



CLAY MOVEMENT IN A SALINE-SODIC SOIL TOPOSEQUENCE

Muhammad Nathan

Ir. (Hons) Department of Soil Science
Hasanuddin University, Makassar, Indonesia

Thesis submitted in the Department of Soil and Water
Faculty of Agricultural and Natural Resource Sciences
The University of Adelaide, South Australia
for the degree of Master of Agricultural Science

July, 2001

Table of Contents

List of figures, tables, appendices & plates	iii
Summary	viii
Declaration of originality	xi
Acknowledgements	xii
Chapter 1. Introduction	
1.1 Background	1
1.2 Objectives and hypotheses	2
Chapter 2. Review of Literature	
2.1 Introduction	5
2.2 Pedogenesis of clays	5
2.3 Factors affecting the dispersibility and mobility of clays	7
2.3.1 Soil sodicity and clay dispersion	7
2.3.2 Soil salinity and clay dispersion	10
2.3.3 pH and clay dispersion	11
2.3.4 Organic matter	13
2.3.5 Mineralogical composition and clay dispersion	15
2.3.6 Particle size & porosity in relation to clay dispersion	16
2.3.7 Hydraulic conductivity and dispersion	18
2.4 Pedogenesis of texture-contrast soils	20
2.5 Environmental implications of sodic, saline or sodic-saline soils	22
Chapter 3. Soil properties, their distribution along the toposequences and soil classification	
3.1 Introduction	25
3.2 Field and laboratory measurements	25
3.3. Results	32
3.3.1 Soil morphological properties and classification	32
3.3.2 Soil physical properties	34
3.3.3 Soil chemical and mineralogical properties	37
3.3.4 Distribution of soil properties in toposequences in eroded and non-eroded areas	40
3.4 Discussion	45
3.4.1 Sodicity-salinity criteria, soil properties and clay dispersibility	45
3.4.2 Relationship between classified soil properties & topography	47
3.4.3 Colour patterns from an aerial photograph in relation to soil properties	52
3.5 Conclusions	52
Chapter 4. Lysimeter, overland flows and stream sediment investigation	
4.1 Introduction	54
4.2 Materials and methods	54
4.2.1 Lysimeter design and installation	54
4.2.2 Overland flow sediments and stream water sampling	56
4.2.3 KBr tracer to investigate lateral flow	57
4.3 Results	59
4.3.1 Colloid concentrations, chemistry and mineralogy of lysimeter solutions	59
4.3.2 Lateral flow of colloids	62
4.3.3 Overland flow and stream water sediments	62

4.4 Discussion	64
4.4.1 Dispersibility and migration of the clays	64
4.4.2 Erosion	69
4.5 Conclusions	74
Chapter 5. General conclusion and land degradation model	
5.1 Introduction	75
5.2 Clay movement and model of land degradation in the Herrmanns sub-catchment	76
References	78

List of figures

Figure 1.1.	Clay dispersion and ESP of Bt horizons along a toposequence from the Herrmanns subcatchment, Mt Lofty Ranges, South Australia (from Fitzpatrick <i>et al.</i> , 1994b).	4
Figure 2.1	Proposed scheme for description of Na-affected soils in terms of physical behavior (dispersibility) and sodicity and salinity (from Sumner <i>et al.</i> , 1998)	8
Figure 2.2	Salinization of soil in an arid region (from Fanning and Fanning, 1989)	11
Figure 2.3.	Various terminologies and criteria of porosity proposed by several authors since 1951 (from Luxmoore, 1981)	19
Figure 2.4.	Locality of the study site (Herrmanns sub-catchment).	24
Figure 3.1.	Schematic locality diagram showing position of profiles P1, P2, P3, PS and two toposequences in relation to landscape position.	26
Figure 3.2.	Profiles showing particle size distributions, mechanically dispersible clay contents and bulk densities for (a) Profile P1, (b) P2, (c) P3 and (d) PS.	35
Figure 3.3	Profiles showing EC, pH, ESP and SAR for (a) Profile P1, (b) P2, (c) P3 and (d) PS	38
Figure 3.4	Profiles showing cation exchange capacities (CEC), exchangeable cations and organic carbon contents for (a) Profile P1, (b) P2, (c) P3 and (d) PS	41
Figure 3.5	XRD patterns for each horizon in (a) Profile P1, (b) P2, (c) P3 and (d) PS.	43
Figure 3.6a.	Soil colour, soil classification, and gravel content of profile P1 along toposequence A-B.	46
Figure 3.6b.	Soil colour and soil classification of profile P2 along toposequence C-D	46
Figure 3.7.	Distribution of EC _{1:5} in relation to topography.	48
Figure 3.8	Salinity-sodicity classification of soils in the study site according to criteria of Sumner <i>et al.</i> (1998) (modified).	48
Figure 3.9.	Correlations of ESP with (a) clay dispersion and (b) pH _e for profiles P1, P2, P3 and PS.	49
Figure 3.10.	False colour aerial photograph (1995), illustrating postulated link between soil texture (inferring soil moisture) and the distribution of vegetation observed in year 2000.	53
Figure 4.1	Design of wick lysimeter used in study site and sketch of position of lysimeter in profile (from Biddle <i>et al.</i> , 1995)	55
Figure 4.2	Installation of sediment collectors in the field	57
Figure 4.3.	Placement of KBr at upslope locations from profiles P1, P2 and P3 lysimeters	58
Figure 4.4.	Powder XRD patterns of colloids collected from the lysimeters in profiles (a) P1, (b) P2 and (c) P3	60
Figure 4.5.	XRD patterns of colloids collected from stream water samples, and overland flows.	63
Figure 4.6	Relationships between clay dispersibility of profiles P1, P2, P3 and clay collected in the lysimeters.	64
Figure 4.7	Relationships between clay dispersibility and hydraulic conductivity of profiles P1, P2 and P3.	65

Figure 4.8.	Clay dispersibility of the soils in profiles P1, P2, P3 and the concentration of colloids of lysimeter solutions	66
Figure 4.9	Relationships between hydraulic conductivity and colloid collected from the lysimeters in profiles P1, P2 and P3.	67
Figure 4.10	Three water flow systems in the Herrmanns subcatchment, Mt Lofty Ranges (from Fitzpatrick, 1994a)	69
Figure 4.11	Aerial photographs of the study site taken at two different times (1945 and 1999).	70
Figure 4.12.	Interpretation of aerial photographs showing the intensity and distribution of erosion in 1945 and 1999.	70
Figure 5.1	Three dimensional view showing the position of the eroded (P1) and non-eroded (P2) areas in the study site, in relation to the landscape of the sub-catchment (modified from Fitzpatrick <i>et al.</i> , 1999).	77
Figure 5.2	A conceptual model illustrating the degraded-nondegraded phenomenon at the Herrmanns sub-catchment.	78

List of tables

Table 3.1	Classification of soils using Soil Taxonomy (Soil Survey Staff, 1999)	33
Table 3.2	Saturated hydraulic conductivity measured at different depths and positions along the toposequence using disc- and bore-hole permeameters.	37
Table 4.1.	Depth of lysimeter installations in the profiles	56
Table 4.2.	The pH and EC of solutions collected from lysimeters in profiles P1, P2 and P3.	59
Table 4.3.	Concentration of suspended colloids from lysimeters in profiles P1, P2 and P3	60
Table 4.4.	Br ⁻ concentration in lysimeter solution before & after injection of KBr into soil	62
Table 4.5.	Chemistry of stream water and overland flows in eroded and non-eroded areas.	62
Table 4.6.	Concentration of colloids collected from stream water and overland flows in eroded and non-eroded areas	64

List of appendices and plates

(all > p.86)

Appendix 1.	Morphological descriptions of a representative saline soil profile (P1) in the Herrmanns sub-catchment
Appendix 2.	Morphological descriptions of a representative sodic soil profile (P2) in the Herrmanns sub-catchment
Appendix 3.	Morphological descriptions of a representative nonsodic-nonsaline soil profile (P3) in the Herrmanns sub-catchment
Appendix 4.	Morphological description of a soil profile on the edge of a stream bank in the Herrmanns sub-catchment.
Plate 1.	Photograph showing degraded and non-degraded areas in Herrmanns sub-catchment. The white patches are salt crusts in the soil surface of eroded area.
Plate 2.	Photograph showing the most severely eroded area in Herrmanns

sub-catchment. The Ap horizon has been completely removed by erosion. The salt crusts (white coloured zones) are evident.

Plate 3. Photograph showing eroded area in the Herrmanns sub-catchment with some gully erosion along the stream bank.

Plate 4. Creek and gully erosion, which did not exist 50 years ago, is now approximately 200 cm depth in some parts.

Plate 5. Soil profile exposed by gully erosion in the deepest part of the stream bank. White patches on the sides of the streambank are salt crusts.

Plate 6. (a) CSIRO disc permeameter and (b) Guelph well permeameter used to measure K_s .

Summary

Sodic soils, cover about 28% (340 million ha) of Australian continent. Many studies have shown that the dispersive propensity of sodic soils causes detrimental changes in soil structure, water logging and soil erosion, leading to severe land degradation. Clay movement from sodic soils has also been recognised as a major cause of turbidity of streams and cause a deterioration of the quality of water.

The Herrmanns sub-catchment in the central Mt. Lofty Ranges (near Mt. Torrens) is one of the regions in South Australia affected by saline-sodic soils. A soil survey of the sub-catchment showed a variation in the levels of salinity and sodicity in relation to topography and that dispersibility of clay also varied along the toposequence. Little is known about the mobility of clays from these soils in relation to the seasonal salinity-sodicity conditions. Consequently, the main objectives of this study were to quantify the interrelationships between clay movement and soil properties (especially salinity and sodicity) down a toposequence.

Three profiles with different levels of salinity and sodicity were excavated and described according to the Australian Soil and Land Survey Handbook (McDonald *et al*, 1990) and classified according to Soil Taxonomy (Soil Survey Staff, 1999). Samples from the profiles were analysed in terms of chemical, physical and mineralogical properties. Two lysimeters were installed in each profile to collect colloids moving vertically. These were embedded in the A/Bt horizon boundary and at the base of the Bt horizon. KBr was applied to assess the possibility of clay being transported laterally by flow of water over the Bt horizon. Overland flow sediment collectors were installed in a typical eroded area of the sub-catchment (a highly sodic area) and in a non-eroded area, which was less sodic. Stream water samples were collected during the periods of rainfall at three locations, chosen randomly. Two toposequences were sampled to characterize the distribution of soil properties (i.e. soil colour, texture, soil EC, gravel contents). The hydrology of the saline ground watertable was inferred from the results of the distribution and value of intensive EC measurements taken from the core samples (up to 1.5 m depth) collected using a systematic grid (50 m x 50 m).

All the soils were classified as Alfisols with Subgroups being dependent on topographic position as follows: Typic Natraqualfs on the eroded areas of the foot slopes, Typic

Natrixeralfs on the waning mid-slopes and stream banks, and Ultic Palexeralfs on the upper waning mid-slopes. Each subgroup had vastly different sodicities. The most sodic was the Typic Natraqualfs (ESP ranged from 9 to 24.3 %), followed by the Typic Natrixeralfs (ESP ranged from 0.4 to 16.5 %) and the Ultic Palexeralfs (ESP ranged from 0.3 to 2.6).

There was a strong positive relationship between clay dispersibility and sodicity in the eroded areas along the foot slopes. For the other profiles (from the mid slope and stream bank), the increase in sodicity was accompanied with a drop in clay dispersibility. In the midslope profiles, the flocculative-dispersive phenomenon was largely controlled by other soil properties (i.e. pH, mineralogical composition, organic carbon, base cation content, etc.) rather than salinity-sodicity effects. In the stream bank profile, the high salinity level caused the clay to flocculate.

Clay dispersibility, hydraulic conductivity and the amount of clay collected in the lysimeters (vertical clay movement) were correlated. Increase in clay dispersibility predisposed clay to move and clog micropores, a process that lead to a decrease in saturated hydraulic conductivity and a decrease in the amount of colloid transported to lower horizons. High hydraulic conductivity correlated with higher amounts of colloids collected in the lysimeters. The Typic Natraqualf in the eroded area along the foot slope was the exception. The profile had the most readily dispersible clay but the hydraulic conductivity remained high. This was thought to be caused the high gravel content (about 35 %), which conferred an appreciable macroporosity in this profile. In this profile, the dispersibility of clay led to an increase in the vertical mobility of clay.

In the Herrmanns sub-catchment, an appreciable amount of clay moved laterally via throughflow of water over Bt horizons (concluded from KBr tracer experiment), and via overland flow. The amount of clay collected via overland flow was 0.425 g/L and 0.034 g/L in eroded (highly sodic) and non-eroded (less sodic) areas respectively. The relatively high concentrations of clay in the overland flow water in the eroded area compared to the non-eroded area was attributed to high potential clay dispersibility, lack of vegetative cover, and the steeper and longer slopes in the eroded area compared to non-eroded area.

The concentration of sediment in the stream water was 0.15 g/L and was less than the concentration of sediment collected from overland flow (0.34 g/L) in eroded area, but

higher than the concentration of overland flow sediment (0.03 g/L) collected from non-eroded area. From this, it was concluded that the area with soils having highly dispersive clays with no vegetation cover was a major contributor of sediment into the stream draining the Herrmanns sub-catchment.

The high electrical conductivity values in soils at depths greater than 50 cm across the study site indicated that a saline water table strongly influenced the nature of the soils along the entire foot slope (i.e. from the contour line at 388 m ASL in the stream up to the contour line 392 m ASL). However, the interplay between incoming water from rainfall and the saline groundwater table in the area along the foot slope provided an explanation for the origin of the various soils down the toposequence (i.e. highly sodic, non or slightly sodic soils).

In summary, in the Herrmanns sub-catchment, soil sodicity was the dominant factor causing clay to disperse in the eroded area along the foot slopes, whereas in non eroded areas of the mid-slopes and on the stream banks, the dispersive power of sodicity was attenuated by the flocculative power of other soil properties. The dispersed clay was more easily transported by flow of water laterally over the soil surface than non-dispersed clay. For vertical movement of clay within profiles, however, the dispersed clay appeared to be less mobile than non-dispersed clay, especially where micropores dominated the soil matrix.

STATEMENT

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

I consent to this thesis being made available for photocopying and loan.

Muhammad Nathan
July, 2001

Acknowledgments

Thanks God, I was finally able to complete this thesis. Many people have given great support in various ways during my studies at the University of Adelaide. Firstly, I wish to thank my supervisors: Ass. Prof. D.J. Chittleborough and Dr C.D. Grant (Department of Soil and Water) and Dr. R.W. Fitzpatrick (CSIRO Land and Water) for assisting me to develop the ideas of the project, for their support, supervision, guidance, and interest throughout the project. Ass. Prof. Chittleborough, in particular, introduced the use of lysimeters to my project and helped me in the interpretation of XRD data. I had a great experience in the field with Dr. Fitzpatrick and his encouragement is invaluable.

I wish also thank all staff members of the Department of Soil and Water for their assistance and help. In particular I wish to thank Mr. C Rivers for his great assistance with my laboratory and field experiments, my fellow student Jon Varcoe for helping me to establish and install the lysimeters in the field, Mr. John Davey for computational assistance, and Ms Alla Baklan for XRD analysis.

Thanks also to Mr. Phil Davies (CSIRO Land and Water) for providing me some aerial photographs of my study site, Dr. Mark Jaksa (Department of Civil Engineering, the University of Adelaide) for assisting me to use a theodolite, my friends (Abdul Munir, Ahmad Saidy, Agus Budiharso, Didik Suharyadi) for their great help during my field work. Most importantly Australian Agency for International Development (AusAID) is greatly acknowledged for the scholarships.

To my wife Ros and my children Ananda and Aulia, thank you for your understanding, love and support during the difficult times of my study. Finally, to my *ibu dan ayah*, thanks for carrying love and prayers. I pay respect and tribute to them.

Chapter I. Introduction

1.1 Background

Land degradation associated with soil sodicity is one of the major environmental issues in Australia. Sodification of soils is always associated with undesirable changes in soil structure (Gupta and Abrol, 1990), waterlogging (Rengasamy and Olsson, 1991), erosion of hillslopes and watercourses and often intensifies the transport of colloids and nutrients to streams (Fitzpatrick *et al.*, 1994b). Transport of mobile colloidal particles (e.g. clays) can cause severe turbidity in water bodies (e.g. streams, rivers and dams), can also carry strongly adsorbed pollutants, such as radionuclides, trace metals, and non polar organic compounds (e.g. Kretzschmar *et al.*, 1994), all of which lead to poor water quality.

In sodic soils, sodium occupies at least 6 % of the clay exchange complex (Northcote and Skene, 1972) and this enhances dispersion, particularly in electrolyte-free water. Dispersed clay migrates through soil layers and may block the pores and channels in the soil matrix leading to restricted water- and air-transport (Chorom *et al.*, 1994). Sodium on the clay exchange complex of Australian sodic soils originates primarily from the leaching of NaCl (the main salt) from the soil profile. Thus sodic soils often occur in association with saline soils in Australia (Rengasamy and Olsson, 1991). It is estimated that sodic soils ($ESP \geq 6\%$) cover 28% of the land area of Australia and extend from the intensively cultivated southern and eastern regions to the rangeland and grazing areas of northern Queensland and Western Australia (Chartres, 1995).

Factors that influence the dispersibility of clays (e.g. sodicity, salinity, pH, and organic carbon) and which eventually influence their mobility have been well documented (Goldberg *et al.*, 1988; Rengasamy and Olsson, 1991; Shainberg and Letey, 1984).

During wetting, sodic soils tend to form surface seals and crusts (Rengasamy and Olsson, 1991), and in dry periods they become hard and difficult to re-wet (Purves and Blyth, 1969). In addition, detailed studies have been conducted in the laboratory to explain the mechanisms of clay transportation in sodic soils (e.g. Rowell *et al.*, 1969; Hallsworth, 1963, Noack *et al.*, 2000).

Several studies have focused on saline-sodic soils in Australia. For example, Fritsch and Fitzpatrick (1994) developed a method to construct soil-water-landscape models by interpreting soil features produced by ancient and modern processes in the degraded landscape of the Herrmanns sub-catchment in the central Mt Lofty Ranges (near Mt. Torrens, South Australia). For this same sub-catchment, Fitzpatrick *et al.* (1994a) developed a diagnostic key to assess the management of the saline and waterlogged areas, plus the development of saline sulfidic soils in the catchment (i.e. non-tidal seepage areas) (Fitzpatrick *et al.*, 1996), and seasonally variable hydrological conditions (Cox *et al.*, 1996). Fitzpatrick *et al.*, (1994b) suggested that significant clay dispersion in some parts of the catchment, presumably from sodic soils might produce mobile colloids that are easily transported by water flow (Figure 1.1). This idea has not yet been confirmed through detailed landscape analysis.

1.2 Objectives and hypotheses

The objective of this study was to quantify the relationships between soil properties in the degraded toposequence in the Herrmanns sub-catchment described by Fitzpatrick *et al.* (1996) and the movement of colloids in that landscape.

Of particular interest was the need to explain where in the landscape the high concentration of colloids and particulate matter in the sub-catchment-stream was coming

from. In view of the presence of a local saline ground water table (rising under hydrostatic pressure), three possible sources were considered likely: (1) stream bank erosion, (2) surface overland flow in the sub-catchment, and (3) lateral flow through the landscape via pathways caused either by discontinuities in the soil profile (e.g. horizon boundary) and variations in soil properties (e.g. clay dispersibility).

Of secondary interest was the need to evaluate the extent to which recent improvements in land management (e.g. re-vegetation of sub-catchment, improved drainage) have halted and/or reversed the degradation of the landscape.

It was hypothesized that the primary sources of stream colloids were in the landscape, but this was not precisely known. Erosion of stream banks and overland flow were hypothesized to be contributors affected by the degree of colloid dispersion in the landscape. It was further hypothesized that the mobile (dispersed) colloids in the landscape were generated by sodification of clays exposed to fluctuating saline ground water. The extent of this effect was unknown but considered significant in preliminary surveys. Finally, it was hypothesized that colloid movement through the soil profile and landscape was a function of colloid dispersion and its consequent effects on the hydraulic properties of the soil, particularly in the vicinity of texture-contrast boundaries (e.g. A horizon and Bt horizon).

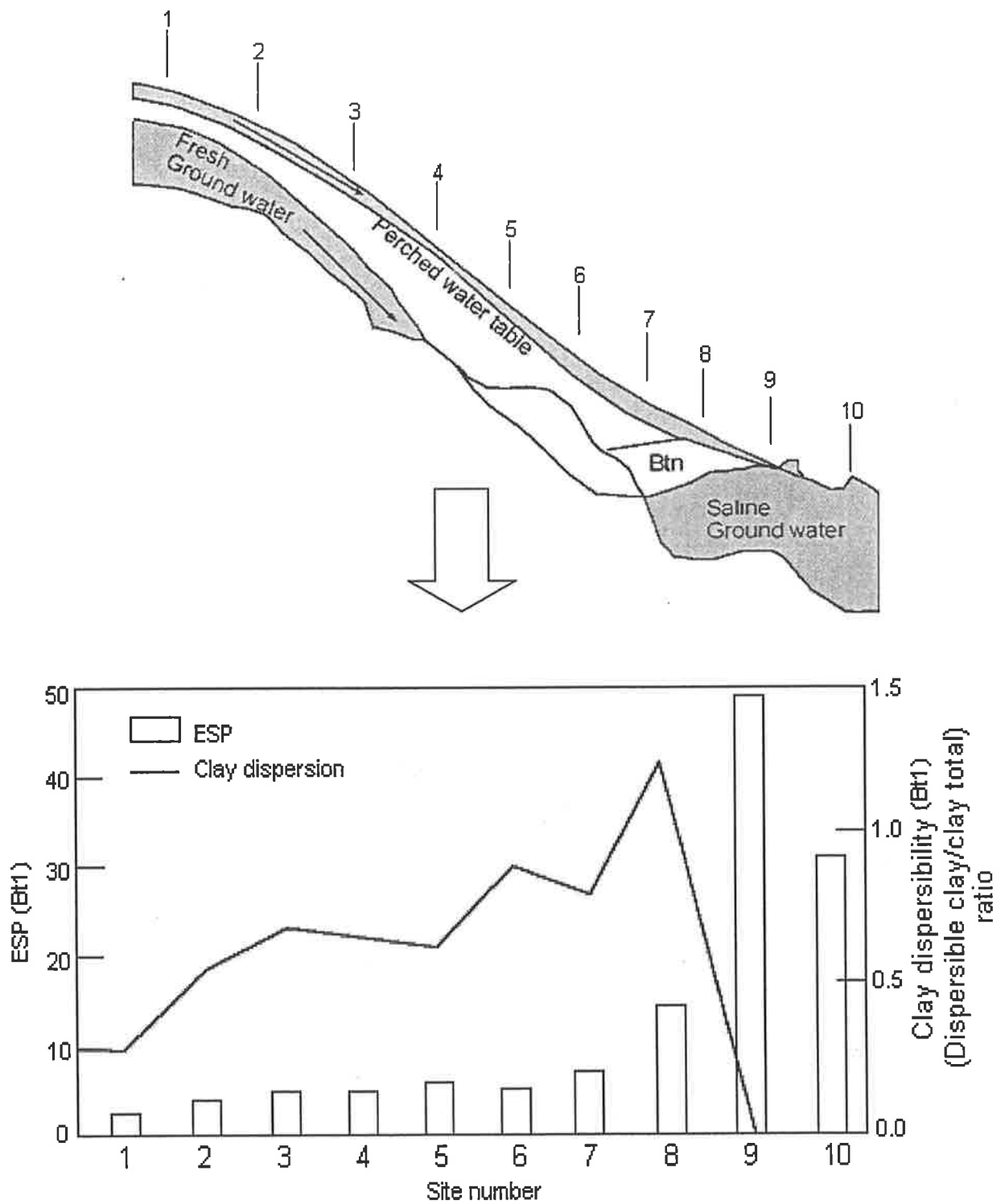


Figure 1.1. Clay dispersion and ESP of Bt horizons along a toposequence from the Herrmanns subcatchment, Mt Lofty Ranges, South Australia (from Fitzpatrick *et al.*, 1994b).

Chapter 2. Review of Literature

2.1 Introduction

This chapter examines the pedogenic movement of clay in a soil profile in relation to the genesis of texture-contrast soils in modern day landscapes. It also reviews a number of factors affecting colloid dispersion and mobility, including sodicity, salinity, physical, chemical, mineralogical and hydrological soil properties. Finally, the literature dealing specifically with the Herrmanns sub-catchment and its problems is examined with a view to placing the present study in context.

2.2 Pedogenesis of clays

Water flows vertically down to the lower horizons in a profile or flows laterally over the soil surface and over subsurface (Bt horizon) and takes with it any mobilised and suspended material (i.e. clay, organic carbon). The movement of large amounts of material from the soil surface and its deposition in subsurface layers is known as clay migration or clay transportation (Thorp *et al.*, 1959; Brydon, 1965). Duchafour (1982) referred to this as lessivage, wherein clay particles $<2 \mu\text{m}$ and other primary minerals $<1 \mu\text{m}$ in diameter are transported downward unchanged. In the present study, the term clay will be used in reference to inorganic colloidal material ($<2 \mu\text{m}$ in diameter), and the terms clay movement, clay translocation or clay migration will be used in reference to the transportation of clay vertically down to the lower horizons or laterally through the profile, either concentrated in biopores, along horizon boundaries, or throughout the profile.

There are two primary interests related to clay movement. The first is of pedogenic interest and leads to the differentiation of textural horizons of soil (Dixit *et al.*, 1975). The

second is of environmental interest and is concerned with water-land degradation processes (i.e. soil erosion, waterlogging, and reduction of stream water quality).

Vertical movement of clay in a pedogenic sense is a slow phenomenon, leading to the creation of argillans or illuvial cutans (Brewer, 1960) on ped surfaces, which is evidence of clay illuviation (Cabrera-Martinez *et al.*, 1989), and is often associated with soils in old landscapes. In general, several factors are responsible for clay movement or clay translocation. Eswaran and Sys (1979) reported that to be able to be transported via water flow, the clay in the overlaying horizon must be dispersible and there must be wet periods interspersed with significant drying (continuous wet periods without drying or vice versa rarely produce argillic horizons (Chittleborough, 1992). Bartelli and Odell (1960) suggested that root channels and pores also play a significant role in clay illuviation, and that where such channels were lacking, minimal colloid movement occurred. Argillic horizons generally developed from clays which carry negative surface charges (Tessen, 1984), and where there is minimal pedoturbation processes (illuviated clay is undisturbed) (Nettleton *et al.*, 1987).

Not all dispersible clay can move vertically to the lower horizons via water flow. Evidence suggests only very fine particles are transported downward, and this requires time (months to hundreds of years) for considerable movement of clay (Pilgrim and Huff, 1983). Because fine clay migrates more easily to lower horizons than coarse clay (McKeague and St.Arnaud, 1969), illuvial horizons often have more fine clay to total clay ratios than overlying and deeper horizons (Cabrera-Martinez *et al.*, 1989). For lateral movement of colloids, however, Lettenmaier *et al.* (1991) found that due to the mechanical power of water flowing over soil surfaces (particularly over long steep topographies), clay that is removed is generally a mixture of fine and coarse particles, and can thus be shown to

be responsible for much of the suspended load in water bodies. Therefore, water contaminated by colloids coming from erosion sources often contains a mixture of fine and coarse particles, whereas water contaminated with colloids from non-erosion sources often only contains very fine clay.

2.3 Factors affecting the dispersibility and mobility of clays

Numerous studies show that dispersion of soil colloids is influenced by factors such as mineralogy, sodicity, pH, CEC, clay-organic interactions, and electrolyte levels (Collis-George and Smiles, 1963; Emerson and Chi, 1977; Frenkel *et al.*, 1978; Oster and Schroer, 1979; Shanmuganathan and Oades, 1983; Gupta *et al.*, 1984).

2.3.1 Soil sodicity and clay dispersion

McNeal & Coleman (1966) found that permeability in soils with 2:1 layer silicates, decreased with increasing sodium adsorption ratio (SAR) and decreasing electrical conductivity (EC). Purves & Blyth (1969) found that exchangeable sodium percentages (ESP) > 15 were sufficient to disperse clay and reduce permeability of soils.

Many methods can be used to assess the dispersive level of clays and soils (Emerson, 1954). Van Olphen (1977) used the critical coagulation concentration (CCC) or the flocculation value. This is the minimum concentration of electrolyte needed to flocculate a given suspension in a given time. To remain dispersed, the EC must be less than the CCC value of the clay. A popular method is to agitate a clay suspension and measure the amount of clay suspended after a certain settling time following Stoke's law.

Because all soils behave differently, there is no exact definition of sodicity (Sumner, 1995). Therefore some different definitions are used and accepted by several authors. For example, in the USA, Brady and Weil (1999) and Kelly (1951), characterized

sodic soils as those with $EC < 4.0 \text{ dSm}^{-1}$ and $ESP > 15$ or $SAR > 13$ with pH between 8.5 and 10 or even higher for several types of soils. In Australia, by contrast, Rengasamy and Olsson (1991), classified sodic soils into alkaline sodic (pH > 8.0), neutral sodic (pH 6.0-8.0) and acidic sodic (pH < 6.0), similar to criteria adopted by the USSL (1954).

Northcote and Skene (1972) adopted the value of $ESP \geq 6$ for sodic soils. Other criteria proposed by Sumner *et al.*, (1998) used a combination of $SAR_{1:5}$ and $EC_{1:5}$ to characterize salt affected soils (Figure 2.1).

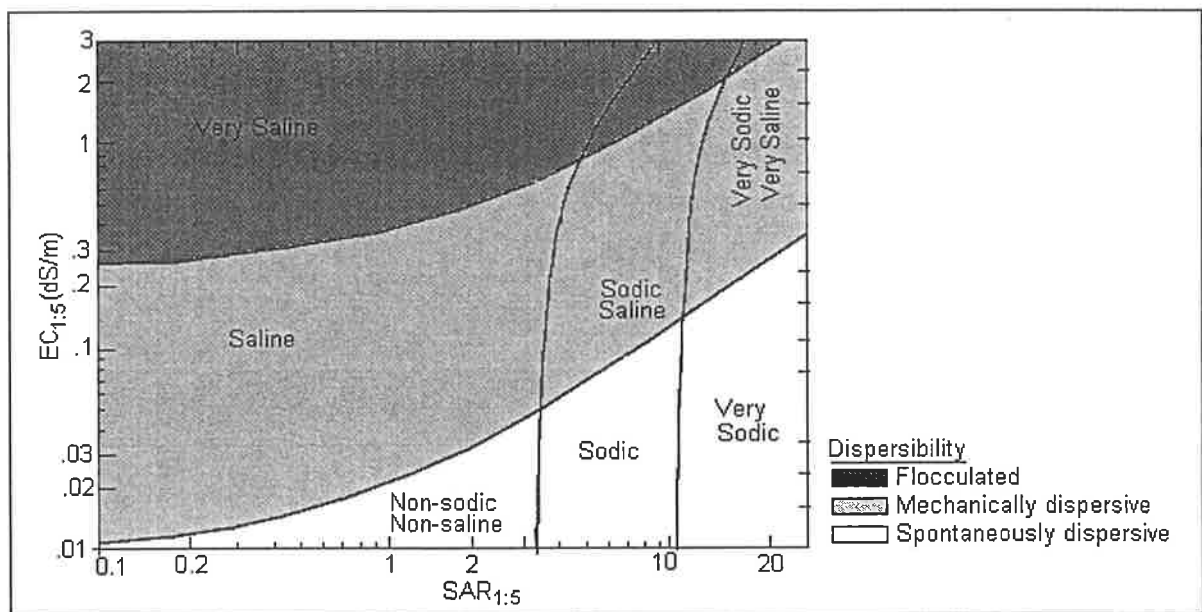


Figure 2.1. Proposed scheme for description of Na-affected soils in terms of physical behavior (dispersibility) and sodicity and salinity (from Sumner *et al.*, 1998)

The recent criteria proposed by Sumner *et al.*, (1998) will be used in the present study of the soils in the Herrmanns sub-catchment. These criteria integrate SAR and EC with the dispersibility of the clay. The ESP and SAR are defined as follows :

$$ESP = \frac{\text{Exchangeable } Na^+, \text{ cmol/kg}}{CEC, \text{ cmol/kg}} \times 100$$

$$SAR = \frac{Na^+}{\sqrt{(Ca^{2+} + Mg^{2+})}}$$

where Na^+ , Ca^{2+} and Mg^{2+} are in mmol/L.

The higher the ESP, the higher the SAR, and the USSL (1954) developed an equation to predict the value of ESP from the SAR paste-extracts:

$$ESP = \frac{[-1.26 + (1.475 \times SAR_e)]}{[0.9874 + (0.0147 \times SAR_e)]}$$

If the data are obtained from 1:5 soil:water solutions, the relationship is:

$$ESP = 1.95 \times SAR + 1.8 \quad (\text{Rengasamy } et \text{ al., 1984}).$$

The processes leading to sodicity are called solonization (Fanning and Fanning, 1988). When fresh water (i.e. during rainy season) flushes away the soluble salts, the exchangeable-Na dominates the physical behaviour of clay (Fritsch and Fitzpatrick, 1994).

Sodium in the soil may originate from a number of sources (Duchaufour, 1982), including (i) saline water tables containing Na^+ , Ca^{2+} or Mg^{2+} , (ii) saline water tables with Na^+ as dominant cation, or (iii) weathering of Na-bearing minerals e.g. feldspar or feldspathoid, and particularly albite (Fanning and Fanning, 1988; Gunn, 1967), and (iv) atmospheric salt accession (Chartres, 1995).

Rising saline water with Na^+ and Ca^{2+} or Mg^{2+} may lead to the formation of non sodic saline soils, because Ca^{2+} and Mg^{2+} tend to compete for spots on the clay exchange complex, so Na^+ rarely exceeds 15% (Duchaufour, 1982). Rising saline water with primarily Na^+ may initiate solonization because sodium is the dominant cation and ESPs exceeding 30 often result. Naturally-occurring Na^+ -bearing minerals may also lead directly to solonization without necessarily any salinisation.

Hydrolysis of Na in soils may raise the pH into the alkaline range (Fanning and Fanning, 1988). However, in high rainfall areas (annual precipitation 550-750), basic cations (Ca^{2+} and Mg^{2+}) have usually been leached from the exchange complex, so any added Na allows sodicity to develop at $\text{pH} < 6$ (Rengasamy and Olsson, 1991). In field conditions, a high pH is induced by the presence of Na-bicarbonate/carbonate minerals (Nakayama, 1970). At pHs > 8.5 both Ca^{2+} and Mg^{2+} precipitate as insoluble carbonates, which severely increases the SAR of solutions (Gupta and Abrol, 1990).

Crops grow poorly in alkaline sodic soils primarily because of a combination of poor nutrient availability (Gupta *et al.*, 1984) and poor soil structure. Sodic soils become too wet in rainy seasons and too hard and dry in the summer, leading to poor root development and crop growth (Shainberg and Letey, 1984).

2.3.2 Soil salinity and clay dispersion

The primary cations and anions associated with saline soils are (i.e. Na^+ , K^+ , Ca^{2+} , Mg^{2+} , Cl^- , SO_4^{2-} , CO_3^{2-} , and HCO_3^-) (e.g. Cass *et al.*, 1996), and the excess salt is easy to recognize from a white crust on the soil surface (Gupta and Abrol, 1990).

In the salinization process, soluble salts accumulate in surface horizons and lead to salic horizons (Fanning and Fanning, 1988). Rainwater may contain 10-30 mg/L soluble salt and sometimes >50 mg/L, but this decreases with distance from the coast (Ghassemi *et al.*, 1995). Figure 2.2 from Fanning and Fanning (1988), illustrates the salinization process and the type of soils that develop under saline conditions.

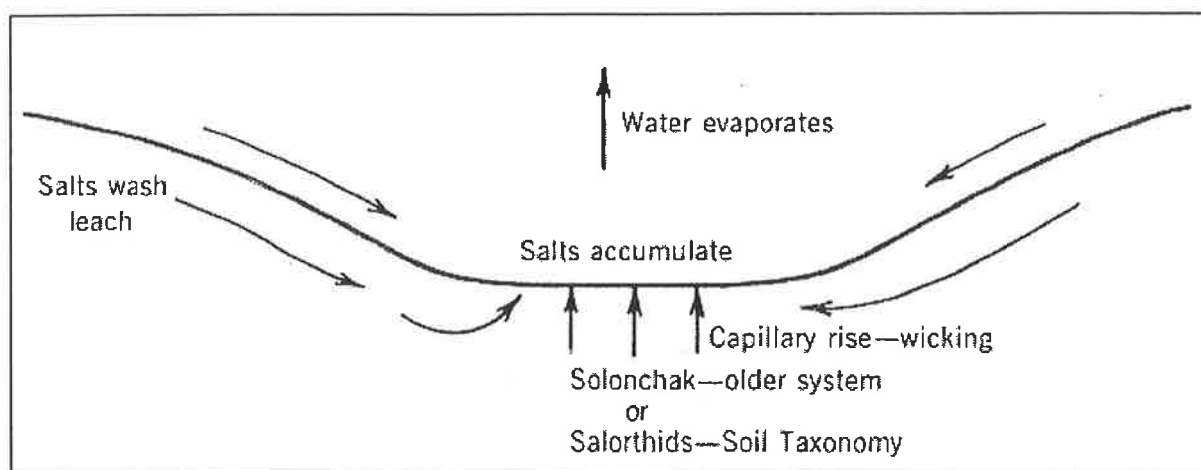


Figure 2.2. Salinization of soil in an arid region (from Fanning and Fanning, 1989)

Upward capillary movement of saline water, usually during summer, can bring a significant amount of salts to surface soil horizons. Capillary rise from a saline groundwater table within 1 m below the soil surface and containing 10,000 mg/L salt, for example, would result in accumulation of 10kg salt/m³ soil (McTainsh & Boughton, 1993).

An important distinction between saline and non-saline sodic soils is that saline soils generally have a good physical condition (clay particles are flocculated). However, saline soils are usually more erodible than non-saline and non-sodic soils, consequently, their flocculated condition is really only a minor improvement. Furthermore, the extent to which this apparent stability affects colloid mobility in the landscape is generally unknown at present.

2.3.3 pH and clay dispersion

The role of pH in dispersing clay has been evaluated in a number of situations. Suarez *et al.*, (1984) reported that dispersion of clay in kaolinitic and montmorillonitic soils in an arid zone increased with increasing pH at a constant SAR and electrolyte level. Gupta *et*

al., (1984), also found that clay dispersion of Na-saturated Indian soils shaken in 0.1 M NaCl for one day increased as pH increased from 6.5 to 10.5. Arora and Coleman (1979) confirmed this trend, but found that maximum dispersion of their clays occurred at pH 8.3. The effect of pH on dispersibility of clay is enhanced by variable-charge minerals such as iron oxides (gibbsite), kaolinite, or organic matter in soils (Seta & Karathanasis, 1996).

A standard way to characterise the dispersive nature of a colloidal material at a given pH is to report the critical coagulation concentration (CCC), which is the minimum concentration of an electrolyte required to flocculate a given dispersible colloid in a given time at a given pH (van Olphen, 1977). The CCC values for Ca-montmorillonite range from 0.17 meq/L (van Olphen, 1977) to 0.25 meq/L CaCl₂ (Greene *et al.*, 1978; Oster *et al.*, 1980), while those for Na-montmorillonite are considerably greater (ranging from 7 to 20 meq/L NaCl), for Na-kaolinite CCC values range from 0 to 5 meq/L NaCl (Arora and Coleman, 1979). Lower CCCs indicate lower spontaneous dispersibility of clay. Soil pH can thus significantly influence whether clays at a given electrolyte concentration are stable or dispersible. The magnitude of pH-dependent charge on clay also has a large impact on the CCC at a given pH. For example, Swartzen-Allen and Matijevic (1976) demonstrated that the CCCs for Na-montmorillonite in NaNO₃ solution increased with increasing pH from 1 meq/L at pH 3.8 to 10 meq/L at pH 10. For Na-kaolinite in NaNO₃ the CCCs increased from 2 meq/L at pH 4.1 to 40 meq/L at pH 10.1. The greater proportion of pH-dependent charge associated with kaolinite, makes their CCCs more pH-dependent than for montmorillonites. Chorom *et al.* (1994) demonstrated that higher net negative charge of soil clays (due to association with organic matter and other constituents) than pure clays of similar mineralogy, was responsible for higher CCCs for soil clays. Rengasamy (1983) and Goldberg and Forster (1990) also reported similar findings.

Sumner (1993) reported that pH dictates clay dispersibility by changing the charge on soil particles, and depending on clay mineralogy and oxide contents, soils might exhibit a net negative or positive charge at a given pH. Schofield & Samson (1954) for example, showed that kaolinite edge faces are positively charged at low pH and negatively charged at high pH. Changes in pH also affect the edge charge on clays and the surface charge of variable-charge minerals such as iron and aluminum oxides. Generally, iron and aluminum oxides experience a surface charge reversal between pH 7 and 9, positively charged below those pHs and negatively charged above, depending on structural composition and degree of crystallinity (Gupta *et al.*, 1984). The pH at which the net charge is zero is called the point of zero charge, ZPC. At lower pHs, edge to face bonding occurs, and bonding between positive iron and aluminum oxides and negative clay surfaces (van Olphen, 1977). As pH increases toward the ZPC, the degree of edge-to-face clay bonding via iron and aluminum decreases. Above the ZPC clay dispersion may result but in soils with appreciable quantities of CaCO₃, the effect of pH on clay dispersion is less dramatic than in soils with little CaCO₃; Ca is thus clearly an effective flocculant (Gupta *et al.*, 1984).

2.3.4 Organic matter

Organic matter, can act either as a bonding agent or dispersing agent for clay particles, depending on the ESP, the nature of the organic materials present, the nature of organo-mineral interactions that occur, and the degree of mechanical disturbance involved (Nelson and Oades, 1998). Long-chain organic anions can bond particles at the positively charged edges of clay plates (Emerson, 1954) or oxide surfaces. In soils, which contain sufficient polyvalent cations, long-chain organic anions link negatively-charged clay particles via cation bridges. Plant roots, mycorrhizal and saprophytic fungal hyphae, plus other particulate organic matter have the ability to stabilise microaggregates, and thus

prevent slaking and dispersion of clay (Nelson and Oades 1998). Transient binding materials that stabilize aggregates against slaking, however, have varying effects influenced by the degree of mechanical disturbance and decomposition. In non-sodic soils, for example, the microaggregates produced by slaking are quite stable and only disperse with considerable input of mechanical energy. In sodic soils, by contrast, once the transient bonds are broken and disaggregation has occurred, organic matter can act as a dispersant (Emerson, 1983). Humic acids, which can be aggregating agents in soil formation (particularly in the presence of cations), can also act as dispersing agents at concentrations >40 mg/L (Visser and Caillier, 1988). The dispersing effect relates to the large net negative charge on the humic materials at most pHs.

Gupta *et al.* (1984) and Baldock *et al.* (1994), for example, all showed that organic matter may increase clay dispersion, and Shanmuganathan and Oades (1983) found that clay of an illitic Australian soil containing adsorbed organic anions was more dispersive when it was simply in the presence of non-adsorbed organic anions. A smaller face-to-face attraction force due to poor contact between edges and terrace surfaces of Na-illite clays (Greene *et al.*, 1978) makes soils containing these clays more susceptible to dispersion than soils containing montmorillonitic clays (Oster *et al.*, 1980).

Kretzschmar *et al.* (1993) demonstrated that clays freed from their humic substances flocculated rapidly in distilled water at pH < 5, whereas a dispersed suspension of the same clays retaining their humic substances remained dispersed at pH 4.5. The CCCs of humic-free clays are invariably lower than those for clays containing humic substances, which suggests humic substances in natural clay systems act as strong dispersing agents. Durgin and Chaney (1984) reported that organic molecules associated with Douglas-fir roots were responsible for dispersion of kaolinite. These organic

molecules were predominantly large aromatic and aliphatic compounds with plenty of carboxylic and hydroxy groups. They also reported that polycarboxylic acids, which are common in forest soils, have greater negative charges than simple carboxylic acids, and so they cause greater dispersion in nature and contribute to erodibility of forest soils (Durgin and Chaney, 1984). Shanmuganathan and Oades (1983) reported that application of industrial organic anions (i.e. fulfate, citrate, oxalate, tatrane, salicylate, aspartate, lactate, and acetate) gave rise to clay dispersion in an Australian soil.

In sodic soils, the combination of high sodium content and the occurrence of organic carbon might even increase the dispersion of the clay. Exchangeable Na gives rise to the dispersibility by increasing the thickness of diffuse double layer (Shainberg and Letey, 1984), while organic anions could reduce the activity of flocculating agents (i.e. Ca^{2+} and Al^{2+}) in solution and enhance the negative charges of the clays, in turn promoting dispersion (Nelson *et al.*, 1999).

2.3.5 Mineralogical composition and clay dispersion

Arora and Coleman (1979) reported that different sodium-saturated clay minerals had different CCCs. Illite, vermiculite, smectite and kaolinite have CCCs of 185, 58, 28-60, and 8 meq/L, respectively. This indicates that illite is the most dispersible mineral studied, and is linked to its greater net-negative charge (also location of charge) and small particle size and shape relative to the other minerals. Smaller edge-to-face attraction forces due to the poor contact between edges and terraced surfaces might be responsible for the higher susceptibility of Na-illite to dispersion (Greene *et al.*, 1978). Churchman *et al.* (1993) predicted that differences in shape, size and/or morphological crystal features between illite and smectite may be responsible for varying tendencies of different minerals towards dispersion. Illite particles are relatively thick and only slightly flexible (Greene *et*

al., 1978), while particles of smectite, can be very thin and highly flexible (Emerson, 1983). This means that edge-to-face attractions between illite particles are weaker and less extensive than between smectite particles (Churchman, *et al.*, 1993).

Furthermore, non-crystalline Al- & Fe-oxides may greatly enhance flocculation of clay-systems. Goldberg and Glaubig (1987) found size and morphological features may be responsible for the flocculative ability of oxide particles. Greater proportions of sub-micrometer particles of Al-oxides and hydroxides in soils than of Fe-oxides generates higher charge densities in soils and thus gives Al-oxides higher flocculating power than Fe-oxide materials (Goldberg & Glaubig, 1987). Greater aggregate stability, permeability, friability, porosity and hydraulic conductivity are evidence of soil structural improvements caused by Al- and Fe-oxides (Goldberg, 1989).

The CCCs of each mineral, however, are also influenced by pH (Arora and Coleman, 1979), which is why El-Swaify (1970) observed that the CCC values for Na-illite increased as pH increased from 2 to 12 in NaCl and why Goldberg and Glaubig (1987) observed that the CCCs for both Ca-montmorillonite and Ca-kaolinite increased as pH increased from 5.5 to 9.3 in CaCl₂. Compared to montmorillonite, the CCCs of kaolinite and illite are more pH dependent, and so their dispersibility is also dependent on pH (Goldberg and Forster, 1990). Furthermore, Goldberg *et al.* (1988) also reported that SAR influenced the CCCs of kaolinite and illite more than it did for montmorillonite.

2.3.6 Particle size and porosity in relation to dispersion

Particle size distribution is related in various ways to the porosity of soils. For example, Bodman and Constantin (1965) found that the minimum porosity of soil occurred for soil with clay content close to 25 %, while Smith *et al.* (1978) and Coughlan *et al.* (1978)

claimed the minimum porosity for soil aggregates occurred at higher clay contents (50%).

It is also likely that the total porosity of the soils in the field is related to shrink-swell potential, root channels, animal burrows, worm channels, and inter-gravel pore space, which create preferential flow pathways (Luxmoore, 1981). For water and solute transmission through soils, the total porosity is less important than its pore-size distribution. Baven and Germann (1982) reported on the significant role of macroporosity in controlling rapid movement of solutes and pollutants through soils. Ehlers (1975) found that the cumulative water intake of untilled soil was dominated by earthworm channels, and that infiltration via the soil matrix was negligible. The amount of water transmitted in 5 min was 10 ml, 120 ml and 300 ml through earthworm channels of 2-5 mm, 5-8 mm and 8-11 mm in diameter, respectively (Ehlers, 1975). Clearly, the amount of water infiltrating could be much greater depending upon the number and size of earthworm channels. Vinten and Nye (1985) reported that most water that contains suspended materials moved through saturated soils in transmission pores $> 50 \mu\text{m}$ in diameter, because this is considerably larger than platelets, or other colloidal material, thus pore blockage was minimal. Noack *et al.* (2000) made similar observations for soil colloids moving through stable soil matrices of different aggregate/pore size distributions.

Smettem and Collis-George (1985) identified 38 macropores of ± 0.45 cm diameter per square meter cross-sectional area, while Wilson and Luxmoore (1988) estimated larger numbers (94 and 183) of macropores >0.15 cm diameter per square meter in two watersheds. Wilson and Luxmoore (1988) also reported the volume of macroporosity of two watershed soils to be 3.10^{-4} and $2.10^{-4} \text{ m}^3/\text{m}^3$, respectively. Ehlers (1975) found that untilled soil contained more (and larger) earthworm channels than tilled soil. Emmanuel (1973 cited in Ehlers, 1975) determined a mean of 55 casts per m^2 on the untilled soil

surface, which correlates well with the number of channels >5mm in the untilled Ap horizon found by Ehlers (1975).

The influence of macropores on water flow through soils is well recognised (Dixon and Peterson, 1971), and so, it is perhaps surprising that the inconsistency to describe and classify soil pores still exists. Luxmoore *et al.* (1990) concluded that at least 9 terminologies have been proposed since 1951 (Figure 2.3).

Each proposed terminology has a different criteria and definition of porosity. Skopp (1981) defined macroporosity as "pore space which provides preferential paths of flow so that mixing and transfer between such pores and remaining pores is limited. Macropores, may consist of interaggregate pore space, shrink and swell cracks, root channels or faunal tunnels". This definition is most useful for understanding the effect of porosity on clay dispersion and mobility (Luxmoore, 1981).

2.3.7 Hydraulic conductivity and dispersion

The hydraulic transport of clay, vertically or laterally through soil horizons, is strongly affected by salinity, sodicity, mineralogy, organic carbon, and pH. McNeal and Coleman (1966) demonstrated that hydraulic conductivity, K_s , invariably increased when electrolyte concentration increased at all ESPs studied (0, 12, 15, 19, 32, 49 and 100 %). Quirk and Schofield (1955) and Shainberg *et al.* (1981) reported similar results, but also showed that higher ESPs could be tolerated without reducing K_s when the EC was greater. When high quality irrigation water (e.g. rain water) is used on soils, ESPs as small as 5 (instead of 15 as proposed by the USSS) are large enough to reduce K_s in Australian soils (McIntyre, 1979) and lead to waterlogging and severe erosion.

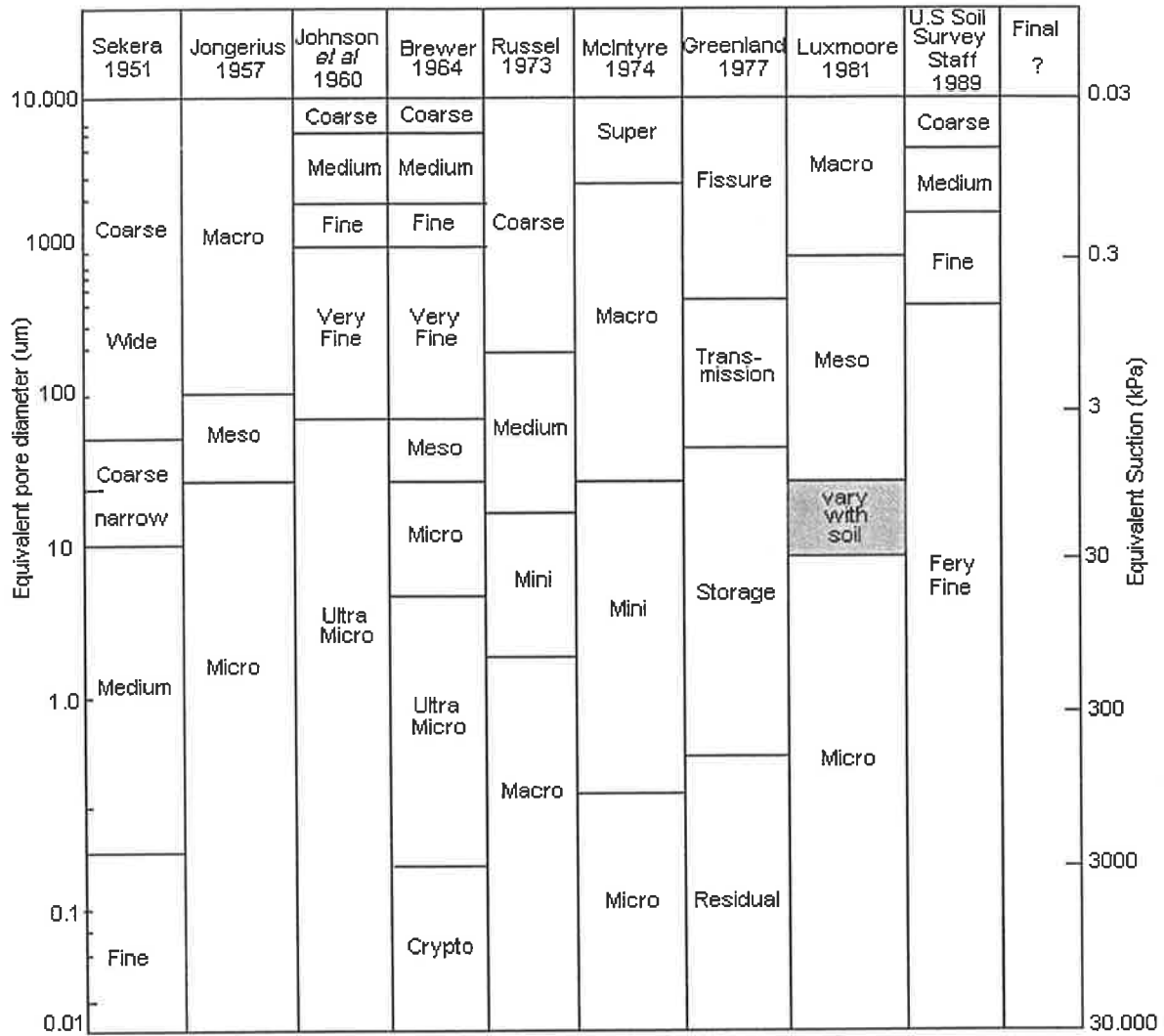


Figure 2.3. Various terminologies and criteria of porosity proposed by several authors since 1951 (from Luxmoore, 1981)

Levy *et al.* (1988) reported that greater exchangeable Mg contents (relative to Ca) reduced K_s at all ESPs studied. Ilyas *et al.* (1993) found no effect of applying gypsum to the soil in the field on K_s , even after 6 months, and this may have resulted from unmeasured Mg effects.

Alperovitch *et al.* (1985) reported strong relationships between K_s and clay mineralogy. Illite, because of its strong dispersive behaviour, tended to have smaller K_s than montmorillonite. El-Rayah & Rowell (1973) reported that Al- and Fe-oxides

maintained high K_s due to their ability to reduce swelling and dispersion. McNeal *et al.* (1968) also reported that clay fractions rich in Al- and Fe-oxides and kaolinites are relatively resistant to sodicity-induced K_s -reductions. Shainberg *et al.* (1981), however, found that a soil dominated with sesquioxides and kaolinites was the most dispersive when treated with exchangeable Na, leading to reduced K_s . They concluded that soils which release salts by weathering in greater concentrations are less susceptible to sodicity-induced clay dispersion and K_s -reduction when they are leached with dilute waters.

Frenkel *et al.* (1978) found little drop in K_s as bulk density increased, but a dramatic reduction in K_s when clay content and ESP increased. Hydraulic conductivity is also influenced by pH. For example, Suarez *et al.* (1984) reported that in solutions of high electrolyte concentrations (100 mmol_c L⁻¹), K_s was essentially independent of pH, but at lower electrolytes concentrations K_s decreased as pH increased.

2.4 Pedogenesis of texture-contrast soils

Duplex or texture contrast soils (Northcote, 1979) are widely distributed throughout Australia. Mt. Lofty ranges in South Australia are predominantly covered with duplex soils. Red, yellow and grey duplex soils (Natraqualfs) of the Mt. Lofty ranges (Fitzpatrick *et al.* 1996), are the soils which have a distinct texture contrast between soft (loose sandy) A horizons and massive or weakly to strongly pedal clayey B horizons in which the uppermost layer at least 15 cm thick is mottled and dominantly yellow or yellow-gray.

Approximately 20% of soils on the Australian continent consist of these texture contrast soils (Chittleborough 1992), and 56 % of the agricultural south-west of western Australia features duplex soils (Tennant *et al.*, 1992). Consequently, much work has been published in Australia concerning these texture contrast soils (e.g. Chittleborough 1992; Gregory *et al.* 1992; Tennant *et al.* 1992; Turner, 1992).

The controversy concerning this pedogenesis remains. In a study on 6 sodic red-brown earths of South Australia, Oertel (1961) concluded that clays in Bt horizons developed *in situ*, and did not come by illuviation. By this mechanism, texture-contrasts would occur where the intensity of weathering in the B horizon exceeded that in the A horizon. Chittleborough (1992), however, argued that weathering intensity was greater in the A horizon of most Urrbrae Red Brown Earths. Lithologic discontinuities or deposition of clay from other sources can also be responsible for abrupt textural-contrasts on sodic soils, rather than genetic processes (Stace *et al.*, 1968). Similarly, Bishop *et al.* (1980) concluded that the high clay content of the B horizons derived essentially by the *in situ* weathering of bedrock, and that the dominant sand fraction of the overlying horizon originated from the lateral surface movement of sandy materials. Bishop *et al.* (1980) concluded that downward movement of coarser grained materials played a significant role in the development of texture-contrasts. However, Chittleborough (1992) demonstrated that “downward movement” applies only to unusual cases, because (i) texture contrast soils only occur in both steep regions as well as on gently undulating-to-flat terrains; (ii) the thickness of A horizons in duplex soils in almost every topographic position of the Flinders Ranges was similar, indicating that the “downward movement” mechanism could not apply there, and (iii) the relationship between mineralogy of soil and underlying bedrock is very strong, indicating that sedimentary origin is incompatible. Fritsch and Fitzpatrick (1994) in their study in the Mt. Lofty ranges indicated that formation of duplex soils was generated by clay illuviation as the dominant processes in the landscape.

To distinguish between pedological and geological processes generating texture-contrast soils, the presence of argillans on ped surfaces (by careful morphological-geological field and laboratory examination) can indicate pedological clay illuviation

(Cabrera-Martinez *et al.*, 1989, Chittleborough, 1992).

The study of duplex soil formation cannot be separated from the study of argillic horizons. Eswaran and Sys (1979) distinguished 3 processes in the illuvial formation of argillic horizons: dispersion, translocation, and accumulation. Dispersion and translocation of clay have been reviewed in earlier sections. Accumulation processes can be largely controlled by physical factors (Duchaufour, 1982). For example, slow water flow leads to precipitation and accumulation of suspended fine particles (mostly $<0.2 \mu\text{m}$). Less soluble minerals (e.g. kaolinite) tend to precipitate more easily than other minerals such as feldspar (Fanning & Fanning, 1988). Precipitation of fine material may start during desiccation, which causes ions and molecules to concentrate and flocculate (Fanning & Fanning, 1988). While the genesis of duplex soils is not the main interest of the present study, the physical properties (e.g. hydraulic conductivity) causing texture-contrast soils is now examined to explain the possible mechanisms of lateral clay movement over Bt horizons.

2.5 Environmental implications of sodic, saline or sodic-saline soils

Poor physical properties of alkali soils induced by high sodicity and high pH invariably restrict water and air movement and thus plant growth (Acharya and Abrol, 1978). Rengasamy and Olsson (1991) also proposed when both salinization and sodification occur together, the destructive effects of sodicity are usually evident after salt-leaching. Fitzpatrick *et al.* (1994b) showed that sodicity may cause soil structure decline, dispersion, waterlogging and tunnel erosion, leading to increased hillslope and downstream water erosion, and contribute to reduced quality of water and reservoir capacity (Durgin and Chaney, 1984).

Several of these problems associated with salt-affected soils occur in the Herrmanns

sub-catchment of the Mt Lofty Ranges (Figure 2.4). Fritsch and Fitzpatrick (1994), for example, determined the link between waterlogging, clay migration, dryland salinity, sulfidic conditions and erosion in this sub-catchment. Fitzpatrick *et al.* (1994a) showed that the undesirable changes in structure of soil surfaces caused by intensive waterlogging created sheet and rill erosion. By contrast, cohesion breakdown in the E (A2) horizon as a result of sand accumulation due to clay migration and lateral discharge from perched water tables created tunnel erosion. Gully erosion in the Herrmanns sub-catchment was caused by lowering the saline ground water table (Fitzpatrick *et al.*, 1994b).

Fitzpatrick *et al.* (1996) also identified saline-sulfidic features (e.g. black sulfidic materials, iron-rich gelatinous precipitates, salt-iron crusts). Saline-sulfidic features found in the Mediterranean climate of Mt. Lofty Ranges was unusual (Fitzpatrick *et al.* 1996) because such saline sulfidic features are usually only common in coastal flood plains (Willet and Walker, 1982; Willet *et al.*, 1992). These soils receive an intensive supply of iron and sulfur from sulfide-bearing minerals (e.g. pyrite), which commonly occur in tidal marsh soils (Fanning and Fanning, 1988). The source of iron and sulfur in the saline-sulfidic features in Herrmanns sub-catchment is derived from the rising saline groundwater table that contains Fe and S.

In summary, land degradation (e.g. soil erosion, waterlogging, and dryland salinity) that currently exists in Herrmanns sub-catchment is associated with sodicity. This study deals with clay movement, the other problem induced by high sodicity of the soils.

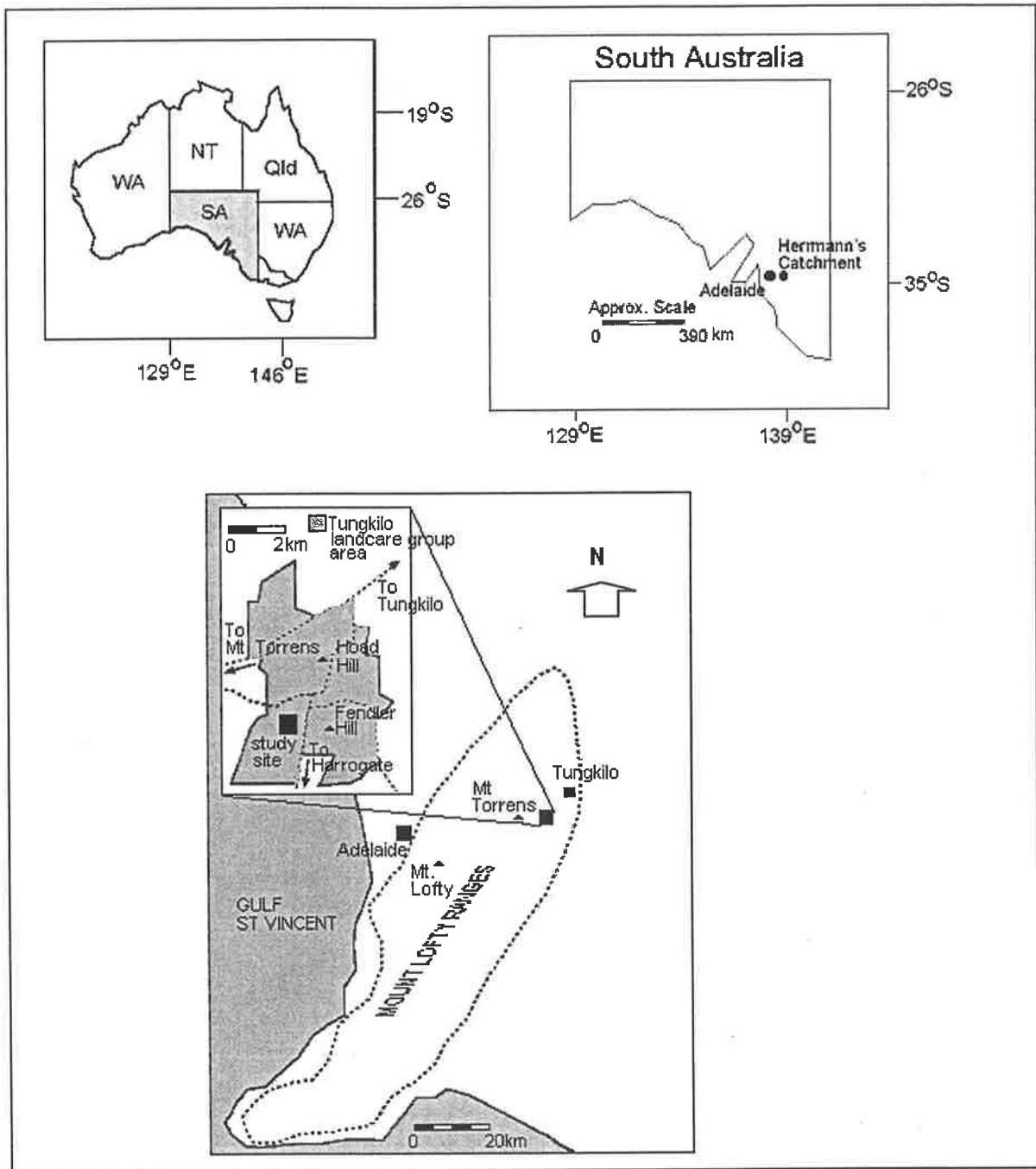


Figure 2.4. Locality of the study site (Herrmanns sub-catchment).

Chapter 3. Soil properties, their distribution along a toposequence, and soil classification

3.1 Introduction

Characterization of soil properties and their surrounding environment is crucial for a better understanding of the mobility and transport processes of clays in the landscape. Detailed characteristics of the soil can be organized into classes of soils with maximum similarity according to the Soil Taxonomy (Soil Survey Staff, 1999). The lower the level of the category, the more detailed is the soil characterization. The objectives of this chapter were to categorize the salt-affected soils using criteria of Sumner *et al.* (1998); to define the properties of the soils (e.g. soil colour, texture, EC, pH, clay dispersibility) and their distribution in the study site in relation to clay mobility along the toposequences; to relate the soil properties (particularly salinity, sodicity) with clay dispersibility; to classify the soils to the category of Family, which is the second lowest category after *series* in the Soil Taxonomy (Soil Survey Staff, 1999) and to investigate the soil distribution along the toposequences previously identified by Fitzpatrick *et al.* (1996).

3.2 Field and laboratory measurements

Samples were collected in three stages. In the first stage, seven soil cores were collected (to a maximum depth of 137 cm) from each of the foot-slope, mid-slope and crest of Herrmanns sub-catchment and analyzed for chemical (i.e. pH, CEC, OC, EC and exchangeable cations: Ca, Mg, K, Na) and physical (i.e. particle size distribution and clay dispersibility) properties. The results of this first stage were used to select locations along the toposequence, to excavate profiles for more detailed study (particularly in relation to hydraulic conductivity) and to install lysimeters (Figure 3.1). In the second stage, three profiles representing saline, sodic and non sodic-non saline areas were chosen. Samples were collected from each soil horizon down the profiles to 180 cm or the C horizon,

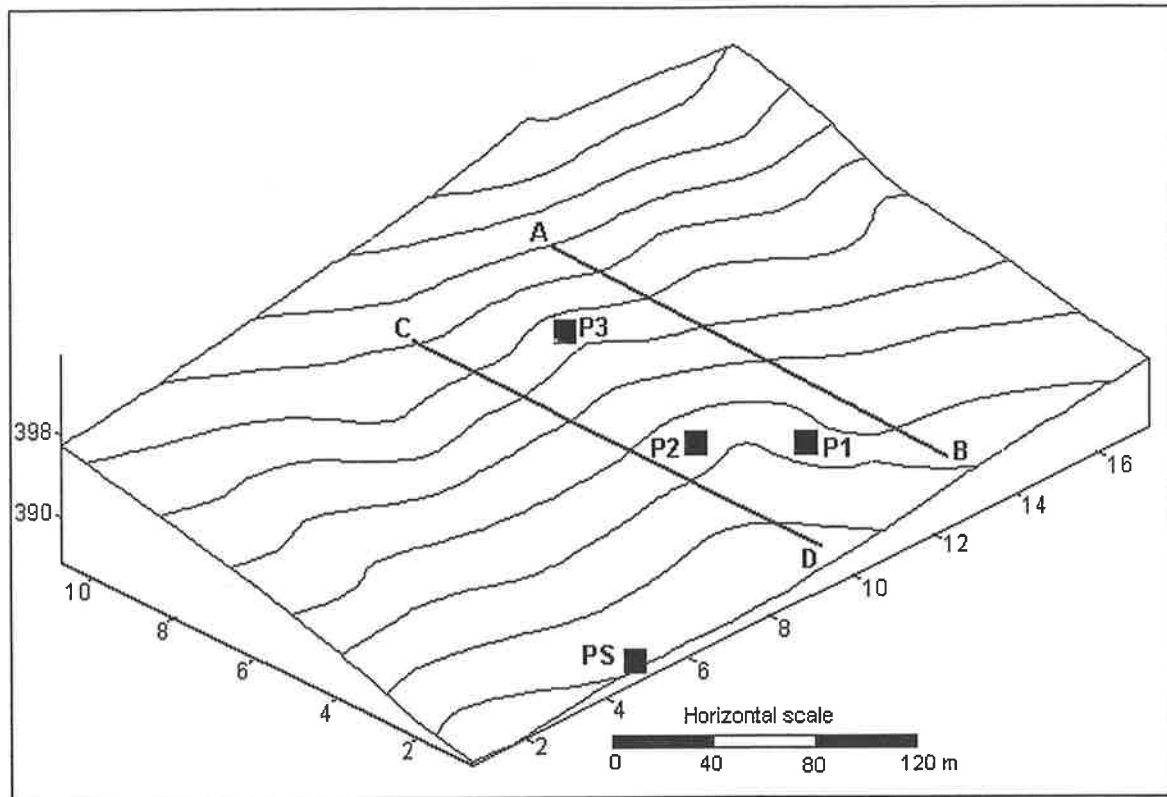


Figure 3.1. Schematic locality diagram showing position of profiles P1, P2, P3, PS and two toposesquences in relation to landscape position.

whichever was shallower. While collecting samples for taxonomic classification and thin section analysis, soil features were described in the field according to the Australian Soil and Land Survey Handbook (McDonald and Isbell, 1990), and classified using Soil Taxonomy (Soil Survey Staff 1999).

In the third stage, two toposesquences were chosen for more detailed field and laboratory investigations on a 50 m x 50 m systematic grid to map the boundaries of the saline and non saline areas. A drill rig was used to collect soil cores to 100 cm depth, and samples were brought into the laboratory to measure pH and EC.

The saturated hydraulic conductivity of each horizon was measured in the field at representative points within sodic, saline and non sodic-non saline soils to evaluate the extent of preferential water flow along the toposesquences. CSIRO disc permeameters

(Perroux and White, 1988) were used for measurements taken on the soil surface, whereas Guelph permeameters (model 2800) were used for all subsurface measurements (Reynolds *et al.*, 1983). Initial and final gravimetric water contents and bulk densities were collected for calculation of volumetric water contents. For determination of bulk density, undisturbed soil cores (89 cm³) were taken approximately 250 mm from the center of the permeameter disk. Saturated surface hydraulic conductivity was calculated using the following formula for the CSIRO disc permeameter:

$$K_o = q/\pi r_o - \frac{4bS_o^2}{\pi r_o(\theta_o - \theta_n)}$$

where r_o is the radius of the infiltration ring, θ_o is the final volumetric water content, θ_n is the initial volumetric water content, b is a constant (~ 0.55), $q/\pi r_o^2$ is the total cumulative infiltration at time, t , and S_o is the sorptivity, calculated from the slope of a plot of cumulative early infiltration versus square root of time t .

Subsurface saturated hydraulic conductivity was measured in the Ap and Bt horizons using a Guelph permeameter. Holes of diameter 6 cm were augered and a coarse-bristled brush was used to remove any smear layers which might block the flow of water out of the bore hole. After the preparation, permeameter assembling and water filling, the measurement was conducted. The first of two measurements at each point used a well-head height of 5 cm, and the second measurement used a height of 10 cm. The average saturated hydraulic conductivity was calculated using the values determined at each well-head height. The appropriate reservoir used was noted (inner reservoir was used if the infiltration rate was slow and both reservoirs were used if infiltration was rapid). The "field-saturated" hydraulic conductivity, K_s , was calculated as follows:

$$K_s = (0.0041)(X)(R_1) - (0.0054)(X)(R_2),$$

where X is the reservoir area constant selected and R_1 and R_2 are the rate of fall of water in the reservoir at the 10 cm and 5 cm well-head height respectively.

The particle size distribution was determined for all samples in the laboratory using a modified Gee and Bauder (1986) method to produce the proportions of clay ($<2\mu\text{m}$), silt (2 - 20 μm), fine sand (20-53 μm), medium sand (53-200 μm) and coarse sand (200-2000 μm). Samples of the < 2 mm fraction (10-20 g) were chemically dispersed using 10 % sodium hexametaphosphate and 0.6 M NaOH in 150 mL distilled water, and agitated for 24 h on an end-over-end shaker. Samples were then decanted and diluted to 500-700 mL (total volume determined subsequently) and allowed to stand for an appropriate time (according to Stoke's law) before the clay fraction was sampled. The sand fraction was collected, dried, sieved and weighed, and the silt fraction was calculated by difference.

All samples were checked qualitatively for signs of spontaneous dispersion in water. A sample of 10g air-dried soil ($< 2\text{mm}$ fraction) was placed in a petri-dish with sufficient distilled water to be completely immersed, and allowed to sit overnight. Because samples showed no sign of spontaneous dispersion, the mechanical clay dispersibility of the samples was then examined using a modification of the method of Nelson *et al.* (1999), as follows. A sample of 8 g air dry soil ($<2\text{mm}$) was placed in de-ionised water in a 50 mL vial. The vial was inverted once, gently, such that the aggregates just slid down the side of the vial and back. After an appropriate settling time, a 30 mL sample was collected to determine the amount of $< 2 \mu\text{m}$ material suspended. This was passed through a 0.45 μm filter and the filtrate dried and weighed to obtain the percent of 0.45 - 2 μm material after minor correction for salt in the filter. To eliminate the influence of salinity, the clay dispersibility was also measured in a low EC (< 0.07 dS/m) solution. A sample of 8 g air dry soil ($<2\text{mm}$) was placed in de-ionised water in a 50 mL vial, shaken in an end-

over-end shaker for 5 minutes and diluted with de-ionised water to the particular volume until the EC became less than 0.07 dS/m. The dispersible clay was measured gravimetrically after final shaking and standing.

Bulk density of soil clods (~ 25 g each) collected throughout the soil profile was determined by the clod method outlined in Blake and Hartge (1986). Masses were corrected for water contents. The wax used to seal the clods had a density of 0.91g cm^{-3} .

Electrical conductivity and pH of all soil samples were determined in 1:5 soil:water extracts (Raymont and Higgins, 1992) and in soil saturation pastes (Rhoades, 1982). For the 1:5 soil-water extracts, 8 g of air dried sample $\leq 2\text{mm}$ was weighed into an extraction jar with 40 mL of water. The jar was then shaken for 1 hour and allowed to settle for 20 minutes before EC and pH were measured in the supernatant.

Saturation paste extracts were determined by placing 150 g - 200 g soil in a 500 mL cup with a plastic spatula (weighed beforehand). Water was then added in small increments, and stirred with the spatula until the soil paste glistened as it reflected light, flowed slightly when the container was tipped, or slid freely off the spatula. After mixing, the cup was covered with film wrap and allowed to stand overnight. The paste checked the following morning to ensure no free water collected on the surface of the paste, and that the paste had not stiffened or lost its glossy state. If it appeared to be wet, a small quantity of air-dried soil was added to the mixture. The total amount of water added to the soil was obtained after re-weighing the cup with its content (i.e. soil paste and the spatula). The pH of the saturation paste extract was measured by immersing the pH electrode directly in the paste. Following pH measurement, the paste was transferred to a filter funnel fitted with highly retentive filter paper (Whatman No. 42), and vacuum suction was applied to separate the filtrate from the soil. Measurement of EC and ICP analyses for Na, Ca,

Mg and K were carried out using this filtrate. The sodium adsorption ratio (SAR) of the saturation-extract filtrate was calculated using the relationship $\frac{Na^+}{\sqrt{(Ca^{2+} + Mg^{2+})}}$, where the concentration of Na^+ , Ca^{2+} , and Mg^{2+} is expressed in mmol/L.

The cation exchange capacity (CEC) and the exchangeable base content were determined using the method described by Rayment and Higginson (1992). For the extraction, 2.5 g of air dry soil and 2.5 g of acid-washed sand were mixed in the leaching column with washed filter pulp placed at the bottom of the column. Then, 12.5 mL of ethanol 60% and 12.5 mL of glycerol 20 % were used to leach the soluble salts. The exchangeable cations were then leached with 50 mL alcoholic 1M NH_4Cl pH 8.5 into 125 mL plastic bottles containing 40 mL of 0.5 M HCl, and the volume increased to 100 mL using 1M NH_4Cl pH 8.5 so as to obtain a soil/solution ratio 1:40. The leachate was analyzed using flame photometry and inductively coupled plasma atomic emission spectrometry (ICPAES) for exchangeable cations (K^+ , Na^+ , Ca^{2+} and Mg^{2+}).

For Soil Taxonomy purposes, the base saturation of the soil (the ratio of the sum of exchangeable bases to CEC) was calculated. The CEC used in this calculation was determined by continuing the method for exchangeable base analysis. The soil sample with excess NH_4Cl was washed with 10 mL of alcoholic 0.1M NH_4Cl and 10 mL of ethanol 60%. The NH_4^+ that occupied the exchange sites in the soil was displaced with K-Ca solution [a solution of 15% KNO_3 plus 6% $Ca(NO_3)_2 \cdot 4H_2O$] into 125 mL plastic bottles containing 40 mL of 0.1 M HNO_3 . The leachate, was transferred to a 100 mL volumetric flask and was made up to volume with a K-Ca displacing solution. An autoanalyzer was used to determine the amount of K-Ca on the exchange sites for the CEC calculation. The CEC obtained from this method was used in the calculation of exchangeable sodium

percentage (ESP) using the following formula: $\frac{\text{Na}^+}{\text{CEC}} \times 100$.

Organic carbon content of all samples used the Walkley-Black rapid titration method (Allison, 1965). One gram (adjusted according to level of C expected) of finely ground air dried sample was placed in a 500 mL conical flask to which 10 mL of 1N $\text{K}_2\text{Cr}_2\text{O}_7$ solution was added, along with 20 mL concentrated sulfuric acid. The flask was shaken gently for 1 minute and allowed to settle for 30 minutes. Then, 200 mL distilled water, 10 mL of concentrated orthophosphoric acid and 0.5 mL o-phenanthroline solution were placed in the flask. Titration of the excess chromic acid with ferrous sulfate solution from a burette was started when the solution returned to room temperature. Titration was stopped when the solution turned from green to red. The organic carbon content was calculated on a dry weight basis after correcting using a calibration standard.

The presence or absence of various clay minerals in soil samples was determined by powder XRD analyses, using a Philips PW1710 X-ray diffractometer combined with a pattern analyzing program, XPLOT (Raven, 1990). The proportion of different minerals was determined semi-quantitatively by measuring peak heights of the XRD patterns and areas relative to the other minerals. Analyses were conducted using samples collected during particle size separation. Clay particles ($< 2\mu\text{m}$) were flocculated by adding NaCl. The supernatant was discarded and the flocculated clay was transferred to a Beckman centrifuge tube (250 mL). Prior to centrifugation at 6000 rpm for 10 minutes, 2 mL 5 M acetic acid was added to remove the phosphate used during the dispersion stage, and the volume was made-up with distilled water. The clay was then saturated with Ca by adding 5 mL 2M CaCl_2 , and again the suspension centrifuged at 6000 rpm for 10 minutes. The clear CaCl_2 solution was poured off and ethanol added to remove excess CaCl_2 . The suspension was mixed using a Bronson ultrasonic probe at high power for 4 minutes, followed by

centrifugation at 6000 rpm for 10 minutes. Again, the alcohol was decanted and the alcohol-washing treatment was repeated to ensure the suspension was free of CaCl_2 . To obtain a clay powder, the tube was placed in an oven at 60°C for 1 week after which the clay was transferred to a small beaker for further drying at 100°C . For un-oriented XRD analyses, a small quantity of powder was suspended in distilled water and dried at 40°C for 20 minutes in a sample holder, ready for XRD analysis.

3.3 Results

3.3.1 Soil morphological properties and classification

Duplex soils were the dominant soils found along the toposequences and these all classified as Alfisols (Soil Survey Staff, 1999) of the following subgroups: Typic Natraqualfs in the eroded area of foot slope (profile P1), Typic Natrixeralfs adjacent to the stream bank (profile PS), Typic Natrixeralfs on the waning mid-slope (profile P2) and Ultic Palexeralfs on the upper waning mid slope (profile P3) (Table 3.1).

The depth of the four profiles studied varied between 119 and 180 cm, with the deepest (180 cm) being profile PS (stream bank) and the shallowest (119 cm) was profile P3 (upper waning slope). The dry colour of the soil surfaces varied between very dark grayish brown (10YR3/2) for profile P3 and light gray (10YR7/2) for profile P1, whereas the subsurface of the soils had lighter colours ranging from strong brown (7.5 YR 6/6) for profile P1 to light brownish gray (2.5 Y 6/3) for the deepest horizon in profile P3 (Appendices # 1, 2, 3 and 4).

None of the soils effervesced with HCl, indicating the soils contained little or no carbonates. A few medium gravel fragments were observed on the surface of Profile P1 and P3, and the amount increased with depth to the lower horizon in profile P1. The morphological structure of the A1 horizons in profiles P1 and P2 is polyhedral, and in

Table 3.1. Classification of soils using Soil Taxonomy (Soil Survey Staff, 1999).

Category	Soil classification	Criteria and evidence
Profile PS (edge of stream bank)		
Order	Alfisol	- Argillic Bt horizon with base saturation 69% (>35%).
Suborder	Xeralf	- Xeric moisture regime (dry in all parts of soil moisture control-section >45 days/y)
Great Group	Natrixeralf	- Natric Bt horizon (ESP=14-16)
Subgroup	Typic Natrixeralf	- No specific properties
Family	Fine, mixed, non calcareous, thermic, Typic Natrixeralf	- clay 39 %, clay mineral kaolin, mica and illite, no effervescence with 25 % HCl, thermic temperature regime.
Profile P1		
Order	Alfisol	- Argillic Bt horizon with base saturation 90% (>35%)
Suborder	Aqualf	- Aquic Bt horizon (chroma ≤ 2 and gley colour)
Great Group	Natraqualf	- Natric Bt horizon (ESP=13 - 26)
Subgroup	Typic Natraqualf	- No specific properties
Family	Fine, kaolinitic, non calcareous, thermic, Typic Natraqualf	- Clay 58%, control section dominant mineral kaolin, no effervescence with 25 % HCl, thermic temperature regime.
Profile P2		
Order	Alfisol	- Argillic Bt horizon with base saturation 95% (>35%).
Suborder	Xeralf	- Xeric moisture regime (dry in all parts of soil moisture control section for >45 days/y)
Great Group	Natrixeralf	- Natric Bt horizon (ESP=13-14)
Subgroup	Typic Natrixeralf	- No specific properties
Family	Clayey, mixed, non calcareous, thermic, Typic Natrixeralf	- Clay 33 %, clay mineral kaolin, mica, illite, does not effervescence with 25 % HCl, thermic temperature regime.
Profile P3		
Order	Alfisol	- Argillic Bt horizon with base saturation 68% (>35%).
Suborder	Xeralf	- Xeric moisture regime (dry in all parts of soil moisture control section for >45 days)
Great Group	Palexeralf	- Clayey particle size on the sub- surface horizons and clay increase with 20 % or more within a vertical distance of 7.5 cm
Subgroup	Ultic Palexeralf	- Base saturation < 75% in argillic horizon.
Family	Clayey, mixed, non calcareous, thermic, Ultic Palexeralf	- Clay 34%, clay minerals kaolin, mica, illite, no effervescence with 25 % HCl, thermic temperature regime.

profiles P3 and PS it is subangular blocky. The dominant soil structure in all B horizons is columnar (Appendices # 1, 2, 3 and 4), reflecting the high sodicity on exchangeable sites of clay. For all soils studied, the exchangeable sodium percentage and/or SAR was found to be critical for classifying Great Groups (Table 3.1), because “natric” subsurface horizons were observed in three out of four profiles (i.e. profiles P1, P2 and PS). For the category of Subgroup, aquic conditions in the Bt horizon was the diagnostic property in profiles P1 and PS, whereas in profile P2 no special properties were found (typic), and base saturation was < 75 % on profile P3.

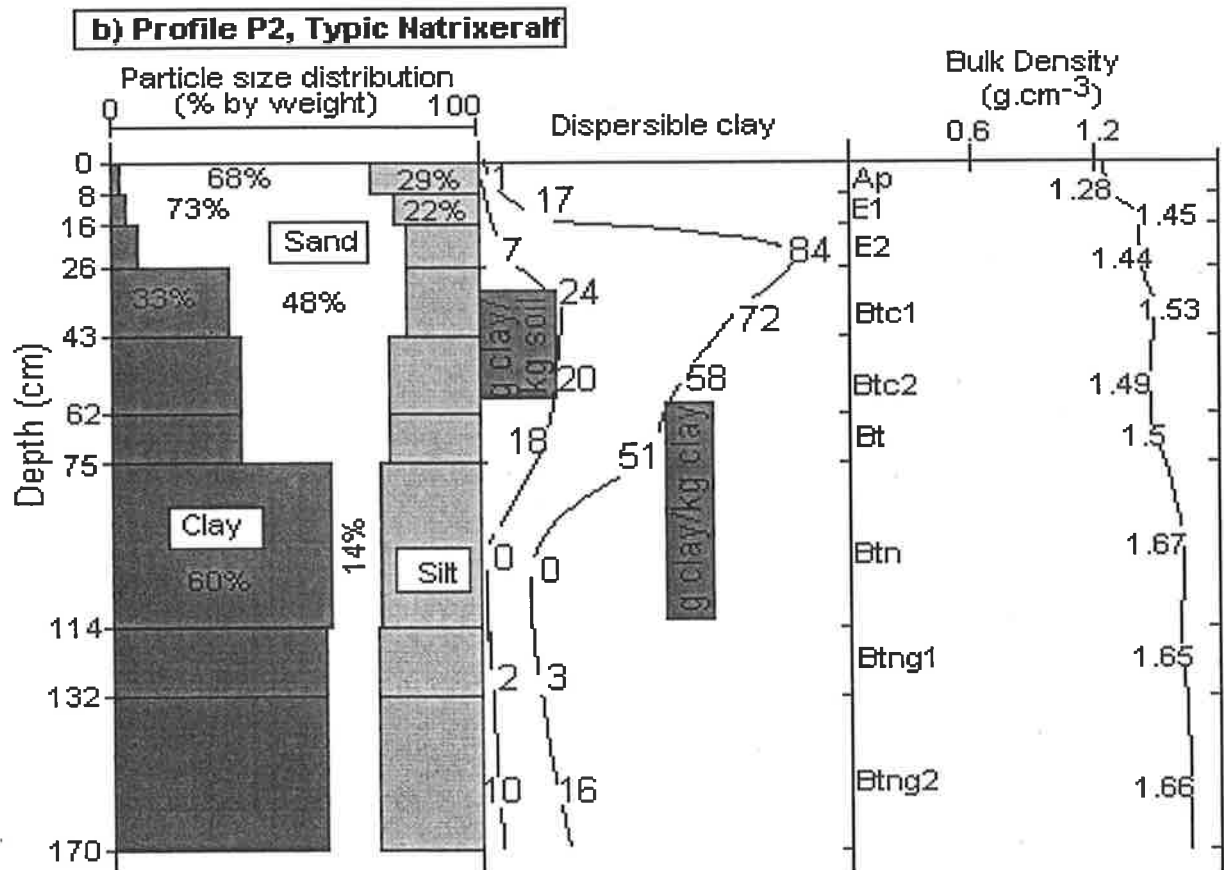
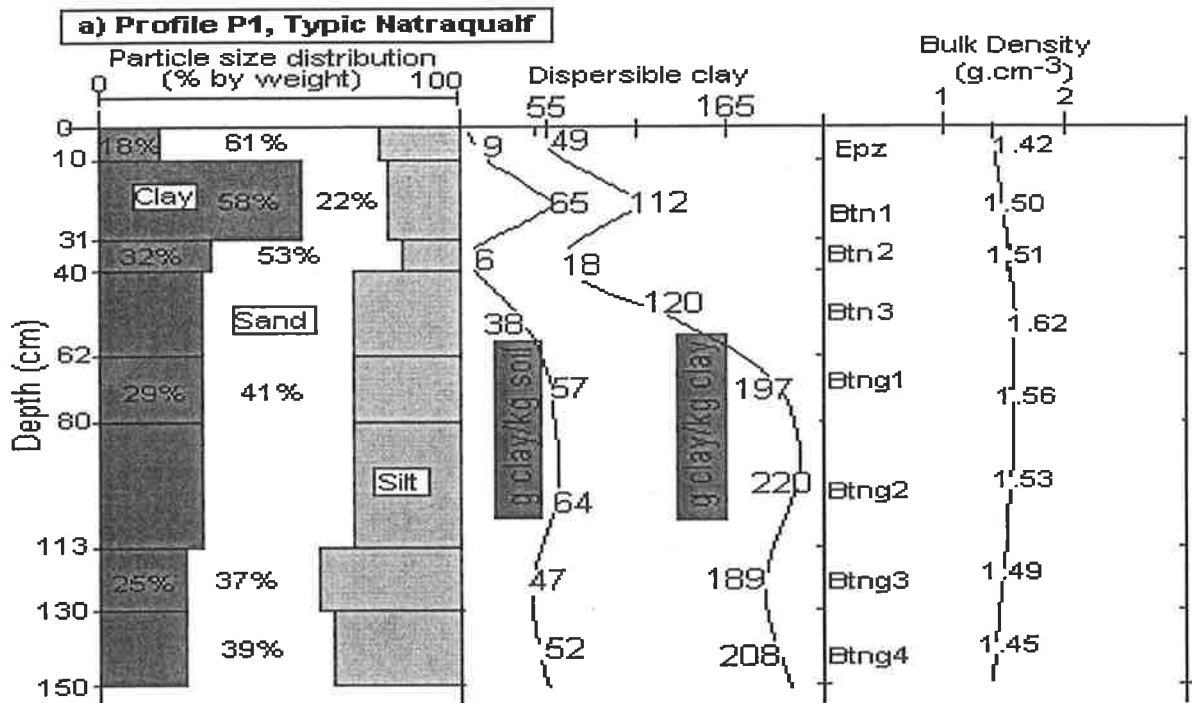
3.3.2 Soil physical properties

Particle size distribution, clay dispersibilities and bulk densities are presented in Figure 3.2 for all the main soil profiles down the toposequence (i.e. profiles P1, P2, P3 and PS). Particle size distribution showed a contrast in texture between surface and subsurface horizons. The sand fraction in all profiles occurred predominantly in the soil surface horizons, but the proportion decreased sharply with depth. This corresponded with a sharp increase of clay content in the subsurface horizons.

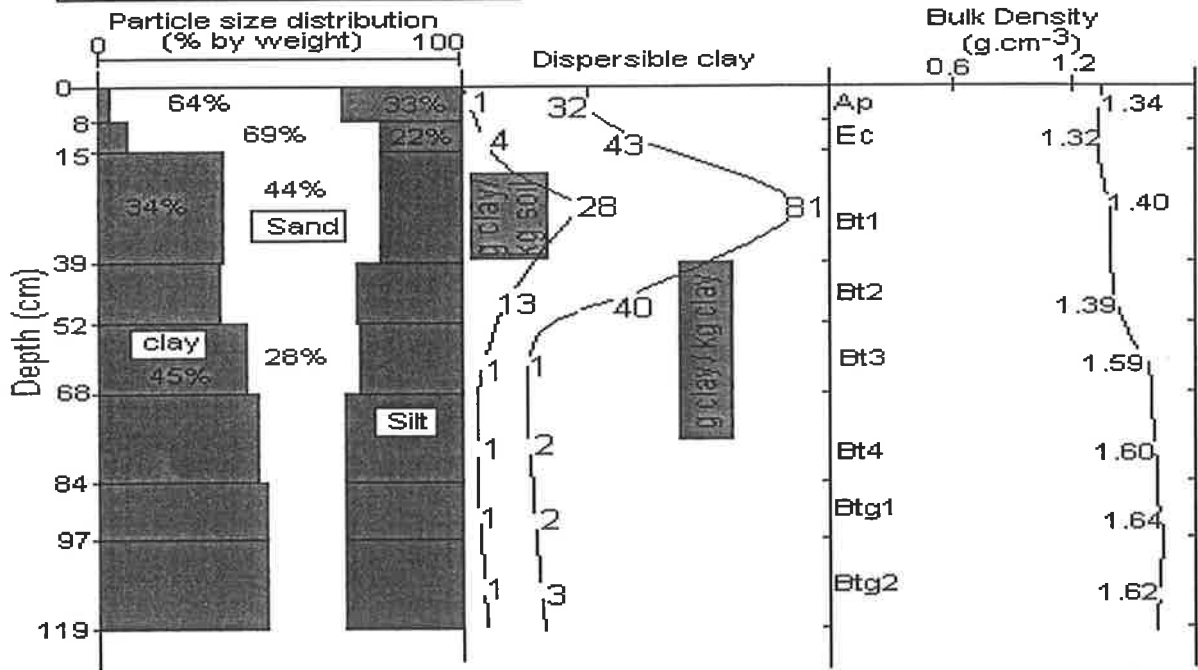
The dispersible clay content in profile P1 tended to increase with depth, whereas the other two profiles (P3 and PS), there was a clear decrease in clay dispersibility with depth. For profile P2, clay dispersibility was highest at depth between 16 to 60 cm.

Bulk density followed a general pattern, with low bulk density in the surface horizon and higher in the subsurface horizon. Bulk densities in all horizons of profile PS were generally lower than in the other profiles (P1, P2 and P3). There was a significant decrease in hydraulic conductivity from the soil surface to the subsurface (Table 3.2). In general, hydraulic conductivity dropped from moderate or moderately slow on the soil surface to moderate or extremely slow in the subsurface.

Figure 3.2 Profiles showing particle size distributions, mechanically dispersible clay contents and bulk densities for (a) Profile P1, (b) P2, (c) P3 and (d) PS.



c) Profile P3, Ultic Palexeralf



d) Profile PS, Typic Natixeralf

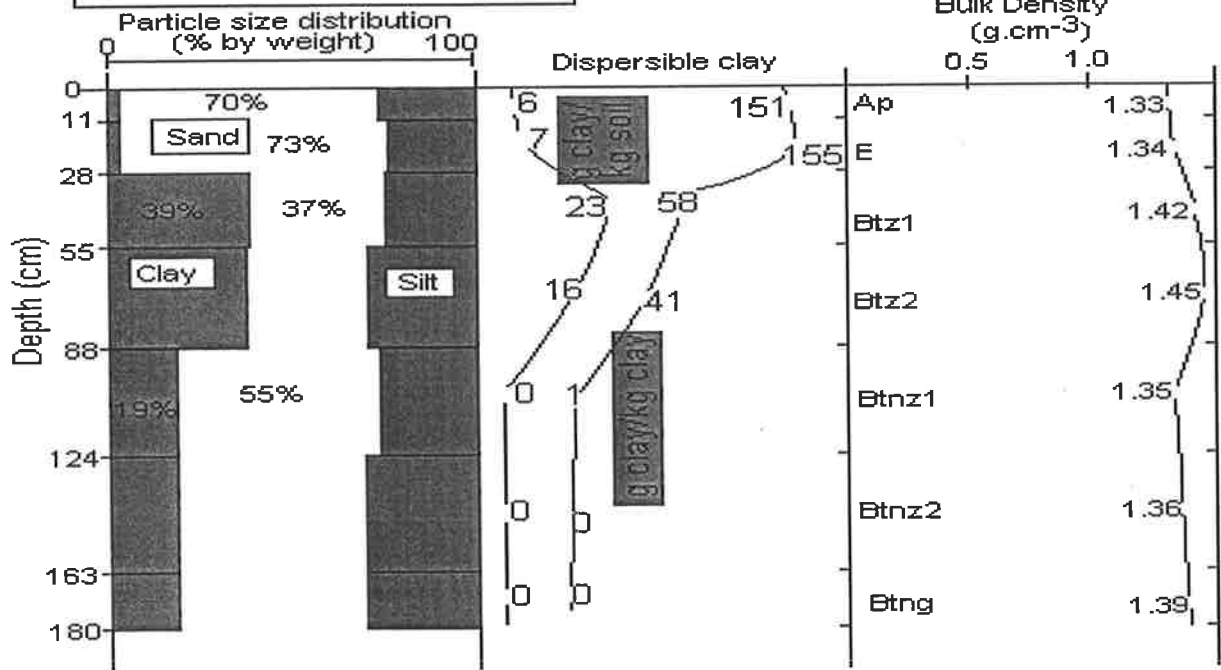


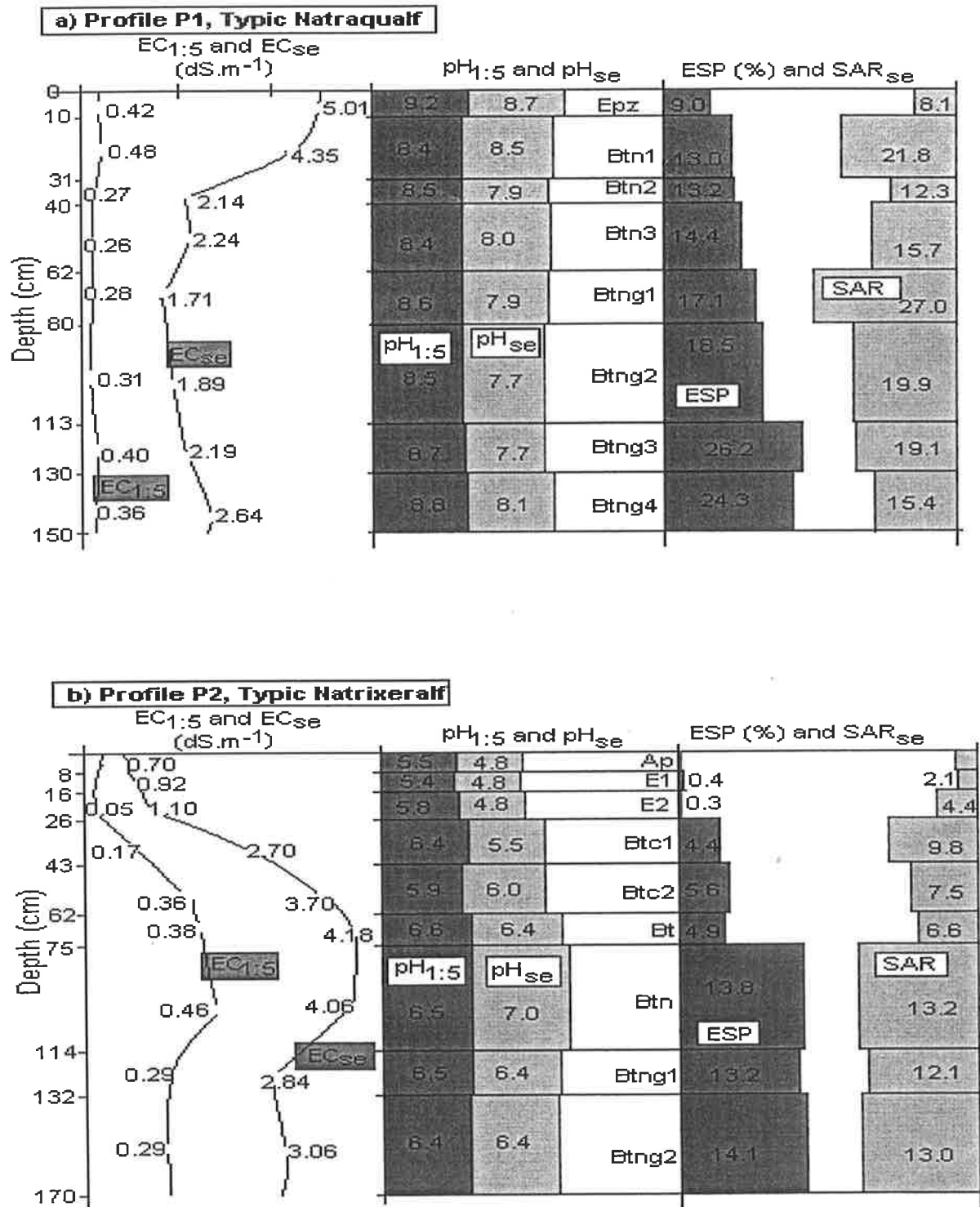
Table 3.2. Saturated hydraulic conductivity measured at different depths and positions along the toposequence using disc- and bore hole-permeameters

Profile	Depth (cm)	K_s (cm.s ⁻¹)	Permeameter	Category (Miller & Gardiner, 1998)
P1	Surface	$3.7 \times 10^{-4} \pm 0.9 \times 10^{-4}$ (n=4)	Disc	Moderately slow
	Subsurface (Btnc1)	$4.8 \times 10^{-6} \pm 0.4 \times 10^{-6}$ (n=3)	Bore-hole	Extremely slow
P2	Surface	$5.1 \times 10^{-4} \pm 0.4 \times 10^{-4}$ (n=4)	Disc	Moderate
	Subsurface (Btnc1)	$7.9 \times 10^{-6} \pm 0.4 \times 10^{-6}$ (n=3)	Bore-hole	Very slow
P3	Surface	$7.5 \times 10^{-4} \pm 0.3 \times 10^{-4}$ (n=4)	Disc	Moderate
	Subsurface (Bt1)	$5.7 \times 10^{-5} \pm 0.6 \times 10^{-5}$ (n=3)	Bore-hole	Slow

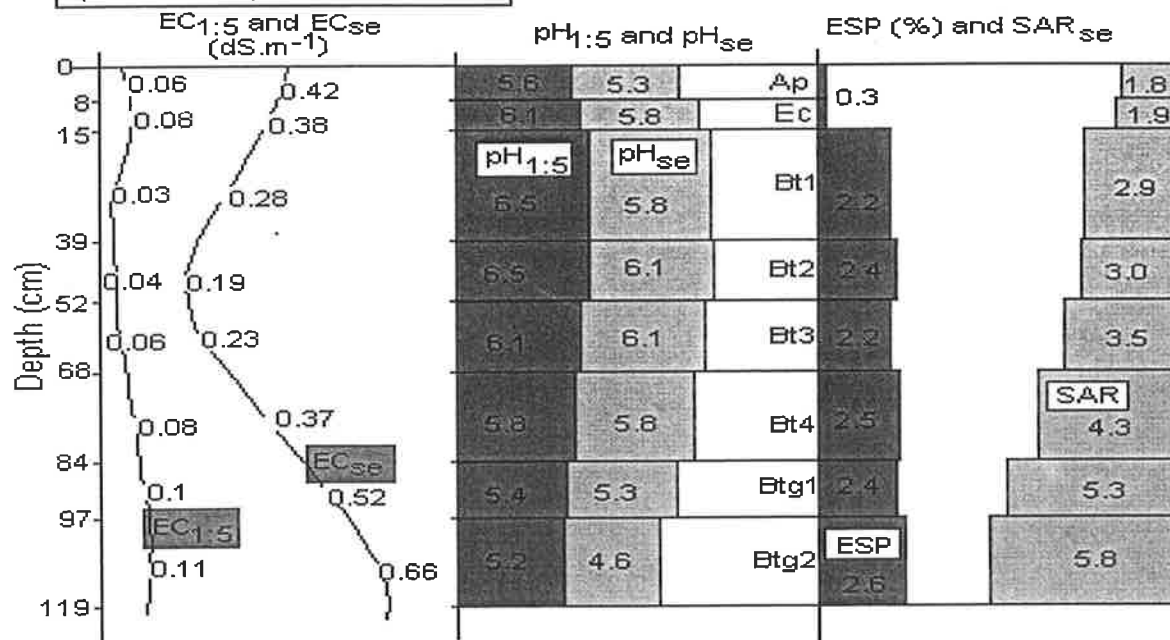
3.3.3 Soil chemical and mineralogical properties

Electrical conductivity, pH, exchangeable sodium percentage (ESP) and sodium adsorption ratio (SAR) of the four profiles are presented in Fig 3.3. Electrical conductivity, measured using the two methods ($EC_{1:5}$, EC_{se}), showed similar trends within the profiles. EC values in profiles P1 and PS were high in the soil surface horizon and dropped slightly in the middle horizon before rising again in the deeper horizons. The EC of profiles P2 and P3 generally increased with depth. The soil pH generally increased with depth in the soil profiles, and the 1:5 values were usually greater than those of the saturation-paste extracts (as might be expected). The greatest differences between 1:5 and saturation-extracts occurred deeper in the profiles (e.g. profile P1, 120 cm $pH_{1:5} = 8.7$; $pH_{se} = 7.7$). The

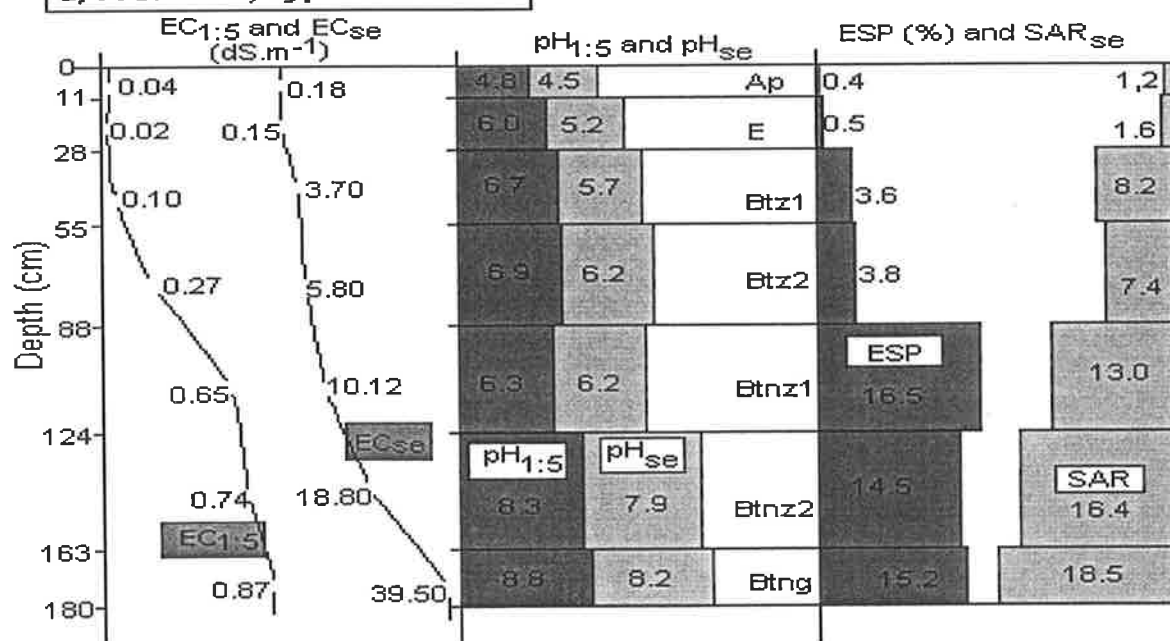
Figure 3.3. Profiles of EC, pH, ESP and SAR for (a) Profile P1, (b) P2, (c) P3 and (d) PS



c) Profile P3, Ultic Paleixeralf



d) Profile PS, Typic Natixeralf



ESP and SAR values for profile P1, were the highest amongst all profiles studied. Profiles P2 and PS had low values of SAR and ESP at the soil surface but these increased markedly at depths >75 cm. The lowest values of ESP and SAR were found in profile P3.

The CEC, exchangeable cations and organic carbon contents are shown in Fig 3.4. The CEC of all profiles was <20 cmol(+) kg⁻¹ and relatively uniform with depth. The increase in CEC for profile P2 was moderate (~5 cmol(+) kg⁻¹ at the soil surface, increasing to ~17 cmol(+) kg⁻¹ in the lower B horizon). The exchangeable cations of all profiles were dominated by Mg followed by Ca, Na and K respectively. Profile P1 contained highest amounts of exchangeable Na, followed by P2, PS and P3.

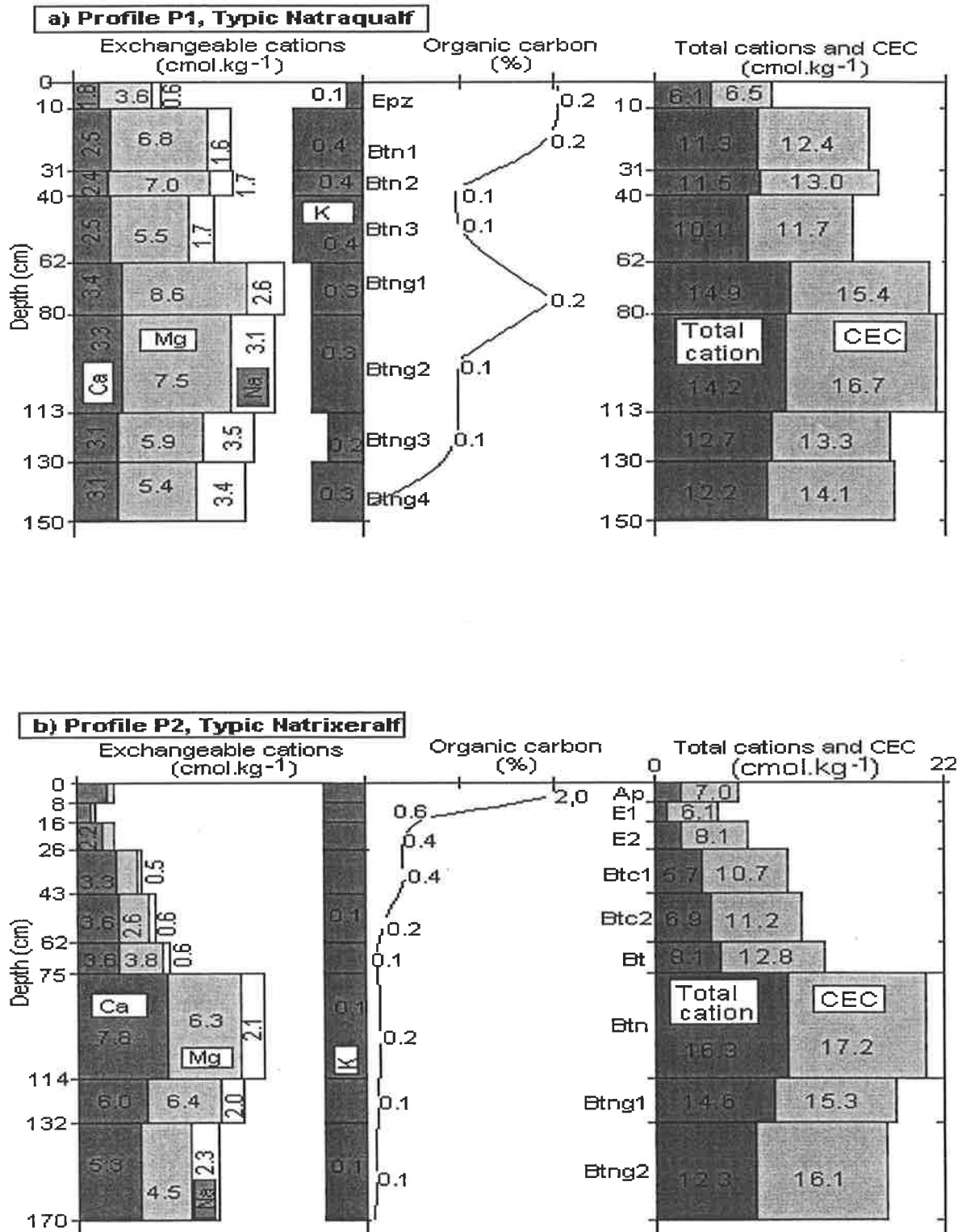
Organic carbon (OC) contents were greatest in all surface or A1 horizons, and decreased markedly with depth. Profile P1 contained very little OC (0.1 - 0.2 %), whereas profiles P2, P3 and PS had relatively higher OC contents in the upper horizons (2.0, 2.7 and 0.9 respectively).

The XRD analyses (Figs 3.5a-d) revealed that kaolin was the dominant layer silicate in all soils. Illite was sub-dominant and an interstratified smectite mineral occurred in profile P1 at >113 cm. The proportion of kaolin to illite varied between profiles and within horizons, but overall, the ratio of kaolin:illite was > 3:1.

3.3.4 Distribution of soil properties in toposequences in eroded and non-eroded areas

In general, soil properties of the two toposequences (eroded and non-eroded areas), combined with the profile data, varied greatly in the foot slope positions of the toposequences (Figure 3.6a and 3.6b). In toposequence A-B (eroded area), soil colour of the soil surface changed gradually from brown (10YR5/3) in the sequence to grayish brown (2.5Y5/2), dark grayish brown (10YR4/2), very dark greyish brown (10YR3/2), and abruptly changed to light gray (10YR7/2) along the foot slope. Soil colour in the sub-

Figure 3.4. Profiles showing cation exchange capacities (CEC), exchangeable cations and organic carbon contents for (a) Profile P1, (b) P2, (c) P3 and (d) PS



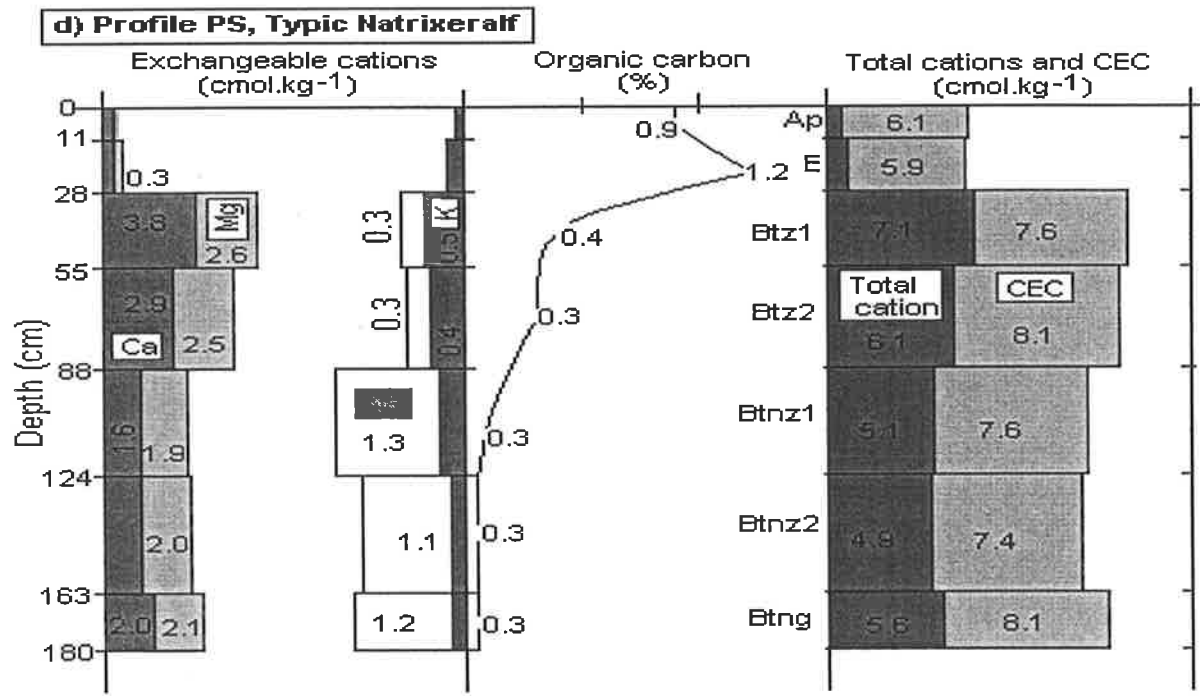
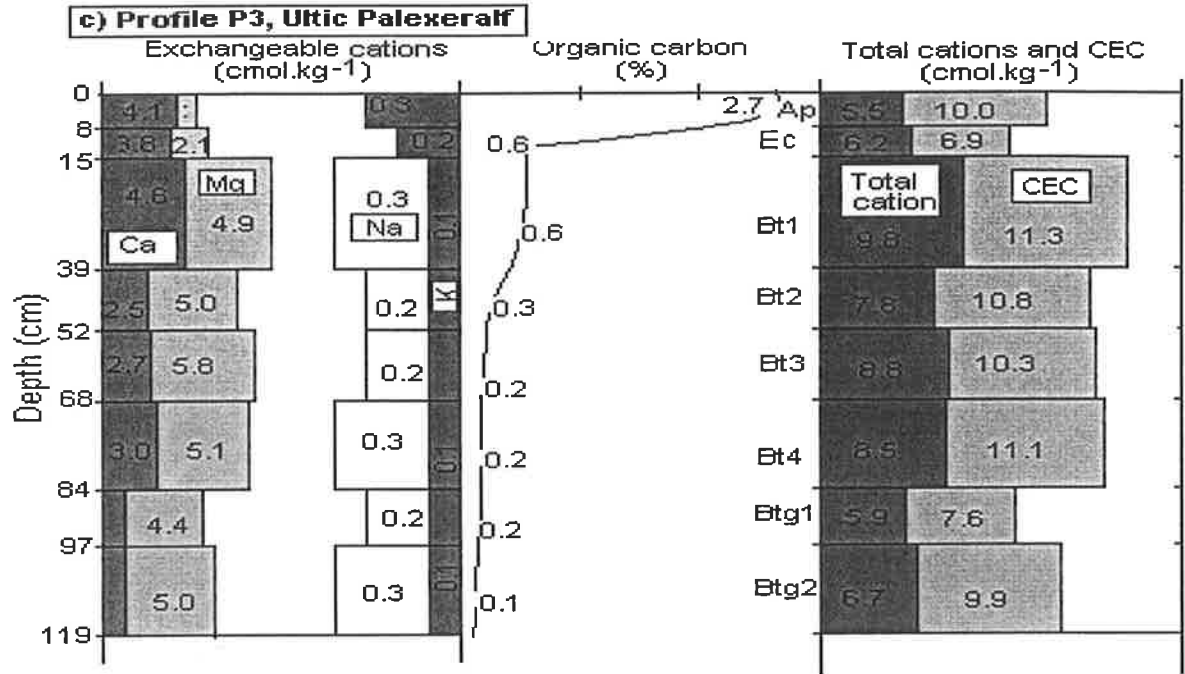
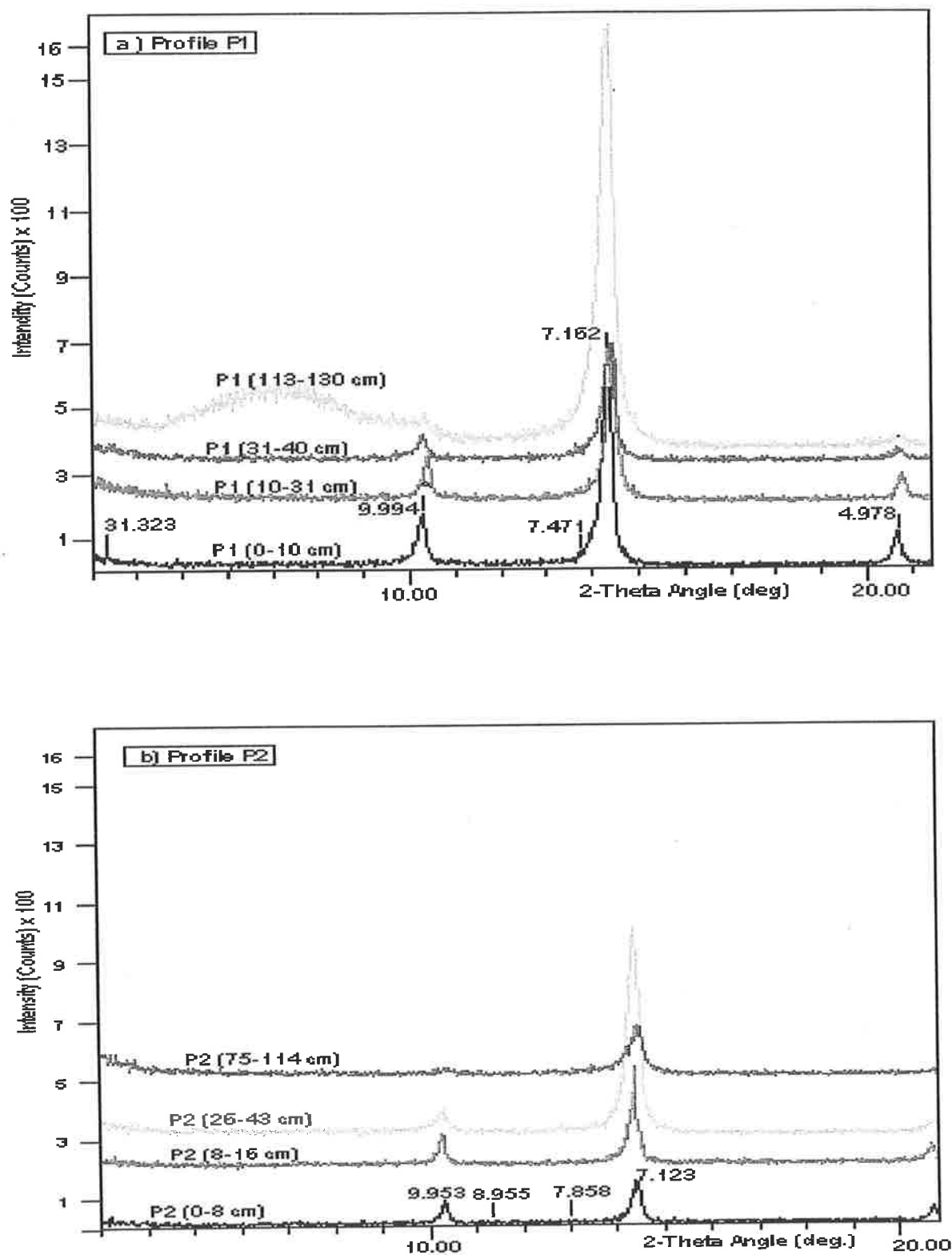
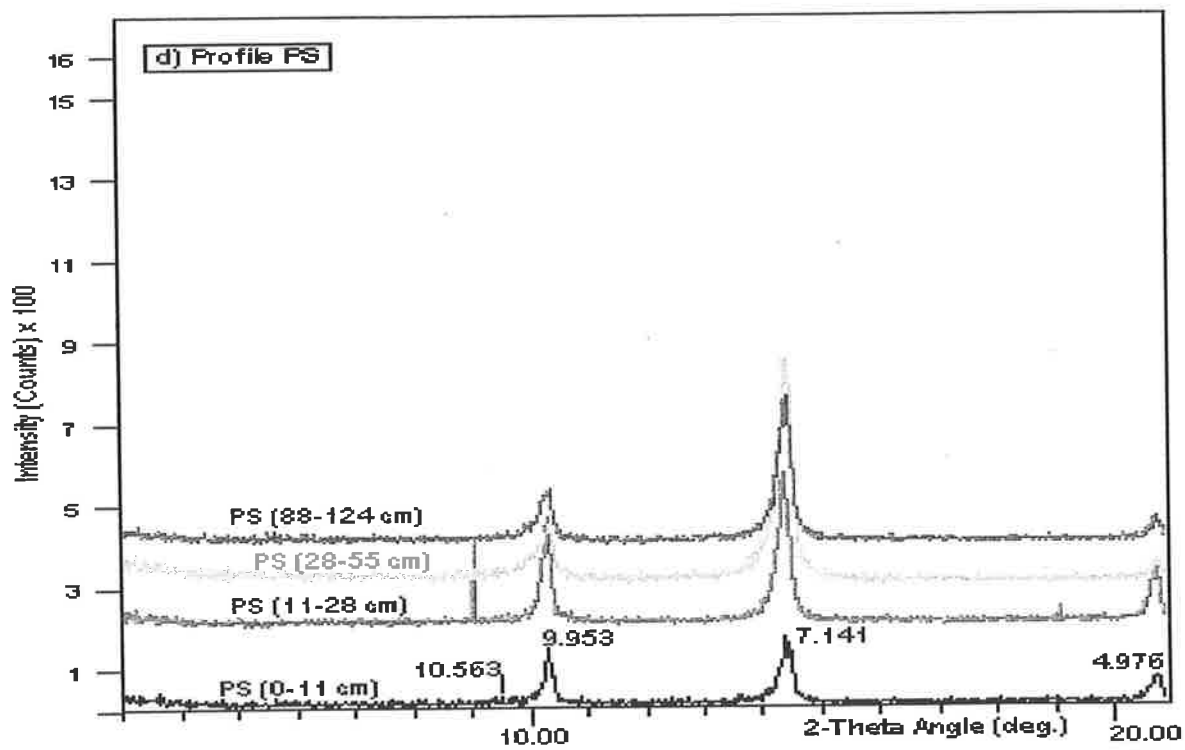
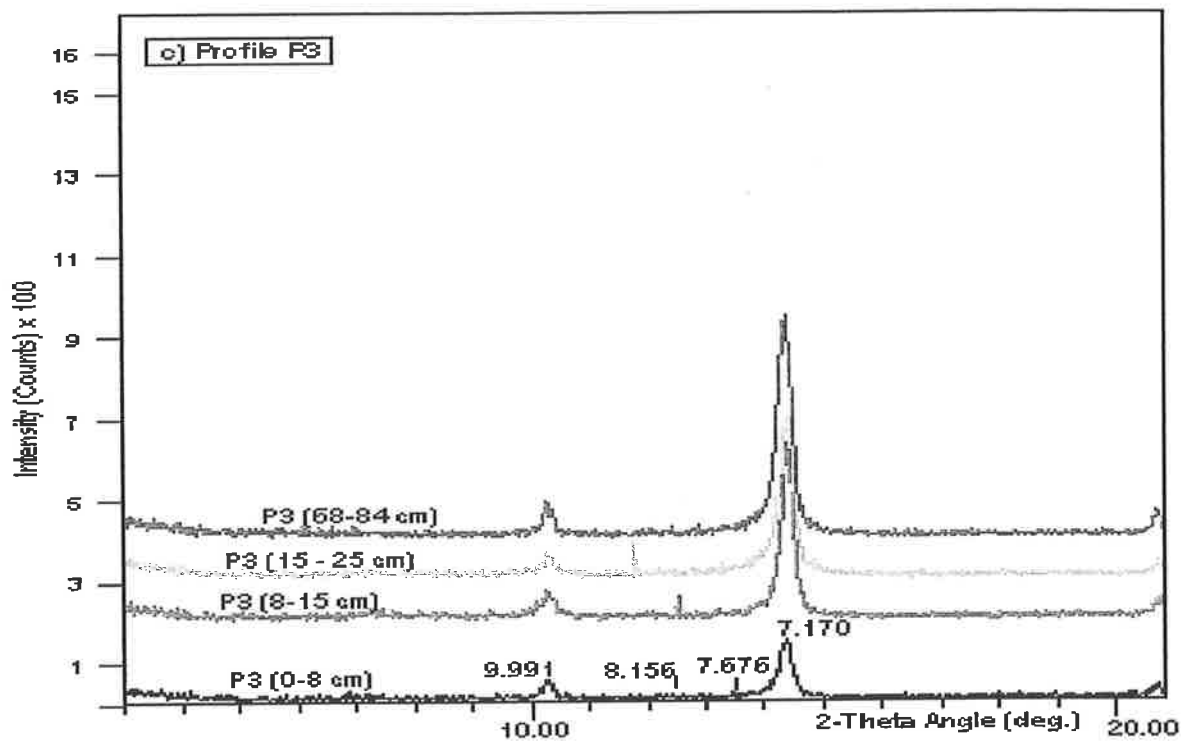


Figure 3.5. XRD patterns for each horizon in profiles (a) P1, (b) P2, (c) P3 and (d) PS.





-surface horizons (Bt) along toposequence A-B changed from red (10R5/8 or 2.5YR4/6) along the upper slope of the sequence to yellowish red (5YR4/6), and abruptly changed to reddish yellow (7.5YR6/6) in the foot slope (Figure 3.6a). A similar pattern of soil colour was observed in the upper slope of the topo-sequence C-D (non-eroded area), except in the foot slope a much darker colour (strong brown, 7.5YR5/8) was observed compared to that shown in toposequence A-B. Gravel contents varied between eroded and non-eroded regions: along the foot slope of toposequence A-B (eroded area) it was 35%, whereas along the foot slope of toposequence C-D it was 5%. Vegetation cover, which could affect the mobility of clay in soil surface, also showed differences: in the eroded area, there was no vegetation cover, which may explain why salt crusts were found on soil surfaces & stream banks in eroded areas, while in non-eroded areas (toposequence C-D) salt crusts were only found on stream banks.

The $EC_{1.5}$ values of the core samples collected using a 50m x 50 m grid system indicated that the soils in the lower slope position had markedly higher $EC_{1.5}$ values compared to soils in the upper slope positions. From the contour line of 387m to 392m above sea level, the soil $EC_{1.5}$ varied between 0.4 and 1.0 dS/m, whereas in the upper slope soil $EC_{1.5}$ ranged from 0.05 to 0.01 dS/m (Fig. 3.7).

3.4 Discussion

3.4.1 Sodicity-salinity criteria, soil properties and clay dispersibility

Using the recently proposed scheme of Sumner *et al.* (1998) for describing of salt- and Na-affected soils in terms of physical behaviour (dispersibility), the soils in the Herrmanns sub-catchment were catagorised as follows (Fig. 3.8):

- Profile P1 was classed as *Sodic Saline*.
- Profile P2 was classed as *Saline* (in upper horizons); *Sodic-Saline* in lower horizons.
- Profile P3 was classed as *Saline*, and...

Figure 3.6a. Soil colour, soil classification, and gravel content of profile P1 along toposequence A-B.

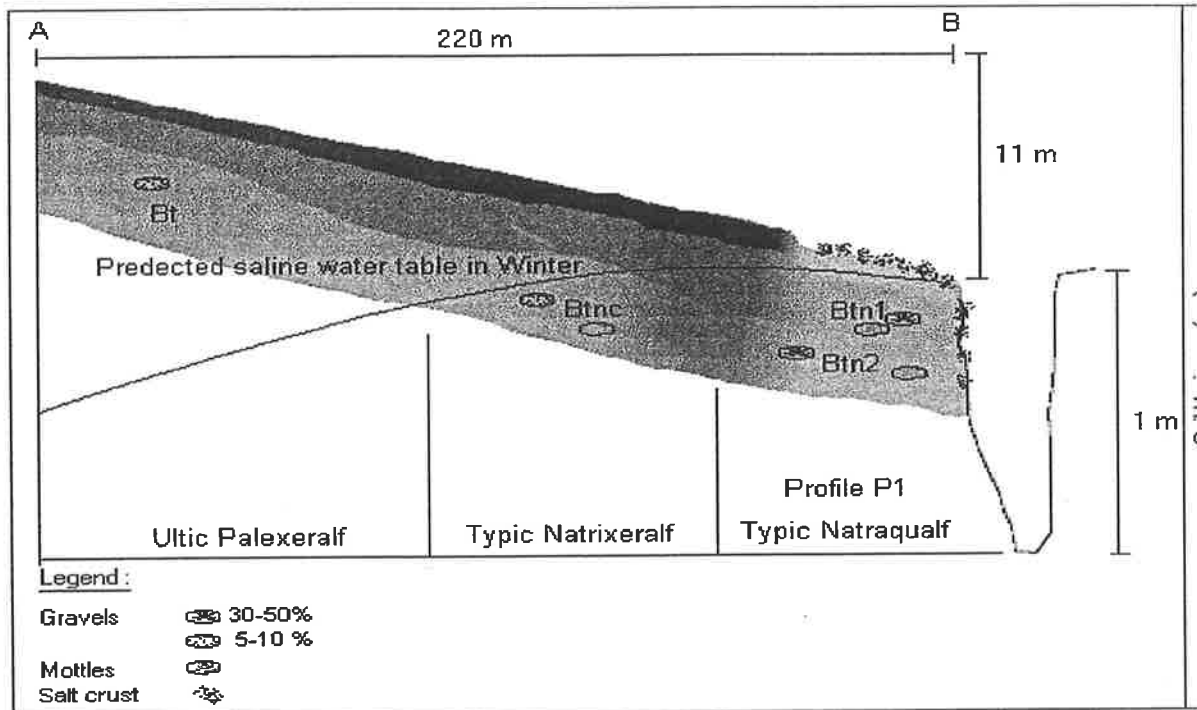
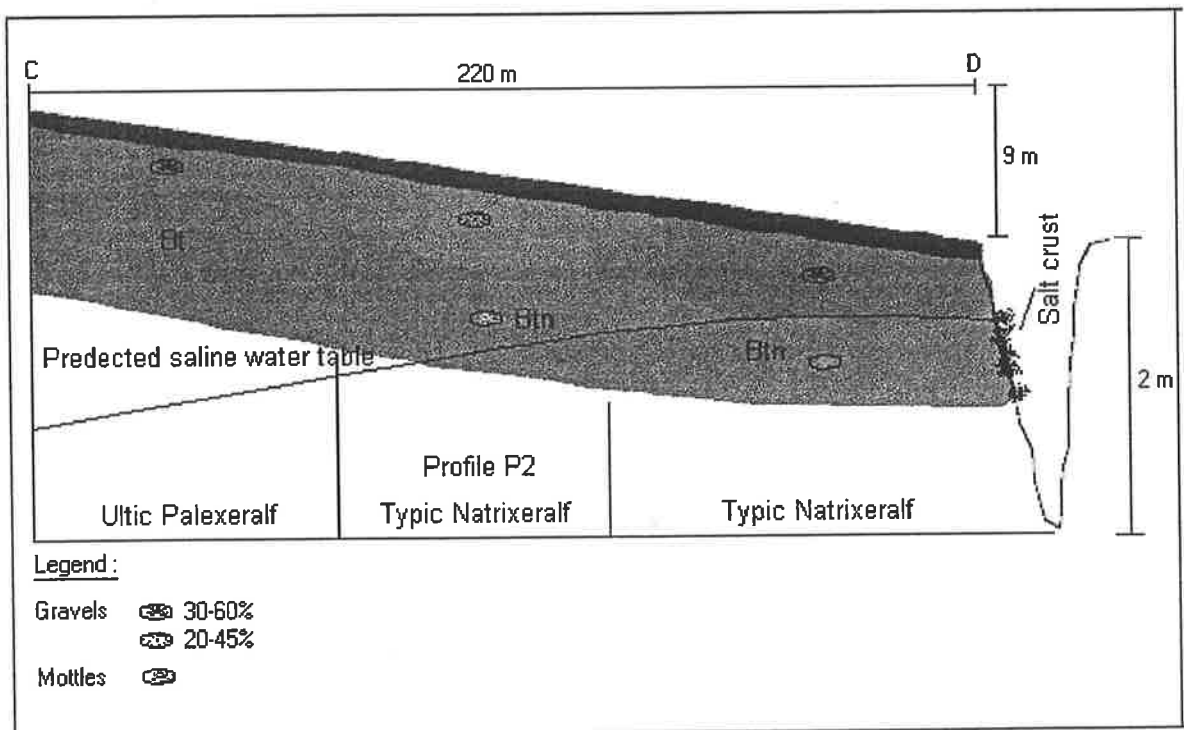


Figure 3.6b. Soil colour and soil classification of profile P2 along toposequence C-D.



- Profile PS was classed as *Saline* (in upper horizons); *Sodic-Saline* (in lower horizons).

With this extent of salinity and sodicity, a relatively high mechanical dispersibility of the clay in all these soils was expected (El-Swaify *et al.*, 1970; Rengasamy and Olsson, 1991; Chorom *et al.*, 1994). Despite the wide range of other important properties that control dispersion (eg. pH and clay content shown in Figs 3-2 and 3-3), clay dispersion generally increased with sodicity (Fig. 3.9a), albeit more so in some profiles than in others. There was also a strong correlation between ESP and soil pH (Fig. 3.9b), with the more sodic soils occurring at higher pHs. This supports the general close association between sodic soils and alkaline soils in Australia.

In this study, the degree of clay dispersibility and its effect on soil hydraulic properties must be considered in light of the appreciable variability of soil properties between and within the soil profiles (eg. EC, pH, sodicity, exchangeable cations, shown in Figs 3.2 and 3.3). The degree of clay dispersibility must also be considered in light of the general uniformity between the profiles of other important soil properties (e.g. organic carbon content, clay mineralogy, cation exchange capacity, shown in Figs 3.4 and 3.5). It is particularly important that the saturated hydraulic conductivities (Table 3.2) showed little variation between the four different profiles, yet showed enormous differences between the surface and subsurface soil horizons (Table 3.2).

3.4.2 Relationship between classified soil properties and topography

This work attempted to relate the type of soil to its position in the salt-affected landscape, and to determine the extent to which a saline water table may have influenced soil development and hydraulic properties. As indicated in Table 3.1, three subgroups of the Soil Taxonomy” were identified. Profiles P1 and PS along the foot slopes were classified as Typic Natraqualfs and Typic Natrixeralfs, respectively. Natraqualfs have

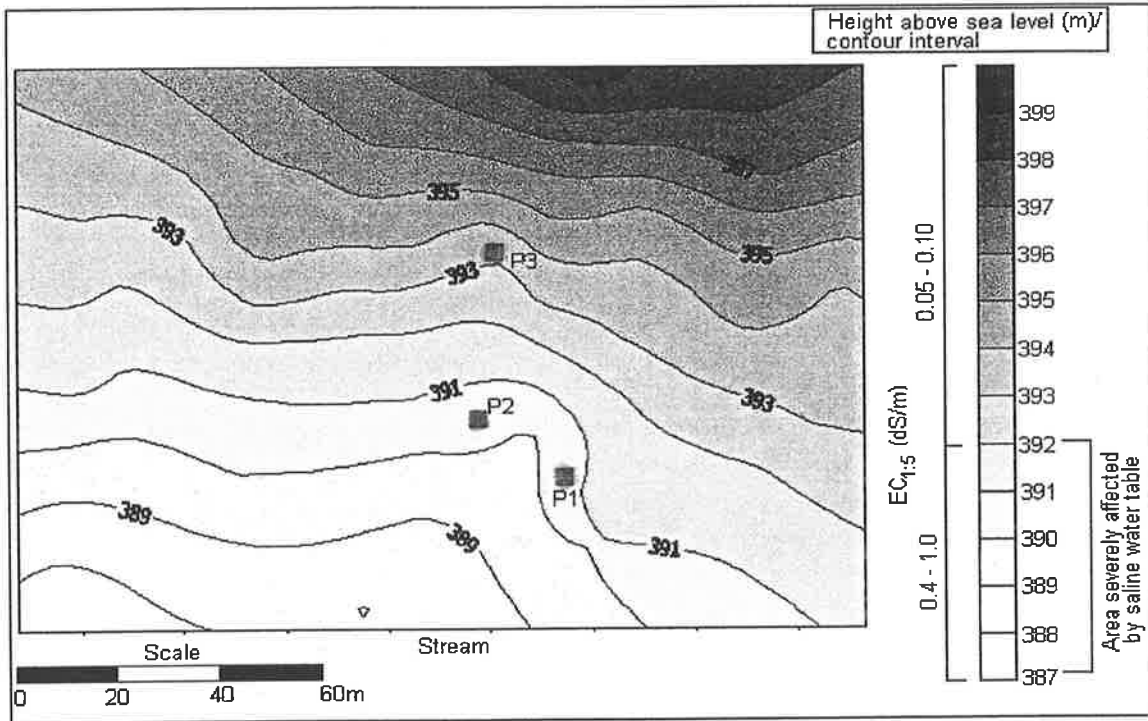


Figure 3.7. Distribution of $EC_{1.5}$ in relation to topography.

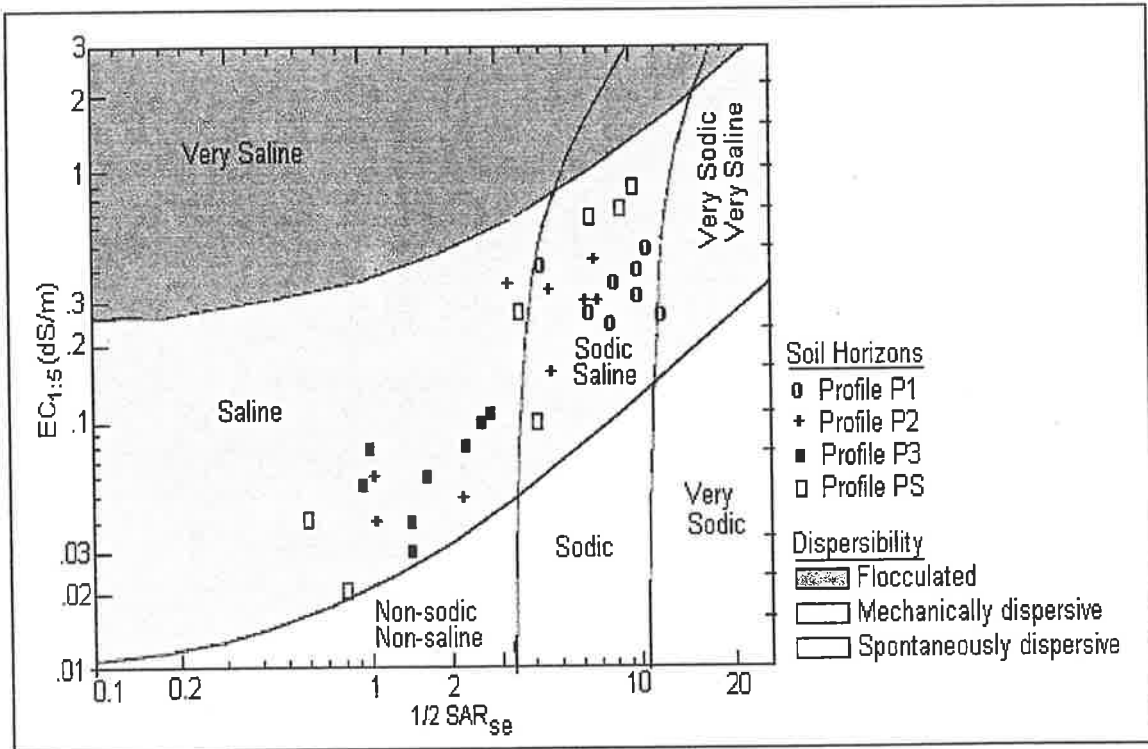


Figure 3.8. Salinity-sodicity classification of soils in the study site according to criteria of Sumner *et al.* (1998) (modified).

natric diagnostic horizons and aquic conditions, which can be attributed to the influence of a saline groundwater table (Table 3.1). Fitzpatrick *et al.* (1996) identified other Natraqualfs at this site, so this work provides some corroborative evidence. Profile P2, located on the upper slope (waning mid slope), was classified as a Typic Natrixeralf, while profile P3, in the upper waning mid slope, was classified as an Ultic Palexeralf.

The observed soil colours observed in this work support the contention of Fritsch and Fitzpatrick (1994) and Fitzpatrick *et al.* (1996) that the soils in this landscape were significantly influenced by both a saline groundwater table plus fresh surface (or perched near-surface) water. The soil colours in profile P1 in the foot slope were greenish gray (6/5G) or (5/10Y) and light gray (2.5 Y7/2) (Appendix 1), which suggests the influence of long periods of waterlogging. Significant waterlogging invariably generates reducing conditions, which generate typical gleyed soil colours. Under prolonged anaerobic conditions, red (Fe^{3+}) mottles produced under variable redox conditions change to bluish-grey and even blue (Fitzpatrick, 1971), as observed here.

In profile PS (foot slope) and profile P2 (waning middle slope), the soil colours were redder than in profile P1, suggesting conditions of better drainage or less prolonged inundation with water. More prevalent greyish brown (10 YR5/2) mottles and fewer yellowish red mottles (5YR4/6) were observed at depth in these two profiles, suggesting lower (more negative) redox conditions persisted for longer periods below the water table. The occurrence of periodic waterlogging produces mottling with colours ranging from grey matrices with red and yellow mottles to yellowish brown (Soil Survey Staff, 1999).

With respect to soil classification, the two categories of Subgroups that I identified (i.e. Aqualfs and Xeralfs) were classified based on the different soil colours found across the toposequences (caused by variable waterlogging). The Great Groups that I

identified (i.e. Natrixeralfs and Palexeralfs) were classified based on the existence or absence of sodic horizons.

Sodic horizons generally develop on parent materials containing appreciable sodium-bearing minerals (e.g. albite), or in landscapes containing saline watertables. The parent material of the sub-catchment in this study was micaschist, which contains sodium, and there is evidence from a groundwater study in an adjacent sub-catchment 500 meters away that a saline water table existed under hydrostatic pressure (e.g. Fitzpatrick *et al.*, 1996; Cox *et al.*, 1996). Thus it is likely the sodic horizons examined in the present study developed under the influence of both an invading saline water table as well as the weathering of Na-bearing parent materials (Fritsch and Fitzpatrick, 1994).

The influence of a saline ground water in profiles P1 and P2 was inferred from the high EC of the solution collected from the lysimeters installed in these profiles. In profile P1, the groundwater table can reach the surface horizon (Epz) as indicated by the high EC_{se} ($5.01 \text{ dS}\cdot\text{m}^{-1}$). In profiles P2 and PS, however, the watertable may also have reached the soil surface but only for restricted periods of time – most of the time, it stayed below ~26 cm from the soil surface. These conclusions come from the marked contrast in EC from the A1 horizons of profiles P1 and PS, suggesting the presence of a saline sulfatic water table staying longer in the Bt horizons of profile P2 and PS. Furthermore, the SARs and ESPs in profile P1 were substantially greater than in P2, suggesting the influence of a saline water table was more intensive in profile P1 (possibly because the groundwater table stayed longer) than in profile P2.

The presence of saline groundwater was also responsible for the development of soil structural features peculiar to salt-affected soils in the landscape. For example, profiles P1, P2 and PS, which were strongly affected by the saline water table, exhibited strong

medium columnar soil structure, whereas profile P3 exhibited weak subangular blocky and strong medium polyhedral soil structure. There is plenty of evidence showing that strongly developed columnar structure is associated with extensive quantities of exchangeable sodium and magnesium (Murtha 1988).

All soils in the study site have either argillic, kandic or natric horizons (Soil Survey Staff, 1999) in the subsoils with high base saturation status (>35%) and relatively low accumulation of organic carbon. The soil features (particularly argillic horizons), are common to soils found in ancient (ie. Holocene or older) landscapes (Buol *et al*, 1997).

3.4.3 Colour patterns from an aerial photograph in relation to soil properties

Field observations suggested that variations in different species of native grasses matched well with variations in tonal (colour) patterns shown in an aerial photograph of the study site taken in the summer of 1995 (Fig. 3.10). The fact that these grasses were native species (not cultivated) suggests their distribution may be related to soil properties affecting their growth. Variations in colour shown in Fig 3.10 may thus reflect variations in vegetative cover that are linked to different soil properties. A systemic grid sampling (50m x 50m) of the soils in the area indicated the distribution of vegetation was indeed related to the measured particle size distributions for the surface soils. In fact, soil texture was the only property that related to colour. For example, the red patches in Fig. 3.10 coincided with heavier textured soils, while the blued coloured patches coincided with lighter textured soils. Because heavier textures retain more water, it was postulated that this had a marked effect on survival of the natural grasses produced that year.

3.5 Conclusions

- Fluctuating saline groundwater tables in the Herrmanns sub-catchment would appear to have generated varying degrees of salt-affected soils. Using criteria of Sumner *et al*.

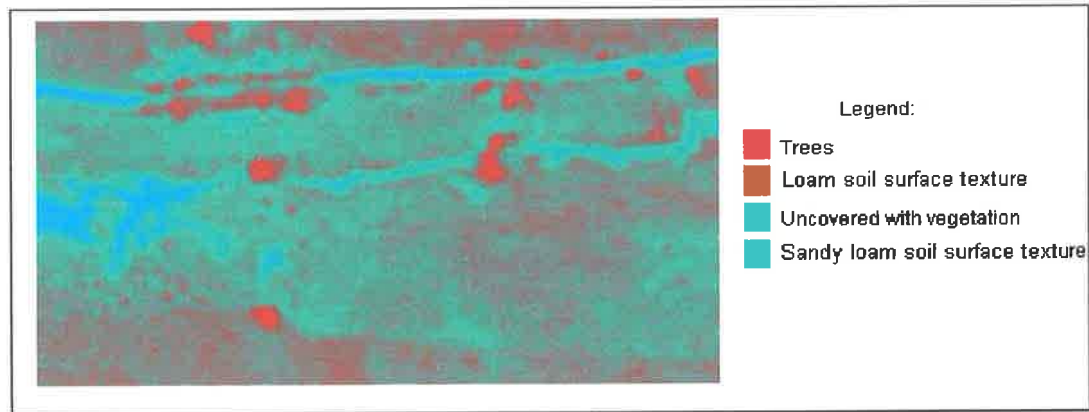


Figure 3.10. False-colour aerial photograph (1995), illustrating postulated link between soil texture (inferring soil moisture) and the distribution of vegetation observed in the year 2000.

...(1988), the four soil profiles in the study site were designated into three main classes: *Sodic-Saline*, *Saline to Sodic-Saline*, and *Saline*.

- Variations in sodicity was a major factor responsible for variations in clay dispersibility. High sodicity generally caused greater clay dispersibility.
- Soils along the foot slope were classed as Typic Natraqualfs, and they were strongly influenced by fluctuating saline groundwater tables.
- Soils in the waning mid slope were classed as Typic Natrixeralfs, and they were subject to less intensive invasions of saline groundwater tables.
- Soils in the upper waning mid slope were classed as Ultic Palexeralfs, and they were least exposed to saline groundwater tables.
- The soils on the stream bank were classed as Typic Natrixeralfs, because while they were exposed to saline groundwater (salt crystals observed on face of profile), they were well drained, and thus exposure was only short-lived and so aquic conditions did not prevail.

Chapter 4. Lysimeter, overland flows and stream sediment investigation

4.1 Introduction

The dispersibility of soil clay in the landscape and soil profiles in relation to soil properties was discussed in the previous chapter to identify potential sources of dispersed clays. The prime objective in this chapter is to investigate the mobility of the dispersed clays within the study site in an attempt to better understand the causes of erosion in the landscape.

In general, a dispersed state is a prerequisite for clays to migrate from their points of genesis. Thus it is reasonable to assume that the degree of clay dispersibility in a soil profile should correlate with its potential mobility. There are many other factors (e.g. topography and vegetation) that might affect clay transport in the landscape. Nevertheless, if external factors are constant, it should be possible to link the mobility of colloids in the landscape to their dispersibility and variations in the permeability of the soil. In this field-based study, the amount of clay migrating vertically within the profile and laterally via overland flow was quantified and linked to soil hydraulic properties and clay dispersibility. Sediment from overland flow was accounted for in this work by collecting samples of overland flow in plastic sediment traps (installed in the ground) during rainfall events. Quantification of sediment load in the stream was also carried out in order to compare the result with the amount of sediment thought to be transported from the landscape of the study site. Biddle *et al.* (1995) has carried out similar investigations, but they did not specifically relate dispersibility to the mobility of colloids.

4.2 Materials and methods

4.2.1 Lysimeter design and installation

The design of capillary wick lysimeters was similar to that of Biddle *et al.* (1995). The materials included 7 x 130 cm of glass wick, 40 mm I.D. PVC tubing of 100 cm, 8 mm

PVC plate of 30 cm x 30 cm , 10 L storage bucket, 8 mm I.D. sampling tube, 45° elbow joints and PVC glues (Figure 4.1)

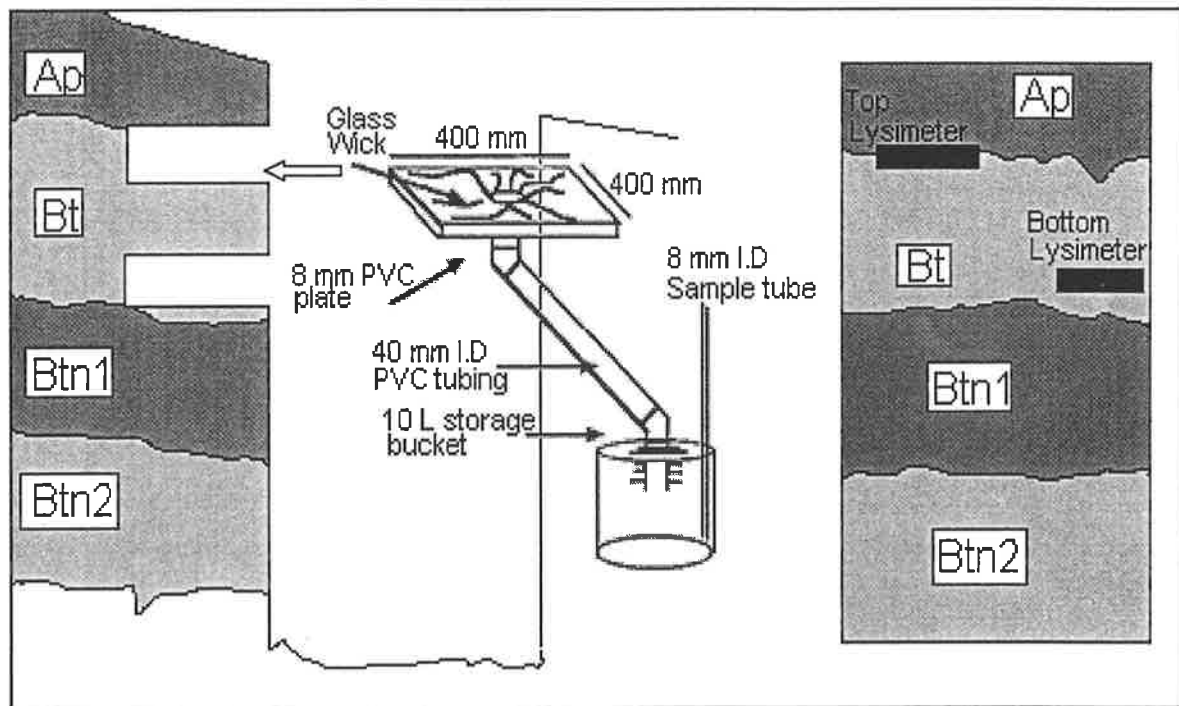


Figure 4.1. Design of wick lysimeter used in the study site and the sketch of the position lysimeter in the profile (modified from Biddle *et al.*, 1995)

Three profiles (100 cm wide, 250 cm long and 200 cm depth) representing saline, sodic and nonsaline-nonsodic areas were excavated and soil sampled for analyses. Two capillary-wick lysimeters were then installed on each side of the excavation (Figure 4.1).

The first lysimeter was installed at the bottom of the Ap horizon by inserting the lysimeter-soil junction plate (top of lysimeter) into the upslope wall of the profile that had been prepared with a slot to fit the plate (Figure 4.1). Similarly, the second lysimeter was placed at about 22 cm at the top of the Btn1 horizon (Table 4.1). Sample-storage buckets were buried at a slight angle so that the withdrawal tubing was fixed at the lowest point to allow all solutions to be extracted. To ensure that the lysimeters were firmly installed, tension against the soil was applied using adjustable supports. After both lysimeters were installed, and before adding soil to fill the profile, sample withdrawal tubing was directed

to the surface, and fixed to wooden peg. Fences were placed around the lysimeters to prevent livestock damaging the soil surface.

Table 4.1. Depth of lysimeter installations in the profiles

	Top lysimeters (cm)	Bottom lysimeters (cm)
Profile P1	10	28
Profile P2	24	42
Profile P3	14	36

Lysimeters were first sampled following rainfall events. After that, depending on weather conditions, samples were collected from the buckets approximately every two weeks, or more frequently if intense rain events occurred. Sample solutions were extracted in the field using a vacuum pump. Water samples were immediately transported to the laboratory and stored in a refrigerator until analysed.

4.2.2 Overland flow sediments and stream water sampling

The overland flow sediment collector used in this trial was a plastic container with dimensions 30 cm wide, 40 cm long and 15 cm in height. The collectors were placed in such a way that they could harvest a large amount of sediment during each rainfall event. In this investigation, the collectors were placed in small depressions. Two collectors were installed in the eroded region (RE), which was barren and the other two collectors were placed in a non-eroded region (RU), which was covered with pasture grasses.

Small pits were excavated to fit the collection containers and the up-slope wall of the pit was blanketed with a plastic sheet to avoid disturbance of the pit wall (Figure 4.2). To ensure that the plastic cover remains fixed to the soil surface and does not obstruct the flow of sediments, it was clamped against the soil surface using small metal cables.

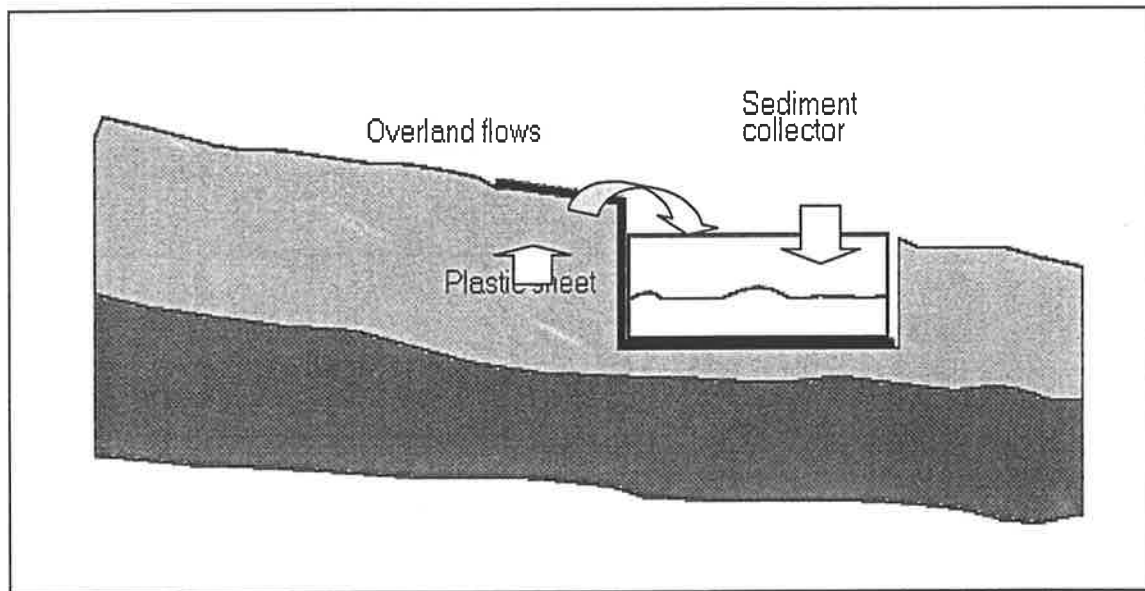


Figure 4.2. Installation of sediment collectors in the field .

Collection of sediment depended on the frequency and quantity of rainfall. However, samples were collected at least every two weeks, usually when solutions were collected in the lysimeters. Overland flow samples were analysed in the laboratory for pH and EC immediately after collection. A 500 mL sample was stored in the fridge for DOC analyses, and the remainder of the sample was filtered using 0.45 μ m filter paper for mineralogical analyses.

Collection of stream water samples was carried out after periods of rainfall at three sampling points, which were chosen randomly. In general, the randomised collection of the stream waters was considered sufficient to represent the sediment of the stream water. Walling (1984) states that in a majority of rivers, uniform distribution of solute concentration is likely because of turbulent mixing of the stream water. The sediments collected were treated similarly to the lysimeter and overland flow sediments.

4.2.3 KBr tracer to investigate lateral flow

KBr was applied to the soil surface to assess whether or not perched fresh water flows laterally over the Bt horizon. KBr was chosen because it has a high water solubility

(53.5g/100ml) (Shaw *et al.*, 2001). Prior to KBr application to the soil, the lysimeter was emptied, followed by the digging of several holes using a hand auger down to a similar depth at which the lysimeters were installed (Table 4.1). The distance from the profiles in which the lysimeters were installed was 550 cm upslope (Figure 4.3). About 20g of KBr was spread into each hole and refilled. After several rainfall events, the solution was extracted from the lysimeter and analysed for Br⁻.

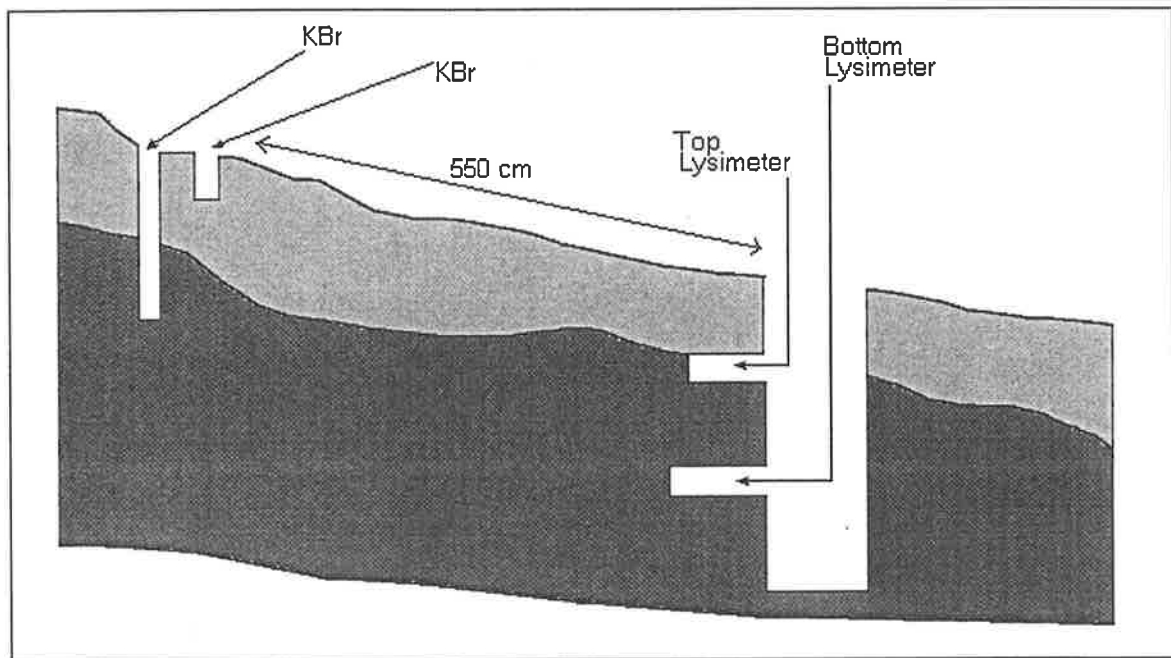


Figure 4.3. Placement of KBr at upslope locations from profiles P1, P2 and P3 lysimeters. Filtration and evaporation are two standard procedures for determining the concentration of suspended sediments (Ward, 1984). Due to the high EC of the suspensions, determination of the concentration of suspended colloids was carried out by suction filtration. Cellulose filter paper (0.45 μm) was used to filter the clays from a 200mL suspension. Afterwards, the filter paper with retained clay and blank filter paper were placed in an oven at 105°C for 24 hours. The mass of colloid was corrected by taking into account the weight of dissolved material contained in water.

The pH and EC were measured on solutions sampled from the lysimeters, as well as from overland flows and stream waters. Dissolved organic carbon was determined using the modified Walkey and Black method (Nelson and Sommers, 1982).

Preparation of clay for un-oriented powder X-ray diffraction (XRD) was carried out by filtering the solution through a 0.45 μm porosity cellulose-nitrate filter paper under vacuum. The clay retained on the filter paper was then removed and collected by washing with small amounts of distilled water. The clay collected from the suspension was then dripped onto a silica wafer for mineralogical analyses by XRD (*cf.* Chapter 3).

4.3 Results

4.3.1 Colloid concentrations, chemistry and mineralogy of lysimeter solutions

The volume of solution collected each time from the lysimeters during three consecutive periods was 10 L on average. This volume was appreciably higher than those measured on a sub-catchment at Mount Crawford by Biddle *et al.* (1995). The data indicate a trend towards more acidic and less saline condition over the period monitoring was conducted. The lysimeters in profile P2, showed slightly higher pH and EC values for the first collection compared to the second and third collections (Table 4.2).

Table 4.2. The pH and EC of solutions collected from lysimeters in profiles P1, P2 and P3.

Date of Collection	pH	DOC (ppm)	EC (dS.m ⁻¹)	pH	DOC (ppm)	EC (dS.m ⁻¹)	pH	DOC (ppm)	EC (dS.m ⁻¹)
	Profile P1			Profile P2			Profile P3		
Top lysimeter									
15-Jun-00	8.42	1.0	6.13	7.96	1.2	8.8	7.9	0.1	0.34
28-Jun-00	8.13	1.3	6.15	7.78	1.9	5.84	6.63	0.1	0.24
31-Jul-00	8.33	1.5	5.91	7.32	0.8	6.12	6.78	0.1	0.25
Bottom lysimeter									
15-Jun-00	8.35	1.7	5.84	7.95	1.7	8.65	7.4	0.1	0.28
28-Jun-00	8.13	1.7	5.9	7.18	1.6	5.77	6.97	0.2	0.25
31-Jul-00	8.12	1.7	5.76	7.54	1.3	5.94	7.31	0.2	0.26

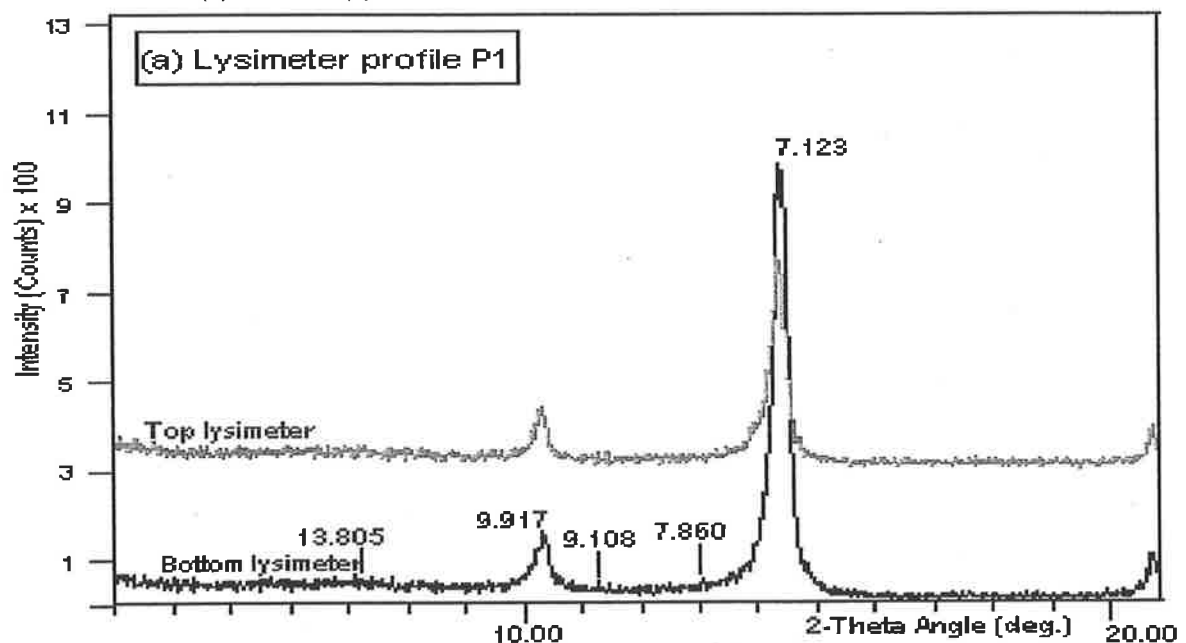
The concentration of colloid showed fluctuations with time of collection, but in general, the top lysimeters in P2 and P3 gained more clays than respective bottom lysimeters (Table 4.3). However, the bottom lysimeter of profile P1 harvested more clay than its top lysimeter (Table 4.3).

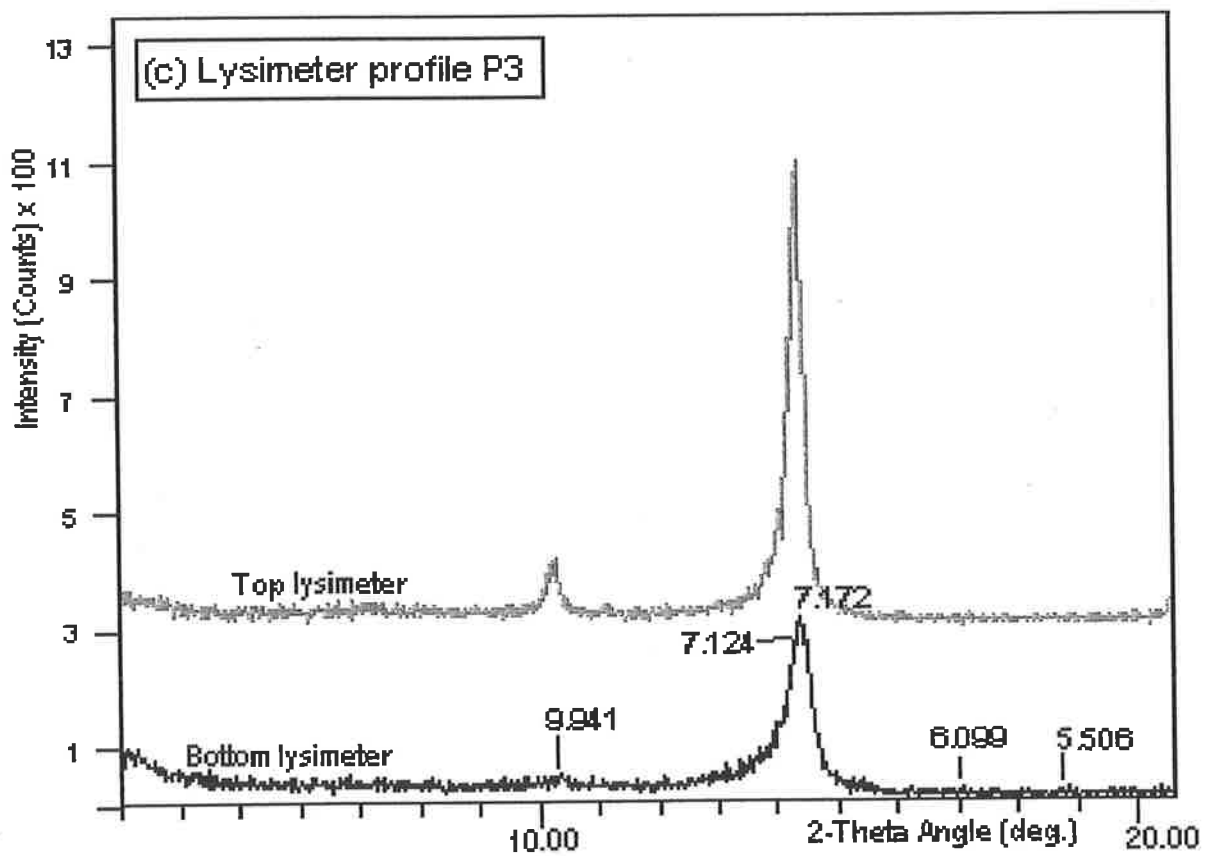
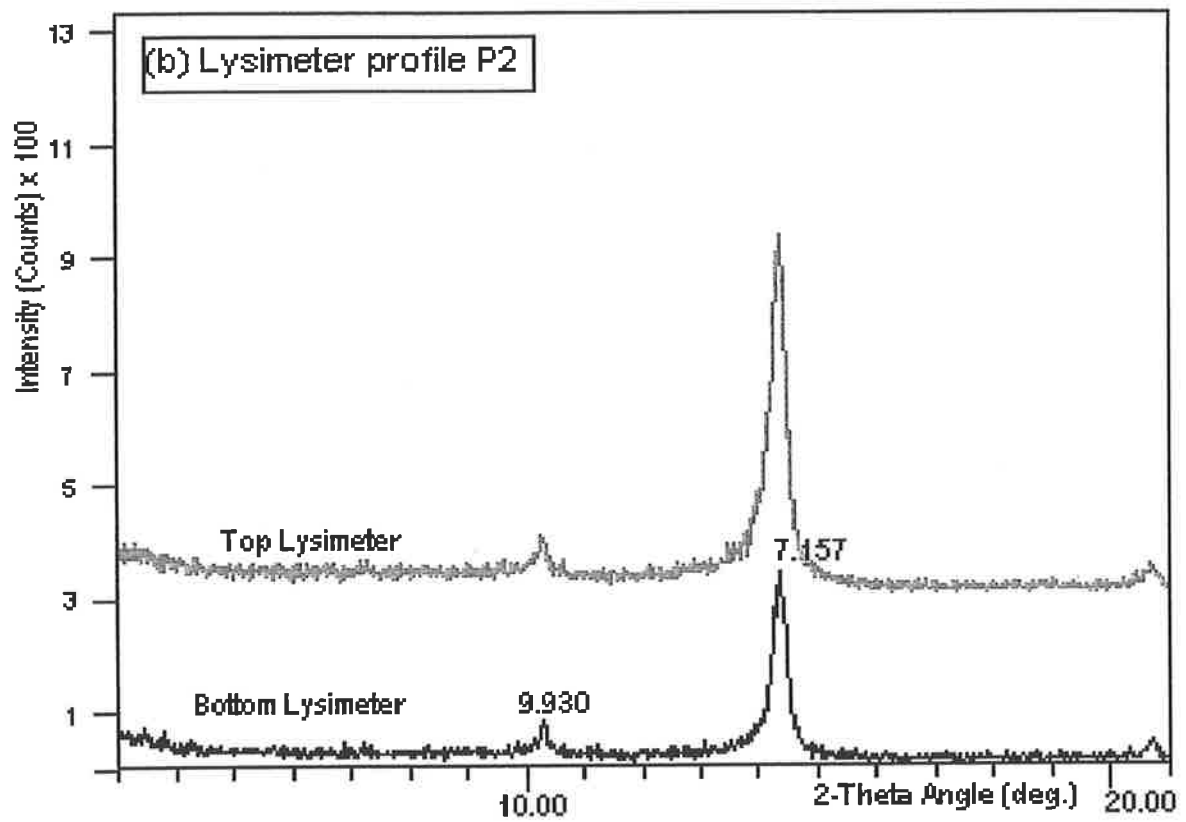
Table 4.3. Concentration of suspended colloids from lysimeters in profiles P1, P2 and P3

Collection Date	----- Total particulates (g/L) -----		
	Profile P1	Profile P2	Profile P3
Top lysimeter			
15-Jun-00	0.016	0.024	0.050
28-Jun-00	0.029	0.028	0.065
31-Jul-00	0.017	0.034	0.053
Bottom lysimeter			
15-Jun-00	0.299	0.019	0.034
28-Jun-00	0.186	0.024	0.020
31-Jul-00	0.300	0.034	0.015

The mineralogical composition of colloids collected from each lysimeter was dominated by kaolinite (Fig 4.4). The kaolinite:illite ratio was 3:1, which is similar to the layer silicate composition of all profiles where lysimeters were installed (see Chapter 3).

Figure 4.4. Powder XRD patterns of colloids collected from the lysimeters in profiles (a) P1, (b) P2 and (c) P3.





4.3.2 Lateral flow of colloids

The high concentration of Br⁻ in solution after KBr application indicated that Br⁻ was transported laterally in through-flow over the Bt horizon (Table 4.4).

Table 4.4. Br⁻ concentration in lysimeter solution before & after injection of KBr into soil

Profile	Br ⁻ (ppm) before placement	Br ⁻ (ppm) after placement
----- Top lysimeter -----		
P1	0.14	8.6
P2	0.38	8.5
P3	0.43	8.3
----- Bottom lysimeter -----		
P1	0.21	4.6
P2	0.39	7
P3	0.44	6.9

4.3.3 Overland flow and stream water sediments

In Table 4.5, the pH, EC and DOC concentrations of overland flows and stream water samples are presented. The EC of overland flow suspensions collected from the eroded region and stream samples were similar and ranged between 4.4 and 6.9 dS.m⁻¹. However, the EC values of water collected from un-eroded area were low (<0.2 dS/m) compared to the stream water. The pHs followed a similar pattern in that the stream water and overland flow in the eroded area were similar (pH ranges from 6.2 to 8.2).

Table 4.5. Chemistry of stream water and overland flows in eroded and non-eroded areas.

Date of Collection	Stream water			Date of Collection	Eroded area			Non-eroded area		
	pH	EC (dS m ⁻¹)	DOC (ppm)		pH	EC (dS m ⁻¹)	DOC (ppm)	pH	EC (dS m ⁻¹)	DOC (ppm)
23-Jun-00	7.7	6.9	2.1	15-Jun-00	8.2	6.4	1.6	6.2	0.1	0.4
28-Jun-00	7.8	6.86	1.7	28-Jun-00	6.6	4.4	1.9	6.2	0.2	0.3
31-Jul-00	6.8	6.4	1.9	15-Jun-00	8.1	6.1	2.2	6.3	0.2	0.3
				28-Jun-00	7.6	6.2	2.8	6.2	0.1	0.3

In general, the mineralogical composition of colloids in all the overland flow suspensions were similar (i.e. eroded region, un-eroded region, and stream water); being kaolinite and illite in the proportion of 3:1 (Figure 4.5). This mineralogical composition is remarkably similar in composition to the clay fraction in the soils and suspensions collected from the lysimeters.

The concentration of colloids collected from stream waters and from overland flows is presented in Table 4.6. As might be expected, the eroded areas produced the highest colloid concentrations, followed by the stream water samples. Colloid concentrations were an order of magnitude smaller in the non-eroded area.

Figure 4.5. XRD patterns of colloids collected from stream water samples, and overland flows.

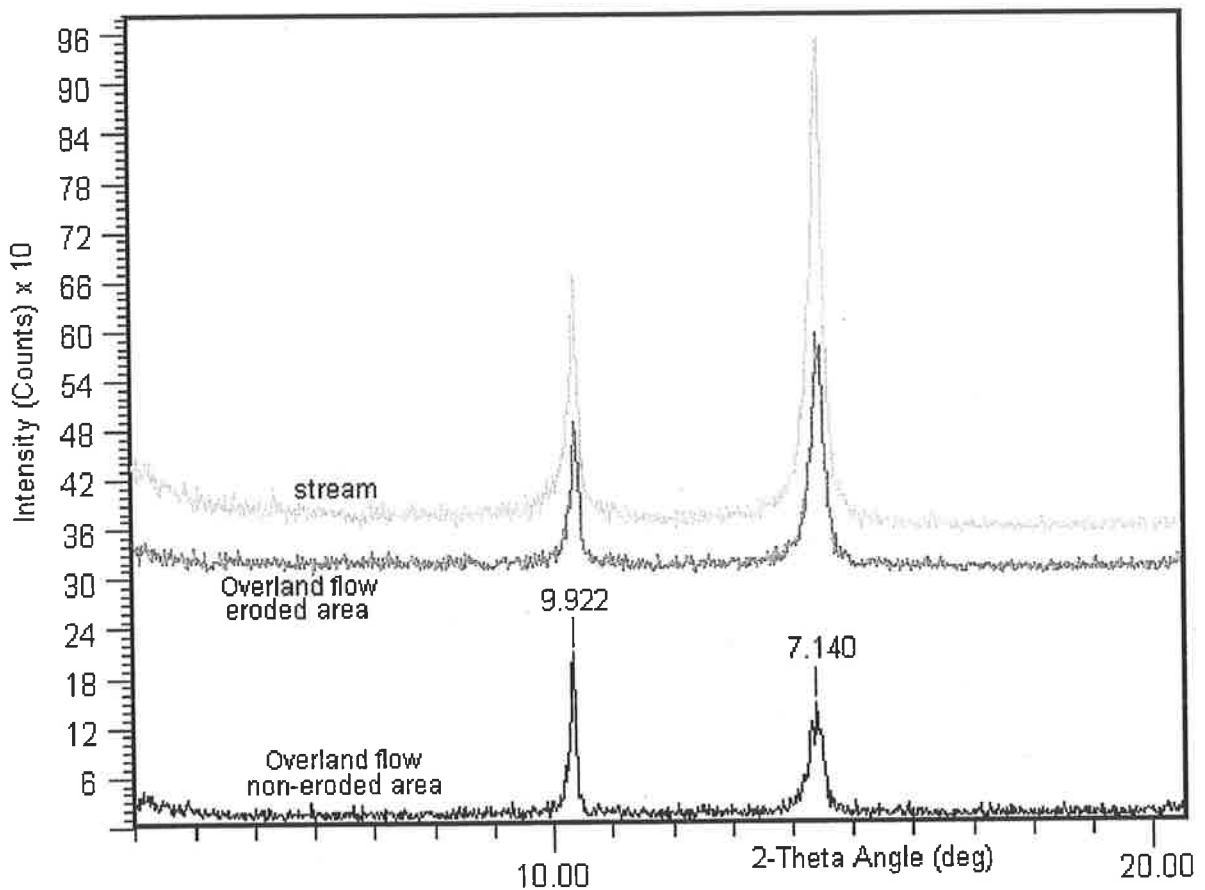


Table 4.6. Concentration of colloids collected from stream water and overland flows in eroded and non-eroded areas

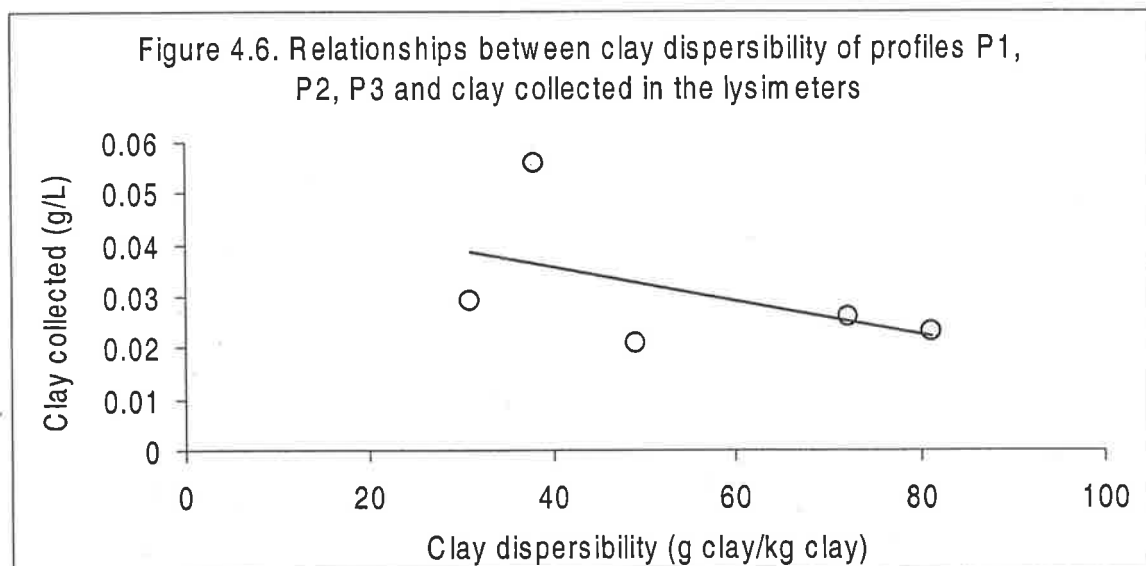
Stream water		Eroded area	Non-eroded area
Date of Sample Collection	Total particulates (g L ⁻¹)		
		Sediment collector 1	
15-Jun-00	-	0.425	0.031
23-Jun-00	0.237	-	-
28-Jun-00	0.121	0.26	0.21
31-Jul-00	0.093	Sediment collector 2	
15-Jun-00	-	0.393	0.034
28-Jun-00	-	0.267	0.022

4.4 Discussion

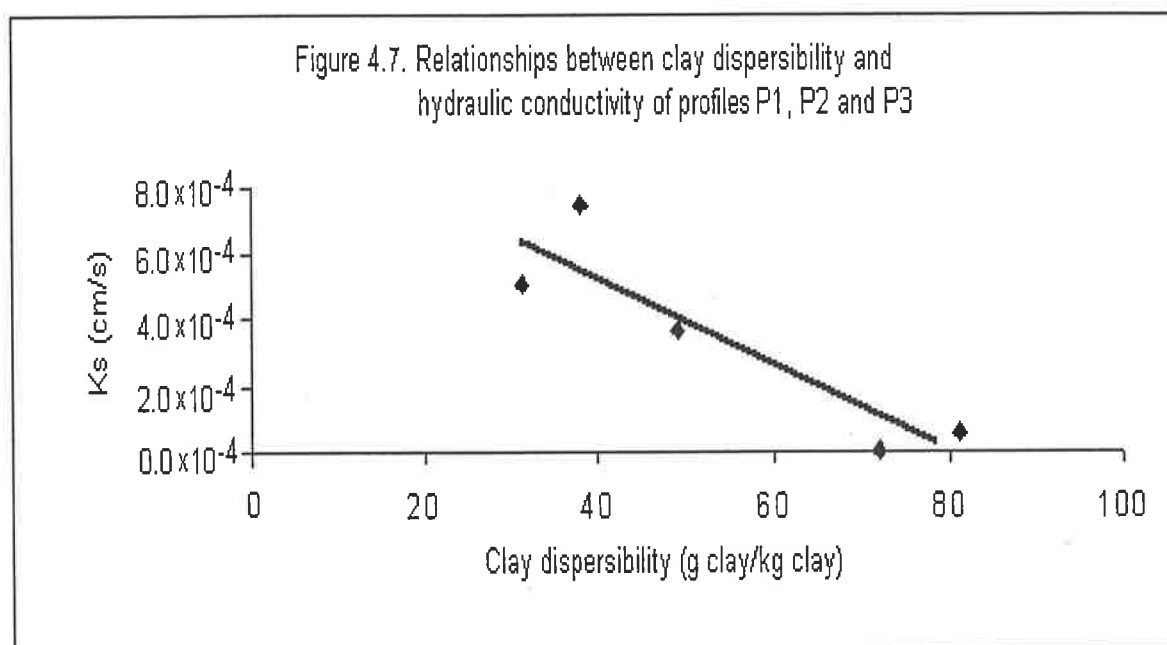
4.4.1 Dispersibility and migration of the clays

The following questions were considered in this chapter. (i) “Does high clay dispersibility always coincide with high clay migration?”, and (ii) “Does clay migrate vertically down the profile or laterally over Bt horizon?”.

Results generally indicated a negative correlation existed between dispersibility of clay and the amount of clay collected in the lysimeters of profiles P1, P2, P3 (excluding subsurface lysimeter of profile P1) (Fig 4.6). This suggested the downward movement of colloids might be less intensive in a dispersive soil compared to a non dispersive one.



Similar non-positive correlations have been reported by Churchman and Weissman (1995) who investigated the relationships between dispersibility (using Emerson test) and mobility of clays in the laboratory. Chorom *et al.* (1994) reported that minimal vertical mobility of dispersive clays could be caused by filling or plugging of pores (particularly micropores) in the soil matrix, leading to restricted water movement and thus downward transport of dispersible clay. This concept is supported by the relationship shown between clay dispersibility and hydraulic conductivity (Fig. 4.7), in which greater clay dispersibility correlated with reduced hydraulic conductivity. The presence of cracks and biopores in the soil profile, however, might allow dispersed clay to become mobile, an idea supported by the large collection of colloids in the subsoil of profile P1 (Fig 4.8).



The role of macropores (i.e. developed by old root channels and gravel content) in triggering clay illuviation has been recognised by Bartelli and Odell (1960). Goss *et al.* (1973) also revealed that the existence of large pores is a necessity for the mobile clay to migrate. Thus, even flocculated clays could be transported downward if macropores exist. In the presence of macropores, clay transportation down to the lower horizons could be even easier if the soil was dominated with dispersible clay such as in profile P1.

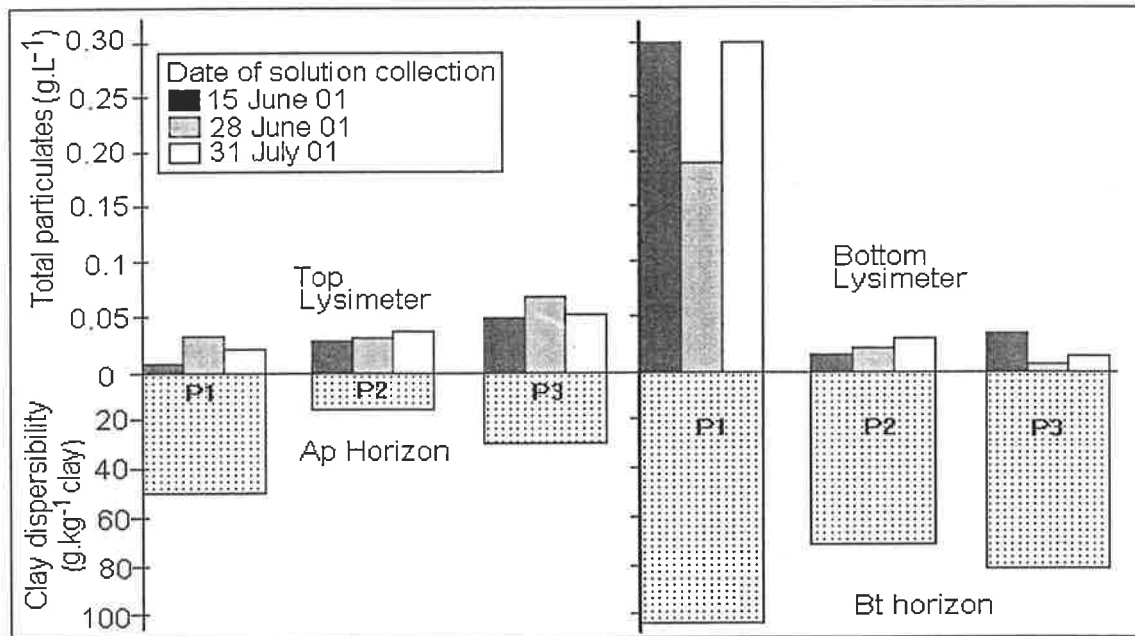
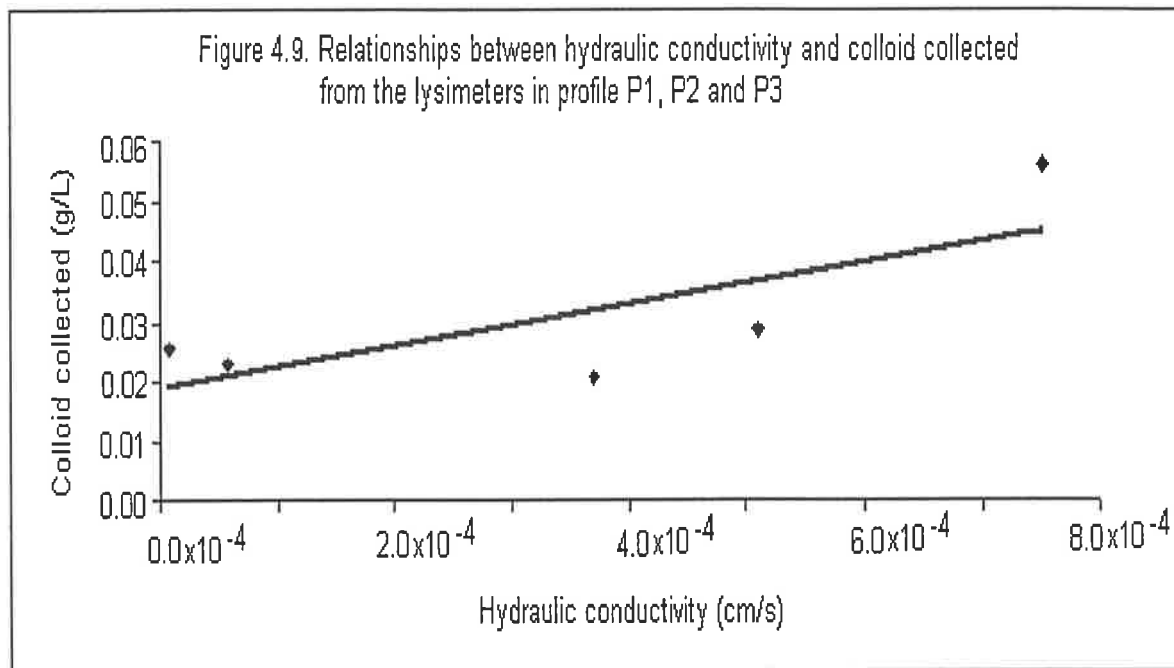


Figure 4.8. Clay dispersibility of the soils in profiles P1, P2, P3 and the concentration of colloids of lysimeter solutions

The fact that the presence of macropores is critical for the vertical transport of clay colloids, is demonstrated by comparing clay collected in the lysimeters in surface horizons and clay collected in the Bt horizons. Lysimeters in profiles P2 and P3 harvested relatively larger amounts of clay from soil surface horizons than from subsurface horizons, despite the higher dispersibility of the subsoils compared to surface horizons (Figure 4.8). In surface horizons, the existence of macropores, which are always accompanied by a coarse texture, small animal burrows and root channels, might explain why the hydraulic conductivity in the surface horizon was higher compared to the lower horizons (Table 3.2). Fig. 4.9 showed that the large quantity of colloid collected in the lysimeter is correlated with high hydraulic conductivity. Again, high hydraulic conductivity is always attributable with the presence appreciable amount of macropores in soil matrix, therefore, irrespective of the dispersibility of clay, clay in surface horizons were much easier to transport vertically to subsurface horizons. Several investigators (e.g. D'hoore, 1974; Dixit, 1978) explicitly defined dispersible clays as mobile clays, and that surface horizons were always found to have more mobile clays than subsurface horizons.

The data (i.e. K_s , colloid collected in the lysimeter and clay dispersibility) of the subsurface horizon of profile P1, were excluded in Figures 4.6, 4.7 and 4.9. The reason is the measured K_s of profile P1 (Table 3.2) was not representative of the real dynamic K_s ,



value for profile P1. Due to difficulties preparing bore-holes in the gravelly area of P1 for K_s measurements, values for profile P1 were not obtained from the area near the lysimeter. Therefore, the real K_s value for profile P1, which had high proportion of gravel (Appendix 1) might reasonably be expected to be greater than the K_s shown in Table 3.2..

Besides the clay migrating vertically (as shown in the lysimeter experiments), the clay in the study site might be partly transported laterally down the sequences. The results of the bromide tracer investigation showed that there were large differences in Br^- content before and after KBr application to the soil (Table 4.4). This indicates that solute transport flows laterally down the soil sequence over Bt horizon. Due to the reduction in saturated hydraulic conductivity in the Bt horizon, it is highly likely that solute preferentially flows over Bt horizon (i.e. through flow). It remained unclear, however, whether the bromide solution collected in the lysimeter in the Bt horizon (Figure 4.3) also originated from

lateral flow over Bt horizon, or Br^- solution derived from lateral flow of the solution through the matrix of the Bt horizon. However, because the K_s of the Bt horizon ranged between very low and extremely low (Table 3.2), it was likely that the Br^- solution in the lysimeter of the Bt horizon originated from flow over the Bt horizon. Deeks *et al.* (1999) who conducted a similar Br^- tracer investigation reported that small concentrations of Br^- tracer (observed in the lateral direction of the Bt horizon), could be an indication of minor lateral migration of solutes. In the field, the slow vertical and lateral movement through the Bt horizon, was indicated by the perched fresh water table (e.g. observed in profile P3), which remained for 2 or 3 days after a rain storm event.

The Br^- tracer experiment is only able to characterise the flow of water, and not the flow of suspended sediment. However, the flow of water may give some indication of the flow of suspended sediments if the diameter of suspended sediment in the solution is smaller than the diameter of pores of the soil matrix. In soils with clayey Bt horizons, with micropores, it will be difficult for the suspended sediment to reach the stream water, because the soil matrix is able to act as a fine filter for suspended sediments. Pilgrim and Huff (1983) reported that suspended colloids in subsurface flow can be retained in soils where the macropores in which the flow was occurring, narrowed or terminated.

Field investigation indicated that the surface soil horizons generally had sandy textures, more organic carbon, more gravel, and more root channels and animal burrows than in the subsurface horizons (Appendices 1, 2, 3 and 4). These macropores allow greater preferential flow of water and colloids, and so the lateral movement of suspended sediments in the soil surface horizons in this study area, could follow, in general, the pattern proposed by Fitzpatrick *et al.* (1994a) as illustrated in Figure 4.10.

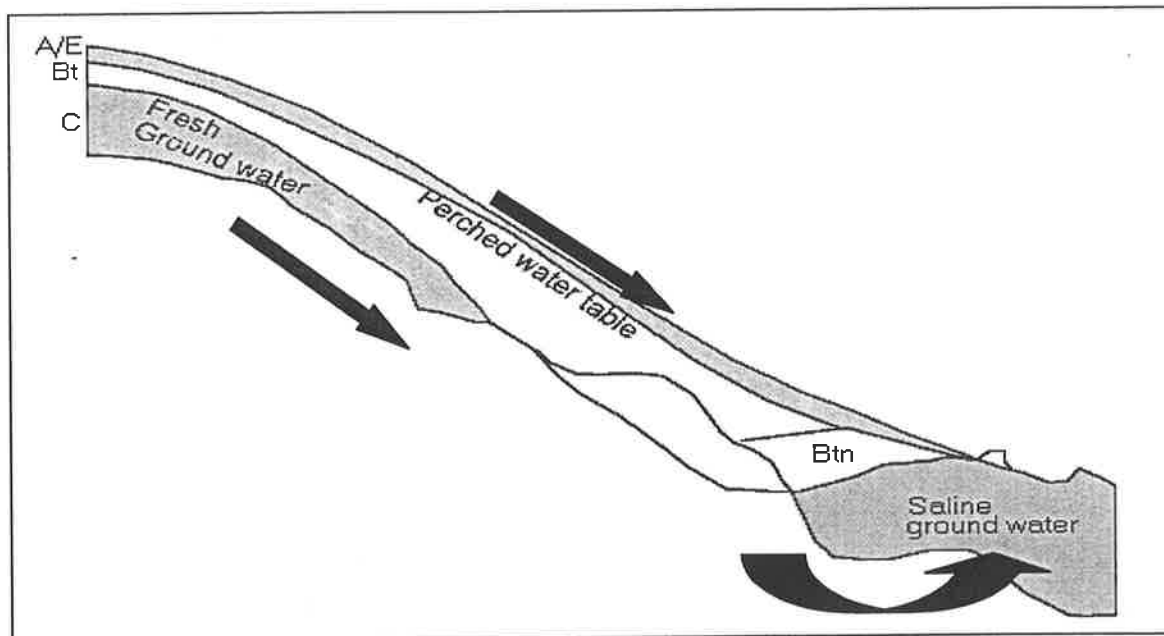


Figure 4.10. Three water flow systems in the Herrmanns subcatchment, Mt Lofty Ranges (from Fitzpatrick, 1994a)

4.4.2 Erosion

The question to be answered in this experiment was; "is there an increase in the eroded area over the last 50 years at the Herrmanns sub-catchment?". The comparison of two sets of aerial photographs, taken at two different times, strongly demonstrated that over the last 50 years major changes had taken place in the study site (Figure 4.11). The creek, which was unidentified 50 years ago, currently is more than 2m in depth in some parts (Plates 4 and 5). The land owner has indicated that the creek has only been in existence for the last two decades. However, the aerial photographs, surprisingly, also indicated that there is a decrease in the area subjected to surface erosion, suggesting that the intensity of the present erosion (at least sheet erosion) might be lower than that compared to 50 years ago (Figure 4.11 and 4.12).

The light pattern on the panchromatic black and white aerial photograph is due to high reflectance, and is associated with a dry soil surface that is bare or eroded. Severely eroded areas were interpreted from the lighter pattern due to high reflectance associated

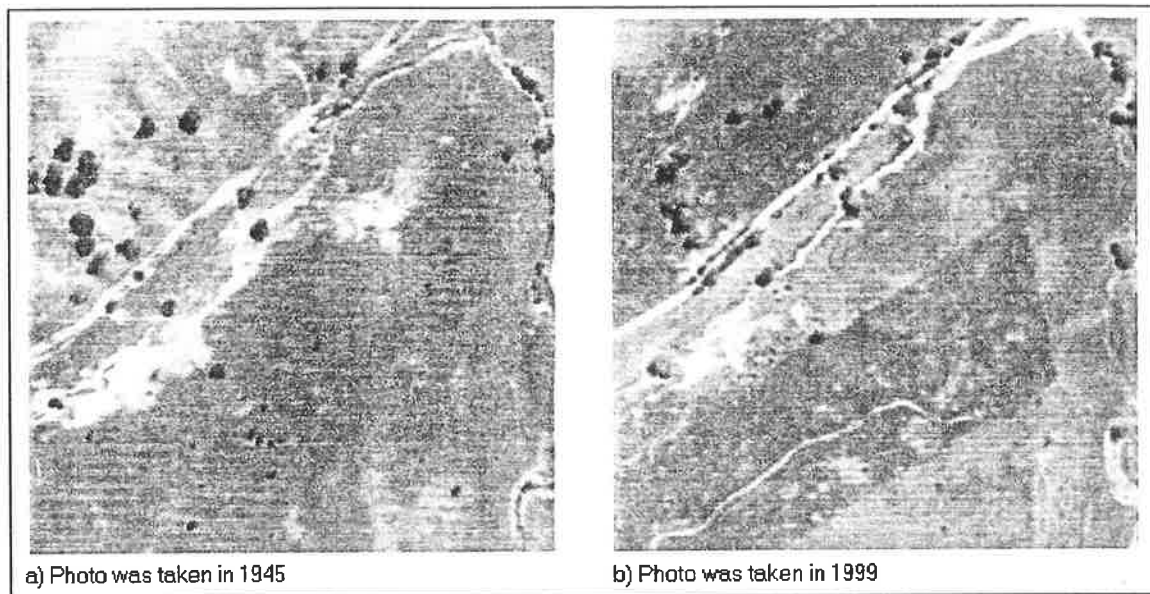


Figure 4.11. Aerial photographs of the study site taken at two different times (1945 and 1999).

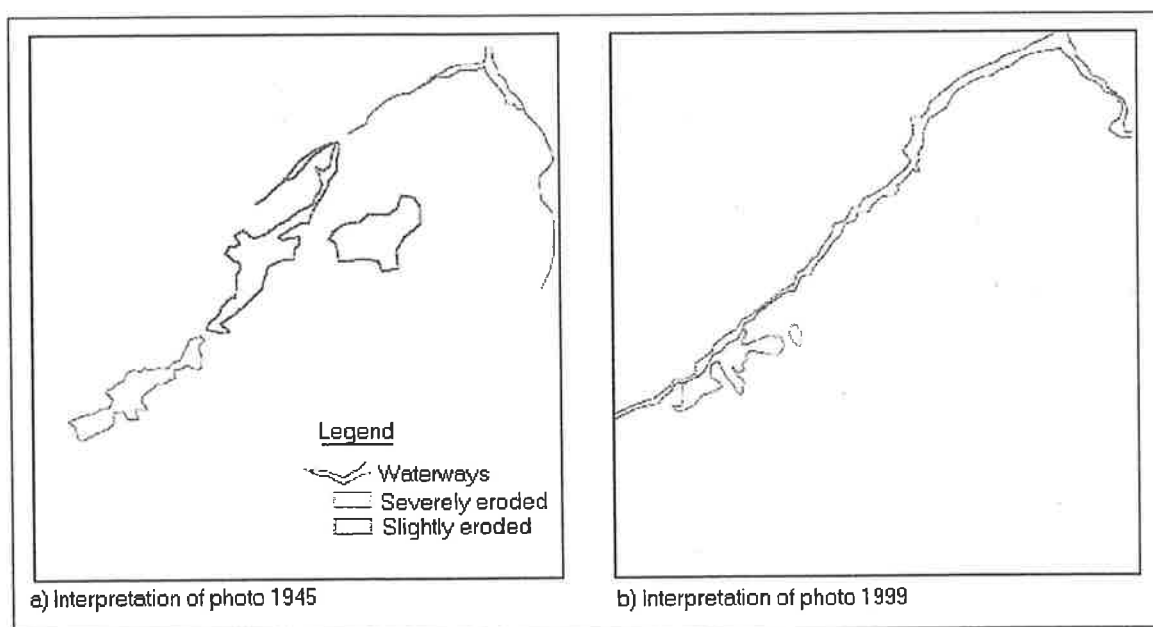


Figure 4.12. Interpretation of aerial photographs showing the intensity and distribution of erosion in 1945 and 1999.

with the presence of the exposed quartz or gravel. Those areas covered by vegetation (mainly grasses) were recognized from low reflectance and thus appeared dark on the aerial photographs (Yanda, 2000). The creek gave a light pattern, which indicated that it was dry (summer) when the photograph was taken.

The amount of sediment transported via overland flow supported the results observed in the aerial photograph that sheet erosion might be lower in more recent times compared to 50 years ago. Where the particulate concentration in the eroded region was 0.425 g L^{-1} (Table 4.6), the maximum clay that could be lost or transported from the landscape was calculated to be $2.55 \text{ t h}^{-1} \text{ y}^{-1}$ (assuming 100 % of the annual 600 mm rainfall was lost from the landscape as run off). In the uneroded area with a concentration of particulate in the overland flow being only 0.034 g L^{-1} , the maximum amount of clay potentially lost from the landscape should be much lower ($\sim 204 \text{ kg h}^{-1} \text{ y}^{-1}$). In reality, 100% of rainfall lost via run off is impossible to achieve. Therefore, the amount of clay lost via overland flow should be lower than that given in the above calculation. Soil loss of $2.55 \text{ t h}^{-1} \text{ y}^{-1}$ appears to be the normal erosion loss in this area, if compared to soil loss via water erosion. On average, soil erosion in Asia, South America and Africa is 6.1, 7.0 and $7.2 \text{ t h}^{-1} \text{ y}^{-1}$, respectively (Zachar, 1982).

Under field conditions, the above calculation could be true (explanation below), but historically the amount of clay transported out from the landscape could have been much greater than at present. The area is severely eroded as evidenced by the complete loss of the Ap horizon in the eroded area (Appendix, Plates 1 and 2).

The possible explanation for the reduction in erosion (particularly sheet erosion) intensity in the clay dispersive region or eroded area (profile P1) is probably the existence of a large drainage channel perpendicular to the toposequence. The channel is effectively reducing the speed of concentrated flow, which is a powerful factor in promoting erosion (Jiongxin, 1999). In the past (around 1945), in the absence of the channel, the amount of water that flowed over the landscape to the eroded area would have been very high, because the position of the eroded area was in the foot slope. With the appreciably higher amount of water that flowed across the landscape compared to the current flows in the

channel, soil loss from the landscape could be considerably greater than predicted from the above calculation. In addition, the greater volume of water that flowed over the landscape historically would have had greater power to disperse and transport the clay, and make the amount of clay loss would be even greater. The existence of the current channel is the reason why the intensity of current surface erosion (as shown in the interpretation of aerial photograph) is lower compared to the intensity of past erosion.

Overland flow investigations in this study demonstrated the appreciable concentration of suspended sediments collected from the sodic eroded area, which was scalded compared with the non-eroded area with dense grass cover (Table 4.6). For the two events of collection, the sediment concentration differed markedly between the first and the second collection.

Owing to the combination of highly dispersive soil and lack of vegetation in the area, it was not surprising the suspended overland flow sediment collected from sodic eroded area was substantially higher compared to the non-eroded area. The determining factors (i.e. high clay dispersibility or bare land) that promote transportation of the clays in the eroded area remained inconclusive. However, low concentration of suspended sediment collected from the un-eroded area, which had the dispersibility status as high as (even higher) than the dispersibility level of eroded area (profile P1), might indicate that that clay dispersibility only provided a small contribution in intensifying clay migration in the study site. The investigation by Pilgrim and Huff (1983) strongly revealed that raindrop impacts play a major role in mobilizing sediments; on the contrary dispersibility of clays does not seem to contribute greatly. In the field there are presumably no clays that are absolutely flocculated (un-dispersed), because the dispersibility of the clays could be controlled either by internal factors (e.g. mineralogical composition, pH, EC, sodicity, organic carbon, etc.) and external factors (level of mechanical stress force applied to clay).

Chittleborough (1982), reported that the amount of clay extracted increased dramatically with the increase in time of shaking. The result of this investigation also implies that the external factors (i.e. mechanical power of rain drop and water flow) play a major role in promoting the mobility of soil, irrespective of the dispersibility of the clay. Therefore, the negative effect of soil sodicity (e.g. in promoting soil erosion) probably can be controlled by controlling the external factors that influence the dispersibility.

Because clay dispersibility was not the dominant factor in triggering clay transport at the study site, the severe water erosion in the past might have been initiated by the exposure of land by land clearing or intensive use of the land or other factors. In the non-eroded area it has been protected from water erosion because of the dense vegetation cover. Inbar (1998) reported the effectiveness of vegetation in protecting the land from erosion. The effect of vegetation in reducing surface erosion could be due to the ability of a vegetation cover to protect the soil surface from raindrop impact. In addition, vegetation has the ability to slow the speed of surface water flow. However, vegetation alone can not effectively reduce the speed of concentrated flows, which is a powerful factor in promoting erosion (Jiongxin, 1999). Therefore, the combination of dense vegetation and the drainage channel that currently exists, might be the other reason for the reduction of intensity of eroded area in the study site.

Because the dispersive-unvegetative area contributed a high amount of sediment to the stream water, one could expect that the concentration of suspended sediment to the stream water would have to be similar in concentration to the sediment in the eroded area. However, the suspended sediment harvested from the stream water was far less than that in the eroded area, but greater compared to the overland flow sediment collected from the non-eroded area (Table 4.6). This suggests that the degraded areas (eroded areas), which

would be the major sediment contributors to the stream, were not wide spread in the sub-catchment.

4.5 Conclusions

- Highly dispersive clay is potentially mobile if external factors (e.g. porosity) are amenable (profile P1). However, the vertical movement of highly dispersive clay could be lower than expected because the propensity of this clay to fill soil micropores, a process that would lead to restricted colloid transport (Profiles P2 and P3).
- The majority of the clay in the study site is transported from the landscape via overland water flow, and part of the clay is potentially transported out of the system via through-flow (over Bt horizon).
- In profiles P2 and P3, there is an increase in clay dispersibility, accompanied with a decrease in K_s . This reduces the potential mobility of dispersible clay to move downwards to lower horizons. In profile P1, the trend was different because this profile contains large amounts of macropores because of the high gravel contents.
- In the Herrmanns subcatchment, the present intensity of sheet erosion is less than it was 50 years ago, in spite of high clay dispersibility. However, the present intensity of gully erosion along the stream banks is very much greater than it was in the past.
- Concentration of suspended sediments collected from the stream in winter (2000) was low compared to that collected from overland flow in the eroded/dispersive area.
- The dominant source of sediment in the stream water in the Herrmanns sub-catchment may not be coming from the soil surfaces, but from severe erosion of stream banks.

Chapter 5. General conclusion and land degradation model

5.1 Introduction

The results of the present study strongly support the previous studies (Fritsch and Fitzpatrick, 1994; Cox et al., 1996; Fitzpatrick *et al.*, 1996, 1999) that all the soils in the foot slope positions have been affected by a saline groundwater table. Therefore, the saline water-table-affected-soils in the landscape are all potentially to become sodic and degraded (i.e. erosion). However, only parts of the landscape have become sodic (in all parts of their soil horizons) and degraded, whereas the other soils remain non-sodic (in the upper horizons) and non-degraded.

Fritsch and Fitzpatrick (1994) and Fitzpatrick *et al.* (1996) have developed conceptual soil-water-landscape models that explain how solonization (leading to sodicity) and salinization occurred in the foot slope positions in the Herrmanns sub-catchment

(particularly in the eroded area). However, the model does not attempt to explain why soil layers in specific parts of the footslope are either saline or sodic or saline-sodic.

This chapter intends to integrate and expand the Fritsch-Fitzpatrick model with a more detailed model in order to explain the factors responsible for the degraded-non-degraded phenomenon occurring in the foot slopes of the Herrmanns sub-catchment. In this chapter it is also intended that the effects of saline water table in influencing the distribution of soil properties along the foot slope of the toposequence be explained. The integration between the spatial distribution in soil properties with the current physical condition of the landscape (i.e topography, vegetation cover, saline water table level), as well as soil conservation practices that have been applied, will be employed to explain the dynamic movement of clay in the Herrmanns sub-catchment.

5.2 Clay movement and model of land degradation in the Herrmanns sub-catchment

There are several possible modes of colloid transport in the Herrmann subcatchment. The clay could migrate vertically down to lower horizons, migrate with through-flow of water (lateral flow) over the Bt horizon or lateral subsurface flow of water, and/or be transported via surface run-off. As a result of the variability of soil properties (i.e. sodicity, clay dispersibility, gravel content) and physical state of the landscape (i.e. intensity of vegetation cover in the soils, drainage patterns of the landscape and topographic condition) the quantity and mode of transport of clay varies greatly across the landscape.

High sodicity levels in soils in eroded areas are responsible for the highly dispersible clay. High gravel contents in soils promote clay transportation vertically to the subsurface horizons. In the surface of the eroded areas, a combination of highly dispersible clay, concentration of water flow in particular areas and lack of vegetation cover have induced appreciable amounts of clay transportation via overland flow.

In non-eroded areas, clay of low dispersion and high vegetation cover may be the factors that have prevented clay transportation via overland flow. Vertical movement of clay to lower horizons also occurs in non-eroded areas but with smaller quantities compared to the vertical movement of clay in the eroded area. Low hydraulic conductivity caused by lack of macroporosity in this area was considered to be responsible for this phenomenon.

Figure 5.4 is a 3D aerial photograph of the Herrmanns sub-catchment showing the eroded (P1) and non-eroded (P2) area. As discussed in Chapter 5, the data indicates that both the eroded (P1) and non-eroded (P2) areas have been subjected to saline groundwater tables, suggesting that both regions may become sodic and degraded. However, as a result of the physical character of the landscape (particularly the drainage patterns), region P1

received flow of perched water more extensively than region P2, which in turn generated the differentiation in sodicity level of the soil between the two regions. Presumably, at the beginning, the saline water table that reached the soil surface in region P1 initiated the destruction of vegetation sensitive to salinity. Therefore, a combination of high sodicity of soil, massive surface run-off and lack of vegetation cover in region P1 initiated intensive clay transport out of the landscape. At the present, surface (Ap) horizon and vegetation in region P1 has completely disappeared, stimulating an increase in upward movement (evaporation) of the saline water table to the soil surface and creating salt crusts at the soil surface with EC_{se} of 5 dS/m.

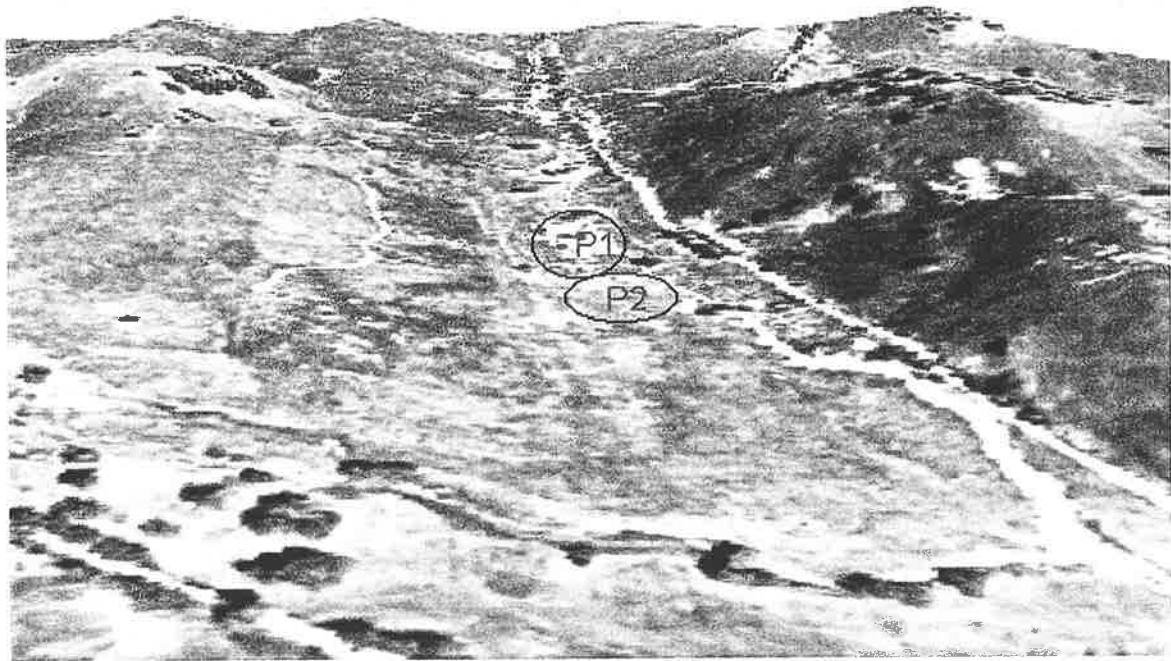


Figure 5.1. Three dimensional view showing the position of the eroded (P1) and non-eroded (P2) areas in the study site, in relation to the landscape of the sub-catchment (modified from Fitzpatrick *et al.*, 1999).

The following model (Figure 5.2) was integrated with the Fritsch and Fitzpatrick (1994) model to explain why land degradation only partly occurred in the sub-catchment.

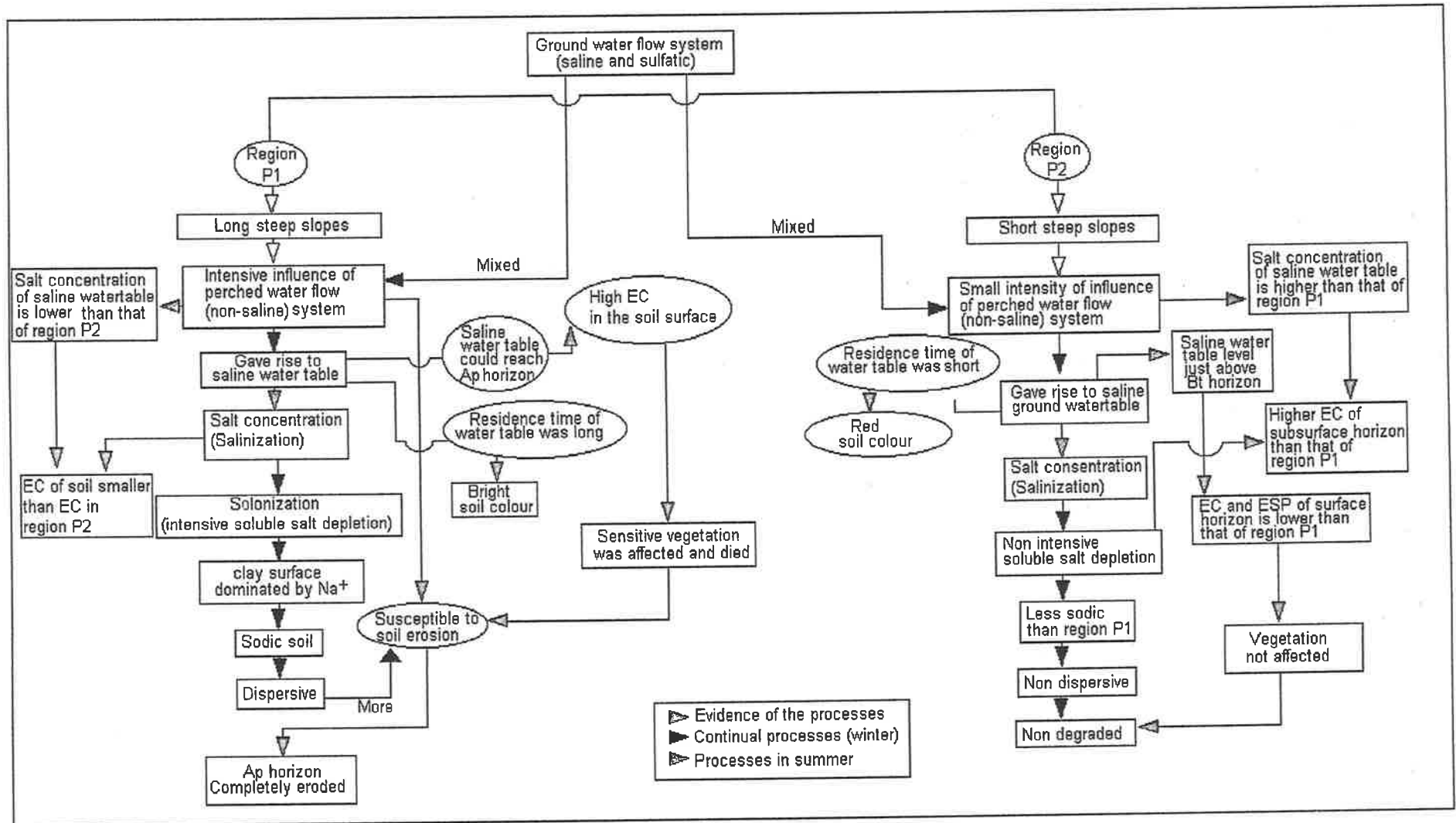


Figure 5.2. A conceptual model illustrating the degraded-nondegraded phenomenon at the Herrmanns sub-catchment

References

- Allison, L.E. (1965). Organic carbon. pp.1367-1378. In Black, C.A., Evans, D.D., White J.L., Ensminger L.L. and Clark F.E. (Eds.), *Methods of soil analysis: Chemical and mineralogical properties*. American Society of Agronomy. Madison, Wisconsin, USA.
- Acharya, C.L., and Abrol, I.P. (1978). Exchangeable sodium and soil water behavior under field conditions. *Soil Science*, 125, 310 – 319.
- Alperovitch, N., Shainberg, I., and Singer, M.J. (1985). Effect of clay mineralogy and aluminum and iron oxides of Hydraulic conductivity of clay-sand mixtures. *Clays and Clay Minerals*, 33, 443-450.
- Arora, H.S. and Coleman, N.T. (1979). The influence of electrolyte concentration on flocculation of clay suspensions. *Soil Science*, 127, 134-139.
- Baldock, J.A., Aoyama, M., Oades, J.M., Susanto, and Grant, C.D. (1994). Structural amelioration of a South Australian Red-Brown Earth using calcium and organic amendments. *Australian Journal of Soil Research*, 32, 571-594.
- Bartelli, L.J. and Odell, R.T (1960). Laboratory studies and genesis of a clay-enriched horizon in the lowest part of the solum of some Brunizem and Gray-Brown Podzolic soils in Illinois. *Soil Science Society of America Proceedings*, 24, 390-395.
- Baven, K. and Germann, P. (1981). Macropores and water flow in soils. *Water Resources Research*, 18, 1311-1325.
- Biddle, D.L., Chittleborough, D.J. and Fitzpatrick, R.W. (1995). Field monitoring of solute and colloidal mobility in a gneissic sub-catchment, South Australia. *Applied Science*, 9, 433-442.
- Bishop, P.M., Mitchell, P.B., and Paton, T.R. (1980). The formation of duplex soils on hillslopes in the Sydney Basin, Australia. *Geoderma*, 23, 175-189.
- Blake, G.R. and Hartge, K.H. (1986). Bulk density. Pp. 363-375. In Klute, A. (Ed.), *Methods of soil analysis: Part 1, Physical and mineralogical methods*. American Society of Agronomy. Madison, Wisconsin, USA.
- Bodman, G.B. and Constantin, G.K. (1965). Influence of particle size distribution in soil compaction. *Hilgardia*, 36, 567-591.
- Brady, C. Nyle and Weil, R. (1999). *The nature and properties of soils*, Twelfth Edition. Prentice-Hall, New Jersey.
- Brewer, R. (1960). Cutans: Their definition, recognition, and their interpretation. *Journal of Soil Science*, 11, 280-291.
- Brydon, J.E. (1965). Clay illuviation in some orthic podzol of Eastern Canada. *Canadian Journal of Soil Science*, 45, 127-138.
- Buol, S.W., Hole, F.D., McCracken, R.J. and Southard, R.J. (1997). *Soil Genesis and Classification*. Iowa State University Press, Ames.

- Cabrera-Martines, F., Harris, W.G., Carlisle, V.W., Collins, M.E. (1989). Evidence for clay translocation in coastal plain with sandy/loamy boundaries. *Soil Science Society of America Journal*, 53, 1108-1113
- Cass, A., Walker, R.R. and Fitzpatrick, R.W. (1996). Vinyard soil degradation by salt accumulation an the effect on the performance of vine. Proceedings. *The Ninth Australian Wine Industry Technical Conference*. 153 - 160.
- Chartres, C.J. (1995). Sodic soils: An introduction to their formation and distribution in Australia. pp. 35-40. In Naidu R., Sumner M.E. and Rengasamy P. (eds.) *Australian sodic soils: Distribution, properties and management*. CSIRO, Australia.
- Chittleborough, D.J. (1982). Effect of method of dispersion on the yield of clay and fine clay. *Autralian Journal of Soil Research*, 20, 339-346.
- Chittleborough, D.J. (1992). Formation and pedology of duplex soils. *Australian Journal of Experimental Agriculture*, 32, 815 - 825.
- Chorom, M., Rengasamy, P., Murray, R.S. (1994). Clay dispersion as influenced by pH and net particle charge of sodic soils. *Australian Journal of soil Research*, 32, 1243-1252.
- Churchman, G.J., Skjemstad, J.O. and Oades, J.M. (1993). Influence of clay minerals and organic matter on effects of sodicity on soils. *Australian Journal of soil Research*, 31, 779-799.
- Churchman, G.J. and Weissmen, D.A (1995). Particle mobility as a parameter in soil dispersibility. pp. 191-194. In Naidu R., Sumner M.E. and Rengasamy P. (eds.), *Australian sodic soils; distribution, properties and management*. CSIRO, Australia.
- Collis-George, N., and Smiles, D.E. (1963). An examination of cation balance and moisture characteristics of determining the stability of soil aggregates. *Journal of Soil Science*, 14, 21-32.
- Coughlan, K.J., Loch, R.J. and Fox, W.E. (1978). Binary packing theory and the physical properties of aggregates. *Australian Journal of Soil Research*, 16, 283-289.
- Cox, J. W., Fritsch, E., and Fitzpatrick, R.W. (1996). Interpretation of soil features produced by ancient and modern processes in degraded landsape: VII. Water duration. *Australian Journal of Soil Research*, 34, 803-824.
- Deeks, L.K., Williams, A.G., Dowd, J.F. and Scholefield, D. (1999). Quantification of pore size distribution and the movement of solutes through isolated soil blocks. *Geoderma*, 90, 65-86
- D'Hoore, J. (1974). Clay dispersion, particle size distribution and colloid stability in high weathered ferralitic soils. *Pedologie* 24, 178-190
- Dixit, S.P. (1978). Measurement of the mobility of soil colloids. *Journal of Soil Science*, 29, 557-566.

- Dixit, S.P., Gombeer, R., and D'Hoore, J., (1975). The electrophoretic mobility of natural clays and their potential mobility within the pedon. *Geoderma*, 13, 325 - 330.
- Dixon, R.M. and Peterson, A.E. (1971), Water infiltration control: A channel system concept. *Soil Science Society of America Proceedings*, 35, 968-973.
- Duchaufour, P. (Translated by Paton, T.R) (1982). *Pedology: Pedogenesis and classification*. George Allen and Unwin, London.
- Durgin, P.B. and Chaney, J.G. (1984). Dispersion of kaolinite by dissolved organic matter from Douglas-fir roots. *Canadian Journal of Soil Science*, 54, 445-455.
- Ehlers, W. (1975). Observations on earthworm channels and infiltration on tilled and untilled losses soil. *Soil Science*, 119: 242-249.
- El-Rayah, H.M.E. and Rowell, D.L. (1973). The influence of Fe and Al hydroxides on swelling montmorillonite and the permeability of a Na-soil. *Journal of Soil Science*, 24, 137-144.
- El-Swaify, S.A., Ahmed, S. and Swindale, L.D. (1970). Effects of absorbed cations on physical properties of tropical red and black earth. II. Liquid limit, degree of dispersion, and moisture retention. *Journal of Soil Science*, 21, 188-198.
- Emerson, W.W. (1954). The determination of the stability of soil crumbs. *J. Soil Sci.* 5:235-250.
- Emerson, W.W., and Chi, C.L. (1977). Exchangeable calcium, magnesium and sodium and the dispersion of illites in water. II. Dispersion of illites in water. *Australian Journal of Soil Research*, 15, 255-262
- Emerson, W.W. (1983). Interparticle bonding. Pp. 477-498. In: *Soils: An Australian view point*. Division of soils, CSIRO, Melbourne, Academic Press. London.
- Eswaran, H. and Sys, C. (1979). Argillic horizon in LAC soils: formation and significance to classification. *Pedologie*, 29, 175-190.
- Fanning, S. D and Fanning, C.B (1988). *Soil: Morphology, genesis, and Classification*. John Wiley and Sons, New York.
- Fitzpatrick, E.A. (1971). *Pedology; a systematic approach to soil science*. Oliver and Boyd, Edinburgh.
- Fitzpatrick, R.W., Boucher, S.C., Naidu, R. and Fritsch, E. (1994b). Environmental consequences of soil sodicity. *Australian Journal of Soil Research*, 32, 1069-1093.
- Fitzpatrick, R.W., Bruce, D.A., Davies, P.J., Spouncer, L.R., Merry, R.H., Fritsch, E. and Maschmedt, D. (1999). Mt Lofty Ranges pilot project, soil landscape quality assesment at catchment and regional scale. Technical report 28/99, July 1999.
- Fitzpatrick, R.W., Cox ,J.W., Fritsch, E. and Hollingsworth, I.D. (1994a). A soil-diagnostic key to manage saline and waterlogged catchment in the Mt. Lofty ranges, South Australia. *Soil Use and Management*, 10, 145-152.

- Fitzpatrick, R.W., Fritsch, E., and Self, P.G. (1996). Interpretation of soil features produced by ancient and modern processes in degraded landscapes: V. Development of saline sulfidic features in non tidal seepage areas. *Geoderma*, 69, 1-29.
- Frenkel, H., Goertzen, J.O., and Rhoades, J.D. (1978). Effect of clay type and content, exchangeable sodium percentage and electrolyte concentration on clay dispersion and hydraulic conductivity. *Soil Science Society of America Journal*, 42, 32-39.
- Fritsch, E. and Fitzpatrick, R.W. (1994). Interpretation of soil features produced by ancient and modern processes in degraded landscapes. I. A new method for constructing conceptual soil-water-landscape models. *Australian Journal of Soil Research*, 32, 889 - 907.
- Gee, G.W. and Bauder, J.W (1986). Particle size analyses. pp. 363-375 in Klute A. (Ed.), *Methods of soil analysis: Part 1, Physical and mineralogical methods*. American Society of Agronomy. Madison, Wisconsin, USA.
- Ghassemi, F., Jakeman, A.J. and Nix, H.A. (1995). *Salinisation of land and water resources: Human causes, management and case studies*. University of New South Wales Press LTD. Sydney.
- Goldberg, S. (1989). Interaction of Aluminum oxides and clay minerals and their effect on soil physical properties. *Communication in Soil Science and Plant Analysis*, 20, 1181-1207.
- Goldberg, S. and Forster, H.S. (1990). Flocculation of reference clays and arid zone soil clays. *Soil Science Society of America Journal*, 54, 714-718.
- Goldberg, S. and Glaubig, R.A. (1987). Effect of saturating cation, pH, and aluminum and iron oxide on the flocculation of kaolinite and montmorillonite. *Clays and Clay Minerals*. 35, 220-227.
- Goldberg, S., Suarez, D.L. and Glaubig, R.A. (1988). Factors affecting clay dispersion and aggregate stability of arid zone soils. *Soil Science*, 146, 317-325.
- Goss, D.W., Smith, S.J., Stewart, B.A. (1973). Movement of added clay through calcareous materials. *Geoderma*, 9, 97-103.
- Greene, R.S.B., Posner, A.M., and Quirk, J.P. (1978). A study of the coagulation of montmorillonite and illite suspensions by calcium chloride using the electron microscope: in *Modification of soil structure*, Emerson W.W., Bond R.D., and Dexter A.R. (eds.). Wiley New York. Pp 35-40.
- Gregory, P.J., Tennant, D., Hamblin, A.P., and Eastham, J. (1992). Components of the water balance on duplex soils in Western Australia. *Australian Journal of Experimental Agriculture*, 32, 845-855.
- Gunn, R.H. (1967). A soil catena on a denuded laterite profiles in Queensland. *Australian Journal of Soil Research*, 5, 117-132
- Gupta, R.K. and Abrol, I.P. (1990). Salt affected soils: Their reclamation and management for crop production. *Advances in Soil Science*, 11, 224-276

- Gupta, R.K., Bhumbra, D.K., and Abrol, I.P. (1984). Effect of sodicity, pH, organic matter, and calcium carbonate on the dispersion behavior of soils. *Soil Science*, 137, 245-251.
- Hallsworth, E.G. (1963). An estimation of some factors affecting the movement of clay in an artificial soil. *Journal of Soil Science*, 14, 360-371.
- Ilyas, M., Miller, R.W. and Qureshy, R.H. (1993). Hydraulic conductivity of saline-sodic soil after gypsum application and cropping. *Soil Science Society of America Journal*, 57, 1580-1585.
- Inbar, M., Tamir, M. and Wittenberg, L.(1998). Runoff and erosion processes after a forest fire in Mount Carmel, a Mediterranean area. *Geomorphology* 24, 17-33.
- Jiongxin, Xu (1999). Erosion caused by hyperconcentrated flow on the Loess Plateau of China. *Catena*, 36, 1-19.
- Kelly, W.P. (1951). Alkali soils: Their formation, properties and reclamation'. Reinhold Publishing Corporation, New York.
- Kretschmar, R., Roberge, W.P., and Amoozegar, A. (1994). Filter efficiency of three saprolites for natural clay and oxide colloids. *Environmental Science and Technology*, 28, 1907-1915.
- Kretschmar, R., Robarge, W.P., and Weed, S.B. (1993). Flocculation of kaolinitic soil clays: Effects of humic substances and iron oxides. *Soil Science Society of America Journal*, 57, 1277-1282.
- Lettenmaier, D.P., Hooper, E.R., Wagner, C., and Faris, K.B.(1991). Trends in stream quality in the continental United States, 1987-1988. *Water Resources Research*, 27, 327-339.
- Levy, G.J., Van-Der-Watt, H.V.H., and Du-Plessis, H.M. (1988). Effect of sodium-magnesium and sodium-calcium systems on soil hydraulic conductivity and infiltration. *Soil Science*, 146, 303-310.
- Luxmoore, R.J. (1981). Micro-, Meso-, and Macroporosity of soil. *Soil Science Society of America Journal*, 45, 671.
- Luxmoore, R.J., Jardine, P.M., Wilson, G.V., Jones J.R., Zelazny L.W. (1990). Physical and chemical controls of preferred pathflows through a forested hillslope. *Geoderma*, 46, 143 - 163.
- McDonald, R.C. and Isbell, R.F. (1990). Soil profiles, pp. 103 - 152, in R.C. McDonald., R.F. Isbell, J.G. Speight, J.G. Walker and M.S. Hopkins (eds). Australian soil and land survey field handbook 2nd ed. Inkata Press, Melbourne.
- McIntyre, D.S. (1979). Exchangeable sodium, subplasticity and hydraulic conductivity of Australian soils. *Australian Journal of Soil Research*, 17, 115-120.
- McKeague, J.A. and St.Arnaud, R.J. (1969). Pedotranslocation: Eluviation-illuviation in soils during quaternary. *Soil Science*, 106, 428-434.

- McNeal, B.L. and Coleman, N.T. (1966). Effect of solution composition on soil hydraulic conductivity. *Soil Science Society of America Journal*, 30, 308-312.
- McNeal, B.L., Layfield, D.A., Norvell, W.A., and Rhoades, J. (1968). Factors influencing hydraulic conductivity of soils in the presence of mixed-salt solution. *Soil Science Society of America Proceedings*, 32, 187-190.
- McTainsh, G.H., and Boughton, W.C. (1993). Land degradation processes in Australia. Longman Cheshire, Melbourne.
- Miller, R.W. and Gardiner, D.T. (1998). Soils in our environment (8th edition). Prentice Hall, New Jersey.
- Murtha, G.G. (1988) in Gunn, R.H., Beattie, J.A., Reid, R.E. and Van de Graaff, R.H.M. (Eds). Australian soil and land survey handbook; Guidelines for conducting surveys. Inkata Press, Melbourne.
- Nakayama, F.S. (1970). Hydrolysis of CaCO_3 , Na_2CO_3 and NaHCO_3 and their combinations in the presence and the absence of external CO_2 source. *Soil Science*, 109, 391-398.
- Nelson, D.W. and Sommers, L.E. (1982). Total carbon, organic carbon, and organic matter. Pp. 539-579. In Page A.L. (editor), Miller R.H. and Keeney D.R. (assoc. editor) Methods of soil analyses Part 2; Chemical and microbial properties second edition. American Society of Agronomy. Madison, Wisconsin, USA.
- Nelson, P.N., Baldock, J.A., Clarke, P., Oades, J.M., and Churchman, G.J. (1999). Dispersed clay and organic matter in soil : their nature and associations. *Australian Journal of Soil Research*, 37, 289 - 315.
- Nelson, P.N. and Oades, J.M. (1998). Organic matter, sodicity and soil structure. pp. 51-76. In Sumner, M.E. and Naidu, R. (Eds.). Sodic soils: Distribution, properties, management, and environmental consequences. Oxford University Press, New York .
- Nettleton, W.D., Eswaran, H., Holzhey, C.S., and Nelson, R.E., (1987). Micromorphological evidence of clay translocation in poorly dispersible soils. *Geoderma*, 40, 37-48.
- Noack, A.G., Grant, C.D., and Chittleborough, D.J., (2000). Colloid movement through stable soils of low cation exchange capacity. *Environmental Science and Technology*, 34, 2490-2497.
- Northcote, K.H. (1979). A factual key for the recognition of Australian soils. Rellim Technical Publication, Glenside, South Australia. 4th edition.
- Northcote, K.H. and Skene, J.K.M. (1972). Australian soil with saline and sodic properties. CSIRO soil publication 27.
- Oertel, A.C. (1961). Pedogenesis of some red brown earths based on trace element profiles. *Journal of Soil Science*, 12, 242-258.
- Oster, J.D. and Schroer, F.W. (1979). Infiltration as influenced by irrigation water quality. *Soil Science Society of America Journal*, 43, 444-447.

- Oster, J.D., Shainberg, I., and Wood, J.D. (1980). Flocculation value and gel structure of sodium/calcium montmorillonite and illite suspensions. *Soil Science Society of American Proceedings*, 44, 955-959.
- Perroux, K.M. and White, I. (1988). Design for disc permeameter. *Soil Science Society of America Journal*, 52, 1205-1215.
- Pilgrim, D.H. and Huff, D.D. (1983). Suspended sediment in rapid subsurface stormflow on a large field plot. *Earth Surface Processes*, 8, 451-463.
- Purves, W.D and Blyth, W.B. (1969). A study of associated hydromorphic and sodic soils on redistributed Karoo sediments. *Rhodesian Journal of Agriculture Research*, 7, 99-108.
- Quirk, J.P., and Schofield, R.K. (1955). The effect of electrolyte concentration on soil permeability. *Journal of Soil Science*, 6, 163-178.
- Raven, M.D. (1990). XPLOTT Version 3.2. User manual. Manipulation Diffraction Data. CSIRO Aust. Div. Soil Tech. Mem. No.1/1989.
- Rayment, G.E. and Higginson, S.E. (1992). Australian laboratory handbook of soil and water chemical methods. Inkata Press, Melbourne.
- Rengasamy, P. (1983). Clay dispersion in relation to changes in electrolyte composition of dialysed red-brown earths. *Journal of Soil Science*, 34, 723-732.
- Rengasamy, P., Greene, R.S.B., Ford, G.W., and Mehanni, A.H. (1984). Identification of dispersive behaviour and the management of Red-Brown Earths. *Australian Journal of Soil Research*, 22, 413-431.
- Rengasamy, P. and Olsson, K.A.,(1991). Sodicity and soil structure. *Australian Journal of Soil Research*, 29, 935-52.
- Reynolds, W.D. Elrick, D.E. and Topp, G.C. (1983). A reexamination of the constant head well permeameter method for measuring saturated hydraulic conductivity above the water table. *Soil Science*, 136, 250-268.
- Rhoades, J.D (1982). Cation exchange capacity. pp. 149-158. In Page A.L. (editor), Miller, R.H. and Keeney, D.R. (assoc. editor) *Methods of soil analyses Part 2; Chemical and microbiological properties* second edition. American Society of Agronomy. Madison, Wisconsin, USA.
- Rowell, D.L., Payne, D. and Ahmad, N. (1969). The effect of the concentration and movement of solutions on the swelling, dispersion, and movement of clay in saline and alkali soils. *Journal of Soil Science*, 20, 176-188.
- Schofield, R.K. and Samson, H.R. (1954). Flocculation of kaolinite due to the attraction of oppositely charged crystal faces. *Discussions of the Faraday Society*, 18, 138-145.
- Seta, A.K. and Karathanasis, A.D. (1996). Water dispersible colloids and factors influencing their dispersibility from soil aggregates. *Geoderma*, 74, 255-266.

- Shainberg, I. and Letey, J. (1984). Response of soils to sodic and saline conditions. *Hilgardia*, 52, 1-57.
- Shainberg, I., Rhoades, J.D., and Prather, R.J. (1981). Effect of low EC on clay dispersion and hydraulic conductivity of a sodic soil. *Soil Science Society of America Journal*, 45, 273-277.
- Shanmuganathan, R.T. and, Oades J.M. (1983). Influence of anions on dispersion and physical properties of the A horizon of a red-brown earth. *Geoderma*, 29, 257-277.
- Shaw, J.N., Bosch, D.D., West, L.T., Truman C.C. and Radcliffe D.E. (2001). Lateral flow in loamy to sandy Kandiuults of the upper coastal plain of Georgia (USA). *Geoderma*, 99, 1-168.
- Skopp, K.J. (1981). Comment on Micro-, Meso- and Macroporosity of soils. *Soil Science Society of America Journal*, 45, 1246
- Smettem, K.R.J. and Collis-George, N. (1985). Statistical characterization of soil biopores using a soil peel method. *Geoderma*, 36, 27-36.
- Smith, G.D., Coughlan, K.J. and Fox, W.E. (1978). The role of texture in soil structure. In: Emerson W.W., Bond R.D. and Dexter A.S (Eds.). *Modification of soil structure*. pp 79-86. Wiley, Chichester, UK.
- Soil survey staff (1999). *Soil taxonomy : A basic system of soil classification for making and interpreting soil surveys*. USDA-SCS Agric. Handbook No. 436 U.S. Govt. Printing office, Washington D.C.
- Stace, H.C.T., Hubble, G.D., Brewer, R., Northcote, K.H., Sleeman, J.R., Mulcahy, M.J., and Hallsworth, E.G. (1968). *A Handbook of Australian Soils*. Rellim, Glenside, South Australia.
- Suarez, D.L., Rhoades, J.D., Lavado, R., and Grieve, C.M. (1984). Effect of pH on saturated hydraulic conductivity and soil dispersion. *Soil Science Society of America Journal*, 48, 50-55
- Sumner, M.E. (1993). Sodic soils: New perspective. *Australian Journal of Soil Research*, 31, 683-750.
- Sumner, M.E. (1995). Sodic soils: New perspective. In Naidu, R., Sumner, M.N., and Rengasamy, P. (eds.). *Australian sodic soils; distribution, properties and management*.CSIRO, Australia.
- Sumner, M.E., Rengasamy, P. and Naidu, R. (1998). Sodic soils: A Reappraisal. pp. 3-17. In Sumner M.E. and Naidu P. (eds), *Sodic soils: distribution, properties, management and environmental consequences*. Oxford University Press, New York.
- Swartzen-Allen, S.L. and Matijevic, E. (1976). Colloid and surface properties of clay suspensions. III. Stability of montmorillonite and kaolinite. *Journal of Colloid Interface Science*, 56, 159-167.

- Tennant, D., Scholz, G., Dixon, J. and Purdie, B. (1992). Physical and chemical characteristics of duplex soils and their distribution in the south-west of Western Australia. *Australian Journal of Experimental Agriculture*, 32, 827-843.
- Tessen, E (1984). Clay migration in upland soils of Malaysia. *J Soil Science*, 35, 615-624.
- Thorp, J., Cady, J.G., and Gamble, E.E. (1959). Genesis of Miami silt loam. *Soil Science Society of America Proceedings*, 23, 156-161.
- Turner, N.C. (1992). Crop production o duplex soils: an introduction. *Australian Journal of Experimental Agriculture*, 32, 797-800.
- U.S. Salinity Laboratory Staff (1954). Diagnosis and improvement of saline and alkali soils. Handbook 60 (U.S Government Printing Office: Washington D.C.)
- van Olphen, H.(1977). An introduction to clay colloid chemistry: 2nd ed., Wiley, NY.
- Vinten, A.J.A. and Nye, P.H. (1985). Transport and deposition of dilute colloidal suspensions in soils. *Journal of Soil Science*, 36, 531-541.
- Visser, S.A. and Caillier, M. (1988). Observation on the dispersion and aggregation of clays by humic substances I. Dispersive effects of humic acids. *Geoderma*, 42, 331-337.
- Walling, D.E. (1984). Dissolved loads and their measurement. pp. 111-177. In Hadley, R.F. and Walling, D.E (eds), *Erosion and sediment yield : Some methods of measurement and modelling*. University Press, Cambridge.
- Ward, P.R.B.(1984) Measurements of sediment yields. pp. 37-70. In Hadley, R.F. and Walling, D.E (eds), *Erosion and sediment yield : Some methods of measurement and modelling*. University Press, Cambridge.
- Willet, I.R. and Walker, P. H. (1982). Soil morphology and distribution of iron and sulfur fractions in a coastal flood plain toposequence. *Australian J Soil Research*, 20, 283-294.
- Willet, I.R., Crockford, R.H. and Milnes, A.R. (1992). Transformations of iron, manganese and aluminum during oxidation of a sulfidic material from an acid sulfate soil. In Skinner, H.C.W. and Fitzpatrick, R.W. (Eds.). *Biomining processes of iron and manganese*. *Catena Supplement*, 21, 287-302.
- Wilson, G.V. and Luxmoore R.J. (1988). Infiltration, macroporosity, and mesoporosity distributions on two Forested watersheds. *Soil Science Society of America Proceedings*, 52, 329-335.
- Yanda, P.Z. (2000). Use of soil horizons for assessing soil degradation and reconstructing chronology of degradation processes: the case of Mwisanga Catchment, Kondoa, central Tanzania *Geomorphology*, 34, 209-225.
- Zachar D. (1982). Soil erosion. Elsevier Scientific Publishing Company, New York.

Appendix 1. Morphological descriptions of a representative saline soil profile (P1) in the Herrmanns sub-catchment.

Topography : Eroded footslope
Classification : Typic Natraqualf

Horizons	Description
Epz 0 - 10 cm;	light gray (10YR7/2) dry and light brownish gray (10YR6/2) moist; mottles many strong brown (7.5YR 5/6) and very few red (2.5YR 4/8); sandy loam; strong medium polyhedral structure; very strong (dry); moderately sticky and moderately plastic; non calcareous; pH _{1.5} 9.2; no roots; gradual wavy boundary.
Btn1 10 - 31cm;	reddish yellow (7.5YR6/6) dry and light brown (7.5YR6/4) moist; mottles many red (2.5YR4/8); clay; strong medium columnar structure; very strong (dry); common coarse macropores; few strong fine manganous nodules; sticky and plastic; non calcareous; pH _{1.5} 8.4; few medium gravel fragments; few fine roots; gradual wavy boundary.
Btn2 31-40cm;	reddish yellow (7.5YR6/8) dry and reddish yellow (7.5YR 6/6) moist; mottles common red (10R4/6) and common light gray (2.5Y7/2); sandy clay loam; strong medium columnar structure; rigid (dry); common coarse macropores; many weak coarse ferruginous nodules and few weak medium manganous nodules; sticky and plastic; non calcareous; pH _{1.5} 8.5; abundant medium gravel fragments; no roots; gradual wavy boundary.
Btn3 40-62cm;	light olive (2.5Y5/4) dry and light brown (2.5Y4/3) moist; mottles very few weak red (10R4/4); clay loam; strong medium columnar structure; very strong (dry); common strong coarse ferruginous nodules; sticky and plastic; non calcareous; pH _{1.5} 8.4; many gravel fragments; no roots; gradual irregular boundary.
Btng1 62-80cm;	dark yellowish brown (10YR4/4) dry and dark grayish brown (10YR4/2) moist; mottles many dark red (2.5YR3/6) and common greenish gray (5/10Y); clay loam; strong medium columnar structure; rigid (dry); few distinct clay skin; common strong medium ferruginous nodules; very sticky and very plastic; non calcareous; pH _{1.5} 8.6; no roots; gradual wavy boundary.
Btng2 80-113cm;	light olive gray (5Y6/2) dry and gray (5Y6/1) moist; mottles many reddish brown (5YR4/4) and common dark yellowish brown (10YR4/6); clay loam; strong medium columnar structure; very strong (dry); few weak coarse manganous nodules; very sticky and very plastic; non calcareous; pH _{1.5} 8.5; no roots; gradual irregular boundary.
Btng3 113-130cm;	pale yellow (2.5Y7/4) dry and light gray (2.5Y7/2) moist; mottles common pale green (6/5G) and many yellowish brown (10YR5/8); clay loam; strong medium columnar structure; very strong (dry); very sticky and very plastic; non calcareous; pH _{1.5} 8.7; no roots; gradual wavy boundary.
Btng4 130-150cm;	pale yellow (2.5Y7/4) dry and light gray (2.5Y7/2) moist; mottles common greenish gray (6/5G) and few yellowish brown (10YR5/8); clay loam; strong medium columnar structure; very strong (dry); very sticky and very plastic; non calcareous; pH _{1.5} 8.8; no root.

Appendix 2. Morphological descriptions of a representative sodic soil profile (P2) in the Herrmanns sub-catchment

Topography : Waning midslope

Classification : Typic Natrixeralf

Horizons	Description
Ap 0-8 cm;	grayish brown (10YR5/2) dry and gray (10YR5/1) moist; sandy loam; weak fine polyhedral structure; very weak (dry); non sticky and non plastic; non calcareous; pH _{1.5} 5.5; common very fine roots; clear wavy boundary.
E1 8-16 cm;	reddish brown (5YR5/4) dry and reddish brown (5YR5/3) moist; loamy sand; weak fine polyhedral structure; weak (dry); non sticky and non plastic; non calcareous; pH _{1.5} 5.4; few very fine roots; diffuse wavy boundary.
E2 16-26 cm;	pale brown (10YR6/3) dry and light brown grayish (10YR 6/2) moist; sandy loam; weak fine polyhedral structure; firm (dry); non sticky and non plastic; non calcareous; pH _{1.5} 5.8; few very fine roots; diffuse irregular boundary.
Btnc1 26-43 cm;	strong brown (7.5YR5/8) dry and strong brown (7.5YR5/6) moist; mottles many grayish brown (10YR5/2) and few yellowish red (5YR4/6); clay loam; weak medium columnar structure; very strong (dry); many strong medium ferruginous nodules; slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.4; few very fine roots; gradual irregular boundary.
Btnc2 43-62 cm;	yellowish red (5YR4/6) dry and reddish brown (5YR4/4) moist; mottles common red (10R4/8) and many grayish brown (10YR5/2); sandy clay loam; strong medium columnar structure; very strong (dry); common strong medium ferruginous nodules; slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.9; few very fine roots; gradual irregular boundary.
Bt 62-75 cm;	strong brown (7.5YR5/6) dry and brown (7.5YR5/4) moist; mottles many weak red (10R4/4); sandy clay loam; strong medium columnar structure; very strong (dry); few prominent clay skin; very few strong fine ferruginous nodules; slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.6; no roots; gradual wavy boundary.
Btn 75-114 cm;	strong brown (7.5YR4/6) dry and red brown (7.5YR4/4) moist; mottles common dusky red (10R3/4) and few grayish brown (10YR5/2); clay; strong fine columnar structure; very strong (dry); common distinct clay skin; sticky and plastic; non calcareous; pH _{1.5} 6.5; few very fine roots; gradual wavy boundary.
Btng1 114-132 cm;	red (10R4/6) moist; mottles few greenish gray (5/5G) and many dark yellowish brown (10YR4/6); clay; strong medium columnar structure; very strong (dry); very sticky and very plastic; non calcareous; pH _{1.5} 6.5; no roots; gradual wavy boundary.
Btng2 132-170 cm;	dusky red (10R3/4) moist; mottles few greenish gray (5/5G); clay; strong medium columnar structure; very strong (dry); sticky and plastic; non calcareous; pH _{1.5} 6.4; no root.

Appendix 3. Morphological descriptions of a representative nonsodic-nonsaline soil profile (P3) in the Herrmanns sub-catchment.

Topography : Upper waning midslope

Classification : Ultic Palexeralf

Horizons	Description
Ap 0 - 8 cm;	very dark grayish brown (10YR3/2) dry and very dark gray (10YR3/1) moist; sandy loam; weak fine subangular blocky structure; very weak (dry); non sticky and non plastic; non calcareous; pH _{1.5} 5.6; few medium gravel fragments; common fine roots; clear irregular boundary.
Ec 8-15 cm;	strong brown (7.5YR4/6) dry and brown (7.5YR5/4) moist; mottles difficult to describe due to high gravel content; sandy loam; weak very fine single grain structure; loose (dry); few strong fine manganous nodules; non sticky and non plastic; non calcareous; pH _{1.5} 6.1; abundant coarse fragments; few very fine roots; clear smooth boundary.
Bt1 15-39 cm;	yellowish brown (5YR4/6) dry and red brown (5YR4/4) moist; mottles few light yellow brown (10YR6/4); clay loam; strong medium polyhedral structure; very strong (dry); very few strong fine manganous nodules; slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.5; common very fine roots; gradual wavy boundary.
Bt2 39-52 cm;	red (2.5YR4/8) dry and dark red (2.5YR3/6) moist; mottles common yellowish brown (10YR5/8); clay loam; moderate medium polyhedral structure; strong (dry); slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.5; few very fine roots; gradual wavy boundary.
Bt3 52-68 cm;	red (2.5YR4/8) dry and dark red (2.5YR3/6) moist; mottles many yellowish brown (10YR5/8); clay; moderate medium polyhedral structure; strong (dry); slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.1; few very fine roots; gradual broken boundary.
Bt4 68-84 cm;	yellowish brown (10YR5/8) moist; mottles many dark red (2.5Y3/6) and common light yellowish brown (2.5Y6/4); clay; moderate medium subangular blocky structure; strong (dry); sticky and plastic; non calcareous; pH _{1.5} 5.8; no roots; gradual broken boundary.
Btg1 84-97 cm;	light yellowish brown (2.5Y6/4) moist; mottles few red (2.5YR4/8); clay; moderate medium polyhedral structure; strong (dry); few weak fine ferruginous nodules; sticky and plastic; non calcareous; pH _{1.5} 5.4; no roots; diffuse smooth boundary.
Btg2 97-119 cm;	light brownish gray (2.5Y6/2) moist; mottles few yellowish brown (10YR5/8) and few yellowish brown (5YR4/6); clay; moderate medium polyhedral structure; strong (dry); sticky and plastic; non calcareous; pH _{1.5} 5.2; no root.

Appendix 4. Morphological description of a soil profile on the edge of a stream bank in the Herrmanns sub-catchment

Topography : Stream bank

Classification : Typic Natrixeralf

Horizons	Description
Ap 0-11 cm;	dark brown (7.5YR4/2) dry and brown (7.5YR5/2) moist; sandy loam; weak fine subangular blocky structure; very weak (dry); non sticky and non plastic; non calcareous; pH _{1.5} 4.8; many very fine roots; gradual wavy boundary.
E 11-28 cm;	gray (2.5Y6/1) dry and light gray (2.5Y7/1) moist; sandy loam; weak fine subangular blocky structure; weak (dry); non sticky and non plastic; non calcareous; pH _{1.5} 6.0; common very fine roots; gradual wavy boundary.
Btz1 28-55 cm;	dark yellowish brown (10YR4/4) dry and dark yellowish brown (10YR4/6) moist; mottles many light gray (2.5Y6/1) and few strong brown (7.5YR4/6); clay loam; strong medium columnar structure; rigid (dry); slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.7; few very fine roots; gradual irregular boundary.
Btz2 55-88 cm;	dark yellowish brown (10YR4/4) dry and dark yellowish brown (10YR4/6) moist; mottles few light gray (2.5Y6/1) and few strong brown (7.5YR4/6); clay loam; strong medium columnar structure; rigid (dry); slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.9; few very fine roots; gradual irregular boundary.
Btnz1 88-124; cm;	dark yellowish brown (10YR4/4) dry and dark yellowish brown (10YR4/6) moist; mottles very few light gray (2.5Y6/1) and few strong brown (7.5YR4/6); sandy loam; strong medium columnar structure; rigid (dry); slightly sticky and slightly plastic; non calcareous; pH _{1.5} 6.3; few very fine roots; gradual irregular boundary.
Btnz2 124-163; cm;	yellowish brown (10YR5/4) moist; very few dark red (10R3/6); sandy loam; strong medium columnar structure; very strong (dry); slightly sticky and slightly plastic; non calcareous; pH _{1.5} 8.3; no root; gradual wavy boundary.
Btng 163-180; cm;	yellowish brown (10YR5/4) moist; mottles common greenish gray (5/10Y); sandy loam; strong medium columnar structure; very strong (dry); slightly sticky and slightly plastic; non calcareous; pH _{1.5} 8.8; no root.

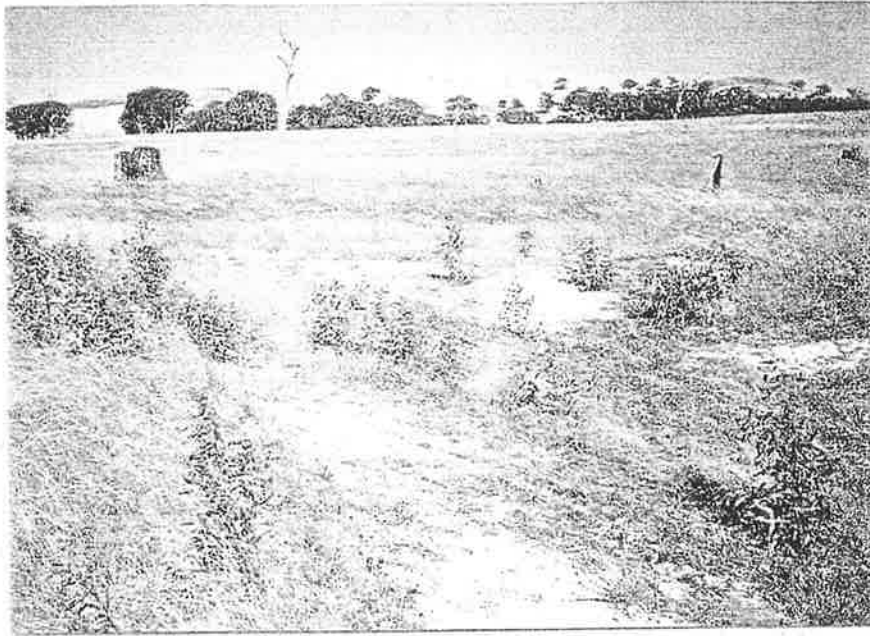


Plate 1. Photograph showing degraded and non degraded areas in Herrmanns sub-catchment. The white patches are salt crusts in the soil surface of eroded area.



Plate 2. Photograph showing the most severely eroded area in Herrmanns sub-catchment. The Ap horizon has been completely removed by erosion. The salt crusts (white coloured zones) are evident.



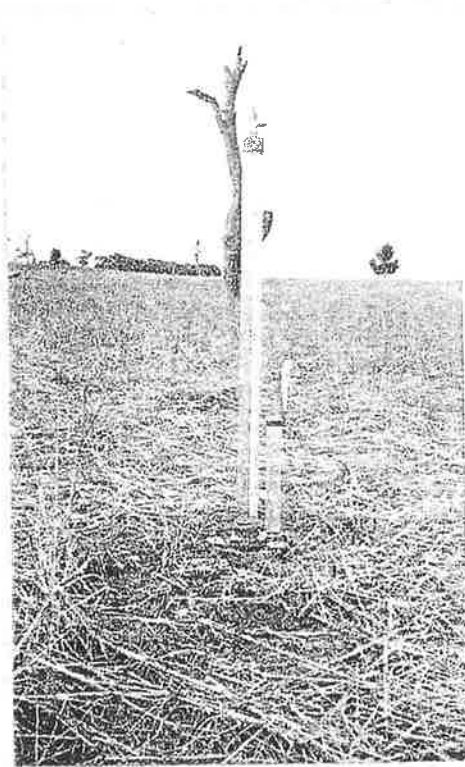
Plate 3. Photograph showing eroded area in the Herrmanns sub-catchment with some gully erosion along the stream bank.



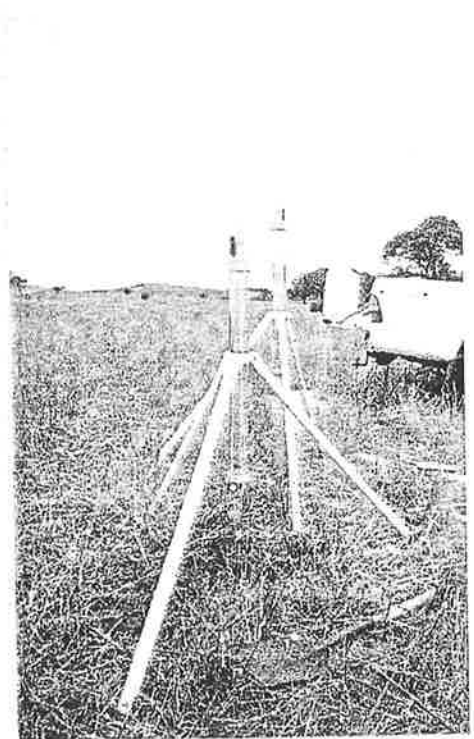
Plate 4. Creek and gully erosion, which did not exist 50 years ago, is now approximately 200 cm depth in some parts.



Plate 5. Soil profile exposed by gully erosion in the deepest part of the stream bank. White patches on the sides of the streambank are salt crusts.



(a)



(b)

Plate 6. (a) CSIRO disc permeameter and (b) Guelph well permeameter used to measure K_s .