degraded land		
Conversion of degraded land to productive land	G	Usually results in increased biomass and hence an increase in SOC. Quantities of between 9-15 Mg C ha^{-1} over 30 years (Powlson *et al*. 2011). Restoring soils previously depleted of SOC provides a higher probability of sequestering C than in soils already saturated with C (Govers *et al*. 2013). Degraded soils depleted of SOC have the potential to sequester SOC at rates between 3.5 and 4.5 Mg C ha^{-1} yr^{-1} following rehabilitation using different fruit tree and grass species combinations (Lenka *et al*. 2012). Global C sequestration potential of degraded land and deserts is 0.8 – 1.3 Gt C yr^{-1} (Lal, 2004a). Conversion of degraded land to agroforestry provides significant opportunity to reduce GHG emissions (Dixon 1995).
Conversion of productive land to degraded land	L	Total land mass affected by soil degradation is 15 % of global land area (Lal, 2004a,b). Soil erosion results in SOC and biomass losses (Lal, 2003). Estimated total SOC loss of 26 ± 9 Gt C has resulted from global soil degradation (Lal, 2004a).
Conversion of scalded soil using waterponding		There has been no previous research into changes to SCS following waterponding. Case Study 2 will investigate the changes.
Land management change		
Reduced soil disturbance – reduced or no-till	G / NC	Changing cropping practices to conservation agricultural practices can result in increases to SOC (Luo *et al.* 2010). Conventional tillage has lower SCS than direct drill except at the 0-10 cm soil layer where there was found to be little difference between conventional and direct drill methods. (Valzano *et al*. 2005). No-till results in increases in SOC up to 15 cm depth (Sanderman *et al*. 2010). Conversion of conventional to no-till leads to an SOC sequestration rate of 0.57 Mg C ha^{-1} yr^{-1} until new equilibrium by 15 years (West and Post, 2002). Conversion to no-till does not increase SCS but can slow the rate of C loss (Page *et al*. 2013). Conversion to conservation tillage estimated to result in SOC sequestration of 0.02-0.04 Gt C yr^{-1} in the US alone (Lal *et al*. 2007).

Grazing management – set stocking to rotational grazing methods	G / NC	Well managed grazing can result in increases, or maintenance of SCS. Reduced grazing can lead to decreases in SCS (Conant *et al*. 2001; Gifford, 2010). No significant difference, but a trend suggesting SCS was higher in pastures being grazed with a rotational grazing system compared with those being grazed continuously (Cowie *et al*. 2013). Pastures grazed using a continuous grazing management system can have higher SCS than pastures grazed using a rotational grazing management method (Orgill *et al*. 2013).
Cultivation of previously untilled grassland	L	Between 20-40 % of SOC can be lost from grassland following cultivation. Up to 40% is lost from the A-horizon and the loss occurs rapidly within the first 5 to 20 years after cultivation (Davidson and Ackerman, 1993).

Changes in terrestrial ecosystems which can lead to a loss of C are influenced by the activities of deforestation, biomass burning, and LUC (Batjes, 1996). The effects of deforestation are twofold; firstly removing the trees reduces the amount of biomass C and secondly the capacity of the introduced vegetation (crops/pasture) to sequester C above and below ground in the longer term is lower than that of forest trees (Gifford *et al*. 1992). Converting native land (undisturbed natural state) to make way for agricultural activities including cropping and pastures can result in reduced SCS (Houghton, 2010; Paustian *et al*. 1997). Additionally, the effects of LUC where land is converted from its natural state (e.g. clearing of native forests) to make way for agriculture also results in the release of CO_2 from the soil into the atmosphere. For example, during the period between 1850-1990 agricultural expansion including cropping and grazing, and the conversion of native grassland and forest to pasture have been the principal LUC activities to result in loss of C to the atmosphere. These activities have been estimated to result in a loss of between 108-188 Gt C to the atmosphere and represent approximately 35% of anthropogenically induced CO_2 emissions (Houghton, 1999; Foley *et al*. 2005; Houghton, 2010). Batjes (1998) estimated that global release of C into the atmosphere due to anthropogenic activity during the 1980s was 7.1 ± 1.1 Gt C $year^{-1}$. Of this release 1.6 ± 1.0 Gt C $year^{-1}$ was attributed to tropical LUC and 5.5 ± 0.5 Gt C $year^{-1}$ due to fossil fuel combustion and cement production. Emissions from agriculture, forestry and other land use (AFOLU) activities during the 20 year period from 1990-2010 increased by 8 % with 7.5 Gt CO_2-e reportedly emitted during the 1990s and 8.1 Gt CO_2e during the 2000s (Tubiello *et al*. 2014).

In Australia, the total amount of C stored in AGB has been estimate by Gifford (2000) to be approximately 12 Gt. Gifford *et al.* (1990) further estimate that the loss of C as a result of tree clearing in Australia for the 200 years since European settlement is approximately equivalent to 7 Gt C as CO_2 released to the atmosphere.

2.3 Pools of carbon relevant to land use change CO_2

In the long term, identifying and developing alternate sources of energy to fossil fuels will be the most effective way to reduce GHG emissions and lower atmospheric CO_2 concentrations. Solar and wind energy are important first steps, but energy efficient ways of replacing fossil fuels will be the most important development for the future. In the meantime it may be possible to mitigate the increase of GHG emissions into the atmosphere by transferring the atmospheric CO_2 into long term sinks. Potential sinks can include the oceans and the terrestrial biosphere.

The oceans are estimated to contain around 39 000 Gt C and are the largest reservoir of C. They contain approximately 50 times more C than the atmosphere which is estimated to contain 780-785 Gt C as CO_2, with 1 ppm of atmospheric CO_2 being equivalent to 4 Gt of C (Janzen, 2004; Lal 2008a).

The terrestrial biosphere includes the vegetation and soil pools of C. Carbon stocks in the biota include forests, grasslands and savannas, deserts, tundra, cropland and wetlands. It is estimated these biomes contain between 400-600 Gt C (Janzen, 2004; Lal, 2008a). Soil is a natural reservoir of organic carbon and is also an important sink for atmospheric CO_2. Soil stores approximately 65% of the total terrestrial C pool as SOC (Schimel *et al.* 1994).

Estimates of the size of the actively cycling global soil C pool vary due to the difficulty in accurately measuring soil C content, the heterogeneous distribution of C in soil and seasonal fluctuations. The quantity of SOC is highest near the soil surface and declines with depth with litterfall and plant roots contributing to the C profile (Schlesinger, *et al.* 2000; Jobbagy and Jackson, 2002).

Janzen (2004) has estimated that between 1 500 to 2 000 Gt C is stored in the top 1 m of soil. Other estimates suggest the soil contains:

- Between 700 Gt to 2 946 Gt, with 1 395 Gt C to 1m (Post *et al*. 1982);
- Between 1 200 to 1 600 Gt C in the top 1m of soil, and 2 376 to 2 456 Gt C in the top 2m (Batjes, 1998);
- 1 502 Gt C in the top 1m of soil (Jobbagy and Jackson, 2002).
- Four pools with 1 550 Gt in the SOC pool, 950 Gt C in the soil inorganic C (SIC) pool, 40 to 80 Gt C in litter and 150 Gt C in peat (Lal, 2008); and
- Up to 2 300 Gt C is stored in soil to a depth of 3m Jagadamma & Lal (2010).

Despite the variation in estimates, soil effectively contains approximately 3 times more C than vegetation, and more than twice that found in the atmosphere. The top 1m of Australia's soil is estimated by some to contain up to 51.8 Gt of SOC, with 10.7 Gt in the top 10cm (Gifford *et al*, 1992).

An explanation of the composition of SOC, the processes that contribute to losses and gains following LUC and the potential of soil to sequester SOC follows.

2.4 Soil organic carbon

There are three forms of C including elemental, inorganic and organic (Table 2-2). The C may be present in soil as either elemental C such as charcoal or coal, inorganic C derived from soil parent material including calcium carbonate ($CaCO_3$), or organic C which is generally termed SOC. Soil C can be measured as either total soil C (TSC) or SOC. TSC refers to the sum of both inorganic and organic C pools of C in soil. Soil inorganic C (IC) is found mainly in soils from arid and semi-arid areas and in soils formed in association with calcareous parent materiel. Carbonaceous materials including $CaCO_3$ and magnesium carbonate ($MgCO_3$) are the dominant forms identified and are usually associated with alkaline soils. Most soil C is associated with SOM, with smaller quantities associated with (IC) and charcoal.

Table 2-2. Forms and sources of C.

Form of C	**Origin**
Elemental	Graphite, diamond, amorphous C
Inorganic	**Primary sources** of carbonate are deposited by dust or derived from parent material weathering including calcite, dolomite, aragonite, siderite **Secondary sources** derived from precipitates commonly in soils with pH range of 7.3 – 8.5 and adequate Ca^{2+} or Mg^{2+} supply
Organic	Soil organic matter (SOM) including plant debris and litter at various stages of decomposition as follows, both living and non-living: • Particulate Organic Matter (labile pool) • Light fraction • Humus (stable pool) • C content of SOM ranges between 40 and 60% by mass (Sanderman *et al.* 2010) Microbial biomass – soil organisms

The CC links all the forms of C. This review will not consider elemental or inorganic forms of soil C since they are seldom able to be manipulated by LUC. Rather it will focus on the organic form of C, the active exchange of CO_2 between the atmosphere and terrestrial biosphere and SOC sequestration.

2.4.1 Factors contributing to SOC loss

Soil is very responsive to different land uses and management practices (Swift, 2001; Ugalde *et al*. 2007; Luo *et al.* 2010; Powlson *et al.* 2011; Cotching *et al.* 2013; Davy and Koen, 2013). The SOC can be lost due to environmental factors, land management, and from temporal and spatial fluctuations resulting from climate variability and soil properties (Badgery *et al*. 2013; Gray *et al*. 2015). The actual mechanisms of SOC loss include mineralisation of SOC, soil erosion, and leaching (Lal, 2001).

Mineralisation of SOC occurs due to heterotrophic respiration which in turn is influenced by litter, soil moisture and temperature, land management practices, and a range of physical, chemical and biological factors. (Paustian *et al.* 1997; Gifford, 1994). Higher rates of C loss

occur following land use conversion in the tropics than from temperate regions (Lal, 2001). This is because where there are high levels of net primary productivity, high rates of litter fall occur. The litter becomes the fuel for heterotrophic organisms, leading to increased heterotrophic respiration and thus higher decomposition rates (Gifford, 1994).

The labile pool of SOC is particularly susceptible to mineralisation, and consequently management changes need to be implemented that either lead to a reduction in the rate of SOM decomposition or increase NPP to enhance the production of SOM (Post and Kwon, 2000; Sanderman *et al*. 2010). Lal (2001) suggests that conversion of natural to agricultural ecosystems can result in SOC losses of approximately 30-50%. However, through increasing the humic fraction which is a more stable pool of SOC, it is possible to slow the decomposition rate. This can be achieved by implementing practices that enhance the physical protection of C in soil through promoting aggregation, enhancing soil biodiversity, changing the vegetation type, and the transfer of humic substances (HS) down through the soil profile (Lal, 2004a). However, other factors including climate and parent material may limit the effectiveness of LUC methods on a temporal or spatial scale.

The SOC losses can be induced by soil aggregate disruption (Rovira and Greacen, 1957; Craswell and Waring, 1972), and therefore cultivation plays a major role in SOC loss. Soil subject to conventional tillage (CT) methods such as those associated with cropping systems has been found to have relatively low levels of SOC. This is due to the disruption of soil aggregates during the tillage process causing changes to soil temperature, moisture, raindrop impact, and aeration, all of which expose additional SOM that otherwise would be inaccessible to decomposition processes, and thereby resulting in higher rates of SOC loss (Tisdall and Oades, 1982; Six *et al*. 1998b; Six *et al*. 1999; Six *et al. 2000*). Post and Kwon, 2000 demonstrate that up to 50 % of SOC is lost from the top 20 cm layer of soils that have been cultivated for up to 50 years. Six *et al.* 1998a found that soil subjected to no-till (NT) had higher levels of SOC than CT soils, but that native soil can have even higher values. Conversion from forest to agriculture can lead to a reduction in SOC by up to 50% due to soil disturbance (Lal, 2005).

In the Australian environment, Dalal and Mayer (1986) found that tillage for cereal cropping operations led to significant losses of SOC, and that a new low equilibrium level of SOC could be rapidly achieved (in one case 1.2%± 0.3 per year). Valzano *et al*. (2005) assessed the effect of tillage management on SCS in Australian sites. They found that SCS was lowest in soil where conventional tillage operations associated with disc ploughing and stubble burning were undertaken when compared with a range of other management activities including no-till and grazing.

Dust storms, which are a common environmental phenomena associated with soil erosion in Australia can be either a source or sink of organic material. Suspended dust in Australia has been shown to contain on average 31% (between 2 and 90%) organic matter (Boon, 1998).

Removal of dust with organic matter from the rural landscape results in productivity losses and environmental degradation. Following entrainment, the dust is deposited elsewhere; often in the oceans. Deposited dusts have been found to contain an average of approximately 34% (range 3-91%) organic matter (Boon, 1998). In estimating the impact of dust storms in terms of economic loss, Gifford (2010) estimated that during a dust storm event in Eastern Australia in 2002 approximately 0.7 Mg C was suspended at any one time. The suspended dust travelled seaward to the east of Australia and therefore was destined to be a net loss of C from terrestrial Australia.

2.4.2 The process of soil organic carbon sequestration

There has been much research in recent years to explore the role soil has in sequestering C as a strategy to mitigate climate change (Schimel *et al*. 1994; Swift, 2001; Janzen, 2004;West and Six, 2007; Lal, 2008a; Baldock *et al.* 2012). SOC sequestration refers to the long term removal of atmospheric CO_2 into the soil through the assimilation of C into the AGB and its subsequent accumulation as SOM. In terms of climate change mitigation, sequestration of C occurs if LUC results in a net increase of atmospheric CO_2 being sequestered in the soil or biomass (Powlson *et al.* 2011).

Sequestration of atmospheric CO_2 into the soil begins with the process of photosynthesis. Photosynthesis is described by Berg and Laskowski (2005) as being the source of almost all organic matter on earth. Plants use photosynthesis to convert CO_2 light energy and water into carbohydrates and oxygen (Equation 1).

$$6CO_2 + 6H_2O \rightarrow \text{light} \rightarrow C_6H_{12}O_6 + 6O_2$$

Equation 1

In photosynthesis the inorganic C from atmospheric CO_2 is fixed by autotrophic organisms into organic compounds. As the plants assimilate the C it is used to synthesize carbohydrates into the structural components of plants. These components form SOM. The SOM is mainly composed of a heterogeneous mix of plant material, animal and microbial residues at various stages of decomposition and humus (Skjemstad *et al*. 1998; Baldock and Skjemstad, 2000; Kogel-Knabner, 2000; Swift, 2001; Clapp *et al*. 2005). It includes material from both living and non-living pools and is the product of microbially and chemically transformed debris (Skjemstad *et al*. 1998) consisting of organic compounds containing C, H, O, N, S and P. The C fraction of SOM generally comprises between 48-60% of the total weight of SOM (Rosell, *et al*. 2001).

To determine the quantity of SOC in SOM it is necessary to identify all the components of SOM which can be a time consuming and expensive process. Consequently a number of conversion factors have been developed to determine the SOC content of SOM. The most commonly used conversion factor is 1.724 although other conversion factors which vary from between 1.4 and 3.3 can be used depending on soil and vegetation types (Broadbent, 1953; Nelson and Sommers 1996; Baldock and Skjemstad, 1999; Rosell *et al.* 2001). Because of the difficulty associated

with measuring SOM, the convention is to measure and report the SOC as a concentration or SCS.

Assimilation of C by photosynthetic organisms is expressed as gross primary production (GPP). Net primary production (NPP) is calculated as the difference between GPP and the loss of C from plant respiration and represents amount of OC present in plants (Roxborough *et al.* 2005). The proportion of daily plant assimilated carbohydrates lost as CO_2 to the atmosphere due to plant respiration is estimated to be in the range of 25% and 70% (Lambers *et al*. 2005). This rate of loss equates to approximately 50-60 Gt C $year^{-1}$ globally (Gifford, 2003). The quantity lost depends on a number of variables including environmental influences, temperature, energy demand, substrate availability, CO_2 concentration and oxygen supply. These variables influence plant respiration rates and therefore the ratio of respiration to photosynthesis. For example, Gifford (2003) shows that different plant species, grown in the same media in a uniform controlled environment have different respiration rates. Further, Mackey *et al*. (2013) show that as forests age their capacity to sequester C declines as their growth rate declines relative to their respiration rate. Thus over time forests evolve from being a sink of atmospheric CO_2 to achieving a state of equilibrium of SCS.

For SOC sequestration to occur the rate of C inputs into the soil must exceed the losses of C as CO_2 from the soil. SOM contains C but the quantity of C within soil depends upon the balance between plant inputs and losses (Jobbagy and Jackson, 2000; Paustian *et al.* 1997). Increases and losses of SOC depend upon NPP of plants and the biological activity of animals. These processes are influenced by climate and soil texture. Humidity contributes to the plant productivity and decomposition rates with higher rainfall resulting in higher rates of both than found in low rainfall areas, whereas lower temperatures can slow decomposition and NPP (Jobbagy and Jackson, 2000; Badgery *et al.* 2013).

A simple Equilibrium Model proposed by Lal (2001) to articulate the complex processes of SOC sequestration is shown at Equation 2.

$$\Delta \text{ SOC} = \text{gains (photosynthesis, plant biomass)} - \text{losses (decomposition, mineralisation)}$$

Equation 2

The equilibrium model demonstrates that when SOC gains exceed losses then SOC is sequestered and when SOC losses exceed gains there will be depletion in SOC concentration. Effectively soils become a sink for atmospheric C when land management practices or environmental factors allow an increase in SOC. Soils can become a source of atmospheric C when OC is lost through decomposition or mineralisation and is transferred to the atmosphere (Powlson *et al.* 2011; Mackey *et al.* 2013).

Another mechanism that can lead to SOC sequestration is the transport and relocation of material that occurs during erosive processes. For example, the alluvial, fluvial and colluvial

movement of material down a hillslope can lead to nutrient enrichment and deposition of SOM low in a soil catena (Rosenbloom *et al*. 2001; Hancock *et al*. 2010). Additionally, SOC can be deposited or accumulated via the process of illuviation or eluviation where material such as colloidal particles are transported in suspension. Such movement is associated with particles on a soil textural basis (Hancock *et al.* 2010). For example Rosenbloom *et al*. (2001) found that approximately 40% of transported SOC is associated with clay, 15% with sand, 25% with silt sized particles. Those authors determined that the remaining SOM is not associated with colloids, but is instead transported with surface litter that moves across the surface. Soil texture is important for SOC storage because it influences the movement of SOC into the stable (passive) pool (Schimel *et al*. 1994). ^{137}Cs analysis can be used to calculate particle movement and erosion rates because the ^{137}Cs is adsorbed onto fine soil particles. The transportation of SOM is effected by terrain and hillslope curvature leading soil depth and SOC heterogeneity due to their associated effects on SOC storage (Yoo *et al*. 2006). The geomorphic processes of erosion and deposition can have a confounding influence on the biological processes of photosynthesis and decomposition with regard to SOC sequestration particularly in relation to the effects of topography, erosion, deposition and the spatial distribution of vegetation.

2.4.3 Carbon sequestration potential of soil

In the first part of this section, different forms of land management and their effects on the potential for SOC sequestration are discussed. In the second part, the role of soil mineralogy in SOC sequestration is reviewed.

Land management systems

The Kyoto Protocol directs Annex I Parties to implement policies and measures to promote sustainable forms of agriculture and to protect and enhance sinks and reservoirs of GHG to manage and reduce GHG emissions. The Kyoto Protocol also requires Parties to meet their commitments by managing changes to emissions arising from anthropogenic LUC and forestry (LULUCF) activities. The Protocol recognises afforestation and reforestation activities for C credits, but does not make the same provision for SOC sequestration. Soil has considerable potential to sequester C. For example, Houghton (2010) has estimated that between 108 – 188 Gt C has been released from soil to the atmosphere as a result of LULUC in the period between 1850 to 2000. Soil with depleted stocks of soil C have reached a low equilibrium and thus are a potential sink of atmospheric CO_2 should adoption of contemporary land management practices aimed at sequestering SOC be implemented (Lal, 2004a).

The SOC does not always increase following LUC. Rather the actual response as either a net loss or gain, can depend upon the previous land use, management, background SOC level, climate, soil properties (clay content and mineralogy), soil depth, vegetation type, soil moisture content, microbial activity and nutrient availability (Batjes, 1998; Schimel *et al.* 1994; Lal, 2008b; Luo *et al.* 2010; Kirkby *et al.* 2011; Gray *et al*, 2015). The balance between NPP, litter

inputs and decomposition rates also contributes to the quantity of SOC in soil (Gifford, 1994; Grace *et al*. 2006; Roxburgh *et al*. 2005).

It is possible to increase the potential to sequester SOC by adopting contemporary agricultural practices such as reduced tillage, stubble retention, introduction of perennial vegetation (including afforestation and adopting an enterprise change from crop to pasture), change vegetation species with the aim of increasing the below ground allocation of biomass, and the use of manure (Batjes, 1998; Lal, 2004a,b). Important increases in sequestration of C can occur in soils that have a high clay content because in this situation there is potential to sequester C in the mineralogically stabilised and humic fractions of SOM (Batjes, 1998). Slattery and Surupaneni (2002) estimated that by increasing the amount of SOM by 10 g C kg^{-1} on 40 M ha of Australian agricultural land in the 0-10 cm soil layer over a 30 year time frame could result in SOC sequestration of 11 Mt of C per year $^{-1}$.

In a meta-analysis into land management induced changes to SOC, Guo and Gifford (2002) found that SCS increased approximately 8% on average when native forests were converted to pasture and up to 18% on conversion of land-use from crop to plantation forest. Conversely they found that up to 10% soil C was lost on conversion from pasture to plantation forest. Conversion of cropland to pastures and refforestation of farmland are LUC methods that can be used to control degradation, increase SCS and sequester C in the AGB (Paul *et al*. 2002a; Sanderman *et al*. 2010; Powlson *et al*. 2011) although the results for soil are variable.

Other agricultural management systems can influence the SCS. For example, West Australian pasture soils responding to application of P fertiliser were found to have accumulated C concentrations at an average rate of 284 ppm per year in the top 12.5cm of Coolup sand depending on the native condition. There were also concomitant increases in N, S and P (Barrow, 1969). Orgill *et al*. (2013) also found that fertiliser use may increase SCS in perennial pastures, but a better SOC sequestration outcome is achieved when the baseline SCS is low. In a study of Victorian soils Robertson *et al*. (2016) found that the application of fertiliser did not significantly increase SCS but instead identified climate as the most important factor influencing SCS. Eyles *et al*. (2015) conducted a review and found that in Australian temperate dryland pastures the application of P may increase SCS but high application rates and persistent application over time was necessary to increase SOM.

Livestock grazing management methods may lead to changes in SCS. However, the trajectory of the change is variable (Eyles *et al*. 2015). For example, continuous grazing management was found to be associated with higher SCS than rotational grazing management practices in southern NSW by Orgill *et al*. (2013), However, Cowie *et al*. (2013) found the difference in SOC between rotationally grazed and continuously grazed sites was not significant, although the authors found a trend suggesting rotational grazing showed higher SCS than continuously grazed pastures on vertisol soils in NSW.

Conversion to no-till farming is most likely to result in a SCS gain only if the previous land use practice involved tillage. But if the land had only recently been converted from grassland, it is likely that C will be lost, at least initially (Sanderman *et al*. 2010; Chan *et al*. 2010; Janzen, 2015). At the very least conversion to no-till can lead to a slowing of SOC loss. However, to effectively achieve SOC gains it is necessary to convert land-use from crop to pastures where disturbance by tillage can be almost eliminated (Page *et al*. 2013).

In the next section the influence of mineralogy on the C sequestration potential of soil is reviewed.

Mineralogy

The C sequestration potential of soil is influenced by the mineralogical compostion of the soil (Churchman and Lowe (2012). For example, sandy soil has a lower capacity to protect SOM from biological attach than clay rich soil and consequently the SOC content between the two soil types can be quite different (Baldock and Broos, 2012). Between 52% and 98% of SOC occurs in clay-organic complexes. Hence, associations between SOM and minerals in the soil are likely to exert control over the uptake and retention of SOM (Churchman, 2010; Churchman and Lowe, 2012).

Baldock and Broos (2012) suggest there are four important characteristics of soil that are involved in the protection of SOC from biological attack. The characteristics include the chemical composition of the mineralogy, the presence of complexes between organic molecules and multivalent cations, adsorption of SOM in relation to soil surface area and the capacity to provide physical protection to SOM from biological attack.

Mechanisms that protect organic material against biological attack and thus influence the SOM content of soil include the type of organo-mineral complexes present and their associations (Chenu and Plante, 2006; Baldock and Broos, 2012). SOC can be associated with almost any secondary mineral and altered primary phase, particularly in poorly crystalline oxides and silicates of Fe and Al if they are present in sufficient concentrations. Further, allophane (a poorly crystalline hydrous aluminosilicate) and ferrihydrite (a poorly crystalline oxyhydroxide mineral) have been found to stabilise SOC effectively most likely due to their high surface areas and due to exchange of ligand with hydroxyl groups (Churchman, 2010; Churchman and Lowe, 2012). The content of Fe oxides can be a useful predictor of C content because mineral associated SOM is typically bound to Fe oxides (Churchman and Lowe, 2012).

Sorption of SOM to mineral surfaces is another important process that allows sequestration of atmospheric CO_2 in soil. However, sequestration of C in soil is constrained by the availability of particles with sufficient soil surface area (SSA) to facilitate sorption of organic matter (Kaiser and Guggenberger, 2003). These authors found that the SSA decreased with increases in sorption of SOM due to sorption taking place close to the opening of the micropores rather than within the micropores. They found that sorption only occurred on certain reactive sites on

mineral surfaces rather than as a continuous coating across the whole surface. Kleber *et al.* (2007) suggest that sorption of SOM occurs on illite at sites where amphoteric Al-OH groups are exposed and also on Fe oxyhydroxides containing highly reactive OH groups. The authors also found that proportion of SSA covered by SOM was higher in topsoils than subsoils. They determined that the cause of this relationship was associated with the amount of OM in the overlying horizon.

During the processes leading to the breakdown of SOM into smaller particles through decomposition the organic material becomes associated with mineral surfaces and is ultimately incorporated into aggregates (Tisdall and Oades, 1982; Lehmann and Kleber, 2015). Aggregation involving microaggregates and primary particles is important both for improving soil structure and productivity as well as being important for sequestering SOC. Aggregation is particularly important for C sequestration because SOM located within aggregates or adsorbed onto the surface of soil minerals is both beyond the reach of microorganisms and their enzymes. The SOM is thus protected from physical disturbance and mineralisation (Chenu and Plante, 2006; Lehman *et al*. 2007). Clay is fundamental to the process of aggregate formation and stability. Lehman *et al.* (2007), used electron microscopy to establish that microbial metabolites were strongly adsorbed onto clay surfaces and formed an important nuclei for subsequent stable microaggregate formation and therefore showed how this interaction is important for protecting SOM. Thus, the greater the proportion of clay in soil, the potential to sequester SOC within stable aggregates is increased.

2.4.4 Soil organic carbon equilibrium

An undisturbed natural ecosystem that has had minimal disturbance over time, and therefore has stable C levels is said to be at a steady state or to have reached an equilibrium. In this case SOC inputs are effectively in balance with outputs (Janzen, 2015). The potential of an ecosystem in equilibrium to sequester additional atmospheric CO_2 is minimal unless a land management practice which aims to increase SOM is implemented (Oades, 1988; Janzen *et al*. 1998b; Gulde *et al*. 2008). A lower steady state of SOC can be achieved when an ecosystem has experienced an intervention of some type. The loss can occur naturally such as occurs infrequently as natural fire, seasonally and with climate variability, or it can occur due to anthropogenic intervention such as through tillage, land clearing or deforestation (Johnson, 1995). Loss can also be reversed when SOC is sequestered.

A conceptual model for the loss and accumulation of SOM can indicate that it is possible to partially restore SOM (A) (Figure 2-2), accumulate SOM to the original level (B) or accumulate SOM in excess of its original level (C) (Johnson, 1995).The original steady state represents the SOM concentration of an undisturbed landscape.

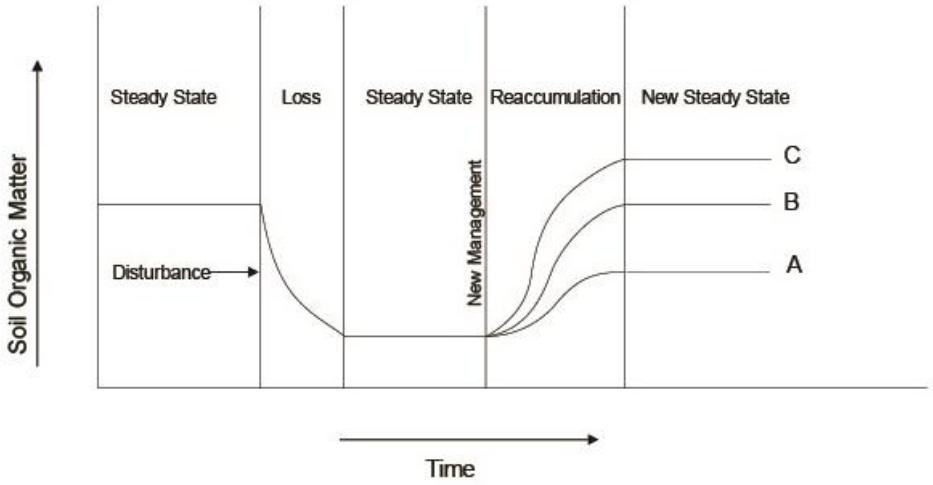


Figure 2-2. Conceptual model of soil organic matter accumulation (after Johnson, 1995).

2.4.5 Management practices causing changes to SOM

Following a management disturbance to the soil such as cultivation there is likely to be a loss of SOM. Initially SOM loss is rapid because decomposition exceeds the rate of NPP. Over time (periods of several decades) the rate of loss slows until a new equilibrium is reached (Oades *et al.* 1988; Johnson, 1995). When the new SOM equilibrium is achieved accumulation and losses of SOM assume balance because the rate of decomposition and NPP are approximately equal. However, the new steady state will most likely be at a lower level of SOM than the original state. This is due to the irreplaceable loss of mineral soil and associated SOC that is removed when all or part of the A-horizon erodes from the landscape via wind or water processes.

Studies show that following adoption of a new land management practice aimed at increasing SOM, increases in SOC can occur in the first few years, and then reach an equilibrium state (Oades *et al.* 1988; Johnson, 1995). Following changes to management it is also possible for agricultural soil to sequester SOC to levels exceeding the natural ecosystem (Johnson, 1995). SCS increase can be achieved by changing tillage regimens (e.g. from conventional to no-till), plant type (e.g. from annual to perennial). Other practices include restoration of degraded agricultural land by establishment of permanent pasture and reforestation, changes to vegetation type, and increased nutrient availability. Increased rates of SOC sequestration can be achieved if an external source of C is applied such as organic amendments, manure or fertiliser (Oades *et al*. 1988; Eyles *et al*. 2015).

Sequestration of C in soil does not occur linearly nor indefinitely. Indeed, the sequestration capacity of soil is constrained by factors such as rainfall, temperature, mineralogy and nutrient availability (Oades, 1988; Stockman *et al*. 2013). Similarly, the rate of SOC sequestration can vary spatially and temporally, fluctuate due to seasonal and climatic conditions, parent material,

soil biology, nutrient availability, plant species, residue management, and land management practices. Oades (1988) suggests that the average turnover time of SOC is approximately 30-40 years but the rate is dependent on a number of variables, including climate (particularly precipitation and temperature), topography, soil parent material and the chemical composition of the organic matter. SOC sequestration in soil with high baseline SCS is difficult and C loss can occur when soil is disturbed such as occurs with LUC. For example Guo and Gifford (2002) found that conversion of land previously used for pasture to plantation, pasture to crop, native forest to crop or plantation generally led to a decline in SCS. However, opportunities to sequester C in soil exist, and SOC sequestration can occur following LUC. For example Guo and Gifford. (2002) found that conversion of agricultural land previously used for cropping to pasture on average increased SCS by approximately 19%. They also found that conversion of land from native forest to pasture, crop to plantation forest and crop to secondary forest generally lead to SCS increases between 8 – 53% on average. Two additional opportunities to sequester SOC via LUC, i.e. converting pastures to BEPs and the use of waterponding, are considered in this thesis.

2.5 Chronosequence approach

The current research project has utilised a paired site chronosequence study design for the two case studies as longitudinally obtained repeated measurements of the sites has not been possible. The use of chronosequences in soil research is a useful approach that provides an opportunity to assess temporal changes to soil properties when studying the effects of LUC where repeated measurements at a single site over a long time period has not been possible (Breuer *et al*. 2006; Walker *et al*. 2010; Davies and Gray, 2015). The use of chronosequences in research can be described as substituting space for time (Johnson and Miyanishi (2008); Walker *et al*. 2010). A shortcoming of the use of chronosequences to study changes to, for example soil following LUC, is that the study sites can at best only have similar site characteristics and as a result, inferences from chronosequence data can be incorrect (Johnson and Miyanishi (2008). The use of chronosequences can be deemed to be pseudoreplication whereby site and or stand treatments may confounded statistical analysis, and that as a results it may be inappropriate to infer that treatments have actually caused an effect (Hurlbert, 2009; Hurlbert, 2013; Davies and Gray, 2015). Therefore, in the case of the current research, differences in data can only be assumed to be associated with LUC, when other features not examined including geological or ecological characteristics may have had a greater influence on the results.

Notwithstanding the shortcomings of chronosequences, they are useful when studying soil changes over time (Walker *et al*. 2010) and there are many examples of studies that have utilised a chronosequence for assessing changes to SOC following land use change; particularly afforestation, e.g., Turner and Lambert (2000) and Vesterdal *et al*. (2002). In these studies the researchers analysed SOC from forests of different tree species and ages in an attempt to

associate change with time. Whilst the studies identified changes to SOC, the influence of site specific factors such as previous land use, site heterogeneity, parent material differences, climate, soil attributes, and topographic variation, amongst others, on SOC were found to be difficult to ascertain, and may have affected research findings (Walker *et al*. (2010). Walker *et al*. (2010) points out that using sites with similar soil attributes is particularly important when very young and very old sites are included in a chronosequence study. For this reason it is useful to undertake repeated sampling over subsequent years to verify research findings. When identifying sites for a chronosequence study in soil research Walker *et al*. (2010) suggests there are five key issues that need to be addressed during site selection. i. evidence of obvious changes to site attributes over time; ii. similar site histories, iii. replication of sites for each age class, iv. obtaining site specific baseline measurements for comparison with repeated measurements over time and v. using standard methods will insure replication of sampling can be undertaken effectively allowing meaningful comparison of results to be achieved.

2.6 Summary

Soil organic matter has a number of important functions including the improvement of soil structure, mitigation of soil erosion, nutrient cycling and sequestration of atmospheric CO_2 into the soil. This review has concentrated on SOC sequestration, and identified several strategies that can be used to improve the C sequestration potential of soil. For example conversion of tillage activities to no till cultivation techniques has been shown to increase SOC in some cases, although there are some conflicting reports. Reforestation, also has been shown to increase soil C in most cases, but not all. This review has provided evidence that SOC sequestration is complex with a number of cycles and pathways involved. Specifically, the chemical properties of the soil, the nature and chemical properties of the OM produced by the vegetation, aggregation, the clay content in soil, disturbances to the soil, climate, topography, C saturation levels and parent material all play a role in determining whether, and how much C, can be sequestered in a particular soil.

This research project explores two land-use change scenarios that are aimed at improving vegetation cover and agricultural productivity to determine their C sequestration potential. In the next chapter methods used are discussed, and then in two chapters two case studies are presented. Case study 1 in Chapter 4, explores the C sequestration potential of agricultural soil in the temperate Southern Tablelands of NSW following establishment of BEPs and further investigates the relationship between AGB growth and SOC sequestration. The second case study in Chapter 5 examines C sequestration in scalded clay pan soils located in the semi-arid rangelands of NSW that have been rehabilitated using a waterponding technique.

Chapter 3: Methods

3.1 Introduction

This research project is designed to explore changes to SOC over time following two LUC activities. Case Study 1 explores changes to SOC following establishment of BEPs and Case Study 2 explores SOC changes following construction of waterponds. Due to the different funding sources and resources available for the two case studies, some laboratory analysis and methods were unique to each project. In some instances analysis was conducted only on a subset of samples due to research resource constraints. The study designs, selection of sampling subsets and methods of analysis are described in detail in the relevant case study chapters.

This chapter provides details of the following laboratory analysis methods that were common to both case studies:

- Soil coring methods;
- Soil bulk density (BD) calculation;
- The pH and EC determination;
- SOC analysis using the Heanes (1984) method;
- SCS calculations;
- Equivalent soil mass (ESM) calculation; and
- Rate of SCS change.

3.1.1 Soil coring

There are several methods that can be used for obtaining soil samples for soil analysis and for determining soil bulk density including coring, the clod methods (McKenzie and Cresswell (2002). Despite known systematic errors involved in extracting soil samples for soil bulk density using a corer, it is the most common method adopted for sample acquisition for the purpose of SCS accounting (Lal and Kimble, 2001; Sanderman *et al*, 2010; Carbon Credits (Carbon Farming Initiative) (Sequestering Carbon in Soils in Grazing Systems) Methodology Determination 2014). Errors in bulk density calculations may arise due to the soil texture, presence of stones in the extracted sample and the soil moisture content at the time of coring (Reynolds, 1970). The coring method involves driving a steel core of a known volume, vertically into the soil profile to extract a soil sample. Because the volume of the core is known it is possible to calculate the soil bulk density (paragraph 3.1.2) and the SCS (paragraph 3.2.4). Potential errors in soil bulk density calculations can occur due to compaction during sampling, presence of gravel and stones, wet or cracking soils (Lal and Kimble, 2001). Sampling was undertaken during winter and spring to ensure samples had adequate consistence, thereby increasing the potential for obtaining an accurate core length.

To minimise the potential of measurement errors occurring during sampling, care was taken to ensure the possibility of extracting a compacted sample was slight by checking the depth of the core hole and comparing it with the length of the soil sample. Where there was a discrepancy the core was rejected. The diameter of the soil core was not checked against the diameter of the core cylinder, although this would have provided greater certainty that compaction of the soil sample during coring had not occurred. Additionally, three replicates for each depth increment at each position at each site were obtained and composited to improve the accuracy of estimating the mean bulk density and SCS.

The collection of soil samples for both case studies was undertaken by soil coring using either a mechanical or manual soil corer. Cores were collected in three ways depending on resource availability and site access as follows:

i. **Case Study 1 (BEPs) mechanical sampling**. A truck mounted "Dingo Ezi Probe" hydraulic soil corer was used to obtain soil cores where vehicular access within BEPs was possible. Each core was obtained in a PVC sheath with an internal diameter of 42 mm, to a depth of 1.0 m (Figure 3-1). The sheath was sealed at each end with a rubber cap to prevent moisture loss in transit. On return to the laboratory the sheath was split open and samples obtained at 0-5 cm, 5-10 cm, 10-20 cm, 20-30 cm and subsequent 10cm increments. Sub samples were obtained for BD analysis before the samples were dried for storage prior to grinding and analysis.

Figure 3-1. Where access was possible within BEPs, soil cores were collected using a truck mounted Dingo Ezi Probe mechanical corer by a PVC sheath.

ii. **Case Study 1 (BEPs) manual sampling.** Where vehicular access was not possible within BEPs, a manual steel soil corer with an internal diameter of 50 mm (Figure 3-2) was used. The steel core had graduations etched at 5cm, 10cm, 20 cm, 30cm and 40cm to gauge depth intervals. A rubber mallet was used to drive the core into the soil to obtain samples from the 0-5 cm and 5 -10 cm increments. For the two deeper increments a steel shaft was placed over the corer and a star stake driver was used to drive the core to the 20cm and 30cm depths. After extraction of the

sample each hole was measured to back check the hole depth and corroborate the core sample size. Cores were taken at individual increments from the same core hole between 0-5 cm, 5-10 cm, 10-20 cm and 20-30 cm. Each sample was stored in a labelled plastic bag. Three replicates were obtained for each sample and depth increment and were bulked together accordingly.

Figure 3-2. Method used to manually extract a soil core within BEPs (photo: C. Vuillot).

iii. **Case Study 2 (Waterponds) mechanical coring.** The waterponds were fully accessible to vehicles, consequently a truck mounted mechanical corer (brand unknown) was used to obtain soil samples to a depth of 40cm.. The core had a cutting head of 40 mm and the core itself was slightly larger to prevent lodging of the core. The steel corer equipment required the soil to be gently pushed through the steel sheath to gain access to the sample. This procedure was done carefully to ensure minimal disruption to the integrity of the sample and minimise sample compaction to enable accurate measurement of the soil BD. The depth of the hole was measured and correlated with the length of the soil sample to ensure the core had not been compressed during the process. The core was sliced into depth increments 0-5 cm, 5-10 cm, 10-20 cm, 20-30 cm, bulked and packaged in an appropriately labelled plastic bag in the field for transport back to the laboratory. Three replicates were obtained for each position and bulked accordingly.

Figure 3-3. The soil sample was gently pushed out of the steel sheath to allow slicing the sample into depth increments

3.1.2 Bulk Density

Although BD is a fundamental measurement in soil science used to calculate soil CS, obtaining accurate BD results can be difficult due to a number of factors including soil compaction during core sampling, differences in soil texture, the presence of rock fragments, abundance and type of vegetation and SOM, soil compaction from vehicles or livestock, and moisture content particularly in shrink-swell soils (Berndt and Coughlan 1976; Manrique and Jones 1991; Vincent and Chadwick, 1994; Lal and Kimble 2001; Post *et al*. 2001; Sanderman *et al*. 2010; Mehler *et al*. 2014).

On return to the laboratory each sample was weighed to obtain a field wet weight. A subsample of approximately 20g was obtained and placed into a pre-weighed tin. The tin and wet subsample were weighed and then dried in an oven at 105ºC for 24 hours to determine the gravimetric moisture content and BD. The oven dried sample was then weighed and sieved to pass through a sieve with 2 mm diameter mesh. The sample that passed through the sieve was taken to be the fine soil fraction. The materiel remaining in the sieve was taken to be gravel. The > 2 mm gravel fraction was weighed to determine the gravel content of the soil sample.

The fine fraction soil BD was calculated as Mg m^{-3} using Equation 3.

$$\text{BD (Mg m}^{-3}\text{)} = \text{Mass of oven dry soil (g) / sample volume (cm}^{3}\text{)}$$

Equation 3

3.1.3 Soil preparation

Following collection of the subsample for BD, the remainder of each sample was air dried for two weeks at ambient room temperature. Any large clods were gently broken down by hand prior to drying. Once the sample was dry, a subsample of approximately half was separated using a quartering method (Rayment and Lyons, 2011:11). Gravel was removed from the subsample which was then ground with a mortar and pestle to pass through a 2mm sieve and stored in an airtight plastic container. The remaining composite unsieved samples were stored in labelled plastic bags.

Approximately 60g of the 2 mm sieved sample was separated using the quartering method and ground with a mortar and pestle to pass through a 500 µ sieve. This subsample was stored in an appropriately labelled, separate airtight plastic container.

3.2 Laboratory methods common to both case studies

The method of analysis used for pH and electrical conductivity (EC), SOC, and calculations for SCS, ESM and rate of SCS change were common to both case studies. Details of the methods used for each analysis follow.

3.2.1 Soil pH and Electrical Conductivity (EC)

The pH and EC for every sample was measured using a 1:5 soil/water extract method (methods 3A1 and 4A1) described by Rayment and Lyons (2011). Briefly, an 8g subsample of the sieved < 2 mm sample was weighed into a 50 ml falcon tube. 40 ml of Mili-Q water was added to the suspension and shaken for 1 hour. The suspension was allowed to settle for approximately 1 hour. Measurements were obtained by placing a calibrated Orion Pacific 5 Star meter probe into the suspension until a stable pH and EC result was displayed. The EC probe used is identified as 013010MD and the pH thermos triode identified as 9107WMMD.

3.2.2 Soil organic carbon concentration analysis using the Heanes, (1984) method

All samples, from both Case Studies were analysed for SOC using the Heanes wet oxidation method (Heanes, 1984). The method has been categorised by Rayment and Lyons (2011) as method code 6B1.

Between 0.25 and 1.0g of finely ground soil (sieved < 500µ) was placed in a 75ml test tube. Sample size was determined on the expected C content. A smaller aliquot was used for samples expected to have a high OC concentration; typically this was the 0-5 cm depth increment.

Next, 6ml of 8% potassium dichromate ($K_2Cr_2O_7$) was added to the test tube and the solution gently swirled. Then 12ml of concentrated sulfuric acid (H_2SO_4) was slowly added and the

solution gently swirled. The test tubes were placed on a digestion block preheated to 135ºC and left for 30 minutes. At 15 minutes the samples were swirled gently. On completion of the digestion process the tubes were removed from the block and placed in cooling racks. Once the sample was cool a flush of Milli-Q water was added and the tubes were then allowed to cool again for approximately 1 hour. A further quantity of Milli-Q water was added to the tube to bring the solution up to the 50ml mark. The tubes were then stoppered and shaken. This same process was used for a set of glucose standard solutions.

The samples were left to settle overnight. 35ml of the sample was poured into a 50ml falcon tube and then centrifuged at 2000 rpm for 10 minutes. Each sample was poured into a cuvette and analysed on a Cary 50 Spectrophotometer with the wavelength set to 625nm. Results were determined on an oven dry soil weight basis.

3.2.3 Total soil carbon using Leco CNS2000 analyser

Analysis for TSC was conducted on all 540 waterpond samples using a Leco CNS2000 dry combustion analyser at the Environmental Analysis Laboratory, Southern Cross University, NSW Australia.

TSC was also conducted on a subset of 96 BEP samples (Figure 4-7) by the University of Melbourne using a Leco CNS2000 dry combustion analyser.

The method used in both case studies was the Dumus high-temperature combustion method 6B2a for TSC (Rayment and Lyons, 2011).

3.2.4 Soil carbon stock

The SCS calculated for every soil sample was calculated on a fixed depth basis using soil BD, depth and SOC concentration results. The following formula to calculate SCS on a fixed depth basis was applied (Hazelton and Murphy, 2007).

$$\text{SCS (Mg ha}^{-1}\text{)} =$$

$$[\text{SOC (\%)} * \text{BD (Mg m}^{-3}\text{)} \times \text{layer depth (m)} \times 10^4 \text{ (m}^2 \text{ ha}^{-1}\text{)}]$$

Equation 4

The soil carbon stock was corrected for gravel content as follows:

Where gravel = 0, the gravel correction = 1.00

Where gravel content > 0, the gravel correction used = 1 – (gravel%)/100

Equation 5

A worked example for calculating a gravel corrected SCS is at Table 3-1

Table 3-1. Worked example for calculating SCS incorporating a gravel correction

Soil layer (cm)	SOC (%)	BD (Mg m^{-3})	Soil mass (layer) (Mg C ha^{-1})	SCS (Mg ha^{-1})	Gravel correction	Gravel corrected SCS for soil layer (Mg C ha^{-1})
5	1.3	1.5	770	10.2	0.9	9.6

3.2.5 Equivalent soil mass

After the SCS had been quantified on a fixed depth basis for each sample (Equation 4 and Equation 5), it was found that the SCS was exaggerated in samples having a high BD and under-represented when the bulk density was low due to the accumulation of organic matter after land use change. It is widely acknowledge that a strong relationship exists between soil bulk density and the quantity of SOM, whereby soils with high levels of SOM typically have higher porosity and hence a low bulk density (Murphy, 2015). As a consequence of this relationship, the results in Case Study 1 showed that the AL often had a higher SCS than the TL and IR position in BEPs because the bulk density was higher, and this was despite the SOC concentration being higher within the BEPs. Similarly, in Case Study 2, the scalded soil was often found to have a higher SCS than the waterponds, and again this arose even although the SOC concentration was higher in the waterponded soil. This counterintuitive bias is common, for example in tilled soil (Ellert and Bettany, 1995), or in cleared landscapes where the soil becomes compacted (Murphy *et al.* 2003).

To overcome the discrepancy a method of calculating an equivalent soil mass adapted from equations by Ellert and Bettany (1995) and Wendt and Hauser, (2013) was used to calculate the SCS for both case studies. Because the method used to pair sites in the BEP case study was different to that used in the waterpond study different methods were used to calculate the SCS on an equivalent soil mass basis. Details of the equations used can be found in the respective methods section of each case study.

3.2.6 Soil Carbon Stock change

The formula used to calculate the annual average difference in SCS is that recommended by IPCC (2006) as follows:

$$\Delta C = \frac{(Ct_2 - Ct_1)}{(t_2 - t_1)}$$

Equation 6

Where:

ΔC = annual SCS change in the pool, Mg C yr^{-1}

Ct_1 = SCS in the pool at t_1, Mg C

Ct_2 = SCS in the pool at time t_2, Mg C.

3.2.7 Relative rate of change in soil carbon stock

The formula used to calculate the relative rate of change, as a percentage, in SCS following waterponding has been calculated using the following formula (Luo, et al. 2010)

$$SCS(t_2) - SCS(t_1) / (t_2 - t_1) * SCSt_1$$

Equation 7

Where:

SCS(t1) = SCS in the scald

SCS(t2) – SCS in the waterpond at waterpond age

3.3 Laboratory methods discrete to each individual case study

Where laboratory analysis was conducted for samples from one case study only, the method used is described in that case study chapter. Details for the methods specifically used in the BEP Case Study are contained in Chapter 4 and for the Waterpond Case Study, Chapter 5. A summary and cross reference detailing those methods is contained in Table 3-2.

Table 3-2. Cross reference for laboratory methods described elsewhere in this thesis. Case study chapters are: Chapter 4 for BEPs and Chapter 5 for Waterponds.

Analysis	Method ID	Case study chapter	Number of samples
TP	1:3 nitric/HCl digest	Waterpond	*n* = 144
TP	Kjeldahl P	BEP	*n* = 830
TN	Kjeldahl N	BEP	*n* = 830
TN	Leco CNS2000	Waterpond	*n* = 540
AP	Olsen P (9C2b)	Waterpond	*n* = 144
TS	1:3 Nitric/HCl digest	Waterpond	*n* = 144
Aggregate stability in Water (ASWAT)	Field *et al*. 1997	Waterpond	*n* = 144
Particle size analysis (PSA)	Laser diffraction	Waterpond	*n* = 16
Mid infrared spectroscopy (MIRS)	Baldock *et al.* 2013b	BEP	*n* = 830
Nuclear magnetic resonance spectroscopy (^{13}C NMRS)	Baldock *et al.* 2013a	BEP	*n* = 12
Soil fractions	Baldock *et al*. 2013a	BEP	*n* = 830
X-ray diffraction (XRD)	Powder diffraction analysis	Waterpond	*n* = 8
Inductively Coupled Plasma-atomic emission spectrometry (ICP-AES)	ICP AES	Waterpond	*n* = 8
Cation exchange capacity (CEC)	15C1 (Rayment and Lyons, 2011)	Waterpond	*n* = 8

3.4 Statistical analysis

Analysis of variance (ANOVA) were calculated using GENSTAT 17th edition (2014) to determine statistically significant effects of changes to SOC and other soil attributes following the establishment of BEPs or the construction of waterponds. Multiple comparison analysis was undertaken using the Tukey test to compare the effects of, and interactions between site, age class and depth with SOC, BD, pH, EC, SOC, SCS and most other attributes measured. Regression analysis was used to identify changes and relationships in given soil attributes. Specific details regarding statistical analysis are contained in the respective case study chapters.

Chapter 4: Biodiverse environmental plantings and soil carbon sequestration

4.1 Introduction

Agroforestry refers to the use of trees in an agricultural context for the purpose of deriving both economic and environmental benefit. Agroforestry includes activities that are conducted in conjunction with other agricultural activities including livestock and crop production. Agroforestry can be associated with small scale subsistence farming practices, intensive agricultural systems, or large scale agricultural operations (Nair *et al*. 2009). The oldest form of agroforestry is reported to have been practiced in southeast Asian fishing communities dating between 13 000 and 9 000 BC (Nair *et al*. 2009). Today, a wide diversity of agroforestry is practiced by farming communities around the world for many applications. Agroforestry can be applied as either mono or multi species crops and there are many scales of agroforestry with activities being undertaken at a large plantation scale, small private native forests, through to woodlots and shelter belts. Torquebiau (2000) classifies agroforestry into the following six structural categories:

1. Crops under tree cover;
2. Agroforests;
3. Agroforestry in a linear arrangement;
4. Animal agroforestry;
5. Sequential agroforestry; and
6. Minor agroforestry techniques.

Establishing agroforestry on land previously used for agricultural production, including cropping and grazing, is recognised as a way to mitigate the effects of erosion, produce a supply of wood for fuel rather than depleting natural forests for this purpose and facilitate cleaner water by reducing nutrient and soil runoff. In recent years agroforestry systems such as environmental plantings have been identified as potentially providing opportunity to reduce anthropogenically caused CO_2 emissions because trees have the capacity to rapidly sequester C into their biomass and less rapidly in soil.

Whilst extensive research has been conducted on both the AGB and SCS potential of forestry, until now few studies have specifically examined changes to SOC associated with specific agroforestry activities. Thus, this case study examines the SOC sequestration potential of BEPs. The term “biodiverse environmental planting” is synonymous with other agroforestry terms used in the literature including: windbreaks (Cleugh *et al.* 2002), shelterbelts (Nuberg, 1998), mixed species environmental plantings (MSEP) (Paul *et al*. 2013; Paul *et al*. 2014), direct seeding for revegetation (Geeves *et al*. 2008), and agroforestry in a linear arrangement (Torquebiau, 2000). BEPs are a particular form of agroforestry with trees laid out in either a

linear or a block arrangement, and are typically established on farms to restore biodiversity on land that has previously been cleared of native woody vegetation to make way for grazing or cropping.

This review will describe the original concept and purpose of BEPs in terms of an LUC, but will also explore the more contemporary use of BEPs as a C sink. An overview of studies that have considered the quantity of SCS that may have been present in Australian soils prior to European settlement in Australia has also been undertaken. The global and Australian policy position on the use of agroforestry and specifically for BEPs as a way of sequestering SOC will also be examined. BEPs produce large quantities of SOM, thus, the review will discuss the changes to SCS and the associated pools of C including particulate organic carbon, (POC), humic organic carbon (HOC) and resistant carbon (ROC) fractions that might occur following afforestation. This review will also detail the development of, and examine the role and application of FullCAM, a modelling tool used by Australia for C accounting in the land sector. Finally, the review will examine the dry combustion and wet oxidation methods used to determine the OC concentration of soil and how these traditional laboratory methods compare with the more recent application of MIRS for C analysis.

4.2 Biodiverse environmental plantings - role and establishment

Much of the productive land in Australia was formerly forest and woodland which was subsequently cleared by the early settlers to make way for agricultural activities including grazing and cropping (Barson *et al*. 2000; SOE 2011). The clearing has contributed to a decline in flora and fauna across the landscape and resulted in soil degradation such as dryland salinity in some geographic areas. This has been instrumental in exacerbating soil structural decline and soil erosion processes. Consequently, there has been considerable effort in recent decades to use trees to revegetate the landscape with the objective of deriving environmental and social benefit. Revegetation of the landscape through establishment of BEPs has helped to restore biodiversity, mitigate against the effects of dryland salinity when planted in recharge and discharge areas and also reduce soil erosion. Organisations such as Landcare and GA have been actively replanting trees and shrubs into the Australian landscape (Vesk *et al*. 2008); indeed their foundation and ongoing role is based on this type of environmental restoration activity. These community based enterprises aim to rehabilitate and revegetate degraded agricultural landscapes across Australia. In so doing these organisations have promoted the benefits of BEPs and this has resulted in a widespread network of plantings being established on private agricultural land.

Reasons that landholders establish BEPs on agricultural land include: to increase livestock survival rates due to the sun and wind shelter they provide, improve farm biodiversity, provide wildlife corridors, reduce soil erosion, manage water quality by trapping pollutants, manage water tables and dryland salinity and occasionally provide a source of wood product for either

commercial or non-commercial use (Corr, 2003; Kavanagh *et al*. 2005) or biofuel feedstock (Franzluebbers, 2015). However, not all these uses are applied in the region where this case study was researched. Tree planting also provides a social service by improving the landscape amenity and can even increase agricultural land property values. The use of windbreaks to provide shelter to stock and crops has been the subject of global research for several decades (Nuberg, 1998). In 1993, a National Windbreak Program (NWP) was launched in Australia to research the role farm trees played in reducing land degradation due to wind erosion, waterlogging and dryland salinity (Cleugh *et al.* 2002).

Establishment of BEPs into agricultural land in the Australian Capital Region has been undertaken since the 1980's. GA in collaboration with private landholders establish BEPs for a number of reasons including to provide wildlife corridors, shelter for livestock and pastures and to revegetate agricultural landscapes that have been previously cleared of native trees. During this time, the methods used to establish BEPs, and the species and stocking composition has evolved to optimise the success and benefits of the plantings (Corr, 2003).

In comparison with other forms of forestry, BEPs are unique in their composition, stocking density, tree species diversity, site layout and environmental benefit. They are established as either linear or block plantings depending on landholder requirements and landscape constraints. They can be directly seeded or established by planting tube stock. Direct seeding involves the use of a single tyne implement to rip the soil and simultaneously drop seed into the broken soil. Planting tube stock refers to placing seedlings that have been grown in a nursery, into the ground. BEPs established by the direct seeding method (as is the case for all the BEPs in this study) are especially unique as they have a high stocking rate and species diversity. They are typically established to stabilise erosion, provide water table and soil salinity control or provide windbreaks. They are commonly sown along the sides of erosion gullies, hilltops, or parallel with fence lines or roads against prevailing winds. Typically linear BEPs have two to four rows of trees and are established along existing fence lines or established along newly constructed new fence lines. Block plantings comprise numerous rows and are commonly placed on degraded hilltops, slopes or erosion gullies. Every BEP has unique site characteristics in respect of elevation, slope, aspect, drainage, soil type and characteristics, tree stocking density and species mix. The stocking density and species mix is often driven by landholder preference, although climatic conditions during the establishment phase of the planting will influence the ultimate composition of each individual BEP (Geeves *et al*. 2008; Semple *et al*. 2010).

In addition to all the aforementioned benefits, the change in land use following afforestation of agricultural land with BEPs also provides opportunity for atmospheric C to be sequestered in both the tree biomass, litter layer and in the soil (Paul *et al*. 2002a,b; Cunningham *et al*. 2012; Hoogmoed *et al*. 2014; Paul *et al*. 2014; Paul *et al*. 2015). Consequently, as a result of the C

sequestration potential reforestation/afforestation has become an approved method of climate change mitigation under the provisions of the Kyoto Protocol (1998), and falls within the Australian Government's Emissions Reduction Fund (ERF) (Carbon Credits Methodology, Determination 2014). The opportunity for BEPs to be a C sink and the constraints limiting their potential are examined in this review.

4.3 Soil C stock levels prior to European settlement

To establish a possible pre-European settlement (baseline) SCS level for the Southern Tableland region of NSW where the current research was carried out, an extensive literature review has been completed. The review demonstrated that the state of vegetation on the Southern Tablelands prior to European settlement comprised *Eucalyptus* spp., forests on hillslopes, woodlands on slopes and grasslands and swamps in the valley floors (Wasson *et al*. 1998). Extensive clearing of Eucalypt woodlands occurred in south eastern Australia to make way for agriculture following European settlement with extensive soil degradation and erosion following (Yates and Hobbs, 1997). Grazing commenced around 1824 and reports of contemporary sheet erosion and deep erosion gullies emerged in 1848. By 1900 the development of incised channels in the region was almost complete. There is also evidence that between 1930 and 1940 sheet erosion on crests and slopes occurred as a result of droughts, overgrazing and rabbit plagues (Wasson *et al*. 1998). With the evidence presented by this history it is likely that there has been a substantial loss of SCS as a result of topsoil loss by erosion.

Estimates of current SCS across Australia have been undertaken by Viscarra Rossel *et al*. (2014) by using soil inventory data for the period 2000 to 2013. More recent work by Viscarra Rossel *et al*. (2015) used archived soil data to develop a three dimensional soil map for Australia showing SOC concentration on a decadal basis since 1930. Due to lack of empirical data it is more difficult to estimate the SCS that existed in Australia generally, and the Southern Tablelands specifically, prior to European settlement. Consequently the results of studies by Wong *et al*. (2008), Miklos *et al*. (2010) and Wilson *et al*. (2011) have been used to develop an estimate of a possible pre-European settlement SCS for the Southern Tablelands for the purpose of establishing an aspirational SOC sequestration target.

The study by Wilson *et al*. (2011) compared SOC on basalt, granitic and meta sediments on the Northern Tablelands of NSW. Their study included sites with land uses described as cultivation, improved pasture, unimproved pasture and woodland. These authors found that woodland environments on basaltic, granitic and meta-sediment soils had a SCS of 101.8, 45.8, and 121.5 Mg C ha^{-1} (calculated on an ESM basis) respectively. Their study also found the SCS on unimproved pastures was 62.3, 35.1 and 78.6 Mg C ha^{-1} on the same soil types respectively. The Northern Tablelands of NSW has a higher average rainfall than the Southern Tablelands. Therefore, the mean SCS results from the Wilson *et al*. (2011) study may overestimate the

baseline SCS typically found in Southern Tableland soils. Consequently, a more conservative baseline estimate indicative of a possible pre-European settlement SCS level for the Southern Tablelands (and not including the Monaro basalt provenance) has been estimated by using the mean SCS results for the unimproved pastures located on granitic and meta-sediment soils only (35.1 and 78.6 Mg C ha^{-1} respectively) (Wilson *et al*. 2011) giving an estimated pre-European SCS for the Southern Tablelands of approximately 57.0 Mg C ha^{-1}. A FullCAM simulation has been undertaken as part of this case study and in a retrospective analysis the simulation shows that the SCS in the region in 1900 was approximately 50 Mg C ha^{-1} and declining (Figure 4-61).

The estimate of 57 Mg C ha^{-1} is reasonably consistent with the findings of other researchers. For example, Gifford (2009) who in a review noted the sparsity of SCS data associated with grazing lands, estimated a mean SCS across Australia of 64 Mg C ha^{-1} (depth unspecified). However, the estimate was prefaced with the proviso that the result has a large but unknown uncertainty. In other research, using a sequence of historical aerial photographs, Wong *et al.* (2008), was able to estimate that salt scald formation commenced in approximately 1945 on their study sites located near Bevondale, NSW (located on the Southern Tablelands, approximately 100km to the north of where the current research project was conducted). Having determined the 1945 baseline, they assumed that the SCS of 55.5 Mg C ha^{-1} measured in a nearby vegetated site was representative of the stock present prior to formation of the scald.

In another study Miklos *et al*. (2010) proposed that travelling stock routes (TSRs) may be a valuable surrogate representative of little disturbed, perhaps pre-European soil conditions, and thus useful for determining baseline SCS. These authors found that the SCS in TSRs was on average 82 Mg C ha^{-1} to 1 m depth, whereas pastures were found to have a mean SCS of 68 Mg C ha^{-1}. The difference between SCS in TSRs and pastures identified by Miklos *et al*. (2010) is likely to be the result of C being eroded from the agricultural land since European settlement. An important difference between the study by Miklos *et al*. (2010) and this current research project is the applicable soil depth (1 m compared with 0.3 m respectively). However, because the quantity of SOC declines with depth, the proportion of SOC in the top 0.3m of the TSR results could well be within the range of 57 Mg C ha^{-1}.

4.4 Agroforestry as a carbon sink

There has been extensive research into the potential of forestry to be a C sink to offset the effects of GHG emissions (IPCC, 2000; Pan *et al*. 2011). The AGB of forests is a productive sink for sequestering atmospheric CO_2 (Houghton *et al*. 1998; Montagnini and Nair, 2004). Therefore, afforestation and reforestation, both of which increase AGB, are viewed as important strategies for reducing the concentration of atmospheric CO_2. This is especially the case in the context of LUC from agriculture to forest due to the relatively rapid growth rate of trees, larger woody biomass and longer turnover than can be achieved by either pastures or crops (Houghton, 2005).

The potential of agroforestry systems to sequester C can be difficult to estimate due to methodological variables associated with estimating the C stock in both AGB and soils (Nair *et al*. 2009). The estimations for AGB can be influenced by the quality of allometric equations and the appropriate application of the equations which may have been developed for natural forestry systems rather than agroforestry systems. Further, the allometric equations may have been developed using location specific data that may not be applicable to other sites. This is because variables such as management, site location, climate, stand age, species composition, stocking density and edaphic factors can influence the C sequestration potential of agroforestry systems (Cunningham *et al*. 2015). For example, Kumar *et al*. (1998) found that under and over estimations of AGB can occur as a result of population based genetic variation, environmental variation, stocking rate and density.

Nair *et al*. (2009) estimate that approximately 1023 million ha of agricultural land is being used for agroforestry production worldwide. The IPCC (2000) also estimate that a further 630 million ha of degraded agricultural land can potentially be converted for agroforestry use. The current proportion of C stock (CS) in global forests has been estimated by Pan *et al*. (2011) to be 383 Gt C in soil to 1 m depth, 363 Gt C in live AGB and below ground biomass (BGB), and 43 Gt C in litter. However, these authors also noted there was considerable variation in gross CS and the annual CS change between boreal, temperate and tropical forests. For example, globally the CS change incorporating biomass, dead wood, litter and soil was estimated to be 1.04 Mg C ha^{-1} yr^{-1}, whereas temperate forests were estimated to achieve a CS change of 0.34 Mg C ha-1 yr^{-1} (Pan et al. 2011).

Nair *et al*. (2009), using examples from around the world estimated that the mean C sequestration rates in agroforestry AGB ranged between 0.29 and 15.21 Mg C ha^{-1} yr^{-1}. Potential, annual SOC sequestration rates are more difficult to quantify given the range of agroforestry activities undertaken, diversity of SCS values and the different depth parameters used Nair (2012). However, in a synthesis of literature from global reports Nair *et al*. (2009) found SCS in forest soils ranged from between 1.25 Mg ha^{-1} to 97.20 Mg ha^{-1} at 0-40 cm depth, and from 24 Mg ha^{-1} to 173 Mg ha^{-1} to 1m depth. The variability these authors found in sequestration rates demonstrates the difficulty in determining global C sequestration rates associated with agroforestry (Craswell *et al*. 2013).

In addition to the C sequestered in the AGB, afforestation of agricultural land can also result in changes to SOC. However, the rate of change to the SOC pool is slower than for the AGB pool (Paul *et al*. 2002a). Trees contribute to the SOC pool because they increase root biomass and litter quantity. They also modify the quality of the litter. The trees can also affect the microclimate of the landscape by reducing wind flows, reducing soil temperature and increasing moisture retention in soil; all factors that lead to modification in the soil fauna and microbial biomass and thus the C sequestration potential of the soil.

Previous researchers have found that afforestation of agricultural land can result in either a loss, gain or no change to SOC. For example, in a global review of the literature Paul *et al*. (2002a) found that time since afforestation and soil depth were important factors determining changes to SOC. Their research found that in the 0-30 cm layer soil lost 0.2 Mg C ha yr^{-1} in forests less than five years old, but in forests older than 31 years SOC increased by 0.3 Mg C ha yr^{-1}. Hoogmoed *et al*. (2012) conducted a meta-analysis of research that investigated changes to SOC and N stocks following afforestation of land previously used for pasture under Mediterranean climatic conditions in Australia. They identified no clear temporal change in either C or N concentrations in plantings up to 30 years of age. They did, however, postulate that because remnant native woodlands had higher SCS there was potential for the plantings to sequester more SOC but more time was required. Post and Kwon (2000) found that when forests were established into land previously used for agriculture an average accumulation rate of 0.3 Mg C ha^{-1} yr^{-1} could be achieved. The average accumulation rate in their meta-analysis was determined from sites located in a range of climatic conditions from cool temperate to tropical wet forests. However, they found that when agricultural land from temperate areas was converted to forest, the SOC sequestration rate ranged from a loss of 0.1 Mg C ha^{-1} yr^{-1} to a gain of 0.7 Mg C ha^{-1} yr^{-1}. A global meta analysis conducted by Guo and Gifford (2002) found that SOC significantly increased on conversion from crop to pasture and from crop to secondary plantation. Finally, conversion of European cropping land to woodland has been estimated by Smith (2004) to have the potential to increase SCS by 0.6 Mg C ha^{-1} yr^{-1}. However, that estimation comes with a > 50 % uncertainty.

4.5 Australian policy position regarding BEPs and C sequestration

Australia ratified the Kyoto Protocol on 12 December 2007 and consequently is bound to achieve GHG emission reduction targets and meet reporting requirements that demonstrate compliance with the Kyoto Protocol. Afforestation and reforestation are recognised under Article 3.3 the Kyoto Protocol through the LULUCF framework as a GHG offset activity. To this end the Australian Government has developed legislation for an offsets scheme entitled the Carbon Farming Initiative (CFI). *The Carbon Credits (Carbon Farming Initiative) Act 2011* underpins the scheme providing incentives for farmers and land managers to store C or contribute to a reduction in GHG emissions and by so doing earn C credits. This voluntary scheme facilitates a C market to allow C credits to be traded for the purpose of offsetting GHG emission liabilities. Australia presently has set an emissions reduction target of 26-28 % below 2005 levels by 2030 (DPMC, 2015). An Emissions Reduction Fund is available to provide incentives for reducing GHG emissions. Amongst the eligible strategies is storing C in forests and soil (COA, 2015).

Australian national policy with specific relevance to C sequestration and BEPs is the *Carbon Credits (Carbon Farming Initiative) Act 2011* (s. 54). This Act refers to GHG abatement in the land sector and defines sequestration projects as those which remove CO_2 from the atmosphere or avoid emitting GHGs by sequestering C in the living biomass, dead OM and/or soil. The Act also defines Kyoto offset projects such as those specified in *The Carbon Credits (Carbon Farming Initiative) Regulations 2011*, and the *Carbon Credits (Carbon Farming Initiative) Rule 2015*. The offsets projects specified in the regulations include permanent plantings established after 1 July 2007, or permanent plantings established prior to 1 July 2007 that have documentary evidence to prove that the primary purpose of the planting was to generate C offsets.

The *Carbon Credits (Carbon Farming Initiative) Regulations (2011)* further define environmental plantings as those consisting of native vegetation species endemic to the local area, from seed sourced within the natural distribution of the species and comprising a mix of trees shrubs and understorey species representative of the local native vegetation.

Each offset project is required to be supported with an approved methodology. The methodology determining the C sequestration potential of Environmental Plantings is the *Carbon Credits (Carbon Farming Initiative) (Reforestation by Environmental or Mallee Plantings – FullCAM) Methodology Determination 2014* (Methodology Determination, 2014). This methodology is applicable to permanent environmental plantings that have been established using either a direct seeding or planting method. To be eligible plantings are required to contain trees with the potential to attain a crown cover > 20 % and a height > 2 m. The plantings are to be established on land previously used for grazing or cropping. The Methodology Determination (2014) requires that FullCAM be used to estimate tree and debris biomass for each approved BEP project. The purpose of using FullCAM is to provide a report for national estimates of net GHG abatement achieved through sequestration of atmospheric CO_2 in tree biomass. The Methodology Determination (2014) details BEP projects in terms of domains associated with geometry (configuration), spacing, stocking density and tree proportion. The Methodology Determination (2014) does not, however, include provision for providing credits for changes to SOC due to both a lack of data and confidence in the changes, sufficient to adequately calibrate FullCAM for soil.

The BEPs studied in the current research fulfil the definition of reforestation under the terms of the Kyoto Protocol (1998) as they were all sown after 1 January 1990, and they are an agroforestry project that sequesters atmospheric C. Afforestation is defined as the planting of forests on land that has not historically contained a forest and reforestation is the planting of a forest on land that has historically contained a forest that was cleared to make way for some other use such as agriculture. Reforestation meets Kyoto Protocol (1998) rules when the trees have been established by humans on land that was clear of forest on 31 December 1989.

Reforestation trees must also be determined to have the potential to grow to a height of at least two meters, have a crown cover of at least 20% and be grown on areas larger than 0.2ha. Reforestation is a recognised C credit earning activity from which credits can be used to offset Australia's Kyoto Protocol (1998) obligations.

4.6 Estimating carbon stocks in environmental plantings using FullCAM

4.6.1 The Australian Full Carbon Accounting Model (FullCAM)

FullCAM was developed by the then Australian Greenhouse Office (AGO) for the National Carbon Accounting System (NCAS) (Richards, 2001). The rationale for developing the model was to provide a cost effective way to monitor changes to the national C inventory rather than using direct measurement methods that are expensive and time consuming. The modelling tool was developed to monitor and account for GHG emissions and C sequestration arising from LULUC, particularly from the forestry and agricultural sectors at project and national scales (Richards, 2001; Brack *et al*. 2006). The model was developed to include all terrestrial C pools and transfers between those pools. Since development of FullCAM, its calibration and validation with empirically derived data has been ongoing. This section details the development of FullCAM for applicable AGB and soil C accounting as used in the current research project.

4.6.2 Above ground biomass

To quantify forest CS it is necessary to estimate both the AGB and BGB C content. Direct measurement of trees or stands can be undertaken but this requires destructive sampling. Destructive sampling requires that trees be felled and each component (stem, crown, dead attached material and dead detached material) weighed to determine fresh weight. However, for large scale estimations destructive sampling is impractical (Snowdon *et al*. 2002) time consuming, resource and labour intensive and consequently inefficient and expensive. From this information is it possible to develop allometric regression equations to predict biomass variables at a species and generic scale (Snowdon *et al*. 2002). The correlation between stem diameter measurements that are made at varying heights up to 1.3 m (DBH) and allometric equations for biomass are generally reported to provide easily obtainable and convertible estimates of tree biomass (Jonson and Freudenberger, 2011). Jonson and Freudenberger (2011) demonstrated that generic equations for biomass estimates are an efficient and economical way of determining biomass, particularly at a regional scale for sites having similar climate and soils. The same principal applies when determining the CS of the AGB in BEPs.

Use of the FullCAM, developed as a modelling tool for the purpose of predicting land sector (forest and agriculture) C sequestration (Richards, 2001; Richards and Evans, 2004; Brack *et al*. 2002; Brack *et al*. 2006), and for reporting land sector GHG emissions (DoE, 2014) is a cost effective and rapid way to monitor CS and flows at a variety of scales. FullCAM uses the following sub-models:

- **RothC -** Rothamsted soil C model version 26.3 (Jenkinson, 1990; Coleman *et al*. 1997) which is used for predicting soil C turnover;
- **GENDEC** – a microbial decomposition model (Moorhead and Reynolds, 1991) used for determining litter decomposition rates;
- **3PG** – a physiological growth model for forests (Landsberg and Waring, 1997) used for forest growth predictions;
- **CAMFor** – a C accounting model for forests (Richards and Evans, 2000) which is used for determining forest, trees, forest debris, inert forest soil and forest products (Paul and Polglase, 2004a,b);
- **CAMAg –** a C accounting model for cropping and grazing systems (Richards, 2001); and
- **Tree Yield Formula –** is the method used to determine tree production (Paul *et al*. 2015).

FullCAM can be configured for C sequestration modelling for a number of agricultural or forest systems and the transition from one system to another. For example Paul and Polglase (2004) carried out calibration and validation of FullCAM by analysing eucalypt and pine litter empirical data. In their study they determined that the CAMFor more accurately predicted litter decomposition than GENDEC, but suggested more research was required to analyse the chemical properties of litter.

The tool also provides modelling specifically for mixed species environmental plantings (MSEP) and has been developed to predict biomass at a national scale. FullCAM relies on the application of reliable growth modifiers to the yield curves used for C accounting. The growth modifiers for MSEP were recently reviewed by Paul *et al*. (2015) to develop biomass accumulation calibrations applicable to BEPs. The work undertaken to develop the new allometrics for above and below ground biomass included direct measurement and indirect estimates of above and below ground biomass, and an analysis of data to determine factors that influence biomass accumulation. The allometric equations developed for AGB specifically for FullCAM achieved an average model efficiency of 93% and were based on stem diameter measurement. The comparison between measured and allometric biomass estimates had an average difference of only 15%. However, the relationship varied between sites, and between life form or growth habit within sites. In addition rainfall was found to affect between-site variation. The site specific biomass allometrics were found to be more accurate than the

allometrics for generic life forms (Paul *et al*. 2014; Paul *et al*. 2015). Overall, FullCAM has been effectively calibrated to predict, within reasonable parameters, the biomass in environmental plantings (Paul *et al*. 2013; Paul *et al*. 2014; Paul *et al*. 2015).

4.6.3 Soil organic carbon and FullCAM/RothC

RothC was developed to model the turnover of SOC in arable soil by using the long-term field cropping and manuring experimental site at Rothamsted (Jenkinson and Rayner, 1977; Jenkinson, 1990). The model was developed to have application to a wide range of soils and environments, and was found to be applicable to 14 of 18 sites examined in Germany, England, USA, the Czech Republic and Australia (Coleman *et al*. 1997). Within this suite of experiments RothC was found to provide variable results for a research site located in Tamworth, NSW. The variability was attributed to the dry soil during summer fallow; something that RothC was not calibrated for at the time.

FullCAM uses the Rothamsted Soil Carbon Model (RothC version 26.3) (Jenkinson and Rayner, 1977; Jenkinson, 1990) to model SOC sequestration following establishment of mixed species environmental plantings. Until now, modelling of SCS under BEPs by FullCAM has not been as vigorously tested as have the calibrations for BEP biomass. Research was conducted by Paul *et al*. (2002b) into the use of RothC for modelling soil C change associated with plantation establishment and management with sites of differing management, soil type ad climate. The authors found that RothC showed sensitivity to SOC decomposition rates for the C pools and to microbial efficiency. As a result of the sensitivity of the model to these properties, these authors recommended further analysis was necessary to calibrate RothC to Australian forest soil conditions. Work undertaken by Paul and Polglase (2004a,b) and Paul *et al*. (2004) calibrated RothC within FullCAM for forest soils using two sites: one site was a second rotation, 10 year old *Pinus radiata* plantation, the second included first rotation *P. radiata* and *Eucalyptus grandis* plantations established on a previously improved pasture site. The authors concluded that decomposition constants used by RothC for agricultural soil were also adequate under the two forestry systems studied.

More recently a collaborative project examining SOC sequestration associated with mixed species environmental plantings was coordinated by CSIRO under a federal program called Filling the Research Gap (FTRG) National Soil Carbon Program (NSCP). The project aimed to incorporate SOC sequestration into the existing CFI Methodology for MSEPs (Methodology Determination, 2014). To achieve the aim calibration of FullCAM-RothC was carried out by using a collation of new and existing datasets of SOC sequestration associated with environmental plantings across Australia (England *et al*. 2016; Madhaven *et al*. in prep). The results from the current research project have been contributed to the pool of existing datasets used by the CSIRO/FTRG project. The results for the calibration of SOC for FullCAM are still being examined. Therefore, the findings between measured and FullCAM predicted results from

this current research project for SOC under BEPs is one of a few studies to have undertaken such a review. Consequently, the current work may contribute to extending the FullCAM based ERF Methodology (Methodology Determination, 2014) for MSEPs to include SOC.

4.7 Classification of soil organic carbon fractions and their application to C modelling

As discussed in section 4.6.3 previous studies have showed that the ability to model the turnover of SOC depends on a knowledge of the constituents of SOC, especially the humic and particulate pools (Jenkinson, 1990; Paul *et al*. 2002b). The proportion of humic to particulate material in soil varies depending on edaphic and land management practices (Oades *et al*. 1988). Thus, the determination of SOC functional pools attributed to changes in land management can be useful in detecting changes to SOC. For example, pasture soils are disturbed infrequently and have higher levels of stable aggregates than cropping soils which are tilled frequently. Tillage is a management technique that disrupts soil aggregates, thereby exposing organic material to decomposition (Tisdall and Oades, 1982; Oades, 1984).

Converting cropped soils to pasture is an effective way of stabilising aggregates and allowing time for humic material to become assimilated with the mineral fraction of soil. Because pasture soils are disturbed less frequently than cropped soil, the conditions and time in pasture are suitable for aggregates to form and for SOC to be sequestered within stable aggregates. Aggregation to form macroaggregates is achieved by increasing decomposable soil organic material such as that produced by pastures. In a study into the effects of afforestation on soil aggregation Six *et al*. (2002) found that forest systems had higher particulate organic matter (POM) than either afforested or agricultural systems for most aggregate size classes between 2mm to < 53µm. Afforested systems had higher levels of POM in the < 53 µm, 250 coarse and 2mm fine fractions than agricultural soil. Six *et al*. (2002) also found that the percentage of large macroaggregates > 2 mm in forested and afforested systems was up to 30-40% higher than agricultural systems. Consequently, given the potential reported by Six *et al*. (2002) for afforested systems to increase aggregation, it is expected that afforestation of pastures with BEPs will also lead to increases in aggregation and thus SOC sequestration.

Section 4.7 will detail the evolution of the concept of SOC fractions and their application to soil C modelling tools such as RothC and FullCAM. Methods used to measure or predict the content of SOC fractions and the correlation of these results with modelled data will then be discussed.

4.7.1 Evolution of conceptual and measured SOC fractions

There are a number of conceptual SOC pools each of which is identified through the size or density of the particles and their turnover time. Each pool has a distinct functional role and association with soil inorganic mineral particles. Jenkinson (1990) identified the following five pools:

1. Decomposable plant material (DPM),
2. Resistant plant material (RPM),
3. Microbial biomass (BIO),
4. Humified organic matter (HUM) and
5. Inert organic matter (IOM).

A conceptual model of SOC dynamics which incorporates the five pools was developed by Jenkinson (2000). The transfer of C through the five pools over varying periods of time underpin the RothC turnover simulation model version 26.3 (Jenkinson, 2000) (Figure 4-1). The model predicts changes to SOC in grassland, forest, pasture and cropping scenarios, including SOC sequestration or loss arising from LUC, management change and climate change (Paul *et al*. 2004). The model incorporates the association between SOC and other edaphic factors including mineralogy, climate, the nature of plant residues and microbial decomposition processes (Skjemstad *et al*. 2004.)

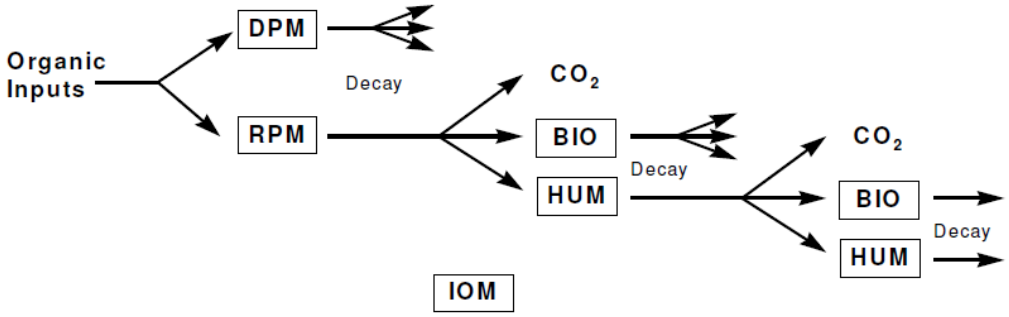


Figure 4-1. The RothC model and the flow of C (see text above for acronyms) (Source: Jenkinson, 2000).

Fundamental to the model is the concept that SOM can be partitioned into labile and stable pools. The labile SOM includes surface litter, undecomposed organic material within the mineral matrix and the microbial biomass including viruses, bacteria, fungi, algae and nematodes. This material originates from plant residue and is represented as DPM and RPM in the RothC model. Once associated with soil, the labile or non-humic constituents of SOM can decompose rapidly; within periods of days to months to years depending on climatic conditions (Jenkinson, 1990; Skjemstad *et al*. 2004). The residues from the decomposed material either transfer to the BIO or HUM pools or are lost to the atmosphere as CO_2. The more recalcitrant

HUM fraction can remain in soil for up to thousands of years. The HUM fraction is the organic matter fraction of soil that has decomposed to the stage that it no longer resembles its original form. HUM is comprised of HS such as humin, humic acid and fulvic acid, and non-humic substances including carbohydrates, proteins, lipids and organic acids (Oades, 1989:95). This fraction is less available to decomposer organisms because of an association with mineral particles which occurs either through adsorption onto mineral particles or by being occluded within soil aggregates (Oades, 1989, Golchin *et al*. 1997). The IOM pool represents organic material that is extremely resistant to decomposition (Jenkinson, 1990) and includes charcoal, so-called biochar or charred plant material.

As a further development of the conceptual SOC pools associated with the RothC model, Skjemstad *et al*. (2004) explored an association between the RothC conceptual pools with measured fractions of SOC for the purpose of verifying modelled predictions with measured data. In this context Skjemstad *et al*. (2004) found the following associations between the RothC pools and measured SOC fractions:

1. > 53 μm POC is the equivalent of RPM (as defined by RothC) and is comprised of labile organic material,
2. < 53 μm charcoal-C fraction (IOM) is biologically inert organic C, and
3. < 53 μm non charcoal C fraction (HUM) is the residue of decomposed organic material.

The methods these authors used to derive the three fractions included wet sieving to extract POC, photo-oxidation and ^{13}C NMR analysis to determine the content of inert material (Skjemstad *et al*. 2004). HUM-C content was determined from the difference of total organic carbon (TOC) less the POC and IOC content.

More recently Baldock *et al*. (2013a) redefined the Skjemstad *et al*. (2004) pool characteristics to derive the proportion of SOC fractions, and have reclassified the fractions as follows:

1. Particulate organic carbon 50-2000 μm (POC),
2. Humus ≤ 50 μm (HOC), and
3. Resistant (≤ 2000 μm) (ROC).

The modified method was developed after comparing the Skjemstad *et al*. (2004) manual wet sieving method, with an automated wet sieving method. The automated method was determined to provide a more efficient and precise fractionation method than the manual method (Baldock *et al*. 2013a). The fractions were then analysed for TOC. ROC content was determined for both the coarse and fine fractions (POC and HOC) by quantifying the poly-aryl structures contained in the samples by ^{13}C NMR analysis. Although labour intensive and expensive, this method provides a reliable measurement of the proportion of fractions within soil, and the results can be applied with confidence to calibrate C modelling tools such as RothC and thus the

FullCAM-RothC framework that supports national C accounting (Baldock *et al*. 2013b). The next section provides a comparison of the traditional methods to analyse SOC with MIRS.

4.8 Comparison of methods to measure SOC and the use of MIR spectroscopy to predict SOC and its component fractions

Measurement of soil C and component fractions has been achieved, until relatively recently, by utilising either wet oxidation or dry combustion methods. Each of the methods is time consuming and requires the use of either toxic chemicals resulting in toxic waste (wet combustion) or with expensive laboratory equipment (dry combustion). More recently MIRS has been used to predict the quantity and fraction composition of SOC. Section 4.8 compares these three methods of SOC analysis.

4.8.1 Traditional methods used to measure SOC

The wet oxidation methods such as the Walkley-Black (WB) (Walkley and Black, 1934) method and the Heanes method (Heanes, 1984) are commonly used to measure SOC. These methods arguably provide direct SOC results rapidly and more efficiently than dry combustion methods (Chatterjee *et al*. 2009). A significant advantage of these methods is that the results are not affected by the presence of carbonates and there is minimal effect caused by the presence of charcoal in soil (Walkley, 1947). However, because the methods use $K_2Cr_2O_7$ a substantial quantity of toxic chromate waste is produced.

The WB indirect oxidative procedure uses $K_2Cr_2O_7 + H_2SO_4$ to oxidise OM and has been found to produce imprecise and variable results (Vogel, 1994; Conyers *et al*. 2011). This is because not all the OC is oxidised. Indeed, Walkley and Black (1934) found the method resulted in an SOC recovery of approximately 76% (range 60 to 86%; se ±5.6 %). Consequently the authors determined that it is necessary to apply a correction factor of 1.32 to determine the OC presence in soil.

The Heanes method is a modification of the WB method that provides improved recovery of OC, with negligible interference by carbonates and charcoal. The method adds heat to the $K_2Cr_2O_7 + H_2SO_4$ digest at 137^{o}C for 30 minutes, thereby improving OC recovery to almost 100% (Heanes, 1984). Although achieving excellent OC recovery the Heanes method still results in the production of large quantities of toxic waste.

The high temperature dry combustion method to analyse soil C uses specialised laboratory equipment such as the Leco CNS2000™ instrument. The instruments are used to combust samples at temperatures > 950^{o}C for the recovery of both IC and OC thereby providing measurements of total soil C (TSC) concentration (Nelson and Sommers, 1996).

Methods to determine SOC using Dumas Catalysed High Temperature Combustion (DCHTC) have also been developed but they require several steps. First, it is necessary to identify samples containing carbonates by using the 'fizz' test. This test involves placing drops of 4 *M* hydrochloric acid (HCl) onto a moistened soil sample and observing whether any effervescence occurs (Nelson and Sommers, 1996). Samples that do not have carbonates require one dry combustion analysis for the TSC measurement, as the sample only contains OC. However, samples identified as containing carbonates need to be pre-treated with H_2SO_4 and $FeSO_4$ (Nelson and Sommers, 1996) or HCl (Chatterjee *et al*. 2009). Two dry combustion determinations are then necessary for each calcareous sample; one to quantify TSC including carbonates, and the second, after the soil has been treated to remove carbonates. SOC is measured by calculating the difference between the two results (Nelson and Sommers, 1996). This process makes the dry combustion method to determine SOC expensive, inefficient and time consuming. A further problem with dry combustion has been identified by Harris *et al*. (2001) who found that soils treated with HCl showed a significant decline in TSC and loss of OC, and therefore warned that the pre-treatment may cause inherent error in the quantification of SOC in calcareous soil. However, when H_2SO_4 is used the $FeSO_4$ present acts as an antioxidant thereby preventing the loss of C from the SOM (Allison, 1960; Nelson and Sommers, 1996).

Vogel (1994) reviewed the methods used to determine SOC and found that the direct dry combustion methods and indirect wet oxidation methods (excluding the WB method) provide comparable OC results following the correction on the dry combustion results due to carbonates.

4.8.2 MIR spectroscopy for soil C analysis

Given the shortcomings of the traditional analytical methods used to determine soil C content, and the labour intensive, time consuming and expensive methods used to determine SOC fractions, there has been much interest and recent research to explore the use of near or MIRS to examine several properties of soil composition including SOC.

MIRS offers advantages over conventional soil C analysis methods. These advantages include lower cost, improved accuracy, improved sample preparation methods, faster time to obtain results and the ability to achieve good quantitative and qualitative analysis of soil C fractions (McCarty *et al*. 2002). MIRS has been found to detect organic and inorganic soil attributes within the MIR wavelengths between 2 500 and 25 000 nm (Viscara Rossel *et al.* 2006) and can be used to qualitatively analyse soils by spectral interpretation and to quantitatively measure TOC, SOC and IC (McCarty *et al.* 2002; Reeves, 2010). It can also be used to predict SOC fractions.

MIRS has also been shown to provide accurate predictions for a range of other soil properties such as pH, CEC, texture, TP and EC (Viscarra Rossel *et al*. 2006) and can also be used to quantitatively predict results for TN, active N, and biomass N. Correlation between MIRS predicted results and conventional laboratory analysis can be greater than $R^2 > 0.9$ (Janik *et al,* 1998; Janik *et al*. 2007) with some researchers obtaining predictions as high as $R^2 = 0.99$ for SOC (Xie *et al.* 2011).

MIRS can provide rapid results through remote sensing, in the laboratory or in the field. Further, it has the added benefits of not requiring pre-treatment for carbonates and not using potentially toxic chemical reagents to perform the analysis and thus overcomes any adverse environmental outcomes arising from the disposal of toxic chemical waste (Viscarra Rossel *et al*. 2006; Janik *et al.* 2007).

In an important study into the effectiveness of MIRS in determining the component fractions of SOC (POC, HOC and ROC) Baldock *et al.* (2013b) compared MIRS predicted results with measured results obtained using wet sieving for fractionation analysis of the C content for each fraction using Leco CNS2000 C and ^{13}C NMR analysis (Baldock *et al*. 2013a). In this research the authors demonstrated that MIRS provided reliable predictions for TOC and the POC, HOC and ROC fractions of soil. From their research they concluded that MIRS with partial least squares regression (PLSR) analysis provides a rapid and cost effective method for determining the SOC and component fractions. Further, the authors propose that MIRS derived data can be used in conjunction with C accounting models such as FullCAM. In more recent work Madhaven *et al* (in prep) have developed MIRS calibration models for predicting changes to C pools under mixed species environmental plantings.

4.9 Research rationale and aim

4.9.1 Research Rationale

To date there has been very little empirical research into the SOC sequestration potential of BEPs established on land previously used for agricultural practices such as cropping and grazing in the temperate zone of NSW, Australia. There has also been very little research into the relationship between BEP AGB growth and SOC sequestration. BEPs are an approved FullCAM based ERF project (Methodology Determination, 2014) and are counted in Australia's national greenhouse gas accounting system. Consequently, there is potential for landholders who establish BEPs to trade in future C markets. It is therefore, important to have empirically determined biomass and SOC data to corroborate modelled data such as that derived using the FullCAM modelling tool. This Case Study therefore uses empirical data to determine the SOC and AGB sequestration potential of BEPs, and compares results obtained from the measured data with FullCAM simulated data.

4.9.2 Aim

The aim of the research in this case study is to identify whether any changes in SOC occur following establishment of BEPs in soil previously used for agriculture. A paired site, chronosequence approach has been used on 20 BEP sites sampled between 1 and 19 years following their establishment.

The research project has two research questions:

1. Does afforestation with BEPs lead to sequestration of SOC; and
2. How do the dynamic properties of BEPs influence the SOC sequestration potential?

In order to address the two research questions, the objectives of the study have been to:

1. Estimate annual rates of SOC sequestration and AGB growth following establishment of BEPs;
2. Compare empirically derived SOC and AGB data with FullCAM/RothC modelled results;
3. Identify any edaphic properties that may influence the SOC sequestration potential of the BEPs, including soil BD, TN and TP, C:N, and C:P ratios; and
4. Use MIRS to identify the soil fractions (HOC, POC and ROC) and validate the results with manual fractionation methods using wet sieving and ^{13}C NMR spectroscopy.

4.10 Study sites

4.10.1 Study area

This case study investigated 20 BEPs on 13 private rural properties located within a 75km radius to the north and north east of the Australian Capital Territory (Figure 4-2). Biogeographically the region is referred to as the Southern Tablelands of NSW, an area located to the west of the Great Dividing Range that has been extensively used for grazing since European settlement.

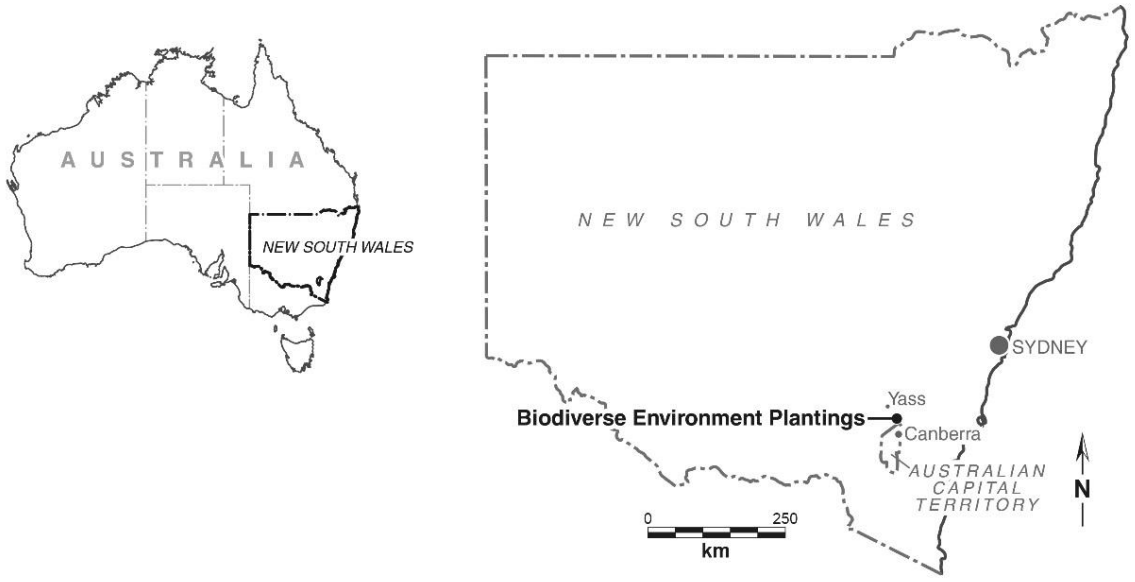


Figure 4-2. Map showing location of BEP study sites.

4.10.2 Site details

Between 1990 and 1996 GA established several BEP sites in the study region by using a direct seeding method. In 1998 GA undertook research at 64 of the sites to examine their species composition and health. At the same time they established permanent transects on the sites. Thirty one of those sites were used in 2008 for further research undertaken by Lowson, (2008); Read, (2008); and Schneeman, (2008). These three complementary projects examined AGB measurements, soil properties and tree species composition and health respectively. Six of these previously studied sites have been included in this current research project.

In consultation with GA, 14 additional sites were selected to allow inclusion of younger and older plantings. These sites were purposely selected based on the following parameters:

1. BEPs to fulfil the definition of reforestation under the terms of the Kyoto Protocol and the ERF Determination 2014 for Reforestation by Environmental or Mallee Plantings – FullCAM; including being sown after 1 January 1990;
2. Site composition could be either block or linear;
3. Similar rainfall, hence the sites were located within the same biogeographic region;
4. A range of aspect, landform and slope elements;
5. Must be fenced to exclude livestock; and
6. Age since planting to ensure at least four sites for each of four age classes were included in the study, where age class was defined as BEPs 1-5 years, 6-10 years, 11-15 years and 16-19 years since establishment.

In total 20 BEP sites on 13 privately owned rural properties were selected for inclusion in the current research project. The location of the BEP study sits is shown in Figure 4-3

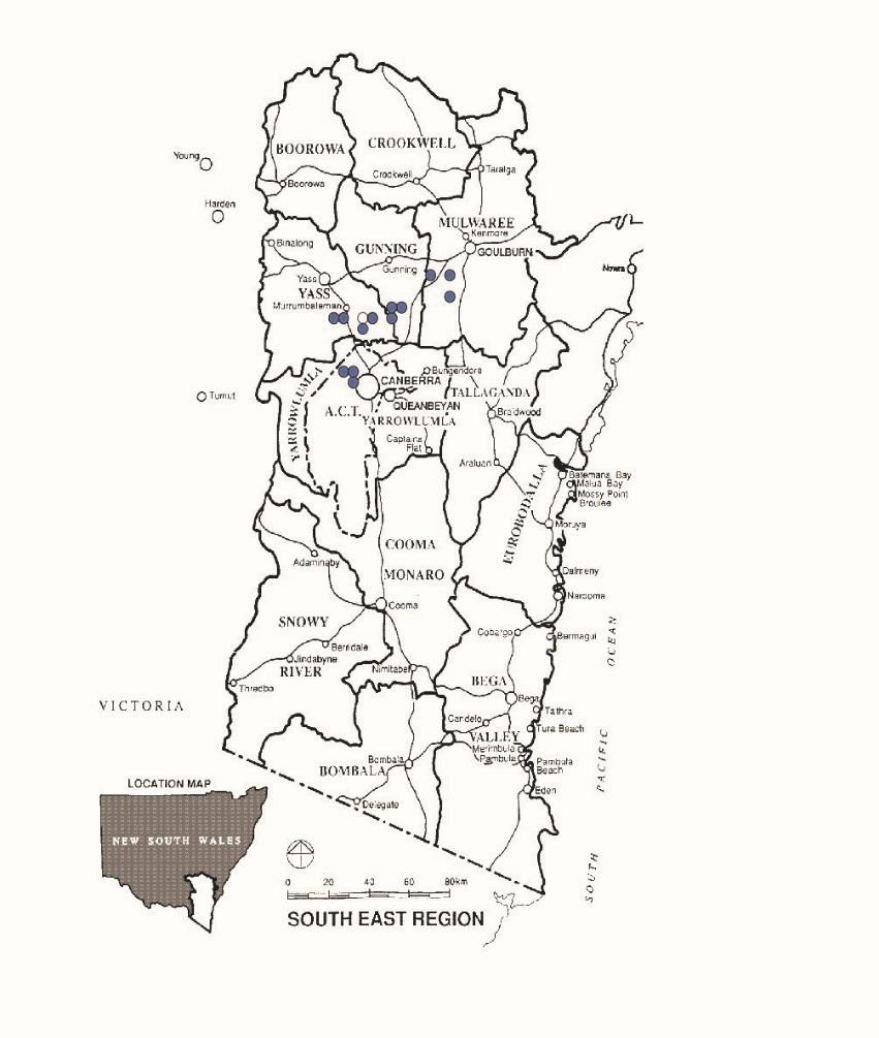


Figure 4-3. BEP study site locations (Source: GA).

The 20 BEP sites used in the current research were all established between 1991 and 2010 using the direct seeding method. All sites contained a mix of eucalypt and acacia species. Younger BEPs also contain a number of native understory species (Appendix 1:). Between 1999 and 2010 approximately 8 574 km rows of treelines had been sown in the Capital Region by GA (pers comm J. Cummings, CEO Greening Australia, Australian Capital Region, 2011).

The composition of the BEPs included in the current research were either linear, typically with 2 to 4 rows of trees, or block plantings comprising several rows (Table 4-1). BEP sites were established on a range of topographic positions within the landscape in terms of elevation, slope, drainage and aspect. Read (2008) identified that sites where grazing was practiced had higher SOC concentrations than those where grazing was excluded. Consequently, the BEP sites were included in this study only if they were fenced to exclude livestock grazing.

Table 4-1. Site details showing the configuration and landform descriptions. (a) Landform descriptions from Speight (2009).

Site ID	Site No	Age at sampling (yrs)	Location	Config-uration	Rows (no)	Aspect	Landform element [(a)]	Slope element [(a)]
anb	1	1	Murrumbateman	Linear	3	N:S	Low	Flat
keb	2	2	Murrumbateman	Linear	5	N:S	Mid	Flat
sha	3	3	Murrumbateman	Linear	4	E:W	Low	Flat
woo	4	4	Murrumbateman	Linear	7	E:W	Low	Waning lower slope
ter	5	6	Wallaroo	Block	4	E:W	Low	Flat
sc	6	9	Gundaroo	Linear	3	N:S	Low	Flat
wic	7	9	Gundaroo	Block	-	E:W	Low	Flat
ws	8	9	Hall	Block	9	N:S	Upper	Crest
cap	9	11	Murrumbateman	Linear	4	N:S	Upper	Waxing upper slope
car	10	11	Hall	Block	-	N:S	Upper	Crest
kea	11	12	Murrumbateman	Linear	4	E:W	Upper	Waxing upper slope
shb	12	12	Murrumbateman	Block	-	N:S	Mid	Simple slope
ana	13	13	Murrumbateman	Block	10	N:S	Mid	Waning mid slope
sta	14	15	Gundaroo	Block	-	N:S	Upper	Crest
wia	15	15	Gundaroo	Linear	4	E:W	Mid	Minimal Mid slope
wib	16	15	Gundaroo	Linear	3	N:S	Mid	Minimal mid slope
haa	17	16	Murrumbateman	Linear	4	E:W	Mid	Minimal mid slope
hab	18	16	Murrumbateman	Linear	4	N:S	Mid	Minimal mid slope
spr	19	18	Tirranaville	Block	-	E:W	Upper	Crest
pyl	20	19	Tarago	Linear	4	E:W	Upper	Waxing Upper slope

An objective of the current research was to seek an explanation as to whether change in SOC over time had a linear correlation with AGB growth within the BEPs. In order to identify any temporal or spatial differences in the SCS a chronosequence of sites based on the age of the BEP was used. The chronosequence approach has been used before for soil C and agroforestry e.g., Nair (2012) and Mudge *et al*. (2014). Consequently 20 BEPs across an age spectrum from between one and 19 years old were selected. Because sites for every year increment between one and 19 were not available four age classes were chosen. To achieve the spectrum of BEP ages, sites were selected to ensure a minimum of four sites were included from each of four nominally categorised age classes as follows:

- BEP age 1-5 years,
- BEP age 6-10 years,
- BEP age 11-15 years, and
- BEP age 16-19 years.

Visually and aesthetically BEPs in this region appear to be in their prime approximately 10 years after establishment. This is due to the acacias being in optimum health. Consequently, eight BEP sites aged 11-15 years were included in the study to allow detailed analysis of growth attributes from within this particular cohort to explore reasons for inter-site variation where appropriate (Figure 4-4).

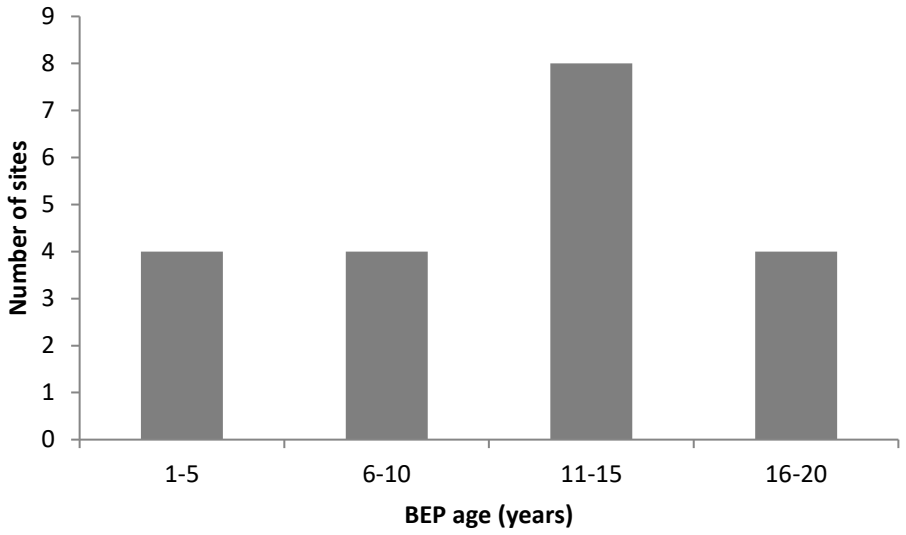


Figure 4-4. Histogram showing the number of BEP sites studied in each age class.

4.10.3 Climate data

Canberra Airport is the closest location to most of the study sites that has a record for mean maximum and minimum temperatures, evaporation and rainfall data (Figure 4-5). Climatic data is shown for the period 1939 – 2010. The mean annual maximum temperature for Canberra Airport during that period was 19.7 ºC, mean minimum temperature was 6.5 ºC and mean annual rainfall was 614 mm and mean annual evaporation is 1764 mm.

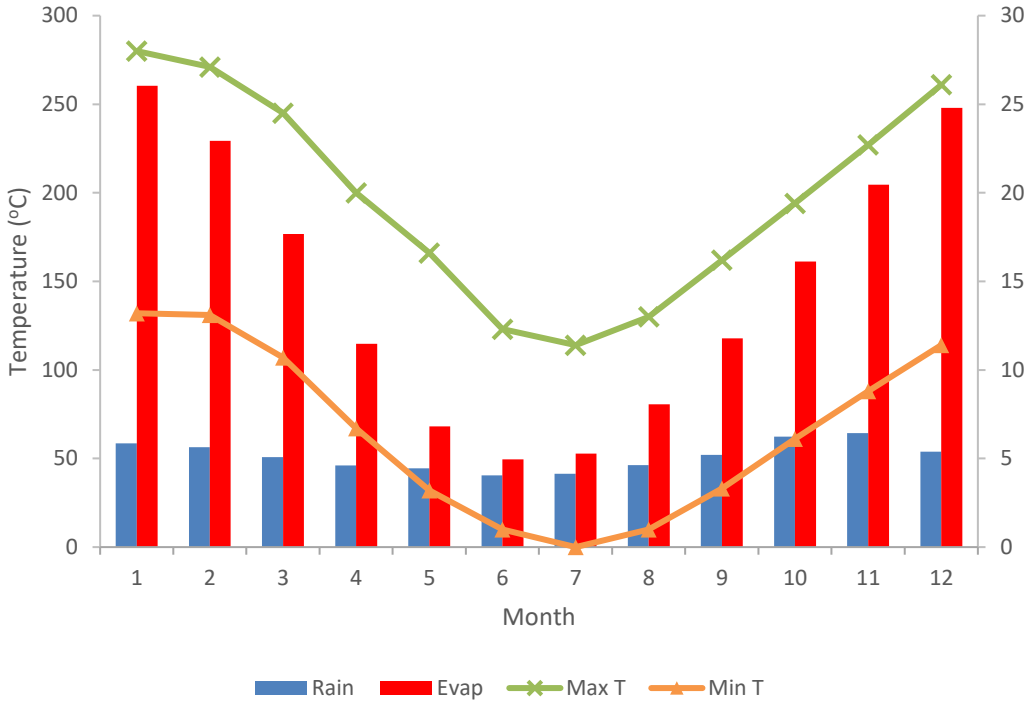


Figure 4-5. Mean rainfall and temperature data for Canberra Airport (1939-2010). Source BOM, 28 May 2016) (mean maximum temperature = green line, mean minimum temperature = orange line, rainfall = blue bar, evaporation = red bar.

The Bureau of Meteorology (BOM) also holds rainfall data for other sites located in close proximity to the study sites. The long-term recorded mean annual rainfall data for these locations (Source: BOM, 2016) is as follows:

Yass	650 mm (1898 – 2011)
Murrumbateman	726 mm (1985 - 2016)
Gundaroo	633 mm (1987 – 2014)
Hall	700 mm (1903 – 2016)
Goulburn (Springfield)	671 mm (1930 – 2016)

Detailed monthly breakdown of rainfall data for each of the above stations during the period of this study (1991-2011) is contained in Appendix 2.

4.10.4 Lithology

The elevation of the study sites ranged between 554 – 743 m. The study area is within the Lachlan Fold Belt. The Lachlan Fold Belt is an orogenic belt with varying tectonic structural zones (Gilligan and Scheibner, 1978). Geological formations in the region are developed from early Devonian felsic volcanics, Silurian – Devonian carboniferous granites, and Ordovician – Silurian metamorphic complexes.

Sites located in the area between the Australian Capital Territory (ACT) and Yass, NSW, adjacent to the Barton Highway (e.g., Murrumbateman, Hall and Wallaroo) are predominantly late Silurian felsic volcanics and include the Laidlaw and Deakin Volcanic suites which are made up from quartz, plagioclase, biotite and other types (Jennings, 1972; Wyborn *et al*. 1982). Sites closer to Gundaroo are late Ordovician to Early Silurian deep marine felsic sediments. Soils ranged from chromosols, through kurosols and rudosols. Soil classification and geological information for each site is detailed in Table 4-2.

Table 4-2. Geology and soil classification of study sites.

Site ID	Age at sampling (yrs)	Location	Geology (* Abel, 1992: #Thomas *et al.* 2013)	Australian Soil Classification (order, sub-order, great-group (Isbell, 2007)
anb	1	Murrumbateman	Mt Painter Volcanics, sandstone, siltstone, metasediments*	Chromosol, yellow, dystrophic
keb	2	Murrumbateman	Mt Painter Volcanics, sandstone, siltstone, metasediments*	Chromosol, red, dystrophic
Sha	3	Murrumbateman	Douro Group, Hawkins Volcanics sandstone, siltstone, mudstone and conglomerate#	Chromosol, brown, dystrophic
woo	4	Murrumbateman	Mt Ainslie Volcanics, metasediments, quartz*	Chromosol, brown, dystrophic
ter	6	Wallaroo	Deakin Volcanics, metasediments*	Chromosol, red eutrophic
sc	9	Gundaroo	Alluvium, dark alluvial soil*	Rudosol, stratic
wic	9	Gundaroo	Acton shale member, quartz, metasediments*	Rudosol, clastic, colluvic
ws	9	Hall	Mt Ainslie Volcanics, metasediments, quartz*	Rudosol, clastic, lithosolic
cap	11	Murrumbateman	Mt Painter Volcanics, sandstone, siltstone, metasediments*	Chromosol, red, dystrophic
car	11	Hall	Mt Ainslie Volcanics, shale, felsic, metasediments*	Rudosol, clastic, lithosolic
kea	12	Murrumbateman	Mt Painter Volcanics, sandstone, siltstone, metasediments*	Chromosol, red eutrophic
shb	12	Murrumbateman	Douro Group, Hawkins Volcanics sandstone, siltstone, mudstone and conglomerate#	Chromosol, yellow, mesotrophic
ana	13	Murrumbateman	Mt Painter Volcanics, sandstone, siltstone, metasediments*	Chromosol, yellow, dystrophic
sta	15	Gundaroo	Pullman Formation, sandstone, siltstone, metasediments*	Chromosol, red, eutrophic
wia	15	Gundaroo	Pullman Formation, sandstone, siltstone, metasediments*	Chromosol, red, dystrophic
wib	15	Gundaroo	Pullman Formation, sandstone, siltstone, metasediments*	Chromosol, red, dystrophic
haa	16	Murrumbateman	Yass Formation, sandstone, shale*	Kurosol, yellow, dystrophic
hab	16	Murrumbateman	Yass Formation, sandstone, shale*	Chromosol, yellow, eutrophic
spr	18	Tirranaville	Quialigo Volcanics, quartz, metasediments*	Chromosol, yellow, dystrophic
pyl	19	Tarago	Woodlawn volcanics, rhyolite, sandstone, quartz*	Chromosol, red, dystrophic

4.11 Study design

A paired site approach was taken to enable a determination of changes to SCS in agricultural land (AL) following establishment of BEPs. Soil attributes such as landscape form, pH, EC, TN, TP and nutrient ratios for C:N:P were also characterised at each paired site. The SCS results for the paired AL position were used as a baseline taken to be representative of the SCS present at time zero when the BEPs were first sown into the pasture. Any difference in SCS after the BEP was established was taken to be indicative of the C sequestration potential of the BEPs.

At each BEP site, three replicate transect positions were randomly selected to locate a 10 m length transect parallel with a tree line (TL) (Figure 4-6). The three replicates were labelled as either T1, T2 or T3 accordingly. A total of $n = 77$ TL transects were sampled. The following attributes of each TL transect were recorded:

1. Aspect of BEP;
2. Position on slope (upper, mid, or low) and elevation;
3. Whether an inner or outer row;
4. Tree genus;
5. Tree species were identified based on the reference Costermans (1983),
6. Form (tree or shrub),
7. Whether the tree or shrub was alive or dead,
8. Position along the transect (cm distance from start of the transect), and
9. Tree measurement (diameter at breast height 1.3 m (DBH), diameter at 30 cm (D30) or diameter at 10 cm (D10) depending on tree size and form).

Adjacent to each of the three TL transects, a paired transect was established in the inter-row (IR) position; that being the mid-point between two TLs (Figure 4-6). These transects were labelled IR 1, 2 or 3 accordingly. Adjacent to the three TL and IR transects, and outside the BEP fence line a further three 10 m long transects were established on the adjoining AL. To avoid interference by soil roots and livestock camps the AL transects were established at least 30 m away from the BEP fence line. The samples were labelled AL 1, 2 or 3 accordingly.

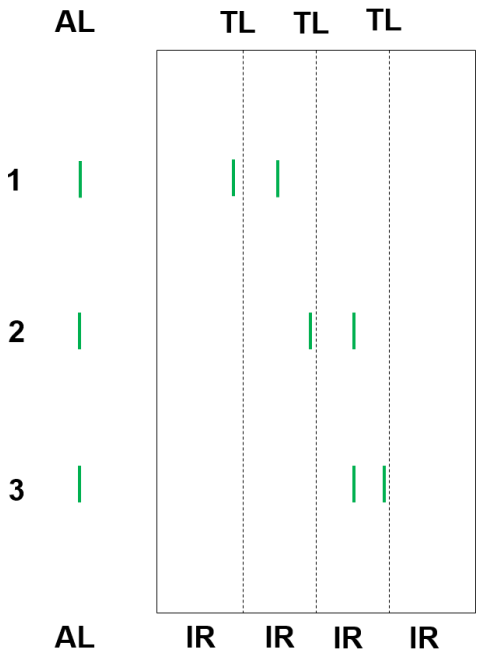


Figure 4-6. Typical transect layout at each site (not to scale) (TL = tree line; IR = inter-row; AL = agricultural land).

At the 5 m point in each TL transect a 0.5 m x 0.5 m quadrat was placed on the ground and, where it was present, the tree litter collected and placed in labelled paper sample bags. Tree litter was present at and sampled from *n* = 53 transects.

At each transect, three soil cores were taken to 30 cm depth at the 2.5 m, 5 m and 7.5 m point along each transect. As described earlier (section 3.1.1), where possible a truck mounted, mechanical soil corer was used, but vehicular access within BEPs was usually not possible, particularly in fenced linear plantings. In sites where vehicular access was not possible manual soil core extraction equipment was used. Each 30 cm soil core was divided into four depth increments i.e. 0-5, 5-10, 10-20 and 20-30 cm. The three soil cores from individual depths obtained from each transect position were bulked together in a labelled plastic bag.

At two sites, additional transects were established and a different sampling strategy was used. The rational for the sampling strategy, sampling method used and site no. are detailed in Table 4-3.

Table 4-3. Sites where an alternative sampling regime was followed.

Site No	Sampling method used	Rationale
9	8 transects for TL and AL only	This site was used as the pilot site to determine the optimum sampling and analysis strategy. Samples from IR were not taken.
14	12 plots used to obtain TL and IR samples only	These 12 plots were in addition to the three paired transects studied for this research project. The 12 plots were included in a project undertaken by CSIRO to calibrate allometrics for FullCAM (Reference: Paul, *et al*. 2013)

4.12 Methods used to analyse changes to soil properties including C and above ground biomass in BEPs

Standard methods used to analyse BD, $pH_{(1:5\ water)}$, $EC_{(1:5\ water)}$, and SOC using the Heanes, (1989) method are described in Chapter 3. The standard analytical methods were applied to all *n* = 830 BEP samples.

The methods used to analyse soil samples for TN and TP, Leco CNS2000 C, wet sieve fractionation, MIRS and ^{13}C NMR analysis are detailed below. The details for the ESM equation used to calculate SCS for BEPs are also covered in this section. Because soil sampling was not undertaken before establishment of the BEPs, the soil attributes from the adjoining AL position has been used as the baseline against which change caused by BEPs can be assessed.

AGB measurements, allometrics and methods of analysis are also outlined in this section. Details of the default data used for the FullCAM simulations are also presented. On overview of the sampling and analysis regime for both soil and AGB used for this case study are shown in Figure 4-7.

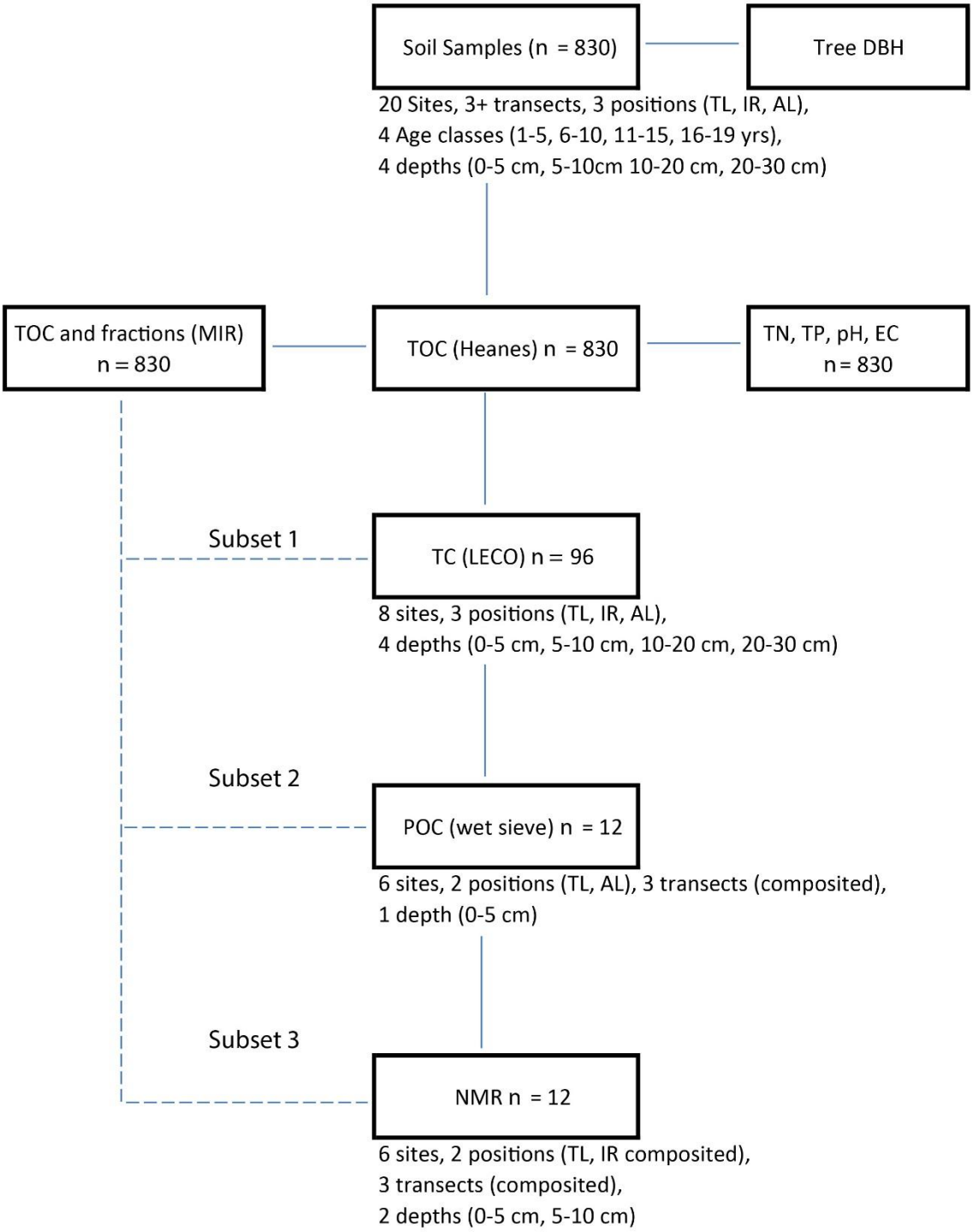


Figure 4-7. Overview of the strategy used for soil sampling, tree measurement and laboratory analyses for the BEP Case Study.

4.12.1 Tree identification and measurement

Trees and shrubs were identified using the botanical key, guide lists, and detailed descriptions contained in Costermans (1983).

Tree measurements were obtained on 1641 live and dead trees or shrubs from *n* = 77 TL transects. All the trees and shrubs located along each 10 m TL transect were measured using a diameter measuring tape. Tree diameters were recorded at DBH, or at 30 cm for trees or shrubs with shorter stems. The diameter at 10 cm was recorded for very small or young trees or shrubs where appropriate.

The tree stocking rate for each transect was calculated by summing the number of live and dead trees/shrubs present in each transect and multiplying by the transect area (ha).

4.12.2 Above ground biomass allometric equations

AGB was calculated from allometric equations for environmental plantings developed by Dr K. Paul, (pers comm, 2012) and from Paul *et al*. (2013), Paul *et al*. (2014) and Paul *et al*. (2015). BGB was estimated on allometric equations developed by Paul *et al*. (2014). These authors constructed the equations from both an analysis of previously developed data and by the direct measurement of the biomass of 3139 eucalypt, acacia and casuarinaceae tree species and acacia and melaleuca shrubs at 22 Australian BEP sites. At eight sites trees were harvested to derive accurate measurements of above-ground biomass and to validate allometric equations (Paul *et al*. 2013a; Paul *et al*. 2014). One of the eight sites (ID = "sta", Table 4-2) was included in the Paul *et al.* 2014 study.

The allometric equations were developed for two purposes:

1. To understand the relationship between general allometric equations for AGB that had previously been developed for BEPs with directly measured results obtained by harvesting trees, and
2. To calibrate FullCAM which is used to report rates of AGB C sequestration as part of Australia's greenhouse gas reporting obligation and includes an estimate of C sequestration by BEPs.

The allometric equations developed by Paul *et al*. (2013a,) and Paul *et al*. (2014) relevant to this current research were provided by Dr K Paul, (pers comm, 2012). They included the site and species specific, generic species, and universal equations developed for BEPs.

4.12.3 FullCAM modelling

FullCAM Version 3.55.6.0781 (Research Edition) was used to obtain predictions for biomass C, litter and soil. This version of FullCAM utilises version 26.3 of RothC. A simulation plot was established for each BEP site, and configured as a forest system. The tree yield formula (TYF) was used to predict tree production. Site simulations were set up to start at the year and month of sowing, and ended at the year and month of sampling. The Determination (Schedule 1) provides specific calibrations for planting type, geometry, stocking density and tree proportion. The Schedule is reproduced at Appendix 1.The initial tree species was set as "Mixed species environmental planting temperate". The regimes for each site were selected based on the actual geometry, measured stocking rate and proportion of eucalypt trees identified in accordance with Schedule 1 of the Determination. The initial value for tree biomass and debris was set as zero. Site specific soil fraction data for POC, HOC and ROC from the AL position was used as the baseline.

FullCAM default values for initial tree conditions were used except as detailed in Table 4-4. Based on calibrations derived by Paul *et al*. (2013) and more recently (K. Paul. pers. comm. 5 Dec 2014).

Table 4-4. FullCAM parameter adjustments made to derive annual turnover %.

Variable	*Turnover % per year*
Branches	10
Bark	10
Leaves	50
Coarse roots	10
Fine roots	80

Turnover inputs for FullCAM representing the turnover or the annual rate at which material is lost (% per year) were adjusted for debris as detailed in Table 4-5. The rates were based on calibration analysis undertaken by Paul *et al*. (2004) and Paul *et al*. (2013a).

Table 4-5. Breakdown percentages adjusted for debris properties.

Variable	*Breakdown %/yr*		*Atmospheric % of breakdown products*	
	Decomposable	*Resistant*	*Decomposable*	*Resistant*
Deadwood	11	11	77	77
Chopped wood	82	82	77	77
Bark litter	50	30	77	77
Leaf litter	100	26	77	77
Coarse dead roots	50	50	77	77
Fine dead roots	100	100	77	77

The initial maximum tree biomass conditions and the proportion of C mass (Mg C ha^{-1}) for the debris components were adjusted to zero % (K. Paul, pers. comm. 5 Dec 2014). The initial conditions for the POC (RPM) and HOC (HUM) pools of soil C were based on the MIRS

predicted data for the AL position at each site; this being taken to be representative of the baseline conditions of each site before the BEP was established. Initial conditions for ROC (inert soil) were taken to be those from the BEP. The MIR data was then added to the FullCAM plot file as shown at Table 4-6.

Table 4-6. Adjustments for initial conditions for each of the soil pools.

SOC Pool	*C mass (Mg C ha^{-1})*
DPM	10
RPM (POC)	Derived from site specific MIR data (AL)
BIO-F	0.66
BIO-S	0.66
HUM (HOC)	Derived from site specific MIR data (AL)
Inert soil (ROC)	Derived from site specific MIR data (BEP)

Once all the calibration adjustments had been made to the FullCAM defaults, a simulation was run for each site. Data for initial conditions at the start of the FullCAM simulation for C mass of trees, soil, litter, DPM, RPM, BIO, HUM, and IOM (Mg C ha^{-1}) and the dry mass of trees (DMT) per hectare was then compared with the results for the end of the simulation to obtain a change over time. MIRS results were also used to verify the FullCAM predicted results.

4.12.4 Litter sampling

Litter was sampled at the TL and IR positions from 53 transects at 11 sites aged between 11 and 19 years old. The sampling method involved placing a 50 x 50 cm quadrat on the ground at the 5.0 m midpoint of the TL and IR transect and collecting the litter located within the quadrat. The litter was separated into the O1 (A_o) and O2 (A_{oo}) horizon in accordance with the classification recommended by McDonald and Isbell (2009). The material classified as O1 horizon was comprised of undecomposed material including leaves and twigs. Material classified as O2 horizon included decomposed organic debris. When collecting the O2 horizon material, care was taken to avoid the inclusion of mineral matter as far as was practicable. The litter samples were returned to the laboratory, weighed and placed in an oven at 60ºC until the weight remained constant. The dry samples were weighed to determine the dry mass (DM) of litter for each transect.

Sites that had grassy or shrub understory but not tree litter were not sampled for litter even although some were in the 11 to 19 year old BEP cohorts. Notably, all sites younger than 10 years old were found to have grass as the dominant ground cover as insufficient time had lapsed for the tree canopy to be dense enough to allow a build-up of tree litter.

FullCAM provides modelled results for the CS of litter. Because analysis for C content of litter was not undertaken for the current research, initially a conversion factor of 54.3% was used to convert the DM result for litter to a CS. This rate was based on the findings of Gifford (2000) who analysed the C % of various components of tissue from 19 eastern Australian tree species, many of which were the same as those included in the current research. Gifford (2000) found

that 54.3 % of naturally senesced surface leaf litter was C (CV 2.5). In further support for the conversion factor adopted for the current research project, Harper *et al*. (2012) found similar results to Gifford (2000) for litter C content in their study examining the effects of afforestation of agricultural land on C storage. They found leaf litter contained 56.3 ±0.2% C and fine debris contained 55.1±1.2% C.

4.12.5 Equivalent soil mass

To determine an unbiased comparison between paired AL and BEPs the SCS was determined on an ESM basis for each individual depth layer tested (i.e., 0-5, 5-10, 10-20 and 20-30 cm layers). The 10th percentile of the mean soil mass for the AL, TL and IR was used to calculate a site reference soil mass (K Paul, pers comm, 2014). The same method was used to calculate the SCS on an ESM for each of the individual soil fractions determined by MIRS analysis, including POC, HOC and ROC. The calculation was carried out by following the steps outlined below.

1. Calculating the SCS (Mg C ha^{-1}) on a fixed depth basis (CS_{fd}) (See Equation 4);

2. Calculating the fine soil mass per ha^{-1} (FSM) on a fixed depth basis (FSM_{fd} Mg ha^{-1}) for each sample as follows

   Soil bulk density (g cm^3) x (soil layer depth (cm)/100) x 10000

   Equation 8

3. Determining a site reference soil mass (SRSM) for an individual soil layer (eg 0-5 cm) from the mean FSM of the TL, IR and AL samples from the site and then by using the 10th percentile of this mean, the reference soil mass for the site was calculated ($SRSM_{10}$ Mg ha^{-1}); eg

   10th Percentile FSM (Mg ha^{-1}) = mean FSM for every sample, from each BEP position at a site, for a given soil layer (eg 0-5 cm)

   Equation 9

4. The ESM was then calculated for each site, transect and soil layer as follows:

   $$ESM = CS_{fd} + (SRSM_{10} - FSM_{fd}) \times TOC_{odw} (\%) / 100$$

   Equation 10

5. To determine the SCS to 30 cm, the sum of the ESM from each depth layer (0-5, 5-10, 10-20 and 20-30 cm) was calculated for each site, position and transect.

A worked example calculating the ESM for an 11 year old BEP site using the above equations is shown at Appendix 4: BEP Equivalent Soil Mass calculation

4.12.6 Total Nitrogen

All *n* = 830 samples were analysed for TN using the Kjeldahl N flow injection analysis (FIA) method to determine the concentration of ammonium N (Bremner, 1996). Between 0.2 to 0.25g of soil ground to < 500 µ was placed into digestion tubes. 1 ml of copper catalyst was added to the sample and left overnight. 2.25g of K_2SO_4, 5.25ml H_2SO_4 and 1 ml H_2O_2 was then added to the tubes which were placed in a digestion block and warmed to 50ºC. The block temperature was increased to 200ºC and then the tubes removed. Once cool a further 1ml of H_2O_2 was added and the tubes returned to the block which was then heated to 350ºC. Once the block reached 350ºC the tubes were again removed, cooled and a final 1m H_2O_2 added. The tubes were again returned to the block at 350ºC until the digested liquid turned green. The samples were then allowed to cool before being diluted with Milli-Q water to the 75ml mark. The tubes were stoppered and allowed to settle overnight.

Calibration standards were prepared following the QuikChem Method 13-107-06-2-D method (Diamond, 2007). Samples were analysed using a Lachat Flow Injection Autoanalyser (FIA).

4.12.7 Total Phosphorus

All n = 830 samples were analysed for TP using a total Kjeldahl P FIA method. The same samples as for the TN analysis were used. Calibration standards were prepared following the QuikChem Method 13-115-01-1-B method (Diamond, 2006). Samples were analysed concurrently with TN analysis.

4.12.8 MIRS analysis

MIRS analysis was undertaken on finely ground soil samples for each of the *n* = 830 BEP samples to obtain soil C fraction data. The analysis was conducted at the University of Melbourne (UM) (T. Baker, pers comm, 2013). The samples were scanned to acquire diffuse reflectance spectra using a Thermo Nicolet 6700 FTIR spectrometer and Pike AutoDiff automated diffuse reflectance accessory (Pike Technologies, Madison, WI, USA). The spectra range used was between 4000 to 450 cm^{-1} baseline corrected spectra. The method used to acquire the spectra is detailed in Madhaven *et al*. (in prep). Briefly, each samples was scanned for one minute to obtain approximately 240 scans. The reflectance spectrum results were then determined as an average of the 240 scans and converted to absorbance spectra.

4.12.9 Soil fractionation using wet sieving

A subset from six BEP sites, comprising 12 samples from the 0-5 cm soil layer of randomly selected BEP sites established for 10 years or more was selected for a detailed study of their fractions. The sites were purposely selected from older BEPs because it was anticipated the BEPs of this age had been established for long enough to produce changes to the soil C pools. The sites were also from similar geographic, geologic and pedologic groups. This subset is named Subset 2, and was constructed to identify whether BEPs had changed the soil C fraction composition of the AL by comparing samples from the TL and paired AL positions. The samples were made up of composites of samples from two or three transects for each site (depending on sufficient soil sample for analysis). Each composite was prepared as detailed in Table 4-7.

Table 4-7. Sites selected for POC determination and composite constituents.

Site	*Position*	*Age*	*Composite made up of transects indicated*
haa	AL		T1 + T2 + T3
haa	TL	16	T1 + T2 + T3
hab	AL		T1 + T3
hab	TL	16	T1 + T3
kea	AL		T1 + T2 + T3
kea	TL	12	T1 + T2 + T3
shb	AL		T1 + T2 + T3
shb	TL	12	T1 + T2 + T3
sta	AL		T1 + T2 + T3
sta	TL	15	T1 + T2
wia	AL		T1 + T2 + T3
wia	TL	15	T1 + T2 + T3

Fractionation was undertaken by wet sieving samples using the automated fractionation method described by Baldock *et al*. (2013b) and Madhaven *et al*. (under review, 2016) at the Burnley Campus, University of Melbourne. Briefly, a 10 g sample from the < 2mm sieved soil was placed into a 50ml plastic tube with 45 ml of 5g l^{-1} of sodium hexametaphosphate. The tubes were placed on a shaker overnight. The dispersed solution was then placed into an automated vibratory sieve shaker using a 50 µm sieve to physically separate the soil into coarse (2000-50µm) and fine (<50µm) fractions (Sanderman *et al*, 2011; Baldock *et al*. 2013b; Madhaven *et al*. under review, 2016). The materials were dried at a low temperature (40ºC) until completely dry and the coarse fraction was then ground to a fine powder texture. Analysis for SOC was undertaken on both fractions using a Leco CNS2000 dry combustion analyser (T. Baker, pers comm, 2013).

4.12.10 ^{13}C NMR spectroscopy

Using the same BEP sites as for the wet sieve soil fractionation analysis (paragraph 4.12.9) a new subset of 12 samples was prepared for ^{13}C NMR spectroscopy using the wet sieving method (Subset 3). The subset was created by combining samples from the TL and IR positions together (to be representative of the C fractions within BEPs). Six samples from the 0-5 cm layer and a further six from the 5-10 cm layer were included in this study. SOC was determined for the coarse (> 50 μm) and fine (< 50 μm) fractions by dry combustion analysis, but each contained ROC which biased the C concentration in each fraction. Consequently, ^{13}C NMR spectra were acquired on both wet sieved fine and coarse samples to determine the concentration of ROC in each fractions. This was achieved by identifying the signal intensity in the aryl (110-145 ppm) and O-aryl (145-165 ppm) regions thereby allowing calculation of the proportion of lignin and ROC in each fraction using Equation 11 (Baldock *et al*. 2013a; Madhaven *et al*. (under review, 2016)).

$$ROC = (aryl\ C - 1.77 * O\text{-}aryl\ C)/0.45$$

Equation 11

The concentration of the POC and HOC fractions were calculated by subtracting the concentration of ROC from the concentration of TOC in the coarse and fine fractions as follows (Madhaven *et al.* under review, 2016):

- POC fraction = TOC of the coarse fraction x 1 – the proportion of ROC in the coarse fraction,

Equation 12

- HOC fraction = TOC of the fine fraction x 1 – the proportion of ROC in the fine fraction

Equation 13

The ^{13}C NMR analysis was conducted by UM at the CSIRO Waite laboratory in South Australia (T. Baker, pers comm, 2014; D. Madhaven, pers comm, 2014).

A flow chart of the steps involved in the preparation of samples and determination of the concentration of ROC, POC and HOC is shown in Figure 4-8. The method used for the ^{13}C NMR analysis is described in detail by Skjemstad *et al*. (1994), Baldock *et al*. (2013a) and Madhaven *et al*. (under review, 2016).

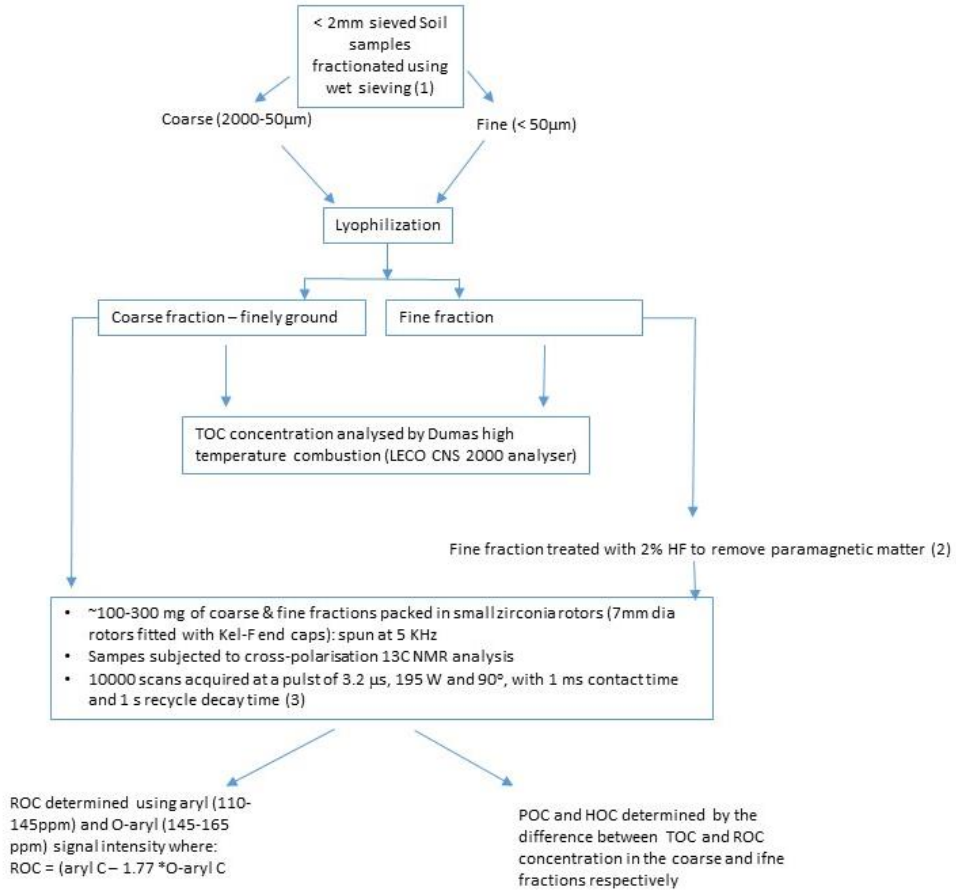


Figure 4-8 Method for NMR determination of ROC, POC and HOC fractions. (1) Madhaven *et al.* (under review, 2016; (2) Skjemstad *et al.* (1994); (3) Baldock *et al.* (2013a).

The results from Subset 3 have also been used to assess the relationship between wet sieved, ^{13}C NMR measured results and MIR predicted results (using the mean of MIR predicted results for TL and IR) for the SOC content in each of the soil fractions.

4.12.11 MIRS calibration model development

Calibration and validation of the MIRS spectra (in the band 6000-450 cm^{-1}) was undertaken to develop MIRS PLSR predictive models. Matlab software version R2013a by MathWorks, Natick, MA, USA was used to undertake principal component analysis (PCA). PLSR analysis was performed using PLS_Toolbox 7.0 (Eigenvector Research Inc) (D Madhaven, 2014 pers comm., 5 Dec 2014). A progressive series of calibrations using three datasets were used to create robust MIRS calibration models for TOC and C fractions for this project.

The calibration model was developed using *n* = 287 samples derived from the Australian National soil carbon research (SCaRP) project (Baldock *et al*. 2013a,b), plus *n*=12 BEP samples (Subset 3) that had been analysed by ^{13}C NMR analysis. The SCaRP samples included soils from around Australia. Initially data were validated using 'leave-one-out' cross validation. Spectral pre-processing for calibration results included multiplicative scatter correction (MSC) and then first derivative, mean centering and auto scaling. Pre-processing of the MIRS SOC data was undertaken by cube root and autoscale to normalise the distributions (D. Madhaven, pers comm 13 November 2013). This dataset provided ^{13}C NMR measured results for C fractions, which were calibrated against the MIR fraction predictions. However, because of the diversity of soils and range of mineralogy and soil types included in the SCaRP National dataset, correlations with the SOC results from the current research project were weak. Consequently, a further calibration model was developed by using the *n*=12 subset (3 samples) from the current research and a subset (*n* = 45) from the SCaRP samples. This subset was selected by regression analysis to identify and exclude outlier SCaRP samples from the model (paragraph 4.12.10).

The summary statistics used for the MIRS-PLSR calibration and leave-one-out cross validation used to develop the PLSR predictive models used to determine the SOC content of the soil fractions are shown in Table 4-8 (D. Madhaven, pers comm, 5 Dec 2014).

Table 4-8. MIRS-PLSR calibration and cross validation statistics.

n	151	55	52	52
Spectra transformation	MSC (mean) 1st derivative Mean center Autoscale	MSC (mean) 1st derivative Mean center Autoscale	1st derivative Mean center	SNV Autoscale 1st derivative
Data transformation	Cube root Autoscale	Cube root Autoscale	Cube root Autoscale	Cube root Autoscale
		-- Calibration --		
Latent variables	6	5	4	4
$\boldsymbol{R^2_{cal}}$	0.99	0.95	0.96	0.95
Bias$_{cal}$	-0.05	-0.11	-0.04	-0.05
RMSEC	2.09	1.83	1.63	0.97
RPD$_{cal}$	10.28	5.09	6.52	9.32
		-- Cross validation --		
$\boldsymbol{R^2_{cv}}$	0.98	0.87	0.94	0.92
Bias$_{cv}$	-0.10	-0.11	-0.01	-0.03
RMSECV	2.80	3.12	2.13	1.28
RPD$_{cv}$	7.70	1.75	3.82	5.33

Outliers in POC, HUM and ROC calibrations excluded based on Q residuals, leverage and Hotelling T^2
MSC (mean) = Multiplicative scatter correction
1st derivative = Polynomial order: 2, window: 15 pt
SNV = Standard Normal Variate scaling
RMSEC/RMSECV = Root Mean Square Error of Calibration/Cross Validation
RPD = Ratio of Performance to Deviation

4.12.12 Calibration and validation of SOC data derived by MIRS, Heanes and Leco CNS2000 analysis

A subset of 96 samples made up of samples from 8 BEP sites and incorporating the three positons (TL, IR and AL), and four depth increments (0-5, 5-10, 10-20 and 20-30 cm) (Subset 1, Figure 4-7) was selected for total soil C analysis utilising the DCHTC method (Raymont and Lyons 2011 method 6B2a) with a Leco CNS2000 analyser. This step was undertaken to allow correlation and validation of the relationship between the SOC results derived from samples that had been previously analysed using the Heanes and MIRS analytical methods. Linear regression analysis was used to assess the strength of the relationship for SOC results obtained from the three methods used in the current research.

4.12.13 Statistical analysis

Statistical analysis was conducted using GenStat 17th Edition 2014. ANOVA was used to compare treatment and block effects. Relationships between SOC with predictor variables including paired positions (AL, TL, IR and BEP), age (being years since establishment of BEP), age class (being 1 = BEPs aged 1-5 yrs, 2 = 6-10 yrs, 3 = 11-15 yrs and 4 = 16-19 yrs), soil depth increments (0-5, 5-10, 10-20 and 20-30 cm or 0-30 cm as appropriate), topography, pH, EC, TN, TP and nutrient ratios, tree stocking rates, planting configuration and edge effect and aspect was carried out. The significance of each variable and interactions between variables were determined and are reported. Tukeys honestly significant difference (HSD) method with an α of 0.05 was used for multiple pairwise comparisons of means when there was significant variation between positions or sites. A non-orthagonal ANOVA was applied because the data sets for some age classes, site configurations, and number of available replicates varied.

Linear regression analysis was used to compare the strength of relationships between factors and this is specified in the appropriate paragraphs contained in this case study. Probability testing of variables was conducted by using generalised linear modelling

4.13 Results

The BEPs included in the current research have been sown into soil previously used for agriculture. At the time of sampling for soil, tree measurements and litter, all the properties were being used for livestock grazing. It was assumed that all the sites were used for grazing at the time the BEPs were established because that is the predominant agricultural activity undertaken throughout the region. Results for soil and tree attributes are presented initially for the AL position, thereby establishing a baseline value to identify changes following establishment of BEPs. Results for the BEP position are presented initially for the TL and IR positions. The results for the BEP position were obtained by calculating the arithmetic mean of results for the TL and IR positions.

Normally, results are presented as a mean for BEPs each age classes as follows:

Age class 1: BEPs 1-5 years since planting,

Age class 2: BEPs 6-10 years since planting,

Age class 3: BEPs 11-15 years since planting, and

Age class 4: BEPs 16-19 years since planting.

Because most change was found to occur in the 0-5 cm soil layer, results from this depth interval in particular have been used to highlight significant differences between AL and BEP. Where appropriate, changes associated with depth increments below 5 cm have also been presented.

Unless otherwise stated error bars on charts represent the standard error of the mean (SEM).

4.13.1 Soil organic carbon (Heanes) concentration in the AL, TL IR and BEP positions

The soil C concentration for the AL position shows variation between age classes (Figure 4-9) and the ANOVA showed age class to be statistically significant (F= 12.1, P<0.001). In the 0-5 cm layer the mean SOC concentration results range between 2.5 g C 100g^{-1} in BEP sites from the 6-10 year old cohort. The highest mean in the 0-5 cm layer was in the 16-19 year old BEP cohort with 3.9 g C 100g^{-1}. Post hoc analysis using Tukey's HSD with an α of 0.05 showed that sites with BEPs aged 16-19 years have a significantly higher SOC concentration in the 0-5 cm layer than sites from age classes 1, 2 and 3. There is no significant difference in SOC concentration in the 0-5 cm layer in the AL position for BEP sites in age classes 1, 2, and 3. The results show a trend suggesting an increase in SOC in the AL position with time. The results also show there is a decreasing concentration of SOC with depth in all age classes. The mean SOC concentration in the AL position in the 0-5 cm layer was found to be 3.0g C 100g^{-1}.

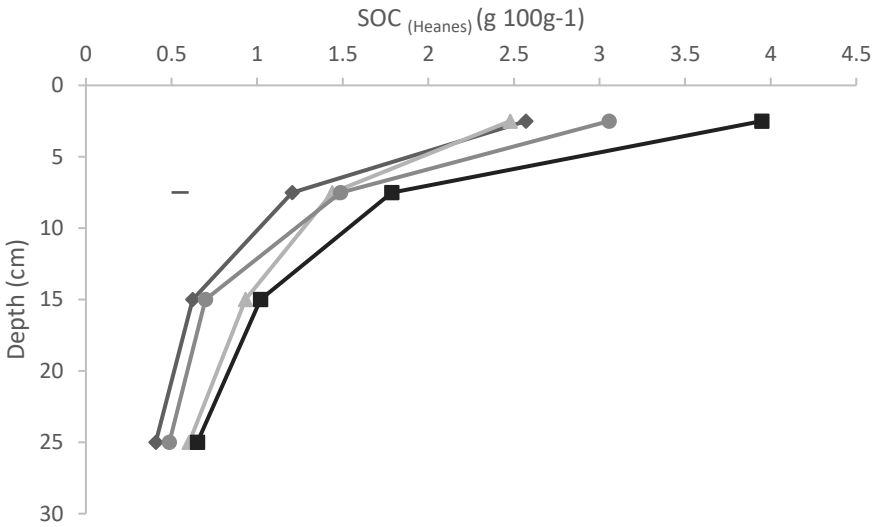


Figure 4-9. SOC concentration trend with depth for the AL position by age class. (Horizontal bar = SEM, n=112) (♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs).

The $SOC_{(Heanes)}$ results for the TL position show that the 1-5 year old BEPs have the lowest mean concentration of SOC at all depths. This age cohort has an average SOC concentration of 2.3 g $100g^{-1}$ at the 0-5 cm depth. There is significantly more SOC at the 0-5 cm depth in the TL position of BEPs aged 6-10 and 11-15 years (3.3 and 4.1 g $100g^{-1}$ respectively) compared with the 1-5 year old BEPs. The 16-19 year old sites have a lower SOC concentration than either the 6-10 and 11-15 of and 3.1 g $100g^{-1}$ in the 0-5 cm depth increment (Figure 4-10) and for all depths. At depth increments below 5 cm, the 6-10 year old cohort has the highest overall SOC concentration. The mean SOC concentration in the TL position in the 0-5 cm layer was 3.2 g $100g^{-1}$. Post hoc test analysis with Tukey's HSD (using an α of 0.05) revealed that the mean SOC concentration in the 0-5 cm depth was significantly higher than in the 5-10 cm layer. There was no statistically significant difference found for the SOC concentration of the TL in the 10-20 and 20-30 cm depths. ANOVA suggested the difference in SOC concentration in the TL by age class to be statistically significant (F=19.7, P<0.001). Post hoc test analysis using Tukey HSD also suggests that the SOC concentration in the TL for BEPs was significantly lower in sites aged 1-5 than the sites aged 6-10, 11-15 and 16-19 years. The TL in BEPs also had significantly higher SOC concentration in sites aged 6-10 and 11-15 years than sites aged 16-19 years. ANOVA suggested the results for an interaction between age and depth was also significant (P < 0.001).

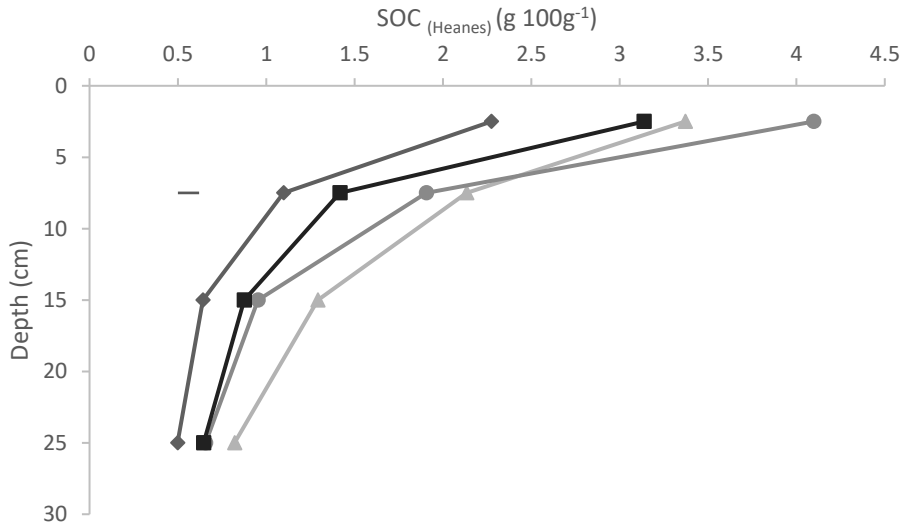


Figure 4-10. SOC concentration trend with depth for the TL position (n=153) (Horizontal bar = SEM, n=153). (♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs).

The mean SOC concentration distribution in the IR shows slightly less variation between age cohorts than seen in the TL position (range 2.8 to 4.0 g C $100g^{-1}$). The difference between age classes in the 0-5 cm layer was suggested to be significant (P = <0.001). The lowest SOC concentration for the 0-5 cm depth increment in the IR position of 2.8 g C $100g^{-1}$ was shown in BEP sites aged 6-10 years. The highest mean SOC concentration of 4.0 g C $100g^{-1}$ is seen in BEPs aged 16-19 years. The overall mean SOC concentration in IR position for the 0-5 cm layer was 3.5 g $100g^{-1}$ (Figure 4-11).

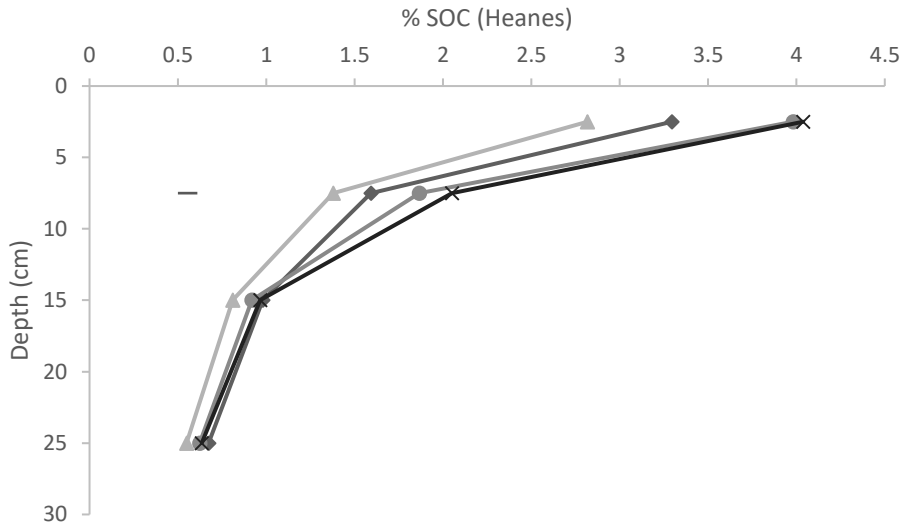


Figure 4-11. SOC concentration for the IR position (Horizontal bar = SEM, *n*=148). (♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs).

When viewed together, the results show that there is considerable variation in mean SOC concentration between the TL, IR and AL positions in the 0-5 cm depth (Figure 4-12). The AL has a higher mean concentration of SOC than the TL for the 1-5 and 16-19 year old BEPs, whereas the TL in the 6-10 and 11-15 year old cohort have a higher SOC concentration than the AL. The AL has a lower average SOC concentration than the IR in all age groups. Both the TL and IR positions have a higher SOC concentration than the AL in the 6-10 and 11-15 year old BEPs. The ANOVA suggested that age class was a statistically significant variable (F=8.84, P<0.001) as was position (F=7.79, P = 0.001). Using Tukey HSD post hoc analysis with an α of 0.05, the mean SOC concentration in the 0-5 cm layer for all age classes in the AL position is 3.0 g C 100g^{-1} which is significantly lower to the mean of the TL and IR positions which have an SOC concentration of 3.2 and 3.5 g C 100g^{-1} respectively.

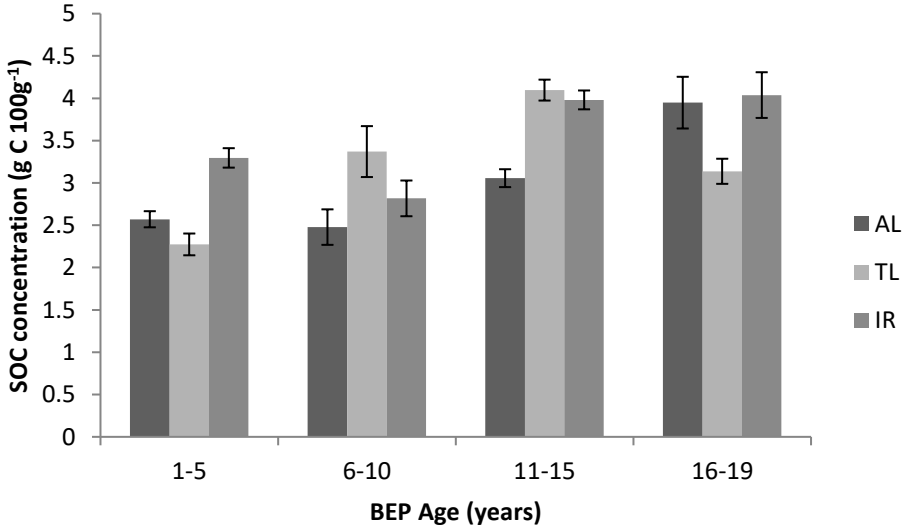


Figure 4-12. SOC concentration at the 0-5 cm layer showing variation between the AL, TL and IR positions by BEP age (error bars = SEM).

The SOC concentration results for the 0-5 cm soil layer for TL and IR position have been averaged to obtain a mean SOC concentration for the BEP, and then compared with the results for the paired age class cohort for the AL position (Figure 4-13). ANOVA suggested that position within age class was statistically significant (F=0.30, P= <0.001). The results show that the BEPs have a higher SOC concentration than the paired AL in the age cohorts between 1 and 15 years. However, BEPs in the oldest cohort (16-19 years) have a lower SOC concentration than the paired AL. The lowest SOC concentration was in the AL position for BEP sites aged 6-10 years with 2.5% (±0.2 SE). The highest SOC concentration was in BEPs aged 11-15 years with 4.1% (± 1.0 SE).

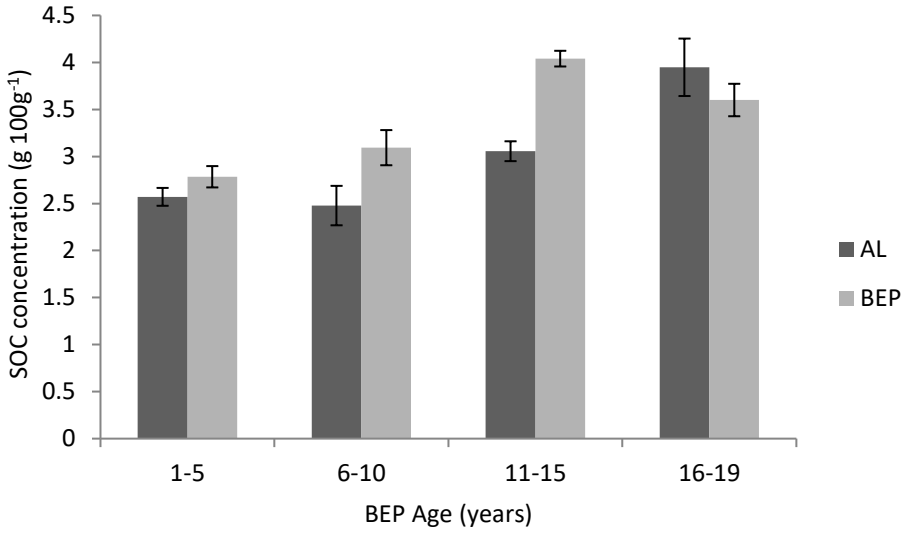


Figure 4-13. SOC concentration (g $100g^{-1}$) for the 0-5 cm depth showing difference between AL and BEP at the 0-5 cm depth (error bars =SEM).

4.13.2 SOC and effects of BEP configuration, topography and direction of planting

By comparing the effects of site configuration the results show that there is a significant difference ($P = 0.02$) in SOC concentration in the 0-5 cm layer between block (> 4 rows of trees) and linear plantings (3-4 rows of trees) (Figure 4-14). The difference in SOC between land use (AL and BEP) is significant (Tukey HSD at α 0.05) in BEPs sown into a block configuration. The interaction between configuration and land use (AL or BEP) was also suggested to be significant ($P = 0.02$). The SOC concentration in the block plantings of BEPs is higher than in the linear plantings (3.8 g $100g^{-1}$ ± 1.0 (SE) and 3.5 g $100g^{-1}$ ± 1.0 (SE) respectively). Conversely the block plantings have a lower C concentration than the linear plantings in the AL position (2.7 g $100g^{-1}$ ± 0.9 (SE) and 3.2 g $100g^{-1}$ ± 1.0 (SE) respectively).

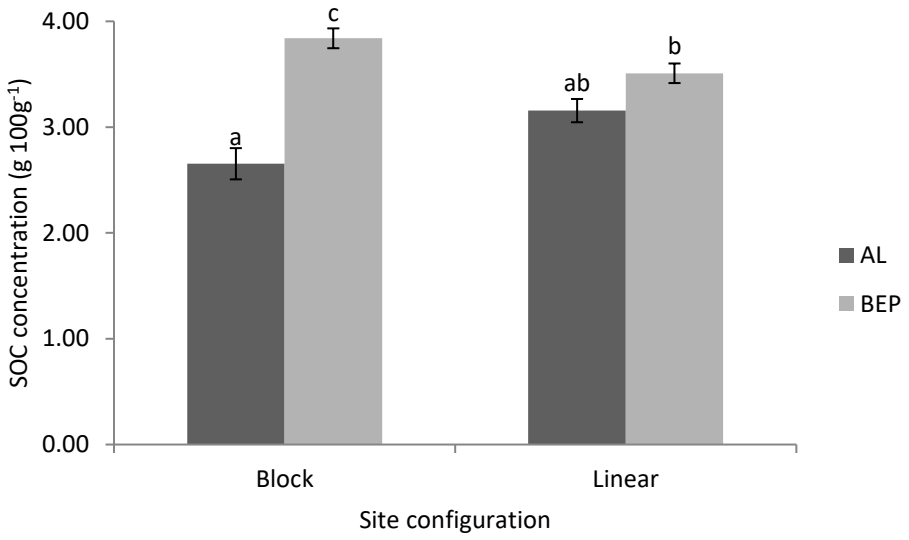


Figure 4-14. SOC concentration in the 0-5 cm layer for the AL and BEP positions (means of all age classes). Significant differences between treatments are indicated by different letters.

The data for the 0-5 cm level was used to determine any relationship between SOC sequestration and the site configuration (block vs linear plantings), land use (AL or BEP), topographic position (crest, mid slope or low) and tree row orientation of BEP planting (North-South (NS) or East-West (EW)) and their interactions (Table 4-9). The effects of configuration, land use, orientation of planting and topographic position were all suggested by ANOVA to be significant.

The interactions between configuration and land use, and between configuration and direction were also significant ($P = 0.02$ and $P < 0.001$ respectively).

Table 4-9. Effects of land use, planting configuration, topographic position and direction of planting

Change	*d.f.*	*s.s.*	*m.s.*	*v.r.*	*F pr*
Configuration	1	5.59	5.59	5.41	0.020
Land use	1	32.94	32.94	31.90	<.001
Tree row orientation	1	8.71	8.71	8.43	0.004
Topographic position	2	55.51	27.76	26.88	<.001
Configuration.Land use	1	5.65	5.65	5.47	0.020
Configuration.Direction	1	26.96	26.96	26.10	<.001
Land use.Direction	1	0.50	0.50	0.48	0.487
Configuration.topographic position	1	1.52	1.52	1.48	0.225
Land_use.Topography	2	0.49	0.25	0.24	0.789
Configuration.Land_use.orientation	1	0.21	0.21	0.20	0.653

The interaction between configuration, land use and site orientation for SOC concentration in the 0-5 cm layer was indicated to be not significant ($P = 0.65$). Block plantings sown in an EW orientation had the lowest SOC concentration, both in the AL and BEP positions (2.3 g $100g^{-1}$ and 2.5 g $100g^{-1}$ SOC respectively). Linear BEPs sown with an EW orientation had a higher SOC concentration in the BEP position (3.7 g $100g^{-1}$ SOC) than BEPs sown with a NS orientation (3.4 g $100g^{-1}$ SOC) (Figure 4-15).

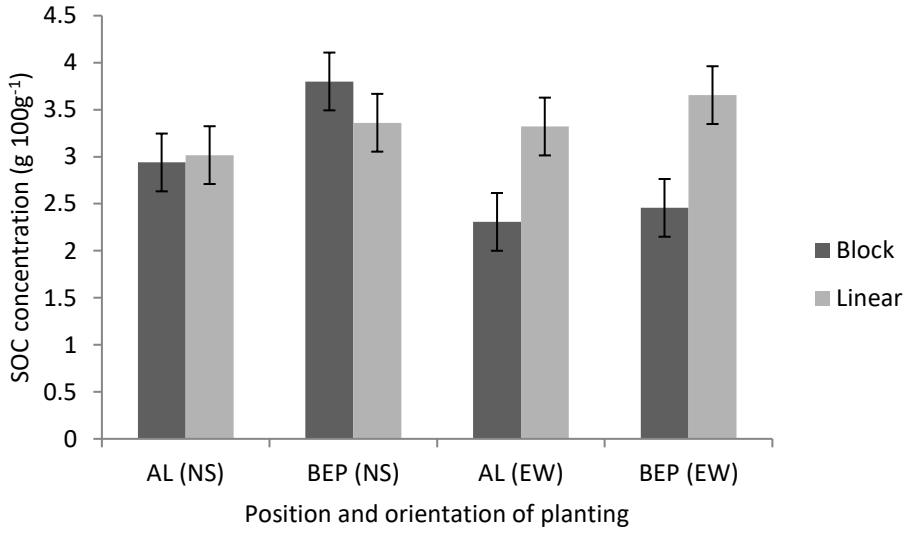


Figure 4-15. SOC concentration for AL and BEP in the 0-5 cm layer, showing relationships between configuration (block/linear) and orientation of planting (NS =north-south, EW = east-west (error bars = SEM).

4.13.3 Soil bulk density

ANOVA results for BD suggest there is a significant difference in relation to age class and depth (P = <0.001) and position (P=0.006). The interactions between age and depth, and age and position, and depth and position are also significant, but the interaction between age and position is not significant. (Table 4-10).

Table 4-10. ANOVA results for soil bulk density.

Source of variation	**d.f**	**s.s**	**m.s.**	**v.r**	**F pr**
Age_class	3	5.89	1.6	54.86	<.001
Depth	3	30.22	10.07	281.59	<.001
Position	2	0.37	0.18	5.15	0.006
Age_class.Depth	9	1.04	0.12	3.24	<.001
Age_class.Position	6	0.44	0.07	2.06	0.055
Depth.Position	6	0.81	0.12	3.76	0.001
Age_class.Depth.Position	18	0.84	0.05	1.3	0.178
Residual	786	28.12	0.04		
Total	833	67.63			

The mean soil BD by soil depth increment for the BEP position (mean of TL and IR) for each age class, and the mean of all AL position age classes combined (Figure 4-16) has is a trend showing BD is higher with depth, and a trend showing BD decreases with BEP age in the 0-5 cm soil layer. The AL and BEPs from the 1-5 year old cohort were found to have the highest soil BD in the 0-5 cm depth layer with 1.12 and 1.14 Mg m^{-3} for AL and 1-5 year old BEP respectively. Overall, BEPs (being the mean of TL and IR) have a temporal trend whereby the mean BD decreases with BEP age (1.2 Mg m^{-3} in the 1-5 year old BEP cohort, decreasing to 0.9 Mg m^{-3} in the 16-19 year old cohort).

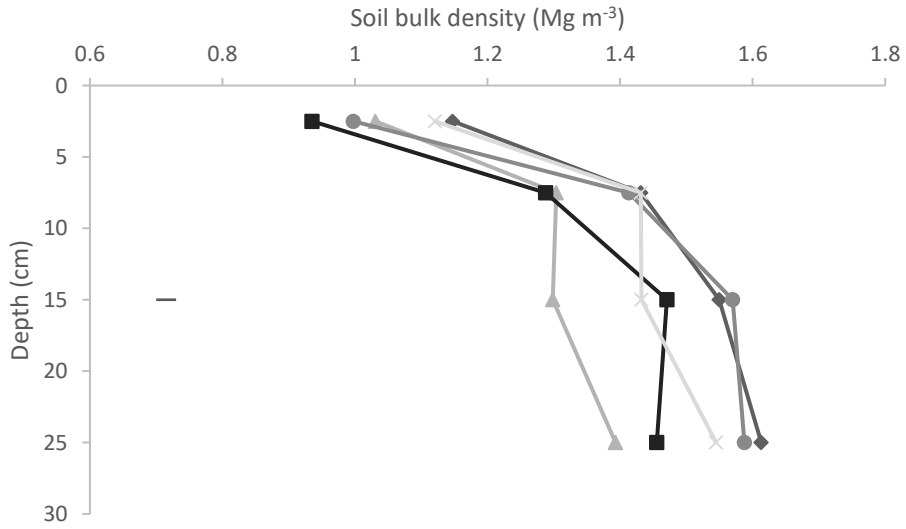


Figure 4-16. Soil bulk density trend with depth comparing differences (P=0.001) for the BEP position (mean of TL and IR) for the four age classes, and the overall mean for the AL position, by depth (Horizontal bar = SEM). (× = AL; ♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs)

When the results for the AL by age class are analysed individually there is no particular trend with age apparent (Figure 4-17). The AL in the 16-19 year old cohort have the lowest BD of 0.9 Mg m^{-3} and the 11-15 year old cohort have the highest BD of 1.26 Mg m^{-3}. When the TL and IR positions are analysed separately the results show there are no apparent trends between positions identifiable. The difference between TL and IR varies between age class with TL having a higher BD than IR in the 1-5 and 16-19 year old BEPs, but lower in the 6-10 and 11-15 year old BEPs.

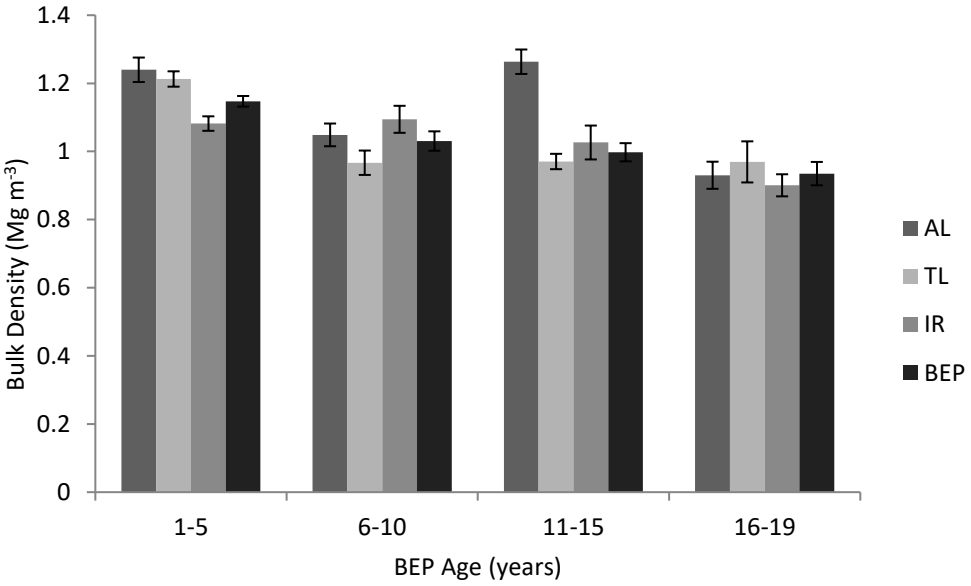
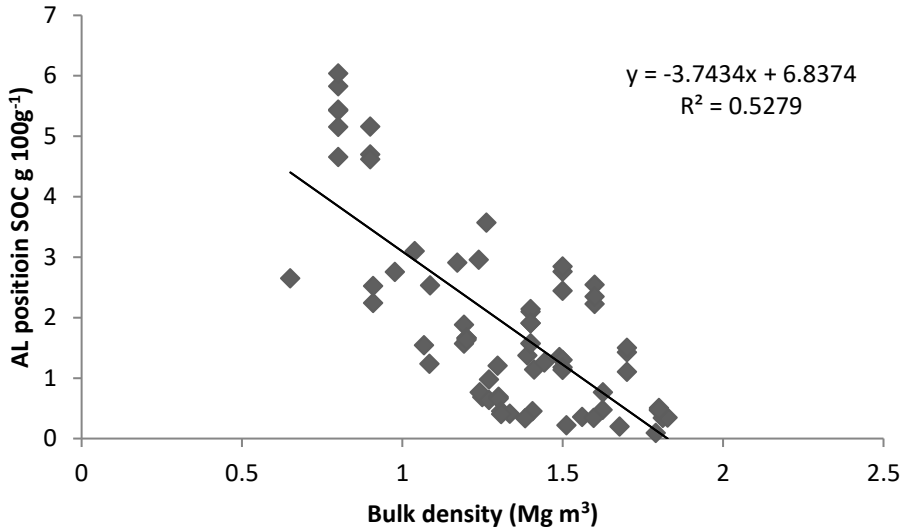


Figure 4-17. Soil bulk density for the 0-5 cm soil layer showing AL, TL, IR and BEP (being the mean of TL and IR) (error bars = SEM).

This study has found there is an inverse relationship between soil bulk density and SOC for both the AL position (Figure 4-18 (a)) and for BEP soils (Figure 4-18 (b)). Correlation analysis showed a moderate negative correlation R^2 = 0.53 and 0.48 for AL and BEP respectively.

(a)



(b)

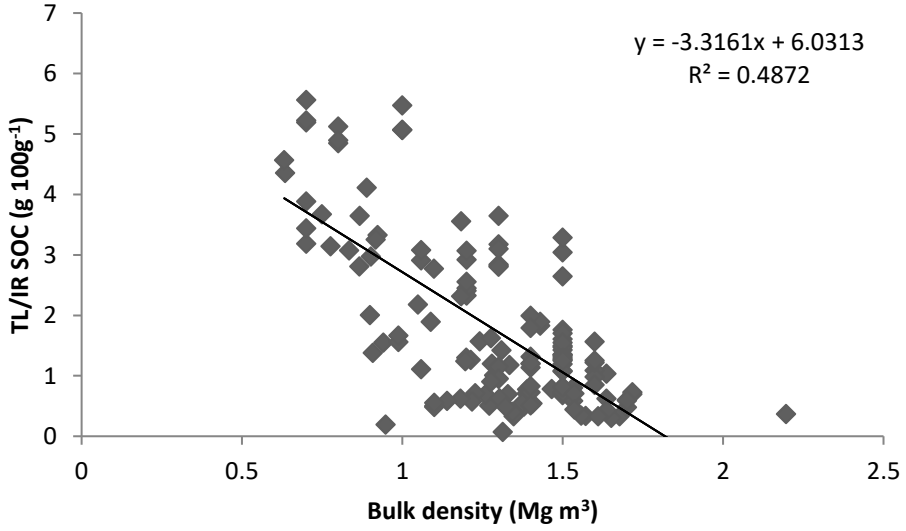


Figure 4-18. Bulk density and SOC concentration in AL (a) and BEP (b); showing an inverse relationship between BD and SOC**.**

The data point located at position 2.2, 0.4 in Figure 4-18(b) above is likely to be an outlier due to soil compaction caused by wildlife or sampling error. Gravel percentage for this point was 2.2 % so is unlikely to be a factor.

4.13.4 Soil organic carbon stock

The SCS to 30 cm depth was first calculated on a fixed depth basis (SCS_{FD}) for the fine fraction (< 2 mm) for the TL, IR and AL positions. There was no particular trend apparent associated with position. In three of the four age cohorts the IR was found to have a higher SCS than either the TL or AL (Figure 4-19). The exception was the 6-10 year cohort where TL had the highest SCS.

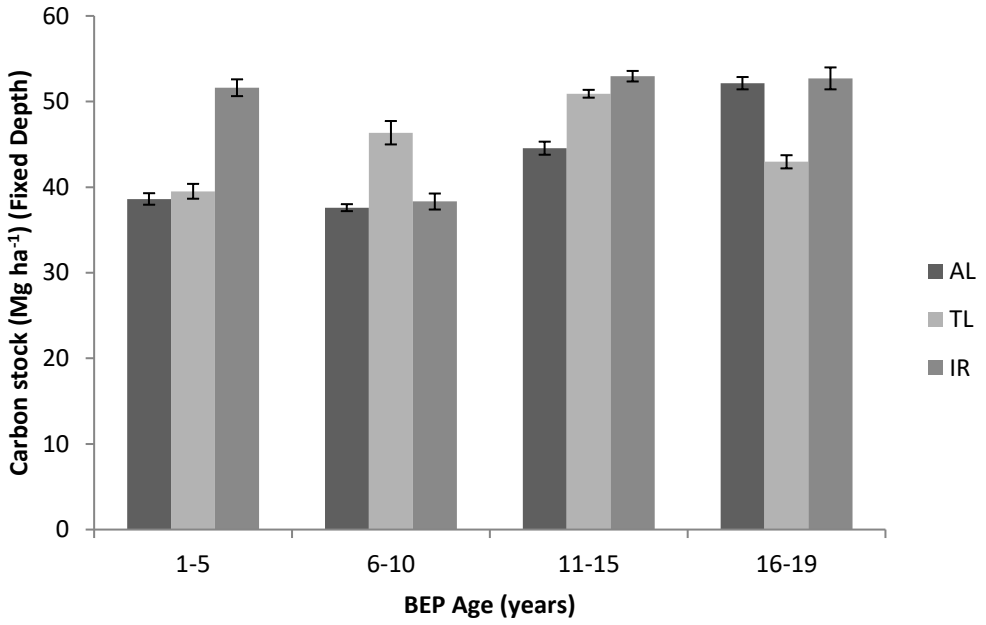


Figure 4-19. SCS $_{(FD)}$ to 30 cm showing the stocks measured for the AL, TL and IR positions by age class. (Error bars = SEM).

The SCS_{FD} results for TL and IR were averaged to give a total BEP SCS_{FD} to 30 cm depth for each age group and each depth increment (Figure 4-20). These results were then compared with the mean age class paired AL results. The results show there is not a linear increase in SCS over time aligned with BEP age. The BEPs have a higher average SCS than AL in sites up to 15 years of age, but in the oldest cohort the SCS is lower in the BEPs than the AL (47.9 Mg C ha^{-1} compared with 52.1 Mg C ha^{-1} respectively). Sites in the age group 6-10 years have the lowest CS_{FD} in the BEP and AL positions (42.3 Mg C ha^{-1} and 37.6 Mg C ha^{-1} respectively).

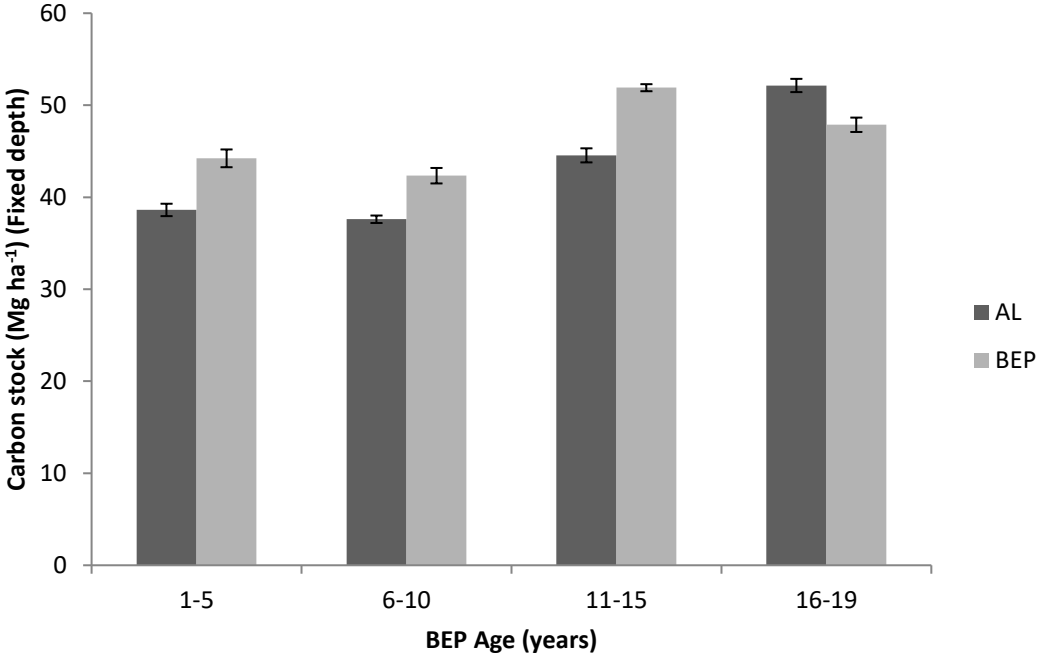


Figure 4-20. SCS $_{(FD)}$ to 30 cm depth showing difference between AL and BEP by age class (error bars = SEM).

Having calculated to SCS results based on a fixed depth, the SCS results for all sites and positions were then adjusted to an ESM (SCS_{ESM}) with the AL. The BEP SCS_{ESM} has been compared with the AL SCS_{ESM} (Figure 4-21). These results show that when all sites are combined the overall average SCS_{ESM} in the AL is 41.1 Mg C ha^{-1}. ANOVA results suggest a significant difference between AL and BEP SCS_{ESM} results (P = 0.037). Age was suggested to be a significant factor for SCS in BEPs (*P* = 0.001).

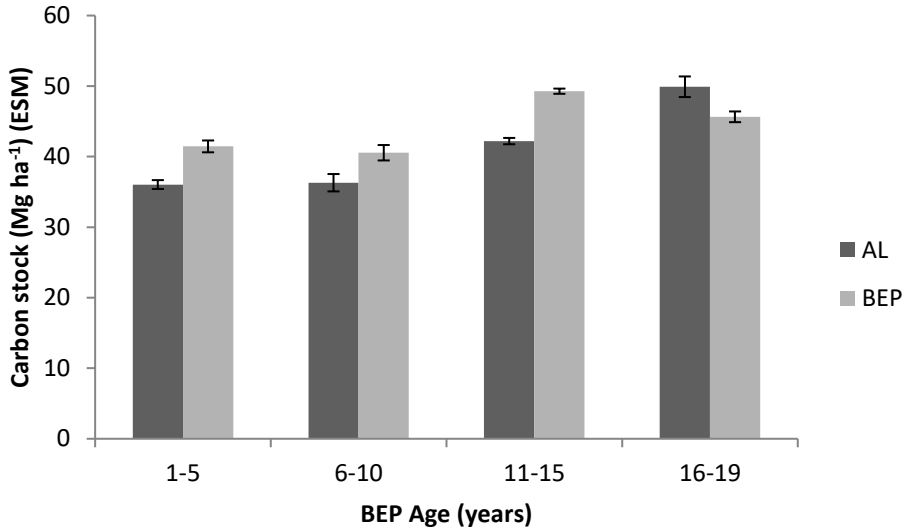


Figure 4-21. $SCS_{(ESM)}$ (Mg ha^{-1}) to 30 cm comparing results for the paired age cohort AL position with results for BEP position (error bars = SEM).

The overall average SCS_{ESM} for BEPs of all ages was found to be 44.2 Mg C ha^{-1}. On average, BEPs were found to have a higher SCS than the paired AL position in BEPs up to 15 years of age. However, in the BEPs in the 6-10 year old cohort, the mean SCS was found to be slightly less than that found for the overall average in the AL position (40.6 and 41.1 Mg C ha^{-1} respectively). The highest CS_{ESM} was found in the 11-15 year old BEP cohort with a mean of 49.3 Mg C ha^{-1}. The oldest BEPs were found to have an average CS_{ESM} of 45.7 Mg C ha^{-1}, which is 4.3 Mg C ha^{-1} lower than the paired AL position but 4.6 Mg C ha^{-1} more than the SCS of the overall mean AL position.

The SCS based on an ESM basis for the AL position from all sites was compared with the SCS for the paired BEPs using regression analysis (Figure 4-22). A weak correlation was achieved by linear regression (R^2 = 0.34, P<0.001), suggesting that overall BEPs tend to have a higher SCS than AL, but not always.

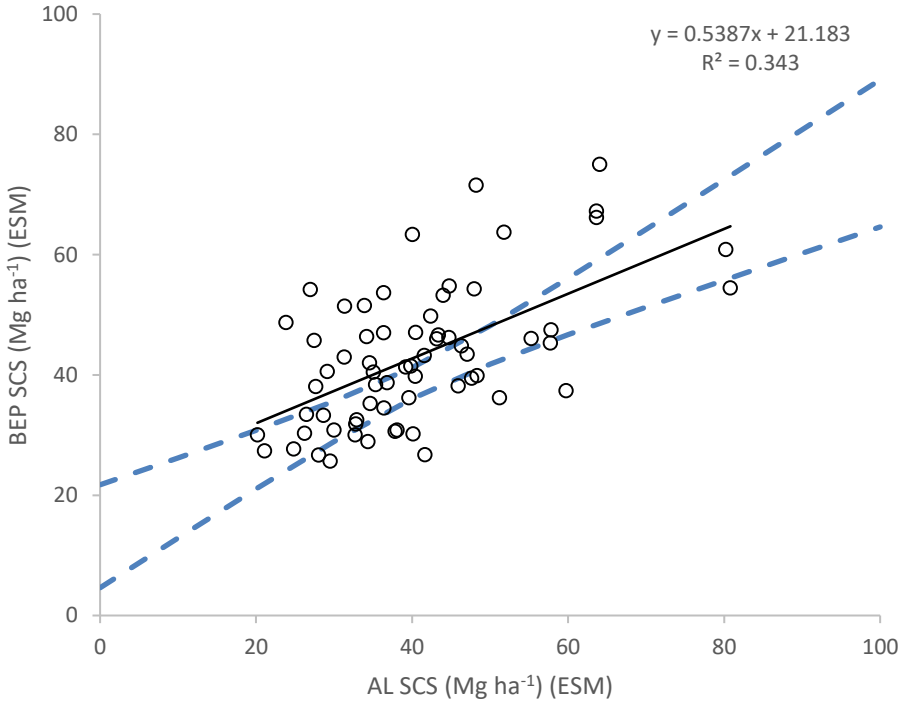


Figure 4-22. Linear regression analysis comparing SCS (ESM) for the AL position with the BEP position at all sites.

4.13.5 Rate of soil C stock change

A comparison between the SCS to 30 cm depth for the paired AL and BEP (n = 77) positions has been used to provide an indication of change on a per site basis (Table 4-11). Results for both fixed depth and ESM SCS is provided for comparison. The Δ represents the difference between the BEP and its paired AL. Sites with a Δ in red indicate the BEP has a lower SCS than the paired AL. The results show that on average BEPs achieve a net increase of 3.54 Mg C ha^{-1} based on ESM results. However, the range of SCS change varies between sites, with the highest increase found to be 16.6 Mg C ha^{-1} and the lowest a loss of -8.33 Mg C ha^{-1} (Table 4-11).

The average annual rate of change in SCS was found to be 0.5 Mg C ha^{-1} yr^{-1} on both fixed depth and ESM basis. The mean annual rate of change when calculated on a fixed depth basis was found to be 0.5 Mg C ha^{-1} yr^{-1}, and the range from a loss of -1.2 Mg C ha^{-1} yr^{-1} to a gain of 4.9 Mg C ha^{-1} yr^{-1}. When calculated on an ESM basis the mean annual rate of change is still 0.5 Mg C ha^{-1} yr^{-1}, with the range from an increase of 4.7 Mg C ha^{-1} yr^{-1} to a loss of -1.7 Mg C ha^{-1} yr^{-1}.

Table 4-11. SCS trends (Mg C ha^{-1}) to 30 cm depth comparing fixed depth (FD) and equivalent soil mass (ESM) results on a site basis (- equals an SCS loss).

Site	*BEP Age (yrs)*	*Δ SCS (Mg C ha^{-1}) FD*	*Δ SCS (Mg C ha^{-1}) ESM*	*Rate of change (Mg C ha^{-1} yr^{-1}) FD*	*Mean annual rate of change (Mg C ha^{-1} yr^{-1}) ESM*
anb	1	-1.2	-1.7	-1.2	-1.7
keb	2	9.8	9.3	4.9	4.7
sha	3	5.0	4.4	1.7	1.5
woo	4	2.8	2.8	0.7	0.7
ter	6	6.7	5.5	1.1	0.9
sc	9	4.5	6.0	0.1	0.7
wic	9	5.5	3.5	0.6	0.4
ws	9	-1.7	3.2	-0.2	0.4
cap	11	-4.3	-3.0	-0.4	-0.3
car	11	-2.5	4.4	-0.2	0.4
kea	12	0.9	2.1	0.1	0.2
shb	12	18.4	16.6	1.5	1.4
ana	13	-3.1	-1.6	-0.2	-0.1
str	15	5.9	8.1	0.4	0.5
wia	15	11.1	8.1	0.7	0.5
wib	15	4.9	7.4	0.3	0.5
haa	16	3.3	3.7	0.2	0.2
hab	16	1.8	2.7	0.1	0.2
spr	18	-2.3	-2.0	-0.1	-0.1
pyl	19	-10.2	-8.3	-0.5	-0.5
Mean		2.8	3.5	0.5	0.5

Using the overall average AL $CS_{(ESM)}$ result of 41.1 Mg C ha^{-1} as representing the baseline SCS to 30 cm depth prior to BEP establishment, the annual rate of SCS increase in BEPs is 0.5 Mg C ha^{-1} yr^{-1} over 15 years (Figure 4-23). However, when BEPs from the 16-19 year old cohort are included in the calculation the change in SCS was found to be lower, with an annual increase of 0.2 Mg C ha^{-1} yr^{-1}. ANOVA and post hoc analysis with Tukey HSD at α 0.05 suggest there was no significant difference for SCS between the overall mean for AL and each of the BEP age classes (F=1.678, P=0.159).

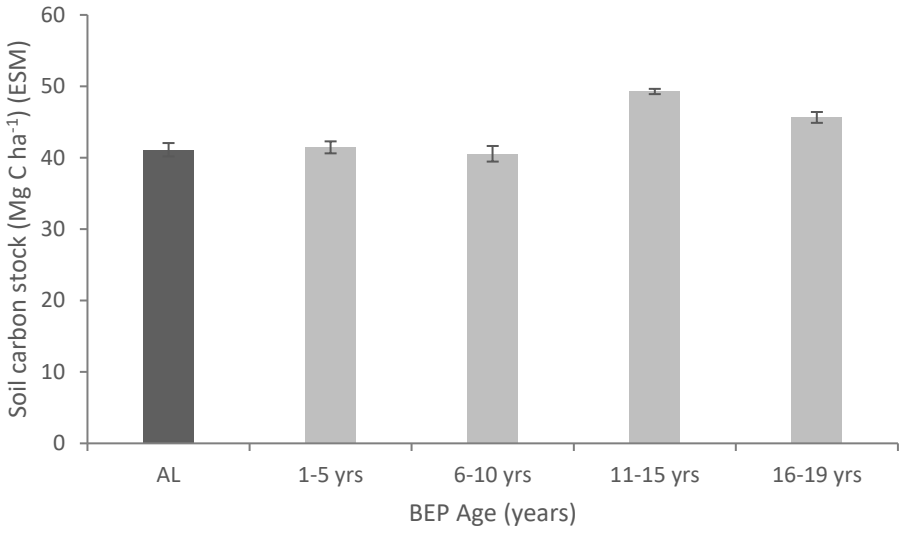


Figure 4-23. Changes to SCS over time using AL as the baseline (error bars = SEM). Dark bar = AL, grey bars = BEP age cohort.

4.13.6 Soil C stock change as influenced by planting configuration, orientation of sowing, edge effect and topographic position

Because BEP configuration, orientation of sowing the tree row, and time of planting all have the potential to influence SCS, each of these factors has been examined.

BEPs planted in a linear configuration (n = 12 sites; 41 paired transects), were found to have a significantly (P = 0.003) higher $SCS_{(ESM)}$ to 30 cm depth than BEPs planted in a block configuration (n = 8 sites; 24 paired transects). The corresponding AL positions also had statistically significantly more $SCS_{(ESM)}$ (Figure 4-24). Therefore, planting configuration was found to have a significant effect on SCS. The difference in SCS_{ESM} between positions (AL and BEP) was found to be significant (P = 0.04) and configuration (block or linear) highly significant (P = 0.003). The interaction between position and configuration was not statistically significant.

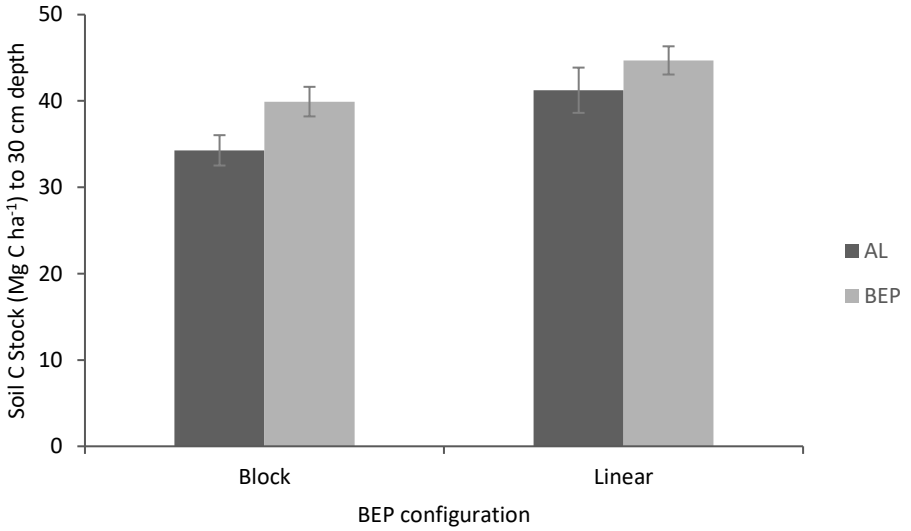


Figure 4-24. A comparison of the SCS in block and linear configured BEPs and relationship with paired AL position. (Error bars = SEM).

When the mean SCS_{ESM} for all $n = 77$ BEP transects regardless of age was calculated based on the direction of sowing; BEPs sown in a NS direction had a significantly higher ($P = 0.02$) mean SCS than BEPs sown in an EW direction (44.9 Mg C ha^{-1} and 38.9 Mg C ha^{-1} respectively) (Figure 4-25). Orientation of tree row has been found to have a significant effect on SCS. Regardless of orientation of the BEPs, linear plantings had a higher SCS than block plantings.

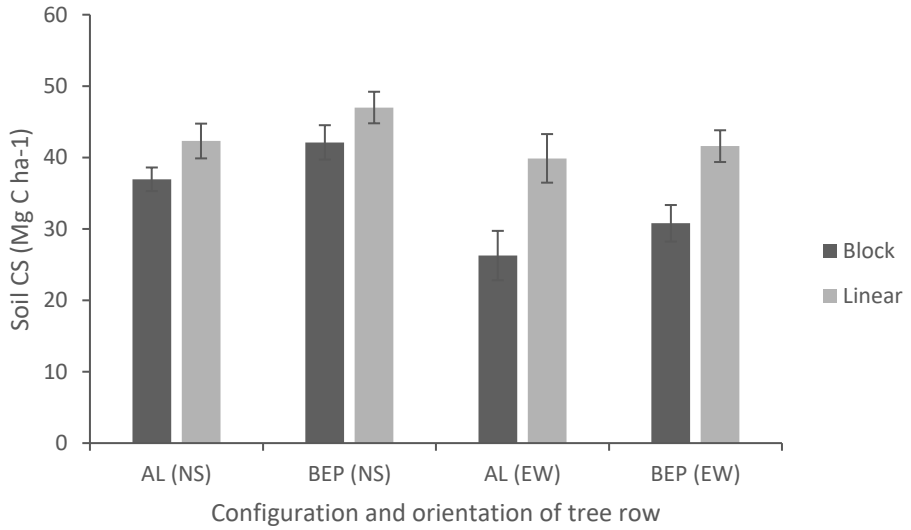


Figure 4-25. SCS and changes occurring due to direction of planting and BEP configuration (error bars = SEM).

The influence of the "edge effect" on SOC sequestration associated with BEPs was determined by comparing the SCS in soil sampled from transects located on inner or outer rows of the BEPs. Edge effect was found to be to be not significant ($P = 0.89$). The mean SCS_{ESM} for transects located on inner rows was 43.2 Mg C ha^{-1} and 41.7 Mg C ha^{-1} for transects located on edge rows.

The relationship between SCS and topographic position of BEPs, at either the crest, mid-slope and low in the catena was also examined. The results show the mean SCS for BEPs located on crests had 43.6 Mg C ha^{-1}, mid-slope 43.8 Mg C ha^{-1} and those located low in the catena had 44.4 Mg C ha^{-1}. Consequently topographic position was found to not have a significant effect on soil C sequestration following establishment of BEPs.

4.13.7 Above-ground tree biomass measurements

The species distribution of all the 1 641 stems in the 77 TL transects is shown at Table 4-12. The numbers include stems measured at DBH (1.3m), D30 (0.3m) and D10 (0.1m) depending on the size of the trees. The ratio of live eucalypt to acacia trees was 1.2. The death rate in the acacia species was higher than for the eucalypts, although it was difficult to identify the species of 263 dead stems.

Table 4-12. Summary of the species distribution for live and dead stems found in the 77 TL transects.

Species	*Live*	*Dead*
Eucalypt	625	23
Acacia	521	66
Shrub	143	
Unknown (Eucalypt/Acacia)		263
Total	1289	352

The frequency distribution for the 551 live eucalypt trees measured at 130 cm (DBH) and having a stem diameter ≥ 2.0 cm across all sites is shown at Figure 4-26. Trees measured at D30 or D10 have not been included. Up to 246 (44 %) of the eucalypt trees measured at DBH, had a stem diameter at DBH of 2 cm or less and 102 (18 %) had a stem diameter of between 2 and 4 cm.

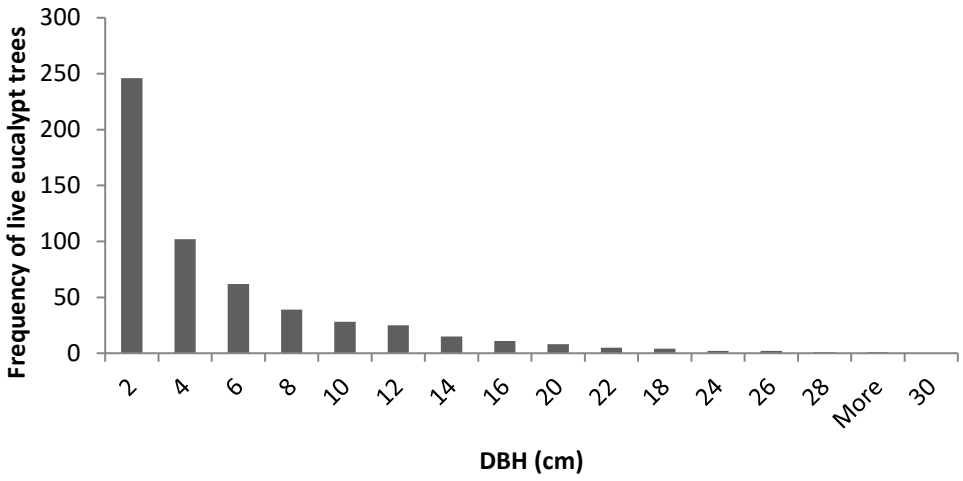


Figure 4-26. Frequency distribution of eucalypt tree stem diameters measured at 1.3 m

The frequency distribution for acacia trees shows that of the 356 live stems measured at DBH (1.3m), 130 or 36 % had a stem diameter of 2 cm or less. 88 live acacia stems, or 24 % had a stem diameter between 2 and 4 cm (Figure 4-27).

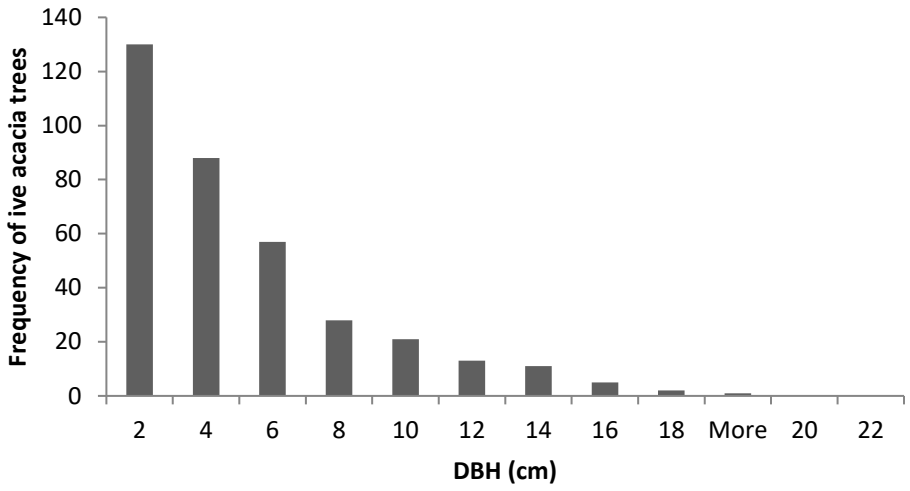


Figure 4-27. Frequency distribution of the stem diameters of live acacia trees measured at DBH (1.3 m).

The 77 BEP transects examined in the current research had a mean stocking rate of 6 375 live stems ha^{-1} with a range between 493 – 26 879 stems ha^{-1} (Figure 4-28). The stocking rate varied between sites with no particular trend apparent with BEP age.

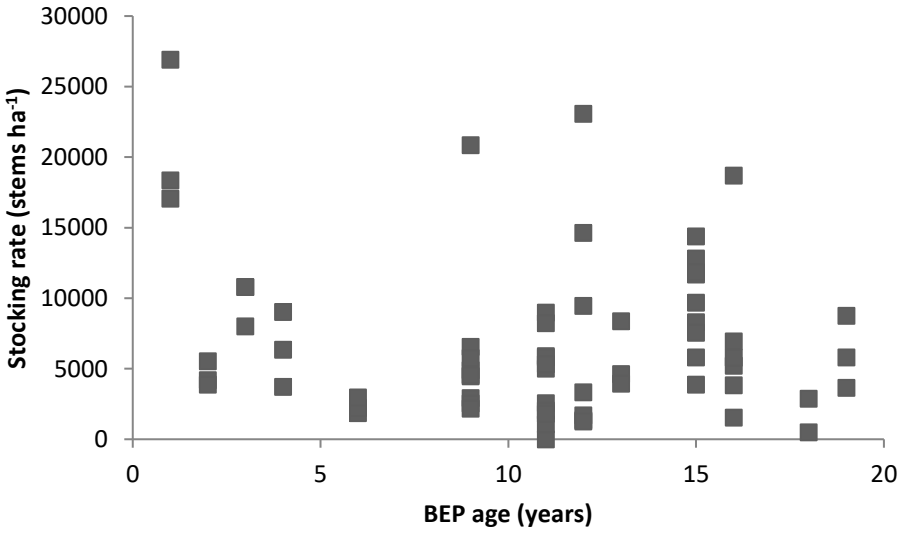


Figure 4-28. Tree stocking rate (stems ha^{-1}) by BEP age for each transect measured. Points represent stocking rates for each individual transect.

When the total live tree stocking rate is averaged for each of the four age classes, the youngest sites from 1-5 years had the highest average tree stocking rate of 10 380 stems ha^{-1} (Figure 4-29) and the oldest BEP sites (16-19 years) had the lowest average tree stocking rate of 5 433 stems ha^{-1}.

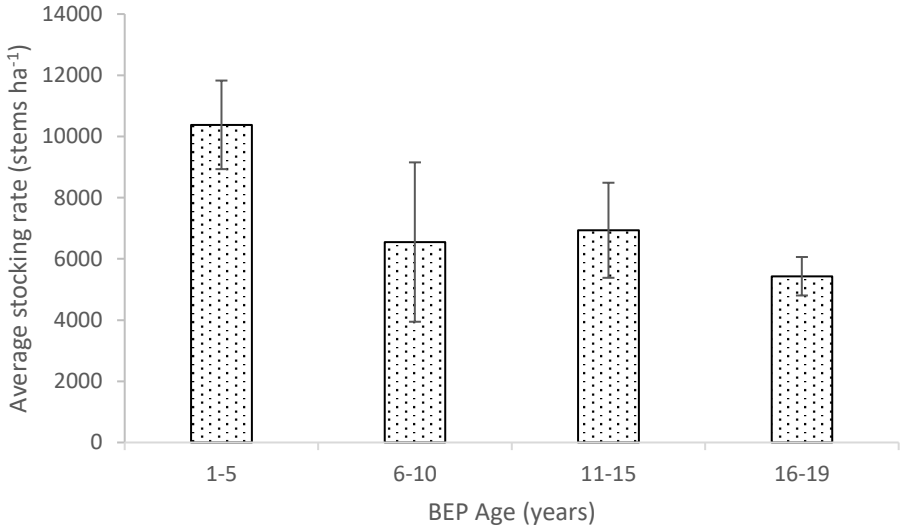


Figure 4-29. Mean live tree stocking rate (stems ha^{-1}) by age class (error bars = SEM).

The live tree stocking rate for BEPs across all age classes with a linear configuration is significantly higher ($P < 0.001$) than for plantings sown in a block configuration plantings (mean 9 239 stems ha^{-1} and 2 753 stems ha^{-1} respectively) (Figure 4-30).

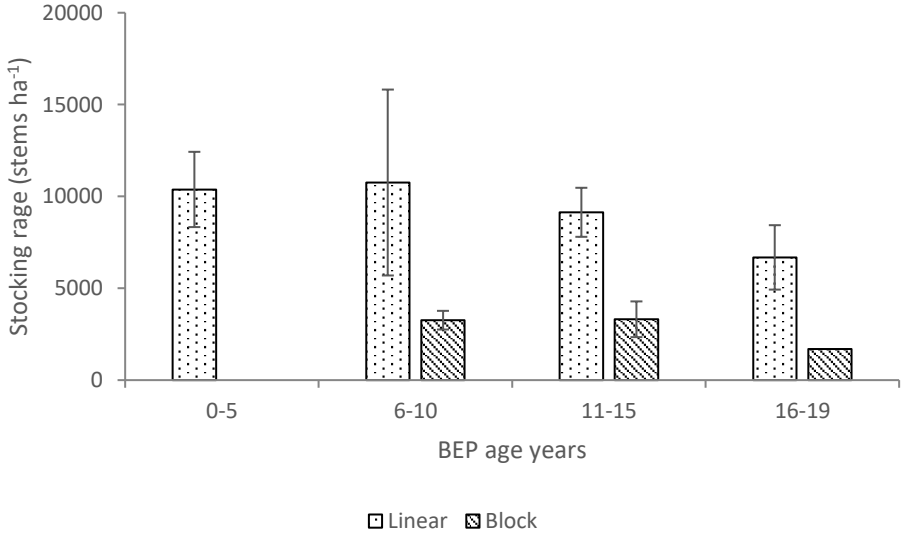


Figure 4-30. Tree stocking rate (live stems ha^{-1}) for block and linear configured BEPs (error bars = SEM).

4.13.8 Allometric measurements for above and below ground biomass C

Allometric equations for generic species and species specific predictions applicable across sites was applied to measured DBH results and used to predict tree biomass for acacia and eucalypt species. The resulting relationship is shown at Figure 4-31. The results show that acacia trees have a larger biomass than eucalypt trees per DBH.

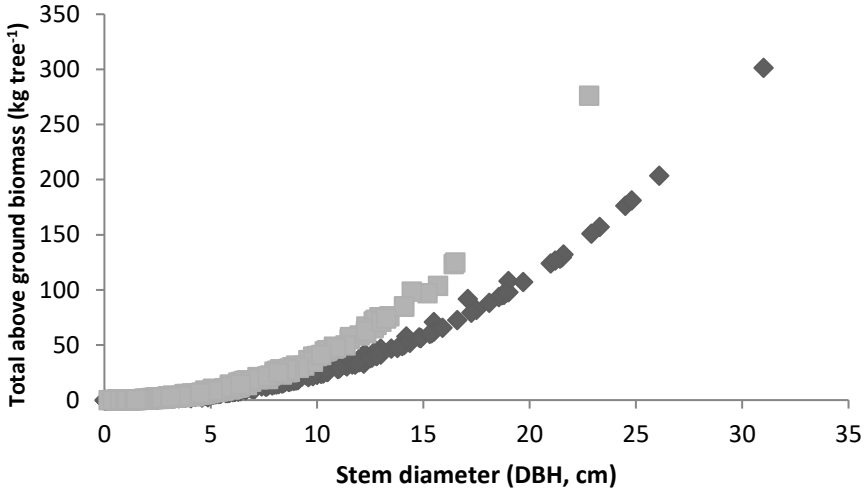


Figure 4-31. Relationship between predicted biomass (kg tree $^{-1}$) and measured stem diameter (DBH) for eucalypt (♦) and acacia (■) trees.

Live AGB C results show that there is very little tree biomass C in sites up to six years of age (mean total AGB is 0.8 Mg C ha^{-1}). The live AGB results for sites established for nine and more years shows significant variation between sites and ages (average 35.7 Mg C ha^{-1}; range between 1.4 and 97.0 Mg C ha^{-1}). The results show there is a very weak correlation for biomass C increase with age since establishment R^2= 0.26 (Figure 4-32).

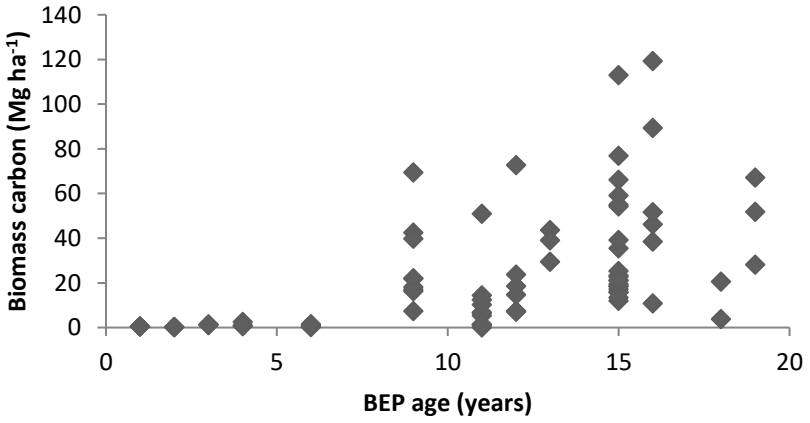
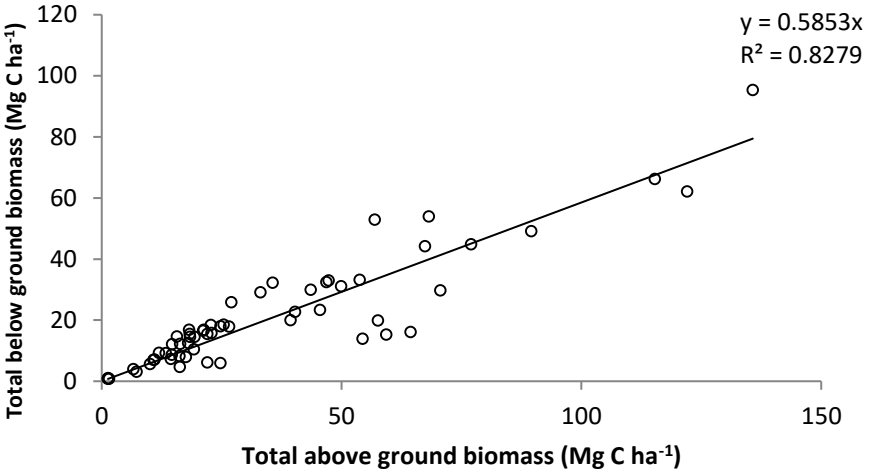
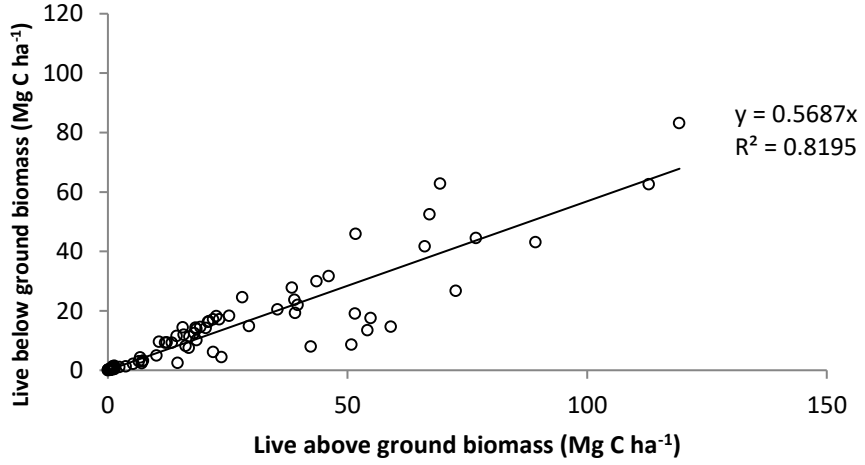


Figure 4-32. Live above-ground biomass C (Mg C ha^{-1}) for BEPs aged 1-19 years.

There is a strong positive correlation between total AGB and below ground biomass (R^2 – 0.83) and live AGB and BGB (R^2 = 0.82) (Figure 4-33). Similar correlations exist between other pools of biomass C derived through allometric equations.



(a)



(b)

Figure 4-33. Strong correlation between (a) total above ground and total below ground biomass, and (b) live above ground and live below ground biomass.

When examined on an age basis the results show no clear trend for the change in live AGB C over the 19 year chronosequence. The average total AGB for all sites aged between 1 and 19 years is 31.4 Mg C ha^{-1} with the range between 0.2 Mg C ha^{-1} in a two year old site (Site 2), and 97.0 Mg C ha^{-1} in a BEP site aged 16 years (Site 18) (Figure 4-34).

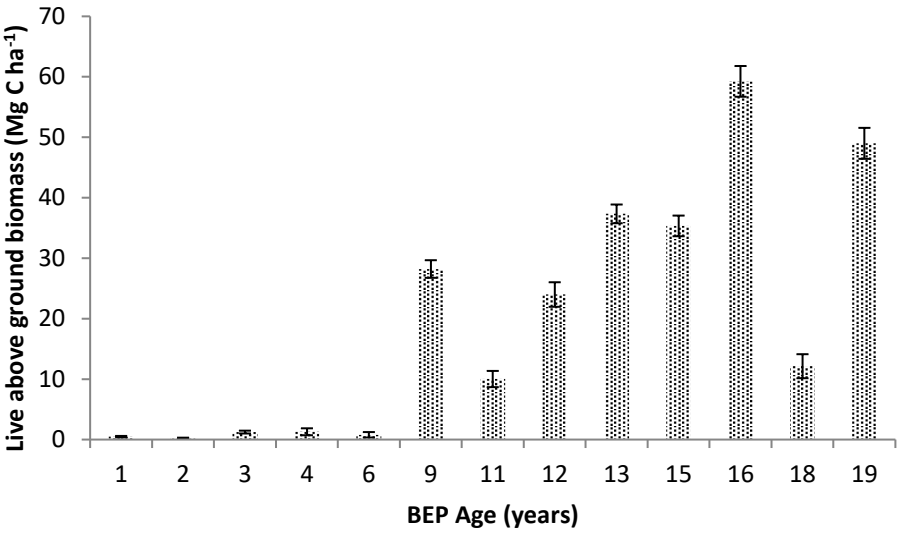


Figure 4-34. Live above ground biomass C by BEP age (error bars = SEM).

When the sites are allocated to their respective age class the site variability is hidden and there appears to be a temporal increase in live AGB C with BEP age (Figure 4-35). BEPs aged between one and five years have a mean total AGB of 0.8 Mg C ha^{-1}. The oldest BEPs, aged between 16 and 19 years have a mean AGB C of 49.3 Mg C ha^{-1}.

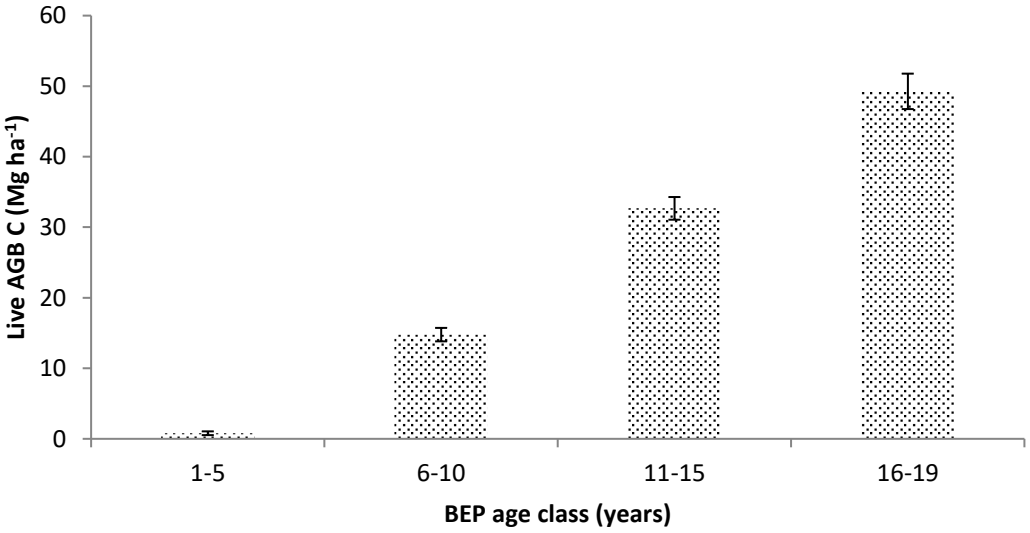


Figure 4-35. Average total (live + dead) AGB C for the four age classes (error bars = SEM).

The average annual rate of increase in AGB C was found to be 2.4 Mg C ha^{-1} yr^{-1}. The lowest rate of change was 0.1 Mg C ha^{-1} yr^{-1} in a two year old site (Site 2), and the highest was 6.1 Mg C ha^{-1} yr^{-1} in a 16 year old BEP (Site 18) (Figure 4-36).

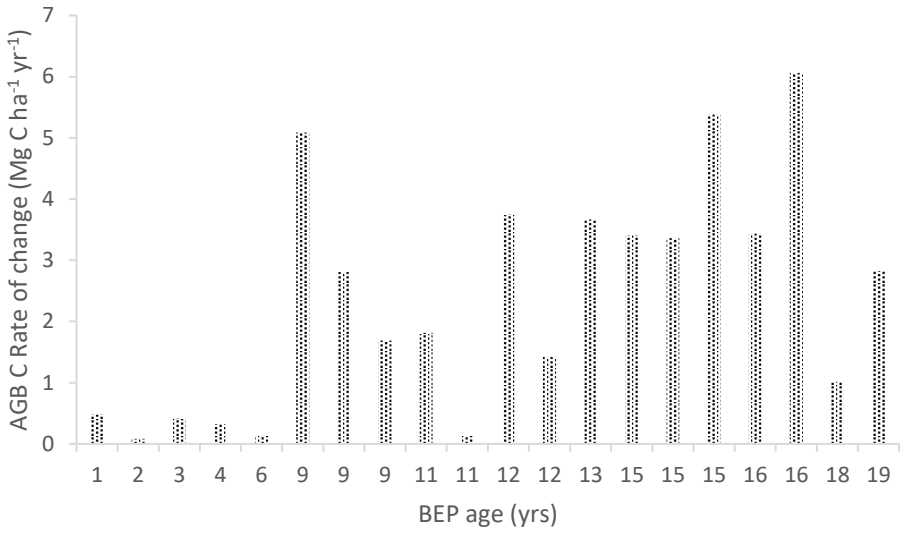


Figure 4-36. Above ground biomass showing annual rate of change by BEP age (Mg C ha^{-1} yr^{-1}) for each site.

4.13.9 Edge effect on eucalypt and acacia growth

The results for edge effect on trees in both block and linear configurations (Figure 4-37) show that eucalypt trees located in the outer rows have a larger average DBH (4.7 cm) than trees located in the inner rows (4.4 cm). Acacias have a slightly smaller average DBH than eucalypts. Similar to eucalypts, acacias located on outer rows have a slightly larger mean DBH (4.2 cm) compared with those located on inner rows (4.1 cm).

When analysed for live AGB and tree stocking rate, the edge effect was found to be not significant ($P = 0.162$ and $P = 0.847$ respectively).

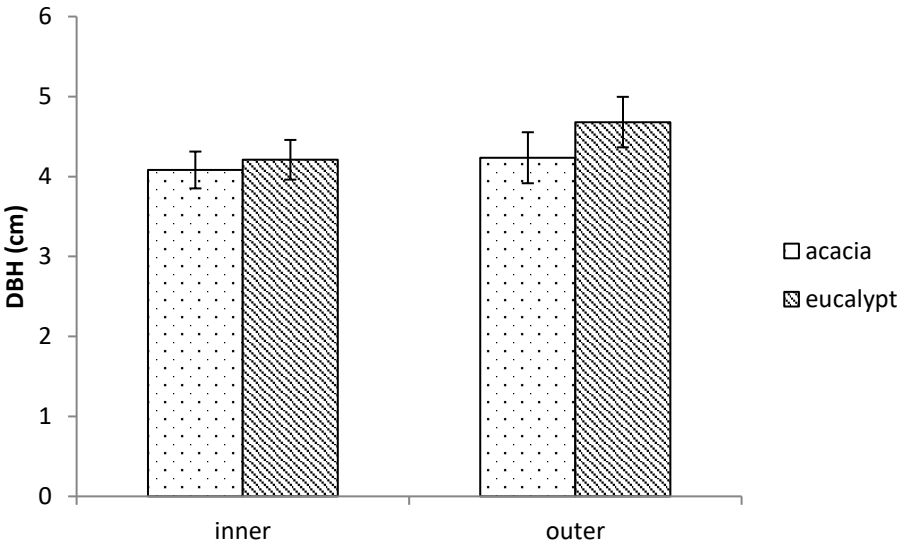


Figure 4-37. Edge effect, showing trees located on outer rows have a larger DBH than trees located in inner rows (error bars = SEM).

4.13.10 Above ground biomass in relation to planting configuration

Linear plantings were found to have more AGB (contribution from both eucalypt and acacia spp.) than block plantings (overall mean 34.0 and 15.8 Mg C ha^{-1} respectively) (Figure 4-38). The results show that both age and the configuration of the planting either block or linear are significant ($P < 0.001$) factors associated with the AGB of the sites.

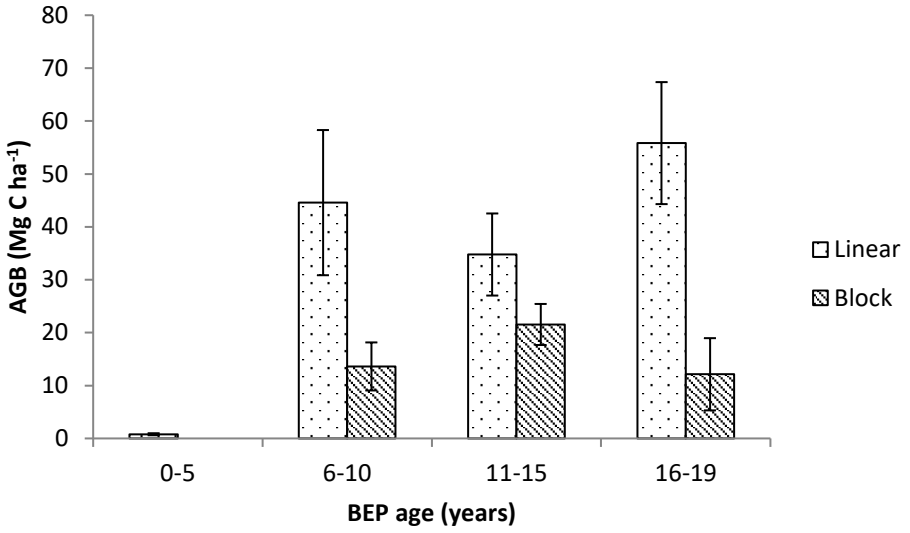
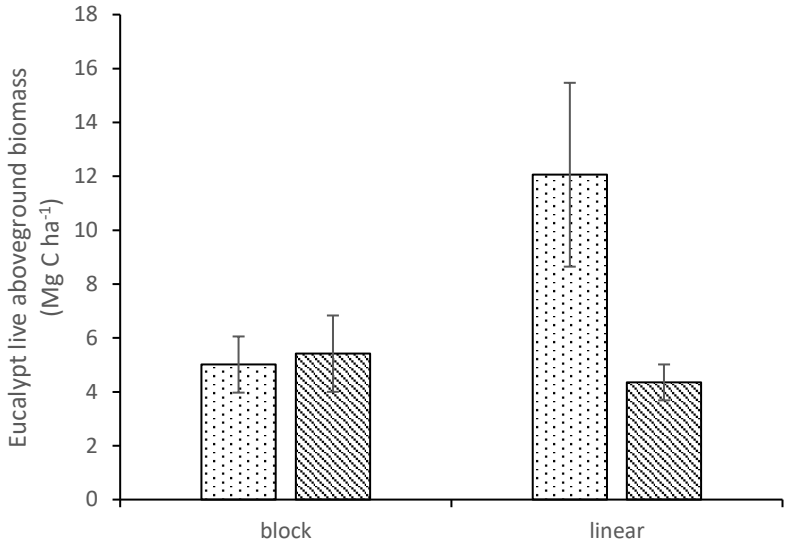


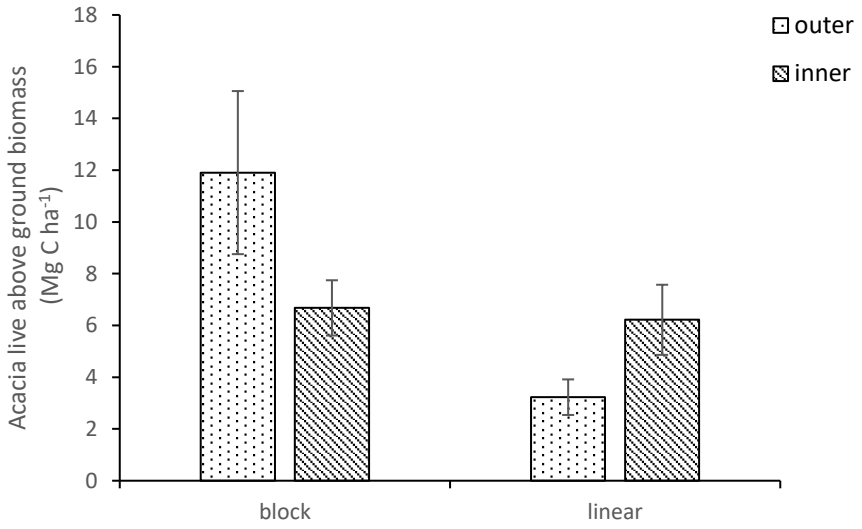
Figure 4-38. Results showing site configuration for aboveground biomass growth with linear sites having more AGB than block configured sites in each age class (error bars = SEM).

With the interaction between site configuration and tree location on either inner or outer rows there was found to be a species effect (Figure 4-39). Eucalypts had a higher AGB when located on the outer row of linear plantings than when located on the inner row (12.1 Mg C ha^{-1} and 5.1 Mg C ha^{-1} respectively). There was little edge effect for eucalypts in block plantings. On the other hand acacias were found to have a higher AGB when they were located on the outer row of block plantings rather than those in linear plantings (11.9 Mg C ha^{-1} and 3.2 Mg C ha^{-1} respectively). There was little difference in acacia biomass in either the inner or outer row of linear plantings.

When the results for species are analysed for site configuration the eucalypts have a higher AGB (12.1 Mg C ha^{-1}) on the outer rows when in a linear configuration than acacias (3.2 Mg C ha^{-1}), whereas acacias have a higher AGB (12 Mg C ha-^{1}) on the outer row in a block configuration than eucalypts (5.0 Mg C ha^{-1}). There is little AGB C difference between eucalypt and acacia species located in the inner rows of block plantings (5.4 and 6.7 Mg C ha^{-1} respectively).



(a)

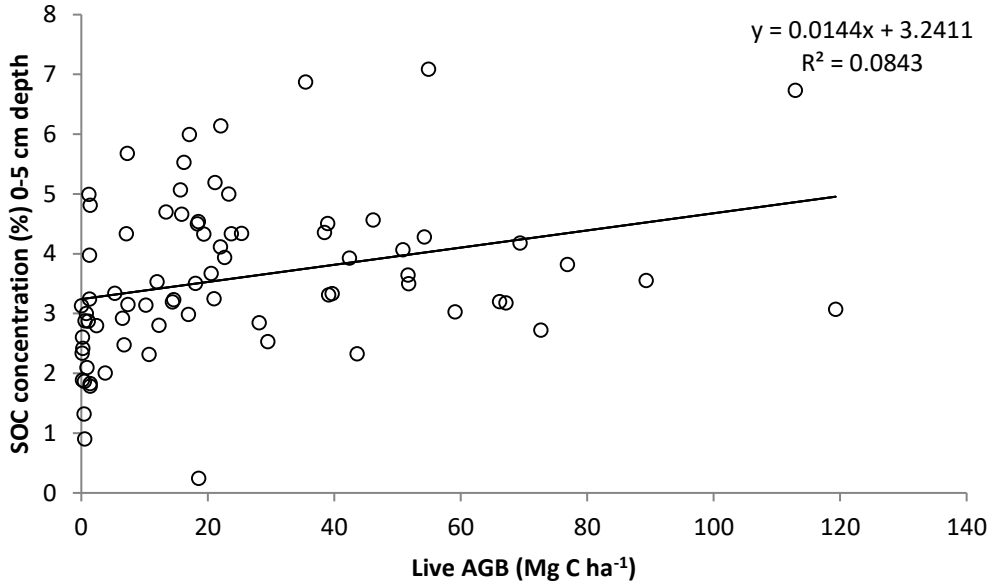


(b)

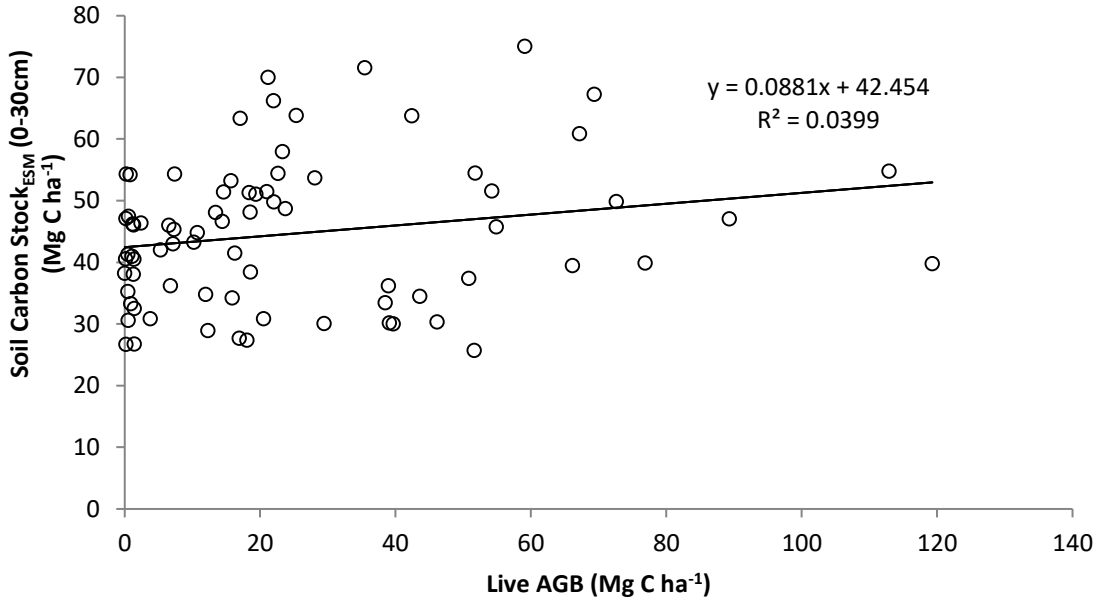
Figure 4-39. Effect of site configuration and edge on species above ground biomass growth (a) = eucalypts, (b) = acacia). Error bars = SEM.

4.13.11 Relationship between above ground biomass C and soil organic C

The relationship between AGB C with SOC in the 0-5 cm depth was found to be weak with the R^2 = 0.07, P=0.1 (Figure 4-40 (a)). The relationship between AGB and SCS is similarly very weak (R^2=0.04, P=0.07) (Figure 4-40 (b)).



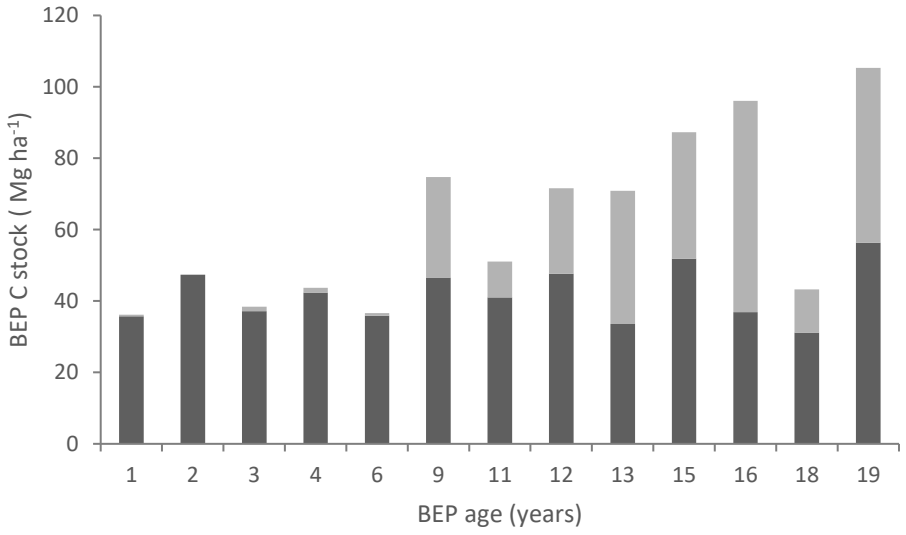
(a)



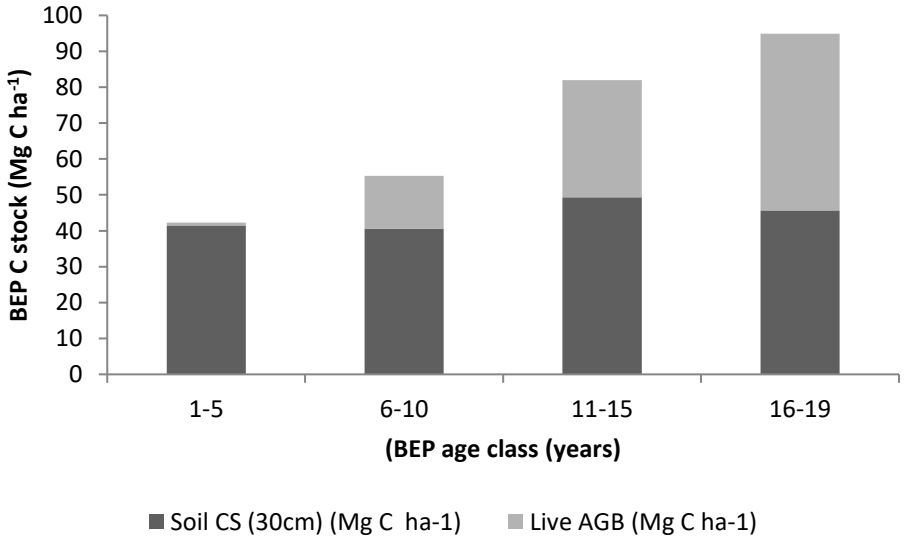
(b)

Figure 4-40. Results show there is no relationship between (a) live above ground biomass C and SOC concentration in the 0-5 cm depth and (b) between AGB C and SCS to 30 cm.

The proportion of C sequestered in the AGB and the soil is shown in Figure 4-41. These results demonstrate that the soil is the principal store for C in BEPs aged up to six years. As the BEPs become established and AGB develops there is a proportional but not linear increase in AGB C. In the oldest cohort of BEPs the AGB is proportionally higher than the SOC (49.2 and 45.6 Mg C ha^{-1} respectively).



(a)



(b)

Figure 4-41. Proportion of CS in the live AGB and the soil (to 30 cm) by (a) age and (b) age class. SCS based on ESM.

A correlation matrix shows there is a weak relationship between SCS and concentration in the 0-5 cm depth with AGB C (Table 4-13). Good correlations were obtained between tree stocking rate and DBH, the basal area, total AGB C and live AGB C. Excellent correlations were obtained for basal area, live AGB C and total AGB C.

Table 4-13. Correlation matrix showing R^2 relationships between above ground biomass C and soil C. (n=77 transects). Results with an R^2> 3 are shown in bold. P values as follows: * = P<0.05, ** = P<0.01, ***P<0.001.

	Tree stocking rate	***DBH (cm)***	***BA (m^2 ha^{-1})***	***Tot AGB (Mg C ha^{-1})***	***Live AGB (Mg C ha^{-1}***	***SOC (%) 0-5 cm***	***SCS to 30 cm (ESM)***
Tree stocking rate	1						
DBH (cm)	**-0.424*****	1					
BA	**0.371*****	**0.380*****	1				
Total AGB	**0.352****	**0.377*****	**0.956*****	1			
Live AGB	**0.363****	**0.386*****	**0.985*****	**0.979*****	1		
SOC %	-0.224*	**0.406*****	0.274**	0.252*	0.290**	1	
CD	0.078	0.118	0.210	0.171	0.200*	**0.413*****	1

4.13.12 Litter biomass and C stock

The CS of litter has been calculated based on two conversion factors. Litter CS calculated on a C content of 54.3% (Gifford, 2000) for both the O1 and O2 horizons is shown in Table 4-14 column A. Litter CS has also been calculated using a conversion factor of 45.4% for the O1 horizon and 27% for the O2 horizon (Table 4-14 column B) (England, pers comm 2015).

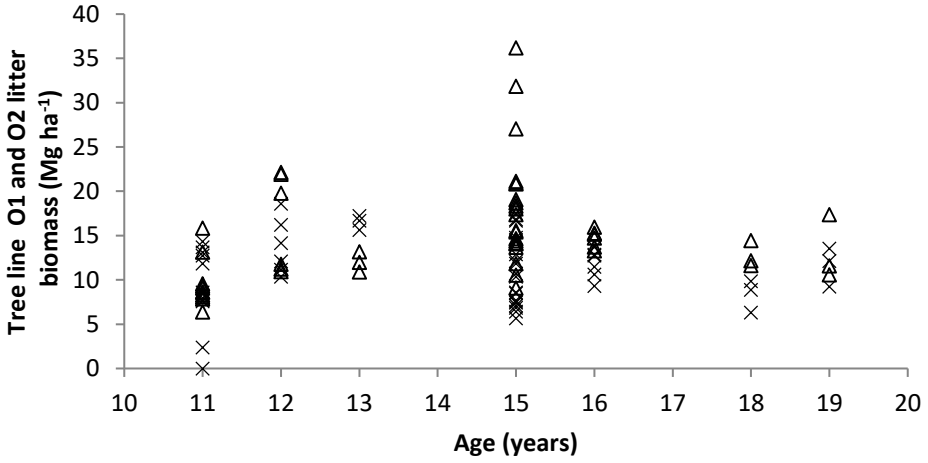
The average litter biomass for the O1 and O2 horizons for both the TL and IR is also shown in Table 4-14. The results show that the O1 horizon, which is comprised of undecomposed leaves and twigs makes a statistically significantly lower (P < 0.001) biomass contribution to BEPs than the O2 horizon (9.3 and 13.4 DM Mg ha^{-1} respectively). Litter biomass and litter C is more abundant in the TL than the IR in both litter layers.

Litter C in BEPs was found to be 4.2 Mg C ha^{-1} in the O1 horizon and 3.3 Mg C ha^{-1} in the O2 horizon, giving a mean total of 7.5 Mg C ha^{-1} in BEP litter (Table 4-14). However, when the litter CS, was calculated using the 54.3 % C as the conversion factor, the CS of litter was found to be higher in the O2 layer of BEPs (6.7 Mg C ha^{-1}) than in the O1 layer (5.1 Mg C ha^{-1}). The overall BEP mean CS is 11.8 Mg C ha^{-1} with the range between 3.6 and 19.3 Mg C ha^{-1}. The mean annual rate of litter accumulation from the measured data was found to be 0.8 Mg C ha^{-1} yr^{-1}.

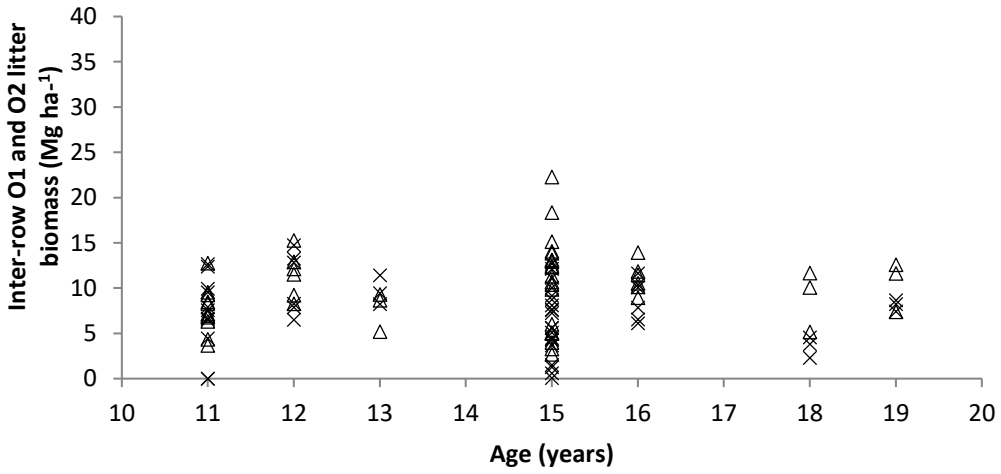
Table 4-14. Mean litter biomass (Mg ha^{-1}) and C stock (Mg ha^{-1}) for TL and IR positions on sites aged between 11-19 years (standard error in parenthesis).

Position	Litter (DM) Mg ha^{-1}	Litter (CS) Mg ha^{-1} (A)	Litter (CS) Mg ha^{-1} (B)
	O1 horizon		
TL	11.3 (0.53)	6.1 (0.30)	5.1
IR	7.3 (0.50)	4.0 (0.28)	3.3
BEP	9.3 (0.47)	5.1 (0.26)	4.2
	O2 horizon		
TL	14.9 (0.79)	8.0 0.43)	4.0
IR	9.9 (0.53)	5.4 (0.29)	2.7
BEP	13.4 (0.55)	6.7 (0.30)	3.3

The weighted mean of DM from the TL and IR was used to determine a litter mass for BEPs. The results show that BEP litter does not have an increasing trend with age either in the TL or IR positions (Figure 4-42). Significant variability is present associated with site age in the TL position in the O1 and O2 layer ($P = 0.038$ and $P = 0.002$ respectively) and in the IR position O1 layer ($P = 0.05$).



(a)



(b)

Figure 4-42. Litter biomass results for the O1 (×) and O2 (Δ) litter layers for both the TL (a) and IR (b) transects by BEP age.

When the measured litter C results calculated using a conversion factor of 54.3% (Gifford, 2000) as the mean of the O1 and O2 layers from the TL and IR positions (representing BEP) and as a mean of the three replicate transects per site are compared with site results derived using FullCAM modelled results (Figure 4-43) there is a weak relationship (R^2 = 0.3, P = 0.04). The relationship is slightly weaker when the measured results for BEPs (mean of TL and IR) from the O1 layer only are used (R^2 = 0.29, P = 0.7). However, When the C content of litter was calculated using conversion factors of 45.4 for the O1 layer and 27% for the O2 layer (J. England, pers com Sep 2015) the correlation with FullCAM predicted results improved slightly with R^2 = 0.34.

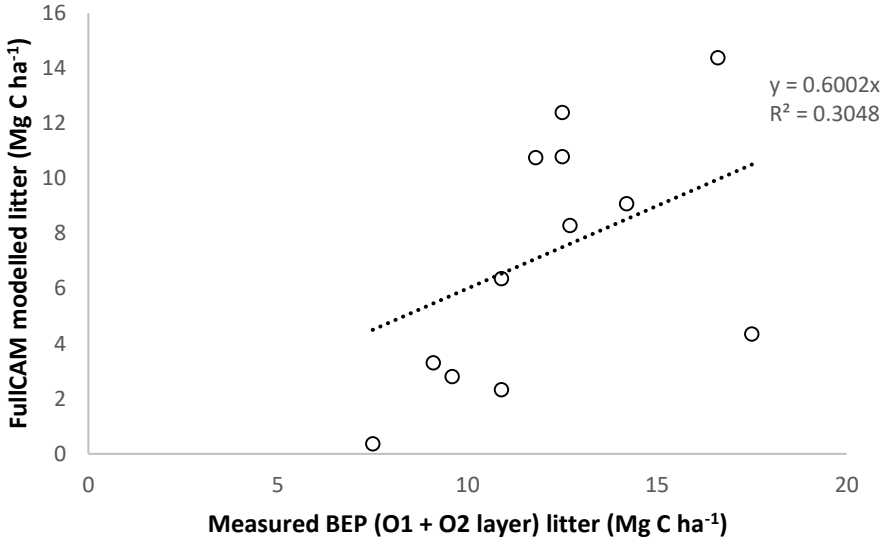


Figure 4-43. Correlation between measured (conversion of 54.3%) and FullCAM modelled data for litter (Mg C ha^{-1}).

4.13.13 Electrical conductivity

The results show that the EC $_{(1:5\ water)}$ values for all sites, positions and depths are low. The EC values of the 0-5 cm layer in the AL position for BEP sites aged at between 1-5 years are the lowest, whereas the other three age classes have similar EC at the surface (Figure 4-44). All age classes EC values decrease with depth.

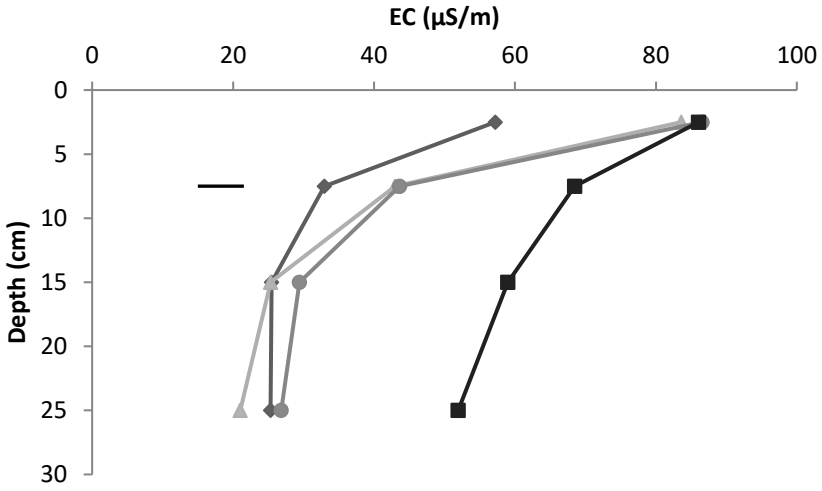


Figure 4-44. The electrical conductivity results for the AL position showing differences based on the age class of the BEPs. Error bar = SEM. (♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs).

The EC grand mean for all sites, positions and depths is 51.6 µS cm^{-1} (Figure 4-45) which is a very low concentration. The ANOVA for depth and age class was suggested to be statistically significant (P<0.001) but position was not significant .The interactions between position and age class, and depth with age class were significant (P=0.006 and P=0.016 respectively). There is a trend indicating the EC is higher in the BEPs older than five years than in the AL.

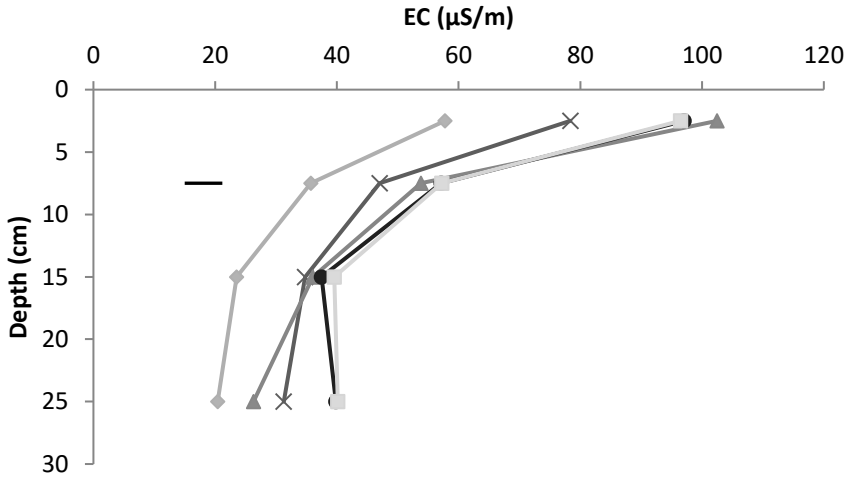


Figure 4-45. EC showing the overall mean for all AL positions, and BEPs by age cohort. Error bar = SEM. (× = AL; ♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs).

4.13.14 pH

The soil pH $_{(1:5\ water)}$ in the AL position and for all age classes shows variation at all depths (Figure 4-46). The soil pH was found to increase with depth. The mean pH for the combined age classes in the AL position at the 0-5 cm depth increment is in the acidic range at 5.3.

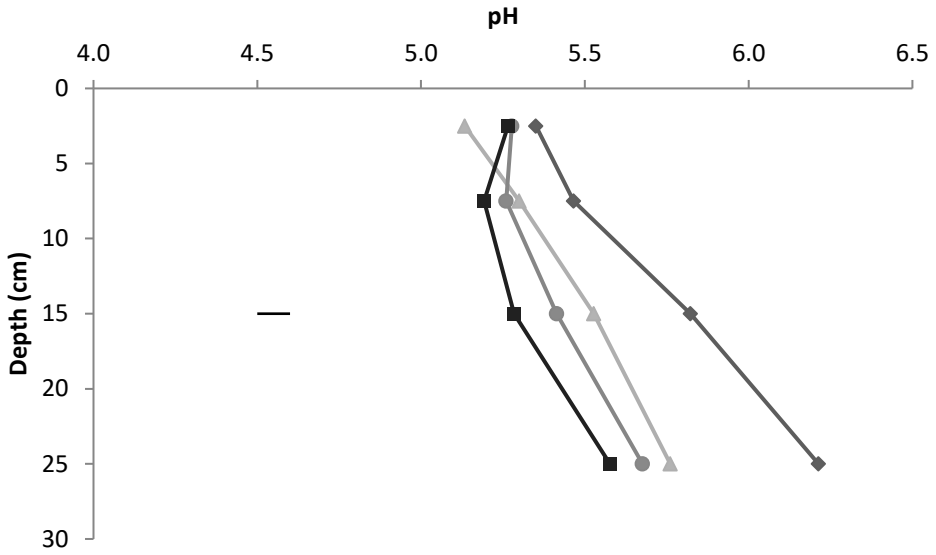


Figure 4-46. Mean soil pH for the AL position for all sites by BEP age class (error bar = SEM) (♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs).

When comparing soil pH in the AL position with soil from the BEP (mean results for TL and IR positions), the AL position had a higher pH for all age classes and depth increments down throughout the 30 cm profile except at the 20 to 30 cm depth increment where sites in the 1-5 year age class had a higher pH (Figure 4-47). There is a strong trend suggesting that as the BEPs get older the pH reduces, becoming more acidic.

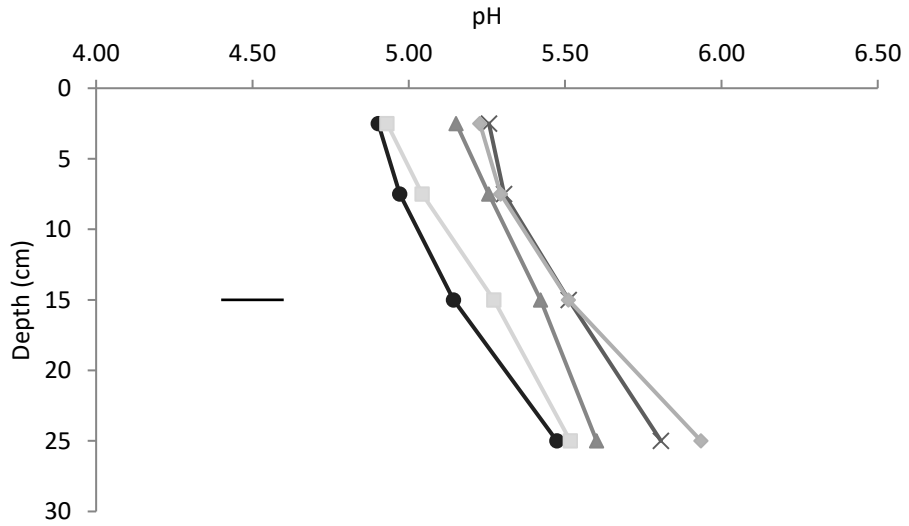


Figure 4-47. Soil pH comparing AL with BEP (mean of TL and IR). Horizontal error bar = SEM. (× = AL; ♦ = 1-5 yrs; ▲ = 6-10 yrs; ● = 11-15 yrs; ■ = 16-19 yrs).

The ANOVA suggests that BEP age, soil depth and position are all highly significant (P<0.001) variables for pH. The interaction between position and soil depth, age class and position was suggested by ANOVA to have a significant effect on pH (P <0.001) (Table 4-15).

Table 4-15. ANOVA analysis for soil pH showing position, depth and age class interactions.

Source of variation	*d.f.*	*s.s*	*m.s*	*v.r*	*F pr.*
Transect position	2	7.81	3.90	23.3	<.001
Transect position . soil depth	9	52.75	4.75	28.35	<.001
Transect position . soil depth . age class	36	17.56	0.49	2.91	<.001

4.13.15 Total nitrogen concentration and C:N ratio

The overall concentration of TN was found to be at the low to medium range typically found in soils, whereby in an agricultural context the TN concentration in the 0-5 cm layer is considered to be low when it is < 1 500 mg kg^{-1}, and be in the medium to high range when the concentration is approximately 2 500 mg kg^{-1} (Moore, 1998).

The concentration of TN in the AL position (Figure 4-48) shows that there is significant variation between depths and between ages (P < 0.001), however the interaction between age class and depth in the AL position is not significant. TN in the AL position is highest in the 16-19 year old BEP sites for all depth increments. At the 0-5 cm soil layer the TN concentration was lowest in the 6-10 year old BEP cohort with 1 282 mg kg^{-1}. The highest TN concentration was in the 0-5 cm layer of in the AL position of BEPs aged between 16-19 years, i.e., a mean of 2 538 mg kg^{-1}.

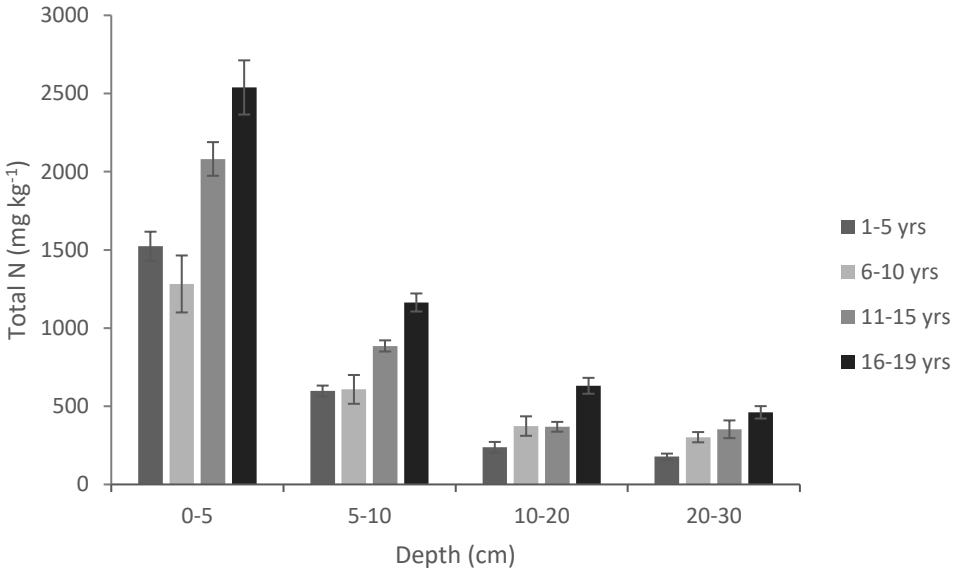


Figure 4-48. Total N (TN) concentration in the AL position showing variation with depth and age. (Error bars = SEM).

The distribution of TN by depth in the BEP position (average TN in both the TL and IR positions) is similar to the AL position (Figure 4-49). There is a decreasing TN concentration with depth. The 11-15 year old BEPs had on average the highest overall TN concentration through the profile, and the highest TN concentration (2 554 mg kg^{-1}) was found in the 11-15 year old BEP cohort in the 0-5 cm soil layer.

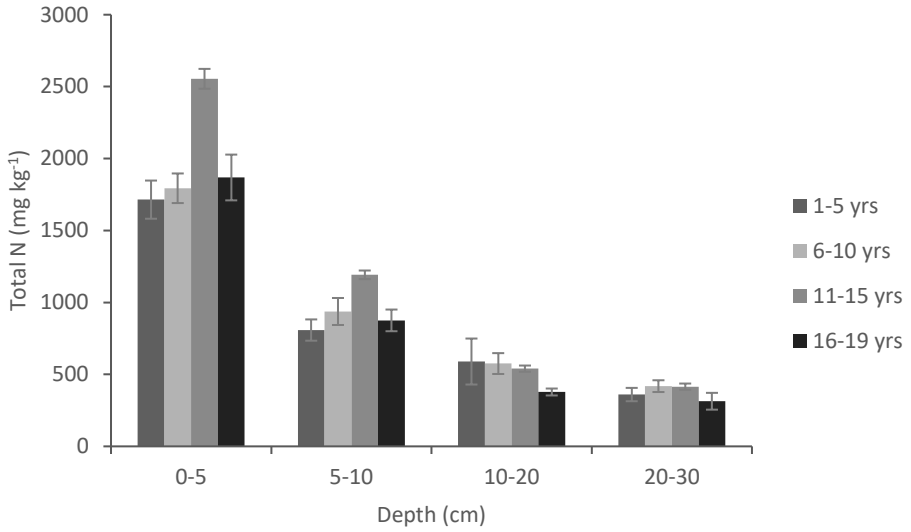


Figure 4-49. Concentration of total N (TN) in BEPs (mean of TL and IR) showing distribution by depth and BEP age. (Error bars = SEM).

To compare the TN concentration between positions (i.e., in AL, TL, IR and BEP), overall, in the 0-5 cm layer of the BEPs the mean TN concentration was highest in the 11-15 year old cohort. The TN concentration was generally found to be higher in the TL, IR and BEP position than in the AL position, except in the BEP cohort aged 16-19 years where the AL was found to have a higher TN concentration (Figure 4-50). The TL had a higher TN concentration than the AL in the 6-10 and 11-5 year old BEPs only. The TN concentration is higher in the IR than in the TL position except in sites aged between six and ten years when the TL has a higher TN concentration than the IR.

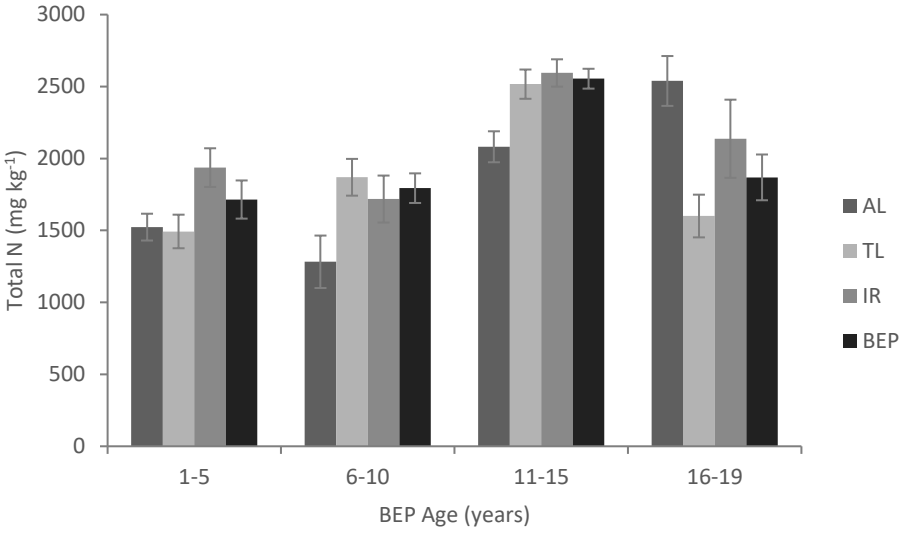


Figure 4-50. Total N in the 0-5 cm soil layer showing the difference between AL, TL, IR and BEP positions. Error bars = SEM.

The C:N ratio results for the 0-5 cm depth increment when comparing all positions (Figure 4-51) the AL position show there is variation in C:N between the age cohorts with the lowest ratio of 16 found in the AL position of the 16-19 year old BEPs and the highest ratio of 23 was found in the AL position of the 6-10 year old BEPs. The 16-19 year old BEPs had the highest C:N ratio of 26. The C:N ratio in the BEPs aged 1-5, 6-10 and 11-15 was 19, 18 and 19 respectively. Overall, the 16-19 year old cohort has on average a higher C:N ratio in the TL, IR and BEP than the younger sites.

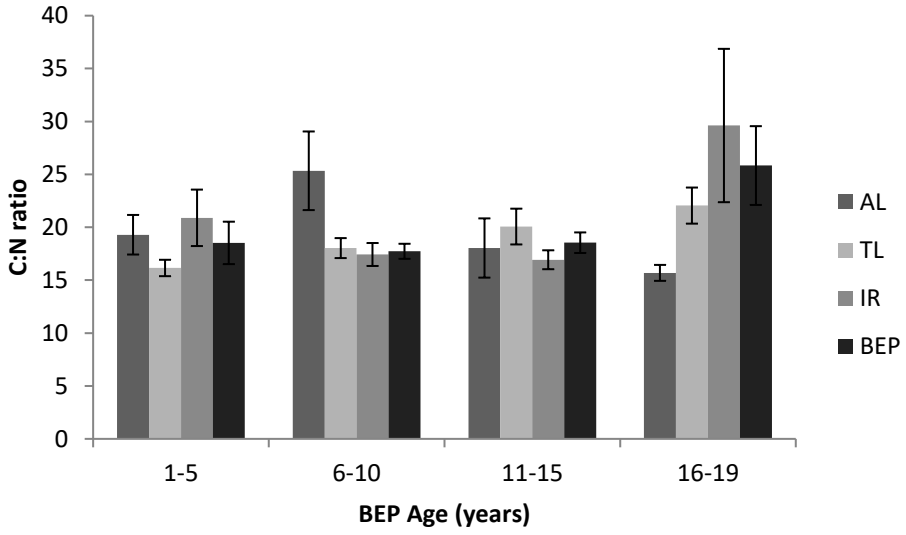


Figure 4-51. C:N ratio in the 0-5 cm soil layer, comparing positions and using a mean of TL and IR for BEP. Error bars = SEM.

4.13.16 Total nitrogen and C:N ratio constraints on SOC sequestration potential

The potential of the AL and the BEPs to sequester C based on the TN concentration and the C:N ratio present in the soil is summarised in Table 4-16. For example, the mean results from the paired AL in the 0-5 cm soil layer of the 11-15 year old BEP cohort show that the AL has a measured SOC concentration of 30 562 mg kg^{-1}, the concentration of TN is 2 081 mg kg^{-1} and the C:N ratio is 15. Based on the measured SOC and TN, and the C:N ratio, the AL position has the potential to sequester up to 31 215 mg kg^{-1} of SOC. Conversely, the BEPs have a measured SOC concentration of 40 407 mg kg^{-1}, a TN of 2 554 mg kg^{-1} and the C:N ratio is 16. The BEPs therefore have the potential to sequester 40 864 mg kg^{-1} of SOC, which is 9 847 mg kg^{-1} greater than the concentration of SOC in the AL position. TN is therefore not limiting the SOC sequestration potential of BEPs in the 11-15 year old cohort.

The ideal C:N ratio for plant growth is 12 (Kirkby *et al*. 2011). However, because the C:N ratio in the AL and BEPs is higher, the SOC sequestration potential of 24 972 mg kg^{-1} in the 0-5 cm layer, is exceeded by both the AL and BEP positions (Table 4-16).

The potential for the younger BEPs to sequester SOC also exists (results not shown). For example, in BEPs aged 1-5 years, the AL in the 0-5 cm layer has a measured SOC of 25 709 mg kg^{-1}, and potential to sequester 25 891 mg kg^{-1}. The 1-5 year old BEPs have 27 846 mg kg^{-1} SOC and their potential cased on the C:N of 16 is for 27 440 mg kg^{-1} SOC. Similarly, the 0-5 cm layer in the AL position for BEPs aged 6-10 years has the potential to sequester 24 358 mg kg^{-1} and the BEPs have the potential to sequester 30 498 mg kg^{-1}. In both these age cohorts the TN is higher in the BEPs than in the AL.

Table 4-16. The SOC sequestration potential of AL and BEPs in the 11-15 year old cohort, based on the mean SOC and mean TN concentration and the C:N ratio. Calculations also included for SOC sequestration potential based an optimal C:N ratio of 12 (Kirkby *et al*. 2011).

Soil layer (cm)	*Measured TN (mg kg^{-1})*	*Measured SOC (mg kg^{-1})*	*C:N ratio*	*SOC Sequestration Potential (mg kg^{-1}) (based on C:N ratio)*	*SOC sequestration potential based on C:N ratio of 12*
			AL		
0-5	2 081	30 562	15	31 215	24 972
5-10	886	14 878	17	15 062	10 632
10-20	369	6 978	19	7 011	4 428
20-30	353	4 872	14	4 942	4 236
			BEP		
0-5	2 554	40 407	16	40 864	30 648
5-10	1 192	18 867	16	19 072	14 304
10-20	540	9 360	17	9 180	6 480
20-30	415	6 406	15	6 225	4 980

The 16-19 year old BEP cohort in the 0-5 cm layer have a mean SOC concentration of 39 484 mg kg^{-1} in the 0-5 cm layer of the AL position based on their C:N ratio of 16, and the TN concentration of 2 539 mg kg^{-1} (Table 4-17). The AL therefore has the potential to sequester 40 624 mg kg^{-1} SOC. The age paired BEPs have 36 006 mg kg^{-1} SOC, 1 869 mg kg^{-1}, and C:N ratio of 19, which is 3 478 mg kg^{-1} less than the paired AL position. The 16-19 year old BEPs have a potential to sequester 35 511 mg kg^{-1} in the 0-5 cm layer based on their C:N ratio of 19, which is less than the potential of the AL. Therefore, these results show that TN is limiting the SOC sequestration potential of BEPs in the 16-19 year old cohort.

Table 4-17. The SOC sequestration potential of the AL and BEPs in the 16-19 year old cohort, based on the TN concentration and C:N ratios. Optimal C:N ratio of 12 is based on Kirkby et al. (2011).

Soil layer (cm)	*Measured TN (mg kg^{-1})*	*Measured SOC (mg kg^{-1})*	*C:N ratio*	*SOC Sequestration Potential (mg kg^{-1}) (based on C:N ratio)*	*SOC sequestration potential based on C:N ratio of 12*
			AL		
0-5	2 539	39 484	16	40 624	30 468
5-10	1 164	17 858	15	17 460	13 968
10-20	631	10 207	16	10 096	7 572
20-30	461	6 506	14	6 454	5 532
			BEP		
0-5	1 869	36 006	19	35 511	22 428
5-10	876	17 335	20	17 520	10 512
10-20	378	9 072	24	9 072	4 536
20-30	314	6 392	20	6 280	3 768

4.13.17 Total Phosphorus

The TP concentration results for the AL position show there is a decreasing concentration with depth in all age classes ($P < 0.001$) (Figure 4-52). The average TP concentration in sites aged between 11-15 years is higher than the other age cohorts at all depth increments. Sites in the 16-19 year old cohort have the lowest concentration of TP at all depth increments except at 0-5 cm depth which has a higher concentration of TP than the TP in the 0-5 cm layer in the 6-10 year age cohort. At the 0-5 cm depth increment, sites in the 11-15 year cohort have the highest mean TP concentration at 512 mg kg^{-1}, and sites in the 6-10 year cohort have the lowest TP concentration at 248 mg kg^{-1}.

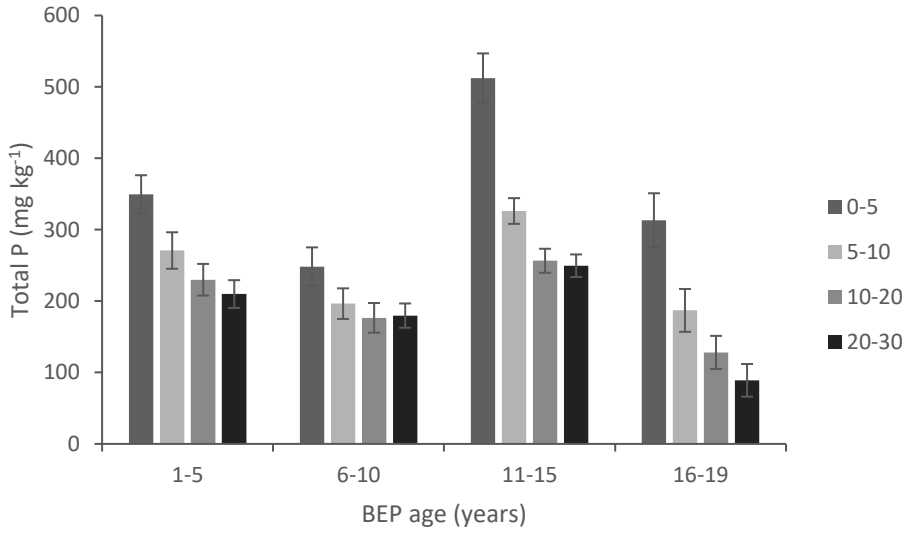


Figure 4-52. Total P for the AL position by soil depth increment (cm). (Error bars = SEM).

In the BEP position (mean of TL and IR) there is a decreasing trend in TP concentration with depth in all age classes (Figure 4-53). The 11-15 year BEP cohort has the overall highest concentration of TP, and the 6-10 year cohort the lowest. In the 20-30 cm depth increment the 11-15 year and 16-19 year age classes have a slightly higher TP concentration than for the 10-20 cm depth increment. There is no particular trend with age, with the results showing variation across age groups.

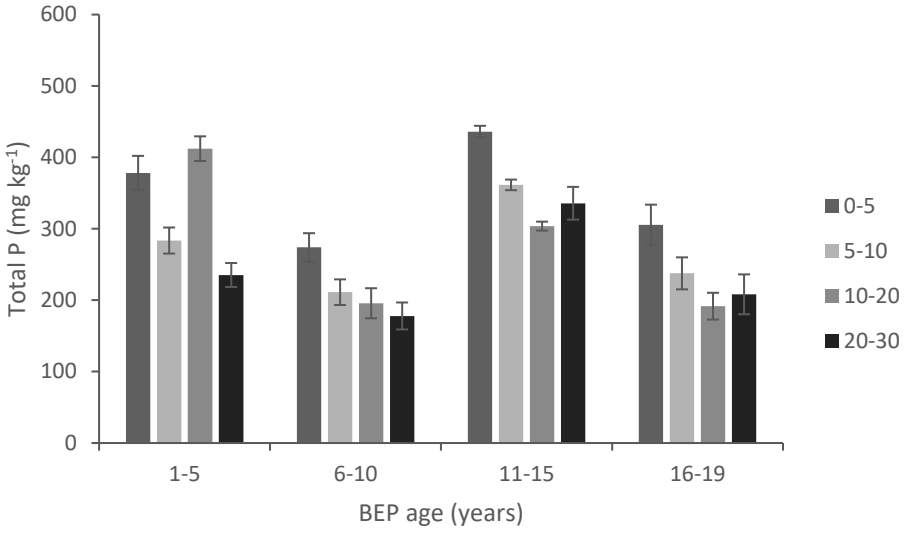


Figure 4-53. Total P for the BEP position by BEP age class and by depth increments (cm). (Error bars = SEM).

When viewed by transect position there is no particular trend for TP (Figure 4-54). When viewed by age class, the average TP in the 0-5 cm soil layer shows no particular trend with age and there is also no trend when comparing the AL position with the TL, IR and BEP positions. The 6-10 year old BEPs generally have the lowest TP across all positions, and the 11-15 year old sites have the highest TP in all positions.

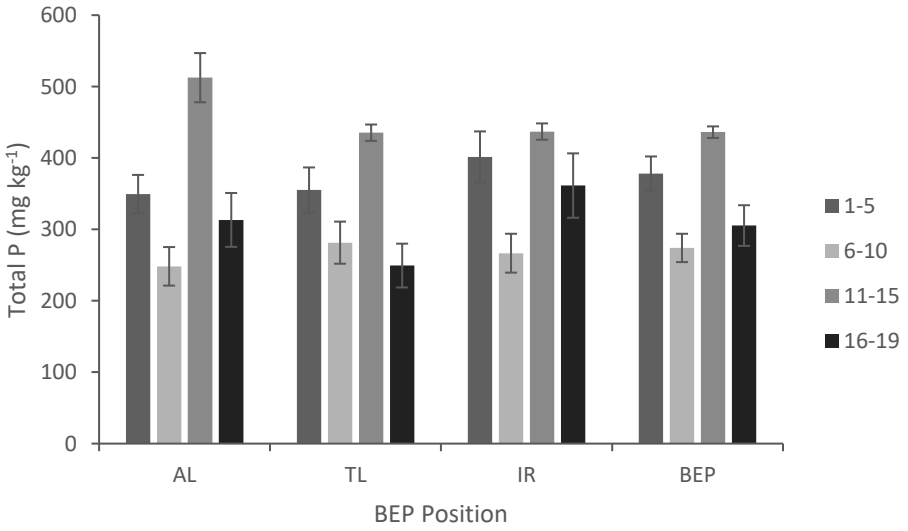


Figure 4-54. Total P in the 0-5 cm layer comparing age classes for the AL with TL, IR and BEP positions. Error bars = SEM.

4.13.18 C:P ratio

The results showing the C:P ratio in the 0-5 cm soil layer for each position (AL, TL, IR and BEP) (Figure 4-55) have no particular trend between position and age, except the oldest BEP cohort has wider C:P ratio particularly in the TL position than in the paired AL position. The variation within the 16-19 year old cohort (large standard error) shows intra-site variability is a factor influencing the results. The SOC:TP (C:P) ratio show a significant narrowing concentration with each successive depth increment (P = < 0.001). The interactions between position, depth and age are similarly significant at P = <0.001).

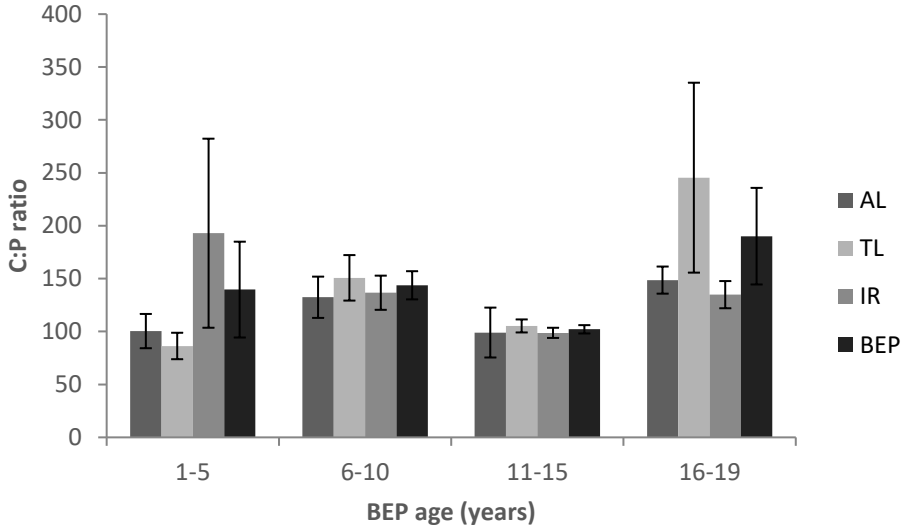


Figure 4-55. C:P ratio in the 0-5 cm soil layer by age class, comparing positions and using a mean of TL and IR for BEP. Error bars = SEM

4.13.19 Total P and C:P ratio constraints on SOC sequestration potential

Analysis of the SOC sequestration potential of the AL and BEPs from the 11-15 year old cohort shows that the 0-5 cm soil layer in the AL position has a measured SOC concentration of 30 562 mg kg^{-1} (Table 4-18), the mean C:P ratio is 60 and the mean TP concentration is 512 mg kg^{-1}. The AL could therefore potentially sequester up to 30 720 mg kg^{-1} of SOC based on the C:P ratio and concentration of TP. The 11-15 year old BEPs have an SOC concentration of 40 407 mg kg^{-1} in the 0-5cm soil layer, which is very similar to the sequestration potential based on the C:P ratio. Therefore, the BEPs have sequestered 9 845 mg kg^{-1} of SOC over the 15 year time since establishment above the baseline AL SOC concentration. Consequently TP is not limiting the SOC sequestration potential of BEPs in the 11-15 year old cohort.

The ideal C:P ratio for plant growth is 50 (Kirkby *et al*. 2011). When the ratio of 50 is used to calculate an optimum SOC sequestration potential for the AL in the 0-5 cm layer, a concentration of 25 600 mg kg^{-1} has been determined. However, the results show that the AL has a concentration of 30 562 mg kg^{-1} which is close to the SOC sequestration potential due to the C:P ratio of 60, of 30 720 mg kg^{-1}.

Table 4-18. The SOC sequestration potential of AL and BEPs in the 11-15 year old cohort, based on the TP concentration and the C:P ratio. Optimal C:P ratio of 50 based on Kirkby *et al*. (2011).

Soil layer (cm)	*Measured TP (mg kg^{-1})*	*Measured SOC (mg kg^{-1})*	*C:P ratio*	*SOC Sequestration potential (mg kg^{-1}) (based on C:P ratio)*	*SOC sequestration potential based on C:P ratio of 50*
			AL		
0-5	512	30 562	60	307 20	25 600
5-10	326	14 878	46	14 996	16 300
10-20	256	6 978	27	6 912	12 800
20-30	250	4 872	19	4 750	12 500
			BEP		
0-5	436	40 407	92	40 112	21 800
5-10	362	18 867	52	18 824	18 100
10-20	304	9 360	31	9 424	15 200
20-30	336	6 406	19	6 384	16 800

The C:TP and SOC potential for the 15-19 year old BEP cohort is summarised in Table 4-19. The AL has an SOC concentration of 39 484 mg kg^{-1} in the 0-5 cm layer whereas the BEP has 35 990 mg kg^{-1}. The C:AP ratio and the SOC sequestration potential based on that ratio implies the TP concentration is insufficient for the sequestration of further SOC in the 16-19 year old cohort of BEPs. The SOC in both the AL and BEPs exceeds the sequestration potential based on the optimum C:P ratio of 50.

Table 4-19. The SOC sequestration potential of AL and BEPs in the 16-19 year old BEP cohort, based on the TP concentration and the C:TP ratio. Optimal C:P ratio of 50 based on Kirkby *et al*. (2011).

Soil layer (cm)	*Measured AP (mg kg^{-1})*	*Measured SOC (mg kg^{-1})*	*C:P ratio*	*SOC Sequestration potential (mg kg^{-1}) (based on measured C:P ratio)*	*SOC sequestration potential based on C:P ratio of 50*
			AL		
0-5	313	39 484	126	394 38	15 650
5-10	187	17 858	95	17 765	9 350
10-20	128	10 207	79	10 112	6 400
20-30	89	6 506	73	6 497	4 450
			BEP		
0-5	305	36 006	118	35 990	15 250
5-10	238	17 335	72	17 136	11 900
10-20	192	9 072	47	9 024	9 600
20-30	208	6 392	30	6 240	10 400

To summarise the results for TN, TP, SOC and nutrient ratios a correlation matrix Table 4-20 shows that the only soil properties to have a strong correlation are SOC and TN where the R^2 was 0.88 (P<0.001).

Table 4-20. Correlation matrix showing R^2 relationships among nutrients and nutrient ratios. Results with an R^2 >0 3 is shown in bold. P values as follows: * = P<0.05, ** = P<0.01, ***P<0.001.

	TN (%)	*TP (%)*	*SOC (%)*	*C:TP ratio*	*C:N ratio*
TN (%)	1				
TP (%)	**0.33** ***	1			
SOC (%)	**0.88** ***	0.27 ***	1		
C:P ratio	0.21 ***	-0.28 ***	**0.33** ***	1	
C:N ratio	- 0.25 ***	-0.21	-0.08 ***	0.03	1

4.13.20 Relationship between Leco CNS2000 and Heanes measured SOC results

Subset 1 (n=96) samples, incorporating eight BEP sites, three positions (TL, IR and AL) and four depth intervals were analysed using Leco CNS2000 dry combustion analysis for TOC. The results were compared with the SOC results obtained from analysis using the Heanes (1984) method. Using linear regression a very strong positive correlation R^2 of 0.92 (P < 0.001) (Figure 4-56) was found.

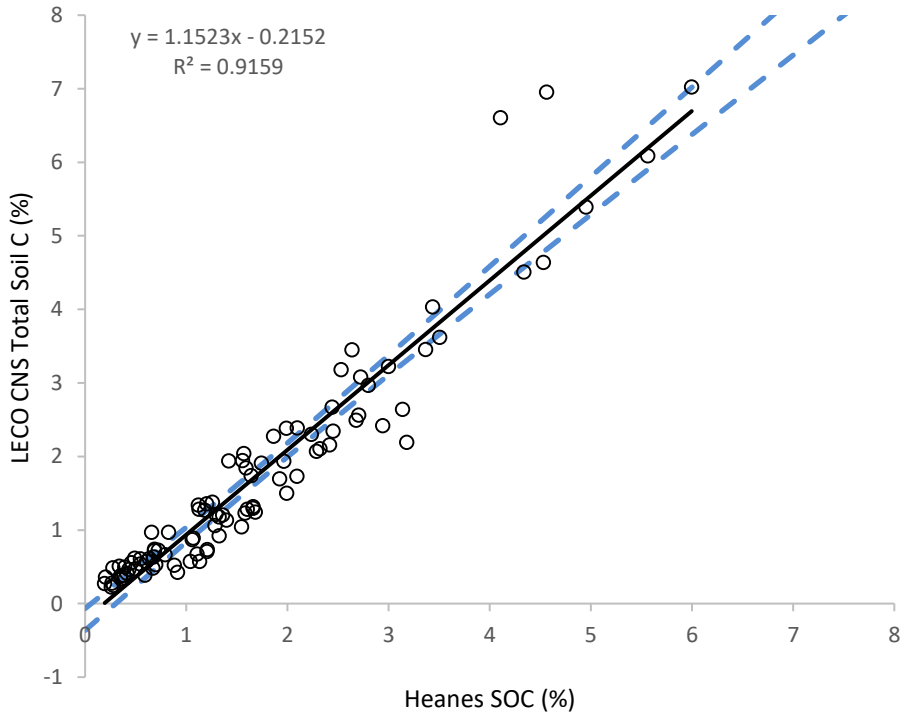


Figure 4-56. Relationship between Leco CNS2000 and Heanes analytical methods for SOC on a subset of 96 samples incorporating 8 BEP sites, 4 depth intervals and three positions (dotted lines = 95 % confidence interval).

4.13.21 Comparison of Leco CNS2000, Heanes and MIR predicted SOC results

Correlation analysis for SOC results obtained by the three different analytical methods used in the current research has been undertaken. The correlation shows that SOC analysis conducted using the SOC_{Heanes} method has a strong correlation with the results obtained using MIRS calibrated predictions (R^2 = 0.93) (Table 4-21). The TOC_{LECO} and predicted MIRS SOC results also had a strong correlation of R^2=0.98. There was a strong correlation for SOC results obtained using both the TOC_{LECO} and SOC_{Heanes} methods (R^2 = 0.92). The strong relationship for SOC concentration identified for the three methods shows that MIRS can reliably be used to predict the SOC concentration of soil samples in both agricultural soils and soils where BEPs have been established.

Table 4-21. Correlation matrix showing R^2 relationships for between SOC analysis methods (Leco CNS2000, Heanes and MIR). Results with an R^2 >0 3 are shown in bold. P-values as follows * = P<0.05, ** = P<0.01, ***P<0.001.

	TOC Leco CNS2000	*SOC Heanes*	*MIR calibrated predicted SOC*
TOC Leco CNS2000	1		
SOC Heanes	**0.917** ***	1	
MIR calibrated predicted SOC	**0.982** ***	**0.927** ***	1

4.13.22 Soil organic C and soil organic C fraction results derived by MIR analysis

The SCS to 30 cm depth for soil fractions derived from MIRS analysis results are shown following calculation using the ESM equation (Equation 10). The results at Figure 4-57 show SCS for the AL position for each site and highlights the variability in each SOC fraction. When the results for the AL position are analysed there appears to initially be an increase in POC, HOC and ROC with age, with a site aged 9 years (Site 8) having the highest SCS in each fraction. The HOC fraction consistently has the highest proportion of SCS.

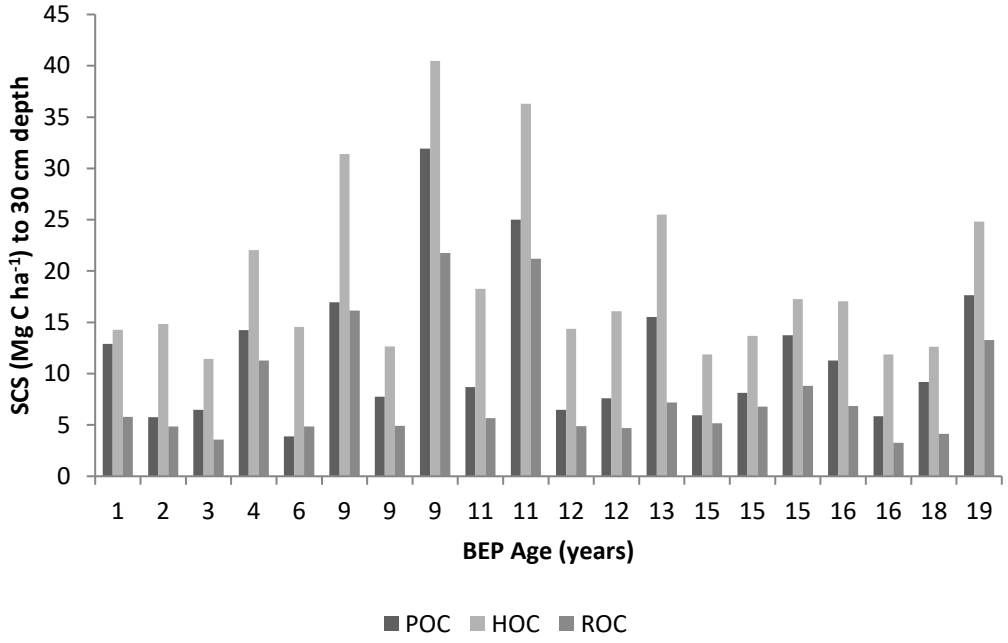


Figure 4-57. SCS to 30 cm depth and allocation to SOC fractions in the AL position for all sites based on BEP age.

The mean results for POC, HOC and ROC fractions for the AL position together with the mean of each fraction for the four BEP age classes is show at Figure 4-58. The SOC $_{(Heanes)}$ ESM results have been included for comparison. The HOC fraction contains the largest C stock for all age classes and positions (AL, TL, IR & BEP) (Figure 4-58). When age was not included as a factor the overall mean HOC was higher in BEP (19.0 Mg C ha^{-1}) than AL (16.7 Mg C ha^{-1} ±1.3) (df = 76, t = 1.15 *P* = 0.25). When broken down by age class, BEPs in the 6-10 year old cohort have the highest mean proportion of HOC (23.1 Mg C ha^{-1}) whereas the 16-19 year old BEPs had the lowest HOC of 14.9 Mg C ha^{-1}. BEPs aged 1-5, 6-10 and 11-15 years had the same or more HOC than the AL position. The 16-19 year old BEPs had a lower average HOC content than the AL (14.9 and 15.3 Mg C ha^{-1} respectively).

The results show the average POC is less than the average HOC for all age classes. The mean (±SE) difference in POC from AL to BEP is 10.2 (±1.0) Mg C ha^{-1} for AL and 12.5 (±1.1) Mg C ha^{-1}) for BEP (df = 76, t = 1.43 and *P* = 0.16). BEPs aged 6-10 and 11-5 had the highest

average proportion of POC than either the AL, or younger or older sites. The 16-19 year old BEPs had slightly less POC than the AL (10.1 and 10.2 Mg C ha^{-1} respectively).

Of the three fractions ROC had the lowest average proportion in all age classes and in the AL position. The mean (±SE) ROC in AL was found to be 7.26 (±0.9) Mg C ha^{-1} and in the BEP 8.4 (±1.1) Mg C ha-1 (df = 76, t = 0.78, *P* = 0.44)

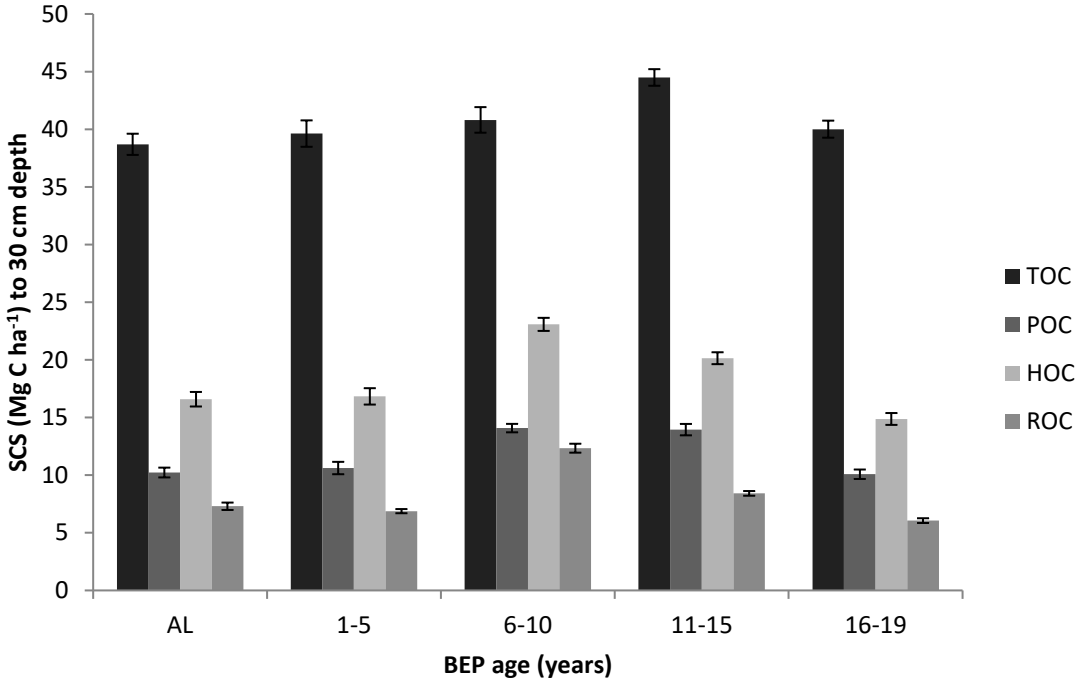


Figure 4-58. Mean SCS for fractions to 30 cm depth, calculated on an ESM basis showing results for each age class (the AL is the mean for all BEP sites (*n* = 20)) (BEP is the mean of TL and IR. Error bars = SEM.

Overall, the results show that none of the fractions in the chronosequence have achieved a linear increase in SOC with age. Initially the results show increase in SOC in all the fractions until the sites are aged 6-10 years. BEPs aged 11-15 and 16-19 have less C in the soil fractions than the 6-10 year old BEPs. BEPs aged 16-19 have the lowest proportion of all fractions, and their results are lower than the mean AL position.

When the results for the 16-19 year old BEPs are compared with their paired AL results, the trend is the same. The AL has a higher HOC, POC, ROC and TOC (15.3, 10.5, 6.3 and 41.1 Mg C ha^{-1} (respectively) than the BEPs (14.9, 10.1, 6.1 and 40.0 Mg C ha^{-1} respectively).

The results for each site showing the difference for each C fraction following conversion from AL to BEP, show there is considerable variation between sites irrespective of age since establishment with some sites having a loss (Table 4-22). The mean change across all sites and ages from AL to BEP for the POC, HOC and ROC fractions is a gain of 2.3, 2.3 and 1.2 Mg C ha^{-1} respectively. The annual average rate of change for the POC, HOC and ROC fractions is 0.3, 0.4 and 0.2 Mg C ha^{-1} yr^{-1} increase respectively. The HOC fraction has the largest annual rate of change.

Table 4-22. SCS to 30 cm for SOC fractions in the AL and BEP (ESM) and showing the change in each fraction following conversion from AL to BEP and the annual rate of change. Results are a mean of three transects per site except * which is a mean of eight transects. "–" indicates C loss.

Site ID	Site Age	POC (Mg C ha^{-1})				HOC (Mg C ha^{-1})				ROC (Mg C ha^{-1})			
		AL	BEP	ΔPOC since LUC to BEP	annual rate of POC change	AL	BEP	ΔHOC since LUC to BEP	annual rate of HOC change	AL	BEP	ΔROC since LUC to BEP	annual rate of ROC change
anb	1	12.6	12.3	-0.3	-0.3	13.9	15.5	1.6	1.6	5.6	5.8	0.2	0.2
keb	2	5.5	7.8	2.3	1.2	12.6	17.7	5.1	2.6	4.4	6.2	1.8	0.9
sha	3	5.9	7.4	1.5	0.5	9.8	12.7	2.9	1.0	3.1	4.2	1.1	0.4
woo	4	13.3	14.9	1.6	0.4	20.0	21.4	1.4	0.3	10.3	11.3	1.0	0.3
ter	6	3.9	6.9	3.0	0.5	14.4	19.7	5.2	0.9	4.8	7.5	2.7	0.4
sc	9	16.3	18.4	2.2	0.2	29.5	31.6	2.1	0.2	15.2	16.3	1.1	0.1
wic	9	7.5	8.3	0.9	0.1	11.8	12.1	0.3	0.0	4.7	5.1	0.5	0.1
ws	9	15.6	22.6	7.0	0.8	23.8	29.0	5.2	0.6	15.1	20.4	5.3	0.6
cap*	11	8.5	7.3	-1.2	-0.1	16.8	15.7	-1.2	-0.1	5.3	4.8	-0.6	-0.1
car	11	18.5	20.6	2.1	0.2	26.9	28.6	1.7	0.2	16.1	18.0	1.9	0.2
kea	12	6.5	10.8	4.3	0.4	13.5	16.8	3.3	0.3	4.7	5.9	1.2	0.1
shb	12	7.7	14.4	6.7	0.6	16.4	20.8	4.3	0.4	4.8	6.8	2.0	0.2
ana	13	15.0	14.9	-0.2	0.0	24.0	24.7	0.7	0.1	6.9	6.7	-0.2	0.0
str	15	5.6	15.6	10.0	0.7	10.7	22.7	12.0	0.8	4.8	9.6	4.8	0.3
wia	15	8.0	13.6	5.6	0.4	12.9	16.1	3.2	0.2	6.7	7.7	1.1	0.1
wib	15	12.7	14.3	1.6	0.1	15.1	15.8	0.7	0.0	7.6	7.8	0.2	0.0
haa	16	10.8	11.6	0.8	0.0	15.5	15.8	0.3	0.0	6.2	7.4	1.2	0.1
hab	16	5.8	6.1	0.3	0.0	11.6	13.6	2.0	0.1	3.2	3.8	0.6	0.0
spr	18	9.0	7.5	-1.6	-0.1	12.2	11.3	-0.9	0.0	4.0	3.4	-0.6	0.0
pyl	19	16.2	15.1	-1.1	-0.1	22.1	18.7	-3.3	-0.2	11.9	9.5	-2.3	-0.1

4.13.23 Proportion of soil fractions determined by the wet sieved method

The wet sieve results for Subset 2 (n = 12) comprising samples from the 0-5 cm layer in the TL and AL positions shows that the TL has on average a higher proportion of TOC, POC and HOC than the AL position (Table 4-23). The results show that the wet sieve derived HOC fraction is proportionally higher than the POC fraction. Wet sieving does not separate samples to differentiate between the HOC and ROC fractions consequently the HOC results are shown as the composite of the HOC + ROC fractions.

Table 4-23. Mean measured wet sieve and MIR predicted soil fraction results for TL and AL positions (subset 2; *n = 12).*

	TL wet sieve *g C kg^{-1}* *n = 6*	*TL MIR* *g C kg^{-1}* *n=6*	*AL wet sieve* *g C kg^{-1}* *n=6*	*AL MIR* *g C kg^{-1}* *n=6*
TOC	50.77	44.59	31.64	30.21
POC	24.52	16.34	10.18	9.16
HOC		20.38		15.22
ROC		7.93		5.12
HOC + ROC	20.98		17.23	

4.13.24 ^{13}C NMR analysis for soil fractions

^{13}C NMR analysis of soil fractions in samples from subset 3 (Figure 4-7) shows that the proportion of TOC and the POC, HOC and ROC fractions is higher in the 0-5 cm soil layer than in the 5-10 cm layer (Table 4-24). Proportionally the concentration POC is greater than HOC in the 0-5 cm layer (21.2 g C kg^{-1} and 16.8 g C kg^{-1} respectively), whereas HOC is greater in the 5-10 cm layer than POC (9.6 g C kg^{-1} and 3.4 g C kg^{-1} respectively). The comparison between the TOC results obtained by Leco CNS2000 analysis and the sum of fractions determined by ^{13}C NMR analysis shows a strong relationship (R^2 = 0.75).

Table 4-24. TOC (Leco CNS2000) and ^{13}C NMR analysis showing soil C fraction results for TL and IR composited samples (n = 6 sites) for two depth increments (0-5 cm and 5-10 cm).

Site ID	TOC g/kg soil $_{(LECO)}$	POC g/kg soil $_{(^{13}C\ NMR)}$	HOC g/kg soil $_{(^{13}C\ NMR)}$	ROC g/kg soil $_{(^{13}C\ NMR)}$	Sum of fractions $_{(^{13}C\ NMR)}$
			0-5 cm		
haa	58.9	31.4	20.5	11.7	63.6
hab	24.5	9.2	11.9	4.8	25.9
kea	28.8	8.8	15.6	5.6	30.0
shb	44.1	14.4	16.6	7.1	38.0
str	48.0	35.6	19.9	12.6	68.0
wia	55.5	27.6	16.3	9.7	53.6
			5-10 cm		
haa	17.2	3.8	9.2	3.1	16.0
hab	14.9	2.0	9.7	2.7	14.4
kea	15.9	2.7	9.3	2.7	14.7
shb	15.8	2.5	8.8	2.6	13.9
str	18.8	2.9	11.1	3.3	17.3
wia	23.0	6.5	9.6	3.7	19.7

4.13.25 Comparison of results obtained by wet sieving with MIR predicted results and ^{13}C NMR analysis

When the MIRS predicted results for SOC from the whole soil (< 2.0 mm sieved sample) were compared with the SOC concentration results derived by Leco CNS2000 analysis on the samples selected for wet sieving, the correlation was found to be very strong (R^2 = 0.95) (Figure 4-59).

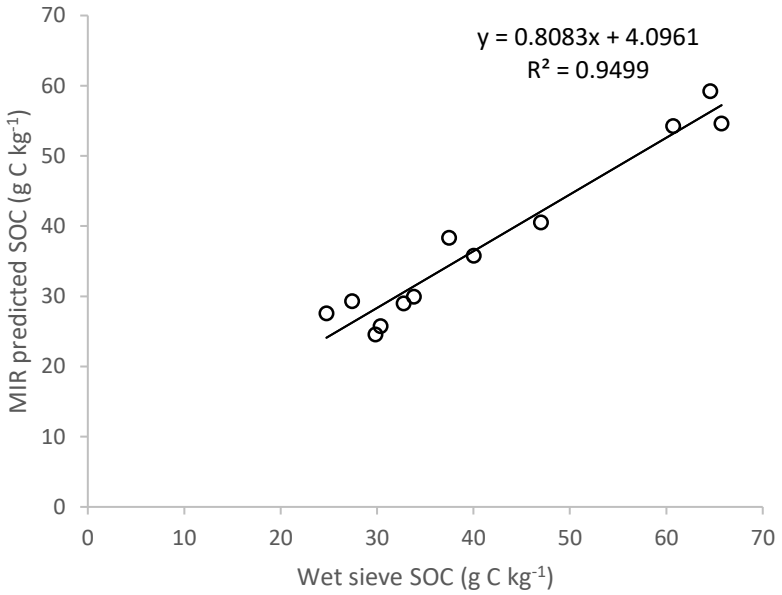


Figure 4-59. Correlation analysis for SOC comparing wet sieved and MIR predicted results.

Linear regression analysis comparing wet sieved, MIRS and ^{13}C NMR fraction results found strong relationships for most fractions. A summary of the correlations is shown at Table 4-25. The strongest correlations were obtained in the TOC fraction and the weakest in the HOC fraction. The overall weakest relationship was determined between MIRS and ^{13}C NMR for the HOC fraction ($R^2 = 0.47$). The correlation between the ^{13}C NMR determined fractions and the MIR predicted fractions (mean of TL and IR) in this subset shows strong relationships between the two determinations. As the ROC fraction was not removed from the wet sieved HOC fraction the correlation between wet sieved HOC + ROC and the MIR predicted HOC fraction was not as strong with an $R^2 = 0.67$. The R^2 for POC, HOC and ROC is 0.84, 0.47 and 0.95 respectively.

Table 4-25. Summary of correlation analysis for SOC fractions comparing results obtained by wet sieving, MIR and ^{13}C NMR analysis (WS = wet sieve, MIR = mid infra-red, NMR = ^{13}C nuclear magnetic resonance).

Analysis	*n*	*r*	R^2	*P*
			TOC	
WS/MIR	12	0.98	0.95	0.000
WS/NMR	6	0.93	0.86	0.008
MIR/NMR	6	0.90	0.81	0.013
			POC	
WS/MIR	12	0.90	0.81	0.000
WS/NMR	6	0.83	0.69	0.041
MIR/NMR	6	0.92	0.84	0.011
			HOC	
WS/MIR	12	0.82	0.67	0.001
WS/NMR	6	0.88	0.71	0.022
MIR/NMR	6	0.68	0.47	0.140
			ROC	
MIR/NMR	6	0.97	0.95	0.001

4.13.26 FullCAM modelled data compared with measured data for above ground biomass C and soil C pools

FullCAM modelling was used to predict AGB C and SOC within the BEPs after the calibrations detailed in Section 4.12.3 had been applied. Figure 4-60 shows an example of a FullCAM graphical simulation output for the AGB C, SCS and soil C pool data for a 13 year old BEP site. The simulation output shows that the C mass of trees starts at zero because the baseline was an AL used for grazing. Over the subsequent 13 years tree C mass increases to 85 Mg C ha^{-1}. The results show that for forest soil the baseline SCS was 48 Mg C ha^{-1} and declined for the first five years, reflecting the loss due to disturbance at time of BEP establishment. After 2005 the FullCAM simulation indicates that SCS gradually increases to 46 Mg C ha^{-1} by 2012.

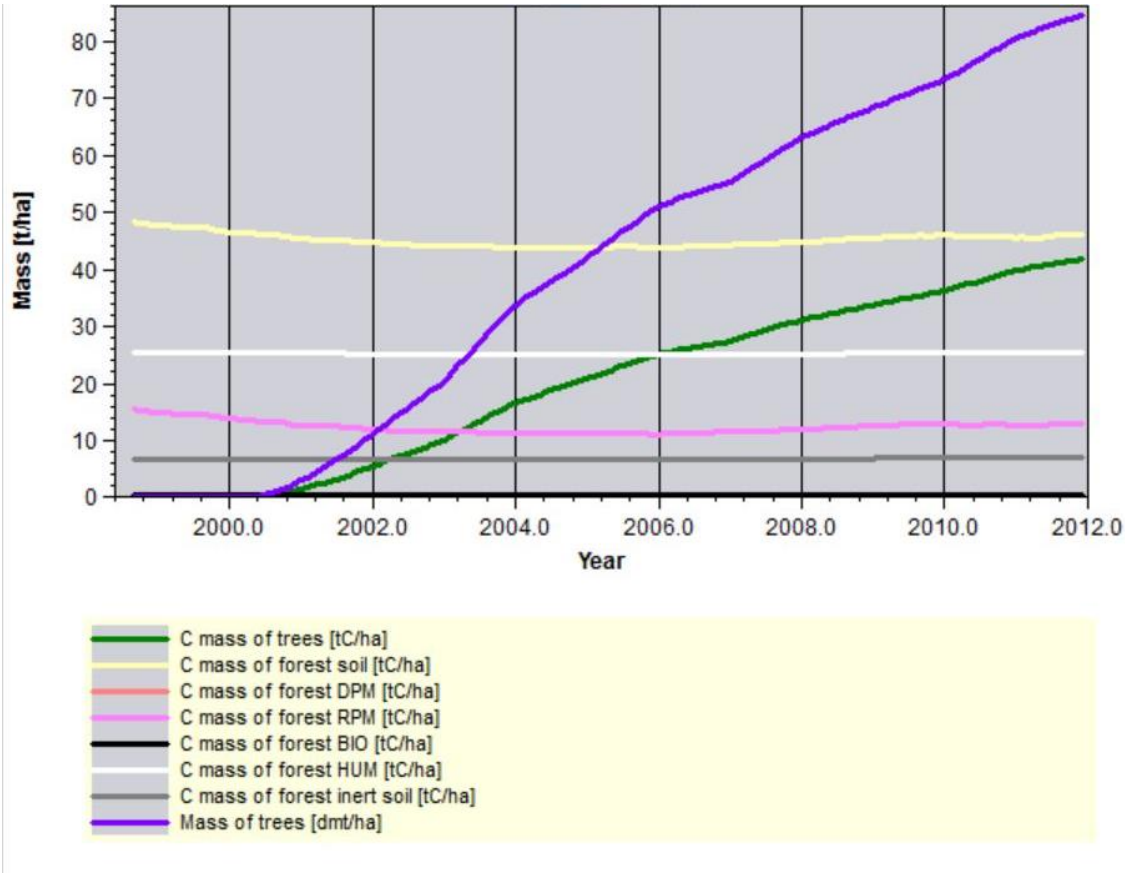


Figure 4-60. FullCAM simulation output for site "ana"; a 13 year old BEP site for the period 1999 to 2012.

Using the same site as the example in Figure 4-60 FullCAM simulations have been run for 100 years prior to establishment and 100 years post BEP establishment (Figure 4-61 and Figure 4-63 respectively).

The simulation for the 100 years prior to BEP establishment shows that SCS, POC (FullCAM terminology = RPM) and HOC (FullCAM terminology = HUM) have probably declined. The stocks of ROC are simulated to have remained constant over time.

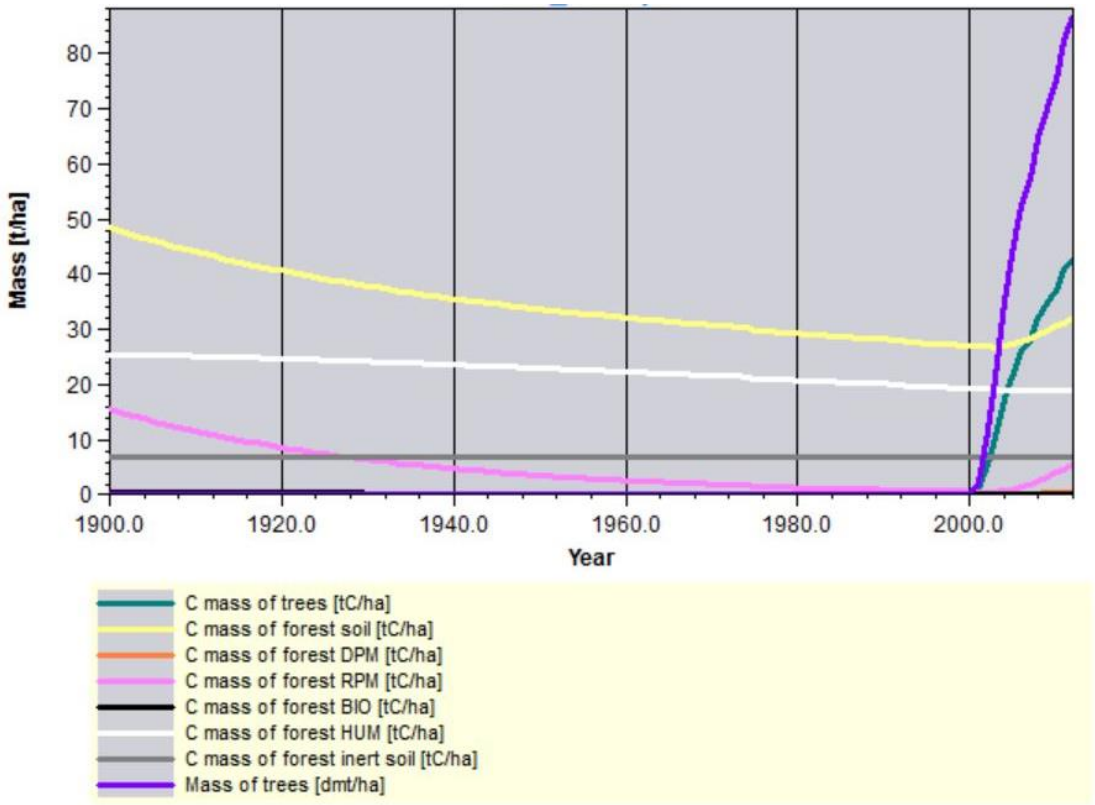


Figure 4-61. A 100 year simulation for BEP site 'ana' showing predicted backyear modelled trends for SOC and AGB C sequestration.

The out-year simulation indicates that the C mass of the AGB is predicted to increase to 65 Mg C ha^{-1} at approximately 40 years after establishment. After an initial decline, the SCS is modelled to gradually increase over the 100 year period reaching 83 Mg C ha^{-1} by 2098. The soil C pools of HUM (HOC) and RPM (POC) also have modelled gradually increasing trends over time.

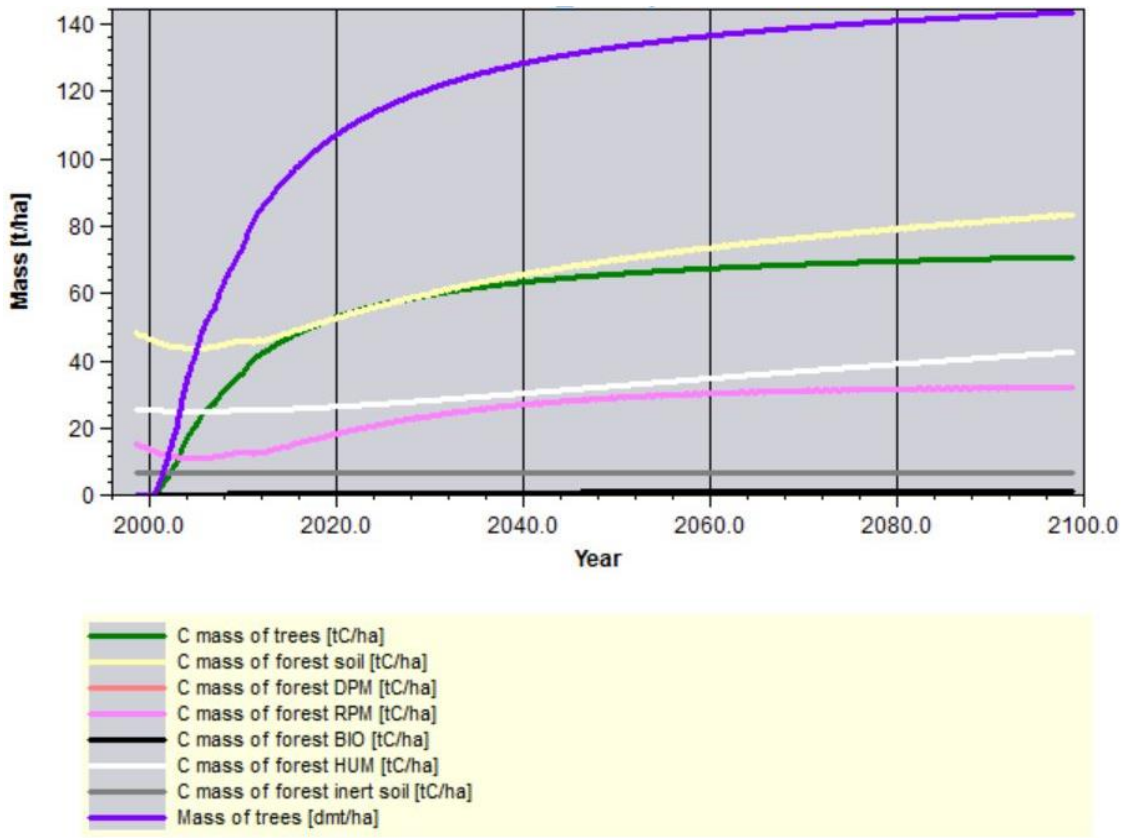


Figure 4-62. A 100 year simulation for BEP site 'ana' showing predicted outyear modelled trends for SOC and AGB C sequestration.

Linear regression analysis results shown in Figure 4-63 compare the modelled FullCAM AGB C data with the measured mean live AGB C for BEPs in each of the 20 sites. A very strong correlation $R^2 = 0.99$ ($P<0.001$) was found between FullCAM modelled data for AGB C and the measured live AGB C data (Figure 4-63). The correlation between FullCAM AGB C and measured mean total AGB C was similarly strong ($R^2 = 0.95$)

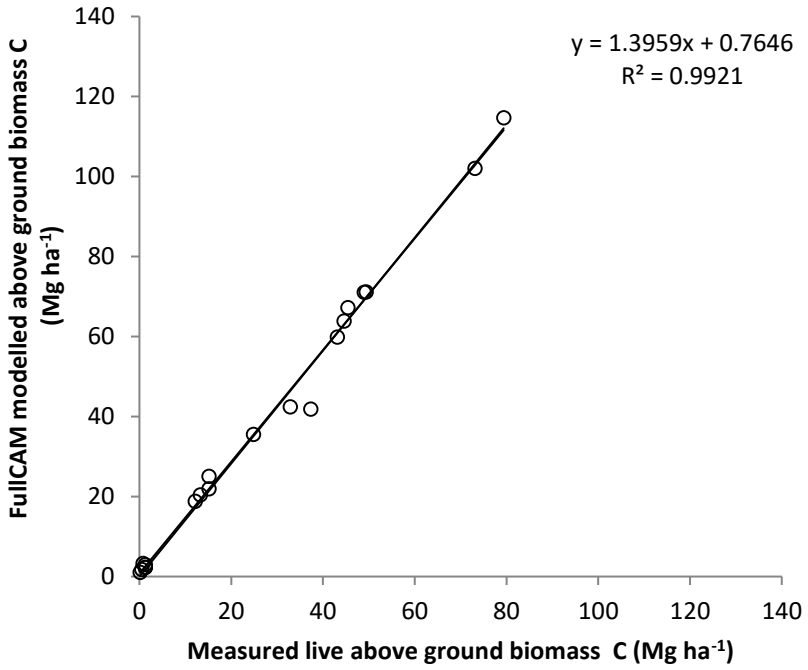


Figure 4-63. Correlation between predicted (FullCAM) and observed live above ground biomass with an R^2 of 0.99, $P < 0.001$.

FullCAM was also used to predict the soil C pools using the RothC, and defaults as provided in Section 4.12.3. A linear regression comparison between measured and FullCAM modelled data was undertaken. The best relationships were obtained when measured results were calculated on a fixed depth basis. Consequently, the correlation between measured and FullCAM results discussed are based on the fixed depth results. The HOC and ROC soil pools showed a strong correlation (R^2 = 0.77, P <0.001 (Figure 4-64) and R^2 = 1 respectively). However, the correlation between FullCAM and measured TOC and POC soil pools had very weak correlations (R^2 = 0.25, P=0.02 and R^2 = 0.17, P=0.07 respectively).

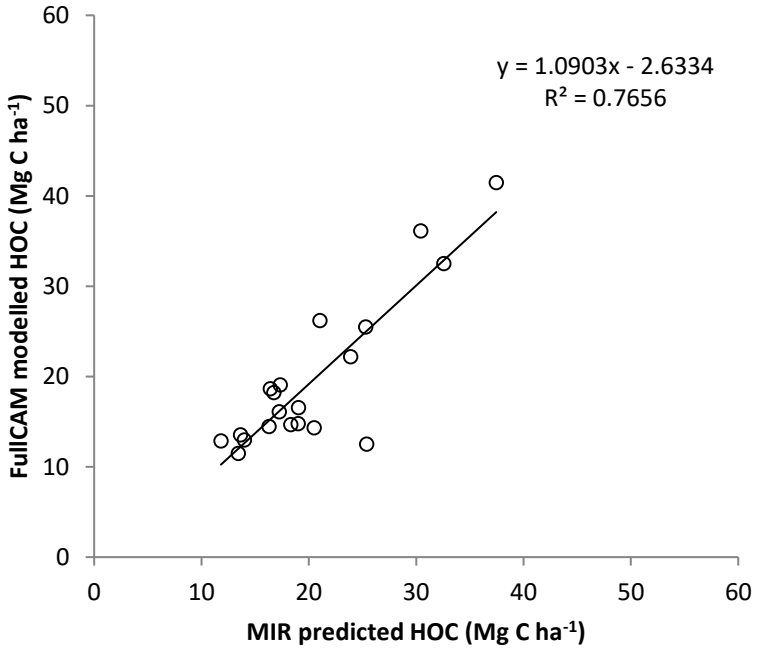
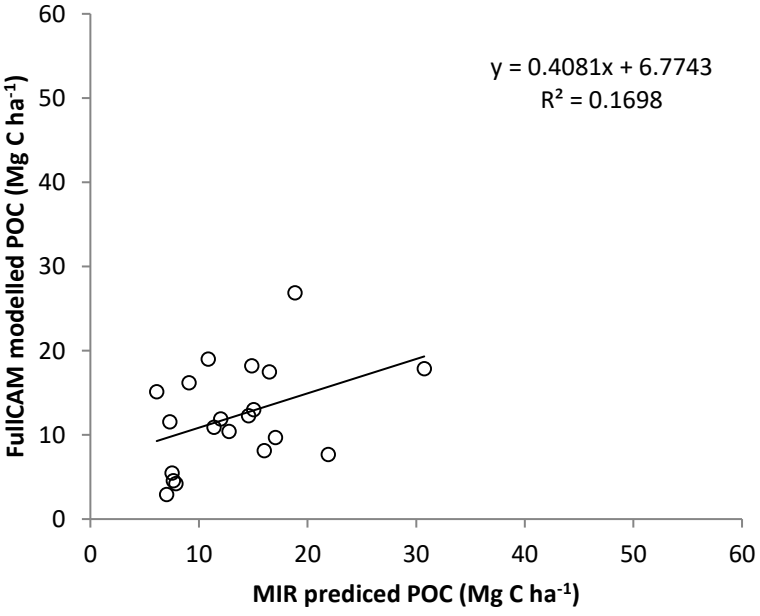


Figure 4-64. Correlation between MIR predicted HOC pool and FullCAM modelled HOC (Mg ha^{-1}) with an R^2 of 0.76 (Slope 0.97).



A temporal breakdown (by BEP age since establishment) of the FullCAM modelled results and the MIRS predicted soil C pools for POC, HOC and ROC (reported on a fixed depth basis) are shown at Figure 4-65 and Figure 4-66. Overall, FullCAM results are slightly higher than the measured data, although there is some variability. The largest discrepancy between modelled and measured data is shown in the POC pool for BEP sites aged 9-12 years. At ten sites FullCAM underestimates the measured POC results, and at seven sites FullCAM overestimates the measured POC results. At 11 sites FullCAM underestimates the measured HOC results, and at seven sites FullCAM overestimates the measured HOC results. The results show there is no particular trend with BEP age to suggest either an increase or decrease in SOC, and this is confirmed by ANOVA that suggests BEP age is not a significant factor for any of the SOC pools.

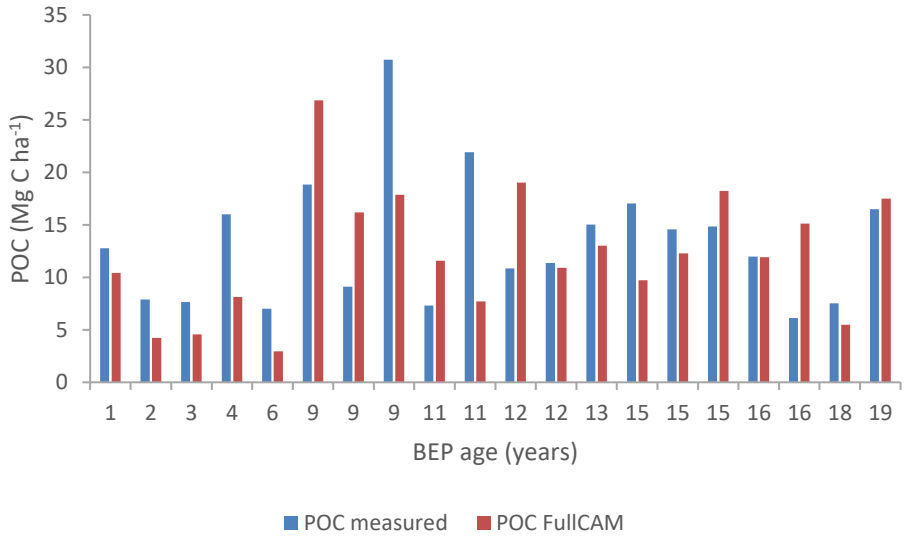


Figure 4-65. Comparison of POC measured (as predicted by MIRS) and POC FullCAM modelled results by BEP age.

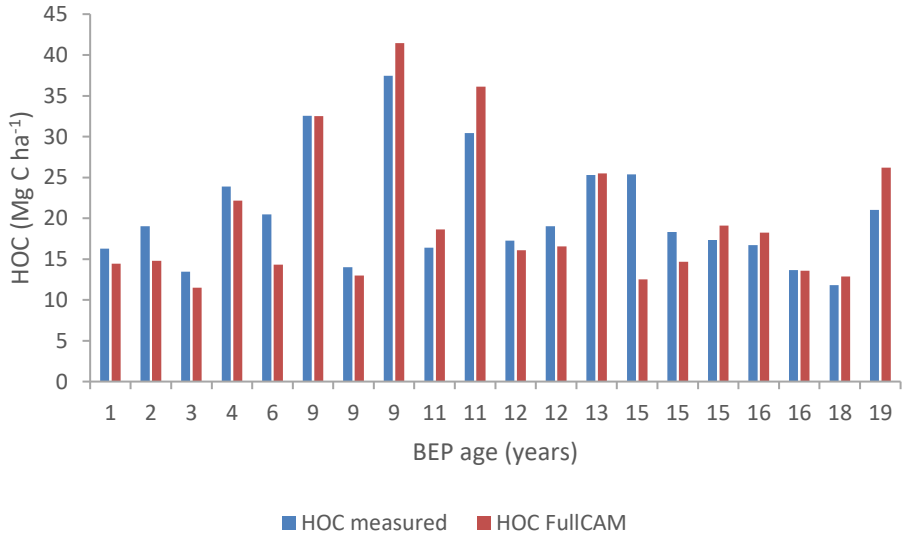


Figure 4-66. Comparison of HOC measured (as predicted by MIRS) and POC FullCAM modelled results by BEP age

The measured data can be compared with the simulated data on a site by site basis. The comparison displayed in Figure 4-67 shows the trends for SOC and the soil C pools to 30 cm depth, and AGB(C) for “ana”, a 13 year old BEP site. The FullCAM simulated data shows that over the first 5 years SCS decreases, and subsequently starts to accumulate, but does not achieve the pre establishment level. The FullCAM simulated data for SOC is higher than the measured data whereas for POC the measured data is higher than the FullCAM simulated data. For the other attributes the measured and simulated data is closely aligned. The measured data, which compares the SCS in the AL (taken to be representative of the baseline) is higher than the SCS in the BEP at age 13 (37.1 and 34.0 Mg C ha^{-1} respectively).

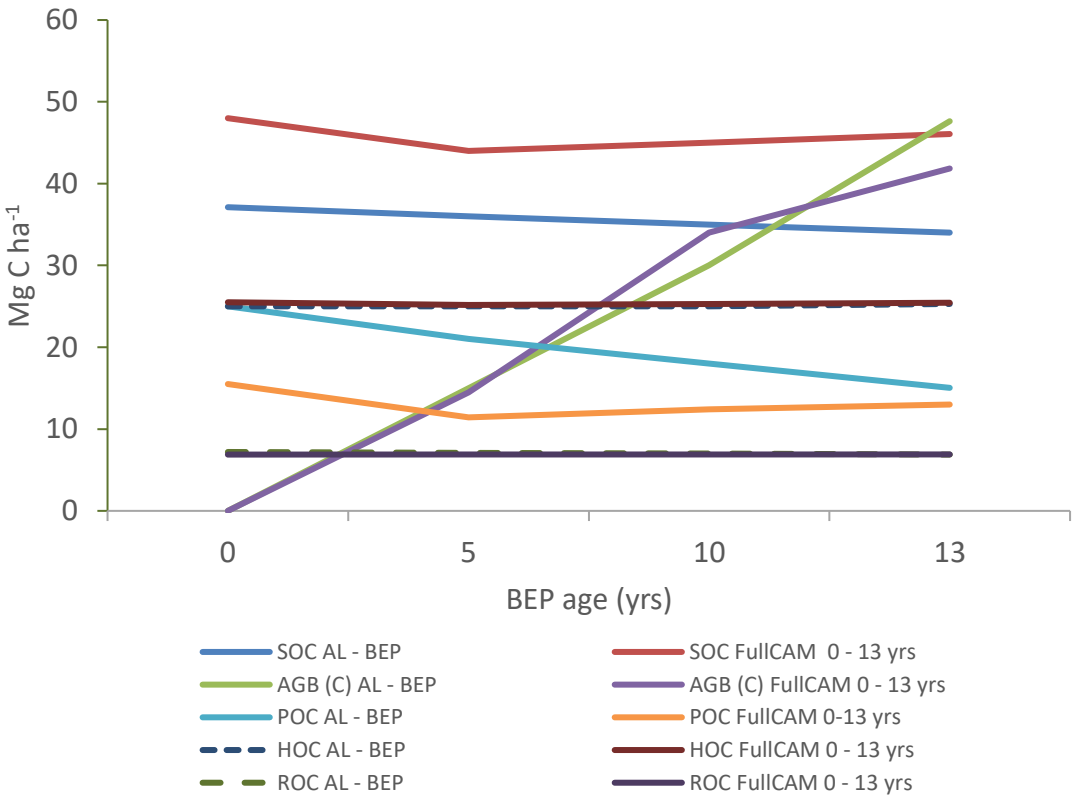


Figure 4-67. Measured data compared with FullCAM simulated data for site "ana" - a 13 year old BEP. Yr 0 represents the C status at the baseline AL. The measured data for years 5 and 10 are hypothetical and consequently give a straight line trend.

4.14 Discussion

It is widely reported that by planting forests on soils that have been depleted of SOC such as occurs following land clearing, extensive cropping, overgrazing and other agricultural activities, may provide an excellent opportunity to sequester C. SOC sequestration has the potential to mitigate increases of atmospheric CO_2 (Janzen, 2015). However, research investigating the conversion of agricultural land into forest shows that this type of LUC can result in either sequestration or loss of SOC (Guo and Gifford, 2002; Paul *et al*. 2002a; Morris *et al*. 2007; Whitehead, 2011). With results in the literature identifying such variable results following afforestation, the current research case study has explored two research questions. The first was to investigate whether afforestation of agricultural land with BEPs can lead to SOC sequestration. The second, to identify how the dynamic properties of BEPs influence the SOC sequestration potential.

The study design has allowed an assessment of the effects that BEPs have on agricultural soil by using a paired site approach on a chronosequence of 20 sites established between 1 and 19 years prior to the time of sampling. Study objectives have included quantifying the rate of SOC sequestration and AGB C and identifying the edaphic properties that might provide opportunities or constraints for the capacity of BEPs to enable tree growth and sequester C. The edaphic properties examined have included soil BD, pH, EC, TN and TP and nutrient ratios. The measured results from this study have been compared with modelled data derived from FullCAM. The results from the MIRS analysis has been used to identify soil C fraction changes, and these results have also been compared with modelled data using FullCAM. This section discusses the findings.

4.14.1 Spatial variation in SOC concentration and soil C stock in the agricultural land; establishment of baseline levels

To establish baseline SOC concentrations and SCS values (to 30 cm depth) the results for the AL position are summarised. The results of the current research suggest there has been an increase in $SOC_{(Heanes)}$ concentration in the AL profile commensurate with BEP age (Figure 4-9), particularly in the 0-5 cm soil layer. The sites with the oldest BEPs were found to have 3.4 g $100g^{-1}$ SOC and the sites with the youngest BEPs have 2.7 g $100g^{-1}$ SOC in the 0-5 cm layer. This is a misleading interpretation because no particular treatment was applied to the AL for the current research other than the business as usual management by the landholder; typically grazing on these sites. Instead, the trend highlights a spatial variation in SOC concentration that is most likely caused by land management and the natural heterogeneity of the landscape as influenced by inherent soil attributes or land management. It is possible that landholders who are motivated to establish BEPs on their properties also adopt changed land management practices to improve their AL, which in turn contributes to higher SCS. Such practices could include rotational grazing, avoided overgrazing, and a switch from sheep grazing to cattle. The

BEP sites included in the current research were positioned high or low on a catena or on the crest of hills. Consequently, the previous land use and condition of the soil, landscape position (topography) and parent material at each site will all have an influence on SOC. In terms of monitoring changes to SOC concentration following establishment of BEPs, the variation in (for example) the 0-5 cm layer of almost 1% for the AL position of the 20 sites SOC is an important finding. The variation can be indicative of either natural spatial variation, study design deficiency or sampling error.

Spatial variability of SOC both at the landscape and plot scale is well recognised, with parent material and soil texture, climate, land use and management, litter quality and quantity, slope and topography, nutrient availability and pH all being important factors that contribute to SOC stocks (Kasel and Bennet (2007); Chan *et al*. (2010); Grüneberg *et al*. (2010); Liu *et al*. (2012); Davy and Koen, (2013); Orgill *et al*. (2013), Wuest (2014)). For example, Jackson and Caldwell (1993) showed the variation in SOM across a 10m x 12m sagebrush-steppe grid ranged between 1.3 g $100g^{-1}$ to 7.4 g $100g^{-1}$, with the variability demonstrated across the whole field site as well as in association with proximity to individual plants. Wuest (2014) has demonstrated that SOC concentration can be variable depending on seasonal influences on plant and root growth and decomposition rates and can contribute approximately 2 to 8% variation around the mean SOC. Wuest (2014) also shows that agricultural management techniques such as till and no-till contribute to spatial variation in SOC. Orgill *et al*. (2013) found a strong relationship between parent material, and soil nutrients, especially TN and extractable S with changes to SCS.

Mineralogy has been widely reported to be an important factor for sequestration and stabilisation of SOC. For example Kasel and Bennet (2007) found that mineralogy and particle size fractions was positively associated with SOC on a number of forest types and land use conversions. To better understand the variation further, SOC analysis incorporating more replicates across the landscape would be necessary.

The mean SCS_{ESM} to 30 cm in the AL position for the 20 sites was found to be 41.12 Mg ha^{-1}. The trend in SCS in AL associated with each BEP age class (Figure 4-21) was consistent with the trend for SOC concentration (Figure 4-9). The SCS to 30 cm depth is within the range found by Chan *et al*. (2010) who undertook a paired site study in Central and Southern NSW to determine the SCS potential of pastures under different management systems including sites that contained P fertiliser and unimproved pastures without P application. Their study found a mean SCS in pastures of 46 Mg C ha^{-1} with a range between 22.4 and 66.3 Mg C ha^{-1}. Consequently, the mean SCS in the AL position found by the current research project is within an expected range for pastures within the biogeographical region.

4.14.2 Changes to SOC concentration following establishment of BEPs

To address the first research question as to whether LUC from AL to BEP leads to sequestration of SOC, the results for SOC concentration in the TL, IR and BEP positions have been compared with those found for the AL position.

Firstly, the results show that there is a significant difference in SOC concentration between the TL and IR positions (Figure 4-10 and Figure 4-11). Most notable is the difference in SOC concentration between TL and IR in the 1-5 year old cohort of BEPs. The TL has 2.2 g $100g^{-1}$ SOC and the IR has more with 3.4 g 100g-1 SOC in the 0-5 cm soil layer. A similar difference in SCS was also found (Figure 4-19).

There are two probable explanations for the difference in SOC concentration and SCS between the TL and IR positions and why the difference changes (Figure 4-12). The first explanation relates to the soil disturbance that occurs along the TL when the soil is ripped to establish the seed bed. Johnson, (1992) discussed results from a literature search to show that site preparation generally resulted in a net loss of soil C, but the extent of the loss depends upon the extent of the disturbance. The soil disturbance associated with establishing BEPs is similar to that which occurs during tillage operations. Direct drilling requires that a tyne implement be used to rip and disrupt the soil to prepare the seed bed. The equivalent agricultural practice of tillage, is widely reported to cause loss of SOC through mineralisation because plant residues and SOM becomes accessible to soil microbial activity and as a consequence of microbial actions CO_2 is released into the atmosphere (Turner and Lambert, 2000; Conant *et al*. 2001; Lal, 2004a). To support the case that minimal soil disturbance can lead to greater increases in SOC, results of a meta-analysis by Laganiere *et al*. (2010) found that when soil disturbance is minimised during preparation for plantations up to 15% more SOC can be sequestered. The soil in the IR position is not disturbed by the soil ripping process. Thus, the results which show SOC is higher in the IR position are a good example of how soil that is exposed to minimal disturbance can sequester more SOC. Secondly, the IR typically has a large grassy biomass present, especially in the younger sites that is the result of its previous land use as a pasture. The pasture retained in newly established BEPs, being enclosed behind a fence, is no longer exposed to grazing, and therefore the input of C into the soil in the IR position is from both the grass, and then over time from the trees.

When the results between TL, IR and AL positions in the 0-5 cm soil layer are compared the mean SOC concentration was found to be lower in the TL than the AL in the BEPs aged 1-5 years and 16-19 years, and higher in the IR than the AL in all age cohorts (Figure 4-12). The results suggest there is a trend whereby SOC increases in the fenced-out BEP sites during the first 15 years of the BEP chronosequence. However, SOC concentration in the TL and IR in the 16-19 year old BEPs is significantly lower than the paired AL position. Similar findings were

observed in a study into the changes of soil C following afforestation by environmental plantings (up to 29 years after establishment) in northern Victoria, (Cunningham *et al.* 2012). These authors also found that SOC concentration was initially slightly higher under the plantings than the adjoining pastures, but that there was an apparent decrease in SOC at the older sites.

When the results for TL and IR are averaged to represent the SOC concentration across the BEP, the results show that the mean $SOC_{(Heanes)}$ concentration in BEPs up to 15 years old is higher than in the paired AL (Figure 4-13). In the oldest BEP cohort (16-19 years) however, the mean SOC concentration is higher in the AL than the BEPs which implies the older BEPs have not increased SOC in the time since establishment; in fact they have lost some SOC, most likely to the AGB pool. One explanation for the loss is the possibility that grazing by feral animals (rabbits) and wildlife that can enter the BEP enclosures (particularly during drought) may be affecting the SOC. However, results in other studies have similarly shown variable results where afforestation of land previously used for agriculture can lead to either a loss or gain of SOC over time. Turner and Lambert (2000) for example found that SOC decreases in plantations remained evident up to 20 years after establishment. Poeplau and Don (2013) found that LUC from grassland to forest in Europe resulted in a loss of SOC in the top 10 cm at three of four sites, with no change in the fourth. Vesterdal *et al*. (2002) also found no relationship between site age and SOC in sites up to 30 years of aged due to a decline in the C content at soil depths below 5 cm. They explained the decrease was due to rapid decomposition of soil C from the products of the previous land use after afforestation and the build-up of grasses that occurs prior to canopy closure. These are features consistent with those of the BEPs particularly in the IR position before canopy closure and are therefore a feasible explanation for the low SOC in the TL and high SOC in the IR. In another study, Hoogmoed *et al*. (2012) conducted a meta-analysis into changes following afforestation of agricultural soil in Mediterranean regions of Australia with plantings up to 30 years of age. These authors also found that remnant forests had significantly higher SOC than either pastures or afforested soils. They suggested that although soil type and species mix contribute to high levels of SOC variation between sites that with time afforestation may eventually achieve SOC levels equivalent to those prior to land clearing. Therefore, it is feasible to expect that in time the BEP sites used in the current research project will eventually sequester SOC. For this to be determined however, an ongoing longitudinal analysis using the same sites would be necessary.

The results show variability in SOC concentration between the BEP positions (TL, IR and AL) (Figure 4-12), and the paired AL and BEP age classes (Figure 4-13). The variability may be associated with different mineralogy within the site (section 2.4.2) and particularly the presence of Fe and Al oxides and the interrelationship with SOM. The variability is most likely being effected by site specific factors such as planting configuration (ie block vs linear) (Figure 4-14) and direction of planting (Figure 4-15). Another factor that may have an effect include previous

land use which is reported to have an influence on the rate of SOC accumulation following reforestation. Although all the sites included in this study were established on farms located in the same biogeographical region, and all on land used for grazing, some legacy effect resulting from the earlier land use history may still be influencing the potential for change. For example, in a global review conducted by Paul *et al*. (2002a) previous land use was found to be an important factor in SOC sequestration. These authors found that reforestation of pasture sites typically resulted in a decline of SOC but resulted in an increase on land previously used for cropping, although time and soil depth also influenced the trajectory and rate of change. Laganiere *et al.* (2010) similarly identified LUC from pasture to forest resulted in the lowest proportion of change in SOC, whereas the change from crop land to forest was found by these authors to provide the best opportunity to sequester SOC.

To minimise the effects of climate for the current research project all the sites chosen were located within the same biogeographic region and thus received a similar rainfall pattern and experience similar temperature ranges. Consequently, climate is not considered to be a factor affecting the C sequestration potential of the sites included in the current research. However, climate generally is an important factor that can influence the sequestration of SOC following afforestation with mean annual precipitation (MAP) and potential evapotranspiration (PET) ratios having an effect on C sequestration rates (Conant *et al*. 2001; Robertson *et al*. 2016). For example, Paul *et al.* (2002) found that afforested sites in tropical and subtropical climatic regions sequestered more C than sites located in continental moist regions, whereas sites located in temperate/ Mediterranean climatic zones typically lost SOC. In another study, Silver *et al.* (2000) found reforestation of tropical soils on sites previously cleared and abandoned accumulated SOC at a faster rate than sites established on land previously used for pasture or agriculture.

However, associated with climate is site aspect which can influence temperature and evapotranspiration rates due to exposure to, or lack of sunlight, especially in the first years after the BEPs have been established i.e., before canopy closure. The results of the current research have found that the direction of planting, and topographic position have a significant effect on SOC concentration ($P = 0.004$ and $P < 0.001$ respectively) (Table 4-9 and Figure 4-15). BEPs planted in a block configuration had higher SOC concentration when the sites had been sown in a north-south (NS) direction than in an east-west (EW) direction. This finding is logical given that plantings established in an EW direction are more likely to receive direct exposure to sun and thus experience high evaporation rates for most of the hot part of the day leading to water deficiency which can result in lower biomass and thus lower SOC. However, in research exploring changes to SOC following conversion of agricultural land to forest Kasel and Bennett (2007) found weak associations between topography and SOC. Consequently, it is likely that a combination of factors, rather than just one (eg. topographic position alone) contribute to SOC concentration at any time.

The use of a chronosequence approach has been suggested to be inadequate for this type of research because different sites may have different edaphic or biophysical attributes and thus any changes or differences noted in a chronosequence may be an artefact of the sites used (Vesterdal *et al*. 2002; Laganiare *et al*. 2010; Sauer, 2015). However when the results of the current research project are evaluated, because a paired chronosequence method was used the apparent loss of SOC in the older BEPs is also seen when the SOC results for the BEP are compared with the AL. Consequently the apparent SOC decline in the older sites has been verified in two ways; by using a chronosequence of sites and by using paired sites.

Therefore, to address the first research question as to whether afforestation with BEPs can lead to SOC sequestration the findings of the current research show that in BEP sites up to 15 years after establishment there is on average a higher SOC concentration than the paired AL position (Figure 4-13). However, the BEPs aged between 16 and 19 years have on average a lower SOC concentration than the paired AL. These results indicate that in the first 19 years at least, BEPs do not achieve a net increase in SOC. However, from evidence presented in the literature it is likely that in the longer term, BEPs have the potential to sequester SOC above that found in the baseline paired AL position.

4.14.3 Soil Bulk Density

Soil BD is an important factor used to calculate SCS (Equation 4). Thus, the change to BD following afforestation with BEPs is briefly discussed here before an examination of SCS change under BEPs is undertaken. Broadly the soil BD results meet the expectation that the BD would be lower in the BEP than in the AL. However, the results show variability in the AL position at BEP sites from different BEP age classes,-suggesting that even though AL is being used as a base line figure against which changes caused by BEPs can be measured, there is natural variability between sites which can be explained by site factors such as soil texture, SOC concentration, management, geography, AGB and land-use history.

Not surprisingly, and consistent with research in the literature, this study has found an inverse relationship between soil BD and the SOC for both the AL and BEP soils with an R^2 of 0.5 and 0.4 for AL and BEP respectively (Figure 4-18). For example, Federer *et al.* (1993) found an inverse relationship between BD and the soil organic fraction of forest soils. The reason for such a relationship is the higher proportion of organic material being laid down by the AGB in the BEPs, to the proportion of mineral soil particles and aggregates, thereby making the soil BD lower.

When the BD of the TL and IR positions are compared with that of the age paired AL position, the results again show variability (Figure 4-17). In the 0-5 cm layer the AL has a higher BD than the TL in sites aged between 1-15 years. AL also has a higher BD than the IR in the 1-5 year cohort and also in sites aged between 11-19 years.

When the BD is averaged for TL and IR to get a weighted mean BEP BD result (Figure 4-17), and then the BEP result compared with the paired ALs, the AL position has a higher BD in sites aged between 1–15 years. This result indicates that there is a change manifest as a lower soil BD in sites up to 15 years. The change is most likely attributable to both the exclusion of livestock grazing which prevents soil compaction, and to increases in SOM provided by litterfall and root biomass in the soil beneath the BEPs. In similar research conducted in Ireland, Peichl *et al*. (2012) found the soil BD was lower in the surface layers of forested sites compared with the adjoining grassland. Those authors also attributed the change in BD to result from changes due to ploughing as part of the site preparation, the presence of tree roots and the absence of stock and machinery trampling in the forested area.

Ritter *et al*. (2003) found a temporal decrease in soil BD associated with forest stand age in the 0-5 cm layer. However, this BEP case study has identified an apparent diminution of the influence of the trees on soil BD over time because the BD in the oldest 16-19 year old cohort is the same in both the AL and BEP positions. Interestingly, the decrease in BD is also apparent in the AL position of the 16-19 year old BEP cohort of sites and thus it is likely that BD change is not only associated with BEPs trees, but is also due to other site factors or LUC within AL.

Broadly, the results of the current research show a trend which indicates a higher BD in deeper layers, but the difference is most pronounced between the 0-5 cm and 5-10 cm layers. This trend is typical of a soil profile (Ritter *et al*. 2003; Peichl *et al*. 2102) and is consistent with the roots, litter and organic material in the surface layer, and also due to soil turnover caused by the bioturbation action of biotic activity in the surface layer.

4.14.4 Soil bulk density and equivalent soil mass

An important consequence of the lower or variable soil BD in the BEP when compared with the AL position is the need to adjust SCS calculations for an ESM (Ellert and Bettany, 1995; Wendt and Hauser, 2013). The ESM calculation was undertaken to determine C mass per unit area representative of the two treatments (AL and BEP) by calculating a site reference soil mass for each layer at each site. Thus, for the current research the mass of C and the soil mass was calculated for each soil layer using Equation 10.

On average the results for both AL and BEP were higher for calculations derived using the fixed depth method than those adjusted using the ESM equation. This outcome has relevance particularly for use in calculations for determining C credits associated with SCS. For example, the Carbon Credits (Carbon Farming Initiative) methodology (Methodology Determination, 2014) for sequestering C in soils in grazing systems requires variation in soil BD to be considered, and SCS calculations adjusted accordingly by calculating the CS on an ESM basis to the 10^{th} percentile (as was used in Equation 10).

4.14.5 Rate of soil carbon stock change following establishment of BEPs in agricultural land

To report changes to the national C inventory, changes to SCS to 30 cm depth is the accepted standard under IPCC guidelines. Therefore, to partially meet objective one of the current research project the rate of SOC sequestration following establishment of BEPs has been used to quantify the change in the SCS following afforestation of AL with BEPs.

The SCS (0-30 cm) results are reported in this section on an ESM basis to take account of the changes to soil BD that have occurred following conversion from AL to BEP (see Section 4.14.4). To establish a baseline SCS (0-30 cm) prior to afforestation with BEPs the results for SCS in the AL position are discussed.

The SCS values (to 30 cm depth) for the AL position suggest there has been an increase in in SCS in the AL profile commensurate with BEP age (Figure 4-23). As explained in relation to the SOC concentration in section 4.14.1, the apparent increase in SCS in the AL with time could be associated with land management differences between the sites. The reasons behind the apparent increase are not clear, however, it can be speculated that the older BEPs were established by innovative landholders who have adopted land management strategies to increase SOC in their pastures, whereas the younger sites may have been established by more mainstream landholders who have not adopted land management practices to increase SOC. Consequently, there may be a systematic explanation for the variation in SCS in the AL position over time.

When the SCS to 30 cm in the AL is averaged between all sites to establish a baseline steady state prior to intervention by BEPs, the mean SCS was found to be 41.1 Mg C ha^{-1} (Figure 4-21 and Figure 4-68). When compared with the estimated SCS present in the region prior to European settlement of 57 Mg C ha^{-1} discussed in Section 4.3, it is possible that the AL has lost approximately 16 Mg C ha^{-1} since European settlement. This level of loss is consistent with modelled estimates proposed by Chappell *et al*. (2015) who calculated that between 0.04 and 0.19 Mg C ha^{-1} yr^{-1} of SOC was lost due to wind and water erosion in the Oceania region. When the results of the C loss estimated by Chappell *et al.* (2015) for the Oceania region are extrapolated over 200 years of land clearing and agriculture in the region, we arrive at the equivalent of a loss of SOC in the range between 8 and 38 Mg C ha^{-1}.

At a national level, Viscara Rossel *et al*. (2014) have estimated that the mean SCS in topsoil (depth unspecified) across Australia is approximately 29.7 Mg C ha^{-1}. Their estimate accounts for all bioclimatic zones, vegetation cover and land-use and accordingly is lower than the SCS in the AL, but supports the SCS measured in the AL as reasonable.

Table 4-11 shows the variation in SCS across all BEP sites. The differences are not temporal in that there is no trend showing an increase in SCS with time since establishment. Two sites in particular stand out – “cap” and “pyl”. These two sites have the largest negative values of -3.0

and -8.3 Mg C ha^{-1} (ESM) since establishment. The most likely explanation for the two sites to have an apparent SCS deficit is associated with the landform on which they have been established (Table 4-1) with both being located on a waxing upper landform (Speight, 2009). The topographic shape of this landform, high in the landscape potentially limits the capacity of the soil to retain water and thus promote plant growth, root development and nutrient cycling necessary for C sequestration.

After BEPs have been established the results indicate there is a higher mean SCS in the BEPs than in the paired ALs in sites aged between 1-15 years (Figure 4-21). For C accounting purposes this is an encouraging outcome. However, in the 16-19 year old BEPs the mean SCS has started to decline and is now lower in the BEP position than in the adjoining AL. As a result of the trend displayed in the older plantings, the long term trajectory of the SOC sequestration potential of BEPs is uncertain.

A conceptual model depicting the various stages of SOC loss and accumulation within the BEPs is shown in Figure 4-68. The model shows the flows of SCS since land clearing, incorporates the period when grazing was conducted on the land, and shows the accumulation of SOC following establishment of BEPs. However, the projection shows an uncertain trajectory for the state of future SCS within the BEPs. Other authors have observed varying results as either gains or loss of SOC following afforestation. For example Paul *et al*. (2002a), in a global review of afforestation, noted that in the first 30 years following afforestation of land previously used for agriculture there is initially typically a decrease in soil C. However, the authors also observed that 30 years after afforestation there is potential for the SCS to be greater than was present in the agricultural soil at the time of planting.

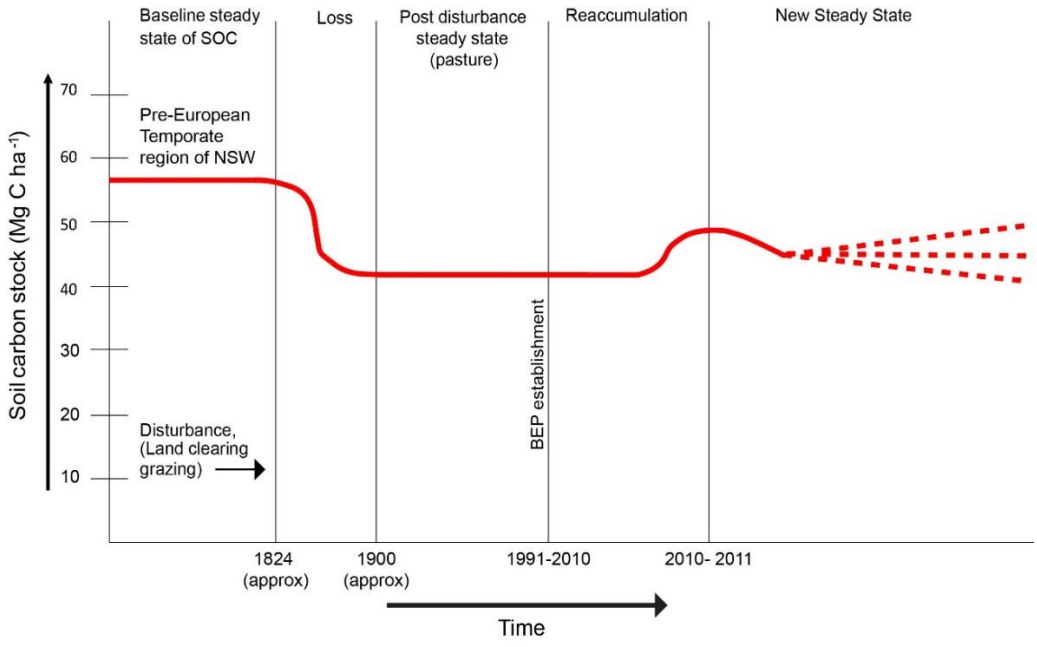


Figure 4-68. Conceptual model of SOC accumulation following establishment of BEPs into AL (After Johnson, 1995). The dashed lines in the New Steady State phase are conceptually based because the actual trajectory is uncertain.

When examined on a site by site basis (Table 4-11), the results show there is significant variation in SCS change between sites and BEP ages. Such a result is not surprising given the variation in SOC concentration and BD discussed previously. The SCS ranges from a loss of 8.33 Mg C ha^{-1} (0-30 cm) in a site aged 19 years, i.e., less SOC in BEPs than the AL; to a gain of 16.64 Mg C ha^{-1} (0-30 cm) in a site aged 12 years. Notwithstanding the significant loss in one 19 year old site the results suggest that overall there has been a net increase in SCS following establishment of BEPs. The results for SCS change show the BEPs included in this study have sequestered on average 3.54 Mg C ha^{-1} over the 19 year period in soils to 30 cm depth (Table 4-11). This result indicates that establishment of BEPs can lead to SOC sequestration, but as previously highlighted, only small and variable quantities can be sequestered depending on site factors and previous land use.

The mean annual SOC sequestration rate across all sites was found to be 0.53 Mg C ha^{-1} yr^{-1} to 30 cm depth over 19 years. In reality, however, the large variation between sites indicates that the averaged result should be viewed with caution as it is not necessarily indicative of the real SCS change over time. On a site by site basis the rate of SOC sequestration was found to be between a loss of 1.73 Mg C ha^{-1} yr^{-1} and a gain of 4.67 Mg C ha^{-1} yr^{-1} (based on ESM calculations) (Table 4-11). Site factors such as tree stocking density, planting configuration and species mix are likely to influence the sequestration rate. Climate, notably rainfall, plays a role in SOC sequestration. For example, Guo and Gifford (2002) found that conversion of pasture to plantation resulted in a significant loss of SOC in higher rainfall areas. In another study Turner and Kelly (1977) identified tree species as important in determining the quality and quantity of litter and root exudate returned to the soil and thus the trajectory of SOC sequestration. Silver *et al*. (2000) found, in a review of research into the SOC sequestration potential of tropical soil following afforestation of land previously used for agriculture and pasture, that soil C accumulated at a rate of 0.41 Mg ha^{-1} yr^{-1} over a 100 year time frame. They also found that during the first 20 years the SOC sequestration rate was 1.3 Mg ha^{-1} yr^{-1} which is a much higher rate than that seen in the BEPs sites in this case study, located in temperate NSW.

When the sites are grouped into the four age classes (Figure 4-21) the results show that the BEPs have on average more SOC to 30 cm than their paired AL for the three youngest age cohorts. However, the 16-19 year old cohort of BEPs has a lower SCS than the AL. This trend suggests that BEPs may contribute to an increase in SCS during the early years of tree growth. However, there are probably other factors influencing the C sequestration for the older sites. The influences could include:

- Adverse effects of water deficiency during the millennium drought (2001-2009), leading to high mortality rate in older trees;
- Site effects (climate, aspect, management, edaphic characteristics);

- Tree stocking rate is lowest in the 16-19 year old cohort of BEPs, consequently, it is possible that the tree stocking rate in these BEPs is not sufficient to lead to SOC sequestration;
- Soil nutrient depletion in the older BEP sites; and
- Widening of soil C:N ratios in the older BEP sites.

Water deficiency has potentially been an issue for the BEP sites included in this study as they were all established prior to or during the millennium drought experienced in the southeast of Australia between 2001 and 2009 (van Dijk *et al.* 2013). Tree deaths in environmental plantings during this period were noted by Geeves, *et al*. (2008) and Semple *et al*. (2010) who attributed the losses due to the high tree stocking rate of the plantings and to rainfall deficiency confounded by shallow soils as the two principal drivers of tree mortality. SOC sequestration occurs when the input of SOM is higher than losses incurred from decomposition.

Details regarding the AGB growth, tree stocking rate and litter in BEPs will be considered in the next section. The effects of edaphic factors, tree stocking rate, nutrient availability and nutrient ratios are discussed separately in subsequent sections.

4.15 Aboveground biomass growth in biodiverse environmental plantings

As discussed in Sections 4.2 and 4.4 reforestation of agricultural land with BEPs provides a wide range of ecosystem services. Additionally, trees provide opportunity to sequester atmospheric CO_2 in both the soil and AGB. BEPs are a unique form of afforestation typically having a high tree stocking rate (compared with other forms of forestry), and being comprised of a mix of native tree and understorey species. They are often established on land forms such as erosion gullies or crests of hills that are not generally suited for productive agricultural activities and therefore the AL may not always provide a good baseline.

For the purpose of establishing a baseline for biomass C, NPP in the AL has been assumed to be zero. Tree growth in the BEPs has been taken to be representative of increases in NPP relative to the AL. The results of the current research have found an increase in aboveground tree biomass associated with BEP consistent with expectations of tree growth. However, the extent of biomass increase was found to vary between sites, stand age and between transects within sites. A temporal increase was apparent when the results were averaged and considered on an age class basis rather than individual site age basis. The rate of tree growth was found to be greater in BEP sites nine years and older than the younger sites. Comparatively, the BEP sites have been found to have a similar AGB growth rate to other mixed species plantings and forests. The overall mean rate of AGB increase in the BEPs of 2.4 Mg C ha^{-1} yr^{-1} was lower than that found by Cunningham *et al*. (2015) of 3.1 ± 0.9 Mg C ha^{-1} yr^{-1}. However, the lower growth rate can be explained by the inclusion in this study of BEP sites younger than five years

old and the absence of sites older than 19 years. In a review of the literature Silver *et al*. (2000) found the mean rate of AGB increase in tropical areas, comprising wet, moderate and dry locations was 2.4 Mg ha^{-1} yr^{-1} over a 60 year period, but up to 6.2 Mg ha^{-1} yr^{-1} during the first 20 years of growth. Several factors including tree stocking rate, site configuration and edge effect were found by these authors to be factors affecting the quantity of biomass on a site by site basis. These factors have been examined for this study and details are presented in this section.

Tree stocking rate was significantly higher in linear plantings than block plantings. The reason for this is the tree row spacing is typically wider in block plantings than in linear plantings. The land space allocated to linear plantings is usually limited so as to take up as little arable farm land as possible. These type of plantings are typically located along fence lines. On the other hand, block plantings are typically established to rehabilitate large areas of degraded land such as eroded rocky hills and crests.

BEPs planted in a block configuration were found to have a lower tree stocking rate (Figure 4-30), basal area and AGB C (Figure 4-38) than those planted in a linear configuration. This finding aligns with the results from a large study of 605 BEP sites across Australia undertaken by Paul *et al*. (2015). However, in earlier work undertaken by Henskens *et al*. (2001) to explore the effects of spacing and layout on tree growth in a farm forestry context block plantings were found to have a higher tree stocking rate and slightly higher stand volume than linear plantings. However, Henskens *et al*. (2001) also found that linear plantings had a significantly higher basal area and tree volume than block plantings. Plantings in linear configurations and isolated trees were found by Henskens *et al*. (2001) to allow trees to gain better access to light than trees in block plantings, and concluded that greater light access translated in to higher tree growth. Consequently, the potential to sequester C in AGB is likely to be better in plantings aligned in a linear configuration than in block plantings.

Edge effect influencing AGB was found to be a significant factor for BEPs with the mean DBH of trees located on outer rows 5.08 cm, and inner rows 3.99 cm. The average AGB C was 7.28 and 5.11 Mg C ha^{-1} on outer and inner rows respectively. A similar difference was also found for tree DBH and live AGB when analysed for eucalypt and acacia species separately (Figure 4-37). The effect of site configuration and trees situated on the outer row shows that eucalypts have higher biomass than acacias in linearly configured sites and acacias have higher biomass on block sites than eucalypts. The finding that eucalypts on outer rows develop higher AGB aligns with the research done on *E. globulus* by Henskens *et al*. (2001) who found wide spacing maximised tree growth due to greater light interception, nutrients and water and thereby become the dominant species in this situation. However, acacia growth is lower on the outer rows than eucalypts in linear plantings. The reason for the different growth is probably because the eucalypt trees achieve dominance over the acacias due to their larger canopy and stem growth

when growing in the heavily tree stocked linearly configured sites, and thus supress acacia growth in this situation. The converse is possibly true for the acacias growing in the block configured sites where the tree stocking rate is lower, thereby allowing acacias to gain access to water, nutrient and light resources necessary for their growth.

4.15.1 Site tree stocking rate

Tree stocking rate represents the number of standing trees, both dead and alive, in a site. The tree stocking rate was observed to decrease with site age (Figure 4-28 and Figure 4-29). The oldest BEPs were found to have the lowest tree stocking rate in both linear and block planting configurations (Figure 4-30). Possible reasons for the lower tree stocking rate in the older sites include:

- High initial tree stocking rate due to seedling germination in the youngest plantings before self thinning occurs;
- Tree death over time (Table 4-12) possibly in association with the millennium drought during 2001-2009;
- Widening of the C:N ratio in the older sites. Wide C:N ratios can limit nutrient cycling (Figure 4-51). BEPs aged 16-19 had a high C:N ratio of 26 whereas the younger sites had a mean C:N ratio of 18.
- Sowing strategy, operator or mechanical variation over time; or
- Other site specific effects such as aspect, land form, elevation, climate, soil type or geology.

Most likely the reason for the high tree stocking rate found in the very young sites occurs because the plantings have been established using a direct seeding method where optimum quantities of seed are used. The seedlings emerge when conditions are optimum for their germination. The timing of emergence can be different for different species, depending on optimal seasonal, soil moisture, temperature and nutrient availability. In an early study by GA into the early germination of seedlings in BEPs the average survival rate was found to be 60 % in eight year old sites and 100 % in two year old sites (GA unpublished data, 1998). The proportion of biomass C contributed by dead trees in the sites included in the current research project is on average 33% in the BEPs older than 11 years, whereas only 1 % - 2 % is of the biomass is contributed by dead trees in BEPs aged between 1 and 10 years. However, it is probable that there is high tree mortality in seedlings and younger trees, but due to their size they leave little evidence of their presence.

4.15.2 Litter accumulation within BEPs

The mean dry mass (DM) for total litter within BEPs was calculated using the sum of the O1 and O2 layers, in the TL and IR positions for 53 transects from 11 BEPs aged between 11 and 19 years. The results found the mean litter mass in the TL position was 26.2 Mg DM ha^{-1} and ranged from 6.4 to 46.2 Mg DM ha^{-1}. The mean total litter mass for the IR position was 17.2 Mg DM ha^{-1} and ranged from 3.4 to 35.0 Mg DM ha^{-1}. The mean of total litter stock for TL and IR was used to calculate litter biomass for BEPs. Thus, the results showed BEPs had on average 21.7 Mg DM ha^{-1} with the range from 6.6 to 35.5 Mg DM ha^{-1}. The average annual rate of litter accumulation was 1.6 Mg DM ha^{-1} yr^{-1} but varied between sites (0.6 - 2.8 Mg DM ha^{-1} yr^{-1}).

The rate of litter accumulation measured in this case study is consistent with, for example, a litter fall study carried out by O'Connell and Menagé (1982) on *E. diversicolor* forests located in south-western Australia. The results of their study showed that litter fall increased commensurately with stand age (1.1 Mg ha^{-1} at 2 years and up to 10.2 Mg ha^{-1} at a mature forest stage). In another study, Pressland (1982) reported an annual litter accumulation of 2.5 and 3.75 Mg ha^{-1} yr^{-1} in a mixed eucalyptus forest and eucalyptus woodland. More recently in a study, in which the results of this case study have been incorporated, England *et al*. (2016) found that the total litter stock from 53 plantings of ≤ 29 year old mixed species plantings (BEPs) ranged between 2.4 and 30.6 Mg DM ha^{-1} and the rate of litter accumulation was between 0.22 to 2.27 Mg DM ha^{-1} yr^{-1}.

Explanations for the high litter content present in the BEP sites may include low moisture content within the BEPs due to canopy interception and uptake of water by the trees which is exacerbated due to the highly stocking density of BEPs. This combination of factors could contribute to a relatively slow decomposition rate and therefore a high rate of litter accumulation. However, in a study undertaken in New Zealand, Guo and Sims (1999) found no evidence to support the hypothesis that tree stocking rate affected the rate of litter decomposition. Instead they noted that litter from different eucalypt species decomposed at different rates and that season also affected the decomposition rate, with the rate of decomposition in autumn being the highest. They also noted that temperature and moisture were optimum during autumn for microbial action and thus litter decomposition. Consequently, it is less likely that tree stocking rate *per se* is a factor influencing the accumulation of litter in BEPs, but rather other associated factors including substrate quality, poor light penetration reducing decomposition, limited moisture which inhibits the abundance of micro-organisms and microbial activity are important.

Evidence for the change is supported with the wider C:N ratio in the older BEPs, suggesting that nutrient cycling and decomposition have been retarded thereby allowing the high rate of accumulation of litter within the BEPs.

The DM content found in the BEP litter measurements was initially converted into a C content by using a conversion factor of 54.3 % (Gifford, 2000). Following the conversion, the mean annual accumulation of litter C determined by using the mean of both the O1 and O2 layers for the TL and IR positions was found to be 0.8 Mg C ha^{-1} yr^{-1} for BEP sites aged between 11 and 19 years. When sampling for the current research project it was observed that sites younger than 11 years had not accumulated tree litter. Instead the understorey was comprised mainly of grasses and shrubs. Paul *et al*. (2003) noted that litter fall became a significant process after canopy closure. A similar observation was made by Vesterdal *et al*. (2002) who, in a European afforestation study found that the O-horizon had begun to accumulate in forest stands older than 8 years. Consequently stand age is the most likely explanation for the dearth of litter present under young BEPs. Vesterdal *et al*. (2002) also found the accumulation differed between the different tree species. For example, oak stands in their study were found to sequester approximately 3.0 Mg C ha^{-1} over 30 years whereas spruce stands sequestered up to 9 Mg C ha^{-1}. These rates are the equivalent of 0.1 and 0.3 Mg C ha^{-1} yr^{-1} respectively and are somewhat lower than the litter accumulation rate found for BEPs. A probable explanation for the difference between the BEP litter accumulation rate identified in this study and those found by Vesterdal *et al*. (2002) is to do with the relatively high tree stocking rate in a small area allowing BEPs to produce and accumulate higher litter density.

In a recent study by England *et al*. (2016), the litter from 65 MSEPs from across Australia (including sites from this study) was analysed, and a mean accumulation litter C rate was found to be 0.4±0.02 Mg C ha^{-1} yr^{-1}. The reason for the different findings between England *et al*. (2016) and those found by this current research, are probably the result of sampling and analytical methods applied. For example, the current research project used 54.3 % as a conversion factor from the DM result and collected litter from 53 BEP transects from sites aged between 11 and 19 years; the conversion has most likely resulted in an over-estimate of the C content of the litter. In comparison, work by England *et al.* (2016) actually analysed litter for C content on 310 samples from 29 environmental planting sites aged between 6 and 29 years. Their analysis found litter from the 2 to 25 mm fraction had a mean C concentration of 45.4% (range 42 - 28%) and analysis of the < 2 mm fraction showed a mean C concentration of 27% (pers comm J. England 3 Sep 2015). When the C % of litter for the 2 to 25 mm fraction was applied to the O1 layer from this study, and 27% C to the O2 layer, the annual average rate of litter C increase in BEPs was calculated to be 0.3 Mg C ha^{-1} yr^{-1}, and consequently is in line with the mean litter C accumulation rate found by England *et al*. (2016). Therefore, it is likely that, for BEPs at least, the percent of C in litter proposed by Gifford, (2000) is an overestimate because it doesn't allow for the two fractions of litter, where the O2 layer is difficult to extract without some mineral soil contamination from the A1 layer.

The relationship between the AGB growth and SCS is considered in the next section.

4.16 Relationship between above ground biomass growth and soil C stock

The second research question sought to explore the dynamic properties of BEPs, the AGB and SCS (0-30 cm). The question underpinned the proposition that tree growth would be linked with increases to SOC. Such a relationship, if it existed, would be useful for predicting changes to SOC in BEPs, and consequently provide a rapid way of accounting for the whole CS in BEPs. The results from the chronosequence of BEP sites studied for the current research have shown that there is no temporal relationship between tree growth and SOC. For example the correlation between AGB and SOC from the 0-5 cm layer was found to be very weak (R^2 =0.29) (Figure 4-40a). Thus it is not possible for the results of the current research project to be used for the development of a C accounting model.

Spatial heterogeneity of SOC concentration and SCS has been widely reported (e.g., Harper *et al*. 2012; Cunningham *et al*. 2012) and is likely to be a major factor inhibiting any correlation between soil C and AGB. For example, Harper *et al*. (2012) suggested that spatial variability of SOC could mask expected differences following afforestation of agricultural land, as could the lapse of insufficient time since reforestation. In another study examining changes to SOC following afforestation of pastures with mixed species plantings, Cunningham *et al*. (2012) similarly found variability between sites.

The case for there being insufficient time for any noticeable increasing trend in SOC since reforestation has been suggested by Paul *et al*. (2002a). These authors proposed that any noticeable change was most likely to become apparent only 30 years after afforestation. They argued this was because pastures already had a relatively high SOC content and thus it would take a long period of time for trees to contribute additional C. Harper *et al*. (2012) suggests that a relationship between biomass and SOC is possible when reforestation occurred > 50 years previously and when sampling was conducted below 30cm due to the long term contribution of coarse roots to the SOC pool.

A further analysis of the proportion of AGB C, SCS and litter C has been undertaken in this current study. The results show that the mean AGB C amongst the 20 sites is 31.4 Mg C ha^{-1}. The mean SCS to 30 cm in BEPs is 44.2 Mg C ha^{-1}, and the mean litter pool is 7.6 Mg C ha^{-1}. Proportionally SCS contributes 53% of the C pool and AGB contributes 38% of the C pool and litter 9% of the SCS in BEPs. (Figure 4-69).

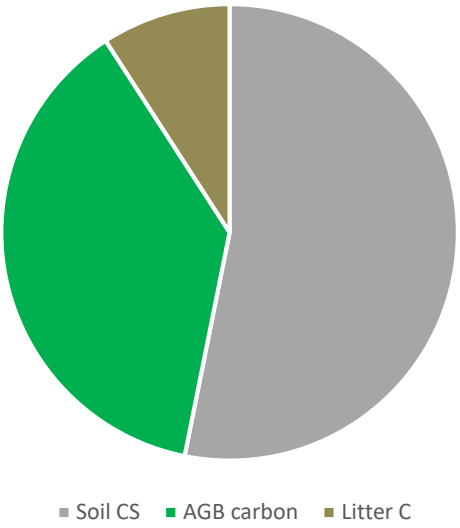


Figure 4-69. Proportion of SCS (Mg C ha^{-1}) in soil, aboveground biomass and litter pools in BEPs age 1-19 years.

The mean result for AGB C shown in Figure 4-69 is relatively low because all the sites, including young sites that have little to no AGB growth, have been included in the calculation. The young sites have been included in the above calculation, because under the rules of the Carbon Credits (CFI) Methodology (Methodology Determination 2014), plantings that have the potential to attain a minimum height of 2 m are eligible offsets projects.

In a study of 117 environmental planting sites across temperate Australia (including the results of the current research), England *et al* (2016) found the mean AGB to be 62.8 Mg DM ha^{-1}. Assuming the carbon fraction of the dry matter is 50%, the mean AGB finding by England *et al.* (2016) is the equivalent of 31.4 Mg C ha^{-1} which is the same mean AGB C result as found in the current research.

In a global meta analysis exploring pools of C in forests, Dixon *et al*. (1994) estimated that the CS of forest vegetation in Australia was 45 Mg C ha^{-1} and Australian soils to 1 m depth had an SCS of 83 Mg C ha^{-1}. The relationship between the measured data from this current research project and the estimated values from Dixon *et al*. (1994) has considerable uncertainty due to different methodologies applied in each study, the different land uses, soil depth, allometrics, and forest types and ages. Not-with-standing the uncertainty, the proportion of SCS to AGB in this study is 58% to 42% respectively (with litter excluded) and in Dixon *et al*. (1995) the proportion was found to be 65% and 35% respectively. In another study exploring the C sink potential of global forests Pan *et al*. (2011) estimated that forests have approximately 861 ±66 Gt C. The proportion of C in forest soil (to 1 m) was estimated to be 44%, with 42% in live above and belowground biomass, 8 % in deadwood and 5 % in litter (Pan *et al*. 2011). Therefore, the proportional distribution of AGB C and SCS within BEPs is similar to that found on a national and at a global scale. Additionally, SCS was calculated to 1 m depth by Dixon *et*

al. (1995) and Pan *et al*. (2011) for their calculations. Their SCS estimates will therefore be higher than was found in this current study which measured to 30 cm depth only.

4.17 FullCAM modelled AGB and soil CS results compared with measured results

The second objective of this case study has been to evaluate correlations between the empirically derived data with modelled results from FullCAM. FullCAM is used to report Australia's land sector greenhouse gas account to meet the reporting criteria of the Kyoto Protocol. Consequently, it is important to review its correlation with measured data to ensure it produces data representative of actual C stocks. Short and long term trends have been modelled for a 13 year old site (Figure 4-60, Figure 4-61 and Figure 4-62) to show the potential gains and losses of AGB C, SCS and C in the soil pools. The simulations show that initially (approximately 5 years), soil loses SOC following the disturbance caused by establishment of BEPs. However over time (modelled for 100 years) the sites are modelled to gain C in both the AGB and soil.

The measured AGB and SCS (0-30 cm) results obtained for the current research have been calibrated against the simulated FullCAM results. The results of the current research have found the correlation between measured and predicted FullCAM data for AGB C yielded a strong relationship with an R^2 of 0.95 (P<0.001). This result is similar to that found by Paul and Polglase, (2004b) who found correlations for AGB between 0.94 and 0.99 in the two scenarios examined. More recent work has been undertaken by Paul *et al*. (2013), Paul *et al*. (2014) and Paul *et al*. (2015) to further refine allometrics used in FullCAM for environmental plantings. Their work has identified that research using additional datasets, longer term studies to improve the tree yield formula for older sites is needed.

The correlation between measured and FullCAM predicted results for soil C pools under BEPs was found to be less convincing. The best correlations were achieved for the HOC and ROC pools with correlations with R^2 = 0.77 and 1 respectively (Figure 4-64). Paul and Polglase (2004a) also found the strongest correlations were made by FullCAM for these two pools. On the other hand, the relationships between the measured results and FullCAM predicted results for TOC and POC were very weak. The second objective of this case study included an aim to compare empirically derived SOC results with FullCAM/RothC modelled results. However, due to the contradictory nature of the results related to the relationship between measured and modelled data the results of this research project are unable to confidently validate the use of FullCAM for soil C pools. Therefore further work is recommended to validate the use of FullFAM/RothC for soil C.

FullCAM is used to calculate Australia's national greenhouse emissions and also is used to underpin the CFI methodology (Methodology Determination, 2014) for reforestation by

environmental or mallee plantings to be used by proponents of offset projects. At present the methodology only allows for inclusion of biomass. From the measured AGB results from the present research project and the good correlations achieved with modelled FullCAM data, the objective of comparing measured AGB data with FullCAM modelled data confirms that FullCAM is a useful tool for modelling AGB. The good agreement is largely due to the extensive allometric development specifically for environmental plantings for use in the FullCAM model developed by Paul *et al*. (2013) and Paul *et al*. (2014).

4.17.1 FullCAM modelled and measured litter comparison

FullCAM uses CAMFor to model forest debris. CAMFor utilises tree growth, management, and climate parameters to estimate the mass of tree components including litter and root decomposition, and soil C turnover (Paul and Polglase, 2004b). In research undertaken to calibrate and validate the sub-models within FullCAM (CAMFor and GENDEC) appropriate for litter decomposition, Paul and Polglase (2004a) found that after calibration CAMFor was able to predict litter decomposition from observed and predicted data from eucalypt and pine debris with a model efficiency of 0.65 and a mean square error of 117 g^2 per 100g^2.

Having determined that the rate of litter accumulation measured in BEPs from this study are similar with the findings of other studies (Paragraph 4.15.2) a comparison of the measured rate of litter accumulation with FullCAM modelled results has been undertaken. FullCAM reports litter as the C mass of litter, consequently correlations have been undertaken using the C equivalents converted from the findings by England *et al*. (2016). The best results were obtained when the measured C mass from the O2 layer were compared with the FullCAM modelled data. However, the results of this comparison show a weak R^2 of 0.34 (Figure 4-43). The finding for the O1 layer was weaker with $R^2 = 0.29$.

Given the weak relationships found between the measured BEP litter data from the current research project and the FullCAM modelled results for litter, further work is recommended in developing calibrations for litter and decomposition rates for BEPs, at least in the biogeographic regions in which this current study was carried out.

4.18 Soil edaphic properties and their influence on SOC sequestration in BEPs

Edaphic properties such as EC, pH, mineralogy and CEC have been shown to influence the SOC sequestration potential of soil (O'Brien *et al*. 2015). Further, afforestation of grasslands has been reported to result in significant increases in Na and ESP and a lowering of pH (Jackson *et al*. 2005). Consequently, the EC, and pH may play a role in the SOC sequestration potential of the AL when converted to BEP. Thus, these properties and their relationship with SOC has been examined here to address the third objective of the current research project which has been to identify soil attributes that may influence the SOC sequestration potential of the BEPs.

4.18.1 Electrical conductivity

The electrical conductivity (EC) was very low in both the AL and in the BEPs and consequently is considered unlikely to have any adverse effect on SOC sequestration and AGB growth rates within the BEPs included. For example, Sun and Dickinson (1995) studied the effect of low, medium and high levels of soluble salt on the growth of *Eucalyptus camaldulensis* and *Casuarina cunninghamiana*. The low EC group in their study had a similar mean EC to that of the BEPs in the current research. Sun and Dickinson (1995) found that trees were taller and had a larger DBH in low and moderate salinity sites than in the high salinity sites.

BEPs older than five years had a slightly higher EC than both the AL and BEPs aged 1-5 years. The specific cause of the higher EC in BEPs is unknown but two possible causes are suggested. Firstly, the transfer of nutrients, salts and water resources through the soil profile will be stimulated by the action of tree roots. Thus, cations such as Na could be transferred from deep in the soil profile to the surface layers. Although analysis of soil cations was not undertaken for the current research, earlier unpublished research by Read (2008) found that exchangeable Na was higher in BEPs than in the AL. Archibald *et al*. (2006) found that the EC increased temporally following afforestation and Berthrong *et al*. (2009) found consistently higher levels of Na following afforestation when compared with control sites. These authors attributed the increase to be a result of enhanced capillary rise, decreased leaching, or tree uptake of ground water from deep in the soil profile thereby transferring the Na higher up the soil profile. In other research Jackson *et al*. (2005) noted that on a global scale, levels of exchangeable Na were significantly higher following afforestation of grasslands. They explained that Ca, Mg and K were taken up by the trees, whereas Na was excluded and thus became concentrated in the soil. The source of the Na is via enhanced weathering due to greater root penetration and organic acid production (Field and Little, 2008). Heuperman (1999) explained this process could be caused by the trees creating a reversal to the hydraulic gradient, thereby leading to salt accumulation in the soil over a period of 20-30 years particularly in areas with a shallow water table.

A possible alternate explanation is that the tree canopy of the BEPs may intercept aeolian dust particles containing soluble salts and transfer them to the soil through throughfall. Shiga *et al*. (2011) analysed aeolian dust samples from eight sites located in south eastern Australia. They reported that dust can contain high levels of salts, especially NaCl that originated from terrestrial sources. However, more research into the composition of dust intercepted by BEPs is required to confirm this hypothesis.

4.18.2 Soil pH

The top 5 cm of the soil profile in the AL position was found to be slightly acidic with an overall mean pH of 5.26 $_{(1:5\ water)}$. Both the AL and BEP soils show a trend where the pH is higher with depth, but this is a natural edaphic attribute not necessarily influenced by the trees

in the BEPs. The pH was seen to decrease with BEP age whereby the youngest BEPs had a mean pH of 5.2 in the 0-5 cm soil layer and the BEPs aged 11-15 and 16-19 years had had a mean pH of 4.93 and 4.90 respectively. This result suggests that as BEPs age the trees are probably having an acidifying effect on the soil.

The finding of the current research that soil pH was lower in BEPs than AL was not unexpected as afforestation is known to lead to soil acidification (Ritter *et al*. 2003). This finding is also consistent with that of Jackson *et al*. (2005) who found that afforestation resulted in a median decrease of 0.3 pH units. They attributed the acidification to a combination of redistribution of soil nutrients within the profile, acidic litter and decomposition of SOM. In other research, Rhoades and Binkley, (1996) found that soil pH declined 0.9 pH units in just eight years after *Eucayptus saligna* was planted. They also found the loss was greater in plantations of *Albizia falcataria* an N fixing plantation species. In a meta analysis into the effects of agroforestry on soil properties Berthrong *et al*. (2009) found that eucalypt spp plantations resulted in a decrease in soil pH. They suggested the trees can influence a number of factors that can lead to the change in soil pH, including organic acid production such as carbonic acid generation, but particularly cation uptake. These authors found that afforestation led to a decrease in Ca, Mg and K and an increase in Na and H^+. These findings are corroborated by earlier unpublished work by Read (2008) who also found that BEPs had lower Ca, Mg and K than AL. Additionally, Na and Al was found to be higher in BEPs than in the paired AL.

There was variation in soil pH values between the TL and IR positions in the 0-5 cm depth increment. The soil pH in the IR position was lower than in the TL in the 1-5 year and 16-19 year cohorts. The IR had a higher pH in the 6-10 year old cohort and there was very little difference for the 11-15 year old cohort. In a study of scattered native trees in NSW Wilson *et al*. (2007) found that the surface soil pH was lower at positions further away from trees and outside the tree canopy. They suggest that organic anion redistribution from litter being deposited on the surface is responsible for the higher surface pH under tree canopies. Their suggestion aligns with the finding in the current research that there is a greater quantity of litter on the TL than in the IR, and that the soil pH is higher in the TL position than the IR position.

Soil pH is important for SOC sequestration because it can influence the availability to plants of important chemical elements such as N, P, K and S and thus the production of SOM. At low pH levels these nutrients can become limiting to plant growth. Soil pH can also influence the solubility of toxic elements such as Al, Mn and Fe. Acidic soils with a pH <5.5 can inhibit plant growth (Murphy, 2015) and can also inhibit processes that promote litter breakdown, which has the flow on effect of restricting nutrient availability. Accumulation of SOM may lead to a lowering of the soil pH and this is consistent with the trend whereby the pH is lower in the older BEPs than the younger BEPs. Consequently it can be concluded that the BEPs are contributing to the acidification of the soil. The lower pH in the older BEPs is therefore an important factor

possibly resulting in the lower SOC concentration, SCS to 30 cm depth and AGB seen in the older BEPs by limiting nutrient availability to the trees.

4.19 Soil nutrients and nutrient ratios: stoichiometric constraints on the C sequestration potential of BEPs

Sequestration of atmospheric CO_2 as SOC is a function of NPP; that is, the balance between above and below ground biomass inputs and losses. The inputs are derived from tree and understorey biomass growth, litter and root turnover, and the losses through decomposition and microbial respiration (Hoogmoed *et al*. 2012). Climate, parent material and nutrient cycling are also factors that have a strong influence upon SOC sequestration. Sequestration of C also depends on the stoichiometric constraints of SOM.

To minimise the influence of site variability as far as practicable the BEP sites included in the current research project are located within the same climatic and geographic region. However, nutrient availability and cycling potential of the soil and litter layer can be site specific and remains an important factor in the C sequestration potential of biomass and soil. Therefore, analysis of the nutrient dynamics within BEPs and the paired AL position to determine differences in SOC, TN and TP concentrations and their ratios has been undertaken. The corollary of this analysis will help determine whether changes to the nutrient dynamics create constraints or opportunities for BEPs to sequester SOC. The analysis of TN and TP and their ratios with C have therefore been used to further address the third objective of the current research case study which is to identify edaphic properties influencing the SOC sequestration potential of BEPs.

The findings on nutrient dynamics are important in the context of biomass production, where nutrient availability is inextricably linked with NPP and thus the production of SOM and SOC. For example, Attiwill, (1979), and Baker and Attiwill (1985) proposed three stages of forest stand growth. These stages are applicable to the nutrient demand and tree growth within BEPs. Firstly, these authors identify that the growth of the living biomass occurs by increasing the canopy foliage and development of the metabolic transport system in the stem bark and sapwood thereby strengthening the structural components of the trees. This typically occurs in forests up to 20 years of age. During this time the residence time of organic matter in the tree biomass is < 5 years (Attiwill, 1979). During the early growth stage the nutritional requirement of forest stands are high because the developing crowns have a high nutritional demand (Miller, 1981). In BEPs it can be argued, based on the tree attributes, this stage lasts up to 15 years. In the second stage Attiwill, (1979), and Baker and Attiwill (1985) suggest there is less diversion of resources to the photosynthetic and transport systems and instead resources are diverted to increase NPP during the development of the heartwood structures. This stage occurs in trees aged between 20 and 45 years. Attiwill, (1979) suggests that during this stage the residence time

of tree organic matter increases from 5 years to 18 years. It can be extrapolated from these two stages of forest stand growth that initially, the young BEPs transfer OM to the soil which has a fast residence time, whereas the older BEPs are nearing stage two where the cycling of OM into the soil is slow. In BEPs this is manifest as an apparent decline in SCS in the 0-30 cm depth in the older sites. The third stage refers to mature forests where heartwood is shed as litter and becomes a source of soil organic matter that can stimulate nutrient cycling. It is likely that the older BEPs have not yet achieved the third stage.

In the following section, the changes to soil TN and TP following incorporation of BEPs into land previously used for agriculture and the ratios between C, N and P are discussed and compared with the nutrient demand and stages of forest growth proposed by Attiwill, (1979), and Baker and Attiwill (1985).

4.19.1 Soil total nitrogen changes under BEPs

TN concentration in the AL position soil was found to be highly variable between sites and age classes. There was a trend suggesting that over time TN increased in the AL position. This trend was consistent down through the four soil layers (Figure 4-48). The difference in TN between the age classes seen in the AL position is most likely attributable to site factors such as legume growth, pasture composition, fertiliser application and season at time of sampling. It was beyond the scope of the current research to record pasture composition and fertiliser history of the AL so these factors have not been incorporated in the analysis. Time of sampling may be a factor because the sites were sampled over a five month period. For example, Wuest, (2015) has shown that fluctuations in organic N occur in response to management (till and no-till), and seasonal variation. It is noted that the trend with SOC concentration in the AL position follows a similar trajectory to TN. A similar relationship was identified by Franzluebbers *et al*. (1994) who showed that SOC was highly correlated with TN. Given that N is an essential nutrient for plant growth, it is not surprising, therefore that there is a higher SOC concentration in the sites having a higher TN concentration.

Another factor that may influence the TN concentration in the AL position over time may include nutrient deposition from livestock that camp in the lee of the BEPs. Livestock dung has been shown to increase TN and SOM (During *et al*. 1973). This is important in regard to BEPs because livestock are known to seek out shelter from wind, rain and sun, and established BEPs can provide each of these protections. For example, Bird *et al*. (2007) found that shelter belts (such as BEPs) can reduce wind speed by up to 60 % on the lee side for a distance of up to six tree heights away. Further, (Bird *et al*. 2002) found that rainfall can be up to 50 % lower up to a distance of 0.3 tree heights to the lee of a windbreak. Additionally, livestock behaviour involving congregation on the lee side of vegetation has been associated with higher soil N and P concentrations (Duncan *et al*. 2008). To minimise the potential of the effect of livestock excrement influencing the nutrient concentrations, soil samples for the AL position were taken

30 m outside the BEP enclosure, or further where there was evidence of stock camps. This precaution was taken because soil nutrients are known to be redistributed by livestock, with large increases in K, Ca, Mg, P and N possible (Whalley *et al*. 1978). Notwithstanding the precaution, it is likely that a combination of each of the aforementioned factors has contributed to the trend in TN seen in the AL position. The variation and trend seen in TN in the AL position further enforces the suggestion previously made in relation to variation in SOC across the landscape (Section 4.14.1), that analysis incorporating more replicates is necessary.

Within BEPs, there is a trend where the TN concentration in the 0-5 cm soil layer appears to increase temporally in sites up to 15 years of age, but decreases in BEPs aged 16-19 years of age (Figure 4-49). A similar trend is associated with each soil layer. The TN concentration in the 0-5 cm layer of BEP soil is higher than in the adjoining AL position in sites up to 15 years (Figure 4-50). However, in sites older than 15 years the results show that the BEPs have a lower TN concentration than the paired AL. The finding that TN decreases in soil following afforestation is consistent with other similar research, for example Groenendijk *et al*. (2002), who observed lower TN concentration following afforestation with conifers in New Zealand. Therefore, the results indicate that several factors are influencing the soil TN concentration. For example the trees decrease TN over time (through take-up into the biomass), change the organic matter composition resulting in changes to nutrient cycling, or there is natural variability between sites.

To further explore the proposition that the soil TN concentration has changed due to the presence of the BEPs a review of the composition of the tree species within the BEPs has been undertaken. The proportion of live *eucalypt* spp. trees was 70 % in the 16-19 year old cohort of BEPs, whereas the proportion of live *eucalypt* spp trees ranged between 35 and 50% in the BEPs aged between 1 and 15 years. The low proportion of acacia species, and absence of leguminous shrubs in the oldest cohort indicates that it is likely that lower levels of N-fixing bacteria are present. Thus the contribution of N to the tree, soil and litter pools especially within the older plantings will be proportionally lower and may be a further explanation as to why there is less SOC in the oldest cohort of BEPs than the paired AL

Conversely, the younger BEPs have a higher TN concentration than the AL position and this is possibly associated with the presence of *acacia* spp. and *casuarina* spp. trees in those BEP sites. Additionally, the younger BEPs have been sown with *daviesia, indogofera and Hardenbergia* spp., all of which are native N-fixing shrubs. *Acacia* spp. are capable of producing atmospheric nitrogen-fixing nodules through their symbiotic relationship with rhizobial bacteria (Kasel *et al*. 2011) and *casuarina* spp. with relationships with actinomycetes, VAM and ectomycorrhizae (Resh *et al*. 2002). N-fixing native shrubs such as those included in the BEPs have also been found to develop symbiotic relationships with rhizobia, although other edaphic factors including pH, salinity and N availability will also contribute to the composition and diversity of the

bacterial communities (Thrall *et al*. 2011). The presence of the N-fixing tree and shrub species within the BEP seed mix may affect the availability of N by releasing it during their decomposition process or through the formation of a symbiotic relationship with mycorrhizae (Forrester *et al*. 2006). The available N provided by the *acacia* spp. trees may lead to higher growth rates for the non N-fixing species (notably the *eucalyptus* spp.) growing within the BEPs. For example, Forrester *et al*. (2004) found that stem volume and AGB was greater in mixed species plantations with *eucalyptus globulus* and *acacia mearnsii* mixtures when compared with the growth of monocultures of *E. globulus* trees.

Although analysis of the nutrient composition of the litter from BEPs was not carried out for the current research project, the composition and turnover rates of litter is an important factor determining nutrient availability for plant growth and thus SOC sequestration. Indeed, litter is an important source of nutrients. Therefore, the nutrient composition and turnover of nutrients from litter in BEPs and its role in SOC sequestration deserves further research. However, the findings of research conducted by Forrester *et al*. (2004) into the composition of nutrients in litter under pure and mixed acacia and eucalyptus stands can be used to explain the nutrient composition of litter within BEPs. Forrester *et al*. (2004) found litter from *A. mearnsii* had a higher concentration of N than litter produced by *E. globulus* monocultures. Forrester *et al*. (2004) also found that litter from *A. mearnsii* stands contained similar P concentrations to *E. globulus* litter. However, the *A. mearnsii* stands cycled higher quantities of P from litter due to production of larger litter pools when acacias were present. The microbial associations stimulated by the presence of the leguminous species facilitate nutrient uptake by eucalypts. The research of Forrester *et al*. (2014) explored the benefits or otherwise of having mixed species plantations in terms of forest productivity. They found that having *A. mearnsii* present in plantations increased the height and diameter growth of *E. globulus* and attributed the greater growth to the larger N availability which was facilitated by improved N cycling provided by *A. mearnsii.* This finding is important for BEP productivity, which although not planted for commercial production, are established for a number of ecological functions and more recently for sequestering SOC in the long term.

4.19.2 Soil total phosphorus

There was no temporal trend in soil TP concentration in the AL position. There was however, a decreasing trend with depth in each of the age cohorts (Figure 4-52). TP was found to be highest in the sites aged 11-15 years in all soil layers and lowest in the sites aged 6-10 years. The results also show a lack of a clear temporal trend for changes to the TP concentration in BEPs (Figure 4-53). The trends suggest that TP is more likely to be influenced by site factors rather than a response to the trees. Hopmans and Elms (2009) found that there was little change in TP between first and second rotation plantings of radiata pine. Indeed, TP has been closely related to the P content in the parent material (Walker and Adams, 1958).

In BEPs aged between 1-10 years there was a slightly higher average concentration of TP in BEPs compared with the adjoining AL position, whereas in sites older than 11 years the mean TP concentration was lower in the BEPs than the AL position (Figure 4-54). This finding corresponds with other research. For example in a study comparing a 19 year old *Pinus* spp. plantation with grassland in New Zealand Chen *et al*. (2000) found that TP was lower in forest soils.

Forrester *et al*. (2004) proposed that when tree growth in mixed species plantations is inhibited due to a lack of soil moisture or available P, the *A. mearnsii* trees may outcompete *eucalyptus* spp. and thereby inhibit productivity. Thus, as well as soil moisture deficiency it is probable that the available P pool is limiting the AGB productivity and SOC sequestration in some BEPs and consequently provides an explanation as to the variability between sites and BEP age.

4.19.3 Soil nutrient ratio changes following afforestation of AL with BEPs

The mean C:N ratio for the four age cohorts in the AL position ranged between 14 and 19, and in the BEPs the mean range is between 15 and 17. These ratios are considerably higher than the optimum of 12 normally required for sequestering SOC (Himes, 1998; Kirkby *et al*. 2011; Kirkby *et al*. 2013; Kirkby *et al*. 2014). The results show variability with no particular trend (Figure 4-51) to the C:N ratio in BEPs during the first 15 years. The C:N ratio in the BEP position in the 16-19 year old BEPs is much wider than in the AL position consistent with the lower mean TN concentration in BEPs (Figure 4-50). Increases in the C:N ratio have previously been noted following afforestation. For example, in a study into the effects of *pinus* spp. plantation stands aged between 17 and 19 years in New Zealand, Groenendijk *et al*. (2002) showed that the C:N ratio in three of the four plantations studied had slightly higher C:N ratios than the adjacent pastures. In another study into the effects of mixed species environmental plantings in Victoria Cunningham *et al*. (2015) found that the C:N ratio widened in plantings over time. These authors detected a temporal increasing trend in C:N in plantings aged 10, 18 and 30 years. They have suggested that the increase in C:N, together with higher recalcitrant C in the plantings could be interpreted as an indication that there is a lower rate of decomposition and thus the plantings are retaining SOC in the stable pool. Their suggestion is supported by other research (eg Degryze *et al*. 2004) who suggest there are three mechanisms important to achieve SOC sequestration, one of which is the need for a low litter decomposition rate such as occurs when there is a high C:N ratio (Finzi *et al*. 1998).

C accumulation in soil is dependent on nutrient ratios (Kirkby *et al*. 2016). Indeed, Frossard *et al*. (2016) has suggested that the ratios of C:N:P in SOM are useful for identifying nutrient deficiencies that inhibit SOC sequestration. Further, Kirkby *et al*. (2011) proposed that it may be possible to determine whether the availability of C, N, P and S could limit sequestration of C into the humic pool. Consequently, the soil C:N ratio present in the AL and BEP has been used

to predict whether TN concentration of BEPs is limiting their potential to sequester SOC (Table 4-16 and Table 4-17). Based on the TN results and soil C:N ratio calculations the BEPs in the both the 11-15 and 16-19 year old cohorts still have capacity to sequester further SOC. That they haven't, indicates that some other factor is limiting the sequestration potential of the BEPs and Frossard *et al*. (2016) has found that whilst C:N:P ratios can be affected by nutrient inputs, the extent that the ratios affect sequestration of C (as stabilisation of compounds of C, N and P) depends on edaphic properties such as soil texture, mineralogy, stability and nutrient status. Therefore, whilst this section considers constraints to C sequestration due to possible stoichiometric deficiencies, other edaphic factors may have a larger constraining influence.

When the C:N ratio increases, SOM becomes more resistant to decomposition and thereby C accumulates at a faster rate than it can be decomposed (Eyles *et al*. 2015). Litter produced by trees is typically high in lignin which decomposes slowly. Finzi *et al*, (1998) has shown that litter with a high lignin:nitrogen ratio will decompose at a slower rate than litter with a low lignin:nitrogen ratio. Consequently it is likely that the composition of the BEP litter is high in lignin which together with the high C:N ratio is limiting N mineralization and thus the potential of the older BEPs to sequester higher amounts of SOC. Further study is required to identify the composition of the litter and the SOM fractions within BEPs to determine whether nutrient availability, nutrient ratios, soil type or other factors are inhibiting SOC sequestration in BEPs.

Fertilisation normally leads to an increase in SOC due to influences on increasing NPP (Zeng and Wang, 2015). However, Johnson (1992) found that the presence of N-fixers was more important than fertiliser in increasing SOC, although N-fixers generally need P and other nutrients for growth. Research reported in the literature demonstrates the response of litter to decomposition may or may not be influenced by the addition of fertiliser (Hobbie, 2000; Resh *et al*. 2002). Further, litter decomposition may be increased, inhibited or not be affected by fertiliser application (Hobbie, 2000). Litter that contains high lignin content may inhibit decomposition. For example the addition of N can result in an increase in leaf litter decomposition when the litter contains low levels of lignin, but there may be no difference in decomposition rate when the litter is high in lignin (Hobbie, 2000; Resh *et al*. 2002). Such relationships have been previously identified, where soils growing N fixing tree species are able to sequester between 20 % and 100 % more SOC than soils growing non N-fixing tree species (Johnson, 1992; Resh *et al*. 2001). However, the SOC concentration and thus the SCS depends on the balance between the rate of decomposition and NPP with climate, nutrient availability and biotic factors influencing both. Further, the ratio of eucalypts to N-fixing species is higher in the older plantings. This ratio would affect the composition of the surface litter layer because eucalypts produce litter that has a high lignin and low nutrient concentration. Further, eucalyptus litter has a high C: nutrient ratio due to the woody composition of the litter (Keith, 1997).

The results show that in the 0-5 cm soil layer the C:P ratio in the AL ranges between 100 and 148. The ideal C:P ratio necessary to sequester C as humic C is 50:1; i.e. 20 kg P is necessary to produce 1000 kg of humic C (Himes, 1998; Kirkby *et al*. 2011). The BEPs have a C:P ratio of between 102 and 190 in the 0-5 cm soil layer which is higher than needed for optimum plant growth and then SOC sequestration.

To address the third objective of this case study, it is clear from the findings of the current research, that changes to edaphic properties, principally soil pH which become lower as BEPs get older, together with a combination of low TN and high C:N ratio, are constraints that are inhibiting the SOC sequestration potential of the older BEP sites in the current research project and may indeed prevent BEPs from sequestering atmospheric CO_2 in the soil in the longer term. Consequently, it is more likely that BEPs provide greater value for their ecosystem services such as shelter for livestock and pastures/crops, increased landscape biodiversity, wildlife corridors, and for sequestering C in the AGB, rather than their potential to sequester C in the soil.

4.20 SOC fraction changes following reforestation with BEPs

In addition to identifying changes to SCS following LUC from AL to BEP, the experimental design of this research project has allowed for analysis of SOC fractions through the use of MIR analysis. The results from MIRS analysis provide predicted data to address the fourth research objective which was to use MIRS as a method for identifying changes to the various SOC fractions comprising POC, HOC and ROC following conversion of AL to BEP. A determination of the change to SOC fractions is important because different fractions have different sensitivities to decomposition. For example, POC has a short residence time because it is comprised of readily decomposable plant material, whereas HOC being humified organic matter is more resistant to decomposition and thus can remain stable for up to hundreds of years and ROC is inert to decomposition. Consequently, to address the fourth objective, this section initially discusses the results of the validation and correlation analysis of MIRS derived data with the results of data obtained from wet sieving technique and ^{13}C NMR analysis. MIRS analysis was undertaken on all samples, and wet sieving and ^{13}C NMR analysis was undertaken on samples from Subsets 2 and Subset 3 respectively (Figure 4-7). The SOC fraction results derived by MIRS analysis are then discussed to determine whether LUC from AL to BEP results in modification and/or stabilisation of SOC fractions.

4.20.1 Validation and correlation of MIRS derived results with Leco CNS2000, wet sieved and ^{13}C NMR analysis results

To validate the MIRS results, several correlations were undertaken. To establish a baseline, an initial correlation analysis was undertaken to assess the relationship between the SOC concentration results obtained using both Heanes and Leco CNS2000 methods on n = 96 samples from subset 1 (Figure 4-7). A strong R^2 = 0.92 (P<0.001) (Figure 4-56) was achieved, giving confidence that the SOC results obtained by Heanes analysis were comparable with the results obtained from LECO analysis. The strong relationship between the Heanes and Leco CNS2000 methods has previously been observed by Conyers *et al*. (2011) who found the coefficient of determination for the linear correlation between the two methods was R^2=0.98.

Having determined confidence in the Heanes result in relation to Leco CNS2000 analysis, and because all samples from the current research project were analysed for SOC using both the Heanes method and MIRS, the next step was to identify the relationship between SOC $_{(Heanes)}$ concentration results with MIRS predicted SOC concentration. This analysis achieved a strong R^2 = 0.93 (P = <0.001) (Table 4-21) linear regression relationship giving confidence that the MIRS model used was able to provide a very good prediction for SOC from the BEP study sites. Similar results were obtained by Baldock *et al*. (2013b) who developed a set of MIRS/PLSR models to predict SOC fractions for a range of different soil types across Australia. Those authors obtained a calibration R^2 of 0.86 and validation R^2 of 0.85 when comparing MIRS predicted OC with measured (by Leco CNS2000 analysis) TOC on n = 13 100 calibration, and n = 6 626 validation samples. In another study into changes of soil fractions for 36 Australian environmental plantings recently undertaken by Madhaven *et al*. (in prep), the leave-one-out cross-validation predictions showed the relationship between measured TOC (again by Leco CNS2000 analysis) and predicted MIRS results was strong with an R^2 of 0.97.

Soil fraction results obtained on n=12 samples (Subset 2, Figure 4-7) by using the manual method of wet sieving was used to compare soil fraction differences between TL and AL. The results show that the TL had a higher proportion of TOC, POC and HOC+ROC than the paired AL samples (Table 4-23).

Having calculated soil fraction data from the wet sieve results, a comparison, using the same samples, was undertaken with the results obtained using MIRS derived predictions. The correlation analysis showed a very strong relationships between data predicted by MIRS analysis with wet sieved determined data. For example TOC: R^2 – 0.95, POC: R^2 = 0.81 and HOC: R^2 = 0.67 (Table 4-25).

Finally, ^{13}C NMR analysis for soil fractions was undertaken on 12 samples (Subset 3, Figure 4-7) to compare the difference in soil fractions between the 0-5 and 5-10 cm soil layers. The results were determined from composite samples made up of subsamples from the TL and IR position (to represent the BEP). The results show that the concentration of POC, HOC and

ROC was higher in the 0-5 cm layer than in the 5-10 cm layer (Table 4-24). This finding is not surprising given that the concentration of SOC and SCS is typically higher in the surface layer than sub-surface soil layers due to there being more SOM closer to the surface.

When the ^{13}C NMR results are compared with the results for the same samples determined by using MIRS predictions strong correlation relationships have been identified for POC and ROC (R^2 = 0.84 and 0.95 respectively). The correlation between ^{13}C NMR and MIRS derived results for HOC was moderate with an R^2 of 0.47 (Table 4-25).

The strength of the correlations between the sampling methods used in this current research confirms similar findings by other authors (e.g., Baldock *et al*. 2013a,b and Madhaven *et al*. (in prep), that MIRS can confidently be used to predict SOC in both agricultural and BEP scenarios and can also be used to detect changes to SOC following LUC.

4.20.2 SOC fraction change after reforestation with BEPs

Using SOC fraction data derived from MIR analysis, the research question as to whether the composition of the C fractions in the AL is changed after establishment of BEPs has been addressed. The results show that POC, HOC and ROC was higher in BEPs than the AL in BEPs aged 1-5, 6-10 and 11-15 years suggesting that integration of BEPs into land previously used for agriculture can increase C in all fractions, at least initially. However, the mean POC, HOC and ROC was lower in BEPs than AL in the oldest BEPs studied (16-19 year old BEPs) in the chronosequence. When examined on a site basis the results show there is considerable variation in SOC fractions between sites, even in the AL position. This is not surprising as there are many accounts in the literature that report variability in SOC and C fractions in both agricultural and afforested landscapes (Cambardella *et al*. 1994; Conant *et al*. 2003; Turner *et al*. 2005).

The HOC fraction shows the largest change following LUC from AL to BEP, with the mean difference incorporating all sites and ages, being a gain of 2.33 Mg C ha^{-1}. POC had a marginally smaller gain of 2.29 Mg C ha^{-1}. The ROC fraction showed the least change across the 19 year chronosequence with a mean gain of 1.16 Mg C ha^{-1}.

In research conducted by Turner *et al*. (2005) the proportion of stable C (the equivalent of HOC) was found to be lower than labile C (the equivalent of POC) 16 years after establishment of *P. radiata* plantations. Turner *et al*. (2005) also found that the change in stable C after testing at 16 and 30 years intervals after afforestation had resulted in very little change (with no fertiliser treatment). Conversely they found that the labile C pool increased from approximately 13 g kg^{-1} to 16 g kg^{-1} at 16 and 30 years respectively when no fertiliser was applied. The increase was greater when 100 kg ha^{-1} of single superphosphate ($Ca(H_2PO_4)_2$) fertiliser was applied.

The results of the current research suggest that LUC following afforestation of AL with BEPs may lead to an accumulation of C in each of the SOC fractions. The change occurs following

the exclusion of grazing and the establishment of trees on land previously used for agriculture. The greatest increase occurs in the HOC and POC pools suggesting that the increase in plant residues following integration of the trees into land previously used for agriculture, together with changes to decomposition processes converts C into two forms:

- Labile form (POC) comprising fresh plant residues readily available for decomposition, although under BEPs some of the litter comprises lignin which is more resistant to decomposition than other forms; and
- Stable humified form (HOC) which in terms of C sequestration is important for mitigating the effects of climate change, because C in this pool has a higher residence time in soil than POC because it is resistant to further decomposition and thus conversion to CO_2.

The finding that there is a higher stock of stable HOC than labile POC following afforestation with BEPs is consistent with findings in other agroforestry studies. For example, Baah-Acheamfour *et al.* (2015), conducted a study in Alberta, Canada and found that agroforestry had a different soil fraction distribution than herb-land that had no tree cover. Their research found that SCS in each fraction (light, occluded and heavy fractions) was higher in forested soil than adjacent herb-land. They also found that different types of agroforestry system influenced the SOC fractions whereby labile C was higher in silvopastural systems than in shelterbelts, whereas shelterbelts had more stable C than silvopastural systems. However, Poeplau and Don (2013) found that afforestation of grassland in Europe resulted in an increase of POM despite an overall loss of SOC. They suggested this change indicated the SOC becomes more labile following afforestation of grassland soils.

The relative proportion of HOC over POC is important in terms of sequestration of atmospheric CO_2. HOC represents the stable, or passive pool of C and therefore the results indicate that BEPs are contributing C to this important pool.

The quantity of C in the POC, HOC and ROC fractions is lower in the older sites that the baseline AL (Figure 4-58 and Table 4-22). This difference indicates a loss of C from the older BEPs. The loss of C from the functional pools in the older BEPs is of concern because the underlying premise of this research has been that BEPs will sequester SOC and that the sequestration could mitigate the effects of anthropogenic CO_2 emissions. A probable reason the BEPs in this study do not gain more resistant C over time is possibly associated with site factors, for example the parent material and proportion of clay minerals in the soils of the older sites not being in a form suitable for stabilisation of the C. This is because clay mineralogy is necessary to support physicochemical protection of SOM and thus stabilisation of SOC (Oades, 1984; Skjemstad *et al*. 1998; Degryze *et al*. 2004).

Paul, *et al*. (2004a) found that initially total SOC decreased after afforestation of agricultural soil, but eventually there was change and the SOC was found to increase gradually over time. The HUM pool however, was found by these authors to decrease in one situation and increase in another.

Skjemstad *et al.* (2004) found that TOC and the POC fraction declined at a faster rate than the HUM fraction in cropped Brigalow catchment samples and Tarlee crop rotations. They also found RothC modelled RPM initially declined faster than the measured POC fraction, consequently they adjusted the rate constant for the RPM pool to achieve better calibrations.

4.21 Summary

The findings of this case study do not provide convincing evidence that incorporation of BEPs into pastures previously used for grazing leads to SOC sequestration. The soil C pool is, however, very important contributing approximately 58 % of the C pool in BEPs. BEPs do lead to an increase C stock as AGB. The reforestation of agricultural land with BEPs was expected to lead to increases in SCS in the 0-30 cm layer commensurate with the growth of AGB, however, the variability in results for both the change in AGB and SOC has led to the conclusion that there is no relationship between the two.

The optimum study design to determine changes to SOC over time would have been to undertake a longitudinal study. However, due to the time scales of the study, and the aim to determine change over time a chronosequence approach was used. A critical shortcoming of using a chronosequence is the identification of an appropriate baseline parameter. The current research project utilised the AL to obtain baseline data with which to compare the results from samples obtained from the adjacent BEP – a land use change. The critical results for SOC concentration and SCS and other parameters found that in some instances that the AL had 'better' results than the BEP. The variation and trends in results for the AL position highlight the probability that factors other than LUC alone (i.e. establishment of BEPs) have influenced the results. The factors would include underlying geology, soil type, mineralogy, topography, previous land use, current land management practices, aspect, climate, the interplay of the millennium drought on tree establishment and growth. Additionally, changes following establishment of the BEPS are most likely influenced by a combination of factors including land management of the AL, natural spatial heterogeneity of SOC and nutrients, soil type, nutrient cycling, tree stocking rate and species composition.

On average, the results for BEP age class demonstrate there is initially an increase in SOC and AGB in BEPs up to 15 years since establishment. However, there is an apparent SCS loss and less AGB in BEPs aged 16-19 years. The results also show that the AGB is lower in the 16-19 year old BEPs. Therefore, the results of the current research project cannot conclusively assert that BEPs up to 19 years of age sequester SOC. Further research is necessary to monitor whether this trend continues or if indeed as the literature suggests, that eventually the BEPs may

sequester SOC to levels above the paired AL baseline level. Overall, the BEPs were found to have sequestered on average 0.53 Mg C ha^{-1} yr^{-1}.

The results of this research show that edaphic and nutrient changes following afforestation of AL with BEPs may become constraints that inhibit the potential of the BEPs to sequester atmospheric CO_2 in the soil.

The results of this research project have shown there is a great deal of variability in both SOC and AGB and there is little trend with age. Consequently, to answer the second research question as to whether increases to SOC are commensurate with AGB growth; that AGB growth is not an indicator of changes to SOC. However, the predictions for AGB growth for the BEP sites derived by FullCAM were found to result in strong correlations with the measured data. The relationship between FullCAM results and measured SOC data was less predictive with only weak correlations for TOC and POC achieved, but good correlations for the HOC and ROC functional pools.

Chapter 5: Carbon sequestration potential of scalded soil rehabilitated by waterponding

5.1 Introduction

Waterponding is a term used to describe a rehabilitation method for highly eroded, scalded soils on clay pans. The method involves the construction of horseshoe shaped or fully enclosed banks which are networked across areas of scalded soil on clay pans in such a way as to trap rainfall that would otherwise run off the scalded surface. The waterponds generally successfully lead to revegetation of the scalded clay pan landscape and after just few years (rainfall dependent) grazing can be reintroduced on a carefully managed basis. In the context of LUC, waterponding is a rehabilitation method that changes the landscape from one with neither productive agricultural value nor ecosystem service capability, to one where grazing can be undertaken and ecosystem services regain functionality.

In one sense the term waterponding could be a misnomer because anecdotally it may conjure up a vision of a wetland. Whilst the waterponds are constructed for the purpose of capturing rainfall runoff water, that water is quickly removed from the surface via infiltration and evaporation. The water that infiltrates into the soil profile initiates a number of physical and chemical changes in the soil which ultimately allow vegetation to reestablish. An important early physical change is establishment of surface cracks in the soil. Once the cracks appear and vegetation becomes established water no longer pools within the waterponds, but instead readily infiltrates into the soil profile.

Earlier research, by Beadle (1948a,b,c) Cunningham, (1987) Ringrose-Voase *et al.* (1989a,b), and Ditchfield (1996) explored clay pan scald formation. These authors also identified a number of vegetative, physical and chemical changes that occur to the scalded soil after waterponding. The early research did not, however, include analysis of the SOC sequestration potential of scalded soil following waterponding. Consequently, the current case study examines the SCS in the 0-30 cm layer of scalded soil and explores the SOC sequestration potential of waterponding by following a 27 year chronosequence. The literature relating to scald formation, the waterponding method and earlier work by other researchers that identified physical and chemical changes to the soil following waterponding is examined first. Some of the physical and chemical changes observed in the earlier research have been re-examined in this current research project to identify the relevance of such changes to the SOC sequestration potential of the clay pan scalds, and also in the context of the chronosequence.

Additional research undertaken here has also extended our knowledge on the mineralogical changes that occur in the scalded soil following waterponding, and suggest the possible origin of vegetated hummocks which are scattered across the scalded landscape.

5.2 Scald formation

Grazing by domestic herbivores has been carried out in the Marra Creek region since the arrival of Europeans (Metcalfe *et al.* 2003). Major regional locations include Walgett which was settled from 1839 and the Bogan River from 1846-47 (Beadle, 1948a) to support grazing enterprises in the area. Indeed, the diary of Sir Thomas Mitchell (1848) refers to an abundance of grasses suitable for livestock production across the plains. Mitchell explored the region between the Macquarie Marshes and the Bogan River in 1846. Notably, in January 1846 he established camp at Cannonba, a short distance to the south west of Marra Creek (Mitchell, 1848). His description of the camp site was that there was

> "…better grass than we had seen near water during the whole journey…" (Mitchell, 1848).

After departing from the camp site for the Macquarie River on the 11th February 1846 Mitchell passed Marra and Duck Creeks and described

> "… plains covered with luxuriant crops of very rich grass…" (Mitchell, 1848).

Expansive grazing of sheep and cattle continued in the region in the years following Mitchell's exploration. By 1880 Green (1989) explains that the whole area was covered by pastoral runs with unsustainably high domestic livestock numbers of up to 15.5 million dry sheep equivalents present during that period. Additionally, land clearing and overgrazing caused by high livestock stocking rates, rabbits, and native herbivores each contributed to widespread loss of vegetation across the landscape. The overgrazing that occurred in the district at this time resulted in changes to the perenniality of the vegetation. For example, Rhodes (1987) found perennial species, such as for example *Atriplex vesicaria*, were readily replaced by annual species including copperburrs (*Sclerolaena* spp). Thus, the resilience of the vegetation in providing sustained protection to the landscape from erosion during periods of drought was diminished. Further, the severe drought of 1898-1902 exacerbated the situation (Beadle, 1948a,b; Green, 1989) by further diminishing the resilience and abundance of vegetation. Once the vegetation is lost, the unprotected, dry, uncohesive soil surface becomes exposed, leading to the erosion of the sandy A_1 horizon (Beadle, 1948a,b,c; Cunningham, 1987) and exposing a shallow silty-loam A_2 horizon. Prior to the erosion, the soils were classified as red duplex soils (Dr1.22) (Ringrose-Voase *et al*. 1989a; Ringrose Voase and McClure, 1994) or Chromosols (Isbell, 2002) having a strong texture contrast between horizons. However, due to the erosion and loss of the majority of the A-horizon, the soils have lost their texture contrast and can be re-classified as eroded Sodosols and Chromosols (Isbell, 2002). Cunningham (1987) explains that the scalds are typically associated with alluvial soils on prior streams and river flood plains.

After the erosion, the residual A_2 soil particles become reorganised and compacted into a thin surface crust (Ditchfield, 1996). The compaction occurs because raindrop impact causes destruction of the clay aggregates. The clay particles remain in suspension and are dispersed

across the surface until the water is evaporated allowing the crust to form. The crust overlies a saline and sodic, clay B-horizon that has a sub-angular blocky nutty structure that has been described as having a nutty appearance (Ringrose-Voase *et al*. 1989; Ringrose-Voase and McClure, 1994). The B-horizon is less susceptible to erosion than the original A-horizon (Beadle, 1948a,b). The residual crusted surface is so extensive that it results in a scalded landscape (Figure 5-1). Several authors explain that scalding occurs after the A_1 horizon has been eroded from duplex soil profiles (Beadle, 1948a,b, Marshall, 1974; Cunningham, 1987, Green, 1989). Beadle (1948b) further suggests that wind erosion is the principal cause of scalding, with sandy soils having little or no vegetation cover being the most susceptible. Warren (1965) on the other hand argues that the primary cause of scalding is water erosion with wind only having a secondary role. Both these mechanisms have occurred in the study area.

Figure 5-1. Scalded soil surface

Beadle (1948a) describes scalding as being widespread and the most serious type of erosion in the west of NSW, with natural regeneration and revegetation being slow, even after livestock have been excluded from a site for years. The scald surface crust effectively protects the soil from ongoing wind and water erosion, however, it also prevents water infiltration and plant growth. For example, in an experiment into the infiltration rate of scalded soil, Beadle, (1948c) found that scalded and crusted soil has a much lower infiltration rate than non-eroded soil due to compaction, crust breakup and low porosity of the exposed sub-soil surface. Consequently, much of the rainwater that falls on the scalded surface is lost through runoff or evaporation making this landscape bare and unproductive. Beadle (1948c,d) also observed that annual plant species are the primary colonisers on soft scalded surfaces but even after good rain they provide less than 1% ground cover. However, Beadle (1948c,d) also observed that some perennial species including *Chloris truncate, Atriplex semibaccatum, A. leptocarpum* amongst others

colonised cracks and depressions on the scalded surface, but that overall, the percentage of plant cover on the scalds was small.

Cunningham (1987), reported that in 1967, approximately 49 515km^2 of land in the Western Plains of NSW was scalded. The extensive areas of scalded soil present across the semi-arid rangelands of the Central West Catchment of New South Wales occur on flat to slightly sloping land on alluvial soils of prior streams, or meander plains and in the western river flood plains. These scalded areas are common across the Lower Macquarie catchment and include the meander plains of Marra Creek where this study was undertaken.

Beadle (1948a,b) describes three types of scald: soft, hard and limestone. Beadle (1948a, b) explains that soft scalds are characterised by a spongey surface with fine cracks and these are formed through clay dispersion on the surface. He further explains that the hard scalds can be either compacted clays or cemented sands. Beadle notes that the compacted clays are typically located in depressions where dispersed clay accumulates, whereas in cemented sands, the clay cements the sand grains together. The majority of sites included in the current research meet the definition of soft scalds (Beadle, 1948a; Jones, 1966), although the position of soft and hard scalds in the landscape may form a mosaic pattern at a small scale (Beadle, 1948b; Warren, 1965). The limestone scalds, which are not present in this landscape, are formed when $CaCO_3$ nodules become incorporated into the crust.

The soils in the Marra Creek area comprise two main soil landscapes, Marra Creek and Bugwah. The Marra Creek landscape group is formed on the most recent alluvial materials deposited within the past 5000 years and are associated with flows of the modern Macquarie River. Soils associated with the Marra group include grey and brown dermosols. Soft scalds are found on these soils (Forbes *et al*. n.d).

The Bugwah soil landscape on the other hand is formed on alluvial materials laid down between 5000 and 15 000 years ago. This landscape consists of Red Chromosols and Red Sodosols on the meander plains and splay outwashes, with Grey and Yellow Vertosols on the backplains. It is the Red soils on the meander plains, especially the Sodosols which are most vulnerable to scalding. The Lower Bugwah meander plain (BLms) is associated with soft scalds and the Bugwah Downstream scalded meander plains (BDms) are associated with hard scalds (Forbes *et al*. n.d). Soils from both groups can be sodic and saline. A satellite image (Figure 5-2) can clearly show the spatial relationship between the soil groups in the region, as well as the extensive waterponding work completed on the scalded soil in the region.

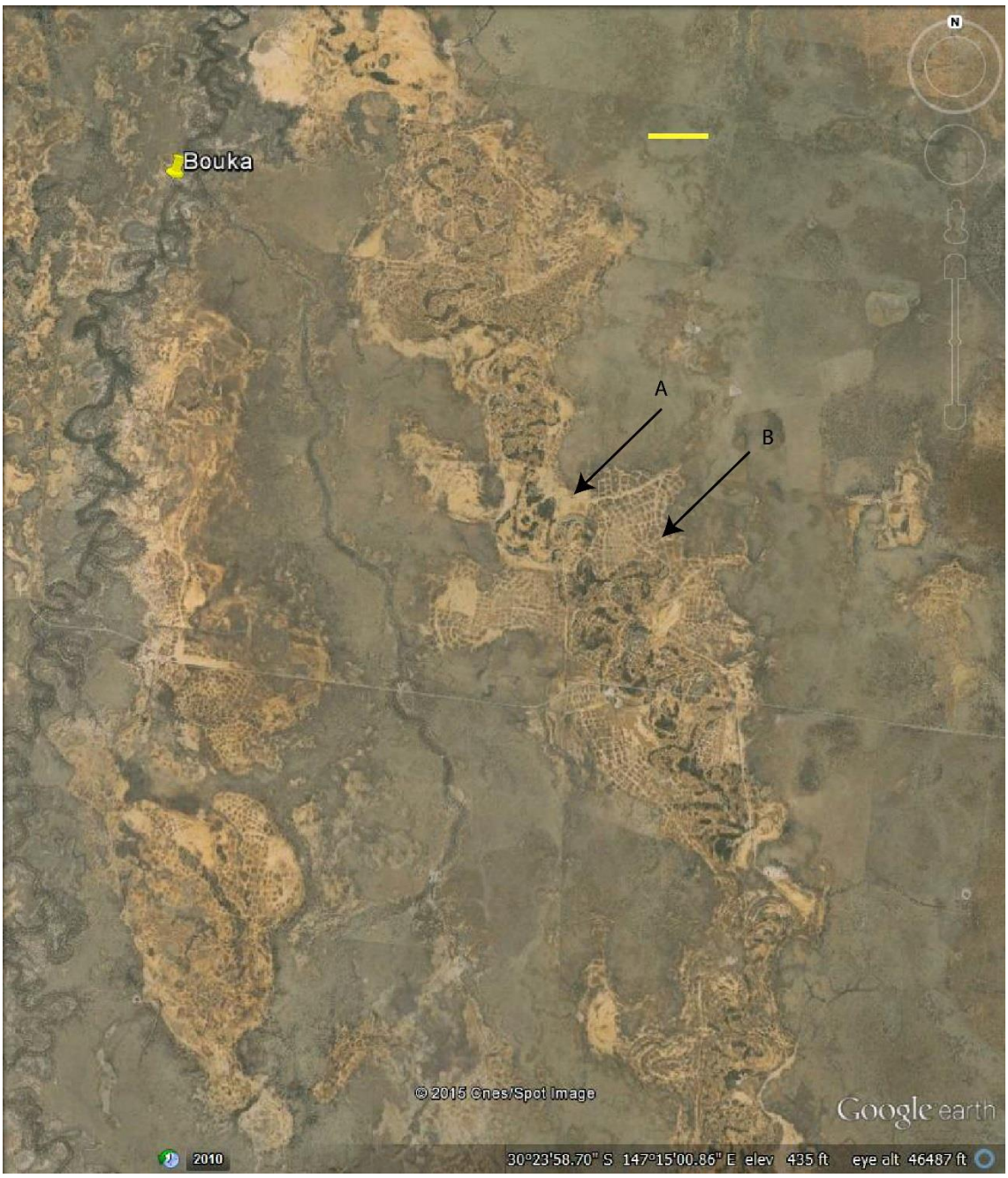


Figure 5-2. Satellite image showing spatial relationship between scalded meander plains rehabilitated by waterponding with adjacent unscalded soil groups. Yellow pin is Site 11, yellow bar = 1km; A = scalded clay pan, B = waterponded clay pan.

5.3 Vegetated hummocks

There are two key indicators across the Bugwah soil landscapes that suggest approximately 30-50 cm of the A horizon has been eroded from the landscape since European settlement.

Firstly, the most obvious visible evidence that a substantial layer of soil has been lost from the landscape is the presence of occasional tree stumps showing roots protruding laterally above the soil surface (Figure 5-3, Figure 5-4). The lateral roots are found to be approximately 30-50 cm above the surface and represent a reliable indication of the former depth of the A horizon.

Figure 5-3. Exposed lateral roots of an old tree stump showing extent of soil loss.

Figure 5-4. Vegetated hummock soil showing vegetation, and colour contrast with scald surface

Secondly, the clay pan scalds have a small number of isolated vegetated hummocks that are approximately 20-30 cm higher than the scalded surface (Figure 5-4). These hummocks were believed to be 'relics' of the former landscape (Ringrose-Voase *et al.* 1989a). The hummocks display distinct colour and texture contrasts and have grassy vegetation cover. Beadle (1948d) identified *Salsola Kali* (roly poly) and *Cryptostemma calendulaccum* (cape weed) amongst other annual species on the hummocks but suggested the plant communities on the hummocks were unstable and detrimentally impacted by grazing. The soil profile of a hummock was examined by Ringrose-Voase *et al.* (1989a,b) and found to have a loamy sand A_1 horizon up to 16cm deep, overlaying a silt loam A_2 which extended to a depth of 28cm. The hummocks trap and retain seed, and are able to support vegetation because the loamy sand texture allows water infiltration (Beadle, 1948c). Warren, (1964) argues that the hummocks form after water erodes the boundaries of residual vegetation, initially producing semi-circular or fan shaped peninsular formations and eventually leaving the isolated hummocks. This being the case, it is possible that the hummocks are representative of the original soil profile prior to erosion. Conversely, Beadle (1984c) explains that the hummocks may be either sand accumulations on the scalded surface or islands of soil not previously eroded but have been covered with accumulated sandy deposits. The conflicting explanations regarding the origin of the hummocks has been explored further in the current case study by utilising ^{137}Cesium analysis of scalded, waterponded and relic soil samples in an attempt to identify the origin of the soil particles (Section 5.7.14).

5.4 Evolution of the waterponding method to restore scalded soil

Green (1989) describes methods of restoration to landscapes affected by both sheet erosion and scalding. To differentiate between the two erosion classes, Green (1989) firstly defines sheet erosion as the situation where the surface layers of the uniform or gradational soil profiles is stripped from the landscape, whereas scalding refers to the loss of topsoil from duplex soil profiles. Surface ripping was found to produce short term improvement on sheet eroded soils but continued grazing and slaking was found to result in reversal of any improvement (Green, 1989). Contour furrowing was also found to be successful on sheet eroded sloping ridge country because the furrowed structures prevent water runoff and allow accumulation of moisture, litter and seed, thereby allowing vegetation to re-establish (Green, 1989). Unfortunately, neither of these techniques can effectively restore scalded clay pan soil.

Over several decades a range of management options have been attempted to restore vegetation on landscapes affected by scalding. Beadle (1948*a*:82) noted that natural regeneration of hard scalds was improbable. He suggested the most likely way to rehabilitate the hard scalds required an accumulation of sand which could only be achieved by trapping wind borne grains along furrows or log obstructions placed on the surface. He further indicated that all attempts to rehabilitate scalded soils had been "disappointing" and would be an extremely slow process (Beadle, 1948*a,d*). Further, these methods would only restore small patches across the landscape. Consequently Beadle (1948*a,d*) suggested that a method was needed that could rehabilitate large swathes of the scalded clay pans.

More recently a variety of reclamation techniques have been attempted to restore the Marra Creek scalded soils including spiral mouldboard and checkerboard furrowing, tyne pitting and ploughing. Each of these methods has had minimal success (Thompson, 2008) because slaking and dispersion lead to the reformation of the surface crust. Destocking has little effect because the crust inhibits water infiltration and lacks suitable niches for seed to lodge, germinate and vegetation to become established.

Through trial and error the only effective method found to reclaim scalded soils is the method of waterponding developed by the NSW Soil Conservation Service during the 1960's (Thompson, 2008). This method achieves substantial increases in vegetative growth and species diversity on the scalded areas on a wide scale. Reclamation is achieved on gently sloping land (<0.5%) by establishing a network of horseshoe shaped banks across the landscape. Waterponds are constructed by using a grader which establishes banks approximately 30 cm high. The waterponds have a surface area of approximately 0.4ha each and allow no more than 10cm water depth to avoid wave action that can break down the banks. Waterponds are completely enclosed on the flatter areas (Figure 5-5) and open up slope on the sloping clay pans. The

waterponds become revegetated from seed that is washed and blown in from endemic species including sea barley grass, annual saltbushes and copperburrs (Rhodes, 1987).

Over time the waterpond banks wear down, but do not need to be replaced because the vegetation and soil interactions make the water holding capacity of the landscape self-sustaining. Instead of pooling on the surface where it will evaporate, water is retained in the rehabilitated landscape because it infiltrates the soil profile. Moisture differences between scalds and reclaimed scalds have been studied by Jones (1966) and Jones (1967). Jones found that the soil moisture percentage of scalds was 9.8 whereas naturally reclaimed scalds had a soil moisture percentage of 25.1 (Jones, 1966). In a second study comparing soil moisture percentage between scalds and waterponded sites Jones (1967) found that depending on the distance from the bank the moisture content of scalds was on average 9.3 % whereas a 12 month old waterpond had between 14.1% or 17.3 % moisture in the 4" soil layer. The findings by Jones (1966, 1967) exemplifies the improved water infiltration capacity and soil moisture content of rehabilitated scalded soil with recharge occurring mostly in the upper 0.5 m of the soil profile.

The revegetation process can be hastened with the broadcasting of other saltbush species (*Atriplex* spp.) and Mitchell grass (*Astebla lappacea*) on the waterpond bank and by direct drilling of these species along the centre of each waterpond (Thompson, 2008). A number of the vegetation species that naturally colonise the rehabilitated landscape and those selected for introduction are suitable for livestock grazing. Cunningham *et al*. (1987) found that waterponding of scalded areas could increase dry matter vegetation by up to 2.4 tonnes per hectare annually. Further work by Thompson (2008) shows an up to 80% increase in groundcover, including perennial native and introduced grasses and saltbush can be achieved following 5-7 average rainfall years.

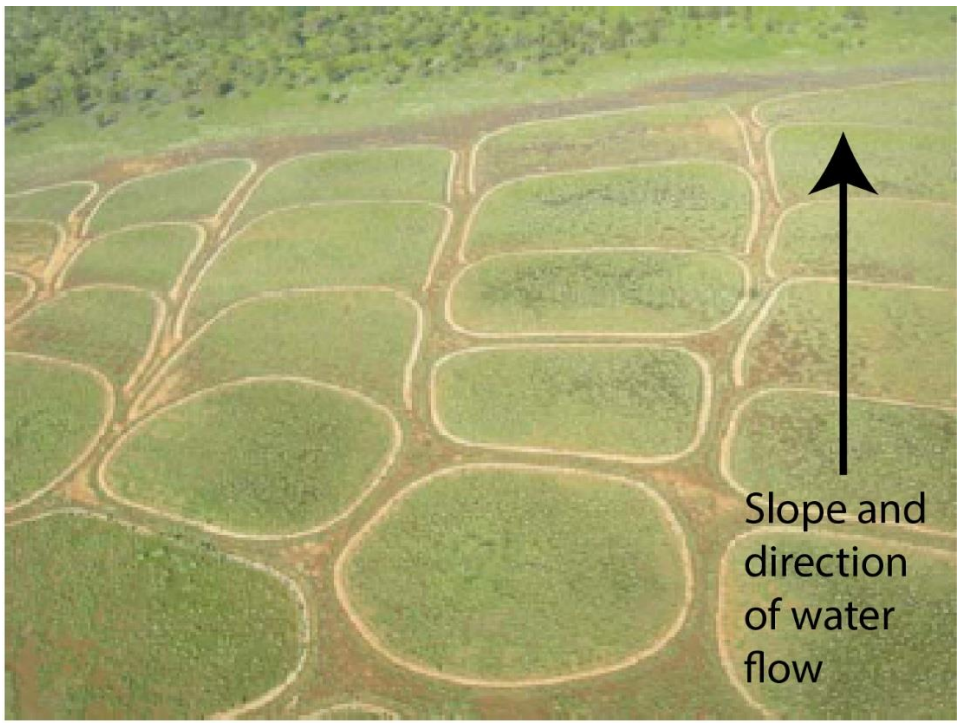


Figure 5-5. Aerial photograph showing a network of horseshoe shaped and fully enclosed waterponds. Fully enclosed waterponds are established where the surface is flat, horseshoe shaped waterponds are established on sloping surface. (Source: R. Thompson).

Waterponding is an effective method for rehabilitating the degraded scalded landscape because it traps surface rain water long enough for it to permeate through the surface crust and enter the soil profile. As water infiltrates the soil profile it leaches soluble salts (Jones, 1967; Ringrose-Voase *et al*. 1989*a*) from the soil profile and restores the shrink swell function of the clay rich subsoil (Ringrose-Voase and McClure, 1994). Surface and deep cracks appear allowing naturally available wind and water borne grass and forb seed to be retained. The seed is then able to germinate and survive in the newly hydrated cracked soil. Further, because water is essential for plant survival, waterponding is important for storing soil moisture and to increase the productivity of this otherwise water poor semi-arid landscape (Noy-Meir, 1973). There are structural constraints in the construction of waterponds. For example, Rhodes (1987) found that the water should not be allowed to pool deeper than 10 cm in the waterponds. This is because the wave action of the water can damage the wall, but also to prevent water tolerant but poor grazing quality plant species from establishing.

Since 1984 over 56 700 waterponds have been constructed around the Marra Ck region, rehabilitating an area of over 28 000 ha (Thompson, 2008). The waterponding method has been used effectively on scalds located in the Kimberly in West Australia, the Northern Territory, Victoria and other regions of NSW.

The length of time for waterponds to rehabilitate a scalded site depends on rainfall, the time taken for soluble salt to be leached from the scald soil profile, and for vegetation to establish. It can take less than five years for waterponds to be revegetated if good seasonal conditions

prevail, but when conditions are drier it may take up to eight or nine years before the scalds are considered to be rehabilitated (Rhodes, 1987). The landscape is considered to be reclaimed after perennial vegetation has become established within the waterponds (Figure 5-6). The re-establishment of vegetation on scalded areas also leads to landscape function by improving nutrient cycling and potential increases in SOC.

Figure 5-6. A 27 year old waterponded site showing extensive ground cover and plant species diversity.

5.5 Aim

The principal aim of this Case Study is to quantify changes to SCS in the top 30 cm of the soil profile that occur as a result of waterponding using a chronosequence approach from an array of sites representing waterponds of different ages.

To support the aim, the following three objectives have been investigated:

1. Quantify the rate of SOC sequestration over time following waterponding;
2. Identify whether any physico-chemical properties including TN, C:N ratio, TP, C:P ratio, AP, C:AP ratio, TS, C:S ratio, pH, EC, ESP and ASWAT, contribute to the SOC sequestration potential of scalded soil following waterponding; and
3. Determine whether waterponding leads to clay mineralogical changes that may affect the C sequestration potential of the soil.

5.6 Study Sites

The study area is located in the Central West semi-arid rangelands of NSW and extends along a transect adjacent to Marra Creek, to the north of Nyngan, between coordinates 30° 55' 53.9S"S, 147° 10' 18.8"E in the south, and 30° 20' 39.6"S, 147° 11' 23.9"E to the north (Figure 5-7). Marra Ck is situated between the Bogan and Macquarie Rivers. The elevation of the study area ranges between 130 and 160m.

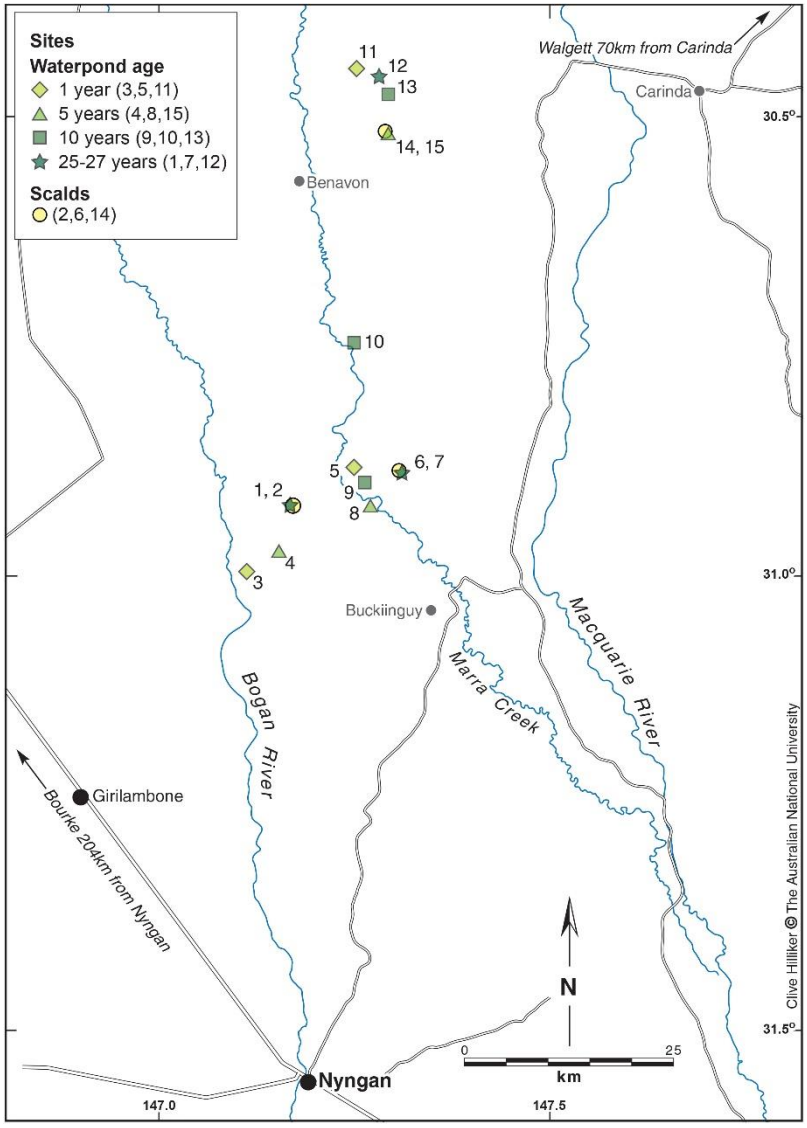


Figure 5-7. Study site locations. Scalds located at Sites 2, 6 and 14.

5.6.1 Physiography

The study region is adjacent and to the east of the Cobar Peneplain bioregion as described by Waters *et al*. (2015). The eastern boundary of the Cobar Peneplain bioregion is the Bogan River. The Marra Creek is east of the Bogan River and to the west of the Macquarie River. The physiography of the region is described in detail by Watkins and Meakin (1996). The area is located on the Riverine Plain geomorphic subdivision, an alluvial landscape with abandoned,

but slightly elevated Quaternary river systems (Watkins & Meakin, 1996). The sites are located almost exclusively on the Bugwah Formation (Qr) (Late Pleistocene to Holocene). The fluvial Marra Creek Formation (Qm) (Holocene) are spatially nearby, tend not to be scalded (Forbes *et al*. n.d). The Bugwah Formation is a mixed-load sediment system and comprises elevated, scalded narrow meander plains, whilst the Marra Creek Formation is described by Watkins and Meakin (1996) as having a modern fine suspended load system with channelized backplains and marshes. The provenance of the sediments for the alluvium is generally the upper catchment of the Macquarie River which drains the rocks of the Lachlan Fold Belt. The origin of the salt in the Bugwah Formation is open to some debate as the levels are higher than those of the older Carrabar and Trangie Formations and is possibly a consequence of the drier period the climate of Australia about 20 000 years ago (Forbes *et al*. n.d).

5.6.2 Climate

The region has a climate characterized as semi-arid. The average annual precipitation is 432 mm which is summer dominant (Figure 5-8) and average annual evaporation is 2023 mm. The annual mean maximum temperature of the location is 26.8 ºC and the annual mean minimum temperature is 12.7 ºC (ANU CLIM, 2014).

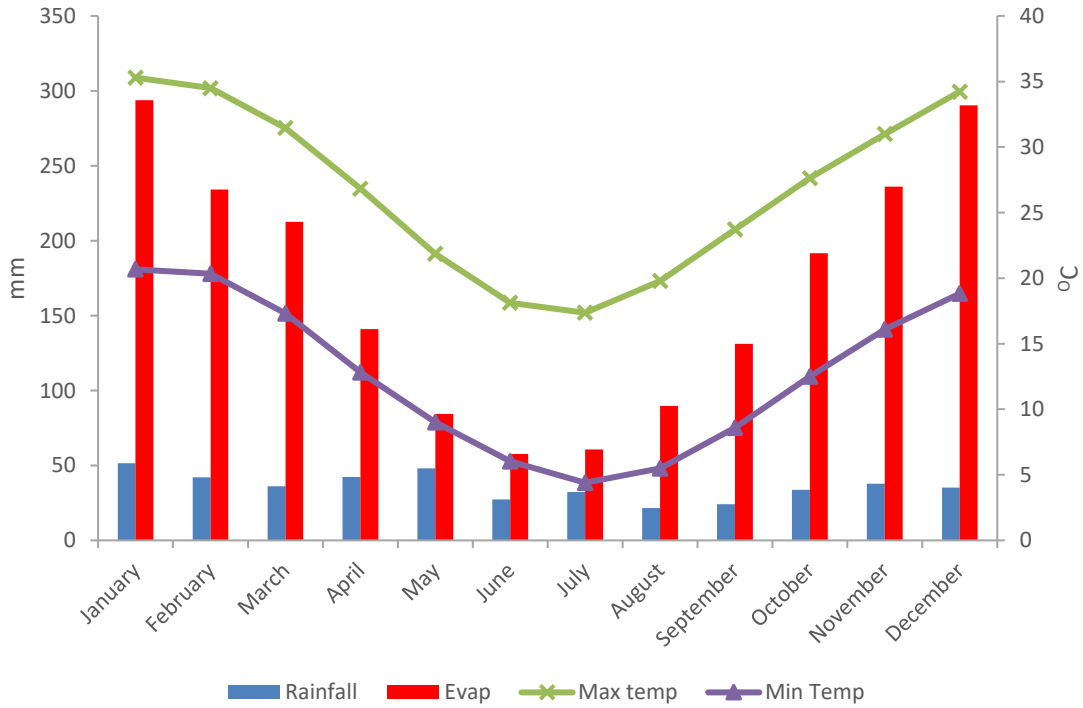


Figure 5-8. Mean climate data for the study sites 1976-2005 (source: ANUCLIM, 2014) (× evaporation; ▲ mean monthly rainfall; red bar monthly maximum temperature; purple bar minimum monthly temperature).

5.6.3 Site details

Twelve privately owned grazing properties were selected by the Central West Catchment Management Authority (CWCMA) for inclusion in this research project. The properties were located along a 60 km transect adjacent to Marra Creek (Figure 5-7). Twelve waterpond sites and three scald sites were included in the study. The age of the waterpond sites ranged between 1 and 27 years since establishment at the time of sampling. The waterpond sites were categorised into four age classes (Table 5-1).

Table 5-1. Age class categories based on years since establishment of waterponds (site numbers in parenthesis).

Age class	*Description*
1 year	Waterponds established for 1 year (3, 11, 5)
5 years	Waterponds established for 5 years (4, 8, 15)
10 years	Waterponds established for 10 years (9, 13, 10)
25-27 years	Waterponds established for 25-27 years (1, 7, 12,).

5.6.4 Soil sampling strategy - waterponds

To minimise bias in selecting waterponds for inclusion in the study, ten waterponds were randomly selected using an aerial photograph of each site and allocated an identification number between one and ten. Then, using a random number generator, three of the ten waterponds were selected from each site to give three replicates per site. A total of 36 waterponds were included in this study.

5.6.5 Within waterpond variation

On visual inspection there appeared to be a vegetation gradient within each waterpond (Figure 5-9). Therefore, it was felt that there may be a commensurate spatial variation in SCS within waterponds. To explore this possibility, each waterpond was divided into three "within waterpond" positions as follows:

- **The 'wall'** position visually displays the highest quantity of biomass. This occurs because most resources including water, seed and soil accumulates downslope in each waterpond, namely close to the bank. (Note: Construction of the waterpond bank is achieved using soil taken from a 'borrow' area approximately 1 m either side of the bank. Soil cores were not obtained from the borrow area);
- **The "mid"** position was located across the centre of the waterpond and was considered to have an 'average' accumulation of resources; and
- **The "top"** position often had sparse and patchy vegetation cover. This position was located close to the opening of the waterpond, and the bank of the next upslope waterpond.

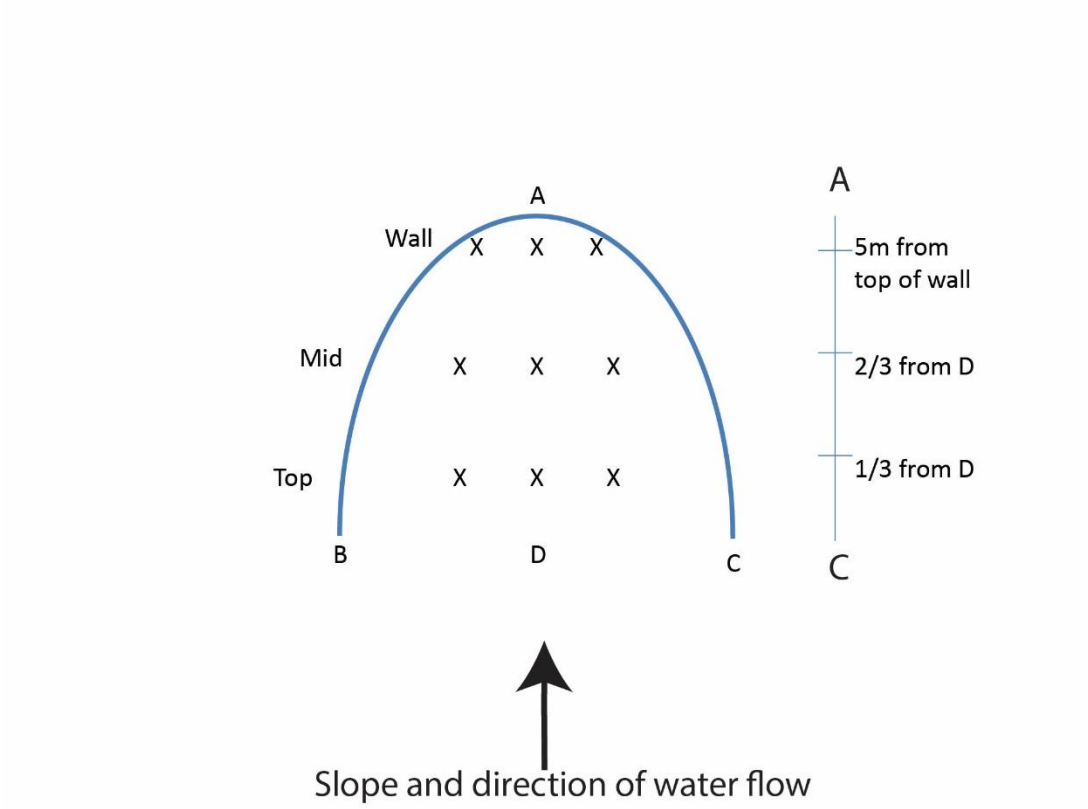


Figure 5-9. Layout of soil coring positions used in each waterpond. Three cores taken from each of the wall, mid and top positions and bulked respectively.

5.6.6 Soil sampling strategy for clay pan scalds

At each clay pan scald site one 25 m x 25 m quadrat was set up and 100 cells established using the method described by Bowman *et al.* (2009). A random number generator was used to select 9 cells. Soil cores were collected to 40 cm depth from these randomly selected cells. An example of the quadrat layout is shown at Figure 5-10. The soil cores from each cell were kept separate to give 9 cores by 4 depth increments for each scald site (total n = 27).

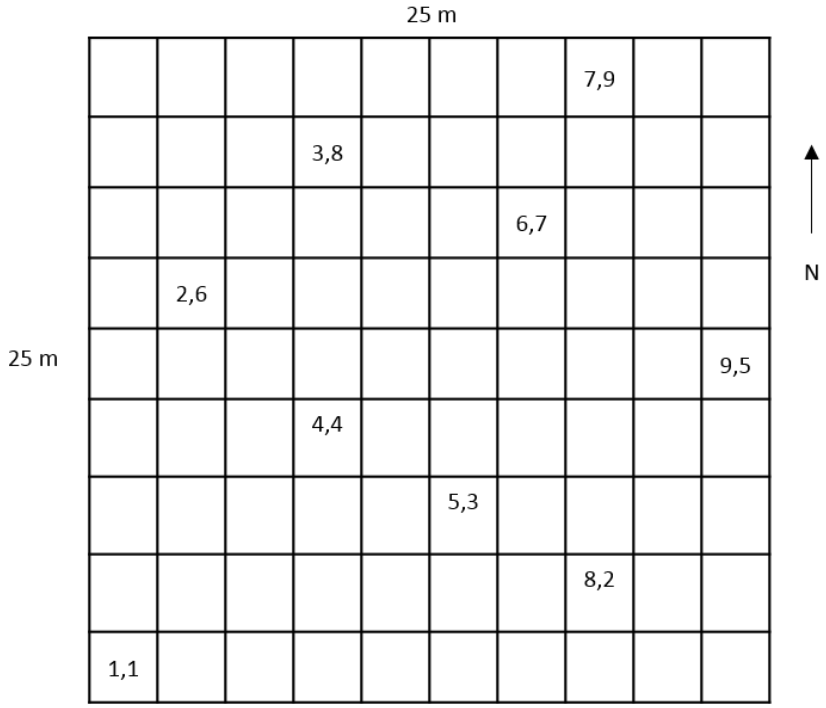


Figure 5-10. Example of soil core location in scalds using a 25 m x 25 m quadrat.

5.6.7 Soil coring method

Soil cores to 40 cm depth were collected in October 2010 and February 2011 using a truck mounted mechanical corer (internal diameter of the core was 40 mm) (Figure 5-11). To maximise the quantity of soil sampled, and to minimise the potential for BD error, three soil cores were taken from each waterpond position and then bulked. For example, a single core was collected from each “X” shown at Figure 5-9; i.e., three cores were collected 5 m in from the waterpond wall and bulked, three laterally across the centre (“mid”) of the waterpond and bulked and three across the top (eg, the opening of the horseshoe shape) of the waterpond and bulked (Figure 5-9). The depth of each core hole was measured to confirm the core depth. Where there was a discrepancy, the sample was discarded and another taken. Each core was carefully placed onto a split PVC pipe, measured and sliced into four depth increments at 0-5 cm, 5-10 cm, 10-20 cm and 20-30 cm depth. The remaining 30-40 cm sample was discarded. Samples were bulked according to site, waterpond number, position and depth. For example, the three samples taken from a given site, from waterpond number 1, wall position in the 0-5 cm depth increment were bulked into one plastic bag.

Figure 5-11. Soil coring rig used for sample collection on scalds and waterponds.

5.6.8 Preparation of samples for analysis and storage

On return to the laboratory the field moist samples were weighed and approximately 20 g sub samples taken to determine soil BD (Chapter 3). The remaining soil was air dried and homogenised. Approximately 100 g of the air dried soil was crushed using a mortar and pestle to pass through a 2 mm sieve. Root and litter fragments > 2mm were discarded Using the sieved 100 g subsample approximately 50 g was ground using a mortar and pestle to pass through a <500 µ sieve. The 2 mm and <500 µ sub samples were stored in separate plastic containers. The remaining un-sieved subsample was stored in sealed plastic bags.

5.7 Laboratory Analysis

Analysis was conducted on either all (n = 540) samples, or from one of three subsets of samples as dictated by resource constraints, as follows:

- Subset 1: (n = 144). This subset was purposely selected based on the proximity of a scald (Site 2) with three nearby waterpond sites (Sites 1, 3 & 4) with ages one year, five years and 25 years since establishment. The proximity criterion was used to minimise extraneous effects due to distance between sites such as climate and soil type. The Subset 1 sites were close enough together to be considered paired and thus allow changes to SOC to be measured on a temporal basis. All samples from all depth increments for these three sites were included in the subset.
- Subset 2: (n = 8). This small Subset comprised one core from a scald (Site 2) and a paired (based on proximity) 25 year old waterpond (Site 1). The Subset was purposely selected from samples included in Subset 1 using an old waterpond which thereby allowed the longest time lapse possible (within the sample set) since establishment for mineralogical change to have occurred. One core with a sample from each of the four soil layers (0-5, 5-10, 10-20 and 20-30 cm layer) from the scald and waterpond. The purpose of this data set was to determine whether waterponding leads to soil mineralogical change and thereby influence the C sequestration potential of the soil. And
- Subset 3: (n = 20). This final Subset was made up of composite samples including all three scald sites with waterpond sites located in close proximity. In addition, samples from two “relic” hummocks were combined and analysed. The aim of this subset was to identify erosion trends and to establish whether the hummocks were relics or more recent deposits.

An overview of the analyses conducted for each subset appears below (Figure 5-12).

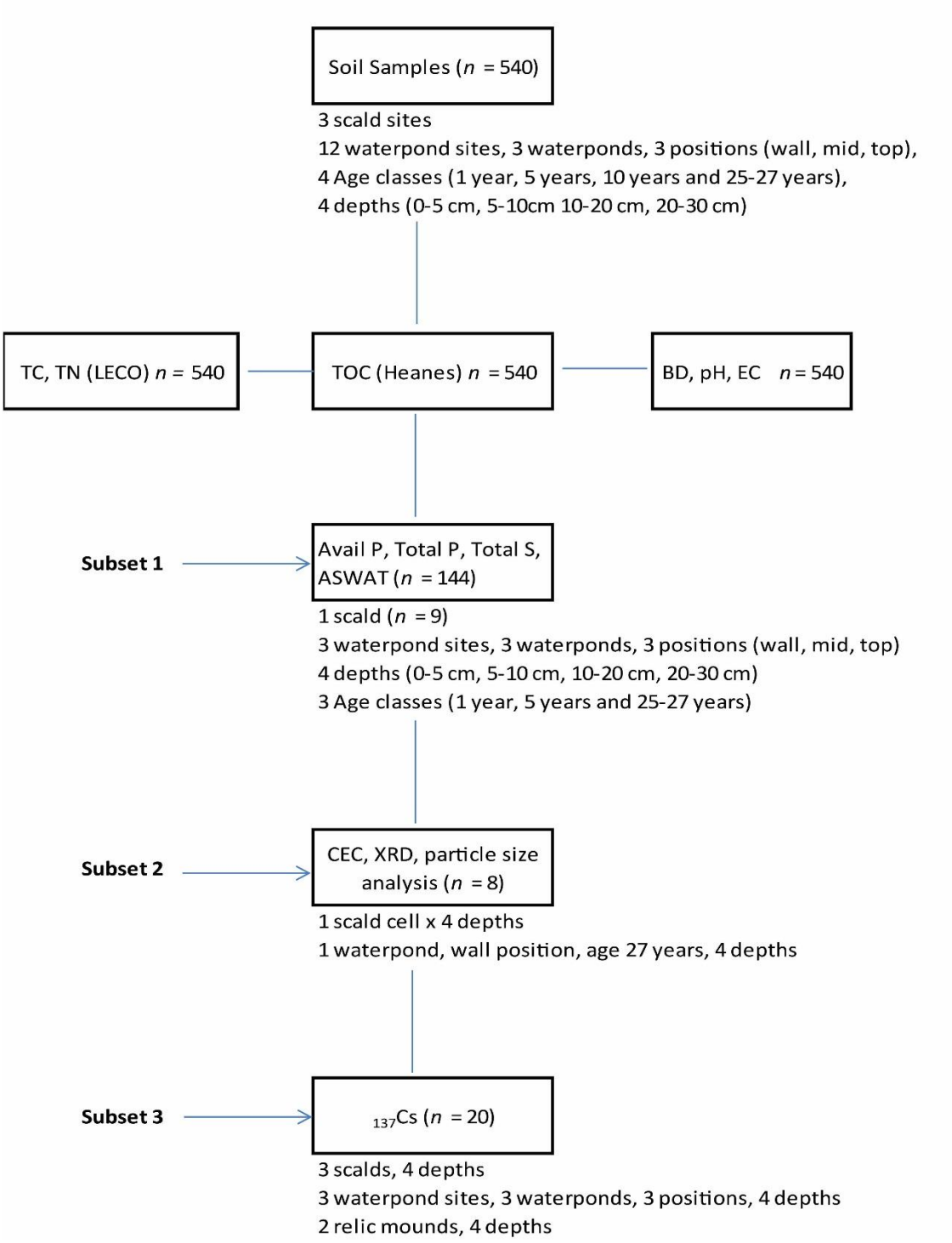


Figure 5-12. The soil analysis strategy showing the composition of each of the three subsets and the analyses carried out on each subset.

The following analysis was conducted on all soil samples (n = 540) obtained for this case study (Figure 5-12):

- soil BD (expressed as Mg m^{-3});
- The $pH_{(1:5\ water)}$ and $EC_{(1:5\ water)}$;
- $SOC_{(Heanes)}$ concentration, is a percentage of C in soil by weight and expressed throughout this report as grams of C per 100 grams of soil (g C 100 g^{-1}); and
- SCS to a fixed depth was calculated to quantify the SCS (expressed as Mg C ha^{-1}) in soil.

Details of the methods and equations used for the aforementioned analyses are provided in Chapter 3. Details of the methods and equations used specifically for this case study including ECe, TSC, TN and the ESM method follow.

5.7.1 Electrical conductivity ($EC_{(1:5water)}$) and electrical conductivity of saturated extracts ($EC_{(sat)}$)

$EC_{(1:5\ water)}$ is a determination of the quantity of soluble salts per unit weight of soil and is measured using a soil water suspension. The method used to determine $EC_{(1:5\ water)}$ is described in Chapter 3. To determine the effects of soluble salts on vegetation the concentration of EC(sat) as kg of salts per litre of soil water has been calculated for scalded and waterponded soil using the appropriate conversion factor as based on soil texture grades recommended by Slavich and Petterson (1993) (Table 5-2).

Table 5-2. Multiplier factors for converting $EC_{(1:5\ water)}$ (dS/m) into EC(sat) (dS/m) (After Slavich and Petterson, 1993).

Soil texture	Multiplier factor
Sand, loamy sand, clayey sand	22.7
Sandy loam, fine sandy loam, light sandy clay loam	13.8
Loam, fine sandy loam, silty loam, sandy clay loam	9.5
Clay loam, silty clay loam, fine sandy clay loam, sand clay, silty clay, light clay, light medium clay	8.6
Medium clay	7.5
Heavy clay	5.8

5.7.2 Total soil carbon and total nitrogen analysis

Analysis for TSC and TN was conducted on all samples using a Leco CNS2000 dry combustion analyser at the Environmental Analysis Laboratory, Southern Cross University, NSW Australia. The method used was DCHTC using Method 6B2a for TC and Method 7A5 for TN, (Rayment and Lyons, 2011).

C:N ratios are recorded as the ratio of TSC and TN as analysed by a Leco CNS2000 dry combustion analyser.

5.7.3 Carbon stock in an equivalent soil mass

To overcome a potential source of error caused by changes in BD following waterponding an ESM equation was used. SCS in an ESM were calculated for each sample to account for the difference in soil mass between scalded soils and waterponded soils. The mean soil mass from the three scald sites was used as the reference soil mass. The equations applied were based on those described by Ellert and Bettany (1995) and Ellert *et al*. (2001):

(1) The mass of soil from each layer:

$$\text{Mass of soil (Mg ha}^{-1}\text{)} = \text{BD (Mg m}^{-3}\text{)} \times \text{thickness of soil layer (m)} \times 10\,000$$

Equation 14

(2) Mass of C per unit area (fixed depth):

$$\text{Mass of C (Mg ha}^{-1}\text{)} = \text{(C concentration (SOC}_{\text{(Heanes)}}\text{) (g 100g}^{-1}\text{)} \times \text{BD (Mg m}^{-3}\text{)} \times \text{thickness of soil layer (m)} \times 10\,000$$

Equation 15

(3) The thickness of the layer required to attain an ESM was calculated by:

$$\text{Thickness to add} = \frac{\text{mass of soil in heaviest layer (Mg ha}^{-1}\text{) – mass of soil in surface layer (Mg ha-1)) x 0.0001 (ha m}^{-2}\text{)}}{\text{Bulk density in sub surface (Mg m}^{-3}\text{)}}$$

Equation 16

(4) Finally, the amount of C in the ESM was calculated by summing the mass of C from successive layers plus the thickness to add calculated by:

$$\text{Mass of C in an ESM} = \text{Mass of C in the surface layer (Mg C ha}^{-1}\text{)} + \text{Mass of C in thickness to add (Mg ha}^{-1}\text{)}$$

Equation 17

A worked example showing the calculation of the equivalent soil mass for a waterpond compared with mean of the three scald sites taken as representing the baseline SCS is shown in Table 5-3. In the example the baseline scald SCS is 18.8 Mg ha^{-1} which is calculated on a fixed depth basis (Equation 15).

Table 5-3. Worked example for the calculation of equivalent soil mass for a 5 year old waterpond (wall position). Number in parenthesis refers to the relevant equation number used for the calculation.

Soil layer (cm)	*Mean SCS of scald soil (Fixed depth) (15) (Mg ha^{-1})*	*Mean mass of scald soil (14) (Mg ha^{-1})*	*Mass of WP soil (14) (Mg ha^{-1})*	*WP SCS (Fixed depth) (15) (Mg ha^{-1})*	*Thickness of additional layer (16) (cm)*	*C mass in ESM (17) (Mg ha^{-1})*
0-5	3.0	760	622	9.7	1.1	10.8
5-10	3.4	778	704	4.7	0.5	5.2
10-20	6.6	1516	1336	9.7	1.3	11.1
20-30	5.8	1548	1346	8.8	1.5	10.3
0-30	18.8	4602	4008	32.9		37.4

5.7.4 Subset 1 analysis (ASWAT, AP, TP and TS)

A range of physico-chemical properties were analysed on the Subset 1 samples, (n = 144) to further characterise the soils:

- Aggregate stability in water (ASWAT);
- Available phosphorus (AP);
- Total phosphorus (TP); and
- Total sulphur (TS).

Details of each of these methods follow.

5.7.5 Aggregate stability in water

Structural stability of soil was assessed using the aggregate stability in water (ASWAT) method recommended by Field *et al*. (1997) on a subset of 144 samples. The ASWAT method assesses the propensity of soil to disperse in distilled water and is a modification of earlier methods developed by Emerson (1967) and Loveday and Pyle (1973). Briefly, the ASWAT method entails gently placing a pea size air dry aggregate into a container containing distilled (as MilliQ) water. After 10 minutes the samples were visually assessed for dispersion and scored with a range of values between 0-4 (Table 5-4).

Table 5-4. ASWAT visual assessment scoring criteria (Field *et al*. 1997).

Score	Visual assessment
0	No dispersion observed within 2 hours
1	Slight dispersion within 2 hours
2	Slight dispersion within 2 hours or strong dispersion within 10 minutes
3	Strong dispersion within 10 minutes or complete dispersion within 2 hours
4	Complete dispersion within 10 minutes

The samples were reassessed after two hours and again scored with values between 0-4 (Figure 5-13). The two scores were added together giving a result between 0-8. For samples that

dispersed the score is added to 8 to give a total score of between 9 and 16. All samples with a score of zero, indicating no dispersion, were remoulded and a pea sized sample placed in distilled water and monitored after 10 minutes and two hours. Scores from the dry aggregate and remoulded sample were added together giving a score of between 0-16. Scores greater than 8 indicate that the soil is susceptible to spontaneous dispersion. Four replicates of the air dry sample and four replicates of each remould were assessed following the recommendation of Field *et al*. (1997).

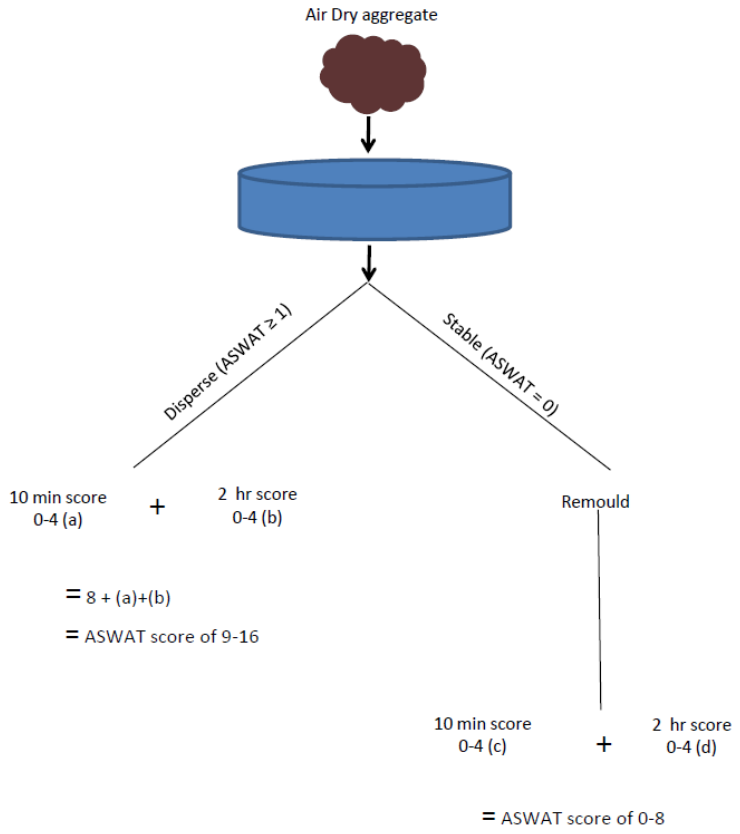


Figure 5-13. ASWAT scoring method (From: Field *et al*. 1997).

5.7.6 Available Phosphorus

A subset of Case Study 2 samples (n = 144) were analysed for AP. The method used to determine AP was based on the Olsen P Method 9C2b where 0.5M $NaHCO_3$ (pH 8.5) is the extracting solution (Olsen *et al*. 1954; Rayment and Lyons, 2011). Reagents and Standards were prepared using the QuikChem Method 12-115-01-1B (Lachat Instruments). Extracts were analysed using a Lachat Flow Injection Auto analyser (FIA).

5.7.7 Total Phosphorus and Total Sulfur

Analysis for TP and TS was conducted by the Environmental Analysis Laboratory, Southern Cross University. The method used was 3:1 Nitric/HCl digest - APHA 3120 using Inductively Coupled Plasma - Optical Emission Spectrometry (ICP-OES). This is method 17B1 (Rayment and Lyons, 2011) pseudo-total elemental analysis of soils and sediments – reverse aqua regia block digestion with determination by atomic spectroscopy.

5.7.8 Subset 2 analysis (PSA, ECEC, XRD, ICP-AES, ESP)

The samples for subset 2 were selected using purposive sampling method (McKenzie and Cresswell, 2002:13). Because resources for this suite of analysis were limited only two cores could be analysed. The cores selected were from a scald (Site 2) and 25 year old waterpond (Site 1, from the wall position) that were situated in close proximity, thereby reducing the possibility of different soil morphological properties between the two samples. The waterpond site was selected because it was one of the older sites thereby ensuring the possibility that sufficient time had elapsed for mineralogical change to become apparent. All four soil layer samples from Subset 2 (n = 8) were analysed to identify whether waterponding leads to any changes to soil mineralogy. The particular analyses used in this study included:

- Particle size analysis (PSA),
- Effective cation exchange capacity (ECEC),
- X-ray diffraction (XRD),
- Inductively coupled plasma – atomic emission spectrometry (ICP-AES), and
- Exchangeable sodium percentage (ESP).

5.7.9 Dispersibility of small aggregates by particle size analysis

An analysis of the particle size distribution of both the scald and waterpond soils was undertaken to determine whether waterponding leads to changes in soil structure. Sample particle size distribution was determined by a PSA process involving both chemical ($NaPO_3$) and ultrasonic destruction of soil aggregates as follows.

Samples from Subset 2, plus two other waterpond positions (mid and top) and from four depth increments of one scald (Site 2) and three positions from one 25 year old waterpond (Site 1) were analysed for particle size distribution using a Malvern Mastersizer 2000 Laser Diffraction particle size analyser. Soil was sieved to pass through a 2 mm sieve. A small quantity of the sieved sample (approximately 1 g) was placed into a 900 ml beaker of tap water and the Mastersizer 2000 was run to determine and record the mean particle size five times. This treatment represents “mild dispersion”. Next, 5 drops of sodium hexametaphosphate $(NaPO_3)_6$ (0.5 g l^{-1} solution) was added to the solution to disperse any aggregates (Mason, 2003). The Mastersizer 2000 was started again to analyse the sample another 5 times. Finally, the solution

was exposed to 30 seconds of ultrasonic vibration to break down any resistant aggregates before the solution was analysed again another five times. These two treatments represent "full dispersion". The full dispersion treatment, which was applied uniformly to the scald and waterpond samples, was designed to break down macroaggregates into small stable aggregates.

5.7.10 Effective cation exchange capacity

Analyses to determine the amounts of the exchangeable cations calcium (Ca^{2+}), magnesium (Mg^{2+}), potassium (K^{+}) and sodium (Na^{+}) and the ECEC were carried out to identify differences in the cation composition of the scalded and waterponded whole soil fraction. The analyses were undertaken at the Environmental Analysis Laboratory, Southern Cross University. Because the scalded soils are saline (EC 1.84 – 3.02 dS/m) the method used to determine ECEC was 15C1 (Rayment and Lyons, 2011). The method uses a pre-treatment to leach out the soluble salts before the exchangeable cations are extracted with 1 M NH_4Cl. Results are reported on an oven dry weight basis.

The ECEC of the soil was also predicted by using the mineralogy determined by XRD for the clay fraction of the scald and waterpond samples used in Subsample 2, and by adopting CEC values for clay minerals proposed by Greene *et al*. (2002), McKenzie *et al*. (2004:13) and T. Eggleton (pers comm, 13 Jan 2016). The CEC of the clay fraction was then converted to a soil CEC using a value of 35% clay content of the soil. The CEC values used for the individual clay minerals was 100 for smectite, 20 for illite and 4 for kaolinite. For the purpose of the calculation, other clay minerals were assumed to have a CEC of 0 $cmol^{+}kg^{-1}$.

5.7.11 X-Ray Diffraction

XRD was used to identify, characterise and compare the minerals in the soil from one scald and one 25 year old waterpond (Sites 1 and 2). These samples were purposely selected as representative of the two extremes (a scald compared with oldest waterpond) from a site where the scald was immediately adjacent to the waterpond. The aim was to identify whether waterponding caused any significant mineralogical changes.

Powder diffraction analysis was undertaken at the X-ray Diffraction Laboratory, Research School of Earth Sciences (RSES), ANU, using a Siemens D501 Bragg-Brentano x-ray diffractometer.

Bulk samples were milled with a McCrone micronizing mill and suspended on a side packed sample holder to avoid preferred orientation. The samples were analysed from 2-70°, 2-theta, with a step width of 0.02° with a scan speed of 1° per minute using CuKα radiation.

Clay fraction samples were separated using the settling method and oriented samples prepared by the millipore filter method (Moore and Reynolds, 1997). Samples were then analysed by XRD following a series of treatments as follows:

1. Samples were saturated with a 0.1M $MgCl_2$ solution while suspended on the Millipore filter. The $MgCl_2$ sample was then x-rayed from 2-42°, 2-theta, with the same step width and speed as the bulk sample;
2. The $MgCl_2$ soaked sample was then saturated with ethelyne glycol at 60°C inside a desiccator. The sample was then x-rayed at 2-32°, 2-theta;
3. The sample was then heated at 350°C in an oven for 1 hour and then x-rayed from 2-28°, 2-theta;
4. The sample was then heated at 550°C for 1 hour in an oven and x-rayed again from 2-28°, 2-theta; and
5. Clay species were identified by the position and potential shift of the 0 01 peak.

Mineral identification of the bulk scan was performed using XRD processing software DIFFRAC^{plus} Eva (2003) using the data base Powder Diffraction File (PDF2) (Nash *et al*. 2013).

Clay mineral quantification using the XRD scan of the bulk sample was performed using SIROQUANT V.3 software (Alves and Omotoso 2009). The database contained in the SIROQUANT V.3 software contains both calculated clay XRD patterns as well as a clay pack of experimentally determined XRD clay patterns which is used for more disordered clays (for example a montmorillite/illite interlayered clay).

Whilst clay quantification is still very challenging, undertaken with many potential pitfalls, and procedural flaws based on suppositions, including the belief that all mineral phases have been identified and the sum of the phases is 100% (Kahle *et al*. 2002), the above sample preparation method and Riedvald refinement programs such as SIROQUANT are the best approach available today (Alves and Omotoso, 2009). The RSES XRD laboratory where this analysis was undertaken has participated several times in the Clay Minerals Society, Reynolds Cup, which is an international round robin test with the aim of improving analytical techniques for clay quantification (Omotoso *et al*. 2006).

5.7.12 Elemental analysis for oxides using ICP-AES

Elemental analysis for major elements (oxides) was undertaken using a lithum borate fusion and Inductively Coupled Plasma atomic emission spectrometry (ICP-AES). The analysis was used to determine the oxide composition of the scalded and waterponded soil samples for the purpose of identifying any mineralogical changes that may have been initiated by waterponding. The oxide analysis was undertaken by Bureau Veritas Minerals Pty Ltd at the Thebarton mineral testing and laboratory service. The major elements were reported as oxides. ICP-AES detection limits (percent) for the elements are: Al_2O_3 (0.01), CaO (0.01), Fe_2O_3 (0.01), K2O (0.01), MgO (0.01), MnO (0.01), Na_2O (0.01) P_2O_5 (0.01) SiO_2 (0.01), TiO_2 (0.005), LOI (0.01). Briefly, the method used included fusing an aliquot of soil with lithium metaborate at high temperature in a

platinum crucible. Nitric acid is used to digest the fused glass to achieve complete dissolution of most minerals. The high fusion temperatures results in loss of volatile elements.

5.7.13 Exchangeable sodium percentage

To evaluate the potential sodicity of scalded and waterponded soil the level of ESP of the CEC was calculated. The ESP was calculated using the following formula:

$$ESP = (100 \cdot \text{exch Na})/ CEC$$

(where units are expressed in $cmol^{+}\ kg^{-1}$)

Equation 18

5.7.14 Subset 3: analysis of the hummock soil

Subset 3 was prepared specifically to undertake caesium-137 (^{137}Cs) analysis to confirm the idea proposed by Beadle (1948a); Warren (1965); Cunningham (1987) and Ringrose-Voase *et al.* (1989a) that vegetated hummocks may be relic soils possibly representing the soil profile prior to erosion or accumulation of material. To do the analysis, composite samples were prepared using soil samples collected from the three scald sites (Sites 2, 6 and 12), two hummock sites (located at Sites 2 and 6), and the three oldest waterpond sites (Sites 1, 7 and 13). Samples were prepared for four depth increments 0-5 cm, 5-10 cm, 10-20 cm and 20-30 cm for the scald and waterpond sites. The hummock samples were divided 0-5 cm, 5-10 cm, 10-30 cm and 30-50 cm depth increments. The results of the combinations gave a total of 20 samples for ^{137}Cs analysis. The analysis was conducted using the method outlined by Li *et al*. (2006) and Li *et al*. (2011).

TN and TSC analysis using the DCHTC method was also undertaken for the four composited hummock soil layers using an Elementar Variomax analyser.

5.7.15 Statistical analysis

Means and SEM were calculated on a depth, age cohort, site and position basis for all chemical analysis undertaken. Genstat ® Release 17 was used to undertake ANOVA statistical analysis of the changes in soil properties arising from waterponding. Treatment effects within and between sites for pH, EC, SOC, TSC and CN data was undertaken. The block structure used was location within age-class, within waterpond within depth. The treatment used was a comparison of scald against waterpond age class, depth and location and their interactions. Treatment means were compared using a Tukey post hoc test (least significant difference at P=0.05) to determine significant differences between scalds and waterponds and waterpond age. Linear regression models were used to determine any relationships between nutrient ratios.

5.8 Results

Initially, the chemical properties of the three scald sites were analysed and recorded individually to identify any variation between the scald sites. Next the results of soil chemical properties for the three scald sites were averaged to establish baseline values representative of the pre-waterponding attributes of scalded soil. Finally, to determine whether waterponding of scalded soil results in SOC sequestration or contributes to other edaphic changes in the soil, a number of physical and chemical properties from each of the four waterpond age cohorts (1 year, 5 years, 10 years and 25-27 years) were compared against the scald baseline values to identify any temporal changes. Physical and chemical properties have been measured for each of four soil layers (e.g. 0-5, 5-10, 10-20 and 20-30 cm) and to a cumulative total of 30 cm depth where appropriate. Unless otherwise stated results are expressed as mean values ± standard error.

5.8.1 Baseline soil properties for scalded soil

The three scalds were found to have a mean pH in the neutral to slightly alkaline range of approximately 7.0. The mean pH from the three sites in the 0-5 cm soil layer was found to be 7.3 and the 20-30 cm layer has a pH of 7.6. The ANOVA identified that pH statistically significantly increases with soil depth and is statistically significantly different between scald sites ($P = 0.002$ and $P<0.001$ respectively). There were no statistically significant differences between soil depth and scald sites ($P = 0.76$).

Table 5-5. Mean $pH_{(1:5\ water)}$ and $EC_{(1:5\ water)}$ for each of the scald sites by depth increment. Standard error in parenthesis. Within rows, variable means followed by the same letter are not significantly different (Tukey HSD using $\alpha = 0.05$).

Soil depth (cm)	Site 2 pH (1:5)	Site 2 EC (1:5) dS/m	Site 6 pH (1:5)	Site 6 EC (1:5) dS/m	Site 14 pH (1:5)	Site 14 EC (1:5) dS/m
0-5	7.5 (0.2)$_a$	1.3 (0.3)$_a$	7.1 (0.1)$_b$	1.9 (0.2)$_{ab}$	7.4 (0.1)$_{ab}$	2.3 (0.3)$_b$
5-10	7.5 (0.2)$_a$	2.6 (0.3)$_b$	6.9 (0.6)$_a$	2.5 (0.1)$_b$	7.2 (1.0)	3.6 (0.3)$_a$
10-20	7.6 (0.2)$_a$	2.7 (0.3)$_a$	6.9 (0.1)$_b$	3.0 (0.1)$_a$	7.4 (0.1)$_a$	3.7 (0.2)$_b$
20-30	8.0 (0.2)$_a$	2.5 (0.2)$_a$	7.3 (0.1)$_b$	3.1 (0.1)$_b$	7.5 (0.1)$_{ab}$	3.45 (0.1)

Scalds were found to be highly saline, with an increasing EC trend with depth. EC results show a mean EC for the three sites of 1.84 dS/m in the surface 0-5 cm soil layer. ANOVA suggested a significant difference between sites and for depth in the scalds ($P < 0.001$, LSD = 0.306 for site; $P < 0.001$, LSD = 0.353 for depth). The overall mean EC for Site 14 was significantly different (higher) than Site 2 and Site 6 (Tukey's 95 % confidence interval). The interaction between soil depth and site was found to be not significant ($P = 0.531$, LSD 0.612). Multiple comparison analysis shows that the 0-5 cm layer is significantly different to the other three soil layers. There is no significant difference between the 5–30 cm layers. The highest EC (3.11 dS/m) was in the 10-20cm depth increment.

The concentration of SOC is based on the amount of C present in soil on a weight basis and is expressed in the current research as g C 100 g^{-1} soil. The $SOC_{(Heanes)}$ concentration in the three scalded soils to 30cm depth is shown in Figure 5-14. There was a significant difference in SOC concentration between sites (P<0.001) and with depth (P=0.005). The SOC concentration was lowest at scald Site 14 (Figure 5-7). The depth effect was statistically significant (P<0.001). The mean SOC concentration for the three sites in the 0-5 cm scalded soil layer was 0.39 g C 100 g^{-1}. The mean concentration of SOC was lower in the 0-5 cm soil layer than in the 5-10 and 10-20 cm soil layers which both had a mean of 0.43 g C 100 g^{-1}. The 20-30 cm depth increment had the lowest SOC concentration with a mean of 0.37 g C 100 g^{-1}.

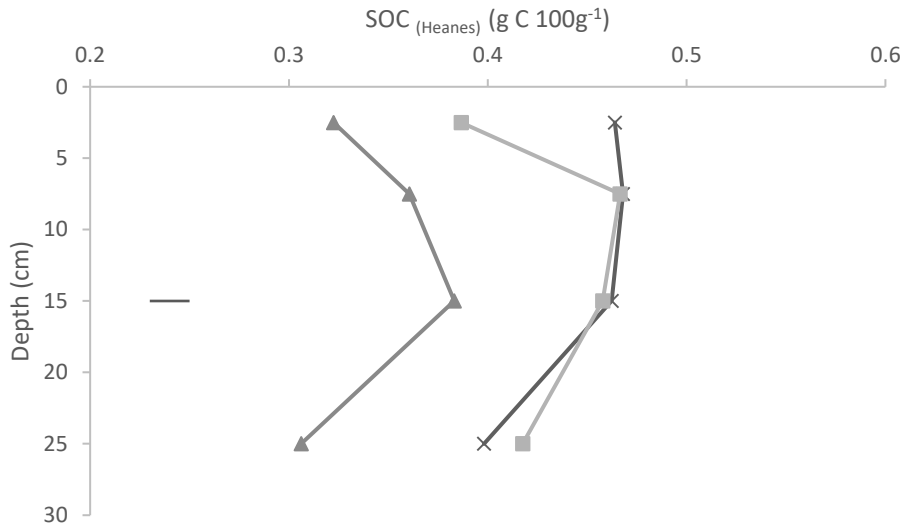


Figure 5-14. SOC $_{(Heanes)}$ (g C 100 g^{-1}) results for three scalded sites (Site 2 ×; Site 6 ■, Site 14 ▲). Horizontal bar = SEM (n=108).

The soil BD of the scalds was found to vary with Site 2 having a particularly high BD in the 0-5 cm soil layer. There was a strong difference in BD between scald sites ($P < 0.001$) (Figure 5-15) with, for example, the difference between Sites 2 and 14 statistically significant at $P = 0.05$ in the 0-5 cm layer. The interaction between site and depth was also statistically significant ($P < 0.001$). Site 2 had the highest mean BD to 30 cm depth (1.7 Mg m^{-3}). Sites 6 and 14 had a mean BD to 30 cm of 1.45 and 1.49 Mg m^{-3} respectively. Scald BD changes with depth was variable and found to be not significantly different ($P = 0.73$). The greatest statistically significant variation between scald sites is expressed in the 0-5 cm soil layer where BD ranges between 1.3 Mg m^{-3} and 1.8 Mg m^{-3}. The results show an overall mean BD for the three sites of 1.5 Mg m^{-3} for the four depth increments. The mean BD in the 0-5 cm layer of the scalds was 1.5 ± 0.06 Mg m^{-3}.

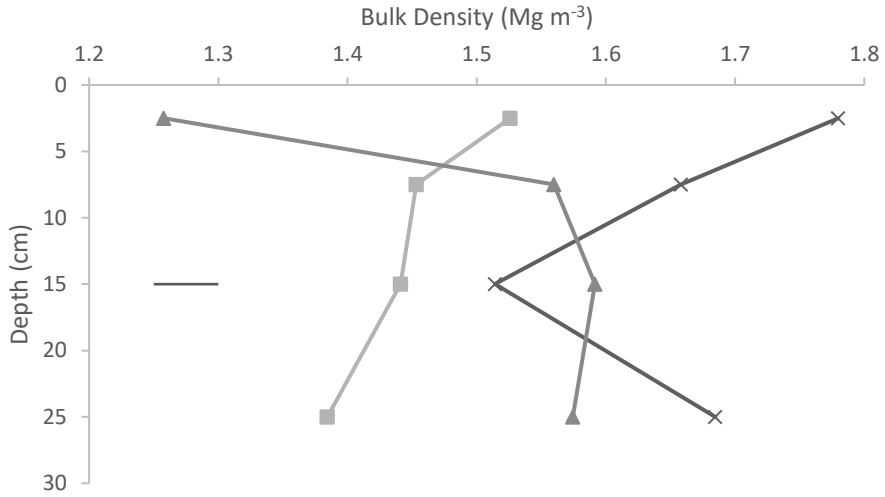


Figure 5-15. Bulk density results for the three scald sites (Site 2 ×; Site 6 ■; Site 14 ▲). (LSD for site 0.0741; Depth 0.0856). (Horizontal bar = SEM (n=108)).

The EC values in the 0-5 cm layer are highest for the Site 14 scald (2.3 dS/m), and lowest at Site 2 (1.3 dS/m) (Table 5-6). The corresponding BD results are lowest at Site 14 (1.3 Mg m^{-3}) compared with 1.8 Mg m^{-3} at scald Site 2. Site 2 was found to have the highest BD and lowest EC, whereas Site 14 had the lowest BD and highest EC (Table 5-6). This relationship indicates that the high EC is promoting high aggregate stability and thus low BD.

Table 5-6. Scald bulk density means (Mg m^{-3}) results. Within columns, means followed by the same letter are not significantly different at P=0.05. EC $_{(1:5\ water)}$ results (italicised) are included for comparison.

Depth (cm)	***Site 2***		***Site 6***		***Site 14***	
	BD Mg m^{-3}	***EC (1:5) dS/m***	***BD Mg m^{-3}***	***EC (1:5) dS/m***	***BD Mg m^{-3}***	***EC (1:5) dS/m***
0-5	1.8d	*1.3a*	1.5bc	*1.9ab*	1.3a	*2.3bc*
5-10	1.7cd	*2.6bcd*	1.5abc	*2.5bcd*	1.6bcd	*3.6e*
10-20	1.5bc	*2.7bcde*	1.4abc	*3.0cde*	1.6bcd	*3.7e*
20-30	1.7cd	*2.5bcd*	1.4ab	*3.1cde*	1.6bcd	*3.5de*

SCS in scalds increases with depth with a mean from the three sites of 3.0 Mg C ha^{-1} in the 0-5 cm layer, and reaching 6.7 Mg C ha^{-1} in the 10-20 cm depth increment. The SCS in the 20-30 cm depth was 5.8 Mg C ha^{-1}. ANOVA of the SCS for the three scald sites suggests that there is a significant difference (P<0.001) for both site and depth values, but the interaction between site and depth is not significant. Of the three sites, Site 14 has the lowest mean SCS of 3.9 Mg C ha^{-1}, Site 6 has 4.7 Mg C ha^{-1} and Site 2 has 5.4 Mg C ha^{-1}, per depth increment (Figure 5-16). Overall, the 0-5 cm depth increment has the lowest mean SCS of 3.1 Mg C ha^{-1}, and the 10-20 cm depth has the highest SCS at 6.6 Mg C ha^{-1}.

When the SCS from the four depths is summed together, the three scald sites have a mean SCS of 18.7 Mg C ha^{-1} (Figure 5-16) to 0.3m soil depth. This value has been determined on a fixed depth basis. SCS to 0.3 m is significantly different (P<0.001) between sites. Site 14 has a significantly lower mean SCS to 0.3 m of 15.70 Mg C ha^{-1} than Site 6 (18.75 Mg C ha^{-1}) and Site 2 (21.69 Mg C ha^{-1}).

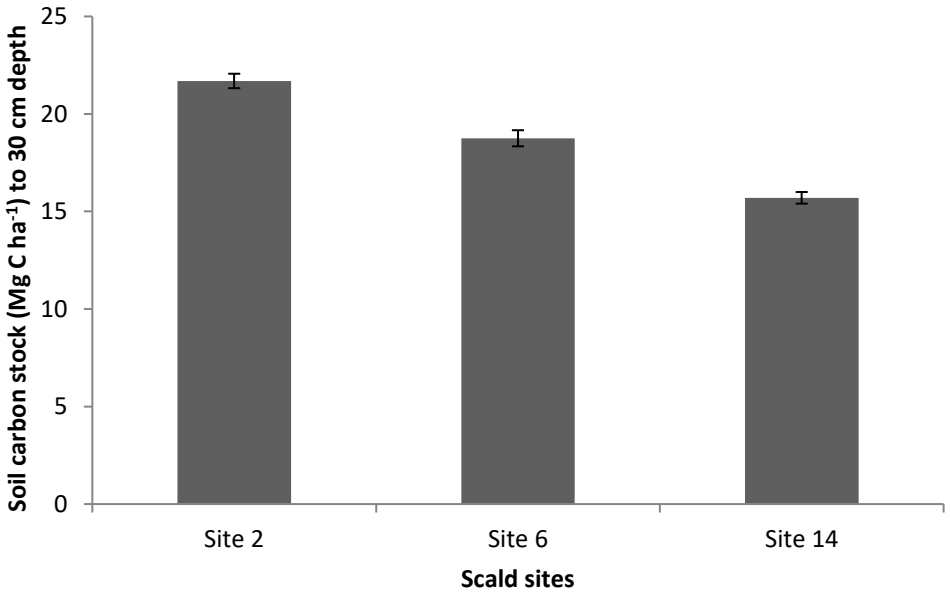


Figure 5-16. SCS (Mg ha^{-1}) to 30 cm depth for three scald sites (error bars = SEM (n=9).

The ANOVA identified no statistical significance for TN in either scald depth, site and the scald depth by site interaction ($P = 0.615$, $P = 0.065$ and $P = 0.286$ respectively. Site 14 had the lowest overall mean TN of 594 mg kg^{-1} incorporating all four depth layers. The mean TN (for four layers) for both Sites 2 and 6 was 650 mg kg^{-1} (Figure 5-17).

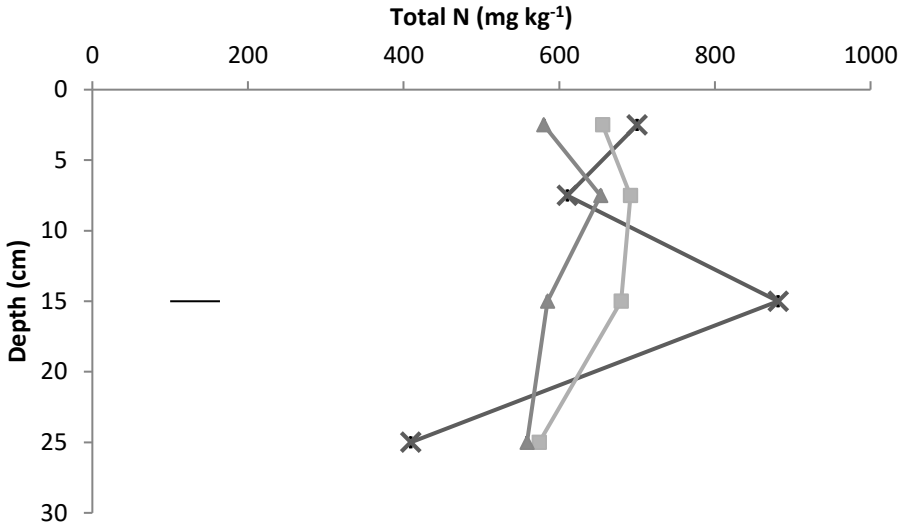


Figure 5-17. TN (Mg kg^{-1}) for the three scald sites (Site 2×; Site 6 ■; Site 14 ▲). Horizontal bar = SEM (n=108).

The C:N ratio was calculated using the results obtained from $TSC_{(LECO)}$ and $TN_{(LECO)}$ analysis. The C:N ratio has a statistically significant ($P < 0.001$) increasing trend with depth (Figure 5-18). The mean ratio was found to be 6 in the 0-5 cm layer and 10 in the 20-30 cm depth. There was no statistical significance found for site or the site by depth interaction.

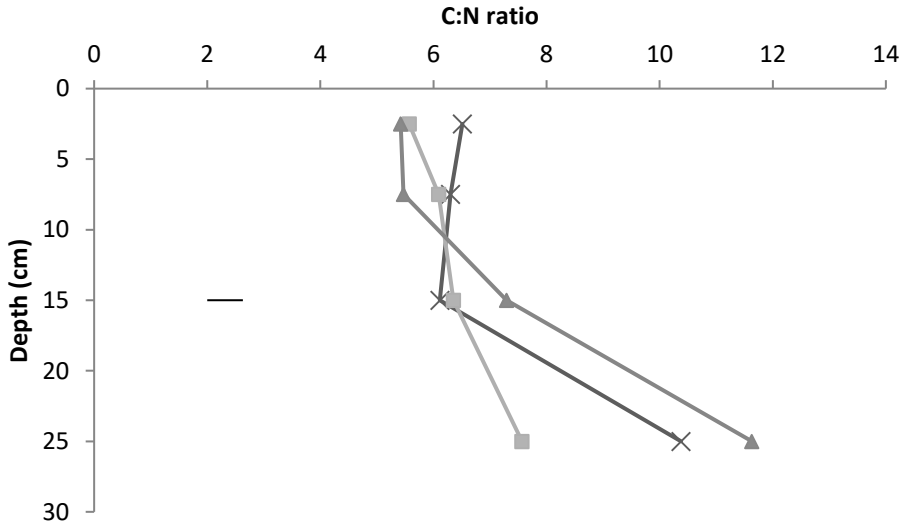


Figure 5-18. C:N ratio for the three scald sites (Site 2 ×; Site 6 ■; Site 14 ▲). Horizontal bar = SEM (n=108).

Soil properties from the scalded soil have been used to represent the baseline soil conditions prior to waterponding. To do this the results from the three scald sites have been averaged. The mean results for pH, EC, SOC, BD, SCS, TN and C:N ratio for the three scald sites are shown (Table 5-7).

Table 5-7. Results of soil properties of the scalded soil (mean of 3 sites and 9 cells per site (n=27). Figure in parenthesis is SE.

Depth (cm)	pH (1:5 water)	EC (1:5 water) (dS/m)	EC(sat) (dS/m)	SOC (g 100g^{-1})	Bulk Density (Mg m^{-3})	SCS (Mg C ha^{-1})	Total N (mg kg^{-1})	C:N ratio
0-5	7.3 (0.07)	1.8 (0.16)	17.5	0.4 (0.02)	1.5 (0.06)	3.0 (0.2)	645 (0.005)	6 (0.27)
5-10	7.2 (0.08)	2.9 (0.18)	24.8	0.4 (0.02)	1.6 (0.03)	3.4 (0.15)	651 (0.001)	6 (0.20)
10-20	7.3 (0.10)	3.1 (0.14)	26.8	0.4 (0.02)	1.5 (0.02)	6. 6 (0.25)	715 (0.007)	7 (0.38)
20-30	7.6 (0.09)	3.0 (0.11)	26.0	0.4 (0.02)	1.6 (0.03)	5.8 (0.26)	513 (0.005)	10 (0.83)

5.8.2 Soil pH changes following waterponding

Post hoc analyses with Tukey's HSD (using an α of 0.05) showed the $pH_{(1:5\ water)}$ was significantly higher in waterponded soil than the scalded soil throughout the 30 cm profile (Figure 5-19). In the surface 0-5 cm, scalds showed an average pH of 7.31 (Table 5-7**Error! Reference source not found.**) whereas the one year old waterponds were found to have a mean pH of 7.75. The oldest sites aged 25-27 years showed a mean pH in the 0-5 cm depth increment of 7.97. In the 20-30 cm depth increment the average scald pH was 7.61, whereas the age class 4 waterponds had an average pH of 8.3. The results show a trend where pH generally increases with waterpond age. ANOVA suggests that both age class and soil depth can have a significant effect on pH (eg age class F=44.987, P < 0.001).

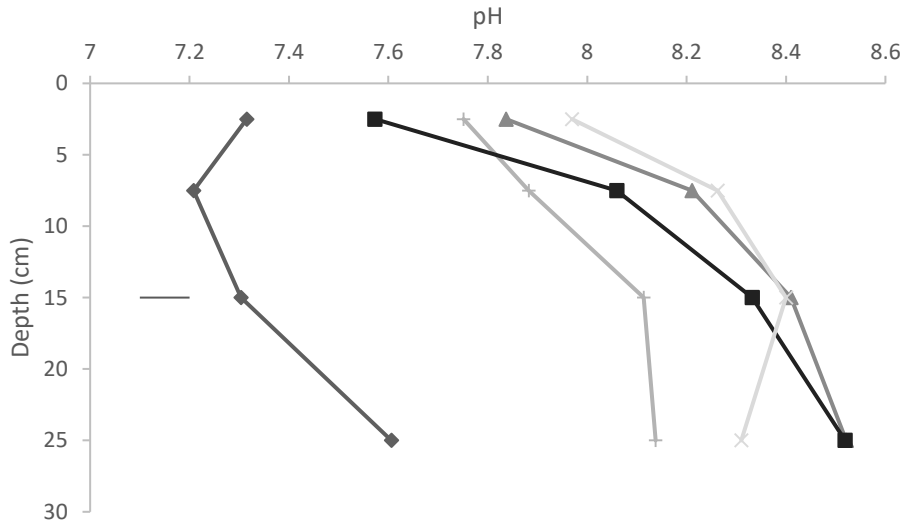


Figure 5-19. Changes to pH following waterponding (♦ scald, + 1 year, ▲ 5 year, ■ 10 year and × 25-27 year old waterponds). Horizontal bar = SEM.

5.8.3 Changes to electrical conductivity following waterponding

Post hoc analyses with Tukey's HSD (using an α of 0.05) showed the $EC_{(1:5\ water)}$ was significantly lower in waterponded soil than the scalded soil throughout the 30 cm profile (Figure 5-20). The significant difference applies also to both depth and age class (Figure 5-20). Within one year following waterpond establishment the surface $EC_{1:5}$ reduced from 1.84 dS/m to 0.27 dS/m. A similar trend followed at depths below 5cm, but the results suggest a time lag whereby the reduction in EC appears to be slower at soil layers below 5cm.

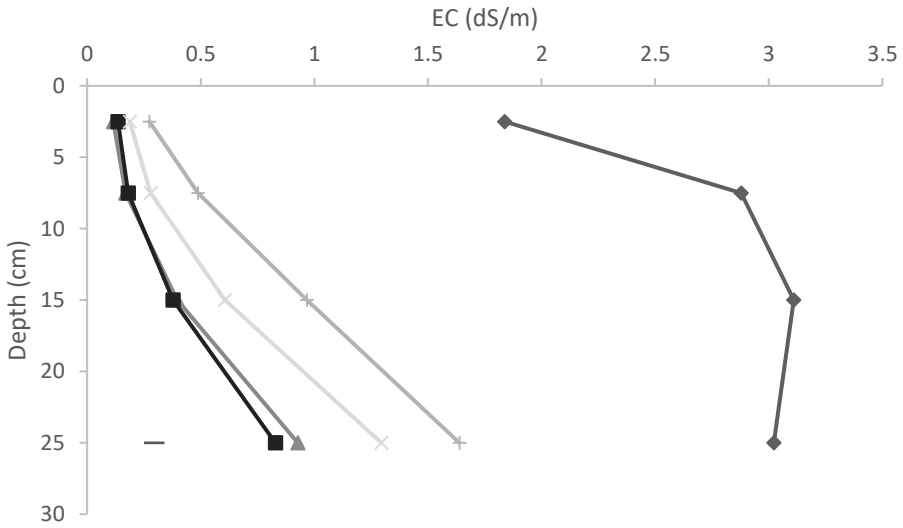


Figure 5-20. Electrical conductivity changes following waterponding (♦ scald, + 1 year, ▲ 5 year, ■ 10 year and × 25-27 year old waterponds). Horizontal bar = SEM (*n*=540)

EC results from this study and soil textures determined by Ringrose-Voase *et al.* (1989b), (namely silty loam in the top 1-3 cm and silty clay below 3 cm of the waterponds, and silt loam in the top 1.5 cm and silty clay below 1.5 cm in the scalds (see Table 5-2) have been used to calculate the EC(sat) dS/m for scalded and waterpond soils (Table 5-8). The results show that the scalds have a mean EC(sat) of 17.5 in the 0-5 cm soil layer and 26.0 in the 20-30 cm layer. As for EC, the waterponded sites have lower ECe results than the scalded sites.

Table 5-8. EC(sat) (dS/m) for scalds and waterponds (by age) determined by using EC results from this study, soil texture (Ringrose-Voase *et al*. 1989b), and EC(sat) conversion factor (Slavich and Petterson, 1993).

Depth	**Scald**		**1 year**		**5 year**		**10 years**		**25-27 years**	
cm	**EC**	**EC(sat)**	**EC**	**EC(sat)**	**EC**	**EC(sat)**	**EC**	**EC(sat)**	**EC**	**EC(sat)**
0-5	1.8	17.4	0.3	2.6	0.1	1.1	0.1	1.3	0.2	1.8
5-10	2.9	24.8	0.5	4.2	0.2	1.5	0.2	1.6	0.3	2.4
10-20	3.1	26.8	1.0	8.3	0.4	3.4	0.4	3.3	0.6	5.2
20-30	3.02	26.0	1.6	14.1	1.0	8.0	0.8	7.1	1.3	11.1

5.8.4 Changes to soil bulk density following waterponding

The within waterpond BD for each waterpond position (wall, mid and top) increases with depth (Figure 5-21) and were statistically significant ($P < 0.001$) (Table 5-9). The results include a significant difference for BD between waterpond age classes (P=0.035). No significant difference in BD was found between the three waterpond positions (P=0.627). The age class by within waterpond position interaction was not significant (P= 0.427), neither was the interaction between position and depth (P= 0.992). There was no significant difference found between the interactions of age class, within waterpond position and depth (P= 0.998).

Table 5-9. Analysis of variance for bulk density considering waterpond age class, position and depth. (n.s Not significant; * $P \leq 0.05$; *** $P \leq 0.001$)

ANOVA term	d.f.	P value	Significance
Age class	3	0.035	*
Position	2	0.627	n.s.
Depth	3	<0.001	***
Age class x position	6	0.497	n.s.
Age class x depth	9	0.874	n.s.
Position x depth	6	0.992	n.s.
Age class x position x depth	18	0.998	n.s.
Residual	384		

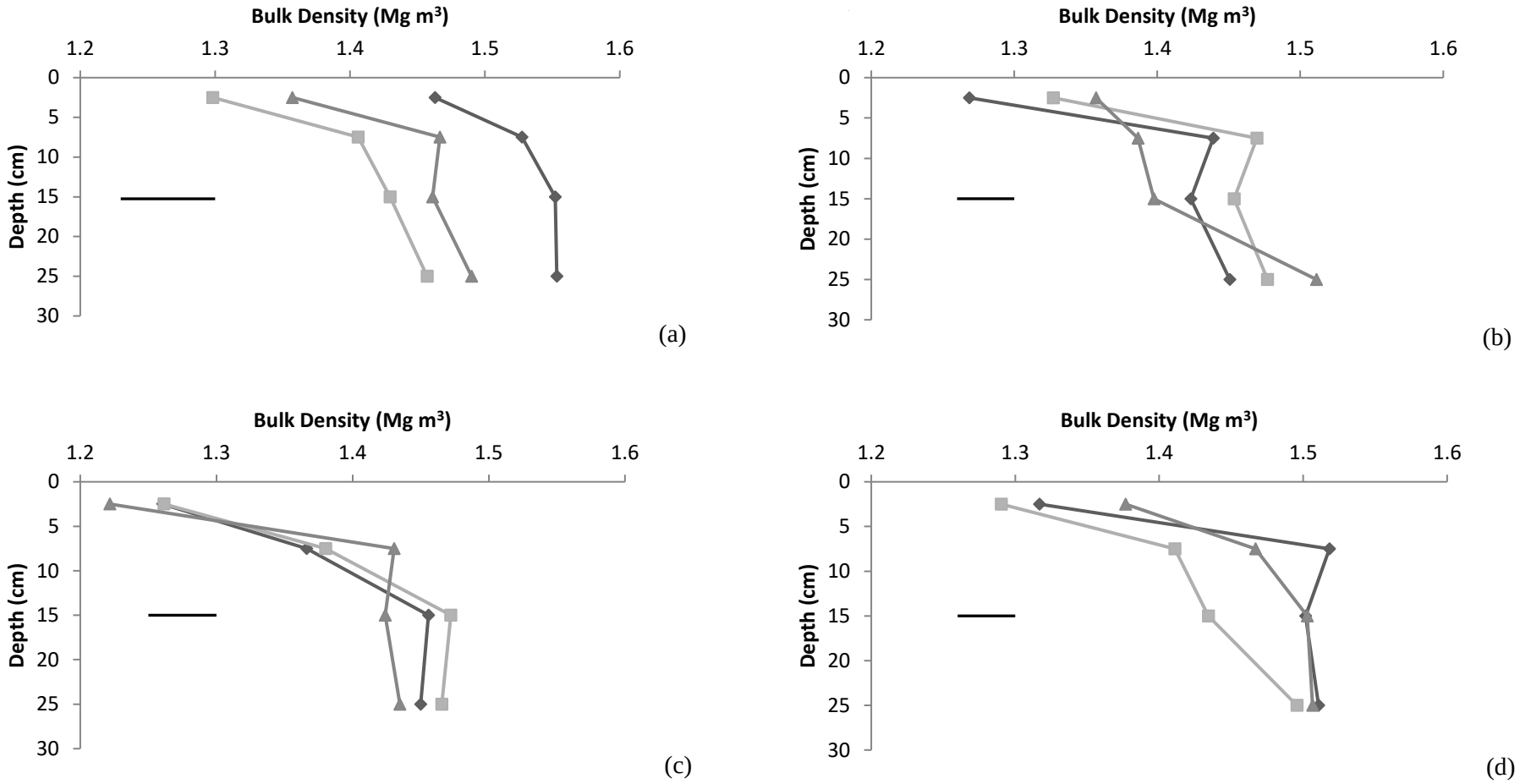


Figure 5-21. Bulk density by waterpond position at age (a) 1 year, (b) 5 years, (c) 10 years and (d) 25-27 years (♦ Wall; ■ Mid; ▲ Top positions). Horizontal bars = SEM (*n*=108).

There is a significant difference in BD between waterponds and scalds ($P < 0.001$) with waterponds lower than scalded soils (Figure 5-22). The decrease is from 1.54 Mg m^3 (mean of all depths) in the scalded soil to 1.43 Mg m^3 (mean of all age classes, positions and depths) in the waterponds. The BD is higher in the subsurface layers (5-30 cm) than the surface layer (0-5 cm) for all waterpond age classes.

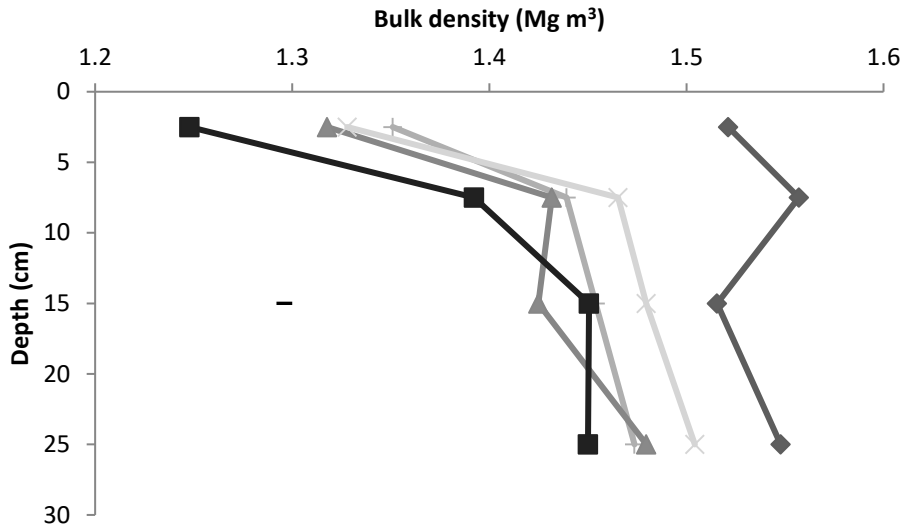


Figure 5-22. Changes to soil bulk density following waterponding (results are means of all waterpond positions) (♦ scald, + 1 year, ▲ 5 year, ■ 10 year and × 25-27 year old waterponds). Horizontal bar = SEM (*n*=540)

5.8.5 Comparison between total soil carbon $_{(LECO)}$ and soil organic carbon $_{(Heanes)}$ results

Because SOC concentration was determined using both Leco CNS2000 dry combustion analysis (for TSC) and Heanes wet oxidation (for SOC) methods a comparison of the results for both methods has been undertaken. When a comparison of the TSC results obtained by the Leco CNS2000 dry combustion method with SOC results obtained using the Heanes wet oxidation method was carried out a good correlation between the two methods was obtained (adjusted R^2 = 0.73, P<0.001), (Figure 5-23). The mean $SOC_{(Heanes)}$ concentration combining results for all depths and positions (n = 540) was 0.5 %. The mean for the same samples was 0.5 %, with the difference between $TSC_{(LECO)}$ and $SOC_{(Heanes)}$ being 0.01 % indicating that the level of soil carbonates in the samples was low. If there was some carbonate the difference between $TSC_{(LECO)}$ and $SOC_{(Heanes)}$ should be positive. The relatively low level of carbonates found is in agreement with that of Ringrose Voase *et al*. (1989) who found carbonates present only in the soil layer between 20-80cm.

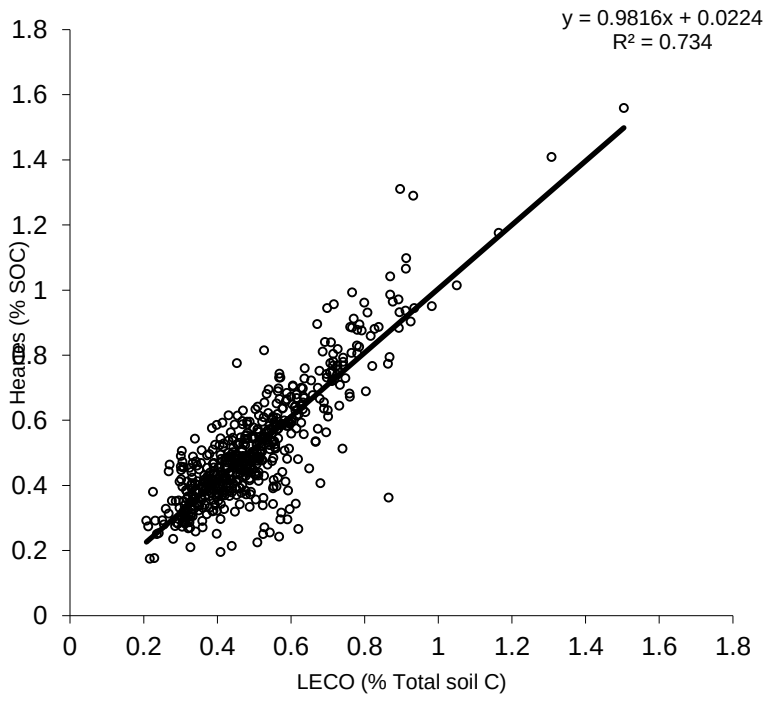


Figure 5-23. Linear regression analysis comparing results analysed for TSC $_{(LECO)}$ with SOC $_{(Heanes)}$ methods.

Cl^- is the most likely anion present in the scalded soil. The presence of Cl^- is reported to interfere with analysis methods that use dichromate such as the Heanes method (Nelson and Sommers (1996) leading to a positive error. Consequently regression analysis comparing $TSC_{(Leco)}$ results with $SOC_{(Heanes)}$ results has been undertaken (Figure 5-24). The analysis shows a very weak R^2 and provides no indication of a positive error bias for scald samples analysed using the Heanes (1984) method.

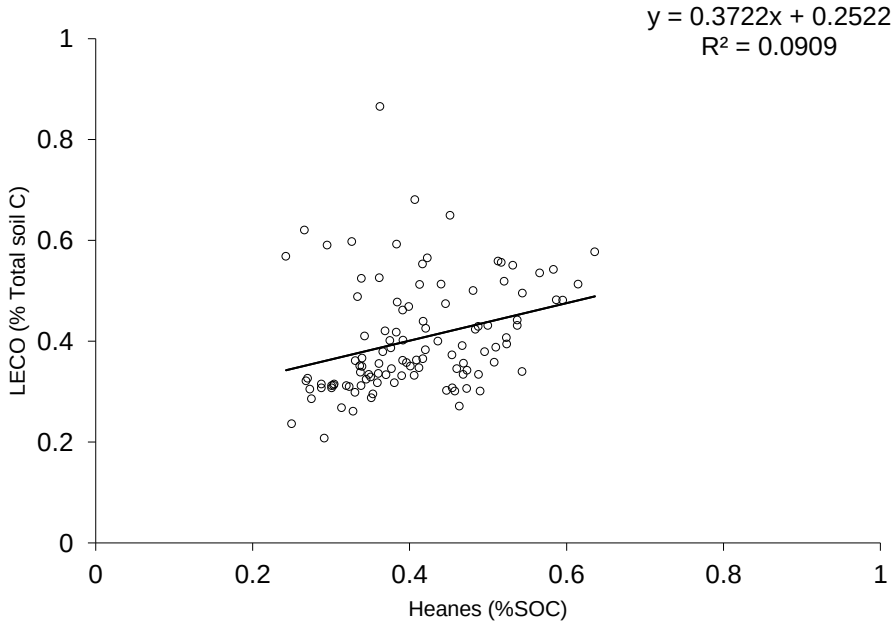


Figure 5-24. Linear regression analysis comparing results analysed for TSC (LECO) with SOC (Heanes) methods for the scald only

5.8.6 Soil organic carbon concentration in waterponds

ANOVA of the waterpond data (Table 5-10) has indicated that there was a significant difference in SOC concentration for both waterpond age (P<0.001) and depth (P<0.001). One year old waterponds had a mean SOC concentration for the four depth increments of 0.45 g 100g^{-1}. There was a significant increase in SOC from the one year old waterponds compared with the SOC for waterponds aged 5, 10 and 25-27 years (0.58, 0.57 and 0.57 g 100g^{-1} respectively). As expected the SOC concentration was highest for the 0-5 cm depth increment and decreased with depth for all waterpond age cohorts.

Table 5-10. Analysis of variance results for SOC considering waterpond age class, position and depth. n.s = not significant; * P≤0.05; ***P≤0.001

ANOVA term	d.f.	P value	Significance
Age class	3	<0.001	***
Position	2	0.837	n.s.
Depth	3	<0.001	***
Age class x position	6	0.701	n.s.
Age class x depth	9	0.039	*
Position x depth	6	0.811	n.s.
Age class x position x depth	18	0.999	n.s.
Residual	384		

The distribution of SOC concentration for the three waterpond positions (wall, mid and top) across the four age classes (Figure 5-25) and the result combining waterpond position by depth gave a mean of 0.55 g 100 g^{-1} at the wall position, 0.54 g 100 g^{-1} at the mid position and 0.53 g 100 g^{-1} at the top position. There was no significant difference between the three positions. However, the interaction between age class and depth was statistically significant (P=0.039) with the greatest increase achieved by the one year and five year old waterponds particularly in the 0-5 cm depth layer and wall position.

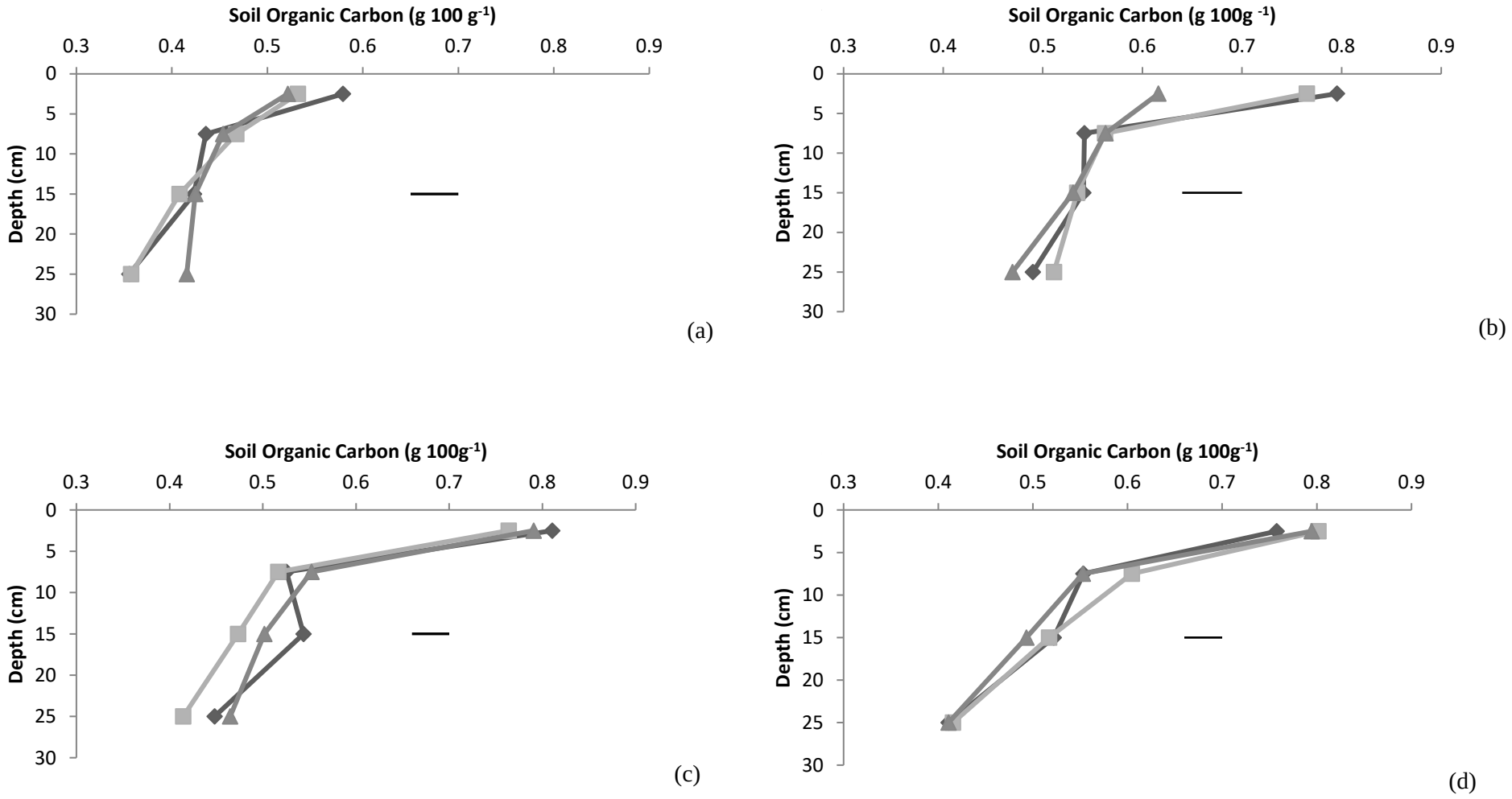


Figure 5-25. Within waterpond position trends in SOC (%) (Heanes) change to 30 cm by age a) 1 year, b) 5 years, c) 10 years, d) 25-27 years (♦ Wall; ■ mid; ▲ top position).

5.8.7 SOC concentration comparing waterponds and scalds

The average SOC concentration for scalds and waterponds from each age class in the four depth increments is shown (Figure 5-26). In the 0-5 cm layer scalds have the lowest SOC concentration. The SOC concentration increases with waterpond age notably in the 0-5 cm layer. However, there is little difference between waterponds aged 10 years and those aged 25-27 years. The temporal trend is less clear in soil layers below 5 cm.

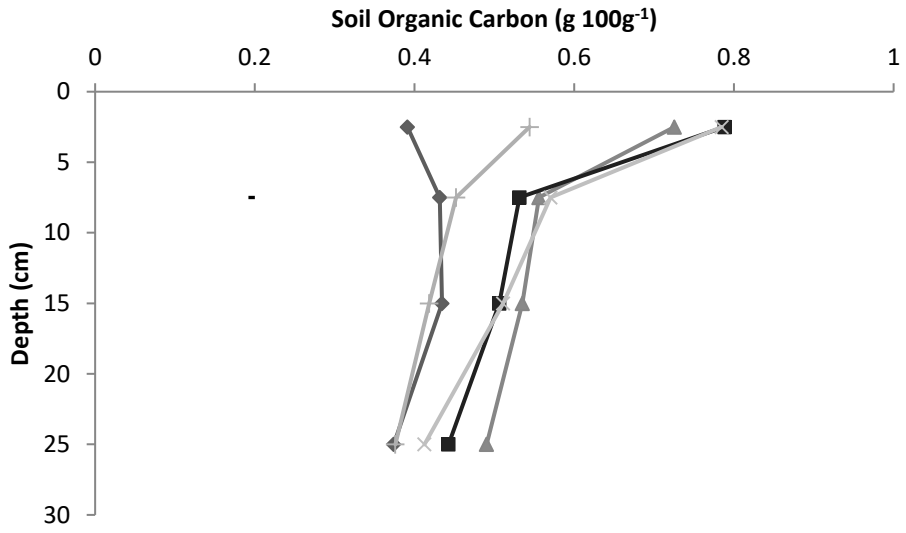


Figure 5-26. Variation in SOC concentration by age-class (♦ scald, + 1 year, ▲ 5 year, ■ 10 year and × 25-27 year old waterponds). Horizontal bar = SEM.

When results for scald and waterpond SOC concentration are compared there is a significant difference for age class, depth and the interaction between age class and depth (P <0.001 respectively). There was a significantly higher SOC concentration in soil to 0.3m depth in waterponds older than five years when compared with scalds and one year old waterponds. Respective mean values for SOC to 0.3 m is shown in Table 5-11.

Table 5-11. Mean SOC values comparing scalds and waterponds. (Means followed by the same letter are not significantly different at P= 0.05)

Age (years)	Mean SOC (%) to 0.3 m
Scald	0.4a
1	0.5a
5	0.6b
10	0.6b
25-27	0.6b

ANOVA results suggest that depth has a significant effect with the highest mean SOC concentration in the 0-5 cm depth (0.7 g C 100g^{-1}). The mean concentration decreases slightly with depth with 0.51 g C 100 g^{-1} at 5-10 cm, 0.48 g C 100 g^{-1} at 10-20 cm and 0.42 g C 100 g^{-1} in the 20-30 cm depth increment.

The age class by depth interaction indicates that the increase in SOC was most strongly expressed in the 0-5 cm depth (Table 5-12). Within one year following the establishment of waterponds there is a significant increase in SOC concentration for the 0-5 cm depth when compared with the mean scald result. The results also indicate that the increasing trend continues until waterponds are five years old. There then appears to be a stabilisation in SOC in the oldest waterpond sites with results showing an average concentration of 0.8% in the oldest waterponds in the 0-5 cm depth.

Table 5-12. ANOVA results suggesting mean SOC concentration (%) results for the interaction between age class and depth. (Means followed by the same letter are not significantly different at P=0.05)

Depth (cm)	Scald	1 year	5 years	10 years	25-27 years
0-5	0.39 a	0.54 bcd	0.73 e	0.79 e	0.78 e
5-10	0.43 abcd	0.46 abcd	0.56 cd	0.53 bcd	0.57 d
10-20	0.43 abcd	0.42 abc	0.54 bcd	0.51 abcd	0.51 abcd
20-30	0.37 a	0.38 a	0.49 abcd	0.44 abcd	0.41 ab

Waterponds show an increase in SOC over time with the most notable increase in the 0-5 cm depth increment. The rate of SOC increase at depths below 5 cm is slower than in the surface 0-5 cm.

5.8.8 Within waterpond soil carbon stock distribution

The mean SCS by depth increment for the three within waterpond positions shows there is no significant difference for results both for fixed depth and ESM calculations. There was also no significant difference in SCS to 0.3 m depth between within waterpond positions on a fixed depth or ESM basis (Table 5-13).

Table 5-13. Within waterpond variation in SCS (mean of all age classes).

Position	**Mean SCS (Mg C ha^{-1}) per depth increment**	**Mean SCS (Mg C ha^{-1}) per depth increment (ESM)**	**Mean SCS (Mg C ha^{-1}) to 0.3m (fixed depth)**	**Mean SCS (Mg C ha^{-1}) to 0.3m (ESM)**
Wall	5.6	6.1	22.4	24.5
Mid	5.5	6.1	21.5	24.3
Top	5.5	6.0	21.8	24.1

The results for SCS to 0.3 m depth calculated on an ESM basis (Table 5-14) shown a significant difference for SCS between age classes and with depth (P<0.001). The one year old waterponds had a significantly lower mean SCS than the other age classes (at $P < 0.05$) (Table 5-14). There was no significant difference in mean SCS between waterponds aged 5, 10 and 25-27 years.

Table 5-14. Within waterpond SCS by depth increment and to 0.3 m. Means followed by the same letter are not significantly different at P=0.05.

Waterpond age	Mean CS (Mg C ha^{-1}) per depth increment	Mean CS (Mg C ha^{-1}) to 0.3 m
1	4.7a	18.7a
5	5.6b	22.3b
10	5.8b	23.1b
25-27	5.9b	23.5b

5.8.9 Changes to carbon stock following waterponding (fixed depth calculation)

A comparison between the SCS of the scalds and the three within waterpond positions by waterpond age (Figure 5-27) shows scalds have a mean SCS to 30 cm of 18.7 Mg C ha^{-1}. This SCS result have been calculated using a fixed depth method (Equation 15). There is no significant difference found for SCS between the three within waterpond positions for any of the four age cohorts. There is also no significant difference for SCS between scalds and waterponds aged one year. However, the results show a significant difference between scalds and waterponds aged one year, with waterponds aged five years. The mean within waterpond value of five year old waterponds is 23.1 Mg C ha^{-1} representing a mean SCS increase of 4.4 Mg C ha^{-1}. The fixed depth results also show the SCS appears to plateau at the five year point with no further increase in SCS apparent in waterponds older than five years.

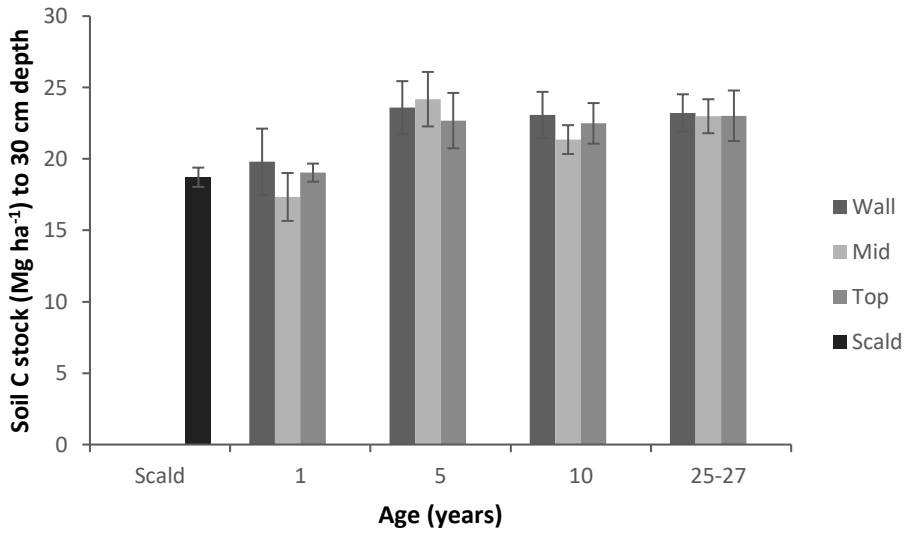


Figure 5-27. Within waterpond SCS to 30 cm depth, comparing scalds and waterponds based on a fixed depth calculation (error bars = SEM).

5.8.10 Soil carbon stock change after waterponding using an equivalent soil mass formula

Changes to SCS following waterponding have also been calculated by using the ESM formula (Equation 17). The ESM SCS has been calculated by using the mean scald soil mass results as the reference soil mass.

The ESM calculation has resulted in larger changes to the SCS (Figure 5-28) in waterponds for the 0-30 cm soil layer than the SCS results obtained using the fixed depth method (Figure 5-27). Linear regression comparing SCS calculated on a fixed depth basis and an ESM basis was R^2 = 0.81, slope – 0.95, and P <0.001.

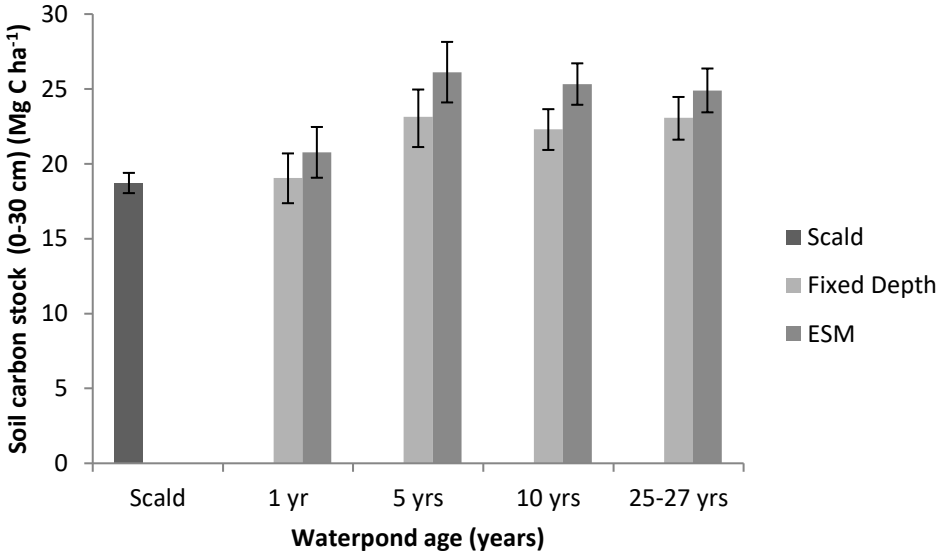


Figure 5-28. SCS (0-30 cm depth) comparing ESM results with fixed depth results. Error bars = SEM.

The mean SCS in the 0-30 cm soil layer for the within waterpond positions across all age groups has been found to be not significantly different ($P = 0.95$). The within waterpond SCS means were 24.5 Mg C ha^{-1} for the wall, 24.3 Mg C ha^{-1} for the mid and 24.12 Mg C ha^{-1} for the top position (Figure 5-29).

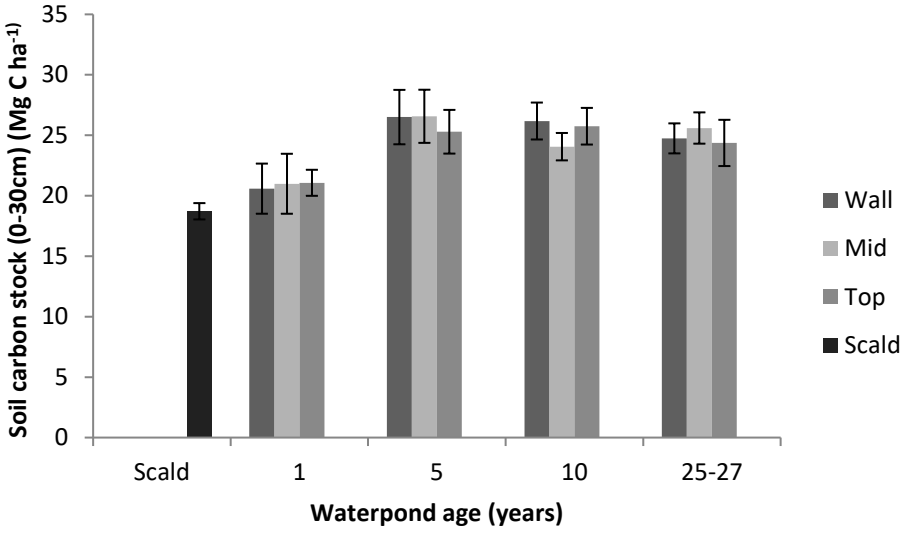


Figure 5-29. Within waterpond position SCS (0-30 cm) showing temporal change to SCS in scalds following waterponding using the equivalent soil mass equation (error bars = SEM).

The SCS of the three within waterpond positions has been averaged to calculate a mean waterpond result. Tukey multiple comparison analysis shows that the mean SCS in the scalds and waterponds aged 1 year on the one hand are significantly different ($P = 0.05$) to waterponds aged 5, 10 and 25-27 years (Table 5-15) on the other. When the SCS in scalds is compared with the three within waterpond positions and waterponds aged 5, 10 and 25-27 years, the scald SCS was found to be significantly lower ($P < 0.001$) (Table 5-16). The highest SCS change is seen after five years with an overall increase of 7.4 Mg C ha^{-1}. There is a slight, but not statistically significant, SCS decrease in waterponds aged 10 and 25-27 years from 26.1 Mg C ha^{-1}to 25.0 Mg C ha^{-1}.

Table 5-15. Mean SCS (Mg C ha^{-1}) to 30 cm depth for scalds and waterponds by age. The results show a significant change following waterponding. Means followed by the same letter are not significantly different at P=0.05.

Scald	*1 year*	*5 years*	*10 years*	*25-27 years*
18.72_a	20.9_a	26.1_b	25.3_b	24.9_b

The SCS (ESM) results have been analysed to identify which depth increments achieve the highest SCS over time (Figure 5-30). Most increase occurred in the 0-5 cm depth layer. The greatest mean increase in SCS of 3.1 Mg C ha^{-1} occurred at the 0-5 cm depth interval between the scalds and the 10 year old waterponds. This represents an increase of 100% in just 10 years. There was also an increase of 2.1 Mg C ha^{-1} in the 20-30 cm layer of the 5 year old waterponds. This represents a 73 % increase. ANOVA suggests that both depth and age are significantly different (P=0.001) but their interaction is not significant (P=0.21).

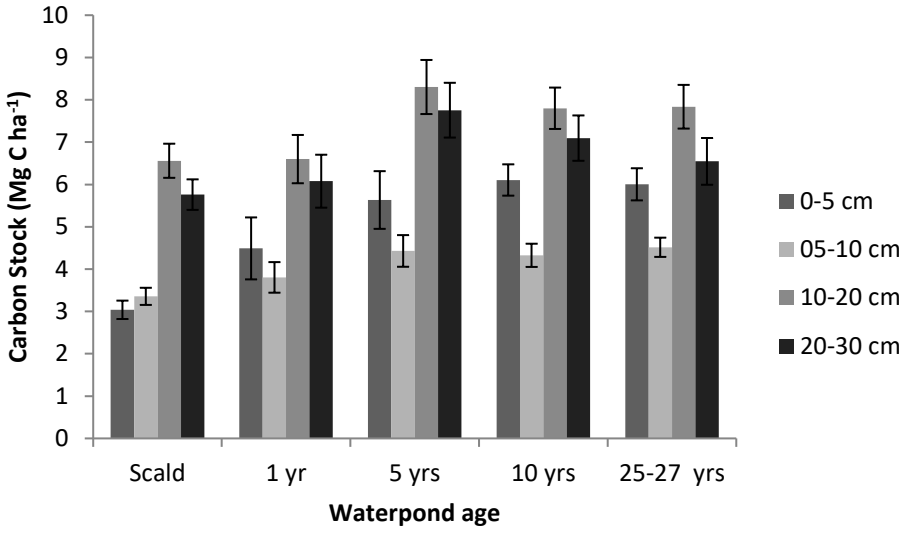


Figure 5-30. Changes to SCS (ESM) by soil depth layer for each waterpond age class. Error bars =SEM.

When the SCS results are separated into depth increments it is apparent that an increasing trend applies to most of the individual depth increments for the first five years (Table 5-16). After five years te SCS is the same or slightly lower in all depth increments. Specifically, the 0-5 cm soil layer shows SCS increases until 10 years. The 5-10 cm depth increment shows a slightly increasing trend over time. There is some variation in results for the 10-20 cm and 20-30 depth increments with the highest result achieved in waterponds aged five years but then slightly decreasing in the older waterponds.

Table 5-16. Mean SCS (Mg C ha-1) differences between depths and waterpond age. Means followed by the same letter are not significantly different at P=0.05.

Depth (cm)	1 year	5 years	10 years	25-27 years
0-5	4.49ab	5.63 bcd	6.11d	6.00cd
5-10	3.72a	4.43ab	4.33ab	4.52abc
10-20	6.00de	8.30f	7.80ef	7.84ef
20-30	6.08d	7.76ef	7.10def	6.55de

5.8.11 Rate of change and relative rate of change to soil carbon stock change following waterponding

Using the mean SCS in the scalds as a baseline Equation 5 has been used to calculate the rate of change in SCS following waterponding. Most change in SCS following waterponding occurs in the 0-5 cm soil layer with the results (Figure 5-31) showing a mean SCS rate of change is an increase for all depth increments and waterpond ages. In the 0-5 cm depth, the mean rate of change is 0.6 Mg C ha^{-1}whereas the mean rate of change below 5 cm is between 0.1 and 0.2 Mg C ha^{-1}. The one year old waterponds have the highest rate of SCS increase of 1.5 Mg C ha^{-1} in the 0-5 cm soil layer, and between 0.4 and 0.04 Mg C ha^{-1} below 5 cm. The 25-27 year old waterponds have the lowest rate of change at 0.1 Mg C ha^{-1} increase in the 0-5 cm soil layer. The five year old waterponds have the largest rate of change co SCS below 5 cm with 0.2 Mg C ha^{-1} increase in the 5-10 cm layer and 0.4 Mg C ha^{-1} in the 20-30 cm depth. The average rate of change for the 0-30 cm layer is 1.5 Mg C ha^{-1} yr^{-1} in the first five years after waterponding.

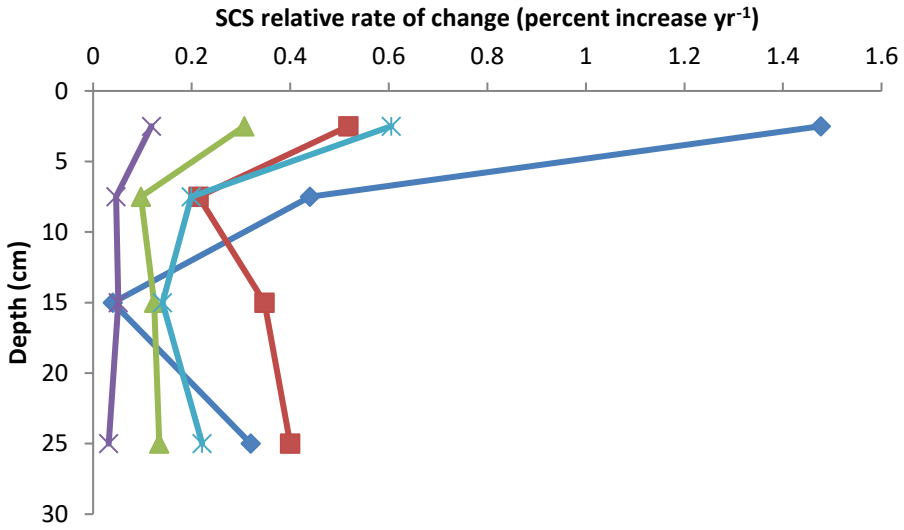


Figure 5-31. Estimated annual rate of SCS change (Mg C ha^{-1} yr^{-1}) depth increment (♦ = 1 yr; ■ = 5 yrs; ▲ = 10 yrs; × = 25-27 yrs; ✳ = mean).

The relative rate of change from scald to waterpond (by age) has been calculated using Equation 7. The highest relative rate of change of 47% occurs in the 0-5 cm layer of the 1 year old waterponds (Figure 5-32. The 5 year old waterponds have a mean relative rate of change of 17% per year. The lowest relative rate of change occurs in the 20-30 cm layer of between 7% and 0.05% per year. On average the relative rate of change of SCS following waterponding in the 0-5 cm layer is 20% per year, and 4% rate of change per year in the 20-30 cm layer.

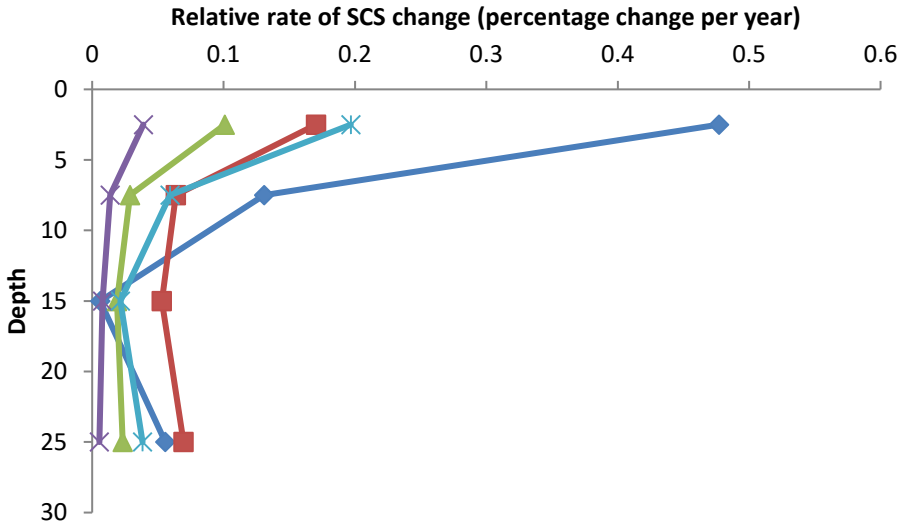


Figure 5-32. Relative rate of SCS change per year (♦ = 1 yr; ■ = 5 yrs; ▲ = 10 yrs; × = 25-27 yrs; ✳ = mean)

5.8.12 Total nitrogen

From the results for *n* = 540 samples depicted in Figure 5-33 it can be deduced that one year following waterponding there is a decrease in TN concentration, probably associated with N leaching and volatilisation following water infiltration into the soil profile. However, TN results for the waterponds older than 5 years, suggest that there is a gradual temporal increase at the surface. The results suggest it takes longer below the surface 0-5 cm for the TN concentration to increase. ANOVA main effects results suggest that there is no significant difference in TN between scalds and waterponds (P = 0.781). However changes with age and depth were both found to be significant (*P*-values <0.001). The interactions between TN in the scalded soil, waterponds, age and depth were significant (*P* < 0.001 and F statistic = 4.93).

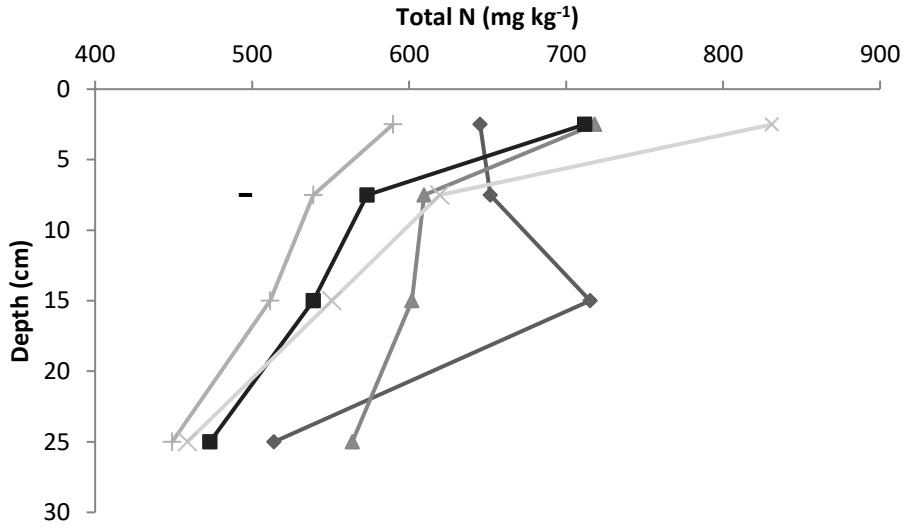


Figure 5-33. Total N results for scalds and waterponds from 4 age classes and 4 depth increments (♦ scald, + 1 year, ▲ 5 year, ■ 10 year and × 25-27 year old waterponds).

5.8.13 Total phosphorus, total sulfur and available phosphorus

Analysis for TP, total sulfur (TS) and available phosphorus (AP) was undertaken on a subset of samples (*n* = 144) including 1 scald site (9 cells), and waterponds from three age classes: 1 year, 5 year and 25-27 years. The results shown (Table 5-17) are the mean of three waterponds and the three within waterpond positions for each age class.

Table 5-17. Mean ± SEM for TSC, SOC, TN, TP, TS, and their ratios (using SOC) for the subset of *n* = 144 samples.

Scald / Waterpond age (yrs)	*TSC (mg kg^{-1})*	*SOC (mg kg^{-1})*	*TN (mg kg^{-1})*	*TP (mg kg^{-1})*	*TS (mg kg^{-1})*	*AP (mg kg^{-1})*	*SOC:TN ratio*	*SOC:TP ratio*	*SOC:AP ratio*	*SOC:TS ratio*
					0-5 cm					
Scald	3 880±0.04	4 556±0.03	700±0.02	157±43.6	73±10.3	19±3.5	8	48	297	82
1	3 937±0.03	4 621±0.06	488±0.01	76±10.7	58± 4.1	8±0.6	9	64	586	80
5	9 715±0.06	10 255±0.06	994±0.02	147±16.1	86± 9. 5	11±3.4	10	72	1055	122
25-27	7 059±0.04	8 003±0.04	901±0.02	118±28.1	90±16.9	7±2.3	9	99	1241	104
					5-10 cm					
Scald	3 869±0.03	3 921±0.02	610±0.01	98±30.0	226±44.3	11±1.8	8	108	515	28
1	4 050±0.02	3 855±0.03	505±0.01	22± 5.0	106±19.2	4±0.3	8	711	988	42
5	6 079±0.04	6 525±0.04	714±0.01	57± 8.5	73± 3.3	5±1.6	9.	133	1434	89
25-27	4 857±0.03	5 771±0.04	648±0.01	60±29.9	112±35.1	3±0.1	9	266	2539	69
					10-20 cm					
Scald	3 919±0.03	4 045±0.03	881±0.03	98±21.6	253±31.5	9±1.3	7	73	593	21
1	4 179±0.02	3 639±0.03	477±0.01	11± 2.2	202±39.2	3±0.3	7	294	1173	24
5	6 240±0.04	6 472±0.04	725±0.01	36± 7.6	106±17.6	4±1.2	9	290	2100	68
25-27	4 710±0.03	5 317±0.04	571±0.01	46±31.3	136±39.7	2±0.8	9	1075	2309	52
					20-30 cm					
Scald	3 891±0.04	4 635±0.03	409±0.01	94±19.6	253±31.5	7±0.6	10	60	588	17
1	4 915±0.03	3 057±0.04	400±0.01	18± 2.6	330±46.3	3±0.3	8	174	982	11
5	6 060±0.04	5 986±0.04	712±0.01	27± 8.7	123±10.8	3±1.0	8	65	2884	54
25-27	4 575±0.03	4 886±0.04	499±0.01	49±31.9	243±48.6	3±0.9	10	466	2074	27

The results for TP (Figure 5-34) show that the scald has a higher concentration than any of the waterpond sites analysed. The one year old waterpond has the lowest concentration. However, there was no apparent trend in TP concentration with age. ANOVA results suggest a significant difference in TP concentration for age class and depth ($P < 0.001$ for both comparisons).

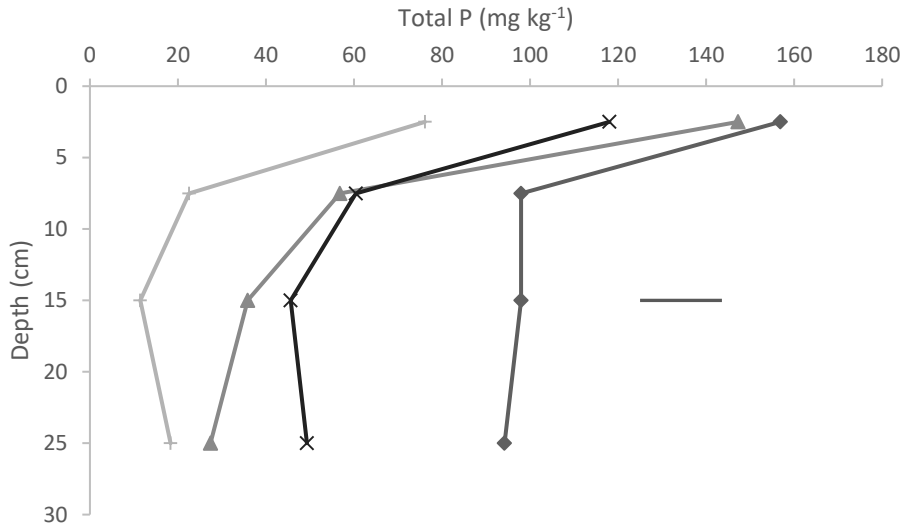


Figure 5-34. TP (mg kg^{-1}) in scalded soils and waterponds, showing a lower concentration in waterponds when compared with scalded soils (♦ scald, + 1 year, ▲ 5 year, and × 25-27 year old waterponds). Horizontal bar = SEM (n=144).

5.8.14 Total sulphur

There is considerable variation in TS between ages and depths with no particular trend apparent (Figure 5-35). The interaction between waterpond and scald, age and depth was found to be significant (P <0.001). There is no apparent trend in TS between scalds and waterponds in the 0-5 cm depth increment. There is a trend showing that the concentration of TS increases with depth in both scalds and waterponds. In the soil depth increments below 5 cm the concentration of TS is greatest in the scalded soils except below 20 cm, where the 1 year old waterpond has a higher concentration. The main effects analysis shows that site, age class and depth are significant (P <0.001, 0.003 and P<0.001 respectively).

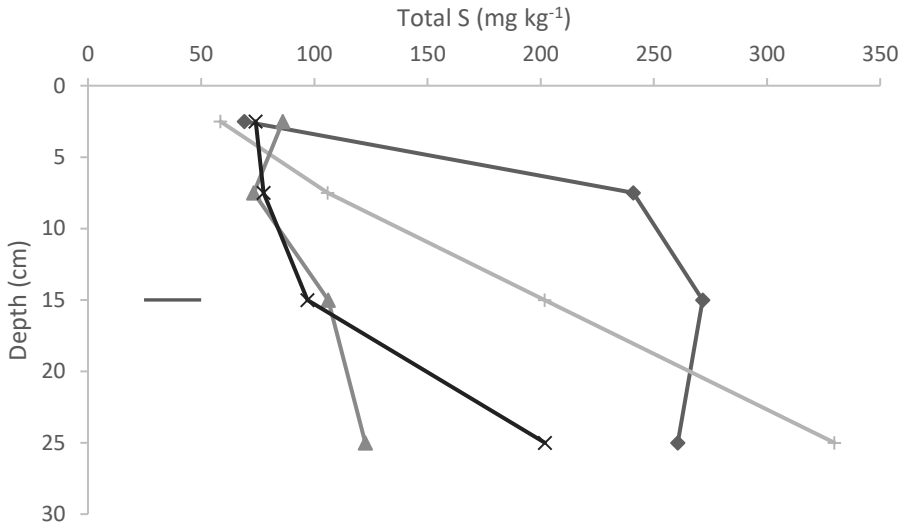


Figure 5-35. The concentration of TS showing that scalded soils have higher concentrations in soil depths below 5 cm than waterpond soils. (♦ scald, + 1 year, ▲ 5 year, and × 25-27 year old waterponds). Horizontal bar = SEM (*n*=144)

5.8.15 Available phosphorus

The scalded soil has the highest concentration of AP in all depth increments (Figure 5-36). In the 0-5 cm depth increment the scald has an AP concentration of 19.5 mg kg^{-1}. Within one year following waterponding AP levels are reduced to 8.2 mg kg^{-1} in the 0-5 cm depth. There is a commensurate decrease in AP concentration with decreasing depth increment and with age of waterpond. The ANOVA results suggest that the interaction between waterpond and scald, age class and depth are significantly different (P = 0.001), with main effects for site, age and depth P = 0.001, 0.104 and 0.001 respectively.

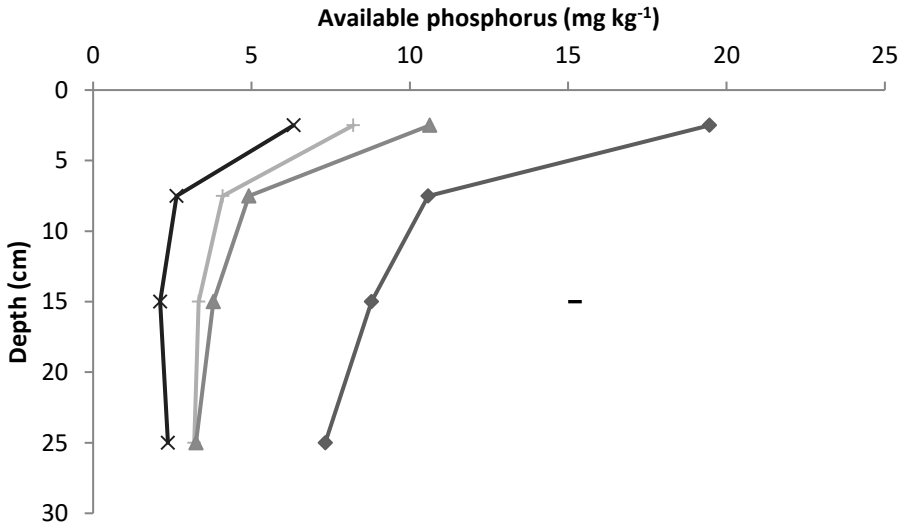


Figure 5-36. Changes to AP following waterponding (subset of three waterpond sites and one scald site). (♦ scald, + 1 year, ▲ 5 year, and × 25-27 year old waterponds). Horizontal bar = SEM (*n*=144)

5.8.16 Carbon to nitrogen, C:P and C:S ratios

Linear regression analysis for $SOC_{(Heanes)}$ and $TN_{(LECO)}$ incorporating the results from all positions, ages and depths was found to have an adjusted R^2 of 0.349, $P<0.001$. The $SOC_{(Heanes)}$ to $TN_{(LECO)}$ (C:N) ratios for scalds and waterponds (Figure 5-37), (n = 540). The differences for C:N ratios with age were significant (F=41.389, P=0.001), and the difference with depth was also found to be significant (F=4.451, P=0.004). The interaction between site (both scald and waterpond), age and depth was found to be significant (F=2.205, P = 0.011). Waterponds from all ages showed an increase in C:N ratio in the top 0-20 cm when compared with scalds. The 20-30 cm increment showed a gradual, but inconsistent change with time since waterponding with scalds having a C:N ratio of 8, 10 year old waterponds showing soil C:N ratio of 9.5, whereas the ratio for the 25-27 year old waterponds is slightly lower at 9. Post hoc analyses with Tukey's HSD (using an α of 0.05) showed the mean C:N ratio was significantly lower in scalded soil than in waterponds aged 1 year. Post hoc analysis also found that the mean C:N ratio for the 1 year old waterponds was significantly lower than for the older cohorts.

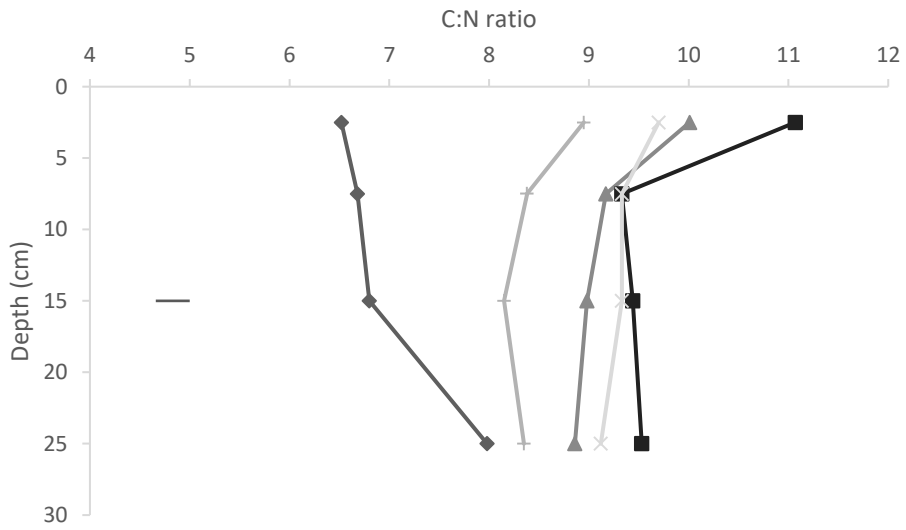


Figure 5-37. SOC:TN ratio showing a significant difference between scalded soil and waterponds (n = 540) (♦ scald, + 1 year, ▲ 5 year, ■ 10 year and × 25-27 year old waterponds). Horizontal bar = SEM (n=540).

The scalds had a significantly lower C:AP ratio than the waterponds, and there was a significant temporal increase with time ($P < 0.001$). The trend with depth shows a general increase with depth to 20 cm except in the oldest waterponds where the ratio is highest in the 5–10 cm layer. ANOVA results imply that the interaction between site (waterpond and scald), depth and age is significant ($P < 0.001$). The main effects for site, age and depth was also suggested to be significant at $P < 0.001$ for each factor.

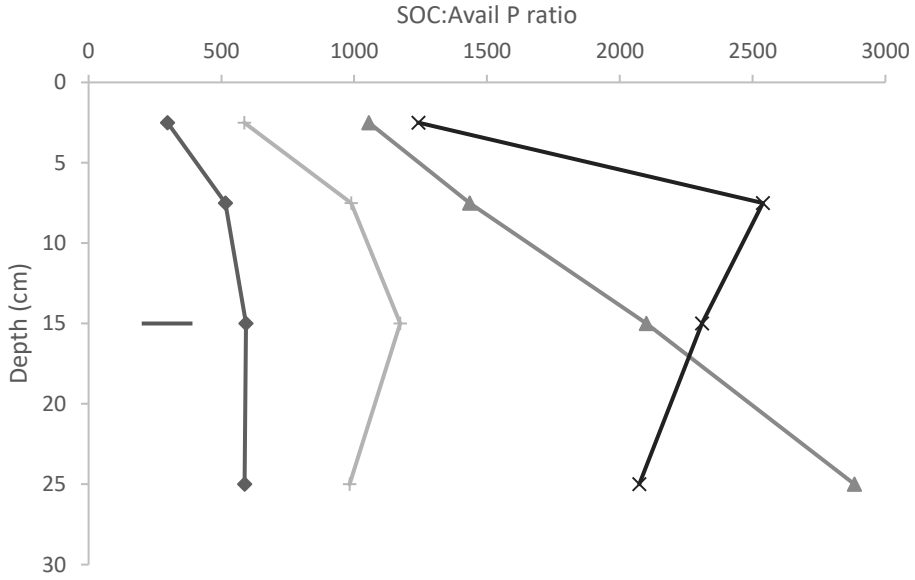


Figure 5-38. The SOC:Avail P ratio shows a commensurate widening of the ratio with the decrease (uptake) of avail P over time (♦ scald, + 1 year, ▲ 5 year, and × 25-27 year old waterponds). Horizontal bar = SEM (n = 144)

The SOC:TP (C:P) ratio depicted in Figure 5-39 for the subset of $n = 144$ samples shows that there is a shift from the scald which has the narrowest C:P ratio compared with the waterponds. There is variability with depth in the waterponds whereas the scalds have a more even distribution of this ratio down through the soil profile. Some outliers with very high ratios have affected the results and imply large increases in this ratio below 5 cm depth. ANOVA results indicate that there was no significant effect for site (scald and waterpond), age and depth on the C:P concentration (P values = 0.03, 0.172 and 0.199 respectively). The interaction between the factors was indicated to be not significant ($P = 0.211$).

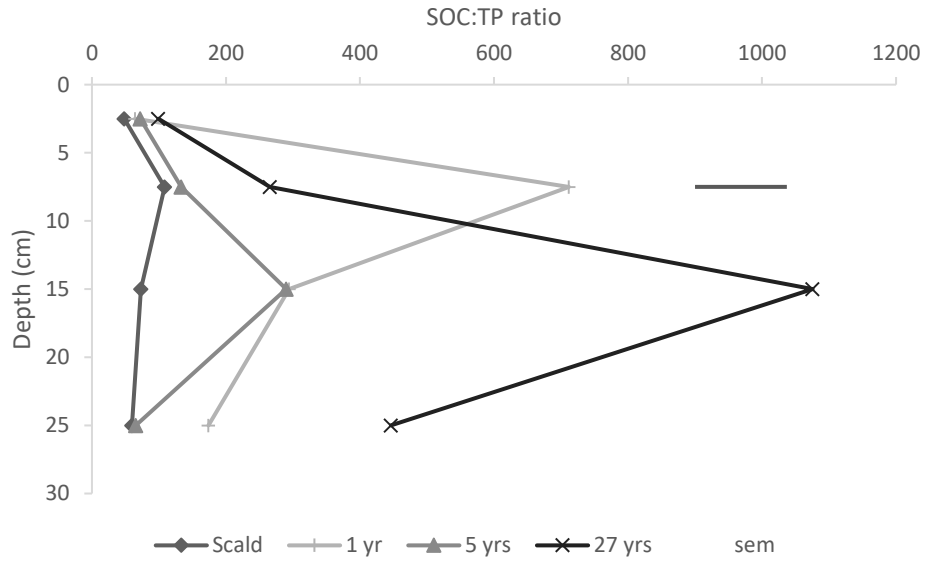


Figure 5-39. C:TP ratio showing waterponds have a wider ratio than scalds ($n = 144$) (♦ scald, + 1 year, ▲ 5 year, and × 25-27 year old waterponds). Horizontal bar = SEM.

There is a significant change ($P < 0.001$) in the SOC to Total S ratio (C:S) after waterponding with the results showing the scalds have the lowest C:S ratio at 82 in the 0-5 cm layer (subset $n = 144$). The main effects model shows a significant difference ($P < 0.001$) between the scalds and the waterpond age classes and for depth. The widest ratio of 122 is achieved in waterponds aged 5 years. The oldest waterponds (25-27 years) have a C:S ratio of 103. In all cases the ratio is lower with depth, however the depth by age class interaction is not significant ($P = 0.83$).

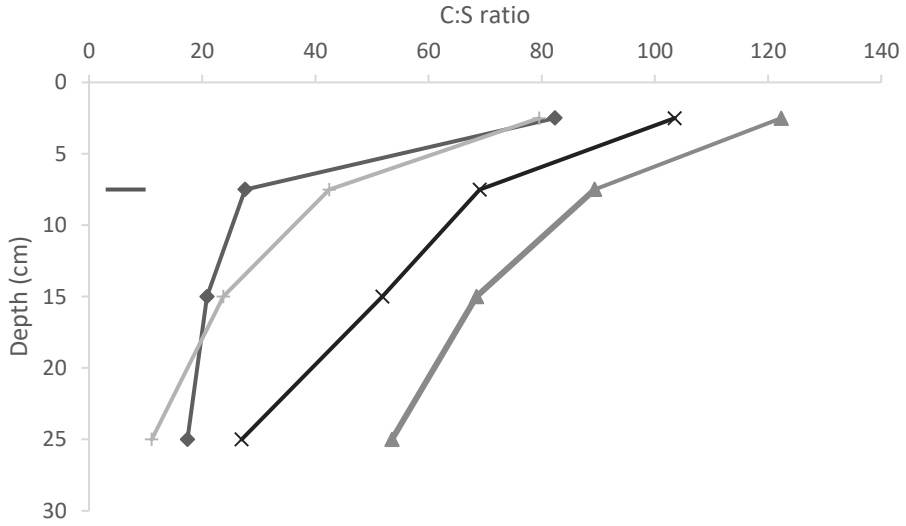


Figure 5-40. C:S ratio showing waterponds have a higher ratio than scalded soil ($n = 144$) (♦ scald, + 1 year, ▲ 5 year, and × 25-27 year old waterponds). Horizontal bar = SEM.

The correlation matrix (Table 5-18) shows strong correlations between C:N and C:AP. ($R^2 = 0.595$, $P<0.001$ and $R^2 = 0.624$, P<0.001 respectively).

Table 5-18. Correlation matrix showing R^2 relationships among nutrients in the 0-5 cm layer incorporating both scald and waterpond samples from subset 1 (n=36). Results with an $R^2 > 3$ shown in bold. P values as follows: * = P<0.05, ** = P<0.01, ***P<0.001.

	SOC (%)	***TN (%)***	***TS (%)***	***AP (%)***	***TP (%)***
SOC (%)	1				
TN (%)	**0.594*****	1			
TS (%)	**0.401****	**0.422****	1		
AP (%)	-0.159	-0.080	**-0.363****	1	
TP (%)	0.296*	**0.348***	0.244	**0.624*****	1

5.8.17 Exchangeable sodium percentage

The results (Table 5-19) show the waterpond soil has a higher ESP than the scalded soil in the 0-5 cm depth increment. However, at depths below 5 cm the scald has a higher ESP than the waterpond. The ESP increases with depth in the scalded soil from 12.9 in the 0-5 cm to 32.1 in the 20-30 cm depth increment. Conversely, the waterpond has a higher ESP at the surface and a decreasing trend with depth. These values can be used to determine the dispersive potential of the soils by using the model proposed by Rengasamy *et al*. (1984).

Table 5-19. Exchangeable sodium percentage (ESP) (%) for one scald core and one 27 year old waterpond core.

Depth (cm)	*Scald ESP (%)*	*Waterpond ESP (%)*
0-5	12.9	22.1
5-10	16.7	15.8
10-20	24.5	15.5
20-30	32.1	16.8

The Rengasamy *et al.* (1984) model has been adapted to show where the dispersive potential of scalded soils and waterponded soil fits (Figure 5-41). The model shows that based on EC and ESP results the scalds fit within the flocculated class 3a, and the waterponds in Class 1 as dispersive soils.

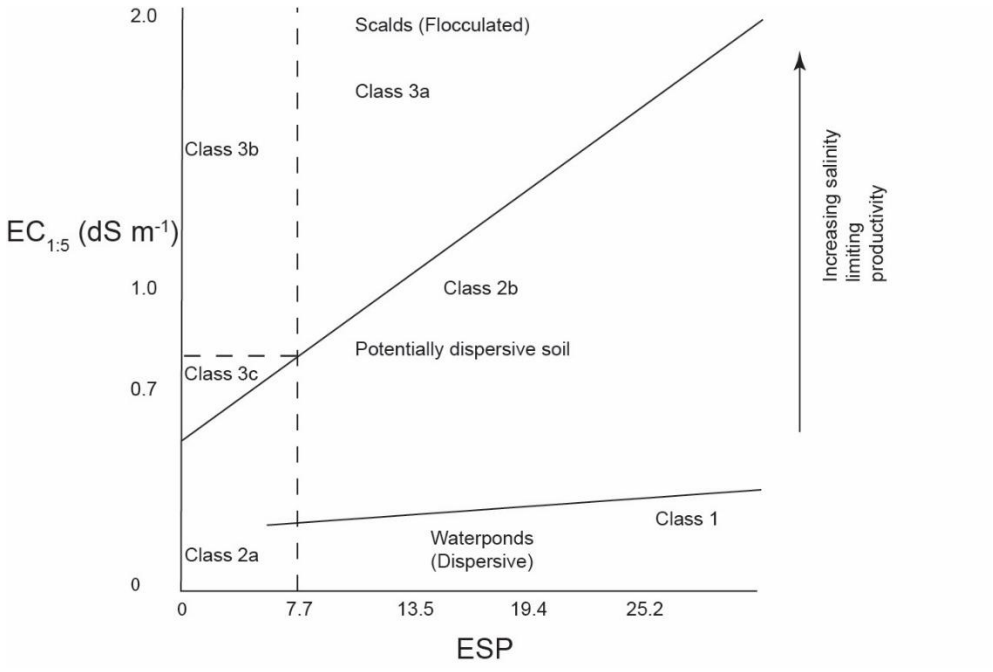


Figure 5-41. Dispersive behaviour of scalds and waterponds in relation to electrical conductivity (EC) and exchangeable sodium percentage (ESP) (after Rengasamy *et al.* (1984); Hazelton and Murphy, 2007).

5.8.18 Aggregate stability

Aggregates were assessed for stability on one scald site, and Waterpond Sites 1, 3 and 4 (*n = 144)*. Dispersion was scored between 0 and 16 (Table 5-20). Scalds showed no visible indication of dispersion and were found to have an average ASWAT score of 0 at 10 minutes and 2 hours. After remoulding the ASWAT score for the scalds was 2 in the 0-5 cm depth, but remained 0 for depths lower than 5 cm.

Waterpond aggregates were found to disperse with the degree of dispersion dependent on age (Table 5-20). The samples from the one year old waterpond (Site 3) were found to be potentially dispersive (Class 2 as defined by Rengasamy *et al*. (1984)) as they required remoulding to initiate dispersion. Samples from the five year old waterpond (Site 4) were found to disperse spontaneously (Class 1), especially in the 0-5 and 5-10 cm soil layer. These samples had an average ASWAT score of 9 in the 0-5 cm soil layer, and a score of 10 in the 5-10 cm layer. Remoulding was not always necessary for samples below the 10 cm depth increment. Remoulding was not necessary for any of the 25 year old waterpond samples (Site 1) as all samples dispersed within the initial 2 hour treatment.

Table 5-20. Aggregate stability in water (ASWAT) scores for waterponds and scalds. Scores > 8 are potentially dispersive.

Depth (cm)	Scald	1 yr	5 yrs	27 yrs
0-5	2.2	6.2	9.0	9.5
5-10	0.1	5.6	10.6	10.8
10-20	0	3.8	7.7	11.0
20-30	0	1.4	4.8	8.4

5.8.19 Effective cation exchange capacity of the whole soil

The results show no significant difference in the ECEC for the whole soil between waterpond and scald (Table 5-21). There is a trend showing that the ECEC decreases with depth increment, and reflects the influence of organic matter on ECEC, with higher ECEC values closer to the surface corresponding with the higher SOC concentration found in the surface layers. Some trends for specific cations are apparent. For example, the waterpond results show a higher concentration of Ca for all depths when compared with the results for the scald. Mg is lower in the waterpond than in the scald between the 0-20 cm depth increments, although the difference between waterpond positions reduces with depth. The Mg concentration decreases with depth, but shows variability between depths within the waterpond. Results for Na indicate an increasing trend in the scald with depth but show that Na generally reduces (leaches) with depth in the waterpond. In the surface 0-5 cm, the scald has lower Na than the waterpond (3.26 $cmol^{+}kg^{-1}$ compared with 5.6 $cmol^{+}$ kg^{-1}) but in depth increments below 5 cm the scald shows higher levels of Na than the waterpond.

Table 5-21. Whole soil effective cation exchange capacity (ECEC) and exchangeable cations ($cmol^{+}$ kg^{-1}).

Depth	Na	Ca	Mg	Ca:Mg Ratio	K	∑ECEC
			Scald			
0-5	3.3	9.1	11.2	0.8	1.5	25.1
5-10	4.0	8.5	10.5	0.8	0.9	23.9
10-20	5.4	7.1	9.0	0.8	0.5	21.9
20-30	6.9	6.3	8.0	0.8	0.4	21.6
			Waterpond			
0-5	5.6	11.0	7.6	1.4	1.1	25.3
5-10	3.7	10.4	8.53	1.2	0.8	23.5
10-20	3.6	10.0	8.8	1.1	0.7	23.0
20-30	3.7	9.5	8.4	1.1	0.6	22.2

The $EC_{(1:5\ water)}$ and $pH_{(1:5\ water)}$ results for the two cores used as Subset 2 (Table 5-22) show that $EC_{(1:5\ water)}$ and $pH_{(1:5\ water)}$ increase with depth in the waterpond, whereas in the scalded soil the reduce. In the upper 0-5 and 5-10 cm layers the pH is lower in the waterpond than in the scalded soil, while below 10 cm the pH is higher in the waterpond. The EC is higher in the scald than the waterpond throughout the soil profile.

Table 5-22. $EC_{1:5}$ (dS/m) and $pH_{1:5}$ for Subset 2 scald and 25 year old waterpond

Depth (cm)	*Scald*		*Waterpond*	
	EC	*pH*	*EC*	*pH*
0-5	2.6	8.21	0.1	7.61
5-10	3.6	8.01	0.1	7.82
10-20	3.5	8.18	0.3	8.51
20-30	3.3	7.96	0.8	8.93

5.8.20 Particle Size Analysis and prediction of the soil ECEC based on the mineralogy of the clay fraction

Samples from the four depth increments and the three waterpond positions (wall, mid, top) from one 25 year old waterpond (Site 1) and one core from the paired scald site (Site 2) were analysed to determine and compare mean particle sizes (PS). The results (Figure 5-42) show the volume weighted mean after partial dispersion (a) and full dispersion (b). The results of this analysis demonstrate that:

- The volume weighted mean PS decreases with depth in partially dispersed samples in all positions;
- Full dispersion treatment results in a large reduction in volume weighted mean PS in both the scald and waterpond samples;
- After full dispersion the scalded soil has a mean PS of 23.2 μm;
- After partial dispersion the mean PS in the 0-5 cm depth increment in all waterpond positions is greater than the scald;
- Variation in mean PS is present in all positions in the waterpond after full dispersion; and
- At depths below 5 cm the scalds and waterponds have a similar volume weighted mean PS.

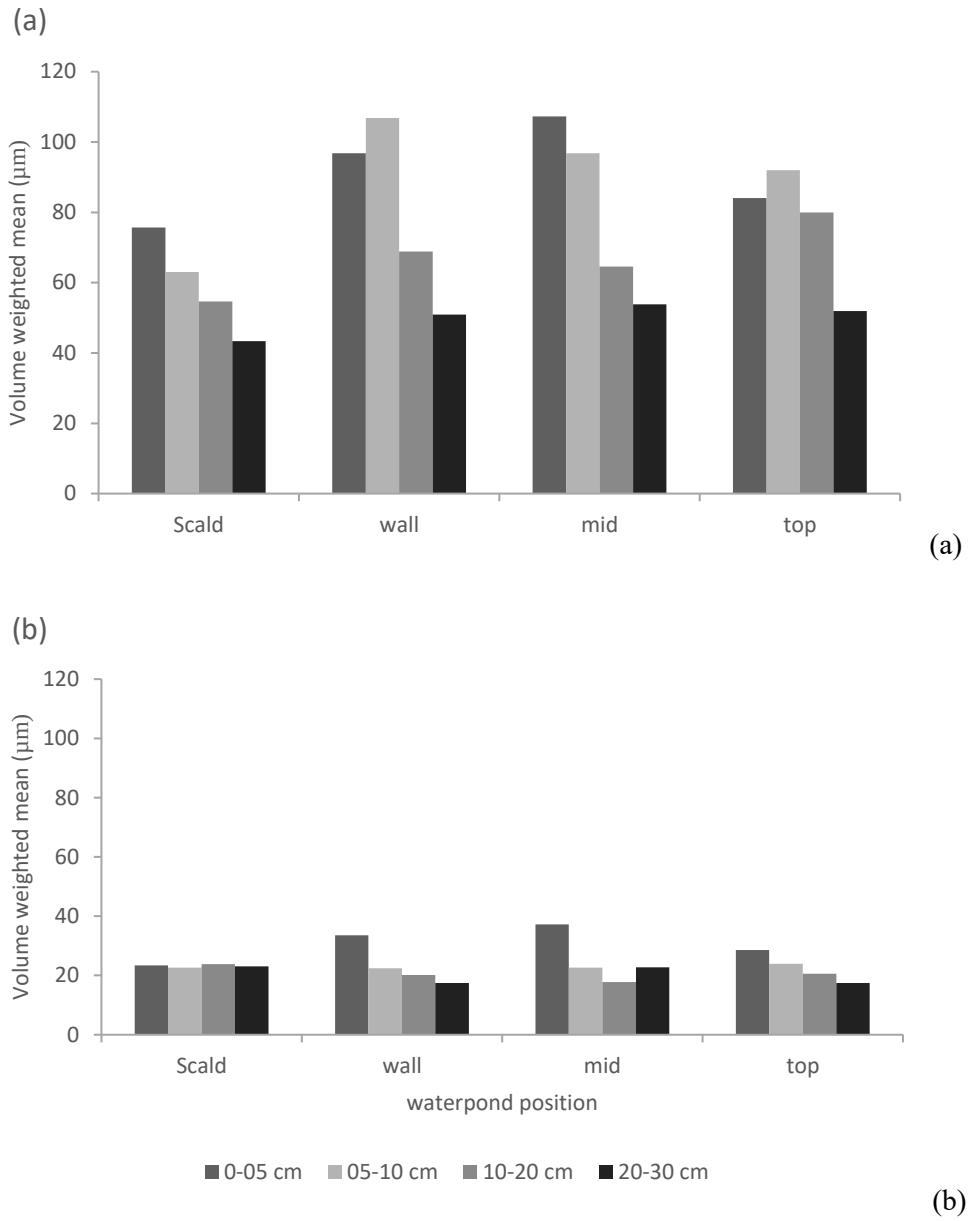


Figure 5-42. Particle size analysis after mild dispersion (a) and full dispersion (b), showing that waterponding results in changes to the physical structure of the soil (no error bars due to no replicates).

Samples that were subject to mild dispersion treatment (water only), show that the scald has a higher percentage of silt sized particles (2µm - 63µm) than any of the waterpond positions (57% for scald and average of 50% for waterpond positions) (Figure 5-43a). The scald also has a slightly lower percentage of sand sized particles (63 µm – 2000 µm) than the waterponds (41% and 48% respectively).

When disturbed with both $NaPO^3$ and ultrasound for 30 seconds to achieve full dispersion (Figure 5-43b), the percentage of sand particles was found to be much lower in both the scald and waterpond samples than the partially dispersed samples. For example, the scald had 41.5 % sand particles after mild dispersion treatment and 6.1% particles in the sand range after full dispersion. The results indicate that the majority of the sand sized aggregates can be broken down into silt sized particles when subjected to full dispersion treatment, so these are micro aggregates.

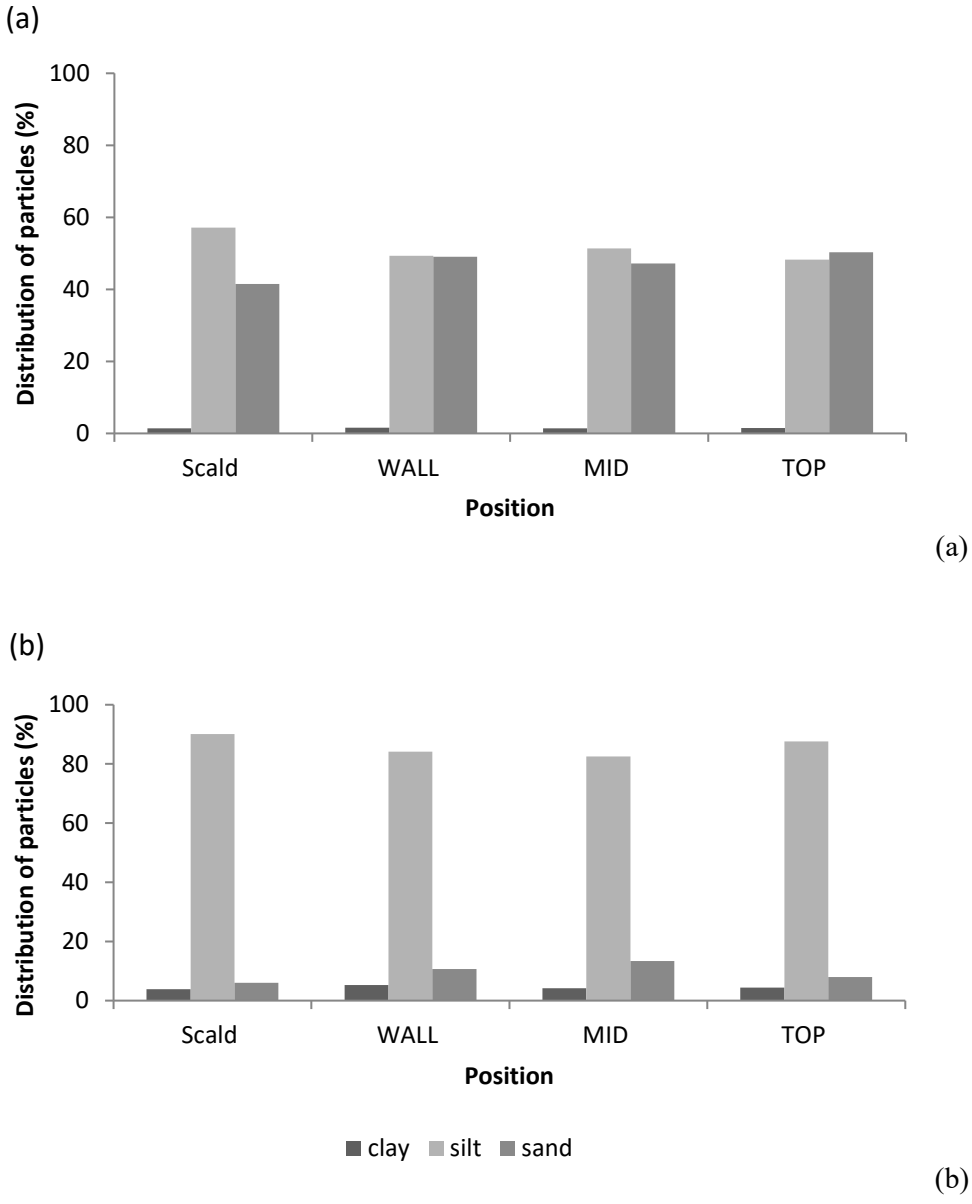


Figure 5-43. Particle size analysis comparing scald and waterpond positions in the 0-5 cm depth increment that have been partially dispersed in water (a), and fully dispersed using both $NaPO^3$ and 30 seconds of ultrasound (b).

The percentage of particles < 2µm, representing the clay fraction of the scald and waterpond soils was found to be less than 10% even after the full dispersion treatment (Table 5-23). The waterpond at the wall position and 5-10 cm layer had the highest clay fraction with 6.1% of particles sized < 2µm. The scald was found to have a lower percentage of clay than the waterpond samples in all depth increments. The PSA results at Table 5-23 suggest both the scald and waterpond soils have a very low clay percentage. This finding is not surprising as previously Mason *et al*. (2003) has demonstrated that PSA using laser diffraction greatly underestimates the < 2 µm material when compared with PSA performed using, for example, physical dispersion with the pipette method (Gee and Bauder, 1986).

Table 5-23. Percentage clay fraction (<2µm) using PSA for the scald and three waterpond positions and CEC (in parenthesis) calculated based on the PSA full dispersion measured clay fraction results

Depth (cm)	Scald	Waterpond Wall	Waterpond Mid	Waterpond Top
0-5	3.8 (2.49)	5.2 (1.19)	4.2	4.4
5-10	4.1 (0.86)	6.1 (1.06)	4.6	5.1
10-20	3.7 (0.1)	5.1 (1.24)	4.4	5.3
20-30	3.5 (0.67)	4.7 (0.94	5.3	5.0

The soil texture has previously been measured by Ringrose-Voase *et al*. (1989b) by using the plummet method (McIntyre & Loveday, 1974) to have a silty loam to silty clay texture with the clay fraction measured between 25% and up to 60 % clay at similar depths to those used in the current research. Therefore, for the purpose of predicting the CEC of the soil in both the scald and waterpond samples based on their mineralogy, it has conservatively been assumed both soils have a clay fraction of 35% at each depth increment. This percentage has been derived based both on the clay content measured by Ringrose-Voase *et al*. (1989b), that determined by field textures in the current study, and the approximate clay content for a silty clay soil in the range of between 35-40% clay (Northcote, 1979).

The method used to predict the soil CEC was to use CEC values for each of the clay minerals, the percentage of each clay mineral from XRD, and the clay percent of the total soil. For example Greene *et al*. (2002) used 120 for smectite, 20 for illite and 10 for kaolinite to calculate CECs of the clay fraction. However, clay minerals have variable CECs based on ranges and therefore the figures used by Greene *et al*. (2002) have been adjusted slightly to obtain a closer relationship with the measured ECEC from the current research (Pers comm, T. Eggleton 13 Jan 2016). Based on the XRD measured percentage of the clay minerals smectite, illite and kaolinite present in the scald and waterponds (Table 5-25) and assuming a 35% clay content of the soil, a mean predicted CEC (for the four depth increments) has been calculated with 38.6 $cmol^{+}\ kg^{-1}$ for the scald and 33.5 $cmol^{+}\ kg^{-1}$ for the waterpond (Table 5-24).

Table 5-24. Measured ECEC (a) compared with predicted CEC (b) using a clay fraction of 35% (typical of a silty clay texture (Northcote, 1979) and calculated by using CEC values for the clay fraction of minerals (100 for smectite (Smc), 20 for illite (I) and 4 for kaolinite (K) (Greene *et al*. 2002; Pers comm T. Eggleton 7 Jan 2016).

Depth (cm)	***Waterpond CEC ($cmol^{+}kg^{-1}$)***			
	(a) Measured ECEC	(b) Predicted CEC (35% clay) for Smc, I, K	(a) Measured ECEC	(b) Predicted CEC (35% clay) for Smc, I, K
0-5	25.1	37.5	25.3	31.5
5-10	23.9	36.9	23.5	33.1
10-20	21.9	36.8	23.0	34.9
20-30	21.6	43.0	22.2	34.6
0-30	23.1	38.6	23.5	33.5

5.8.21 X-ray Diffraction results

XRD analysis of the clay fraction shows that the 25 year old waterpond (Site 1) has more quartz, and kaolinite than the scald (Site 2). The waterpond was also found to have less illite than the scald (Table 5-25). Calcite is barely present in the waterpond samples but is present in the scalded soil. Smectite + chlorite is present in the scald samples but only at the 20-30 cm depth in the waterpond. The waterpond samples were found to have less than 5% smectite + vermiculite present in the upper 20 cm and trace amounts at 20-30 cm, whereas the scald was found to have none. Illite, a 10Å layer silicate was detected in both the scald and waterpond with higher percentage in the waterpond than the scald. Illite was also detected interlayered with 14Å clays. In this form there was a higher percentage in the scald than the waterpond. Spectra for the bulk samples and clay identification for the 0-5 cm layer of the scald and waterpond is shown in Appendix 3: X-ray diffraction spectra for a scald and a waterpond.

Table 5-25. X-ray diffraction clay mineral quantification (%) ± standard deviation for site 2 (Bugwah Scald (cell 2,2)) and site 1 (Bugwah 1986 waterpond (P2, wall)). nd = not detected.

Depth (cm)	*Scald*	*Waterpond*
	Quartz	
0-5	38 ± 0.6	43 ± 0.4
5-10	39 ± 0.7	41 ± 0.5
10-20	40 ± 0.7	42 ± 0.3
20-30	43 ± 0.7	45 ± 0.4
	Halite (NaCl)	
0-5	<1	nd
5-10	<1	nd
10-20	<1	nd
20-30	<1	nd
	Plagioclase	
0-5	6 ± 0.4	5 ± 0.3
5-10	6± 0.4	5 ± 0.5
10-20	5 ± 0.4	5 ± 0.3
20-30	6 ± 0.4	6 ± 0.3
	K-feldspar	
0-5	1 ± 0.1	1 ± 0.1
5-10	1 ± 0.1	1 ± 0.2
10-20	1 ± 0.1	1 ± 0.1
20-30	1 ± 0.1	1 ± 0.1
	Calcite	
0-5	2 ± 0.2	<1 ± 0.2
5-10	1 ± 0.2	<1 ± 0.2
10-20	3 ± 0.2	<1 ± 0.2
20-30	1 ± 0.2	<1 ± 0.2
	Hematite	
0-5	1 ± 0.2	2 ± 0.1
5-10	1 ± 0.2	2 ± 0.2
10-20	1 ± 0.2	1 ± 0.1
20-30	1 ± 0.2	1 ± 0.1
	Gypsum	
0-5	1 ± 0.2	1 ± 0.2
5-10	1 ± 0.3	1 ± 0.3
10-20	nd ± 0.2	<1 ± 0.2
20-30	<1 ± 0.2	<1 ± 0.2
	Kaolinite	
0-5	8 ± 0.5	11 ± 0.3
5-10	8 ± 0.5	10 ± 0.5
10-20	7 ± 0.5	11 ± 0.3
20-30	6 ± 0.4	8 ± 0.3
	Illite	
0-5	13 ± 0.7	20 ± 0.5
5-10	14 ± 0.8	19 ± 0.8
10-20	14 ± 0.7	18 ± 0.5
20-30	12 ± 0.6	14 ± 0.5
	Illite interlayered with 14Å clays	
0-5	26 ± 0.9	13 ± 0.9
5-10	25 ± 0.9	17 ± 0.2
10-20	24 ± 0.9	17 ± 0.9
20-30	22 ± 0.9	20 ± 0.9
	Smectite + Chlorite interlayered with Illite	
0-5	5 ± 0.9	nd
5-10	5 ± 0.9	nd
10-20	5 ± 0.9	nd
20-30	8 ± 0.9	5 ±0.4
	Smectite + Vermiculite interlayered with illite	
0-5	nd	4 ± 0.8
5-10	nd	4 ± 0.8
10-20	nd	5 ± 0.9
20-30	nd	Possibly present

5.8.22 ICP-AES analysis for total Oxides

Oxide analysis results show that the scald core (Site 2) has a higher CaO, MgO and Na_2O concentration than the core from the 25 year old waterpond (Site 1) (Table 5-26). The results also show the waterpond has a higher concentration of Al_2O_3 and Fe_2O_3 below the 5 cm depth increment than the scald. There is little difference between the scald and waterpond cores at all depth increments for K_2O, MnO, P_2O_5, SiO_2 and TiO_2.

Table 5-26. Oxide analysis (%) showing differences between one scald and one 27 year old waterpond sample.

Depth (cm)	*Al_2O_3*	*CaO*	*Fe_2O_3*	*K_2O*	*MgO*	*MnO*	*Na_2O*	*P_2O_5*	*SiO_2*	*TiO_2*
					Scald					
0-5	13.1	1.3	5.85	1.81	1.72	0.17	1.12	0.09	63.7	0.82
5-10	12.8	1.92	5.76	1.80	1.69	0.16	1.17	0.08	63.3	0.80
10-20	12.4	2.6	5.62	1.78	1.71	0.14	1.17	0.09	63.1	0.80
20-30	12.7	1.87	5.72	1.82	1.72	0.12	1.25	0.08	64.8	0.83
					Waterpond					
0-5	13.6	0.42	5.88	1.8	1.12	0.19	0.66	0.08	66.3	0.86
5-10	14.6	0.49	6.51	1.93	1.27	0.17	0.69	0.08	63.1	0.86
10-20	14.7	0.49	6.36	1.82	1.33	0.15	0.75	0.07	62.5	0.87
20-30	13.9	0.90	6.08	1.78	1.44	0.14	0.90	0.08	64.2	0.88

5.8.23 Hummock soil

Soil samples were taken to 50 cm in the hummock soil to reflect the loss of the A-horizon from the scald. Effectively the equivalent of the 30-50 cm soil layer of the hummock is taken to be the equivalent of the 20-30 cm in the scalded soil. The hummock soils were found to have low levels of TOC (Table 5-27). Soil below 10 cm depth was found to have a higher concentration of TOC (mean 0.4 g C $100g^{-1}$) than soil in the 0-5 and 5-10 cm layers (mean 0.31 g C $100g^{-1}$). The concentration of total N was highest in the 30-50 cm layer at 470 mg kg^{-1} and lowest at 300 mg kg^{-1} in the 5-10 cm layer.

Table 5-27. SOC and TN concentration (g C $100g^{-1}$) for composite hummock sample)

DEPTH (cm)	**TOTAL SOC (mg kg^{-1})**	**TOTAL N (mg kg^{-1})**	**C:N RATIO**
0-5	3 200	370	9
5-10	3 000	300	10
10-30	4 200	400	11
30-50	3 800	470	8

The results show there was no ^{137}Cs activity in the top 30 cm of the hummock soil (Table 5-28). There was however, ^{137} Cs activity identified below 30 cm. This indicates that soil particles that have not previously been exposed to fallout have been deposited at the relic sites. ^{137}Cs activity has been detected in the top 10 cm of the scalded soils but none below this depth. There is more ^{137}Cs activity detected in the mid and wall positions of the waterpond than the top position.

Table 5-28. ^{137}Cs activity (Bq/kg)

Depth (cm)	*Relic Hummock*	*Scald*	*Top*	*Mid*	*Wall*
0-5	0	0.9	0.88	1.17	1.14
5-10	0	0.97	0	0	0.78
10-20	0	0	0.62	0.68	0.57
20-30	0	0	0	0	0
30-50	0.51	na	na	na	na

5.9 Discussion

The overarching hypothesis of this research is that establishment of waterponds on scalded clay pan soils can lead to SOC sequestration. Consequently, this case study has aimed at quantifying temporal changes to SCS (0-30 cm depth) that can be attributed to establishment of waterponds.

In order to investigate the SOC sequestration potential of the scalds the experimental design enabled comparison of SOC in both scalded and waterponded soil. A chronosequence of waterponds established between one and 27 years prior to sampling was used. Assessment of a number of physico-chemical characteristics was undertaken to determine their role in modifying the capability of the scalded soil to sequester SOC. To test the hypothesis that waterponding leads to SOC sequestration, initially the soil BD was determined and then the SOC concentration was measured. These two data sets allowed calculation of changes to SCS following waterponding and hence the rate that C sequestration occurred.

The results indicate that there are a number of differences between scalded soil and waterponded soil caused by the retention of water in the waterponds. Additionally, there are a number of unique processes related to SOC sequestration in waterponds compared with changes typically observed following LUC. The scalded soil has a relatively low, but stable SOC equilibrium because there is sparse vegetation, no disturbance to the profile such as occurs via water infiltration or mechanical intervention, or other disturbances capable of causing fluctuations to SCS. Waterponding, however, initiates a dynamic sequence of physical and chemical changes to the scalded soil which lead to SOC sequestration. Initially, soluble salts are leached from the soil, and then, following wetting and drying the natural shrink-swell function of the soil leads to the development of cracks within the soil profile. These cracks further enhance the infiltration of water into the soil profile, as well as enhancing the lodgement of seeds. Following these changes, vegetation is established thereby allowing SOM to accumulate in the soil profile. The accumulation of SOM leads to a gradual increase in SOC over a relatively short, five year time frame, until a new, higher, equilibrium level is reached. The processes involved in the distribution of SOC through the waterpond soil profile include the influence of soil BD, nutrient cycling and the potential permanence of the SOC following waterponding.

5.10 Changes to EC and pH following waterponding

SOC sequestration could not occur in scalded soil without changes to the physical and chemical properties of the soil profile. Important findings from the current research can largely be related to changes to the EC and pH following waterponding.

5.10.1 Leaching of soluble salts stimulates favourable conditions for vegetation growth and succession in waterponds

The scalded soil was found to be both saline having a high soluble salt concentration, and sodic having a high exchangeable sodium level. The scalds were found to have an $EC_{(1:5\ water)}$ of 1.84 dS/m in the 0-5 cm soil layer. In the soil layers below 10 cm the $EC_{(1:5\ water)}$ was greater than 3.0 dS/m. These results are consistent with the findings of Jones (1966); Jones (1967); Ringrose-Voase *et al*. (1989a, b) and Ditchfield (1996).

Soil $EC_{(1:5\ water)}$ results represent the quantity of salt per unit of soil whereas $EC_{(sat)}$ is a measure that represents the concentration of salt in soil water (Slavich and Petterson, 1993). The results (Table 5-8) show the $EC_{(sat)}$ of the scalded soil is 17.45 dS/m in the 0-5cm soil layer and reach a maximum of 26.75 dS/m in the 10-20 cm layer. Soils are considered saline when the $EC_{(sat)}$ is > 4 dS/m. At this concentration crop yields can be reduced (Munns and Tester, 2008). Therefore the very high soluble salt concentration found in the scalded soil is an important factor making the scalds inhospitable to many plant species.

The soluble salts present in the scalded environment are most likely to have originated during the landscape evolutionary processes of primary mineral weathering and aeolian cycling and deposition. The aeolian deposition most likely occurred prior to European settlement (Gunn and Richardson, 1979; Chartres, 1993). Indeed, the influx of salt into the region most likely occurred during the glacial age arid period between 18 000 – 16 000 years ago. There is evidence that during this period aeolian activity intensified across the arid interior region of Australia. Such evidence includes low precipitation, dry lake beds, formation of lunettes, and the expansion of dune fields. These events occurred following changes to climatic oscillations during the Quaternary (Bowler, 1976; Gunn and Richardson, 1979). The dominant anion in the scalds is most likely to be chloride (Cl^-). In his study Jones (1969) found scalds had a mean Cl^- of 0.4% and the waterponds 0.05%.

The waterponds were found by the current research and others (e.g. Jones, 1969; Ringrose-Voase *et al*. 1989a; Ringrose–Voase and McClure, 1994; Ditchfield, 1996), to have a lower EC than scalded soil. Specifically, the results of the current research show that the mean $EC_{(1:5\ water)}$ in the 0-5 cm depth increment in one year old waterponds is 0.27 dS/m; a reduction of 1.56 dS/m when compared with the scalds. Further, the EC in the same waterponds in the 5-20 cm

layer was found to be > 2 dS/m lower than the scalded soil. The EC is even lower in the older waterponds when compared with the scalds.

The $EC_{(1:5\ water)}$ is lower in waterponds because the banks trap rainfall water on the clay pan scald surface. Once trapped, there is sufficient time for water to break down the surface crust, infiltrate and leach the soluble salts from the soil profile. Gupta and Abrol (1990) have suggested that up to 80% of soluble salt can be leached from the root zone (upper 50 cm of the soil profile) after infiltration of 50 cm of water. Thus, the rapid decrease in EC in the young waterponds is in line with their proposition.

Soils affected by high concentrations of soluble salt can only sustain low levels of plant growth due to the inability of plants to take up saline water. The limitation on plant growth can be caused by the salts decreasing the osmotic potential of soil, causing Cl^- or Na^+ ion toxicity or can be caused by a nutritional impediment due to cation imbalance (Bernstein, 1975; Naidu and Rengasamy, 1983; Wong *et al*. 2008). Scalds typically have between 0 to 10 % vegetative ground cover (Thompson, 2008). Thus, the very low SCS (to 30 cm depth) found in the scalds profile is a direct result of the low plant growth. Further, C cannot be sequestered in the soil without first removing the soluble salts. Consequently, because waterponding initiates leaching of soluble salt from the soil profile, a wide range of annual and perennial plant species can grow, thereby establishing a succession of plants in the waterponds, and resulting in restoration of ecological diversity to the landscape (Thompson, 2008).

Beadle (1948c), Rhodes (1987) and Thompson (2008) have listed plant species that naturally inhabit the waterponds. Indeed, Thompson (2008) claims that over 80 % of the scald surface can become revegetated following waterponding. Much of the revegetation occurs naturally from seed that is washed or blown across the surface. However, some plant species are introduced into the waterponds by broadcasting or drilling methods during their construction. Thompson (2008) for example, explains that *Atriplex nummularia, A. vesicaria* and *Astrebla lappacea* seed is broadcast or sown across each waterpond and on the bank as it is constructed to hasten revegetation of these desired species. Generally, however, the resulting plant species composition depends on the diversity of the natural seed source, the rainfall intensity and season of occurrence, and wind direction during the period immediately following waterpond construction (Rhodes, 1987).

The rate and extent that soluble salts are leached from the scalded soil is an important factor in determining which species colonise waterponds because some plant species are more tolerant of saline soil conditions than others. Some of the dominant waterpond plant colonisers and their tolerance level of saline conditions can be listed (Table 5-29).

Table 5-29. Some vegetation species found in waterponds known to tolerate waterlogging and saline soil conditions (L = low; M = moderate, H = high, NA information not available) After Rhodes, 1987; FFI CRC) .

Botanical name	***Common name***	***Water logging tolerant***	***Salt tolerance***
Astrebla lappacea	Curly Mitchell Grass	NA	M
Atriplex nummularia	Oldman Saltbush	L	M
Atriplex semibaccata	Creeping saltbush	L	M
Chloris truncate	Windmill Grass	L	L
Hordeum leporinum	Barley grass	M	M
Hordeum marinum	Sea Barley	M	H
Maireana pentagona	Hairy bluebush	NA	H
Medicago polymorpha	Burr Medic	LM	LM
Portulaca oleracea	Common Pigweed	L	M
Sclerolaena muricata	Black Rolly-poly	L	M

Although some of the plant species that become established in the waterponds have low salt tolerance, others are halophytic; a characteristic that allows plants to either tolerate or exclude the uptake of Na^+ and Cl^- from saline soil water. For example, Garthwaite *et al.* (2005) determined that *Hordeum marinum*, one of the species found on the waterponds (Table 5-29) has a high salt tolerance, because it can exclude Na^+ and Cl^- in soils with an EC up to 45 dS/m, yet still retain the capacity to produce relatively high quantities of dry biomass.

Rhodes (1987) showed the relationship between leaching of salt and an increase in perenniality of vegetation in waterponds; perenniality being important for sustainable increases in SCS. Rhodes found that initially halophytic plant species populate the waterponds including *Atriplex spp, sclerolaena spp* and *Hordeum spp* amongst others (Rhodes, 1987). These species are able to tolerate the levels of soluble salt still present in the profile in the period after waterponds are initially constructed. Over time there is an increase in plant species diversity including some species that have low salt tolerance such as *Chloris Truncata* (Cunningham, 1987). Rhodes (1987) described these changes in the context of landscape rehabilitation as depicted in Figure 5-44. After the scalded soil becomes revegetated it is possible to increase grazing livestock numbers and therefore enterprise profitability, but also as shown in the current research, sequester SOC.

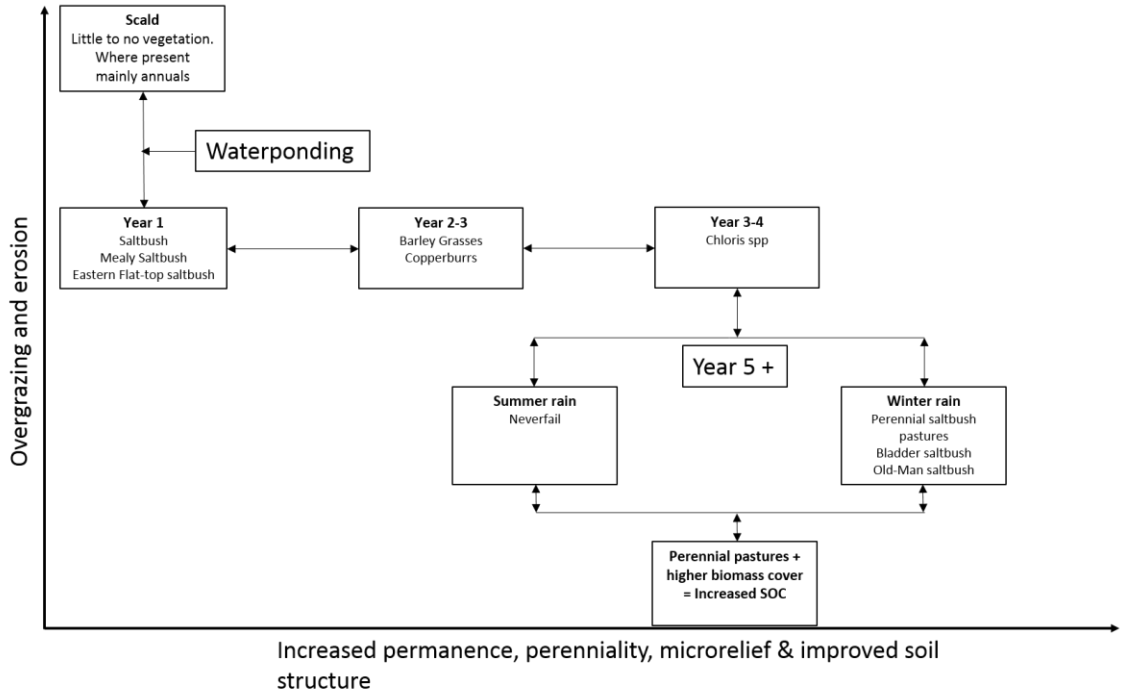


Figure 5-44. Succession of scald rehabilitation and SOC sequestration following waterponding by years post establishment. Arrows are two-way because reversion due to overgrazing is possible (After: Rhodes, 1987).

In the terms of the current research project, the succession of plant species that become established on waterponds after soluble salts are leached from the soil profile is an important contributing factor, leading to a commensurate temporal increase in SOC following waterponding of scalded soil. Consequently the model of succession developed by Rhodes (1987) (Figure 5-44) has been adapted to show that the vegetation succession leads to increasing SCS in waterponds.

5.10.2 Soil pH

The current research has found the $pH_{(1:5\ water)}$ in both the scalds and waterponds is within the neutral to alkaline range. The scalds were found to have an average pH of 7.3 at the surface whereas the 25-27 year old waterponds were found to have an average pH of 8.0 at the same depth. This is as expected because arid zone soils are generally found to have a neutral to alkaline pH and this is often associated with high EC values and $CaCO_3$ (Yuan *et al*. 2007), therefore the pH range of the scald and waterpond soils of the semi-arid region where this study was undertaken is as expected.

The change, whereby the results show a temporal increase in soil pH following waterponding is most likely a result of leaching soluble salts from the soil profile. For example, Peech (1965) explains that the pH of a soil suspension is reduced when the concentration of soluble salt in the suspension is increased and the corollary occurs when the solution is diluted with water. Although Peech's work explained the relationship between pH and soluble salts in a saturated paste and soil solution, it is reasonable that a similar behaviour can occur in the field whereby

water that is trapped in the waterponds leaches the soluble salts thereby causing the pH to increase. The equilibrium pH in the presence of Na^+ salts is lower than that in the presence of Ca^{2+}, Mg^{2+} or K^+ salts (Field and Little, 2008).

Further evidence that changes to soil pH can be caused by leaching is associated with natural variation in response to seasonal change. Typically, the pH is lowest during hot dry seasons and highest during cool wet seasons. Seasonal changes to pH are closely associated with leaching of soluble salts (Peech, 1965; Thomas, 1996) and changes to H^+, NO_3^- and Mn_2^+ (Conyers *et al*. 1995). Additionally, in a study exploring the pH response to various wetting and drying cycles, Paul *et al*. (1999) found that soil pH increased in continuously dry soil whereas the pH decreased in soils that were maintained in a continuously wet condition. Their study also found that the pH decreased in soils subject to moist-dry cycles. They attributed the lowering in soil pH to increases in NO_3^- levels resulting from nitrification of SOM.

The mechanism for change to the soil pH following waterponding and leaching of the soluble salts is hydrolisation of exchangeable Na^+ which results in an increase in the OH^- concentration of the soil solution (Chorom *et al*. 1994). After soluble salts are removed from the solution displacement of H^+ and Al^{3+} ion by other cations occurs on the cation exchange sites (Bloom *et al*. 2005). Consequently the finding that the pH is lower in scalds which have a high EC, and higher in waterponds after the soluble salts have been leached from the profile (Table 5-22) is consistent with accepted pH/electrolyte relationships and commensurate changes to the composition of exchangeable cations on clay particles.

In both scalds and waterponds there is a general trend of increasing pH with depth. For example below the 5 cm depth increment the soil pH in the waterponds ranges between 8.0 and 8.5. This trend is a feature of soils located in areas with drier climates (McKenzie *et al*. 2004). The pH trend with depth is consistent with the findings of other authors who have also researched the scald and waterpond environment (Jones, 1969; Ringrose-Voase *et al*. 1989a; Ringrose–Voase and McClure, 1994; Ditchfield, 1996) and is a characteristic associated with the soil reaction trend in soil profiles of duplex and gradational soils as used by Northcote (1979), for soil classification.

Because changes to pH can change the net negative charge of clay particles and hence the interaction between clay particles, increases to pH can result in a commensurate increase in dispersive behaviour (Rengasamy and Olsson, 1991; Sumner, 1993; Chorom *et al*. 1994). Gupta *et al*. (1984) found that when the soil pH increases the propensity for soil to disperse can also increase. Thus the higher pH in waterponds is a factor influencing dispersion of the waterponded soil. The factors affecting the flocculation and dispersion properties of scalds and waterponds are discussed in detail in Section 5.15.

5.11 Waterponding of scalds and changes to C concentration

Scalds were found to have a mean SOC $_{(Heanes)}$ concentration of 0.39 (±0.02) g 100g^{-1} in the surface 0-5 cm, 0.43 (±0.02) g C 100 g^{-1} from 5-20 cm, but decreasing to 0.37 (±0.02) g C 100 g^{-1} in the 20-30 cm depth (Table 5-7). The decreasing concentration with depth, although relatively small, was expected due to the diminishing nutrient and water resources normally available with depth. Such a trend is consistent with other research reported in the literature (e.g. Skjemstad *et al*. 1998; Baldock and Skjemstad 1999; Olsen *et al*. 2014). The SOC concentrations found in the scalds are rated as extremely low (i.e. Hazelton and Murphy (2007) rate < 0.40 g C 100 g^{-1} as an extremely low concentration of SOC). The low concentration in the scalds was anticipated for three reasons:

1. Soil landscapes typically have most SOC concentrated in the A-horizon, and have a decreasing concentration associated with soil depth, particularly in the B-horizon. However, as the A-horizon has been almost entirely eroded from the scalded soil, only the depauperate, SOC poor B-horizon remains;
2. There is little if any vegetation to be a source of SOM on the scalded soil; and
3. The scalded soil studied for this research project are located in the semi-arid rangelands where high temperatures, low rainfall and high evaporation rates enable only sparse vegetation growth and low SOM inputs.

Statistically, there was significant variation in SOC concentration between the three scald sites. Scald Site 14 (Figure 5-7) located to the north of the transect had a significantly lower ($P <$ 0.01) concentration through the 0-30 cm soil profile than the other two scald sites (Figure 5-14). This finding is not unexpected because there is approximately 60 km distance between the three scald sites and therefore a range of factors could exist to influence the soil physical and chemical attributes. Numerous authors have demonstrated heterogeneity for SOC concentration across the landscape even at very small spatial and temporal scales. The variation is influenced by factors such as landscape morphology, parent material, climate, vegetation and land use (Skjemstad *et al*. 1998; Lal, 2002; Schuman *et al*. 2002; Gray *et al*. 2015; Xiong *et al*. 2015). However, with little over 0.1 g C 100g^{-1} difference in SOC concentration between the sites (± SE 0.02) the similarity in SOC concentration between the scald sites is remarkable. The most likely reasons for the apparent homogeneity, despite the considerable distance between sites, are the similar soil morphology, lack of vegetation to stimulate landscape heterogeneity and very little variability in climatic characteristics (average rainfall 432 mm, SD ± 16; average temperature 34.22 ^{o}C, SD ± 0.16; and average evaporation 2 023 mm, SD ± 3.55) between the sites.

The one year old waterponds had a mean SOC concentration of 0.54 (± 0.1) g C $100g^{-1}$ at 0-5 cm (Figure 5-26). This represents a 38% increase (0.15 g C $100g^{-1}$) in SOC concentration just one year after waterponding. However, this is still rated as a very low concentration. For example, Hazelton and Murphy (2007) suggest that between 1.0 and 1.8 g C $100g^{-1}$ is a moderate level of SOC in terms of contribution to soil physical properties. The results of the current research also show SOC is higher in the 5-10 cm depth in the waterponds than in the scalds, and this difference is observed in waterponds just one year after waterponding. However, the difference is less apparent below the 10 cm soil layer, indicating more time is required for the SOC to become sequestered deeper in the soil profile.

After approximately five years the mean SOC concentration within the waterponds in the 0-5cm depth was found to be 0.73 (±0.1) g C $100g^{-1}$. This represents an 87 % increase in SOC and is the equivalent of an increase of 0.4 g C $100g^{-1}$. The increasing trend over the relatively short five year time period supports the hypothesis that waterponding effectively sequesters SOC. However, there is no evidence from the results to show that waterponding sequesters additional SOC beyond five years.

Interestingly, the increase in SOC is commensurate with the increase in groundcover following waterponding. Thompson, 2008 measured changes to vegetative groundcover following waterponding. His results found that scalds typically have on average 10% groundcover. The study showed there was a 20% annual increase in chenopod grassland groundcover up to 3 years after construction of waterponds but mainly annual species. When waterponds are around 5-7 years old they have on average 80% ground cover. The plant species mix by this time is comprised of native pastures and chenopods with the majority comprised of perennial plant.

5.11.1 Within waterpond variation in SOC

The difference in SOC concentration between scalds and waterponds was statistically significant ($P < 0.001$). However, the difference within waterpond positions was found not to be significant ($P = 0.77$). The interaction between within waterpond position (wall, mid and top) and age was also found not to be significant ($P= 0.27$). However, the depth interaction was significant ($P < 0.001$). Although position was a significant factor, by examining changes to the SOC concentration in the within waterpond positions some trends were identified. For example, it was found that the greatest increase in SOC concentration in the 0-5 cm depth occurs initially at the wall position of the waterponds where the SOC concentration was 0.58 g C $100g^{-1}$ compared with 0.52 g C $100g^{-1}$ at the top position. This suggests a pattern of spatial distribution of resources within waterponds where most seed and water resources accrue downslope at the wall position leading to commensurate increases in SOC. After five years the mean SOC concentration was 0.8 g C $100g^{-1}$ at the wall position, but only 0.61 g C $100g^{-1}$ for the top position. The temporal and spatial increase in SOC concentration across the waterpond surface suggests that the natural slope of the landscape initially facilitates movement of seed and water

resources into the lowest position of the waterpond. Later, the top position also begins to sequester SOC but it takes longer because the top position is up slope from the wall position and therefore resources take longer to accrue.

Gradually, as the waterponded soil regains functionality, resources are spread across most of the waterponded surface, and eventually waterponds are almost completely covered with vegetation. Approximately ten years after establishment the variation in SOC concentration across the three waterpond positions was found not to be significant with the range between 0.76 g C $100g^{-1}$ and 0.81 g C $100g^{-1}$ SOC (Figure 5-25). The results show that SOC concentration reaches an equilibrium at the 0-5 cm depth increment with no significant difference between the ten year old waterponds and the 25-27 year old waterponds (Figure 5-25).

5.11.2 Distribution of SOC with depth

The waterponds were found to have a higher SOC concentration than scalded soil at all the depth increments examined (Figure 5-26). However, the difference between waterponds and scalds reduced with depth. The results show that the greatest increase in SOC concentration occurs following waterponding in the 0-5 cm depth increment. This is consistent with plant growth in waterponds, root development and increases in SOM. SOC is normally concentrated at the soil surface due to proximity to plant roots and the higher microbial activity that occurs near the surface. For example, Jobbagy and Jackson, (2000) examined vertical changes to SOC in relation to vegetation, precipitation and soil texture. These authors concluded that vegetation was the most important factor in determining the vertical distribution of SOC, due to the heterogeneous patterns of root and plant distribution. Therefore, it is most likely that the higher SOC concentration at 0-5 cm compared with the deeper increments is consistent with the establishment of vegetation within the waterponds and the distribution of roots in the upper soil profile.

5.11.3 SOC concentration summary

The overall increase in SOC across the 27 year waterpond chronosequence can be directly attributed to the LUC achieved by converting scalded soil into grassland and shrubland following waterponding. In a meta-analysis into SOC sequestration in 17 countries following changes to grassland management Conant *et al*. (2001) found that management improvement and land use conversion generally result in an increase of SOC concentration. Notably Conant *et al*. (2001) found that in 76 % of studies SOC increased following management change. The authors also found that SOC concentration increased by an average of 29 % following management change. Further, those authors found that for the studies where an increase in soil C occurred, the mean increase was actually 51 %. Therefore, the 87 % increase in SOC

concentration following waterponding is relatively high when compared with the average increases reported (Conant *et al*. 2001).

The current research also found that scalded soils have the potential to sequester SOC in a relatively short period of time following rehabilitation based on the increase in SOC concentration following waterponding. The quantity of SOC sequestered depends on the amount of SOM input into the soil profile and the quantity lost via decomposition (Jobbagy and Jackson, 2000; Lutzow. *et al*. 2006). Further, highly degraded soils, such as the scalded soils, that have been eroded and depleted of SOC have the greatest potential to rapidly sequester SOC at rates exceeding 1 Mg C ha^{-1} yr^{-1} (Lal, 2004b; Govers *et al*. 2013). However, such rates can only be achieved and sustained where sufficient nutrient supply exists (Govers *et al*. 2013). In addition, the unique soil physical and chemical attributes of the scalds viz. the salinity and sodicity, need to be modified before plant growth can occur thereby allowing SOC to be sequestered. In the scalded soil physical and chemical modification is achieved following leaching of soluble salt which provides an environment suitable for vegetation to become established. Consequently the results of the current research corroborate the proposition by Lal, (2004b) and Govers *et al*. (2013) that degraded soils such as the scalds have potential to sequester C at a relatively rapid rate, when compared with LUC activities in soils with already moderate levels of SOC. However, sequestration can only occur following modification of soil physical and chemical constraints together with adequate nutrient availability.

5.12 Changes to soil bulk density following waterponding

The results of the current research show a number of changes to soil BD following waterponding. Notably there was found to be:

- Significantly lower (P<0.001) soil BD in the waterponds compared with the scalds, indicating that waterponding leads to a reduction in soil BD;
- Variation in BD between the three scald sites most likely due to the presence of either soft or hard scalds depending on the provenance of the soil;
- Significant differences (P=0.04) in BD between the waterponds of different ages. However, the results do not demonstrate a clear temporal trend with variable results between age classes identified. For example, in the surface 0-5 cm, the BD trend for age classes decreased in the order:
    - one year old waterponds > 25-27 year old waterponds > five year old waterponds > ten year old waterponds; and

- Within waterpond position was found to be not a significant factor (*P*=0.63). There was, however, a trend showing that by 25-27 years of age the top position has a higher BD at most depth increments than the wall and mid positions. The trend is feasible in view of the distribution of vegetation, SOC, water and surface cracks within waterponds.

The establishment and increased growth of biomass with time in the waterponds is considered to be the most likely cause for the lower soil BD in the waterponds than in the scalded soil. It is widely accepted that there is a strong relationship between SOM and soil BD, whereby higher levels of SOM generally leads to lower BD; principally because SOM promotes aggregation and is associated with plant roots that enhance porosity.

Different factors influence the soil BD of the scalded soil rather than the waterponded soil and they are addressed separately below.

5.12.1 Scald soil bulk density

The three scald sites were found to have a mean soil BD to 30 cm of 1.54 Mg m^3. The results show variation between the three scald sites particularly at the 0-5cm depth where the range was found to be between 1.26 and 1.78 Mg m^3. Because there is a complete lack of vegetation on the three scalded sites SOM is limited and therefore not likely to be a factor affecting soil BD. The surface crust on the scalded surface prevents water infiltration and hence inhibits crack formation. Consequently, soil cracks are not a factor to disrupt core uniformity on the scalds. However, there are four other factors associated with the scalded soil that could contribute to the variation to soil BD in scalds including:

- The subangular blocky or “nutty” structure (Ringrose-Voase *et al*. 1989) of the scalded soil profile causing variability in inter-aggregate pore space;
- Moisture content of the soil at time of coring whereby moist samples will be more cohesive than dry samples. During this project sampling was undertaken in October when there was moisture in the soil profile and in February when the soil was very dry;
- Different soil morphologies noting there is approximately 60 km difference between the most northern and the southern sites; and
- The presence of either soft or hard scalds depending on soil provenance.

It is likely that a combination of each of the four factors contributed to some extent in density variability between scald sites.

5.12.2 Waterpond soil bulk density

When the results of the within waterpond positions (wall, mid and top) were averaged to obtain a mean waterpond soil BD result, a temporal decline in soil BD was not identified. Instead the BD varied between ages as follows:

- 1year > 5years > 10 years (Figure 5-22).

The 25-27 year old waterponds, however, had a mean soil BD below the 5 cm depth increment almost as high as the 1 year old waterponds.

The lower soil BD in waterponds less than ten years of age than in scalds, and the declining trend with time following waterpond establishment was expected due to a commensurate increase in vegetation growth and SOM.

The reasons for the apparent increase in mean BD for the oldest waterpond sites is likely to be the consequence of two key factors:

(i) Soil compaction due to stock trampling; and
(ii) Soil dispersion (Table 5-20) leading to changes to the soil structure.

Murphy *et al*. (2003) conducted research on sites located within the same catchment as this study and found variation in BD between sites which they attributed to heavy grazing leading to compaction. Soil compaction caused by grazing livestock has also been observed by Willatt and Pullar (1984) and Steffens *et al*. (2008) amongst others. Consequently, livestock grazing on the waterponded landscape provides a feasible explanation for the within waterpond variability in soil BD. Another probable cause of the higher BD within the oldest waterponds is the reorganisation of soil particles due to the greater tendency of their soils to disperse with increasing age. This characteristic is discussed further at Section 5.14.

When the 12 waterpond sites were initially inspected a gradient in the density of vegetation biomass was observed within each waterpond; i.e. there was visibly more vegetation closer to the 'wall' position than in the mid position and particularly less at the top position being the most upslope position in the clay pan waterponds. The gradient was particularly noticeable in the younger waterponds. These observations were consistent with expectations that more water and seed resources accumulate closer to the wall of the waterpond (the downslope position) due to water and wind transport mechanisms. Given this observation it was expected that the soil BD would be lower near the wall of the waterponds and higher closer to the top commensurate with the vegetation abundance. However, the within waterpond BD results (Figure 5-21) show a greater spatial variability than expected with no apparent relationship with the within waterpond vegetation gradient.

There are two other factors that probably contribute to the within waterpond variation in BD observed in this study. The first includes the actual establishment of waterponds that leads to

soil structural change and which occurs after the soluble salts have been leached from the scald soil profile. After the salts have been leached, the nutty peds previously present in the scalded soil disperse due to water infiltration into the profile. Then, following a series of wetting and drying events and shrinking and swelling, the waterponded soil develops a prismatic structure, interspersed with soil cracks. This process most likely occurs faster at the wall position or in low areas of micro-relief, than at the top position of the waterponds. Therefore, variability in the rate and distribution of soil structure modification could contribute to the variability in BD that has been identified across the waterpond. Secondly, livestock which are most likely to preferentially graze where the most abundant and greenest vegetation is present can cause soil compaction whilst grazing, especially where and when the soil is moist. Consequently, because most biomass appears to grow closer to the wall position where water and seed resources are most likely to accumulate, livestock compaction of the soil is a probable cause of the higher BD in the wall position.

In summary, , it is likely that in the waterponded soils dispersion, soil cracking, vegetation and SOM and stock trampling have affected soil BD. In the scalded soil the nutty soil structure and soil moisture probably have had an effect on the soil BD results. Finally the use of the soil coring method and the potential for compaction during core drilling probably all contributed to some extent in the variation found in the BD results obtained for the current research. Whilst every endeavour was taken to minimise variability during sampling and processing, the consequence of each of these factors affecting BD is a commensurate effect on SCS calculations.

5.13 Soil carbon stock change following waterponding

The scalds have a mean SCS (ESM) to 30 cm depth of 18.7 Mg C ha^{-1} with a range of between 15.7 and 21.7 Mg C ha^{-1}. This is a very low SCS when compared with average stocks found elsewhere in NSW agricultural soil. For example, in a study into SCS of different pasture treatments from central and southern NSW Chan *et al*. (2010) found there was a wide range of between 22.4 and 66.3 Mg C ha^{-1} to 30 cm depth. In a second study Chan *et al*. (2011) found that SCS to 30 cm under different cropping treatments in NSW ranged between 33.6 and 48.0 Mg C ha^{-1}. Even the lowest SOC results found by Chan *et al*. (2010) and Chan *et al*. (2011) were higher than the SCS present in scalded soil. In another study, Murphy *et al*. (2003) found that uncleared Bimble Box grassy woodland on a red kandosol (red earth) on the Cobar Peneplain near Girilambone, a locality with a similar climate and in close proximity to Marra Creek, had approximately 32 Mg C ha^{-1} to 30 cm depth (based on ESM) on the uncleared sites, whereas the paired crop rotated with a lucerne site had 24.9 Mg C ha^{-1}. The results from Murphy *et al*. (2003) suggest that a representative baseline SCS (to 30 cm depth) in the study

region prior to erosion can reasonably be taken to be approximately 32 Mg C ha^{-1}, or almost double that found in the scalded soil.

The results of the waterponding research show there is approximately 29% more SCS in the 0-30 cm depth increment in the five year old waterponds than in the scalds to the same depth. Indeed, of all the age cohorts the five year old waterponds were found to have the highest SCS at the 0-30 cm depth layer with 26.3 Mg C ha^{-1} (based on an ESM).

The ten year old and 25-27 year old waterpond cohorts were found to have a slightly lower SCS than the five year old waterponds with an average of 25.3 and 24.9 Mg C ha^{-1} respectively. The slight temporal downward trend in SCS is not mirrored in the changes to SOC concentration. Instead C concentration results show a temporal increase following waterponding. A probable reason for the lower SCS result in the older waterponds is due to them having a lower soil BD (Section 5.12) than the younger waterponds. The reasons for the changes to soil BD following waterponding have been discussed in detail above, but briefly include the combination of vegetation growth, increased SOM, and formation of soil cracks. To take account of the lower BD in the waterponds, the SCS has been calculated and reported using an ESM equation (Equation 17) based on the concept proposed by Ellert and Bettany (1995).

The relatively constant SCS found between the older waterpond cohorts suggests that the C carrying capacity of the soil has achieved a new natural equilibrium level. The concept of steady state or SOC equilibrium has previously been discussed in Chapter 2 (Figure 2-2). The SOC equilibrium level in waterpond soils is set by plant growth which is constrained by nutrient availability; particularly AP which is rapidly depleted due to water leaching and grazing. Constraints on plant growth from nutrient availability following waterponding is discussed in detail in Section 5.14.

The flow of SCS losses and gains and equilibrium since European settlement is shown at Figure 5-45. Prior to the formation of the scalded soil, the SOC was most likely in a state of equilibrium, with any fluctuations in SCS being caused mainly by natural forces rather than human intervention. Overgrazing and drought were underlying factors that contributed to erosion of the A-horizon from the landscape. A key driver of soil loss is wind erosion associated with frontal weather systems. Westerly frontal systems play a key role in entrainment of surface sediments transporting them off the east coast of Australia (Baddock *et al*. 2015). There is also substantial evidence of massive soil loss following reports of extreme dust storm events and long distance transport of dust. Reports of dust storms, soil loss, sand drifts and associated landscape degradation commenced around 1880 and persisted particularly during the Federation Drought of 1895-1902, the 1920's and the 1930s (Cattle, 2016). As a result there are relatively few undisturbed soils in the region so it is difficult to quantify the SCS that existed prior to European settlement. However, Murphy *et al*. (2003), who conducted research on the Cobar Peneplain Bimble Box Woodlands located at Girilambone, in the same region as Marra Creek,

found the undisturbed soils contained approximately 32 Mg C ha^{-1} to 30 cm depth. Therefore by using their figure as a baseline it is possible that the quantity of SCS lost due to erosion of the A-horizons is in the order of 14 Mg ha^{-1} leaving the new low steady state of approximately 18 Mg SOC ha^{-1} in the scalded soil. After the crust forms on the scald there is little probability of further SOC sequestration without intervention. The new management regime initiated by constructing the waterponds has allowed an increase in SOM and thereby facilitated a commensurate increase in SOC until the new steady state of approximately 26 Mg C ha^{-1} is achieved. When the waterponding results are applied to the Johnson (1995) conceptual model of SOM accumulation it can be concluded that although waterponding has probably not restored SOC levels to the pre-disturbance level it has however, led to SOC sequestration and achieved a new steady state higher than the scalds (Figure 5-45). The dotted line in the new steady state period has been applied due to uncertainty in ongoing retention of SCS beyond the 25-27 year time span of the chronosequence of sites used in this research project.

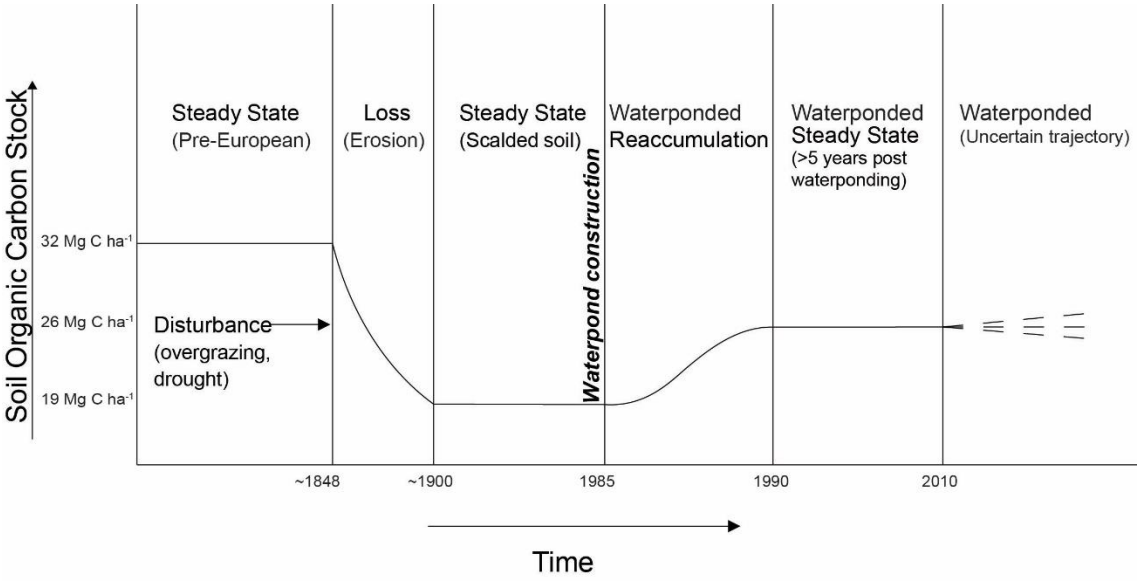


Figure 5-45. Conceptual model of SOC accumulation following waterponding (After Johnson, 1995). The steady state (pre disturbance) SCS figure is based on Murphy *et al*. (2003). The dotted line reflects the uncertainty as to retention of SCS beyond the 25-27 time frame.

5.13.1 Waterponding and C sequestration at depth

This research has shown conclusively that the increase in SOC following waterponding occurs as a direct result of SOM inputs from vegetation growth. Consequently, SOC sequestration typically occurs close to the surface where most plant litter accumulates and roots develop. However, the results also show SOC is increasing at depths below 10 cm (Figure 5-26). This sequestration is beneficial because the C is less exposed to environmental or land management processes that initiate decomposition, and thus the sequestration is more permanent.

The C in the waterponds is potentially being stabilised due to two possible mechanisms:

- SOM interaction with clay minerals thereby becoming physically protected from decomposition; and
- SOM being sequestered at depths below which decomposer organisms can live.

The quantity of SOC sequestered depends on the amount of SOM input into the soil profile and the quantity lost via decomposition (Jobbagy and Jackson, 2000; Lutzow. *et al*. 2006). OM, such as roots and fungal hyphae, form associations with soil mineral particles to form stable aggregates where the C can be physically protected from decomposition (Tisdall & Oades, 1982; Baldock & Skjemstad, 2000; Six *et al*. 2002). In the waterponds physical reorganisation of mineral particles via wetting, drying, dispersion and sufficient shrink - swell behaviour to promote soil cracking (Ringrose-Voase & McClure, 1994) together with the increased SOM each contribute to soil structural change and formation of soil aggregates. Transportation of SOM down the soil profile can also be facilitated by the shrink/swell action of the clay (Skjemstad *et al*. 1998), and as dissolved organic C with water infiltration. Gray *et al*. (2015) also show that the proportion of SCS in the 30-100 cm soil layer of east Australian soils located in dry climatic zones is 54%. This relatively high proportion further highlights the importance of SCS at depth for C accounting purposes. In other research Orgill *et al*. 2017 shows that SOC can be sequestered at depths to 0.5 m following application of nutrients and Newey (2005) found that the rate of decomposition was slower at depths below 50 cm than in the top 10 cm.

The finding of the current research that C may be sequestered at depths to 30 cm has highlighted the extraordinary potential of waterponds to sequester stable C. The mechanisms involved in the sequestration of SOC at depth, and the quantity being sequestered below 30 cm has not been empirically examined by this study, but deserves further research.

5.13.2 Estimated annual rates of change to soil C stock

Estimated rates of SOC sequestration vary depending on baseline SOC content, management, LUC, soil type, climatic variability, clay content, the use of fertiliser or manure, decomposition rates and net primary production (NPP). Prior to erosion of the A-horizons and scald development, it is likely that the baseline CS of native soils in the Marra Creek region were at least 32 Mg C ha^{-1} (Murphy, *et al*. 2003). However, the scalded soil has been measured to have

a mean SCS of approximately 18.7 Mg C ha^{-1}. The low SCS in the scalds indicates that up to 13 Mg C ha^{-1} has been lost from the soil profile. The most likely cause of the loss is from soil erosion that has resulted in the complete removal of the A-horizon soil, or the equivalent of at least 30 cm of soil. The current research has not included a calculation of the historical rate of loss of C. However, it is most likely the majority of the erosion occurred rapidly and most likely in the late 1800's following European settlement and grazing in the district.

Waterponding results in SOC sequestration due to an increase in the presence of water and plant growth. The results of the current research show that in the first five years following construction of waterponds, the new SCS was found to be on average 26.3 Mg C ha^{-1}. This change indicates a mean rate of SCS increase to 30 cm depth of 1.7 Mg C ha^{-1} yr^{-1}. A conceptual model outlining this series of changes is shown at Figure 5-45.

In Figure 5-31 the relative change in SCS over time following waterponding with a mean relative increase in the 0-5 cm depth increment of 81% is shown. The only age/depth to show a relative loss was the one year old waterponds in the 10-20 cm depth increment. All other age and depth interactions resulted in a relative increase in SOC. The waterponding results show the relative change of SOC increases commensurately with time in the 0-10 cm depth. The results also indicate that the five year old waterponds have the most SOC below 10 cm.

Other researchers have identified varying increases or losses of SCS following LUC, depending on the activities undertaken. For example, in a paper summarising potential SOC sequestration rates in the USA, Lal *et al*. (2007) suggest that improved pasture and rangeland management practices in arid/semi-arid grazing lands can effectively sequester between <0.1 and 0.5 Mg C ha^{-1} yr^{-1}. In semi-arid prairie soils VandenBygaart *et al*. (2008) recorded a rate of SCS change from between 0.02 and 0.22 Mg C ha^{-1} yr^{-1} on conversion from conventional till to no-till farming. Freibauer *et al*. (2004) reported potential rates of SOC sequestration following revegetation of abandoned arable land in Europe of between 0.3 – 0.6 Mg C ha^{-1} yr^{-1}. Vagen *et al*. (2005) estimated a potential SOC sequestration rate of 0.05-0.36 Mg C ha^{-1} yr^{-1} in severely degraded sub-Saharan Africa land use systems, but the authors noted that to achieve this rate would require addition of manure, crop residues and no-till practices. Watson *et al*. (2000) suggested that restoration of degraded land could lead to a potential SOC sequestration rate of 0.1 to 7 Mg C ha^{-1} yr^{-1} with a default estimate of 0.25 Mg C ha^{-1} yr^{-1}. These authors estimated it would take around 30 years from commencement of the restoration activity before a new SCS equilibrium could be achieved.

Whilst none of the global scenarios outlined above exactly match the waterponding example, they represent potential SOC sequestration rates in degraded land. Therefore, the SOC sequestration rate of 1.65 Mg C ha^{-1} yr^{-1} achieved by waterponding during the first five years after construction is at the upper end of sequestration potential. The SOC sequestration rates achieved following waterponding support the proposition by Watson *et al*. (2000) that the rate

of SOC gain will be higher in landscapes that have previously been depleted of C. The clay pan scalds are an excellent example of such a landscape. Further, Govers *et al.* (2013) suggest SOC sequestration rates exceeding 1 Mg C ha^{-1} yr^{-1} are most likely only in degraded landscapes previously depleted of SOC by erosion.

5.14 Nutrients, nutrient ratios and their implication for C sequestration in waterponds

All SOM is comprised of C, O, H, N, P and S and is therefore an important reservoir of plant nutrients. This study undertook soil analysis of nutrients essential for plant growth and humus formation including N, P and S. The results show these nutrients are present in adequate concentrations in the scalded soil. However, nutrient concentrations are reduced, and their ratios with SOC change following waterponding. The changes arise because leaching, nutrient uptake by plants and/or changes to decomposition rates occur.

Changes to land management including fertiliser application, tillage and cropping rotations are known to have an impact on nutrient dynamics in soil (Parton *et al*. 1988). However, in the context of the scald/waterpond environment the application of fertiliser is impractical and uneconomical due to the extensive area affected. Similarly, tillage, and other soil disturbance techniques have been shown to be ineffective at permanently restoring scalded soil (Beadle, 1948d,e; Cunningham, 1987; Thompson, 2007). Therefore, any changes to soil nutrient dynamics occurs principally because of physico-chemical modifications to the soil that are initiated following the construction of waterponds, and subsequently through water retention which facilitates plant growth and leads to C sequestration.

SOM is comprised of HS and POM. HS typically accounts for 60-85% of the SOM pool (Sanderman *et al*. 2010). Decomposition is a biological process involving humification and mineralisation of SOM; processes that occur simultaneously and are important for maintaining a supply of plant available nutrients (Lavelle *et al*. 1993). Humification is the process whereby plant residues are transformed into stable forms of organic matter or HS through microbial activity. Humic transformation is important in terms of SOC sequestration because resistant organic polymers can be sequestered in soil and persist there for decades and even centuries, whereas particulate material that is more easily decomposed or mineralised is more dynamic and much less persistent. Mineralisation refers to the transformation of organic forms of C, N, P and S by microbes into inorganic compounds including CO_2, (Zech, *et al*. 1997) and ions of nitrate, phosphate and sulphate. Indeed, decomposition can account for up to 90 % of plant available N and P (Sanderman *et al*. 2010). However, the rate of SOM decomposition is influenced by a number of factors that regulate microbial activity including climate, mineralogy, resource quality, macro and microorganisms, pH and C:N ratios (Lavelle *et al*. 1993; Zech *et al*. 1997). In waterponds decomposition and nutrient cycling are likely to be stimulated because water can be retained leading to increased moisture levels in the soil profile. The improved

moisture level together with microbial activity, pH and nutrient availability can lead to increases in SOM.

The scalded soil was found to have moderate levels of N, P and S in quantities adequate to sustain plant growth. Indeed, Beadle (1948e) found a similar result and suggested that this was because the scalds were made up of B-horizon soils which normally have higher nutrient levels than the A horizon. Using the scalds as a baseline the this research has found that waterponding results in changes (both increases and decreases) to the quantity of SOC, TN, TP and TS and the C:N:P:S nutrient ratios over time although the results show variability. The presence of sufficient soil nutrients including N, P and S and their respective ratios with C can be used to indicate optimum plant growth and equilibrium conditions. From this data it is possible to make predictions of NPP, the production of humus, rates of decomposition and consequently the maximum quantity of SOC that can be sequestered (Himes, 1998; Kirkby *et al*. 2011) by scalded soil. The ideal ratios of C:N:P:S for biomass and SOM production, and humification are typically of the order 1000:83:20:14 respectively (Dalal, 1977; Himes, 1998, Kirkby *et al*. 2011). This research has found the mean C:N:P:S ratios in the scalded soil in the 0-5 cm soil layer to be 1000:125:20:12. The 27 year old waterponds were found to have a mean C:N:P:S ratio of 1000:111:15:13 in the 0-5 cm soil layer.

5.14.1 Total Nitrogen (TN)

Changes to TN are identified here by measuring the difference in concentration between scalded soil and that of waterponded soil. TN analysis provides a measure of all organic and inorganic pools of soil N (Bronson, 2008; Rutherford *et al*. 2008). The most abundant form of N measured by TN analysis is the organic form with typically less than 10 % in the inorganic pool (Bronson, 2008). Measurement of TN is useful for monitoring temporal changes and spatial differences in soil TN, but doesn't provide an indication of the concentration for the individual forms of nitrogen in soil (e.g. N_2O, nitrogen (N_2), nitrate (NO_3^-), ammonium (NH_4^+) and nitrite (NO_2^-)). TN, therefore, is of limited use for determining fertiliser needs to improve productivity, nor for monitoring N_2O emissions.

Analysis of results for all n = 540 scald and waterpond samples shows that in the 0-5 cm soil layer the mean TN concentration in the scalds is 645 mg kg^{-1} (Figure 5-33). The scalded soil showed no consistent pattern in N distribution with depth. The concentration of TN in the scalded soils is higher in the 5-10 cm and 10-20 cm depth layers than at the surface. This is probably because there is insufficient water infiltration to cause leaching of the N from the scald soil profile. Additionally, the trend is commensurate with SOC (Figure 5-26) where there is more SOC in the 5-10 and 10-20 cm soil layers than at the surface, confirming the relationship between SOM, C and N.

Waterponds aged one year have a slightly lower mean TN concentration of 590 mg kg^{-1} in the 0-5 cm layer. The low TN concentration in the scalds and youngest waterponds is most likely the result of the low levels of organic material. SOM is a major source of essential inorganic nutrients for plant growth and N for the soil biological community (Lavelle and Spain, 2005). The absence of SOM is likely to inhibit decomposition and C sequestration processes in the scalds and young waterponds. The lower TN concentration in the young waterponds probably indicates a loss associated with leaching of nitrate (NO_3^-) due to the retention and infiltration of water within the waterponds. There is no evidence that the N is relocated down through the 0-30 cm soil layer in the depth increments examined in the current research in the young waterponds.

The waterponds aged five and ten years have a similar TN concentration in the 0-5 cm layer (mean 715 mg kg^{-1}) and the 25-27 year old waterponds have a mean of TN 831 mg kg^{-1}. The increasing temporal trend suggests that organic N is most likely being added to the waterponded soil from plant residues and decomposing soil organic matter including humic compounds, plant roots and microbial biomass (Manzoni and Porporato, 2009). Other sources include atmospheric deposition and perhaps small quantities from animal manure. Although the organic N in the plant material is initially not available for plant uptake, it is mineralised through microbial and faunal decomposition of organic material (Manzoni and Porporato, 2009) to produce plant available forms of N including NH_4^+ and NO_3^-. However, the rate of organic material decomposition depends on the C:N ratio of SOM (Eyles *et al*. 2015), climate (temperature and moisture conditions), chemical (quality of SOM), physical (clay mineralogy) and biological (micro and macro organisms) factors and their interactions (Lavelle and Spain, 2005). The presence of livestock and their dung and urine may be a further source of soil N on the waterponded soil. However, the management practice of running very low numbers of grazing livestock in this environment suggests this would be a very small source.

In the waterponded soils the results show there is a significantly lower TN concentration with depth ($P = 0.001$). This trend is consistent with the lower concentration of SOC found with depth, because both C and N are sourced from SOM as plant biomass (Bronson, 2008).

The increase in soil moisture content in the waterponded soil is a potential concern as in waterlogged conditions NO_3^- can denitrify to N_2O an important GHG (Giltrap *et al*. 2014; Huang *et al*. 2004; Farquharson and Baldock, 2008). Thus, the increased soil moisture content together with increases to the SOC concentration that occurs following waterponding could potentially stimulate N_2O emissions. The relationship between mineralisation of plant residues and N_2O emissions depends on the C:N ratio (Eyles *et al*. 2015), and Huang *et al*. (2004) identified a negative correlation between N_2O emissions and increasing plant residue C:N ratios. This provides further concern because the scalds and waterponds have relatively low soil C:N ratios (between 6 and 11). These ratios are in a range where decomposition rates (and thus N_2O emissions) can be high. The current research project did not include N_2O emission monitoring

on the scalds and waterponds. However, because of the potential of N_2O emissions to offset the SOC sequestration benefit of waterponding, further research in this field should be pursued.

5.14.2 Carbon to nitrogen (C:N) ratio of scalds and waterponds

The scalds have a C:N ratio of 7:1 which is considerably lower than the optimum of 12 normally required for sequestering SOC (Himes, 1998; Kirkby *et al*. 2011; Kirkby *et al*. 2014) The C:N ratio values of the waterponds is between 9:1 and 11:1 in the surface 0-5 cm and these results are consistent with the ratios found in other studies conducted in semi-arid rangelands, including those of China and Brazil (Yong-Zhong *et al*. 2005; Sousa *et al*. 2012). The reason for the widening of the ratio with waterponding is a consequence of the increase in plant growth on the waterponds which is subsequently converted into SOM through root development and litter incorporation.

The C:N ratio in the scalds ranges between 6 tand7 in the top 20 cm and increases to 8 below 20 cm. The waterponds have a lower C:N ratio with depth and this trend is consistent with other findings where the C:N ratio in most soils is generally found to be narrower with depth (Stevenson, 1959). The C:N ratio for all waterpond ages and all depths ranges between 8 to11.

Research has previously been undertaken to identify an optimum C:N ratio for plant growth. For example Wang *et al*. (1996) found a C:N ratio of between 13.1 and 15.6 on Tasmanian basaltic ferrosols in the 0-10 cm soil layer under previous land uses of mixed forest, forest, and pasture soil. These authors found that ex pine forest soils had a wide C:N ratio and found the soils provided a less than favourable environment for microbial activity. The authors concluded that the narrower C:N ratio in the forest and pasture soils indicate those soils have a higher fertility level and consequently provide a more suitable environment for microbial activity. Oades (1995) found that Australian Krasnosems (ferrosol soil) have a C:N ratio of approximately 14 and described the soils in this context as having high fertility and providing a favourable habitat for soil flora and fauna. Himes (1998) proposed an optimal C:N ratio of 12 was necessary for increasing the humic composition of soil. Finally, Kirkby *et al*. (2011) suggests that an optimum C:N ratio for productive agricultural soils should be in the range between 10:1 and 12:1 (or 1 000:83). That the waterponded soil reaches this ratio in the surface 0-5 cm is an indication that soil functionality is being restored to the landscape as a result of the increase in SOM and thereby allowing optimum humification conditions to prevail.

Because the C:N ratio is less than 12 in both the scalds and waterponds, it is likely that N is not a limiting nutrient for SOC sequestration. The C:N ratio can be used as a crude measure to predict the potential of the scalded soil to sequester SOC. For example the scalds have 4 600 mg kg^{-1} SOC in the 0-5 cm level; the C:N ratio is 8 and the TN concentration is 700 ppm (Table 5-30). If the C:N ratio reaches an optimum of 12, the waterponds have the potential to sequester 8 400 mg kg^{-1} SOC in the 0-5 cm soil layer. The oldest waterponds have 8 000 mg kg^{-1} SOC,

which when compared with the scalded soil indicates they have sequestered 3 400 mg kg^{-1} in the 0-5 cm soil layer over the 27 years since establishment. Consequently, because the potential SOC is higher than the measured SOC, it can be deduced that a nutrient other than N is limiting the sequestration potential of the waterponds.

Table 5-30. SOC sequestration potential of scalded soils based on the availability of N and the C:N ratio.

Soil layer (cm)	*Measured TN (mg kg^{-1})*	*Measured SOC (scald) (mg kg^{-1})*	*Measured SOC (27 year old waterpond) (mg kg^{-1})*	*SOC Sequestration Potential of scalded soil (mg kg^{-1}) based on Kirkby et al. 2011 ratio*	*Actual ΔC after waterponding (mg kg^{-1})*
0-5	700	4 600	8 000	8 400	3 400
5-10	610	4 700	5 800	7 320	1 100
10-20	881	4 600	5 300	10 572	700
20-30	409	4 000	4 900	4 908	900

5.14.3 Total phosphorus

TP is a measure of the concentration of organic and inorganic P in soil. Approximately 50-75% of TP is inorganic and is derived from weathered parent material (Sims and Pierzynski, 2008). Normal ranges vary, however, between 20 and 80 % of P in soil is organic in origin (Dalal, 1977; Sims and Pierzynski, 2008). The variation can depend on soil type (Sims and Pierzynski, 2008). The sources of organic P are plant, animal and microbial detritus (Condron *et al*. 2005). Plants absorb P as inorganic P ions from the soil solution (Holford, 1997). Low levels of TP can restrict plant growth and consequently lead to low C sequestration rates. As the P from the soil solution is taken up by plants, it needs to be replaced to ensure a continued supply for ongoing plant growth (Holford, 1997). If the replenishment does not occur through natural mineralisation processes, application of P by fertiliser is necessary.

Australian soils are typically low in P (Dalal, 1977; Holford, 1997) and the scalded soils are no exception. In the case of agricultural production a TP concentration is considered low at levels less than 200 mg kg^{-1} (Moore, 1998). The results show that the scalds have a mean TP concentration of 157 mg kg^{-1} in the 0-5 cm soil layer. The concentration of TP is lower at depths below 5 cm. The mean TP in the waterponds is lower than the scald with concentrations ranging between 76 and 147 mg kg^{-1} in the 0-5 cm soil layer. The temporal results are variable and show no apparent trend for waterpond age. The reason for lower TP in the waterponds is possibly associated with biomass uptake and mineralisation of the organic P pool. The mineralisation process releases P into solution thereby making it available for the plants to take up and hence lowers TP. Replenishment of P via mineralisation in the waterponded environment is likely to be very low, and due to economic constraints and the physical barriers caused by the banks of the waterponds application of fertiliser is impractical in the waterponding scenario.

Consequently P is, and will continue to be a limiting factor restricting further increases in NPP and SOC in waterponds.

5.14.4 Carbon to total phosphorus (C:P) ratio

The trend for C:TP (C:P) ratio shows that the difference between scalds and waterponds in the 0-5 cm depth widens temporally, but in the soil layers below 5 cm there is variation in the results and no particular trend apparent. The ideal C:P ratio necessary to sequester C as humic C is 50:1; i.e. 20 kg P is necessary to produce 1 000 kg of humic C (Himes, 1998; Kirkby *et al*. 2011). The scalds have a C:P ratio of 48 in the 0-5 cm soil layer and thus potentially have an ideal ratio for C sequestration.

P concentration and the C:P ratio can be used to determine whether TP is limiting the SOC sequestration potential of the clay pan scalds. The data (Table 5-31) show that the scalds have 4 600 mg kg^{-1} SOC in the 0-5 cm soil layer. By using the measured TP concentration from the scald and using the 'ideal' C:P nutrient ratio of 1000:20 (or 50) (Himes, 1998; Kirkby, *et al*. 2011), it is possible to predict the C sequestration potential of the scalded soil. The predicted values have been calculated as 7 850 mg kg^{-1} C in the 0-5 cm soil layer whereas the measured results show the oldest waterponds have 8 000 mg kg^{-1} C in the 0-5 cm soil layer. Thus it can be said that over the 27 year time frame C is being sequestered up to and above the limit of the TP concentration present in the baseline scalded soil. Because the quantity of C being sequestered is in excess of the potential possible based on the TP concentration and the C:P ratio, it is probable that the waterponded soils are obtaining P from another source. Such sources could include aeolian deposition, overland flow or uptake from soil layers below 30 cm by roots. Another source would be fertilisation, but this is not carried out on the waterponded soils.

Table 5-31. The actual and potential C sequestration potential of waterponds given the measured C:total phosphorus (C:P) ratio.

Soil layer (cm)	*Measured TP (mg kg^{-1})*	*Measured SOC (scald) (mg kg^{-1})*	*Measured SOC (27 year old waterpond) (mg kg^{-1})*	*SOC Sequestration Potential of scalded soil (mg kg^{-1}) based on Kirkby et al. 2011 ratio*	*Actual ΔC after waterponding (mg kg^{-1})*
0-5	157	4 600	8 000	7 850	3 400
5-10	98	4 700	5 800	4 900	1 100
10-20	98	4 600	5 300	4 900	700
20-30	94	4 000	4 900	4 700	900

Dust deposition of P from aeolian deposits and overland flow has been calculated to be approximately < 1 kg ha^{-1} yr^{-1} (pers comm, R. Greene 29 Aug 2015). Thus, the most likely source is considered to be uptake of P from lower in the soil profile because plant roots relocate the P through the profile and root hairs increase soil volume and available P (Kovar and Claassen, 2005).

5.14.5 Available phosphorus

The concentration of AP in soil represents the amount of P in solution that is available to plants for their growth (Olsen *et al*. 1954; Tiessen and Moir, 2008). This form of P is highly reactive and readily converted to non-available forms through sorption and precipitation reactions (Guppy and McLaughlin, 2009). P becomes available to plants through desorption or dissolution of inorganic P or through mineralisation of organic P (Tiessen and Moir, 2008). The concentration of AP in the scalded soil is 19 mg kg^{-1} in the 0-5 cm soil layer which is within the moderate range of between 10 and 30 mg kg^{-1} (Moore, 1998). The concentration of AP is quickly reduced, with waterponds aged 1 year having 8.2 mg kg^{-1}. Because there is limited water supply in the subsurface of the scalded soil, the mass flow and diffusion of AP is relatively limited (Kleinman *et al*. 2011). Therefore, relatively moderate levels of AP are retained in the scalded profile. However, following waterponding, AP may be leached from the profile or be taken up by the biomass.

Plants preferentially absorb $H_2PO_4^-$ (Kovar and Claassen, 2005). However, because the scald and waterpond soils have soil pH > 7 it is probable that the dominant P ion in solution is $HPO^2_4^-$ (Sims and Pierzynski, 2005) and in this form is less soluble. Therefore leaching and plant availability of P may decrease as the waterponds age and the pH increases leading to immobilisation of P. Increases in soil BD, such as is seen in the older waterponds can inhibit the diffusive flux of P and thereby add another layer of complexity to the potential of SOC sequestration of waterponding.

5.14.6 Carbon to available P ratio

The C:Avail P (C:AP) ratio shows that as plant growth continues under waterponding the ratio widens. This is because, as the plants grow they become the source of SOM and thus contribute to the SOC pool. However, plants also take up the AP. Because only small quantities of AP can be added to the soil via plant residues, livestock dung and labile organic P, the loss through plant uptake is greater than that which is being supplied. Consequently, low levels of AP is the major factor contributing to the widening of the C:AP ratio. Thus, the depletion of AP in the soil profile following waterponding is the most likely factor contributing to the apparent equilibrium seen in the SOC concentration and SCS results in the oldest waterponds, and will probably continue to be a limiting factor for plant growth over time. Therefore, it is expected that waterponds older than 25+ years are unlikely to sequester additional C.

5.14.7 Total sulfur

The results for the concentration of TS in the 0-5 cm soil layer are variable, showing no particular trend with time since waterponding. However, at depths below 5 cm the scalds generally have a higher concentration of TS than the waterponds with the exception of the 20-30 cm soil layer.

Soil S can be present in soil in both an organic and inorganic form, and continuously cycles between these. The transformation occurs through mineralisation of organic S into inorganic sulfate (SO_4^{2-}), or immobilisation of SO_4^{2-} in SOM (Churka Blum *et al*. 2013). The normal ratio of organic to inorganic S in soil is approximately 95:5. Because the majority of S is present in soil in an organic form changes to concentration are correlated with fluctuations in SOC and TN. The quantity of TS in soil depends on the presence of SOM, parent material, fertiliser application and atmospheric deposition (Kovar and Grant, 2011). The organic form of S is present in organic residues, microbial biomass and humus and it is not normally present in a form available to plants until it is mineralised (Nguyen and Goh, 1990; Kovar and Grant, 2011). Optimum conditions for mineralisation occur when the pH is > 6, soil water content is > 60% of field capacity and when the soil temperature ranges between 20 and 40^{o}C (Eriksen, 2009: Kovar and Grant, 2011). The important pool of S for plant growth is the inorganic pool. Plants absorb inorganic S that is present in the soil solution as SO_4^{2-} through their root systems. The concentration of inorganic S in the soil profile varies throughout the year depending on seasonal conditions, fertiliser application, plant uptake and leaching. Low SO_4^{2-} concentration is typical in winter and spring due to leaching, plant uptake and low temperatures (Eriksen, 2009). Plants can also utilise atmospheric SO_2 that has been metabolised as plant residue. In soils where the pH is greater than 6.5 the majority of SO_4^{2-} is available in the soil solution (Scherer, 2009). Consequently, because both the scalds and waterponds have pH > 7.0 it is probable that most SO_4^{2-} in the soil is available for plant growth.

5.14.8 Carbon to sulfur ratio

The results show that waterponding leads to a widening of the C:S ratio at all soil depths measured, although there is no apparent linear increase with time. The C:S ratio is lower with depth.

The scalds have a C:S ratio of 83 in the 0-5 cm soil layer, and the five year old waterponds have a ratio of 122. Globally, the mean C:S ratio for soils identified by Kirkby *et al*. (2011) is in the range between 54 and 132. Himes (1998) demonstrated that a C:S ratio of 70 (1000:14) can contributed to an increase in humus of 0.7 %.

By using the C:S ratio of 70 (Himes, 1998; Kirkby *et al*, 2011) it can be predicted that the scalded soil which has a C:S ratio of 82, has the potential to sequester up to 5 110 mg kg^{-1} C in the 0-5 cm soil layer (Table 5-32). However, because only 3 400 mg kg^{-1} has been sequestered in the 27 year old waterponds it is probable that TS is not the limiting nutrient for SOC sequestration in the waterponds.

Table 5-32. C:S ratio used to predict potential C sequestration of scalded soil following waterponding

Soil layer (cm)	*Measured TS (mg kg^{-1})*	*Measured SOC (scald) (mg kg^{-1})*	*Measured SOC (27 year old waterpond) (mg kg^{-1})*	*SOC Sequestration Potential of scalded soil (mg kg^{-1}) based on Kirkby et al. 2011 ratio*	*Actual ΔC after waterponding (mg kg^{-1})*
0-5	73	4 600	8 000	5 110	3 400
5-10	226	4 700	5 800	15 820	1 100
10-20	253	4 600	5 300	17 710	700
20-30	253	4 000	4 900	17 710	900

5.14.9 Nutrient availability summary

The nutrient concentration in the scalded soil can explain the increase in SOC concentration in the younger waterponds. The results suggest that as waterponds age the availability of P and N reduces, and the ratios for P, N and S all widen when compared with scalded soil. This suggests that after the waterponds are established and vegetation begins to grow, the nutrients are taken up by the plants. The only sources of soil nutrient replenishment are plant residue decomposition, stock dung and weathering of parent material. Each of these sources contributes minimal amounts of nutrient and what is supplied is not necessarily in a form immediately available to plants. Consequently, nutrient depletion may become a concern for the long-term sustainability of the landscape.

5.15 Sodicity, aggregate stability and ESP changes after waterponding

The scalded soil is both saline and sodic whereas the waterponded soil is sodic but no longer saline because soluble salts have been leached from the soil profile. A consequence of this difference is the propensity for aggregates in the scalded soil to be stable but disperse in the waterponded soil. In Figure 5-46 a summary of the differences in EC, ESP and pH between the scalded and waterponded soil are shown. How the changes occur, their inter-relationship and their influence on soil aggregate stability is discussed in this section.

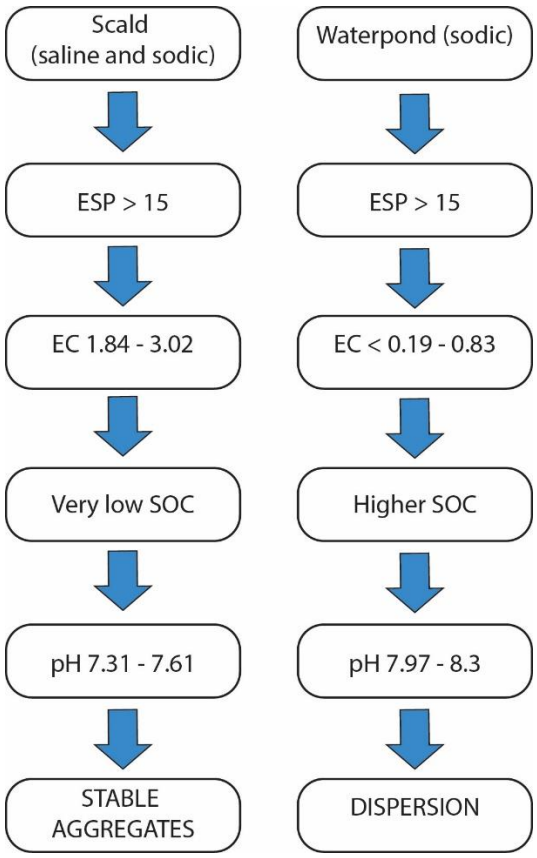


Figure 5-46. The changes to EC, pH, ESP and aggregate stability following waterponding.

The scald profile included in Subset 2 was found to have an ESP of 13 in the 0-5 cm soil layer increasing to 32 in the 20-30 cm layer (Table 5-19), where ESP refers to the concentration of adsorbed sodium (Na^+) in soil. In Australia, soils with an ESP > 6 are classified as sodic meaning they are potentially dispersive (Figure 5-47) (Rengasamy *et al*. 1984). Dispersion occurs in sodic soil because they have a high concentration of exchangeable Na^+ that results in a thick diffuse double electrical layer in clay particles. The electrostatic repulsive energy within the double layer of the clay particles is greater than the attraction force (Sumner, 1993). The thickness of the double layer is dependent on the concentration and valency of ions present in the soil solution. Because the negatively charged double layer clay particles have a weak bond

with the Na^+ when water is added, such as occurs after waterponding, the ions become hydrated and the clay particles are forced (dispersed) apart (Sumner, 1993).

Despite the scalded soil having an ESP > 6, the results show the scalds have an ASWAT score of 2 in the 0-5 cm soil layer and 0 at depths below 5 cm (Table 5-20). This means the scalded soil is sodic but shows no evidence of dispersion. The $EC_{(1:5\ water)}$/ESP results showing the dispersive behaviour of both the scald and waterpond cores (0-5, 5-10, 10-20 and 20-30 cm depth increments) sampled as Subset 2, has been overlayed onto the model proposed by Rengasamy and Olsson (1991 (Figure 5-47).

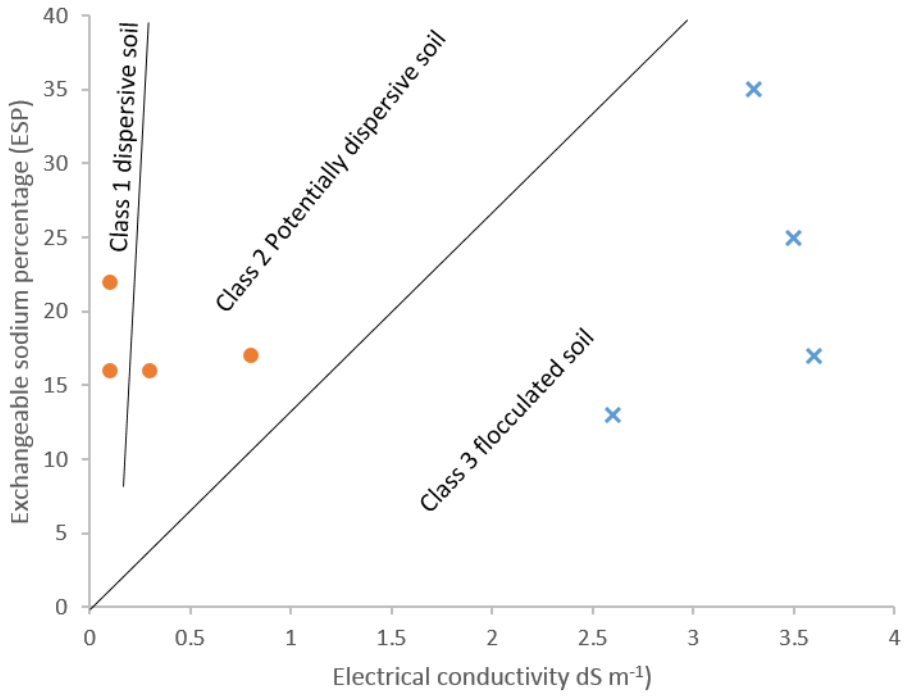


Figure 5-47. Dispersive behaviour of scalds and waterponds in relation to electrical conductivity (EC) and exchangeable sodium percentage (ESP). × = scalded soil, • = 27 year old waterpond (after Streeton *et al*. 2013)

The results show that the scalded soil fits within the Class 3 flocculated soil parameters. Mg is the dominant cation in the scalded soil, ranging from 45% at the surface down to 37 % in the 20-30 cm layer (Table 5-21). The Ca:Mg ratio is 0.8 for all soil layers in the scald. Normally soils that have Mg as the dominant cation tend to disperse. However, the reason the scalded soil doesn't disperse is due to the high soluble salt content (EC >2.5 dS m^{-1}) throughout the 30 cm profile. Effectively, the high soluble salt content in the scalds counteracts the effects of sodicity (ESP >6) and therefore promotes flocculation and stable aggregates (Gupta *et al*. 1984; Rengasamy *et al*. 1986; Gupta and Abrol, 1990; Rengasamy and Olsson, 1991).

The waterpond soil has samples that sit within both Class 1 (dispersive) and Class 2 (potentially) dispersive parameters. However, all the waterpond soil samples from this subset were found to disperse spontaneously with ASWAT scores of between 9 and 10. The reason why two of the samples sit within the Class 2 category (potentially dispersive) although they were found to disperse spontaneously is associated with the percentage of Mg and the Ca:Mg ratio of the soils. Firstly, to differentiate the samples from Class 1 and Class 2, the two Class 2 samples are from the 10-20 and 20-30 cm soil layer and have a soluble salt content of 0.3 and 0.8 dS m-1 respectively. The soluble salt content is lower in the 0-5 and 5-10 cm layers with an $EC_{(1:5\ water)}$ of 0.08 and 0.09 dS m-1 respectively. The Ca:Mg ratio for both of the potentially dispersive samples is 1.1 whereas the ratio is higher for the 0-5 and 5-10 cm layers. The ESP of the two potentially dispersive samples was found to be 16 and 17 respectively. The percentage of Mg in the waterponded soil is 30, 36, 38 and 36 for the 0-5, 5-10, 10-20 and 20-30 cm layers respectively. When the percentage of Mg in soil increases, the propensity for dispersion in sodic soils also increases (Rengasamy *et al*. 1986). So, whilst the ESP for both the scald and the waterpond is within the dispersive range, the Ca:Mg ratio and the percentage of Mg all sit within a dispersive range, the high concentration of electrolytes in the scalded soil causes flocculation, thereby overcoming the dispersive qualities of the scalded soil.

Soil aggregates in the youngest waterponds were found to disperse after disturbance, which was simulated by remoulding the aggregates. Undisturbed aggregates were found to disperse spontaneously in the five and 27 year old waterponds. The ASWAT scores were 6.2 in the 0-5 cm layer of the one year old waterponds and 9.5 in the 27 year old waterponds. Field *et al*. (1997) demonstrated that an ASWAT test score of 6 was approximately equivalent to an ESP of 6 ($ESP = 1.07(ASWAT) + 0.99$, $r^2 = 0.54$). However, the authors noted other factors including $CaCO_3$ content, CEC, clay content EC and OM may influence the ESP of soil. This point is also made by Murphy (1995). The dispersion in the waterponds was found to occur only after the soluble salts had been leached from the soil profile (Figure 5-47). However, even after leaching the waterpond soil remains sodic having an ESP of ≥ 15 at all depth increments. Therefore the waterponds fit the classification by Rengasamy *et al*. (1984) of Class 1 soils which are subject to spontaneous dispersion (Figure 5-47).

Clay swelling and dispersion is normally considered to be a detrimental soil characteristic associated with soil structural degradation following aggregate breakdown (Oster *et al*. 1999; Wong *et al*. 2010). Dispersion can result in surface crusting due to rearrangement of clay particles at the soil surface and hence cause a deterioration of soil hydraulic conductivity and water infiltration. However, in the case of waterponding dispersion is an important and beneficial feature. The soil physical changes, especially at depth (i.e. the redevelopment of shrink-swell behaviour of the clays at depth) following the reorganisation of soil colloids allows the surface crust to break down and then subsequently form deep cracks in the soil profile.

These changes transform the surface and soil profile, allow moisture to penetrate into the profile and thus create an environment suitable for seed to establish and plants to grow.

Given the higher concentration of SOC and higher plant biomass in the waterponds the dispersive characteristic of the waterponded soil is somewhat counterintuitive. Typically, clay dispersion is negatively correlated with increasing SOM content (Dong *et al*. 1983; Nelson *et al*. 1999; Bronick and Lal, 2005) because transient SOM in the form of fine roots and hyphae bind clay particles into stable aggregates and thereby prevent dispersion (Tisdall and Oades, 1982; Qadir and Schubert, 2002). However, SOM does not always lead to aggregate stability. For example, Visser and Caillier (1988) found that at low concentrations (approximately 40 mg l^{-1}) humic substances can contribute to dispersion. The contribution of the SOM to form aggregates depends on the type of OM, whether the soil has achieved SOC saturation, and physical changes following disturbance which can influence whether increases in SOM result in a commensurate increase in water-stable aggregates (Tisdall and Oades, 1982; Qadir and Schubert, 2002). Churchman *et al*. (1993) further noted that dispersion was possible in soil that had low levels of OM when Na^+ was present on the exchange sites. This is because Na^+ is an ineffective bridging cation and easily be solvated by water molecules (Rengasamy and Olsson, 1991).

The dispersive action of the sodic waterponded soil may result in either a net positive or negative effect on the SOC sequestration potential. For a positive effect to occur the surface crust of the scald needs to be broken down. Next, the shrink-swell effect and the dispersive function needs to be restored to allow vertical cracks to develop and thereby allow water infiltration down through the profile to leach the soluble salts. These changes create an environment suitable for plant establishment and subsequent SOC sequestration. As is apparent from the results there is an increase in SOC in the waterponded soil. The rate of increase is rapid within the first 5 years, but reaches an apparent equilibrium, with little, if any change to SOC concentration in subsequent years. This could be due to the soils reaching a C equilibrium between plant inputs and decomposition and therefore no longer having the capacity to sequester additional C (Chung *et al*. 2008; Orgill *et al*. 2017).

Dispersion may also contribute to increased rates of SOC mineralisation due to the disturbance it causes to soil aggregates. Consequently with each wetting and drying cycle it is possible that some of the newly sequestered SOC is mineralised, thereby having a negative effect on SOC sequestration. Mineralisation can also be increased due to increasing alkalinity found in the older waterponds. For example Laura (1976) suggested that higher alkalinity enhances the availability of organic matter to microbial activity due to the dissolution, dispersion or chemical hydrolysis of the SOM. Also, soils with a high ESP such as occurs in the waterponds may have high rates of OM mineralisation (Laura, 1976). Consequently higher rates of dispersion, alkalinity and ESP may all be constraining factors that prevent the older waterponds sequestering additional SOC.

5.16 Mineralogical changes to scalded soils following waterponding

The overarching aim of this thesis has been to examine the potential of scalded clay pan soil to sequester C following waterponding. Soil mineralogy plays an important role in the potential of the soil to sequester C (Churchman and Lowe, 2012) and C turnover time (section 2.4.3). For example, Hassink (1997) found that silt and clay particles provided protection for C against microbial attack. In other research Torn *et al*. (1997) found that non-crystalline or amorphous minerals such as Fe and Al oxides are associated with a higher abundance of stabilized C than crystalline minerals. In addition, the clay mineralogy influences the shrink-swell properties and the cation exchange capacity of soil (Ross, 1978). Consequently an objective of the current research has been to attempt to identify any mineralogical changes that may have occurred to the soil following waterponding and whether those changes influence the C sequestration potential of the soil. This brief analysis is a first attempt to explore whether there is any mineralogical change associated with waterponding, and if so, whether there is a relationship between the change and the potential influence on SOC sequestration.

To investigate mineralogical change, the composition of the paired scald and a 25 year old waterpond core (Subset 2) at four depth increments was analysed. PSA was undertaken to assess differences in soil particle size between the scald and waterponded soils. PSA results have been expressed as a volume weighted mean for each particle size group (clay, silt and sand). XRD analysis was used to ascertain differences in the mineralogical composition of the two soil cores. ICP-AES analysis was used to identify elemental oxides present in the two cores and ECEC was used to detect differences in the cation composition. Finally ^{137}Cs analysis was undertaken to explore the origin of the vegetated hummocks that are present in the clay pans on the scalded soil.

The important mineralogical differences, possible explanations for the differences and their impact on the potential of the scalded soil to sequester C following waterponding are discussed below.

5.16.1 Particle size analysis

Scalds

When the scald samples were exposed to mild dispersion (in water only) during PSA, the volume weighted mean particle size was found to decrease with soil depth (Figure 5-42a) (e.g. the mean size in the 0-5 cm layer was 76 µm and 43µm in the 20-30 cm layer). The majority of particles (be they micro aggregates or individual mineral particles) in the mildly dispersed samples were found to be in the silt (2-63µm) and sand size range (> 63µm) (Figure 5-43a). This trend with depth is consistent with duplex soils such as occurs in this study. Soils typically

have more clay present lower in the soil profile than at the surface. This occurs due to illuviation of the fine clay particles through the profile.

Following treatment with $NaPO_3$ and ultrasonic vibration to stimulate full dispersion of soil aggregates there was a large decrease in the volume weighted mean diameter from, for example in the 0-5 cm depth layer, 75µm to 23µm. The fully dispersed sample results also showed very little difference in the volume weighted mean particle size between the four depth increments in the scald soil (range between 22.6 and 23.7 µm) (Figure 5-42b). The average volume weighted mean diameter for the four depth increments in the scald core is 23.2 µm, with the proportion of clay to silt and sand sized particles (3.8%, 90.1% and 6.1% respectively) (Figure 5-43b), with particles in the silt size range being dominant. The smaller volume weighted mean particle size after full dispersion during the PSA process means that aggregation was strong in the scalded soil.

The aggregation in scalded soil could be associated with inorganic or organo-mineral associations. For example, Tisdall and Oades (1982) show that organo mineral associations are related to aggregates < 250µm. These authors explain that aggregates in the 20-250µm size range are water stable due to both their small size and binding by organic material, oxides and aluminosilicates. The current research demonstrates that the scalded soil is indeed stable and not susceptible to dispersion, with the likely cause of the stability being the presence of a high concentration of soluble salt. Tisdall and Oades (1982) found that aggregates in the 20-250µm size range can be broken down when exposed to ultrasonic vibration and the current PSA results confirm that such a breakdown has occurred. It is possible that $CaCO_3$, which is known to limit clay dispersion (Tisdall and Oades, 1982), may also be the primary cementing agent for the current scald soil particles. $CaCO_3$ nodules can be seen deeper in the soil profile (Ringrose-Voase and McClure, 1994).

If particle binding in the scalded soil is assumed to be associated with organic agents then it is most likely the aggregation occurred prior to the loss of the A-horizon and was in association with organic material deposited by vegetation. Thus, the aggregate binding in the scalds most likely has been caused by the presence of persistent organic material that become encapsulated by clay particles in the past. This explanation correlates with the presence of, albeit a small concentration, of SOC present in the scalded soil. Because the scalded landscape has little or no vegetation to resupply fresh SOM, the SOC that is present has probably been in the soil, undisturbed for a long period of time. With further analysis by using fractionation techniques, XRD and scanning electron micrographs (SEM) such a hypothesis could be confirmed. Unfortunately resources to undertake fractionation and SEM analysis were not available for the current research project. However, an understanding of the origin and stability of the C in the scalded profile would be useful for estimating the SOC content of this soil in carrying out National Carbon Accounting (NCA). These investigations would also be helpful in determining

the potential of the soil to continue to sequester SOC and for it to be sequestered in the resistant pool.

Waterponds

An important difference between the scalds and waterponds is the resumption of the wetting and drying cycle following waterponding. These cycles, and dispersion as a result of salts being leached to lower in the profile, can accelerate aggregate disturbance and breakdown, and therefore may be responsible for some SOC loss from the landscape (Skjemstad *et al*. 1998) in the early years following waterponding, particularly until plant growth occurs and NPP increases (Luo *et al*. 2010). There is evidence that structural change occurs following waterponding. For example, in an examination into the fabric changes of soil following waterponding Ringrose-Voase and McClure (1992) identified a nutty, sub-angular blocky soil structure in scalded soil that was modified following waterponding into a prismatic structure. These authors used SEM to show reorganisation of the clay fabric. They found that quartz grains were embedded in clay in the scald profile, whereas the clay matrix of a 22 year old waterpond had clay coated quartz grains with pore spaces between the grains. Ringrose-Voase and McClure (1992) also found that the structural changes had been initiated by dispersion that followed after soluble salts were leached from the waterpond soil profile.

The PSA results show that there is a higher percentage of aggregates comprising sand sized particles in the waterponds than the scalds in both the mildly and fully dispersed samples (Figure 5-43). Additionally, when the full dispersion treatment with a dispersant $(NaPO_3)_6$ and ultrasound are applied to the samples the percentage of particles in the sand fraction size range (> 63 μm) is decreased from approximately 50% to 10% and the percentage of silt particles increases from approximately 50% to 85%. This trend is apparent in both the scald and waterpond samples. Full dispersion also results in a slight increase in clay sized particles in both the waterponded and scalded soil.

Disturbance by wetting and drying processes and clay dispersion resulting from leaching of salt are typically associated with aggregate breakdown. Consequently it was expected that the waterponds would have a smaller number of aggregates than the scalded soils. However, the PSA results show that when compared with the scald, the 25 year old waterpond has a larger, but not significant ($t = 0.16$, $df = 6$) volume weighted mean particle size in the mildly dispersed samples (Figure 5-42). The increased mean particle size indicates that aggregation is occurring in the waterponds. The result also corroborates the XRD analysis (Table 5-25) showing a higher percentage of quartz, kaolinite, and illite in the waterponds than scalds. However, due to the small data set used the results cannot be used to predict how long after waterponding such changes occur. Consequently, additional PSA is recommended to identify any temporal trend in the relationship between waterpond ages, NPP and particle size as a surrogate for aggregation.

There are two possible explanations for the different mean particle sizes between the scald and waterpond. Firstly, as previously discussed in relation to the scalded soil, the process of organo-mineral particle binding could be occurring in the waterponded soil due to the resumption of plant growth promoting aggregation. A second explanation could be the action of saltation whereby particles are transported by wind across the surface and become trapped within the waterponds. These two explanations are discussed below.

It is widely reported that organo-mineral particle binding can occur due to the presence of vegetation especially in the top few centimetres of the soil profile where root growth occurs. For example, the possibility of particle binding is consistent with the findings of Tisdall and Oades (1982) whose conceptual model of the soil aggregate hierarchy demonstrated the important role of organic matter in the formation and stabilisation of aggregates. Fine roots and fungal hyphae have been found to enmesh soil particles and transient binding agents such as microbial polysaccharides which become persistent due to their interaction with clays (Tisdall and Oades, 1982). Several other authors provide evidence to support the Tisdall and Oades (1982) aggregate hierarchy model (e.g. Cambardella and Elliott, 1992; Six *et al.* 1999). These researchers have shown that when agricultural soils are not disturbed by cultivation the abundance of stabilised aggregates increases due to root and hyphae binding mechanisms.

The second explanation refers to deposition of sand grains in waterponds. On a windy day there is plenty of visual evidence that particle movement is occurring across the scalded clay pan surface (Figure 5-48).

Figure 5-48. Dust swept up from scalded surface (Photo R. Greene).

Saltation is the transport of particles across a landscape via wind or water and depositing them at a new position on the surface. Saltation ultimately results in the permanent loss of fine

material containing nutrients and organic matter from the landscape (Shao *et al*. 1996). From the history of erosion in this landscape already discussed (Section 5.2), the scalds are particularly susceptible to wind erosion due to the lack of protection from vegetation, lack of moisture and micro-topography. Waterponds, their banks, soil cracks and vegetation provide structures and material that can entrap sand being transported by creep or saltation (Zobeck and Van Pelt, 2011). Consequently, saltation across the scald surface and entrapment of particles in the waterponds is another feasible explanation for the larger volume weighted mean diameter of particles, and larger abundance of stable sand sized particles found in the waterponded soil than the scalded soil. The XRD analysis results show there is a higher abundance of quartz material in the waterpond core than the scald core (Table 5-25). Thus, the quartz may be saltated material trapped within the waterponds. However, this movement and deposition does not imply that a new A-horizon is forming.

The PSA results show that waterponding leads to changes, although not significant, in the mean particle size distribution of the soil profile. In relation to the research objective to determine whether mineralogical change following waterponding is influencing the C sequestration potential of soil it can be inferred from the higher proportion of sand sized particles that aggregation may be occurring in association with plant growth. In which case, it is probable that clay binding mechanisms may be increasing the abundance of resistant C in the soil. This project has not had the resources to explore this hypothesis experimentally, but is certainly a matter that deserves further exploration in terms of soil C accounting.

5.16.2 Mineralogical differences between scald and waterpond soil

Forbes *et al*. (n.d.), describe the geology of the region as comprising mainly Quaternary alluvium. They explain the provenance of the alluvium is variable being derived from a wide range of parent materials including granites and Palaeozoic metasediments originating from the Molong Rise and Hill End Trough and from areas of Tertiary volcanics. Earlier in this chapter, physical evidence present in the scalded landscape was used to demonstrate how the A-horizon may have been totally eroded by wind and water erosion. Consequently, the mineralogical analysis of the soils analysed in this case study has been undertaken on what is effectively the upper 30 cm of the remaining B-horizon soil formed during the Quaternary.

To establish a baseline mineralogical composition of the soil an XRD analysis was undertaken for each of the four soil increments (0-5, 5-10, 10-20 and 20-30 cm) from the scald core. Analysis of an adjacent 25 year old waterpond core has been used to ascertain whether there may be a mineralogical difference between the scald and waterpond and thus determine whether waterponding may change the mineralogy of the scalded soil. From these results the effect of waterponding on the C sequestration potential of the scalded soil has been explored.

In earlier research, Ringrose-Voase *et al*. (1989) performed XRD analysis to identify the mineralogy for two depth increments of one scald core only. Those authors found the scald profile included roughly equal quantities of illite, kaolinite and smectite (range 21-30% each in the 5.5-16.5 cm layer) and a lesser quantity of quartz (range 11-20% in the 5.5-16.5 cm layer). This current study is the first to compare the mineralogy of a scald with that of an adjacent waterpond.

The mineralogical composition for the four depth increments of the scald and waterpond soil cores examined in the current research is summarised (Figure 5-49). XRD results for the scald showed quartz was dominant, followed by illite interlayered with 14Å clays and 10Å illite, kaolinite, plagioclase, and smectite + chlorite. (Table 5-25). These results of this research differ slightly to those found by Ringrose-Voase *et al*. (1989). In the current research a higher percentage of quartz and less kaolinite and smectite was identified. The percentage of illite was similar to that identified by Ringrose-Voase *et al*. (1989). When XRD mineralogy of the scald core was compared with the waterpond core a number of differences were identified. For example, the scald was found to have a larger quantity of calcite, illite interlayered with 14 Å clays, smectite + chlorite, and halite, a slightly larger quantity of plagioclase, but have slightly less quartz, haematite, kaolinite, 10Å illite than the waterponded soils. Smectite + vermiculite was absent in the scalds. XRD analysis identified little difference in the percentage of K-feldspar and gypsum between either the scald or waterpond cores. Quartz, 10Å illite, illite interlayered with 14 Å clays and kaolinite were the most abundant minerals contributing 90 % of the mineralogical composition of the scalded soil and 98 % of the waterponded soil in the 0-5 cm soil layer.

ICP-AES analysis shows that the most abundant oxides contributing approximately 83% of the composition of both the scald and waterpond samples were SiO_2, Al_2O_3 and Fe_2O_3. Minor chemical differences include a greater abundance of CaO, MgO and Na_2O in the scalds than in the waterponds (Table 5-26). There was little difference found between scalds and waterponds for the other elements analysed.

Results of the cation analysis found Ca^{2+} and Mg^{2+} to be the dominant exchangeable cations in the four soil layers between 0-30 cm in both the scalds and waterponds. The decreasing concentration with depth of Ca^{2+}, K^+ in both the scalds and waterponds (Table 5-21) may be explained by eluviation, as can the decrease in Mg in the scalds and Na^+ in the waterponds. However, Mg^{2+} was found to increase slightly with depth in the waterponds but this can also be explained as an eluvial process. Na^+ was found to be higher with depth in the scalds, which is most likely an illuvial transfer that has occurred over a long period of time.

The measured ECEC results show there is very little difference for the whole soil fractions between the scald and waterpond core at each of the four depth increments (Table 5-21). The scalds have a whole soil ECEC of 25.1 in the 0-5 cm layer and the waterponds 25.3 $cmol^+ kg^{-1}$.

The ECEC results measured on the whole soil samples both for the scald and waterpond samples appear to be relatively even and both show a slight decrease with depth (Table 5-21). In the 0-5 cm layer for example, both the scald and waterpond were found to have an ECEC of approximately 25 $cmol^{+}\ kg^{-1}$, but at the 20-30 cm layer the ECEC was approximately 22 $cmol^{+}kg^{-1}$ for both the scald and waterpond soils. The ECEC results found in the current research were found vary in some cases, and be roughly equivalent in others, to the CEC range of between of approximately 9.7 and 30 $cmol^{+}\ kg^{-1}$ for similar depth increments found in earlier work by Ringrose-Voase *et al*. (1989b).

Previously, Greene *et al*. (2002) undertook an analysis to compare measured CEC with the CEC predicted by using the sum of the CEC of the clay fraction present in soil (as analysed by XRD). Those authors determined clay mineralogy could provide a realistic explanation for CEC results. The predicted CEC for the current research using the clay mineralogical CEC values applied by Greene *et al*. (2002) over-estimated the CEC for the scald and waterpond samples by approximately 50%. Consequently, a further calculation was undertaken to predict the ECEC for the current research based on the mineralogy of the clay fraction (estimated to be 35%) of the soil sample and using the cation values of 100 for smectite, 20 for illite and 4 for kaolinite. Using these values, the mean predicted ECEC was 38.6 $cmol^{+}\ kg^{-1}$ for the scald and 33.5 $cmol^{+}\ kg^{-1}$ for the waterpond (Table 5-24). The final predicted ECEC results show an over prediction (approximately 40%) for the scalded soil (approximately 30% for the waterpond soil) implying that XRD analysis is possibly underestimating the clay mineral composition. Even after the clay mineralogical CEC calculations were applied the predicted results for the current research for the scalded soil were found to be consistent with the measured results (range 38.7 – 41.7) reported by Ringrose-Voase *et al*. (1989b).

Although the difference in measured ECEC between the scald and waterpond samples is slight, some differences have been identified in the mineralogy and chemical composition of samples from different depths. Some of the differences can be attributed to plant growth, leaching or weathering that occurs following waterponding. Each of the mineralogical differences and explanations for the underlying reasons for the differences is summarised in Figure 5-49 and discussed in detail as follows.

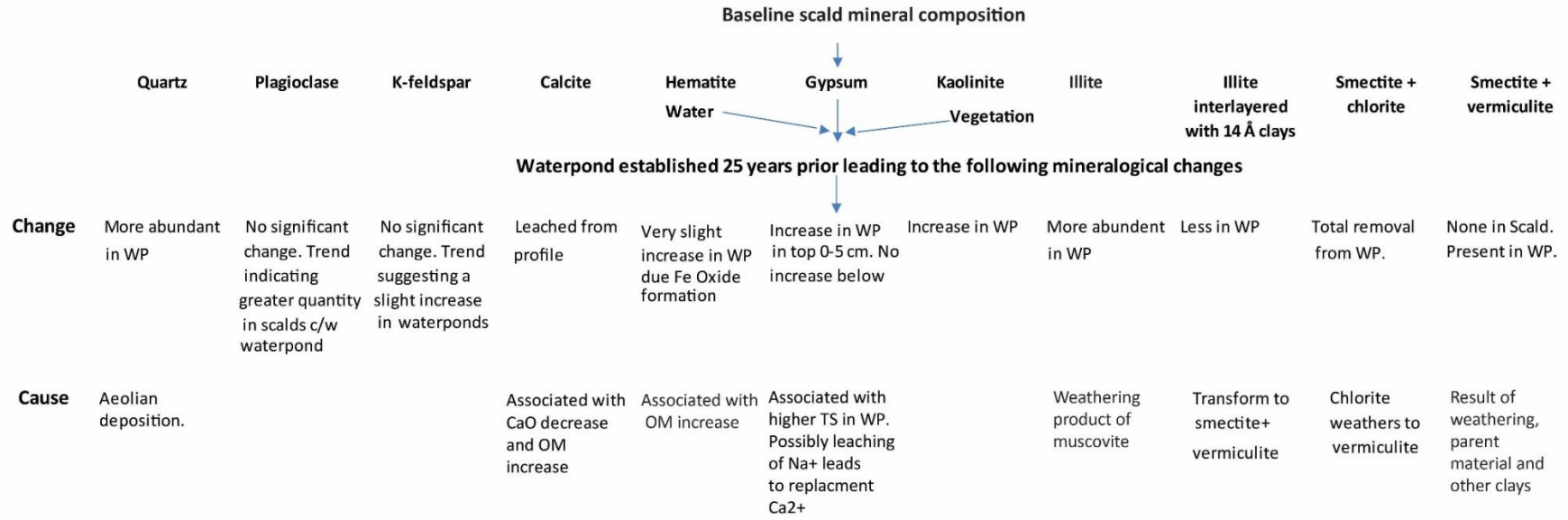


Figure 5-49. Overview of mineralogical changes to scalded soils following waterponding.

Quartz

XRD analysis shows that quartz makes up approximately 40% of the scald (38%) and waterpond soil (43%) with the waterpond having a slightly higher percentage. The abundance of quartz slightly increases with depth in the scald, but is more variable in the waterponded soil. The ICP-AES oxide analysis for SiO_2 shows similar trends to quartz. These results are corroborated by the larger particle size profile of the waterponds identified using PSA (paragraph 5.16.1), and the ^{137}Cs analysis. PSA of the waterponds, particularly in the 0-5 cm soil layer of the mid and wall positions were found to have a volume weighted mean of 34 and 37 μm respectively.

Quartz is readily transported by wind and water across the landscape. Walker *et al*. (1988) state that sand particles < 100 μm are associated globally with loessial deposits, and that particles in the 62-31 μm size fraction have been associated with parna material located at Wagga Wagga in NSW. Therefore, it is possible that the higher quantity of quartz present in the 0-5 cm layer of the waterponds is associated with aeolian transportation of quartz grains, entrapment by plants and lodgement within soil cracks. Further characterisation of the quartz using $\delta^{18}O$ could provide further evidence of the aeolian origin of the material (Chartres *et al*. 1988).

The presence of the higher quantity of quartz in the waterponds has occurred relatively rapidly; i.e. in the 25 years since the waterponds were constructed. However, given the development of soil cracks, vegetation growth and the waterpond bank itself, it is not surprising that some textural change is being detected.

Illite

After quartz, the next most abundant mineral present in the scald or waterpond soil, identified by XRD. Illite has been identified by the 10Å peak. Illite interlayered with 14 Å clay was also identified. In total, illite, makes up approximately 39% of the mineral composition of the scald in the 0-5 cm layer and only 31% of the waterpond soil. The slightly lower percentage of illite in the waterponds suggests that illite is broken down as a result of waterponding.

Illite is a common soil clay mineral in arid environments originating from the weathering of micas including muscovite (Velde and Meunier, 2008; Churchman and Lowe, 2012) and typically having a ECEC of between 10-40 $cmol^{+}\ kg^{-1}$ (McQueen and Scott, 2008). Illite, a secondary mineral, can be transformed into vermiculite in association with vegetation (Koch *et al*. 1992; Velde and Meunier, 2008) or become interstratified with illite/smectite (hence the interlayered composition of illite detected in the current research). XRD analysis provides evidence that smectite+vermiculite is not present in the scalded soil but is present in the waterponded soil (abundance 4.2%). It is probable that some of the illite has been transformed into smectite+vermiculite and may be associated with vegetation which can be a weathering agent through the alteration of trioctahedral micas (Churchman and Lowe, 2012). This type of

weathering transformation is possible in the waterpond soils following the growth of plants but the extent of the alteration depends on drainage and nutrient depletion caused by plants (Hughes *et al*. 1994). The transformation is possible because H^+ from plant roots can both degrade clay minerals and decrease K^+ in illitic interlayers and concentration in the soil solution (Tributh *et al*. 1987). The hypothesis that weathering is a feasible explanation for the lower quantity of illite interlayered with 14Å clays and the increase in smectite in the waterpond is supported by the ECEC results where there is slightly less K^+ in the waterponds than the scalds in the 0-5 and 5-10 cm layers (Table 5-21). Thus, it is feasible that the K^+ has been removed by plant uptake or leaching following waterponding.

Illite is an expandable clay (Churchman and Lowe, 2012) hence its presence is important because the shrink-swell nature of the scalded soils is a vital process that makes waterponding an effective method to restore scalded soils and thus sequester SOC.

Kaolinite

Kaolinite is a stable 1:1 lattice clay that is formed through the weathering of primary rock minerals including plagioclase feldspar (Chartres *et al*. 1988; Nesbitt and Markovics, 1997) and mica, but only in humid, acidic conditions. Because kaolinite is a stable phylosyllicate, it is unlikely to have weathered in-situ. Kaolinite develops in tropical areas, therefore its presence in the semi-arid scalded landscape is most likely due to it being a relic that developed during a different palaeoclimatic time (Thiry, 2000).

Kaolinite contributes 7.6 % of the scald and 10.5% of the waterpond mineralogy in the 0-5 cm layer, and is the fourth most abundant mineral present in the profile. By contrast, this result is significantly lower than percentage of kaolinite found by Ringrose-Voase *et al*. (1989) in the scald. These researchers found kaolinite provided 21-30 % of the clay composition of the scald in the 5.5-16.5 cm layer and 31 to 40 % in the 35-80 cm layer.

The processes that probably contribute to the greater abundance of kaolinite in the waterpond include:

- Spatial variation of a relic mineral from a more humid period; or
- Aeolian deposition.

Churchman and Lowe (2012) suggest kaolinite can be interstratified with smectite, and that smectite (and vermiculite smectite) can provide nucleation sites for kaolinite formation. Thus, the higher percentage of kaolinite in the waterponds may be as a result of smectite weathering. The authors note that XRD analysis may not identify the interstratification peaks and therefore they may be identified as kaolinite incorrectly. Such an error in identification may be one reason that the predicted CECs are not particularly close to the measured CECs shown in Table 5-24.

Kaolinite is not an expanding mineral (Bhattacharyya *et al*. 1997). Consequently, it would not contribute to the shrink-swell action of the soil. Further, increases in kaolinite and interactions

with Fe and Al oxides has been found to be positively associated with increasing aggregate stability (Six *et al*. 2000). Although there are trends from the results of the current research indicating kaolinite, Fe and Al increase after waterponding, the changes are small. However, there is no indication from the ASWAT results that the dispersive behaviour of the old waterponds is diminishing because the waterponded soils tend to disperse spontaneously, whereas the scalds are stable. The reasons the dispersion occurs in the waterponded soil have already been discussed in detail, but in brief are associated with leaching of soluble salt from the profile. Thus the soil is transformed from a saline to sodic characteristic due to leaching of Na^+ by H^+ from plant roots. Consequently, any improvement in soil stability or aggregation due to possible increases in kaolinite are not observable from this sample set.

5.16.3 Minerals with a minor contribution but demonstrating a large effect from waterponding

The minerals that demonstrate waterponding has contributed to a change in the clay composition following waterponding include halite, calcite, smectite + chlorite and smectite + vermiculite. Explanations for the differences between scalds and waterponds are discussed below.

Halite is highly soluble and calcite mildly soluble and both readily weather from physical and chemical processes. These non-silicate minerals are found in arid regions where leaching and weathering processes are minimal. However, waterponding has led to the addition of water which has resulted in leaching of NaCl and $CaCO_3$ from the soil profile. Specifically:

- Halite is often found in soils from arid regions but due to high solubility is easily lost through dissolution. Halite was found to be present in the scalded soil but there was none identified in the waterponded soil. This finding correlates with the higher EC, and NaO_2 present in the scalds than the waterponds. Therefore, leaching of the NaCl from the profile is the most likely mechanism for the complete absence of halite in the waterpond.
- Calcite was visibly present in the form of carbonate nodules in the lower soil layers of the scalded soil particularly below 10 cm depth. Ringrose-Voase *et al*. (1989b) similarly noted the minimal presence of carbonate concretions. XRD identified a greater abundance of calcite for all layers analysed in the scalds than in the water ponded soil. For example there was 2% in the 0-5 cm layer of the scald and only 0.1 % in the equivalent layer in the waterpond. The results also show a lower percentage of CaO in the waterponds than the scalds. Leaching following waterponding is the most likely cause of loss of calcite and CaO from the soil profile. Even though calcite is considerably less soluble than halite, it is readily broken down during physical and chemical weathering processes (Schulze, 2005) resulting from plant growth in the waterponds.

Smectite clays

Smectite is an important component of soil that contributes to the success of the waterponding method to reclaim scalded soil because of shrink-swell characteristics and thus a role in crack formation (Reid-Soukup and Ulery, 2002). Ringrose-Voase *et al*. (1989b) identified between 21-30% smectite present in the scalded soil. The high proportion of smectite provides a valid explanation for the appearance of cracks in the soil following water retention in the waterponded soil. However, analysis of the samples for the current research has only found approximately 5% smectite in both the scald and waterpond soil. Specifically, the scalds were found to have smectite+chlorite present, with an increasing quantity with depth. The waterponds only had smectite+chlorite in the 20-30 cm layer indicating that weathering in association with SOM, water and shrink-swell disturbance has been occurring progressively down the profile since waterponding, but either there has not been enough time or the conditions have not been suitable for the complete loss of chlorite from the profile. Chlorite is typically a trioctahedral mineral of soil parent material, but is not common in soil as they are unstable, readily lose interlayer hydroxides of Mg, Fe and other cations. Chlorite can also weather to chlorite/vermiculite interstratifications during weathering (Churchman and Lowe, 2012) and this is consistent with the presence of vermiculite in the waterpond, but not the scald soil.

Smectite may be formed by transformation of mica and chlorite, or by neogenisis from alkaline solutions (Wilson, 1999; Churchman and Lowe, 2012); it can be inherited from transported parent material and is also reported to be ubiquitous with semi-arid environments (Odom, 1984). Wilson (1999) explains the example of transformation is illite – to – vermiculite – smectite occurs vie a process of decreased layer charge and depletion and exchange of interlayer K. Neogenisis of smecitite occurs through partial hydrolysis (Wilson, 1999). Smectite can be formed as a result of weathering leading to the transformation into 2:1 phyllosilicate minerals from mica following the loss of K from the interlayer (Tributh *et al*. 1987; Reid-Soukup and Ulery, 2002). The chlorite is often the first weathering product of mafic minerals, but it can be a primary mineral with a complex 2:1 layer structure and may weather to form vermiculite and smectite (Wilson, 1999; Schulze, 2005; Churchman, 2010). For example, Hughes *et al*. (1994) state that chlorite is the first clay mineral to undergo transformations as soils weather. They suggest the change occurs due to plant extraction of Mg^{2+} which results in alteration of chlorite/vermiculite to vermiculite. Hence it is feasible to interpret the difference in composition between the scald and waterpond as being the transformation of smectite + chlorite to smectite + vermiculite as a result of weathering and OM mineralisation under waterponding.

Smectite and smectite interlayer clays (with vermiculite and to a lesser extent chlorite and 14Å clays) contribute to the shrink-swell function of soil due to its expanding lattice layer, and therefore presence in the mineralogical composition of the scalded soil was thought to contribute to the re-establishment of cracks in the waterponds through swelling and dispersive

properties (Churchman *et al*. 1993). However, swelling and dispersive behaviour depends upon the particular smectite clay species and the Ca:Mg ratio (Odom, 1984). Due to the low percentage of smectite (and smectite interlayers) found in the soil profile in the current research, other processes, such as the removal of soluble salt, increases to ESP, changes to the Ca:Mg ratio, changes to soil structure during wetting, dispersion and drying are sufficient to form soil cracks after waterponding. To emphasise this finding, Ditchfield (1996), found that dispersion and swelling increased following the reduction in soluble salt from the profile. He also found that the soluble salt reduction contributed to an increase in linear shrinkage due mainly to the loss and gain of water in the interlaminar surfaces; a process described as inter-particle porosity (Odom, 1984; Chertkov, 2003).

5.16.4 Minerals with minor differences following establishment of waterponds

Waterponding was found to have little if any effect on the abundance of K-feldspar, plagioclase, hematite and gypsum. The lack of replicates may well mean this variation is purely natural heterogeneity.

K-feldspar

The results show very little difference in the presence of K-feldspar between the waterpond and scald samples. There is perhaps a slight trend indicating more feldspar in the waterpond which may be being deposited as the result of aeolian transport. Alternatively the loss of one may lead to a relative increase in another. For example, K-feldspar is a primary mineral and has been found to weather to illite and to smectite (Banfield and Eggleton, 1990), but because it is present in the scald and waterpond and appears to be in equilibrium, there is little evidence to suggest deposition or weathering.

Plagioclase

There was slightly less plagioclase in the waterpond than the scald in all layers. Plagioclase is a feldspar dominated by Na and Ca. Plagioclase weathers to form smectite or illite clay and during the process Ca, Na and K are released (Banfield and Eggleton, 1990). Given the loss of plagioclase following waterponding, it is possible that it is being transformed into smectite. Because the proportion of illite is lower in the waterpond, than the scald, it is unlikely the plagioclase is currently being weathered into an illite.

Haematite

The XRD and oxide analysis found little difference in the abundance of haematite and Fe_2O_3 between the waterpond and scalded soil. Haematite is a relatively stable Fe oxide originating from primary mineral weathering that is common in arid and semi-arid environments and is responsible for giving soil its red colour (Bigham *et al*. 2002). Goethite is the most common iron oxide and is responsible for giving soil a yellow/brown color; haematite is usually present

in association with it (Schulze, 2002). Stace *et al* (1968) described all iron oxides determined by XRD in Australian soil as goethite/haematite. As goethite gives a distinct peak near the first quartz peak it can be distinguished and none of the characteristic goethite peaks were identified on the XRD analysis for the current research. However, goethite may be present as nano particles which do not give rise to XRD patterns. The quantification of haematite was based on the haematite signal and was therefore distinguishable from goethite (pers comm U. Troitzsch 17 June 2016).

Phosphate fixation by Fe oxides, where P is occluded within the haematite structure, has been reported by Galvez *et al* (1999). However, this behaviour does not explain why the scalds have been found to have a higher TP and AP than the waterponds.

Fe oxides are the product of oxidation of Fe(II) present in most soils. The occur in primary minerals or as weathering products as Fe(III) (Churchman and Lowe, 2012). Fe oxides play an important role in cementation of aggregates and adsorption of SOM and so are important for C sequestration. This research has not explored the way that waterponding of scalded soil may change the role Fe oxides have on the C sequestration potential of soil. This deficiency should be explored further in future research.

Gypsum

Gypsum is often present in soils located in arid climates and was occasionally observed in both the waterpond and scald samples collected for the current research. Gypsum was also observed by Ringrose-Voase *et al*. (1989b) in the scald and waterpond at depths below 80cm. XRD analysis shows both the scald and waterponded soil has approximately 1% of gypsum in the 0-10 cm layers and < 1% below 10cm (Table 5-25). Gypsum is made up from two important, but not major plant nutrients and hence the distribution within the 30 cm profile may simply reflect plant uptake and cycling (Field and Little, 2008). Gypsum can also form through evaporate formation in arid and semi-arid environments. For example, (Eswaran and Zi-Tong, 1991). Chen, (1997) explain that dehydrated forms of sulphates can rehydrate into gypsum when water is present. Accordingly it is possible that this has occurred in the upper layer of the waterpond soil because water is able to infiltrate the profile. However, gypsum can also be transported by water and wind (Chartres, 1982; Chen, 1997). Consequently, there is a slight possibility that the modified micro-relief within the waterponds allows gypsum originating from salt lakes located in central Australia or from nearby prior alluvial systems (Chen, 1997) (of which Marra Creek is one) to be transported and deposited onto the waterpond surface.

This preliminary study examining the combination of PSA, aggregate PSA, mineralogical and chemical composition differences between scald and waterponded soil has provided some insight into the possible origin and fate of the materials present in the landscape. Given the very small size of the study it is difficult to draw robust conclusions as to the affect the changes may have on the C sequestration potential of the soil following waterponding. That said, this study

has provided evidence that waterponding may result in mineralogical change to the soil through leaching of soluble constituents and weathering sequences such as the transition of into plagioclase into smectite and illite into smectite vermiculite. Finally, there is evidence to suggest that aeolian deposition of quartz and illite may have occurred. It is likely that these changes would not have occurred without the contribution of water and plant growth in the waterponds.

Whilst SOC sequestration *per se*, cannot be directly attributed to any of the mineralogical changes that have been identified here, organo mineral associations and the associated physical functions including stabilisation of SOM and aggregate formation are factors that can contribute to improving the stability of the organic material and thus the long term sequestration of SOC. Therefore, the relative increase in kaolinite, or even the increase in the percentage of clay sized particles (< 2µm) identified by PSA suggest that waterponding is leading to organo-mineral changes to the soil. Such changes can be interpreted as being improvements that will provide resources or mechanisms leading to the physical protection of organic material by forming micro aggregates (Tisdall & Oades, 1982).

Finally, the hypothesised aeolian deposition of material such as quartz and illite may be leading to the accumulation of biochemical resources such as K for plant growth (Wilson, 2004) and enhancing the physical structure of the soil. The presence of these wind borne materials therefore represent the embryonic stage of soil formation and perhaps even suggest the development of an aeolian and biologically based A-horizon.

5.17 Cesium-137 analysis to predict erosion and deposition in the scald landscape

Historical and physical evidence exists to indicate that scald sites have been severely eroded over time (Figure 5-50). There are also physical attributes of the older waterponds suggesting they are areas of deposition. The physical attributes include changes to soil texture and vegetation biomass and mineralogical change. All these attributes have been explored in the current research and have previously been discussed. However, the origin of the hummocks is more problematic. The hummocks have been reasonably suggested by Beadle (1948); Warren (1965), Cunningham (1987) and Ringrose-Voase *et al.* (1989a), to be either relics of the original soil profile or modern deposits. Indeed, the suggestion that they may be relic soils was accepted as quite feasible and consequently it was not initially the intent of the current research to explore the origin of the hummocks further. However, ^{137}Cs analysis of a small subset of the scald, waterpond and hummock samples taken for the current research project (Subset 3) was undertaken to test the hypothesis that the hummocks were relics of the original A-horizon.

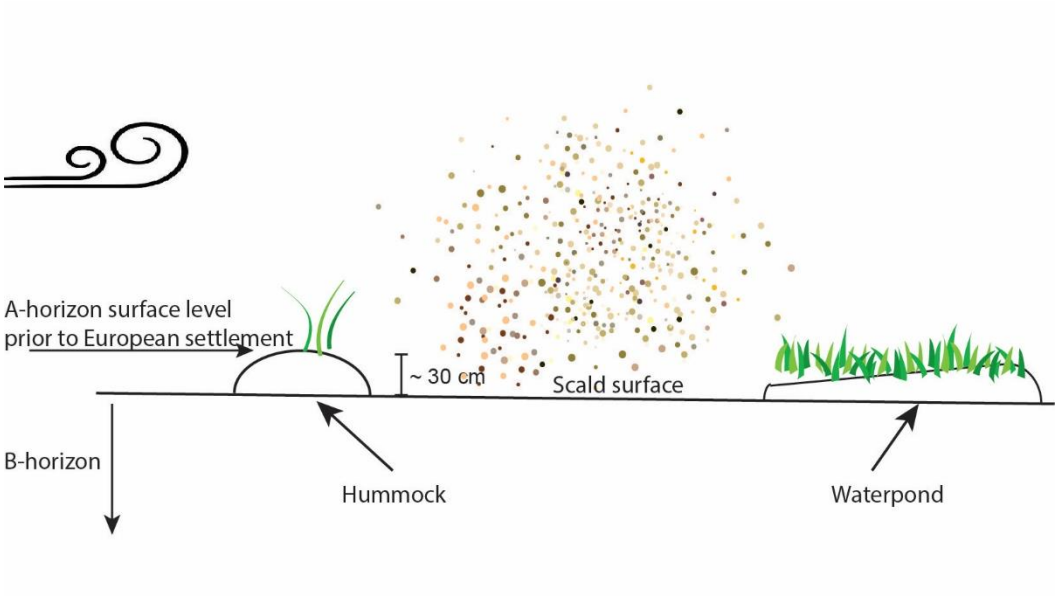


Figure 5-50 Diagrammatic representation of the hummock and waterpond in relation to the level of the A horizon prior to European settlement, soil loss through wind erosion and deposition onto the surface of the waterponds.

^{137}Cs is a radioactive isotope that is not naturally present in the environment but is found in soil as a consequence of radioactive fallout from high yielding thermonuclear tests that took place between 1954 and the 1970's (Nouira *et al*. 2003). The other notable source of ^{137}Cs in soil is from nuclear accidents such as Chernobyl and Fukoshima (Ritchie and McHenry, 1990). The ^{137}Cs is deposited in soil from the atmosphere via rainfall, bioturbation, vegetation decomposition and foliage deposition. Most ^{137}Cs is contained in the top 20 cm of the soil profile, with a diminishing quantity descending through the profile (Nouira *et al*. 2003). There is a definitive baseline differentiating the presence or absence of ^{137}Cs in soil dating back to 1952 ± 2 years. There is also a strong spatial distribution of ^{137}Cs related to rainfall patterns within latitudinal zones. Therefore, because ^{137}Cs is readily and strongly adsorbed onto the cation exchange sites of clay particles, and has a known half-life of approximately 30 years, its abundance in soil can be used to determine the quantity of soil physically redistributed across the landscape due to wind and water erosion over approximately the past 45 years (Chappell, 1999). Early research on uneroded sites showed that the abundance of ^{137}Cs in the upper levels of the soil profile was equal to ^{137}Cs deposition following fallout. Additionally, movement of ^{137}Cs within the soil profile is consistent with soil turnover caused by cultivation (Nouira *et al*. 2003) or bioturbation. Therefore, uneroded sites are useful for determining baseline concentrations on a horizontal and vertical plane across the landscape. Such sites can be compared with sites susceptible to erosion or deposition and thus enable determination of soil redistribution attributable to soil erosion or deposition.

The current results show that although there is variation in the ^{137}Cs concentration between sites, positions and depths, there is a trend suggesting vertical and spatial accumulation of ^{137}Cs enriched material in the waterponded soil, especially in the lower part of the waterpond closer to

the wall position. The result indicates that soil particles contaminated with ^{137}Cs are being transported and deposited into lower parts of the landscape. The likely mechanism for this is by aeolian and alluvial redistribution of the particles labelled with ^{137}Cs.

There is evidence of ^{137}Cs activity in the 30-50 cm layer, but no evidence in the upper 30 cm of the hummock profile. This pattern indicates that the soil particles in the 30-50 cm soil layer have been exposed to ^{137}Cs activity, whereas the soil in the 0-30 cm level of the hummock have not been exposed to radioactive fallout. It is likely, that the 30-50 cm layer was actually on the soil surface and thus received the ^{137}Cs label when the radioactive fallout occurred. The particles that are now located in the 0-30 cm layer of the hummock soil were previously situated low in a soil profile during radioactive fallout episodes and therefore avoided exposure to ^{137}Cs. The ^{137}Cs free particles have subsequently been eroded and have been transported by aeolian or water deposition, to be trapped by vegetation and thus form the modern vegetated hummocks. Thus, the hypothesis that the hummocks are representative of relic soils is not proven, but rather the hummocks are an accumulation of relatively recent transported material.

However, two alternative explanations for the lack of ^{137}Cs labelling on the upper 30 cm of the hummock profile include:

- Due to the light texture of the soil and lack of clay, it is possible that the ^{137}Cs has leached out of the surfer layers of the hummock profile; or
- The erosion events that removed the A-horizon from around the hummocks, possibly also removed particles from the surface layer of the hummocks that contained ^{137}Cs labelling and all that remains are the unlabelled particles that form the hummock.

Consequently, given the relatively small data set examined it remains difficult to conclusively attribute the origin of the vegetated hummocks.

The SOC concentration in the hummock soil is lower than the concentration in the scald soil. The mean SOC concentration is 0.32 g C 100 g^{-1} in the 0-5 cm layer of the hummock and 0.39 g C 100 g^{-1} in the scalds. The SOC concentration in the 30-50 cm layer of the hummock soil is 0.38 g C 100g^{-1} soil, and therefore similar to the 0-5 cm layer of the scald. The relative similarity in SOC concentration between the 30-50 cm layer in the hummock soil and the 0-5 cm layer in the scald provides further evidence that the 30-50cm layer of the hummocks most likely has covered the scald surface. The results also indicate that despite the presence of vegetation the hummock soil does not sequester SOC. However, because this was a very small subset of samples further research is necessary to establish the extent of land covered by hummock soil and to determine its SOC sequestration capability.

5.18 Summary and conclusion

Waterponding triggers a sequence of physico-chemical events that improve the potential of scalded soil to sequester SOC. The series of events commences after the waterponds have been constructed thereby allowing water to be trapped within the waterpond soil structure. The trapped water infiltrates the soil profile and leaches soluble salts from the saline and sodic scalded soil. Once the electrolytes have largely been removed from the soil, the soil is no longer classified as saline, and the dispersive properties of the sodic soil, which has an ESP of > 15 %, are reinitiated. The dispersion allows reorganisation and wetting of clay particles which results in restoration of the shrink swell capacity of the soils. The soil pH increases in the waterponded soil reflecting the change from Na^{+} to Ca^{2+} and this change also has an effect on the dispersive properties of the soil. After these chemical changes, conditions become more conducive for seed to germinate and plants to grow on the formerly denuded landscape. Over time the landscape regains sufficient vegetation, including palatable forage species, to enable livestock grazing to resume. Ultimately, as the plant growth increases the SOC concentration also increases, leading to C sequestration, until a higher equilibrium is achieved. The current research has suggested changes to soil mineralogy whereby leaching and weathering is causing physical and chemical changes modifying nutrient availability in the soil. Aggregate and particle size changes, possibly from aeolian deposition provide further evidence that as the landscape becomes revegetated not only are water and seed resources trapped, but mineral particles being transferred by wind and water erosion, are accumulated within the waterponds and retained in the landscape.

This example of LUC from a degraded to a more functional landscape demonstrates that sometimes intervention by engineering, in this case construction of waterponds on clay pan scalded soils is necessary to restore a degraded landscape. The results show that waterponding is an effective method that facilitates SOC sequestration in a relatively short period of time (approximately 5 years) and suggest that after around five years an equilibrium level of SOC is achieved. The results also suggest that after the C has been sequestered the SCS levels persist at a new equilibrium level over time.

However, a limiting factor against further SOC sequestration could be nutrient availability with demonstrable decreases of AP and TP, a reduction in TS at depth and variable TN results. The nutrient ratios of C:N:P:S suggest that although the C:N ratio becomes optimal following waterponding, the C:P ratios widen so that it is probable that these resources are limiting further vegetation growth as they are being taken up by biota, or possibly leached from the landscape. The long term impact of nutrient availability and the effect on SOC sequestration on this landscape will require ongoing examination. Other edaphic factors that may constrain SOC sequestration in the older waterponds are the propensity for dispersion to occur and the higher alkalinity and ESP. These are characteristics that can lead to C mineralisation of SOC, and

therefore may be inhibiting the older waterponds from sequestering additional SOC. Finally, the mineralogy of the soils, changes to mineralogy due to waterponding and the presence of Fe and Al oxides are important characteristics that most likely influence the potential of the soil under waterponding to sequester SOC.

Chapter 6: Synthesis

6.1 Introduction

The principal aim of the current research project was to study the C sequestration potential of soil over time following LUC by using two case studies from two very different environments in NSW Australia; one located in the temperate, moist Southern Tablelands of NSW and the second in the semi-arid rangelands of central western NSW. To achieve the aim a chronosequence approach was used in both studies to demonstrate spatial and temporal effects of LUC on soil SCS to 30cm depth. This enabled an investigation in both studies of the increase in SCS following LUC. In the first case study paired sites comparing 20 BEPs with the adjoining AL were used. In the second case study a comparative assessment of three clay pan scalds with 12 waterpond sites was undertaken.

The results from these two case studies have provided statistically significant empirical data to satisfy the principal research aim by identifying the potential of agricultural soil that has undergone LUC to sequester C. The results identified large spatial and temporal variability in SCS between and within the case studies and even suggested that in some situations (i.e., BEPs) large long term SOC sequestration following LUC may not be achieved. The results also suggest that LUC will most likely not bring SCS back to pre European settlement levels of SCS. This research has also provided convincing evidence that soils highly degraded and depleted of SOC have a greater capacity, at least initially, to sequester C than soils that have a proportionally higher baseline SCS present prior to the LUC intervention. Similar conclusions were previously reached by Paul *et al.* (2002), Cunningham *et al.* (2015) and Prior *et al*. (2015) working with environmental plantings and by Chan *et al.* (2010), Chang *et al*. (2011), Badgery *et al*. (2013) and Orgill *et al*. (2013) for example, working with grazing, pasture and cropping soils.

This Synthesis chapter presents a summary of the results pertaining to SOC sequestration following LUC for both cases studies. A “state-and-transition” (SAT) (Westoby *et al*. 1989) framework has been used to explain changes to the state of SCS following LUC for each case study. The SCS results have then been applied to an ecosystem response curve to demonstrate changes to landscape functionality encompassing degradation, resilience, stability and sustainability of the landscape (Tongway and Ludwig, 2007) following the LUC activities in each case study. The chapter concludes by presenting a synthesis of SCS change by amalgamating the results from the two case studies onto the model of SOM accumulation proposed by Johnson (1995) (Section 2.4.4 and Figure 2-2) to show the C sequestration potential of different soils following LUC.

6.2 Summary of the carbon sequestration potential of soil following the land use change activities of BEPs and waterponding

When the results of the two case studies are compared, not surprisingly little similarity exists. For example, the mean baseline SCS of AL in the temperate Southern Tablelands region of NSW was found to be 41.1 Mg C ha^{-1} (Figure 6-1). However, the baseline mean SCS of the scalded clay pan soil in the semi-arid rangelands of the central west of NSW was found to be only 18.7 Mg C ha^{-1}. The large difference between the two case studies stems from the disparate climatic, geological, geomorphological and land use histories of the two locations.

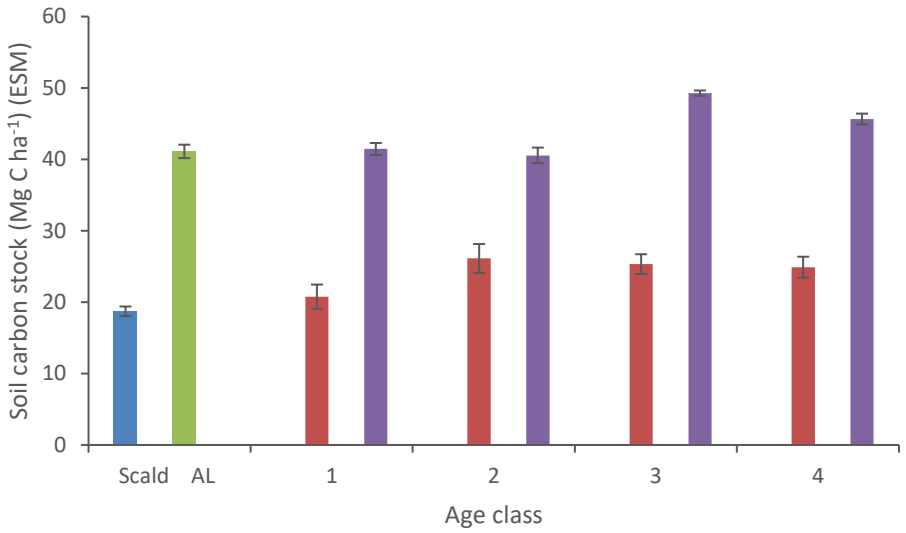


Figure 6-1. Synthesis of SCS results for BEPs and waterponds showing AL has a higher baseline than scalds, and BEPs have more SCS than waterponds. (Blue bar = scald; green bar = AL; red bar = waterponds; purple bar = BEPs; Age classes: 1 = waterponds aged 1 yr and BEPs aged 1-5 yrs; 2 = waterponds aged 5 yrs and BEPs aged 6-10 yrs; 3 = waterponds aged 10 yrs and BEPs aged 11-15 yrs; 4 = waterponds aged 25-27 yrs and BEPs aged 16-19 yrs). Error bars = SEM.

Following LUC intervention each case study was found to have different temporal and spatial responses. In the first, the SOC sequestration potential of BEPs established on pastures previously used for grazing was examined. The SCS results showed spatial and temporal variation between sites and positions within sites. The results also show a trend where the oldest cohort of BEP sites aged 16-19 years (Age Class 4) had a lower SCS on an ESM basis (45.7 Mg C ha^{-1} to 30 cm depth) than the soil in the paired AL position (49.9 Mg C ha^{-1}) (Figure 4-21), and also less than in the younger cohort of BEP sites aged 11-15 years (49.3 Mg C ha^{-1}). When the baseline amount of SCS to 30 cm depth for the 20 sites is taken as the average of that under all 20 sites in the AL position (Figure 6-1) the mean SCS is 41.1 Mg C ha^{-1}. The BEPs aged between 11 and 15 years have a SCS of 49.3 Mg C ha^{-1} which is the equivalent of an annual C sequestration rate of 0.5 Mg C ha^{-1} yr^{-1} over 15 years. However, the annual rate of SOC sequestration including all BEPs up to 19 years of age was 0.2 Mg C ha^{-1} yr^{-1}.

The results obtained for the second case study showed an increase to SCS following waterponding of scalded soil, albeit coming from a low baseline of 18.7 Mg C ha^{-1}. After 5 years waterponds had an average SCS of 26.1 Mg C ha^{-1} indicating they had sequestered on average up to 7.5 Mg C ha^{-1}, which is an SOC sequestration rate of 1.5 Mg C ha^{-1} yr^{-1}. By 27 years (Age class 4) the waterponds had sustained the SCS at an average of 24.9 Mg C ha^{-1}, but had sequestered no additional C during the intervening 20 year period. The mean annual SOC sequestration rate following waterponding is 0.23 Mg C ha^{-1} yr^{-1} over 27 years which compared very closely with the rate of 0.2 Mg C ha^{-1} yr^{-1} over 19 years achieved with BEPs. However, because there is an eight year difference in time frame between the two Case Studies the relative change is not comparable. Additionally, the SCS sequestered in waterponds has most likely reached an equilibrium level and is unlikely to sequester additional SOC, whereas there remains potential for BEPs to sequester more SOC in the longer term.

6.2.1 Transferability of results to other soils sequestration and retention of SOC in soil following LUC

The research was conducted on sites from two disparate climatic regions and land management activities, and also encompassed a large range of soil types. It was therefore beyond the scope of the research to differentiate SOC sequestration potential following LUC between soil types. Overall, the findings from the waterponding case study suggest that positive additions of SOC is achievable regardless of soil type especially, when the starting level of SOC is very low such as found in the scalded soil.

The transferability of the results found by this research to other soil types and management change scenarios depends on soil factors and properties such as mineralogy, soil surface area, sorption and the role of aggregation and mineralogy in attracting and retaining SOC.

The transferability of the findings from this research to other SOC sequestration scenarios with soils comprising similar soil types encountered in the case studies is not feasible due to the diverse range of soil types studied.

6.3 The use of ecological frameworks to show changes to soil CS following LUC

Plant community climax (Clements, 1916) eventuates, following a series of ecological processes that allow the composition of the vegetation to be in equilibrium in the absence of grazing or other pressures. The sequential nature of the processes leading to plant community change has been termed “succession”. The evolution of the succession model has been summarised by Clements, (1916). SAT frameworks have built on Clements’ original succession model and have been used as a simple way to explain changes to rangeland vegetation caused by natural (eg climatic) or management changes (Brandon *et al*. 2003; Cingolani *et al*. 2004). Westoby *et al*. (1989) proposed the state-and-transition framework as an alternative to the traditional

succession model whereby the "state" of vegetation depended on responses to rangeland management and grazing. "State" in this context refers to the species composition of a plant community and "transition" implies the pathway from one plant community (state) to another. A change in state is typically triggered by an event resulting in either degradation or improvement to species composition. The processes triggering a change in state is explained by a series of "transitions. Westoby *et al*. (1989) suggest transitions are opportunities or threats to rangeland vegetation success that may be stimulated by events such as weather (e.g., rainfall or temperature), fire or management strategies. Indeed, the concept of transitions that have a positive or negative dynamic to ecological communities was seen by Westoby *et al*. (1989) and Briske *et al*. (2005) as overcoming the seemingly one way continuum of vegetation succession towards climax plant communities, by allowing for reversion to earlier states or fluctuation between states.

Practical examples explaining the application of state-and-transition models to describe the pathways related to changed rangeland vegetation communities includes that by Ludwig and Tongway (2002). These authors applied the state-and-transition framework to their research on savannah pastures in northern Australia. Their study sites were managed using either fire, chain clearing or natural regrowth strategies, or by thinning using ring-barking or poisoning methods. Ludwig and Tongway (2002) describe four different states of community in their study: State I, unaltered savanna remnants; State II, chained but not cleared savannahs; State III, cleared and then seeded savannah and State IV, grazed savannah leading to high modification. They found the "transitions" were prompted by management activities (treatments) that included chaining, clearing, seeding or grazing. They also found the productive potential of the soil was substantially modified by the applied treatments. Alemseged *et al*. (2011) used an SAT framework on the Cobar pediplain in the semi-arid rangelands of the central west of NSW to describe four vegetation states associated with changes to management. The authors also provide a catalogue of seven transition phases that could arise from different management interventions including clearing activities as well as cropping and grazing. They concluded that by using an SAT model it could be established that transition pathways cause by natural or management intervention may result in either a positive change or a reversion back to a less desirable vegetation composition or state.

In a further development of the SAT model, and building on the theory by Noy-Meir (1973), Ludwig *et al*. (1997) developed a conceptual model for explaining rangeland functionality referred to as the trigger-transfer-reserve-pulse (TTRP) framework. These authors argued that the traditional state-and-transition framework was limited because it could not be used to understand landscape dynamics. Instead they argued that the state and transition framework classified rangeland communities as being stable at a particular point in time, and required transitional processes to initiate a change of state (Ludwig and Tongway, 1997). Consequently, Ludwig et al. (1997) designed the TTRP framework to recognise both temporal and spatial

components of the landscape to show run-off / run-on, sediment transfer and mineral deposition processes (Figure 6-2). The TTRP model uses the terms “Trigger” to refer to an event such as rainfall, and “Transfer” as the spatial redistribution of resources including water, nutrients, soil particles and litter resources across the landscape that have been initiated by the Trigger event. The resources accumulate in the soil and are retained in “Reserve” until suitable conditions trigger a growth response referred to as a “Pulse”. The effect of disturbance regimes was to initially alter the runoff and infiltration dynamics so that the availability of water was changed, altering the habitat suitability for certain plants, and ultimately the species composition of the landscape. The TTRP framework first proposed by Ludwig *et al*. (1997), and shown at Figure 6-2, enables an examination of new ecosystem dynamics across a range of changes, leading to ecosystem predictability instead of merely recording changes to state after they have occurred.

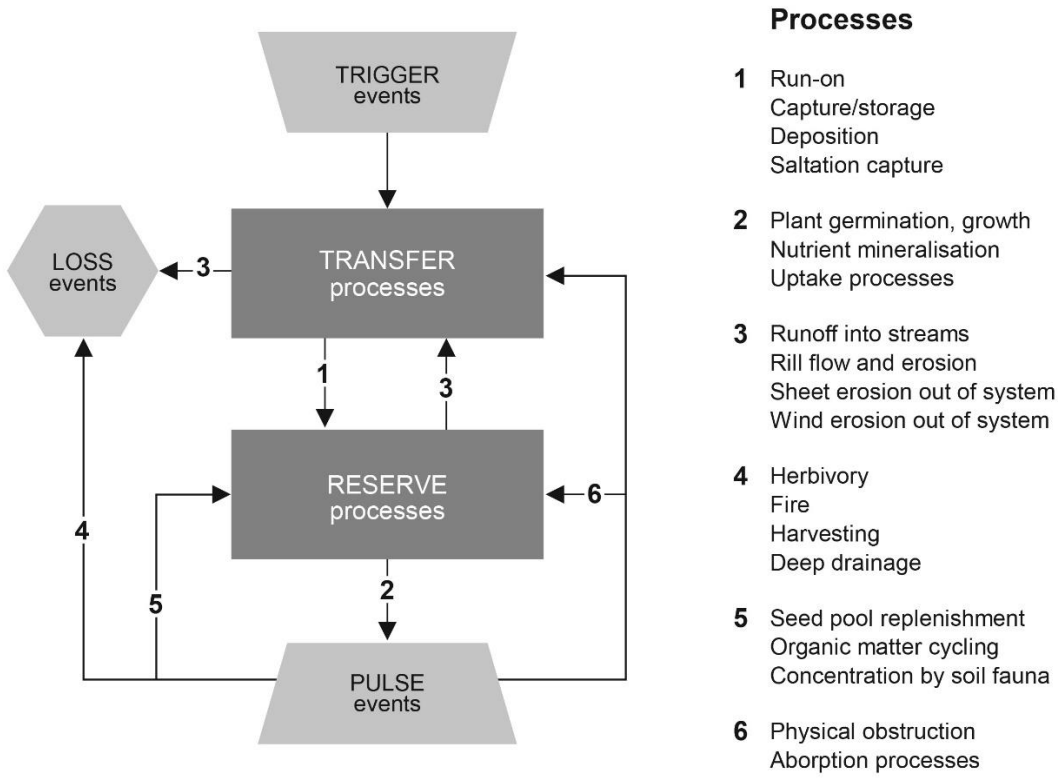


Figure 6-2. The Trigger-Transfer-Reserve-Pulse framework and the six processes that operate at different stages of the framework. (after Tongway & Ludwig, 2007)

The TTRP framework guides an understanding of landscape functionality, where functionality relates to the capacity of the landscape to capture and retain rainfall and soil, nutrient and seed resources (Ludwig and Tongway 1997). The TTRP conceptual framework underpins the landscape monitoring analysis tool of Landscape Function Analysis (LFA) (Tongway and Hindley, 2004b). LFA is a procedure that uses 11 soil surface indicators to undertake rapid visual assessment and quantification of biogeochemical properties and processes, functionality and responses to stress and disturbance associated with impacts on the landscape. Impacts can result in either degradation or rehabilitation outcomes. The results of the 11 indicators are used to calculate a quantifiable score for three landscape function indices; stability, infiltration and

nutrient cycling. The scores can be used to evaluate the processes and characteristics of the landscape, and for monitoring changes to processes as an indicator of landscape functionality following an intervention such as LUC. An ecological sigmoidal response curve is used as the interpretational framework for the LFA indicator values that can be used to describe landscape resilience to stressors and help explain whether a landscape is, or is more or less functional or dysfunctional. The response curve is useful for showing the resilience and capacity of an ecosystem to respond to stress and disturbance actions such as drought or grazing may have on vegetation or soil ecosystems (May, 1977; Williams, 1993; Scheffer *et al.* 2001; Tongway and Hindley, 2004a,b; Tongway and Ludwig, 2007).

Critical thresholds on the response curves (Figure 6-3) show where the landscape capacity to recover to a former state following a stress or disturbance such as a significant natural occurrence or a change in management. Robust landscape systems (Figure 6-3: A) have a relatively grater "low" critical threshold and thus the intensity of change necessary to achieve recovery from stress or disturbance is less than that required for a fragile system. Robust systems therefore can quickly respond to rehabilitation efforts (Tongway and Hindley, 2004a; Tongway and Ludwig, 2007). Fragile systems (Figure 6-3: B) on the other hand, have a lower critical threshold and a long, slow rate of landscape recovery.

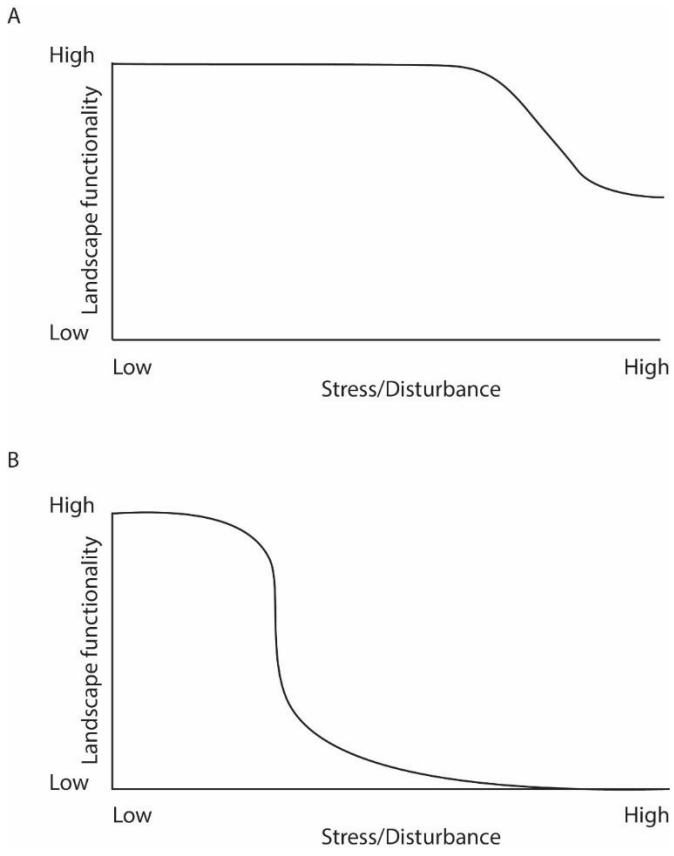


Figure 6-3. Sigmoidal landscape stress and disturbance response curves for robust (A) and fragile (B) landscapes. (After: Tongway and Ludwig, 2007).

Recent work by Read *et al*. 2016 shows that for BEP sites located in the temperate southern tablelands of NSW the AL has on average LFA scores that site just below or at the threshold of potential concern indicating they may be susceptible to degradation if subjected to disturbance or stress. The BEPs sit midway between the upper threshold and the threshold of potential concern for the three indices of soil stability, infiltration and nutrient cycling suggesting that they may be able to withstand environmental or management stresses. In the case of scalded soil, it sits below the critical threshold for all the three indices: stability, infiltration and nutrient cycling. Both young and old waterponds have improved functionality but remain below the threshold of potential concern which means that although their functionality is better than that of the scalded soil, it remains fragile to stresses such as drought and overgrazing (Figure 6-4).

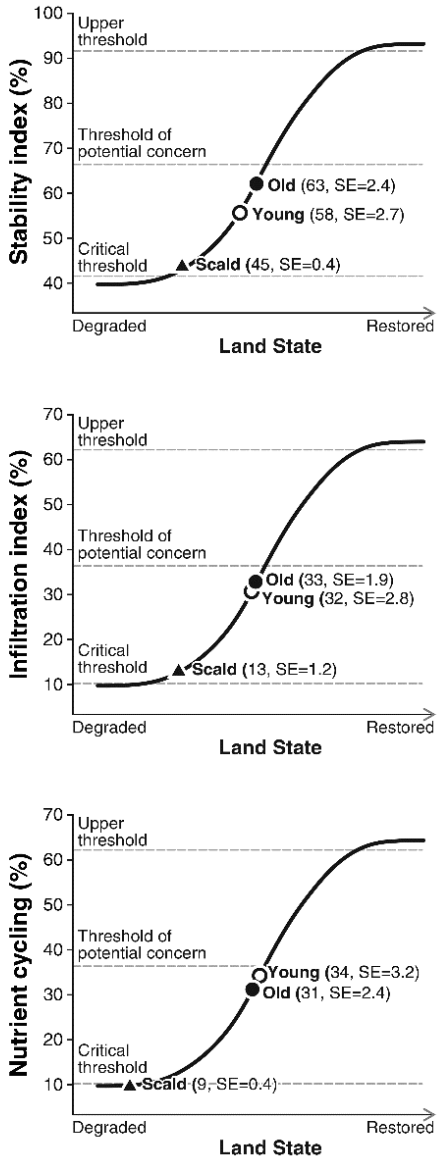


Figure 6-4. Sigmoid response curves for scalds and waterpond soils (source Read *et al*. 2016).

Plant communities in a state of climax are associated with soils that are also in a state of climax (Cain, 1939; Jenny, 1941). Consequently, because of the inextricable link between plant associations, soil formation and SOC, the SAT framework has been adapted to show the pathways between, and the successive states of, SCS in both the BEP and waterponding case studies. In these cases studies each State is taken to represent the SCS at a specified point in time and transition refers to the pathway between each SCS state. The SAT frameworks for the BEP and waterponding case studies are detailed in the next two sections respectively.

6.3.1 State and transition framework for BEPs

The reforestation of grazing pastures in the Southern Tablelands of NSW with BEPs has been shown to increase SCS (to 30 cm depth) in some cases. However, the results show variability in temporal and spatial scales with some sites sequestering SCS to quantities above their paired AL, whilst others BEPs have a lower SCS than their paired AL. An SAT framework has been applied to this type of reforestation to show discrete states of SCS and the transition phases between the states associated with LUC (Figure 6-5).

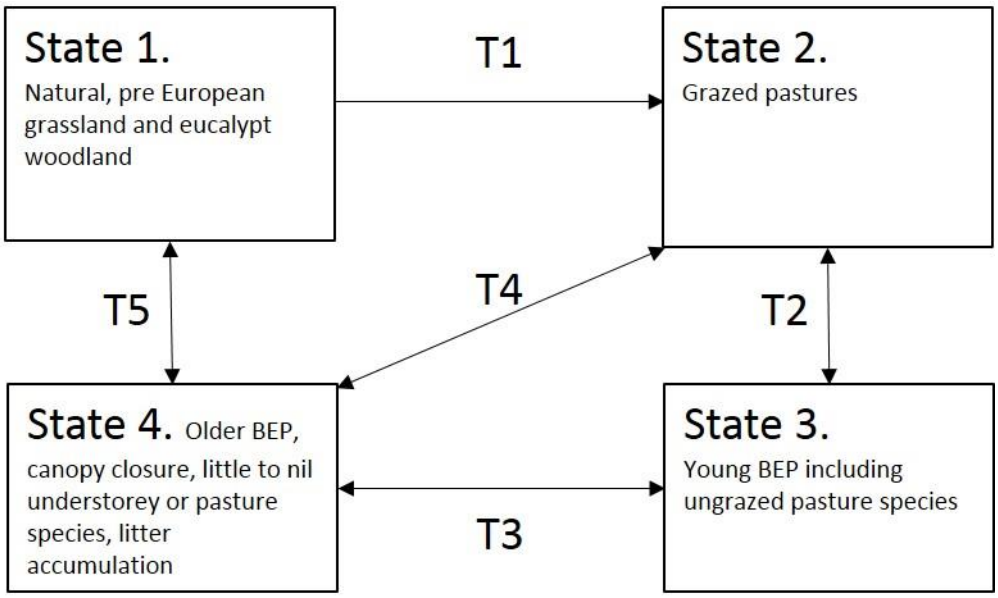


Figure 6-5. State-and-transition framework showing changes to SCS after reforestation of agricultural land using BEPs.

There are four distinct states of SCS present prior to and after establishment of BEPs. Each state is described in detail in Table 6-1. The first state represents the baseline SCS level presumed to have existed prior to land clearing by European settlement. State 2 represents the SCS state in the AL after losses caused by soil disturbance following European settlement and interventions involving land clearing and agricultural practices. State 3 represents the SCS in the first five years (Age Class 1) following LUC to BEP and State 4 considers the changes to SCS on sites that have been established between 16-19 years (Age Class 4).

Table 6-1. Catalogue of states for BEPs commencing with SCS in the pre European settlement state, and progressing to the final state where BEPs have been established in pastures previously used for grazing from up to 16-19 years earlier.

State	*Description*
State 1	Refers to the undisturbed state of SCS that existed prior to European settlement (and the natural grasslands and eucalypt woodland). The SCS in this state was most likely stable and in equilibrium. As explained by Jenny, (1941) the mature soil properties likely to have been present at this time were in a climax association with the endemic plant community (Jenny, 1941). Estimated SCS in the Southern Tablelands of NSW at this time was 57 Mg C ha^{-1} (Section 4.3).
State 2	Exists following clearing of native grasslands and eucalypt woodland, and subsequent conversion to perennial grassland for grazing. Results of this intervention include the loss of SCS due to removal of deep rooted perennial trees and native perennial understorey plants, soil disturbance following tillage during pasture establishment, change from perennial to annual vegetation, grazing of pastures, overgrazing by all species, ground cover loss and erosion of the A-horizon. Reversion to the SCS extant in State 1 is unlikely, without intervention, due to the loss of A-horizon and associated SOC. The mean SCS of 41.1 Mg C ha^{-1} found in the AL position is taken as being representative for this state. This landscape is robust when managed appropriately – i.e., avoiding overgrazing and maintaining ground cover to prevent soil erosion.
State 3	Describes the state of the landscape in the first 1-5 years following establishment of BEPs. In this state a fenced enclosure is established, trees are sown using a direct sowing implement, trees germinate and become established (after naturally occurring selective thinning). At this stage the pasture grasses from State 2 are able to grow without competition from grazing herbivores and the trees begin to grow and therefore contribute to the overall SCS in the BEP enclosure. During State 3 the trees are producing living tree biomass to support the photosynthetic biomass and consequently are in Stage 1 of forest development proposed by Attiwill (1979) and Baker and Attiwill (1985). The landscape functionality is higher following livestock exclusion allowing grass and BEP tree growth.
State 4	Describes that part of the agricultural landscape that has been reafforested with BEPs. In this state, there has been an active LUC intervention to restore woody vegetation back into the landscape to improve biodiversity. C is sequestered in the woody biomass, but the results for SOC sequestration is less conclusive. Failure of BEPs can lead to passive reversion back to State 2 or 3 where grasslands can become dominant. In this state, the BEP trees provided substantial landscape protection and thus contribute to a high LFA score. The older BEPs may be in or approaching the second stage of forest development, which is the formation of heartwood (Attiwill (1979) and Baker and Attiwill (1985)). The BEPs produce additional SCS than AL in sites up to 15 years. However, BEPs older than 15 years have a lower SCS than the paired AL. Additionally, the BEPs contribute to the landscape resilience through the growth of large quantities of AGB and BGB. The AGB and BGB also sequesters large quantities of C which overall makes the BEPs an important sink of atmospheric CO_2.

There are five transition pathways that lead to the SCS levels found in the older BEPs (Table 6-2). Broadly, each transition represents an active intervention that causes change to the SCS following an LUC intervention. Apart from Transition 1, all the other Transitions are reversible.

Table 6-2. Transition catalogue showing LUC interventions and their effects on SCS within BEPs.

Transition	*Description*
Transition 1	Active LUC intervention by tree felling and ring barking followed with ploughing, pasture establishment and fertiliser application. Reversion to State 1 may be possible via the passive seed pool and regeneration by coppicing and/or root suckering, when progress to subsequent States is not achieved.
Transition 2	The second active LUC intervention is the introduction of cloven hooved animals for grazing. This activity results in soil compaction, overgrazing and loss of landscape resilience.
Transition 3	Reforestation of the AL with BEPs can lead to increased CS in the biomass and in the soil.
Transition 4	Reversion to the SCS levels present in the AL is possible if the BEPs fail.
Transition 5	In the absence of any further intervention, it may be possible for SCS to eventually reach pre-European levels. However, the evidence from the results of this research suggest that BEPs are unlikely to achieve this level.

6.3.2 State and transition framework for scalded clay pan soil rehabilitated by waterponding

A succession model for vegetation change following waterponding has previously been shown (Figure 5-44). An SAT model for changes to SCS following LUC (Figure 6-6) in the waterpond case study, has five states each with a level of SCS. In this case study, State 1 depicts the initial SCS level of the grassland that existed prior to European settlement and the introduction of livestock grazing. In this state, the plant community was in climax and the SCS in equilibrium. However, overgrazing and mismanaged drought created deleterious conditions conducive to soil erosion and thus led to the degrading landscape and SCS present at State 2. If grazing management had been conducted in a sustainable way (as we know today), to minimise soil erosion, a return of SCS to levels seen at State 1 may have been achievable. Regrettably, however, the overgrazing and mismanaged droughts resulted in the complete loss of the A-horizon and subsequent development of scalded soil (State 3). The SCS in the scalded clay pan soil represents the lowest amount of SCS found in the current research project. Active engineering intervention via the waterponding method has led to water and seed resource retention, which in turn has led to rapid vegetation growth, and commensurate increases in SCS. However, despite waterponding intervention, pre European levels of SOC are unlikely to be achieved because the A-horizon has been completely eroded. Instead, SOC sequestration is only occurring in the B-horizon, which is typically less resource rich than the A-horizon. The loss of the A-horizon is therefore a major constraint on the SOC sequestration potential of the scalded soil.

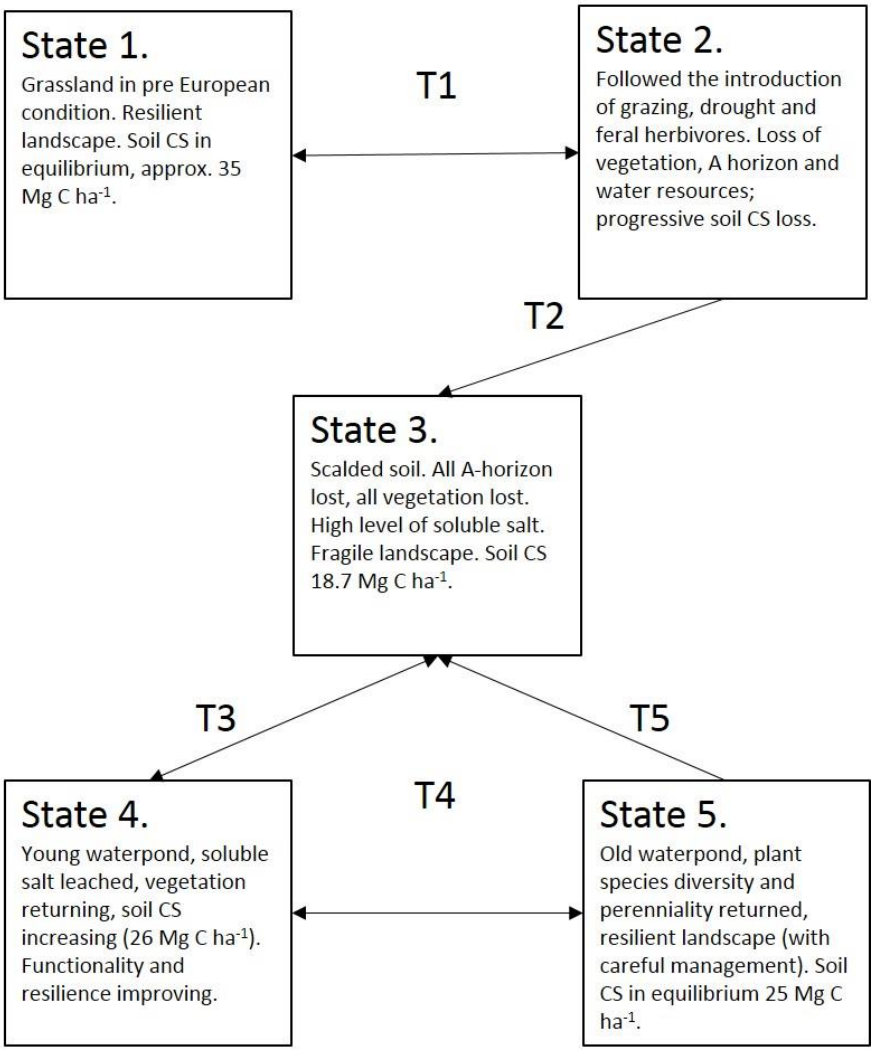


Figure 6-6. State and transition framework for SOC sequestration of clay pan scalded soil following waterponding

To summarise the changes to vegetation and thus SCS arising from grazing, erosion, scalding and then restoration by waterponding, the catalogue of five states depicted in Figure 6-6 is explained in detail in Table 6-3.

Table 6-3. Catalogue of States detailing the condition of vegetation and SCS for each of the five states for the waterpond case study.

State	***Description***
State 1	Describes the robust state of SCS prior to European settlement. In this state the landscape was a dynamic grassland (Mitchell, 1848); SCS was most likely in climax (maximum carrying capacity) and thus in equilibrium, with C inputs and losses in balance. The landscape was robust with resilience against natural disturbance. Based on a study by Murphy *et al*. (2003) it is likely the SCS to 30 cm depth was approximately 35 Mg C ha^{-1} at that time.
State 2	Refers to the state of the landscape after the introduction of grazing, adverse impacts on plant species composition, degraded resilience of vegetation and the landscape, against the impact of drought and overgrazing. Progressive loss of vegetation and soil A-horizon leading to insidious loss of SOC. In this stage resilience of the landscape and potential to revert to pre European condition (State 1) is possible but progressively less likely as conditions worsen. Water and seed resources are permanently lost from the landscape. SCS degraded.
State 3	Describes the scalded state where vegetation loss is complete, SCS is at its lowest level (18.7 Mg C ha^{-1}), and there are high soluble salts in the soil profile. Reversion to pre European condition (State 1) unlikely. This landscape is now fragile and functionality has fallen below the critical threshold for recovery. Active intervention is needed to retain water and stimulate vegetation growth.
State 4	Refers to young waterponds (< 5 years of age). This state occurs following active intervention through engineering construction of waterponds on scalded soil. Then, following a rainfall event (trigger) waterponding results in the transfer of water resources which in turn results in leaching of soluble salt. Seed can be retained allowing natural germination of annuals leading to an increase in SOC. Gradual increase in diversity, perenniality and abundance of plants. In this state SCS is approximately 26 Mg C ha^{-1}.
State 5	Represents older waterponds where there is low salinity, high plant species diversity including a range of annuals and perennials, and a sustainable level of SOC. In this state the waterponds have a store of resources that stimulate a pulse leading to germination and seeding. After waterponds are established the landscape has regained resilience against stress and disturbance. Although regaining State 1 is desirable, it is unlikely. Old waterponds have an average SCS of 25 Mg C ha^{-1}.

As in the BEPs, there are five transitions that lead to the SCS levels found in the older waterponds. The catalogue of five transitions in the semi-arid rangelands of NSW initially follow a trajectory towards total landscape degradation (i.e. Transitions 1 and 2). In this State the climax community becomes a scalded soil surface on clay pans with little or no vegetation. However, active intervention through construction of waterponds at Transition 3 show rehabilitation of clay pan scalded soil is possible, and that SOC sequestration follows that intervention (Table 6-4).

Table 6-4. Catalogue of transition pathways encompassing SCS change following post European settlement, development of scalded soil and rehabilitation by waterponding.

Transition	***Description***
Transition 1	Introduction of livestock grazing and rabbits, leading to degradation of plant species diversity and quantity.
Transition 2	Overgrazing, drought, wind and water erosion leading to loss of the A-horizon and loss of SOC. Scald development and an increase in soluble salts in the soil profile. Reversion to State 1 is unlikely from States 4 or 5 despite plant growth and SOM accumulation, because the A-horizon has been lost, the soils are saline and the surface crust is impermeable to water infiltration.
Transition 3	Active intervention (use of grader) to construct waterponds, (including direct seeding of some vegetation species i.e. *Atriplex* spp.) leading to retention of water and other resources within the landscape and gradual increase in SOC. Eventually passive rehabilitation occurs from the self-replacing germination of annuals and perennial species. It is possible for either the active intervention or passive rehabilitation to fail and thus reversion to State 3 may occur if grazing is commenced too early in the rehabilitation process.
Transition 4	Soluble salt completely leached from the upper soil profile, crack formation and ongoing passive establishment of annual and perennial plant species leading to an increase in SOC. No further intervention besides sustainable grazing (of all herbivores) is required. Passive germination of plant species will continue if the landscape is carefully managed.
Transition 5	There is a risk that reversion to States 4 or 5 is possible if the rehabilitated landscape is mismanaged through overgrazing, thereby leading to a loss of resources; i.e., the threshold for disturbance remains low.

In the waterponding example, a transition pathway to improve the SCS found in State 4 to the level most likely to have existed at State 1 is desirable but unlikely to ever be achieved. The A-horizon has been lost from this landscape, therefore any improvement to SCS is occurring in the remnants of the upper B horizon soil and is constrained by the available resources. Despite the success of waterponding to restore the landscape, reversion from States 4 and 5, back to State 3 is possible if unsustainable land management practices are resumed. Therefore, it is important to maintain low livestock numbers and implement sustainable grazing management practices for other herbivores to protect plant species diversity, and perenniality to promote landscape resilience. Such management will provide conditions suitable for the long-term viability and resilience of both the vegetation and the landscape and thus retention of SOC.

6.4 LUC and effects on degradation, stability, resilience and sustainability of the landscape

The characterisation of the states and transitions for the two case studies has allowed a detailed examination of the changes to SOC states caused by LUC. However the SAT model does not provide a clear representation of the different states in relation to a spatial or temporal pattern. Consequently the sigmoidal response curves (Figure 6-3) have been adapted in a conceptual way to demonstrate how the LUC has impacted on SCS and concomitantly the landscape function. This has been achieved by applying an ecosystem response curve to the two case studies.

6.4.1 Ecosystem response curves for LUC in BEPs

The ecosystem response curve showing changes to and critical thresholds of landscape functionality following LUC in the BEP case study is shown at Figure 6-7. The four states of SCS have been overlayed onto the response curve to show where LUC interventions modify the landscape functionality. In the period prior to European settlement (State 1) the landscape would have been stable and resilient against most natural stressors and thus above the critical threshold level for self-sustainability. After European settlement and land clearing it is likely that the functionality of the landscape changed, leading to a reduction in the stability and reduced resilience of the landscape. However, the grazing pasture land is also relatively stable and resilient if managed carefully (State 2). There is no evidence that the agricultural land has been degraded sufficiently for it to completely lose resilience against the effects of stressors such as drought. Consequently, the agricultural land is able to remain above the critical level of landscape resilience. After establishment of the BEPs (State 3), the trajectory of the landscape function is upward over time, because in addition to changes to SOC, the AGB is also contributing to the functionality of the landscape by improving the indices. As the BEPs become more established (State 4) the trees become the dominant vegetation thus leading to improved landscape function but not necessarily to increases in SCS in the top 30 cm of the soil profile. Whether BEPs can lead to the same level of landscape functionality and SOC as were present prior to land clearing is uncertain because the results for this case study do not provide convincing evidence that the rate of SOC sequestration is adequate for such a target to be met.

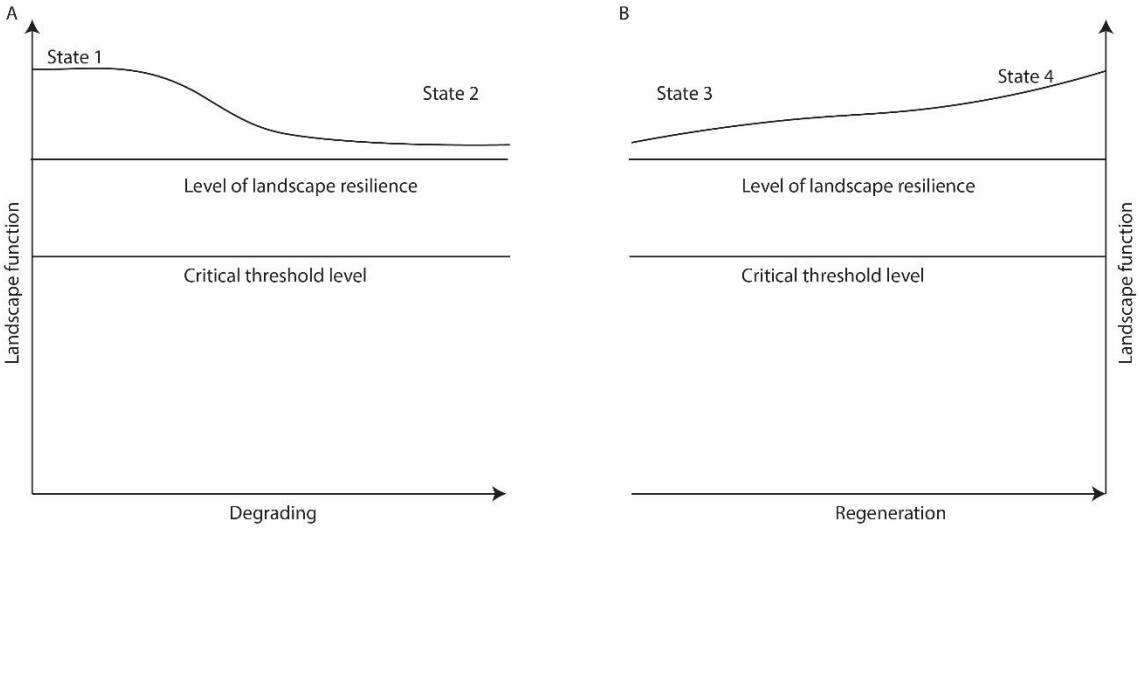


Figure 6-7. Ecosystem response curves for SCS and landscape functionality in the BEP case study where A is the change following LUC from a grassland and eucalypt woodland community to a grazing system and B is the change in landscape function after incorporation of BEPs into the AL. After both LUC interventions the landscape retains resilience against stress and disturbance and remains above the critical threshold for self-sustainability.

6.4.2 Ecosystem response curves for LUC in the waterponding Case Study

Ecosystem response curves showing the trajectories towards degradation and then recovery for the waterponding case study are shown (Figure 6-8). The landscape prior to the introduction of grazing was most likely above the level of landscape resilience and able to recover the effects of stress and disturbance without intervention (State 1) (Figure 6-8 (A)). At this state, vegetation could recover from stress because the soil and SCS (as a surrogate for SOM) was relatively undisturbed. Following the introduction of grazing (State 2), loss of vegetation, drought and erosion, the SCS declined and landscape functionality degraded, eventually falling below the critical threshold for sustainability and thus self-recovery. The amplitude of the degradation was steep, reflecting the relatively short time taken (in landscape terms) for changes of state to occur, and thus for the capacity of the system to recover without intervention. During the early period of degradation, there were possibly some management strategies that could have been used to allow the landscape to regain functionality equivalent to State 1. However, at the time, the need to drastically and rapidly reduce herbivore numbers from the region during drought, before the degradation became irreversible, was not recognised. Therefore, the landscape degraded below the critical threshold for recovery until the scalded landscape that is present in the region developed as State 3. At State 3, the scalded surface is highly resistant to self-recovery, except perhaps at a very slow rate taking perhaps hundreds or thousands of years during which time absolutely minimal further disturbance to the landscape would be required.

In a way, at State 3, the degraded landscape is resilient to any further stress and disturbance because the surface has been effectively sealed by the crust.

The application of the waterponding method has not only led SOC sequestration, it has also changed the state as well as trajectory of the ecosystem response curve. In Figure 6-8 (B), State 4 represents waterponds aged 1 year. At this point the trajectory of landscape functionality changes due to the opportunity for the landscape to become revegetated. The amplitude of the curve at this time is steep due to the relatively rapid increase in SOC to a peak when waterponds are approximately 5 years old. During this period the critical threshold level is surpassed. With careful grazing management to ensure plant cover is retained, the landscape has regained a degree of self-sustainability. At this time, the waterponded landscape reaches the new climax State 5 with sustainable SOC and a variety of perennial plant species. It is expected that over time this landscape will continue to achieve higher levels of resilience provided management is sustainable, but a long geological time to regain the resilience originally present at State 1.

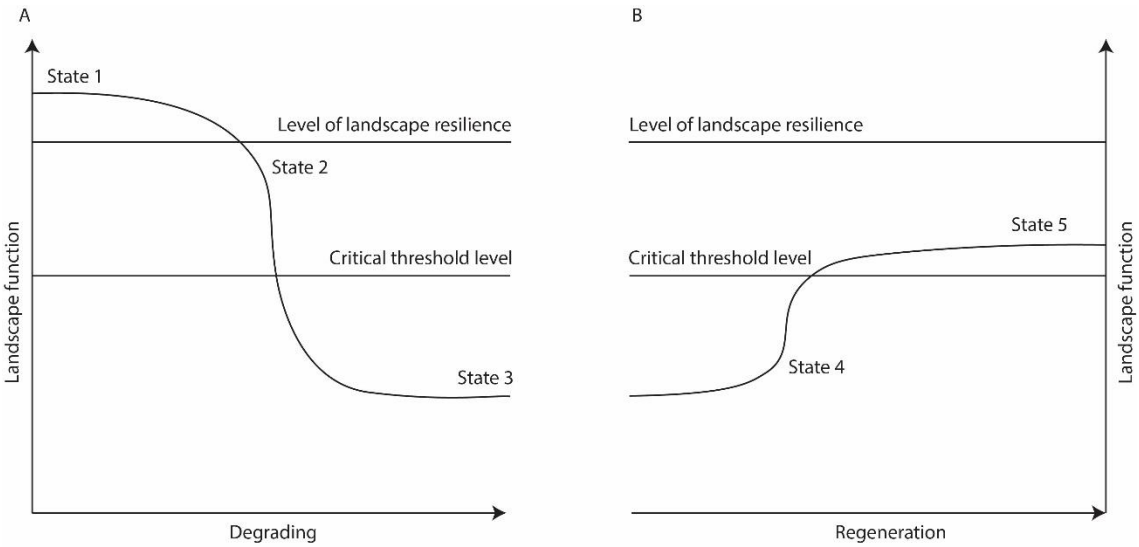


Figure 6-8. Two conceptual ecosystem response curves showing landscape functionality changes for the waterponding case study. The five states are shown in relation to points of stress, disturbance and regeneration of the landscape. 'A' has landscape degradation below the critical threshold level for self-sustainability that occurs following the introduction of grazing and 'B' has improved landscape function with regeneration after active intervention via waterponding. Threshold levels of landscape resilience, and the critical threshold level below which natural regeneration is unlikely, are also shown.

6.5 Model of soil C stock change following LUC

The two environmental rehabilitation activities studied in this research; the reforestation of AL with BEPs and waterponding of scalded claypan soil, have been researched in the expectation that the respective activities will result in sequestration of atmospheric CO_2 in soil. The effect of different LUC methods on SOC sequestration can be concisely summarised by using the conceptual model developed by Johnson (1995) (Figure 2-2). Johnson's conceptual model shows that SOM can be gained or lost depending on land management practices. The empirically measured and quantified SCS data from the current research project has been incorporated onto Johnson's model to demonstrate the effect of anthropogenic LUC interventions on SCS and has enabled a comparison of the results from each Case Study (Figure 6-9). At different states the soil from each case study has been shown to be either a source or sink of atmospheric CO_2.

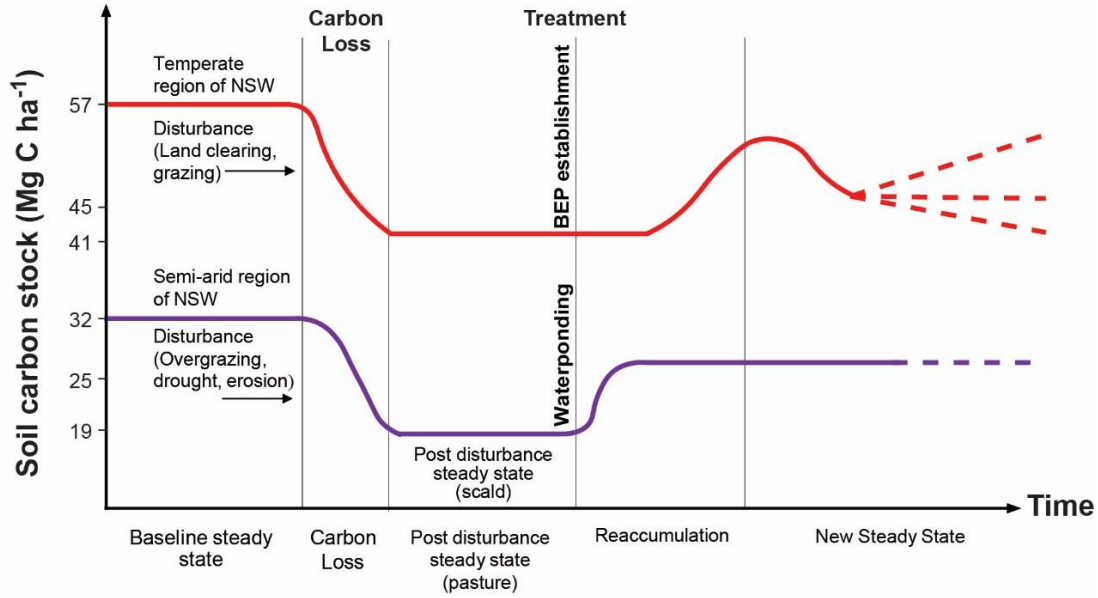


Figure 6-9. Ecosystem response curves showing the flows over time of SCS to 30 cm depth change following LUC for both the BEP (solid red line) and waterponding (solid purple line) case studies (after Johnson, 1995). The dashed red lines represent three possible trajectories for SOC sequestration under BEPs. The dashed purple line represent a hypothetical steady state trajectory for the waterpond case study.

Baseline steady state. The initial baseline steady states (Figure 6-9) represent the quantity of SCS likely to have existed in each case study scenario prior to European settlement. Information about and estimates of pre-European settlement SCS levels are scarce (Gifford, 2010). Notwithstanding the limited available data, possible SOC baseline estimates have been derived from the literature as has been discussed in detail in the respective case study chapters (Sections 4.3 and 5.13). The baseline SCS in the Southern Tablelands of NSW has been estimated to be approximately 57 Mg C ha^{-1} and in the semi-arid rangelands of the Central West of NSW approximately 32 Mg C ha^{-1}. However, both estimates are likely to have high uncertainty. The two case study landscapes were climatically, geographically and biologically dissimilar locations. However, prior to European settlement in Australia, both ecosystems were most likely

to have adapted to, and become resilient against natural disturbances such as fire and native herbivore grazing. At this state the landscapes would have been in a state of ecological and SCS climax. Fluctuations to SCS in this effectively undisturbed landscape would most likely have been in equilibrium with the environment and the CC until the balance between C inputs and decomposition was disrupted following European settlement.

C loss. The C "loss" phase depicted (Figure 6-9) refers to a transition phase incorporating an active intervention which disturbed the baseline soil. In brief, in the Southern Tablelands of NSW, the intervention was removal of trees from grassy woodlands, and in the Central West of NSW the activity was overgrazing of native grasslands and shrublands by livestock and rabbits, exacerbated by drought; this actively led to erosion of the A-horizon. Effectively the transition activities may have resulted in CS losses of up to 16 Mg C ha^{-1} from soils located in the temperate region on the southern tableland soils, and 13 Mg C ha^{-1} from the areas in the semi-arid rangelands that became scalded.

Post disturbance steady state. The mean post disturbance steady-state results from the AL and scalded soils measured in the current research are assumed to represent the lowest SCS possible under the "business as usual" agricultural activities being practiced at the time of sampling. In the case of BEPs, business as usual represents grazing of improved pastures. However, no agricultural productivity or ecological sustainability is possible on the scalded soil. The mean SCS in the temperate region of NSW was found to be 41 Mg C ha^{-1} and in the semi-arid rangeland scalded soil 19 Mg C ha^{-1}. Apart from natural variability including seasonal fluctuations, there is little evidence to suggest that substantial improved SCS is possible for the AL and scalded soil unless some form of active intervention is implemented to initiate a transition towards an increase.

Reaccumulation. Active intervention, (i.e., establishing BEPs and waterponding), is the fourth Transition phase for both case studies and is expected to result in a period of SOC accumulation following establishment of BEPs, or construction of waterponds. Whilst sustainable increases in SCS following establishment BEPs is not apparent, the results following waterponding suggest that SCS accumulation reaches a plateau after approximately five years. With appropriate management it is likely that the SCS levels of the waterponded soil could remain constant, and thereby meet the critical carbon sequestration objective of permanence (Figure 6-9).

New steady state. The results show that waterponding sequesters SOC and that the quantity of C sequestered remains stable, probably permanently, as long as the vegetation is retained. The outcome is less clear for the SCS in BEPs, where the results indicate variability with an increase in sites aged 11-15 years, and then a downward trajectory towards C loss in sites older than 15 years (Figure 6-9). The literature acknowledges the variability in SOC sequestration following afforestation of AL especially in the initial years following afforestation (e.g. Read, 2008; Poeplau and Axel, 2013; Cunningham *et al*. 2015; England *et al.* 2016; Prior *et al*. 2015; Paul *et al*. 2016) with some research showing a loss and others a gain following afforestation. Research

by Paul *et al*. (2002a) show that initially there may be a small gain in soil C in the top 30cm following afforestation with environmental plantings, however, additional C sequestration may take in excess of 30 years before it really becomes apparent.

6.5.1 Summary

This synthesis of the results from two diverse LUC activities has used SAT models to show that changes to SCS states follow a series of transitions in response to LUC activities. The concept of landscape resilience has been used to demonstrate the favoured end point for landscape rehabilitation.

The current research has assumed that SOC is likely to have been lost following European settlement in both environments studied. Active LUC intervention to restore SCS in the top 30 cm of the soil profile by establishing BEPs or waterponds was expected to result in SOC sequestration. The results of the BEP case study do not conclusively support the overarching hypothesis of the current research that LUC will result in sequestration of atmospheric CO_2 in soil, because the long term trajectory of the SCS remains uncertain. The SAT framework has shown the changes in SCS state following a series of LUC interventions. Following the LUC interventions the resilience of the AL and BEPs has remained above critical stress thresholds.

In contrast to the BEP study, the results for the waterponding case study indicate that sequestration of C in soil is possible and is likely to be stored permanently. This case study therefore supports the hypothesis that waterponding is a LUC intervention that will likely result in sequestration of atmospheric CO_2 in soil as long as the rehabilitated landscape is subsequently managed sustainably to prevent deterioration below a critical threshold level and therefore reversion to the scalded state (State 3).

Chapter 7: Conclusion

7.1 General

The sequestration of C in soil is one strategy to mitigate the rising levels of atmospheric CO_2 caused by an ever increasing quantity of anthropogenic emissions. Consequently, the overarching aim of this thesis has been to determine the C sequestration potential of soil that has previously been used for agricultural production, and that has recently undergone LUC. This aim has largely been achieved. The results from the two case studies have demonstrated that SOC sequestration can occur to varying degrees under the two LUC activities (i) reforestation of grazed agricultural land through establishment of BEPs, and (ii) rehabilitation of a highly degraded scalded landscape following waterponding.

Prior to this research project, there was a lack of empirical data on the SOC sequestration potential for the two scenarios explored. Consequently the empirical data presented by the current research project has addressed knowledge gaps pertaining to establishment of BEPs in grazed agricultural land, and to waterponding of scalded clay pan soils. The data from the two case studies has also demonstrated their respective contribution to SCS.

Both case studies had the following common principal aims:

1. To identify the C sequestration potential of agricultural soil following each LUC activity; and
2. To identify biogeochemical factors that contribute to the C sequestration potential of soil after each of two LUC methods have been implemented.

This research has found that the capacity for C to be sequestered in soil is largely dependent upon the previous land use, soil type, the baseline SOC concentration and site effects. In both case studies SOC was shown to increase following LUC. However, the rate of sequestration and final amount sequestered was quite different for each case study. A critical finding of the current research has been the results that clearly show that the magnitude of SCS change is dependent on the baseline SOC level prior to the LUC intervention. Each study was found to present different opportunities and constraints influencing the potential of soil to sequester C following LUC.

7.2 Key findings

The key findings from the two case studies are identified and discussed individually.

7.2.1 Case study 1 – Biodiverse environmental plantings

BEPs are established normally to meet social aspirations and achieve environmental landscape goals and benefits, such as providing shelter to livestock and pastures, sequestering C in the biomass and improving biodiversity; sequestering C in soil is a more recent rationale for BEPs.

The case study addressed the following two research questions:

1. Does afforestation with BEPs lead to sequestration of SOC; and
2. How do the dynamic properties of BEPs influence the SOC sequestration potential?

To address the first research question as to whether BEPs lead to SOC sequestration, the change to the SCS in the 0-30 cm soil layer between the paired AL and BEP was determined. On average, BEPs aged 1-15 years were found to have a higher SCS than the soils in the paired AL position. However, BEPs aged 16-19 years were found to have on average 4.3 Mg C ha^{-1} less SCS compared to their paired AL position. This result suggests that during the first 15 years BEPs sequester SOC, but older sites may not. Total C for BEP sites is discussed later.

The overall mean SCS for the AL in all n=20 sites was determined to be 41.1 Mg C ha^{-1}. BEPs aged 1-5 years had 41.5 Mg C ha^{-1}, the 6-10 year old BEP cohort had 40.6 Mg C ha^{-1}, 11-15 year old BEPs had 49.2 Mg C ha^{-1} and the oldest cohort had 45.6 Mg C ha^{-1}. The trend identified by these data shows that BEPs initially sequester C, but because the SCS (0-30 cm) is lower in the 16-19 year old BEPs than in the 11-15 year old cohort, the long term sequestration potential of BEPs remains uncertain.

With the results of this case study showing variability in SCS between land uses (positions) and BEP age, the changes to SCS are most likely influenced by specific site effects. These effects have been addressed by the second research question which was to investigate the way in which the dynamic properties of BEPs influence the C sequestration potential of soil following incorporation of BEPs into land previously used for agriculture.

The key findings addressing the second research question and four objectives of this case study include:

Objective 1. *Estimated annual rates of SOC sequestration and AGB growth following establishment of BEPs.*

The average rate of SCS change to 30 cm was found to be 0.52 Mg C ha^{-1} yr^{-1} after BEP establishment. However, the results were variable with a one year old BEP achieving the largest loss of 1.7 Mg C ha^{-1} yr^{-1} and a two year old BEP achieving the largest increase of 4.7 Mg C ha^{-1} yr^{-1}.

The annual rate of increase in AGB was found to be 2.4 Mg C ha^{-1} yr^{-1}. The lowest rate of AGB change was 0.1 Mg C ha^{-1} yr^{-1} in a two year old BEP and the highest rate of increase was 6.1 Mg C ha^{-1} yr^{-1} in a 16 year old BEP.

The average AGB C for sites established for < six years was 0.8 Mg C ha^{-1}. Sites aged between nine and 16 years had on average 35.7 Mg C ha^{-1} but with a wide range of values from between 1.4 and 97 Mg C ha^{-1}.

To answer objective 1, the SOC and AGB following establishment of BEPs has been quantified. The average total BEP C stock incorporating SOC, AGB and litter has been calculated to be 83.2 Mg C ha^{-1}. The mean AGB for the 20 BEP sites is 31.4 Mg C ha^{-1}. The mean SCS (to 30 cm) is 44.2 Mg C ha^{-1} and the mean litter pool is 7.6 Mg C ha^{-1}. Proportionally, of the total SCS in BEPs, AGB contributes 38%, SCS contributes 53% while litter contributes 9% of the total C stock in BEPs.

Objective 2. *Compare empirically derived SOC and AGB data with FullCAM/RothC modelled results.*

The correlation between measured BEP SCS results and FullCAM predicted SCS results for MSEPs was weak. The best correlations were achieved for the HOC and ROC pools with correlations of $R^2 = 0.77$ and $R^2 = 1$ respectively. On the other hand predictions for SCS and POC were very weak.

FullCAM modelled results showed good correlations with measured data for AGB ($R^2 = 0.95$, $P<0.001$). This is not surprising due to the work that has been carried out to refine the allometrics for MSEP (e.g., Paul and Polglase, 2004; Paul *et al*. 2013).

The correlation between measured litter and FullCAM modelled results were weak with an $R^2 = 0.29$ for the O1 layer and $R^2 = 0.34$ for the O2 layer.

Therefore to meet Objective 2 for the current research, the findings show that FullCAM provides modelled data for AGB in MSEP that is closely aligned with measured data for BEP AGB and therefore the model is an effective tool for calculating AGB. However, the weak correlations achieved in the current research when comparing FullCAM/RothC modelled data with measured SOC, the functional pools of C, and litter, indicates that further calibration work with FullCAM parameters for soil C pools and litter is required.

Objective 3. *Identify any edaphic and AGB properties that may influence the SOC sequestration potential of the BEPs, including soil EC and pH, BD, TN and TP, C:N, and C:P ratios, landscape features and BEP configuration.*

Soil $EC_{(1:5\ water)}$ was found to be very low in both the AL and BEP and consequently is not considered to affect the SOC sequestration potential of BEPs. However, soil pH becomes lower as BEPs become older. The progressive acidity may be a factor in reducing nutrient availability for tree growth and SOC sequestration.

TN is higher in the BEPs aged between 1 and 15 years than the paired AL. The BEP trees have *Acacia* spp. trees and shrubs which are N fixers and therefore may be contributing to N to the soil. However, in BEPs aged 16-19 the TN concentration is lower than the paired AL. The reduction could be due to plant uptake of N or the lower proportion of live acacia trees within the oldest BEP cohort. The C:N ratio in the AL was found to range between 16 and 25, whereas the BEPs have a range of between 18 and 26. These ratios are considerably higher than the optimum C:N ratio of 12 recommended for sequestering SOC (Himes, 1998; Kirkby *et al*. 2011). Based on the TN and C:N ratio results TN is limiting the SOC sequestration potential of BEPs in the 16-19 year old cohort, but not in the younger BEPs.

In the 0-5 cm layer of BEPs aged between 1 and 10 years the mean TP concentration is slightly higher (325 mg kg^{-1}) than in the paired AL (299 mg kg^{-1}). In sites older than 11 years on average the TP is lower in the BEPs (370 mg kg^{-1}) than the AL (412 mg kg^{-1}). The mean C:P ratio results in the 0-5 cm layer of BEPs is 143 and in the AL it is 119. The results indicate that TP is not limiting the SOC sequestration potential of BEPs.

BEPs established in a linear configuration were found to have higher SCS, tree stocking rate and AGB C, than those laid out in a block configuration. Additionally, linear BEPs sown in a NS direction were found to have more SOC than those sown in an EW direction. Trees located on outer rows were found to have a mean DBH of 5.08 cm, whereas those located in the inner rows had a smaller mean DBH of 3.99 cm. The AGB for tress on outer rows was 7.28 Mg C ha^{-1} and trees located on inner rows had a biomass of 5.11 Mg C ha^{-1}.

Objective 4. *Use MIRS to identify the soil fractions (HOC, POC and ROC) and validate the results with manual fractionation methods using wet sieving and ^{13}C NMR spectroscopy.*

The POC, HOC and ROC fractions in BEPs aged between 1 and 15 years was higher in the BEPs than AL. However, in BEPs aged between 16 and 19 years the POC, HOC and ROC was lower in the BEPs than the AL. The HOC fraction shows the largest change following LUC from AL to BEP, with the mean difference incorporating all sites and ages, being a gain of 2.33 Mg C ha^{-1}. POC had a marginally smaller gain of 2.29 Mg C ha^{-1}. The ROC fraction showed the least change across the 19 year chronosequence with a mean of 1.16 Mg C ha^{-1}. Therefore it can be inferred that BEPs are sequestering C in the HOC stable form which is a positive outcome for efforts to mitigate rising atmospheric CO_2 levels.

Validation and correlation of soil fraction data derived by using MIRS with $TSC_{(LECO)}$, wet sieving and ^{13}C NMR analysis was undertaken. Strong correlation was found beatween MIR and $TSC_{(LECO)}$ with $R^2 = 0.93$. A comparison between MIRS and wet sieving found strong correlation for TOC ($R^2 = 0.95$), POC ($R^2 = 0.81$), and HOC ($R^2 = 0.67$). Finally a comparison between MIRS results with 13 C NMR results found strong correlations for POC and ROC ($R^2 = 0.84$ and $R^2 = 0.95$ respectively, and a moderate correlation for HOC ($R^2 = 0.47$). From these results MIRS is a useful way to accurately, rapidly and cheaply predict changes in soil C pools following LUC.

7.2.2 Case study 2 – Waterponding of scalded clay pan soils

Scalded clay pan soils have no vegetation and very low levels of SOC, and when left untreated they are unlikely to ever have the capacity to sequester atmospheric CO_2. Earlier efforts to rehabilitate scalded soil included livestock exclusion, tyne ripping and ploughing. Each of these methods has been shown to be ineffective at permanently restoring vegetation on the scalded soil. Indeed, waterponding has been found to be the only method that can effectively enable revegetation of scalded soil.

The principal aim of this case study was to quantify changes to SCS to 30 cm depth that occur as a result of waterponding, using a chronosequence approach from an array of sites representing waterponds of different ages. The results have shown that starting from a very low baseline SCS in the scalded soils (18.7 Mg C ha^{-1}), waterponding is an efficient and effective way of sequestering SOC, with waterponds aged five years having on average 26.3 Mg C ha^{-1}. Overall, waterponding leads to a sustainable SCS increase of approximately 7.6 Mg C ha^{-1}. Therefore, the sequestration of SOC in scalded soil following waterponding is an effective opportunity to contribute towards the mitigation of atmospheric CO_2 emissions.

To support the aim, the following three objectives have been investigated:

Objective 1. *Quantify the rate of SOC sequestration over time following waterponding.*

Most of the SCS increase was found to occur in the 0-5 cm layer with a mean of increase of 0.8 Mg C ha^{-1}. The rate of SCS increase in the 30 cm layer is approximately 1.5 Mg C ha^{-1} yr^{-1} during the first five years after waterponds have been established. The results found that waterponds older than five years did not increase SCS and thus are most likely in a state of equilibrium. The rate of SCS sequestration during the first five years after construction is at the upper end of SOC sequestration rates. However, the increase in SCS during the first five years following waterponding has been shown to be sustainable with an equilibrium SCS being reached approximately five years following establishment of the waterponds.

Objective 2. *Identify whether any physico-chemical properties including TN, C:N ratio, TP, C:P ratio, AP, C:AP ratio, TS, C:S ratio, pH, EC, ESP and ASWAT, contribute to the SOC sequestration potential of scalded soil following waterponding.*

TN was found to increase in waterponded soil. For example the TN concentration in the 0-5 cm layer of the scalded soil was 645 mg kg^{-1}, and in the 25-27 year old waterponds the mean TN concentration was 831 mg kg^{-1}. The scalds were found to have an average C:N ratio of 7 in the 0-5 cm layer, whereas the waterponds had a C:N ration of between 9 and 11 in the 0-5 cm layer The increase in TN concentration and C:N ratio following waterponding has been attributed to organic N being added to the waterponded soil from plant residues and decomposition of SOM. The increases are thus a result of waterponding and subsequent vegetation growth and indicate an overall improvement in nutrient cycling and landscape functionality.

TP was found to be higher in the 0-5 cm layer of scalds than the waterponds (157 mg kg^{-1} and between 76 and 145 mg kg^{-1} respectively). The reason for the lower TP in the waterponds is possibly associated with biomass uptake and mineralisation of the organic P pool. The scalds were found to have a mean C:P ratio of 48 in the 0-5 cm layer and waterponds had a mean C:P ratio of between 64 and 99. The lower TP concentration and wider C:P ratio indicate that P deficiency is possibly limiting C sequestration in the waterponds.

The reservoir of AP in the scald soil provides essential nutrients for plant growth following waterponding. The results suggest AP is rapidly taken up following waterponding with scalds having a concentration of 19.5 mg kg^{-1} and waterponds aged 1 year having only 8.2 mg kg^{-1}. The lack of AP in the older waterponds, and widening of the C:AP ratio are likely to be a constraints on further increases to SOC sequestration in waterponds in the longer term.

The concentration of TS was found to be adequate in both scalds and waterponds. The mean C:S ratio was 82 in the 0-5 cm layer of the scalded clay pan soil and waterponds aged 5 years had the highest C:S ratio in the 0-5 cm layer of 122. Waterponds aged 25-27 years had a C:S

ratio of 103. Overall, the C:S ratio was found to not limit the long term sequestration of SOC following waterponding.

The scalded soil is saline and sodic. The $EC_{(1:5\ water)}$ was significantly lower in waterponded soils than in scalded soils. This change occurs because after the waterponds have been constructed, water is trapped and infiltrates the soil profile thereby leaching soluble salts. After soluble salts have been leached, the soil becomes more sodic with an ESP that is > 15. The waterponded surface is able to support a diversity of perennial plants thereby revegetating the scalded surface. The plants provide fodder for grazing livestock and result in the sequestration of SOC if managed sustainably.

Soil $pH_{(1:5\ water)}$ was within the neutral to slightly alkaline range and significantly higher in waterponds than in scalds. The pH increase in the waterponds has been attributed to the leaching of soluble salts from the soil profile, the net effect of which results in an increase in the OH^- concentration of the soil solution. The change to pH can change the net negative charge on clay particles and result in a commensurate increase in the dispersive behaviour of soil.

The scalded soil has an ESP > 15 but does not disperse due to the high soluble salt concentration. After waterponding the soluble salts are leached and the saline and sodic clay pan scald soil becomes non-saline sodic with the ESP remaining > 15. The ASWAT results show that waterponded soils disperse spontaneously. Although dispersion is normally considered a detrimental soil characteristic associated with soil structure degradation, in the case of the waterponds it is an important positive characteristic enabling the surface crust to break down. In addition the leaching of soluble salts restores the shrink swell behaviour of the soil and thereby allows deep soil cracks to form in the profile. These changes enable water to infiltrate into the profile, cracks to form, seed then becomes lodged in the cracks and can grow. Therefore, the waterponding of scalded clay pan soil is important for changing the dispersive behaviour and shrink swell behaviour of the soil and consequently improving the SOC sequestration potential of the landscape.

Objective 3. *Determine whether waterponding leads to clay mineralogical changes that may affect the C sequestration potential of the soil.*

On a very small sample subset, PSA has shown that waterponds have a higher percentage of sand sized aggregates than the scalded soil. XRD analysis also shows that a higher percentage of quartz and illite is present in the waterponds. The change occurs after only 25 years and possibly may be caused by transport of particles or aggregation via organo mineral particle binding. These changes contribute to soil structural improvement in the waterponded soil which will provide resources that could enable ongoing SOC sequestration.

There may be evidence (given the sample size), that weathering sequences are initiated by waterponding which cause transitions of, for example, plagioclase into smectite, and illite into

smectite + vermiculite. Calcite and halite are probably leached down the profile. They are present lower in the waterpond soils than on the scalds due to their solubility and subsequent leaching which is initiated by water infiltration in the waterponds.

SOC sequestration in the waterponds cannot be directly attributed to the mineralogical changes that occur to scalded soil following waterponding. However, the changes can be viewed more as contributing to soil structural improvements, which over time may provide resources or mechanisms important for SOC sequestration and its long term physical protection of that C.

7.3 Limitations and future areas of research

The methods used in this research project have been successful in demonstrating temporal and spatial changes to SOC in each of the case studies. In particular, the methods used to measure and analyse the SOC results have given a useful indication of the SOC sequestration potential of the two LUC methods studied. The research has however, highlighted a number of areas of uncertainty in the C sequestration of soil following LUC. In addition, as a result of the current research some key areas of uncertainty regarding the outcomes of LUC on SOC sequestration have emerged. Some of the limitations, suggested improvements and areas of study that would benefit from further research are discussed below.

7.3.1 Biodiverse environmental plantings

(1) The BEPs included in this case study were all established using a direct drill method which often results in a high tree stocking rate. BEPs planted with tube stock typically result in lower tree stocking rates than seen in direct seeded plantings. Therefore, the results of this research are not to be assumed to be representative of BEPs established using tube stock or other establishment methods. Additional research encompassing different methods to establish BEPs should be undertaken to determine whether there is any resultant difference in tree stocking rates. This would have relevance for calibrations with FullCAM and National C accounting.

(2) An important part of the study design for this research was to use a chronosequence of paired sites in the absence of being able to undertake a longitudinal study. Paired site strategies are commonly used in research when baseline, or pre-treatment analysis was not made and / or is not possible. However, they have limitations where the pairing may not be ideal due to geological, geomorphic, landuse or land tenure constraints. Additionally, the variation in SCS in the AL must be explored further because the effects of LUC on the AL following establishment of the BEPs is unknown. For example, changes to management practices, such as fertilizer additions, may have caused the apparent temporal increase in SCS in the AL. A change to land ownership that often results in establishment of new land uses such as BEPs is often accompanied by other changes such as animal stocking rates, types of livestock and / or other

changes that allow the AL soil to evolve. A follow up longitudinal study using the same sites will confirm whether the apparent trend whereby SOC is lower in the BEP than the adjoining AL in the oldest cohort of BEPs, is an outcome of deficiencies in the study design of the current research project, or is indeed representative of what is actually occurring.

(3) An original objective of the current research project was to develop a simple model for use by landholders to predict changes in SOC by measuring the AGB. The results have shown that there is little or no correlation between AGB growth and SOC sequestration. This outcome may simply be a result of SOC heterogeneity across the landscape. However, the use of randomly located, 10 m tree transects, replicated 3 times at each site has provided insufficient data to draw any useful comparison between AGB and SOC. It is likely that the study design and the selection of a suitable range of sites, constrained by the number of sites present in the biogeographic region contributed to the weak AGB and SOC relationship. For example, the final combination of site replicates encompassing age, configuration (block or linear), and topographic position (crest, mid or low slope), was constrained by the number of sites available for each combination, and the resources available to sample and analyse additional sites. This constraint led to the need to undertake analysis on an age-cohort basis rather than an individual age basis. Ideally, a larger number of each site configuration, replicated, would have given stronger statistical outcomes. Future studies should attempt to obtain access to a larger number of sites with similar ages, configurations and topographic position to gain a better understanding of the effects of spatial variability and environmental heterogeneity on the SOC sequestration potential of BEPs.

(4) The temperate Southern Tablelands region of NSW had been under drought conditions for almost 10 years immediately prior to the current research being undertaken. The drought most likely would have contributed to a depletion in water resources and probably constrained the nutrient availability and biotic activity for the BEPs during that time. This may be an explanation for the lower tree stocking rate in the younger sites, and might have limited AGB growth rates and SOC sequestration in the older sites. Therefore, drought may have had a detrimental effect on the SOC sequestration rate found in the older BEP sites. This could be determined with follow-up analysis of the same sites at 5 yearly intervals.

(5) This research identified variation in the changes to SCS following afforestation of agricultural land with BEPs especially in the older sites. Some of the variability can be attributed to nutrient resource availability. Consequently, new research into the

application of fertilisers to determine whether increased nutrient availability can promote further SOC sequestration is recommended.

(6) Correlations between measured data for SOC, soil fractions and litter were found to be generally weak. Consequently, at least in the biogeographic region in which this study was undertaken, additional calibration and validation is required before FullCAM modelled data for these parameters can be used with confidence.

(7) Further research involving a dust interception study needs to be conducted to identify and quantify dust being intercepted by BEPs to determine whether the dust is contributing to changes in nutrient, SOC or EC properties within the BEPs.

(8) Correlation analysis was undertaken for SOC concentrations obtained using LECO CNS2000 dry combustion analysis, and Heanes wet oxidation analysis. Validation between the methods would be improved with inclusion of analysis using Kjeldahl N and Dumas N.

(9) Analysis of the C and N concentration of the sampled litter from the BEPs was not conducted during this research. To allow determination of the C content and therefore the contribution to the NPP of trees in the BEPs it would be useful to conduct C and N analysis and C:N ratio of the litter pool in future research.

(10) The aim of the research has been primarily to determine the SOC sequestration potential of BEPs by calculating the difference in SOC between BEP and the adjacent AL. Consequently the specific area of the TL and IR positions within BEPs was not calculated. Further, such a measurement would be extremely difficult due to the variability in canopy cover within BEPs due to plot age, tree size and distance between rows. The area of a BEP per se was calculated based on the overall dimensions of the plot. This approach has been used for by other researchers who have examined the C sequestration potential of mixed species environmental plantings referred to throughout this thesis (eg Paul *et al.* 2014; England, *et al.* 2016; Cunningham *et al.* 2012 & 2015).

(11) A limitation of the research has been that the plant biomass and its relative relationship with SOC in the AL position was not measured. Future research should include the allocation of plant biomass measurements for the AL as well as the BEPs.

(12) A limitation of the research has been that landholders were not questioned about their management strategies and fertiliser history. This could have provided an explanation for the trend with time whereby the SCS was higher the AL position in the older BEP sites. Future research should include exploration of management practices that could be influencing the C sequestration potential of the soils being studied.

(13) The research project obtained C analysis results using different analytical methods at a variety of analytical laboratories (Sections 3.2.2 and 3.2.3). The results are deficient in that no standard reference material was utilised. Therefore comparative analysis of the

C concentrations between the different methods includes a level of uncertainty. In future research, standard reference material should be utilised to ensure validity of analytical results.

7.3.2 Waterponds

Sixteen important study design limitations have been identified for this case study.

(1) The scald and waterpond sites included in this case study were spread along a 60 km long transect. Therefore, it is likely that some climatic and soil physical, chemical and biological variation is present across that distance. This approach, although allowing the best use of limited resources, is a shortcoming of the current research project;

(2) The soil samples were obtained from three scald sites only, whereas there were 12 waterpond sites. Consequently the current research was conducted as a comparative rather than paired site study. A more robust approach would have been to use paired sites. The latter would provide more certainty with regard to the spatial difference and temporal rate of change in SCS, and other physico-chemical changes following waterponding;

(3) The SOC sequestration potential of the waterponded soil may be better understood if baseline (analogue) SCS results are obtained from surrounding tree lines along prior streams, relic soils and uneroded sites. Analysis of these analogue sites will help to determine a baseline C content representative of pre-European settlement SOC levels;

(4) The results have found that SOC equilibrium has been achieved by the older waterponds. Further analysis as to the long term storage of the SOC in waterponds is recommended for National C Accounting purposes; and

(5) A chronosequence approach was used to enable calculation of temporal rates of SCS change. Chrono-sequencing is a useful method, but has limitations. It assumes that all other soil factors are equal when compared across time. However, it is widely acknowledged that most physical, chemical and biological soil factors are extremely heterogeneous spatially and temporally, and in the case of SOC, amongst others, seasonal fluctuation are omnipresent.

(6) The use of a relatively small sub-set sample sizes was used for some analysis which may have constrained the statistical vigour of some results. For instance, the mineralogical changes in particular, were only conducted on one scald core and one waterpond core. To overcome the deficiency in replicates, the comparison was done on the two extremes to maximise the potential of identifying any change e.g., (i) a scald core was taken as representing the baseline conditions, and (ii) one core from a closely paired old waterpond was use to find out whether sufficient time had elapsed to show mineralogical change. Therefore, any conclusions drawn have been based on

a most likely cause and effect. Future mineralogical analysis of scalded and waterponded soils should be undertaken on a chronosequence, or longitudinal study basis involving a larger number of replicates. There are considerable resource implications for such a recommendation.

(7) As stated, mineralogical changes to scalded soil following waterponding were only investigated on two soil cores representing a paired site between a scald and a 27 year old waterpond, and the study was not replicated. A more meaningful study using a larger number of replicates and incorporating preferably a longitudinal study or alternatively a chronosequence of sites would allow a more robust examination of the changes. Additionally, mineralogical analysis would have been enhanced with the inclusion of SEM analysis to visually identify changes.

(8) A study exploring the mineralogical changes is important as it relates to the long term stability of the sequestered C within the soil profile and because permanence of the sequestration is needed to ensure the long-term reduction of atmospheric GHGs. Of particular relevance would be an investigation of the role poorly-crystallised Fe and Al have on stabilising SOC following waterponding. This is because Fe oxides play important roles in SOC sequestration including contribution to soil aggregation and cementation, and because of their large surface area they also have an important role in sorption of SOM.

(9) Soil BD is a critical measure used to calculate SCS. The nutty structure of the scald soil, the presence of soil cracks in the waterponded soil and the soil core sampling method used, all contributed to some degree towards the variability in soil BD results. Preferably, in future research on these soils, a more robust sampling method should be used to provide greater confidence in the soil BD results.

(10) To further clarify the soil sequestration potential of scalded soils on clay pans across Australia it would be useful to measure the extent of scalded soil in Australia. This assessment would help to determine the extent of restoration and potential SOC sequestration for C accounting purposes.

(11) This research has not measured potential GHG emissions, including NO_2 and CH_4 that can occur in conjunction with the trapping of water and presence of organic material within waterponds. It is important to understand both the gains and losses of GHGs to determine whether waterponding successfully leads to a net gain of SOC or whether the emissions result in an offset against the amount of CO_2 sequestered in the soil after waterponding.

(12) The results show that there can be up to a 73% increase in soil C concentration in the 20-30 cm layer. Consequently, there is potential for sequestration of C deeper than 30 cm where it is probably stored in the stable C pool and can be protected from decomposition and disturbance. Although outside international C accounting

parameters, storing C at this depth has important implications for C sequestration and therefore deserves further study.

(13) PSA was conducted on a small number of samples only, but has exposed a trend whereby the presence of coarser grained particles in waterponds may arguably be the preliminary stage of formation of an A-horizon. The soil texture change should be explored further to gain a better understanding of aggregate development associated with waterpond age. Characterisation of the quartz using $\delta^{18}O$ could provide evidence of the aeolian origin of the quartz; and

(14) Further ^{137}Cs analysis incorporating scalds, analogue sites and hummock soils would help derive a clearer understanding of the erosion processes occurring in the region.

(15) The 20 subset 3 samples composited for ^{137}Cs analysis were also analysed for TC using an Elementar VarioMax analyser. The results were not comparatively analysed using a LECO CNS2000 analyser. For optimum comparison, it is desirable to use the same analyser throughout a single research project. Where this is not possible (as was the case in this research project), an alternative compromise should include verification of consistency between the analytical methods.

(16) A water balance study to highlight the effect of waterponds on soil water, evaporation and transpiration for waterponds was not undertaken as it was beyond the scope of the research. To link the finding of the current research project that waterponding leads to a rapid increase in SCS until a quasi-equilibrium after approximately five years of waterponding it is suggested future research consider including a water balance study.

(17) TS has been identified in the current research as a component of the salts in both the scald and waterpond soil. Jones, 1969 also identified Chloride as a major component whereas the current research identified sulfate as being more dominant. Future research should include analysis to identify the presence of sulfate and choloride and changes brought about following waterponding.

7.4 Concluding remarks

Sequestration of C in soil can be an effective strategy to mitigate the effects of anthropogenic atmospheric CO_2 emissions. Adoption of LUC activities can be used by land managers who want to increase SOM for on-farm productivity benefits and also to gain environmental improvements. Often LUC has the added benefit that increases to SCS can be traded for C credits under the Commonwealth Governments Emissions Trading Scheme (ETS). However, much research has been undertaken that shows LUC can result in either a nett loss, nett gain or steady state of SCS, often with variability at temporal and spatial sales. The empirical data provided by this research will be valuable for providing evidence of the importance of LUC and outcomes in terms of SOC sequestration. The implications of the research results are important, informing decision makers responsible for programmes and policies; mitigating against climate change through SOC sequestration and for land managers seeking to trade on the C market.

References

Abell, R.S., 1992. Canberra (1:100 000 scale geological map 8727 1st Edition). Bureau of Mineral Resources, Canberra.

ABARE-BRS 2010, *Land use of Australia,* Version 4, 2005-2006. Australian Bureau of Agricultural and Resource Economics and Sciences, Canberra.

Albrecht, W.A., 1938. Loss of soil organic matter and its restoration. Soils and Men, in *Yearbook of Agriculture,* USDA, Washington, DC, pp 347-360.

Alemseged, Y., Hacker, R.B., Smith, W.J. and Melville, G.J., 2011. Temporary cropping in semi-arid shrublands increases native perennial grasses. *The Rangeland Journal*, **33**, 67-78.

Allison, L.E., 1960. Wet-combustion apparatus and procedure for organic and inorganic carbon in soil. *Soil Science Society of America Proceedings,* **24**, 36-40.

Alves, M.E. and Omotoso, O. 2009. Improving Rietveld-based clay mineralogic quantification of oxisols using Siroquant, Soil Science Society of America Journal, **73**, 2191–2197.

Anjani, J.I., Keith, H., Blakers, M., Mackey, B.G. and King, H.P., 2013. Comprehensive carbon stock and flow accounting: a national framework to support climate change mitigation policy. *Ecological economics,* **89**, 61-72.

Archibald, R.D., Harper, R.J., Fox, J.E.D. and Silberstein, R.P., 2006. Tree performance and reet-zone salt accumulation in three dryland Australian plantations. *Agroforestry Systems*, **66**, 191-204.

Attiwill, P.M., 1979. Nutrient cycling in a Eucalyptus oblique (L'Herit.) Forest. III* Growth, biomass and net primary production. *Australian Journal of Botany,* **27,** 439-438.

Baah-Acheamfour, M., Chang, S.X., Carlyle, C.N. and Bork E.W., 2015. Carbon pool size and stability are affected by trees and grassland cover types within agroforestry systems of western Canada. *Agriculture, Ecosystems and Environment*, **213**, 105-113.

Baddock, M., Parsons, K., Strong, C., Leys, J. and McTainsh, G. 2015. Drivers of Australian Dust: a case study of frontal winds and dust dynamics in the lower lake Eyre basin, *Earth Surface Processes and Landforms*, **40,** 1982-1988.

Badgery, W.B., Simmons, A.T., Murphy, B., Rawson, A., Andersson, K.O., Lonergan, V.E. and van de Ven, R., 2013. Relationship between environmental and land-use variables on soil carbon levels at the regional scale in central New South Wales, Australia. *Soil Research,* **51**, 645-656.

Baker, T.G. and Attiwill, P.M., 1985. Above-ground nutrient distribution and cycling in *Pinus Radiata* D.Don and *Eucalyptus obliqua* L'Herit forests in southeastern Australia. *Forest Ecology and Management*, **13**, 41-52.

Baldock, J.A. and Broos, K. 2012. Soil Organic Matter. In *Handbook of Soil Sciences Properties and Processes*, second edition, pp. 11-1 to 11-52, (eds P.M. Huang, Y. Li, and M.E. Sumner). CRC Press, Boca Raton, FL, USA.

Baldock, J.A. and Skjemstad, J.O., 1999. Organic soil carbon/soil organic matter. In '*Soil analysis: An interpretation Manual*'. (Eds KI Peverill, LA Sparrow, DJ Reuter) pp 159-170. (CSIRO Publishing: Collingwood, Victoria).

Baldock, J.A. and Skjemstad, J.O., 2000. Role of the soil matrix and minerals in protecting natural organic materials against biological attack, *Organic Geochemistry*, **31**, 697-710.

Baldock, J.A., Wheeler, I., McKenzie, N. and McBratney, A., 2012. Soils and climate change: potential impacts on carbon stocks and greenhouse gas emissions, and future research for Australian agriculture, *Crop and Pasture Science*, **63,** 269-283.

Baldock, J.A., Sanderman, J., Macdonald, L.M., Puccini, A., Hawke, B., Szarvas, S. and McGowan, J., 2013a. Quantifying the allocation of soil organic carbon to biologically significant fractions. *Soil Research,* **51,** 561-576.

Baldock, J.A., Hawke, B., Sanderman, J. and Macdonald, L.M., 2013b. Predicting contents of carbon and its component fractions in Australian soils from diffuse reflectance mid-infrared spectra. *Soil Research,* **51**, 577-595.

Banfield, J.F. and Eggleton, R.A., 1990. Analytical transmission electron microscope studies of plagioclase, muscovite and K-feldspar weathering. *Clays and Clay Minerals,* **38**, 77-89.

Barrow, N.J., 1969. The accumulation of soil organic matter under pasture and its effect on soil properties. *Australian Journal of Experimental Agriculture and Animal Husbandry*, **9**, 437-444.

Barson, M., Randall, I. and Bordas, V., 2000. *Land cover changes in Australia. Results of the collaborative Bureau of Rural Sciences – State agencies' Project on Remote Sensing of Land Cover Change*. Bureau of Rural Science, Canberra.

Batjes, N.H., 1992. Organic matter and carbon dioxide. In: *A review of soil factors and processes that control fluxes of heat, moisture and greenhouse gases,* pp 97-148 (eds N.H. Batjes and E.M. Bridges). Technical Paper 23, International Soil Reference and Information Centre, Wageningen.

Batjes, N.H., 1996. Total carbon and nitrogen in the soils of the world, *European Journal of Soil Science, June 1996, **47**,* 151-163.

Batjes, N.H., 1998. Mitigation of atmospheric CO2 concentrations by increased carbon sequestration in the soil, *Biology and Fertility of Soils,* **27,** 230-235.

Beadle, N.C.W., 1948a. The vegetation and pastures of western New South Wales with special reference to soil erosion. Department of Conservation of New South Wales, Sydney.

Beadle, N.C.W., 1948b. Studies in wind erosion I; The origin, general appearance and classification of scalds. *Journal of Soil Conservation, New South Wales,* **4**, 30-35.

Beadle, N.C.W., 1948c. Studies in wind erosion II; Some characteristics of scalded surfaces. *Journal of Soil Conservation, New South Wales,* **4**, 64-68.

Beadle, N.C.W., 1948d. Studies in wind erosion III; Natural regeneration on scalded surfaces. *Journal of Soil Conservation, New South Wales,* **4**, 123-134.

Beadle, N.C.W., 1948e. Studies in wind erosion IV; Reclamation of scalds: general

Berg, B. and Laskowski, R., 2005. Litter decomposition: A guide to carbon and nutrient turnover, *Advances in Ecological Research*, **38**, 1-17.

Berndt, R.D. and Coughlan, K.J., 1976. The nature of changes in bulk density with water content in a cracking clay. *Australian Journal of Soil Research,* **15**, 27-37.

Bernstein, L., 1975. Effects of salinity and sodicity on plant growth. *Annual Review of Phytopathology,* **13**, 295-312.

Berthrong, S.T., Jobbagy, E.G. and Jackson, R.B., 2009. A global meta-analysis of soil exchangeable cations, pH, carbon, and nitrogen with afforestation. *Ecological Applications,* **19**, 2228-2241.

Bhattacharyya, T. Pal, D.K. and Deshpande, S.B., 1997. On kaolinitic and mixed mineralogy classes of shrink-swell soils. *Australian Journal of Soil Research,* **35**, 1245-1252.

Bigham, J.M. Fitzpatrick R.W. and Schulze, D.G., 2002. Iron oxides, in *Soil Mineralogy with Environmental Applications,* SSSA Book Series no. 7, Soil Science Society of America, Madison, USA, pp 323-366.

Bird, P.R., Jackson, T.T., Kearney, G.A. and Williams, K.W., 2002. Effect of two tree windbreaks on adjacent pastures in south-western Victoria, Australia. *Australian Journal of Experimental Agriculture,* **42**, 809-830.

Bird, P.R., Jackson, T.T., Kearney, G.A. and Roache, A., 2007. Effects of windbreak structure on shelter characteristics. *Australian Journal of Experimental Agriculture,* **47**, 727-737.

Bloom, P.R., Skyllberg, U.L. and Sumner, M.E., 2005. 'Soil acidity' in *Chemical Processes in Soils*, Soils Science Society of America Book Series, No. 8. Madison, WI, USA, pp 411-459.

Boon, K.F., Kiefert, L., and McTainsh, G.M., 1998. Organic matter content of rural dusts in Australia. Atmospheric Environment, **32**, 2817-2823.

Bowler, J.M., 1976. Aridity in Australia: Age, origins and expression in aeolian landforms and sediments. *Earth-Science Reviews,* **12**, 279-310.

Bowman, G. Murphy, B. Chapman, G. Wilson, B. Jenkins, B. Koen, T. Gray, D. Morand, D. Atkinson, G. Murphy, C. Murrell, A. and Milford, H., 2009. *Protocols for NSW Soil Condition and Land Capability Monitoring: Assessment of soil erosion, acidification, salinity, carbon, structure, ground cover, land and soil capability, and land management.* Department of Environment and Climate Change NSW, February 2009.

Brack, C.L. and Richards, G.P., 2002. Carbon accounting model for forests in Australia, *Environmental Pollution*, **116**, 187-194.

Brack, C.L., Richards, G. and Waterworth, R., 2006. Integrated and comprehensive estimation of greenhouse gas emissions from land systems. *Sustainable Science*, **1**, 91-106.

Brandon, T., Bestelmeyer, J.R., Brown, K.M., Havstad, R.A., Chavez, G. and Herrick, J.E., 2003. Development and use of State-and-Transition models for rangelands. *Journal of Range Management*, **56**, 114-126.

Bremner, J.M., 1996. Nitrogen-Total in Methods of Soil Analysis. Part 3. Chemical Methods (SSSA Book Series no. 5 (pp 1085-1119).

Briske, D.D., Fuhlendorf, S.D. and Smeins, F.E., 2005. State-and-transition models, thresholds and rangeland health: a synthesis of ecological concepts and perspectives. *Rangeland Ecology and Management,* **58**, 1-10.

Broadbent, F.E., 1953. The soil organic fraction, *Advances in Agronomy* **5,** 153-183.

Bronson, K.F., 2008. Forms of inorganic nitrogen in soil. In: Nitrogen in Agricultural Systems, Agronomy Monograph 49. American Society of Agronomy, Madison, WI, pp 31-55.

Bronik, C.J. and Lal, R., 2005. Soil structure and management: a review. *Geoderma*, **124**, 3-22.

Cain, S.A., 1939. The climax and its complexities. *The American Midland Naturalist*, **21**, 146-181.

Cambardella, C.A. and Elliott, E.T., 1993. Carbon and nitrogen distribution in aggregates from cultivated and native grassland soils. *Soil Science of America Journal*, **57**, 1071-1076.

Cambardella, C.A., Moorman, T.B., Novak, J.M., Parking, T.B., Karlen, D.L., Turco, R.F. and Konopka, A.E., 1994. Field-scale variabaility of soil properties in central Iowa soils. *Soil Science Society of America Journal*, **58**, 1501-1511.

Cattle, S.R. 2016. The case for a southeastern Australian Dust Bowl, 1985-1945. *Aeolian Research*, **21**, 1-20.

Chan, K.Y., Cowie, A., Kelly, G., Singh, B. and Slavich, P. 2008. Scoping Paper: Soil organic carbon sequestration potential for Agriculture in NSW. NSW DPI Science and Research Technical Paper, September 2008. NSW Department of Primary Industries.

Chan, K.Y. Oates, A. Li, G.D. Conyers, M.K. Prangnell, R.J. Poile, G. Liu, D.L. and Barchia, I.M., 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, K.Y., Conyers, M.K., Li, G.D., Helyar, K.R., Poile, G., Oates, A. and Barchia, I.M., 2011. Soil carbon dynamics under different cropping and pasture management in temperate Australia: Results of three long-term experiments, *Soil Research,* **49**, 320-328.

Chappell, A., Baldock, J. and Sanderman, J., 2015. The global significance of omitting soil erosion from soil organic carbon cycling schemes. *Nature Climate Change*, DOI: 10.1038/NCLIMATE 2829. 26 Oct 2015 online.

Chartres, C.J., 1982. The pedogenesis of desert loam soils in the Barrier Range, western NSW 1. Soil parent materials. *Australian Journal of Soil Research,* **20**, 269-281.

Chartres, C.J. Chivas, A.R. and Walker, P.H., 1988. The effect of aeolian accessions on soil development on granitic rocks in south-eastern Australia. 11.* Oxygen-isotope, mineralogical and geochemical evidence for aeolian deposition. *Australian Journal of Soil Research,* **26**, 17-31.

Chartres, C.J., 1993. Sodic soils: an introduction to their formation and distribution. *Australian Journal of Soil Research,* **31,** 751-60.

Chatterjee, A., Lal, R., Wielopolski, L., Martin, M.Z. and Ebinger, M.H., 2009. Evaluation of different soil carbon determination methods, *Critical Reviews in Plant Science*, **28**, 164-178.

Chen, C.R., Condron, L.M., Davis, M.R. and Sherlock, R.R., 2000. Effects of afforestation on phosphorus dynamics and biological properties in a New Zealand grassland soil. *Plant and Soil*, **220**, 151-163.

Chen, X.Y., 1997. Pedogenic gypcrete formation in arid central Australia. *Geoderma,* **77**, 39-61.

Chenu, C. and Plante, A.F. 2006. Clay-sized organo-mineral complexes in a cultivation chronosequence: revisiting the concept of the 'primary organo-mineral complex'. *European Journal of Soil Science*, **57**, 596-607.

Chertkov, V.Y., 2003. Modelling the shrinkage curve of soil clay pastes. *Geoderma*, **112**, 71-95.

Chorom, M., Rengasamy, P. and Murray, R.S., 1994. Clay dispersion as influenced by pH and net particle charge of sodic soils. *Australian Journal of Soil Research*, **32**, 1243-1252.

Chung, H., Grove, J.H. and Six, J., 2008. Indications for soil carbon saturation in a temperate agroecosystem. *Soil Science Society of America Journal*, **72**, 1132-1139.

Churchman, G.J. 2010. Is the geological concept of clay minerals appropriate for soil science? A literature-based and philosophical analysis. *Physics and Chemistry of the Earth*, **35**, 927-940.

Churchman, G.J. and Lowe, 2012. Alteration, Formation, and Occurrence of Minerals in Soils. In *Handbook of Soil Sciences Properties and Processes*, second edition, pp. 20-1 – 20-72, (eds P.M. Huang, Y. Li, and M.E. Sumner). CRC Press, Boca Raton, FL, USA.

Churchman, G.J. Skjemstad, J.O. and Oades, J.M., 1993. Influence of clay minerals and organic matter on effects of sodicity on soils. *Australian Journal of Soil Research,* **31**, 779-800.

Churka Blum, S., Lehmann, J. and Solomon, D., 2013. Sulfur forms in organic substrates affecting S mineralization in soil. *Geoderma*, **200-201**, 156-164.

Cingolani, A.M., Noy-Meir, I. and Diaz, S., 2005. Grazing effects on rangeland diversity: a synthesis of contemporary models. *Ecological Applications*, **15**, 757-773.

Clapp, C.E., Hayes, M.H.B., Simpson, A.J. and Kingery, W.L., 2005. Chemistry of Soil Organic Matter. In: *Chemical Processes in Soils*, SSSA Book Series No 8.

Clements, F.E., 1916. Plant succession: an analysis of the development of vegetation. Carnegie Institution of Washington. Publication No. 242.

Cleugh, H., Prinsley, R., Bird, R.P., Brooks, S.J., Carberry, P.S., Crawford, M.C., Jackson, T.T., Meinke, H., Mylius, S.J., Nuberg, I.K., Sudmeyer, R.A., and Wright, A.J., 2002. Australian National Windbreaks Program: overview and summary of results, *Australian Journal of Experimental Agriculture*, **42,** 649-664.

COA, 2015. Commonwealth of Australia. Fact Sheet: Emissions Reduction Fund; available at: http://www.dpmc.gov.au/sites/default/files/publications/Fact%20Sheet%20-%20Emissions%20Reduction%20Fund.pdf Last accessed 6 Oct 2015.

Coleman, K., Jenkinson, D.S., Crocker, G.J., Grace, P.R., Klir, J., Körschens, M., Poulton, P.R. and Richter, D.D., 1997. Simulating trends in soil organic carbon in long-term experiments using RothC-26.3. *Geoderma*, **81**, 29-44.

Conant, R.T., Paustian, K. and Elliott, E.T., 2001. Grassland management and conversion into grassland: effects on soil carbon, *Ecological Applications*, **11**, 343-355.

Conant, R.T., Smith, G.R and Paustian, K., 2003. Spatial variability of soil carbon in forested and cultivated sites: implications for change detection. *Journal of Environmental Quality*, **32**, 278-286.

Condron, L.M. Turner, B.L. and Cade-Menun, B.J., 2005. Chemistry and dynamics of soil organic phosphorus. In: *Phosphorus: Agriculture and the Environment*, Agronomy Monograph no 46. American Society of Agronomy, Madison, WI, pp 87-121.

Conyers, M.K. Uren, N.C. and Helyar, K.R., 1995. Causes of changes in pH in acidic mineral soils. *Soil Biology and Biochemistry,* **27**, 1383-1392.

Conyers, M.K., Poile, G.J., Oates, A.A., Waters, D. and Chan, K.Y., 2011. Comparison of three carbon determination methods on naturally occurring substrates and the implication for the quantification of 'soil carbon'. *Soil Research*, **49**, 27-33.

Corr, K., 2003. Revegetation techniques; a guide for establishing native vegetation in Victoria. Greening Australia Victoria.

Costermans, L., 1983. *Native trees and shrubs of south-eastern Australia*. Reed New Holland, Sydney.

Cotching, W.E., Oliver, G., Downie, M., Corkrey, R. and Doyle, R.B., 2013. Land use and management influences on surface soil organic carbon in Tasmania, *Soil Research, Soil Research*, **51,** 615-630.

Cowie, A.L., Lonergan, V.E., Fazle Rabbi, S.M., Fornasier, F., Macdonald, C., Harden, S., Kawasaki, A. and Singh, B.K., 2013. Impact of carbon farming practices on soil carbon in northern New South Wales. *Soil Research*, **51**, 707-718.

Craswell, E.T. and Waring, S.A., 1972. Effect of grinding on the decomposition of soil organic matter – 1. The mineralization of organic nitrogen in relation to soil type, *Soil Biology and Biochemistry*, **4**, 427-433.

Craswell, E.T., King, H., and Read, Z., 2013. Carbon sequestration in agricultural soils: Separating the muck from the magic, *Proceedings Manuscript IAEA-CN-191-10*, pp 199-204.

Cunningham, G.M., 1987. Reclamation of scalded land in western New South Wales: A review. *Journal of Soil Conservation, New South Wales,* **43** (2): 52-61.

Cunningham, S.C., Metzeling, K.J., MacNally, R., Thomson, J.R. and Cavagnaro, T.R., 2012., Changes in soil carbon of pastures after afforestation with mixed species: sampling, heterogeneity and surrogates. *Agriculture, Ecosystems and Environment*, **158**, 58-65.

Cunningham, S.C. Cavagnaro, T.R., MacNally, R., Paul, K.I., Baker, P.J., Beringer, J., Thomson, J.R. and Thompson, R.M., 2015. Reforestation with native mixed-species plantings in a temperate continental climate effectively sequesters and stabilizes carbon within decades. *Global Change Biology*, **21**, 1552-1566.

Dalal, R.C., 1977. Soil organic phosphorus. *Advances in Agronomy,* **29**, 83-117.

Dalal, R.C. and Mayer, R.J., 1986. Long-term trends in fertility of soils under continuous cultivation and cereal cropping in southern Queensland. II Total organic Carbon and its rate of loss from the soil profile. *Australian Journal of Soil Research,* **24,** 281-292.

Davidson, E.A. and Ackerman, I.L., 1993. Changes in soil carbon inventories following cultivation of previously untilled soils. *Biogeochemistry*, **20**, 161-193.

Davies, G.M. and Gray, A. 2015. Don't let spurious accusations of pseudoreplication limit our ability to learn from natural experiments (and other messy kinds of ecological monitoring). *Ecology and Evolution*, **5**, 5295-5304.

Davy, M.C. and Koen, T.B., 2013. Variations in soil organic carbon for two soil types and six land uses in the Murray Catchment, New South Wales, Australia, *Soil Research, Soil Research*, **51,** 631-644.

Degryze, S., Six, J., Paustian, K., Morriss, S.J., Paul, E.A. and Merckx, R., 2004. Soil organic carbon pool changes following land-use conversions. *Global Change Biology*, **10**, 1120-1132.

Diamond, D., 2006. Quik Chem method 13-115-01-1-B – Determination of total Kjeldahl phosphorus in soils and plants by flow injection analysis, Lachat Instruments.

Diamond, D., 2007.Quik Chem method 13-107-06-2D – Determination of total Kjeldahl nitrogen in soils and plants by flow injection analysis. Lachat Instruments.

Ditchfield, R., 1996. The effects of waterponding on the reclamation of degraded scalds in the semi-arid rangelands of N.S.W. Honours thesis, the Australian National University.

Dixon, R.K., Brown, S., Houghton, R.A., Solomon, A.M., Trexler, M.C. and Wisniewski, J., 1994. Carbon pools and flux of global forest ecosystems. *Science*, **263**, 185-190.

Dixon, R.K., 1995. Agroforestry systems: sources or sinks of greenhouse gases? *Agroforestry Systems*, **31:** 99-116.

DOE, 2015a. Department of the Environment. *National Inventory Report 2013 Volume 1*. Commonwealth of Australia, 2015.

DOE, 2015b. Department of the Environment. *Quarterly Update of Australia's National Greenhouse Gase Inventory: June 2015*. Commonwealth of Australia, 2015.

Dong, A. Chesters, G. and Simsman, G.V., 1983. Soil dispersibility. *Soil Science*, **136**, 208-212.

DPMC, 2015. Department of the Prime Minister and Cabinet http://www.dpmc.gov.au/taskforces/unfccc las accessed 6 Oct 2015.

Duncan, D.H., Dorrough, J., White, M. and Moxham, C., 2008. Blowing in the wind? Nutrient enrichment of remnant woodlands in an agricultural landscape. *Landscape Ecology*, **23,** 107-119.

During, C., Weeda, W.C. and Dorofaeff, F.D., 1973. Some effects of cattle dung on soil properties, pasture production and nutrient uptake. *New Zealand Journal of Agricultural Research,* **16,** 431-438.

Ellert, B.H. and Bettany, J.R., 1995. Calculation of organic matter and nutrients stored in soils under contrasting management regimes, *Canadian Journal of Soil Science*, **75,** 529-538.

Ellert, B.H. Janzen, H.H. and McConkey, B.G., 2001. Measuring and comparing soil carbon storage. In: *Assessment Methods for Soil Carbon* (eds Lal R, Kimble JM, Follett BA, Stewart BA), pp. 131-146. Lewis Publishers, Boca Raton, FL.

Emerson, W.W., 1967. A classification of soil aggregates based on their coherence in water. *Australian Journal of Soil Research,* **5**, 47-57.

England, J.R., Paul, K.I., Cunningham, S.C., Madhavan, D.B., Baker, T.G., Read, Z.J., Wilson, B.R., Cavagnaro, T.R., Lewis, T., Perring, M.P., Herrmann, T., and Polglase, P.J. 2016. Previous land use and climate influence changes in soil organic carbon following reforestation of agricultural land with mixed species plantings, *Agriculture, Ecosystems and Environment*, **227**, 61-72.

Eriksen, J., 2009. Soil sulfur cycling in temperate agricultural systems. In: *Advances in Agronomy*, **102**, pp 55-89.

Eswaran, H. and Zi-Tong, G., 1991. 'Properties, genesis, classification and distribution of soils with gypsum, in *Occurrence, Characteristics and Genesis of Carbonate, Gypsum, And*

Silica Accumulations in Soils. SSSA Special Publication No. 26, Soil Science Society of America, Madison, WI, USA, pp 89-119.

Eyles, A., Coghlan, G., Hardie, M., Hovenden, M. and Bridle, K., 2015. Soil carbon sequestration in cool-temperate dryland pastures: mechanisms and management options. *Soil Research,* **53**, 349-365.

Farquharson, R. and Baldock, J., 2008. Concepts in modelling N_2O emissions from land use. *Plant Soil*, **309**, 147-167.

Federer, C.A., Turcotte, D.E. and Smith, C.T., 1993. The organic fraction – bulk density relationship and expression of nutrient content in forest soils. *Canadian Journal of Forest Research,* **23**, 1026-1032.

FFI CRC, Future Farm Industries Co-operative Research Centre. Saltland Geni. Available from: http://www.saltlandgenie.org.au (Accessed 26 June 2015).

Field, J. and Little, D., 2008. Regolith and biota, in *Regolith Science*, eds K.M. Scott and C.F. Pain, CSIRO Publishing, Victoria, pp 175-217.

Field, D.J. McKenzie, D.C. and Koppi, A.J., 1997. Development of an improved Vertosol stability test for SOILpak. *Australian Journal of Soil Research,* **35**, 843-852.

Finzi, A.C., van Breemen, N. and Canham, C.D., 1998. Canopy tree-soil interactions within temperate forests species effects on soil carbon and nitrogen. *Ecological Applications,* **8**, 440-446.

Foley, J.A., DeFries, R., Asner, G.P., Barford, C., Bonan, G., Carpenter, S.R., Chapin, F.S., Coe, M.T., Daily, G.C., Gibbs, H.K., Helkowski, J.H., Holloway, T., Howard, E.A., Kurcharik, C.j., Monfreda, C., Patz, J.A., Prentice, I.C., Ramankutty, N. and Snyder, P.K., 2005. Global consequences of land use. *Science*, **309**, 570-574.

Forbes, B. Duncan, D. Welch, A. Murphy, B. Brennan, N. Kew, G. Wooldridge, A. and Banks, R., N.D. *Detailed legend for the Walgett 1:250 000 Soil Landscape Map: Soil information package for the soil resources of the lower floodplain of the Macquarie River in the Central West Catchment of NSW*. Report for the Central West Catchment Management Authority.

Forrester, D.I., 2014. The spatial and temporal dynamics of species interactions in mixed-species forests: from pattern to process. *Forest Ecology and Management*, **312**, 282-292.

Forrester, D.I., Bauhus, J. and Khanna, P.K., 2004. Growth dynamics in a mixed-species plantation of *Eucalyptus globulus* and *Acacia mearnsii*. *Forest Ecology and Management*, **193**, 81-95.

Forrester, D.I., Bauhus, J. and Cowie, A.L., 2006. Mixed-species plantation of *Eucalyptus* with nitrogen-fixing trees: A review. *Forest Ecology and Management*, **233**, 211-230.

Franzluebbers, A.J., Hons, F.M. and Zuberer, D.A., 1994. Long-term changes in soil carbon and nitrogen pools in wheat management systems. *Soil Science Society of America Journal*, **58**, 1639-1645.

Franzluebbers, A.J., 2015. Farming strategies to fuel bioenergy demands and facilitate essential soil services. *Geoderma,* **259-260**, 251-258.

Freibauer, A. Rounsevell, M.D.A. Smith, P. and Verhagen, J., 2004. Carbon sequestration in the agricultural soils of Europe. *Geoderma,* **122**, 1-23.

Frossard, E., Buchmann, N., Bunemann, E.K., Kiba, D.I., Lompo, F., Oberson, A., Tamburini, F., and Traore, O.Y.A. 2016. Soil properties and not inputs control carbon:nitrogen:phosphorus ratios in cropped soils in the long term. *Soil*, **2**, 83-99.

Galvez, N. Barron, V and Torrent, J., 1999. Preparation and properties of hematite with structural phosphorus. *Clays and Clay Minerals,* **47**, 375-385.

Garnaut, R., 2011. The Garnaut Reviews 2011: Australia in the Global Response to Climate Change. Cambridge University Press, Melbourne, Australia.

Garthwaite, A.J., von Bothmer, R. and Colmer, T.D., 2005. Salt tolerance in wild *Hordeum* species is associated with restricted entry of Na+ and Cl- into the shoots. *Journal of Experimental Botany,* **56 (419)**, 2365-2378.

Gee, G.W. and Bauder, J.W., 1986. Particle size analysis. In: Klute, A. (ed.), Methods of Soil Analysis, Part 1. American Society of Agronomy, Madison, WI, pp. 383-411.

Geeves, G., Semple, B., Johnston, D., Johnston, A., Hughes, J., Koen, T. and Young, J., 2008. Improving the reliability of direct seeding for revegetation in the Central West of New south Wales. *Ecological Management and Restoration,* **9,** 68-71.

Gifford, R.M., 1994. The global carbon cycle: a viewpoint on the missing sink, *Australian Journal of Plant Physiology*, **21**, 1-15.

Gifford, R.M., 2000. Carbon contents of above-ground tissues of forest and woodland trees. National Carbon Accounting System, Technical Report no. 22, Australian Greenhouse Office, Canberra.

Gifford, R.M., 2003. Plant respiration in productivity models; conceptualisation, representation and issues for global terrestrial carbon-cycle research, *Functional Plant Biology*, **30**, 171-186.

Gifford, R.M., 2010. Carbon sequestration in Australian Grasslands: Technical and Policy Issues. In Grassland carbon sequestration: management, policy and economics. Proceedings of the workshop on the role of grassland carbon sequestration in the mitigation of climate change. Rome April 2009, 33-56.

Gifford RM., Cheney N.P., Noble J.C., Russell J.S., Wellington A.B. and Zammit C., 1992. Australian land use, primary production of vegetation and carbon pools in relation to atmospheric carbon dioxide concentration. In R.M. Gifford and M. Barson, eds. *Australia's renewable resources, sustainability and global change: International geosphere-biosphere programme Australia planning workshop.* Bureau of Rural Resources Proceedings No. 14, Oct 2-4, 1990, pp.151–187.

Gilligan L.B. and Scheibner, E., 1978. Lachlan Fold Belt in New South Wales. *Technophysics,* **48**, 217-265.

Giltrap, D.L., Berben, P., Palmada, T. and Saggar, S., 2014. Understanding and analysing spatial variability of nitrous oxide emissions from a grazed pasture. *Agriculture, Ecosystems and Environment,* **186**, 1-10.

Golchin, A., Baldock, J.A. and Oades, J.M., 1997 A model linking organic matter decomposition, chemistry and aggregate dynamics. In *Soil processes and the carbon cycle*. Eds R Lal, JM Kimble, RF Follett, BA Stewart, CRC Press, Boca Raton, FL, pp. 297-307.

Goldberg, S., Lebron, I., Seaman, J.C. and Suarez. D.L., 2012. Soil colloidal behaviour, in *Handbook of Soil Sciences, Properties and Processes*, Volume 2, eds P.M. Huang, Y.Li and M.E. Sumner, CRC Press, Boca Raton, FL, Chapter 15, pp 15-3.

Govers, G., Merckx, R., Van Oost, K. and van Wesemael, B., 2013. Managing Soil Organic Carbon for Global Benefits: A STAP Technical Report. Global Environment Facility, Washington DC.

Grace, P.R., Ladd, J.N., Robertson, G.P. and Gage, S.H., 2006. SOCRATES – A simple model for predicting long-term changes in soil organic carbon in terrestrial ecosystems, *Soil Biology and Biochemistry*, **38**, 1172–1176.

Gray, J.M., Bishop, T.F.A. and Wilson, B.R., 2015. Factors controlling soil organic carbon stocks with depth in eastern Australia. *Soil Science Society of America Journal*, **79**, 1741-1751.

Green, D.R., 1989, Rangeland restoration projects in western New South Wales. *Australian Rangeland Journal*, 11(2), 110-116.

Greene, R.S.B., Eggleton, R.A. and Rengasamy, P., 2002. Relationships between clay mineralogy and the hardsetting properties of soils in the Carnarvon horticultural district of Western Australia. *Applied Clay Science*, **20**, 211-223.

Groenendijk, F.M., Condron, L.M. and Rijkse, W.C., 2002. Effects of afforestation on organic carbon, nitrogen and sulfur concentrations in New Zealand hill country soils. *Geoderma,* **108**, 91-100.

Grüneberg, E., Schöning, I., Kalko, E.K.V. and Weisser, W.W., 2010. Regional organic carbon stock variability: A comparison between depth increments and soil horizons. *Geoderma,* **55**, 426-422.

Gulde, S., Chung, H., Amelung, W., Chang, C. and Six, J., 2008. Soil carbon saturation controls labile and stable carbon pool dynamics, *Soil Science Society of America Journal,* **72,** 605-612.

Gunn, R.H. and Richardson, D.P., 1979. The nature and possible origins of soluble salts in deeply weathered landscapes of eastern Australia. *Australian Journal of Soil Research*, **17**, 197-215.

Guo, L.B. and Gifford, R.M., 2002. Soil carbon stocks and land use change: a meta analysis, *Global Change Biology,* **8**, 345-360.

Guo, L.B. and Sims, R.E.H., 1999. Litter decomposition and nutrient release via litter decomposition in New Zealand eucalypt short rotation forests. *Agriculture, Ecosystems and Environment,* **75**, 133-140.

Guppy, C.N. and McLaughlin, M.J., 2009. Options for increasing the biological cycling of phosphorus in low input and organic agricultural systems. *Crop and Pasture Science,* **60**, 116-123.

Gupta, R.K., Bhumbla, D.K., and Abrol, I.P., 1984. Effect of sodicity, pH, organic matter, and calcium carbonate on the dispersion behaviour of soils. *Soil Science*, **137**, 245-251.

Gupta, R.K. and Abrol, I.P., 1990. Salt affected soils: their reclamation and management for crop production, *Advances in Soil Science*, **11,** 224-288.

Harmon, M.E., Ferrell, W.K. and Franklin, J.F., 1990. Effects on carbon storage of conversion of old-growth forest to young forests, *Science*, **247,** 699-702.

Harper, R.J., Okam, A.E.A., Stilwell, A.T., Tibbett, M., Dean, C., George, S.J., Sochacki, S.J., Mitchell, C.D., Mann, S.S., and Dods, K., 2012. Reforesting degraded agricultural landscapes with *Eucalypts*: Effects on carbon storage and soil fertility after 26 years. *Agriculture, Ecosystems and Environment*, **163**, 3-13.

Harris, D., Horwath, W.R., and van Kessel, C., 2001. Acid fumigation of soils to remove carbonates prior to total organic carbon or carbon-13 isotopic analysis, *Soil Science Society of America Journal,* **65**, 1853-1856.

Hartmann, D.L., Klein, A.M.G., Tank, M., Rusticucci, L.V., Alexander, S., Brönnimann, Y., Charabi, F.J., Dentener, E.J., Dlogokencky, D.R., Easterling, A., Kaplan, B.J., Soden, P.W., Thorne, M., Wild and Zhai, P.M., 2013. Observations: Atmosphere and Surface. In: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group 1 to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Stocker, T.F. D. Qin, G.K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley (eds)]*. Cambridge University Press, Cambridge, United Kingdom and New York, NY. USA.

Hassink, J., 1997. The capacity of soils to preserve organic C and N by their association with clay and silt particles. *Plant and Soil*, **191**, 77-87.

Hazelton, P. and Murphy, B., 2007. Interpreting Soil Test Results: What do all the number mean? CSIRO Publishing, Victoria.

Heanes, D.L., 1984. Determination of total organic-C in soils by an improved chromic acid digestion and spectrophotometric procedure. *Communications in Soil Science and Plant Analysis,* **15**, 1191-1213.

Heberger, M., 2011. Australia's millennium drought: impacts and responses, in: *The Worlds Water*, Volume 7. Ed: P.H. Gleik. Island Press / Centre for Resource Economics, pp 97-125.

Henskens, F.L., Battaglia, M., Cherry, M.L. and Beadle, C.L., 2001. Physiological basis of spacing effects on tree growth and form in *Eucalyptus globulus*. *Trees*, **15,** 365-377.

Heuperman, A., 1999. Hydraulic gradient reversal by trees in shallow water table areas and repercussions for the sustainability of tree-growing systems. *Agricultural Water Management,* **39**, 153-167.

Himes, F.L., 1998. Nitrogen, sulphur and phosphorus and the sequestering of carbon. In R.Lal, J.M. Kimble, R.F. Follett and B.A. Stewart (Eds), *Soil Processes and the Carbon Cycle* (pp. 315-319) Taylor and Francis, Boca Raton, FL.

Hobbie, S.E., 2000. Interactions between litter lignin and soil nitrogen availability during leaf litter decomposition in a Hawaiian montane forest. *Ecosystems*, **3**, 484-494.

Holford, I.C.R., 1997. Soil phosphorus: its measurement, and its uptake by plants. *Australian Journal of Soil Research,* **35**, 227-239.

Hoogmoed, M., Cunningham,S.C., Thomson, J.R., Baker, P.J., Beringer, J., Cavagnaro, T.R., 2012. Does afforestation of pastures increase sequestration of soil carbon in Mediterranean climates? *Agriculture, Ecosystems and Environment,* **159,** 176-183.

Hoogmoed, M., Cunningham, S.C., Baker, P.J., Beringer, J. and Cavagnaro, T.R., 2014a. Is there more soil carbon undr nitrogen-fixing trees than under non-nitrogen-fixing trees in mixed species restoration plantings? *Agriculture, Ecosystems and Environment,* **188**, 80-84.

Hopmans, P., and Elms, S.R., 2009. Changes in total carbon and nutrients in soil profiles and accumulation in biomass after a 30-year rotation of *Pinus radiata* on podzolized sands: Impacts of intensive harvesting on soil resources. *Forest Ecology and Management*, **258**, 2183-2193.

Houghton, R.A., 1999. The annual net flux of carbon to the atmosphere from changes in land use 1850 – 1990, *Tellus*, **51B**, 298-313.

Houghton, R.A., 2005. Aboveground forest biomass and the Global Carbon Balance. *Global Change Biology*, **11**, 945-958.

Houghton, R.A., 2010. How well do we know the flux of CO_2 from land-use change? *Tellus*, **62B**, 337-351.

Houghton, R.A., 2013. The contribution of land use and land use change to the carbon cycle, in *Land Use and the Carbon Cycle: Advances in Integrated Science, Management and Policy*, eds. D.G. Brown, D.T. Robinson, N.H.F. French and B.C. Reed, Cambridge University Press, pp. 52-76.

Houghton, R.A. Davidson, G.A. and Woodwell, G.M., 1998. Missing sinks, feedbacks and understanding the role of terrestrial ecosystems in the global carbon balance. *Global Biogeochemical Cycles*, **12**, 25-34.

Hoyle, F.C., D'Antuono, M.D., Overheu, T. and Murphy, D.V., 2013. Capacity for increasing soil organic carbon stocks in dryland agricultural systems. *Soil Research*, **51**, 657-667.

Huang, Y. Zou, Jianwen. Zheng, Xunhau. Wang, Yuesi. and Xu, Xingkai., 2004. Nitrous oxide emissions as influenced by amendment of plant residues with difference C:N ratios. *Soil Biology and Biochemistry,* **36**, 973-981.

Hughes, R.E. Moore, D.M. and Glass, H.D., 1994. Qualitative and quantitative analysis of clay minerals in soils, in *Quantitative Methods in Soil Mineralogy,* SSA Miscellaneous Publication, Soil Science Society of America Madison, WI: Chapter 11.

Hurlbert, S.H. 2009. The ancient black art and transdisciplinary extent of pseudoreplication, *Journal of Comparative Psychology*, **123**, 434-443.

Hurlbert, S.H. 2013. Pseudofactorialism, response structures and collective responsibility. *Austral Ecology*, **38,** 646-663.

IPCC., 2000. Land Use, Land-use Change and Forestry. Cambridge University Press, Cambridge, 377 pp.

IPCC., 2003 Good practice guidance for land use, land-use change and forestry, [J. Penman, M Gytarsky, T. Hiraishi, T. Krug, D. Kruger, R.Pipatti, L. Buendia, K. Miwa, T. Ngara, K. Tanabe, and F. Wagner (eds)], Institue for Global Environmental Strategies for the IPCC, Japan.

IPCC., 2006. IPCC guidelines for national greenhouse gas inventories, Volume 4. Agriculture, Forestry and other Land Use, Institute for Global Environmental Strategies, Hayama, Japan. [on-line] Available from: http://www.ipcc-nggip.iges.or.jp/public/2006gl/pdf/4_Volume4/V4_14_An2_SumEqua.pdf (Accessed 16 July 2014).

IPCC., 2007. Summary for Policymakers. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M.Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA

Isbell, R.F., 2002. *The Australian Soil Classification* (revised edition). CSIRO publishing, Melbourne.

Jackson, R.B. and Caldwell, M.M., 1993. Geostatistical patterns of soil heterogeneity around individual perennial plants. *Journal of Ecology*, **81**, 683-692.

Jackson, R.B., Jobbagy, E.G. Avissar, R., Baidya Roy, S., Barrett, D.J., Cook, C.W., Farley, K.A., le Maitre, D.C., McCarl, B.A. and Murray, B.C., 2005. Trading water for carbon with biological carbon sequestration. *Science*, **310**, 1944-1947.

Jagadamma, S and Lal, R., 2010. Integrating physical and chemical methods for isolating stable soil organic carbon, *Geoderma,* **158**, 322-330.

Janik, L.J., Merry, R.H. and Skjemstad, J.O., 1998. Can mid-infrared diffuse reflectance analysis replace soil extractions? *Australian Journal of Experimental Agriculture*, **38,** 681-696.

Janik, L.J., Skjemstad, J.O., Shepherd, K.D. and Spouncer, L.R., 2007. The prediction of soil carbon fractions using mid-infrared-partial least square analysis. *Australian Journal of Soil Research,* **45**, 73-81.

Janzen, H.H., 2004. Carbon cycling in earth systems – a soil science perspective, *Agriculture, Ecosystems and Environment*, **104,** 399 – 417.

Janzen, H.H., 2015. Beyond carbon sequestration: soil a conduit of solar energy, *European Journal of Soil Science,* **66**, 16-32.

Janzen, H.H., Campbell, C.A., Gregorich, E.G., and Ellert, B.H., 1998a. Soil carbon dynamics in Canadian Agroecosystems. P 57-80. In: Lal, R *et al* (eds.), *Soil processes and the carbon cycle.* CRC Press, Boca Raton, FL.

Janzen, H.H., Campbell, C.A., Izaurralde, R.C., Ellert, B.H., Juma, N., McGill, W.B. and Zentner., 1998b. Management effects on soil C storage on the Canadian prairies, *Soil and Tillage Research,* **47**, 181-105.

Jenkinson, D.S. and Rayner, J.H., 1977. The turnover of soil organic matter in some of the Rothamsted classical experiments. *Soil Science*, **123**, 298-305.

Jenkinson, D.S., 1990. The turnover of organic carbon and nitrogen in soil. *Philisophical Transactions of the Royal Society: Biological Sciences,* **329**, 361-368.

Jennings, J.N., 1972. The age of Canberra landforms. *Journal of the Geological society of Australia*, **19**, 371-378.

Jenny, H., 1941. Factors of soil formation: A system of quantitative pedology. McGraw-Hill, New York.

Jobbagy E.G. and Jackson, R.B., 2000. The vertical distribution of soil organic carbon and its relation to climate and vegetation. *Ecological Applications,* **10**, 423-436.

Johnson, D.W., 1992. Effects of forest management on soil carbon storage. *Water, Air and Soil Pollution,* **64**, 83-120.

Johnson, E.A. and Miyanishi, K. 2008. Testing the assumptions of chronosequences in succession. *Ecology Letters,* **11**, 419-431.

Johnson, M.G., 1995. The role of soil management in sequestering soil carbon, in *Advances in Soil Science: Soil Management and Greenhouse Effect*, eds R. Lal, J. Kimble, E Levine and B.A. Stewart, Lewis Publishers, pp 351-363.

Jonson, J.H. and Freudenberger, D., 2011. Restore and sequester: estimating biomass in native Australian woodland ecosystems for the carbon-funded restoration. *Australian Journal of Botany,* **59,** 640-653.

Jones, R.M., 1966. Scald reclamation studies in the Hay District, Part 1 – Natural reclamation of scalds. *Journal of Soil Conservation NSW,* **22**, 147-160.

Jones, R.M., 1967. Scald reclamation studies in the Hay district, N.S.W. Part III-reclamation by ponding banks. *Journal of Soil Conservation NSW,* **23,** 57-71.

Jones, R.M., 1969. Scald reclamation studies in the Hay district Part IV-scald soils: their properties and changes with reclamation. *Journal of Soil Conservation NSW,* **25**, 104-120.

Kahle, M., Kleber, M. and Jahn, R. 2002. Review of XRD-based quantitative analyses of clay minerals in soils: the suitability of mineral intensity factors, *Geoderma* **109**, 191-205.

Kaiser, K. and Guggenberger. 2003. Mineral surfaces and organic matter, *European Journal of Soil Science*, **54,** 219-236.

Kasel, S. and Bennett, L.T., 2007. Land-use history, forest conversion, and soil organic carbon in pine plantations and native forests of south eastern Australia. *Geoderma*, **137**, 401-413.

Kasel, S., Singh, S., Sanders, G.J. and Bennett, L.T., 2011. Species specific effects of native trees on soil organic carbon in biodiverse plantings across north-central Victoria, Australia. *Geoderma,* **161**, 95-106.

Kavanagh, R., Law, B., Lemckert, F., Stanton, M., Chidel, M., Brassil, T., Towerton, A. and Herring, M., 2005. *Biodiversity in eucalypt plantings established to reduce salinity*. A report for the RIRDC / Land & Water Australia FWPRDC / MDBC Joint Venture Agroforestry Program, viewed 3 December 2014, https://rirdc.infoservices.com.au/items/05-165

Keith, H., Jacobsen, K.L. and Raison, R.J., 1997. Effects of soil phosphorus availability, temperature and moisture on soil respiration in *Eucalyptus pauciflora* forest. *Plant and Soil*, **190**, 127-141.

Kirkby, C.A., Kirkegaard, J.A., Richardson, A.E., Wade, L.J., 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.

Kirkby, C.A. Richardson, A.E. Wade, L.J. Batten, G.D. Blanchard, C. and Kirkegaard, J.A., 2013. Carbon nutrient stoichiometry to increase soil carbon sequestration. *Soil Biology and Biochemistry, 60,* 77-86.

Kirkby, C.A. Richardson, A.E. Wade, L.J. Passioura, J.B. Batten, G.D. Blanchard, C. and Kirkegaard, J.A., 2014. Nutrient availability limits carbon sequestration in arable soils. *Soil Biology and Biochemistry,* **68**, 402-409.

Kirkby, C.A.; Richardson, A.E., Wade, L.J., Batten, G.D., Blanchard, C.L., McLaren, G., Zwart, A.B. and Kirkegarrd, J.A. 2016. Accurate measurement of resistant soil organic matter and its stoichiometry. *European Journal of Soil Science*, **67**, 695-705.

Kleber, M., Sollins, P. and Sutton, R. A conceptual model of organo-mineral interactions in soils: Self-assembly of organic molecular fragments into zonal structures on mineral surfaces. *Biogeochemistry,* **85**, 9-24.

Kleinman, P.J.A., Sharpley, A.N., Buda, A.R., McDowell, R.W. and Allen, A.L., 2011. Soil controls of phosphorus in runoff: management barriers and opportunities. *Canadian Journal of Soil Science*, **91**, 329-338.

Koch, C.B. Bentzon, M.D. Larsen, E.W. and Borggaard, O.K., 1992. Clay mineralogy of two ultisols from central Kalimantan, Indonesia. *Soil Science,* **154**(2), 158-167.

Kogel-Knabner, I., 2000. Analytical approaches for characterizing soil organic matter. *Organic Geochemistry*. **31**, 609-625.

Kovar, J.L. and Claassen, N., 2005. 'Soil-root interactions and phosphorus nutrition of plants', in: *Phosphorus: Agriculture and the Environment,* American Society of Agronomy. Agronomy Monograph no. 46. Madison WI.

Kovar, J.L. and Grant, C.A., 2011. 'Nutrient Cycling in Soils: Sulfur', in JL Hatfield, and TJ Sauer (eds), *Soil Management: building a stable base for Agriculture,* American Society of Agronomy and Soil Science Society of America, Madison, WI, USA.

Kumar, B.M., George, S.J., Jamaludheen, V. and Suresh, T.K., 1998. Comparison of biomass production, tree allometry and nutrient use efficiency of multipurpose trees grown in wood lot and silvopastoral experiments in Kerala, India. *Forest Ecology and Management*, **112**, 145-163.

Kyoto Protocol., 1998. Kyoto protocol to the United Nations Framework Convention on Climate Change. United Nations.

Laganiere J., Angers, D.A. and Pare, D., 2010. Carbon accumulation in agricultural soils after afforestation: a meta-analysis. *Global Change Biology,* **16**, 439-453.

Lal, R., 1998. 'Land use and soil management effects on soil organic matter dynamics on alfisols in Western Nigeria', in R Lal. (ed), *Soil processes and the carbon cycle.* . pp. 109-126. CRC Press, Boca Raton, FL.

Lal, R., 2001. World cropland soils as a source or sink for atmospheric carbon, in *Advances in Agronomy,* **71**, 145-192.

Lal, R., 2002. Soil carbon dynamics in cropland and rangeland. *Environmental Pollution*, **116**, 353-362.

Lal, R., 2003. Soil erosion and the global carbon budget. *Environment International,* **29,** 437-450.

Lal, R., 2004. Soil carbon sequestration to mitigate climate change. *Geoderma,* **123**, 1-22.

Lal, R., 2005. Forest soils and carbon sequestration. *Forest Ecology and Management*, **220**, 242-258.

Lal, R., 2008a. Sequestration of atmospheric CO2 in global carbon pools, *Energy, Environment and Science,* **1,** 86-100.

Lal, R., 2009. Challenges and opportunities in soil organic matter research. *European Journal of Soil Science, **60**, 158-169.*

Lal, R., 2010a. Soil erosion impact on agronomic productivity and environment quality. *Critical Reviews in Plant Sciences*, **17**, 319-464.

Lal, R., 2010b. Managing soils and ecosystems for mitigating anthropogenic carbon emissions and advancing global food security. *BioScience*, **60**, 708-721.

Lal, R. Follett, R.F. Stewart, B.A. and Kimble, J.M., 2007. Soil carbon sequestration to mitigate climate change and advance food security. *Soil Science,* **172**, 943-956.

Lal, R. and Kimble, J.M., 2001. Importance of soil bulk density and methods of its measurement in *Assessment Methods for Soil Carbon,* eds R. Lal, J.M. Kimble, R.F. Follett and B.A. Stewart, CRC Press Boca Raton, pp 31-44.

Lambers, H., Robinson, S.A. and Ribas-Carbo, M., 2005. Regulation of respiration in vivo. In: H. Lambers and M Ribas-Carbo (eds.), *Plant respiration: from cell to ecosystem,* Springer, Dordrecht, pp 1-15.

Landsberg, J.J. and Waring, R.H., 1997. A generalised model of forest productivity using simplified concepts of radiation-use efficiency, carbon balance and partitioning. *Forest Ecology and Management*, **95**, 209-228.

Laura, R.D., 1976. Effects of alkali salts on carbon and nitrogen mineralization of organic matter in soil. *Plant and Soil*, **44,** 587-596.

Lavelle, P., and Spain, A.V., 2005. *Soil Ecology*. Springer, Dordrecht, The Netherlands.

Lavelle, P. Blanchart, E. Martin, A. and Martin, S., 1993. A hierarchical model for decomposition in terrestrial ecosystems: Application to soils of the humid tropics. *Biotropica*, **25(2)**, 130-150.

Lehman, J. and Kleber, M. 2015. The contentious nature of soil organic matter, *Nature*, **528**, 60-68.

Lehmann, J., Kinyangi, J. and Solomon, D. 2007. Organic matter stabilization in soil microaggregates: implications from spatial heterogeneity of organic carbon contents and carbon forms. *Biogeochemistry,* **85,** 45-57.

Lemus, R. and Lal, R., 2005. Bioenergy Crops and Carbon Sequestration. *Critical Reviews in Plant Sciences*, **24**, 1-21.

Lenka, N.K., Choudhury, P.R., Sudhishri, S., Dass, A. Patnaik, U.S., 2012. Soil aggregation, carbon build up and root zone moisture in degraded sloping lands under selected agroforestry based rehabilitation systems in eastern India. *Agriculture, Ecosystems and Environment,* **150**, 54-62.

Leys, J.F., Heidenreich, S.K., Strong, C.L., McTainsh, G.H. and Quigley, S., 2011. PM10 concentrations and mass transport during "Red Dawn" Sydney 23 September 2009. *Aeolian Research*, **3**, 327-342.

Li, Y. Zhang, Q.W. Reicosky, D.C. Bai, L.Y. Lindstrom, M.J. and Li, L., 2006. Using 137 Cs and 210 Pbex for quantifying soil organic carbon redistribution affected by intensive tillage on steep slopes. *Soil and Tillage Research*, **86**, 176-184.

Li, Y. Geng, X.C. Yu, H.Q. and Wan, G.J., 2011. Effects of the composition of standard reference material on the accuracy of determinations of 210Pb and 137Cs in soils with gamma spectrometry. *Applied radiation and Isotopes,* **69,** 516-520.

Liu, Z.P., Shao, M.A. and Wang, Y.Q., 2012. Large-scale spatial variability and distribution of soil organic carbon across the entire Loess Plateau, China. *Soil Research*, **50,** 114-124.

Loveday, J. and Pyle, J.C., 1973. The Emmerson dispersion test and its relationship to hydraulic conductivity. CSIRO Australia. Division of Soils Technical Paper No. 15.

Lowson, C., 2008. Estimating Carbon in Direct Seeded Environmental Plantings, Honours Thesis, Australian National University.

Ludwig, J. and Tongway, D., 1997. A landscape approach to landscape ecology. Chapter 1 in Ludwig, J., Tongway, D., Freudenberger, D., Noble, J. and Hodgkinson, K. (eds). Landscape Ecology, Function and Management: Principles from Australia's Rangelands, CSIRO, Melbourne.

Ludwig, J., Tongway, D., Freudenberger, D., Noble, J. and Hodgkinson, K. (eds)., 1997. Landscape Ecology, Function and Management: Principles from Australia's Rangelands. CSIRO Publishing, Melbourne.

Ludwig, J. and Tongway, D., 2002. Clearing savannas for use as rangelands in Queensland: altered landscapes and water-erosion processes. *The Rangeland Journal*, **24**, 83-95.

Lugo, A.E. and Brown, S., 1993. Management of tropical soils as sinks or sources of atmospheric carbon. *Plant and Soil*, **149**, 27-41.

Luo, Z., Wang, E. and Sun, O.J., 2010. Soil carbon change and its responses to agricultural practices in Australian agro-ecosystems: A review and synthesis, *Geoderma,* **155**, 211-223.

Luo, Z., Wang, E., Sun, O.J., Smith, C.J. and Probers, M.E., 2011. Modelling long-term soil carbon dynamics and sequestration potential in semi-arid agro-ecosystems, *Agricultural and Forest Meteorology,* **151**, 1529-1544. *Nature Climate Change*, **3**, 552-557.

Lützow, M.V., Kogel-Knabner, I., Ekschmitt, K., Matzner, E. and Guggenberger, G., 2006. Stabilisation of organic matter in temperate soils: mechanisms and their relevance under different soil conditions – a review. *European Journal of Soil Science,* **57,** 426-445.

MAAF., n.d. Join the $4^0/_{00}$ Initiative: Soils for food security and climate. Ministry of Agriculture, Agrifood and Forestry. Available from http://newsroom.unfccc.int/media/408539/4-per-1000-initiative.pdf (Last accessed: 16 Jan 2016).

Mackey, B., Prentice, I.C., Steffen, W., House, J.I., Lindenmayer, D., Keith, H. and Berry, S., 2013. Untangling the confusion around land carbon science and climate change mitigation policy. *Nature Climate Change*, **3**, 552-557.

Madhaven, D.B., Baldock, J.A., Read, Z., Murphy, S., Cunningham, S., Perring, M.P., Herrmann, T., Lewis, T., Cavagnaro, T.R., England, J.R., Paul, K.I., Weston, C.J. and Baker, T.G., Rapid estimation of biologically significant soil organic carbon fractions in reforested soils using mid-infrared spectroscopy. (under review, 2016).

Manrique, L.A. and Jones, C.A., 1991. Bulk density of soils in relation to soil physical and chemical properties. *Soil Science Society of America Journal,* **55,** 476-481.

Manzoni, S. and Porporato, A., 2009. Soil carbon and nitrogen mineralization: Theory and models across scales. *Soil Biology and Biochemistry*, **41**, 1355-1379.

Marshall, J.K., 1974. Effects of stock on microenvironments in Australian rangelands, in *Plant morphogenesis as the basis for scientific management of range resources,* Proceedings of the Workshop of the United States-Australia Rangelands Panel, Berkely, Calif. March 29 - April 5, 1971; Miscellaneous Publication No 1271, Agricultural Research Service, Department of Agriculture.

Mason, J.A., Jacobs, P.M., Greene, R.S.B. and Nettleton, W.D., 2003. Sedimentary aggregates in the Peoria Loess of Nebraska, USA. *Catena,* **53,** 377-397.

May, R.M., 1977. Thresholds and breakpoints in ecosystems with a multiplicity of stable states. *Nature*, **269**, 471-477.

McCarty, G.W., Reeves III, J.B., Reeves, V.B., Follett, R.F. and Kimble, J.M., 2002. Mid-Infrared and Near-Infrared diffuse reflectance spectroscopy for soil carbon measurement, *Soils Science Society of America Journal*, **66**, 640-646.

McDonald, R.C. and Isbell, R.F., 2009. Soil Profile. In *Australian soil and land survey field handbook* (3rd edn). CSIRO Publishing: Melbourne.

McIntyre, D.S., and Loveday, J., 1974. Bulk Density, in Methods for Analysis of Irrigated Soils, Technical Communication No. 54., J. Loveday (ed), Bureau of Soils, South Australia, pp 38-42.

McKenzie, N. and Cresswell, H., 2002. Field sampling, in *Soil Physical Measurement and Interpretation for Land Evaluation,* eds N. McKenzie, K. Coughlan and H. Cresswell, CSIRO Publishing, Collingwood.

McKenzie, N., Jacquier, D., Isbell, R. and Brown, K., 2004. Australian soils and landscapes; an illustrated compendium, CSIRO, Collingwood.

McQueen, K.G. and Scott, K.M., 2011. 'Rock weathering and structure of the regolith', in K.M. Scott and C.F. Pain (eds), *Regolith Science*, CSIRO Publishing, Collingwood, Australia, pp. 105-126.

McTainsh, G., Chen, Y., McGowan, H., Leys, J. and Tews, K., 2005. The 23rd October 2002 dust storm in eastern Australia: characteristics and meteorological conditions. *Atmospheric Environment*, **39**, 1227 – 1236.

Mehler, K., Schoning, I. and Berli, M., 2014. The importance of rock fragment density for the calculation of soil bulk density and soil organic carbon stocks. *Soil Science Society of America Journal,* **78**, 1186-1191.

Metcalfe, L., Siversten, D.P., Tindall, D. and Ryan, K.M., 2003. Natural Vegetation of the New South Wales Wheat-belt (Cobar-Nyngan-Gilgandra, Nymagee-Narromine-Dubbo 1:250 000 vegetation sheets). *Cunninghamia* 8(2), 253-284.

Mewett, J., Paplinska, J., Kelle,y G., Lesslie, R., Pritchard, P and Atyeo, C. 2013. *Towards national reporting on agricultural land use change in Australia*, ABARES technical report 13.06, Canberra.

Miklos, M., Short, M.G., McBratney, A.B., and Minasny, B., 2010. Mapping and comparing the distribution of soil carbon under cropping and grazing management practices in Narrabri, north-west New South Wales. *Australian Journal of Soil Research,* **48**, 248-257.

Miller, H.G., 1981. Forest fertilization: some guiding concepts. *Forestry,* **54**, 157-167.

Mitchell, T.L., 1848. Journal of an Expedition into the Interior of Tropical Australia. In Search of a Route from Sydney to the Gulf of Carpentaria, eBooks @Adelaide2007,

Montagnini, F. and Nair, P.K.R., 2004. Carbon sequestration: an underexploited environmental benefit of agroforestry systems. *Agroforestry Systems*, **61**, 281-295.

Moore, G., 1998. *Soil Guide. A handbook for understanding and managing agricultural soils,* Department of Agriculture, Western Australia, Bulletin No 4343.

Moore, D.M., and Reynolds, R.C., 1997. X-ray diffraction and the identification and analysis of Clay Minerals. Oxford University Press, Oxford, New York.

Moorehead, D.L. and Reynolds, J.F., 1991. A general model of litter decomposition in the north Chihuahuan Desert. *Ecological Modelling*, **56**, 197-219.

Morris, S.J., Bohm S., Haile-Mariam, S. and Paul, E.A., 2007. Evaluation of carbon accrual in afforested agricultural soils. *Global Change Biology*, **13,** 1145-1156.

Mudge, P.L., Schipper, L.A., Baisden, W.T., Ghani, A. and Lewis, R.W., 2014. Changes in soil C, N and $\delta^{15}N$ along three forest-pasture chronosequences in New Zealand. *Soil Research*, **52**, 27-37.

Munns, R. and Tester, M., 2008. Mechanisms of salinity tolerance. *Annual Review of Plant Biology,* **59,** 651-681.

Murphy, B., 1995. Relationship between the Emerson aggregate test and exchangeable sodium percentage in some subsoils from central west New South Wales, in *Australian Sodic Soils; Distribution, properties and management*, eds R. Naidu, M.E. Sumner and P. Rengasamy, CSIRO Melbourne, pp 101-105.

Murphy, B.W. 2015. Impact of soil organic matter on soils properties – a review with emphasis on Australian soils. *Soil Research,* **53** 606-635.

Murphy, B., Rawson, A., Ravenscroft, L., Rankin, M., and Millard, R., 2003. Paired site sampling for soil carbon estimation – New South Wales, National Carbon Accounting System Technical Report No. 34, Australian Greenhouse Office.

Murty, D., Kirschbaum, M.U.F., McMurtrie, R.E. and McGilvray, H., 2002. Does conversion of forest to agricultural land change soil carbon and nitrogen? A review of the literature. *Global Change Biology,* **8,** 105-123.

Naidu, R. and Rengasamy, P., 1993. Ion interactions and constraints to plant nutrition in Australian sodic soils. *Australian Journal of Soil Research,* **31**, 801-819.

Nair, P.K.R., 2012. Carbon sequestration studies in agroforestry systems: a reality-check. *Agroforestry Systems,* **86**, 243-253.

Nair, P.K.R., Kumar, B.M. and Nair, V.D., 2009. Agroforestry as a strategy for carbon sequestration. *Journal of Plant Nutrition and Soil Science*, **172**, 10-23.

Nelson, D.W. and Sommers, L.E., 1996. 'Total carbon, organic carbon and organic matter', in *Methods of Soil Analysis, Part 3, Chemical Methods,* Soil Science Society of America Inc, Wisconsen, USA, pp. 961-1010.

Nelson, P.N., Baldock, J.A., Clarke, P., Oades, J.M. and Churchman, G.J., 1999. Dispersed clay and organic matter in soil: their nature and associations. *Australian Journal of Soil Research*, **37**, 289-315.

Nesbitt, H.W. and Markovics, G., 1997. Weathering of granodioritic crust, long-term storage of elements in weathering profiles, and petrogenesis of siliciclastic sediments. *Geochimica et Cosmochimica Acta,* **61**, 1653-1670.

Newey, A. 2005. Decomposition of plant litter and carbon turnover as a function of soil depth. PhD thesis, Australian National University.

Nguyen, M.L. and Goh, K.M., 1990. Accumulation of soil sulphur fractions in grazed pastures receiving long-term superphosphate applications. *New Zealand Journal of Agricultural Research,* **33**, 111-128.

NOAA, 2015. http://research.noaa.gov/News/NewsArchive/LatestNews/TabId/684/ArtMID/1768/ArticleID/11153/Greenhouse-gas-benchmark-reached-.aspx last accessed 24 Sep 2015.

NOAA National Centres for Environmental Information http://www.ncdc.noaa.gov/indicators/ last accessed 13 Oct 2015.

Noble, J.C. & Tongway, D.J., 1986. 'Pastoral settlement in arid and semi-arid rangelands', in J.S. Russell and R.F. Isbell (eds), *Australian Soils: the human impact,* pp. 217-242, St Lucia, Qld, University of Queensland Press in association with Australian Society of Soil Science.

Northcote, K.H., 1979. *A factual Key for the Recognition of Australia Soils,* 4th edn. Rellim Technical, Adelaide.

Noy-Meir, I., 1973. Desert ecosystems: environment and producers. *Annual Review of Ecological Systems*, **4,** 25-51.

Nuberg, I.K., 1998. Effect of shelter on temperate crops: a review to define research for Australian conditions, *Agroforestry Systems* **41,** 3-34.

Oades, J.M., 1984. Soil organic matter and structural stability: mechanisms and implications for management, *Plant and Soil,* **76**, 319-337.

Oades, J.M., 1988. The retention of organic matter in soils, *Biogeochemistry*, **5**, 35-70.

Oades, J.M., Waters, A.G., Vassallo, A.M., Wilson, M.A. and Jones, G.P., 1988. Influence of management on the composition of organic matter in red-brown earth as shown by ^{13}C nuclear magnetic resonance, *Australian Journal of Soil Research,* **26**, 289-299.

Oades, J.M., 1989. An introduction to organic matter in mineral soils, in *Minerals in Soil Environments* (second edition) Soil Science Society of America Book Series no. 1, pp. 89-159.

Oades, J.M., 1995. Krasnozems – organic matter, *Australian Journal of Soil Research,* **33,** 43-57.

Obrien, S.I., Jastrow, J.D., Grimley, D.A. and Gonzalez-Meler, M.A., 2015. Edaphic controls on soil organic carbon stocks in restored grasslands. *Geoderma*, **251-252,** 117-123.

O'Connell, A.M. and Menagé, P.M., 1982. Litter fall and nutrient cycling in karri (*Eucalyptus diversicolor* F.Muell.) forest in relation to stand age. *Australian Journal of Ecology,* **7** 49-62.

Odom, I.E., 1984. Smectite clay minerals: properties and uses. *Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences,* **311**, 391-409.

Olivier, J.G.J., Hanssens-Maaenhout, G. and Peters, J.A.H.W., 2012. Trends in global CO_2 emissions 2012 Report, PBL Netherlands Environmental Assessment Agency, The Hague/Bilthoven, 2012. Available from: http://www.pbl.nl/sites/default/files/cms/publicaties/PBL_2012_Trends_in_global_CO_2_emissions_500114022.pdf Last accessed 24 Sep 2015.

Olsen, S.R. Cole, C.V. Watanabe, F.S. and Dean, L.A., 1954. *Estimation of available phosphorus in soils by extraction with sodium bicarbonate.* Department of Agriculture Circular 939, USDA, Washington DC.

Olson, K.R., Al-Kaisi, M.M., Lal, R. and Lowery, B., 2014. Experimental consideration, treatments and methods in determining soil organic carbon sequestration rates. *Soil Science Society of America Journal,* **78**, 348-360.

Omotoso, O., McCarty, D.K., Hillier, S. and Kleeberg, R., 2006. Some successful approaches to quantitative mineral analysis as revealed by the 3rd Reynolds Cup contest. *Clays and Clay Minerals*, **4**, 748-760.

Orgill, S.E., Condon, J.R., Conyers, M.K., Greene, R.S.B., Morris, S.G. and Murphy, B.W., 2013. Sensitivity of soil carbon to management and environmental factors within Australian perennial pasture systems. *Geoderma,* **214-215***,* 70-79.

Orgill, S.E., Condon, J.R., Kirkby, C.A., Orchard, B.A., Conyers, M.K., Greene, R.S.B. and Murphy, B.W. 2017. Soil with high organic carbon concentration continues to sequester carbon with increasing carbon inputs. *Geoderma*, **285**, 151-163.

Oster, J.D., Shainberg, I. and Abrol, I.P., 1999. 'Reclamation of salt-affected soils', in *Agricultural Drainage*, Agronomy Monograph No. 38, American Society of Agronomy, Crop Science Society of America, Soil Science Society of America, Madison, WI, USA, pp 659-691.

Page, K.L. Dalal, R.C. Pringle, M.J. Bell, M. Dang, Y.P. Radford, B. and Bailey, K., 2013. Organic carbon stocks in cropping soils of Queensland, Australia, as affected by tillage management, climate, and soil characteristics. *Soil Research,* **51,** 596-607.

Pan, Y., Birdsey, R.A., Fan, J., Houghton, R., Kauppi, P.E., Kurz, W.A., Phillips, O.L., Shvidenko, A., Lewis, S.L., Canadell, J.G., Ciais, P., Jackson, R.B., Pacala, S.W., McGuire, A.D., Shilong, Pl, Rautianen, A., Sitch, S. and Hayes, D., 2011. A large and persistent Carbon sink in the world's forests. *Science*, **333**, 988-933.

Parton, W.J. Stewart, J.W.B. and Cole, C.V., 1988. Dynamics of C, N, P and S in grassland soils: a model. *Biogeochemistry*, **5**, 109-131.

Paul, K.I., Black, A.S. and Conyers, M.K., 1999. Influence of moist-dry cycles on pH changes in surface soils. *Australian Journal of Soil Research,* **37**, 1057-1072.

Paul, K.I., Polglase, P.J., Nyakuengama, J.G. and Khanna, P.K., 2002a. Change in soil carbon following afforestation. *Forest Ecology and Management*, **168**, 241-257.

Paul, K. Polglase, P., Coops, N., O'Connell, T., Grove, T., Mendham, D., Carlyle, C., May, B., Smethurst, P. and Baillie, C., 2002b. *Modelling change in soil carbon following afforestation or reforestation: preliminary simulations using GRC3 and sensitivity analyses.* National Carbon Accounting System Technical Report No. 29, Australian Greenhouse Office, Canberra.

Paul, K.I., Polglase, P.J. and Richards, G.P., 2003. Predicted change in soil carbon following afforestation or reforestation, and analysis of controlling factors by linking a C accounting model (CAMFor) to models of forest growth (3PG), litter decomposition (GENDEC) and soil C turnover (RothC). *Forest Ecology and Management*, **177**, 485-501.

Paul, K.I. and Polglase, P.J., 2004a. Calibration of the RothC model to turnover of soil carbon under eucalypts and pines. *Australian Journal of Soil Research*, **42**, 883-895.

Paul, K.I. and Polglase, P.J., 2004b. Prediction of decomposition of litter under eucalypts and pines using the FullCAM model. *Forest Ecology and Management*, **191**, 73-92.

Paul, K.I., Polglase, P., Bauhus, J., Raison, J. and Khanna, P., 2004. *Modelling change in litter and soil carbon following afforestation or reforestation: calibration of the FullCAM 'Beta' model*. National Carbon Accounting System Technical Report No. 40, Australian Greenhouse Office, Canberra.

Paul, K.I., Roxburgh, S.H., England, J.R., Ritson, P., Hobbs, T., Brooksbank, K., Raison, J.R., Larmour, J.S., Murphy, S., Norris, J., Neumann, C., Lewis, T., Jonson, J., Carter, J.L. McArthur, G., Barton, C. and Rose, B., 2013. Development and testing of allometric equations for estimating above-ground biomass of mixed-species environmental plantings. *Forest Ecology and Management*, **310**, 483-494.

Paul, K., Roxburgh, S., Raison, J., Larmour, J., England, J., Murphy, S., Norris, P., Ritson, P., Brooksbank, K., Hobbs, T., Neumann, C., Lewis, T., Read, Z., Clifford, D., Kmoch, L., Rooney, M., Freudenberger, D., Jonson, J., Peck, A., Giles, R., Bartle, J., McAurthur, G., Wildy, D., Lindsay, A., Preece, N., Cunningham, S., Power, Tl, Carter, J., Bennett, R., Mendham, D., Sudmeyer, R., Rose, B., Butler, D., Cohen, L., Fairman, T., Law, R., Finn, B., Brammar, M., Minchin, G., van Oosterzee and, P., and Lothian, A., 2014. *Improved estimation of biomass accumulation by environmental plantings and mallee plantings using FullCAM*, Report for Department of Climate Change and Energy

Efficiency dated 31st October 2013 (Up-dated on 26th May 2014). CSIRO Sustainable Agriculture Flagship. Canberra, Australia.

Paul, K.I., Roxburgh, S.H., England, J.R., de Ligt, R., Larmour, J.S., Brooksbank, K., Murphy, S., Ritson, P., Hobbs, T., Lewis, T., Preece, N.D., Cunningham, S.C., Read, Z., Clifford, D., and Raison, J., 2015. Improved models for estimating temporal changes in carbon sequestration in above-ground biomass of mixed species environmental plantings. *Forest Ecology and Management,* **338,** 208-218.

Paul, K.I., Cunningham, S.C., England, J.R., Roxburgh, S.H., Preece, N.D., Lewis, T., Brooksbank, K., Crawford, D.F. and Polglase, P.J., 2016. Managing reforestation to sequester carbon, increase biodiversity potential and minimize loss of agricultural land. *Land Use Policy,* **51**, 135-149.

Paustian, K., Andren, O., Janzen, H.H., Lal, R., Smith, P., Tian, G., Tiessen, H., Van Noordwijk, M. and Woomer, P.L., 1997. Agricultural soils as a sink to mitigate CO_2 emissions, *Soil Use and Management*, **13**, 230-244.

Peech, M., 1965. 'Hydrogen-Ion activity', in A.G. Norman (ed) *Methods of Soil Analysis. Part 2. Chemical and Microbiological Properties*, American Society of Agronomy: pp. 914-926.

Peichl, M., Leava, N.A. and Kiely, G., 2012. Above- and belowground ecosystem biomass, carbon and nitrogen allocation in recently afforested grassland and adjacent intensively managed grassland. *Plant and Soil*, **350**, 281-296.

Poeplau, C. and Don, A., 2013. Sensitivity of soil organic carbon stocks and fractions to different land-use changes across Europe. *Geoderma*, **192**, 189-201.

Polglase, P.J., Paul, K.I., Khanna, P.K., Nyakuengama, G., Gwinyai, J., O'Connell, A.M., Grove, T.S. and Battaglia, M., 2000. Change in soil carbon following afforestation or reforestation. National Carbon Accounting System Technical Report No. 20, Australian Greenhouse Office, Canberra.

Post, W.M., Emanuel, W.R., Zinke, P.J. and Stangenberger, A.G., 1982. Soil carbon pools and world life zones, *Nature,* **298,** 156-159.

Post, W.M. and Kwon, K.C., 2000. Soil carbon sequestration and land-use change: processes and potential, *Global Change Biology*, **6**, 317-327.

Post, W.M., Izauralde, R.C., Mann, L.K. and Bliss, N., 2001. Monitoring and verifying changes of organic carbon in soil. *Climatic Change*, **51**, 73-99.

Powlson, D.S., Whitmore, A.P. and Goulding, K.W.T., 2011. Soil carbon sequestration to mitigate climate change: a critical re-examination to identify the true and false, *European Journal of Soil Science*, **62**, 42-55.

Pressland, A.J., 1982. Litter production and decomposition from an overstorey of *Eucalyptus* spp. on two catchments in the New England region of NSW. *Australian Journal of Ecology*, **7**, 171-180.

Prior, L.D., Paul, K.I., Davidson, N.J., Hovenden, M.N., Nichols, S.C. and Bowman, D.J.M.S., 2015. Evaluating carbon storage in restoration plantings in the Tasmanian Midlands, a highly modified agricultural landscape. *The Rangeland Journal,* **37**, 477-488.

Qadir, M. and Schubert, S., 2002. Degradation processes and nutrient constraints in sodic soils. *Land Degradation and Development*, **13**, 275-294.

Quilty, J.R. and Cattle, S.R., 2011. Use and understanding of organic amendments in Australian agriculture: a review. *Soil Research*, **49**, 1-26.

Raupach, M. and Fraser, P., 2011. Climate and greenhouse gases, in *Climate Change: science and solutions for Australia*, eds. H. Cleugh., M. Stafford Smith., M. Battaglia, and P. Graham. CSIRO publishing: Chapter 2.

Rayment, G.E. and Lyons, D.J., 2011. Soil Chemical Methods – Australasia, CSIRO Publishing, Collingwood, Victoria.

Reeves III, J.B., 2010. Near- versus mid-infrared diffuse reflectance spectroscopy for soil analysis emphasizing carbon and laboratory versus on-site analysis: Where are we and what needs to be done? *Geoderma,* **158**, 3-14.

Read, Z.J., 2008. Direct seeded native vegetation belts on the Southern Tablelands, NSW: Their effects on soils, Honours Thesis, Australian National University.

Read, Z.J., King, H.P., Tongway, D.J., Ogilvy, S., Greene, R.S.B. and Hand, G. 2016. Landscape function analysis to assess soil processes on farms following ecological restoration and changes to grazing management. *European Journal of Soil Science*, **67**, 409-420.

Reid-Soukup, D.A. and Ulery, A.L., 2002. Smectites, in, *Soil Mineralogy with Environmental Applications,* Soil Science Society of America Book Series no. 7, Madison, WI, pp. 467-499.

Rengasamy, P., 1983. Clay dispersion in relation to changes in the electrolyte composition of dialysed red-grown earths. *Journal of Soil Science,* **34**, 723-732.

Rengasamy, P., Greene, R.S.B., Ford, G.W. and Mehanni, A.H., 1984. Identification of dispersive behaviour and the management of red-brown earths. *Australian Journal of Soil Research,* **22**, 413-431.

Rengasamy, P., Greene, R.S.B. and Ford, G.W., 1986. Influence of magnesium on aggregate stability in sodic red-brown earths. *Australian Journal of Soil Research,* **24**, 229-237.

Rengasamy, P. and Olsson, K.A., 1991. Sodicity and soil structure. *Australian Journal of Soil Research,* **29**, 935-952.

Resh, S.C., Binkley, D. and Parrotta, J.A., 2002. Greater soil carbon sequestration under Nitrogen-fixing trees compared with *eucalyptus* species. *Ecosystems*, **5**, 217-231.

Rhodes, D.W., 1987. Vegetation response and productivity improvements under waterponding. *Journal of Soil Conservation NSW,* **43** (2), 62-67.

Rhoades, C. and Binkley, D., 1996. Factors influencing decline in soil pH in Hawaiian *Eucalyptus* and *Albizia* plantations. *Forest Ecology and Management* **80**, 47-56.

Richards, G.P., 2001. The FullCAM carbon accounting model: development, calibration and implementation for the national carbon accounting system. National Carbon Accounting System Technical Report No. 28. Australian Greenhouse Office, Canberra.

Richards, G.P. and Evans, D.M.W., 2004. Development of a carbon accounting model (FullCAM Version 1.0) for the Australian continent. *Australian Forestry*, **67**, 277-283.

Ringrose-Voase, A.J., Rhodes, D.W. and Hall, G.F., 1989a, Reclamation of a scalded, red duplex soil by waterponding, *Australian Journal of Soil Research.* **27,** 779-795.

Ringrose-Voase, A.J., Thompson, R., Beatty, H.J., Chapman, G. and Rhodes, D.W., 1989b. Soil properties of scalded, red duplex soils before and after reclamation by waterponding. CSIRO Aust. Div. Soils Tech. Mem. 4/89.

Ringrose-Voase, A.J. and McClure, S.G., 1994. Fabric changes during reclamation of a scalded, red duplex soil by waterponding. In: A.J. Ringrose-Voase and G.S. Humphreys (Editors), Soil Micromorphology: Studies in Management and Genesis. Prox. IX Int. Working Meeting on Soil Micromorphology, Townsville, Australia, July 1992. Developments in Soil Science 22, Elsevier, Amsterdam, pp. 777-786.

Ritter, E., Vesterdal. L. and Gundersen, P., 2003. Changes in soil properties after afforestation of former intensively managed soils with oak and Norway spruce. *Plant and Soil*, **249**, 319-330.

Robertson, R., Crawford, D., Partington, D., Oliver, I., Rees, D., Aumann, C., Armstrong, R., Perris, R., Davey, M., Moodie, M. and Baldock, J., 2016. Soil organic carbon in cropping and pasture systems of Victoria, Australia. *Soil Research,* in press. http://dx.doi.org/10.1071/SR15008.

Rosell, R.A., Gasparoni, J.C. and Galantini, J.A., 2001. Soil organic matter evaluation in *Advances in Soil Science: Assessment Methods for Soil Carbon*, eds R. Lal, J.M. Kimble, R.F. Follett and B.A. Stewart, Lewis Publishers, pp 311-322.

Ross, G.J., 1978. Relationships of specific surface area and clay content to shrink-swell potential of soils having different clay mineralogical compositions. *Canadian Journal of Soil Science*. **58**, 159-166.

Rovira, A.D. and Greacen, E.L., 1957. The effect of aggregate disruption on the activity of microorganisms in the soil. *Australian Journal of Agricultural Research,* **8**, 659-673.

Roxburgh, S.H., Berry, S.L., Buckley, T.N., Barnes, B. and Roderick, M.L., 2005. What is NPP? Inconsistent accounting of respiratory fluxes in the definition of net primary production, *Functional Ecology*, **19**, 378-382.

Rupach, M. and Fraser, P., 2011. Climate and Greenhouse Gases. In *Climate Change Science and Solutions for Australia*, eds H. Cleugh, M. Stafford Smith, M. Battaglia and P. Graham. CSIRO publishing, pp 15-34.

Rutherford, P.M., McGill, W.B. and Arocena, J.M., 2008. Total Nitrogen, in *Soil Sampling and Methods of Analysis,* Eds M.R. Carter and E.G. Gregorich, Canadian Society of Soil Science, CRC Press, Boca Raton FL, pp 239-250.

Sanderman, J., Baldock, J., Hawke, B., Macdonald, L., Massis-Puccini, A., and Szarvas, S., 2011, National soil carbon research programme: field and laboratory methodologies, CSIRO Land and Water, Urrbrae SA

Sanderman, J., Farquharson, R., and Baldock, J., 2010. Soil carbon sequestration potential: A review for Australian agriculture, in: A report prepared for Department of Climate Change and Energy Efficiency, CSIRO.

Sauer, D., 2015. Pedological concepts to be considered in soil chronosequence studies. *Soil Research*, **53**, 577-591.

Scheffer, M., Carpenter, S., Foley, J.A., Folke, C. and Walker, B., 2001. Catastrophic shifts in ecosystems. *Nature,* **413**, 591-596.

Scherer, H.W., 2009. Sulfur in soils. *Journal of Plant Nutrition and Soil Science.* **172**, 326-335.

Schimel, D.S., Braswell, B.H., Holland, E.A., McKeown, R., Ojima, D.S., Painter, T.H., Parton, W.J. and Townsend, A.R., 1994. Climatic, edaphic and biotic controls over storage and turnover of carbon in soils, *Global Biogeochemical Cycles,* **8** (3), 279-293.

Schimel, D.S., House, J.I., Hibbard, K.A., Bousquet, P., Ciais, P., Peylin, P., Braswell, B.H., Apps, M.J., Baker, D., Bondeau, A., Canadell, J., Churkina, G., Cramer, W., Denning, A.S., Field, C.B., Friedlinstein, Pl, Goodale, C., Heimann, M., Houghton, R.A., Melillo, J.M., Moore 111, B., Murdiyarso, D., Noble, I., Pacala, S.W., Prentice, I.C., Raupach, M.R., Rayner, P.J., Scholes, R.J., Steffen, W.L. and Wirth, C., 2000. Recent patterns and mechanisms of carbon exchange by terrestrial ecosystems, *Nature,* **414**, 169-172.

Schlesinger, W.H., Winkler, J.P. and Megonigal, J.P., 2000. Soils and the Global Carbon Cycle. In 'The Carbon Cycle'. (Eds, T.M.L Wigley and D.S. Schimel.) pp 93-101. (Cambridge University Press, USA).

Schneeman, B.S., 2008. Change in structure and composition of direct deeded revegetation in the Southern Tablelands, NSW, Honours Thesis, Australian National University.

Schulze, D.G., 2005. Clay minerals, in *Encyclopaedia of Soils in the Environment*, ed. D. Hillel, Volume 1, Elsevier, Academic Press, Boston, pp. 246-254.

Schuman, G.E. Janzen, H.H. and Herrick, J.E., 2002. Soil carbon dynamics and potential carbon sequestration by rangelands, *Environmental Pollution.* **116**, 391-396.

Semple, B., Rankin, M., Koen, T. and Geeves, G., 2010. A note on tree deaths during the current (2001-?) drought in south-eastern Australia. *Australian Geographer,* **41,** 391-401.

Shao, Y. Raupach, M.R. and Leys, J.F., 1996. A model for predicting aeolian sand drift and dust entrainment on scales from paddock to region. *Australian Journal of Soil Research,* **34**, 309-342.

Shiga, Y., Greene, R.S.B., Scott, .M. and Stelcer, E., 2011. Recognising terrestrially-derived salt (NaCl) in SE Australian dust. *Aeolian Research*, **2**, 215-220.

Silver, W.L., Ostertag, R. and Lugo, A.E., 2000. The potential for carbon sequestration through reforestation of abandoned tropical agricultural and pasture lands. *Restoration Ecology,* **8**, 394-407.

Six, J., Callewaert, P., Lenders, S., DeGryze, S., Morris, S.J., Gregorich, E.G., Paul, E.A. and Paustian, K., 2002. Measuring and understanding carbon storage in afforested soils by physical fractionation. *Soil Science Society of America Journal,* **66,** 1981-1987.

Six, J. Conant, R.T. Paul, E.A. and Paustian, K., 2002. Stabilisation mechanisms of soil organic matter: Implications for C-saturation of soils. *Plant and Soil,* **241,** 155-176.

Six, J., Elliott, E.T., Paustian, K. and Doran, J.W., 1998a. Aggregation and soil organic matter accumulation in cultivated and native grassland soils. *Soil Science Society of America Journal,* **62,** 1367-1377.

Six, J., Paustian, K., Elliott, E.T., and Combrink, C., 1998b. Soil structure and organic matter: 1. Distribution of Aggregate-size classes and aggregate-associated carbon, *Soil Science Society of America Journal,* **64**, 681-689.

Six, J., Elliott, E.T. and Paustian, K., 1999. Aggregate and soil organic matter dynamics under conventional and no-tillage systems, *Soil Science Society of America Journal,* **63**, 1350-1358.

Six, J. Elliott, E.T. and Paustian, K., 2000. Soil structure and soil organic matter 11. A normalised stability index and the effect of mineralogy. *Soil Science Society of America Journal,* **64**, 1042-1049.

Sims, J.T. and Pierzynski, G.M., 2005. Chemistry of phosphorus in soils. In *Chemical Processes in Soils,* Soil Science Society of America Book Series, no. 8. pp 151 – 192.

Skjemstad, J.O., Janik. L.J. and Taylor, J.A., 1998. Non-living soil organic matter: what do we know about it? *Australian Journal of Experimental Agriculture, **38:** 667-680.*

Skjemstad, J.O., Spouncer, L.R., Cowie, B. and Swift, R.S., 2004. Calibration of the Rothamsted organic carbon turnover model (RothC ver. 26.3), using measurable soil organic carbon pools. *Australian Journal of Soil Research,* **42**, 79-88.

Slattery, W.J. and Surapaneni, A., 2002. Effect of soil management practices on the sequestration of carbon in duplex soils of south eastern Australia. In: Kimble J.M., Lal,

R., Follett, R.F. (eds) Agricultural practices and policies for carbon sequestration in soils. CRC, New York, pp 107-117.

Slavich, P.G. and Petterson, G.H., 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.

Smith, P., 2004. Carbon sequestration in croplands: the potential in Europe and the global context. *European Journal of Agronomy,* **20,** 229-236.

Snowdon, P., Raison, J., Keith, H., Ritson, P., Grierson, P., Adams, M., Montagu, K., Bi, Hui-quan., Burrows, W., and Eamus, D., 2002. *Protocol for sampling tree and stand biomass*, National Carbon Accounting System Technical Report No. 31, Australian Greenhouse Office, Canberra, 66 pp.

SOE., 2011. State of the Environment 2011 Committee. Australia State of the Environment 2011. Independent report to the Australian Government Minister for Sustainability, Environment, Water, Population and Communities. Canberra: DSEWPaC, 2011.

Sousa, F.P., Ferreira, T.O., Mendonca, E.S., Romero, R.E. and Oliveira, J.G.B., 2012. Carbon and nitrogen in degraded Brazilian semi-arid soils undergoing desertification. *Agriculture, Ecosystems and Environment*, **148**, 11-21.

Speight, J.G., 2009. Landform, in *Australian Soil and Land Survey Field Handbook* (3rd edn). National Committee on Soil and Terrain, CSIRO Publishing, Melbourne.

Steffens, M. Kolbl, A. Totsche, K.U. Kögel-Knabner, I., 2008. Grazing effects on soil chemical and physical properties in a semiarid steppe of Inner Mongolia (P.R. China). *Geoderma,* **143,** 63-72.

Stevenson, F.J., 1959. Carbon-nitrogen relationships in soil, *Soil Science*, **88**, 201-288.

Stockman, U., Adams, M.A., Crawford, J.W., Field, D.J., Henakaarchchi, N., Jenkins, M., Minasny, B., McBratne, A.B., de Remy De Courcelles, V., Singh, K., Wheeler, I., Abbott, L., Angers, D.A., Baldock, J., Bird, M., Brookes, P.C., Chenu, C., Jastrow, J.D., Lal, R., Lehmann, J., O'Donnell, A.G., Parton, W.J., Whitehead, D., and Zimmermann, M., 2013. The knowns, known unknowns and unknowns of sequestration of soil organic carbon. *Agriculture, Ecosystems and Environment*,**164**, 80-99.

Streeton, N.A., Greene, R.S.B., Marchiori, K., Tongway, D.J. and Carnegie, M.D., 2013. Rehabilitation of an incised ephemeral stream in central New South Wales, Australia: identification of incision causes, rehabilitation techniques and channel response, *The Rangeland Journal,* **35**, 71-83.

Sumner, M.E., 1993. Sodic soils: New perspectives. *Australian Journal of Soil Research,* **31,** 683-750.

Sun, D. and Dickinson, G.R., 1995. Salinity effects on tree growth, root distribution and transpiration of *Casuarina cunninghamiana* and *Eucalyptus camaldulensis* planted on a saline site in tropical north Australia. *Forest Ecology and Management* **77,** 127-138.

Swift, R.S., 2001. Sequestration of carbon by soil, *Soil Science,* **166** (11), 858-871.

Thiry, M., 2000. Palaeoclimatic interpretation of clay minerals in marine deposits: an outlook from the continental origin. *Earth-Science Reviews*, **49**, 201-221.

Thomas, G.W., 1996. 'Soil pH and soil Acidity' in DL Sparks (ed), *Methods of Soil Analysis Part 3 – Chemical Methods*, Soil Science Society of America Book Series 5, Madison, WI, USA.

Thomas, O.D., Scott, M.M., Warren, A.Y.E., Sherwin, L., Pogson, D.J. and Vassallo, J.J., 2013. *Gunning 1:100 000 Geological Sheet, 8728.* Geological Survey of New South Wales, Maitland.

Thompson, R., 2008. Waterponding: reclamation technique for scalded duplex soils in western New South Wales rangelands. *Ecological Management and Restoration*, 9(3), 170-181

Thrall, P.H., Laine, A.L., Broadhurst, L.M., Bagnall, D.J. and Brockwell, J., 2011. Symbiotic effectiveness of rhizobial mutualists varies in interactions with native Australian Legume Genera. PLoS ONE **6**(8): e23545. Doi:10.1371/journal.pone.0023545.

Tiessen, H., Stewart, J.W.B. and Hunt, H.W., 1984. Concepts of soil organic matter transformations in relation to organo-mineral particle size fractions, *Plant and Soil,* **76:** 287-295.

Tisdall, J.M. and Oades, J.M., 1982. Organic matter and water-stable aggregates in soils. *Journal of Soil Science,* **33,** 141-163.

Tiessen, H. and Moir, J.O., 2008. 'Characterisation of available P by sequential extraction', in M.R. Carter and E.G. Gregorich (eds) *Soil Sampling and Methods of Analysis,* Canadian Society of Soil Science, CRC Press, Boca Raton FL, pp 293-306.

Tisdall, J.M. and Oades, J.M., 1982. Organic matter and water-stable aggregates in soils. *Journal of Soil Science,* **33,** 141-163.

Tongway, D. and Hindley, N., 2004a. Landscape function analysis: a system for monitoring rangeland function. *African Journal of Range and Forage Science,* **21,** 109-113.

Tongway, D. and Hindley, N., 2004b. Landscape function analysis: Procedures for monitoring and assessing landscapes with special reference to mine sites and rangelands. Version 3.1. CD. Canberra, ACT: CSIRO Sustainable Ecosystems.

Tongway, D. and Ludwig, J., 2007. Landscape function as a target for restoring natural capital in semiarid Australia. In: *Restoring Natural Capital: Science, Business and Practice*, pp 76-84. (eds J. Aronson, S.J. Milton & J.N. Blignaut). Island Press, Washington, D.C.

Torn, M.S., Trumbore, S.E., Chadwick, O.A., Vitousek, P.M. and Hendricks, D.M., 1997. Mineral control of soil organic carbon storage and turnover. *Nature,* **389**, 170-173.

Torquebiau, E., 2000. A renewed perspective on agroforestry concepts and classification, *C.R. Academy of Science, Paris, Life Sciences*, **323,** 1009-1017.

Tributh, H., Boguslawski, E., Lieres, A., Steffens, D. and Mengel, K., 1987. Effect of potassium removal by crops on transformation of illitic clay minerals. *Soil Science*, **143,** 404-409.

Tubiello, F.N., Salvatore, M., Condor Golec, R.D., Ferrara, A., Rossi, S., Biancalani, R., Federici, S., Jacobs, H. and Flammini, A., 2014. Agriculture, forestry and other land use emissions by sources and removals by sinks, 1990-2011 Analysis. FAO Statistics Division Working Paper Series ESS/14/02.

Turner, J. and Kelly, J., 1977. Soil chemical properties under naturally regenerated *eucalyptus* spp. and planted Douglas-fir. *Australian Forest Research*, **7**, 163-172.

Turner, J. and Lambert, M., 2000. Change in organic carbon in forest plantation soils in eastern Australia. *Forest Ecology and Management*, **133**, 231-247.

Turner, J., Lambert, M.J. and Johnson, D.W., 2005. Experience with patterns of change in soil carbon resulting from forest plantation establishment in eastern Australia. *Forest Ecology and Management*, **220**, 259-269.

Ugalde, D., Brungs, A., Kaebernick, M., McGregor, A. and Slattery, B., 2007. Implications of climate change for tillage practice in Australia, *Soil and Tillage Research,* **97**, 318-330.

UNFCCC., 2015. http://unfccc.int/ghg_data/items/3825.php Last accessed 13 Oct 2015.

Vagen T.G., Lal, R. and Singh, B.R., 2005. Soil carbon sequestration in sub-Saharan Africa: A review. *Land Degradation and Development,* **16,** 53-71.

Valzano, F., Murphy, B. and Koen, T., 2005. The impact of tillage on changes in soil carbon density with special emphasis on Australian conditions. National Carbon Accounting Technical Report No. 43. Australian Greenhouse Office, Commonwealth of Australia.

VandenBygaart, A.J., McConkey, B.G., Angers, D.A., Smith, W., de Gooijer, H., Bentham, M., and Martin, T., 2008. Soil carbon change factors for the Canadian agriculture national greenhouse gas inventory. *Canadian Journal of Soil Science,* **88**, 871-680.

Van Dijk, A.I.J.M., Beck, H.E., Crosbie, R.S., de Jeu, R.A.M., Liu, Y.Y., Podger, G.M., Timbal, B. and Viney, N.R., 2013. The Millennium Drought in southeast Australia (2001-2009): Natural and human causes and implications for water resources, ecosystems, economy and society. *Water Resources Research,* **49,** 1040-1057.

Velde, B. and Meunier, A., 2008. *The origin of clay minerals in soils and weathered rocks.* Springer-Verlag, Berlin.

Vesk, P.A., Nolan, R., Thomson, J.R., Durrough, J.W. and MacNally, R., 2008. Time lands in provision of habitat resources through revegetation. *Biological Conservation*, **141**, 174-186.

Vesterdal, L., Ritter, E. and Gundersen, P., 2002. Change in soil organic carbon following afforestation of former arable land. *Forest Ecology and Management,* **169,** 137-147.

Vincent, K.R. and Chadwick, O.A., 1994. Synthesizing bulk density for soils with abundant rock fragments, *Soil Science Society of America Journal,* **58,** 455-464.

Viscarra Rossel, R.A., Walvoort, D.J.J., McBratney, A.B., Janik, L.J. and Skjemstad, J.O., 2006. Visible, near infrared, mid infrared or combined diffuse reflectance spectroscopy for simultaneous assessment of various soil properties, *Geoderma,* **131**, 59-75.

Viscarra Rossel, R.A., Webster, R., Bui, E.N. and Baldock, J.A., 2014. Baseline map of organic carbon in Australian soil to support national carbon accounting and monitoring under climate change. *Global Change Biology*, **20**, 2953-2970.

Viscarra Rossel, R.A., Chen, C., Grundy, M.J., Searle, R., Clifford, D. and Campbell, P.H., 2015. The Australian three-dimensional soil grid: Australia's contribution to the *GlobalSoilMap* project. *Soil Research*, **53**, 845-864.

Visser, S.A. and Caillier, M., 1988. Observations on the dispersion and aggregation of clays by humic substances, 1. Dispersive effects of humic acids. *Geoderma,* **42,** 331-337.

Vogel, A.W., 1994. *Compatibility of soil analytical data: determinations of cation exchange capacity, organic carbon, soil reaction, bulk density and volume percent of water at selected pF values by different methods.* Working Paper and Preprint 94/07, ISRIC, Wageningen.

Walker, P.H., Chartres, C.J. and Hutka, J., 1988. The effect of aeolian accessions on soil development on granitic rocks in south-eastern Australia. 1. Soil morphology and particle-size distributions, *Australian Journal of Soil Research,* **26**, 1-16.

Walker, T.W. and Adams, A.F.R., 1958. Studies on soil organic matter: 1. Influence of phosphorus content of parent materials on accumulations of carbon, nitrogen, sulfur and organic phosphorus in grassland soils. *Soil Science*, **85**, 307-318.

Walkley, A., 1947. A critical examination of a rapid method for determining organic carbon in soils – effect of variations in digestion conditions and of inorganic soil constituents, *Soil Science*, **63**, 251-264.

Walkley, A., and Black, I.A., 1934. An examination of the Degtjareff method for determining soil organic matter, and a proposed modification of the chromic acid titration method, *Soil Science,* **37**, 29-38.

Walker, L.R., Wardle, D.A., Bardgett, R.D. and Clarkson, B.D. 2010. The use of chronosequences in studies of ecological succession and soil development. *Journal of Ecology*, **98**, 725-736.

Wang, X.J., Smethurst, P.J. and Holz, G.K., 1996. Nitrogen mineralisation indices in ferrosols under eucalypt plantations of north-western Tasmania: association with previous land use. *Australian Journal of Soil Research,* **34**, 925-935.

Wasson, R.J., Mazari, R.K., Starr, B. and Clifton, G., 1998. The recent history of erosion and sedimentation on the Southern Tablelands of south eastern Australia: sediment flux dominated by channel incision. *Geomorphology,* **4**, 291-308.

Warren, J.F., 1965. The scalds of western New South Wales – A form of water erosion. *Australian Geographer*, **9**, 282-292.

Waters, C.M., Melville, G.J., Orgill, S.E. and Alemseged, Y., 2015. The relationship between soil organic carbon and soil surface characteristics in the semi-arid rangelands of southern Australia. *The Rangeland Journal,* **37,** 297-307.

Watson, R.T., Noble, I.R., Bolin, B., Ravindranath, N.H., Verardo, D.J., and Dokken, D.J (eds), 2000. Land Use, Land Use Change and Forestry. Cambridge University Press, Cambridge, England.

Watkins, J.J. and Meakin, N.S., 1996. Explanatory Notes Nyngan and Walgett 1:250 000 Geological Sheets SH/51–15 & SH/55–11. Geological Survey of NSW.

Wendt, J.W. and Hauser, S., 2013. An equivalent soil mass procedure for monitoring soil organic carbon in multiple soil layers. *European Journal of Soil Science,* **64**, 58-65.

West, T.O. and Post, W.M., 2002. Soil organic carbon sequestration rates by tillage and crop rotation: a global data analysis. *Soil Science Society of America Journal*, **66**, 1930-1946.

West, T.O. and Six, J., 2007. Considering the influence of sequestration duration and carbon saturation on estimates of soil carbon capacity, *Climate Change*, **80**, 25-41.

Westoby, M., Walker, B. and Noy-Meir, I., 1989. Opportunistic management for rangelands not at equilibrium. *Journal of Range Management*, **42**, 266-274.

Whalley, R.D.B., Robinson, G.G. and Taylor, J.A. 1978. General effects of management and grazing by domestic livestock on the rangelands of the northern tablelands of New South Wales. *Australian Rangeland Journal*, **1**, 174-190.

Whitehead, D., 2011. Forests as carbon sinks – benefits and consequences. *Tree Physiology*, **31**, 893-902.

Williams, J., Helyar, K.R., Greene, R.S.B. and Hook, R., 1993. Soil characteristics and processes critical to the sustainable use of grasslands in arid, semi-arid and seasonally dry environments. In, Proceedings of the XV11 International Grassland Congress, pp. 1335-1350.

Willatt, S.T. and Pullar, D.M., 1984. Changes in soil physical properties under grazed pastures. *Australian Journal of Soil Research*, **22,** 343-348.

Wilson, M.J. 1999. The origin and formation of clay minerals in soils: past, present and future perspectives. *Clay Minerals*, **34**, 7-25.

Wilson, B.R. Growns, I. and Lemon J., 2007. Scattered native trees and soil patterns in grazing land on the Northern Tablelands of New South Wales, Australia. *Australian Journal of Soil Research,* **45**. 199-205.

Wilson, B.R. Koen, T.B. Barnes, P. Ghosh, S. and King, D., 2011. Soil carbon and related soil properties along a soil type and land-use intensity gradient, New South Wales, Australia. *Soil Use and Management,* **27**, 437-447.

Wilson, M.J., 2004. Weathering of the primary rock-forming minerals: processes, products and rates. *Clay Minerals*, **39**, 233-266.

Wong, V.N.L. Murphy, B.W. Koen, T.B. Greene, R.S.B. and Dalal, R.C., 2008. Soil organic carbon stocks in saline and sodic landscapes. *Australian Journal of Soil Research,* **46**, 378-389.

Wong, V.N.L. Greene, R.S.B. Dalal, R.C. and Murphy, B.W., 2010. Soil carbon dynamics in saline and sodic soils: a review. *Soil Use and Management*, **26**, 2-11.

Wuest, S.B., 2014. Seasonal variation in Soil Organic Carbon. *Soil Science Society of America Journal,* **78**, 1442-1447.

Wuest, S.B., 2015., Seasonal variation in soil bulk density, organic N, available phosphorus, and pH. *Soil Science Society of America Journal,* **79,** 1188-1197.

Wyborn, D., Owen, M., Compston, W. and McDougall, I., 1982. The Laidlaw Volcanics: a Late Silurian point on the geological time scale. *Earth and Planetary Science Letters,* **59**, 90-100.

Xie, H.T., Yan, X.M., Drury, C.F., Yang, J, Y. and Zhang, X.D., 2011. Predicting soil organic carbon and total nitrogen using mid- and near-infrared spectra for Brookston clay loam soil in southwestern Ontario, Canada. *Canadian Journal of Soil Science,* **91**, 53-63.

Xiong, X., Grunwald, S., Brenton Myers, D., Kim, J., Harris, W.G. and Bliznyuk, N., 2015. Assessing uncertainty in soil organic carbon modelling across a highly heterogeneous landscape. *Geoderma,* **251-252,** 105-116.

Yuan, B.C., Li, Z.Z., Liu, H., Gao, M. and Zhang, Y.Y., 2007. Microbial biomass activity in salt affected soils under arid conditions. *Applied Soil Ecology*, **35**, 319-328.

Yates, C.J. and Hobbs, R.J., 1997. Temperate eucalypt woodlands: a review of their status, processes threatening their persistence and techniques for restoration. *Australian Journal of Botany*, **45**, 949-973.

Yong-Zhong, S., Yu-Lin, L., Jian-Yuan, C. and Wen-Zhi, Z., 2005. Influences of continuous grazing and livestock exclusion on soil properties in a degraded sandy grassland, Inner Mongolia, northern China. *Catena*, **59**, 267-278.

Zech, W. Senesi, N. Guggenberger, G. Kaiser, K. Lehmann, J. Miano, T.M. Miltner, A. and Schroth, G., 1997. Factors controlling humification and mineralization of soil organic matter in the tropics. *Geoderma,* **79**, 117-161.

Zeng, W. and Wang, W., 2015. Combination of nitrogen and phosphorus fertilization enhance ecosystem carbon sequestration in a nitrogen-limited temperate plantation of Northern China. *Forest Ecology and Management*, **341**, 59-66.

Zobeck, T.M. and Van Pelt, R.S., 2011. Wind erosion, in *Soil management: building a stable base for agriculture*, eds. J.L. Hatfield and T.J. Sauer, American Society of Agronomy and Soil Science Society of America, Madison, USA: Chapter 14.

Appendix 1: Tree and shrub species found in BEPs

	Genus	Tree/shrub
A. baileyana	Acacia	Shrub/small tree
A. boormani	Acacia	Shrub
A. buxifolia	Acacia	Shrub
A. cardiophylla	Acacia	Shrub/small tree
A. cultriformis	Acacia	
A. dealbata	Acacia	Shrub/tall tree
A. decurrens	Acacia	Small tree
A. implexa	Acacia	Small tree
A. mearnsii	Acacia	Tree
A. melanoxylon	Acacia	Small/large tree
A. parramattensis	Acacia	Tall shrub/small tree
A. pravissima	Acacia	Shrub/small tree
A. retinoides	Acacia	Shrub/small tree
A. rubida	Acacia	Shrub/small tree
A. terminalis	Acacia	Shrub/small tree
A. verniciflua	Acacia	Shrub/bushy tree
C. Violaceus	Callistemon	Shrub
D. genistifolia	Daviesia	Shrub
D. cuneata	Dodonea	Shrub
D. viscosa	Dodonea	Shrub
E. albens	Eucalyptus	Tree
E. albens	Eucalyptus	Tree
E. blakelyi	Eucalyptus	Tree
E. bridgesiana	Eucalyptus	Tree
E. camaldulensis	Eucalyptus	Tree
E. camphora	Eucalyptus	Tree
E. cinerea	Eucalyptus	Tree
E. crenulata	Eucalyptus	Tree
E. dives	Eucalyptus	Tree
E. goniocalyx	Eucalyptus	Tree
E. macarthurii	Eucalyptus	Tree
E. macrorhyncha	Eucalyptus	Tree
E. mannifera	Eucalyptus	Tree
E. melliodora	Eucalyptus	Tree
E. mircrocarpa	Eucalyptus	Tree
E. ovata	Eucalyptus	Tree
E. pauciflora	Eucalyptus	Tree
E. polyanthemos	Eucalyptus	Tree
E. polygalifolium	Eucalyptus	Tree
E. rossii	Eucalyptus	Tree
E. rubida	Eucalyptus	Tree
E. sideroxylon	Eucalyptus	Tree
E. viminalis	Eucalyptus	Tree
I. australis	Indigofera	shrub
K. ericoides	Kunzea	Shrub/small tree
L. flavescens	Leptospermum	Shrub/small tree
L. juniperinum	Leptospermum	Shrub/small tree
L. lanigerum	Leptospermum	Shrub/small tree
M. ericifolia	Melaluca	Shrub

Appendix 2: Annual average rainfall data for 1992-2011

Yass Monthly Rainfall data 1992 - 2011

	Jan	*Feb*	*Mar*	*Apr*	*May*	*Jun*	*Jul*	*Aug*	*Sep*	*Oct*	*Nov*	*Dec*	*Annual*
1992	123.4	97	26.4	25	39	51	31	120.4	78.6	136.4	82.4		
1993	35.6	25.4	98.8	1.4	45.8	48.2	188	38.2	113.2	123.4	92.6	64.2	874.8
1994	39.2	49.8	45.8	56.2	16.2	75.8	45.2	23.8	13.2	43	81	36.2	525.4
1995	133.4	0.8	0	21.2	184.6	71.8	116.4	15	78.8	71.2	137.6	70	900.8
1996	116.4	33.8	36.6	19.2	52	79.2	128.4	71	130.6	64	69.6	81.2	882
1997	23.8	13.8	n/a	0	46.4	103	27.6	n/a	119.2	29.2	48.4	9.8	
1998	41.4	44.4	1.4	84.6	29	128.4	114.6	129	n/a	n/a	88	15.8	
1999	75	33.2	101	41.6	n/a	49.8	32.6	54.6	58.6	149.8	36.2	120	
2000	36.4	15.4	50.6	46.6	73.8	53.4	n/a	110.6	66	64.6	102.6	15	
2001	23.8	54.4	95	23.2	5.2	n/a	48.2	66.8	47.8	63	n/a	22.6	
2002	10	135.6	27.2	11.2	52.8	63.2	47.8	30	82.2	3	4.8	39.6	507.4
2003	15.6	58.2	53.4	46.8	23.4	101.6	54	95.8	68.4	89.4	90.4	145.4	842.4
2004	52.8	26	4.6	10.6	14	67.8	49	70.2	57.2	62.8	90.2	99	604.2
2005	48.6	56	20.4	4	0	97.2	125.4	81.2	114.4	56.6	74.6	22.2	700.6
2006	62.8	4.8	n/a	14	5.2	54	70.9	37.7	36.2	0	20.6	97.3	
2007	7.2	57.5	62.9	46.8	51.2	n/a	54	6.6	14.7	30.9	91.6	n/a	
2008	63.8	49.4	34.3	31.5	19.9	62	n/a	44.5	63.4	66.5	80.3	n/a	
2009	82.4	16.8	13.5	78.3	7.5	73	67.1	38.1	68.1	55.3	77	117	694.1
2010	16	123.1	86.4	45.5	64.2	27	82.8	103.6	83.1	71.4	103.2	138.5	944.8
2011	36.4	129.1	54.6	32.7	48.6	32.1	64	61	65.2	38.3	110.4	61.5	733.9

Yass (Linton Hostel) 1898-2011 average 650 mm

Murrumbateman monthly rainfall 1992-2011

	Jan	*Feb*	*Mar*	*Apr*	*May*	*Jun*	*Jul*	*Aug*	*Sep*	*Oct*	*Nov*	*Dec*	*Annual*
1992	120	63.5	40	30.5	27.5	38.5	24	110	73	99	74	82	782
1993	73.4	33	115.2	13	30	26.9	178.5	32	108.5	129.5	97.7	49	886.7
1994	19	59.5	24.5	55.5	9.5	80	31.5	15.5	11	69.5	103.5	46.5	525.5
1995	164.5	10	0	23	160.5	59	142.5	11	80.5	86	150	94.5	981.5
1996	106	35	78.7	24	60.5	72.5	111	n/a	120	65	72	69.7	
1997	30	10.7	33.7	0	44.9	n/a	26.5	42	147.1	28.5	22.8	10.9	
1998	38	40	3.5	68.9	31.4	145.2	113.7	115.4	97.4	79.6	94.9	7.7	835.7
1999	63.9	25	97.1	53.8	43.8	49.6	23.6	63.6	58	159.9	40.1	124.6	803
2000	29	13.8	46.3	47.8	91.1	50.5	60.8	113.1	68.2	87.6	103.3	22.8	734.3
2001	20.3	112.9	84.3	12.5	6.5	65.7	40.2	86.1	51.4	53.4	41.2	21	595.5
2002	n/a	163.5	35.7	14.4	46.4	67.4	38.3	37.5	67.2	7.5	7.8	30.2	
2003	21	65.4	50	27	20.1	56.2	53.8	97.3	59.2	74.4	84.4	79.6	688.4
2004	57.4	26.4	22.6	4	8	49.8	34.6	n/a	45.6	59.4	90.6	96.4	
2005	35.4	58.4	24	7.2	0.8	92.4	115.6	78.8	110	65.4	93.4	27	708.4
2006	73.2	12.8	26.2	21.4	10.6	64.4	53.6	15.9	7	4.8	21.6	23.9	335.4
2007	4.1	36.8	31.4	31.6	n/a	81	51.8	12.8	21.2	21.6	110.2	89.4	
2008	51.6	36.4	42	25.2	12.2	48.2	66	40.8	46.4	51	70.6	85.4	575.8
2009	59.2	16.4	14	79.8	6.2	56.8	55	31.4	71.6	49	26.6	92.8	558.8
2010	50.2	150.8	50	28.4	58.9	37.8	80.8	102.7	69.6	61	168	153.7	1011.9
2011	37	98.2	42.8	19.2	35.1	20.9	40.4	54.2	41	44	132.5	112.6	677.9

Average annual rainfall 1985-2016 = 726 mm

Gundaroo monthly rainfall 1992-2011

	Jan	*Feb*	*Mar*	*Apr*	*May*	*Jun*	*Jul*	*Aug*	*Sep*	*Oct*	*Nov*	*Dec*	*Annual*
1991	81	20.2	20.8	17.2	22.6	90.6	120.2	77	55.8	27.2	40.6	63.2	636.4
1992	107.8	70.1	49.2	28.6	25	33	21.7	87	79.8	74	72.6	63.4	712.2
1993	87.4	20.8	113.4	4	22	25	126	18	89.2	119.6	96.2	26	747.6
1994	16.8	99.4	37.8	76.8	6.8	35.2	22.7	5.7	8.4	71	113.8	41.8	536.2
1995	132.4	8.6	1.2	9	149	42.6	95.4	5.8	59.6	58.8	97.6	65.2	725.2
1996	89.8	53.8	82.2	19.6	54.2	42.6	80.6	52	88.4	48.2	86.6	61.8	759.8
1997	28.8	14.2	24.6	0	27.4	130.2	19	24.6	115.8	28.8	19	11.8	444.2
1998	48.6	22.8	7	53	52	115.7	64.4	105	53.8	69.2	84.8	10.6	686.9
1999	56.2	18.6	64.8	41	38.8	32.6	20.6	43	55.6	n/a	n/a	120.4	
2000	18.6	7.6	61.2	35.6	52.6	32.4	48.4	80.6	61.2	61	116	22.6	597.8
2001	51.6	67.6	52.2	n/a	2.2	43.8	40.6	74	52.4	n/a	n/a	10.2	
2002	11.4	151.4	20.8	3.8	27.6	51.4	24.4	30.8	60	5.4	11.2	38.8	437
2003	20.2	54.8	n/a	12.4	18.6	63.1	58	58.8	38.8	83.9	42.6	101.2	
2004	31.6	22.6	13.2	5	6.4	26	26	60.8	47.4	60.4	71.8	66.2	437.4
2005	28.2	48.8	28.4	7.6	0.2	67.6	103.4	62.8	99.8	51.6	n/a	16	
2006	n/a	15	24.2	15.2	5.8	87	40	9.6	9	4	31.4	22.6	
2007	9.6	28.2	30.6	28.2	47.8	95.8	34	7.8	22.2	43.4	66	71.8	485.4
2008	65.6	38.2	31.2	20.2	13	47.8	50.6	38.6	45.2	48	110.6	78.4	587.4
2009	82	12.8	21.2	53.4	10.4	42.4	40	23	57.4	64	12.6	101.4	520.6
2010	14.4	149.9	50.9	20	81	39	70	81	65.2	56.4	102.8	144.6	875.2
2011	n/a	108	35.6	18.6	22.8	22.8	34.6	42.2	34.2	33.4	158.4	125.1	

Average annual rainfall 1987-2014 = 633 mm

Hall monthly rainfall 1992-2011

	Jan	*Feb*	*Mar*	*Apr*	*May*	*Jun*	*Jul*	*Aug*	*Sep*	*Oct*	*Nov*	*Dec*	*Annual*
1991	74.2	8.4	22.8	10.2	40.8	120.6	110.4	83.4	62.4	24.4	30.2	65.6	653.4
1992	105.2	75.8	76	45.3	27.8	31	17.8	115.6	71	100.8	125	67.6	858.9
1993	94.2	26	87.4	16	28.2	29.6	164.4	20.8	99.6	86	68.8	33.8	754.8
1994	9	49.2	40.4	62	3.6	38.5	24.8	5.6	4.6	38.8	80.6	63.4	420.5
1995	203.8	5.6	6.4	18	169.2	63.6	91	8.2	61.4	87.4	101	81.8	897.4
1996	123.4	39	101.6	20.6	69	88.8	71.6	60.5	86.4	52.2	67.6	58.8	839.5
1997	25.8	34.6	12.2	0	34.2	82.4	22.2	38	109.4	31.4	28.6	11	429.8
1998	25.8	35.8	13.2	59.8	37	145.4	110.8	130.8	53.2	75.8	88.4	5.8	781.8
1999	97.2	14.8	66.6	59.6	31.2	43.4	n/a	57.6	88.6	154.8	44.4	150	
2000	32.8	15	42.2	84	74.8	35	41.6	n/a	76.4	75.4	126.4	26	
2001	47.6	89.6	55.6	7.2	2.4	52	40.4	n/a	n/a	n/a	34.4	6.2	
2002	13.6	195.6	52.4	12.8	36.6	49.8	34	23.6	64.8	n/a	11.8	n/a	
2003	14.4	82.2	n/a	33.2	31.6	66.6	57.1	83.4	44	n/a	97.8	83.8	
2004	37.4	29	3.8	3.2	12.2	31.2	24.8	95.9	40.8	70.6	73.4	91	513.3
2005	43.8	60.2	30.6	8.2	0	99.2	105	60.4	124.2	70.4	87.4	n/a	
2006	67.6	17.8	31.8	12.6	11.2	78.4	n/a	25.6	18	5.6	31.8	18	
2007	13.4	n/a	39.2	38.2	52.6	81	n/a	8	21.8	18.8	85	150.4	
2008	41.6	37	38.2	23.4	10.8	36	53.3	43.6	57.4	39.4	67.2	103.2	551.1
2009	51	7	13.8	82	8.4	44.6	47.8	38.6	96.8	64.8	24	86.2	565
2010	13.6	146	81.2	28	71.4	30.7	89.6	81	86.6	101.4	144.4	186.4	1060.3
2011	51.8	156.6	45.4	13.6	21	19.9	41.8	57.6	54	47.4	148	53.2	710.3

Average annual rainfall 1903-2016 = 700mm

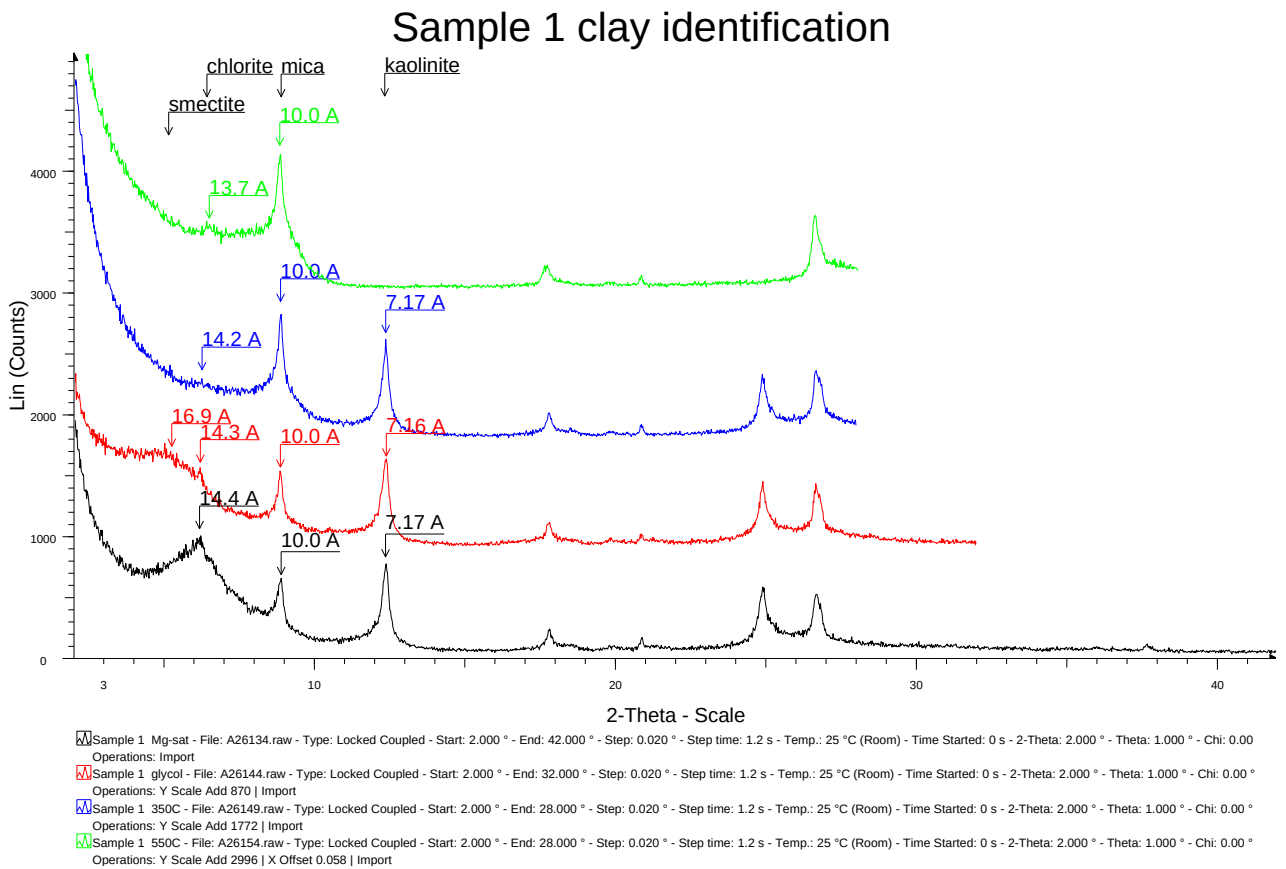
Goulburn (Springfield) monthly rainfall 1992-2011

Year	*Jan*	*Feb*	*Mar*	*Apr*	*May*	*Jun*	*Jul*	*Aug*	*Sep*	*Oct*	*Nov*	*Dec*	*Annual*
1991	100.5	36.2	11	31.5	35.8	130.4	92	83.2	45.6	33	55	76.8	731
1992	152.6	89.2	61	20.8	35.2	36.6	15.6	68	60.6	68.6	73.2	68.4	749.8
1993	70.6	28.2	85.6	5.2	34.8	29.4	118	11.2	48	75.6	61.3	52.4	620.3
1994	15.2	103	59	114.7	6.8	46.2	23.8	8.2	4.2	51.2	87.4	47.6	567.3
1995	142.6	13	14	4	119.8	39.4	57.4	5.8	58.6	76.2	98.6	64.8	694.2
1996	172.5	40.8	21.4	35.4	71.8	45	71.6	54	78.4	76.4	77.6	39.6	784.5
1997	47.2	22.4	42.4	0	31.2	156.7	10.2	21.5	115.2	26.2	19	15.2	507.2
1998	84.2	26.4	0	59.1	55.2	89.2	65	127.4	n/a	55	77.8	10.2	
1999	138.4	32.6	79.6	21.8	24	25.2	39	52.6	48.6	164.2	33.2	166.7	825.9
2000	20.4	14.4	74.8	n/a	46.8	38.2	n/a	n/a	61.4	53	99	16	
2001	81.8	129	67.3	14.2	n/a	29.2	53.9	n/a	51.2	53.8	62	11.2	
2002	20	164.8	20.4	22.9	27.2	31	29.4	30.8	51.6	5.2	8.6	60	471.9
2003	24.4	79.2	60.2	28	30.7	57.4	28.6	52.2	41.9	72.2	78.4	49	602.2
2004	34.2	56.8	26.6	5.9	3.9	15.8	32.6	36.4	64	83.2	72.4	54.6	486.4
2005		57.4	57	12.6	2	71.2	97.4	42.6	84.2	59.7	127.2	29	
2006	101.8	8	18.7	7.8	14.2	112.2	45	15.6	20.1	0	25	51.7	420.1
2007	1.9	89.3	37	35.5	30.6	183.2	23.2	9.2	19	23.7	146.8	n/a	
2008	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a	106.8	
2009	52.8	22.8	4.9	59.8	13.6	25.6	35.6	24.8	38.8	43.7	18.4	99.8	440.6
2010	28	133.2	41.2	9.4	74.2	13	58.2	66	41.4	31	95.4	227.9	818.9
2011	29	118.3	56.6	16.5	14.6	20.2	24.2	69.8	32.6	40.6	86.2	67	575.6

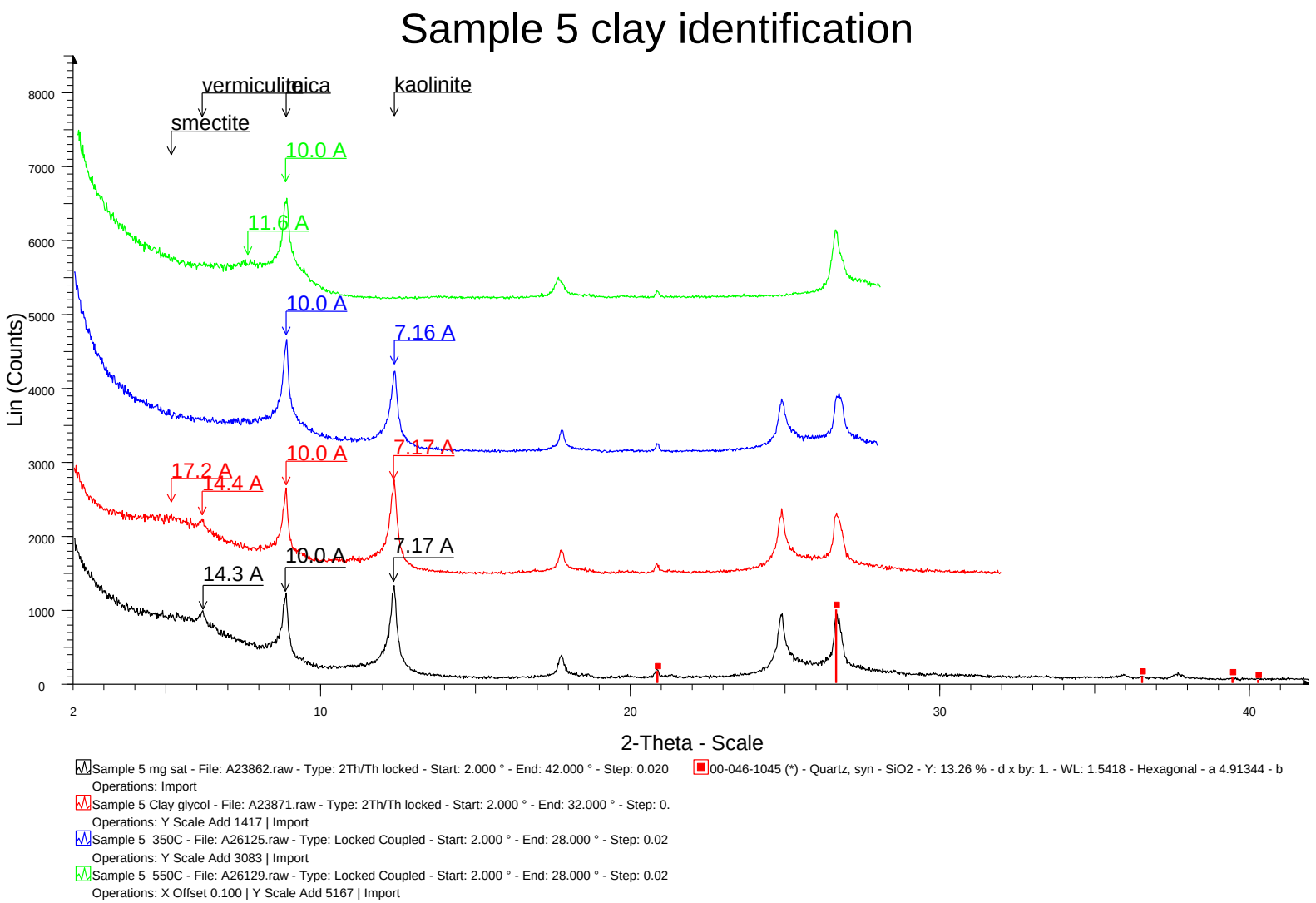
Average annual rainfall Goulburn (Springfield) 1930-2016 = 671mm

Appendix 3: X-ray diffraction spectra for a scald and a waterpond

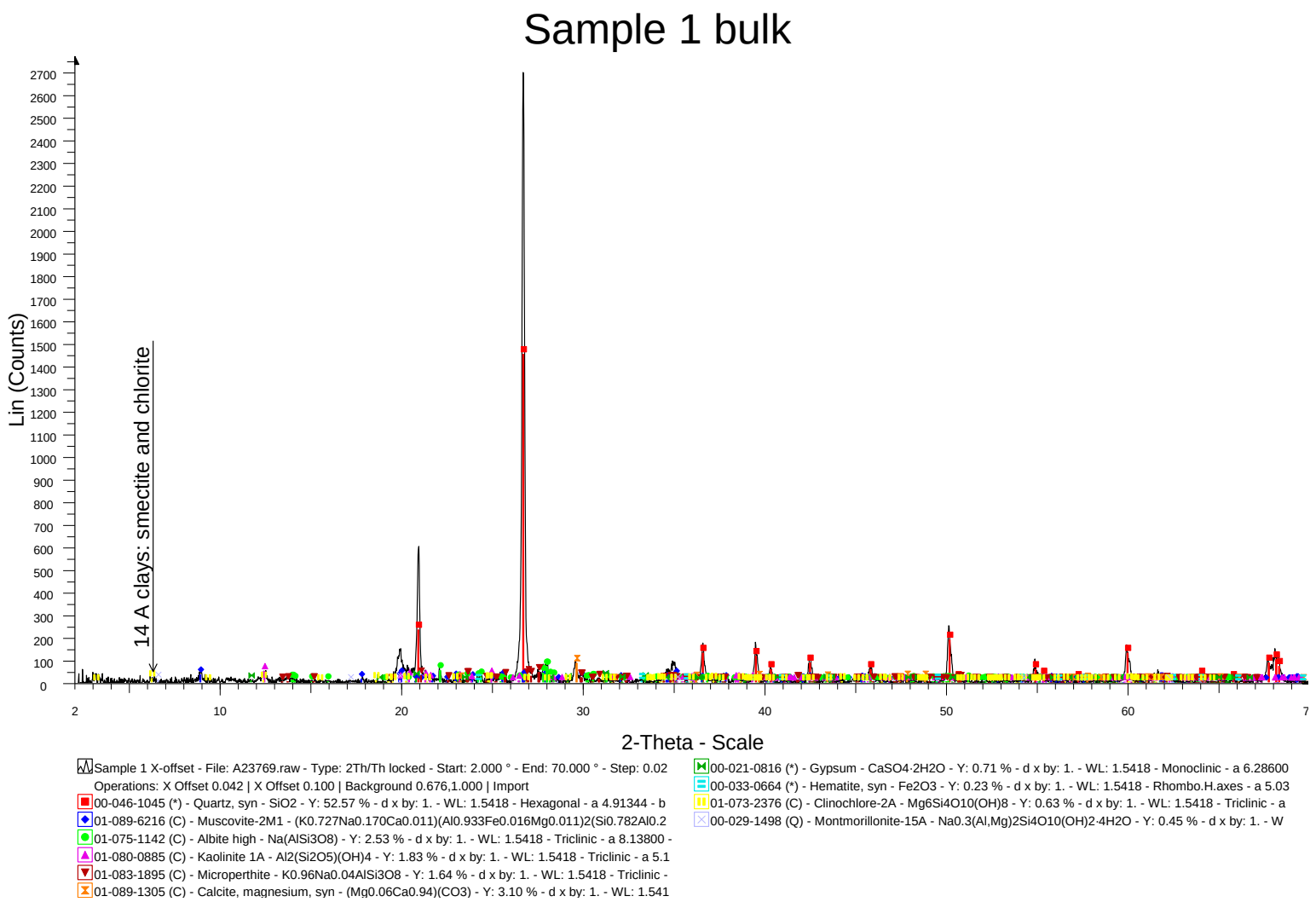
Sample 1: Scald, 0-5 cm layer



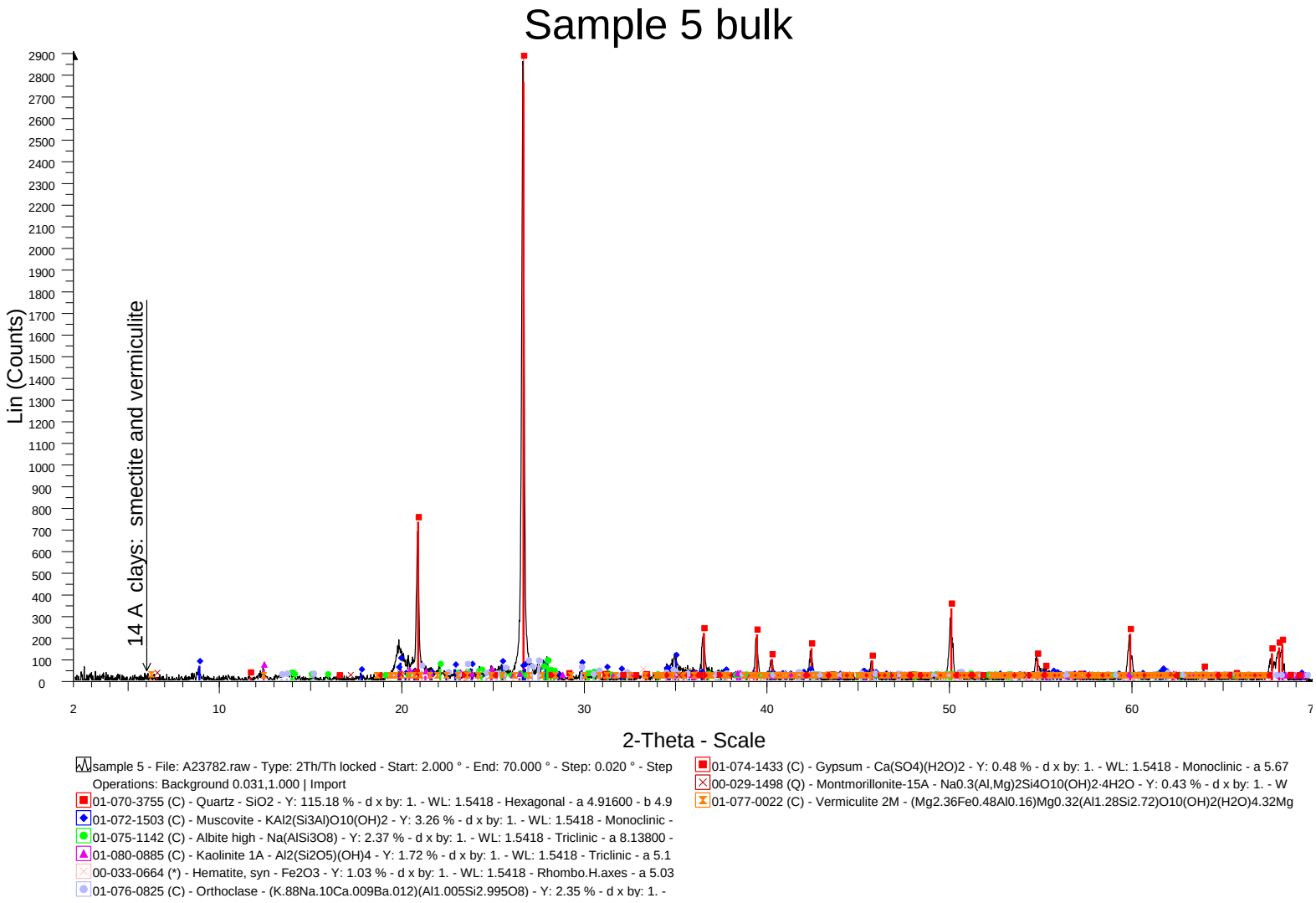
Sample 5: 27 year old waterpond 0-5 cm layer



Sample 1: Scald, 0-5 cm layer



Sample 5: 27 year old waterpond 0-5 cm layer



Appendix 4: BEP Equivalent Soil Mass calculation

Position	*Transect*	*Depth (cm)*	*Bulk Density (g cm^3)*	*SOC (%)*	*SCS (Fixed depth) (Mg C Ha^{-1})*	*Soil mass (fixed depth) (Mg ha^{-1})*	*Reference soil mass (0-5 cm) (Mg ha^{-1})*	*SCS (ESM) (0-5 cm) (Mg C ha^{-1})*	*Soil Mass (fixed depth (0-10 cm) (Mg ha^{-1})*	*Reference soil mass (0-10 cm) Mg ha^{-1})*	*SCS (ESM) (5-10 cm) (Mg C ha^{-1})*	*Soil Mass (fixed depth (0-20 cm) (Mg ha^{-1})*	*Reference soil mass (0-20 cm) Mg ha^{-1})*	*SCS (ESM) (10-20 cm) (Mg C ha^{-1})*	*Soil Mass (fixed depth (0-30 cm) (Mg ha^{-1})*	*Reference soil mass (0-30 cm) Mg ha^{-1})*	*SCS (ESM) (20-30 cm) (Mg C ha^{-1})*	*SCS (Fixed depth (0-30 cm) Mg C ha^{-1})*	*SCS (ESM) (0-30 cm) (Mg C ha^{-1})*
BEP	T1	0-5	0.8	3.2	12.9	412	371	12.6											
BEP	T1	5-10	1.3	1.1	6.9	619			1031	1006	7.1								
BEP	T1	10-20	1.5	0.7	9.4	1371						2402	2302	9.2					
BEP	T1	20-30	1.6	0.4	5.3	1415									3817	3637	5.0	34.5	33.9
BEP	T2	0-5	0.8	3.8	14.8	393	371	14.5											
BEP	T2	5-10	1.3	1.4	9.1	643			1035	1006	9.3								
BEP	T2	10-20	1.6	0.5	7.8	1517						2552	2302	7.1					
BEP	T2	20-30	1.5	0.3	4.6	1430									3983	3637	4.3	36.4	35.3
BEP	T3	0-5	0.9	2.8	12.0	436	371	11.3											
BEP	T3	5-10	1.4	1.2	8.3	673			1109	1006	8.7								
BEP	T3	10-20	1.5	0.4	5.7	1284						2392	2302	6.0					
BEP	T3	20-30	1.6	0.4	5.0	1261									3653	3637	5.5	31.1	31.4
AL	T1	0-5	0.8	2.8	11.8	422	371	11.1											
AL	T1	5-10	1.5	1.4	10.7	754			1176	1006	10.3								
AL	T1	10-20	1.6	0.7	11.1	1584						2761	2302	11.2					
AL	T1	20-30	1.6	0.2	3.5	1548									4309	3637	3.0	37.1	35.6
AL	T2	0-5	0.9	3.5	16.2	460	371	15.1											
AL	T2	5-10	1.6	1.2	9.9	795			1255	1006	9.8								
AL	T2	10-20	1.5	0.5	7.1	1470						2725	2302	6.0					
AL	T2	20-30	1.5	0.5	7.4	1352									4077	3637	7.3	40.6	38.2
AL	T3	0-5	1.0	3.1	16.0	514	371	14.5											
AL	T3	5-10	1.5	1.0	7.8	750			1264	1006	8.4								
AL	T3	10-20	1.6	0.4	5.4	1519						2783	2302	5.0					
AL	T3	20-30	1.6	0.3	4.4	1535									4318	3637	3.8	33.6	31.7