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Rapid Soil Development in Response to Land Use Change

Case studies from the northern New South Wales Slopes, Plains and New England Tablelands

Robert Gordon Banks

BSc Hons, Dip Bus, CPSS

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Abstract

Soil development and soil formation processes are often considered to take long periods of time from hundreds to millions of years. This thesis investigates the potential for land use change to drive whole soil profile development processes in unexpected ways and at unexpected speeds in Eastern Australia. The major objectives of this thesis were to identify and document locations where rapid soil profile changes occur at a land use boundary, to investigate the nature of soil changes across that boundary and propose mechanisms for observed changes.

Alternating arid and humid phases in the Quaternary, coupled with contemporary Holocene and Anthropocene sub-decadal climate variability, makes availability of soil water in Eastern Australia a critical driver to soil development processes. In contrast, many northern climates are constant or at least seasonal and moist.

An initial study (Chapter 3) was undertaken using archival soil survey data from the Liverpool Plains region in the north-west slopes and plains of New South Wales (NSW). The desktop study examined key soil fertility properties in relation to land use through an historical soil survey dataset to determine whether large-scale European land use impacts on whole soil profiles could be observed. Land uses compared were cropping, native pasture, improved pasture and woodland. Soil type, land use, and soil × land use were considered in a mixed model restricted maximum likelihood (REML) analysis using general fertility attributes of available water holding capacity (AWC), soil phosphorus (Bray P), cation exchange capacity (CEC), dispersion percentage (DP), salinity (EC_e), sodicity (indicated by exchangeable sodium percentage, ESP) and soil pH. Results showed that fertility attribute variation depended on soil type reflecting land selection for agriculture. Soil organic carbon (SOC) values were considered a product of fertility but were most strongly associated with land use with SOC being lowest in cropping regardless of soil type. This approach did not illuminate any man-induced whole soil profile changes. General soil survey data was found to be potentially limited by the method of land use and site history recording.

The second study (Chapters 4 and 5) investigated physical and chemical soil profile changes associated with managed tropical pastures compared with volunteer native pastures on sodic-duplex soils in north-western NSW. These soils are limited by low fertility topsoils, poor soil profile drainage and clay-rich B horizons which native vegetation roots and soil water do not readily penetrate. Results indicated that deeper, more abundant tropical pasture roots had initiated changes in soil profile porosity, structure and chemistry. Macroporosity (pores > 30µm) of critical infiltration and root growth limiting upper B horizons had increased in tropical pasture soils to a point where potential water through flow was 81 fold higher than native pasture upper B horizons.

B horizons under tropical pastures were more aggregate stable than under native pastures. Soil stability was investigated through dispersion and flocculation experiments with varying concentrations of synthetic soil water based on tropical pasture topsoil water chemistry. Higher aggregate stability of upper B horizon material was associated with increased ionic strength and composition of soil water from topsoils in tropical pastures. A feedback mechanism was proposed that introduction of a pasture species which utilises deeper subsoil resources created extra macroporosity. Macropores were stabilised by the chemistry of soil water from managed tropical pastures allowing further development of root structures and potential water storage deeper in the profile.

A third study (Chapter 6) was undertaken at Tenterfield in the NSW New England Tablelands. The study utilised a boundary between relatively undisturbed native forest and land which was cleared for grazing 140 years prior. Using methods developed in Chapter Four, soil physical and chemical attributes were compared between native forest and managed pasture along the forest-pasture boundary. All forested soil profiles were Red Dermosols, whilst all pasture soils were texture contrast soils including Kurosols and Chromosols. A thin, often discontinuous A2 (E horizon) was apparent in the pasture soils with very clear boundaries between A and B horizons. Forested soils had gradational soil texture profiles and a well-developed AB horizon.

The existence of megapores (mean diameter 2 cm) throughout forest soil profiles was a driver of potential soil profile drainage maintaining Dermosols in a freely drained state. A feedback mechanism is proposed whereby clearing and grazing results in compaction of topsoils, infilling/collapse of tree root megapores, and consequent drainage impedance. Drainage impedance creates favourable conditions for clay illuviation and subsequent formation of a texture contrast texture profile. The time frame for development of A2 horizons through clay illuviation was likely to have occurred within 140 years instead of millennia that the literature suggests.

The significance of this work is that rapid and deep soil changes can be initiated by man simply through changing plant root architecture and thence water penetration to soil. A change in philosophy of crop and pasture development towards maximising root abundance and penetration is recommended so that benefits of greater subsoil access by plants are realised, especially in landscapes that have low productivity because deep soil resources have been unavailable. Beneficial soil changes

such as these should be considered crucial to improving soil resilience as climate change impacts on Australia’s grazing systems.

Declaration by author

This thesis is composed of my original work, and contains no material previously published or written by another person except where due reference has been made in the text. I have clearly stated the contribution by others to jointly-authored works that I have included in my thesis.

I have clearly stated the contribution of others to my thesis as a whole, including statistical assistance, survey design, data analysis, significant technical procedures, professional editorial advice, financial support and any other original research work used or reported in my thesis. The content of my thesis is the result of work I have carried out since the commencement of my higher degree by research candidature and does not include a substantial part of work that has been submitted to qualify for the award of any other degree or diploma in any university or other tertiary institution. I have clearly stated which parts of my thesis, if any, have been submitted to qualify for another award.

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Publications included in this thesis

No publications included.

Submitted manuscripts included in this thesis

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Other Publications during candidature

Jing YZ, Huang, J, Banks RG, and Triantafilis, J. (2017). *Scope to map soil landscape units at the district level from remotely sensed γ-ray spectrometry and proximal sensed EM induction data.* Soil Use and Management. 4: 538 – 552.

Banks, RG (2017) *Tropical pastures make beneficial changes to poor soil.* Grasslands Society of NSW Newsletter. 32: 3. 5 – 7.

Banks, RG, Wendling, L, Basford, K, Ringrose-Voase, A. (submitted) *Beneficial soil profile changes induced by tropical grass pastures on sodic-texture contrast soils in Northern NSW.* Submitted to The Rangelands Journal October, 2018.

Contributions by others to the thesis

This thesis is the original work of Robert Gordon Banks. Initial concept development was by Robert Banks with advice from Dr Gunnar Kirchhoff. Additional contributions in a review/advisor/editor capacity were made by Dr Laura Wendling, Dr Anthony Ringrose-Voase, and Professor Kaye Basford.

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- Statistics: Professor Kaye Basford, Dr Vivi Arief
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The University of Queensland School of Agriculture and Food Science analytical services staff were responsible for 100% of soil chemistry analysis.

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Statement of parts of the thesis submitted to qualify for the award of another degree

No works submitted towards another degree have been included in this thesis.

Research Involving Human or Animal Subjects

No animal or human subjects were involved in this research.

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Keywords

Man induced soil development, pedogenesis, soil porosity, soil structure, soil chemistry, tropical pasture, native pasture, moist forest, texture contrast soil formation, Liverpool Plains, New England Tablelands.

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List of Abbreviations

ASWAT	Aggregate stability in water
AWC	Available water holding capacity
BD	Bulk density
BP	Before present
CEC	Cation exchange capacity
DIW	Deionised water
DM	Dry matter
DP	Dispersion percentage
EC	Electrical conductivity
EC_e	Electrical conductivity of saturated soil paste extract (approximate)
ENSO	el Nino-Southern Oscillation
ESP	Exchangeable sodium percentage
FC	Field capacity
IOD	Indian Ocean Dipole
MLA	Meat and Livestock Australia
NPN	Native pasture non-sodic soil or soil solution
NPS	Native pasture sodic soil or soil solution
NW LLS	NSW North West Local Land Services Board
REML	Restricted maximum likelihood mixed model statistical analysis.
SALIS	NSW Soil and Land Information System
SOC	Soil organic carbon
SSW	Synthetic soil water
TCC	Total cation concentration
TN	Total nitrogen
TOC	Total organic carbon
TPN	Tropical pasture non-sodic soil or soil solution
TPS	Tropical pasture sodic soil or soil solution

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Chapter 1 General introduction

This study was inspired by 28 years of professional field experience of the author in soil survey, geomorphology and salinity research, in Eastern Australia and South East Asia. The benefit of extensive experience with many different soil types across a large range of climates and geologies is the realisation that there are exceptions to long held concepts of processes of soil development or soil formation that have formed the basis of widely used soil classifications. The discovery of locations where land use change appears to have changed soil type or character substantially and to great depth within the short European history of Anthropocene Australia have become a point of fascination for the author and has resulted in the preparation of this thesis.

The fact is that changes in soil profile characteristics occurring at man-made fence-line boundaries is poorly documented and has not been thoroughly investigated in Australia. The implication is that human or natural activity differences sufficient to significantly alter soil characteristics or even soil types across a fence-line boundary that separates them. The concept of highly significant soil order change within a human lifetime is not normally considered.

1.1 Aims and scope of this study

The thesis tests the hypotheses that:

1. Changes in land use may result in significant differences in soil physical properties to great depth in soil profiles,
2. Resultant physical changes can create the circumstances for further physical and or chemical change in the profile, and
3. Soil profile development is caused by plant initiated feedback processes which are consequent upon land use change.

The aims of this study are to:

- Conduct desktop and field based studies at specific sites of examples of soil profile change in response to land use in Northern New South Wales,
- Use the data to test the hypothesis that the observed differences are related to human activity,
- Determine the nature of and quantify the differences, and

- And to propose mechanisms which may explain soil characteristic changes within shorter than normally accepted timeframes.

1.2 Outline and rationale

This thesis presents three separate studies with the common theme of identifying soil profile changes or soil order changes associated with land use change. A general literature review is presented in Chapter 2 providing contextual background information on Australian Quaternary to Anthropocene history in terms of climate, vegetation, soil, European settlement and land use change. Processes and timelines of soil formation are discussed (mostly derived from soil classification studies) and exceptions to the commonly accepted rules that drive these processes are discussed. Concepts for methodology development in research studies are discussed. Specific study methodologies are developed in further chapters.

This thesis investigated the following observations of soil change associated with land use change:

1. Of note are the development of topsoil and subsoil structure and increase in productive function of soils with the introduction of tropical pastures in north west NSW on normally very difficult to manage sodic-texture contrast soils (Sodosols and sodic Chromosols). These changes have apparently not occurred with land clearing, a long history of cultivation, and subsequent regional retirement to volunteer native pastures.
2. The development of A2 and $A2_e$ horizons on cleared forest margins of the NSW New England Tablelands. It appears these soils were formerly Dermosols with a gentle progression of textures down the profile and since clearing have developed a texture contrast and A2 horizons.

Most studies of soil changes in Australia have traditionally been from an agricultural viewpoint where soil is viewed as an element for plant growth that can be fed and modified in near surface layers for production. Loss of soil in Australia has been well documented since Europeans arrived in Australia and enthusiastically cleared and developed lands without regard to climate or soil limitations. Deep soil profiles of most soil types are generally perceived as non-dynamic in their characteristics below surface horizons. Despite this, examples exist of whole soil profile change which can be considered beneficial or detrimental in an agronomic sense. This thesis offers an important insight into how man can be a substantial driver of soil change and development with unexpected consequences on whole soil profiles which may have landscape-wide implications in terms of agricultural sustainability and security.

Chapter 2. Literature Review

2.1 Introduction

This thesis proposes that land use changes by man, by forest clearing to plant pastures or introduce a new suite of vegetation species, can initiate soil development processes. These changes drive soil development in a new direction, which leads to relatively rapid and significant soil profile changes. Many soil types or soil Orders (per Isbell and NCST (2016)) are generally perceived to be relatively static, similar to geological classifications of rock types. However, an obvious exception to this are soil types with Acid Sulfate Soil features, such as when hypersulfidic material is exposed to air it can rapidly transform to sulfuric material often within hours to days (e.g. (Fanning *et al.*, 2017).The study of soil taxonomy and pedology has shown that soils change with time and can transition from one soil type to another, generally over long periods of time, and often in a prescribed direction towards an endpoint that is in equilibrium with local parent material, biology, geochemistry (e.g. acid sulfate soils) and climate.

Unlike northern Europe, Asia and America, the Eastern Australian region has not been exposed to extensive glaciation after which soil has formed in newly exposed land surfaces and parent materials. Many Eastern Australian soils are polygenetic in nature, having been exposed to arid and humid climate extremes and associated vegetation through the Quaternary period. An understanding of aspects of the Quaternary history of native grasses and herbivore interactions provides important pre-suppositions for part of this research.

Extensive agricultural production, including both grain and tuber production, has recently been documented for Australia during pre-European times, certainly existing for at least 6–12 000 years (Pascoe, 2014). The landscape scale impacts of this unique agricultural system are presently unknown. No studies have yet been conducted to gauge the effect of Aboriginal agriculture on the landscape. The 1800’s saw an influx of intensive European-style cultivation agriculture and clearing of lands for grazing in Eastern Australia, and this led to substantial documented impacts on soil and landscape stability.

An understanding of accepted soil formation processes and timelines is a useful framework for assessing soil changes observed at the field sites in this study. A review of exceptions to these accepted processes demonstrates that the accepted timelines involved may not be appropriate in the context of this study.

To provide context to the studies that follow, this review outlines:

- The evolution of the Eastern Australian vegetation and landscape,
- The timeline and impacts of humans on the Eastern Australian soil environment,
- A comparison of native Australian and East African grasses and grazing and how differing evolution potentially impacts on soil,
- Well recognised and accepted soil formation and development processes, and
- Examples of soil formation which are more rapid than the accepted processes suggest.

2.2 Quaternary to current evolution of Australian landscapes and vegetation

This section reviews Eastern Australia in recent geological time in relation to the evolution of its vegetation types and how climatic shifts, followed by the arrival of humans, have potentially impacted on soil forming processes in Eastern Australia.

To understand the nature of the Eastern Australian soil environment, it is necessary to understand how vegetation, fauna and climate have interacted in recent geological timeframes. The Quaternary period (12 thousand to 2.6 million years BP) has seen vegetation, fauna, climate, geomorphological processes and potential soil environments in a state of flux prior to substantial change by humans. Current stability is a relative state dependent partly on this co-evolution as well as the impact of more recent human history.

Despite the age of many exposed land surfaces in Australia, the soils appear to be younger than the land surfaces, with many soil profiles of Eastern Australia being of the order of thousands to only several million years old (Gray and Murphy, 1999).

The preceding Tertiary period (2.6 to 65 million years before present, BP) in Australia was a period of relatively moist, stable, warm conditions with widespread mesic vegetation, consisting of many species with rainforest origins and affinities. Many areas were dominated by closed forest with few very tall eucalypts and rainforest understory (Martin, 1994). During this time, there was a general paucity of grass species altogether (Strömberg, 2011), indicating that soil development processes may have operated largely in the absence of grass and grazing systems for much of this period. Some very high rainfall environments such as the edge of the eastern fall near Tenterfield still retain similar types of forest.

The Quaternary period a period of climatic instability, characterised globally by expansion (glacial period) and contraction (interglacial period) of polar and continental ice sheets at least 20 times in the past 2.3 million years (Chappell and Grindrod, 1983). The effects of locking up water in ice caps during glacial maxima, increased the continentality of Australia with sea levels more than 100 m lower at these times relative to present day, resulting in decreased rainfall and lower temperatures. Interglacial periods were as moist as or moister than the present. The previous interglacial to the Holocene (0 – 12,000 years BP) had vegetation with more affinities with rainforest as according to pollen core evidence (Kershaw *et al.*, 1991; Hope, 1994).

There is a growing consensus that dominance of Eucalypts in Eastern Australia appears to have only occurred widely in the previous 5,000 – 6,000 years (mid Holocene) at the cessation of sea level rise following the last glacial maximum (Kershaw *et al.*, 1991; Crowley, 1994). This could be highly significant in terms of the recent findings of Verboom and Pate (2013), who concluded that some mallee species of Eucalypt can change soils from acidic sandy Tenosols (deep acid sands) into Sodosols through the generation of clay from its chemical components (Al and Si) and the creation of a clay rich and alkaline B horizon (discussed in section 2.6 of this review).

The widely held belief that Australia is an old continent where little has changed for many tens of thousands to millions of years has been challenged by the elucidation of Australia's Quaternary and Holocene history (Kershaw *et al.*, 1991; Crowley, 1994; Strömberg, 2011). Post glacial Eastern Australia represents a relatively new vegetation and climate regime which draws some parallels with North America, Northern Europe and Northern Asia, where all flora and fauna have moved in since the retreat of the continental ice (Denton, 1981; Velichko and Isavea, 1992). Unlike Northern America and Northern Europe, most of Eastern Australia was not glaciated so the effects of shifting climates on vegetation and soils have not been entirely obliterated, but superimposed upon many times throughout the Quaternary leaving a legacy of polygenetic soils.

The constant change in climate and vegetation through the Quaternary in Eastern Australia may have left some soil types at a threshold of change, with potential to transition to another soil type. Remnant vegetation may offer some degree of physical protection which has prevented that soil type transition from occurring. Dramatic land use changes in these areas such as forest clearing or changing the dominant plant assemblages, may alter soil exposure to the elements and internal soil water regimes, pushing through a theoretical climate/vegetation threshold and result in associated soil change.

2.3 The two land use change events in Australia

The human history of Australia spans three geological timeframes: the late Quaternary (characterised by large glacial and interglacial events); the Holocene, the most recent interglacial or humid and warm period (not as wet as previous ones) which began 11700 years ago; and the Anthropocene, beginning in 1800, the time when human activity became a dominant influence on world climate and environment (Crutzen and Steffen, 2003; Griffiths, 2018).

The first human induced land use change in Australia

Concurrent with the last interglacial period (30,000 – 120,000 BP) was the arrival of the first humans to the Australian environment between 60,000 and 65,000 years BP (Roberts *et al.*, 1994; Clarkson *et al.*, 2017). Eastern Australia, including the NSW Slopes and Plains, is thought to have been inhabited by about 36,000 years BP (Fillios *et al.*, 2010).

Carbon dates of around 18,000 BP are given for the last evidence of living *Diprotodon* sp. (an iconic and very large wambatid like browser) by Wright (1986). Wright (1986) controversially suggested that humans coexisted with the native flora and megafauna of Australia up to this time and proposed that land management practices of burning and hunting may have had a preparatory effect on the large extinctions and vegetation changes which were about to occur in conjunction with the aridity of the last glacial maximum. Fillios *et al.* (2010) studied the occurrence of megafauna and human artefacts, and confirmed that humans and megafauna coexisted and interacted for at least the first 30,000 years of Aboriginal land use in Australia. The implication of this co-existence is that the last interglacial was relatively stable and food resources including megafauna were probably well managed. The study also found that there was a paucity of more modern grass eating kangaroos which are Australia’s only true obligate grazing animals.

The effects of the last glacial maximum (18,000 years BP) included the disappearance of most rainforest affinity species from much of Eastern Australia in response to landscape drying (Kershaw *et al.*, 1991). Concurrent with this loss of vegetation was the final extinction of most megafauna which relied on browsing. This was likely brought about by a coincidental combination of climate change and Aboriginal land management (Wright, 1986). Aboriginal land management including yam and grain production, and the increased use of fire to maintain open lands, may also have been partly responsible post 18,000 years BP for the greater abundance of grassy woodlands which were observed when Europeans arrived in the late 1700's (Grey, 2012; Pascoe, 2014).

Since the mid Holocene (6,000 years BP), a relatively stable climate, vegetation and land use system had developed in Eastern Australia (Dodson and Mooney, 2002). Unlike many European countries, this stability is characterised by alternating drought and flood regimes driven by the El Nino-Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD) where ocean temperature variations to the east and west of the Australian continent dictate severity and duration of weather patterns (Ummenhofer *et al.*, 2011).

The second human induced land use change in Eastern Australia

Coincident with the global Industrial Revolution and the start of the Anthropocene, permanent European settlement in Australia commenced in the 1780's – 1850's and heralded the largest and most rapid changes in human land use that Australia had experienced to date (McKenzie *et al.*, 2004). During this period most lands in South Eastern Australia were taken up or affected by European settlement. Much of the woodland and forested lands were cleared and either cultivated (lighter tablelands soils and those soils near settlements which became towns) or grazed, predominantly with sheep and cattle. In terms of land use induced vegetation changes and impacts on soils, very little has been documented aside from erosion studies, and agricultural experimentation designed for crop or pasture production.

As early as the 1870's, one of the effects of tree removal from catchments in the Hunter Valley of NSW was noted to be a mass release of soil water through springs as well as increased runoff (Abbott, 1880). The mass release of water through springs may have been be related to the cessation of hydraulic lift employed by trees to bring water to near surface soils (Banks, 2006) thus releasing water from this cycle to shallow aquifers which subsequently drained from the landscape.

There is evidence that erosion rates substantially increased across Eastern Australia from the Broken Hill district in the west to the Tablelands and the Western Slopes in the east. Fanning (1999) demonstrated that once-stable valley fills north of Broken Hill were substantially scoured and gullied shortly after the arrival of European livestock. Similarly, Tablelands and Western Slopes environments experienced severe gully erosion of once-stable valley fills and flood plains within years of the introduction of European livestock (Wasson *et al.*, 1998; Gale and Haworth, 2005; Haworth *et al.*, 2010; Dotterweich, 2013).

Dodson and Mooney (2002) showed in their analysis of multiple palynological (pollen core) studies that the magnitude of the impact of European land use was so large that it had the same results regardless of individual site, catchment and climate characteristics over a very large area of Eastern Australia. The theme of all of these studies is that shortly after an area was colonised by European farmers, there were 30 – 50 years of active and severe gully erosion which slowly stabilised to an erosion rate that is generally slightly higher than before European colonisation. These large scale stripping and infilling with sediment events are surface processes which are a component of soil formation discussed in Section 2.6.

The erosion history of Eastern Australia that is concurrent with European colonisation demonstrates one extensive impact that European settlement had on soils. The greatest cause of the mass erosion event was land clearing and the introduction of grazing animals such as cattle and sheep. Lang (1990) and Shellberg *et al.* (2010) showed that reduced groundcover associated with clearing and European grazing practices significantly increased catchment runoff coefficients which contributed to the acceleration of both sheet and gully erosion processes.

2.4 Co-evolution of grasses and grazers: Australia vs Africa

As illustrated in sections 2.2 and 2.3, European settlers colonised relatively delicately managed land which had few natural grazing animals and co-evolved grasses. This makes Australia different from its European, Asian, African and American counterparts. Evolutionary grazing history and coevolution of grazing animals and grasses has meant that Australian grasses do not have the same evolutionary survival strategies, or performance in grazing systems, as exotic grasses and it is these differences which may have impacts on soil development processes.

Concurrent East African grass and grazer evolution

Strömberg (2011), reported that the late Miocene (7 million years BP) in East Africa saw the evolution of the broad grassland expanses that are known today. Leakey and Harris (2003) demonstrate that modern herd grazing animals (obligate grazers) such as wildebeest (*Connochaetes sp.*) were established in East African by the late Miocene (5 – 7 million years BP). In terms of survival mechanisms for heavy grazing regimes, the African co-evolution of grasses and grazing animals has resulted in grasses that have adapted to a heavy, pulsed grazing regime. The term “pulsed” is used here to indicate a period of intense grazing, with plants eaten and trampled, enriched with dung, followed by rest periods between grazing events.

The current migration of grazers in the Maasai Mara (East Africa) includes herds of 1.7 million wildebeest (*Connochaetes tuarinus mearnsi*), 400,000 Thomson's gazelle (*Eudorcas thomsonii*), 300,000 zebra (*Equus quagga*) 12,000 eland (*Taurotragus oryx*) and associated predators such as lions (*Panthera le*o) (Serneels and Lambin, 2001; Anon, 2015). The migrating herds of grazers follow climatic patterns, and are generally kept moving by their associated predators.

Some of these African grasses are well suited to modern grazing systems in Australia due to their resilience to intense pulsed grazing, their productivity, palatability, and their evolution in a climatically similar environment (Harris *et al.*, 2014). The role of African tropical grasses in modern grazing systems in South Eastern Australia will be discussed in section 2.5.

Australian grass and grazer evolution

In the south eastern Australian Tertiary landscape there was a well-developed assemblage of browsing and arboreal marsupial mammals which utilised the abundance of herbage, and few if any grass dedicated grazers (Strömberg, 2011). Large vertebrate species assemblages in Australia through the Quaternary were dominated by browsers and poorly represented by grazers. Archaeological sites at Cuddy Springs in northern NSW detail typical animals used as food by early humans and show a much higher incidence of browsers such as *Diprotodon*, *Palorhestes* (marsupial Tapir), *Phascolonus* (giant wombat), and specialised predators such as *Pallimimnarchus* sp. (land crocodile) (Fillios *et al.*, 2010). This dominance of browsers over modern native dedicated grazers such as kangaroos is significant in terms of the types of survival mechanisms that Australian grasses may have, or importantly, may not have developed for heavy grazing by herd animals. Grass species were generally understory species in the interglacial times or in shaded woodlands unlike the large expanses of grassland in East Africa. In eastern Australia, grasses were likely sparingly grazed as part of a mixed herbaceous and grass diet, rather than grazed heavily.

The only large marsupial grazers common in Eastern Australia post 18,000 years BP have been grey kangaroos (*Macropus giganteus*) and red kangaroos (*Macropus rufus)*, with the remainder of the wallabies (other smaller macropods) generally being browsing animals. The browsing marsupials had a mixed diet of shrubs and understorey grasses, rather than solely relying on grass for fodder. Kangaroo mobs are generally small (30 – 100) unlike African migratory grazers (Kaufmann, 1975).

Expansive grazing landscapes and grasses in open woodland or grassland environments are relatively new in south eastern Australia and may only extend back a few hundred thousand years (Strömberg, 2011). Chiefly, most Australian grass species are likely more adapted to light grazing by predominantly mixed browsers and not adapted to pulsed, or even sustained heavy grazing such by African or Asian bovine species.

Summary of African and Australian grazers and grasses

Survival mechanisms, soil profile exploitation (rooting depth), and aggressiveness of African grasses are likely to be different to many Eastern Australian grass species. The coevolution of heavy grazing with grasses in Africa and the absence of heavy grazers in Australia are the likely drivers of these differences.

2.5 African tropical grass pastures in Northern New South Wales

The use of African pasture species in improved pasture systems is relevant to soil formation processes because of their much greater feed production (Harris *et al.*, 2014) and therefore their potential to accelerate biological, physical and chemical processes. In addition, their greater root access to deep subsoils (Murphy *et al.*, 2010; Murphy, 2014; Murphy *et al.*, 2018) offers increased potential for cycling of subsoil elements compared with native pastures. The greater root densities and depths of African pasture grasses may also create a different internal soil hydrology through the creation of a greater number of soil macropores.

The use of African tropical pastures in grazing systems in Northern NSW (north west slopes and plains) has become more common in recent decades with an estimated 400,000 ha of established tropical pastures in 2014 (Harris *et al.*, 2014). The four most commonly used tropical grasses used in north west NSW are listed in Table 2.1. Harris *et al.* (2014) report that tropical pastures are generally planted for an increase in feed production with reported annual yields as high as 19 tons of dry matter per ha (t DM/ha), whereas native grasses traditionally used for grazing have a maximum yield of 9.7 t DM/ha (NSW western slopes and plains data). The water use efficiency of tropical pastures (DM/mm/ha) is greater than that for tropical pastures (Harris *et al.*, 2014). The tropical species are also reported to be very persistent once planted and are a highly valuable feed source. Tropical pastures are also known to be deeper rooted and facilitate both better water capture and deeper moisture storage than native pastures (Murphy *et al.*, 2010).

In terms of soil organic carbon (SOC), sown tropical pasture species may increase SOC in formerly cropped lands. However, in many broad field sampling or trial conditions, native pastures had slightly more SOC or the same SOC as tropical pastures (Young *et al.*, 2005; Wilson *et al.*, 2010; Wilson *et al.*, 2011; Schwenke *et al.*, 2013). A problematic issue with many of these comparisons is that most studies were confined to the top 30 cm of soil, and the time of residence of native pastures as opposed to sown tropical pastures was in question in some cases (Murphy, *pers comm*). Ideally, an absolute timeline from a zero point, (i.e. time since cultivation) should be the same for both native and tropical pastures as much as is practical when making such comparisons, particularly in agricultural trials.

Table 2.1*: Most commonly used tropical grasses of the north west slopes NSW and plains. After Cook et al. (2005).*

Common name	**Scientific name**	**Origin**
Katambora Rhodes Grass	*Chloris gayana* cv. Katambora	Eastern and central Africa
Premier Digit Grass	*Megathyrsus maximus* cv.Premier	All of Africa except arid zone
CONSOL Lovegrass	*Eragrostis curvula* var. *conferta*	Eastern and southern Africa
Bambatsi Panic	*Panicum colouratum* var. *markarikariense*	Predominantly Botswana

Sown tropical grasses are grown with appropriate winter and spring active legumes. As with most pastures, N is limiting and winter active legumes are grown to provide the soil N (Harris *et al.*, 2014). The legumes require P fertilisation to maximise N fixation. In some higher production grazing systems of the Liverpool Plains, graziers take spring soil N measurements and add N fertiliser if winter N fixation by legumes through has been low. Soil N levels are highly correlated with SOC content (Barrow, 1969).

Because tropical pasture grasses have deeper and more aggressive roots, access to deeper soil water and higher above ground fodder and SOC production, their use may represent a sufficiently significant change of land use to initialise soil formation processes in a new direction.

2.6 Soil formation and soil formation processes

There are different philosophical concepts of the processes of soil formation with much of the initial process groups being related to the bureaucratic need to classify and map soils for taxation purposes in Europe (notably Germany) (Bockheim *et al.*, 2005). This section reviews the development of three points of view with regard to soil formation processes and proposes that simplification and inclusion is better than prescription and exclusion of potential processes which may add to our present understanding of soil formation and how static or active it could be in Eastern Australia.

Soil formation factors – a basis for all soil formation

The volume of material written on soil formation and the factors of soil formation is immense, but the basic concepts have remained the same over time indicating that the framework has survived some robust criticism.

Paton (1978) provides a good summary of the five basic soil formation factors, being lithospheric material, topography, the biosphere, climate and time. Jenny (1941) proposed the "Fundamental Soil Equation" for soil formation:

$$S = f\,(cl, o, p, t, ...)$$

where S is the state of the soil (degree of profile development), f means function, cl is climate, o is organic activity, p is parent material and t is time (Jenny, 1941). There is room at the end of the equation for additional factors. The equation is illustrative and does not attempt to imply magnitude of importance of their interactions, nor invoke specific processes. The equation does imply that given sufficient time and consistency, the interaction of these five factors in some way will result in some sort of equilibrium or end point.

Given the changes in climatic and vegetation conditions experienced in Australia through the Quaternary it would appear that neither climate nor vegetation have been stable during this period. The results of multiple phases of vegetation and climate changes may confound the equation by continually resetting the time condition for a process to commence or end. This is consistent with the statement by Johnson *et al.* (1990) (p306), that "soils are open systems and rarely, if ever, achieve equilibrium with their environments." In the case of Eastern Australia, opportunities for soil formation to approach a stable endpoint of soil formation must be limited to very small areas where climate and consequent vegetation swings have had least effect.

Using the factors Jenny (1941) defined, this would imply that the five factors of soil formation are not constant, and this assertion has important implications for this thesis based on the proposal that significant soil profile changes may occur in short time frames driven by land use change. Considering Eastern Australia, in view of the continually changing variables in the equation, soil bodies and soil forming processes are not static, but dynamic and should be responsive to climate and ecological changes.

Traditionally accepted processes of soil formation – Northern Hemisphere

Bockheim *et al.* (2005) provide a summary of the dominant processes involved in soil formation (or soil profile development) with respect to soil classification and soil genesis. These concepts have been developed over a century of soil science research predominantly in the northern hemisphere and constitute a tremendous contribution to the understanding of soil formation and distribution today (Bockheim *et al.*, 2005; Bockheim and Gennadiyev, 2009). These processes are the basis of and intimately linked to genetic soil classifications such as the German Soil Classification (Albrecht *et al.*, 2005), World Reference Base (WRB) (IUSS, 2015) and Soil Taxonomy (USDA, 2014) and reflect the accepted "norms" of soil formation processes.

The processes described in Bockheim and Gennadiyev (2000) are summarised in Table 2.2 with timeframes for various processes from Yaalon (1971). It should be noted that the processes in Table 2.2 work only in the vertical dimension of the solum and mostly from the top down. The concept of additions and depletions through erosion and deposition processes or development from *insitu* weathering of parent material is not raised (Paton *et al.*, 1995). The processes are best suited to an understanding of soil that results from a well-known time zero and an easily understood parent material, such as the retreat of a continental glacier, exposing till and moraine materials on which soils form.

Table 2.2: *Conventional processes of soil formation and brief descriptions after Bockheim et al. (2005). Timeline after Yaalon (1971)*

Process	Brief Description	Timeline
Argilluviation	Movement of clay in the solum. Often resulting in a depleted layer above (the E or $A2_e$ horizon)	Thousands of years
Biological enrichment of base cations	Generally, the enrichment of soil layers with base cations through the action of plants taking up cations and redistributing enriching soil with them through litter/leaf fall and microbiological activity	Ten thousand to a million years
Andisolization	Dominated by amorphous compounds with low bulk density and high Al and Fe. Generally restricted to volcanic ash environments high in fine volcanic glass.	Thousands of years
Paludization	Deep accumulation (> 40 cm) of organic matter.	Hundreds of years

Process	Brief Description	Timeline
Gleization	Presence or development of redox related features such as gleying, mottling and deposition of Fe and Mn compounds.	Hundreds of years (mottle formation or obliteration)
Melanisation	Accumulation of well humified organic matter in the upper 18 cm of the soil.	Thousands of years
Ferralitization	Concentration of Al and Fe and loss of Si as a result of mineral weathering.	Ten thousand to a million years
Podsolization	Process of eluviation of base cations, weathering transformation of Fe and Al compounds and transport of these compounds to a spodic horizon (B_s) where they complex with leached organic matter weathering products. Generally most effective in silty to sandy materials.	Thousands of years
Base cation leaching	Opposite of biological enrichment. Leaching of base cations – eluviation of Ca, Mg, K, and Na from the solum or part of the solum.	Ten thousand to a million years
Vertization	Development of high shrink swell properties such as cracking and slickensides.	No time suggested
Cryoturbation	Permafrost effects of frost stirring caused by freeze thaw effects.	No time suggested
Salinization	The natural accumulation of highly soluble Na, Ca, Mg, and K salts (as chlorides, carbonates, sulphates or bicarbonates) Often confused with human induced soil salinity processes	Hundreds of years
Calcification	Accumulation of secondary soil carbonates and gypsum as $CaCO_3$ or $CaSO_4$.	Thousands of years to tens of thousands of years
Solonization	Or alkalinisation – where soils subjected to salinization are leached and clayey soil materials become highly sodic and a strong alkaline reaction trend develops.	Thousands of years
Solodization	Argilluviation of dispersed clays and development an acid A horizon and a clay enriched B_{tn} horizon	Thousands of years
Silicification	Accumulation of secondary Si from seasonally high water tables.	Ten thousand to a million years
Anthrosolization	The sum of human induced soil formation processes from agriculture and industry – generally referring to constructed soils or soils that have become unrecognisable through land use.	?

Walker and Butler (1983), investigating a chronosequence of alluvium on the Riverine Plain in S NSW, found that 5,000 years are required for some pedological organisation (development of soil fabric and horizonation) to develop in alluvium and 15,000 years for a texture contrast soil profile to form. Similar work on subsoil structure formation in the United States suggested a time frame of 15,000 – 20,000 years (White, 1967). These time frames are within those given in Bockheim and Gennadiyev (2000) and Yaalon (1971). This work indicates that the timelines and processes summarised in Table 2.2 are valid within the region in which they were measured and in at least some parts of Australia.

The prevailing view of staged and process driven soil profile development is often referred to as the "zonal" model of soil formation (Paton, 1978). This model infers that within a geographical and climatically defined zone, distinct soil formation processes will operate and distinct soil types will occur as endpoints of soil formation processes. This is potentially a circular argument essentially proposing that for a given soil type, certain processes must have contributed to soil formation. This is likely the case in closed systems and under relatively static environmental conditions, but may not be universally applicable to open natural soil systems that are exposed to a number of uncontrolled and changing, interacting and even synergistic ecological factors.

The World Reference Base (WRB) (IUSS, 2015) is highly Euro-centric in its origins and carries these genetic soil formation implications to locations beyond Europe where other processes may be significant, or to where similar soils may have evolved despite different processes operating in these environments.

In a potentially polygenetic environment such as Australia, many processes affecting soil formation are likely operating simultaneously or at different times according to long and short term (within decadal) climate extremes. Despite the polygenetic possibilities and confounding timeframes that Australian environments offer, there are clear indicators that the conventional model of soil forming processes and time scales are applicable in some areas.

Criticism of Northern Hemisphere zonal processes (the Paton era)

Paton (1978) and Paton *et al.* (1995) provided a simpler model of soil formation processes unconstrained by Northern Hemisphere concepts of soil formation processes. Paton's interest was in soil formation processes in areas beyond the Northern Hemisphere or which did not conform to the zonal perspective. He critically assessed the assumption that soil formation factors operate only in a vertical dimension and originate solely from the land surface to create genetically related horizons, namely, the A and B horizons which constitute the solum. Paton was critical of the assumption that the formation of a B horizon was a result of the eluviation of fine materials (clays) from the near surface to concentrate as an illuvial subsurface layer. In consideration of many Australian duplex or texture contrast soils, there was no or little evidence to indicate that the clay in B horizons had come from the surface horizons (Paton *et al.*, 1995). However, during the course of soil development of the Urrbrae Red Brown Earth, a period of 30 000 to 100 000 yr, weight loss has been <2% and volume 2.5%. Horizon development has been caused by redistribution of materials (mostly fine clay) not by the formation of clay by weathering (Chittleborough and Oades, 1980; Chittleborough, 1989, 1991; Fitzpatrick and Chittleborough, 2002).

Weathering of parent materials or the weathering of bedrock materials is recognised in the European model (Table 2.2) as a pre-pedological process creating parent material. The creation of saprolite or C horizon material and subsequently soil is shown as an important process in soil formation, though generally operating over a longer time frame than near surface processes. In this way soil can be regarded as being built from both above, below and horizontally rather than from simple vertical processes as implied in the European model.

Paton (1978) and Bishop *et al.* (1980) demostrated that for some areas of Jurassic Sandstone in Sydney, duplex soils were formed by slope wash depositing the A horizons. These soils have a texture contrast and often an A2 or $A2_e$ (E horizon) giving the appearance of being the end product of argilluviation and possibly base cation leaching. The authors noted that there were no cutans (illuviated clay coatings on subsoil aggregates) within B horizons, and that soil materials of the A1 and A2 horizons showed strong evidence of being transported into place over the clay rich B horizon. A2 and $A2_e$ (E) horizons were more likely to be formed by interflow running down slope above a texture contrast rather than vertical eluviation. Paton (1978) and Bishop *et al.* (1980) argued that geomorphic processes were not merely a factor in soil formation (topographic), but should be considered a dominant soil formation process. In addition to erosion and deposition processes they considered that bioturbation (excavation or mixing and sorting by ants, birds, other invertebrates and even tree fall) is intimately involved with the development of texture contrast soil profiles.

Paton (1978) and Paton *et al.* (1995) proposed that some of the factors of soil formation (Jenny, 1941) were also processes of soil formation. Paton *et al.* (1995) also proposed that the climate factor should be considered in terms of the availability of water. Water supply, according to these authors, determines the types and intensities of geomorphic, biological and chemical processes. They proposed that water supply and biology are effectively an on/off switch to pedogenesis and therefore an "initiating factor" whilst lithosphere (parent material) and topography are determinative factors. Temperature was not considered to be as important as water supply.

Paton *et al.* (1995) summarise that lithospheric materials at a site (the determinative factors) require the presence of water and the biosphere (the initiating factors) so that epimorphism (transformations) and near surface processes can operate over time to produce soil material.

Paton's critique of the accepted factors and zonal processes of soil formation provide an interesting perspective on how a scientific, process-oriented dogma may have restricted the study of pedology and the present understanding of soil formation.

The current Australian model

Building on the simple theory of soil formation based on additions, removals, transformations and translocations posed by Simonson (1959), but incorporating surface and zonal processes, McKenzie *et al.* (2004) propose soil formation processes which are outlined in Table 2.3.

The processes posed in Table 2.3 include the surface processes of Paton (1978) and Paton et al. (1995) along with a simplification of some of the vertical zonal processes of the Northern Hemisphere given in Bockheim et al. (2005) and outlined in Table 2.2. McKenzie et al. (2004) were far less specific in their process descriptions than Bockheim and Gennadiyev (2000) (Table 2.2). The move towards simplification and more general statements on processes may be helpful in moving away from potentially dogma-directed research. The timelines for processes given in McKenzie et al. (2004) are very similar to those of Yaalon (1971). McKenzie *et al.* (2004) also comment on early thoughts of Verboom's theory or phytotaria (plant communities that can generate soil specific to their own needs) (Verboom and Pate, 2013) discussed later in this section.

Table 2.3*: Soil formation processes after McKenzie et al. (2004)*

Process	Brief description
Soil production, erosion and deposition	Soil production in this sense is used to describe the weathering of bedrock to form new soil material, then the balance of erosion and deposition processes that move these materials in the landscape
Formation of clay minerals	Chemical weathering of rocks and minerals to form clays
Textural differentiation	Development of texture contrast soils through faunal activity, clay movement, differential weathering or sedimentary processes

Process	Brief description
Soil organic matter accumulation	As title states
Translocation of organic matter, Fe and Al in leaching environments	Equivalent to podsolization as described in Table 2.2
Accumulation of soluble salts	Including calcium carbonate, gypsum and halite. Equivalent to calcification and salinization above
Iron enrichment	Transformation of soluble Fe^{2+} iron to insoluble Fe^{3+} compounds through dissolution and wetting and drying (Redox) cycles
Wet soils and drainage	Equivalent to gleization process above, but specifically including both groundwater and surface waters, the development of soils in swamp environments as well as those soils with impeded drainage and A2 or $A2_e$ horizons
Cracking clays and gilgai development	Vertisation as described in Table 2.2

Recent international revision of some processes

A revision of texture contrast soil formation processes in the USA was recently published by Bockheim (2016), who examined the USDA soil data set for texture contrast soils. Bockheim (2016) found that the processes which create texture contrast soils in the USA likely include:

1. Sedimentary layering – where coarser textured topsoils have been deposited on top of finer textured subsoils (or buried soils);
2. A lithological discontinuity - where water “hangs” in the $A2_e$ horizon, causing greater weathering in the B horizon than the $A2_e$ horizon;
3. Hydromorphism – where lateral subsurface flow or interflow in the $A2_e$ horizon caused by a low hydraulic conductivity B horizon and slope removes clayey materials from the $A2_e$;
4. Vertical translocation – as described above where argilluviation of clay materials downwards occurs to concentrate clay in the B horizon;
5. Lateral translocation – similar to sedimentary layering but is more dominated by slope wash on slopes of > 10%; and

6. Bioturbation where clay is preferentially moved upwards by termites, although traditionally it has been maintained by Paton (1978) and Paton *et al.* (1995) that coarse materials were preferentially moved upwards by ants or termites.

This revision is important as it indicates an inclusion of geomorphic and biogenetic processes that were previously mostly excluded from US and European models of soil formation. This revised model is aligned with or inclusive of models of soil formation proposed by Paton (1978), Bishop *et al.* (1980) and Paton *et al.* (1995), and recognises that many texture contrast soils are a result of deposition of coarser material over existing clayey soil material.

Other processes – plants creating soil

Pate and Verboom (2009), Verboom *et al.* (2010) and Verboom and Pate (2013) demonstrated that some mallee eucalypt species in Western Australia are capable of creating clay from its dissolved chemical components in the groundwater and soil solution of a quartz rich sand dune, to form a Sodosol soil. In terms of the conventional understanding of soil formation processes this finding is unusual and warrants consideration in the frame of reviewing and adding to conventional concepts of soil formation. Pate and Verboom (2009) call their concept “Phytotarium” and suggest that the process occurs in “instances where targeted changes in soil profile characteristics appear to optimise control of resources of water and nutrients by the principle plant players involved” (Pate and Verboom, 2009) (p 673).

Pate and Verboom (2009) determined that formation of dense, sodic clay B horizons, with associated carbonates and silicates had occurred post Aboriginal occupation of Australia, as artefacts were present in the clay layer. Substantial pH increases from strongly acid to very strongly alkaline subsoils were associated with the development of plant-generated clays. The researchers proposed that this type of soil formation process, namely neoformation of clays by a “phytotarium” (group of specialised plants) probably occurs within the life cycle of individual trees.

The proposed mechanisms of B horizon construction by a phytotarium is essentially that certain eucalypts and other Australian tree species utilise “xylem-mediated transport of required minerals scavenged from lower regions of a profile and employ these for bioprecipitation around specialised lateral roots, and involve dedicated synthesising activities of microorganisms” (Verboom, pers. com.). The dominant factor in soil formation in this case is biological. The time frames given for conventionally understood soil formation processes in Yaalon (1971) imply that B horizon formation, and the accumulation of carbonates should occur in the order of hundreds of thousands of years.

Prior to the work of Pate and Verboom (2009), clayey B horizon construction as a result of biological epimorphism had not been considered before and has not been accounted for in the generally accepted zonal model as summarised by Bockheim *et al.* (2005) (2.1) nor in Paton *et al.* (1995).

The suggestion of an active involvement of vegetation in texture contrast soil formation from sand is unique and highly significant. It demonstrates that soil formation may take an unexpected pathway to arrive at given endpoint; in this case a Sodosol. Pate and Verboom (2009), Verboom *et al.* (2010) and Verboom and Pate (2013) demonstrated that soils which are conventionally understood to form over a timescale of tens of thousands to hundreds of thousands of years (Yaalon, 1971; Bockheim *et al.*, 2005) can develop in decades to centuries.

2.7 Examples of rapid soil development (including human activity)

Aside from the Verboom and Pate (2013) example, there are other examples of soil formation that appear to be more rapid than conventionally accepted, and which may have followed conventionally understood soil formation processes at an accelerated rate. Despite conventional timelines for soil formation postulated in Yaalon (1971), rapid soil change or rapid changes in the direction of soil formation processes have been observed in many different locations in the world. The most obvious examples are: (i) Podsols (Podosols ASC),(ii) Hydrosols, soils which change quickly under rice cultivation, when new land is converted to flooded paddy fields (Kölbl *et al.*, 2014), (iii) soils resulting from mine rehabilitation projects and acid sulfate soils (Fanning *et al.*, 2017).

Podsolization following disturbance or on fresh sand deposits

Bormann *et al.* (1995) demonstrated through investigation of a 350 year chronosequence, that Podsols horizons can develop in south eastern Alaskan forests in less than 150 years following a tree fall and associated windrow disturbance of the soil. This was achieved by investigating tree falls of known age where the soil had been mixed, then tracing the downward movement of carbon, Al and Fe, which returned to full control site levels and positions in the profile after 350 years.

In a post sand mining rehabilitation chronosequence, Prosser and Roseby (1995) found that re-formation of a Podsol's diagnostic horizons was complete after 17 years. The process of leaching of Fe, Al and organic matter from disturbed material resulted in the development of an $A2_e$ horizon (E horizon), and the accumulation of leached materials as a B_h horizon had occurred at the top of a water table within this period. This is in stark contrast to Yaalon (1971), who ranks these processes as taking thousands of years. The thickness of the B_h horizon may continue to grow with time, but the establishment of a clear Podsol soil profile was clear at 17 years in the study by Prosser and Roseby (1995).

Verboom and Pate (2013) further demonstrated that some members of the Proteaceae family of plants (including some *Banksia* species) modify clean deep sands to form Podsols within the life span of the plant (< 100 years).

Examples from the rice industry

Studies on tropical and subtropical rice soils are important with respect to soil formation processes as these studies illustrate the consequences of interrupting the soil water regime by the conversion of a natural *in situ* soil to a flooded rice paddy. The most obvious impacts of conversion from any type of soil to rice paddy is the use of long term flood irrigation which induces anaerobic conditions, as well as high level disturbance due to the practices of flooding and puddling (or wet cultivation) of the soil. Soil puddling is undertaken to reduce water loss through vertical infiltration. Kirchhof *et al.* (2000) established that increased puddling regimes, or more intensive wet cultivation, had significant yield reductions on some dry season legume crops grown in rotation with rice. Yield decline of dryland legumes was related to topsoil and upper subsoil structural decline, which interfered with root ingress to the soil as well as seedling establishment due to overly wet conditions.

Kölbl *et al.* (2014) studied a chronosequence progression through Fluvisols to Cambisols then Anthrosols (IUSS, 2015) in marshland soils in the Zhejiang province of eastern China. The paddy sites showed clear and continued changes over 2,000 years. The paddy soils showed a staged change through the chronosequence. The stages observed were:

1. Desalinisation and loss of stratification of sediments and the development of a structured B horizon (therefore becoming a Cambisol within 0 –50 years);
2. Decalcification, decrease in pH in topsoils (100 – 300 years);
3. Decalcification, decrease in pH in subsoils (100 – 700 years); and
4. Fe redistribution and crystallisation, appearance of reducing conditions in subsoils evidenced through colour changes and mottles (700 – 2,000 years).

Organic carbon significantly and consistently increased in topsoils in paddy fields over the 2,000 year chronosequence.

Kölbl *et al.* (2014) concluded that the types of formation processes which should occur according to the model in Table 2.1 did happen and flowed through a logical progression as described in the World Reference Base (WRB) Soil Classification (IUSS, 2015). Although their results appear to confirm this progression, there is potential circularity in the conclusions which are based on processes. The use of the word rapid in the title is probably relative to conventional understanding but is likely to be more common in such man made environments than previously thought.

Moormann and Van Breemen (1978) further evaluated soil formation processes in flood irrigated rice lands. Of note are their comments on the development of redoximorphic materials in these soils. Essentially flooding causes conditions which generate soluble Mn^{2+} and Fe^{2+} which increase to high levels in soil water within a few weeks of saturation. Drying of the soil results in the reoxidation of reduced Fe and Mn and the formation of amorphous Fe and Mn oxide and oxyhydroxide minerals. Fe and Mn compounds also reoxidise near rice roots which conduct oxygen from the atmosphere. In all but the lower landscape positions, Fe and Mn also reoxidise at the base of the puddled layer because the underlying soil is often aerobic. Repeated wetting and drying cycles result in the accumulation of mottles and segregations. Mottles in soil aggregates generally develop as ferrous Fe (yellow and gley) migrates from the inside of peds to the outer zone of peds where it becomes relatively immobile ferric hydroxide (red).

Moormann and Van Breemen (1978) also note that the accumulation of both Fe and Mn hydroxides is maximised at moisture boundaries, which are found where a saturated $A2_e$ horizon (E horizon) meets a B horizon that is mostly unsaturated, which occurs at both field sites for this thesis. The gradient that is created in this instance results in the transport of reduced Fe and Mn to the oxidised zone in the top of the B horizon. A similar process occurs at the base of many soils where there is a periodically high groundwater table and Fe and Mn hydroxides form at the interface of wetter and drier soils. In terms of Table 2.1 timeframes, Moormann and Van Breemen (1978) suggest that some of the mottling effects of the redox reactions can occur in days rather than hundreds of years. In work unrelated to the rice industry, Bourman (1993), Phillips (2000) and Banks (1998) showed that the generation of ferricretes (or cemented Fe, Mn and Al hydroxides), thought to take place in geological time, can occur rapidly within a few years to a few decades when peaty swamp environments containing large quantities of reduced cations are drained and rapidly oxidised. Studies in Australia show that soil phosphorus is rapidly immobilised under flood irrigated rice (Willett and Higgins, 1978; Willett *et al.*, 1978).

In summary, the introduction of flood irrigated rice in many soils can result in a re-setting of soil formation processes through the introduction of large soil structure changes, leaching, and intense redox phases which change soil properties. An understanding of soil changes under flooded rice cultivation has the potential to contribute to the study of rapid soil formation in Eastern Australia. Substantial changes in soil water regime may have occurred in response to clearing of lands. In addition, long periods of both intense saturation and drying outside of seasonal frameworks are associated with Australia's ENSO driven climate (Ummenhofer *et al.*, 2011). Rainfall and temperature extremes within years such as occur in eastern Australia may have a highly significant impact on both the speed and direction of soil formation processes.

Mining industry

The mining industry in Australia and overseas is a useful place to look for indications of rapid soil formation. The aim of site rehabilitation after open cast mining is to cover mine spoil in a relatively stable form to allow revegetation and prevent erosion or to attempt to recreate the pre-mining surface conditions. Consistent with previous observations, rapid development of a Podsol soil profile in a post sand mining environment is possible within 17 years under the conditions described in Prosser and Roseby (1995).

Courtney *et al.* (2013) showed that bauxite mining residues (red mud) which are highly sodic and alkaline can be treated to improve aggregate stability with a much lower pH by the application of large quantities of gypsum and compost. The reported time for formation of soil structures was one year. This type of site remediation amounts to a complete adjustment of a soil or mine spoil materials chemistry and could be considered engineered soil formation rather than accelerated natural soil formation.

Ward (2000) showed that after stripping and re-laying of soil materials (mostly gravelly sands) following bauxite mining in Western Australia, the soil chemistry returned to pre-mining condition within 8.5 years. The rapidity of this "restoration" may well be due to the coarse nature of the material used and the generally depauperate soil fertility of the soils prior to mining. Banning *et al.* (2008), reported similar findings with return of organic matter and N cycling to a near normal state for the same region of bauxite mine rehabilitation. It should be noted that in the studies by both Ward (2000) and Banning *et al.* (2008), the soil had been stripped from mining land, stored and replaced in a rough topsoil subsoil order, so they may not represent true soil formation studies, but rather investigations of soil recovery from disturbance.

Schwenke *et al.* (2000a, 2000b) worked on the rehabilitation and soil formation occurring at Weipa in north Queensland. The original soils at the site were Brown Kandosols (ASC) that were significantly more texturally and structurally complex than the sands of Ward (2000) and Banning *et al.* (2008). Soil chemistry (organic C and total N), bulk density (BD) and biological activity were compared for different rehabilitation techniques, and against an undisturbed, unmined area (Schwenke *et al.*, 2000a, 2000b). A chronosequence of rehabilitation was subsequently investigated. These studies found that soil profile reconstruction and success in rebuilding soil structure and organic matter levels to near "natural levels" was highly variable. Pastures were generally the most efficient and rapid generators of organic matter in the first few years following soil rehabilitation. Soil organic C and total N and soil structure were still improving at the end of a 22 year chronosequence, but remained less than that exhibited under native vegetation. A limitation of the work by Schwenke *et al.* (2000a) and Schwenke *et al.* (2000b) is that all research was conducted on the top 20 cm of soil, and no account was given concerning the status of B horizon material. As with Ward (2000) and Banning *et al.* (2008) above, the results of by Schwenke *et al.* (2000a, 2000b) may be considered representative of recovery from disturbance study rather than a study of soil formation.

Sokolov *et al.* (2015) found that non-rehabilitated coal mine tailings in Siberia showed signs of soil formation, predominantly through the accumulation of organic matter. A 30 year chronosequence study of five mine sites, ranging in location from humid to very arid environments, showed the continued accumulation of organic matter on waste rock material that was originally < 30% soil sized particles. Resultant proto soils were called "Embryozems". Sokolov *et al.* (2015) further demonstrated some breakdown of waste rock material into silt-sized aggregates. Following the model of soil formation given in Table 2.2, the authors then proposed the type of soil expected to form on the tailings materials through the eluviation of clay materials. In terms of comparison with Australian conditions, this work has little relevance to the soil formation processes that may occur rapidly in Australia. The dominant feature of soil formation in the work of Sokolov *et al.* (2015) was low temperature, allowing organic matter accumulation even in arid conditions.

Silva *et al.* (2013) studied a 14 year chronosequence at a rehabilitated gravel quarry in Brazil. Original soils were Cambisols. Re-shaping mixed subsoil and topsoil materials and topdressing with 100 t/ha of sewage sludge was the pre-treatment to observation of vegetation and SOC recovery. The results showed a high level of soil SOC accumulation under non-native invasive pastures. Much of the study by Silva *et al.* (2013) focussed on vegetation and only had soil sampling in 5 cm increments to 20 cm. The study concludes that "soil-plant feedbacks regulate the amount of available resources, determining successional trajectories and alternative stable equilibria” (Silva *et al.*, 2013)(p1345). Although this statement was made with regard to vegetation communities, it could equally apply to soil formation processes, since many of them are plant-controlled.

In terms of soil formation, the addition of large quantities of sewage sludge to the reformed mine site in the work of Silva *et al.* (2013) renders comparison with soil formation in unamended native soils difficult.

Acid Sulfate Soils

Acid sulfate soils, generally coastal hydrosols, contain reduced sulphur minerals which stay reduced whilst the water table is stable which keeps the minerals in a reduced state. Drainage of low lying coastal lands with potential acid sulfate soils can rapidly oxidise these minerals, within hours to days, creating an abundance of sulphuric acid, which rapidly lowers soil pH can destroy both organic materials and clay minerals in the process (Fanning *et al.*, 2017)

2.8 Rapid Ferrosol formation on Late Quaternary and Holocene basalts of South Australia

Pillans (1997) measured reverse polarity magnetism in hematite preserved in Ferrosol subsoils, which correlates to world polar magnetism reversals, and showed that soil formation rates of Ferrosols in north Queensland has been very slow (0.3 m per million years). The Mt Gambier region in south eastern South Australia has several examples of Late Quaternary to Holocene basalts, with the youngest of these being extruded approximately 5,000 years BP (Van Otterloo and Cas, 2013). Soil formation rates on Holocene and late Quaternary basalts at Mt Gambier are not reported, but by inference, there are Ferrosols developed on parent material ages between 6,000 and 100,000 years, which are deeper than 30 cm (Lowe, 1992). This implies that either soil formation rates reported by Pillans (1997) are incorrect or that the rates of soil formation vary greatly according to local conditions.

The Australian Ferrosol can be classified as either a Nitisol or Ferralsol in the World Reference Base classification (IUSS, 2015), which states that these soils form under tropical conditions. The climate of Mt Gambier is temperate and has strong cool oceanic influences, and is generally considered Mediterranean (Webb *et al.*, 2010). Ferrosols formed in the Mt Gambier region on very young basalts, therefore have formed under significantly different conditions to those described in IUSS (2015), and/or have formed at rates which are far more rapid than any published estimates or measurements. This indicates that local conditions may be more important than generalised processes.

2.9 Land use and soil physical changes

As shown in Section 2.3, the rapid increase of erosion in response to land use change upon European settlement indicates that soil infiltration rates were reduced and that run-off coefficients of catchments increased significantly. Change of land use from native forest to pasture generally results in increased soil bulk density (BD), decreased soil porosity, decreased hydraulic conductivity and lower overall plant-available moisture. This trend is common worldwide (Greenwood *et al.*, 1998; Greenwood and McKenzie, 2001; Piñeiro *et al.*, 2010; de Souza Braz *et al.*, 2013) and increased BD has been shown in the vicinity of the slopes and plains research site for this study by Young *et al.* (2005). Land use change to cultivation can further increase BD and decrease soil structural integrity and water holding capacity (Bridge and Bell, 1994).

Wilson *et al.* (2009) found that non-basaltic soils with sown or improved pastures consistently had higher BD than those under native pastures which in turn had higher BD than forested or wooded areas. There was little difference between native and sown pastures on basaltic soils in terms of BD. There is much evidence to indicate that clearing land and subsequent cultivation or grazing can reduce the qualities of a soil’s physical attributes in terms of porosity, density, drainage, and water holding capacity. Conversely, these attributes can be changed in a direction more favourable to agricultural or forestry production. All of these land uses affect water infiltration and availability, root growth and the redox state of the soil.

2.10 Potential soil chemistry change associated with land use change

Soil chemistry and field physical properties are factors that facilitate definition of soil types in many soil classifications. Most classifications use some soil chemical characteristics in the interpretation of soil group. The soil Orders of Isbell and NCST (2016) are defined by a mixture of field assessable characteristics and/or chemistry. A Sodosol, for example, must have a sharp textural contrast between the A and upper B horizons, have neutral to alkaline lower B horizons, and have an exchangeable sodium percentage (ESP) of at least 6 in the upper 20 cm of the dominant B horizon. This very large Order of soils within the ASC are known for their impeded drainage as a result of their dispersive upper B horizons (McKenzie *et al.*, 2004).

The defining feature differentiating a Chromosol from a Sodosol is upper B horizon sodicity (ESP> 6 sodic, ESP < 6 non-sodic) Isbell and NCST (2016). In soils where the total CEC of the upper B horizon is very low, small changes in cation balance in soils of either of these orders could easily change the soil order to which they belong and hence their hydrological behaviour, which is of importance to the land user. In this case, the simple addition of lime or superphosphate may provide sufficient calcium to the cation exchange suite to transform a Sodosol into a Chromosol. Conversely, the removal of calcium by plants from this layer may cause the reverse transformation of a Chromosol into a Sodosol.

Soil organic carbon and land use

The accumulation or depletion of organic matter is determined by the speed of biological processes together with temperature and water availability. It follows that soil formation processes in part can be influenced by the amount of organic C in the soil or by a change in the amount of organic C.

Interest in SOC, both as an element of soil health, and as a potential store of atmospheric C, has become a focus of intense research in the recent decades. Total soil carbon response to land use change will be the focus of this section of the review. Soil organic matter is simply defined in McKenzie *et al.* (2004) (p16) as: “the sum of all natural and thermally altered biologically derived organic materials found living in the soil”.

Fine organic matter (humus), with very high cation exchange capacity and a high water holding capacity, can function in a similar way to clay and tends to have a relatively long residence time in soil. Particulate organic matter is still generally identifiable from its origins (grasses/litter fall) and quantities in soil can fluctuate relatively quickly (on a timescale of years). Inert organic matter, is generally charcoal and has longest residence time of all organic carbon fractions in soil (McKenzie *et al.*, 2004).

The SOC in the humus fraction has a crucial role in stabilising soil aggregates through bridging of silt sized particles as well as assisting in the aggregation of clays (Six *et al.*, 2002). Syswerda *et al.* (2011) showed that there is potential under certain conditions for agricultural soils to store C in deep soil layers, where even more volatile organic C can have a lengthy residence time.

SOC can change relatively rapidly in response to land use change and to increases or decreases in available water. As such, SOC is a good indicator of both initial and prolonged effects of land use change on soil. In both natural and agricultural systems, SOC is generally highest in low silica (clayey) environments and lowest in silica rich (sandy) environments, and most favoured by a cooler, moist climate (Badgery *et al.*, 2013). Conant *et al.* (2001) showed that worldwide, managed grasslands can have greater SOC than native forests and woodlands. This statement does not take into account the loss of C in above ground biomass removed in the case of clearing forested or wooded lands for grazing. In the study by Conant *et al.* (2001), the greatest increases in soil C were observed in what were formerly woodland or grassland biomes, rather than more arid or moist biomes.

In eastern Australia, studies show a significant soil C difference between grazing and cropping systems (Cotching, 2012). Comparisons of major land use and land management systems have been conducted across a wide range of soils and climates in Australia on a variety of dominant soil types. Comparisons of cropping and grazing systems for three dominant soil types in Tasmania found that SOC was higher in grazing systems than in cropping systems for each particular soil type (Cotching *et al.*, 2001, 2002b; Cotching *et al.*, 2002a). Cotching (2012), using a large existing data set from Tasmania, found that of the total soil C in the top 1m of soil, 66% was in the top 30 cm in agricultural soils, compared with 50% in native forests.

Orgill *et al.* (2014), working in the Monaro and Boorowa region of NSW, measured SOC to 70 cm in soils under native pasture and improved pastures across a range of soils. They found no difference between the amount of SOC between native and improved pastures, but found that soils on basaltic geologies had a far greater portion of their total C in the 30 – 70 cm depth range. For basalts, Orgill *et al.* (2014) reported 43% of total profile SOC in the 30 – 70 cm of the profile whereas soils in granitic geologies had only 28% of total profile SOC in 30 – 70 cm region. The authors surmised that these values indicate that much C accounting, which concentrates on the upper 30 cm of the soil profile, may significantly underestimate SOC stocks and potential sequestration.

In north west NSW, much work has been done on comparing land use and SOC over a large range of climates and soil types. Young *et al.* (2005), Wilson *et al.* (2008), Wilson *et al.* (2010) and Wilson *et al.* (2011) consistently observed that for the top 20 – 30 cm of soil, those soils under cropping systems contained less SOC than soils under grazing systems and that grazing system soils had less SOC than soils under native trees (woodland or forest remnants).

Young *et al.* (2005) also found significant subsoil differences in SOC to 100 cm depth in Vertosols and Red Chromosols, when comparing SOC of woodland soils with cropping soils. When comparing pasture soils with woodland or pasture soils with cropping soils, these subsoil differences became far less significant across a range of sites. In similar comparisons of tropical pastures and native pastures, Schwenke *et al.* (2013) found that significant increases in SOC had occurred in Vertosols in the top 10 cm of soil, but that no such difference was found for Chromosols.

Barrow (1969) demonstrated that SOC (in the top 12.5 cm) had increased 50% from a virgin (uncleared) state following 33 years of managed and fertilised improved pasture use in dairy farms in Western Australia. There was concomitant enrichment with Ca in the cation exchange suite. Barrow's (1969) study on coastal sand plain sands is interesting as it has some parallels with the north west NSW slopes and plains research site for the present study at Towri, which has sandy topsoils up to 60 cm thick.

Tree plantation measurements on SOC generally conclude that SOC increases associated with plantations are slow and often do not reach native forest levels for hundreds of years (Turner and Lambert, 2000; Vesterdal *et al.*, 2002; Richards *et al.*, 2009). Some studies indicate that introduced grasslands or pastures accumulate SOC more rapidly and effectively than agroforestry (Davis and Condron, 2002; Richards *et al.*, 2009).

Incorporation of organic matter through subsoil manure application has been shown to be an effective way to offset the effects of sodicity and increase plant root penetration of dense subsoils (Gill *et al.*, 2009), although Gourley and Sale (2014) demonstrate that this does not always result in higher pasture or crop production. This may be a reflection of the root size, structure and root strength of the plant-based production systems investigated.

Large changes in SOC are possible with change in land use. Gain or loss of SOC and related soil structural effects or effects on cation exchange capacity could affect soil formation processes and directions.

Soil chemical elements and cycling

An understanding of chemical balances or imbalances form parts of many soil classification systems. The need to classify soil to understand a given soil's qualities and limitations for human uses such as engineering and agriculture are the reason for this focus. Alteration of soil chemistry is known to facilitate physical changes in soil. Examples of how soil chemistry change through land use change may result in rapid soil type changes are discussed below.

Sodicity impacts on soil

Soil sodicity has strong implications for soil structure and drainage. McKenzie *et al.* (1993) define a sodic soil as having an exchangeable sodium percentage (ESP) of greater than 5. Isbell and NCST (2016) define a sodic soil material as having an ESP greater than 6 and strongly sodic soil material as having an ESP greater than 15. The USDA definition of sodicity is an ESP greater than 15 (Richards, 1954).

Although these threshold numbers vary slightly between different classifications and interpretations of soil chemistry, sodicity is diagnosed to indicate that a soil may be dispersive when wet, set hard when dry, and have soil structural limitations for cultivation. Consequently, many soil surveys refer to sodicity and dispersion in the same category of limitation (Banks, 1995, 1998, 2001). Banks (1995) described Grey Vertosols in the bed of Goran Lake near Gunnedah with ESP greater than 40. However, these soils generally did not disperse. McKenzie *et al.* (1993) similarly found that ESP is often a very poor indicator of dispersion and that soils with ESP < 6 can disperse. They determined that soil dispersion is also partially dependent on the electrolyte composition of soil water and is also affected by land management. Hulugalle and Finlay (2003) proposed that electrochemical stability index (ESI) or the ratio of electrical conductivity ($EC_{1:5}$) of soil to ESP is a better chemical measure to predict soil dispersion.

Some soils may become increasingly sodic as organic matter is depleted, were the physical amount of sodium doesn't change but the proportion relative to other cations does. Organic matter generally increases the overall non-sodium cation component of CEC (So and Aylmore, 1993). In the case of light textured Sodosols, such as are found at Towri (research site for this thesis), sandy clay upper B horizons have a total CEC of 2 – 10 cmol(+)/kg (which is very low), and an ESP average of 20 – 40%. Although the ESP is a high percentage, it is equivalent to only 0.4 – 4.0 cmol(+)/kg) and the net quantity of sodium equivalents is very low. As discussed above, small changes in the cation exchange suite that increase divalent cations such as Ca^{2+} (e.g. as a result of liming or P fertilisation), or an increase in SOC which encourages aggregation can reduce the potentially dispersive impact of exchangeable sodium.

A common agronomic means to offset dispersive effects of sodicity is the addition of Ca ions through the application of gypsum or lime (Hazelton and Murphy, 2007). The resultant change in cation balance causes previously sodic soil materials to remain flocculated instead of dispersing when wetted. The practical effect of these ameliorants is often temporary with soils remaining non-dispersive whilst gypsum and/or calcium carbonate are acting as an electrolyte, both increasing soil electrical conductivity (EC) and displacing sodium with calcium, promoting soil structural stability. The effect generally declines with time as the electrolyte leaches down the profile.

Magnesic soils are soils where dispersion is caused by an overabundance of Mg relative to Ca. They are defined by Isbell and NCST (2016) as having a Ca/Mg ratio < 0.1. These soils are generally ameliorated in the same way as sodic soils and have similar management and ameliorant requirements in agriculture. Soils diagnosed as being Magnesic do not always disperse.

Calcium

Calcium has many roles to play in plant nutrition, and creation or stabilisation of soil structure. Almost all plants form relatively insoluble Ca oxalate crystals in their leaves which are then cycled through litter to topsoils. Calcium oxalate is recognised as an important mode of Ca replenishment of topsoil materials, particularly in forest studies, where trees have been shown to move Ca from deep subsoil layers within the root zone to the surface through leaf and stem fall (Cailleau *et al.*, 2014; Dauer and Perakis, 2014; Crowther *et al.*, 2015). The contribution to Ca cycling by different grass species is variable (Jones and Ford, 1972). Some species of introduced tropical pastures are known to have high calcium oxalate contents and are potentially toxic to livestock (Harris *et al.*, 2014). Although Ca oxalate crystals are relatively insoluble, bacterial processes and decomposition of organic matter, or highly acidic topsoils can break down Ca oxalate and return Ca^{2+} ions to the soil (Dijkstra and Smits, 2002).

The native topsoils and upper subsoils at the Towri site in this thesis have relatively low Ca in their CEC suite, yet often have a large quantity of stored Ca in the form of carbonate (pedogenic or bedrock weathered lime) in deeper subsoil horizons (Banks, 2005; Pengelly, 2009). This is presumably in response to greater cation leaching relative to calcium cycling over time.

Plant based adjustment of the Na - Ca balance

Johnson *et al.* (1990) propose profile rejuvenation (or “youthering”) as the process by which traditionally held soil formation processes may be reversed. For example, increasing inputs of base rich organic matter (from high above ground production) or introducing more divalent cations to a system (through deeper access to and mobilisation of soil Ca stores) may reverse the processes which create a leached soil profile. The introduction to or removal of dominant plant species from an ecosystem may have an impact on the type of definitive soil chemical processes operating at a site.

This can be understood as a loss or gain of Ca associated with the architecture and rooting depth of new dominant plants in an area.

2.11 Potential cumulative effects of land use change on physical and chemical attributes

The functions of soil organic matter, soil physical structures and soil chemistry can be shown to be complex with feedback pathways. An increase in organic matter can improve soil structure which favours plant growth which in turn affects soil structure and promotes increased root exploration of the soil mass. Increased root exploration allows more water to penetrate which in turn allows more root penetration and greater organic matter accumulation. Changes in sodicity can influence all of the above attributes of water penetration, soil structure, root access and organic matter. These feedback mechanisms can push soil formation processes over “threshold” values and into new directions (Johnson *et al.*, 1990). The introduction or removal of dominant plant species may represent an aggressive change in the direction or equilibrium of soil formation, and if the change is significant enough, then a cascade of the pedogenic feedback processes of Johnson *et al.* (1990) may come into play.

Basic soil chemistry changes, such as described above, may be facilitated by the simple process of removing a factor which is maintaining soil equilibrium. Simple removal of the physical cover and microclimate created by tall, moist forest and associated large root porosity may result in a cascade of soil changes if the soil is perched near a threshold that is determined by the conditions created by the forest.

2.12 Discussion and synthesis of literature review

The concepts of soil formation and development that dominate the literature have grown from the pragmatic need to classify soils for mapping purposes for environmental, agricultural or administrative reasons. The use of a logical genetic soil classification system from Europe (World Reference Base or WRB) (IUSS, 2015) or North America (Soil Taxonomy) (USDA, 2014) has led to a certain degree of circularity of understanding soil formation and has come under some criticism for this reason (Paton *et al.*, 1995). The WRB (IUSS, 2015) particularly appears to have developed to a stage where the presence of certain soils is always attributed to a set of fixed processes and environmental conditions which operate in a prescribed timeframe.

The recent revision of texture contrast soil formation in the USA by Bockheim (2016) shows that the concepts of soil formation and development processes are becoming more open-ended in regions where these more traditional views have dominated. In the light of unusual and unexpected findings, such as those of Verboom and Pate (2013), and other examples of relatively rapid soil formation related to human or natural activity, it is necessary for soil formation process and timeline concepts to be viewed as more open and flexible than conventionally believed.

The differing history of soil classification in Australia has generally avoided intrinsic soil genetics. The Australian Soil Classification (ASC) and its predecessor the Factual Key (Northcote, 1979) were not genetic soil classifications by design, and used field soil profile characteristics (textures and texture contrasts, horizonation, colour pH, surface and site conditions and simple chemistry) to classify soils. For communication purposes, these classifications both allow a soil to be named and something of its characteristics to be understood from that name. The ASC and Factual Key classifications allow for soils which may have formed by different processes to arrive at the same endpoint. The endpoint or soil Order yields a classification that is practical in terms of pragmatic demands on a soil classification, such as providing an understanding of agricultural, engineering or environmental qualities or limitations of a soil.

Paton *et al.* (1995) suggest that an overriding factor in soil development is water, its availability and how it interacts with the other factors of soil formation. Water has not been specifically addressed herein, yet it provides the driver for all of the above processes of soil development or formation. Water (and native nutrient) availability determines which plants grow, how plants and roots grow, dissolves solutes, allows the building or destruction of structure and organic matter, and drives the redox reactions that precipitate or dissolve soil components. Water availability is likely to moderate the speed at which most soil development processes operate and determine which direction these processes with take to form a particular soil. Parent material clearly plays a role in the endpoint for a given soil, but the same parent materials can produce different soils depending on water availability. For example, basalts producing Vertosols or Ferrosols, or sandstones producing Kandosols or texture contrast soils like Sodosols under different moisture regimes (Gray and Murphy, 1999).

The eastern Australian environment and vegetation have been highly unstable through the Quaternary period and relatively modern climatic stability has really occurred only over the past 6,000 – 12,000 years. The stability is but relative and still dominated by short term drought and flood regimes. In terms of vegetation, stability has likely been achieved only in the past 6,000 years (Kershaw *et al.*, 1991; Crowley, 1994). This is important in terms of the availability of water and the nature and potential of the biota to drive soil formation processes. It is also likely to be the reason that Gray and

Murphy (1999) found that most of non-arid eastern Australia has soil ages averaging only thousands to hundreds of thousands of years, rather than millions.

Because of the timeframes involved in studying soil change, agricultural trials which have an average life of five years in Australia are generally not suitable to observe potentially deep or large whole profile soil changes. Studies of pedological development and processes generally use a space for time substitute in the form of chronosequences (Chittleborough *et al.*, 1984; Busacca, 1987; Pillans, 1997; Huggett, 1998; Schwenke *et al.*, 2000a; Sauer, 2015). Although chronosequences can offer limited replication opportunities, paired sites with well-known site history (either in human terms or geological time) are often the only alternative available for such studies. For example, the effect of long-term farming on chemical properties (e.g. cation exchange capacity, pH and soil solution composition), organic carbon content, physical properties (e.g. aggregate stability) was investigated using contrasting soils from 4 paired sites in the Mid North of South Australia by Naidu *et al.* (1996) At all the paired sites the undisturbed profiles still had native Eucalyptus woodland. The adjacent farmed profiles have been under a cropping rotation mainly with wheat-barley-grain legume for at least 50 years. Two sites were cleared earlier this century but have been under grassland and grazed at irregular intervals for at least 30 years. The dispersion index, a measure of aggregate stability, increased with farming, indicating an enhanced sensitivity of these sodic duplex soils to dispersion. The greater stability of the undisturbed soils was attributed to the stabilising effect of soil organic matter.

Similarly, Parker and Chartres (1983) studied the effects of changing land use on morphological, physical, and chemical properties of “red podzolic soils under pasture and adjacent / paired forest from four soil catenas north-west of Sydney. They reported that the pasture soils, which had lower organic matter contents also had reduced aggregate stability in the A horizons compared with their forest counterparts. The use of paired sites or simple chronosequences is widespread in making comparisons of cultivated lands or pastured lands with virgin (uncleared) lands (Blank and Fosberg, 1989; Piñeiro *et al.*, 2010).

The irregularity of the current climate of Australia may accelerate soil formation processes due to the constant exposure to extremes of both temperature and wet and dry periods, which are unlike those experienced in Northern Europe and North American countries where climatic regimes are generally seasonal rather than episodic. In effect, this could be considered a constant re-setting of both soil formation timelines and processes.

Arrival of Europeans superimposed on irregular climate has resulted in additional massive changes in catchment hydrology and erosion, accompanied by large changes in the ability of soils to take in water, through compaction or structural decline, or the removal of groundcover and topsoil, exposing subsoils with lower infiltration capacity (Abbott, 1880; Dodson and Mooney, 2002; Haworth *et al.*, 2010; Dotterweich, 2013). The impact of the land use change from Aboriginal agriculture to heavy grazing and cropping has been so immense that it has transgressed all physiographic and climate regions within Eastern Australia (Dodson and Mooney, 2002).

Yaalon and Yaron (1966) and Richter (2007) make the proposal that "*human influence creates a new reference system for soil formation, a new pedogenesis altogether, and in the technical jargon of pedologists, a new* t_0 *from which a new wave of polygenesis has begun* (Richter, 2007) (p962). In terms of eastern Australia, an already confounded and multidirectional set of potential soil development processes which have ebbed and flowed with climate unpredictability and the removal or replacement of vegetation may well represent the new time zero, allowing the suite of soil formation processes operating within the soil to change and take soil development in a new direction in some areas.

2.13 Conclusions

An examination of the factors and processes of soil formation or development shows that there are potentially many exceptions to the conventionally understood processes and timelines. European and USA soil classifications with a genetic base are most useful in areas that have relatively newly exposed parent materials and a reliable climate, which is often much cooler and moister than those in most of Eastern Australia. In far northern hemisphere regions, subsoils generally do not entirely dry and moisture is in surplus for much of the year.

Land use change in much of eastern Australia has been rapid and aggressive. There are many areas of eastern Australia where soil types have essentially remained unchanged since settlement except for widely reported erosion and decline in organic matter. It is possible that there are certain combinations of parent material, chemistry, climate and topography that have worked together to form soils which are on a threshold of potential change. In these areas, the introduction of a new vegetation type, or the removal of a once dominant vegetation type, could change the equilibrium and initiate a cascade of processes that change whole soil profiles, and not solely soil surface condition.

Whilst Paton *et al.* (1995) considered water supply as the main on and off switch to soil formation, it seems that simple addition of water will only drive processes in one way. An example is changing the redox state of a newly saturated soil for flood irrigated rice production, where the cumulative effects of profile leaching and saturation take many years to change soil (Kölbl *et al.*, 2014). Turning the switch on and off more rapidly may have unexpected effects. Eastern Australia's alternating drought and flood regimes equates to potential for alternately saturating and deeply drying profiles rapidly compared with northern Europe or North America.

The impacts of land use change on eastern Australian soils can be described in terms of a change in soil physical properties, loss of structure, increasing density and lowering porosity, hence the large scale runoff erosion events coincident with the influx of European settlement. Substantial impacts of land use change on soils may be initiated through simple physical changes. Clearing of a well-developed forest to establish pasture, for example, results in exposure of soil at ground level to higher temperature, loss of the large root mass that helps maintain porosity, and increased BD. All of these factors may change the internal drainage of the soil profile, which could drive soil development processes in a new direction.

In terms of the introduction of new plants, replacing native pastures with introduced tropical pastures results in documented changes in above ground biomass production, rooting depths and densities, and consequent soil moisture storage and use. The initiation of greater porosity through the action of more aggressive roots, may drive soil development processes in a new direction if soil is wetting and drying to a greater depth than under native vegetation.

The simple act of adding agricultural lime or fertiliser to pastures may be sufficient in itself to transgress some soil classification boundaries where only a slight change in pH or sodicity is required to assign a soil order.

In the context of the present study, simply extending or shortening rooting depth by introducing new plants or removing existing plants may result in large changes of soil moisture regime and soil moisture dynamics. This could then initiate different physical processes, alter mineral and nutrient cycling, and change soils beyond the surface horizons. Coupled with irregular, short- or long-term partial or complete wetting and drying, soil development in response to anthropogenic changes could be more rapid than expected using soil formation models developed on simple parent materials in more stable or constant climate areas of the world.

Chapter 3. Study One: Using existing soil survey data to investigate soil changes resulting from European land use change. A case study from the Liverpool Plains region of New South Wales.

3.1 Introduction

This study considered a relatively large existing soil data set from soil surveys in the Liverpool Plains region in NSW to determine whether deep soil profile changes may have occurred in response to land use change from undeveloped to grazed or cropped. In terms of the scope of this thesis, it provides a review of available raw data relevant to Study 2 (a comparison of soils under native pastures with soils under improved tropical pastures) and Study 3 (a comparison with of soils under cleared pasture lands with those under adjacent tall eucalypt forest). This study also assessed the usefulness or otherwise of this type of soil survey data to assess impacts of land use change which otherwise might require detailed studies using chronosequences or long-term monitoring trials.

Changes in soil physical and chemical characteristics including SOC content, fertility and salinity in response to agricultural land use have significant impacts on global environmental issues such as climate change, agricultural productivity and surface and groundwater quality (Oldeman *et al.*, 1991; de Moraes *et al.*, 1996; de Moraes Sá *et al.*, 2017; Robertson *et al.*, 2018). Anthropogenic land use change from a natural state to agricultural use, or from a less intensive use such as sparse grazing to frequent cropping has impacts that are well documented, especially for SOC (West and Post, 2002; Lopez-Garrido *et al.*, 2011; Badgery *et al.*, 2013; Badgery *et al.*, 2014; Rabbi *et al.*, 2014; Jones *et al.*, 2016). Most of these studies cited herein covered large areas with a diverse climate range and land use, and found that climate also significantly affects SOC storage along with soil type and land use. These studies also indicated that soils under intensive land uses such as cropping are generally more depleted in SOC.

Locality and history of study area

The Liverpool Plains area of New South Wales (Figure 3.1) is an important and generally reliable dryland and irrigated cropping and grazing region of Australia. Though small in area (1. 2 million ha), summer and winter agricultural production (grains and cotton) is disproportionately higher than most other Australian farming regions. Most of the Liverpool Plains' regional rainfall is in the range of 590 – 680 mm with annual average maximum temperatures in excess of 23°C and is considered to be a dry subhumid region (Banks, 1995).

The Liverpool Plains area was a focal region for soil salinity research and soil survey from 1990 to 2005 (Ringrose-Voase *et al.*, 2003). Regional soil surveys during this period generated 1095 soil profiles with associated laboratory data and more than 1200 additional soil profiles as soil descriptions only. Soil profile records for this period are stored in the NSW Soil and Land Information System (SALIS). SALIS is administered by the NSW Office of Environment and Heritage (OEH) and contains > 70,000 soil profile descriptions, a large number of which have associated horizon-based laboratory characterisation from many government soil surveys across NSW. These represent a substantial financial investment and are an invaluable resource for regional assessments of soil and landscape qualities and limitations.

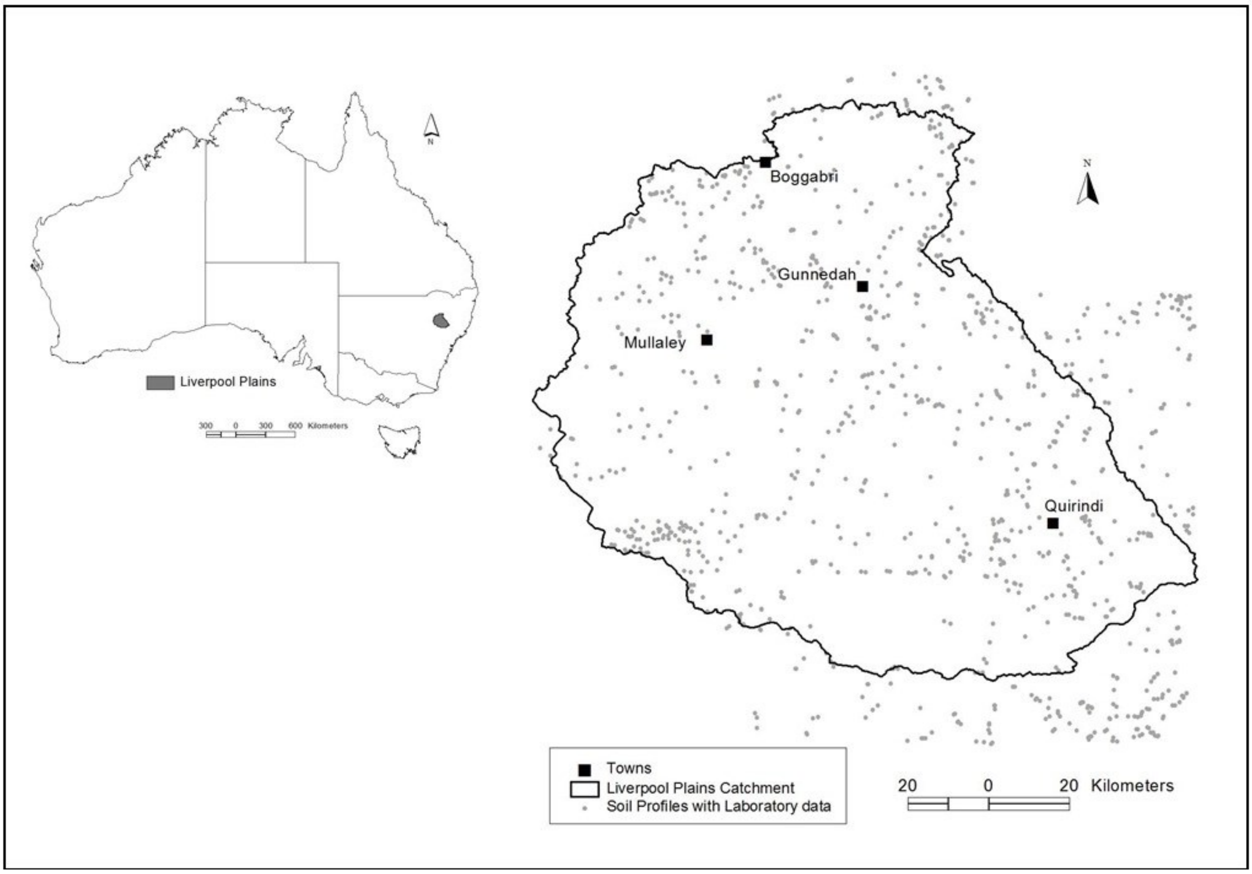


Figure 3.1*: Location of the Liverpool Plains in Australia and location of soil profile sampling points in the broader Liverpool Plains region*

SOC and land use effects

Estimates of carbon storage in pre-European NSW have been made using soil data from relatively undisturbed sites and extrapolated to whole landscape mapping units (Banks and McKane, 2002) or by digital modelling using control points for validation (Gray *et al.*, 2016). Modelling by Gray *et al.* (2016) showed that the largest declines in SOC since Europeans arrived in NSW have occurred in the intermediate rainfall zones and where cropping is the dominant land use. Gray *et al.* (2015) used an annual precipitation (mm) to annual daily maximum temperature (°C) ratio (P/Tm) to classify areas as “Dry < 30, Moist (30 – 60) or Wet (> 60)”. Further modelling of south eastern Australian soils by Liu *et al.* (2016) confirmed losses of SOC in cropping systems were exacerbated by annual average maximum temperatures in excess of 20°C.

Internationally, Piñeiro *et al.* (2010) reviewed 67 published paired site comparisons of grazing versus grazing exclusion, and found that grazing systems in intermediate rainfall areas (400 – 850 mm) have the most variable SOC response (accumulation or loss) to changed management. In terms of climate, the Liverpool Plains area is mostly on the boundary of the “dry” and “moist” category of Gray *et al.* (2015) and within the variable SOC response zone of Piñeiro *et al.* (2010). The relatively homogenous climate of the broader Liverpool Plains area offers a large area within a relatively uniform climate on gentle slopes (therefore limiting aspect effects) in which to study the effects of land use and soil type on SOC content and other soil attributes.

Detailed comparative land use and soil studies in the northwest slopes and tablelands of NSW have been conducted through a range of tightly defined agricultural field trials studying land use effects on soils. Examples include the study by (Young *et al.*, 2005); Young *et al.* (2009) and broader soil type specific comparisons of landscape scale soil characteristics with different agricultural land uses and woodland landscapes such as those reported in Badgery *et al.* (2013), McLeod *et al.* (2013) or Schwenke *et al.* (2013). These studies reveal a general pattern of soil carbon storage showing that SOC is highest under forest or woodland, intermediate in pastures (whether sown or natural) and lowest in cropping systems. Rabbi *et al.* (2014, 2015) examined carbon and land use effects over a large area of NSW, and found a poorer relationship of land use with SOC and soil type than with climate, indicating that water availability and temperature were significant determinants of SOC sequestration in large studies where climate thresholds were crossed.

Fertility factors influencing SOC sequestration

Fertility can be considered a mixture of or interaction between physical properties, that affect soil water storage and access and soil structure, and chemical properties that control plant nutrient availability or may have toxic effects on plants. Water supply is the driver of SOC sequestration in any environment. Extremely moist environments or saturated soil environments sequester carbon as peat, and extremely dry or arid soil environments sequester little carbon, because plant production is limited and soil biological activity is greatly reduced. Given the sporadic nature of rainfall in the Liverpool Plains area (which is unevenly distributed and often unseasonal), the ability to store soil moisture for long time periods is essential for agricultural production systems. In terms of soil attribute selection for agricultural land use, high plant available water holding capacity (AWC) is desirable.

Soil structure influences the hydrological characteristics of soils and to some extent how biologically active a soil is (Bronick and Lal, 2005). Soils which are naturally dispersive (high DP) through elevated exchangeable sodium percentage (ESP) tend to have less well-developed soil structure, poor drainage, and are often hard setting which is reflected in restricted plant growth (So and Aylmore, 1993). Sodic and dispersive soils are associated with lower or inhibited microbiological activity (Naidu and Rengasamy, 1993). Well-structured soils tend to protect mineralisable SOC within stable aggregates (Pulleman and Marinissen, 2004).

Soil chemical fertility is derived mainly from mineral and biological weathering of soil parent material. Gray *et al.* (2015) used parent material silica content (how mafic the parent material is) as an index of chemical fertility in modelling SOC sequestration in eastern Australia. More mafic parent materials generally weather to higher clay content soils which serves to protect SOC from mineralisation and controls soil fertility and nutrient supply (Gray *et al.*, 2015).

The soil's ability to hold positively charged ions or its cation exchange capacity (CEC) is a key characteristic influencing soil structure stability, nutrient availability, soil pH and the fate of fertilisers and other ameliorants in soil (Hazelton and Murphy, 2007). Elevated salinity (indicated by increased effective electrical conductivity, EC_e) may interfere with natural soil fertility by limiting plant growth through toxic osmotic effects (Bernstein, 1975).

Hartemink (2006) used sequential soil pH, organic C, total N, available P, CEC, and exchangeable cation measurements from soil survey data to prepare guidelines for soil fertility decline assessment in sub-Saharan Africa. Although SOC is part of soil fertility assessment, its sequestration in soil can also be considered an outcome of sufficient soil fertility. Soil chemical fertility positively affects plant biomass production, and residual biomass production available for microbial incorporation to soil as SOC (Oades, 1984; Kallenbach *et al.*, 2016).

In relatively undisturbed non-waterlogged environments, such as forest and well-managed pasture lands within a narrow climatic range, carbon sequestration is limited by soil fertility (Oren *et al.*, 2001; Kirkby *et al.*, 2014). It follows then that the most fertile soils with best chemical and physical attributes in this study, where climate and aspect have limited variability, should have the highest SOC.

Objective and hypotheses

The objective of this study was to examine the relationships of land use with soil type and with a range of soil fertility parameters.

The study tested two hypotheses:

1. That changes in land use may result in significant differences in soil physical and chemical properties deep in soil profiles, and
2. That land use is an overriding driver of change in SOC content compared with soil type and land use × soil type interactions.

3.2 Materials and methods

Data selection from SALIS database

1095 soil profiles were extracted from SALIS for analysis. Topographically, 97.5% of the soil profiles with laboratory data for the Liverpool Plains area were described on slopes of < 9. 5 % (or 5. 4°). Of 1095 soil profiles, 791 soil profiles were extracted from SALIS (SALIS, 2017) for the greater Liverpool Plains region. Each of the selected soil profiles was required to have a full complement of laboratory characterisation data. All selected soil profiles were on slopes of < 9.5% to minimise the influence of aspect. Topography (slope/aspect) and climate (rainfall and temperature) values for the selected data set were relatively homogenous, and were not considered here.

Soil profiles had site and soil descriptions (i.e. location, land use category, colour, field texture, field pH, horizonation, structure and soil fabric) per NCST (2009) and Milford *et al.* (2001). Soil profiles were sampled by soil horizon for laboratory analysis. Laboratory data for each profile included available AWC (McIntyre, 1974), Bray P (Vimpany *et al.*, 1987), CEC, ESP (Pleysier and Juo, 1980), DP (Richie, 1963), EC_e, (Rayment and Higginson, 1992), soil pH in water (1:5 solid/liquid ratio) (Rayment and Higginson, 1992) and SOC (Black, 1965). Field soil colour or Munsell notation posed limitations for direct statistical comparison of traditionally described and sampled soil profiles and was not used in subsequent analyses.

Some land use categories, including hardwood and softwood plantation, urban, industrial, mining and other were excluded from the analysis due to limited representation in the data-set. The remaining land use categories from SALIS were grouped into simplified land use categories: cropping, improved pasture, native pasture (directly from NCST (2009) and Milford *et al.* (2001)) and woodland, which included National/State Parks, timber/scrub/unused and logged native forest from Milford *et al.* (2001) and NCST (2009). Four soil profiles representing two soil Orders, the Hydrosols and Organosols (found in swamps with reducing conditions), were removed from the dataset (4 profiles), due to limited geographic distribution in the wider Liverpool Plains region. Land use groups were comprised of up to 11 different soil Orders according the Australian Soil Classification (ASC) (Isbell and NCST, 2016) (Figure 3.2). Broad relationships between the ASC with the World Reference Base (IUSS, 2015) and Soil Taxonomy (USDA, 2014) classification systems are given in Table 3. 1

Potential limitations of SALIS data

Soil profile data from SALIS are generally described at points of easy access, rather than specific targeted points within a landscape or soil mapping unit. Limitations of the data include sampling by horizons which are highly variable in thickness and depth, and the use of often poorly defined categorical site data for land use groups, that lack site history. Despite the site and location limitations, the soil profile descriptions are of high quality, performed to a consistent standard (NCST, 2009) and the analytical data were generated through a National Association of Testing Authorities (NATA) accredited laboratory.

Soil profile descriptions and measurements generally lack measured BD data and consistency of depth and thickness of sampling as a result of highly variable soil horizons. These factors posed limitations for direct statistical comparison of traditionally described and sampled soil profiles.

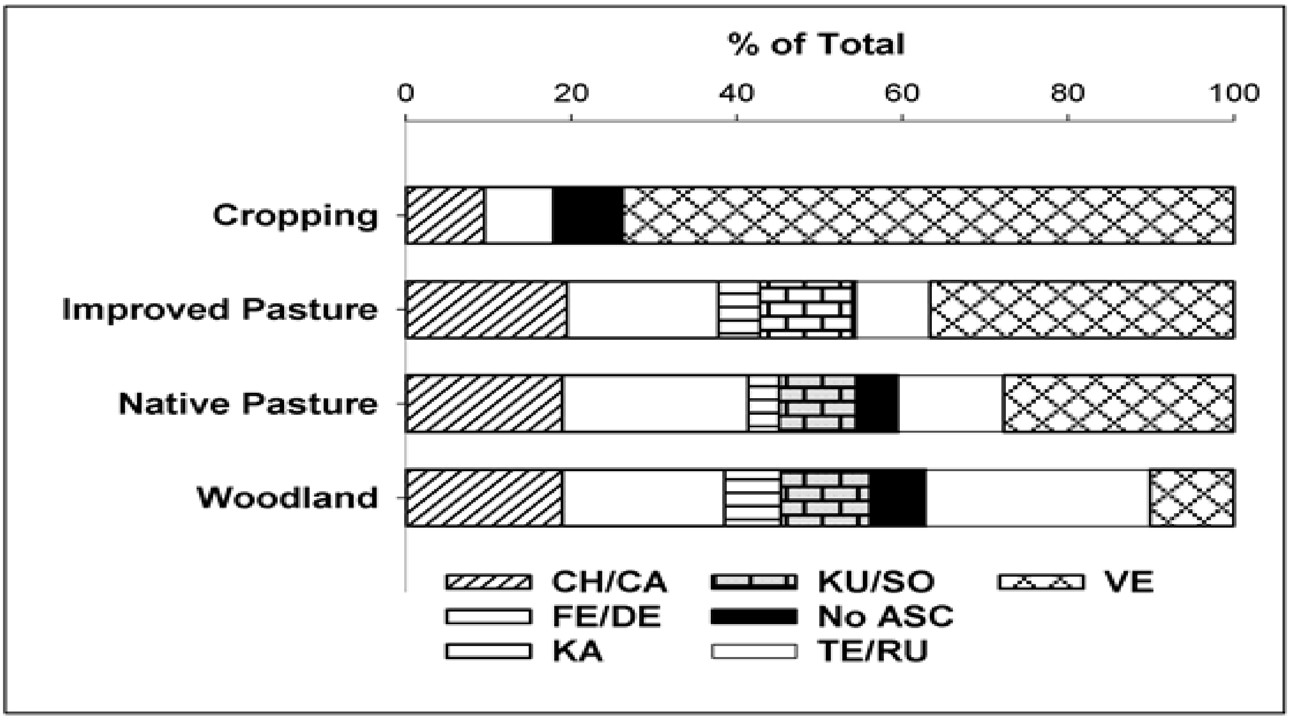


Figure 3.2: Distribution of Australian Soil Orders within land use groups. CA (Calcarosols), CH (Chromosols), DE (Dermosols), FE (Ferrosols) KA (Kandosols), KU (Kurosols), No ASC (No Class Assigned) RU (Rudosols), SO (Sodosols), TE (Tenosols), VE (Vertosols). Cropping n = 76, Improved Pasture n = 174, Native Pasture n = 424, Woodland n = 117. Total n = 791.

Soil groupings

Four soil groups were created from the aggregation of soil classes for analysis in this study based on the broad fertility/productivity classes of Charman (1978). They were:

1. Chromosols (Chromosols and Calcarosols),
2. Dermosols/Ferrosols (Dermosols and Ferrosols),
3. Other Soils (Kandosols, Kurosols, Rudosols, Sodosols and Tenosols) and
4. Vertosols.

Table 3.1: *Broad relationships of ASC Isbell and NCST (2016) with World Reference Base (IUSS, 2015), and Soil Taxonomy (USDA, 2014).*

ASC Order	World Reference Base	Soil Taxonomy Order
Calcarosols	Calcisols	Alfisols
Chromosols	Alisols/ Luvisols	Alfisols
Dermosols	Chernozems/	Mollisols, Alfisols, Ultisols
Ferrosols	Nitisols/Ferralsols	Oxisols, Alfisols
Hydrosols	Histosols	Mollisoils, Alfisols, , Inceptisols
Kandosols	Cambisols	Alfisols, Ultisols
Kurosols	Acrisols/Lixisols	Alfisols, Ultisols
Rudosols	Regosols/Leptosols	Entisols
Sodosols	Acrisols/Lixisols/Solonetz	Alfisols, Aridisols
Tenosols	Arenosols/Regosols	Inceptisiols, Aridisols, Entisols
Vertosols	Vertisols	Vertisols

Data preparation

Bulk density estimation

BD measurements were required to convert mass based SOC and AWC based measurements to volumetric estimates. Only 79 data points in the data set had measured BD. Bulk density was therefore estimated for the whole data set. The BD for all data points was calculated using the adapted Adams-Stewart multiple linear regression pedotransfer function assessed in Tranter *et al.* (2007). The function considered mineral packing structures (p_m) and soil structure (Δ_p) to derive a predicted BD (P_b) as follows:

$$P_b = p_m + \Delta_p$$

$$p_m = a + b \times \text{sand} + (c - \text{sand})^2 \times d + e \log \text{depth}$$

$$\Delta_p = w + y \log \text{OC} + z \log \text{depth}$$

Where: sand is in %; depth in cm; a = 1. 35; b = 0. 045; c = 44. 7; $d = -6 \times 10^{-5}$ and e = 0. 060;

w = -0. 217; y = -0114; and z = -0. 077.

A regression of predicted BD against measured BD was performed for the 79 data points available. The regression explained 63% of the variation in the predicted BD data (Figure 3.3) which was slightly higher than that reported in Tranter *et al.* (2007). This level of precision was considered reasonable for calculation of volumetric measures.

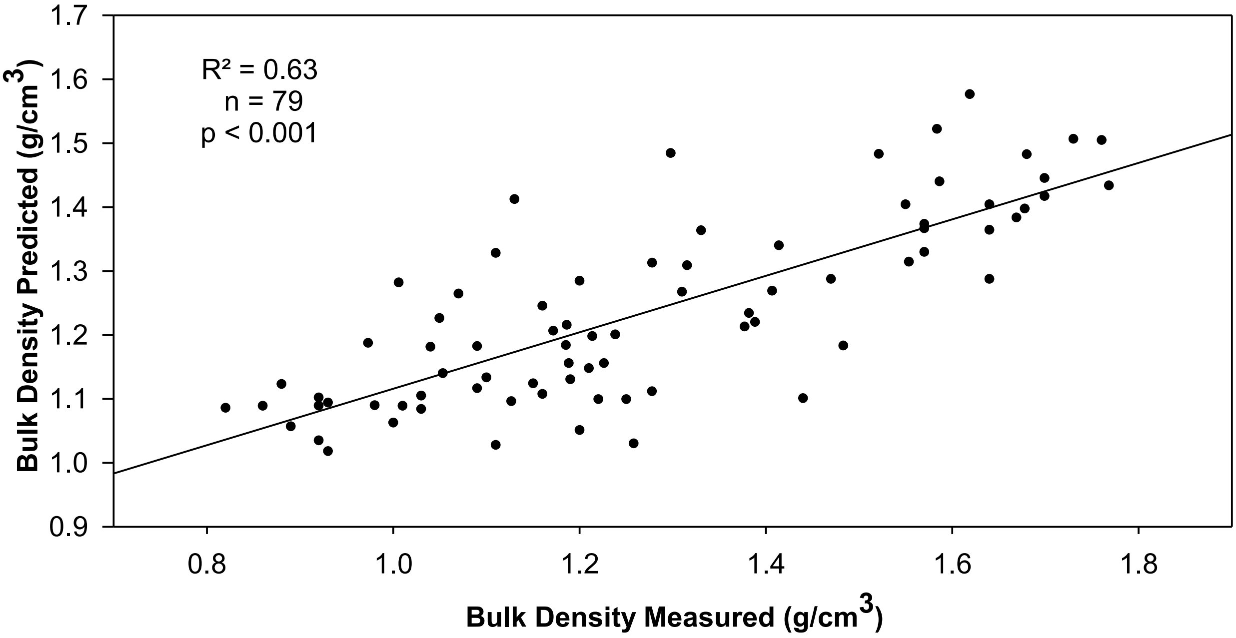


Figure 3.3*: Regression of measured BD with predicted BD as per Tranter et al. (2007).*

Quadratic splining of horizon data

Studies of soil carbon stocks have been done where soil horizon data was relatively consistent (Wiesmeier *et al.*, 2012) or where data were collected at specified and comparable depth intervals. Soil sampling for this study was by horizon, which were highly variable in texture, thickness and depth in the soil profile, making direct comparison of soil attributes problematic (Bishop *et al.*, 1999; Malone *et al.*, 2009). Continuous soil information to specified depths was required for reasonable statistical comparison. To overcome this challenge, the CSIRO equal area quadratic spline program "Spline Tool" (Jacquier and Seaton, 2012) was used to generate data in depth increments of 0 – 10, 10 – 30, 30 – 50, 50 – 70, 70 – 100, and 100 – 200 cm, respectively, for each soil property. Data for 797 individual soil profiles were splined, creating 3864 depth slices for comparison.

Statistical testing

Simple ANOVA or non-parametric testing of the generated depth slice data was considered inappropriate in this instance. Splined soil profile data were analysed using a restricted maximum likelihood (REML) mixed model analysis using ASReml-R (Butler *et al.*, 2009). The model chosen was:

$$y_{ijk} = \mu + S_i|D_k + L_j|D_k + (S_i \times L_j)|D_k + \varepsilon_{ij}|D_k \ ,$$

where y_{ijk} is the splined soil profile for Soil Group i, Land Use j and depth k; μ is the grand mean; $S_i|D_k$ is the effect of Soil Group i for depth k; $L_j|D_k$ is the effect of Land Use j for depth k; $(S_i \times L_j)|D_k$ is the interaction between Soil Group i and Land Use j for depth k; and $\varepsilon_{ij}|D_k$ is the residual effect for depth k.

All terms, except μ, were fitted as random. The % of variance explained was calculated by dividing each residual component for Soil Group, Land Use, and Soil Group × Land Use, by the sum of the residuals excluding those for depth alone and expressing the resultant value as a percentage. The significance of the variance of the random terms was tested using the log-likelihood ratio test. All results where $p \leq 0.\ 05$ were considered significant (Table 3. 2).

Simple fertility measures selected for this study were soil pH (controls nutrient availability and toxicities), CEC (general indicator of soil fertility), EC_e (broad salinity), Bray P (a measure of available P, an essential plant nutrient), AWC (ability of a soil to store and release water to plants), ESP and DP (each a measure of soil structure stability, dispersibility, potentially impeded drainage and subsoil constraints to plant growth) (McKenzie *et al.*, 2004). SOC is considered to be a product of fertility in this study (Hartemink, 2006; Kirkby *et al.*, 2014). No soil nitrogen values were available for any profiles in the data set.

Attributes were selected because they were available for almost all points in the data set and all have a known effect on physical or chemical fertility. The significant terms (best linear unbiased predictors or BLUPs) from the model were presented graphically with their standard error intervals in Figures 3.4 – 3.11. BLUPs from this analysis were considered to be equivalent to a mean and are treated as such in the results and discussion.

3.3 Results

The residual variances varied across depth for all traits (Table 3.2). The residual variances tended to increase with the increasing depth, except for Bray P and OC where the shallower depths tended to have larger residual variances than the deeper ones. Significant interaction between Soil Group and Land Use was only detected for Bray P, CEC, and OC (Table 3.2). AWC and pH showed significant effects for both Soil Group and Land Use, whereas DP, EC_e and ESP only showed significant effects for Soil Group (3. 2).

Table 3.2*: Variance components for effects of Soil Group, Land Use, interaction of Soil Group and Land Use and residuals for each depth.*

TRAIT	SOURCE OF VARIABILITY	VARIANCE COMPONENT	% VARIANCE EXPLAINED	STD ERROR	P - VALUE
AWC	Soil Group | Depth	19. 26	70. 83	6. 11	0. 000
	Land Use |Depth	7. 60	27. 96	2. 82	0. 000
	Soil Group × Land Use |Depth	0. 33	1. 21	0. 52	0. 611
	Residual for Depth_5	53. 70	-	2. 72	
	Residual for Depth_20	52. 23	-	2. 65	
	Residual for _40	70. 30	-	3. 65	
	Residual for Depth_60	84. 51	-	4. 70	
	Residual for Depth_85	74. 24	-	4. 58	
	Residual for Depth_150	86. 60	-	6. 67	
BRAY P	Soil Group | Depth	56. 35	62. 08	24. 76	0. 000
	Land Use |Depth	0. 00	0. 00	NA	-
	Soil Group × Land Use |Depth	34. 42	37. 92	13. 65	0. 000
	Residual for Depth_5	1808. 47	-	92. 20	
	Residual for Depth_20	1473. 63	-	74. 96	
	Residual for _40	1243. 42	-	64. 16	
	Residual for Depth_60	1250. 20	-	67. 75	
	Residual for Depth_85	2290. 93	-	140. 90	
	Residual for Depth_150	2975. 22	-	239. 02	
CEC	Soil Group | Depth	129. 21	89. 44	39. 76	0. 000
	Land Use |Depth	7. 49	5. 18	4. 44	0. 018
	Soil Group × Land Use |Depth	7. 78	5. 38	3. 45	0. 000
	Residual for Depth_5	242. 14	-	12. 27	
	Residual for Depth_20	234. 95	-	11. 90	

TRAIT	SOURCE OF VARIABILITY	VARIANCE COMPONENT	% VARIANCE EXPLAINED	STD ERROR	P - VALUE
	Residual for _40	246. 70	-	12. 82	
	Residual for Depth_60	252. 96	-	14. 10	
	Residual for Depth_85	277. 07	-	17. 07	
	Residual for Depth_150	431. 16	-	33. 06	
DP	Soil Group | Depth	26. 89	93. 58	8. 65	0. 000
	Land Use |Depth	0. 00	0. 00	NA	-
	Soil Group × Land Use |Depth	1. 84	6. 42	1. 03	0. 007
	Residual for Depth_5	112. 99	-	5. 77	
	Residual for Depth_20	113. 31	-	5. 74	
	Residual for _40	164. 33	-	8. 34	
	Residual for Depth_60	208. 85	-	10. 88	
	Residual for Depth_85	266. 77	-	15. 56	
	Residual for Depth_150	347. 92	-	26. 97	
ECE	Soil Group | Depth	0. 92	89. 67	0. 31	0. 000
	Land Use |Depth	0. 09	8. 87	0. 06	0. 026
	Soil Group × Land Use |Depth	0. 01	1. 46	0. 03	0. 568
	Residual for Depth_5	3. 78	-	0. 20	
	Residual for Depth_20	6. 47	-	0. 33	
	Residual for _40	5. 65	-	0. 29	
	Residual for Depth_60	7. 30	-	0. 41	
	Residual for Depth_85	13. 78	-	0. 85	
	Residual for Depth_150	23. 17	-	1. 77	
ESP	Soil Group | Depth	7. 00	100. 00	2. 22	0. 000
	Land Use |Depth	0. 00	0. 00	NA	-
	Soil Group × Land Use |Depth	0. 00	0. 00	NA	-
	Residual for Depth_5	27. 60	-	1. 41	

TRAIT	SOURCE OF VARIABILITY	VARIANCE COMPONENT	% VARIANCE EXPLAINED	STD ERROR	P - VALUE
	Residual for Depth_20	37. 59	-	1. 90	
	Residual for _40	44. 78	-	2. 32	
	Residual for Depth_60	58. 97	-	3. 28	
	Residual for Depth_85	68. 72	-	4. 22	
	Residual for Depth_150	91. 77	-	7. 03	
OC	Soil Group | Depth	2. 40	2. 98	1. 42	0. 020
	Land Use |Depth	74. 67	92. 85	23. 06	0. 006
	Soil Group × Land Use |Depth	3. 35	4. 17	1. 19	0. 000
	Residual for Depth_5	305. 28	-	15. 40	
	Residual for Depth_20	106. 96	-	5. 41	
	Residual for _40	46. 49	-	2. 46	
	Residual for Depth_60	36. 33	-	2. 09	
	Residual for Depth_85	30. 89	-	1. 95	
	Residual for Depth_150	22. 29	-	1. 75	
PH	Soil Group | Depth	0. 20	49. 13	0. 07	0. 000
	Land Use |Depth	0. 21	50. 87	0. 07	0. 000
	Soil Group × Land Use |Depth	0. 00	0. 00	NA	-
	Residual for Depth_5	0. 54	-	0. 03	
	Residual for Depth_20	0. 56	-	0. 03	
	Residual for _40	0. 77	-	0. 04	
	Residual for Depth_60	0. 89	-	0. 05	
	Residual for Depth_85	0. 85	-	0. 05	
	Residual for Depth_150	0. 86	-	0. 07	

Fertility factors

Available water holding capacity (AWC)

Soil Group explained 71% of the variation in AWC data with Land Use having a lesser effect on AWC than Soil Group (28%) (Table 3.2). Vertosols had the highest AWC down the profile with significantly higher values than all other Soil Groups (Figure 3.4 B). AWC for Land Use groups showed that woodland had the lowest AWC, and cropping the highest with both types of pasture intermediate between cropping and woodland (Figure 3.4 A). There was no interaction between Land Use and Soil Group, although the pattern of significant separation of Land Use groups within Soil Group was clearly indicated with cropping having the highest AWC and woodland the lowest AWC within all Soil Groups (Figures 3.4 C–F). Native and improved pastures could not be separated on the basis of AWC and both Land Uses exhibited intermediate AWC between woodland and cropping for all analyses.

Bray P

Land Use did not explain a significant amount of variation in Bray P values (Table 3.2), hence predicted Bray P were the similar for all Land Use categories (Figure 3.5 A). Soil Group explained 62% of the variation in the predicted Bray P (Table 3.2), with Vertosols having higher Bray P than Ferrosols/Dermosols, and Chromosols and Other Soils having similar response patterns (Figure 3.5B). Vertosols had significantly higher P than Chromosols and Other soils. The interaction between Land Use and Soil Group explained 38% of the variation in predicted Bray P (Table 3.2) with Vertosols under cropping showing significantly higher P in the upper 50 cm of the soil than all other Land Use groups within Vertosols (Figure 3.5 F). All other Soil Group × Land Use groups showed no significant difference between Land Use categories for Bray P (Figure 3.5 C – F).

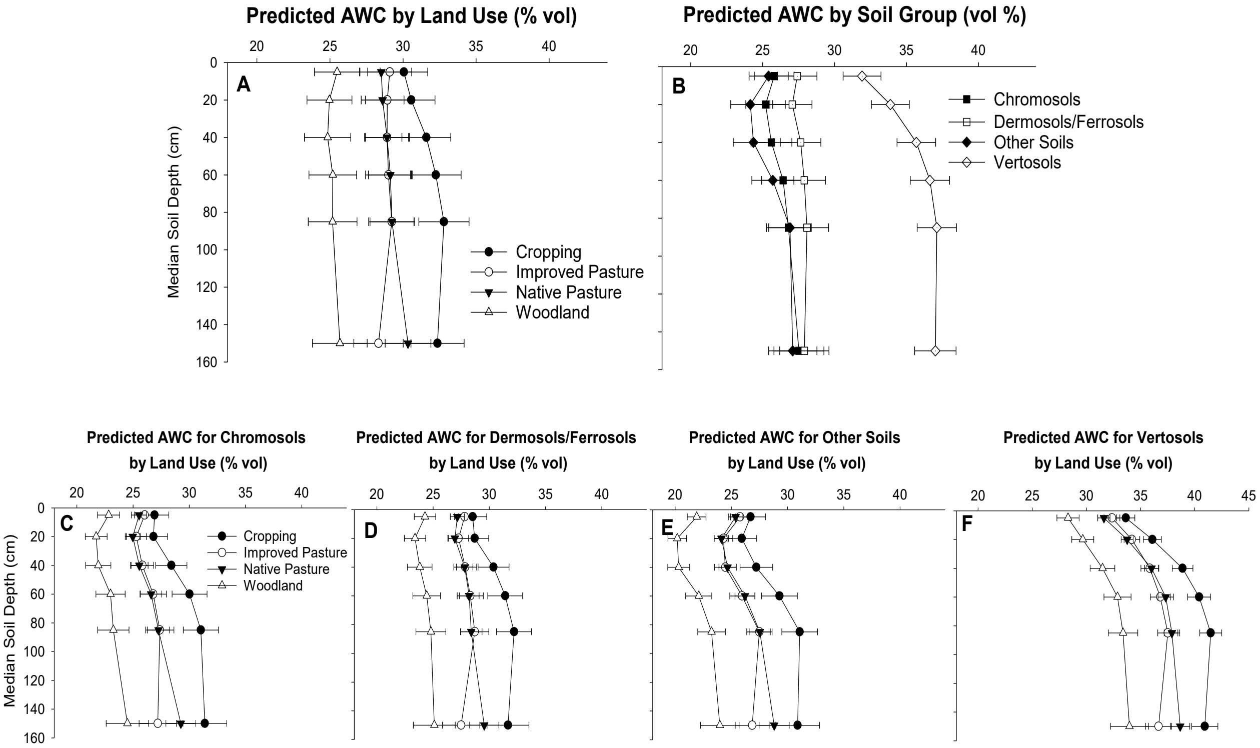


***Figure** 3.4: Predicted available water holding capacity by: A. Land Use, B. Soil Group and C – F Soil Group × Land Use Interaction. Error bars represent one standard error of predicted means.*

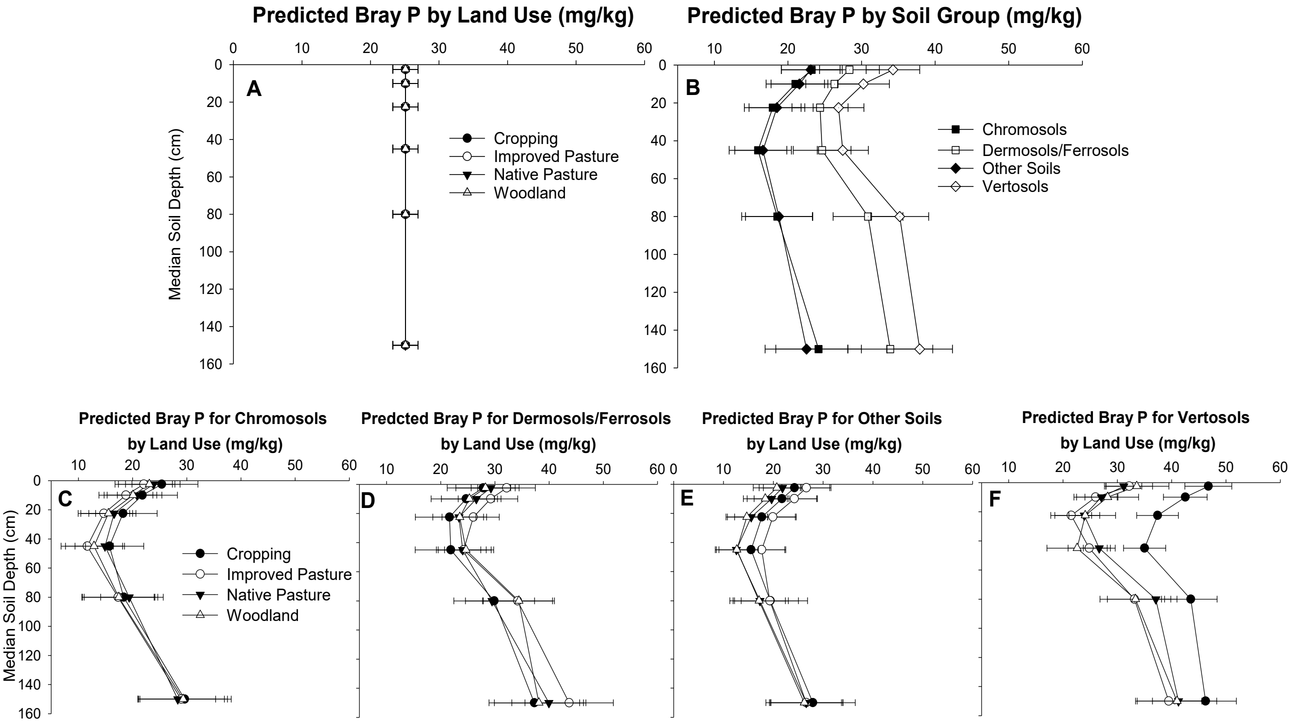


***Figure 3.5**: Predicted Bray P by: A. Land Use, B. Soil Group and C – F Soil Group × Land Use Interaction. Error bars represent one standard error of predicted means.*

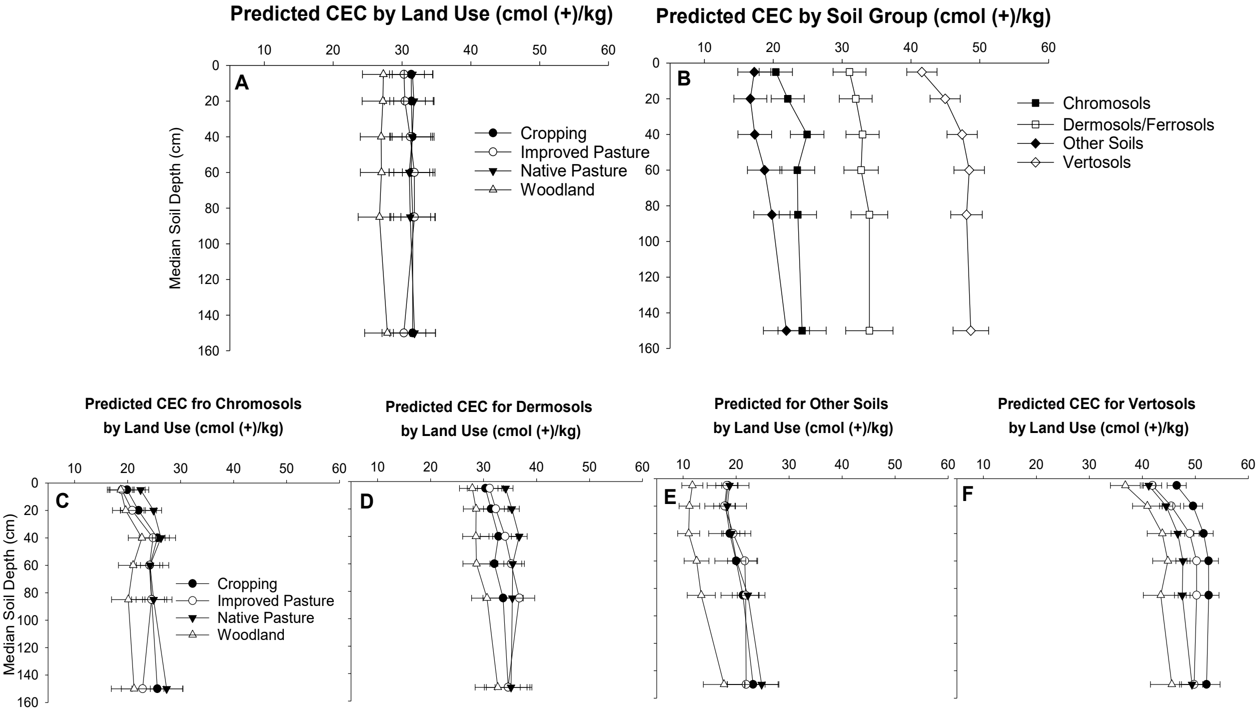


Figure 3.6*: Predicted Cation Exchange Capacity (CEC) by: A. Land Use, B. Soil Group and C – F Soil Group × Land Use Interaction. Error bars represent one standard error of predicted means.*

Cation exchange capacity

Land Use had a limited main effect on CEC (Table 3.2, Figure 3.6 A). Soil Group accounted for 89% of the variation in CEC ((Table 3.2, Figure 3.6 B), with Vertosols and Dermosol/Ferrosol soil groups exhibiting significantly higher CEC than Chromosols and Other Soils. Predicted CEC was highest in Vertosols, intermediate in Dermosols/Ferrosols and lowest in the Chromosol/Other Soil groups. There was little variability due to interaction of Land Use and Soil Group with regard to CEC, although woodland CEC was consistently the lowest for all soil group × land use assessments (Figures 3.6 C – F).

Dispersion percentage (DP)

Land Use explained none of the variance in DP (Table 3.2, Figure 3.7 A); however, Soil Group explained 94% of variation in DP (Figure 3.7 B). DP for Dermosols/Ferrosols and Vertosols was the lowest in near surface layers, with Dermosols/Ferrosols DP remaining at low levels down the profile. The DP in Vertosols increased steadily down the profile and was significantly higher than Dermosols/Ferrosols below 80 cm depth. Chromosol DP was consistent down the profile, with Other Soils having the highest overall DP, significantly higher than all other soil groups at 22. 5 – 80 cm depths, and significantly higher than Dermosols/Ferrosols at 150 cm. Soil Group × Land Use explained a very small percentage of the variation in DP, which is why graphs for interactions of Soil Group × Land Use (Figure 3.7 C – F) are almost identical to the main effect graph for Soil Group.

Salinity (EC_e)

Land Use explained a small amount of the variation in predicted EC_e (Table 3.2, Figure 3.8 A). Soil Group explained 90% of the variation in EC_e data. Surface salinity was similar for all Soil Groups, with salinity increasing significantly at depth in Vertosols compared with the other three Soil Groups (Figure 3.8 B). There was no interaction between Soil Group and Land Use with regard to ECe, with all Land Use types showing a gradual increase in predicted EC_e down the soil profile (Figures 3.8 C – F).

Exchangeable sodium percentage (ESP)

Land Use explained a small amount of the variation in ESP (Figure 3.9 A). Soil Group explained 100% of variation in ESP (Figure 3.9 B) which is why the graphs of Soil Group × Land Use (Figure 3.9 C – F) are exactly the same as the graph for Soil Group. There was no interaction between Soil Group and Land Use. Dermosols/Ferrosols and Chromosols had the lowest in profile ESP, which increased very slightly with depth. Other Soils had the highest in profile predicted ESP ranging from 5. 8 at the surface to 10 at depth. ESP for the Other Soil group was significantly higher for all depths than both Dermosols/Ferrosols and significantly higher than Vertosols to 0 – 22 cm. ESP in Vertosol in topsoils was similar to Dermosols/Ferrosols and Chromosols, which increased almost linearly to the same range as the Other Soils group at 100 cm and 150 cm.

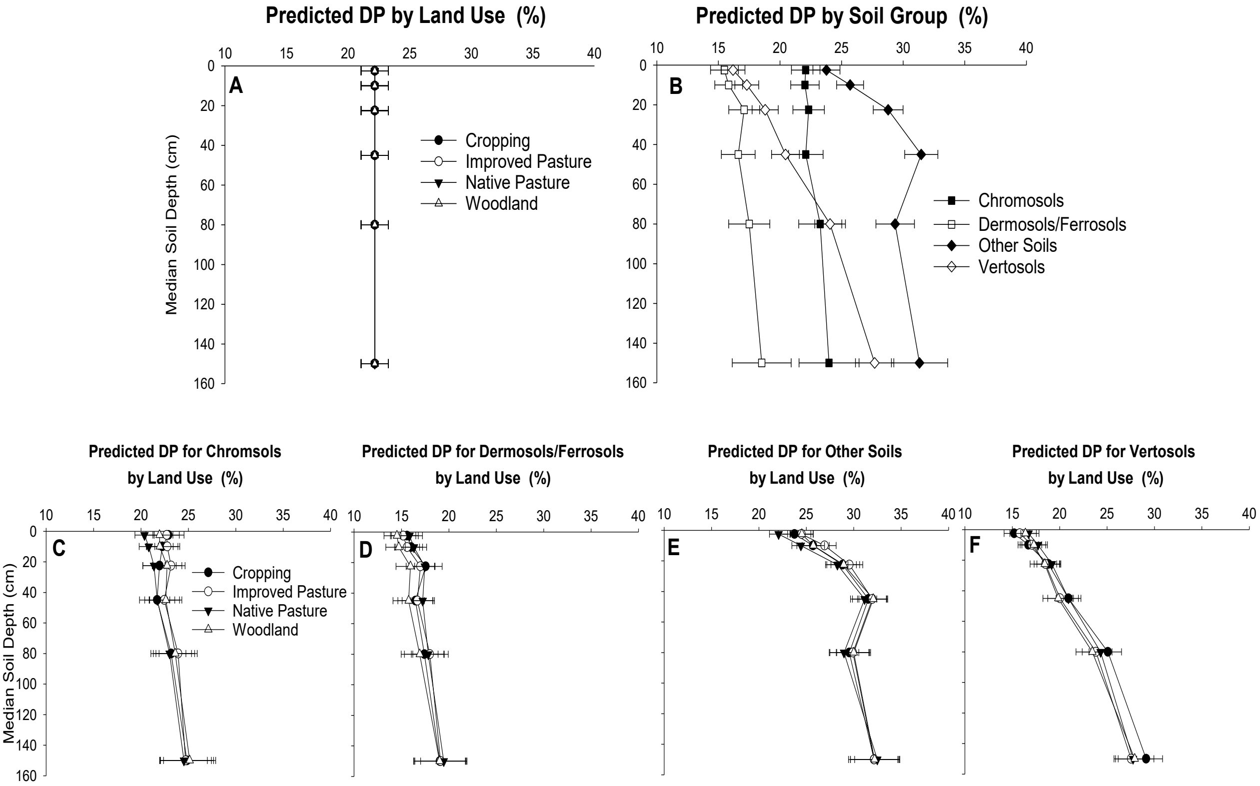


***Figure 3.7**: Predicted dispersion percentage (DP) by: A. Land Use, B. Soil Group and C – F Soil Group × Land Use Interaction. Error bars represent one standard error of predicted means.*

Soil pH

Fifty-one percent (51%) of the variation in soil pH was explained by Land Use and 49% was explained by Soil Group, with no interaction between Soil Group and Land Use (Table 3.2). Soil pH was significantly lower in woodland (Figure 3.10 A) and highest in cropping with both pasture land uses being intermediate between the two. Soil pH by soil type showed that Vertosols had significantly higher pH down the profile than all other soil groups, which had a similar trend down the profile. Soil pH of both pasture types approached that of Vertosols at 150 cm depth. The effect of land use in influencing pH was reflected in Figures 3.10 C – F, which showed that woodland had significantly the lowest pH for all soil types and that cropping had the highest pH, with both pasture types being intermediate between the two.

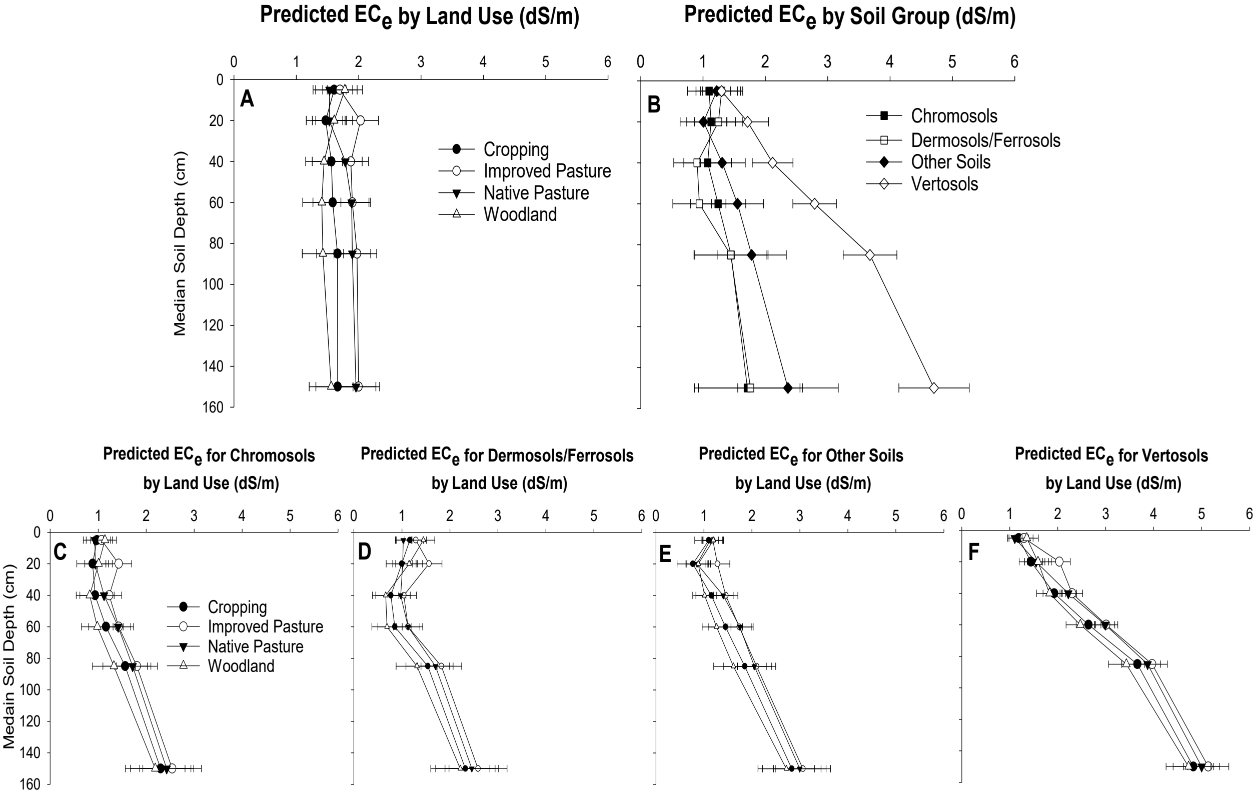


Figure 3.8: *Predicted salinity (EC$_e$) by A. Land Use, B. Soil Group and C - F Soil Group× Land Use Interaction. Error bars represent one standard error of predicted means.*

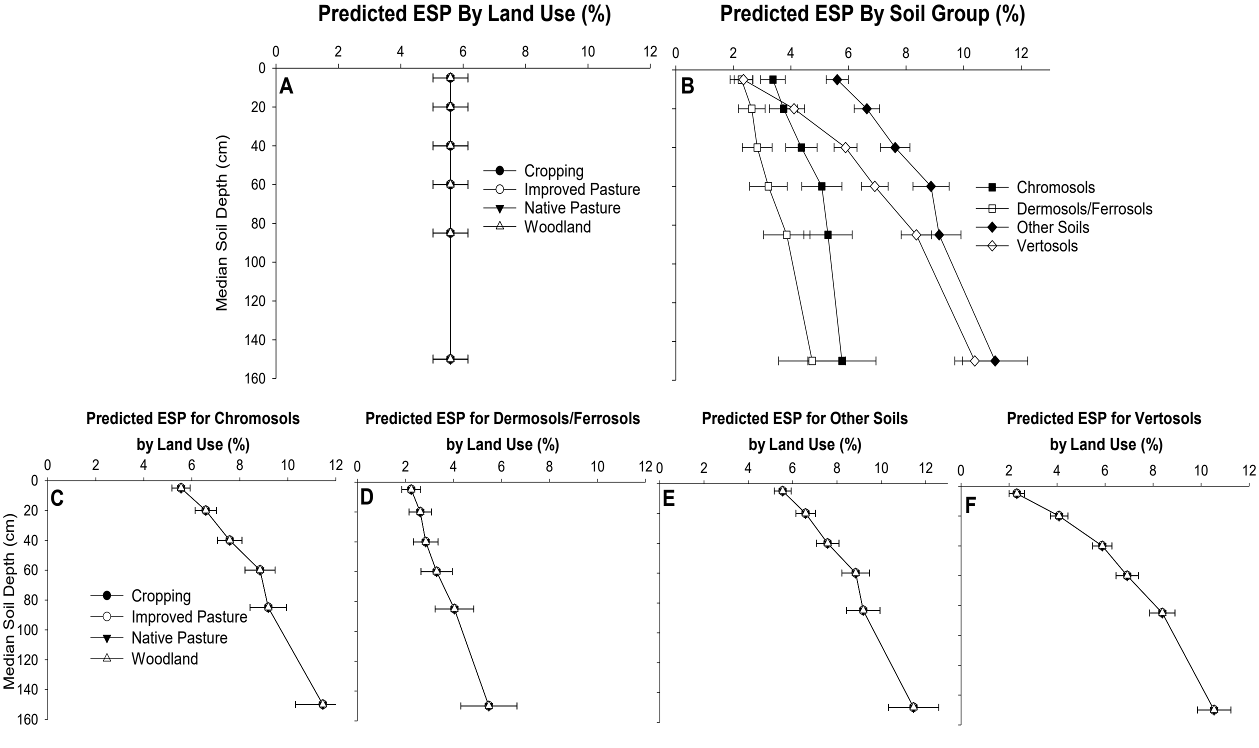


Figure 3.9: *Predicted Exchangeable Sodium Percentage (ESP) by A. Land Use, B. Soil Group and C – F Soil Group × Land use Interaction. Error bars represent one standard error of predicted means.*

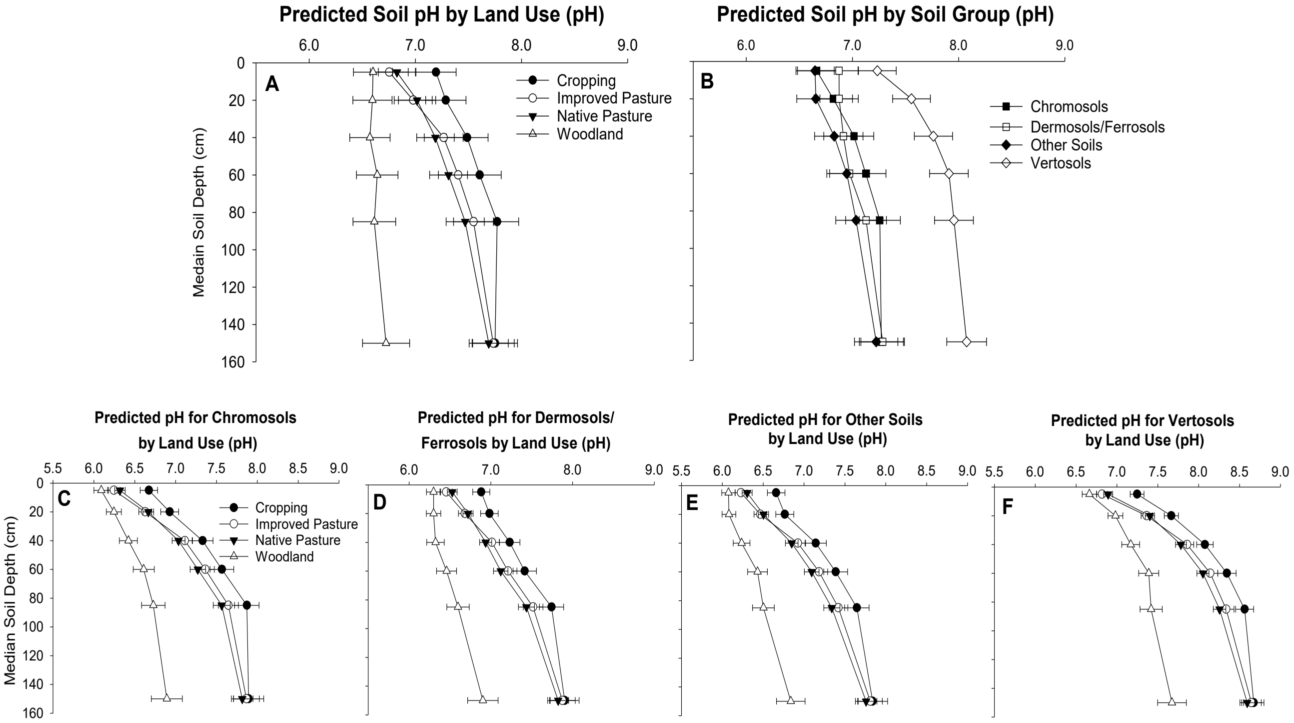


Figure 3.10: *Predicted soil pH by A. Land Use, B. Soil Group and C – F Soil Group × Land use Interaction. Error bars represent one standard error of predicted means.*

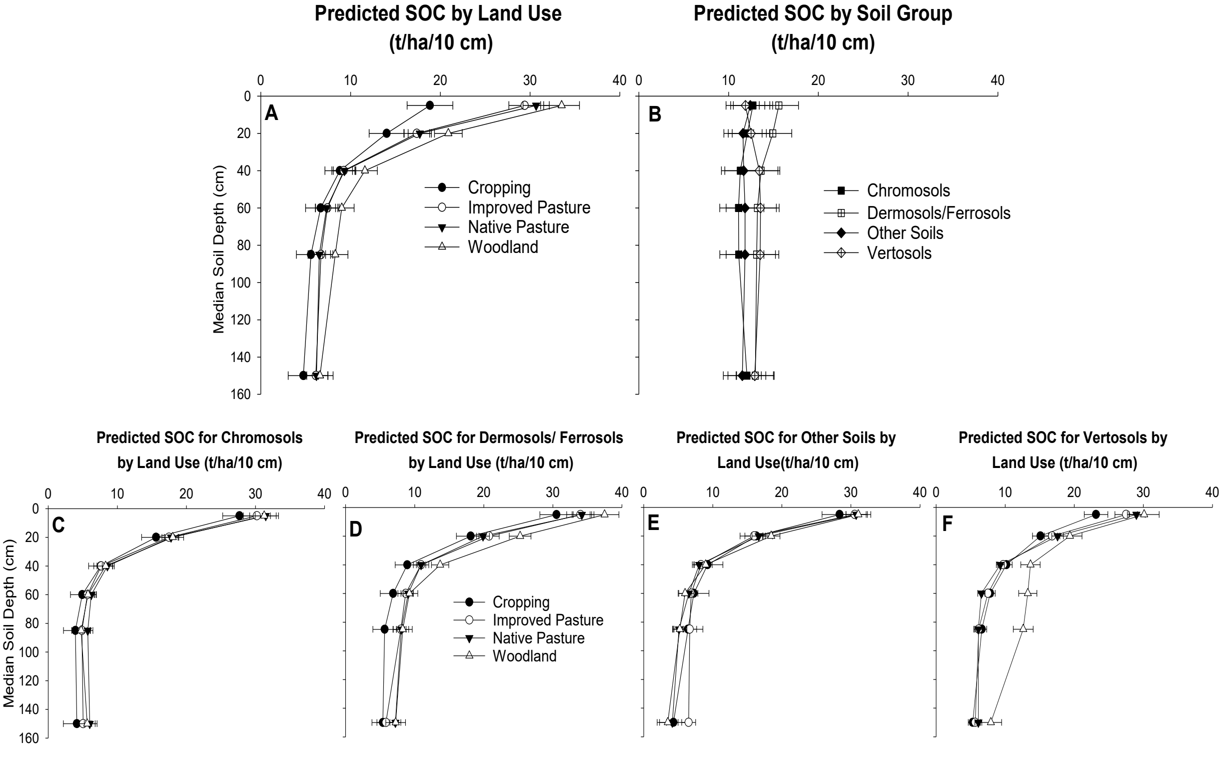


Figure 3.11: *Predicted soil organic carbon (SOC) by A. Land Use, B. Soil Group and C – F Soil Group × Land use interaction. Error bars represent one standard error of predicted means.*

Product of fertility factors

Soil organic carbon (SOC)

Ninety-three percent (93%) of the variability in SOC was explained by Land Use, with limited contribution from Soil Group (3%) or the interaction of Soil Group by Land Use (4%) (Table 3.2). Cropping had significantly lower SOC than pastures or woodland with a greater than 30% reduction in SOC (Figure 3.11 A). Although Soil Group and the interaction of Soil Group and Land Use explained little variation in the data (Figures 3.11 B – F), the distribution of SOC within Vertosols land use groups was significantly different for woodland compared with other Land Uses (Figure 3.11 F). Vertosols in woodland had significantly more carbon than Vertosols with other Land Uses for 20 to 85 cm median depth.

3.4 Discussion

This study used existing soil survey data to determine if major differences in soil profile physical and chemical parameters of the soil profiles were influenced by land use. The analysis based on land use as recorded in the SALIS dataset yielded no significant difference between native pasture and improved pasture. It may have been that the relatively short or highly variable period of time since a land use change, along with the overly broad definition of pastures, masked any fine detail differentiating native and improved pastures.

The results show that most differences in fertility parameters examined can be attributed to land use selection of favourable soils types and selection of better soils within soil groups for agriculture. If a plausible causal relationship were found between land use and soil fertility factors, then more reliance could have been made on SALIS data for further research on the impacts of native versus improved pastures on soils in the Liverpool Plains region. The most significant finding herein was that variation in SOC was overwhelmingly related to land use, and that cropping was strongly related to SOC decline across all soil groups. In terms of land use across all soil groups, woodland had 86% more predicted SOC than cropping, and both pasture types had 63% more SOC than cropping at 0 –10 cm depth. This is irrespective of the SOC in woodland (considered as an original value) for each soil group.

Use of simple fertility measures

The approach of using fertility factors to model soil fertility decline has been used on a broad scale with some success (Hartemink, 2006). Integrating common factors of plant nutrition (Bray P, CEC, pH, EC_e) and factors potentially effecting water supply to plants (soil depth, AWC, ESP and DP) with Soil Group was used for statistical purposes to examine Land Use effects on soil. Most of the nutrient parameters were determined predominantly by Soil Group or an interaction of Soil Group and Land Use, which does not support the first hypothesis for this study. An exception to this was pH which was equally affected by Land Use with woodland always having the lowest pH within each Soil Group. Vertosols always had the highest pH. In terms of simple fertility factors, Soil Groups could be ranked from lowest to highest in the following order: Other Soils, Chromosols, Dermosols/Ferrosols, and Vertosols. Unsurprisingly the most fertile lands (e.g. those with the highest AWC, P and CEC and with low surface ESP and EC_e) have been chosen for the more intensive land uses within each of the Soil Groups considered. Woodlands within each Soil Group consistently had the lowest fertility indicating that land selection has occurred not only for more the fertile Soil Groups, but also for more fertile lands within each Soil Group.

A preference for cropping on Vertosols is clear. This selection is commonly reflected in intuitive selection of land for agriculture or in more formal land and soil capability classification systems such as Murphy *et al.* (2004) or OEH (2013) where land suitability or quality is determined by both soil characteristics and topography. This is a reflection of requirements of the land use practices and shows that the European settlers chose the higher quality land for higher intensity and more productive land uses.

Soil carbon as a product of soil fertility

In terms of the above factors of soil fertility considered in this study, SOC should be highest in Vertosols and lowest in Other Soil, with carbon sequestration limited by soil fertility (Oren *et al.*, 2001; Kirkby *et al.*, 2014). Land use overrode the effects of soil type on SOC. In this analysis, woodland and both pasture uses were not distinguishable but cropping overrode the other factors as it had a lower SOC for each soil group. The depth of significantly greater SOC storage in Vertosols in woodlands is an important finding of this study and indicates that woodland on Vertosols potentially store 34% more SOC than any other soil in woodland to a depth of 85 cm or more. The deep SOC content for Vertosols in woodland are consistent with Mathieu *et al.* (2015), who found that Vertosols under woodland or forest had a deeper SOC storage capacity worldwide and, more locally, with Young *et al.* (2005) who found deeper storage of SOC in Vertosols compared with Chromosols.

The present analysis yielded similarities with published findings for SOC, specifically that native and improved pastures are difficult to distinguish from one another, and that cropping and woodland categories stand out from one another in terms of SOC in the topsoil layers (Ford and Wilson, 1981; Young *et al.*, 2005; Wilson *et al.*, 2008; Wilson *et al.*, 2010; Wilson *et al.*, 2011; Badgery *et al.*, 2013; Schwenke *et al.*, 2013; Rabbi *et al.*, 2014). These studies often show that the major differences in SOC are mostly confined to the top 10 cm of soils. Highly controlled studies of SOC on individual soil types often conclude that SOC in pasture (whether native or improved) is between cropping and woodland, and that effects of managed pasture type are difficult to distinguish. The coarser analysis and types of data used in this study showed that SOC in woodland and both types of pastures were statistically indistinguishable with the available data despite the spread of fertility ranges seen across the Soil Groups.

In terms of the broad fertility factors of AWC, Bray P, CEC, pH, Vertosols had the highest basic inherent fertility. Gray *et al.* (2016) and Banks and McKane (2002) mapped approximations of pre-European levels of SOC across NSW. Examination of both of these maps shows that the Vertosols were likely underrated in their ability to store SOC in the Liverpool Plains wider region. This may be because the modelling of Gray *et al.* (2016) was limited to 0 – 30 cm depth, above the point in the profile where the Vertosols under woodland have a higher capacity to store carbon at greater depths. Both maps clearly show higher SOC in the poorer (mostly sodic-texture contrast) soils of the western Liverpool Plains, where most of the soils are classified in this present study as Other Soils.

Recent land use changes

Cropping, which is predominantly carried out on the finer textured soils of the Liverpool Plains (mostly Vertosols), is relatively new to these soil types. Coarser textured soils were targeted for cropping from 1880 until the 1960's when sufficient mechanical horsepower required to cultivate Vertosols became available (Banks, 1998). Many of the coarser textured soils in areas with favourable slopes were cultivated for cereal crops for almost 100 years, before being progressively retired to voluntary native or improved pasture. Conversion of available Vertosol dominated soil landscapes to cultivation was virtually complete by the late 1980's. Interestingly, the late region-wide conversion to cropping on Vertosols and the reversion to both native and improved pastures on lighter soils is not reflected in this analysis, indicating that there has likely been some degree of recovery of SOC from a cropped land use to a pasture land use.

Of note within the Vertosol soil group is that the cropping soils have the highest Bray P. The surveys which contributed to this data set were generally completed prior to 2003. Phosphorus fertilisers were not used on cropped Vertosols at this time as the soils were thought to be sufficiently high in native P (Banks, 1995). In this case the higher P is once more a reflection of selection of better soils within the Vertosols for cropping.

Limitations of study

The dataset for this study is confounded with site selection issues, the broad land use definitions used, and inclusion of a large range of soil types. Unfortunately, the land use definitions from NCST (2009) and Milford *et al.* (2001) do not include site history. The length of time from native vegetation to cropping, from native vegetation to cleared pastures, from copping back to pastures, or whether the original vegetation was grassland or woodland are unknown factors that are likely to be highly variable, with timeframes from a few years to in excess of 100 years.

Transformation of original data firstly through a pedotransfer function to allocate BD values, then by splining to create equal and comparable depth intervals also has potential issues of creating data that may not represent the on-site conditions. It is likely that slight but significant differences, such as may exist between native or improved pastures, may be “drowned in regional scale data sets” such as the SALIS data set used (Peri *et al.*, 2016, p73).

Despite the apparent limitations of these data, there are highly significant differences between soil characteristics such as fertility factors and SOC which are related to land use selection, or caused by land use. The importance of the SOC differences is that they are sufficiently large to transgress soil type and that SOC can be explained using land use in a causal framework with simple data.

Future directions

The results of this study are limited to gross differences in SOC and, to a lesser extent pH, in response to land use change. In terms of understanding the effects of native and improved pastures on soils it is desirable to look for the detail that may be lost in regional scale datasets. In view of the published detailed land use comparison that has been made regionally on soil characteristics and land use, it appears that native pasture and improved pastures are seldom separated. This may be because trial periods are too short in duration, or they are focussed on production or solely on carbon sequestration, or there is no difference. This work lays the foundation for an in-depth comparison of soils in managed improved pastures and native pastures using a short chronosequence on a site with well-known history, rather than a trial.

3.5 Conclusions

The finding in this study that land use overrides the effects of soil type and minor climate variation in terms of SOC is highly significant. It shows the importance particularly of retaining undeveloped woodland and implies that re-establishment of trees on any soil type may result in higher soil carbon sequestration than any other activity. Of note, SOC is stored much more deeply in woodland Vertosols than in any other land use, with potential to store an additional 30% SOC in the upper 85 cm as compared with other soil types. The focus of most SOC research on 0 – 30 cm of soil may need to be reviewed in areas with extensive Vertosols where a woodland environment exists or is planned through strategic plantation.

The values determined in the present study, and in more detailed trials which examine SOC for pasture and woodland soils, could potentially be used as a regional benchmark for cropping systems in the Liverpool Plains wider region, given that the cropping lands possess potential to store more C than the relatively stable pasture or woodland soils in the region.

Most of the analysis shows that land has been selected for the most favourable basic soil fertility factors, with soil group and mixed effects dominating the analysis. Land use selection is reflected not only in selection of the most fertile soil group (the Vertosols) but also selection within all soil groups of the most fertile lands for the most intensive land uses.

Whilst other studies address specific soils and land uses, this more generalised approach shows the extent of the overall impact of European land use on a large region in terms of SOC, independent of soil type considerations.

Chapter 4. Study Two, Part One: Soil profile changes induced by tropical grass pastures on sodic-duplex contrast soils in Northern NSW

4.1 Introduction

Sodic-duplex soils with a strong-texture contrast between A and B horizons including Sodosols and Sodic Chromosols (Australian Soil Classification, ASC) Isbell and NCST (2016) are extensive in the north west NSW slopes and plains on Permian to Cretaceous sandstones (Figure 4.1). Equivalent soil classes in the World Reference Base are Solonetz, Lixisols or Luvisols (IUSS, 2015) or soils with a Natric Great Group in the United States of America classification (USDA, 2014).

The A horizons in these soils are typically acidic loamy sands and sandy loams, overlying a light to medium clay with an abrupt texture contrast. Deep subsoils are generally neutral to highly alkaline (Banks, 1995; Pengelly, 2009). A strongly bleached, light textured $A2_e$ (E) horizon, characteristic of the Sodosols and Sodic Chromosols reflects both the dispersive nature and heavier texture of the upper B2 horizon. The dense upper B2 horizon slows or prevents drainage and thus soil water infiltration and plant root penetration to deeper soil layers (Rengasamy *et al.*, 1995; Cotching *et al.*, 2001; McKenzie *et al.*, 2004; Jassogne *et al.*, 2007). As a consequence, the upper B2 horizon restricts wetting of deeper B horizons. The light-textured A horizons generally have low plant available water and are prone to both waterlogging under wet conditions and water deficit in dry conditions. These characteristics translate to low inherent resilience in farming systems in a hot and dry climate with irregular rainfall.

In comparison with similar soils in the Northern Hemisphere or tropics which have seasonal conditions of waterlogging and slowly permeable B horizons, the sodic-duplex soils discussed in this study are subjected to much more random weather patterns. When these soils are cored directly following rare prolonged rainfall or flooding events, they are often encountered with no apparent soil moisture in B horizons and liquefied A horizons. For these reasons, it is often recommended that these soils be left uncleared or undeveloped (IUSS, 2015).

Upper B2 and underying B2 horizons of sodic-duplex soils typically have large columnar or prismatic ped structure with few secondary peds (Walsh and Humphreys, 2008). The columnar ped fabric is dominant under both uncleared native vegetation and in volunteer native pastures and cropping systems. Roots of native vegetation do not readily penetrate through the upper B2 horizons and tend to follow topsoil filled spaces between peds to depth, rather than penetrate the dense clays of sodic B horizons (Walsh and Humphreys, 2008).

Sodic-duplex soils and pastures in Northern New South Wales

The Namoi Catchment has approximately 1,800,000 ha of sodic-duplex soils on slopes of less than 10%, of which an estimated 1 million ha is cleared grazing land (Figure 4.1) (SoilFutures, 2008). These soils were extensively cleared in the mid to late 19th century for winter cereal cropping in rotation with lupins or longer lucerne fallows. Owing to declining productivity, poor crop performance and unreliable climate, most of the sodic-duplex soils in the Namoi Catchment were retired from cropping to voluntary native pasture in the 1970's and 1980's (Banks, 1995).

As a result of this legacy of large scale clearing and cropping, most native grass pastures on sodic-duplex soils in the Namoi Catchment are volunteer pastures. There are very few sites on low sloping lands containing native pastures that can be considered never to have been cultivated or otherwise disturbed. Woodland sites with sodic-duplex soils are relatively common in State Forests and National Parks of the region. For this reason, this research takes a space-for-time substitution approach to detailing observed differences in soils under native and managed tropical pastures. As such, this study can be compared with a short-term chronosequence of soil development (Huggett, 1998; Sauer, 2015).

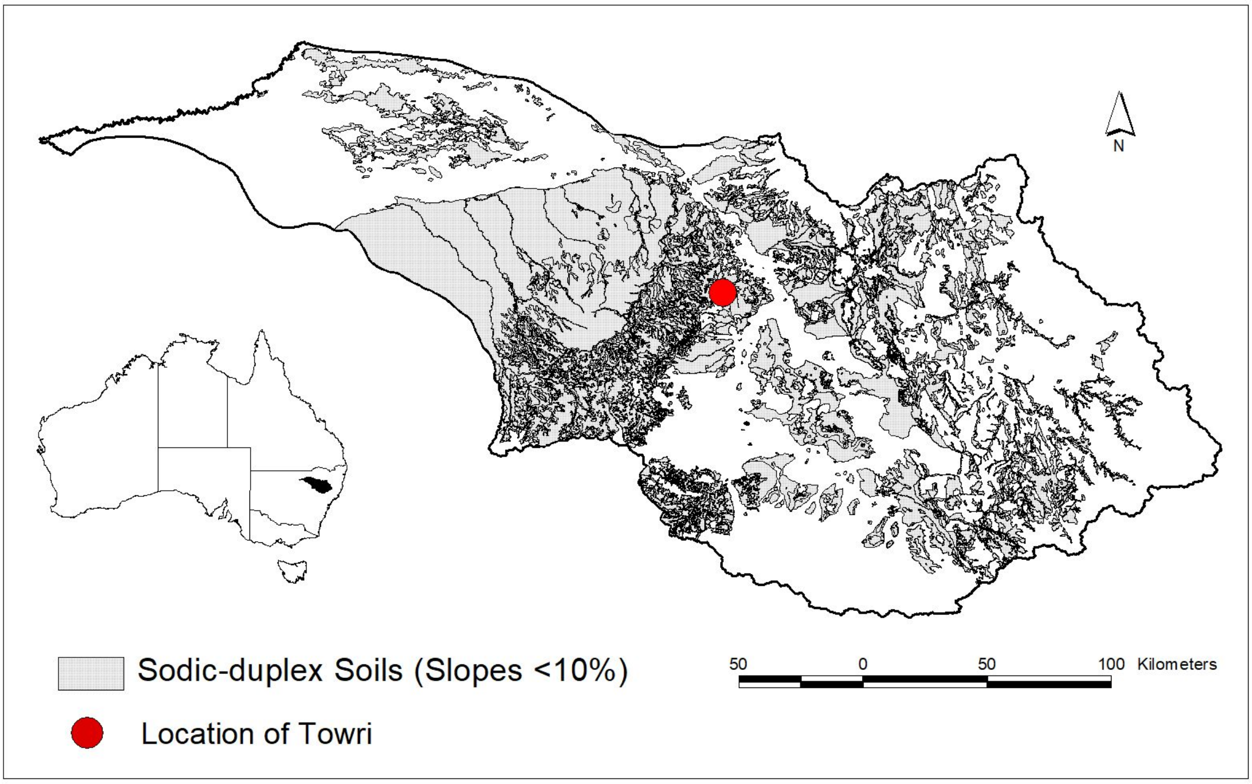


Figure 4.1*: Sodic-duplex soils of the Namoi Catchment and location of Towri study site.*

Benefits of tropical grass pastures

Tropical grass pastures with winter active legumes have increasingly been grown on sodic-duplex soils in north NSW since the early 1990's due to the superior performance and feed value of these species (Harris *et al.*, 2014). Production advantages of tropical grass pastures include dry matter production of 11 – 19 t/ha/yr compared with 4 – 9 t/ha/yr for native pastures and a greater proportion of digestible protein (Harris *et al.*, 2014). Tropical grass pastures typically have significantly deeper roots (Murphy, 2014; Murphy *et al.*, 2018). Murphy *et al.* (2010) showed that rooting depth under sown tropical grass pastures was related to deeper soil water storage and use which coupled with greater water use efficiency (herbage produced per unit of rain) and feed value has enhanced production in grazing systems.

Despite reports of deeper root systems and increased soil water storage in tropical grass pastures, comparative studies have shown little difference in deep soil physical and chemical attributes between soils under native and tropical grass pastures and only minor differences in SOC and BD in surface and near surface soil layers (Young *et al.*, 2005; Wilson *et al.*, 2008; Wilson *et al.*, 2010; Wilson *et* *al.*, 2011). Most of these studies have concluded that native and tropical or “improved” pastures exhibited an intermediate equal rank between cropping (low) and woodland (high) in terms of SOC.

International comparisons of SOC and BD under managed grazing systems show highly variable results for these attributes, particularly in the intermediate rainfall zones (400 – 850 mm) of the world (Piñeiro *et al.*, 2010). In Australia, comparisons of cropping and grazing systems for three dominant soil types in Tasmania found that SOC was significantly greater in grazing systems than in cropping systems, and that soil physical properties of BD were significantly smaller in grazing systems with consequently higher porosity for each soil type although the soils were examined only to 30 cm depth (Cotching *et al.*, 2001, 2002b; Cotching *et al.*, 2002a).

Orgill *et al.* (2014) determined that the focus of much research on the top 30 cm of soil may be misleading as significant carbon stocks can occur below 30 cm in some soils. Similarly, investigation of near surface soils cannot indicate changes in physical structures or chemistry deeper in the profile. Many studies of the effects of pastures on soil structure have concentrated on the surface 0 – 30 cm, where rapid changes (within 10 years) in structure from poor or massive to highly aggregated, pedal and porous are often observed following establishment of sown pastures (Amézquita *et al.*, 2004; Daniells, 2012; Kösters *et al.*, 2013).

Conversion of land to pasture from cropping and concomitant increase in organic carbon is associated with formation of increasingly stable aggregates even in light texture topsoils. Cementing of individual soil (sand and silt particles) in these soils is associated with binding agents from fungal hyphae, the development of dense fine roots and complex organics such as carbohydrates, polysaccharides and aliphatics (Tisdall *et al.*, 1978; Haynes and Swift, 1990; Dinel *et al.*, 1992; Debosz *et al.*, 2002; Mardhiah *et al.*, 2014; Stumpf *et al.*, 2016).

Management and changing upper B horizons in sodic-duplex soils

Changes in the subsoil characteristics of Australian sodic subsoils can be achieved using expensive amelioration processes. For example, deep ripping Dynamic Lifter®, a chicken manure based organic fertiliser, into upper B horizons, has been shown to result in significantly increased macroporosity (pores > 30 μm), saturated hydraulic conductivity and root penetration (Gill *et al.*, 2009). A macroporosity of < 0.1 m^3/m^3 limits root penetration which may otherwise initiate structural or porosity changes (Engelaar *et al.*, 2000). Cresswell and Kirkegaard (1995) found no evidence for macropore creation by root systems (or “biological drilling”) of crop roots into B horizons of Sodosols. McCallum *et al.* (2004) subsequently found that lucerne and phalaris pastures can have a lasting effect on subsoil porosity in Sodosols through the formation of large biopores from decaying larger roots. However, structural changes or changes in ped size in sodic subsoils associated with deep rooted tropical grass pastures have not been reported.

Clark *et al.* (2007) demonstrated that addition of soluble C and N to sodic subsoils significantly increased biological activity. Increased biological activity, particularly mycorrhizal fungi associated with plant roots has been shown to both build and increase stable soil structure (Bronick and Lal, 2005; Daynes *et al.*, 2013).

Improvement of upper B2 horizon and deeper B horizon structures in association with the introduction of tropical grass pastures to previously farmed land or volunteer native pastures land have been observed in the field and recorded at individual paired soil profiles on sodic-duplex soils for more than ten years in the Liverpool Plains region of NSW (Banks, 2005; SoilFutures, 2005). No detailed studies of deep soil structural changes associated with the introduction of tropical grass pastures have been done in this region.

Study focus, objectives and hypotheses

The objectives of the present study were to describe the differences between sodic-duplex soils under established tropical and volunteer native grass pastures. Using common soil assessment procedures such as soil profile field description and measurements, soil physical and chemical measurements, tropical pasture induced soil profile changes in sodic-duplex soils were evaluated. This study tested the hypothesis that: "tropical grass pastures can change whole soil profiles through the stepwise sequence:

1. Development of more abundant and deeper root systems;
2. Changes to soil pore space and soil structure down the profile as a consequence of increased root activity; and
3. Consequent changes to soil chemical characteristics, including pH, exchangeable cations, macronutrient distribution in the soil profile (N and P), and SOC."

Site selection and history

The Towri property is in the mid Namoi Catchment in the Northern Slopes and Plains NSW, Australia, located on Jurassic-Cretaceous Pilliga sandstone (Figure 4.1). The site has a dry sub-humid climate and an annual rainfall of approximately 600 mm with annual evaporation exceeding rainfall by a factor of three. Climate records for the nearby Gunnedah Research Centre show average annual rainfall of 642 mm and evaporation of 1937 mm, with large seasonal variability (Banks, 1995). The region has hot dry summers with average summer temperatures in January ranging from 19° C to 32° C, and winter temperatures in July from 3.5° C to 17.8° C (Pengelly, 2009). Rainfall is almost never distributed evenly or seasonally. Consequently, it is estimated that the topsoils at the "Towri" study site may only be wetted to the depth of the upper B horizons on 2 – 3 occasions per year. Some years this is unlikely to happen at all. Deep water penetration and storage at the site is limiting to both pasture and crop production.

The Towri site is on a simple slope of < 3% and dominated by deep to very deep (1.5 – 5 m) Brown Sodosols and Bleached Brown Chromosols (with sodic deep subsoils). The region including Towri was cleared in the 1880's and cultivated continuously for winter cereal crops with occasional lucerne or lupin rotations (Willala Bicentenary Committee, 1988) until the year 2002 (Landholder Records, nd) when cropping ceased. In 2002 the site was subdivided and fenced with the southern field retired to voluntary native pasture and the northern field sown to a mix of tropical grass pasture species. The history of land use at the site and the general area is well documented and representative of regional historical land use.

The site on Towri was selected for its known management history and has a good side by-side comparison of established sown tropical grass pastures and volunteer native pastures. The low and straight slope (neither concave nor convex) and uniform aspect of the site is favourable for observing potential soil changes over time as it limits lateral movement processes and preferential erosion (Sauer, 2015).

The layout of the site offered the ideal opportunity to study changes. Unfortunately, it is unrealistic to set up an experiment at this opportunistic site. Instead, the site offers an ideal space for time substitution study.

Towri pasture species

Tropical sown pasture species in the northern field of the site included Premier digit (*Digitaria eriantha* cv. 'Premier'), Katambora Rhodes grass (*Chloris gayana* cv. 'Katambora'), Bambatsi panic (*Panicum coloratum* var. *makarikariense*) and CONSOL lovegrass (*Eragrostis curvula* var. *conferta* cv. "CONSOL"). Winter legumes included Serradella (*Ornithopus sativus*), Arrow leaf clover

(*Trifolium vesiculosum*), Hykon rose vlover (*Trifolium hirtum* var. *hykon*), and various sub clovers (*Trifolium subterraneum*).

Native volunteer pastures in the southern field of the site included slender bamboo grass (*Austrostipa verticillata*), windmill grasses (*Chloris* spp.), wallaby grasses (*Austrodanthonia* spp.), three-awned spear grass (*Aristida vagans*), red grass (*Bothriochloa macra*), pitted blue grass (*Bothriochloa decipens*), love grasses (*Eragrostis* spp.), barbed-wire grass (*Cymbopogon refractus*), wire grasses. (*Aristida* spp.), Queensland blue grass (*Dicanthium sericeum*) and panic grasses (*Panicum* spp..). Native legumes were dominated by medics (*Medicago spp.*), with some voluntary invasion of the improved pasture legume, Serradella (*Ornithopus sativus*). This native species mix was typical of light textured topsoils in the region (Banks, 1995; Pengelly, 2009).

Both suites of grass species listed above are generally considered to be summer growing grasses.

Fertiliser History and Pasture Management

The tropical pastures are responsive to phosphate fertiliser and had been fertilised with single superphosphate (600 kg/ha) and chicken litter (2,000 kg/ha) since establishment. This is equivalent to addition of 69 kg/ha P, 154 kg/ha Ca, 164 kg/ha S, 29.4 kg/ha K and 78 kg/ha TN (Rose, 2004; Griffiths, 2011). Volunteer native pastures had not been fertilised. Fertilisation of native grass pastures including *Bothriochloa* spp. and *Aristida* spp. has been shown to have limited or detrimental effect on native pasture production in the region (Chalmers *et al.*, 2005).

Both native and tropical grass pastures sites are rotationally grazed at the same stocking rate, and managed to maintain 100% groundcover. Observed effects of tropical pastures cannot necessarily be separated from the fact that they are fertilised. As this is typical management for the region it is considered a reasonable comparison of two managed systems.

4.2 Methods

The hypotheses that tropical grass pastures may significantly change sodic-duplex soil was examined at Towri property where immediately adjacent sites with tropical and native pastures were available on sodic-duplex soils. Due to a common site history, soils were assumed to possess the same characteristics and classification prior to pasture establishment.

Comparison with reference soil profile data on sodic-duplex soils

One potential criticism of this work is that the soils chosen to represent native pastures may not be typical of the region. To asses this potential the NSW Soil and Land Information System (SALIS) was accessed to obtain comparable field data for the same ASC classes in the greater region surrounding Towri (Figure 4.2). Soil profiles selected from the database had sandy or sandy loam topsoils, a conspicuously bleached A_2 horizon (E horizon) and were classified as Sodosols or Bleached Chromosols.

This enabled comparison of similar volunteer native pasture and woodland sites on the same parent material within the same climatic zone as the Towri site. Soil description data for roots, field pH, and clay content were compared to establish that the soils at Towri were representative of the region. Clay contents were allocated to field texture grades using NCST (2009). Owing to most of the soil descriptions for sodic-texture contrast soils in SALIS being done on 50 mm cored soil profiles, gross soil structures were unable to be compared reliably. Managed tropical pasture soils were not compared with reference soils as there was insufficient data on such pastures represented in the SALIS database.

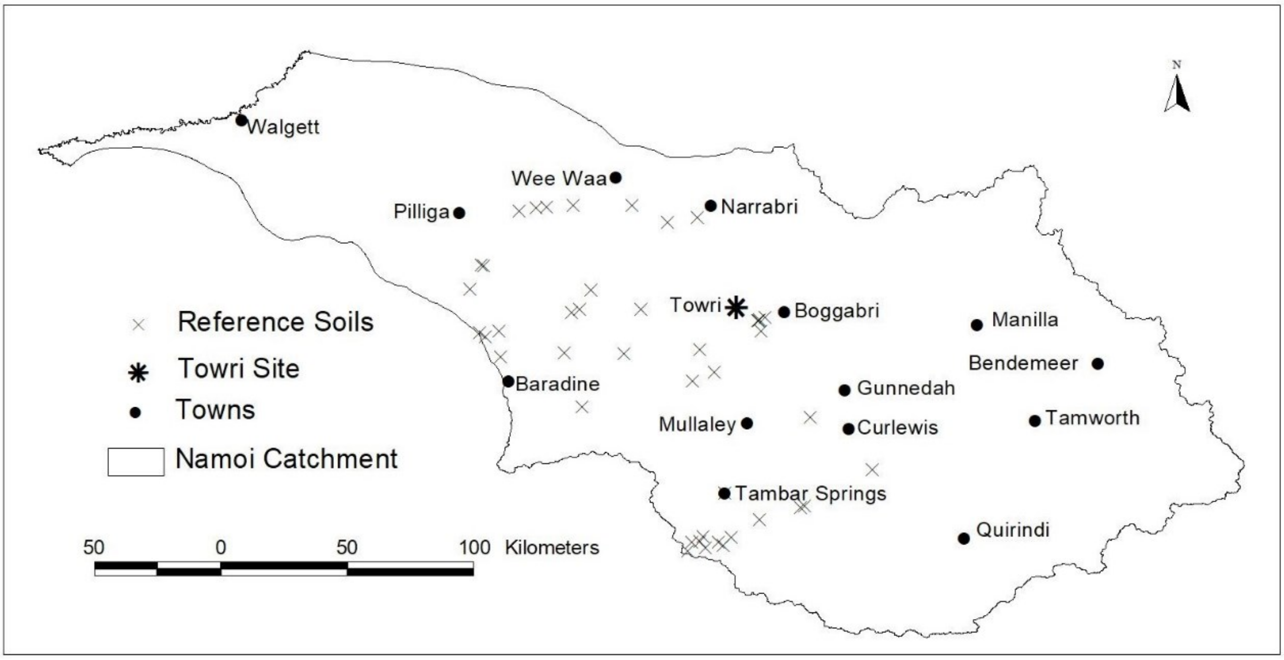


***Figure 4.2**: Location of reference sodic-duplex soil profiles*

Field data description techniques

A comparison of existing soil profile descriptions for the same soil type was made, to ensure that the soils investigated on the main site for this paper, Towri, are representative of the wider regions' sodic-texture contrast soils. Soil profile description methods were consistent with Towri facilitating a simple comparison (NCST 2009).

Focus on upper B2 horizons

The upper B2 horizon was considered the rate limiting horizon for deep soil water penetration in the sodic-duplex soils and a considerable barrier to root penetration of many crops and perennial plants (Heng *et al.*, 2001; Jassogne *et al.*, 2007). Unfortunately, the natural variability of overlying horizon thicknesses prohibits strict incremental sampling from accurately defining characteristics of this important horizon. Potential changes in this horizon may have large potential impacts on the whole soil profile. Therefore, clear unmixed samples were required to characterise this horizon for soil porosity, chemistry and bulk density.

Site location and sampling pattern – paired soil profiles

Paired observations were made across the fence line separating each pasture type at distances up to 1.2 km from the western end of the fence. Ten pairs of soil pits were excavated to 1.5 m in February 2016 in each land use and numbered as shown in Figure 4.3. As there is a road separating the tropical grass pasture from the native pasture at the site, cross-fence photographs were not possible to obtain. Figures 4.4 and 4.5 show typical fence line comparisons of the pasture types from nearby on the property. These sites did not have the 14 year history of the site chosen for this research but offer a good visual comparison of the pastures.

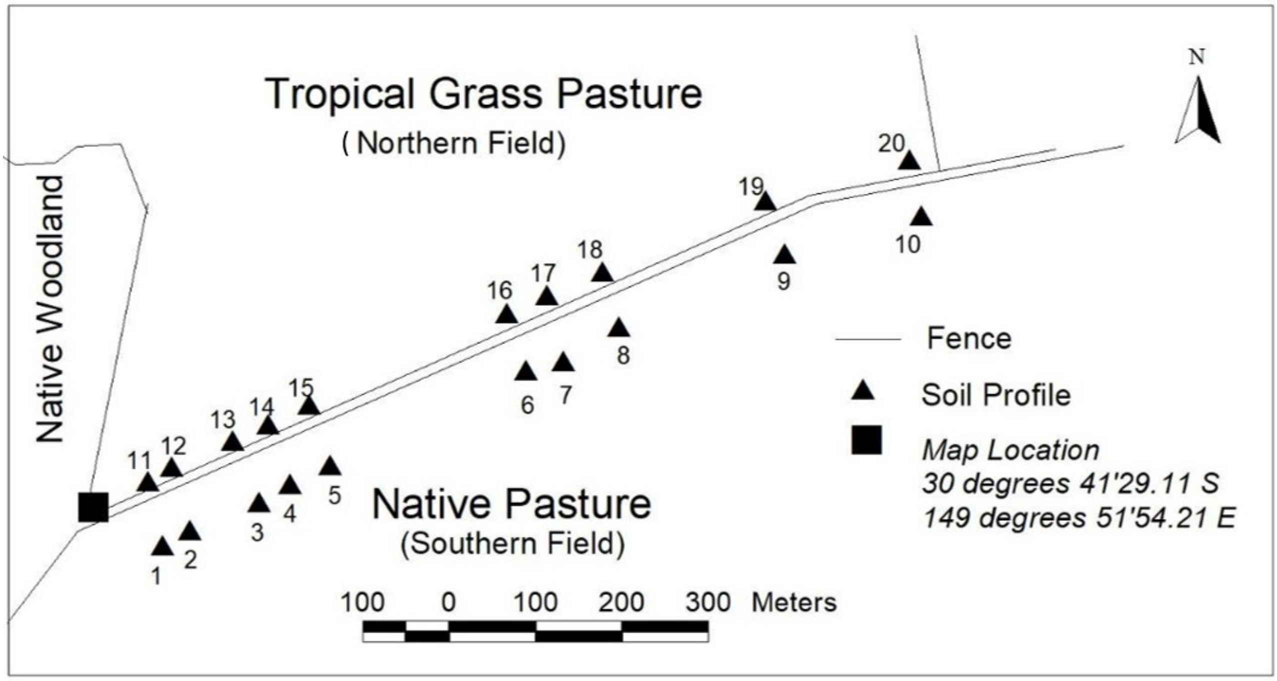


Figure 4.3: *Location of soil profiles on Towri Study Site*

Figure 4.4: *Tropical grass pasture foreground compared with native pasture (background over fence) at Towri.*

***Figure 4.5**: Native pasture (right hand side of fence) compared with tropical grass pasture (left hand side of fence) at Towri.*

Soil profile description and field data recording (0 – 150 cm)

Note that although whole profile sampling for chemistry was originally done in January – February 2016, initial sample integrity was unfortunately compromised at the laboratory and more limited soil chemistry sampling had to be repeated in July 2016. Sampling for soil chemistry to full depth of pits could not be undertaken due to heavy rain and pit collapse.

During the February sampling period, soil profiles were dry. During the July sampling period, the surface horizons of soils under native pasture were waterlogged, having little bearing strength and highly fluid A horizons overlying dry B horizons, whilst soils under tropical pastures were moist throughout but not waterlogged and were highly trafficable. B horizons under tropical pastures were moist and A horizons had firm consistency. B horizons below the upper B in native pastures were dry with A horizons becoming fluid when cored.

To ensure that soil profile description was comparable with other available reference soil profile information, standard profile description techniques were used for field characteristics. Each site and soil profile was described by horizon according to the National Committee on Soil and Terrain (NCST, 2009). The Australian Soil Classification (ASC) was used to classify the soil profile (Isbell and NCST, 2016). Each pit was sampled for aggregate stability in water (ASWAT) testing to 150 cm.

Particular care was taken to accurately record dominant ped size classes for each soil horizon (NCST, 2009), and visual fine (< 2 mm) root abundance (NCST, 2009). Root abundance was recorded per 10×10 cm (100 cm^2) within horizons in the following categories 0, 1- 10, 10 – 25, 25 – 100, and > 100 roots/10×10 cm. Dominant ped size and root count ranges were converted to median values within their range. Field pH was measured using the Raupach indicator (Raupach and Tucker, 1959) and recorded by horizon.

Soil physical properties

Dispersion – aggregate stability in water (ASWAT)

Soil samples for dispersion were taken directly from the soil profile in 0 – 10, 10 – 20, 20 – 30, 30 – 50, 50 – 70, 70 – 90, 90 – 110, 110 – 130 and 130 - 150 cm increments and stored in open weave calico bags for transport and air drying. ASWAT tests were conducted on air dried samples from soil pits to 150 cm depth, ranking soils from 0 (non-dispersed, aggregate stable) to 16 (highly dispersive soil) (Field *et al.*, 1997).

Bulk density and porosity

Sampling was restricted by equipment refusal for BD in subsoils at the time of sampling with each subsoil BD sample taking up to 2.5 hours to collect. Soil samples for BD determination were collected using intact cores of approximately 12.4 cm diameter and 8.6 cm length driven vertically into platforms cut into pit walls with a drop hammer. Paired BD cores were taken from 0 – 30 cm as this was expected to have higher variation than subsoils. Paired BD cores were collected for depths of 0 – 10, 10 – 20, 20 – 30 cm with an additional single core taken at 65 – 75 cm. Soil cores were oven dried at 105° C, moisture content recorded, and BD determined as per McIntyre and Loveday (1974).

Total soil porosity was calculated using the formula

$$\phi = (1 - \rho_b/\rho_s)$$

where ρ_b is the known dry BD and ρ_s is an assumed particle density for quartz of 2.65 g/cm^3 (or 2.60 g/cm^3 for organic topsoils) (Danielson *et al.*, 1986). The lesser particle density was used for the 0 – 10 cm depth slice where organic matter was highest. Any error in calculation of porosity does not impact on relative differences between paired sites.

Pore size distribution

Tempe cell samples were used to determine low pressure water release curves to derive the distribution for pores > 3 μm. Sampling for Tempe cells was limited by number of cells available and it was decided to have duplicate samples of potentially highly variable topsoils and a single sample of the top of the upper B2 horizon. Intact paired Tempe Cell cores (approximately 8.6 cm diameter and 6 cm length, volume 348 cm^3) were obtained for at 0 – 6 cm for each pit. A single core was collected from the top of the upper B2 horizon (depth variable) at each pit. Coring rings were carefully driven into the soil using a hammer driven dolly designed to fit the cores and allow for surplus soil to extrude through the top of the cell for trimming if necessary. Tempe cell samples were wrapped with polythene to provide stability during transport.

Laboratory analysis of Tempe cells were based on the procedure described by Flint and Flint (2002). Briefly, intact soil cores were removed from polythene wrapping and carefully placed into the base of a Tempe Cell. Open Tempe cells were placed in a bath of deionised water with the porous ceramic plate submerged. Water was allowed to rise through the ceramic plate until soil samples in the cores reached saturation typically after 48 – 72 hours. Samples which failed to wet thoroughly during this period were transferred to a water bath in a partial vacuum to enhance water uptake and ensure saturation.

Tempe cells with saturated cores were sealed and attached to a pressure manifold to extract water up to 1 bar (10 m water column). Extracted water was collected using volumetric flasks placed under the core to collect the outflow from each cell. Saturated cells were subjected to decreasing pressure potential from -20, -50, -80, -100 and -200, -300, -500 and -1,000 cm height of water column). Reliable potentials (> -200 cm) were obtained using a 2.4 m water column. The pressures were adjusted by inserting a bleed tube to measured depths below the water surface in the water column. Lower potential (< -200 cm water column) were obtained using standard fine-resolution pressure gauges. Pressures were maintained until soil water content of the cores reached equilibrium (i.e., outflow ceased).

Once cores had equilibrated at 1 bar, cores were weighed and oven dried (105° C) to determine dry mass for the calculation of BD and pore size distribution. Field capacity or soil water content at -100 cm water column (Vimpany *et al.*, 1987) was calculated as the sum of the water extracted from the Tempe Cells between 100 and 1,000 cm of water, and the remaining soil water in the core (water loss at 105° C).

Equivalent pore sizes for each potential were calculated using the capillary rise equation given in Brady and Weil (2013). Porosity ranges for macro, meso and micro-porosity are defined in this study as > 30 µm (above field capacity) for macropores, 3 – 30 µm (Engelaar *et al.*, 2000) for mesopores and < 3 µm for micropores.

Porosity groupings were calculated using the following pressure increments: 0 to -80 cm pressure measured > 30 µm (above field capacity) (macropores); -100 to -1,000 cm pressure measured 3 – 30 µm (mesopores); and the remaining water held in Tempe cells was used to calculate < 3 µm (micropores). Air filled porosity, or the difference between porosity measured in Tempe cells, and that calculated from BD measurements, was added to the micropore measurement. The air filled porosity was assumed to be in very large pores which would fill by rainfall from above but not from contact with a porous plate from below as in a Tempe cell.

Relative clay content determination of upper B2 Horizon

As the upper B2 horizon in sodic-duplex soils is the rate limiting horizon for soil water infiltration to deeper horizons, it was necessary to check that upper B2 samples had similar textural characteristics. It was crucial to ensure that differences in other physical and chemical parameters such as porosity were not due to textural variation.

In addition to field soil texture description given in section 2.3, laser refraction was used to determine the particle size distribution in upper B2 horizons of both the native and tropical grass pasture. No pre-treatment was required for organic matter or carbonates to obtain dispersion (Bowman and Hutka, 2002). Paired samples from upper B2 horizon samples were air dried, ground and passed through a 2 mm sieve with the > 1mm fraction weighed. Approximately 5 grams of each sieved sample was suspended in 0.5 L of deionised water, using a 750 watt sonifier for 15 minutes. Dispersed soil was analysed using a Malvern Mastersizer 3000® to obtain a continuous particle size output ranging from 0.01 – 1000 µm particle diameter (Ryżak *et al.*, 2007; Sochan *et al.*, 2012). These data were added to the 1 – 2 mm fraction from the sieves and converted into a standard Australian particle size distribution per Hazelton and Murphy (2007).

Soil chemical properties

Incremental soil samples (0 – 90 cm) for chemical analysis were taken in July 2016 following an unusually wet period with 230 mm of rainfall during a period of less than 8 weeks. Twenty surface samples (0 – 10 cm) were collected with a tiling spade within a 5 m radius of each soil pit. The spade was used to cut a 10 × 10 cm column of soil, which was then sliced at 10 cm lower depth. Topsoil samples were bulked to obtain one composite sample representing the pit from which the soil was described earlier. Because pits were flooded, a 50 mm diameter corer was used to collect samples to 90 cm depth 5 m from soil pits. Three cores were taken for each pit and sectioned into 10 – 20, 20 – 30, 30 – 50, 50 – 70, and 70 – 90 cm depth increments and bulked for each depth slice. Additional core samples were taken from the top 20 cm of the first B horizon (upper B2 horizon) in each pit (depth variable). Soils were transported in sealed plastic bags, then air dried (40° C) for analysis.

Dried samples were ground, then sieved to pass through a 2 mm mesh in preparation for soil chemical analysis. All chemical analyses were carried out by the Analytical Services Unit at The University of Queensland. Procedures were based on the methods described by Rayment and Lyons (2011) in the handbook "Soil chemical methods: Australasia". Soil testing included soil pH in water (1:5), electrical conductivity (EC 1:5), Colwell P, exchangeable bases (Ca, K, Mg, Na, and Al). and cation exchange capacity (CEC), total organic carbon (TOC) and total nitrogen (TN). TOC and TN measures were multiplied by BD to provide the stocks of TOC and TN on a volumetric basis. EC (1:5) was converted to the equivalent of EC (saturated extract) or EC_e by multiplication through a conversion factor related to field texture (Slavich and Petterson, 1993).

Conversion of horizon data to equivalent depth intervals

Ped size, root abundance and field pH for Towri and SALIS soil profiles were recorded by soil horizons which have highly variable depths and thicknesses (NCST, 2009). These data were difficult to statistically compare with the depth increment data for soil physical and chemical properties. Both the Towri data and SALIS soil profile descriptions were processed using a data splining technique. Bishop *et al.* (1999) and McBratney *et al.* (2000) have demonstrated that this limitation can be overcome by using an equal-area quadratic spline, which uses polynomial equations to generate incremental depth soil data from horizon data. Horizon based ped size, clay content, root abundance and field pH were transformed to incremental depth data using "Spline Tool" developed by Jacquier and Seaton (2012).

Statistical analyses

Soil properties were analysed by treatment (native and tropical grass pastures) and distance from the western end of the fence using a General Linear Model. Treatment was considered a fixed effect and distance along the fence a random effect in the analysis. These statistical procedures assume equal variance, however some variables appeared to have unequal variance (e.g. average ped size). An F-test was used to test for equal variance and where that assumption was violated, a paired t-test was used instead of ANOVA.

A non-parametric Sign test (Dixon and Mood, 1946) was conducted on the differences between paired individual measurements between native and tropical pastures for some analytes to determine if there was a constant direction of difference where comparison of means using ANOVA had not found a significant difference (usually due to large variation about the means). Sign tests results are given in the text in those cases and indicate that one measurement is always higher or lower on one side of the fence compared with the other despite large variation about the means on each side of the fence.

All statistics were calculated using Minitab 16®. All depth increments, as well as upper B2 horizon (variable depths) were summarised as graphs in Figures 4.6 - 13. P values ≤ 0.05 were considered significant. If $0.05 < p \leq 0.10$ then it was considered that a trend was observed.

4.3 Results

Comparison with reference sites

A comparison of field measurements in reference soil profiles from the SALIS database with the Towri native pasture soils shows that the sodic-duplex soils at Towri are similar.

Soil clay contents in reference soil profiles was equivalent in near surface depths (0 – 20 cm) and up to 20% higher for depths 30 – 100 cm (Figure 4.6 A). The texture contrast for all soil profiles was highly visible down the soil profile for both reference and Towri soil profiles. The Towri B horizons had higher clay than undisturbed reference sites, but all data show that the soils are still texture contrast soils.

Soil field pH had the same distribution pattern down the profile to 110 cm for native pastures in the reference set and at Towri, and was slightly more acid in woodland reference sites (Figure 4.6 B). The reference soil field pH below 110 cm become more acid, however this is a reflection of encountering parent material in many of the reference sites at this depth (SALIS database notes).

Towri native pastures had a slightly greater abundance of roots than reference pastures, and a significantly higher root abundance than in woodland reference sites (Figure 4.6 C). This was assumed to be a reflection of highly variable sampling times in the public database, and management in terms of pastures, and a competition artefact in woodland soils where tree roots compete with grasses for moisture.

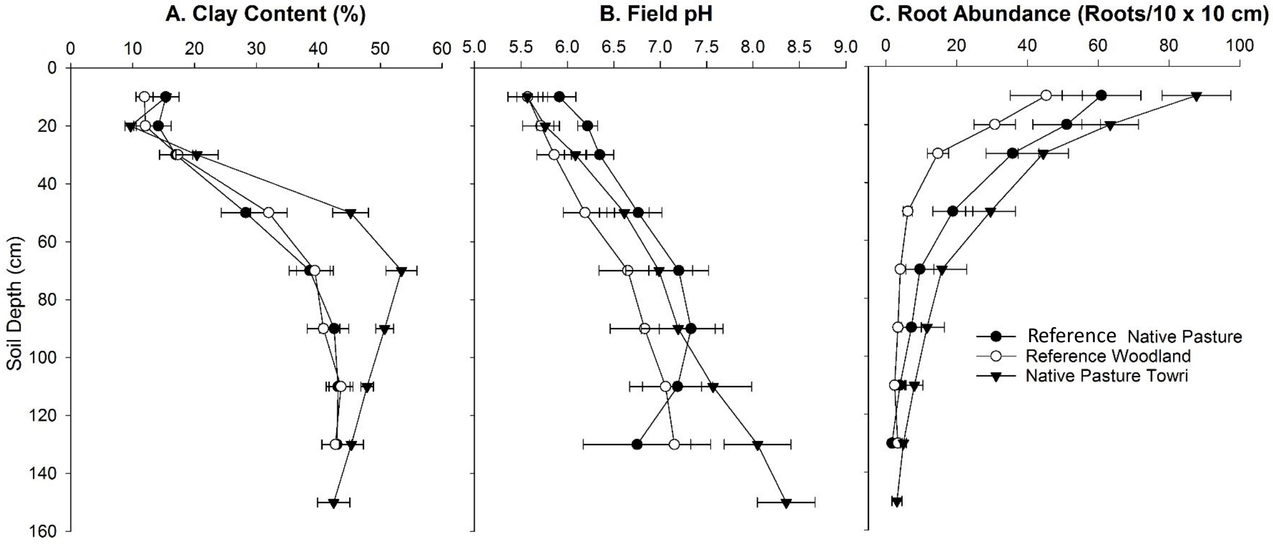


Figure 4.6: Comparison of field measures at Towri with reference soils. One standard error of the mean as a measure of variation within a treatment is shown as horizontal bars

Field properties

Soil profile descriptions were recorded on NSW Soil Data Cards and entered into the NSW Soil and Land Information System (SALIS). The full field data set is registered in NSW as Soil Survey number 1005323, where a permanent online record of soil profiles is stored. The full dataset of field records and soil profile descriptions may be downloaded from the NSW soil data portal, E-Spade at http://www.environment.nsw.gov.au/espde2webapp.

Soil Profile Description

Soils in both the tropical and native pastures included a mix of Vertic Mottled Subnatric Sodosols and Bleached sodic Brown Chromosols (Isbell and NCST, 2016). An example profile pair from Towri is given in Figure 4.7. The ESP of the upper 20 cm of the B horizon was generally non-sodic or slightly sodic (ESP > 6 and < 15) (Isbell and NCST, 2016). All soils were sodic at depth. Both the Sodosols and Chromosols had similar physical appearance with a strongly bleached A2$_e$ (E) horizon and a sharp to abrupt (< 5 mm – 20 mm) texture contrast from sand or clayey sand A2$_e$ to medium-heavy clay B horizon dominated by large columnar peds. The Vertic subgroup in the ASC was included because of the presence of slickenside ped coatings in subsoils. However, incidental observation when wetting Tempe cell cores showed no volume expansion.

Of particular note during field description is that all native pasture A1 horizons had a massive structure, where all tropical pasture sites had a weakly to moderately well-developed fine pedal structure (Figure 4.7).

Roots

Visual root abundance was significantly greater in tropical grass pasture soils for depths of up to 70 cm (Figure 4.9 A). Sign test results show that roots were significantly more abundant for tropical pastures 10 – 90 cm at every sampling point along the boundary fence. Whole profile comparison of root abundance showed that root abundance to 150 cm was greater in soil under tropical grass pastures, with an average of 23 roots/10 ×10cm in native pastures and 42 roots/10 ×10 cm in tropical grass pasture (0 – 150 cm). These measures are equivalent to 4600 m root/m^3 of soil in native pasture and 8400 m root/m^3 in tropical pasture (Ringrose-Voase, 1991). 63% of native pasture roots and 54% of tropical grass roots were concentrated at 0 – 30 cm in the profile. Root counts for the upper 20 cm of the upper B2 horizon were significantly higher in tropical grass pastures with 38 roots/10 ×10 cm in tropical pastures and 15 roots/10 ×10 cm under native grass pastures.

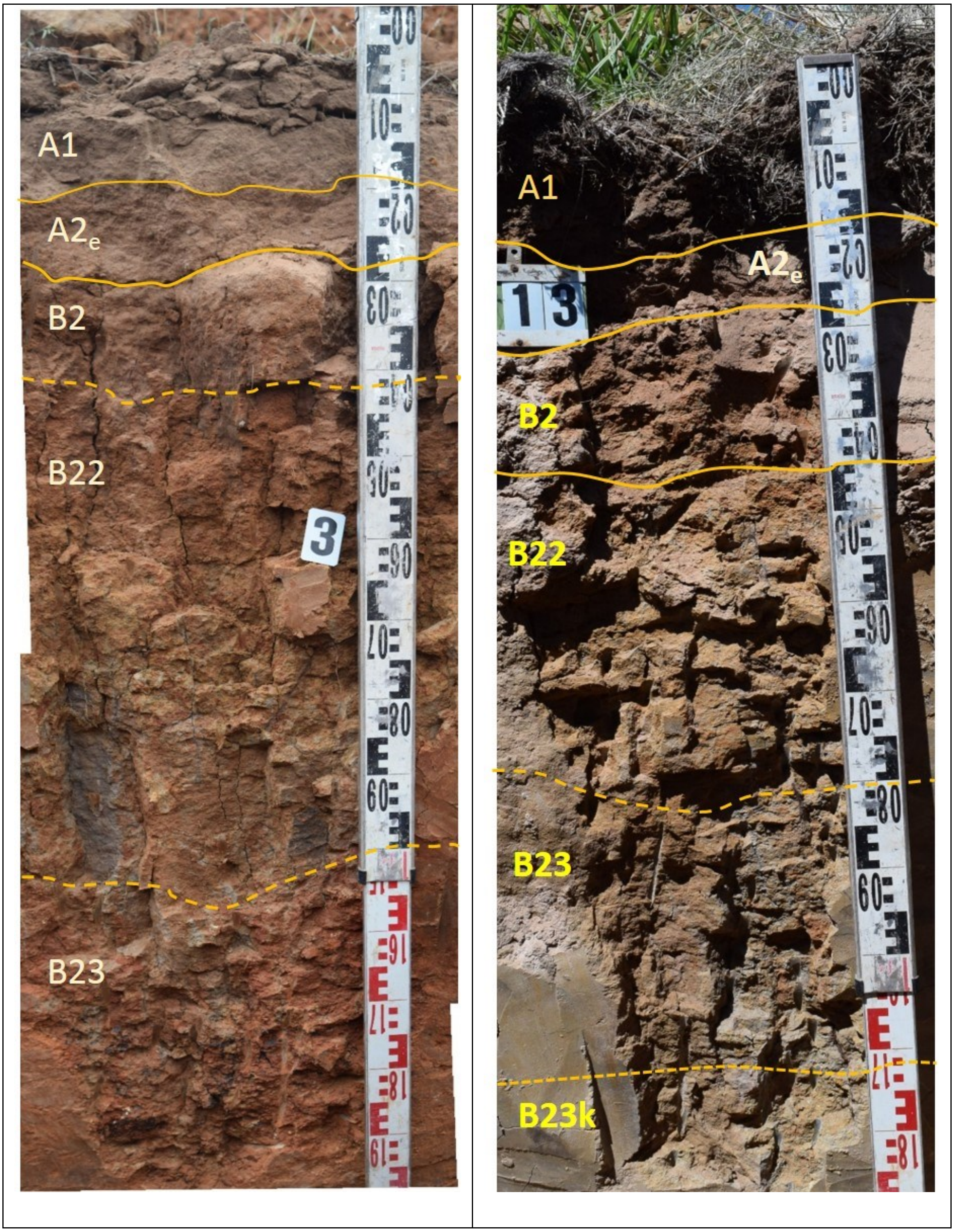


Figure 4.7: Typical paired Sodic Chromosol profiles at Towri.

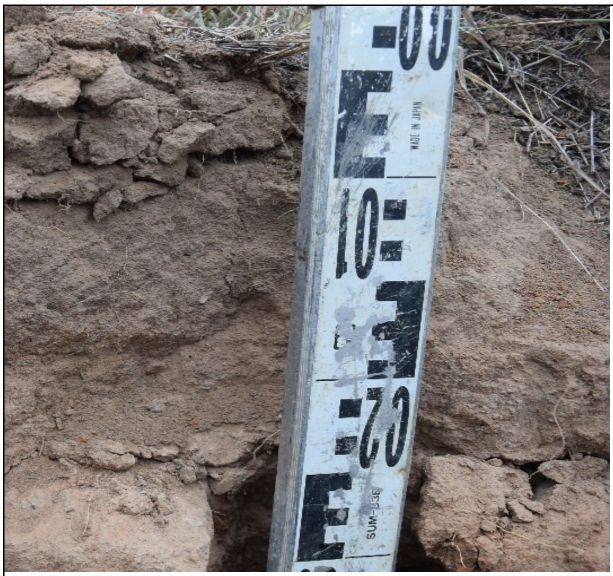
	
Topsoil Structure: Native Pasture. Profile 3	Tospoil Structure: Tropical Pasture. Profile 13

***Figure 4.8**: Illustrating the massive and earthy fabric of native pasture topsoils and the development of fine peds in tropical pasture topsoils at Towri.*

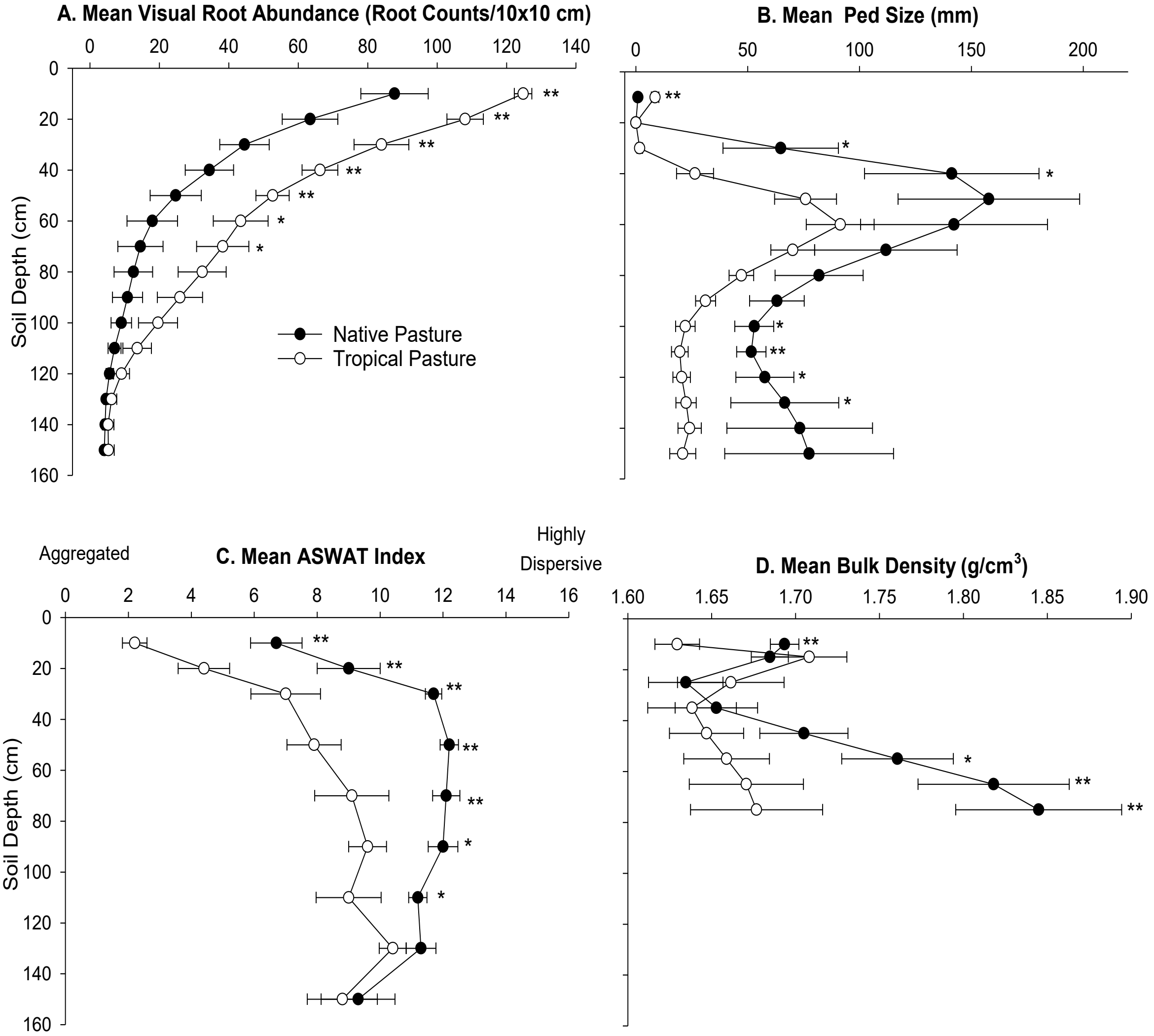


***Figure 4.9**: Field and Physical measurements down the profile for native and tropical pastures. A: Mean visual root abundance ±se (mean) (roots/10×10 cm). B: Mean ped size ±se (mean) (mm). C: Mean ASWAT index ± se (mean) (from 0 (aggregated) to 16 (highly dispersive)). D: Mean bulk density ± se (mean) (g/cm^3). Asterisks indicate levels of significance: $p \leq 0.05$*; $p \leq 0.01$**.*

Soil structure (pedality and ped size)

Native pasture topsoils (0 – 10 cm) had an apedal, massive structure, which was significantly different to tropical grass pastures topsoils which had a weakly to moderately pedal fine (< 2 mm) polyhedral structure (Figure 4.9 B). Tropical pasture soil structures were significantly smaller at 20 – 30 cm, and 70 – 110 cm depth, with no difference at 30 – 70 cm. A2$_e$ horizon data (generally 10 – 20 cm) were not significantly different between pasture types. Non-parametric Sign tests showed that peds were significantly smaller in tropical pasture soils from 0 – 30 cm and from 70 –150 cm for sample pairs along the fence line.

Soil physical properties

Dispersion (ASWAT)

Dispersion was significantly lower in tropical grass pastures from 0 – 110 cm (Figure 4.9 C). Whole profile comparisons showed that dispersion in tropical grass pasture soils was significantly less than in native pasture soils with comparative total soil profile mean ASWAT scores of 7.6 and 10.6, respectively. Dispersion was significantly lower for upper B2 horizons under tropical grass pasture with an ASWAT score of 9.1while it was 12.1 under native pastures.

Bulk density

Bulk density was significantly lower in tropical grass pastures at 0 – 10 cm, but not significantly different at 10 – 40 cm depth. Subsoil BD was significantly lower in tropical grass pastures at 55 – 75 cm (Figure 4.9 D). Specific mean BD measurements of B1horizons show that BD was significantly less under tropical grass pastures (mean BD of 1.67 g/cm^2) compared with under native pasture (mean BD 1.81 g/cm^2).

Field capacity and porosity distribution

Mean field capacity (FC) was not different between native and tropical pasture topsoils with a total mean FC of 15.5% for topsoils for both treatments. Upper B horizons were close to significantly different ($p = 0.062$) with a total mean FC of 23.2% for native pastures and 18% for tropical pastures.

Mean porosity was significantly less in native pasture than tropical grass pasture topsoils with values of 0.345 m^3/m^3 and 0.382 m^3/m^3, respectively. The pore size distribution in topsoils (Figure 4.10 A) was significantly different in the macropore range (> 30 μm) with 0.161 m^3/m^3 and 0.224 m^3/m^3 porosity in this range for native and tropical grass pastures, respectively.

Mean porosity of upper B2 horizons was significantly different between native and tropical pastures with mean porosity of 0.324 m^3/m^3 and 0.370 m^3/m^3 respectively, and there were significant differences in pore size distribution. Upper B2 horizon pore size distribution (Figure 4.10 B) within the macropore range (> 30 μm) was significantly lower with 0.088 m^3/m^3 porosity in native pasture compared with 0.24 m^3/m^3 in tropical grass pasture soils. A horizon porosity in the mesopore range

(3 – 30 μm) was not significantly different. Porosity in the micropore range (< 3 μm), was significantly higher for native pastures with upper B2 horizons possessing 0.21 m^3/m^3 in native pastures and 0.093 m^3/m^3 in tropical pasture.

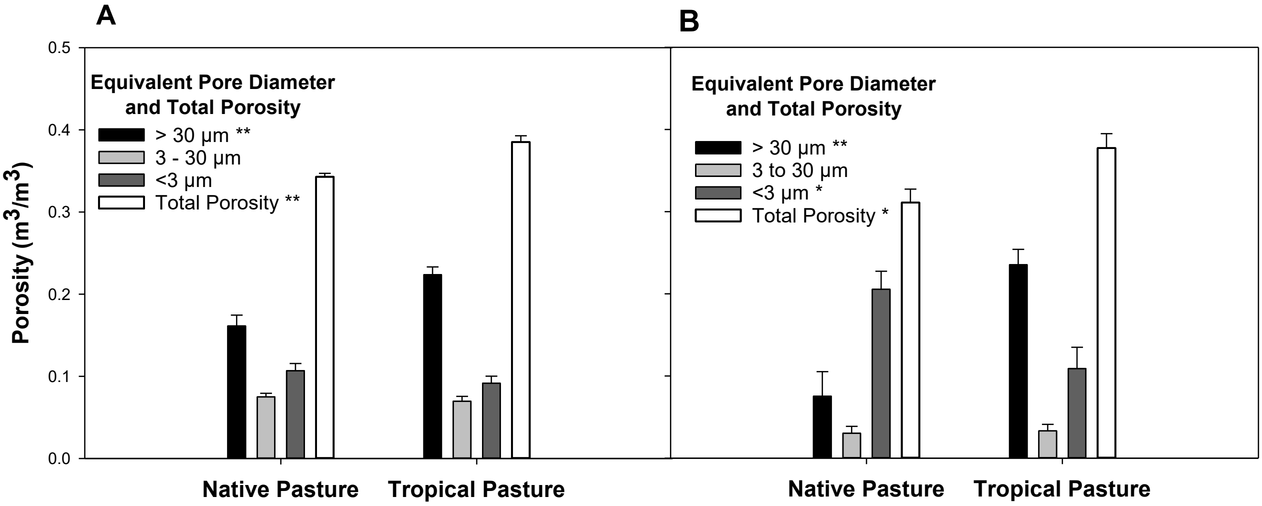


Figure 4.10: *Mean + se (mean) (m^3/m^3) equivalent pore diameter and total porosity for native and tropical pastures. A: Topsoils (0 – 10cm) B: Upper B2 Horizons (depth variable). Asterisks indicate levels of significance: $p \leq 0.05$*; $p \leq 0.01$**.*

Comparative clay content of upper B2 Horizon

There was no significant difference between clay and sand content of upper B2 horizons under native or tropical grass pastures (Figure 4.11). There was a significant but very small difference in the silt fraction with 25.8% and 22.7% silt in native and tropical pasture soils, respectively. Despite the slight difference, the texture classification of the upper B2 horizons under both pastures was light to medium clay (Hazelton and Murphy, 2007).

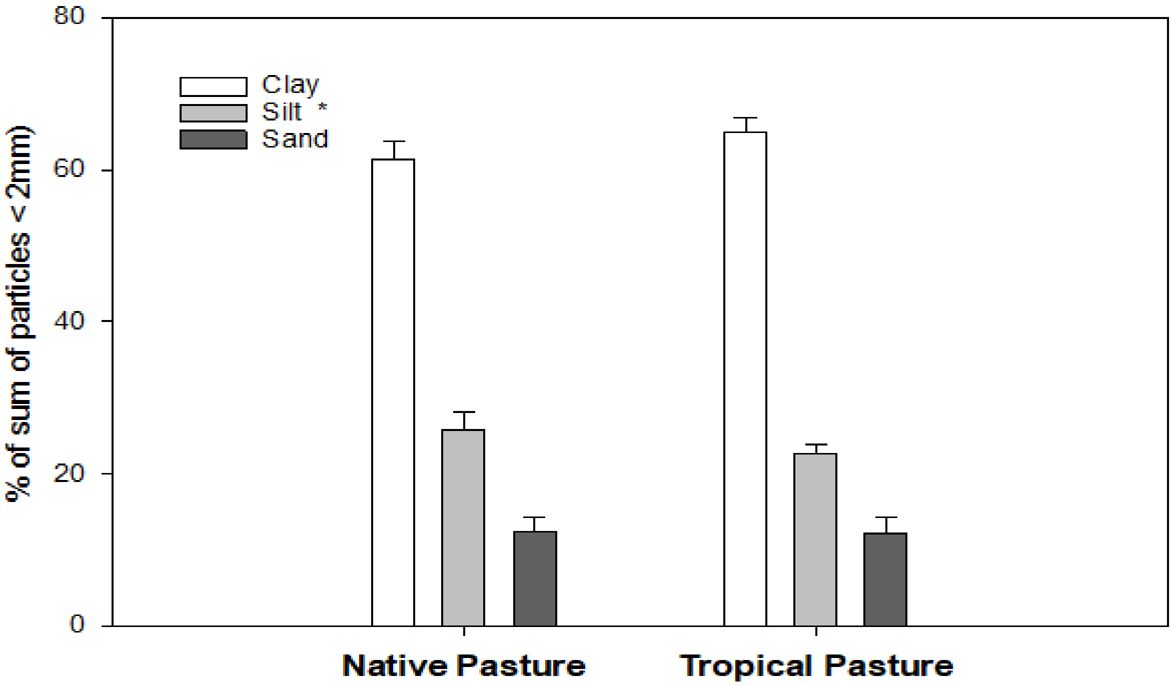


***Figure 4.11**: Particle size distribution across clay, slit and sand, expressed as mean + se (mean) percentage of sum of particles < 2mm in the Upper B2 horizons for native and tropical pastures at Towri. Asterisks indicate levels of significance: $p \leq 0.05$*; $p \leq 0.01$**.*

Soil chemical properties

pH$_w$ and salinity (EC$_e$ 1:5 H_2O)

Soil pH$_w$ (1:5 H_2O) was significantly lower in tropical grass pastures soils from 10 – 30 cm and significantly lower in tropical grass pasture soils in whole profile comparisons (Figure 4.12 A). Soil pH$_w$ values confirmed the patterns for incidental field pH measures. Soil field pH was significantly lower in tropical grass pasture soils from 0 – 30 cm (Figure 4.12 A). Collection of soil field pH was incidental to soil profile descriptions and conformed to the same general pattern shown by pH$_w$ down the profile to 150 cm.

Salinity at 0 – 10 cm depth (EC$_e$ 1:5 H_2O) was 0.6 dS/m in native pasture and significantly greater at 0.9 dS/m in tropical grass pastures with no significant trend below this depth (Figure 4.12 B).

CEC and exchangeable cations

CEC was significantly higher in native pastures from 10 – 30 cm with a trend of higher CEC at 30 – 50 cm but no significant whole profile difference (Figure 4.12 C).

Exchangeable Ca showed a trend of being greater at the surface in tropical grass pastures and significantly less at 10 – 30 cm compared with native pasture soils. Exchangeable Ca was not significantly different below 30 cm depth (Figure 4.12 D). Exchangeable K was similar under native or tropical grass pastures (Figure 4.1 2E). Exchangeable Mg was significantly greater in native pastures at 10 – 30 cm, with no further significant differences at lower depths (Figure 4.12 F). Exchangeable Na showed a trend of being lower in tropical grass pasture soils at 10 – 30 cm, but there were no significant differences down the profile (Figure 4.12 G).

Exchangeable Al was significantly greater in tropical grass pasture topsoils (0 – 10 cm) with 0.01 cmol/kg in native pasture topsoils and 0.05 cmol/kg in tropical grass pasture topsoils. Exchangeable Al was significantly greater in tropical grass pasture subsoils at 80 – 90 cm depth, compared with native pasture soils. Whole profile comparison showed that Al was significantly greater in tropical grass pasture soils (Figure 4.12 H).

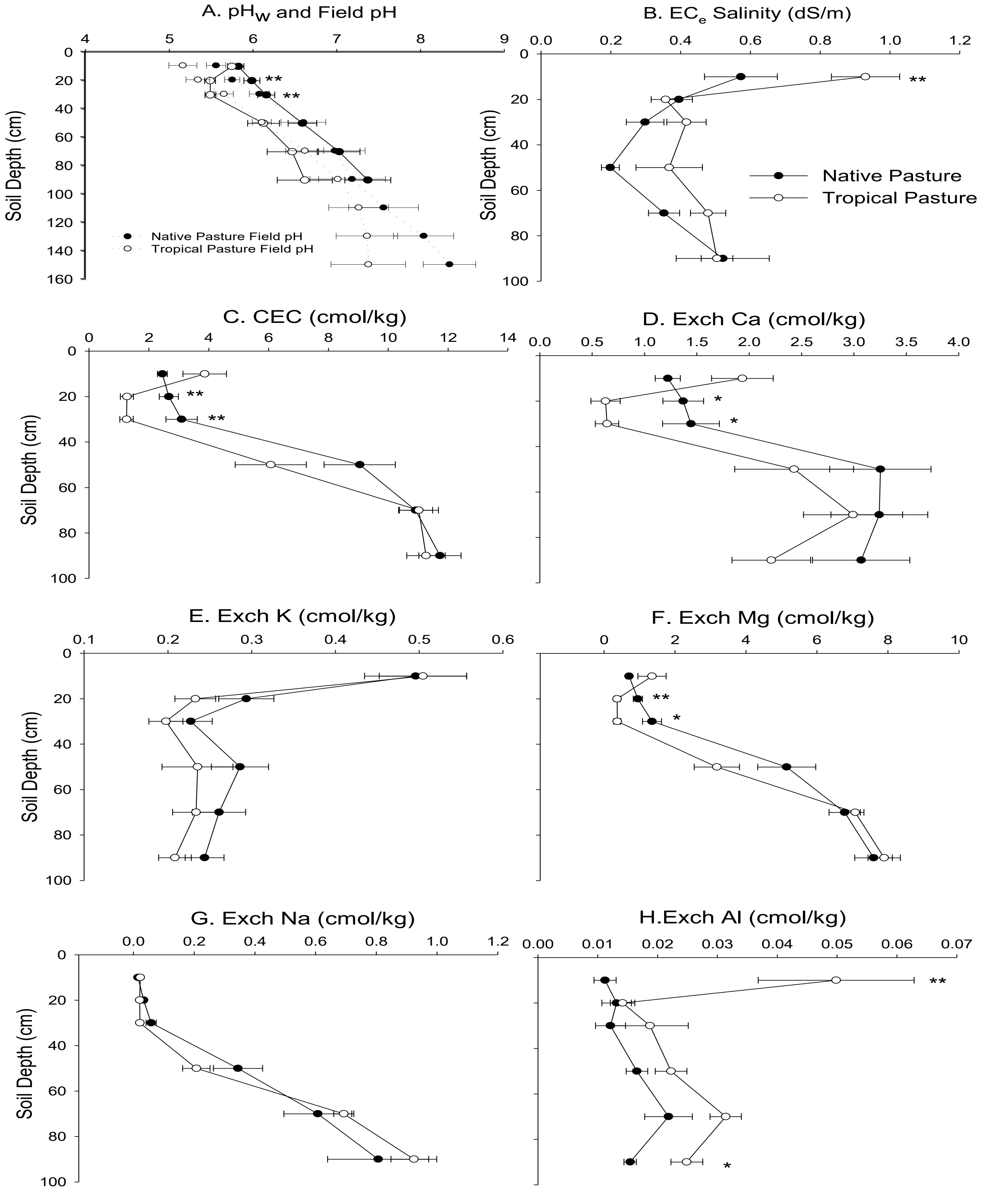


***Figure 4.12**: Soil chemistry down the profile for native and tropical pastures at Towri. A: soil pH ±se (mean). B: Salinity ±se (mean) (dS/m). C: Cation Exchange Capacity ±se (mean) (cmol/kg). D: Exchangeable Ca ±se (mean) (cmol/kg). E: Exchangeable K ±se (mean) (cmol/kg). F: Exchangeable Mg ±se (mean) (cmol/kg). G: Exchangeable Na ±se (mean) (cmol/kg). H: Exchangeable Al ±se (mean) (cmol/kg). The asterisks indicate levels of significance: $p \leq 0.01$ **; $p \leq 0.05$ *.*

Colwell P

Available P was significantly greater in tropical grass pasture soil at 0 – 10 cm (90 mg/kg) measured compared with native pasture (29mg/kg). P was also significantly greater in tropical grass pasture soil at 30 – 50 cm (Figure 4.13 A). Sign tests showed that tropical pastures had significantly more P than native pastures for paired sites at all depths.

Total organic carbon

TOC was significantly greater in tropical grass pasture topsoils (Figure 4.13 B) (31.9 t/ha TOC) compared with native pastures (10.6 t/ha TOC), at 0 – 10 cm. There was a trend for higher TOC in tropical grass pastures at 10 – 20 cm, but no further significant difference was observed between native and tropical grass pasture soils down the profile. The sum of TOC down the soil profile to 90 cm was significantly higher in tropical grass pastures with an average whole profile TOC of 84.9 t/ha C under tropical grass pastures and 58.4t/ha in native pasture. The bulk of the difference in the profile means was caused by the near surface layers.

Total nitrogen

TN was significantly greater in tropical grass pasture soils than native pasture soils at 0 – 20 cm. A trend of higher N was observed in tropical grass pasture soils at 20 – 30 cm (Figure 4.13 C). Sign test results indicated that N was significantly greater in tropical grass pasture to 30 cm and at 90 cm for the profile pairs.

C/N ratio

Due to the high variability of data, ANOVA showed no significant difference between tropical and native pasture means for C/N ratios. Sign tests indicated that the C/N ratio was significantly greater for native pastures from 50 – 90 cm for the profile pairs along the fence line (Figure 4.13 D).

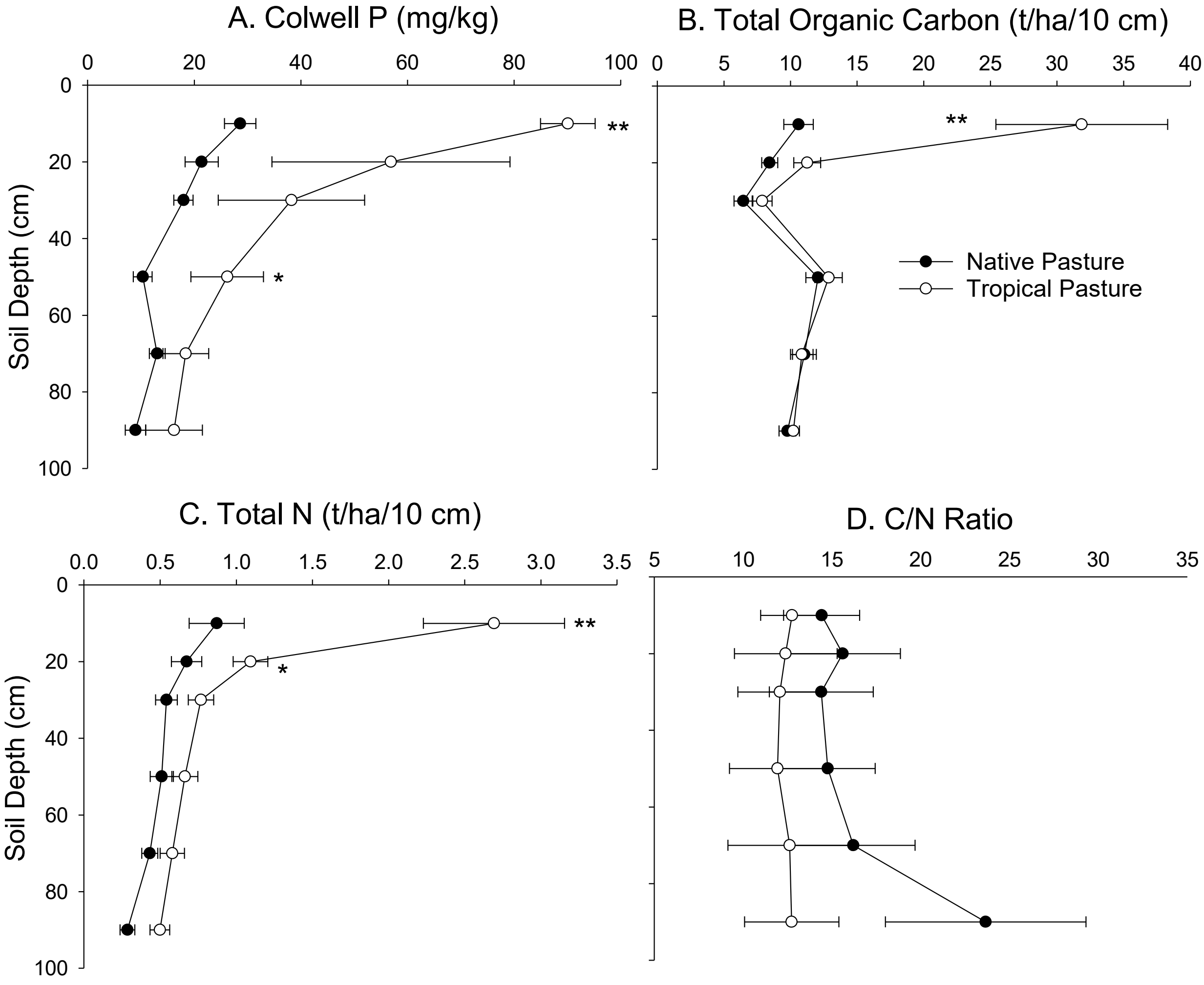


***Figure 4.13**: Soil chemistry down the profile for native and tropical pastures at Towri. A: Colwell P ±se (mean). B: Total organic carbon P ±se (mean). C: Total nitrogen P ±se (mean). D: C/N ratio P ±se (mean). The asterisks indicate levels of significance: $p \leq 0.01$ **; $p \leq 0.05$ *. One standard error of the mean as a measure of variation within a treatment is given.*

4.4 Discussion

Comparison with regional reference soil profiles

Comparison of Towri's native pasture sites with described reference sites confirms that sodic-texture contrast soils with sandy textured topsoils are comparable with those more regionally. The Towri native pasture sites have a higher clay content of B horizons based on field texture, and a higher pH at depths > 90 cm. Root abundance is similar except for 0 – 10 cm in native pasture reference sites, where Towri has higher root abundance. This is more likely a reflection of active groundcover management, given that most of the database soil profiles are opportunistic profiles on public land such as grazing reserves which have limited monitoring.

In terms of soil classification (Isbell, 2002; Isbell and NCST, 2016), the native pasture soils at Towri were equivalent to reference sites. Physically they all had sandy topsoils, an $A2_e$ horizon and a strong duplex. The use of the reference site helped to confirm that the Towri native pasture site had comparable sodic-duplex soils to others in the region.

Given that there are no known native pasture sites which have not been cropped at some stage, this comparison helps to confirm that Towri is a suitable site for comparison of the effects of tropical pastures on these soils. Although detailed examination of soils in this study is confined to one location, effects of managed tropical pastures on sodic-duplex soils were strongly indicated.

Rooting depth and abundance

The observed greater abundance of roots and deeper penetration of tropical grass pasture roots is likely to be an initiating factor in a series of feedback mechanisms which subsequently changed soil characteristics. The findings in this paper are supported by other comparative studies that showed greater root abundance, soil penetration by roots, and water storage in deep subsoil by tropical grass pasture species in comparison with native pastures in north west NSW (Murphy *et al.*, 2010; Murphy, 2014; Murphy *et al.*, 2018). McCallum *et al.* (2004) have shown that the increased abundance of roots under introduced temperate pastures, such as phalaris, has increased porosity of sodic subsoils in southern Australia. At other local sites similar to Towri, root penetration of sown tropical grass pastures has been observed at depths of greater than 1.3 m within 15 months of establishment (SoilFutures, 2005).

Bulk density, porosity and soil structure

It is important to note that the textures of the upper B horizons were not significantly different, and that differences in soil structure and porosity distribution are therefore likely to be related to management rather than natural variation.

Measured BD differences between native and tropical grass pastures showed a marked difference in topsoil (0 – 10 cm) and B horizon (75 cm) BD with a significant reduction in BD of 0.2 g/cm^3 in subsoils under tropical grass pastures. This was reflected in marginally higher porosity values in tropical pasture topsoils (0 – 10 cm) which had 0.037 m^3/m^3 more total porosity than native pasture topsoils, but with a significantly higher component of macropores. Of note is that tropical pasture total porosity was similar for both topsoils and upper B horizons, and the dominance of macropores in both topsoils and upper B horizons within tropical pastures compared with native grasses which are dominated by mesopores and micropores. Significant gains in functional porosity under tropical grass pastures occurred via addition of extra macropores, with a significant reduction in micropore values.

The importance of the differences in upper B2 horizon macroporosity is that macroporosity < 0.1 m^3/m^3 (> 30 μm pores) is considered limiting to root penetration (Engelaar *et al.*, 2000). Native pastures had 0.08 m^3/m^3 porosity in macropores in upper B2 horizons, whereas tropical grass pastures had 0.24 m^3/m^3. This higher macroporosity indicates that root penetration is less likely to be impeded.

Using the Hagen-Poiseuille law for pipe flow it is possible to interpret the increased number of macropores in terms of increased potential water flow. Flow capacity (Q) is proportional to the fourth power of the radius (Chen and Wagenet, 1992) using the equation:

$$Q = \pi r^4/8nl$$

where n is fluid viscosity and l is length of pore or pipe.

If the length of the pore and the fluid viscosity remain the same but large pores are tripled, then the potential flow through the pores would be increased by 3^4, or 81-fold. This represents a considerable increase in potential soil water flow through the macropores in the upper B2 horizon of tropical pastures.

Both the pasture types had similar sodicity at the depth of the upper B horizon, yet ASWAT scores showed that all B horizons under tropical pastures were much more stable. The native pasture upper B2 horizons disperse without re-working and the tropical pasture upper B2 horizons only disperse after reworking. This significant difference in aggregate stability contributes to the stability of extra macroporosity and consequently the potential for water flow through this horizon.

Although specific porosity distribution measurements were only taken at two points in the profile, the fact that soil structures (peds) were significantly smaller in B horizons below 70 cm indicated that the trend to greater porosity continued down the profile under tropical pastures. The large, dense columnar peds common under native pastures have few voids, whereas the smaller prismatic and blocky peds under the tropical pastures have a much more open matrix with greater inter-ped space

(Southard and Buol, 1988; Vervoort *et al.*, 1999). The change in pedality therefore represented a much greater surface area within the soil to absorb and release both nutrients and water.

Detailed soil moisture measurements were not taken through the course of field work, but the waterlogged condition of the native pastures and the trafficable and non-waterlogged condition of the tropical pastures during July sampling indicate greater soil profile drainage. The observed significant porosity shift to macropores and significantly higher stability (lower dispersion) resulted in increased water flow through infiltration rate limiting upper B2 horizons and consequently more water in storage in deeper clayey B horizons which then became available to the deeper tropical grass pasture roots. The shift towards greater macroporosity of upper B2 horizons supports the findings of Murphy *et al.* (2010), Murphy (2014) and Murphy *et al.* (2018) of greater root penetration resulting in greater subsoil moisture storage for tropical pastures.

Soil chemistry differences

It is important to recognise that many of the soil chemistry differences are a result of comparing a managed system which has a beneficial response to applied P fertiliser with a managed native pasture system that does not receive P fertiliser. Native pastures do not respond favourably to fertiliser application (Chalmers *et al.*, 2005). The cause of measured chemistry differences cannot therefore be separated from the combination of the choice of species and fertiliser response.

Melland *et al.* (2008) established that soluble forms of P can be leached down profiles and found in soil water at depth in the profile. Although leaching of P is not a common occurrence it has been reported for Sodosols in Australia (Adeloju *et al.*, 2016). Fertilised grass pastures are associated with increased herbage production and dung containing P which is soluble in water and available for leaching (Rowarth *et al.*, 1988; Sharpley and Moyer, 2000). P profiles under tropical grass pasture in this study indicated partial leaching of applied P to 50 cm depth which is consistent with the findings of Adeloju *et al.* (2016) and also consistent with a system where P is applied as fertiliser.

Total N gives an estimate of the N that can be mineralised and used by plants (Rayment and Lyons, 2011). Total N profiles under tropical pastures were significantly greater to 90 cm depth than under native pastures, indicating either N leaching or fixation of N by legumes to this depth. Once again, the greater total N value is an indication of response to fertilisation in combination with the pasture type and management

Clark *et al.* (2007) demonstrated that addition of soluble C and N to sodic subsoil material significantly increased biological activity in sodic subsoils. Increased biological activity, particularly mycorrhizal fungi associated with roots, has been shown to both build and increase stable soil structure (Bronick and Lal, 2005; Daynes *et al.*, 2013). Mungai and Motavalli (2006) demonstrate that higher speed of C mineralisation is related to higher N and lower C:N ratios.

Whilst there is great variation in the C:N ratios measured down paired profiles, the Sign test indicated that the native pastures had significantly higher C:N ratios at 30 – 90 cm along the fence line. The observed differences were significant and in the same direction, even though the paired test on the mean of the differences was not significant. Significantly lower C:N ratios in tropical grass pasture soil profiles to 90 cm implies potential for greater biological activity in subsoils, C mineralisation and potential for soil structure development in the tropical grass pasture soil profiles.

The greater exchangeable Al and lower pH in tropical grass pasture soils relative to native pasture soils observed in this study are a common effect of fertilised pasture topsoils containing legumes (Ridley *et al.*, 1990; Peoples *et al.*, 1995). Lowering of soil pH alone in dispersive clay soils can have a similar effect in increasing soil structure development and reduction in dispersion (Chorom *et al.*, 1994). Further work on the stability of subsoils in relation to the ionic strength of soil water will be presented in Chapter 5.

Increased SOC in topsoils, lower profile C:N ratios and lowering of soil pH, are associated with structural soil structure and stability improvements in sodic-duplex soils. This is supported in part by the structural (pedality) improvements and lower dispersion with greater porosity observed under tropical pasture in this study.

Parallels with other studies

The results reported herein have parallels with those of Gill *et al.* (2009) who found similar root growth, macroporosity and soil structure improvements in lucerne and cereals by ripping Dynamic Lifter® organic fertiliser (pelletised chicken manure) into upper B horizons of sodic-texture contrast soils in South Australia. Gill *et al.* (2009) attributed physical changes of macroporosity, structural change and root penetration to the supplementation of the upper B2 horizon zone with nutrients and organic matter which promoted microbial activity and offset the impacts of sodicity. In contrast to Gill *et al.* (2009), the current study attributes soil change to the initial action of tropical grass pasture roots with subsequent structural and physico-chemical changes.

McCallum *et al.* (2004) demonstrated that temperate perennial pastures of lucerne (*Medicago sativa*), and phalaris (*Phalaris aquatica*) could increase macroporosity in previously cropped Sodosol subsoils. As such, the findings of the current research suggest that tropical grass pasture roots can utilise the "biological drilling" processes described by Cresswell and Kirkegaard (1995) to enhance soil macroporosity.

In terms of climate, north west NSW is generally unsuitable for temperate pastures and the "biological drilling" potential of tropical grass pastures in this study are likely to have similar effects of increasing subsoil structure and infiltration of soil water. Deeper rooting and storage of soil water in tropical pastures has already been demonstrated on similar soils in replicated trials (Murphy *et al.*, 2010; Harris *et al.*, 2014; Murphy *et al.*, 2018).

Although long-term monitoring of soil moisture was not carried out in this study, site trafficability in the July sampling is a good indicator of soil profile drainage. Both paddocks previously had a history of low wet bearing strength following prolonged rainfall before tropical pastures were established. This implies that the soils under tropical pastures had better profile drainage, which is consistent with the increase in observed macroporosity. Future studies may consider using redoxymorphic measurements such as soil colour and mottles to better characterise soil profile drainage.

4.5 Conclusions

Inclusion of well managed and fertilised tropical grass pasture systems has had significant positive consequences on sodic-duplex soils at the Towri site in a relatively short time frame of 14 years. Tropical pastures chosen for response to fertiliser and fodder production have resulted in deeper, more abundant, more aggressive penetration of tropical grass roots which was associated with the significant creation of greater macroporosity in both surface soils and upper B2 horizons. The higher macroporosity is a reflection of root penetration by tropical grass pastures. Higher aggregate stability in subsoils under tropical pastures offset pore blocking associated with sodic dispersed clays when wetted up. Pore blocking is more likely to occur in native pastures which have substantially lower macroporosity and higher dispersion in upper B2 horizons.

Considering that the natural or undeveloped state of the B horizons in these soils is large structures (very large domed columnar peds), the significant decrease in ped size below 90 cm is an important subsoil improvement. Topsoil structure improvement from massive to pedal has additionally helped to stabilise tropical pasture topsoils.

This is a process which may enable further root development and consequent subsoil development over time. Increases in SOC in the surface, decreased profile pH, increased profile N and lower C:N ratios may be related to the development of smaller, stable peds in deep B horizons. The combination of factors would assist in overcoming limitations of impeded drainage in sodic-duplex soils enabling improved soil water storage and greater pasture production.

Although this study was a “point in time observation”, and a comparison of an actively fertilised system with one that wasn’t actively fertilised, it suggests, along with earlier observations and studies, that tropical pastures can significantly improve deep soil structure whilst increasing fodder production. As the native grasses do not respond well to fertiliser, the confounding of the tropical pastures with the fertiliser regime required to sustain a high production system does not need to be separated.

Given the limitations of traditional temperate pasture species in northern NSW and increasing summer temperatures, these results are promising for productive expansion of tropical grass pastures and soil improvement on these difficult to manage soils.

Chapter 5. Study Two Part Two: Potential chemical mechanism for stabilisation of sodic B horizons with tropical pasture soil water

5.1 Introduction

The previous chapter showed that tropical pastures at the Towri research site were associated with more stable, less dispersive, sodic upper B2 horizons and significantly greater stable macroporosity. The higher macroporosity and stability in water potentially allows for substantially greater water flow through this normally restrictive upper B2 horizon. This chapter tests the hypothesis that "cation enriched soil water from tropical pastures will decrease dispersion and promote flocculation and hence stability of sodic upper B2 horizons."

Summary of previous chapter relevant to this study

Site selection and the impact of pasture species on soil physical and chemical properties were discussed in detail in Chapter 4. Briefly, the tropical grass pasture topsoils (0 – 10 cm) had, in comparison with native pastures: significantly increased root abundance, fine pedal structure as opposed to massive structure, lower BD, and significantly greater porosity with a significantly greater abundance of macropores. Additionally, tropical pasture topsoils had significantly greater salinity (but were not saline), greater CEC and greater Al, Ca, SOC, total N, and available P content. Some of these values (such as P and N) were assumed to be a reflection of P fertiliser application and legume activity of the managed tropical pasture system.

The upper B2 horizons under tropical grass pasture had significantly greater macroporosity, greater root abundance and lower dispersion than under equivalent native pasture horizons. Exchangeable sodium percentage (ESP) values were similar for upper B2 horizons under both pasture types, indicating that tropical pasture sodic upper B2 horizons were being stabilised in a manner that did not affect overall ESP. Field textures of these horizons were sandy clay to medium clay. Subsequent quantitative XRD analysis performed by CSIRO Mineralogical and Geochemical Services (Adeliaide, SA) indicated that the clays of upper B horizons at Towri are comprised of 56% kaolinite and 40% smectite.

McCallum *et al.* (2004) demonstrated that macroporosity of sodic subsoils could be increased with temperate fibrous rooted pastures, but these pastures are considered unsuitable for northern NSW and south-east QLD due to high summer temperatures. Tropical grass pastures have been shown by Murphy *et al.* (2010) to facilitate greater deep moisture storage and use in sodic-duplex soils due to increased root abundance and rooting depth. Although specific soil water measurements were not obtained during the study described in Chapter 4, it was assumed that the potential for soil water to access deeper subsoils was substantially increased by the observed increase in stable macroporosity. Although these differences are reported in Chapter 4, no mechanism was offered to explain the apparent stability of upper B2 horizons under tropical pasture.

Sodic B horizon characteristics which prevent deep water storage

Dispersed clay from sodic upper B2 horizons can effectively restrict vertical water and air movement through the soil matrix by clogging pores and channels when this horizon becomes wet (Chorom *et al.*, 1994). Clogging results in restricted vertical soil profile drainage, and contributes to the bleaching of the $A2_e$ horizon through ponding of water above the B horizons.

The pH of subsoils can have a controlling effect on dispersion. The Australian Soil Classification distinguishes between neutral to alkaline sodic-texture contrast soils (Sodosols) and acidic sodic-texture contrast soils (Kurosols) because 1:5 $pH_w < 5.5$ tends to have a limiting effect on dispersion due to potential for release of trivalent Al^{3+} ions which contribute to stabilisation of clays. This effect, however, is far from consistent and many acid sodic soils exhibit dispersion (Isbell and NCST, 2016; Morand, 2018).

Soil water characteristics

Soil water and solute movement down the profile in sodic-texture contrast soils is complex, and can be related to antecedent moisture (Hardie *et al.*, 2011), the strong texture contrast between A and B horizons (Cox *et al.*, 2002), as well as dispersion of upper B horizons. The composition of soil water chemistry is often reported to be highly complex and highly variable, sometimes at a scale of only a few centimetres of soil (Giesler and Lundström, 1993; Cox *et al.*, 2002).

Cox *et al.* (2002) extracted and analysed soil water from sodic-texture contrast profiles in South Australia over a period of more than one year. The soil water chemistry of topsoils varied widely with season, with pH changes of two units occurring between dry and wet seasons. Seasonal changes in pH of this magnitude will have marked impacts on the type and availability of ions present in soil solution. Soil water pH is known to affect dispersion of sodic soils with dispersion becoming more likely with increasing pH (Churchman *et al.*, 1995). Kopittke and Blamey (2016) demonstrated that presence of potentially toxic levels of Al^{3+} ions are only available in soil solution $pH < 5.0$, and complexation of Al in forms that do not affect soil structure or plant function occur at higher pH values.

Changes in soil moisture content have been observed to elicit substantial changes in the pH and type and availability of free ions in soil solution (Dyer *et al.*, 2008). As such, soil water ionic strength and ion composition are expected to be highly variable and dependent on season and initial moisture content. As water moves through soil, its chemistry can be expected to change as ions are sorbed and desorbed along its path. In this regard, it is difficult to identify individual ions as being responsible for having a potentially stabilising or destabilising effect on clays in sodic B horizons. Total ionic strength or total cation concentration (TCC) of the soil solution is likely to be of greater consideration in terms of the complexities of interactions that are likely to occur between soil water and sodic clays.

Common soil dispersion amelioration methods

Experiments with sodic soil materials and alum ($KAl(SO4)_2 \cdot 12H_2O$) demonstrate that dispersive clays can be transformed to stable flocculated clays with sufficient alum addition (Jafari *et al.*, 2012). Alum is commonly used to flocculate suspended clay sediments to clarify drinking water supplies because of the high affinity of trivalent Al^{3+} ions to attach to clay particles and reduce intraparticle repulsion (NHMRC, 2011). Use of alum is considered an engineering solution to soil stability or water clarity and not employed in agriculture due to the high risk of inducing aluminium toxicity in soil.

Dispersion of sodic soils can be offset in an agricultural setting with addition of organic matter or calcium in the form of gypsum ($CaSO_4 \cdot 2H_2O$) or lime ($CaCO_3$), which chemically or biologically stabilises dispersive clays (McKenzie *et al.*, 1993; Gill *et al.*, 2009; Gourley and Sale, 2014). Dispersion of sodic subsoils can also be reduced by injection of large volumes of organic fertiliser (Gill *et al.*, 2009). In general, the introduction of large quantities of soil amendments is considered impractical and costly for most land uses on dispersive sodic soils in northern NSW and south-east QLD.

Mechanisms of dispersion

Sodicity which results in dispersion is generally caused by the sodium ions forming a wide net of positive effective charge in the outer margins of the diffuse double layer of cations which form around negatively charged clays (known as the counter ion atmosphere) (Menzies *et al.*, 2015). Replacing sodium (Na^+) cations with higher valence cations such as calcium (Ca^{2+}) or aluminium (Al^{3+}) causes compression of the counter ion atmosphere because the higher valence ions have a higher charge and bond more tightly or closely to clays, thereby reducing the distance over which the repulsive force acts between suspended or dispersed clays (Menzies *et al.*, 2015). The repulsive forces between dispersed clays are inversely proportional to the sixth power of the valency of counter ions so exchanging Na^+ ions for Al^{3+} ions reduces this force by approximately 99.86% (Sumner, 1993).

Flocculation of clays resulting from ameliorants is generally caused by an increase in the electrolyte solution of soil water and chemical displacement of Na^+ from clay surfaces (Field *et al.*, 1997; Hulugalle and Finlay, 2003). The effects of electrolyte strength on clay dispersion and flocculation has been demonstrated by van Olphen (1977) in laboratory experiments which showed dispersed clay being treated with increasing amounts of electrolyte was concurrent with increasing clay flocculation. These experiments are a repeat of classical experiments initially performed by Hardy (1900).

The power of cations to induce flocculation in relation to their valency was originally shown by Hardy (1900) who developed the Schulze-Hardy rule which states that "*The coagulative power of a salt is determined by the valency of its ions. The prepotent ion is either negative or the positive ion, according to whether the colloidal particles move up or down the potential gradient. The coagulating ion is always the opposite electrical sign to the particle*" (van Olphen, 1977, p 24).

The present study investigated the potential for soil water from the ion-enriched A1 horizons of managed tropical grass pastures to affect sodic and dispersible upper B2 horizons from native pastures, using an adaptation of methods originally established by Hardy (1900).

5.2 Methods

Background to methodology development

The theory behind the following methodologies was based primarily on observations of the ionic strength of soil water from tropical and native pasture A1 horizons which was incidental to the porosity distribution measurements described in Chapter 4. Those results suggested a difference in soil water chemistry.

Establishing a connection between soil water chemistry and soil stability could be a partial mechanism for explaining the apparent stability of sodic B horizon material under tropical pastures. The present examination of soil water chemistry of tropical pastures and its potential influence on subsoil structural stability employed a stepwise process of:

1. Extraction of soil water from topsoils of tropical pastures;
2. Analysis of the basic chemistry of that soil water;
3. Creation of synthetic soil water in volumes sufficient to conduct soil structural experiments;
4. Conduct of simple dispersion experiments in synthetic soil water on a range of upper B2 horizon material from Towri using synthetic soil water; and
5. Use of the results of the dispersion experiments to select samples for flocculation experiments in a range of synthetic soil waters.

Topsoil water EC and pH from Tempe cells

Tempe cells are short soil columns used for determining soil water retention curves. The column is contained in a pressure housing to which pressures up to 1 bar can be applied. The resultant water retention (or release) curves can be used to calculate intact soil sample porosity distributions. A subset of Tempe cell water from porosity measurement experiments in Chapter 4 was tested for pH and EC. Ten samples of topsoil water from native pastures were taken and nine from native pasture for further analysis. These tests were incidental to the porosity work, but these differences in basic water chemistry form the foundation concepts for this chapter. Samples for porosity distribution were taken in February 2016 as outlined in Chapter 4.

Selection of upper B2 horizon material for dispersion/flocculation testing

Using data specifically for upper B2 horizons from the Towri site, additional soil samples were selected from the most sodic and least sodic upper B2 horizons under native pasture and tropical pasture for further analysis. These samples were classified as NPN (native pasture, non-sodic) NPS (native pasture, sodic), TPN (tropical pasture, non-sodic) and TPS (tropical pasture, sodic) (Table 5.1).

Table 5.1*: Soil Characteristics of Selected upper B2 horizons at Towri.*

Profile	Class	Land Use (Sodicity)	Upper Depth (cm)	Lower Depth (cm)	pH	ECE (dS/m)	Ca %	K %	Mg %	ESP %	Al %	CEC cmol(+)/kg
1	NPS	Native Pasture (Sodic)	38	58	7.73	0.35	31.02	3.54	58.12	7.15	0.16	13.59
3	NPN	Native Pasture (non-Sodic)	35	55	6.91	0.15	54.22	3.79	41.28	0.37	0.33	10.27
12	TPS	Tropical Pasture (Sodic)	35	55	6.11	0.41	14.08	2.45	73.77	9.36	0.34	10.01
20	TPN	Tropical Pasture (non-Sodic)	38	58	7.15	0.26	47.67	3.29	46.05	2.69	0.29	9.03

Selection of A1 horizon soil samples and extraction of soil water

New topsoil samples were taken using Tempe cell cores (approximately 8.6 cm diameter and 6 cm length, volume 348 cm^3) under both tropical pastures and native pastures in July 2016. Samples were paired across a fence line from native to tropical pasture. Two paired sample sites were randomly selected from ten available profile pairs. Only one Tempe core was taken for each native pasture site (as a simple reference), as total ionic strength had previously been established in measurements incidental to porosity distribution work (see Chapter 4). Multiple cores were taken for the two tropical pasture sites. Tempe core samples were obtained and prepared as described in Chapter 4.

Use of Tempe cells to extract soil water

Preparation of Tempe cells and soil samples

Tempe cells are normally used to derive low pressure water release curves from soil and to determine pore size distributions for pores > 3 μm (Flint and Flint, 2002). To avoid cross contamination from previous samples being used in the apparatus, the water holding capacity of the porous plate in a Tempe cell was determined by saturating the plate, then oven drying it and determining the mass difference. This was done to two plates only because the plates were expensive and the process potentially destructive to the plates. Both plates were found to hold 5.2 g of water when saturated. Plates selected for soil water extraction were placed in empty Tempe cells and flushed with 150 mm of deionised water (DIW) at -1 bar to reduce potential contamination from prior samples.

Intact soil samples were placed in open Tempe cells in a bath of DIW with the porous ceramic plate submerged. Water was allowed to rise through the ceramic plate until soil samples in the cores reached saturation, typically after 48 – 72 hours.

Soil water collection

Tempe cells were sealed and connected to a pressure line and brought to -100 cm height of water column (-0.1 bar). An exact pressure was achieved by inserting an air bleed tube from the Tempe cell air-line to 1 m in a water column of 2.4 m height.

The first 10 mL of soil outflow at -0.1 bar was discarded to further reduce contamination or dilution potential and ensure saturation of the ceramic plates with soil water. Pressure was then increased to -1 bar using a standard gauge. Soil water was collected in clean, acid washed conical flasks until outflow ceased. Soil water yield at -1 bar was 40 – 90 mL of water per core, representing water from pores with equivalent diameters > 3 μm.

Soil water sample preparation

Soil water samples were split into two subsamples of equal volume. Unfiltered subsamples were analysed for EC, pH and chloride (Cl^-) content. The remaining 10 mL subsamples were filtered through 0.025 μm Millipore® filters to remove potential fine organic particles which are commonly complexed with cations such as Al^{3+} or $Fe^{2/3+}$ which may otherwise affect the soil water chemistry (Menzies *et al.*, 1991b, 1991a). Filtered samples were inoculated with 20 μL of 1:1 $HCl:H_2O$ to fully dissociate cations following filtering.

Chemical analysis

EC, pH and Cl^- content of unfiltered water subsamples were analysed according to Rayment and Higginson (1992) (Section E1b, p231). Filtered samples were analysed for Al, B, Ba Ca, Cu, Fe, K, Mg, Mn, Na, P, S, Si, Sr and Zn using inductively coupled plasma mass spectrometry (ICP). Equivalent total cation concentration (TCC) in millimoles charge equivalent (mmol(+)/L) was estimated from the EC per Hazelton and Murphy (2007). Analyses were carried out at The University of Queensland School of Agriculture and Food Sciences Analytical Laboratory. Samples of DI water were also analysed to check background levels of cations for potential contamination (negative control).

Creation of synthetic soil water (SSW)

Soil water chemistry results were examined and the examples of soil water with the highest TCC were selected as a guide for further experimentation. Synthetic soil waters (SSW) were created by staff at the University of Queensland Analytical Laboratory for use in dispersion and flocculation experiments. Varying quantities of inorganic salts were dissolved in DI water and the pH adjusted using ammonium hydroxide (NH_4OH). Cations included as inorganic salts included copper (as $CuSO_4$), iron (as $(NH_4)2Fe(SO4)_2\cdot6H_2O$), manganese (as $MnCl_2$), sodium (as $NaHCO)_3$, calcium (as $CaCl_2$), potassium (as KCl), magnesium (as $MgSO_4$), strontium (as $Sr(NO_3)2$), barium (as $Ba(NO_3)2$), phosphorus (as H_3PO_4) and boron (as H_3BO_4). Ion concentrations were adjusted in SSW such that concentrations of all analytes were as similar to actual soil water from tropical pastures as practically possible.

Three batches of soil water were prepared. The first (SSW1) had a pH before dilution of 3.99, and the second a pH of 4.99 (SSW2). The third solution (SSW3) was prepared with as little Al, Cu, Fe, Mn and Zn as possible but equivalent amounts of other analytes as found in tropical pasture topsoil water. The third solution was used to determine whether flocculation could occur in the absence of flocculating ions such as Al^{3+} and Fe^{2+}.

Dispersion experiments

Dispersion in variable pH water

To determine whether soil dispersion was related simply to pH, a range of waters were mixed with pH ranging from 4.0 to 6.8 in increments of 0.2 pH units. Three small intact peds from each of NPN, NPS, TPN and TPS were introduced into the solutions in plastic coated cupcake trays to observe speed and degree of dispersion (Figure 5.1). Samples were observed over a 24 hour period and photographed at increasing time intervals to evaluate the rate and extent of dispersion.

Dispersion was ranked from 0 (aggregate stable) to 4 (fully dispersed) using an adaptation of the method of Field *et al.* (1997). Re-worked samples were not used in this test as the focus of this study was stability of intact peds from upper B2 horizons, which normally are neither exposed nor disturbed.

Figure 5.1*: Dispersion in varying pH NPS at 5 min. Sample 1 has pH 4. Subsequent samples have pH rising in increments of 0.2 pH units to sample 12 at pH 6.4. Experiment continued on next tray.*

Dispersion in variable salinity water

Salinity can affect dispersion of sodic soil materials. Increasing electrolyte concentrations from salts is known to reduce soil dispersion (Quirk and Schofield, 1955). The above dispersion test was repeated with DIW with NaCl added to represent the varying salinity of the soil waters tested. Deionised water was salinised using laboratory grade NaCl to adjust salinity to 0.5, 0.4, 0.3, 0.2, 0.1 0.05 and 0.00 dS/m. Three small intact peds from each of NPN, NPS, TPN and TPS were introduced into the solutions in plastic cupcake trays to observe the rate and degree of dispersion as described above.

Dispersion and concentration of synthetic soil water

The above dispersion experiment was conducted again with synthetic soil water (SSW 1 and SSW 2) diluted to concentrations of 100%, 80%, 50%, 40%, 30%, 20%, and 10% of the original using three small intact peds from NPN, NPS, TPN and TPS samples with results recorded as above.

Flocculation experiments

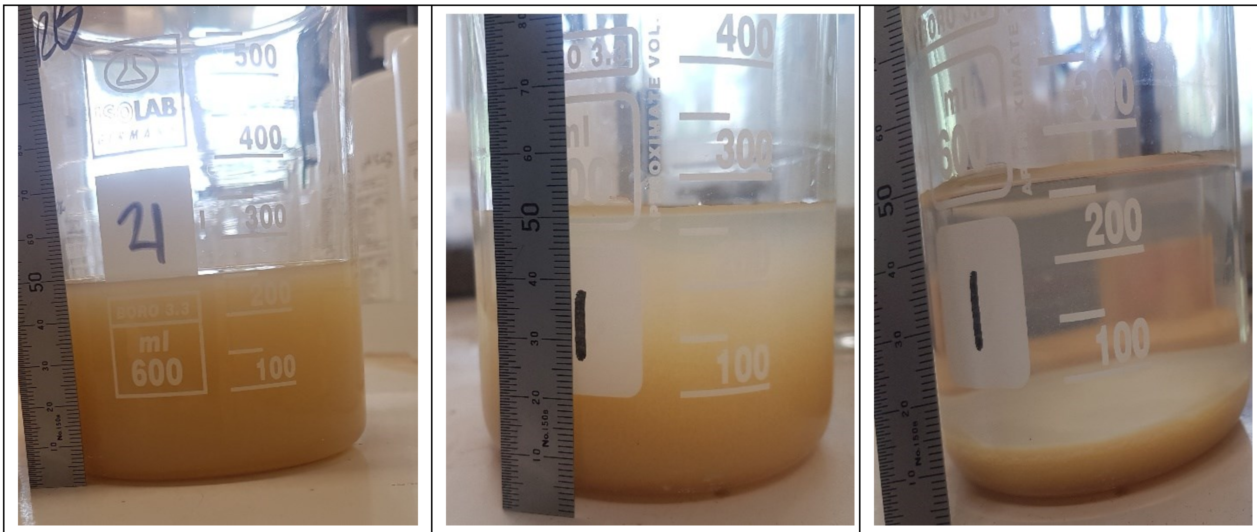
The most dispersive upper B2 horizon material, based on the results of the dispersion tests, was selected for subsequent flocculation tests.

Preparatory dispersion of soil material in water

A 50 g subsample of air dried (40° C) soil was crushed, sieved through a 2 mm mesh and sonically dispersed in 500 mL of DIW for 20 minutes immediately prior to each flocculation experiment.

Flocculation in synthetic soil water

Triplicate 250 mL aliquots of SSW1 concentrations of 50, 40, 30, 20, 10 and 0% were prepared in 500 mL beakers. Similarly, triplicate 250 mL aliquots of SSW2, for SSW 3 at concentrations of 100%, 50%, 40%, 30%, 20% 10% and 0% were prepared. Ten millilitres of the dispersed soil (containing approximately 1 g of dispersed soil with approximately 50% clay) was introduced into each beaker. The rate and extent of resultant flocculation was observed over a 24 hour period by measuring the clear portion of water that developed in the beakers and recording each beaker photographically. The water depth was measured and compared with the depth of flocculated mud developed in the base of the beakers (Figure 5.2). The depths of clear supernatant above flocculated or settled soil material was used to calculate the degree of flocculation.



Fully dispersed: DI water after 24 hrs.

Partially (20%) flocculated: SSW2 at 5 min

Fully flocculated: SSW2 at 188 min

***Figure 5.2**: Measurement of flocculation*

Flocculation in water of varying pH and salinity

To assess the effects of pH in the absence of most cations, the experiment was repeated a fourth time in DI water with pH increasing from 4.0 to 6.8. To assess the effects of salinity in the absence of other cations (as measured by EC), DIW was mixed with laboratory grade NaCl to conductivities of 0.5, 0.4, 0.3, 0.2, 0.1, 0.05 and 0 dS/m and the flocculation experiment repeated.

5.3 Results

Topsoil EC and pH from previous Tempe cell experiments

Incidental measurements of EC and pH for Tempe cell outputs of soil water in Chapter 4 showed that soil water pH did not differ significantly between native and tropical pastures. Solution ionic strength or total cation count (TCC), calculated as ten times the EC of the soil water (Hazelton and Murphy, 2007), was significantly higher in tropical pasture soil water than from native pastures. It should be noted however that these results were from porosity experiments where potential cross contamination of ceramic plates from one sample to other was not controlled. Mean ionic strength of native pastures was 1.4 mmol $L^{-1}$and for tropical pastures it was 6.7 mmol L^{-1}, an almost five-fold difference (Figure 5.3).

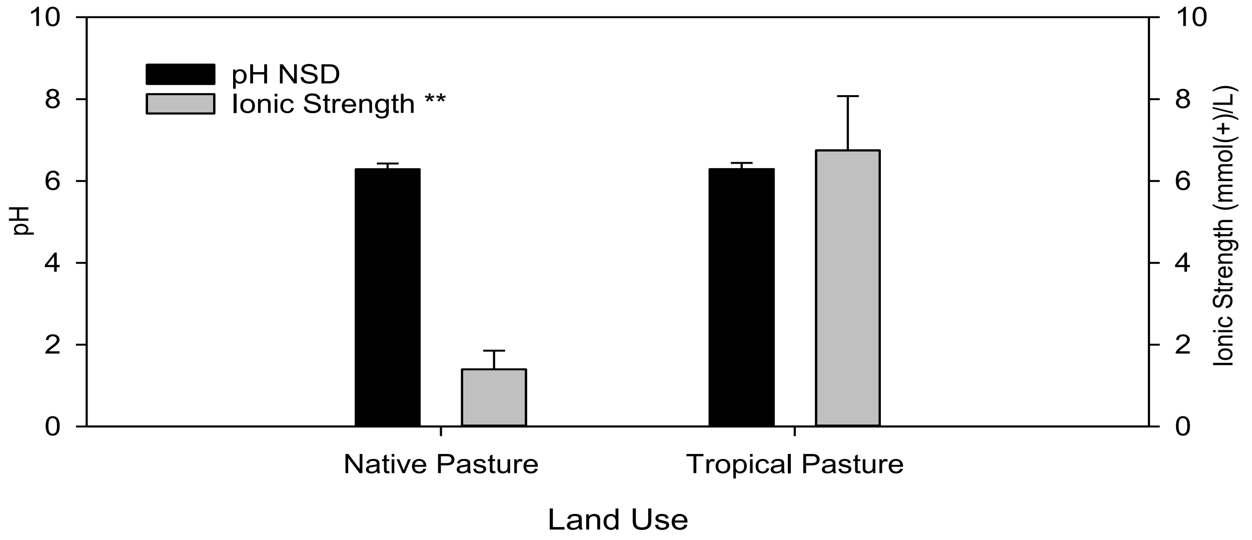


***Figure 5. 3**: Summary of pH and ionic strength of soil water from native (n = 10) and tropical pasture (n = 9) topsoils. Standard error bars show variation from mean.*

Topsoil water analyses from current experiment

Topsoil water analyses are summarised in Figure 5.4. Soil water from tropical pastures was highly variable in composition, generally more acid than native pastures, and had a higher salinity. The large variation in soil water chemistry is interesting considering many Tempe cores were taken within 10 cm of one another at the study site. Tropical pasture soil water had higher mean ion concentrations except for B and K than for native pasture samples. The maximum values of soil water chemical analytes were all higher for tropical pastures. Tropical pasture soil water had a lower pH and a generally higher EC/TCC than native pasture soil water. The lowest pH of soil water in tropical pasture was 4.9.

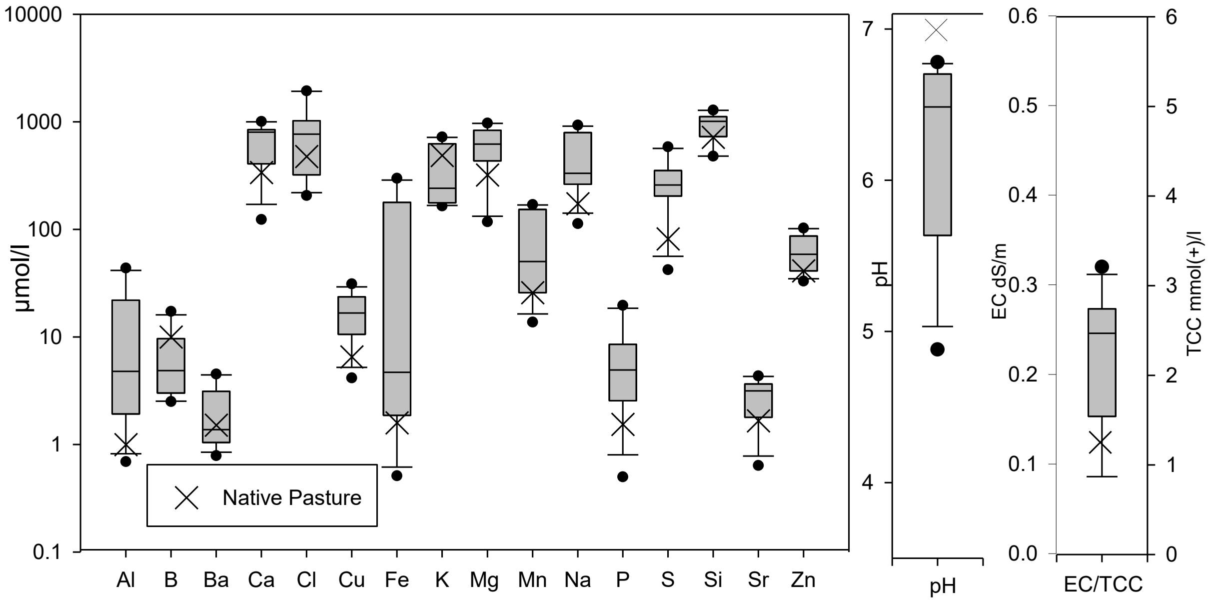


Figure 5.4: *Results of soil water testing from field samples. Box plots represent tropical pasture topsoil water analyte concentration on a log scale. Horizontal lines within box plots are mean tropical pasture values. Solid dots are maximum and minimum values. Native pasture reference values are superimposed on the box plots with an ×.*

Manufacturing synthetic soil waters

Two attempts were made to emulate soil water from tropical pasture topsoils based on the results presented in Figure 5.1. The first synthetic soil water produced (SSW 1) had a pH of 3.9, less than that measured for soil water from tropical pasture topsoils. As SSW1 had very low pH and high Al^{3+} content, it was only used at a maximum concentration of 50%, which was approximately representative of the range of analyte concentrations in the sampled soil water. The second SSW (SSW 2) was produced with a pH at the lower end of that measured for soil water from tropical pasture topsoils. As SSW was diluted to make up treatments during the following flocculation experiments, the pH increased.

A third SSW (SSW 3) was made with greatly reduced $Al^{3+,}$ Cu^{2+}, $Fe^{2/3+}$ and Zn^{2+} concentrations to determine whether the absence of these cations affected flocculation experiments. It was difficult to emulate correct EC for all SSWs. Analysis of SSWs is provided in Figure 5.5.

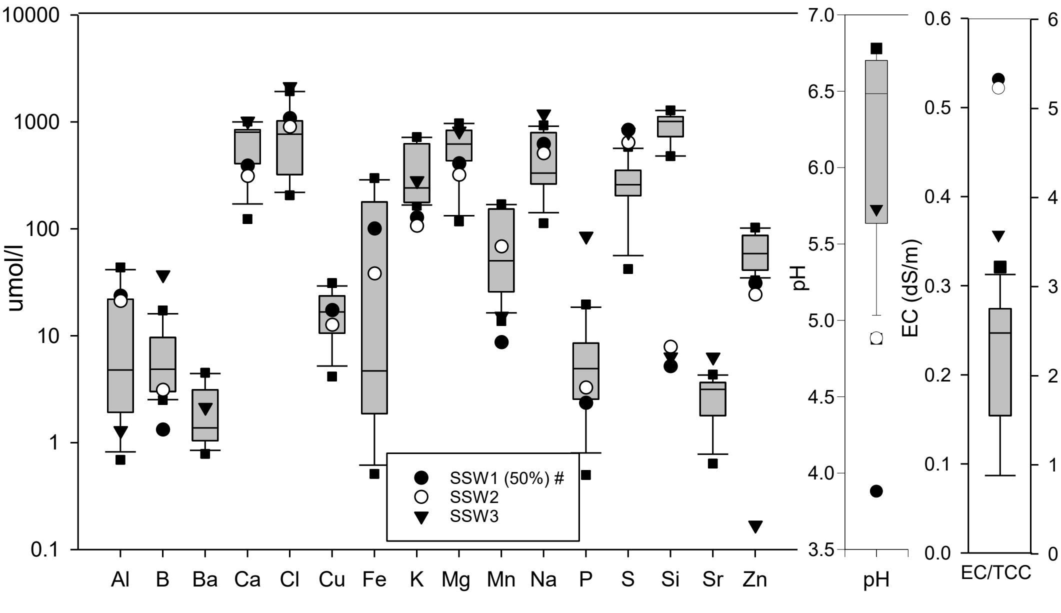


***Figure 5.5**: Chemical characteristics of synthetic soil waters (SSW1-3, represented by data points) and measured soil water from tropical pasture topsoils (analyte concentrations represented by boxplots). Note: SSW1 was not used at full strength due to its high Al3+ concentration and low pH. Box plots represent tropical pasture topsoil water concentration range. Horizontal lines within box plots are mean tropical pasture values. Solid squares are maximum and minimum values.*

Dispersion results

Dispersion and solution pH

Nearly complete (score of 3) or complete (score of 4) dispersion occurred within 24 hours for NPS soil at all pH values examined, but dispersion was slow to commence at pH of 4.4. The speed of NPS soil dispersion increased with increasing pH. Both NPN and TPN soils exhibited limited dispersion (score of 1) at pH of 6.2 or higher. The TPS samples followed the same pattern as NPN and TPN with limited dispersion occurring at pH of 6.0 or higher. TPN and TPS showed limited dispersion (score of 1) at pH 4.2 – 4.6. NPS soil dispersion rates are displayed in Figure 5.6.

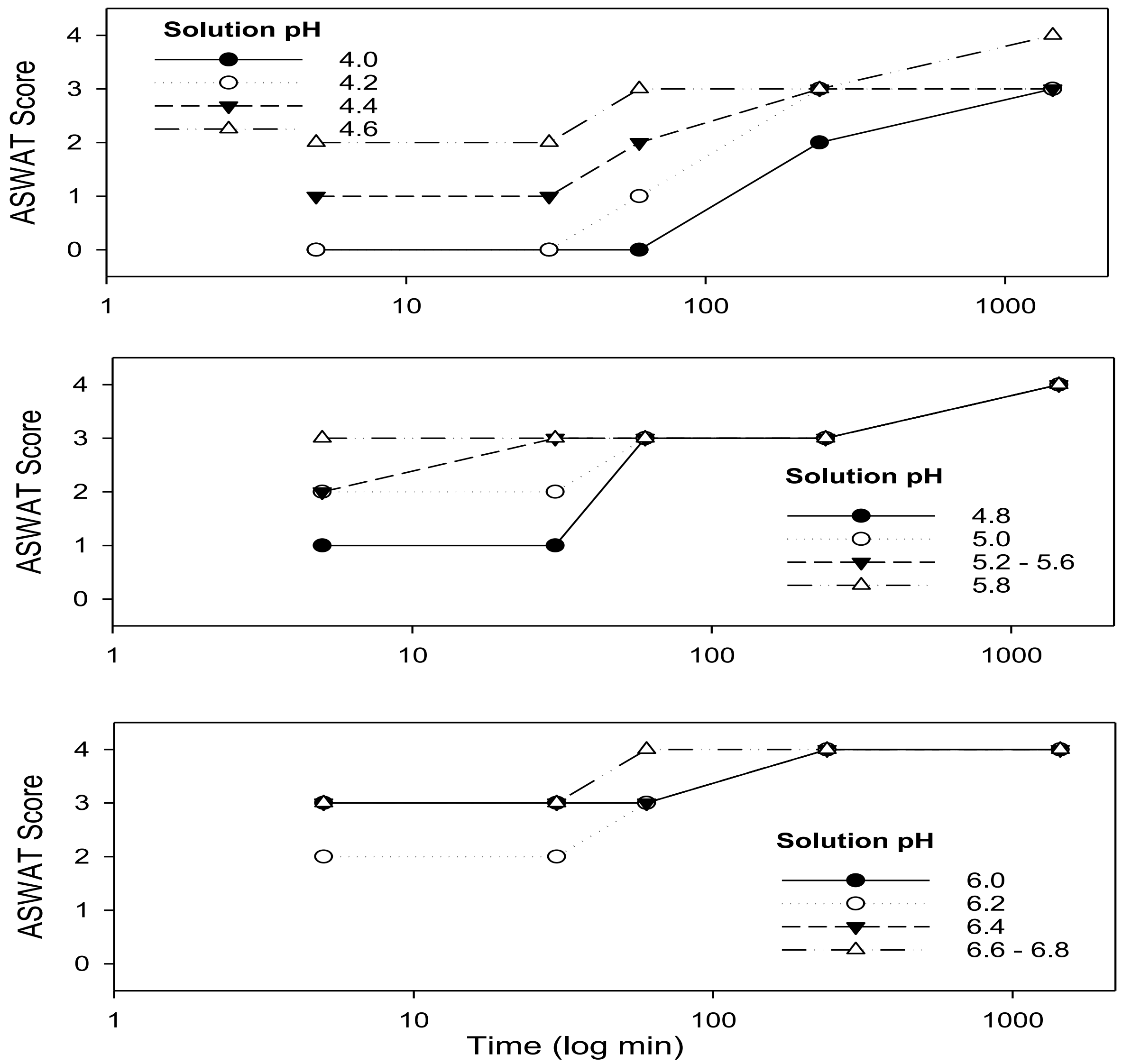


__Figure 5.6__: Water pH and dispersion time of native pasture sodic (NPS) subsoil. ASWAT Score: 0 = aggregated; 1 = very slight dispersion; 2 = slight dispersion; 3 = moderate dispersion; 4 = completely dispersed.

Dispersion and salinity

Dispersion results for varying NaCl salinity levels showed dispersion occurring only in NPS soil samples in all salinities from 0 – 0.5 dS/m (Figure 5.7). No dispersion was observed for TPN, TPS or NPN soils.

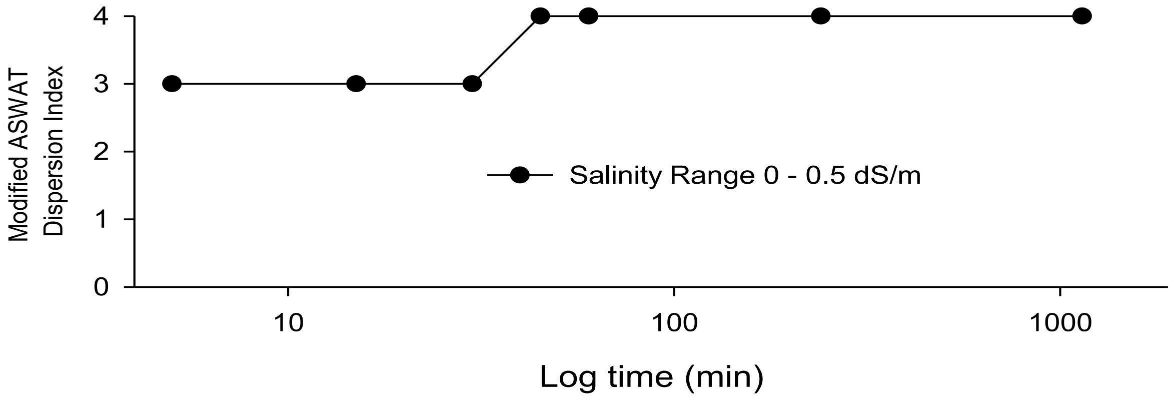


Figure 5.7*: Water salinity and dispersion time of native pasture sodic (NPS) subsoil. ASWAT Score: 0 = aggregated; 1 = very slight dispersion; 2 = slight dispersion; 3 = moderate dispersion; 4 = completely dispersed.*

Dispersion in synthetic soil water

The NPN, TPN and TPS soil samples did not disperse in any concentration of SSW, indicating that they all possess relatively stable aggregates. The NPS soil dispersed rapidly at 0% and overnight at 10% soil water concentration (both yielding a dispersion score of 4 – fully dispersed) (Figure 5.8). The NPS showed no dispersion at SSW strengths greater than 10%. Results indicate that the NPS soils were the least aggregate stable, but that aggregate stability could be affected/increased by the addition of SSW which emulated soil water from tropical pasture topsoils. Because the NPS samples were the least stable soil materials according to dispersion experiments, the NPS soil materials were selected for subsequent flocculation experiments.

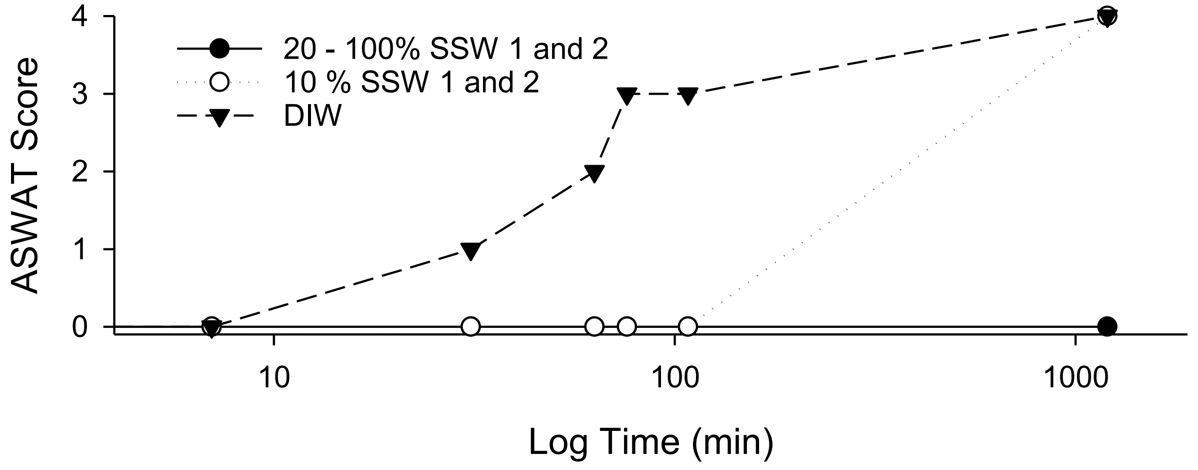


Figure 5.8*: Dispersion time of native pasture sodic (NPS) subsoil in synthetic soil waters 1 (SSW1) and 2 (SSW 2). ASWAT Score: 0 = aggregated; 1 = very slight dispersion; 2 = slight dispersion; 3 = moderate dispersion; 4 = completely dispersed.*

Flocculation results

Flocculation in varying pH and salinity

Neither pH nor EC exhibited individual effects on flocculation of the NPS soil. All samples remained dispersed over a 24 hour period (Figure 5.9).

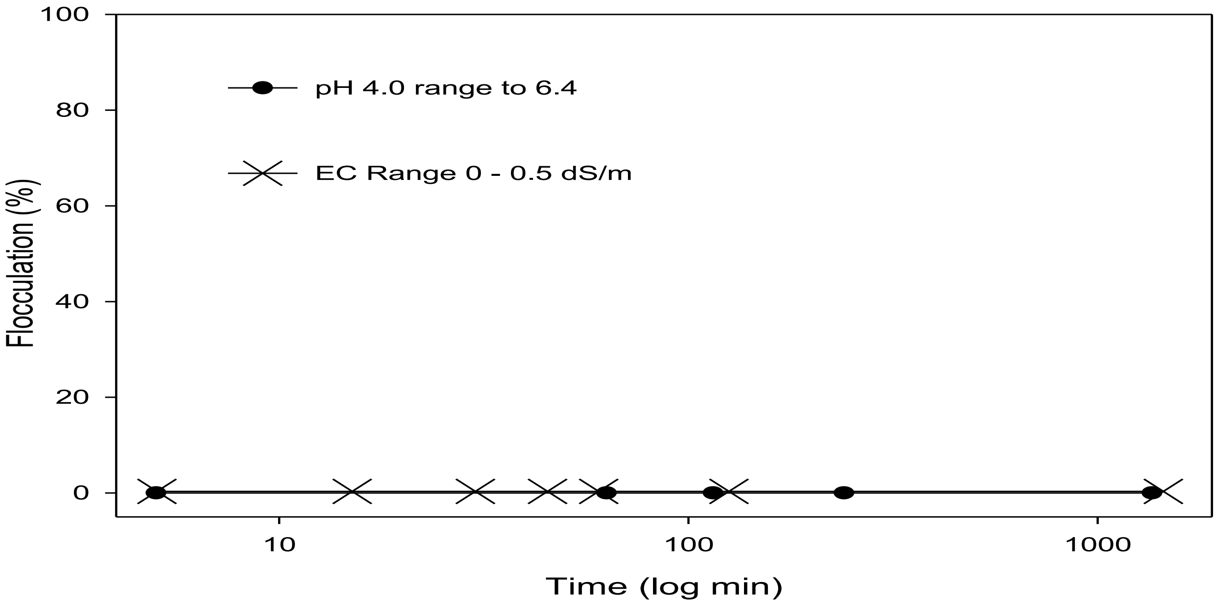


Figure 5.9*: Flocculation rate of Native Pasture Sodic (NPS) subsoil as a function of varying pH from 4.0 – 6.8 and varying salinity water from 0 – 0.5 dS/m.*

Flocculation in Synthetic Soil Water

Rates of NPS soil flocculation in SSW 1, SSW 2 and SSW 3 are summarised in Figure 5.10. Half strength (50% concentration) SSW1 and full strength (100% concentration) SSW2 resulted in complete soil flocculation within 120 minutes. Decreasing concentrations of SSW 1 and SSW 2 to 30% yielded complete flocculation within 24 hours. At 20% concentration SSW 1 and SSW 2 caused 65 – 97.5% flocculation over 24 hours but the supernatant was not entirely clear, indicating that some soil particles remained in suspension. Soil water concentrations below 20% did not yield observable flocculation.

Synthetic soil water with major cations removed (SSW 3) yielded full flocculation at 100% solution strength within 120 minutes. At 50% strength, very slow flocculation occurred with 18% flocculation at 4 hrs, and 89% flocculation at 24 hrs. No further flocculation occurred within 24 hours at SSW 3 concentration below 50%. The experiment was continued for a further 5 days (a total of 144 hours) during which time flocculation occurred at concentrations as low as 30% SSW3. Although this was recorded it is not comparable with the other SSW due to the time difference in monitoring duration.

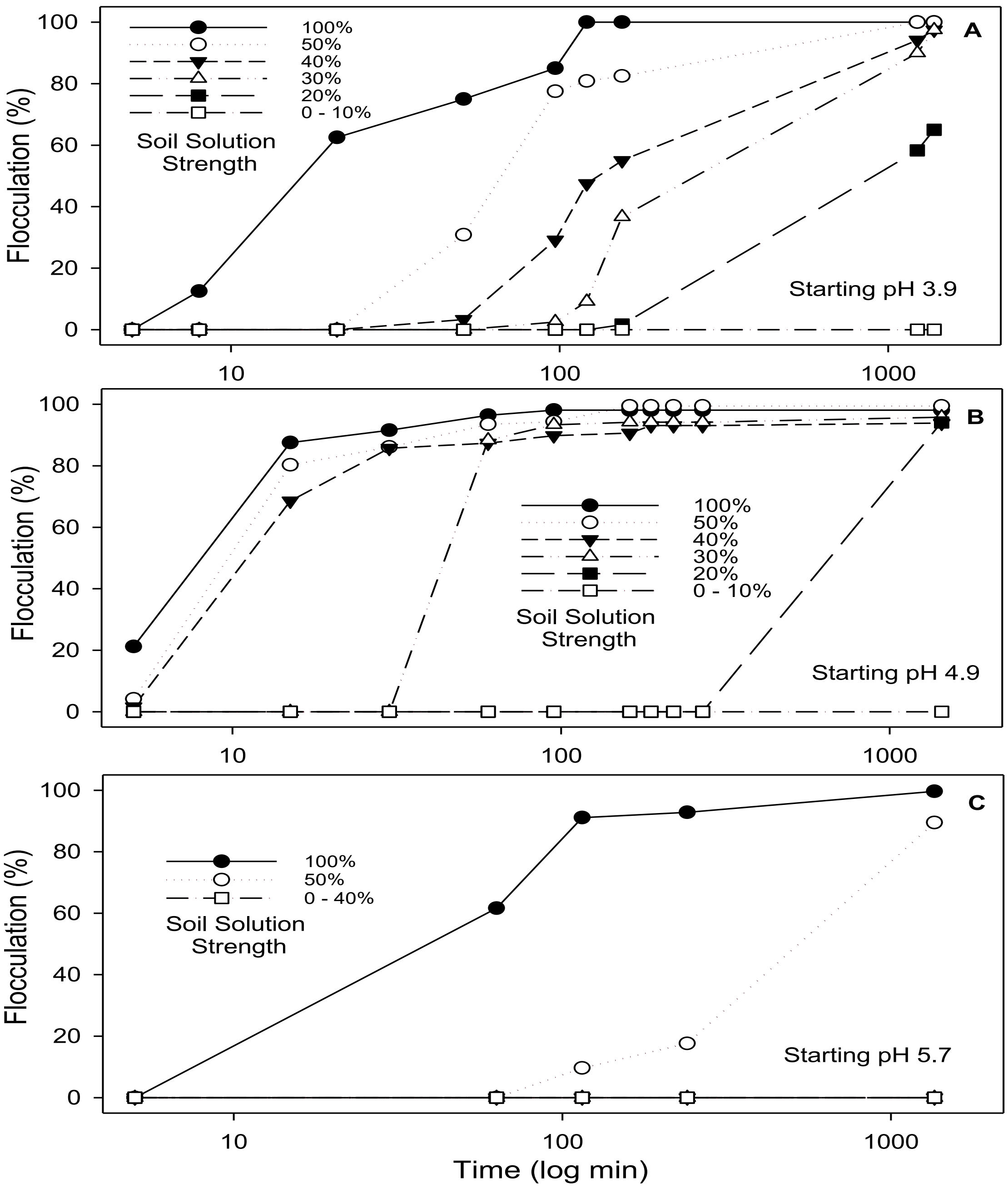


Figure 5.10*: A. Flocculation rate in SSW 1 at initial pH 3.9; B. Flocculation rate in SSW 2 at initial pH 4.9; C. Flocculation rate in SSW 3 with major cations removed initial pH 5.7.* ***Note****: flocculation occurred at time 120 hours (5 days) with SSW3 at concentrations as low as 30% (data not shown).*

Broad chemistry changes TCC and pH

Examination of the soil solution TCC indicated that flocculation occurred at a TCC of 1.12 for SSW1, 1.10 for SSW2, and 1.90 for SSW3 which had no appreciable Al content. The TCC remained the same before and after flocculation in all solutions.

The pH of SSWs increased at all concentrations with the addition of dispersed soil in the flocculation experiments as shown in Figure 5.11.

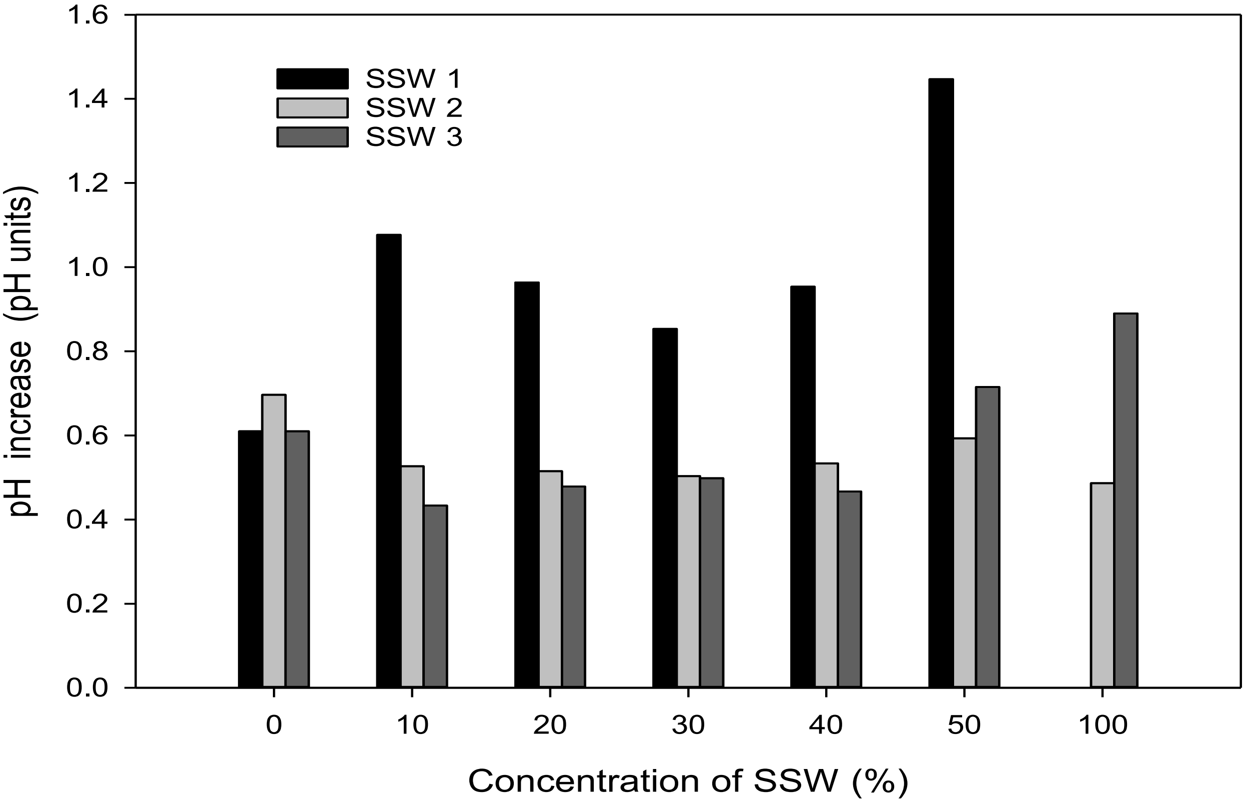


Figure 5.11: *Increase in pH occurring in synthetic soil waters (SSW) following flocculation experiments.*

5.4 Discussion

The pH of pore waters from tropical pasture topsoils was approximately the same as that of pore waters from native pasture topsoils in soil waters collected in February 2016 for assessment of pore size distribution. The pH of the tropical pasture topsoil pore waters from the July 2016 samples had a pH range of 4.9 – 6.7, whereas the range for the summer samples was 5.4 – 6.9. The lower pH observed in the tropical pasture topsoil pore water samples may be a reflection of improved Tempe cell preparation for pore water chemistry analysis. Regardless of the observed seasonal differences, the range of pH in pore waters from the July 2016 soil indicates substantial variation in the total content and composition of cations in solutions from tropical pasture (Kopittke *et al.*, 2016) and that this also could potentially change as soil moisture content changes (Dyer *et al.*, 2008).

Dispersion of the unworked soil samples of TPS, TPN and NPN upper B2 horizon materials was very low or negligible at all DI water pH examined. The dispersion rate of NPS soil was regulated to some extent by the pH of the DI water into which it was placed. Lower pH waters exhibited the slowest rate of dispersion of NPS over 24 hours; however, complete or nearly complete dispersion of NPS occurred at all pH values tested. This implied that mechanism(s) other than pore water pH affected the stability and development of stable macroporosity of sodic upper B2 horizons under tropical pastures.

No dispersion or only slight dispersion of the TPS, TPN and NPN soil samples occurred in any salinity of soil water in the range from 0 – 0.5 dS/m, however NPS soils dispersed completely in all salt concentrations. This is consistent with the findings of Quirk and Schofield (1955) who found that for various clays including kaolinite, illite and montmorillonite, the amount of NaCl in solution required for flocculation was 1.4×10^{-2} to 3×10^{-1} mol/L. The maximum NaCl in DI water concentration used in this chapter was 0.5 dS/m (5.1×10^{-3} mol/L). This indicates that weak solution salinity with NaCl alone is not responsible for apparent stabilisation of sodic upper B2 horizons under tropical pastures. It indicates that the combination of solution salinity and the particular type and combination of cations is more important than simple NaCl salinity. These results are similar to those of Mosley *et al.* (2017) and Dang *et al.* (2018) where exact concentrations of individual electrolytes were difficult to establish.

Both flocculation and dispersion of NPS soil material was regulated by the strength of synthetic soil water (SSW1 and SSW2). Soils flocculated rapidly at full strength SSW and more slowly at 20% SSW strength. Flocculation did not occur below 20% SSW. Conversely, NPS soil material did not disperse at concentrations above 20% soil solution strength water. Although solution pH had regulated dispersion speed, it did not cause flocculation in the range of pH 4.0 – 6.8. This indicated that the soil solution chemistry was offsetting sodicity-induced instability even at relatively dilute concentrations.

Flocculation occurred more slowly and in SSW3 where some dominant cations were absent, but only at 50% and 100% solution in the first 24 hours. This indicates that the electrolyte content of SSW 3 with the known flocculation-inducing cations of Al^{3+}, Cu^{2+}, $Fe^{2/3+}$, Mn^{2+} and Zn^{2+} removed, remained sufficient to flocculate a small amount of soil given adequate time. The TCC of + charges of ions that caused flocculation in SSW 3 was double that for SSW 1 and SSW 2, indicating that the lower divalent and trivalent cation content in SSW 3 was responsible for the significantly slower flocculation in much higher concentrations of artificial soil water. A potentially confounding factor in this experiment is described by Schofield's Law, which showed that the displacement of sodium is not constant with varying cation concentrations, and that cations such as Ca^{2+} (which causes flocculation or prevents dispersion) can be adsorbed to clay mineral surfaces more preferentially as its concentration in solution decreases (Schofield, 1947; Salmon, 1964). Nonetheless, the decreasing ionic strength of the diluted solutions eventually starves the solution of flocculating cations so that flocculation cannot occur.

The potentially toxic Al^{3+} cations were elevated in tropical pasture soil water but they were well below the low toxicity threshold of 5 mmol/L which inhibits root extension (Menzies *et al.*, 2015). Given the measured pH range of tropical pasture soil water, and its potential interaction with B horizon materials of higher pH, free Al^{3+} ions were unlikely to be commonly present in most soil waters under natural conditions. The filtered samples more than likely contained Al^{3+} bound to small organic ligands rather than existing as free Al^{3+}. Even in the case of the most acid soil waters in this study (pH 4.88), the interaction of soil and water caused an increase in the pH of soil water which would quickly immobilise Al^{3+}. The cumulative interaction between all cations in soil water and the soil solid surfaces is therefore more important than specific ions in this study, as the respective availabilities of specific cations vary with changes in soil water pH.

Stabilisation of sodic and dispersible subsoils can be achieved with amendments such as gypsum, lime (both Ca^{2+}sources), alum (an Al^{3+} source), or organic matter (Greene *et al.*, 1988; Gill *et al.*, 2009; Jafari *et al.*, 2012; Gourley and Sale, 2014)). Given the large number of potential flocculating cations in the soil solution from Towri, in combination with the observed changes in soil solution pH, and the changes in pH and solution chemistry which occur on interaction with soil solids, it was not possible using these data to identify a single cation which would be solely responsible for the observed flocculation in the present study

In summary, soil water from tropical pasture topsoils can both offset dispersion of NPS soil material and promote the flocculation of suspended NPS soil material. With respect to the stable macroporosity measured under tropical grass pastures, the soil solution from topsoils in tropical pastures possesses sufficient TCC and an appropriate mix of cations to confer stability to otherwise sodic and dispersible clays. Importantly, experiments show that stability is conferred to dispersible sodic B horizon material even at quite weak concentrations (20%) of soil water from tropical pasture topsoils. Experimental outcomes at lower soil water concentrations may be particularly relevant to soil stability at relatively high initial moisture conditions, preferential storm flow through macropores, and dilution of soil solution occurs as soil wets up during a rainfall event.

5.5 Conclusions

This experiment has been conducted in a laboratory with no attempt to simulate the topsoil overburden pressures that constrain sodic B horizons of Sodosols and Sodic Chromosols *in situ*. The experiments were conducted in a wet environment using small quantities of soil, which is quite different to normal field conditions. In this regard, the results are indicative only.

The results show that the cation-enriched topsoil solution from tropical grass pastures compared with native pastures has the capacity to stabilise sodic and dispersive B horizon soil materials, potentially through stabilisation of pore linings and ped surfaces. Even with selected major cations removed from solution, sufficient cations remain to facilitate partial stabilisation of sodic upper B2 horizons given sufficient time.

In this regard, it is reasonable to propose that the anti-dispersion and flocculation effects of soil water from tropical pasture topsoils may be a significant stabilising mechanism for sodic and dispersive subsoils. Chapter 4 demonstrated that upper B horizon macroporosity is significantly increased in relation to the greater proliferation (depth and density) of pore generating roots in tropical pastures. The greater macropores created are clearly stable or would not be able to be measured using saturation in Tempe cells. This mechanism potentially explains the macropore stability as well as the work of previous authors who found greater soil water penetration and storage under tropical grass pastures.

Chapter 6 Study Three: Soil Order changes at man-made fence line boundary associated with clearing native forest for grazing. New England Tablelands, NSW, Australia

6.1 Introduction

Background

This chapter is based on observations that some well-structured soils in the New England Tablelands of NSW with gradational texture profiles appear to have developed texture contrast profiles, A2 horizons, and coarser structure in approximately 150 years since they were cleared for grazing on managed pastures. Literature on soil type change from one soil Order to another is scant, but existing literature suggests that such soil Order changes should take millennia.

Reconnaissance soil survey (scale 1:250,000) of the Northern Tablelands of NSW was undertaken for the Northern Comprehensive Resource Assessment Project (NCRA) in the late 1990's (DLWC, 1999). Prior to this time, the area lacked consistent soil resource data and the resultant NCRA soil survey maps remain the only broad scale soil resource data set still available for the Northern Tablelands region of NSW. Members of the NSW soil survey staff (including the author of this study) working on this project made observations of the occurrences of A2 and $A2_e$ (E) horizons in cleared land within meters of the cleared side of many forest boundaries where there were gradational soil profiles in the forests. These observations were not specifically recorded in the survey, but created speculative discussion at the time. The importance of these observations is that there were no parent material or topographic changes at these man-made boundaries, and it is likely that soils on both sides of the fence were the same prior to clearing.

Personal observations of soil profile changes were made both at Tenterfield where Red Dermosols transitioned immediately to Kurosols at forest-pasture boundaries and at Uralla where Yellow Dermosols or Kandosols transitioned rapidly to Sodosols or Kurosols at forest-pasture boundaries. This transition appeared to only occur with moderately mafic parent materials and not in silica-rich parent materials such as quartz sandstones or coarse grained granites.

The aforementioned observations yielded the hypothesis that A2 horizons and texture contrast soil profiles can develop in response to land use change within relatively short periods where forest is cleared and pasture is established, and consequently soil Order (as per Isbell and NCST (2016) can be transgressed.

The objectives of this study are to:

1. Verify formation of A2, $A2_e$ or $A2_j$ horizons within recent historical timeframes along a forest to pasture boundary with no apparent change in parent material, slope, or surface drainage;
2. Using methods established in Chapter 4, and soil microbiological measurements, describe and compare the physical, chemical and biological attributes of soils in a forest and soils in directly adjacent cleared lands; and
3. Explore potential mechanisms for the apparently rapid development of texture contrast soil profiles and an A2 horizon.

Land use and soil changes

The introduction or removal of dominant plant species and management systems to or from an ecosystem has been shown to have an impact on the type of definitive soil physical and chemical processes operating at a site. Clearing land and subsequent cultivation or grazing can change a soil's physical attributes in terms of structure, porosity, dispersion, density, drainage, water holding capacity, and root length of plants over time (So and Aylmore, 1993; Greenwood and Hutchinson, 1998; Greenwood *et al.*, 1998; Greenwood and McKenzie, 2001). Commonly reported changes in chemistry include change in pH, CEC, P, N and a decline in SOC over time (Piñeiro *et al.*, 2010; de Souza Braz *et al.*, 2013; de Souza Celarino and Ladeira, 2017). Changes in soil physical and chemical attributes can substantially affect water infiltration and availability, runoff, the redox state of the soil and hence availability of nutrients and toxins, root growth and subsequent above ground plant production. Conversely, these attributes can be changed in a direction more favourable to agricultural or forestry production.

Yaalon and Yaron (1966) proposed that human activity can cause soil formation processes to take a new direction, establishing a new time zero (t_0) from which soil formation changes the pre-existing soil. With the setting of a new t_0, a new set of pedogenic processes may begin. The introduction or removal of dominant plant species may or may not represent an aggressive change in the direction or equilibrium of soil formation, but if the change is significant enough, then a cascade of the pedogenic feedback processes may come into play (Johnson *et al.*, 1990).

Studies on the rate of formation of texture contrast soil profiles and A2 horizons are rare and often based on observations in chronosequences where multiple sediment deposits of known ages are represented in terraces in alluvium. Walker and Butler (1983), working on terraces on the riverine plain in Southern NSW found that a texture contrast profile could form in approximately 15,000 years from unconsolidated alluvium. Yaalon (1971) suggested that argilluviation or the formation of E horizons requires thousands of years. Using weathering indices, Chittleborough *et al.* (1984) did not query ages for texture contrast soil formation and reported it to take 6,000 – 100,000 years. Most available studies show the transition of soils from undifferentiated alluvium to soil and ultimately to a texture contrast profile with a well–developed A2 or E horizon. The present study explored the potential transition of an already developed soil profile from a gradational texture profile to a texture contrast texture profile.

Site selection

The Tenterfield region was settled from 1836 – 1850 following the discovery of gold and other metallic mineral resources. The area was declared a municipality in 1871 (Tenterfield District Historical Society, 1949). The area including Woodbine property was cleared by ringbarking and established as fully cleared grazing land in the 1880's (landholder records). This is typical of much of the NSW New England Tablelands (Halliday, 2004; Reid, 2008; Haworth *et al.*, 2010).

Clearing of forested lands in the New England Tablelands of NSW was nearly completed to modern levels by the 1880's, with cleared land converted to a mixture of volunteer native and introduced pastures and legumes. Land clearing was incomplete due to the creation of State Forests and, later, National Parks, which provide reserves of native forest in a relatively undisturbed state adjacent to long term improved pasture lands.

Known sites of apparent transition from gradational to texture contrast soil profile forms at forest to pasture boundaries are relatively rare. Tenterfield in the Northern New England Tablelands of NSW was chosen as the location for this study because it possessed a site of apparent rapid soil change at a forest boundary that was noted in the NRCA survey of 1999 (DLWC, 1999).

Woodbine property is located on the northern edge of Forestland State Forest 15 km south-east of Tenterfield. Woodbine lies on the edge of the eastern fall into the Clarence River catchment (Figure 6.1). The geology of the forest and adjacent cleared lands is a Permian acid to intermediate porphyritic rhyodacite ring dyke formation (Colqulourn *et al.*, 2015).

The Woodbine site elevation is 1060 – 1120 m with a cool temperate climate and predicted annual rainfall range of 1100 – 1200 mm as shown in rainfall contours created from ANUCLIM (a spatial climate model developed by the Australian National University) (Hutchinson *et al.*, 2002). The site's elevation and position at the top of the catchment ensure an abundance of orographic rainfall from coastal air. Weather records for nearby Tenterfield (elevation 850 m) show January average summer temperatures ranging from 14.6 °C to 26.7 ° C, and winter temperatures in July from 0.8° C to 14.8° C (Weatherzone, 2017).

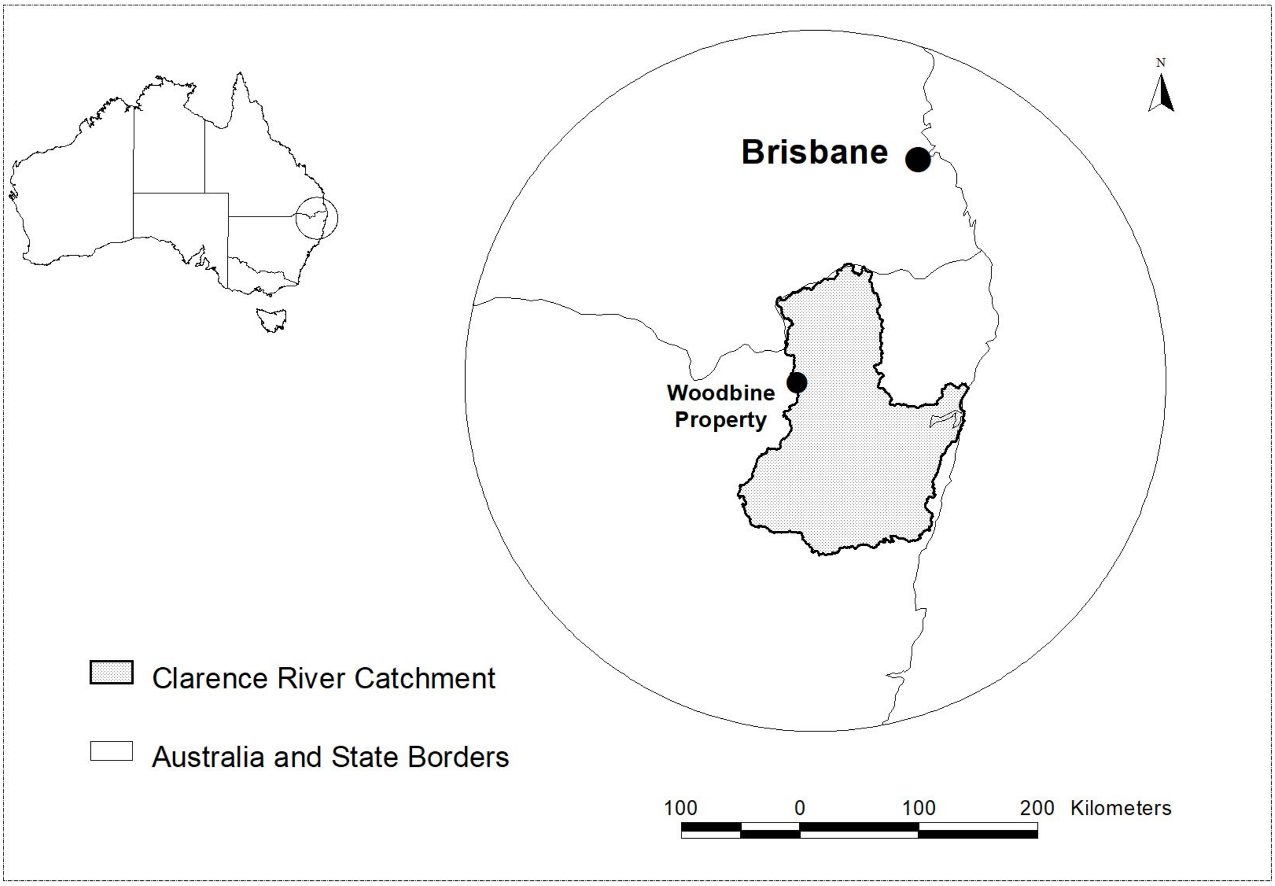


Figure 6.1*: Location of Woodbine Site in National Context.*

The average daily temperate for Tenterfield is 14.7° C. Temperatures for Woodbine are likely to be significantly cooler than Tenterfield because of the elevation difference (adiabatic lapse rate calculation of 0.6 to 0.98° C per 100 m) (Holton and Hakim, 2012). Evaporation is not measured at Tenterfield, but evaporation is likely to exceed rainfall for much of the year.

The Woodbine site's history is relatively well known and there is no step-down or indication of significant erosion on the cleared side of the forest boundary, and slopes of potential paired sites are less than 10% as recommended by Sauer (2015) for chronosequence studies. The known site history and the lack of erosion on-site make it an excellent site to assess decadal effects of land use change and management on soils. It can be considered a medium-term chronosequence (Huggett, 1998; Sauer, 2015).

Pasture species and management

Pastures at Woodbine are a mixture of volunteer native species and introduced species including: cocksfoot (*Dactylis glomerata*), microlena (*Microlaena* sp.), Parramatta grass (*Sporobolus creber*), paspalum (*Paspalum dilatatum*), kikuyu (*Pennisetum clandestinum*), Demeter fescue (*Festuca arundinacea*), Victorian rye grass (*Lolium rigidum*), silky browntop (*Eulalia aurea*), wallaby grasses (*Danthonia* spp.) and snow grasses (*Poa sieberiana* and *P. labillardieria*). Blady grass (*Imperata cylindrica*), bracken fern (*Pteridium esculentum*) and spiny-headed mat-rush (*Bursaria spinosa*) occur as low palatability native invaders of the established grass pastures.

Sown legumes include Haifa White and New Zealand White clover (both cultivars of *Tifolium repens*), and USA red clover (*Trifolium pratense*).

Superphosphate and lime have been applied to the soil since the 1950's to encourage legumes. Local landholders state that the effectiveness of superphosphate has been greatly reduced after the initial applications and clover species are difficult to maintain. The pasture at Woodbine has received a total of 2.2 t/ha of lime, and a minimum of 1 t/ha of superphosphate since the 1950's.

Forest type and species present

Native vegetation adjacent to Woodbine in the Forestland State Forest consists of a mixture of tall wet sclerophyll forest with shrub understory and more open tall forest with a grassy understory, which may be an artefact of historical fire regimes. Both vegetation types alternate along the forest boundary (NPWS, 1999). The forest is used for light grazing as well as a source of native hardwood trees. It is relatively undisturbed along main forest access roads and boundaries with no recent large scale clearing having taken place in these zones.

Tree species present include messmate (*E. obliqua*), white stringybark (*E. umbra*), with Sydney blue gum (*E. saligna*), New England blackbutt (*E. campanulata*) and occasional silvertop stringybark (*E. laevopinea*).

Where there is a shrub understory, species include tree heath (*Trochocarpa laurina*), hairy psychotria (*Psycotria loniceroides*) and scentless rosewood (*Synoum glandulosum*) with a ground layer dominated by bracken fern (*Pteridium esculentum*) and the common ground fern (*Calochlaena* *dubia*). Other groundcover species include spiny-headed mat-rush (*Lomandra longifolia)*, *Dianella caerulea*, *Smilax australis*, *Viola hederacea*, *Geranium potentilloides*, *Poranthera microphylla*, *Dichondra repens*, *Glycine clandestina*, *Coprosma quadrifida*, *Hydrocotyle laxiflora*, *Hibbertia scandens*, *Hibbertia dentata*, *Calochlaena dubia* and *Blechnum cartilagineum*.

Forest with more open grassy understory often have individual shrubs or clumps of blackwood (*Acacia melanoxylon*) and *Leucopogon lanceolatus*. Groundcover is very dense and often dominated by snow grasses (*Poa sieberiana* and *P. labillardieria*), bracken (*Pteridium esculentum*) and spiny-headed mat-rush (*Lomandra longifolia*).

6.2 Methods

Site location and sampling pattern – paired sites

A space for time substitution (Richter Jr and Markewitz, 2001) approach was employed for sampling and the site established as a small chronosequence. This approach is often used in ecological studies with different aged stands of vegetation because of the lengthy time frame that would be required which would exceed those of a conventional field trial, which in Australia is normally less than 5 years. A randomised block design was established with nine paired soil pits across the fence line. Soil Profiles 1 – 9 were in pasture, and Profiles 11 – 19 were in forest. Randomisation was achieved using randomly generated distances from the south western end of the site, forming a 1.5 km long pair of parallel transects from southwest to northeast (Figure 6.3). An additional opportunistic site (Profile 20) was described in forested land at the northern end of the property.

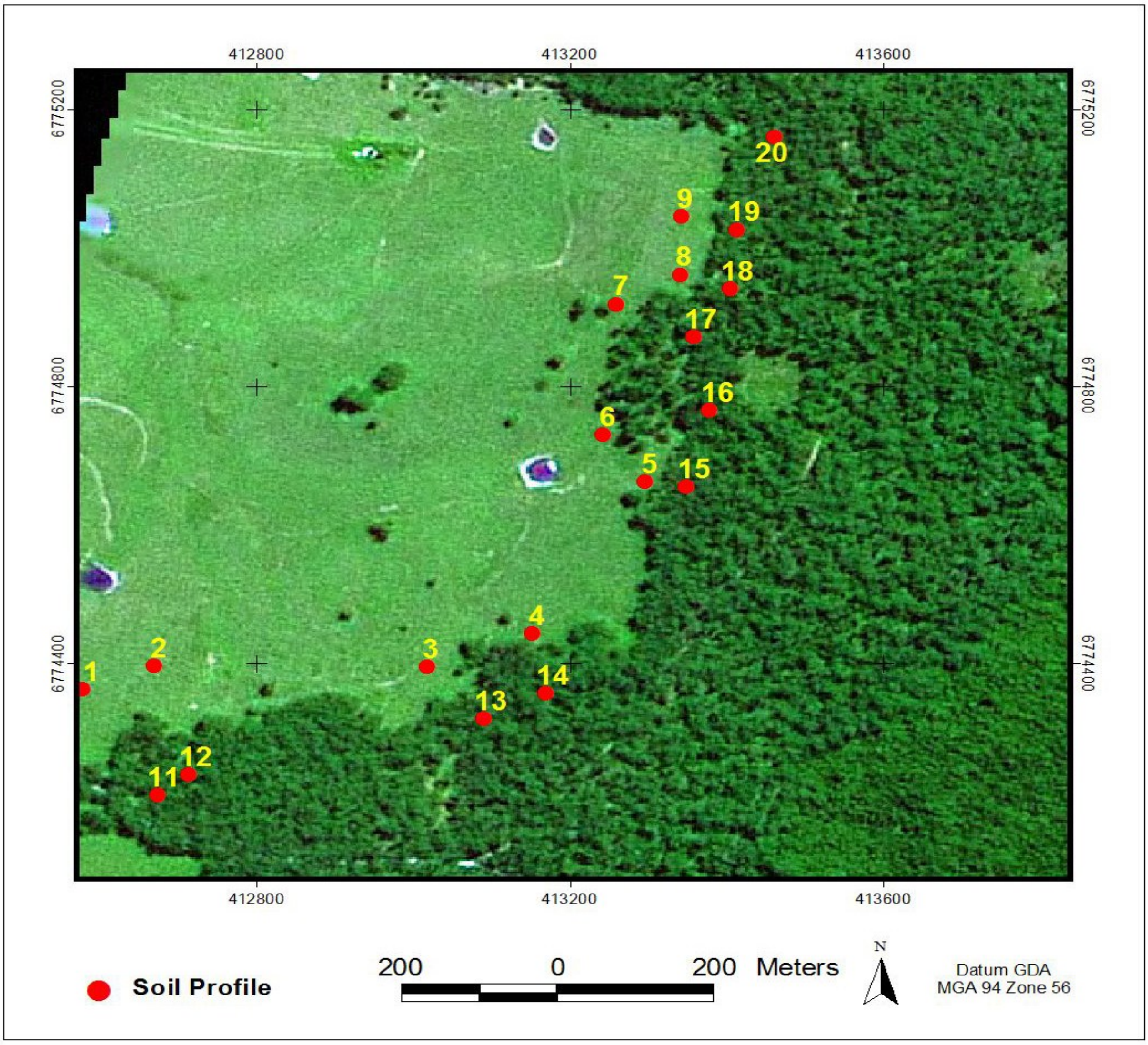


Figure 6.3: Location of field sampling sites – Woodbine. Source: SPOT 2002 Image.

Soil profile description and field data recording (0 – 150 cm)

Soil pits were constructed using an excavator to 1.5 m in January to February 2017 in each land use and numbered as shown in Figure 6.3. Each site and soil profile was described by horizon according to the National Committee on Soil and Terrain (NCST, 2009). The Australian Soil Classification (ASC) was used to classify the soil profiles (Isbell and NCST, 2016)

Particular care was taken to accurately record dominant ped size classes for each soil horizon, visual fine (< 2 mm) root abundance, and visual tree root abundance (NCST, 2009). Root abundance was recorded per 10 × 10 cm quadrat (100 cm^2) within horizons in the following categories 0, 1 – 10, 10 – 25, 25 – 100, and > 100 roots. Roots > 2 mm diameter were dominated by tree or shrub roots with an average diameter of 2 cm and their abundance was recorded as 0, 1 – 2, 2 – 5, and > 5 roots per 10 × 10 cm quadrat (100 cm^2).

Dominant ped size and root count ranges were converted to median values within their range or accorded a top value of 120/10 × 10 cm for roots < 2 mm, and 6/10 × 10 cm for tree and shrub roots (> 2 mm diameter). Field pH was measured using Raupach indicator (Raupach and Tucker, 1959) and recorded by horizon.

Sub-samples were collected from each pit for soil chemical analysis and aggregate stability in water (ASWAT) (Field *et al.*, 1997) testing to 150 cm. Soil sampling increments were 0 – 10, 10 – 20, 20 – 30, 30 – 50, 50 – 70, 0 – 90, 90 – 110, 110 – 130 and 130 – 150 cm. Twenty surface samples (0 – 10 cm) were collected with a spade within a 5 m radius of each soil pit. These samples were bulked to obtain one composite sample representative of the surface depth increment for each pit.

Scoping survey

An exploratory survey was carried out on road cuttings in the Forestland State Forest to determine the consistency of soil types through the forested landscape (Figure 6.2). Four soil observations were made on hillcrest, midslope and open depression (drainage line) landform elements. Continuous observation whilst walking the road cuttings between soil profiles showed little variation in soil type.

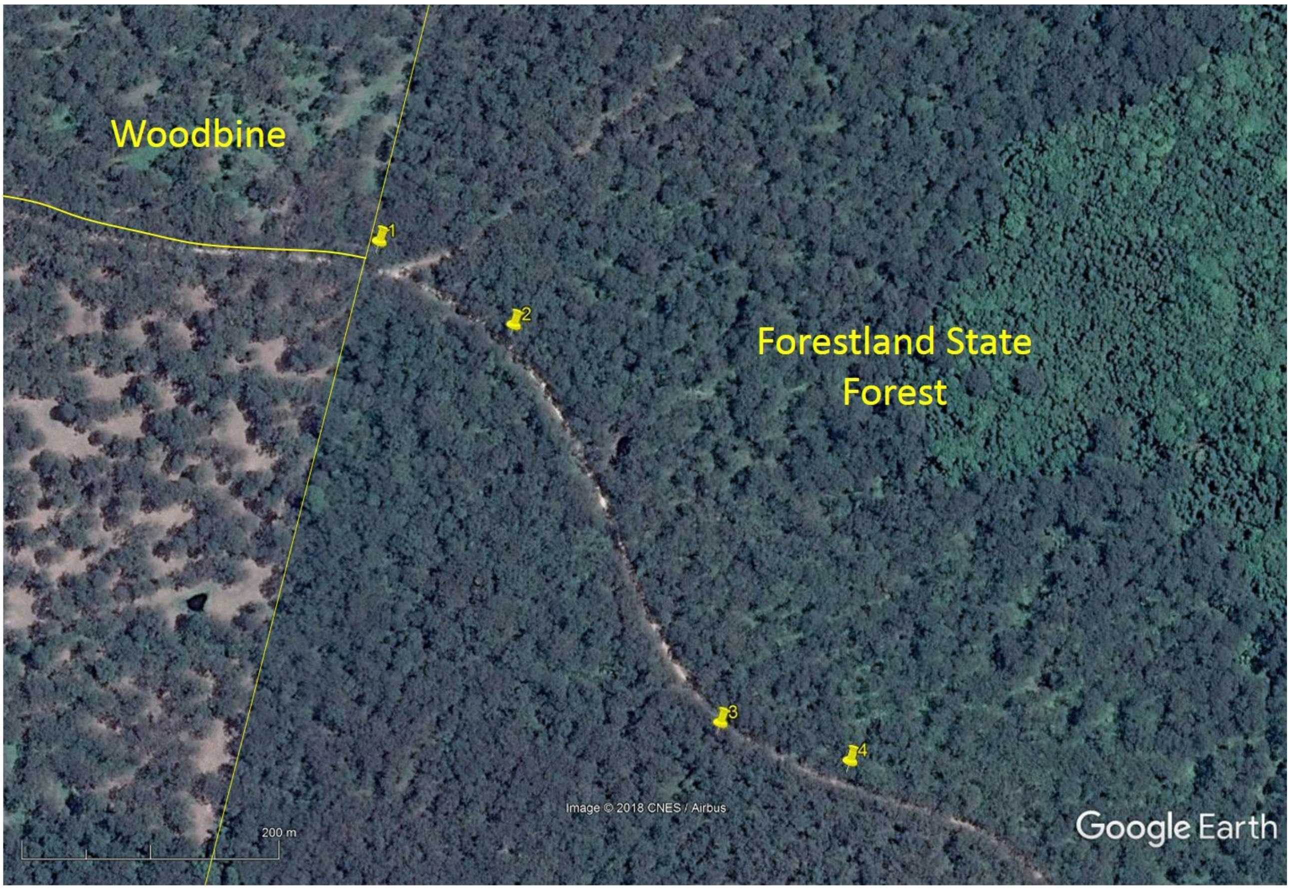


Figure 6.2: *Location of scoping study soil profile observations in Forestland State Forest.*

Soil physical properties

Particle Size Analysis

Particle size analysis (PSA) was carried out on subsamples for profile pairs 1 and 11, 3 and 13, and 8 and 18 for depths increments to 0 – 10, 10 – 20, 20 – 30, 30 – 50, 50 – 70 and 0 – 90 cm. PSA was undertaken by East West Laboratories in Tamworth (NSW) using the hydrometer method described in ASTM (2002). Pre–treatment with H_2O_2 was carried out on all samples to remove organic matter prior to measurement of particle distribution as per Bowman and Hutka (2002). As samples were taken incrementally, but measured differences reflected horizon thicknesses (not aligned exactly to horizon thickness), the results were presented graphically per profile pair. It was not possible to carry out valid statistical analysis for PSA because of the variable depths of texture changes associated with horizonation.

Bulk density

Paired bulk density (BD) cores were collected for depths of 0 – 10, 10 – 20, 20 – 30 cm with additional single cores taken at 40 – 50 and 80 – 90 cm. Storage, transport and handling for BD cores is described in Chapter 4. These cores were used to measure BD and soil water content (McIntyre and Loveday, 1974). BD and porosity was determined as described in Chapter 4.

Aggregate stability in water (ASWAT)

ASWAT testing was carried as described in Chapter 4 on soil sub-samples from all depth intervals collected.

Pore size distribution and potential flow through pores

Paired intact Tempe cell cores were taken from each profile at 0 – 10 cm, with a single Tempe cell core taken at 50 – 60 cm. Each core was polythene wrapped for transport and the pore size distribution and field capacity was measured as described in Chapter 4. Field capacity was derived from water content at -100 cm depth of water column (Vimpany *et al.*, 1987) as per Chapter 4. Pore size abundance was measured using Tempe cells for < 3, 3 – 6, 6 – 10, 10 – 15, 15 – 30, 30 – 37.5, 37.5 – 60, 60 – 150 and > 150µm. Porosity ranges for macro, meso and micro-porosity are defined in this study as > 30 µm (above field capacity) for macropores, 3 – 30 µm (Engelaar *et al.*, 2000) for mesopores and < 3 µm for micropores.

In addition to using Tempe cells, very large pores resulting from rotted-out tree roots were recorded separately because they were not adequately represented by the Tempe cells. It was necessary to account for these large pores in the pore size distribution because they have a disproportionate effect on water flow. These very large vertical pores were recorded by horizontal excavation of at least 1 m^2 at 50 cm depth in each pit, and counting the number of megapores, created by rotted out tree roots and recording their diameter. These large pores were referred to as “megapores” and were not practical to sample in Tempe cells because of their extreme diameter which would have confounded results for the main soil mass. The mean diameter of megapores was 2 cm. Using the Hagen-Poiseuille law for pipe flow, it was possible to provide a modelled interpretation of the impact of the increased number of macropores in terms of increased theoretical water flow. Flow capacity (Q) is proportional to the fourth power of the radius (Chen and Wagenet, 1992) using the equation as per Chapter 4 in this thesis:

$$Q = \pi r^4 (P_1 - P_0) / (8nl)$$

where r is pore radius;

$P_1 - P_0$ is the height difference between the surface and bottom of ponded water expressed in Pascals (pa) (10 cm of water = 981 pa);

n is fluid viscosity of water (assumed to be 1.01 for this interpretation); and

l is length of pore or pipe (where pores were assumed to be 10 cm long).

The subsoil macropore distribution from Tempe cells, was considered along with the additional potential flow effects of megapores. Potential flow rates (Q_{max}) were calculated for each per macropore pore size category (determined using Tempe cells) with a 10 cm head of water. Calculations assumed macropore radius equal to half the diameter for pore size groups from including 30 – 37.5, 37.5 – 60, and 60 – 150. The > 150 μm diameter pore group was assumed to have a pore radius of 150 μm and the megapores (recorded separately from Tempe cells) a radius of 10,000 μm (1 cm).

Average number of pores per pore size group per m^2 of soil at 50 – 60 cm was calculated by dividing the total area occupied by each pore group within 1 m^2 and dividing that area by the theoretical cross sectional area of an individual pore within that group. Theoretical Q_{max} was then calculated by multiplying the calculated flow through one pore in each category by the number of pores in each category in 1 m^2

Soil chemical properties

All preparation for chemical analysis was carried out as described in Chapter 4 by the Analytical Services Unit at The University of Queensland. Soil testing included soil pH in water (1:5), electrical conductivity (EC 1:5), Colwell P, exchangeable bases (Ca, K, Mg, Na, including Al) and cation exchange capacity (CEC), total organic carbon (TOC) and total nitrogen (TN). TOC and TN measures were multiplied by BD to estimate the stocks of TOC and TN, respectively, on a volumetric basis. EC (1:5) was converted to the equivalent of EC (saturated extract) or EC_e by multiplication through a conversion factor related to field texture (Slavich and Petterson, 1993). Soils were additionally analysed for free iron content at 50 cm to determine if the ASC class may have been a Ferrosol (with > 5% free iron) (Isbell and NCST, 2016) using the citrate-dithionate extraction method of Rayment and Lyons (2011)

Conversion of horizon data to equivalent depth intervals

Ped size, root abundance, and soil texture as represented by median clay content from field textures were recorded by soil horizon (NCST, 2009). These data were splined using the equal area quadratic spline technique described in Chapter 4 in order to compare attributes measured by horizon with specific depth increment sampled data.

Soil microbiology phospholipid fatty acid analysis

Phospholipid fatty acid analysis (PLFA) was used to measure microbial biomass and broadly evaluate microbial community composition (Buyer and Sasser, 2012; Willers *et al.*, 2015). Samples for PLFA were taken using a 12 mm wood chisel that was flame sterilised between sites using a small blow torch. Three samples were taken in each pit at 5 cm depth, and a further 3 samples taken at 50 cm depth. Samples were placed in sterile 50 mL falcon tubes and frozen at – 18ºC for transport. They were transported on ice from Tenterfield to The University of Queensland, then stored at -80ºC prior to processing. PLFA samples were then freeze dried prior to transport to Italy for PLFA.

Freeze dried PLFA samples were sent to The University of Palermo, Italy for PLFA analysis. Measurement of PLFA was undertaken on 3 g subsamples of prepared soil using the method described in detail in Laudicina *et al.* (2016). Results were reported both in mol and in mg PLFA/g soil. The provenance of PLFAs was assigned in accordance with previous reports (Frostegård *et al.*, 1993) (Table 6.1).

Statistical analyses

Soil properties were analysed by treatment (pasture and native forest) and distance from the western end of the fence using a General Linear Model. Treatment was considered a fixed effect and distance along fence a random effect in the analysis. These statistical procedures assume equal variance on each side of the fence, although some variables appeared to have unequal variance (e.g. average ped size). An F-test was used to test for equal variance and where that assumption was violated, a paired t-test was used instead of ANOVA.

All statistics were calculated using Minitab 16®. All depth increments, as well as B1 horizon (variable depths) were summarised as graphs. P values ≤ 0.05 were considered significant. If $0.05 \leq p \leq 0.10$ then it was considered that a trend was observed.

Table 6.1: *Microbial groups indicated by Phospholipid fatty acids. Compiled from Laudicina et al. (2016) .*

Fatty Acid Name	**Microorganism group**	**References**
14:0, 16:0, 17:0, 18:0, i17:0, 15:0, a15:0, i15:0, i16:0, 16:1ω7c, cy17:0, C18:1ω9t, 16:1ω5c, 18:1ω9c, 18:2ω6,9c.	Total Microbial biomass	Frostegård *et al.* (1993) Leckie (2005)
14:0, 16:0, 17:0, 18:0, i17:0, 15:0, a15:0, i15:0, i16:0, 16:1ω7c, cy17:0, C18:1ω9t,	Total Bacterial Biomass	
15:0, a15:0, i15:0, i16:0	G-positive (G+) bacteria	Zelles (1997) Zogg *et al.* (1997)
16:1ω7c, cy17:0, C18:1ω9t	G-negative (G -) bacteria	Zelles (1997) Zogg *et al.* (1997)
16:1, 5c; 18:1w9c; 18:2W6,9C.	Fungal biomass	Frostegård and Bååth (1996)

6.3 Results

Field properties

Soil profile descriptions were recorded on NSW Soil Data Cards and entered into the NSW Soil and Land Information System (SALIS). The full field data set is registered in NSW as Soil Survey number 1005336, where a permanent online record of soil profiles is stored. The full dataset of field records and soil profile descriptions may be downloaded from the NSW soil data portal, E-Spade at http://www.environment.nsw.gov.au/espde2webapp.

Forest soil observations

Examination of road cuttings in the Forestland State Forest showed that all soils visible in continuous road cuttings on crests or upper slopes (< 10% slope) and midslopes (< 25%) were Red Dermosols. One soil observation (Profile Observation 3) was in an incipient drainage line and was a bleached Red Kurosol. These soil observations were used to create a representative cross section illustrating the dominance of Red Dermosols in the forest, and the change to a texture contrast soil in pasture at the forest margin (Figure 6.4).

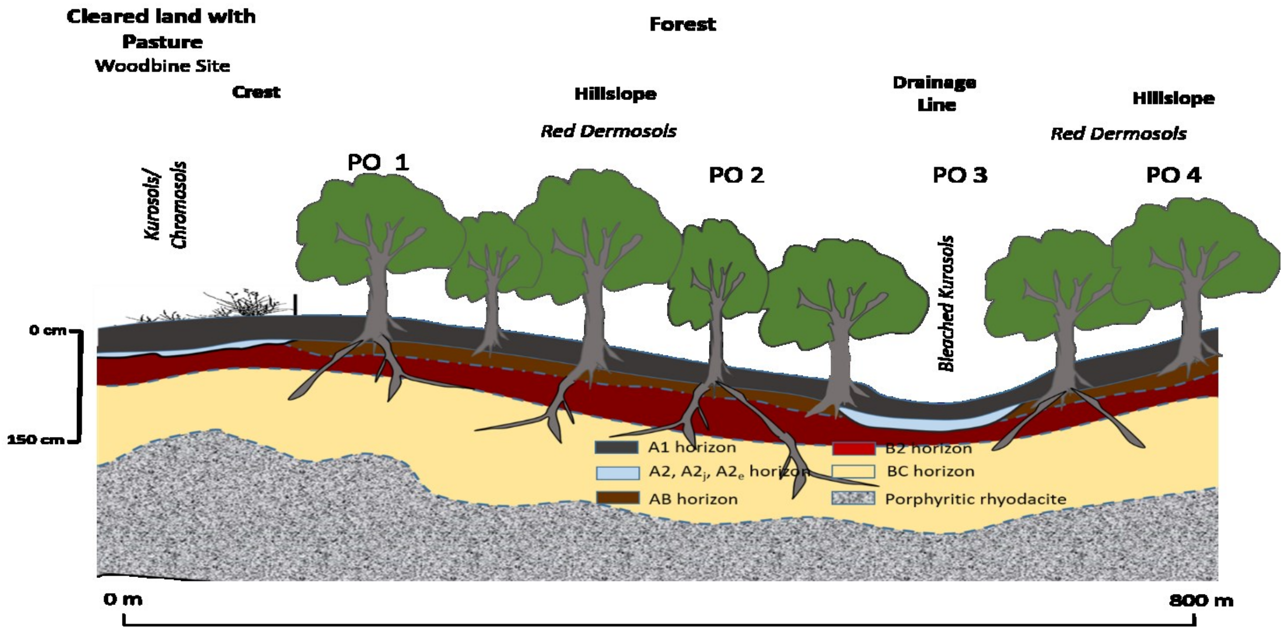


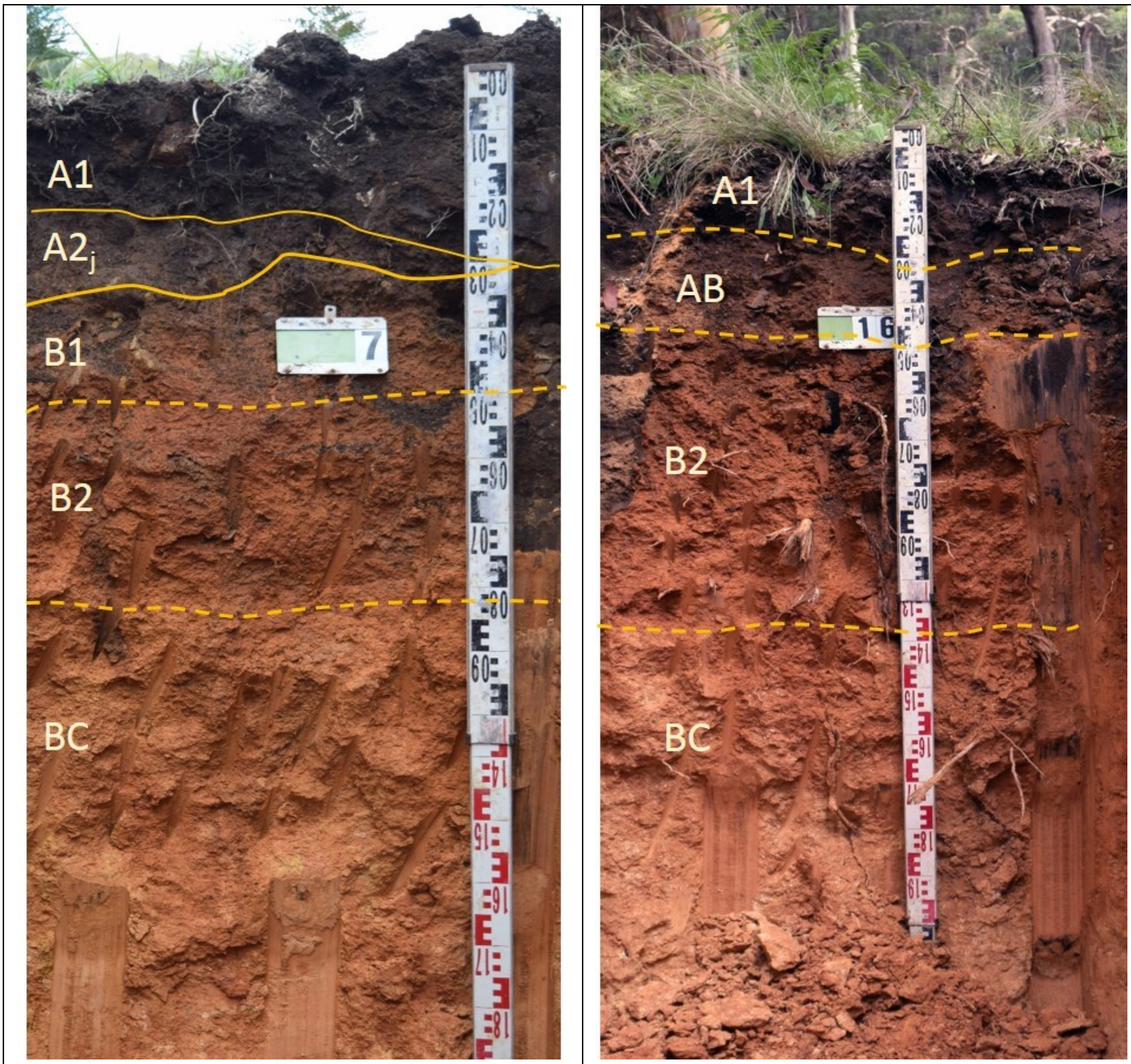
Figure 6.4: *Schematic cross-section from pastured land into uncleared forest; illustrating the dominance of Dermosols in the forest and the change to Kurosols/Chromosols with some A2 horizon development at the boundary of cleared land into pasture. Solid lines represent (< 5 – 50 mm) sharp to clear soil horizon boundaries. Dotted lines represent gradual or diffuse soil horizon boundaries. PO = profile observations in forest shown in Figure 6.2 above.*

Paired sites along forest/pasture boundary

Soils under pasture ranged in classification from Red and Brown Kurosols to Red and Brown Chromosols. All of the pasture soil profiles were acidic texture contrast soils and had similar appearance with slight colour and pH variation down the profile yielding the different classifications. All pasture profiles exhibited a varying degree of A2 horizon development from a discontinuous A2 horizon, an $A2_j$ horizon or an $A2_e$ horizon. All forest soils were Red Dermosols and were characterised by the presence of an AB horizon (Table 6.2 and Figure 6.5).

Table 6.2*: Summary of Soil Types at Woodbine*

Land Use	**Australian Soil Classification**	**Profile Number**
Pasture	Humose Dystrophic Red Kurosol	9
	Humose Dystrophic Brown Kurosol	6
	Humose Mesotrophic Brown Kurosol	1,7
	Humose Subplastic Red Chromosol	2,5,8
	Humose Subplastic Brown Chromosol	3,4
Forest	Acidic Subplastic Red Dermosol	11, 12
	Humose-Acidic Subplastic Red Dermosol	13 – 20



Profile 7: Pasture soil profile showing A2$_j$ horizon in moist condition.

Profile 16: Forest soil profile showing AB horizon or mixing horizon of A and B. Note large roots.

Figure 6.5*: Representative pasture (left, profile 7) and forest (right, profile 16) soil profiles showing soil horizonation. Solid yellow lines denote sharp to clear soil horizon boundaries (< 5 – 50 mm). Dotted lines denote gradual to diffuse soil horizon boundaries (> 50 mm).*

Roots

There was no significant difference in fine (< 2 mm) root counts between the pasture and forest soils examined (Figure 6.6 A). Total root counts which included all roots (< 2 mm and > 2 mm) were converted to cross sectional area of roots in cm^2 per 10 × 10 cm (100 cm^2). The total root abundance was significantly higher throughout the soil profile in forest as compared with pasture (Figure 6.6 B).

Figure 6.7 illustrates the marked difference in root sizes in near surface soils between pasture and forests. fix diagram

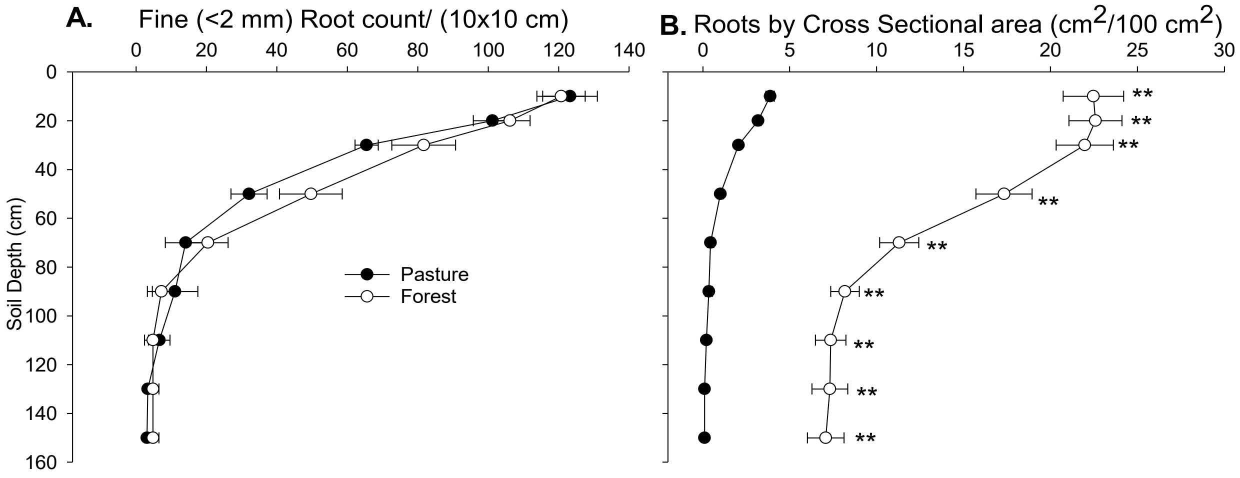


***Figure 6.6**: Root abundance. A. Roots < 2 mm diameter. B. Cross sectional area of all roots counted. Error bars represent one standard error of the mean. Asterisks indicate levels of significance: p ≤ 0.05*; p ≤ 0.01**.*

Profile 9: Pasture. Fine roots, predominantly less than 2 mm in diameter	Profile 20: Forest. Fine roots as on left but also tree roots averaging 2 cm diameter.

***Figure 6.7**: Representative pasture (left, profile 9) and forest (right, profile 20) near-surface profiles showing the differing root abundance and diameter of roots between pasture and forest soils.*

Soil structure (pedality and ped size)

Ped size under both pasture and forest increased with depth, but was significantly larger under pasture at depths from 20 to 150 cm. Mean ped size under pasture increased from 1 mm at 0 – 10 cm to 19 mm at 130 – 150 cm depth, whilst under forest ped size increased from 1 mm at 0 – 10 cm to 8 mm at 150 cm (Figure 6.8 A).

Comparative clay content using field texture

Clay contents estimated from field texture were significantly different, with lower clay content at 0 – 20 cm depth and significantly greater clay content at 50 – 150 cm under pasture compared with forest (Figure 6.8 B). The clay contents averaged over the whole profile were 41% under forest and 46% under pasture.

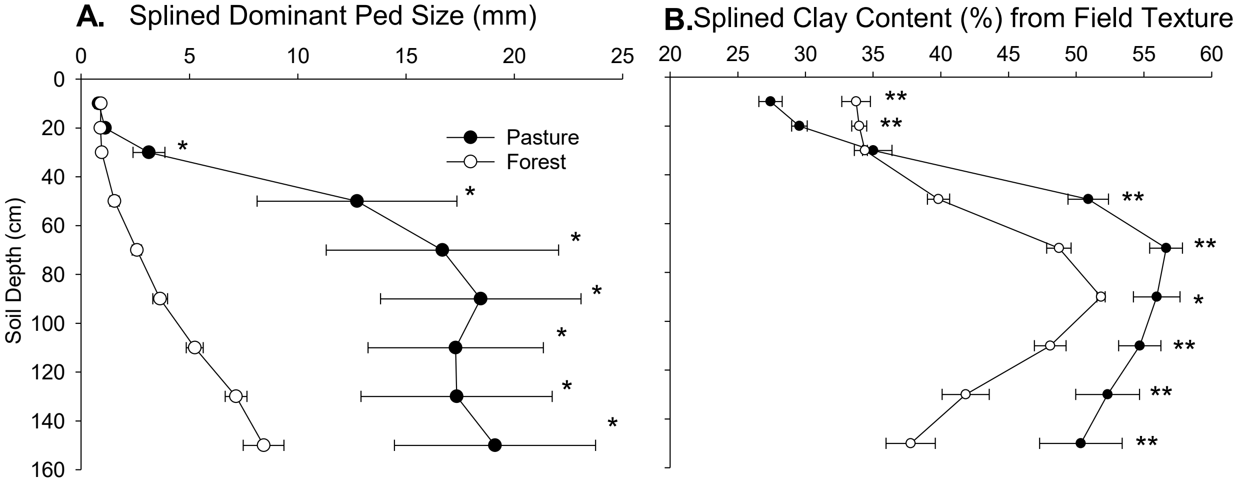


***Figure 6.8**: A. Dominant ped sizes with increasing depth down soil profile. B. Estimated clay content from field textures with increasing depth down soil profile. Error bars represent one standard error of the mean. ** $p \leq 0.01$. *$0.01 < p \leq 0.5$.*

Soil physical properties

Paired particle size analysis

Particle size analysis down the three pairs of profiles differed with horizon thickness, but reflected the more gradational nature of the profiles under native forest. Pasture soil clay content increased suddenly at the top of the B horizon reflecting the texture contrast. The average clay content over 90 cm depth under pasture and forest was almost equal for each pair of profiles with a maximum difference of 2% clay (Figure 6.9 A – C).

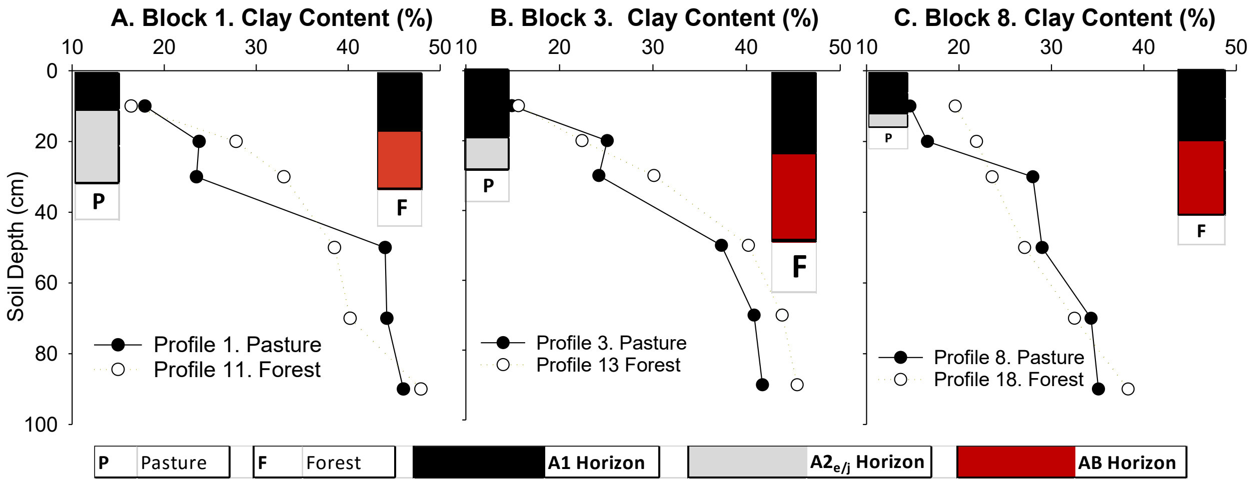


Figure 6.9*: Comparison of clay content of paired profiles using hydrometer method. Vertical bars show individual A horizon thicknesses.*

Bulk density

Bulk density was significantly less at 0 – 20 cm in forest soils compared with pasture soils. Mean bulk densities at 0 – 10 cm were 1.01 and 0.85, and at 10 – 20 cm were 1.26 and 1.17 for pasture and forest soils, respectively. From 20 – 90 cm, there was no significant difference in BD between soils under forest and pasture (Figure 6.10 A).

Dispersion (ASWAT)

All soils had to be re-worked for any dispersion to be observed. There was a slight trend for pasture soils to be more dispersive at 90 and 110 cm, but this was not significant at $p < 0.05$ (Figure 6.10 B).

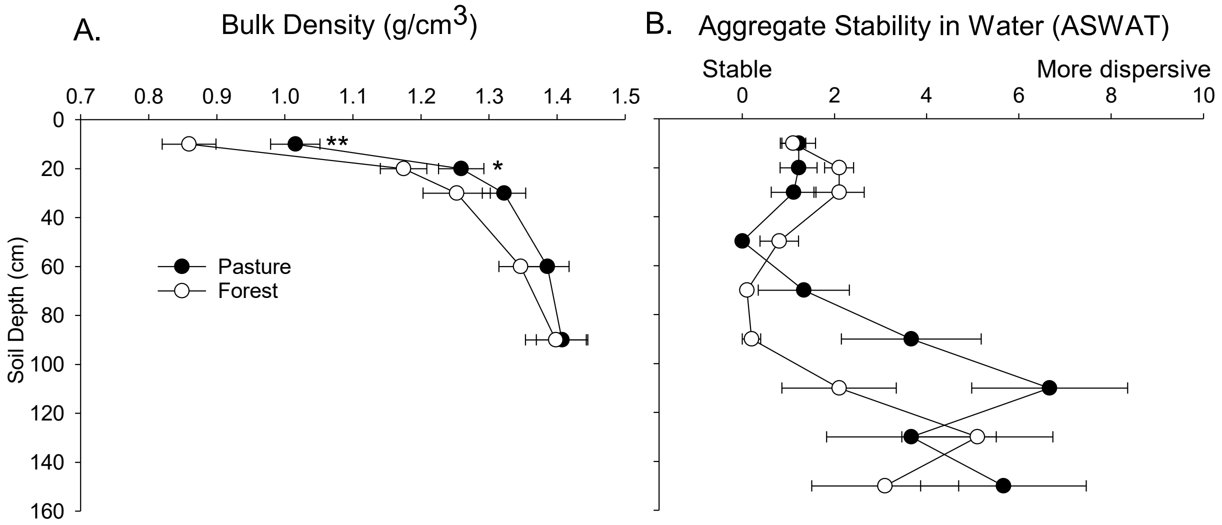


Figure 6.10*: A. Mean bulk density along a depth gradient to 90 cm. B. Mean ASWAT dispersion along a depth gradient to 150 cm. Error bars represent one standard error of the mean. ** p ≤ 0.01. *0.01 ≤p ≤0.5.*

Field capacity

Mean FC of topsoils was 50.7% for pasture, which was significantly higher than 40.7% determined for forest soils (p < 0.000). Mean subsoil FC was not significantly different with both pasture and forest having a mean FC of 36.4%.

Porosity, distribution and potential water flow

Mean total porosity of topsoils as measured by Tempe cells was significantly greater in forests (0.66 $m^3/m)^3$ than pasture soils (0.62 $m^3/m^{3)}$. These measurements excluded megapores that were observed separately. The pore size distribution was significantly different for all pore size groups, with forest topsoils exhibiting twice the number of macropores (> 30 μm), fewer mesopores (3 – 30 μm) and fewer micropores (< 3 μm) compared with pasture topsoils (Figure 6.11 A).

Mean total porosity of soil at 50 – 60 cm depth was not significantly different between forest and pasture soils, with means of 0.48 m^3/m^3 and 0.49 m^3/m^3, respectively. Pore size distribution as measured using Tempe cells was not significantly different between forest and pasture subsoils (Figure 6.11 B)

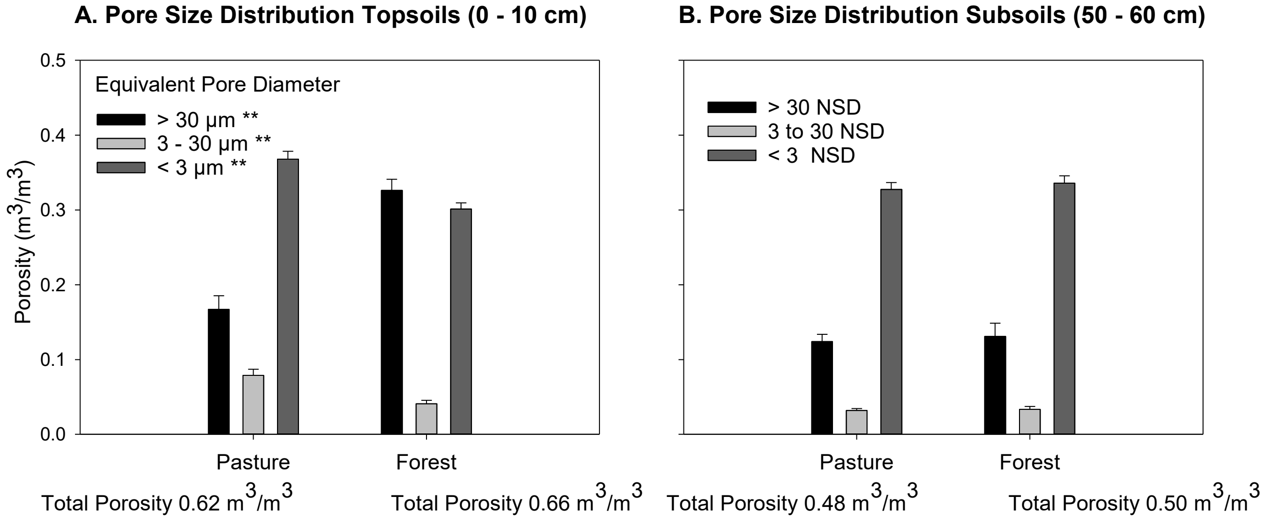


***Figure 6.11**: A. Pore size distribution of topsoils excluding megapores. B. Pore size distribution of subsoils excluding megapores (50 – 60 cm). ** $p \leq 0.01$. NSD. No significant difference. Error bars represent one standard error of the mean.*

Megapores created by tree roots, with an average diameter of 2 cm, were common in the forest soils but completely absent from pasture soils (Figure 6.12). Megapores occurred at a density of 10 megapores/m^2 in the forest. The overall contribution of these megapores to total subsoil (50 – 60 cm) porosity and macroporosity was negligible, but theoretical potential flow rate calculated including the addition of megapores was highly significant.

The predicted theoretical potential maximum flow rate of water through the forest soil at 50 – 60 cm depth was two orders of magnitude greater than that for the flow of water through the pasture subsoils. Total potential maximum water flow through subsoil macropores measured with Tempe cells, was the same for both forest and pasture at 0.225 m/day. Total potential water flow through the megapores (average diameter for 2 cm) was calculated to be 33.6 m/day (Figure 6.13).

***Figure 6.12**: Large vertical void (megapore) from rotted tree root in forest soil profile.*

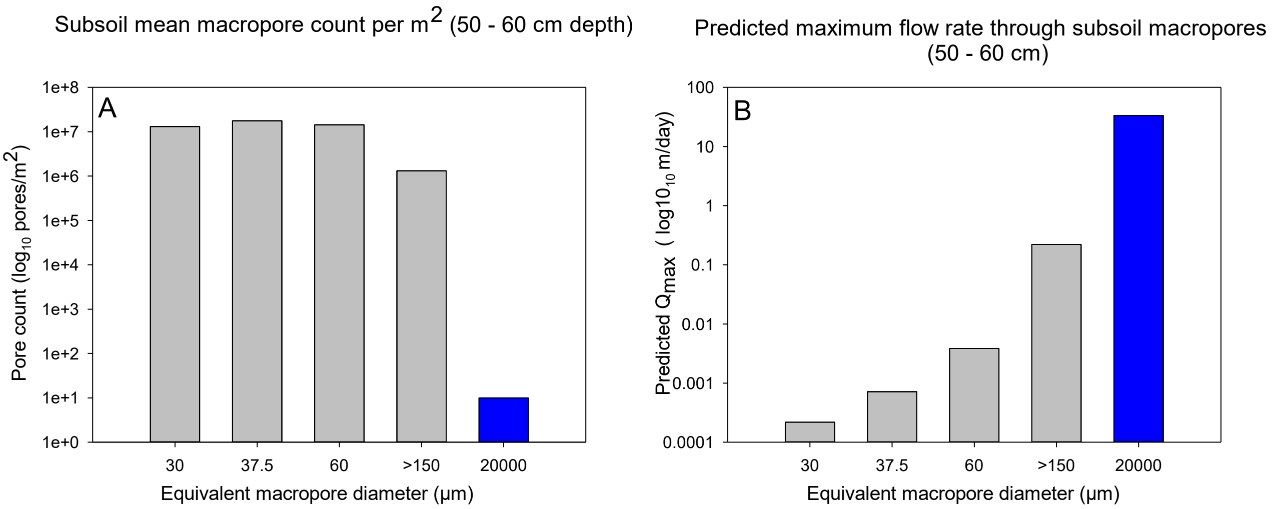


***Figure 6.13**: A pore count per m^2 for pores in macropore range. B. Predicted water flow rate (Q_{max}) for each pore group with 10 cm head of water at 50 – 60 cm depth in soil profile.* ⬜ *Macropores common to pasture and forest subsoils.* 🟦 *Megapores with mean diameter of 2 cm, in forest only.*

Soil chemical properties

pHw and Salinity (ECe 1:5)

Soil pH_w was not significantly different between forest and pasture at 0 – 30 cm depth, but was significantly lower for forest soils at 30 – 50, 50 – 70 cm depth; and significantly lower at 90 – 110 and 110 – 130 cm depth (Figure 6.14 A). Where differences occurred, they were equivalent to 0.2 pH units (30 – 70 cm) and 0.4 pH units (90 – 130 cm). There was a trend for higher pH in deep subsoils under pastures at 130 – 150 cm.

Soil salinity was significantly greater for pasture soils at 0 – 10 cm depth only, with a mean EC_e of 0.7 to 1.2 dS/m measured in forest and pasture soils, respectively (Figure 6.14 B).

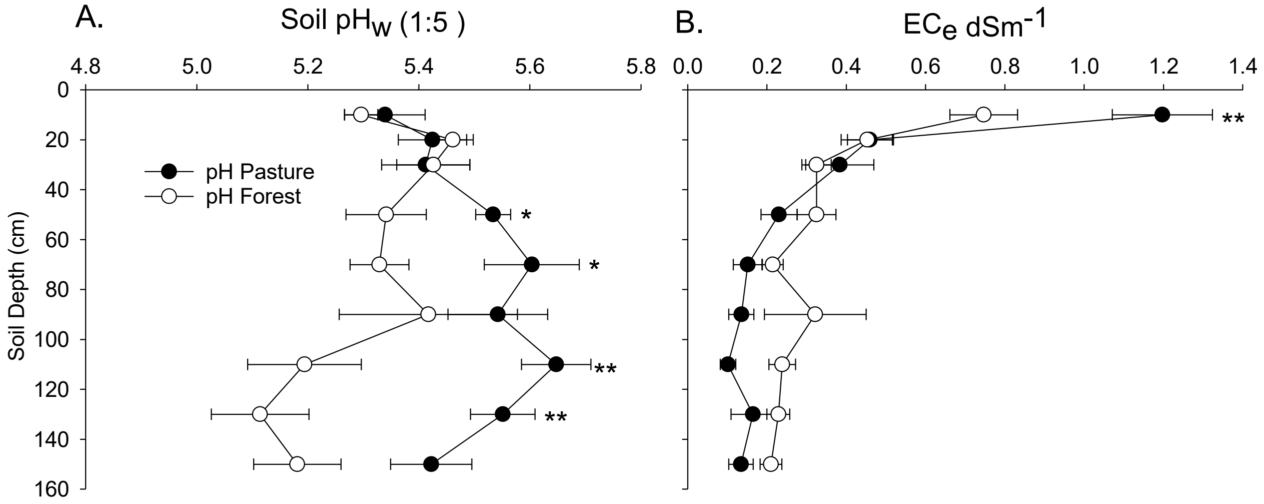


Figure 6.14: *A. Soil pH with depth down profile to 150 cm. B. Salinity as measured using EC_e with depth down soil profile. Error bars represent one standard error of the mean.* $** p \leq 0.01$. $*0.01 \leq p \leq 0.5$.

CEC and exchangeable cations

Cation exchange capacity was not significantly different between forest and pasture except at the 110 – 130 cm depth interval, where pastures had slightly higher CEC than forest. Overall CEC was low to very low down all soil profiles (Figure 6.15 A).

Exchangeable Ca was significantly greater in pasture from 30 – 130 cm depth by a factor of almost two, but was equal to that of forest soils at 0 – 30 cm depth (Figure 6.15 B). Exchangeable Mg was significantly greater in forest soils from 0 – 30 cm depth with no significant difference down the profile (Figure 6.15 C). Exchangeable K was significantly greater in pasture soils at 0 – 10 cm depth and greater in forest soils at 50 – 90 cm depth with no further significant differences through the profile (Figure 6.15 D).

Exchangeable sodium percentage (ESP) was significantly greater in forest soils down most of the profile, with the lower reaches of the forest soils approaching sodic (ESP > 6) (Figure 6.15 E). Exchangeable Al was not significantly different between forest and pasture soil profiles but was substantially more variable in pasture soils below 70 cm (Figure 6.15 F).

Calcium to magnesium (Ca:Mg) ratios were significantly higher in pasture soils throughout the profile from 0 – 150 cm depth (Figure 6.15 G).

Free iron measured at 50 cm in all profiles was not significantly different between pasture and forest with a mean of 2.17% (wt), and a minimum of 1.52% and a maximum of 2.97% (not represented as a Figure).

Colwell P

Available P was significantly greater at 0 – 10 cm and at 50 – 70 cm in pasture soils compared with forest soils (Figure 6.15 H).

Soil organic carbon

Volumetric SOC was significantly greater at the 30 – 50 cm depth in forest soils as compared with pasture soils (Figure 6.16 A). Mass based SOC was significantly greater in forest at 0 – 10 cm and at 30 – 50 cm than in pasture (Figure 6.16 B). Figure 6.16 C shows that the difference in cumulative SOC between pasture and forest soils continued to increase to 50 cm depth, after which no further the rate of increase of cumulative SOC was constant in both pasture and forest soils. This extra storage of SOC between 30 and 50 cm is a reflection of the presence of the AB horizon (or transitional mixing horizon) which was found only in the forest soils.

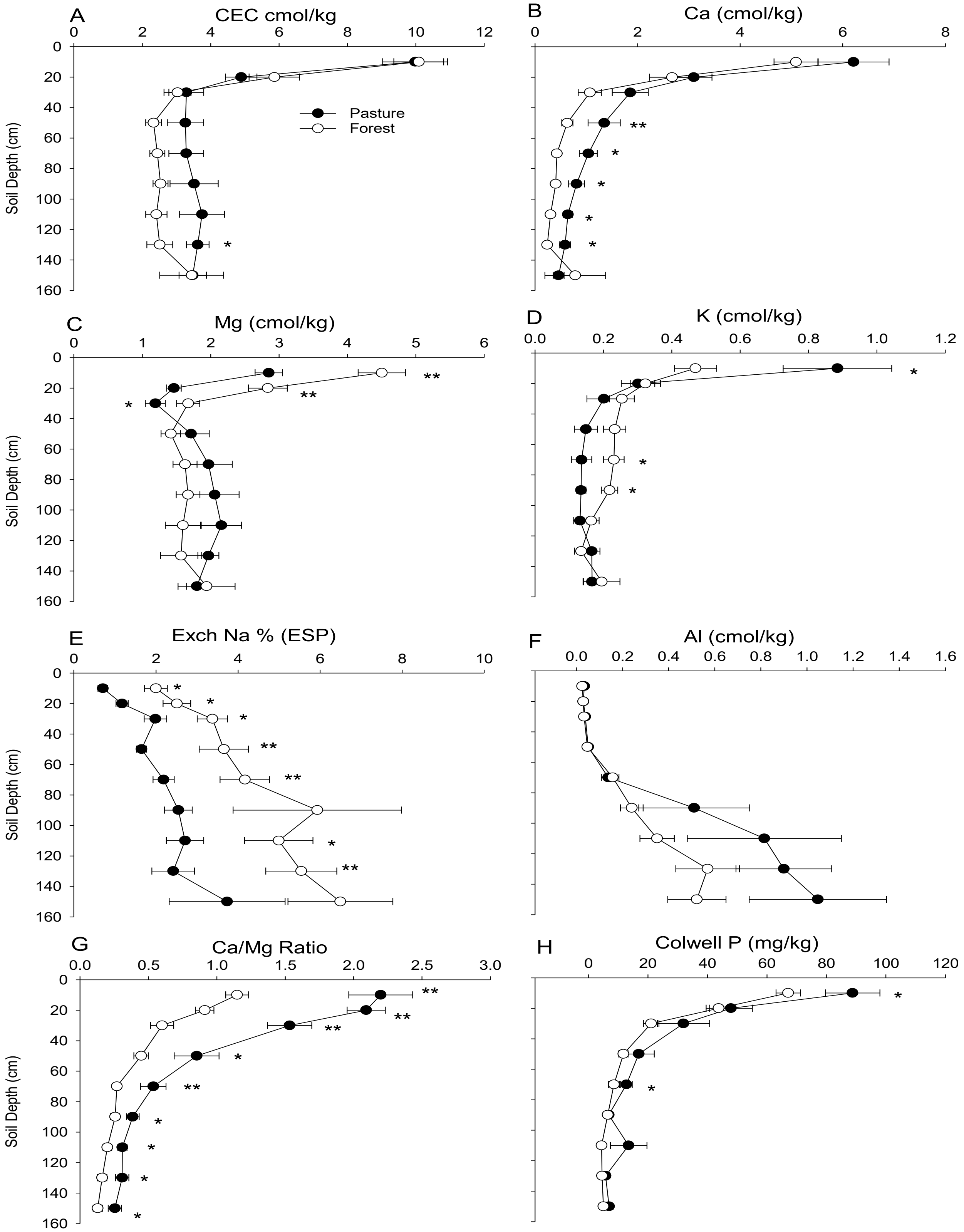


***Figure 6.15**: Soil chemical characteristics including: A. cation exchange capacity (CEC); exchangeable Ca; C. exchangeable Mg; D. exchangeable K; E. exchangeable sodium percentage (ESP); F. exchangeable Al; G. Ca/Mg ratio; H. Colwell P. ** $p \leq 0.01$. *$0.01 \leq p \leq 0.5$. Error bars represent one standard error of means.*

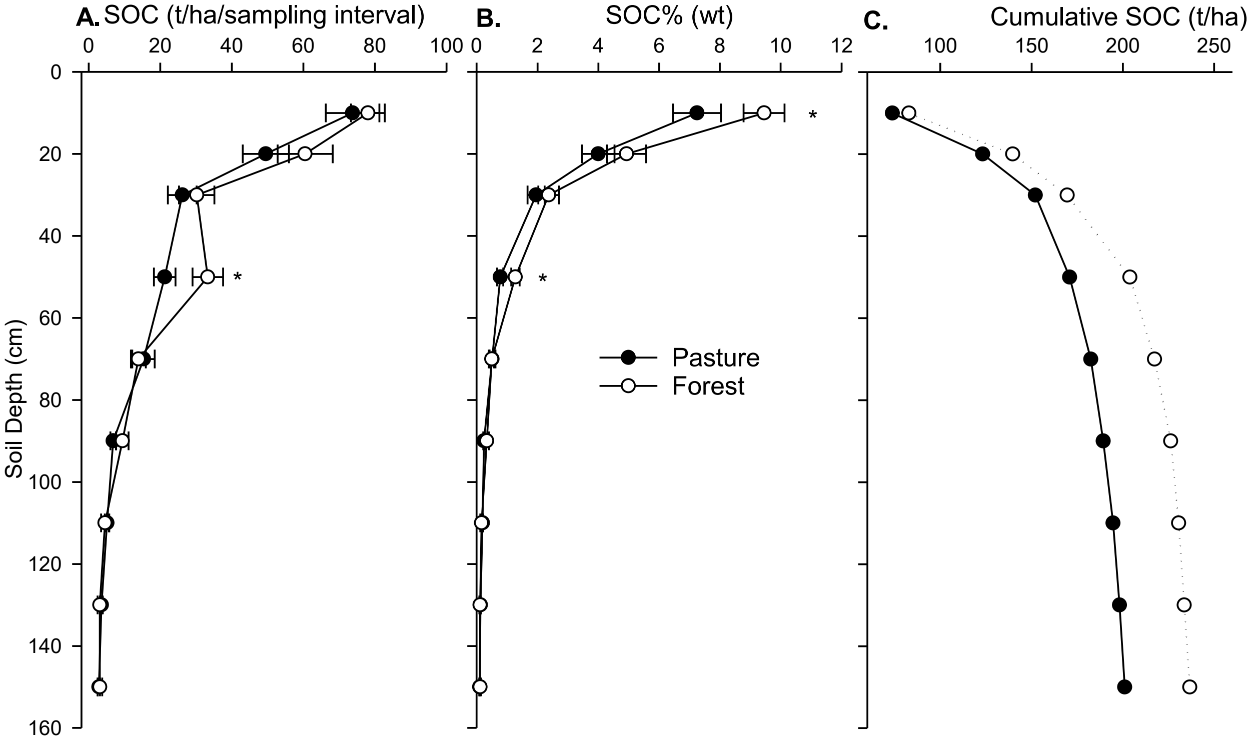


*__Figure 6.16__: A. Total SOC per sampling interval; B. Total SOC % by weight; C. Cumulative SOC down profile. Error bars represent one standard error of the mean. *0.01 $\leq p \leq 0.5$.*

Total nitrogen

Total nitrogen was only significantly greater in pasture soils compared with forest soils at 0 – 10 cm depth (Figure 6.17A).

C/N ratio

C:N ratios were significantly lower in pasture soils at 0 – 50 cm depth compared with forest soils, with no further significant differences in C:N ratio observed to 150 cm depth (Figure 6.17B).

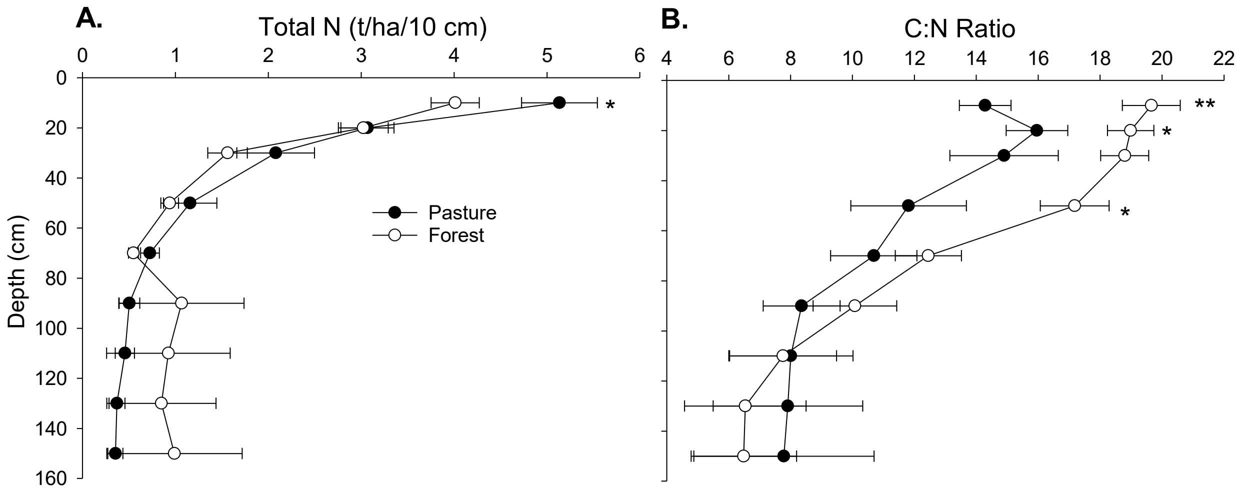


***Figure 6.17**: A. Total nitrogen. B. C:N Ratio. * 0.01 ≤p ≤ 0.05. ** p ≤ 0.01. Error bars represent one standard error of means.*

PLFA Analysis

There were no significant differences between PLFA groups which represented groups of organisms at 0 – 10 cm and at 50 cm, however depth was always significant, indicating a decline in biomass with depth (Figure 6.18 A – E). Mean total microbial biomass from 0 – 10 cm was equivalent to 147 kg/ha in both pasture and forest. Fungal:bacterial ratios and Gram positive:Gram negative ratios were not significantly different at 10 cm and 50 cm between forest and pasture (Figure 6.18 F – G).

Individual lipid analysis showed that there were some significant differences between PLFA's at 10 cm. C14.0, which is a general bacterial biomarker was significantly more abundant in forest soils than in pasture soils (Figure 6.19 A). C17.0cy and C18.1w9t, both gram negative bacterial PLFA's had slight but significantly greater biomass in forest as compared with pasture (Figures 6.19 I and M, respectively). C17.0, representing Gram positive bacteria had a significantly higher biomass in forest compared with pasture (Figure 6.19 H).

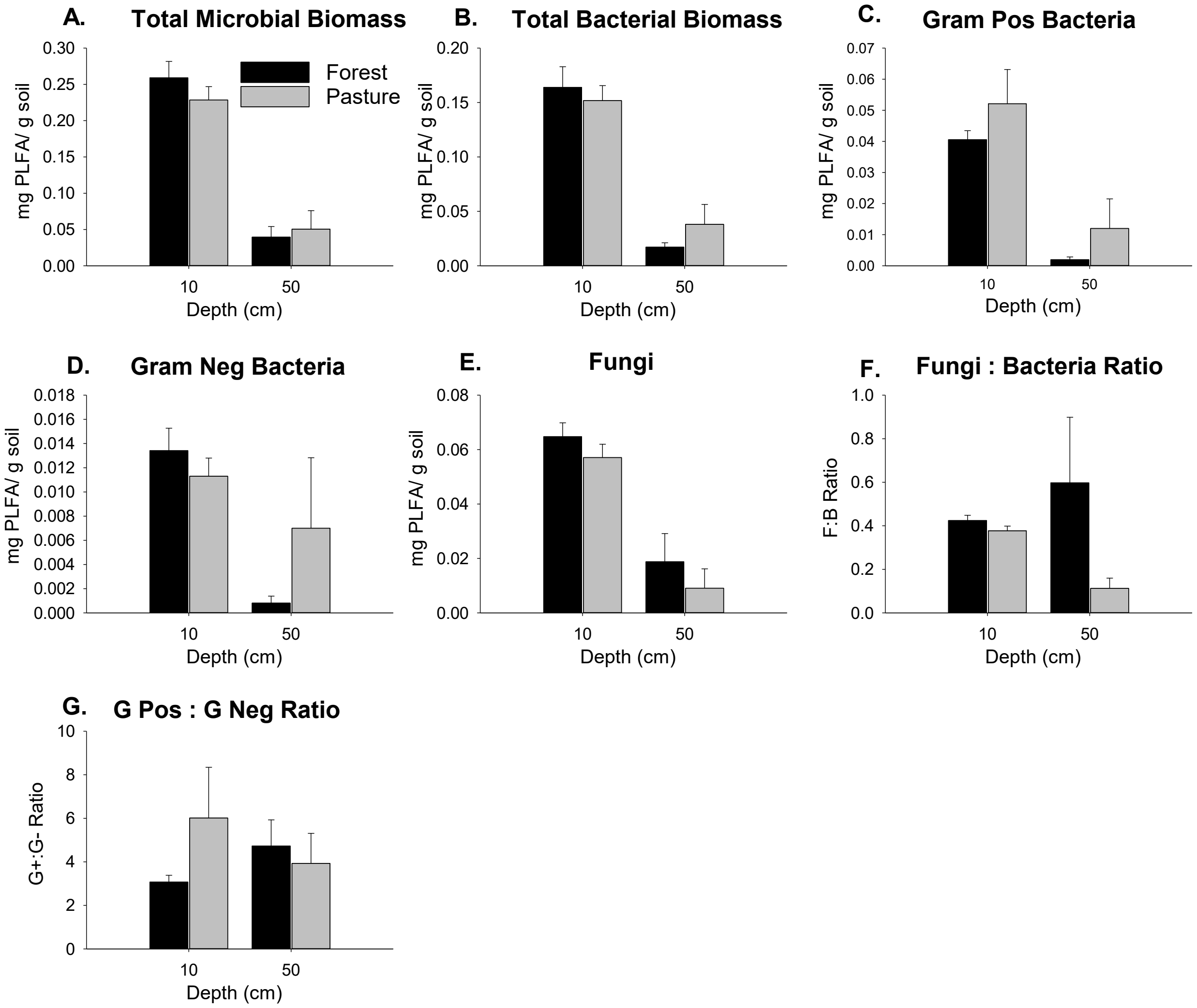


Figure 6.18*: A. Total microbial biomass, B – D. Biomass groupings reflecting microorganism groups. F. Fungal:Bacterial ratio. G. Gram Positive:Gram Negative Bacteria ratio. Error bars represent one standard error of their means.*

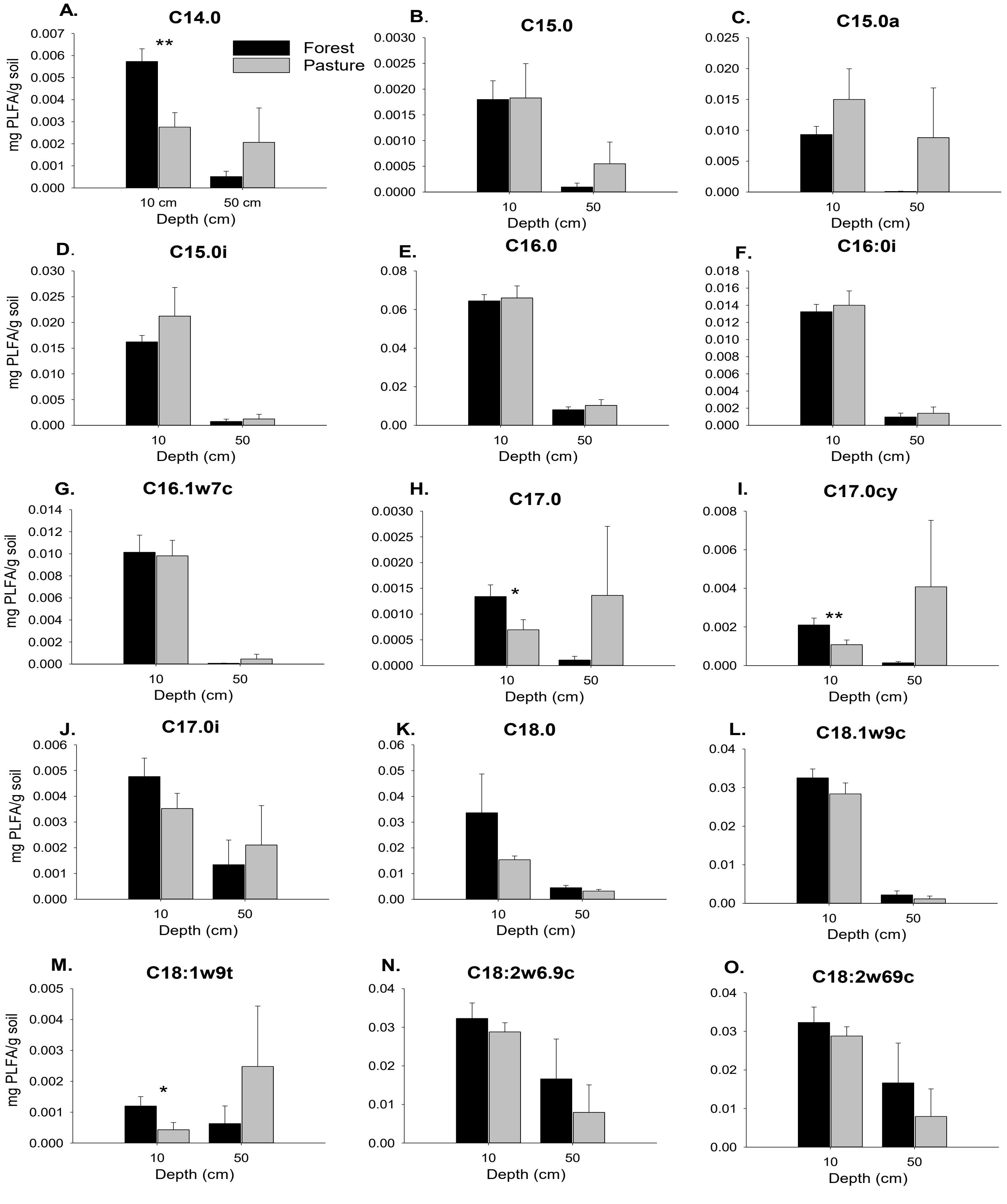


***Figure 6.19**: A – O. Individual PLFA's in forest and pasture. * 0.01 ≤ p ≤ 0.05. ** p < 0.01. Error bars represent one standard error of their means.*

6.4 Discussion

Field measurements

Soil type variation in forest

Soil observations along the road cutting-based transect showed that Red Dermosols dominated the forest except in extreme lower slope and open depression positions such as drainage lines where Kurosols were found. This survey verified that texture contrast profile forms were unlikely to occur on the upper and mid-slope positions in the landscape on the same parent material as the forest, and that Dermosols were likely to have been the pre-existing soil on in cleared pasture lands.

Soil type variation at study site

The soil type for all of the fence-line soil profiles in the forest was a Red Dermosol, characterised by a gradual increase in texture with depth down the soil profile with no clear or abrupt boundaries between horizons.

Soils under pasture were all texture contrast soils and were classified as both Red and Brown Kurosols and Chromosols. A texture contrast soil has an increase in clay content by field texture of a minimum of 20% with a clear to sharp (< 50 mm thick) boundary between A and B horizons (Isbell and NCST, 2016). The distinction between Kurosols and Chromosols was made on the basis of the pH of the upper 0.2 m of the B2 horizon (Isbell and NCST, 2016). The Kurosols are defined by a pH of less than 5.5 in this part of the soil profile. The difference in soil Order in this case was not dramatic because the pH which dictates one Order or the other at each site ranged over a narrow band of 0.3 units (5.3 – 5.6) that straddled the threshold between the Orders. It is likely that the application of agricultural lime to the pasture soils was sufficient to generate the slight pH difference required for the difference in soil Order. Therefore, the use of the generic term ‘acid texture contrast soils’ is appropriate in this case.

Of particular note were the dark AB horizons (a zone of biological horizon mixing or bioturbation) in the forest Dermosols compared with the clear or abrupt boundaries (0 – 50 mm thick) of the acid texture contrast soils under pasture, with all of the latter having some level of A2 horizon development. The $A2_j$ horizon, a weakly developed sporadically bleached A2 horizon (NCST, 2009), was a common feature of the pasture soil profiles that became more visible as they dried out. A2, $A2_e$ and $A2_j$ horizons are indications of impeded soil profile drainage as well as partial eluviation of organic matter and clay down the soil profile (Zinck, 2013).

Root depth and abundance

Fine (< 2 mm) root counts for both pasture and forest were not significantly different. Inclusion of large roots converted to cross-sectional area of roots showed that root abundance in the forest was significantly greater in terms of potential volumes of soil occupied. The potential for the root mass as a whole to impact on soil physical properties through creation of voids is related to both the abundance and the size of roots. Comparison between the abundance of different sized roots in forest and pasture soils using simple root counts showed minimal differences. However, conversion of root numbers to cross sectional areas showed a more than five-fold increase in soil volume occupied by roots in forest as compared with pasture topsoil (0 – 10 cm). This substantial difference in soil volume occupied by roots continued down the soil profile and a 72-fold difference in root cross sectional area was observed in the deepest subsoil (130 – 150 cm) in forests as compared with pasture. The potential for biological drilling of roots of all size categories down the soil profile is thus greatly enhanced by the larger roots present in forest soils (McCallum *et al.*, 2004). The large proportion of soil mass occupied by roots in forest soils yields potential for megapore creation which can have significant impacts on water flow rates or soil profile drainage (Chen and Wagenet, 1992).

Apparent clay content difference and PSA

Clay content from field texture classes are proximal in their relationship with laboratory particle size analysis, however they are critical to field soil classification in Australia (Northcote, 1979; Isbell and NCST, 2016). Isbell and NCST (2016) argued strongly for the use of field texture classes to classify soils as laboratory determined particle size analyses do not reflect the field condition of the soil. The removal of soil organic matter as recommended in Bowman and Hutka (2002) alters the apparent field texture, potentially leading to misinterpretation of field soil conditions.

The estimated mean whole soil profile clay content of forest and pasture soils were not significantly different, however the distribution of clay down the profile differed substantially between forest and pasture soils. The approximate clay contents shown in Figure 6.8 B show that clay appears to have elluviated from the surface layer (0 – 30 cm) to accumulate at greater depth along the profile in pasture soils. A more comprehensive laboratory particle size analysis was used to determine whether argilluviation was evident.

Particle size analysis in comparison with other approximations of texture gave the clearest results. Particle size analysis results for paired sites showed that a clear texture contrast was present in the pasture soils but not in the forest soils. Mean clay contents of whole paired profiles determined by PSA differed between forest and pasture soils by a maximum of only 2% from 0 – 90 cm depth. The lack of change in overall mass of clay or percent clay content within the soil profile to a depth of 150 cm was interpreted as an indication that clay illuviation occurred in pasture soils within the previous 140 years since clearing.

Soil physical properties

BD was significantly lower in forest soils compared with pasture soils at 0 – 20 cm depth, with correspondingly greater porosity. These differences are likely an effect of soil compaction by stock and heavy equipment used during occasional tillage for pasture sowing. Livestock stocking rates in Forestland State Forest have always been very low compared with the pasture, and the impacts of hooves and vehicles on increasing soil BD are well understood (Drewry *et al.*, 2000, 2004).

Soil pore size distribution in topsoils (0 – 10 cm) from forest and pasture areas were significantly different for macro, meso and micropores. Forest soils had twice as many macropores as pasture soils and half as many mesopores with a significant reduction in micropores. Using the Hagen-Poiseuille law for flow through a pipe, a doubling of pore size represents a potential 2^4 or a 16-fold potential increase in potential water flow rate.

Although porosity and pore size distribution in the forest and pasture subsoils (50 – 60 cm) were not significantly different, the Tempe cell data did not account for megapores which were clearly visible in profiles in the forest. These very large root holes had an estimated density of $10/m^2$ and were continuous though forest soil B and BC horizons. A theoretical model comparison of potential maximum flow rate through the normal macropores and through megapores showed an increase in drainage potential of two orders of magnitude. Although this was modelled data with assumed parameters, the impact of the megapores in forest soils has been shown to be substantially increased internal soil profile drainage. Where theoretical pore length and head of water were assumed to be the same for all pores, the relative hypothetical flow through forest subsoils was more than 149 times greater than that for pasture subsoils due to the existence of megapores in the forest.

Although the megapores would only function when there is free water after heavy rain, Woodbine has very high rainfall and is subject to prolonged wet periods. It was observed during the field sampling period that the pasture soils were often boggy, with reduced bearing strength for vehicles, compared with forest soils. The additional drainage provided by megapores would be a reasonable explanation for this difference in trafficability.

Soil chemistry

Most of the observed differences in soil chemistry are likely to be artefacts of the managed nature of the pasture system. Since establishment, the pasture has received 2.2 t/ha of lime, and at least 1 t/ha of superphosphate since the 1950's. The slightly higher pH of pasture subsoils compared with those of the forest, as well as the greater abundance of calcium and higher Ca:Mg ratio in the pasture subsoils reflect the liming history of the property. Greater N and P in pasture soils at 0 – 10 cm was likely a result of legume activity fixing atmospheric N. It is interesting to note that the difference in pH, which is a likely consequence of liming of the pasture soils, is sufficient to breach a pH boundary between soil Orders, showing that the Kurosols can potentially become transient Chromosols until the effect of liming has run its course.

Small differences or no difference in SOC are not uncommon for land use comparisons in lower temperature, high rainfall areas (Hoyle *et al.*, 2016). This was observed in the present study in the similar soil total carbon profiles for both pasture and forest soils. The main difference across the boundary between forest and pasture was that forest soils had significantly greater organic carbon stored at 30 – 50 cm depth. This reflects the greater supply of carbon provided through the abundant tree roots at this depth interval in the forest soils. Plant root activity is also reflected in the significantly greater C:N ratios of forest soils to 50 cm depth.

Possible mechanisms

Formation of texture contrast soils

The consistency of soil type under the forest in the paired profile comparison, and in the road cutting survey into the forest indicated a high likelihood that soil characteristics were consistent on both sides of the fence prior to land clearing. The classification of the forest soil as a Red Dermosol indicated that it is similar in physical characteristics to a Ferrosol except for having a free iron content of less than 5% in B horizons (Isbell and NCST, 2016). Free iron in soil conveys structural integrity and stability (Goldberg, 1989; Isbell, 1994). The Red Dermosols are consequently less structurally stable than Ferrosols and therefore potentially more likely susceptible to clay eluviation.

Soil compaction by livestock is well recognised; the introduction of first sheep and then cattle to Woodbine post-clearing likely account for differences in soil BD and porosity between forest and pasture soils despite the highly organic topsoils of the property (Betteridge *et al.*, 1999; Houlbrooke *et al.*, 2009). Previous studies have shown that the impacts of compaction and poaching by sheep and cattle can extend up to 20 cm below the soil surface, and generally result in a loss of macroporosity (Drewry *et al.*, 2000; Drewry *et al.*, 2004). The observed increase in BD, decrease in porosity and decrease in macroporosity observed in the top 20 cm in pasture soils in this study were consistent with compaction by stock.

Although the abundance and volume of soil occupied by fine (< 2 mm) roots were not significantly different between forest and pasture soils, larger roots (average diameter 2 cm) only occurred in the forest. The presence of large tree roots yielded a legacy of megapores at sufficient density to allow more than 100 times greater drainage potential through the forest soil mass relative to pasture soil. Removal of very large pores through age, collapse and potential infilling with surface soil, coupled with a reduction in macroporosity of pasture topsoils resulted in potential for periodic waterlogging of pasture surface soils as evidenced by the development of weak A2 and $A2_j$ (or E) horizons in pasture soil profiles.

Eluviation and illuviation of clays can occur when soils go through relatively intense wetting and drying cycles (Chittleborough *et al.* 1984). The removal of megapores from soils in pastures had the potential to reduce profile drainage dramatically, allowing for waterlogging to occur in near surface layers of soil. The removal of trees, altered soil hydrology and less transpiration results in more water that needs to runoff or drain (Burch *et al.*, 1987; Mohr *et al.*, 2013). The combination of physical compaction and minor dispersion as a result of mechanical compaction in the moist condition would release some clay from the surface soils for translocation down the profile as it drains. Paired forest and pasture texture profiles using both field texture and PSA indicate that this has occurred at the Woodbine site in soils under pasture following clearing of forest.

The results of microbial biomass analysis using PLFA show few differences between forest and pasture soils, although one difference was a subgroup of the Gram positive bacteria as measured using PLFA C17.0. The gram positive bacteria form spores as a survival mechanism in response to stress (Ringo, 2004). Soil moisture variation is a recognised stress factor in causing spore formation in this group of bacteria (Brockett *et al.*, 2012) and under the right circumstances some spores can be viable for millions of years (Cano and Borucki, 1995). The greater biomass of gram positive bacteria (PLFA C17.0) in the forest, is likely to be a reflection of wetting and drying cycles being more favourable for spore forming bacteria. This may also conversely imply that the pasture topsoils are generally moister than in the forest where soil water is used more rapidly.

The proposed mechanism is consistent with the work of Chittleborough *et al.* (1984, p.45), who found that the initiation of textural differentiation was a feedback process "produced through intrinsically morphological processes." They found that water moves more freely through A horizons than B horizons as soil morphology is altered and soil hydrology begins to change, creating temporary waterlogging during wetter seasons. This in turn induces reducing conditions that accelerate clay migration and mineral weathering. Poaching and compaction from stock initially break the continuity of megapores and macropores, so that they no longer serve to transport water from the A to the B horizon. The eventual reduction in megapores observed in this study may be sufficient to initiate this process as indicated by the large difference in theoretical modelled Q_{max} or potential flow through B horizons.

Although the timelines of existing studies such as Yaalon (1971), Walker and Butler (1983), Chittleborough *et al.* (1984) indicate that texture contrast soil formation requires to the order of thousands of years, most of these studies concentrated on soil formation from originally unconsolidated alluvium in terraced chronosequences. In this regard, the forest soils in the present study have already formed with a gradational texture profile. Based on the results of this study, only minor changes in soil physical characteristics were sufficient to have initiated the development of a textural contrast and an A2 horizon in varying levels of development within 150 years.

Separation of Kurosols and Chromosols

It is clear that texture contrast soils occur in all pasture soil profiles in this study, and that there is varying soil pH, hence the soil Orders of Kurosols and Chromosols. It is likely that the slightly higher pH Chromosols have developed purely in response to liming, as soil formation from a gradational profile to a texture contrast profile has progressed.

Potential implications

The changes to soil characteristics observed in this study were subtle and required close study of soil field and physical attributes. If the process of texture contrast soil formation were to continue, and upper B horizons were to become more dense, developing a strong $A2_e$ (E horizon), this would further limit potential water access to and storage within deep soil horizons. Areas experiencing this effect would become increasingly runoff dominated, and stored subsoil moisture, which is crucial for pasture production, would likely become increasingly limited as the process continued, thereby threatening managed pasture production.

In terms of potential groundwater recharge, Mohr *et al.* (2013) proposed that deep groundwater recharge under forest probably occurs through continuous macropores and megapores in the whole soil mass, and reported that deep recharge was absent from cleared lands due to loss of these large pores. The present study is complementary to the work of Mohr *et al.* (2013) demonstrating the existence of megapores and enhanced macroporosity in soils under forest, also indicating that soil profile internal drainage is substantially reduced following clearing. Mohr *et al.* (2013) had proposed the existence of megapores as an explanation for recharge and runoff changes following clearing.

Lack of soil profile drainage may alternatively lead to saturation of deep subsoils for prolonged periods of time, which in turn can have impacts on erosion risk in terms of mass movement potential. The morphologically similar Ferrosols on the NSW North Coast are prone to mass movement occurring as shallow rotational slumps on relatively low slopes when cleared (Jenkins and Morand, 2002). Steeper (> 15%) midlsope areas of texture contrast soils below the pasture soil description sites at Woodbine have examples of this type of slumping resulting from profile waterlogging. No slumping was observed within the forest on similar or steeper slopes (Figure 6.20).

Figure 6.20: *Rotational slump on hillslope (15% slope) in pasture soils. Woodbine.*

The New England Tablelands have also experienced significant dieback of trees over the past 100 years. This has been related to tree isolation and insect attack as well as changes in soil moisture regimes resulting from clearing (White, 1986). This study shows how soil water access to B horizons has been restricted by simple physical changes which have favoured clay argilluviation. A well-developed texture contrast soil will have impeded near surface drainage as well as restricted subsoil moisture compared with a Dermosol. If the processes observed in this study have been more widespread following clearing, then they may have been a possible contributor to eucalypt dieback.

6.5 Conclusions

Investigation of forest and pasture soils representative of a medium-term chronosequence showed that soil type change occurred at a straight fence-line boundary of a forest with cleared land in response to land clearing and grazing on managed pastures. Changes in the BD and pore size distribution of the soil profile associated with clearing and grazing, particularly the destruction of megapores, allowed clay eluviation to occur in upper soil layers of pasture soil. The development of a texture contrast soil texture profile and partial development of an A2, $A2_j$ or $A2_e$ horizon both represent artefacts of the change in soil hydraulic characteristics or drainage as well as diagnostic criteria by which the classification of these soil profiles has transitioned soil order from Dermosols to acid texture contrast soils. In addition to this structural transition, the subsoil pH difference which separates Kurosols and Chromosols is likely to have been created by repeated agricultural liming, thus lifting subsoil pH in Kurosols sufficiently to become Chromosols.

Chapter 7. Discussion and Conclusions

7.1 Introduction

The objectives of this thesis were to determine if soil formation or soil development processes can be changed substantially, take a new direction, or be accelerated by anthropogenic activity. This research involved several studies which had complementary or similar results to the work of other authors but also found unique results which add to the understanding of soil formation in Eastern Australia. These similar and complementary results are discussed briefly herein followed by new results. New results are mostly related to soil physical processes which initiate change are discussed in terms of their contribution to the understanding of soil development processes, implications at a site and landscape level in agricultural systems and the challenges faced in an uncertain climate future.

Study 1: Use of existing data

The examination of existing data for the Liverpool Plains region in Chapter 3 of this thesis did not show significant whole soil profile changes or development related to land use. However, the study did have similar or complementary findings to many other similar studie,s both locally and worldwide. The study described in Chapter 3 considered existing chemical and physical soil data as fertility factors, which both characterised soil groups and showed that SOC was a product of the fertility factors. Soil fertility factors were primarily related to soil group. Within soil groups, there was evidence that the higher fertility soils had been selected for cultivation, the soils of intermediate fertility for grazing and the poor fertility soils left in woodland.

The main findings were that SOC declines with intensity of land use so that forest or woodland, generally has the highest SOC irrespective of soil type and that cropping causes the greatest decline in SOC. This is well known and the subject of much research (West and Post, 2002; Lopez-Garrido *et al.*, 2011; Badgery *et al.*, 2013; Badgery *et al.*, 2014; Rabbi *et al.*, 2014; Jones *et al.*, 2016). The use of a large data set from the Liverpool Plains area in Northern NSW within a constrained climate range showed that the significance of land use overrides the wide range of native SOC levels in diverse soils within the landscape. SOC partitioning in the soil profile was found to be largely within the top 0 – 20 cm of all soil groups except the Vertosols under woodland, which had significantly higher C stores than all other soil groups under woodland to 85 cm depth.

Using whole soil profile fertility factors, the findings conflicted with the work of Banks and McKane (2002) and Gray *et al.* (2016), who predicted very high pre-development SOC in the western Liverpool Plains Region on poor soils (soil group: Other Soils in Chapter 3) and intermediate levels of SOC in Vertosols. The statistical results from the present study show that the opposite was probably the case pre-clearing. It is possible that a focus of these previous studies solely on near-surface layers, as well as personal field exposure to soils in their modern condition, skewed the methodologies and hence the results. To improve the estimations of those by Banks and McKane (2002) and Gray *et al.* (2016), a model based on fertility and potential SOC sequestration resulting from fertility, constrained by climate, may provide more useful results.

Early estimates of original soil carbon by Banks and McKane (2002) were based almost entirely on dominant soil type within a soil landscape and the field experience of the NSW soil survey team. Soil landscapes within Bio-regions were ranked into groups ranging from low to high SOC to create the pre clearing SOC maps for NSW (Webb, 2002). This expert panel approach did not use any statistical processes.

Gray *et al.* (2016) used existing SOC values from undisturbed sites, along with soil formation factors of mean annual rainfall, mean daily temperature (climate), silica index (parent material), relief (slope, wetness and aspect), weathering index as an indication of age, and a combination of land use disturbance index and vegetation cover (biota). These data were used in a multi linear regression model to map pre clearing SOC. If soil fertility factors were considered as variables with the SOC dataset used, the results may have been more precise.

Study 2: Towri – North-west Slopes and Plains

The study of sodic-texture contrast soils at Towri in northern NSW investigated the validity of observations made of subsoil structure changes resulting from the incorporation of tropical grass pastures and associated legumes into these soils. Regionally these soils are known for their lack of production under cropping or grazing, yet some areas have been brought into stable high production grazing systems through use of fertilised tropical pastures. Soil characteristics beneath volunteer native pasture and tropical pastures of known age were measured along a fence line and physical and chemical attributes of paired whole soil profiles were compared. The results of this study indicated that fertilised tropical pastures could initiate feedback processes in the soil which improve the soils' water holding depth as well as productive capacity.

Sodic-duplex soils have been the object of much research from a crop and pasture production perspective in Australia because their subsoil constraint-based limitations often result in poor production (Rengasamy *et al.*, 1995; Cotching *et al.*, 2001; McKenzie *et al.*, 2004; Jassogne *et al.*, 2007). Improvements in subsoil macroporosity of water and root penetration limiting upper B horizons has been achieved in the past with direct injection of organic matter rich fertiliser (Gill *et al.*, 2009) or by selection of cool temperate pasture species which "biologically drill" via penetration of roots (McCallum *et al.*, 2004) in cooler southern areas of Australia (here climate is more favourable to occasional cropping).

There has been relatively little investigation of sodic-texture contrast soils in the sub-humid areas of Eastern Australia. Large areas of sodic-texture contrast soils had originally been cleared for cropping, but a combination of climate variation and soil limitations has seen cropping mostly abandoned on these soils since the 1970's (Banks, 1998). Volunteer native vegetation including pastures do not utilise the subsoils, with most roots concentrated in the light textured topsoils. Increased soil moisture of B horizons in sodic-texture contrast soils associated with tropical pasture roots, and consequently greatly enhanced production from these soils, has been documented in other work (Murphy *et al.*, 2010; Murphy, 2014; Murphy *et al.*, 2018). The mechanisms of this increase were not fully investigated previously.

This current study showed comparable results with those which studied temperate pastures in southern Australia. The tropical grass pasture and legume mix at Towri, with deeper and more aggressive roots than native vegetation, was related to greatly increased macroporosity and potential rate of water throughflow in upper B horizons that previously limited both root and water penetration. A significant reduction in ped size through the subsoil below 70 cm indicated that the increase in macroporosity has continued to deep subsoil, representing a much greater surface area within the soil to absorb and release both nutrients and water.

Of particular interest is that the infiltration rate-limiting upper B horizons under both native pasture and tropical pasture had similar sodicity and similar structure. However, the aggregates in the upper B horizons under tropical pastures were stable, whereas those under native pasture were dispersive. This implied that a change had occurred beyond the creation of macropores through this layer. Simple creation of more macropores in dispersive material would normally result in collapse of pores through loss of soil structural integrity upon wetting (Chorom *et al.*, 1994). The soil surrounding or lining macropores would need to be stable for the pores to effectively allow water to flow downward to deeper horizons.

Subsequent dispersion and flocculation experiments were undertaken using simulated soil waters based on that extracted from topical pasture topsoils. These experiments determined that tropical pasture topsoil water could induce structural stability in sodic and dispersive material from upper B horizons from native pasture sites, even at concentrations as low as 20% of the strength of soil water. For pores and peds to be stabilised, only the linings of the pores and ped surfaces would need to be in contact with such soil water, not the whole soil mass. The macroporosity is both greatly increased by increased root abundance and stabilised by soil waters from tropical pastures, with a potential > 80-fold increase in potential water permeability.

Soil chemistry differences, though often slight, were meaningful. The enrichment of the soil profile to 90 cm under tropical pastures with P indicates that leaching has occurred through a profile that does not normally facilitate leaching. Although the means were not significantly different (using a paired t-test), the C/N ratios were always significantly lower under tropical pasture soils than native pasture soils to 90 cm (using a Sign test). This is an indicator of greater biological activity which facilitates soil structure development at these depths in the soil, consistent with the increase in the occurrence of finer peds at 70 – 150 cm depth in the profile.

This study indicated that the mechanism of rapid subsoil structural and physical improvement was firstly a development of deeper and more abundant root architecture in soils sown with tropical pastures, and consequent development of greater macroporosity in upper B horizons. Macropores were then stabilised by interaction with the enriched cation solution draining through tropical pasture topsoils. This is a positive feedback mechanism which is likely to continue to improve soil and facilitate deeper root penetration and higher root density, etc.

In terms of accepted soil formation processes, columnar peds are thought to be the endpoint of a long process of solonization which is meant to take thousands of years (Yaalon, 1971). These B horizons are well known for being hard and impermeable. Dense columnar B horizons are often not considered part of the soil profile in Western Australia where they are referred to as clay pavements equally frequently as B horizons (Verboom and Pate, 2013). Studies on the processes and timeframes for development of subsoil structures are few, often based on chronosequences, and not drawing a timeline for development (White, 1967; Southard and Buol, 1988; Vervoort *et al.*, 1999). The timeline of 14 years at Towri for B horizons to change from large domed dispersible peds, to smaller and aggregate stable peds is notable. This rapid change could potentially be considered accelerated cation enrichment (Bockheim and Gennadiyev, 2009) or soil rejuvenation through plant roots (Johnson *et al.*, 1990), both of which have been accelerated by the action of man choosing a different plant species and adding required fertilisers for these plants to maximise growth.

It is well-documented that tropical pastures have greater production and feed value on sodic-texture contrast soils (Harris *et al.*, 2014). The relatively rapid change of pore size distribution and change of subsoil structure in sodic-texture contrast soils associated with the use of these pastures has landscape scale implications.

Firstly, the obvious implication is that incorporation of tropical pastures into these traditionally poorly productive soils will increase grazing system resilience through potential increased water storage, greater biomass and organic matter production, and consequent increased carrying capacity under grazing.

Secondly, the catchment implications are that large scale incorporation of deeper rooted pastures that store more water will reduce runoff substantially, retaining water in the soil for use by plants. The trade-off is that farms relying on surface water supplies, particularly in lower areas of the catchment, may experience fewer runoff events to recharge surface water stores. However, resultant deep hydration of the landscape may ultimately extend periods of base-flow in smaller streams following years of high rainfall (Boulton, 1999).

Thirdly, climate change is expected to result in more extreme drought and flood regimes in Australia (Herold *et al.*, 2018). Storing more water in the landscape is desirable for local climate stability as well as production, and provides a buffer against flooding because the landscape takes longer to saturate to the point of runoff.

In NSW, there is an abundance of land with sodic-texture contrast soils or similarly poor but deep soil types which have been cleared for agriculture and are now degraded or simply of low productivity. Incorporation of more tropical pastures into these landscapes could result in regionally significant soil improvements in substantial economic production increases.

Study 3: Woodbine - Tenterfield

This study focused on a location that had been previously observed to have a rapid soil Order change at a fence line boundary between Forestland State Forest and Woodbine, a property South-east of

Tenterfield in Northern NSW. The boundary is from tall wet forest to managed pastures, on a plateau with no lithology, climate or slope changes that would otherwise cause a different soil type to naturally occur.

Forest soils were all Red Dermosols with very similar characteristics in the field to a Ferrosol, except for the required free Fe content. All pastures adjacent to the forest had been established and maintained over 140 years, with the addition of lime and phosphate fertiliser from the 1950's, and top dressing with a variety of pasture seed through this period. All pasture soils were Red or Brown texture contrast soils, falling into either the Chromosol or Kurosol soil Order depending on soil pH. The slight variation in subsoil pH which distinguished these two soil Orders may have been due to uneven application of lime across the pasture, demonstrating that soil type could potentially be changed simply through liming.

The difference between the gradational Dermosols in the forest and the texture contrast soils under pasture was assumed to have occurred since clearing. Fine root counts were similar in both forest and pasture, and comparison of total (all diameters) root cross sectional area within the profile showed a dominance of very large roots in the forest compared with pastures.

Although macroporosity was higher at the surface in forest soils than in pasture soils, there was no difference in pore size distribution at 50 – 60 cm. The observation and recording of megapores (rotted out tree root channels with mean diameter of 2 cm) in forest soils, however, allowed for a comparison of potential profile drainage between forest and pasture to be made using pipe-flow theory (Chen and Wagenet, 1992). There was an estimated 139-fold greater potential water flow through forest soils compared than through pasture soils. The megapores in this case function to facilitate bypass flow in wet periods through the profile to underlying parent material.

In wet forest areas, it was proposed by Mohr *et al.* (2013) that megapores provide essential drainage to prevent waterlogging and contribute to localised aquifer recharge, and that the removal of megapores results in profile waterlogging and increased runoff. Evidence of impeded profile drainage in the pasture lands is given by low slope mass movement (slumping) and low trafficability following prolonged rainfall which was not observed in the forest. A slight dominance of spore forming gram-positive bacteria in forest topsoils is a further indicator of frequent wetting and complete drying in the forest.

Textural assessment and particle size analysis in the present study demonstrated the high likelihood of clay eluviation having occurred from upper soil layers in the pasture since clearing. The total amount of clay in the soil profile had remained unchanged, however the clay content of individual soil depth increments had changed, with pasture soils having formed a distinct texture contrast texture profile.

A mechanism was proposed for the change from a Dermosol to texture contrast soil at Woodbine as a result of clearing and compaction, the removal of large tree roots, and the collapse of large continuous megapores derived from tree roots. This was sufficient to initiate periodic waterlogging of soil which is consistent with conditions whereby clay eluviation from surface layers can occur. Consequent with waterlogging conditions, a texture contrast developed and incipient or even fully developed $A2_e$ or E horizons could form.

Formation of a texture contrast soil from unconsolidated sediments is thought to require 15,000 years or more (Yaalon, 1971; Walker and Butler, 1983; Chittleborough *et al.*, 1984). It would appear that in the case of the Woodbine site, tree roots which left a legacy of megapores in the forest were sufficient to maintain drainage of the soil below the threshold for texture contrast soil development which occurs with waterlogging of upper soil layers. The loss of root and large pore architecture with the clearing of land for pasture was sufficient to remove a vegetation-determined threshold, and this resulted in a relatively rapid change in soil order.

The initiation of these feedback processes is purely physical or morphological in nature as proposed by and consistent with known processes of eluviation (Bockheim and Gennadiyev, 2000). Once initiated, the conditions become favourable for further clay eluviation from near surface layers, promoting further waterlogging, which becomes a feedback process on further clay eluviation. The further differentiation of resultant texture contrast soils by subsoil pH into Chromosols or Kurosols may be accelerated by artificial base cation enrichment of some soils through the irregular application of lime which alters topsoil and subsoil pH having potential long term impacts on subsoil dispersion.

The implications of the Woodbine study are that this type of soil change may be more widespread in the New England Tablelands area of NSW and southern Queensland. Through understanding the processes that have driven these changes, it may be possible to improve soil profile water usage as well as pasture production in these pastured areas by selecting pasture species which mimic the native vegetation and generate much greater macroporosity or even megaporosity in subsoils, thereby allowing the profile to become beneficially moist, but not saturated.

Prevention of near surface waterlogging in this way may not reverse the texture contrast that had formed through clay eluviation, but would prevent it developing further, whilst ensuring higher production outcomes. Alternatively, removing the cause of waterlogging could potentially result in reactivation of bioturbation processes as per (Paton, 1978) which could re-mix the near surface profile back to a gradational state.

7.2 Conclusions

Soil development processes that are observed products of timelines in natural environments may be accelerated through anthropogenic activity. It is likely that the assignation of long timelines in natural environments for these processes is partially a result of the long periods of time required for vegetation and climate to change naturally (for instance during the Holocene in northern Europe), thus changing the water regime of a soil. In addition, soil development processes in far northern latitudes of the world, which have generally cool, moist and seasonal climates, are likely to be slowed by the lack of deep, within-decadal moisture fluxes that can occur in eastern Australia. The study and development of soil forming processes was initiated in the cooler landscapes of the world, where deep soil moisture content is relatively constant in comparison with the relatively rapid rate deep changes in soil water status (water flux) that occur in eastern Australia.

The present investigation demonstrates that a change in root architecture associated with replacing native vegetation with a pasture environment can have both positive and negative outcomes in terms of desirable agronomic soil characteristics. In the case of tropical pastures at Towri, the introduction of a deeper and more aggressive root system facilitated large changes in deep soil structure as well as soil stability. Other authors have established that these pastures facilitate deeper storage of moisture (Murphy *et al.*, 2010; Murphy, 2014; Murphy *et al.*, 2018).

In the case of Woodbine, removal of large root systems, and hence the eventual collapse of megapores, changed the moisture flux within the pastured soil profiles such that rapid wetting, waterlogging and drying in the slightly compacted upper layers has developed, resulting in accelerated soil formation by eluviation of clay from upper layers.

Water supply was proposed as being more important than climate, both as an initiating factor of soil formation and a process by (Paton *et al.*, 1995). With regard to this study, water supply to the soil surface had not changed significantly since European land use began. However, what had changed was the availability of water within the soil profile to drive physical, chemical and probably biological feedback processes which resulted in significant changes in soil type or character. If climate alone was considered as a dominant factor in these processes, then no change would have been exhibited in the soils examined.

In summary, the present study indicates that the mechanisms by which rapid soil change can occur are initiated by (human) replacement of pre-existing plant roots with new plant roots of different architecture, and that this physical change alters the potential water balance of the soil and creates a feedback process leading a new stability. Human-induced changes to vegetation root structure and composition, along with the use of supplementary fertilisers, can far exceed the speed at which this occurs naturally, thereby accelerating soil forming processes.

7.3 Implications

Implications of the studies described herein are that it is possible, by managing suites of carefully selected plants, to facilitate not only greater and more sustained production but to substantially improve the quality of subsoils within a decade. Other benefits include greater moisture storage and nutrient access, repair/improvement of surface and B horizon soil structures, decreased erosion potential, substantially reduced runoff and consequentially higher SOC levels. Using water where it falls to produce biomass from a given landscape is far more efficient than using irrigation. This offers an opportunity to realise agricultural production on lands previously considered of low value because they did not possess the potential for deep water storage.

The simple relationship between root architecture change by man and accelerated deep soil development processes is a positive finding. That roots can initiate such large changes in some circumstances offers an opportunity for development of new grazing systems that not only improve soil, but are more productive and more resilient in dry periods. That soil development can be driven in a negative or positive direction from an agronomic viewpoint is also a useful result. It indicates that simple plant selection (for maximum use of available soil profile depth) and management (of cover and fertilisation) may be all that is required to bring about positive landscape-wide production increases and beneficial hydrological change.

The results presented in this study will enhance global knowledge on plant-soil interactions where vegetation has been used to play an important role in regulating eco-engineering conditions for plants (Jones *et al.*, 1994; Ehrenfeld *et al.*, 2005), to modify soil environments (Lambers *et al.*, 2009).

7.4 Future directions

Most soil related research in pastures and cropping is largely agronomic and has limited focus on deep soil profile qualities and impacts of agricultural practices. The focus on surface soils only, and on crop or livestock productivity may well be a limiting factor in the effectiveness of human land use systems in Australia. A philosophical change of mindset should be encouraged for researchers and agriculturalists whereby people in these industries see themselves as managers of soil first. This would require that land managers and researchers focus on managing and using soil within its limitations, or finding plant varieties that have the capacity to overcome soil limitations whilst improving soil simultaneously.

The research presented in this thesis suggests that it is possible to match plants and soils in such a way that production is greatly increased at the same time as improving a soils productive capacity substantially. The bulk of the positive and negative changes measured in this study are entirely related to how water interacts with soil layers, how deep water penetrates and how much of that water the plant systems can access.

Much research into poor soils in the world is based on identifying topsoil and subsoil limitations. Perhaps a re-focus on these soil limitations as opportunities for finding useful plants which grow and thrive in them will lead agriculture in new and more resilient direction to match our need for both more sustainable yet more productive landscapes in Australia and around the world.

The research presented in this thesis could be expanded with proper consideration of soil field, physical, chemical and biological characteristics integrated into more replicated chronosequences or into the first stage of properly replicated long term trials. Additional work should also include regular monitoring or recording of volumetric soil moisture, E horizon water tables, and soil water flux in relation to plant use of soil water. Results from further work would be useful in understanding and developing 3 dimensional interactions of land use, soil and hydrological characteristics in the challenging Australian environment.

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