QUANTITATIVE ASSESSMENT OF THE SPATIAL VARIABILITY OF MORPHOLOGICAL AND PHYSICAL PROPERTIES OF SOME ALLUVIAL SOILS AT LINCOLN COLLEGE, CANTERBURY

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The complex soil distribution across part of a Lincoln College Farm relates to the previous history and pattern of alluvial deposition. Depth to mottles (DM), depth to gravels (DG), and thickness of loamy sand and/or coarser-textured layers (TS) are used to classify the soils and delineate the area into Eyre, Templeton and Wakanui soil-series simple mapping units.

Geostatistical analyses of the grid data reveal that values of each morphological parameter are spatially dependent, though to different extents. Most variation of DM, for instance, occurs between 30 m and 430 m, whereas a large amount of the variation in TS is present within less than 30 m. The morphological parameters also vary anisotropically, with the direction of maximum variation for DM and DG being NE-SW across a major abandoned channel hollow. Similar patterns are reflected in the soil maps of the study area and of adjacent larger regions, where mapping units are elongated in a NW-SE direction.

Geostatistical methods are more efficient than conventional in determining optimal sampling strategies for future soil survey and variability studies: less samples are needed to achieve the same level of precision.

The morphologically-based soil classification system is generally effective in separating soils into (series) taxonomic units in terms of soil physical properties. Examined hydraulic properties [e.g. "field-saturated" hydraulic conductivity (K_{fs})] differ between typical profiles of each taxonomic unit: these observations are statistically substantiated by data from taxonomically-pure "window areas" of the three soil series. The differences are mainly attributable to spatial changes in soil texture and pore-size distributions. Different amounts of variation in physical properties, however, are still present within each taxonomic unit.

The variation in physical properties amongst the combined window areas is reduced, though, to differing extents, by the classification and delineation into separate taxonomic units. More than half of the variance in moisture content at both topsoil and subsoil depths amongst Templeton and Wakanui taxonomic units, for instance, is accounted for by the classification, and is thus due to differences between the two soils. Little contribution, however, is made by the classification in reducing the heterogeneity of K_{fs} in topsoils. The classification is particularly effective in separating Wakanui from Templeton taxonomic units in terms of subsoil K_{fs}, an important property controlling water movement, storage and related soil-forming processes.

KEYWORDS: soil spatial variability; quantitative assessment; geostatistics; conventional statistics; morphological properties; physical (hydraulic) properties; alluvial soils; soil classification; soil survey; sampling strategies.

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CHAPTER 1

INTRODUCTION

Soil is a three-dimensional body which varies spatially in accordance with the interaction of different environmental factors. Many soil morphological, physical and chemical properties differ markedly in rates of variation. Physical properties, especially hydraulic characteristics, are particularly variable in alluvial soils as a result of the frequent lateral and vertical changes in texture inherited from the parent materials (Butler, 1958; Mausbach et al., 1980; Drees and Wilding, 1973; Wilding and Drees, 1983). Such variations have significant impact on land management and agricultural productivity.

Soil classification and mapping provides the most common means of partitioning soil variation across an area. This traditional approach splits up the soil mantle into individual units which differ in terms of a few easily-measured diagnostic (normally morphological) characteristics. The aim is normally to isolate mapping units which are spatially uniform in these properties and thus equivalent to taxonomic units. Taxonomic impurities, however, are present within most mapping units: amounts will vary according to mapping scale and spatial complexity of property distributions. Accessory properties of taxonomic (and mapping) units are assumed to vary in similar ways to the diagnostic properties: variances so defined within taxonomic units are therefore minimized over all properties. Such important assumptions, however, have rarely been justified by any authors. Non-definitive soil properties may spatially vary at different rates to definitive properties. Soils grouped together in terms of a few definitive characteristics may therefore differ substantially in other non-definitive soil properties. Similarly, soils separated into different classes may differ in the diagnostic characteristics, yet resemble each other in most other properties (Beckett and Webster, 1971; Giltrap et al., 1983). More work is needed to quantitatively assess the effectiveness of classification and mapping in partitioning the variability of non-diagnostic soil properties.

Conventional statistical approaches to soil variability studies assume that variations in soil properties are randomly distributed within sampling units, and that soil property values at unsampled locations can be estimated by sample means and associated confidence limits. Soil properties, however, are often continuous variables and tend to be spatially correlated over vertical or lateral dimensions (Burgess and Webster, 1980a; Trangmar et al., 1985; Warrick et al., 1986). Generally, soil samples close together tend to be more alike than samples far apart. The conventional approach therefore is inadequate for interpolation of spatially-dependent variables as it does not take into account the spatial correlation and relative location of samples; the estimation error is thus unnecessarily large. Sampling strategies determined on the basis of the conventional estimation errors are often conservative with a result of over-sampling and unnecessary effort (McBratney and Webster, 1983).

The regionalised variable theory (geostatistics) developed in the mining industry (Krige, 1966; Matheron, 1963, 1965, 1971) has recently been introduced to soil variability studies (Burgess and Webster, 1980a, b; Webster and Burgess, 1980; Burgess et al., 1981; Trangmar et al., 1985). The theory takes into account both the random and structured characteristics of spatially-correlated variables, and provides a quantitative tool for assessing the spatial dependence of soil properties. Its main uses include the quantification of soil spatial dependence by means of semi-variograms, estimation of soil properties at unsampled locations and production of soil maps by kriging, and determination of sampling strategies based on kriging errors. The application of geostatistics in soil variability studies is relatively new, however, and more studies are needed to verify this approach and assess its value.

Soils around Lincoln College on the Canterbury Plains are developed on a series of alluvial sediments deposited by rivers flowing eastward from the Southern Alps. Previous studies have separated soils in adjacent regions into a number of taxonomic units, e.g. Eyre, Templeton, Wakanui and Temuka series, according to their morphological features (texture and mottling patterns) (Cox, 1978). A more recent study on these soils, however, indicates that the morphological features, e.g. subsurface textural layers and mottling patterns, are extremely variable and the morphologically-based soil classification scheme is unsatisfactory in separating some of the soils (Karageorgis, 1980). This study also showed that crop growth is clearly influenced by the different soil-moisture regimes associated with each soil series. An assessment of the soil classification scheme in terms of soil hydraulic properties is essential not only for improving the usefulness and applicability of the scheme itself, but also for a better understanding of the relationships between morphological and hydraulic properties.

The major aims of this study are twofold.

(1) To describe, explain and quantitatively assess (using geostatistics) the spatial variability of morphological properties of some alluvial soils in Canterbury. Conventional and geostatistical techniques of soil mapping will be compared and optimal sampling strategies for future soil survey and variability studies determined.

(2) To describe and quantitatively assess the variability of soil physical properties of hydraulic significance between and within morphologically-defined soil series taxonomic units. A qualitative or semi-quantitative comparison of a range of physical properties from typical profiles of each taxonomic unit will first be undertaken. This will be followed by conventional statistical analyses of replicated measurements of key soil physical properties from within relevant taxonomically-pure areas. The results will allow an assessment of the

overall effectiveness of the soil classification system in partitioning soil-physical-property variability.

The thesis is divided into seven chapters. General concepts of soil variability, terminology and methodology relevant to this study are reviewed and outlined in Chapter 2. Chapter 3 describes the general physical environment of the Canterbury Plains and provides more detailed background to the study area. The methods, results, discussion, and conclusions concerning objective 1 are presented in Chapter 4. Chapters 5 and 6 are concerned with the profile studies and quantitative comparisons of taxonomic units respectively (objective 2). The overall conclusions of the study are summarised in Chapter 7.

CHAPTER 2

LITERATURE REVIEW: SOIL VARIABILITY

2.1 Introduction

This review first considers the general concepts and causes of soil variability. Traditional methods of variability assessment, using classification and field mapping techniques, are then discussed. The theoretical basis and role of both conventional statistics and geostatistics in assessing soil variability are outlined. The review concludes with a brief discussion of different sampling strategies used in the evaluation of soil variability.

2.2 Components and causes of soil variability

2.2.1 General concepts

Soil properties may change gradually or suddenly, from time to time, or from place to place. Both soil physical properties (e.g. soil temperature and moisture) and soil chemical properties (e.g. nutrient availability) may exhibit temporal variation, in that they vary considerably at a site within any one year, month or even day. Temporal variations in soils, however, are not considered in this study: it is concerned with soil spatial variability. This is the variation of soil properties as a function of distance in either the vertical or lateral dimension.

The changes in soil properties down a vertical section of the soil, known as the soil profile, have long been recognised and described. A soil profile can be divided into a number of horizons (e.g. A, B, and C) on the basis of vertical changes in such soil properties as colour, texture, structure and consistence. These vertical changes are important from not only academic, but also practical viewpoints. For instance, plant roots cannot grow well in a horizon that is devoid of nutrients such as an E horizon; neither can they penetrate and prosper in a compacted subsoil horizon such as a fragipan. Textural changes down the profile greatly influence water movement: water drainage is often impeded by a fine-textured horizon, thus causing the overlying soil material to become waterlogged in the wet season.

Vertical changes of soil properties are normally caused by numerous soil-forming processes which may take place simultaneously, or in sequence. For example, the strongly leached E horizons and underlying Bs horizons of sesquioxide accumulation in some profiles are formed by the processes of eluviation and illuviation. Other features of vertical variation, such as the textural layers in alluvial soils, may be inherited directly from the parent materials.

The soil-forming processes that are responsible for the various soil horizons or features are, in turn, governed by environmental parameters. The soil profile at any location is the product of the interaction of five soil-forming factors: parent material, topography, biotic elements, climate, and time (Jenny, 1941). Human activities are another important contributor to the production of soil characteristics.

Although there is a certain amount of interdependence between the environmental factors, it is common for some of them to change independently of others from place to place. This has a consequent effect on soil processes and leads to considerable variation in soil profile form and properties within the lateral dimension. This lateral variability of soil is the main concern of the present study: unless specified otherwise, the term "<u>soil variability</u>" will be used to signify the <u>spatial variation</u> of soil properties, or whole profiles, within the <u>lateral</u> dimension.

Soil-forming processes and soil characteristics are often determined by a combination of all the environmental factors. In some cases, however, there is one dominant environmental element that governs the formation and distribution of soils in a specific region, i.e. soil variation may be depicted in terms of a governing soil-forming factor.

Precipitation and temperature are the main climatic variables that affect soil development, particularly in the way they determine the intensity of weathering and leaching processes. Vegetation is the chief biotic component that may influence soil variability. Under comparable climatic conditions, for instance, organic matter contents are higher and more uniformly distributed with depth in grassland than forest soil profiles (Foth, 1984). On a global scale, however, climate and natural vegetation cover are inter-related and soils are often distributed in a zonal pattern in accordance with biotic and climatic zones. This forms the basis for some genetic soil classification systems (cf. Section 2.3).

Parent material is another important factor that causes great variation in soil properties. A soil developed in basalt parent material, for instance, differs considerably in mineralogy and nutrient content from a soil on greywacke. Soils formed on transported materials tend to be more variable (less uniform) than those formed by weathering of bedrock in situ (Kantey and Morse, 1965). Considerable short-range variations in soil morphological, physical, chemical and hydrological properties are particularly evident in soils developed in alluvial parent materials. This aspect is discussed in more detail in Section 2.2.2.

Soil formation takes place over time with soils increasing in degree of development as they become older. Soils formed on old high alluvial terraces, for example, normally have well-developed horizons, whereas soils on newly-formed low terraces or floodplains tend to display little horizon differentiation (cf. Section 2.2.2). Soil variation due to other factors diminishes as time factor becomes dominant. Consequently, soils on older landscapes often exhibit less variability than soils on younger, dynamic landscapes where there is a range of depositional and erosional surfaces of different ages.

Topography influences soil formation and variation by the way it affects soil and water movement and modifies temperature, and moisture regimes (Birkeland, 1984). For instance, soils often vary in a zonal pattern in accordance with biotic and climatic changes with increase of altitude. The degree of soil development differs between south- and north-facing slopes because of the differences in temperature and moisture regimes. In the southern hemisphere, the northern aspect has higher temperatures and lower moisture than the southern aspect. Upper parts of slopes are often eroded and consist of shallow soils; the materials are deposited lower down forming relatively thick or buried soils (Birkeland, 1984). Within each large-scale topographic pattern, short-range changes in micro-relief may also give rise to frequent changes in soil properties, e.g. the hydrological changes down a slope resulting from lateral water movement (Beckett and Webster, 1971).

In most environments, a sixth factor needs to be considered when examining soil variability - human modification. Management practices (e.g. ploughing, subsoiling, grazing, and rotational cropping) introduce additional, normally short-range, spatial changes in soil physical or chemical properties. The degree to which management affects soil variability differs according to individual soil properties. Beckett and Webster (1971) suggested that those properties least affected by management are sand, silt and clay contents, plastic limits, and horizon thickness. More easily-modified properties include available P, Mg, Ca and K.

The general spatial variation in soil properties can sometimes be directly related to simple changes in a single factor. Soil property variation, however, is often determined by complex interactions of the factors, and it is frequently therefore a complicated and difficult task to analyse and interpret soil spatial variability for any specific areas.

Wilding and Drees (1983) divided the spatial variability of soil into two broad categories: systematic and random. Systematic variability is considered as a gradual or marked change in soil properties as a function of soil forming factors or management activities. The zonal distribution of soils related to biotic and climatic factors is an example of systematic variability. The operations of soil survey and pedological investigations can be facilitated by recognition and understanding of soil systematic variability.

Some soil properties also vary in a way that cannot be related to, or interpreted by, any known factors at the given investigation stage. Such changes are termed random, or chance, variation. The differentiation of systematic or random variation, however, is dependent upon investigation intensity and the knowledge of understanding about the soil studied. When the soil is studied in more detail, part of the variation formerly regarded as random, may become systematic and vice versa.

2.2.2 Soil variability within alluvial landscapes

Alluvial soils are renowned for their large lateral and vertical changes in texture, topographically-induced drainage patterns and differing degrees of development associated with different age surfaces. This variability is directly related to, and largely a function of the original alluvial deposition pattern. As this study is concerned with soils developed from alluvial sediments, it is necessary to consider the causes of such variation in more detail and, in particular, discuss alluvial deposition patterns.

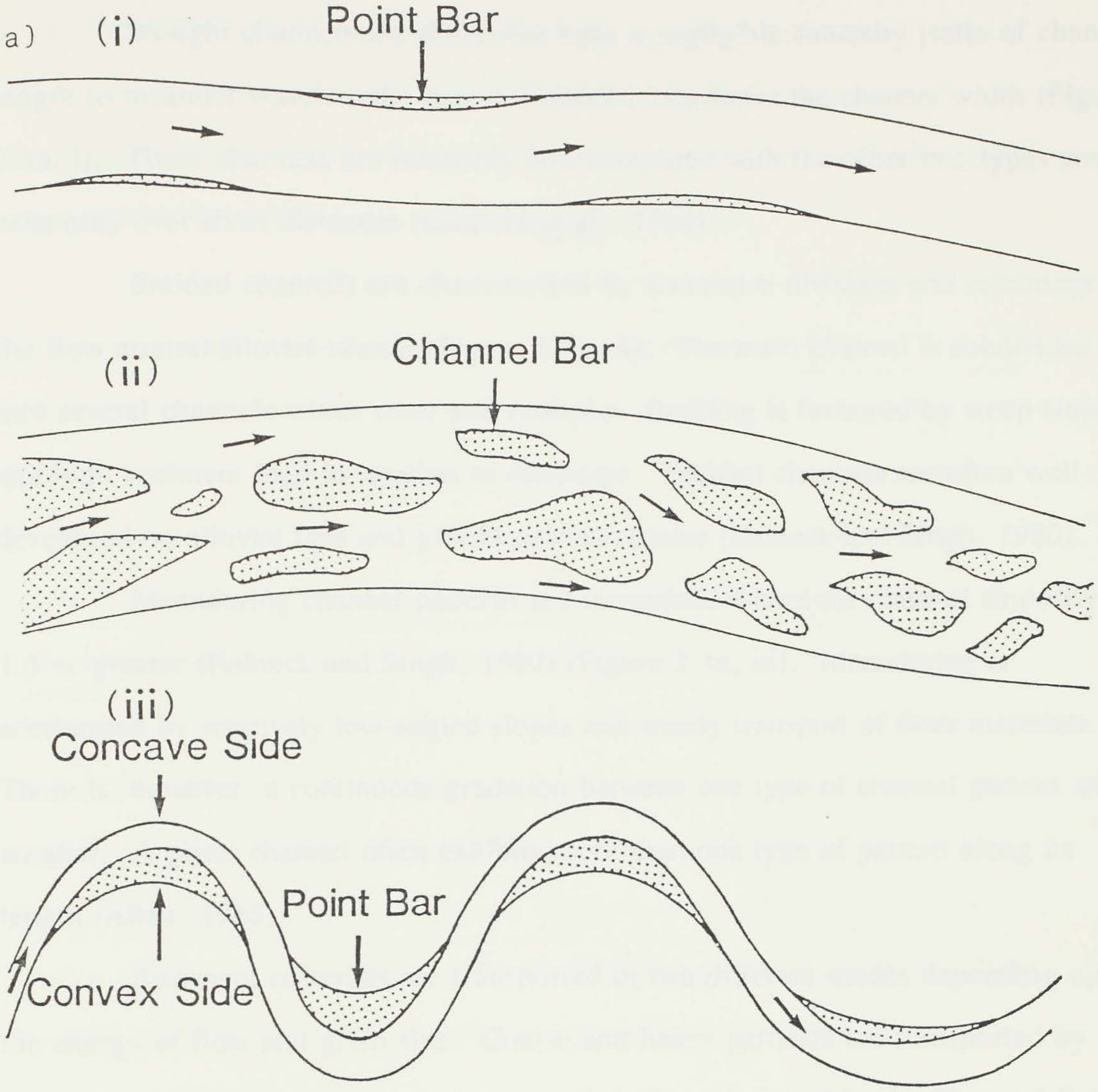
Rivers generally progress through three stages from their catchments to the coast: young, mature and old (Reineck and Singh, 1980). The young stage represents the beginning of the channel system and normally occurs in mountainous regions where small streams meet and grow into larger channels. Sediment materials are added and eroded at this young stage. The mature phase is regarded as the 'transfer' stage of sediments from catchments to the coasts; deposition also occurs at this stage and floodplains develop. Several floodplains of different channel systems meet together during the old stage in the coastal region, and the channels become smaller through repeated divisions. Delta deposits which are usually composed of fine materials may occur at the entrance of rivers into the sea or lakes. Fluvial deposition mainly takes place in these mature and old stages of the channel systems.

Channels flow in different patterns, depending on factors such as amount of water, velocity, sediment concentration and particle size. Three drainage patterns are commonly recognized: straight, braided and meandering (Allen, 1965; Reineck and Singh, 1980) (Figure 2.1).

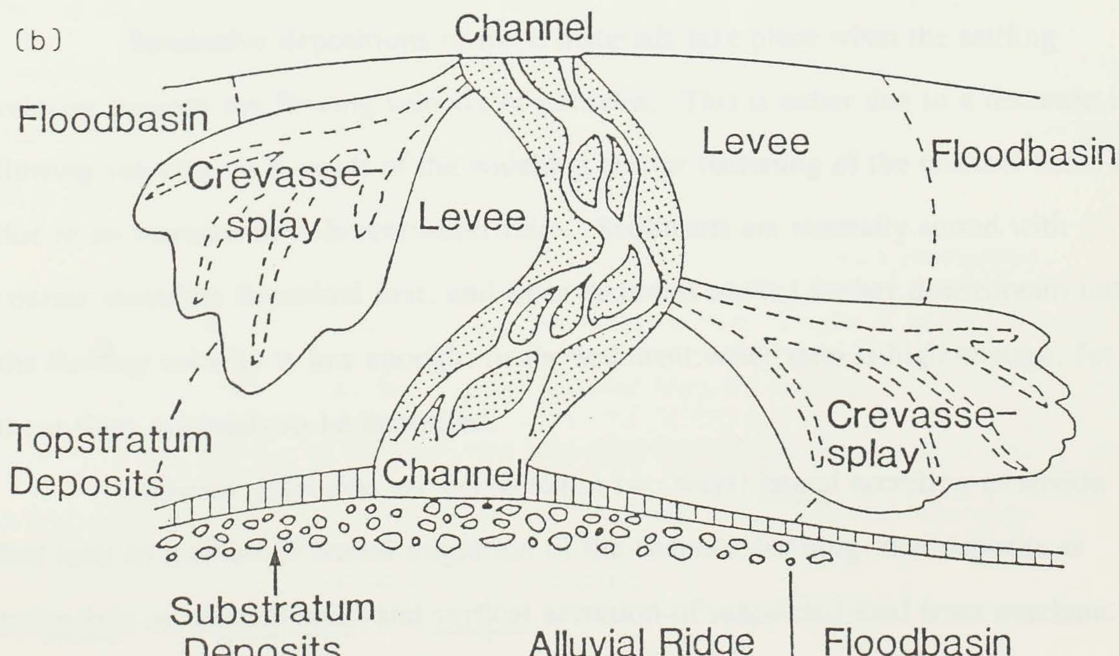
Figure 2.1 (a) Idealised river channel systems: (i) straight; (ii) braided; (iii) meandering (after Reineck and Singh, 1980)

(b) Spatial relationships of different alluvial deposits (after Karageorgis, 1980)

(a)


(i)
Point Bar
(ii)
Channel Bar
(iii)
Concave Side
Point Bar
Convex Side

(b)


Channel
Floodbasin
Crevasse-splay
Levee
Levee
Floodbasin
Crevasse-splay
Topstratum Deposits
Channel
Substratum Deposits
Alluvial Ridge
Floodbasin

Straight channels are those that have a negligible sinuosity (ratio of channel length to meander wavelength) over a distance many times the channel width (Figure 2.1a, i). These channels are relatively rare compared with the other two types and exist only over short distances (Leopold et al., 1964).

Braided channels are characterised by successive divisions and rejoinings of the flow around alluvial islands (Figure 2.1a, ii). The main channel is subdivided into several channels which meet and redivide. Braiding is favoured by steep slopes and high sediment load in relation to discharge. Braided channels are often well-developed on alluvial fans and glacial outwash plains (Reineck and Singh, 1980).

Meandering channel patterns are recognised where the channel sinuosity is 1.5 or greater (Reineck and Singh, 1980) (Figure 2.1a, iii). Meandering is accelerated by relatively low-angled slopes and steady transport of finer materials. There is, however, a continuous gradation between one type of channel pattern and another. A given channel often exhibits more than one type of pattern along its length (Allen, 1965).

Sediment materials are transported in two different modes depending upon the energy of flow and grain size. Coarse and heavy particles are transported by creeping, rolling or saltation processes, and are known as bed load. Fine and light particles are carried in suspension, and are termed the suspended load.

Successive depositions of these materials take place when the settling velocity exceeds the flowing velocity of particles. This is either due to a decrease in flowing velocity, as a result of the widening and/or flattening of the channel beds, or due to an increase in sediment:water ratio. Sediments are normally sorted with coarser materials deposited first, and finer materials carried further downstream until the flowing velocity is low enough, or the sediment:water ratio is high enough, for these finer materials to be deposited.

Alluvial sediments are deposited in two ways: lateral accretion of stream bed load as a result of lateral migration of the channel, forming such deposits as point bars or channel bars, and vertical accretion of suspended load from overbank floods, creating levees, crevasse-splays and floodbasins (Allen, 1965). These sediments are classified into three major categories and eight subordinate types as shown in Table 2.1. The various terms used in the table are comprehensively discussed by Allen (1965) and Reineck and Singh (1980).

Table 2.1 Classification of alluvial sediments (after Allen, 1965)

Environment of deposition	Deposit	Categories
Channel floor	Channel-lag deposit	Channel or
Point bar	Point bar deposit	substratum
Channel bar	Channel bar deposit	deposits
Point bar swale or abandoned braided stream channel	Swale-fill deposit	Overbank or topstratum
Levee	Levee deposit	deposits
Crevasse-splay	Crevasse-splay deposit	
Flood basin	Flood basin deposit	
Within abandoned channel	Channel-fill deposit	Transitional deposit

Point bar deposits are derived from the lateral accretion of sediments on the convex side of channel meanders (Figure 2.1a, iii). They are the most conspicuous sedimentation feature of meandering channels with textures varying from clay to gravels. Grain size often decreases upward in a point-bar sequence, changing from gravels through sand to silty or clayey textures.

Channel bars are created by lateral and vertical accretion of braided channels together with channel cutting and abandonment (Figure 2.1a, ii). Coarse-textured (e.g. pebbles) and fine-grained channel bars are commonly recognised. The slope of a channel bar in the upstream direction is normally steeper than that in the downstream direction, and often has a pool in front of it. Channel bars also commonly exhibit a fining-up textural sequence.

Natural levee deposits are formed by deposition of sediments when flood water overtops the river banks. Coarser sediment (e.g. sand) is deposited in the form of ridges near the channel and the particle-size fines away from the channel. grading into flood basin deposits.

Flood basin deposits occur in the lowest part of alluvial plains where the suspended fine sediment settles down from overbank flows. The sediments tend to be dominated by silt and clay particles. As topography changes from the coarse-textured levee ridges to the fine-textured floodbasins. morphological features (e.g. mottling) and hydrological properties change accordingly.

Crevasse-splay deposits are formed during high flood stages of rivers when large quantities of water and transported load cut through levees and divert into adjacent floodbasins. The particles in crevasse-splay deposits are as coarse as. or even coarser than. the associated natural levee deposits. The crevasse-splay deposits extend across the levees as sandy tongues and into the floodbasin. The three types of deposits form an interfingering spatial pattern (Figure 2.1b) and provide a texturally-variable soil parent material.

Channel-fill deposits are the only type of transitional deposit (Table 2.1): they are due to sedimentation in channels that have been abandoned by a stream or river. The abandonment of channels may be due to filling up as a consequence of extreme increases in sedimentation rates. Alternatively. it may be associated with cut-off processes which occur whenever a meandering stream can shorten its course and locally increase its slope (Reineck and Singh. 1980). Two different types of cut-off processes are commonly recognized: chute cut-off where a stream shortens its course by taking up a new channel along a swale of its convex side. and neck cut-off. where a stream cuts a new channel through the narrow neck of the meander loops. The abandoned channels may then be filled up in various ways including by overbank deposition.

The distribution of deposited materials is complicated by changes in river flow characteristics and channel migration over time. causing erosion and burial or overlap of different types of sediments. Considerable vertical and lateral variation in texture therefore occurs in soils developed in such parent materials. This, in turn, induces rapid changes in related soil physical properties, e.g. bulk density, hydraulic conductivity and moisture content.

Several workers have demonstrated that soils developed in alluvial parent materials exhibit great variability in soil properties. Mausbach et al. (1980), for instance, observed that alluvial soils were more variable in physical properties than soils in loess and glacial drift parent materials. Similar conclusions were made by Drees and Wilding (1973) and Wilding and Drees (1983). Studies by Karageorgis (1980) on soils developed on the Canterbury alluvial plains indicated that soil morphological features (e.g. subsurface textural layers) were extremely variable.

Tectonic uplifting of the land or eustatic lowering of sea level, causes rivers to incise in order to maintain their longitudinal profiles. Terraces are depositional and/or erosional surfaces of old river beds and floodplains which have been relatively uplifted to higher positions because of the downcutting of the rivers. Terraces often display a height-age relationship with the surfaces becoming progressively younger with decreased elevation above the river channel. Terraces on both sides of the rivers, however, are not necessarily symmetrical; unequally paired terraces may be produced by river channel migration, or erosion of terraces on one side. The number of terraces may also differ among the different reaches of rivers. Soils developed on river terraces of different ages often comprise a chronosequence: high-terrace soils are usually more strongly developed than soils on lower terraces (Gerrand, 1981).

In summary, the variability of alluvial soils can be ascribed to three main factors: sediment composition, topography and age. Most soils developed from alluvial sediments display great lateral and vertical variation in soil texture, characteristics which are a function of the complex depositional environment. Textural and topographic changes cause other soil physical and morphological properties to vary accordingly. Soils developed on surfaces (terraces) of different ages vary in terms of their degree of development. Under uniform climatic conditions, soil formation and distribution within alluvial landscapes is largely a function of the interaction of these three factors. They form the basis of numerous soil classification schemes for soils developed in alluvial parent materials (cf. Sections 3.2.4 and 3.3).

2.3 Soil classification

Soil classification involves the identification of soil individuals within the soil mantle and the allocation of these individuals to classes with common characteristics (Bridges and Davidson, 1982). The soil mantle is a continuum that extends over a range of conditions and its properties vary accordingly. It often lacks sharp discontinuities and, therefore, it is not always clear what exactly are the objects, or entities, that are to be identified and classified (Butler, 1980).

Traditionally, the basic unit employed in soil classification and soil survey is the soil profile, which comprises a vertical section from the surface to the parent material. In the English system of soil classification, for example, the soil profile is defined as the soil mantle up to about one square metre in cross sectional area, ranging from the ground surface to a maximum depth of 1.5 m (Avery, 1973).

As a soil profile is only two-dimensional, it theoretically cannot be used to make a soil map because soil individuals are within the soil continuum, and ideally should be three-dimensional. A new concept, called a pedon has therefore been adopted in U.S. Soil Taxonomy (Soil Survey Staff, 1975). The pedon is described as the smallest volume which may be called a soil and is defined as having lateral dimensions large enough to include the natural variation of the horizons present. The area of a pedon ranges from 1 to 10 m^2 depending upon the variability. Pedons are grouped into polypedons. A polypedon is defined as one or more contiguous similar pedons that are bounded on all sides by "not soil" or by pedons of unlike character (Soil Survey Staff, 1975). The pedon is the unit of sampling and study, whereas the polypedon is the unit of classification.

It has been claimed by some authors that the pedon concept gives a more acceptable basic unit of study than the soil profile (Bridges and Davidson, 1982). Nevertheless, as Butler (1980) stated, such discussion has had little effect on soil survey operations. Most studies on soil classification and survey are still based upon soil profiles.

Individual soil profiles studied in the field are grouped into a number of soil profile classes: each should ideally be defined by a set of morphological properties, though more subjective landscape or environmental factors have in the past often been given equal or greater importance. With further characterisation and formalisation, both in the field and laboratory, a soil profile class can be designated as a taxonomic unit, one of the hierarchical classes within a traditional soil classification system. A soil taxonomic unit is normally defined by a number of diagnostic soil properties, in terms of a modal profile form. The properties used for definitions are normally those relatively permanent characteristics that are not readily modified by management.

There are two kinds of soil classification systems: technical and natural. A technical system is one that is designed for a single purpose (e.g. irrigation) and only considers those properties relevant to that purpose. The natural system stresses the origin and relationships between classes, and makes use of as many known properties of the soils as possible without a specific objective (Cutler, 1977; Bridges and Davidson, 1982). Most of the classification systems currently being used in the world belong to the second group.

One of the important assumptions made in soil classification is that soil differences can be adequately characterised by relatively few chosen attributes. The properties that are used to define categories and taxa of soil classification systems are those (hopefully) which have greatest independence of variation from each other, but are highly correlated with many other accessory properties. The variances within taxonomic units so defined are minimised over all properties. Successively lower levels of hierarchical classification systems produce more homogeneous classes as a result of the more specific definitions for lower categories. Such important assumptions, however, have rarely been justified by researchers.

Environmental and pedogenetic criteria have been used to define soil categories and taxa at certain levels in some soil classification systems such as in U.S.S.R. (Rozov and Ivanova, 1967), France (Duchaufour, 1982), New Zealand (Taylor, 1948), and Marbut's prewar soil classification system in the U.S. (Buol et al., 1980). Soil classes defined in terms of environmental parameters or pedological processes, however, have been criticised as being ambiguous and lacking quantitative and objective criteria: they are, more often than not, a classification of environment rather than soils (Kellogg, 1963; Smith, 1968). Consequently, in most recent soil classification systems, such as the U.S. Soil Taxonomy (Soil Survey Staff, 1975), the Canadian (Canada Soil Survey Committee, 1978) and English systems (Avery, 1973), soil properties which are observable and measurable in the field, or in the laboratory, are used as differentiae for definitions of soil classes. The diagnostic criteria used have been criticised as being too subjective and arbitrary, however, since little work has been done to justify the use of few soil characteristics as the basis of predicting all other soil properties (Butler, 1980). Soils grouped together in terms of the chosen diagnostic properties might be clearly different in some other accessory soil properties.

Numerical methods of soil classification are increasingly being tried to create classes and demonstrate relationships (Webster, 1977; Bridges and Davidson, 1982). This approach is based on numerical analyses of soil properties and mathematical determination of appropriate relationships between individual soil classes. The advantage of this approach over the traditional method lies in the large volume of soil data that can be integrated and generalised by computers; quantitative criteria can therefore be readily used as differentiae for classification. The rapid development of computer software and the recognised desire to provide a more quantitative basis to soil classification suggests that numerical techniques may play an increasingly important role in future soil classifications. There are, however, a number of technical problems still to be overcome before it becomes universally accepted.

2.4 Soil mapping

2.4.1 Aims

Soil classification provides the theoretical basis for soil survey operations. The most important product of soil survey is the soil map, in which the landscape is resolved into areas (blocks or parcels) that can be managed uniformly for the purposes to be served by the map. According to Dent and Young (1981), "the practical purpose of soil survey is to enable more numerous, more accurate and more useful predictions to be made for specific purposes than could have been made otherwise". Therefore, the task of soil survey is to analyse the pattern of the soil mantle and divide the pattern into relatively homogeneous units, to map the distribution of these units so that "the properties of soils over any area can be predicted, and to characterise the mapped units in a way that useful statements can be made about the land use potential and response to changes in management" (Dent and Young, 1981).

The relatively "uniform" soil bodies delineated on soil maps comprise soil mapping units. These mapping units represent real geographical areas, and are thus clearly different from soil taxonomic units which are conceptual and normally have no particular spatial connotations in themselves. The variability in most soil properties, in particular, soil diagnostic properties within each mapping unit should be substantially less than that of the whole region, so that each unit can be managed uniformly.

2.4.2 Methods

Soil survey procedures consist of three stages: preliminary work, field survey, and preparation of soil maps and reports. Detailed discussions of these procedures are given by Taylor and Pohlen (1979), Dent and Young (1981), and Bridges and Davidson (1982).

For obvious practical reasons, a soil map has to be compiled from a limited number of direct observations, whether they be pits, road exposures or auger holes. Consequently, one of the most important decisions to be made during field survey operations concerns the location of observation points. There are basically two strategies: grid survey or free survey. Grid survey is where observations are made on the intersections of a grid or at fixed intervals along a line. This approach is especially useful in the situations where the area to be surveyed is covered by thick forest and therefore the landscape features are not visible, or where the sampling area does not contain any surface features to guide the delineation of soil boundaries. This technique is normally considered as most suitable for soil surveys at scales of 1:10,000 or greater (Bridges and Davidson, 1982). Free survey necessitates direct observations to be made at sites determined by the surveyors' comprehension of the environmental factors and the soil-landscape relationships. In this case, many of the soil boundaries are predicted and the observations located to check the boundary predictions, and to characterise the properties of individual units. The great advantage of this method is that the surveyor is free to change the intensity of observations in accordance with the complexity of the soil pattern.

Soil surveys are carried out at a number of different scales. In New Zealand the three main types are referred to as general surveys (1:253,440), district surveys (1:126,720), and detailed surveys (1: 31,680) (Taylor and Pohlen, 1979). Small-scale soil surveys depict broad variations in soils that are related to environmental features. Local or short-range variations of soil properties can only be distinguished by more detailed soil surveys. The scale adopted in specific surveys is dependent on many factors, such as the survey purpose, the size of the area to be surveyed, the complexity of soil distribution and the amount of effort that is afforded. The type of soil mapping unit employed often differs according to the scale of survey.

2.4.3 Soil mapping units

The terms soil series and soil type have been used in the literature to represent both taxonomic and mapping units. taxonomic units. however. are conceptual units defined at any level or category in a soil classification system. The soil series taxonomic unit used in New Zealand is a group of soils with similar modal profiles. similar temperature and moisture regimes. and the same or very similar parent materials (Taylor and Pohlen. 1979). Soil types are subdivisions of soil series. They differ from each other in such properties as texture. slope. stoniness. degree of erosion. and topographic positions. Soil mapping units. on the other hand. are real soil areas that are distinguished and delineated on soil maps to represent relatively homogeneous soil bodies (Dent and Young. 1981). Soil profiles within any mapping unit will normally conform to the definitions of its designated taxonomic unit. though. there will often be discrepancies. Templeton series as a soil mapping unit. for instance. is a real soil area dominated by soil profiles which fulfill the criteria of a Templeton series taxonomic unit. It may. however. include small areas in which other soil taxonomic units (e.g. Eyre and Wakanui series) occur. Thus. any soil mapping unit may be spatially variable in accordance with the accepted range of diagnostic properties and other changes in accessory properties of its designated taxonomic unit(s). Further variation in soil properties within the mapping unit. however. is caused by the presence of taxonomic impurities.

Soil series and soil types are simple mapping units in which there is only one dominant taxonomic unit. They may include small areas - 10-15% - of other taxonomic units (Taylor and Pohlen. 1979). The amount of inclusions allowed in simple mapping units differs from country to country (cf. Section 2.4.4).

Compound mapping units contain appreciable amounts of two or more taxonomic units. They are used in places where the soil pattern cannot be depicted by simple mapping units. because of survey scale. field observation density afforded or the spatial complexity of soil property distribution. Three compound mapping units are used in New Zealand.

(1) Soil set: compound mapping units devised for general soil surveys (scale 1:256,400) of New Zealand (Cutler, 1977). They consist of soils with similar profiles occurring in distinct landscape units.

(2) Soil association: group of geographically associated soils, each of which is confined to a particular facet of the landscape and which occur in a repeating and predictable pattern (Cutler, 1977; Dent and Young, 1981).

(3) Soil complex: Compound mapping unit which contains a mixture of two or more taxonomic units that do not occur in a predictable pattern, and therefore cannot be separated at the survey scale used (Taylor and Pohlen, 1979). The soil complex is of limited predictive value, and is therefore used only as a last resort (Dent and Young, 1981).

2.4.4 Mapping unit purity

The precision of any generalisation or prediction of soil properties within a mapping unit depends largely on the amount of impurities within the unit. The usefulness of mapping units in making statements upon land uses is often judged by mapping unit purity and how seriously the impurities affect land uses. The purity of simple mapping units is defined as the average percentage of the area of each unit which is occupied by its eponymous taxonomic unit (Beckett and Webster, 1971).

Mapping unit purity is, to some extent, implied by the kind of mapping units used in the soil map, i.e. simple mapping units are more "pure" than compound mapping units. The amount of inclusions allowed in simple mapping units, however, varies between different countries. For example, a purity of 85% is attempted by the U.K. (Avery, 1964) and New Zealand (Taylor and Pohlen, 1979), whereas a 70% or more purity is required in The Netherlands (Buringh et al., 1962). In the U.S.A. (Soil Survey Staff, 1980), a 85% purity is required if the inclusions limit land use and management, whilst a 75% purity is permitted when the inclusions do not provide any such limitations. The actual impurity often exceeds the desired limits because only a few field observations per mapping unit can be afforded. Beckett and Webster (1971) concluded that soil series and type mapping units were generally about 50% pure, though some taxonomic inclusions differ only in minor definitive features from the dominant taxonomic unit, and thus not all the 50% impurity requires different management. Studies by Adams and Wilde (1980) indicate that the purity of the Westmere silt loam mapping unit in the Wanganui district of New Zealand is only about 58%; the requirement of 85% purity was considered unrealistic. They suggested that a 50% purity is more reasonable in most mapping units at the series or type level in New Zealand. The relative proportions of the members within compound mapping units (e.g. associations or complexes) are sometimes, but not always, recorded.

The definition of simple mapping unit purity, as discussed above, is based on the relative proportion of taxonomic units within a mapping unit. Soil taxonomic units are defined in terms of certain definitive properties. Non-definitive soil properties, however, may vary spatially at different rates to the definitive soil properties of taxonomic units. Beckettt and Webster (1971), for instance, found that the variability of non-definitive properties within a mapping unit is not necessarily as wide as the range of definitive properties: conversely, some non-definitive properties may be more variable than the definitive properties and thus do not change in line with taxonomic units. Giltrap et al., (1983) also concluded that different properties may display very different patterns of variation. Homogeneity of any area in terms of one set of properties (e.g. morphological), does not imply that the same area is homogeneous in other soil properties (e.g. chemical). Miller et al. (1979) suggested that mapping unit purity is not a proper measure of quality or precision for soil survey: there is an increasing demand by users of soil survey for quantitative appreciation of spatial variability with known confidence limits for specific soil properties and soil performance within mapping units.

2.5 Conventional statistics

2.5.1 Introduction

Conventional statistics is often used to assess the precision of mean values as estimates of soil properties at unsampled locations within sampling units, to quantify the variability of soil properties within or between soil taxonomic and mapping units, and to assess the effectiveness of soil classification and quality of soil survey. Conventional statistics assumes that the variation of soil properties is randomly distributed within sampling units, i.e. there is no spatial dependence between observations, and the parametric statistical analyses assume normal probability distribution. Soil properties that exhibit drastic departures from normality need to be transformed into normal distributions prior to the parametric statistical analyses, or to be analysed using the less efficient method of non-parametric inference where the normality of distribution is not required. Comprehensive discussions of the applications have been presented by Beckett and Webster (1971), Webster (1977) and Wilding and Drees (1983). Only the basic concepts that are relevant to this thesis will be dealt with in the following sections.

2.5.2 Estimation of soil properties

Estimates of soil property values in specific areas, and the degree of confidence that can be achieved by such estimates, are often required by soil surveyors, researchers and land users. The estimation of soil properties using conventional statistics is based on the assumption "that the expected value of a soil property z at any location x within a sampling area is

$$z(x) = \mu + \varepsilon(x) \qquad (2.1)$$

where μ is the population mean or expected value of z, and $\varepsilon(x)$ represents a random, spatially uncorrelated dispersion of values about the mean. Deviations from the population mean are assumed to be normally distributed with a mean of zero and a variance of σ^2" (Trangmar et al., 1985). The variance is defined as follows and its square root is known as the population standard deviation (σ):

$$\sigma^2 = 1/N \sum_{i=1}^{N}(x_i - \mu)^2 \tag{2.2}$$

$$\sigma = \sqrt{[1/N \sum_{i=1}^{N}(x_i - \mu)^2]} \tag{2.3}$$

where N is the number of observations, x_i is the ith observation and μ is the population mean.

Since soil is a continuous mantle, observations and measurements can only be made on limited numbers of sites within the soil population. Thus, a mean derived from all members of a population cannot be obtained directly from observations. Instead, the sample mean ($\bar{x}$) is used to estimate the population mean (μ) and to represent values of soil properties at unsampled locations within the sampling area. With a set of observations, x_1, x_2,.... x_n, the mean $\bar{x}$ is defined as

$$\bar{x} = (1/n) \sum_{i=1}^{n} x_i \tag{2.4}$$

where $\bar{x}$ denotes the sample mean, and n is the number of observations (Snedecor and Cochran, 1980).

Statistical theory shows that if repeated random samples of size n are drawn from any population with mean μ and standard deviation σ, the frequency distribution of the sample mean in these repeated samples has mean μ and standard deviation $\sigma/\sqrt{n}$. The sample mean $\bar{x}$ is therefore an unbiased estimator of μ under random sampling (Snedecor and Cochran, 1980).

Furthermore, the frequency distribution of $\bar{x}$ in repeated random samples of size n tends to become normal as n increases irrespective of the shape of the frequency distribution of the original population (Snedecor and Cochran, 1980). This explains why the normal distribution, and results derived from it, are so commonly used with sample means, even when the original population is not normal.

The standard deviation of $\bar{x}$, $\sigma/\sqrt{n}$, is often called the standard error of $\bar{x}$. It supplies information about the amount of error in $\bar{x}$ when it is used to estimate μ. It can be used to determine the range within which the true population mean μ lies with any desired degree of confidence. Assuming that the sample is large (generally $n \geq 30$) and σ is unknown, the sample standard deviation s can be used to replace σ.

$$s = \sqrt{[1/n-1 \sum_{i=1}^{n}(x_i - \bar{x})^2]} \tag{2.5}$$

and the confidence limits for the population mean are estimated as

$$\bar{x} - z\, s/\sqrt{n} \leq \mu \leq \bar{x} + z\, s/\sqrt{n} \tag{2.6}$$

where the quantity z is the value of the normal deviation for the desired level of confidence, and can be obtained from tables listed in most statistics books. Frequently used values (Webster, 1977) are:

Confidence (%)	75	80	90	95	99
z	1.15	1.28	1.64	1.96	2.58

For example, one can be 95% confident that the true mean μ lies in the following range when the sample mean is used to predict values of soil properties at any location within the examined area.

$$\bar{x} - 1.96\, s/\sqrt{n} \leq \mu \leq \bar{x} + 1.96\, s/\sqrt{n}$$

This approach is based on the fact that the sample mean is approximately normally distributed as $N(\mu, \sigma/\sqrt{n})$, when n is large. In many circumstances where experiments are costly and time-consuming, however, measurements can only be taken from samples of limited size. When the sample size n is small, and σ is not available and has to be replaced by s, the confidence interval cannot be determined using the above method because the sample means may change substantially from a normal distribution. In this case, the determination of confidence interval is based on the "student's t-distribution" (Bhattacharyya and Johnson, 1977). As a consequence, the quantity z in equation 2.6 is replaced by student's t, which is also listed in most statistics books. The confidence interval now becomes

$$\bar{x} - t\, s/\sqrt{n} \leq \mu \leq \bar{x} + t\, s/\sqrt{n} \qquad (2.7)$$

The specific confidence intervals required are dependent on many factors, such as the property in question, the magnitude of the mean, and the risk one is willing to take in making an error in judgement (Wilding, 1985). A confidence level of 99% or 95% is common in many studies. Wilding (1985), however, suggested that a confidence level of 70 to 80% is probably more realistic in soil surveys in terms of time and money inputs that are practical to a sampling scheme.

2.5.3 Assessment of soil variability

One of the most commonly used estimates of soil variability is the coefficient of variation (C.V.):

$$C.V. = s/\bar{x}\ .\ 100\% \qquad (2.8)$$

This relative measure, expressed as a percentage, is often used to contrast the variability of different soil properties within similar sampling units, or of the same soil properties between different sampling entities (e.g. taxonomic units, mapping units, or experimental plots).

Wilding and Drees (1983) provided a comprehensive summary of the magnitude of variability in soil morphological, physical, and chemical properties in terms of C.V. values within similar sampling units, i.e. pedons, series taxonomic and mapping units. They found that soil chemical properties such as exchangeable Ca, Mg and K tend to be extremely variable (mean C.V. values = 50-70%). Bulk density and water content are commonly much less variable (C.V. = 10-20%) than other soil physical properties such as soil hydraulic conductivities (C.V. = 50 - 150%). Wilding and Drees (1983) divided the magnitude of soil variability expected within soil series mapping units of a few hectares, or less, into three categories. The properties with least variability (C.V. < 15%) included soil pH, and thickness of A horizons. Soil properties such as total sand or clay content, and soil structure were considered as moderately variable parameters (C.V. = 15-35%). The most variable soil properties (C.V. > 35%) included depth to mottling, organic matter content and hydraulic conductivity.

Beckett and Webster (1971) and Wilding and Drees (1983) concluded that C.V. values for any soil diagnostic property increased as the sampling entity changed from pedon to series taxonomic unit and finally to the corresponding mapping unit.

Caution, however, should be taken in interpreting the significance of C.V. values. When there is a direct relationship between the magnitude of x (property value) and s (standard deviation) (i.e. they covary), then C.V. is an invalid index. Problems also occur with log-transformed data, or where data may have both positive and negative values with a consequent mean of zero (Wilding and Drees, 1983).

The comparison of variability of certain soil properties between different sampling units can also be achieved by calculating variance ratios and using the <u>F-test</u>.

Assume that $x_1, \ldots, x_{n1}$ and $y_1, \ldots, y_{n2}$ are two independent random samples from two sampling areas and the sample variances are

$$s_1^2 = \sum_{i=1}^{n_1} (x_i-\bar{x})^2/(n_1-1) \qquad (2.9)$$

$$s_2^2 = \sum_{i=1}^{n_2} (Y_i-\bar{Y})^2/(n_2-1) \qquad (2.10)$$

the variance ratio can be calculated (equation 2.11) and compared with tabulated F values (Webster, 1977; Bhattacharyya and Johnson, 1977).

$$F = s_1^2/s_2^2 \qquad (2.11)$$

The advantage of the F-test over the method of comparing C.V. values is that it provides a test of significance between the variances from different sampling areas. Thus, one sampling unit is more variable than the other in the examined properties if the test is significant at an accepted level.

2.5.4 Effectiveness of soil classification and mapping

The aim of soil mapping is to isolate areas which are individually more homogeneous than the region as a whole. Ideally, the variability of most soil properties within delineated areas should be significantly less than the variability between mapping units or the variability of the whole area. In addition, mean values of most soil properties should differ significantly between mapping units. Two factors determine whether these aims actually materialise: the effectiveness of soil classification and the quality of soil survey. An effective classification is one where soils classified in terms of a few chosen properties differ significantly in most other soil properties, whilst a high quality survey should ensure that each mapping unit contains as few taxonomic units as possible.

The significance of differences in mean soil property values between individual taxonomic or mapping units can be assessed using the t-test. If $\bar{x}_1$ and $\bar{x}_2$ represent the mean values of soil properties from two sampling units, and n_1, n_2 are the sampling numbers, the t value can be computed from the following equation and compared with the tabulated t values to determine whether the compared mean values are significantly different at specified confidence levels (Bhattacharyya and Johnson, 1977).

$$t = (\bar{x}_1 - \bar{x}_2)/s_{pooled}\sqrt{(1/n_1 + 1/n_2)} \qquad (2.12)$$

where s_{pooled} is the combined estimation of variances from the two samples. It is defined as

$$s_{pooled} = \sqrt{[(n_1-1)s_1^2 + (n_2-1)s_2^2/(n_1+n_2-2)]} \qquad (2.13)$$

Equations (2.12) and (2.13) are for samples of small sizes. They are based on the assumptions that both distributions are normal and the population variances σ_1^2 and σ_2^2 are equal. Where both sample sizes are large, the assumptions for small samples are no longer necessary, and the calculation can be done using the following equation (Bhattacharyya and Johnson, 1977):

$$z = (\bar{x}_1-\bar{x}_2)/\sqrt{(s_1^2/n_1+s_2^2/n_2)} \qquad (2.14)$$

Analysis of variance is often used to partition and compare the components of variances from different sampling units. The total variance from the whole sampling area, the variance within classes (taxonomic or mapping units) and the variance between classes are first computed (Table 2.2). Then the F-ratio is calculated (equation 2.15) and compared with the tabulated F-values to determine whether the classification or mapping is effective (Webster, 1977). If the F-test is significant at a certain desired level, there are significant differences in the examined properties between the classes. This F-test can be used to compare more than two classes. In situations where there are only two classes, the F-test is equivalent to the t-test discussed above.

Table 2.2 Analysis of variance (after Webster, 1977)

Source	Degrees of freedom	Sum of squares	Mean squares
Between classes	k-1	$\sum_{i=1}^{k} n_i(\bar{x}_i-\bar{x})^2$	$1/(k-1)\sum_{i=1}^{k} n_i(\bar{x}_i-\bar{x})^2 = B$
Within classes	N-k	$\sum_{i=1}^{k}\sum_{j=1}^{n_i}(x_{ij}-\bar{x}_i)^2$	$1/(n-k)\sum_{i=1}^{k}\sum_{j=1}^{n_i}(x_{ij}-\bar{x}_i)^2 = W = s_W^2$
Total	N-1	$\sum_{i=1}^{k}\sum_{j=1}^{n_i}(x_{ij}-\bar{x})^2$	$1/(N-1)\sum_{i=1}^{k}\sum_{j=1}^{n_i}(x_{ij}-\bar{x})^2 = T = s_T^2$

where N is the total numbers of the sample

k is the number of classes, each contains n_i observations

$\bar{x}$ is the mean of the whole area

$\bar{x}_i$ is the mean of the ith class

B, $W(s_W^2)$, and $T(S_T^2)$ are symbols for the three mean squares.

$$F = B/W \tag{2.15}$$

where B and W (calculated from Table 2.2) are the between and within class variances respectively.

Based on the calculations in Table 2.2 the effectiveness of soil classification and mapping can be further assessed using the following expression (Webster, 1977):

$$1 - s_W^2/s_T^2 \tag{2.16}$$

The value derived from the above expression is regarded as the proportion of total variance accounted for by classification or mapping. A more complicated assessment is given by the following equation (Webster, 1977):

$$r_i = s_B^2/(s_W^2 + s_B^2) \tag{2.17}$$

where s_B^2 is defined as

$$s_B^2 = (B - s_W^2)/n \quad (2.18)$$

It should be noticed that s_B^2 is slightly different from B in Table 2.2 in that the within-class variance is removed from B values, and therefore s_B^2 only accounts for the variance derived from the class means.

These two approaches produce very similar results (Webster, 1977). Theoretically, r_i could have a maximum value of 1, which means each class is uniform ($s_W^2 = 0$) and all the variances are accounted for by classification or mapping. The best soil classification system or soil map, therefore, will be the ones with the largest value of r_i in most soil properties.

Wilding et al. (1965) used the analysis of variance method and found that differences in horizon thickness between mapping units were not significant for A horizons but were significant for B horizons in their study area. Beckett and Webster (1965a, b) tested a soil classification system using this technique, and concluded that there were significant differences between classes in the plastic limit of soils. In the same studies, Beckett and Webster also found that a simple soil classification based on profile morphology, physiography or geology could account for approximately half of the variance in the physical properties of soil in a particular region (i.e. r_i approximates to 0.5). Subsequent studies have shown that accessory soil chemical properties are not as easily differentiated as soil morphological or physical properties by soil classification. For instance, Webster and Beckett (1968) reported r_i values of 0.06 for available K, 0.09 for available P, and 0.33 for pH.

It is clear that r_i values will differ substantially for each soil property. This is a problem for soil surveyors or land users wishing to interpret accessory soil properties, or to predict soil responses to land use practices, on the basis of diagnostic properties used for soil classification and mapping. In order to improve the quality of classification and mapping, so that the predictions and interpretations can be made more precisely, it is essential to have some appreciation of the relationships between the different soil properties.

The degree of dependence between different soil properties can be expressed by a parameter known as the correlation coefficient (Webster, 1977). As stated earlier, the diagnostic soil properties used for classification and mapping should ideally be those that have least correlation to each other, but which are highly correlated with accessory soil properties. Soil classification and mapping undertaken on these premises should require the least effort, but give the best result.

The correlation coefficient r is defined as

$$r = c/\sqrt{s_1^2 \, s_2^2} \qquad (2.19)$$

where s_1^2 and s_2^2 are the variances of the two examined properties and c is the covariance of the two properties (Webster, 1977):

$$c = 1/(n-1) \sum_{i=1}^{n} (x_{i1}-\bar{x}_1)(x_{i2}-\bar{x}_2) \qquad (2.20)$$

The correlation coefficient has a value between +1 and -1. The two soil properties are said to be perfectly correlated if r equals +1 or -1. There is no correlation between the two variates if r equals zero. Studies by McKeague et al. (1971), Moore et al. (1972) and Webster and Butler (1976) have shown that correlation coefficients between properties recorded in routine surveys range from 0.3 to -0.3: few r values exceed 0.5. If two properties are strongly correlated, only one of them should be recorded. Banfield and Bascomb (1976) reported a study where the correlation coefficient was 0.34 between dithionite extractable Fe and matrix chroma. Bulk density showed no correlation with pores of size observable with a hand lens. Clay content, however, was slightly correlated with consistency. It often has difficulties, however, to quantitatively correlate morphological properties with analytical as the former is normally qualitatively described.

2.6 Regionalised variable theory

2.6.1 Introduction

Conventional statistics assumes that variations in soil properties are randomly distributed within sampling units, and that the sampling mean can be used to predict values of soil properties on any unsampled sites within the units. Soil properties, however, are often continuous variables and tend to be correlated over vertical and horizontal space (Burgess and Webster, 1980a; Trangmar et al., 1985; Warrick et al., 1986). Soil samples close together tend to be more alike than samples far apart, because soil properties exhibit spatial dependence within some localised region. Thus, the conventional model is inadequate for interpolation of spatially dependent variables, because it takes no account of spatial correlation and relative location of samples.

The recently developed regionalised variable theory takes into account both the random and structured characteristics of spatially correlated variables, and provides a quantitative tool for assessing the spatial dependence of soil properties. This new statistical theory, also known as geostatistics, was developed by Matheron (1963, 1965, 1971) and Krige (1966) for the estimation of ore reserves in the mining industry. It was only recently that the theory has been applied to soil variability studies. The main applications here involve the quantification of soil spatial dependence (by means of semi-variograms), interpolation or extrapolation of soil properties (kriging) and the determination of soil sampling strategies (Burgess and Webster, 1980a, b; Webster and Burgess, 1980; Burgess et al., 1981; Webster, 1985; Trangmar et al., 1985).

2.6.2 Assumptions

The regionalised variable theory is based on a number of assumptions, and these are outlined below.

If the expected value of a random variable $z(x)$ is the same throughout the study region, then the variable is said to be first-order stationary (Trangmar et al., 1985), i.e.

$$E[z(x)] = m \tag{2.21}$$

where m is the mean and

$$E[z(x) - z(x + h)] = 0 \tag{2.22}$$

where h is the vector of separation (distance or direction) between sample locations. If the mean is constant and independent of position and the covariance C(h) of each z(x) and z(x + h) pair only depends upon the separation vector h. i.e.

$$C(h) = E[z(x) - m]\,[z(x + h) - m] \tag{2.23}$$

the variable is said to be second-order stationary (Gutjahr. 1985; Oliver and Webster. 1986).

When $|h| = 0$. equation 2.23 defines C(0). which is the variance (s^2). In this circumstance. the autocorrelation function holds and is defined as

$$r(h) = C(h)/s^2 \tag{2.24}$$

where r(h) is the autocorrelation among samples at distance of separation. or lag h. A plot of the autocorrelation values r(h) versus the lag is called the autocorrelogram. The value of r(h) decreases with increasing separation vector h. The distance at which r(h) no longer decreases defines the range within which samples of the variable are spatially dependent.

The autocorrelation technique has been used to describe the changes in field-measured soil properties over distance. and the degree of dependency among neighbouring observations (Webster and Cuanalo. 1975; Webster. 1978; Vieira et al.. 1981). Vieira et al. (1981). for example. applied the technique in their field-measured infiltration rate studies. and found that the autocorrelogram was a useful tool in determining the maximum sampling distance over which the observations are spatially correlated. Russo and Bresler (1981) also used the technique and concluded that the distance of spatial dependence for soil moisture characteristics was greater in surface horizons than in subsurface horizons. Spatially-distributed variables. however. often do not show second-order stationarity; the finite variance or covariance required by an autocorrelation function (equation 2.24) cannot be defined. because the variance or covariance of many soil properties tend to vary infinitely as the size of the study area is extended. This has led to the development of a weaker assumption of stationarity known as the intrinsic hypothesis (Matheron. 1965).

The intrinsic hypothesis requires that the expected value of z at any place x is the mean and that for any vectors of separation. h. the variance of difference [z(x) - z(x + h)] is finite and independent of position within the localised region (Webster. 1985).

$$E[z(x)] = \mu \tag{2.25}$$

$$\begin{aligned} \mathrm{Var}\,[z(x) - z(x+h)] &= E\{[z(x) - z(x+h)]^2\} \\ &= 2\gamma(h) \end{aligned} \tag{2.26}$$

This assumes the following model of soil variation:

$$z(x) = \mu + \varepsilon(x) \tag{2.27}$$

where z(x) is the value of the property at position x within a region. μ is the mean value in that region. and $\varepsilon(x)$ is a spatially dependent random component.

The quantity $\gamma(h)$ in equation 2.26. which is half the variance of the differences between values at places separated by h. is called the semi-variance.

It is important to realise that the second-order stationarity encompasses the intrinsic hypothesis. but not the converse. Therefore. the semi-variance is valid under both assumptions, whereas the autocorrelation can only be applied when the second-order stationarity holds.

Given intrinsic hypothesis, the semi-variance at a given lag, h, can be estimated as the average of the squared differences between all observations separated by the lag (Webster and Burgess, 1983; Trangmar et al., 1985).

$$\gamma(h) = 1/2N(h) \sum_{i=1}^{N} [z(x_i) - z(x_i + h)]^2 \qquad (2.28)$$

where there are N(h) pairs of observations. Examples of how observation points along transects are paired at three different lags for estimation of semi-variances, are illustrated in Figure 2.2.

The semi-variance between any two locations in the region depends only on the distance, or direction, of separation, and not on their geographic locations. The plot of semi-variance $\gamma(h)$ versus lag h is called the semi-variogram. The semi-variogram is more widely used than autocorrelation for assessing the spatial variability of soil properties because of the former's weaker assumption of stationarity.

2.6.3 Semi-variograms

Figure 2.3a shows the principal features of a well-structured semi-variogram. In most instances, it is found that the semi-variance $\gamma(h)$ increases with increasing separation vector h, and reaches a maximum at which it levels out (Burgess and Webster, 1980). The maximum semi-variance is known as the sill and its value is approximately equal to the sample variance s^2 if the variable is second-order stationary, or meets the intrinsic hypothesis (Trangmar et al., 1985). The lag a, at which the sill is reached is called the range. The semi-variance may increase continuously without showing a definite sill and range (Figure 2.3b). This is interpreted as due to the presence of regional trend effects (local stationarity) (Trangmar et al., 1985). In this case, the distance at which the semi-variance equals

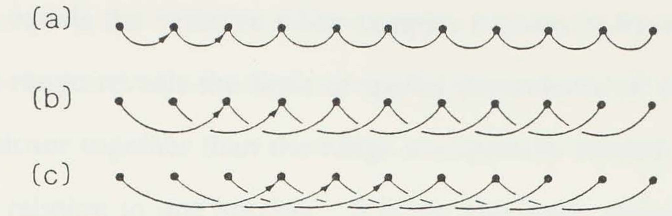


Figure 2.2 Examples of paired comparisons for observation points separated at (a) lag 1, (b) lag 2, and (c) lag 3 along transects for estimation of semi-variograms (after Webster, 1985)

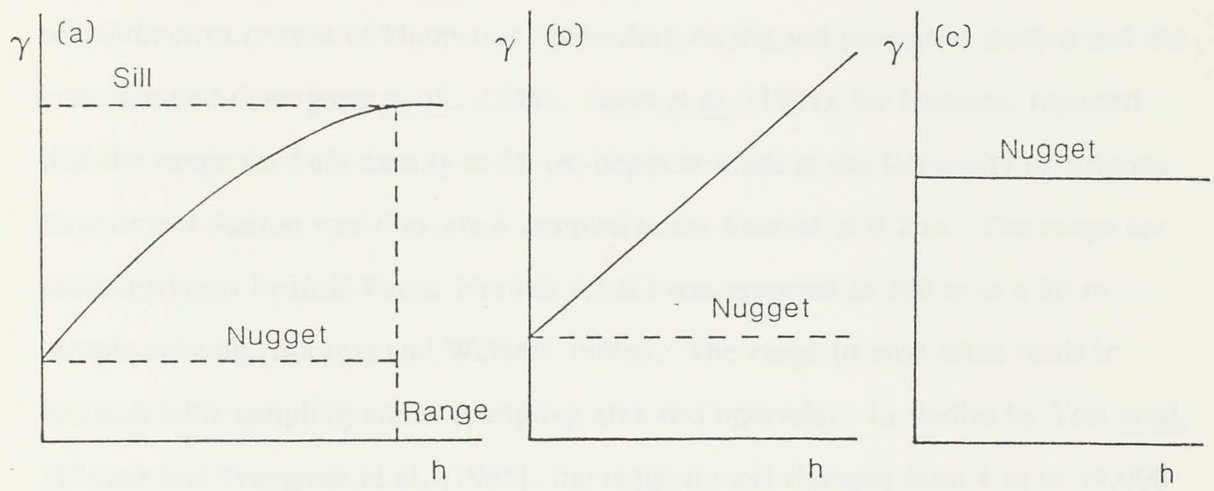


Figure 2.3 Elements of semi-variograms, (a) a well-structured spherical semi-variogram, (b) a linear semi-variogram, (c) pure nugget effect (after Webster, 1985)

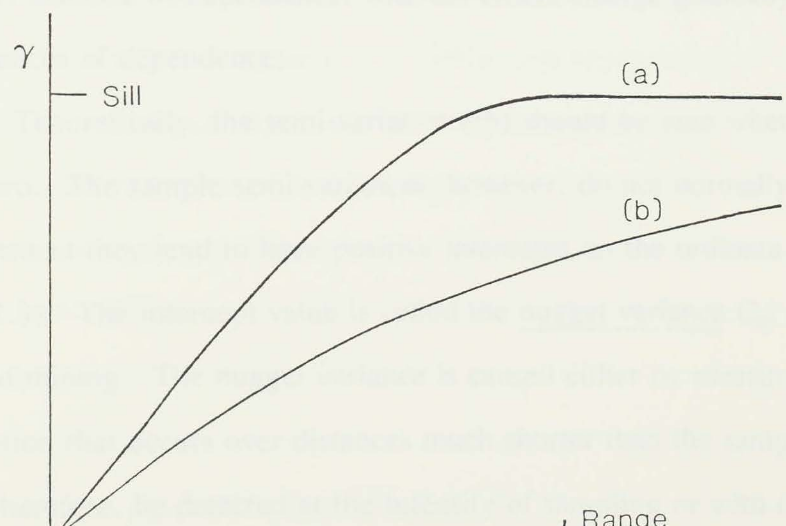


Figure 2.4 Spherical (a) and exponential (b) models with the same range and sill (after Clark, 1979)

the variance (s^2) is often taken as the range. This is based on the assumption that the semi-variance equals the variance when samples become independent.

The range reveals the limit of spatial dependence of soil properties. Those observations closer together than the range are spatially related, whereas soils further apart bear no relation to one another. It is the maximum sampling distance over which neighbouring observations are spatially correlated, and defines the maximum radius within which the neighbouring samples are considered for interpolation by kriging (cf. Section 2.6.4).

Studies have shown that the values of the range vary significantly from tens of centimetres to tens of kilometres, depending on the soil properties studied and the area sampled (Trangmar et al., 1985). Gjem et al. (1981), for instance, reported that the range for bulk density at 50 cm depth in a soil at the University of Arizona Experiment Station was 6 m when sampled at the interval of 0.2 m. The range for loam thickness in Hole Farm, Norfolk (U.K.) was reported as 100 m at a 20 m sample spacing (Burgess and Webster 1980a). The range (if any) often tends to increase with sampling scales (sampling area and intervals). In studies by Yost et al. (1982a) and Trangmar et al. (1985), the range for pH changed from 4 m to 14,000 m when sampled at intervals of 0.5 m and 1000 m respectively. These values clearly demonstrate that some soil properties change rapidly with distance, and therefore have short distance of dependence, whereas others change gradually with consequent long distances of dependence.

Theoretically, the semi-variance $\gamma(h)$ should be zero when the lag itself equals zero. The sample semi-variances, however, do not normally pass through the origin; instead they tend to have positive intercepts on the ordinate when $h = 0$ (Figure 2.3). The intercept value is called the nugget variance C_0, a term derived from gold mining. The nugget variance is caused either by measurement error, or soil variation that occurs over distances much shorter than the sampling intervals. It cannot, therefore, be detected at the intensity of sampling or with the accuracy of the technique used. The nugget variance can be reduced by sampling at closer intervals.

The measurement error is thus defined when the nugget variance no longer decreases with closer sampling. The total component of spatial covariance is defined by the difference between the sill and the measurement error. If the semi-variogram exhibits pure nugget effect, that is the semi-variance $\gamma(h)$ equals the sill or variance (s^2) at all distances of separation h (Figure 2.3c), there is no spatial correlation between samples at the sampling intensity used. The nugget effect is important in kriging as it limits the precision of interpolation (Section 2.6.4).

Spatial dependence of soil properties often varies in different directions, and this can be quantified by comparing semi-variograms of samples taken from transects in different directions. The presence or absence of anisotropic spatial variation is revealed by the difference in slopes of semi-variograms derived from different directions (Webster and Burgess, 1980; Burgess and Webster, 1980a). If soil properties vary at the same rate with distance in all directions, the variation is said to be isotropic. The anisotropic variation occurs when the semi-variogram in one direction is steeper than the others i.e. the variation at a given distance of separation h in one direction is equivalent to the variation at a distance kh in another direction. The parameter k is called the anistropy ratio which has a value of 1 (isotropic) or greater (anisotropic). Burgess and Webster (1980a), for instance, found that anisotropic ratio for stone content in the surface horizon of the soils at Plas Grogerddan, U.K., was 5.42. Most anisotropy ratios reported, however, are in the range of 1.3-4.0 (Trangmar et al., 1985). An appreciation of the anisotropic nature of soil spatial variation is particularly important when designing sampling schemes or establishing experimental plots.

There is no general mathematical formula to describe the shape of the different semi-variograms. The most commonly used model is the linear function (Burgess and Webster, 1980a; Hajrasuliha et al., 1980; Vauclin et al., 1983).

$$\gamma(h) = C_0 + wh \qquad \text{for } h > 0$$
$$\gamma(0) = 0 \qquad (2.29)$$

where w is the slope and C_0 is the nugget variance.

Another model that has been used to fit many semi-variograms of soil properties is the spherical model (Burgess and Webster, 1980a; Trangmar et al., 1985; Webster, 1985; Oliver and Webster, 1986). It is defined as

$$\gamma(h) = C_0 + C[3h/2a - 1/2\ (h/a)^3] \qquad \text{for } 0 < h \leq a$$

$$\gamma(h) = C_0 + C \qquad \text{for } h > a$$

$$\gamma(0) = 0 \qquad (2.30)$$

where a is the range, C_0 is the nugget variance and $C_0 + C$ is the sill.

The semi-variogram of the spherical model reaches a sill at a finite range. There are other situations, however, where the semi-variogram approaches the sill asymptotically, that is there is no absolute range (Figure 2.4). In this case, an exponential model has been recommended (Yost et al., 1982a; Clark, 1979; Webster, 1985):

$$\gamma(h) = C_0 + C\ [1 - \exp(-h/r)] \qquad \text{for } h > 0$$

$$\gamma(h) = 0 \qquad (2.31)$$

where r is a distance parameter. The range is estimated from an approximation a' = 3r, where a' is the lag when γ(a') is approximately equal to $C_0 + 0.95C$.

It is most important to choose the appropriate model for fitting the semi-variogram since different models yield different values for the range and nugget variance. Both of these two parameters are critical for interpolation by kriging.

2.6.4 Kriging

One of the important uses of semi-variograms is for kriging, a technique for making optimal and unbiased estimates of regionalised variables at unsampled locations. In conventional statistics, the sample mean is used to represent the property value at any unsampled locations within a study region. The regionalised variable theory, however, interpolates, or extrapolates, each soil property value at desired locations by taking into account neighbouring observations and their spatial dependence expressed by the semi-variograms. The term was named after D.G. Krige, who applied the method in the South African goldfields (Webster and Burgess, 1983). It is a method of local estimation in which each estimate is a weighted average of neighbouring observed values. The weights for each observed value are chosen so as to give unbiased estimates and, at the same time, to minimise the estimation variance (kriging variance). The estimates can be for points (punctual kriging) or for an area (block kriging). Punctual kriging, however, can be treated as a special case of block kriging. The estimation of the regionalised variable z for block B is achieved by using the following equation (Trangmar et al., 1985; Webster, 1985):

$$\hat{z}(B) = \sum_{i=1}^{n} \lambda_i z(x_i) \qquad (2.32)$$

where $\hat{z}(B)$ is the kriged value of z for block B, n is the number of observations within the neighbourhood weighted for estimation, and λ_i is the weight associated with the ith observation. The neighbouring observations are weighted in a way to minimise the estimation variance with the constraint that the sum of λ_i equals 1 in order to give unbiased estimates, i.e.

$$\sum_{i=1}^{n} \lambda_i = 1 \qquad (2.33)$$

and

$$E[z(B) - \hat{z}(B)] = 0 \qquad (2.34)$$

where z(B) is the true value of z at place B.

The particular weights for sampling points used for estimation depend on the semi-variogram, the configuration of sampling points and the place to be predicted. The weights take account of the spatial dependence expressed in the semi-variogram and the geometric relationships among the observed points. In general, points near the interpolation place carry more weight than distant points. This means that kriging is essentially a local estimation and that the semi-variograms need to be well-fitted with the model only over the first few lags. Most points far from the estimation place (point or block) can be omitted from consideration without serious consequence since the weights λ_i decrease as the distance between observation points and estimation place increases (Burgess and Webster, 1980a). Generally speaking, the nearest 16 to 25 points are adequate to give an accurate estimate (Burgess and Webster, 1980a). The range represents the maximum radius within which the sampling points are weighted for interpolation.

The minimised estimation variance (σ_k^2) for block kriging is obtained from the following expression:

$$\sigma_k^2 = \sum_{i=1}^{n} \lambda_i \bar{\gamma}(x_i, B) + \Psi_B - \bar{\gamma}(B, B) \qquad (2.35)$$

where $\bar{\gamma}(x_i, B)$ is the average semi-variance between the observation points x_i in the neighbourhood and the points within the block B, $\bar{\gamma}(B, B)$ is the average semi-variance between all points within the block, i.e. the within-block variance, and Ψ_B is the Lagrange parameter associated with minimisation (Webster and Burgess, 1983).

In the case of punctual kriging, the last term in equation 2.35 is zero as there is assumed to be no variance at a point. The quantity $\bar{\gamma}(x_i, B)$ becomes the semi-variance between the sampling points x_i and the point to be estimated (x_0), i.e.

$$\sigma_k^2 = \sum_{i=1}^{n} \lambda_i \gamma(x_i, x_0) + \Psi \qquad (2.36)$$

The estimation variance of block kriging is always less than that of punctual kriging, because the within-block variance is removed from the error term in block kriging. In other words, a certain amount of error is buried within the block.

The estimation variance is calculated for each estimated value, providing a measure of the reliability of interpolation. The estimation variance depends on the semi-variogram and the configuration of the data locations in relation to the estimated place, but not on the observed values themselves (Burgess and Webster, 1980a). This is of great importance for sampling design because, provided the semi-variogram is known, the interpolation error can be calculated for a particular sampling scheme before the survey is made. This aspect is considered in more detail in section 2.7.

Punctual and block kriging have been used for the production of isarithmic maps of soil properties, and for designing soil sampling schemes for further studies (Burgess and Webster, 1980a, b; Vieira et al., 1981; Burgess et al., 1981; McBratney and Webster, 1983). The kriging process for isarithmic mapping involves the prediction of soil property values for a fine grid of points or blocks and then contours are drawn based on the kriged values. Van Kuilenburg et al., (1982) compared results from kriging and three other methods (mean values for soil mapping units, proximal and weighted average interpolation), and concluded that kriging was the most precise technique for estimation of moisture content. Laslett et al. (1987) compared several spatial prediction methods and found that kriging gave a better estimate of soil pH than other methods, including that of the conventional statistical mean.

Punctual and block kriging are the basic methods for local estimation or interpolation. With the development of the regionalised variable theory, however, other kriging techniques have been introduced; these include co-kriging and universal kriging.

The spatial distribution of a given soil property is often closely related to that of other properties. That is, some soil properties are co-regionalised and are spatially dependent on one another. In these circumstances, the principle of optimal estimation using regionalised variable theory for a single property can be extended to two or more co-regionalised properties. One soil property that has not been sufficiently sampled can be predicted by another co-regionalised property. This is particularly important in situations where properties that are cheap or easy to measure are co-regionalised with others which, although of importance are less easily determined. The theory of co-kriging and its applications in soil science have been summarised by Vauclin et al. (1983) and Yates and Warrick (1987). Both studies showed that co-kriging could become a useful means of providing unbiased estimates of an under-sampled soil property based on its relationship with other sufficiently sampled properties.

One of the inherent assumptions in kriging is that the data are stationary or meet the intrinsic hypothesis, which means that the difference between any two samples depend only on the distance of separation, but not on the geographic locations in the region. In some circumstances, however, strong local trends (the expected value of the random function z is not always constant within the neighbourhood and is no longer equivalent to the mean) exist in the region and this, theoretically, makes the ordinary kriging process inadequate for interpolation (Trangmar et al., 1985).

Universal kriging, as described by Webster and Burgess (1980), is a technique of interpolation that takes account of local trends. The presence of trends or drifts, as they are termed, is identified and quantified by structural analysis and then removed from the actual semi-variograms. The resulting semi-variograms are then used for interpolation. The evidence of non-stationarity is apparently indicated in the semi-variogram. If the semi-variogram increases concave upward, and does not level out to approach the population variance at large distances, then it is said to be non-stationary (Trangmar et al., 1985; Webster, 1985).

The need for universal kriging in soil science has been controversial. Webster and Burgess (1980) concluded that universal kriging appeared to be neither generally acceptable, nor of particular benefit. Studies by Yost et al. (1982b) also suggested that universal kriging resulted in very little improvement over ordinary kriging. It was shown that ordinary kriging is quite robust even in the presence of strong trends. Therefore, the scope of universal kriging in soil survey seems to be limited.

2.7 Sampling strategies

2.7.1 Introduction

All methods of partitioning and assessing soil variability are based upon a limited number of samples. There is a major problem, however, in deciding how many samples should be examined and where the observations should be located in the field. The appropriate sampling scheme is dependent on the inherent variability of the soil and the level of precision required. Ideally, the sampling should aim to meet the requirements of both efficiency and accuracy. This section outlines the methods of optimising the efficiency of sampling schemes.

2.7.2 Conventional methods

The conventional approaches to the partitioning of soil variability assume that variation of soil properties within sampling units is solely random, i.e. spatially uncorrelated. Therefore the sample mean is the best estimate of a soil property at any location (point or area) within the sampling area and the estimation precision is characterised by the conventional statistical parameters such as the variance, standard deviation, standard error and confidence limits (cf. Section 2.5). The sampling strategy is determined as such that it provides the best estimate of mean soil property values within a sampling area with limited effort.

As far as the configuration of observations in a sampling area is concerned, Random sampling, whereby every sampling unit has an equal chance of being drawn, has been regarded as an unbiased and statistically sound technique widely used in characterising soil mapping units (Wilding et al., 1965; McCormack and Wilding, 1969). Random samples, however, tend to cluster spatially: the density of observations per unit area and the dispersion of the sites over the region are not uniform (Wilding and Drees, 1983). Many observations may occur along the boundary of the delineations. Studies have shown that, for the same number of observations, the precision attained by random sampling can almost always be bettered by systematic sampling at regular intervals along a transect or on an grid (McBratney and Webster, 1983). Studies by Cochran (1946) and Quenouille (1949) showed that systematic sampling gave the most precise estimates for a given effort. Similar conclusions were reported by Webster (1977).

The minimum sample size n needed for estimating the mean values of a soil property within a sampling unit is a function of both the estimation precision desired and the amount of variance that occurs within the sampling area (Cline, 1944):

$$n = t_{\alpha}^{2}s^{2}/(x - \mu)^{2} \qquad (2.37)$$

where n is the number of observations needed for the estimation of population mean μ with a tolerable deviation of $x - \mu$ if the variance is s^2. The quantity t_{α} is student's t at the chosen level of confidence.

Many soil properties, however, reveal spatial dependence among observations. The use of mean as the estimate of soil property values at unsampled locations is inadequate if observations are spatially correlated within a sampling area, as the estimation variance (or estimation error) tends to be very high. The sample size determined on the basis of the conventional estimation variance which is derived under random assumption, using equation 2.37 therefore is often so large that investigators have to either give up the sampling scheme or sacrifice the desired precision.

2.7.3 Geostatistical methods

The regionalised variable theory (Matheron, 1965, 1971) provides an alternative tool for solving sampling problems, as it does take into account the spatial dependence of soil properties in the sampling regions. The estimation variance (or estimation error) for predicting soil property values at unsampled locations within a sampling unit by punctual (for point) or block (for area) kriging depends on the degree of spatial dependence, which is expressed in the semi-variograms, and the configuration of observation points in relation to the point or block to be estimated. If the semi-variogram is known, then the estimation variance of any regular scheme can be determined beforehand (cf. Equations 2.35 and 2.36). Given a desired precision level, the determination of the sampling density necessary to provide the required precision can be achieved by solving Equation 2.35 (for block estimation) or Equation 2.36 (for point estimation). McBratney and Webster (1983) applied the theory to regional soil sampling and suggested that the actual efficiency achieved in their studies was 3 to 9 times greater than that estimated by the classical methods: much fewer samples were needed to achieve the same level of precision using geostatistics than conventional methods. Burgess et al. (1981) pointed out that the conventional approach, since taking no account of spatial dependence among observations, often resulted in oversampling and unnecessary cost. Unfortunately, few other studies of this kind have been made to substantiate these claims.

The kriging variance is minimised if the sampling is conducted on a grid basis (Trangmar et al., 1985). Burgess et al. (1981) and McBratney and Webster (1983) concluded that sampling on an equilateral triangular grid gives slightly more precise estimates than a square one, providing the variation in the region is isotropic. A square or rectangular grid, however, may be preferable in practice because of its convenience.

2.8 Summary and conclusions

Soil is a three-dimensional body which varies spatially in response to the interaction of environmental factors and human activities. The various soil morphological, physical and chemical properties tend to spatially change at different rates. Physical properties are particularly variable in alluvial soils, due to the many lateral and vertical changes in texture that are inherited from the parent material.

Conventional methods of soil classification and mapping have played an important role in partitioning soil variation, and delimiting relatively homogeneous soil bodies for the purposes of making more precise statements and accurate predictions about their inter-relationships, behaviour and land use potentials. The traditional approach to soil classification and mapping, however, relies upon the similarities or differences in a few easily-measured soil properties. The assumption is made that other accessory properties vary in a similar fashion to the definitive properties. Conventional statistical assessment of homogeneity (or variability) within and between taxonomic or mapping units, however, has confirmed and quantified the different rates of spatial variations associated with different soil properties. Soils grouped together in terms of the few definitive characteristics may therefore differ substantially in other non-definitive soil properties: some of these, such as hydraulic behaviour, may be critical to management practices. Similarly, soils separated into different classes (taxonomic or mapping units) may differ in the diagnostic characteristics, yet resemble each other in most other properties. More work still needs to be done to assess the effectiveness of soil classification and the quality of soil mapping.

Soil property values at unsampled locations in a region are conventionally estimated by sample means and associated confidence limits. This approach is inadequate in situations where observations are spatially dependent, because the estimation errors are unnecessarily large. Sampling strategies determined on the basis of the estimation error of the mean, therefore, are conservative with a result of over-sampling and unnecessary effort.

The spatial dependence of soil properties and the components of variations are revealed and characterised by semi-variograms. Soil properties at any unsampled locations within the sampling region can be predicted by kriging, with a minimised estimation error, based on the spatial relationship between the predicted values and their neighbouring observations. Detailed soil isarithmic maps can be readily produced through interpolation by kriging with the aid of computers. Sampling strategies for estimating soil properties in intensive soil studies, determined on the basis of the estimation error by kriging, require less effort than those derived by conventional methods. The application of the regionalised variable theory in soil variability studies is relatively new, however, and more studies are needed to verify the approach and assess its value to soil science studies.

CHAPTER 3

LOCATION AND PHYSICAL ENVIRONMENT OF THE STUDY AREA

3.1 Introduction

The study was conducted on a 10 ha block of the Lincoln College Research Farm, located on the Canterbury Plains in the South Island of New Zealand (Figure 3.1). This chapter outlines the general physical environment of the Canterbury Plains and the characteristics of the soils occurring on and around the Lincoln College properties. It concludes with more detailed environmental information on the study area itself.

3.2 Physical environment of the Canterbury Plains

3.2.1 Physiography

The Canterbury Plains are bounded in the east by the sea and extend inland to the foot-hills of the Southern Alps (Figure 3.1), reaching an altitude of about 350 m (Wilson, 1985). The region covers about 7537 km^2 and consists of a series of overlapping fans, composed of generally coarse-textured glacial outwash and alluvial sediments, with flat to gently undulating surfaces. They have been deposited during the Quaternary by eastward-flowing rivers draining down from the Southern Alps. The rocks in the catchments of these rivers are dominantly greywacke with some argillite.

3.2.2 Climate

The climate in the region is sub-humid and cool temperate. The mean annual rainfall is relatively low compared with other parts of New Zealand, varying

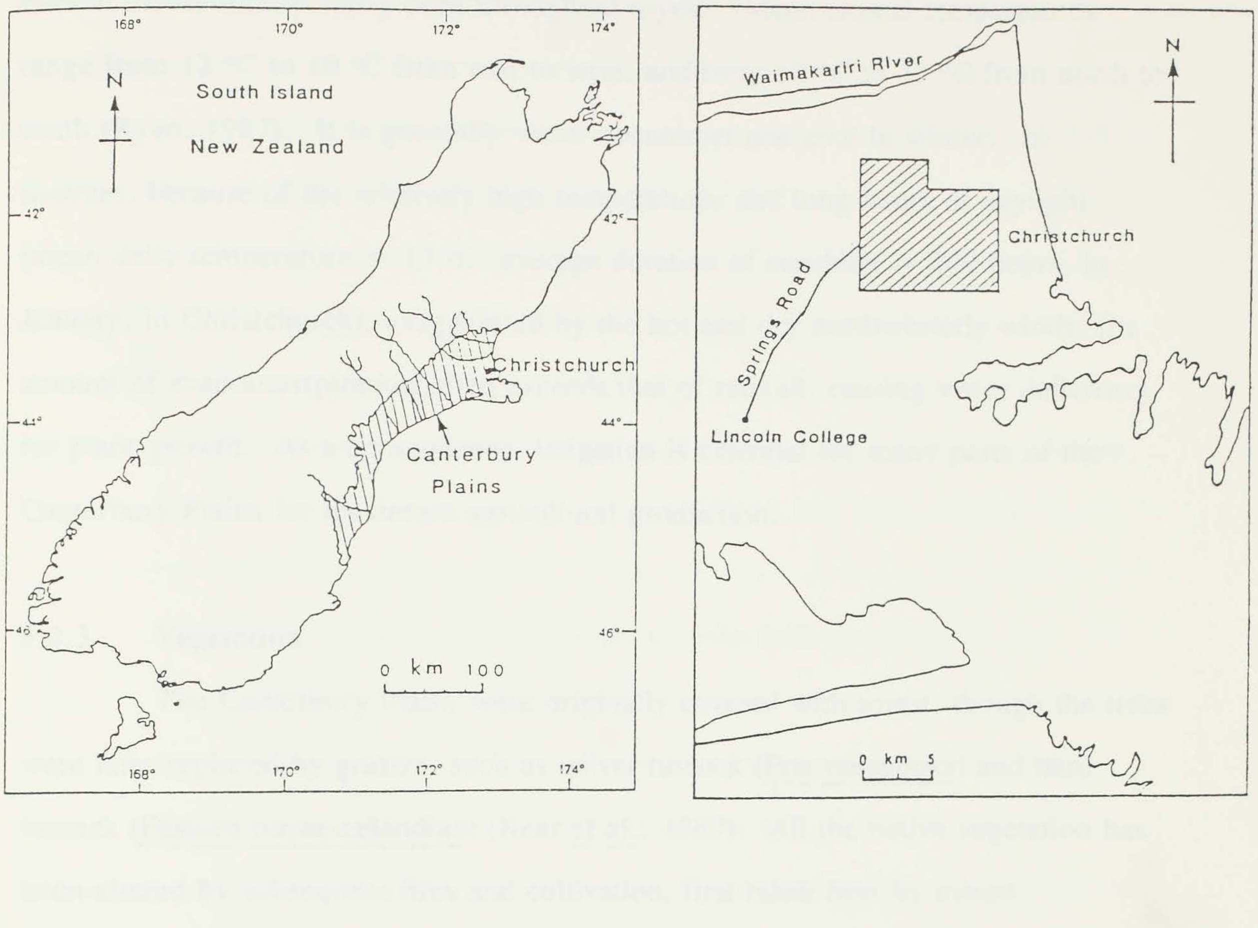


Figure 3.1 Location map of the study area

from 650 mm near the coast to about 1000 mm at the western edge of the Plains; rainfall distribution is fairly even throughout a year. Mean annual temperatures range from 12 °C to 10 °C from east to west, and from 12 °C to 11 °C from north to south (Ryan, 1987). It is generally warm in summer and cool in winter. In summer, because of the relatively high temperatures and long hours of daylight (mean daily temperature = 17 °C, average duration of sunshine = 209 hours, in January, in Christchurch), exaggerated by the hot and dry northwesterly winds, the amount of evapotranspiration often exceeds that of rainfall, causing water deficiency for plant growth. As a consequence, irrigation is essential for many parts of the Canterbury Plains for maximum agricultural production.

3.2.3 Vegetation

The Canterbury Plains were originally covered with forest, though the trees were later replaced by grasses, such as silver tussock (Poa caespitosa) and hard tussock (Festuca novae-zelandiae) (Kear et al., 1967). All the native vegetation has been altered by subsequent fires and cultivation, first taken over by mixed communities of native and exotic plants and then almost completely replaced by pasture and other crops.

3.2.4 Soils

The main soil groups found on the Plains include Recent Soils, Yellow Grey Earths (YGE), Recent-YGE intergrades and YGE-Yellow Brown Earths (YBE) intergrades (Kear et al., 1967). The distribution of zonal soils correspond generally with climatic zones: YGE with subhumid climate in the east and YGE-YBE intergrades with subhumid to humid climate in the western part of the Plains. These soils are intensively used for agriculture and horticulture. Forestry is mainly relegated to shallow gravelly soils.

Soils are developed in greywacke alluvium parent materials which have been transported down from the Southern Alps. Soil properties, particularly texture, vary considerably within any one climatic zone across the Plains in accordance with the expected complex spatial changes in alluvial sedimentation (cf. Section 2.2.2). The distribution pattern of soils on the Plains is further modified by deposition of loess blown by the northwesterly winds: large areas of the Plains are covered with a loess mantle derived from the adjacent rivers to the north.

Traditional methods of classification and mapping have led to the recognition of four broad units differentiated according to age and degree of soil development (Cox. 1978).

(1) soils of the Lismore age group (>20.000 years) on the high terraces.

(2) soils of the Templeton age group (3000-10.000 years) on the intermediate terraces.

(3) soils of the Waimakariri age group (700-2400 years) on the low terraces.

(4) soils of the Selwyn age group (<300 years) on the flood plains.

Each of these age groups consists of several soil series that reflect differences in thickness of textural layers within the alluvial parent material and/or drainage.

3.3 Soils of the Templeton age group

Lincoln College and adjacent regions are located on intermediate terrace levels with soils developed in post-glacial sediments of the Templeton age group. The soils of this group are divided into five soil series. mainly according to the thickness of fine materials over gravels and assumed drainage status as determined through field descriptions of soil mottling patterns (Table 3.1).

Soils on the wind-blown sand dunes are Halkett soils with rolling surfaces and brownish subsoils. Eyre soils are recognised where the gravels are covered by only shallow depths of fine-textured materials. Templeton. Wakanui and Temuka soils are developed on deep fine alluvium over gravels. The Templeton soil is moderately weathered and well-drained with a yellowish brown subsoil colour. and only faint or no mottles. Wakanui soils are characterised by strong prominent brown mottling against matrix colours in subsurface horizons: such mottling characteristics signify an imperfect drainage condition. The subsoil matrix colour for the less imperfectly-drained Wakanui soils is brown to yellowish brown with many strong brown mottles and some dark brown concretions. A light brownish grey subsoil matrix, with abundant strong brown mottles and some dark brown or hard black concretions, is expected in the more imperfectly-drained Wakanui soils. The poorly-drained Temuka soil generally has a subsoil matrix colour of olive grey or grey with abundant yellowish brown and strong brown mottles and dark brown or black concretions. The most poorly-drained Temuka subsoils, however, may be almost uniform grey with very diffuse yellowish brown mottles and few or no concretions.

Distinctions between these soil series are summarised in Table 3.1. The criteria used to further subdivide these series into types are summarised in Tables 3.2 - 3.5.

Table 3.1 Classification of the Templeton age group soils (after Cox, 1978)

Fine materials over gravels (cm)	<46	>46			Eolian sands
Drainage	Excessive	Good	Imperfect	Poor	Good
Soil Series	Eyre	Templeton	Wakanui	Temuka	Halkett

Table 3.2 Classification of Eyre soil series (after Cox, 1978)

Topsoil texture	Silt loam			Fine sandy loam		Sandy loam	
	No stones	Few stones	Stony	No stones	Few stones	Stony	Very stony
Depth to gravels (cm)	25-46	10-25	<25	25-46	10-25	<25	<25
Sub-division	E_1	E_2	E_3	E_4	E_5	E_6	E_7

E_1: Eyre shallow silt loam

E_2: Eyre very shallow silt loam

E_3: Eyre stony silt loam

E_4: Eyre shallow fine sandy loam

E_5: Eyre very shallow fine sandy loam

E_6: Eyre stony sandy loam

E_7: Eyre very stony sandy loam

Table 3.3 Classification of Templeton soil series (after Cox, 1978)

Thickness of fine materials over gravels (cm)		>60		46-60
Thickness (cm) of loamy sand and/or coarser-textured layers within 1m profiles		<30	>30	
Topsoil texture	Silt loam	T_1	T_2	T_3
	Fine sandy loam	T_4	T_5	T_6

T_1: Templeton silt loam

T_2: Templeton silt loam on loamy sand

T_3: Templeton silt loam, moderately deep phase

T_4: Templeton fine sandy loam

T_5: Templeton fine sandy loam on sand

T_6: Templeton fine sandy loam, moderately deep phase

Table 3.4 Classification of Wakanui soil series (after Cox, 1978)

Thickness of fine materials over gravels (cm)	>60		46-60
Thickness of loamy sand and /or coarser-textured layers within 1m profiles (cm)	<30	>30	
Subdivisions	WK_1	WK_2	WK_3

WK_1: Wakanui silt loam

WK_2: Wakanui silt loam on loamy sand

WK_3: Wakanui shallow silt loam

Table 3.5 Classification of Temuka soil series (after Cox, 1978)

Depth to gravels (cm)	>60							<60
Topsoil texture	Silt loam				Clay loam		Peaty silt loam	Silt loam
Subsoil texture	Silt loam or coarser		Clay loam (>15cm)		Clay loam to sandy loam		Silt loam	Silt loam on gravels
Subsoil colour	Brown-grey, many mottles	Grey, few mottles	Olive grey, many mottles	Grey, few mottles	Grey, strong mottles	Grey, few mottles	Grey, faint mottles	Olive grey, many mottles
Subdi-visions	TK_1	TK_2	TK_4	TK_5	TK_7	TK_8	TK_3	TK_6

TK_1: Temuka silt loam

TK_2: Temuka silt loam, strongly gleyed phase

TK_3: Temuka silt loam, peaty phase

TK_4: Temuka silt loam on clay loam

TK_5: Temuka silt loam on clay loam, strongly gleyed phase

TK_6: Temuka shallow silt loam

TK_7: Temuka clay loam

TK_8: Temuka clay loam, strongly gleyed phase

Soil texture and mottling patterns therefore form the basis of the classification of soils on these intermediate terraces. It is assumed that these two properties satisfactorily differentiate soils in terms of other accessory properties. Studies by Karageorgis (1980), however, indicate that the classification scheme (Cox, 1978) is unsatisfactory in separating Templeton and Wakanui soils. He concluded that the soil mottling characteristics are extremely variable and unsuitable for use as differentiating criteria. His results showed that the productivity of Kopara wheat does not respond to such morphological variations, though crop growth is clearly strongly influenced by soil moisture regimes. An assessment of this classification scheme in terms of soil hydraulic properties is essential not only for improving the usefulness and applicability of the scheme itself but also for a better understanding of the relationships between morphological and hydraulic properties.

3.4 The study area

The study area consists of 14 paddocks (10 ha) within part the Lincoln College Research Farm, situated at the intersection of Ellesmere Junction and Weedon Roads, west of Lincoln College (Figure 3.1). The paddocks are currently used for animal grazing experiments by the Animal Science Department of Lincoln College. Most of the paddocks have not been cultivated in the last five years (Hughes, personal communication). A remnant channel hollow extends across the eastern part of the area in a NW-SE direction (cf. Figure 4.1). The channel hollow, has presumably been altered by subsequent depositional or erosional processes, as it is asymmetrical with a gentle slope of 0.83° on the northeast side and a steeper slope of 1.5° on the southwest side. The maximum difference in elevation in the area, i.e. the difference between the bottom of the channel hollow and the top of the southwest bank is about 0.6 m. The rest of the area is relatively flat with a low-angled (0.33°) slope down from the top of the channel bank towards the SW.

The soils in the area have been mapped on a small-scale soil map (1:126,720) as Paparua stony silt loam and Wakanui silt loam (Kear et al., 1967). A larger-scale map (N.Z. Soil Bureau, unpublished) indicates that the area encompasses various Paparua, Templeton and Wakanui soil types, where Paparua is equivalent to the Eyre soil in this region (Kemp, personal communication).

A detailed climatic record is available as the Lincoln College Meteorological Station is located within the study area (Figure 3.1). The mean daily air temperature in January is 16.6 °C and in July is 5.7 °C with an annual average of 11.4 °C (1951-1980) (New Zealand Meterological Service, 1983). The annual rainfall is about 681 mm and the annual evapotranspiration is 867 mm (1941-1984) (New Zealand Meterological Service, 1986). The distribution of rainfall and evapotranspiration in different months is illustrated in Figure 3.2. It is evident that the amount of evapotranspiration in summer exceeds that of the rainfall, and plants therefore suffer water deficiency. In winter, however, surplus water may occur.

3.5 Summary and conclusions

The soils on the Canterbury Plains are developed on a series of greywacke alluvial sediments that have been deposited by the major rivers flowing down from the Southern Alps. These soils exhibit great variation in morphological properties, particularly textural, many of which have been inherited from the parent materials. More appropriate criteria are required for classification of the soils on the intermediate terraces, as the current scheme has been found to be inadequate in differentiating some soils.

Plants suffer water deficiency in summer in most parts of the Canterbury Plains, and irrigation is required for maximum agricultural production. There is a need for a better understanding of the variation in soil hydraulic properties within, or between, those soil taxonomic units defined in terms of morphological features. Such knowledge is important for the evaluation of the effectiveness of the classification scheme.

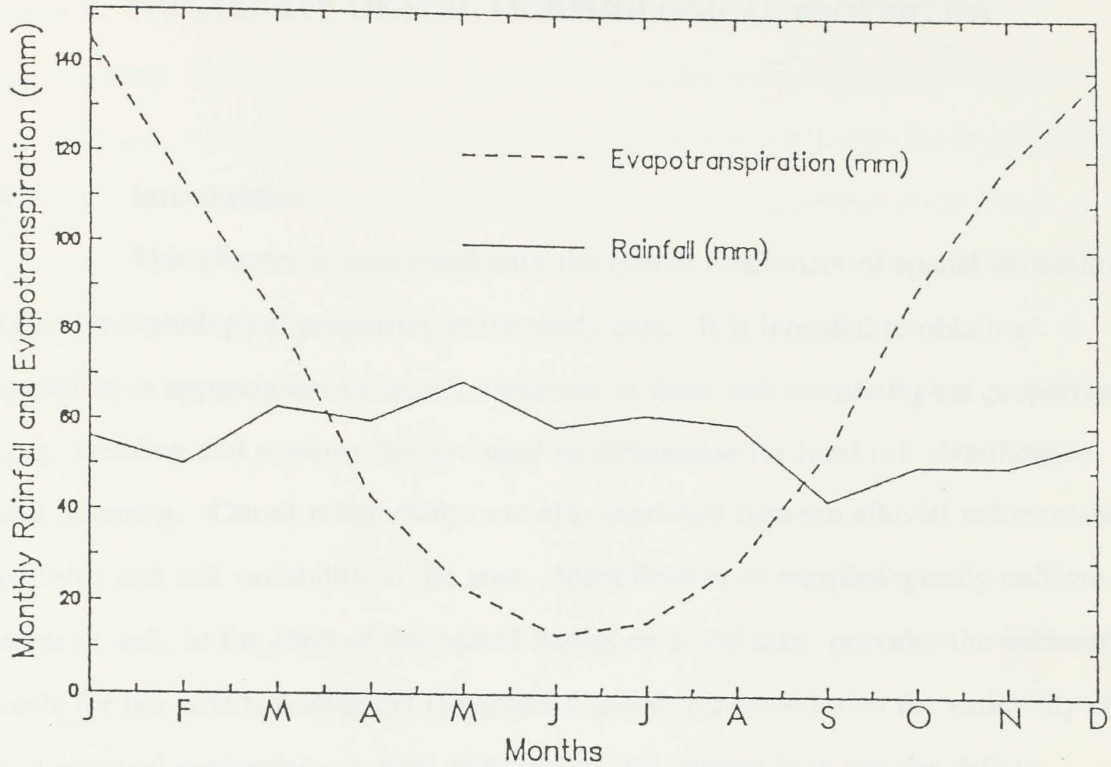


Figure 3.2 Annual water balance in the study area

CHAPTER 4

VARIABILITY OF SOIL MORPHOLOGICAL PROPERTIES

4.1 Introduction

This chapter is concerned with the overall assessment of spatial variability of soil morphological properties in the study area. It is intended to obtain a quantitative appreciation of spatial variations in those soil morphological properties (e.g. mottling and texture) that are used as differentiae for local soil classification and mapping. Causal relationships are also examined between alluvial sedimentation patterns and soil variability in the area. Identification of morphologically-uniform areas of soil, in the form of delineated bodies on a soil map, provides the necessary basis for the next two chapters (Chapters 5 and 6) concerned with the variability of soil physical properties. A final objective of this chapter is to use the data to consider optimal sampling strategies for future work of this kind.

The next section outlines the methods of soil survey and data analysis. For convenience of comparison and interpretation, results are presented and discussed together within the following sections: analysis of soil spatial variability, soil classification and mapping, and optimal sampling strategies.

4.2 Methods

4.2.1 Soil survey

The soil survey was conducted in two stages. An initial observation interval of 30 m was determined on the basis of previous studies in the region (Karageorgis, 1980) and consideration of the amount of effort that could be afforded. A 30 m x 30 m grid auger survey was therefore first carried out on the whole study area. Further observations were subsequently made at 15 m intervals in two areas of apparent spatial complexity (Figure 4.1). The data from this second survey was only used for the compilation of the conventional soil map.

A screw auger was used at each grid intersection and the soil described down to 1 m depth where possible, or to shallower depths where gravels were encountered. The morphological properties recorded were texture, mottling and concretions. All terms and classes used for the descriptions follow those outlined by Taylor and Pohlen (1979). Eight textural classes were recognised in the field: gravels, gravelly sand, sand, loamy sand, sandy loam, fine sandy loam, silt loam, and silty clay loam. Mottles were recorded in terms of their abundance (non, few, many, or abundant), size (fine, medium, or coarse), and contrast (faint, distinct, or prominent). The abundance and size of concretions were described in a similar manner.

4.2.2 Data analysis

The following morphological parameters were obtained from the survey data and used for subsequent analyses:

(1) depth to strong mottles (DM) (maximum = 1 m)

(2) depth to gravels (DG) (maximum = 1 m)

(3) thickness of loamy sand and/or coarser-textured layers (TS) within the top 1 m profiles (including gravelly sand)

These soil properties have obvious hydraulic significance and are those used diagnostically within the current soil classification system of the region. The data for each survey point are summarized in Appendix 1.

Contour maps and three-dimensional block diagrams were drawn based on the 30 m grid survey, using the C3D programme (Baird, 1986), to illustrate and summarize the overall spatial variation in these three morphological properties. The variability was then partitioned using geostatistical methods. Semi-variances were computed and semi-variograms plotted using the VAR2 program (Yost et al., 1986).

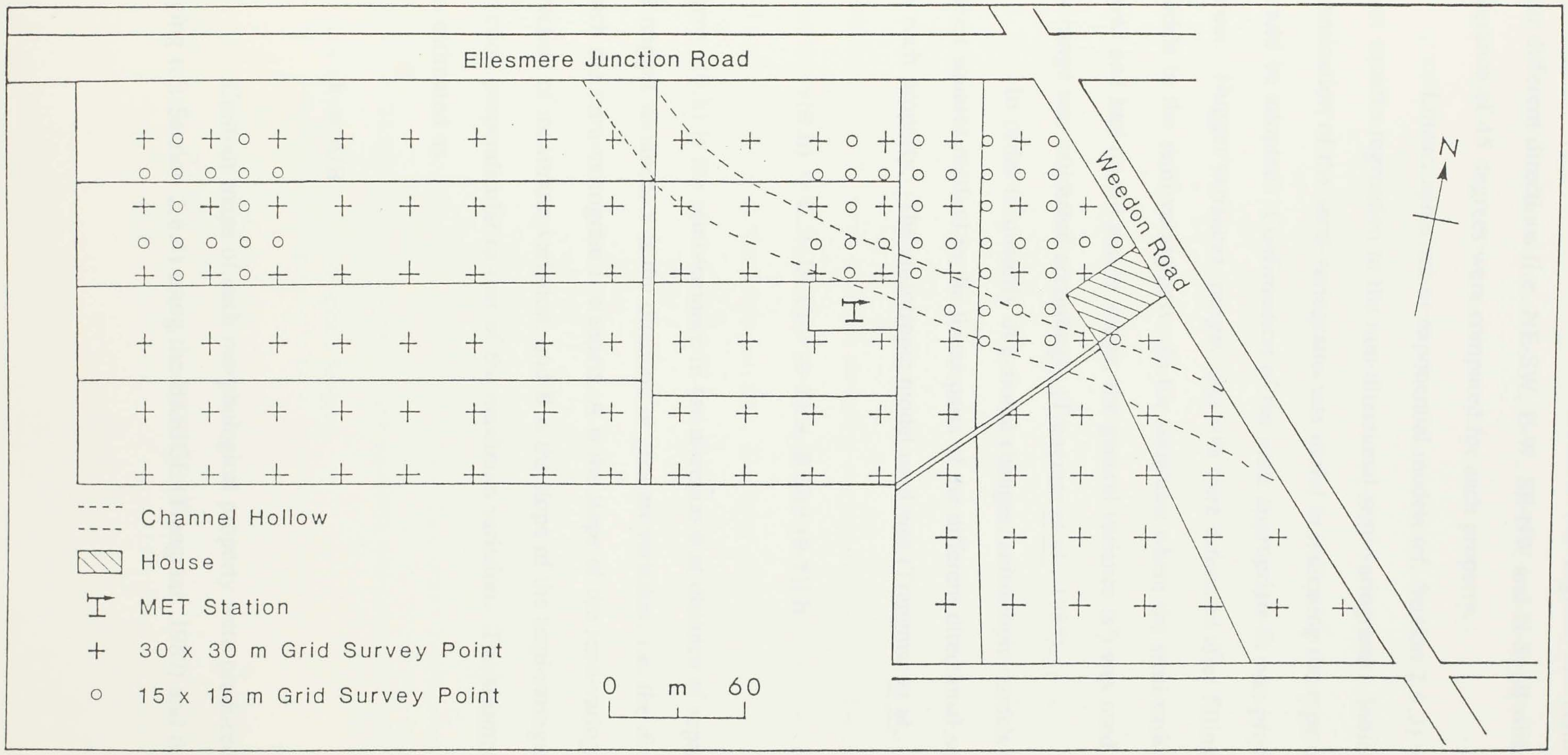


Figure 4.1 Locations of observation points within study area

All terms are defined and discussed in Chapter 2. Non-directional semi-variograms (i.e. those plotted with a direction tolerance of 180 degrees) and semi-variograms in four different directions (i.e. NE-SW, E-W, SE-NW and N-S) all with a direction tolerance of 45 degrees were computed for each property.

Linear, spherical or exponential models (cf. Section 2.6.3) were fitted by least squares regression to the non-directional semi-variograms. Initial visual examination of the semi-variograms was useful in indicating the type of models that should be adopted. Confirmation of the most appropriate fit was provided by the r^2 values. Nugget variances, ranges, and sills were estimated after fitting appropriate models to the semi-variograms. In the situations where the semi-variograms were linear and had no sill and/or range, the general variance (s^2) was used as the sill and the range was estimated accordingly (Trangmar et al., 1985).

In order to quantify directional changes, anisotropic models were fitted by a least squares method to the linear parts of the different directional semi-variograms for each property. The anisotropic model used was (Trangmar et al. (1985)

$$\gamma(\Theta,h) = C + [A\ \mathrm{Cos}^2\ (\Theta\text{-}\Psi) + B\ \mathrm{Sin}^2\ (\Theta\text{-}\Psi)]\ h \tag{4.1}$$

where $\gamma(\Theta,h)$ is the semi-variance in the direction Θ at distance of separation h, C is the nugget variance, Ψ is the direction of greatest variation, i.e. the direction in which the semi-variogram is steepest, A is the slope of the semi-variograms in the direction of maximum variation, and B is the slope of the semi-variogram in the direction perpendicular to that of the maximum variation. The anisotropic ratio k was estimated as

$$k = A/B \tag{4.2}$$

Contour maps of each morphological property were produced by block kriging (cf. Section 2.6.4) using the BKRIGE (Trangmar, 1987) and the C3D programmes. A total of 476 (34×14) blocks were kriged for each property. The kriged blocks were 15 m $\times$ 15 m and the observation points within a 100 m neighbourhood were weighted for estimation. The semi-variogram parameters, i.e. nugget variance, sill and range were estimated from the non-directional semi-variograms. Anisotropy was not taken into account for kriging.

Traditional methods of soil classification and mapping were employed to partition soil variation in the study area. The area was divided up into apparently homogeneous units on the basis of series and type criteria of the morphologically-based soil classification system of Cox (1978) (cf. Section 3.3). The boundaries between these delineated units were manually interpolated between observation points from both the 30 m and 15 m grid. This conventionally-derived soil map was compared to a block kriged soil map of the same area produced by Dr B. Trangmar (Soil Bureau, DSIR, Lincoln). The kriged map was based upon the 30 m grid data and the same classification criteria (series level) as used for the conventional map.

The survey data of the three properties were further analysed to provide guidelines as to the most efficient sampling strategy for future soil survey and variability studies (cf. Section 2.7). Estimation errors by block kriging (cf. Equation 2.35 in Section 2.6.4) were computed for blocks of different sizes and for different sampling intervals in the direction of maximum variation with the interpolation points being centered at the middle of grid cells. Kriging standard errors were calculated for a range of observation numbers. All computations were undertaken using the FORTRAN program by McBratney and Webster (1981). Graphs were produced illustrating the relationships between sampling interval, sampling number and kriging standard errors (kriging SE). Conventional standard errors (conventional SE) of mean were calculated for different sample numbers and compared with those standard errors derived from block kriging.

4.3 Analysis of soil spatial variability

4.3.1 Qualitative assessment and interpretation

The contour maps and three-dimensional block diagrams for the three soil morphological properties. DM (depth to strong mottles). DG (depth to gravels). and TS (thickness of loamy sand and/or coarser-textured layers). are presented in Figure 4.2.

Materials in the study area and neighbouring region were deposited by migrating braided rivers and/or distributive channels. Soil morphological features within the upper 1 m soil profile are partly related to the final depositional phases. This study is confined to a very small area: it is only therefore possible to make tentative interpretations about the alluvial deposition pattern as it partly reflects the influences of the depositional environment extending across the region as a whole.

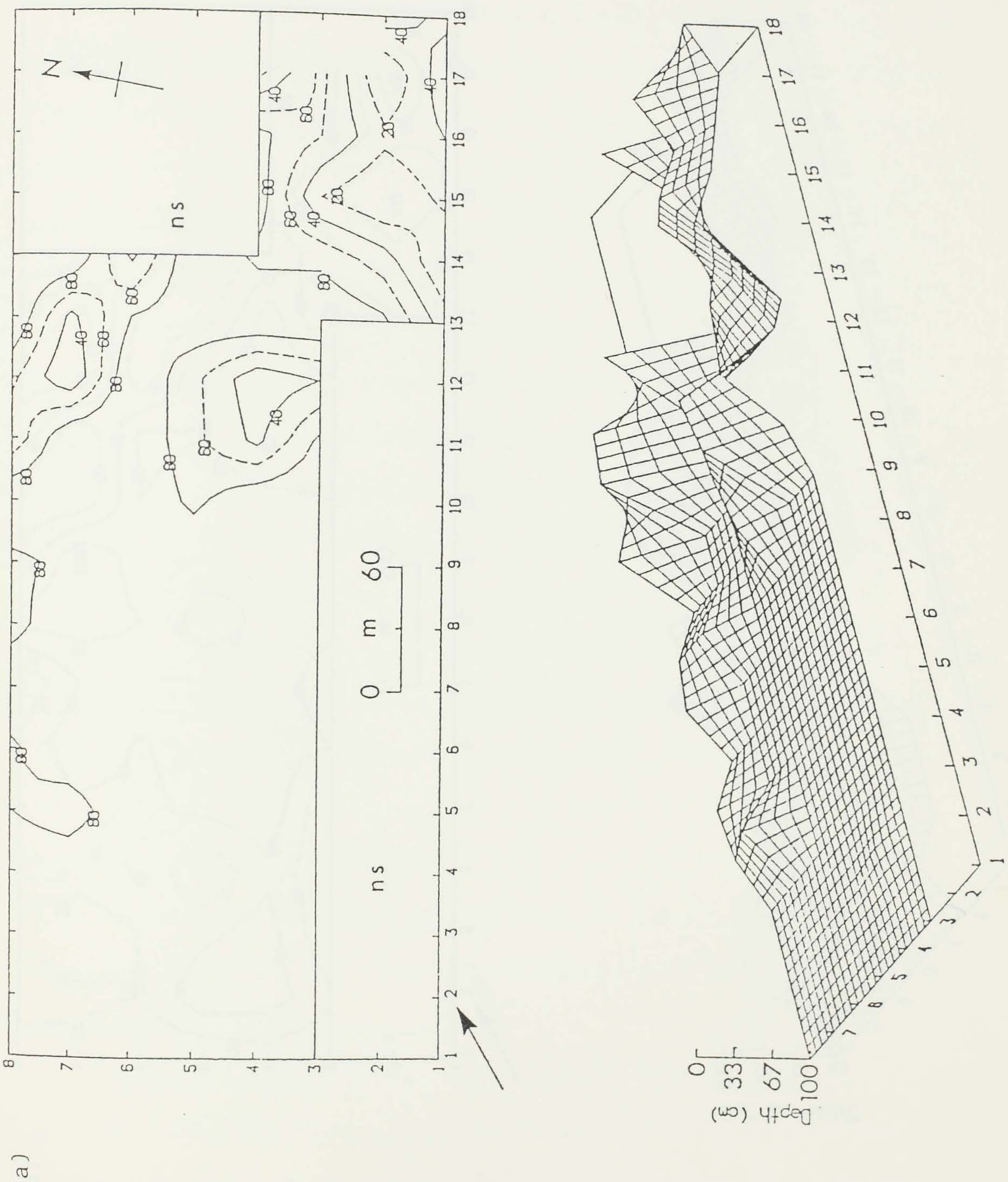
Figure 4.2a shows the distribution of gravels across the study area. Shallow gravels are encountered on the eastern part (right-hand side of the diagrams) adjacent to the channel hollow. No gravels are found within the 1 m profiles further away from the channel hollow in the southwest part of the region (lower left corner of the diagrams). The distribution pattern suggests that these gravels may be channel bar deposits left by braiding rivers that migrated across the region. The remnant channel hollow could represent the last position of one of the migrating channels that were responsible for the deposition of materials in the region. This remnant channel hollow may alternatively be a more recent distributive channel that cut into previously-deposited sediments.

The distribution of TS is illustrated in Figure 4.2b. The two diagrams show that there are thick coarse-textured layers on the central and eastern parts of the area. These layers become thinner towards the southwest and are gradually replaced by silt loam and clay loam textures. though. occasional sandy bands still occur in places. The coarse-textured layers are also thin within the channel hollow where subsurface horizons have silty clay loam textures.

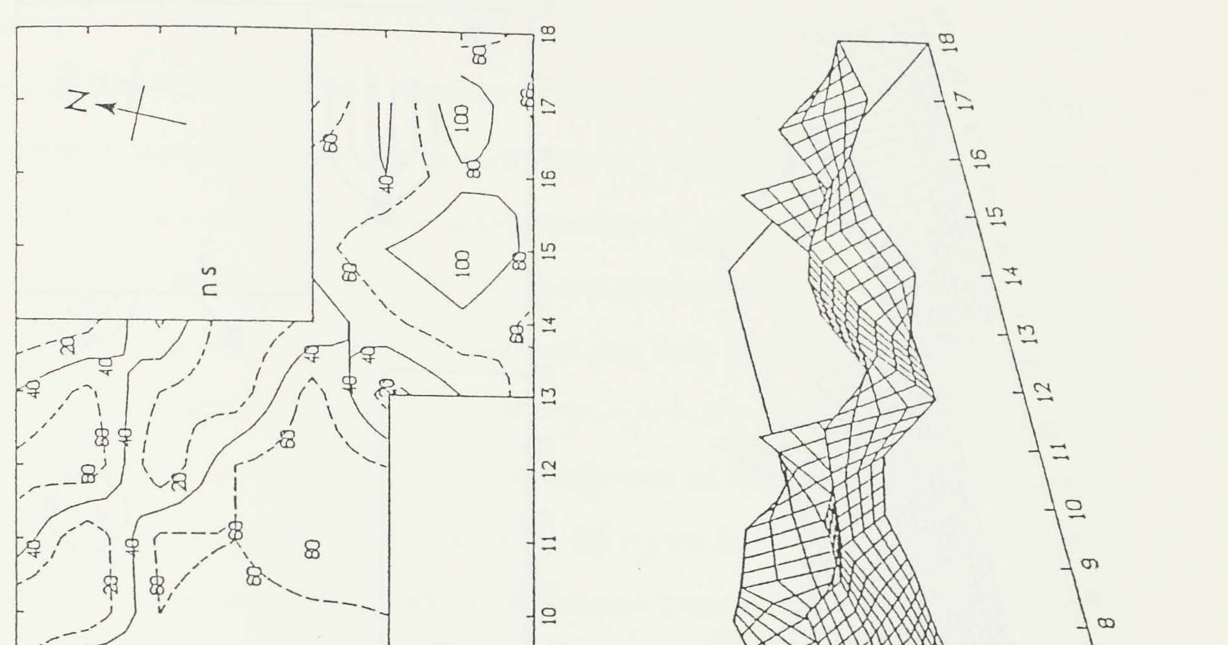
The overall patterns of TS together with DG distributions in the region suggests that the gravels have been buried by finer levee deposits. The shape and location of the original channels that are responsible for the deposition of these

Figure 4.2 Contour maps and three-dimensional block diagrams illustrating the variation in soil properties across the study area: (a) depth (cm) to gravels (DG). (b) thickness (cm) of loamy sand and/or coarser-textured layers (TS). (c) depth (cm) to strong mottles (DM)

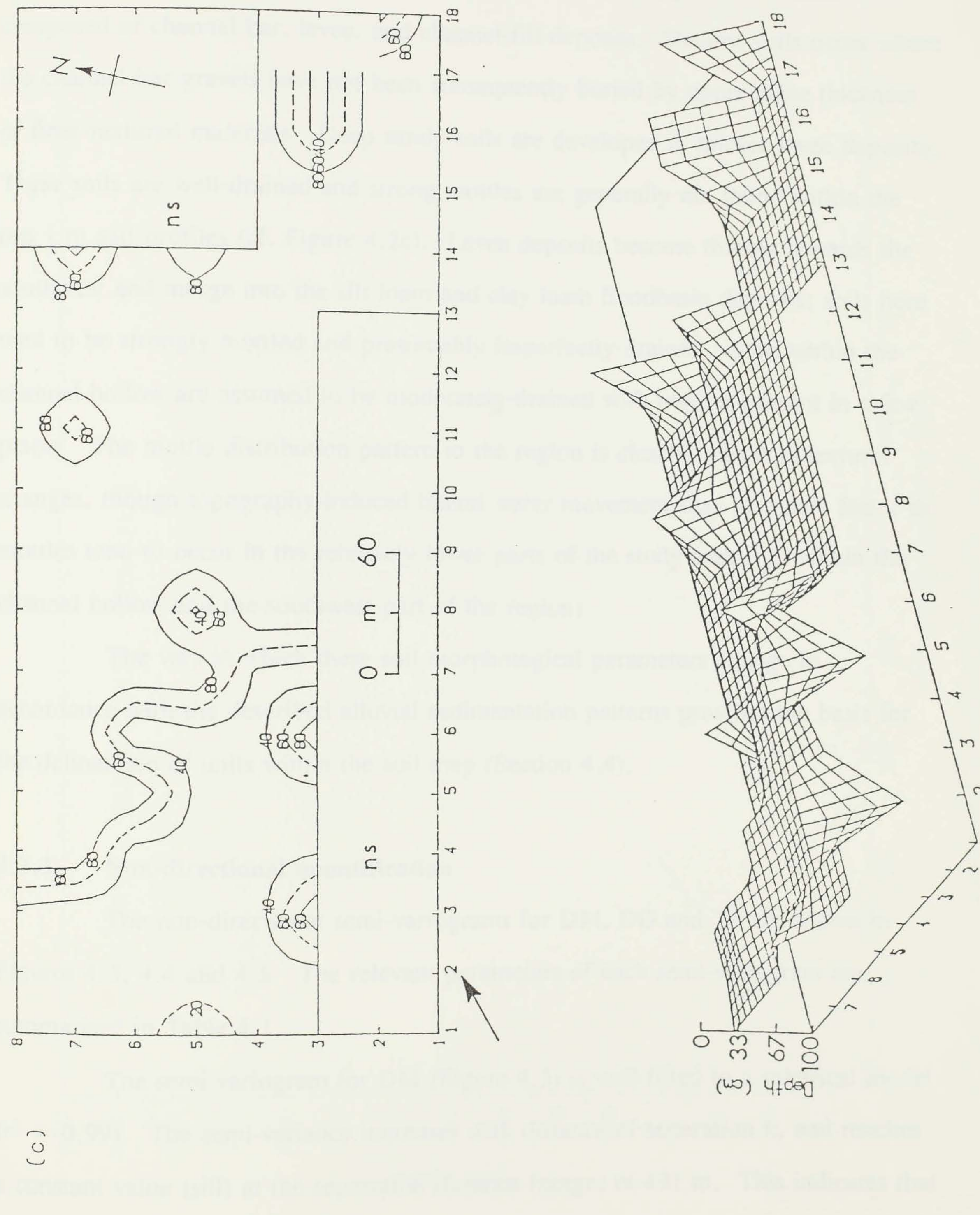
The arrows below the contour maps indicate the direction of viewing for the block diagrams. The 30 m spaced observation points on the maps and block diagrams are numbered along the axes for ease of reference. Areas which were not surveyed are indicated (ns).



(a)



(b)



(c)

gravels have been modified by more recent depositional or erosional processes. Levee deposits, however, thin away towards the southwest and grade into floodbasin deposits. Inclusions of channel-fill deposits may occur within the channel hollow. Some materials in the surface horizons may also be derived from loess deposition.

The parent materials in the study area therefore appear to be mainly composed of channel bar, levee, and channel-fill deposits. Shallow soils occur where the channel-bar gravels have not been subsequently buried by appreciable thickness of finer-textured materials. Deep sandy soils are developed in thicker levee deposits. These soils are well-drained and strong mottles are generally not found within the top 1 m soil profiles (cf. Figure 4.2c). Levee deposits become thinner towards the southwest and merge into the silt loam and clay loam floodbasin deposits: soils here tend to be strongly mottled and presumably imperfectly-drained. Soils within the channel hollow are assumed to be moderately-drained with mottles present in a few places. The mottle distribution pattern in the region is clearly related to textural changes, though topography-induced lateral water movement may also be a factor as mottles tend to occur in the relatively lower parts of the study area (i.e. within the channel hollow and the southwest part of the region).

The way in which these soil morphological parameters change in accordance with the described alluvial sedimentation patterns provides the basis for the delineation of units within the soil map (Section 4.4).

4.3.2 Non-directional quantification

The non-directional semi-variograms for DM, DG and TS are shown in Figures 4.3, 4.4 and 4.5. The relevant parameters of each semi-variogram are summarised in Table 4.1.

The semi-variogram for DM (Figure 4.3) is well fitted to a spherical model (r^2 = 0.99). The semi-variance increases with distance of separation h, and reaches a constant value (sill) at the separation distance (range) of 431 m. This indicates that samples closer than 431 m apart are spatially related, whereas observations further

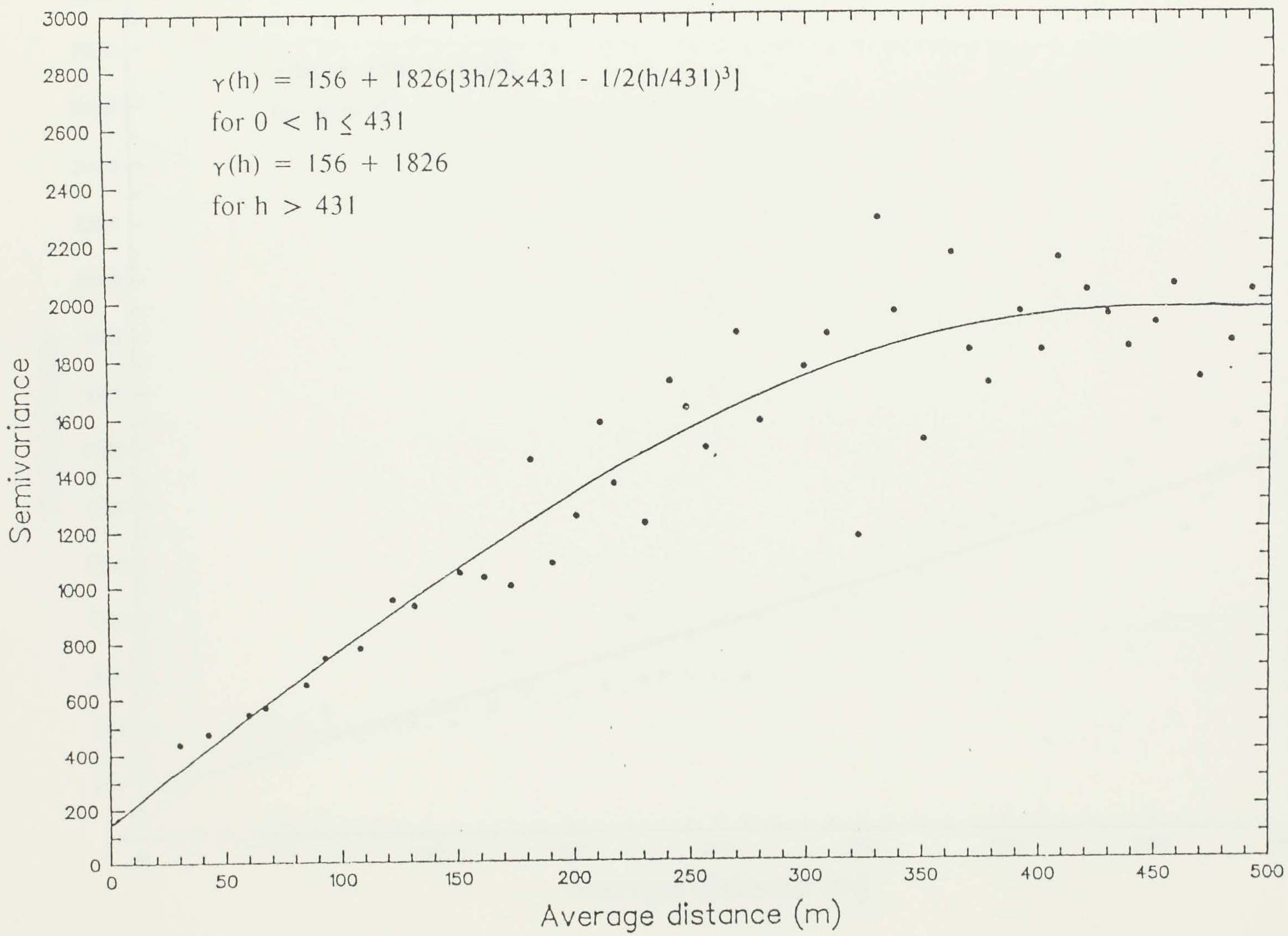


Figure 4.3 Non-directional semi-variograms for DM

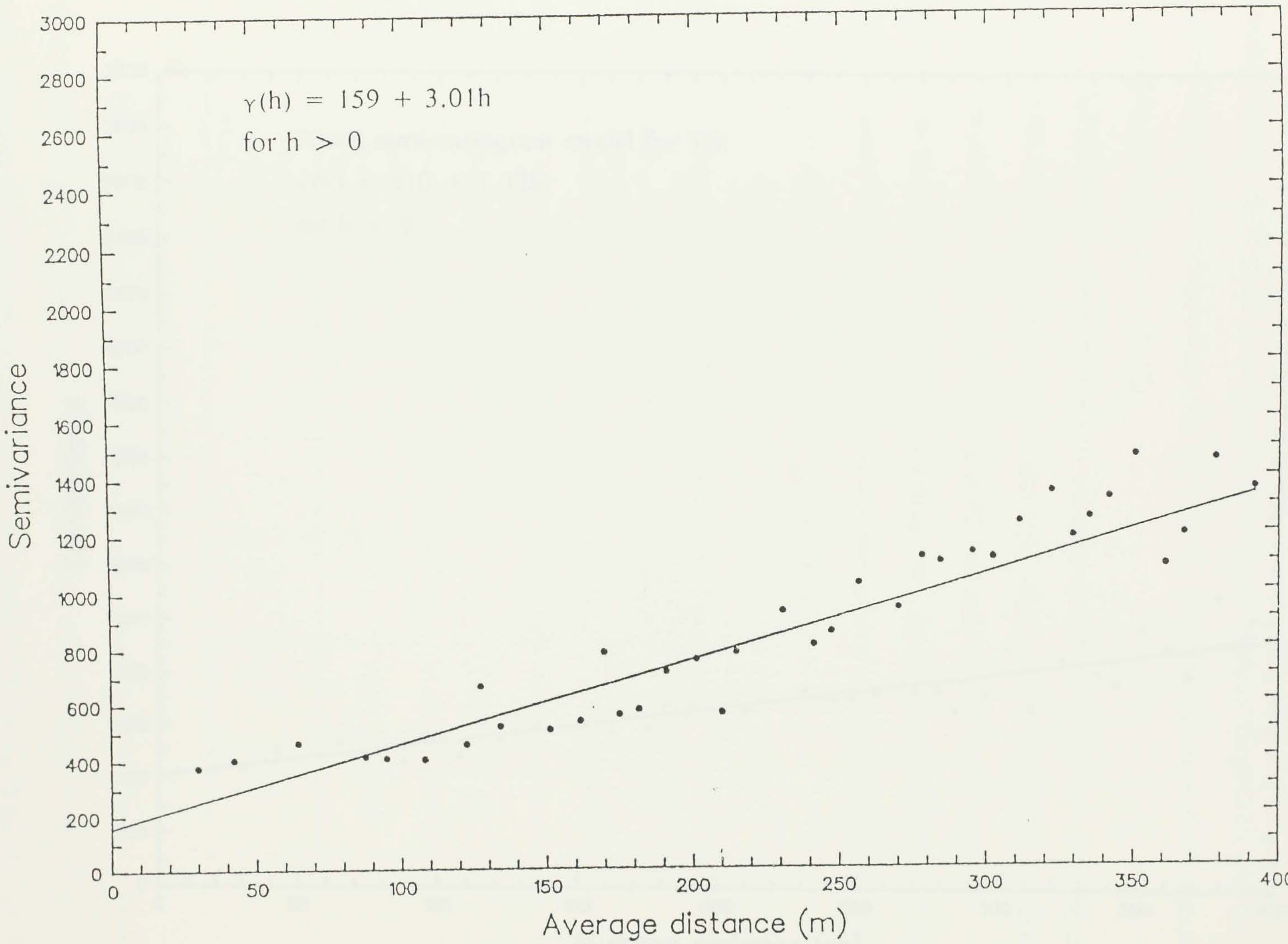


Figure 4.4 Non-directional semi-variograms for DG

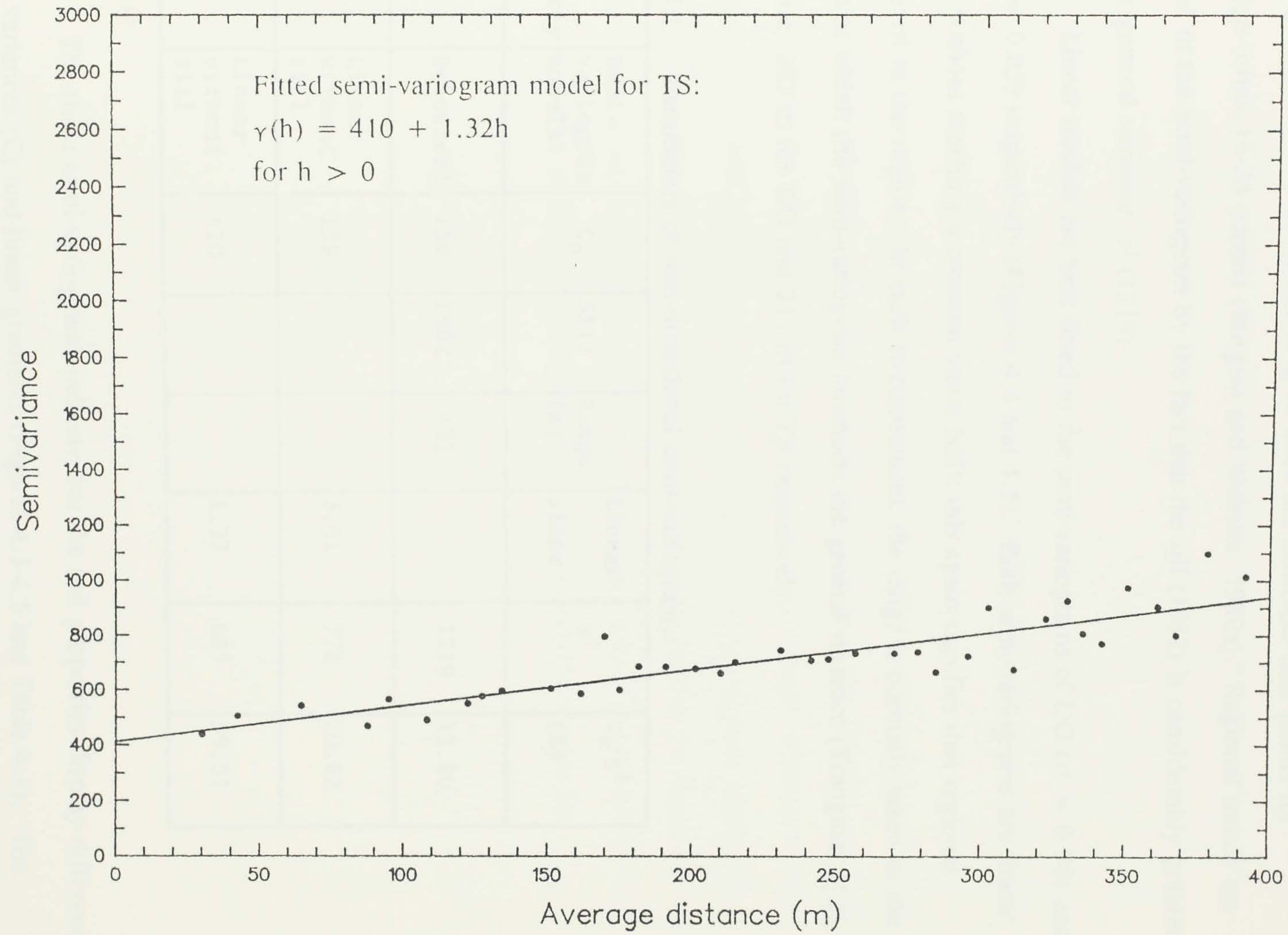


Figure 4.5 Non-directional semi-variograms for TS

apart are spatially independent. The range also defines the maximum radius within which observations may be weighted for estimation by kriging. The kriging radius, however, does not necessarily have to be as large as the range: a shorter distance of searching radius can be chosen as long as there are sufficient data points for estimation (often 16-25 points) (Burgess and Webster, 1980a). Regional trends are indicated in the semi-variogram by the fact that the sill (1982) is considerably greater than the general variance, s^2 (1219).

Linear models are best fitted to the semi-variograms of DG (r^2 = 0.90) and TS (r^2 = 0.85) respectively (Figures 4.4 and 4.5). Both semi-variograms are linear upward without reaching a constant value (sill): this again signifies that regional trends exist in the region. In such circumstances, the range is normally taken as the distance at which the semi-variogram intersects the general variance (Trangmar et al., 1985), i.e. 203 m for DG and 211 m for TS respectively.

Table 4.1 Parameters of non-directional semi-variograms

Soil property	Semi-variogram models	C_0	Sill	Range (m)	Linear slope	s^2	C_0/s^2 (%)
DM	Spherical	156	1982	431		1219	12.80
DG	Linear without sill	159			3.01	771	20.62
TS	Linear without sill	410			1.32	689	59.51

The three semi-variograms indicate that the soil properties display different nugget variances (C) and linear gradients (Figures 4.3-4.5 and Table 4.1). The nugget variances for DM and DG are similar, yet considerably lower than that of TS. The gradient of the linear part of the semi-variogram of DM is larger than the linear slopes of the other two semi-variograms. Most variation in DM therefore occurs between 30 m (sampling interval) and 431 m (range) with only a fraction of the variation (C/s^2 = 12.80%) accounted for within distances shorter than current 30 m sampling spacing. This is opposite to that of the variation in TS: a large proportion of the variation (C/s^2 = 59.51%) here exists within a distance less than the 30 m sampling interval. The manner of variation in DG is intermediate between these other two properties (C/s^2 = 20.62%).

4.3.3 Directional quantification

The semi-variograms in four different directions for each of the three soil properties are presented in Figures 4.6, 4.7 and 4.8.

The four semi-variograms for each property differ considerably in their slopes, particularly for DM and DG. In some directions (e.g. the NE-SW and E-W directions of DM) the semi-variograms show a finite sill and range, whereas in other directions (i.e. SE-NW and N-S directions of DM) the semi-variograms exhibit linear functions without sill and range. Anisotropic models were fitted to the linear parts of these directional semi-variograms in order to quantify the anisotropic variability of each property. The parameters estimated from these anisotropic models are summarised in Table 4.2, and the fitted anisotropic linear models are shown in Figures 4.9, 4.10, and 4.11.

Figure 4.6 Semi-variograms in four different directions for DM

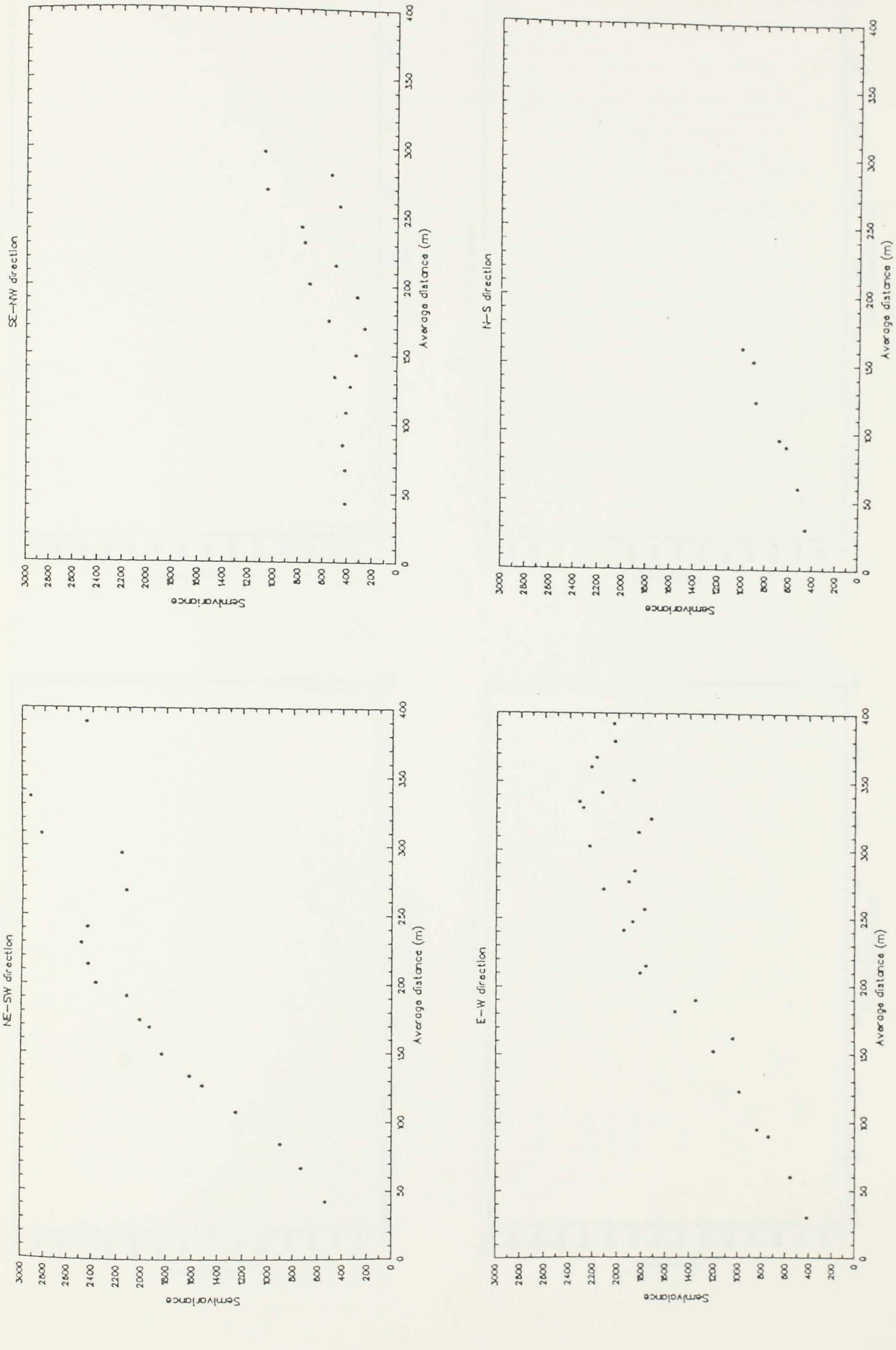

SE–NW direction
N–S direction
NE–SW direction
E–W direction
Semivariance
Average distance (m)

Figure 4.7 Semi-variograms in four different directions for DG

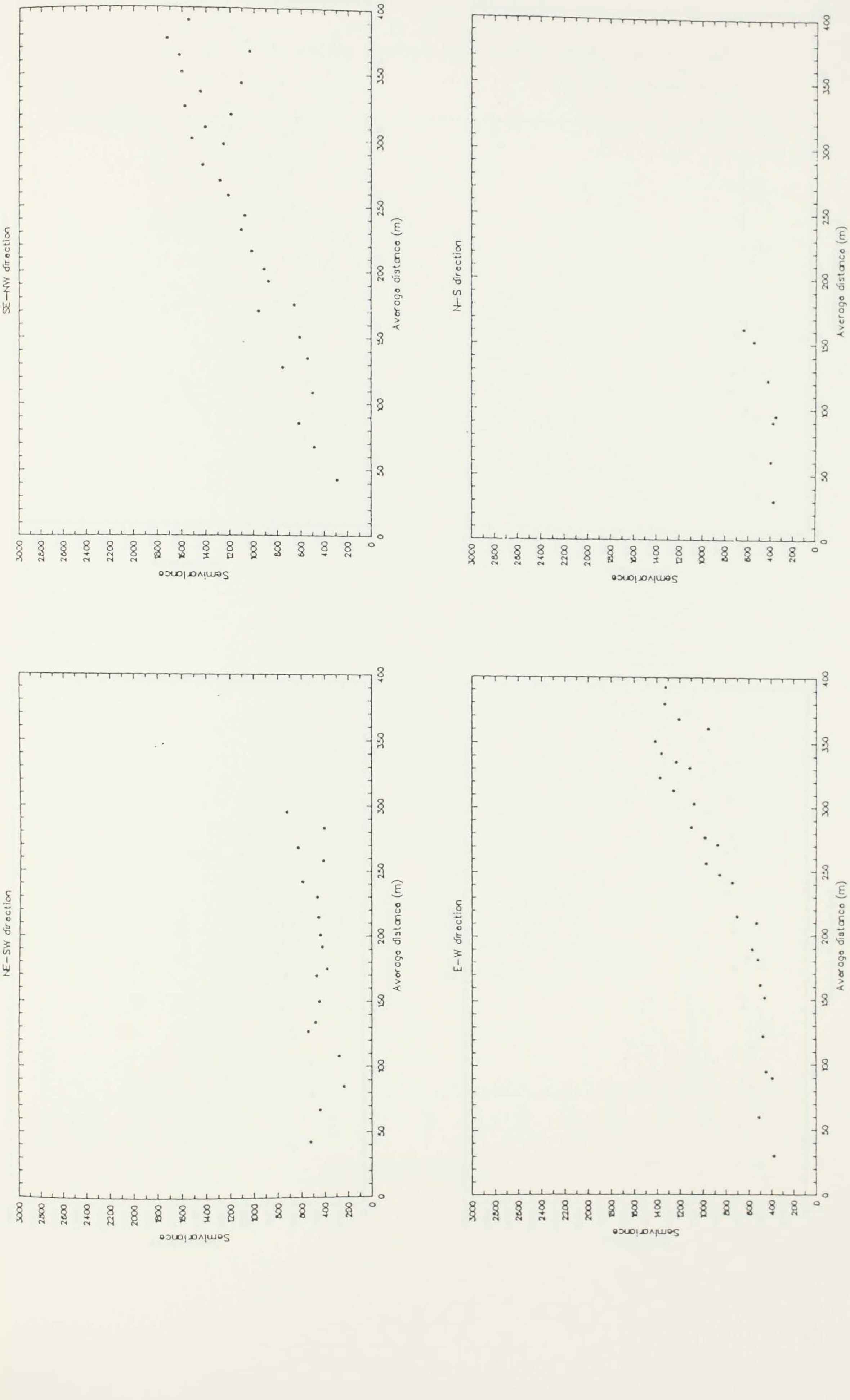

SE-NW direction
N-S direction
NE-SW direction
E-W direction
Average distance (m)
Semivariance

Figure 4.8 Semi-variograms in four different directions for TS

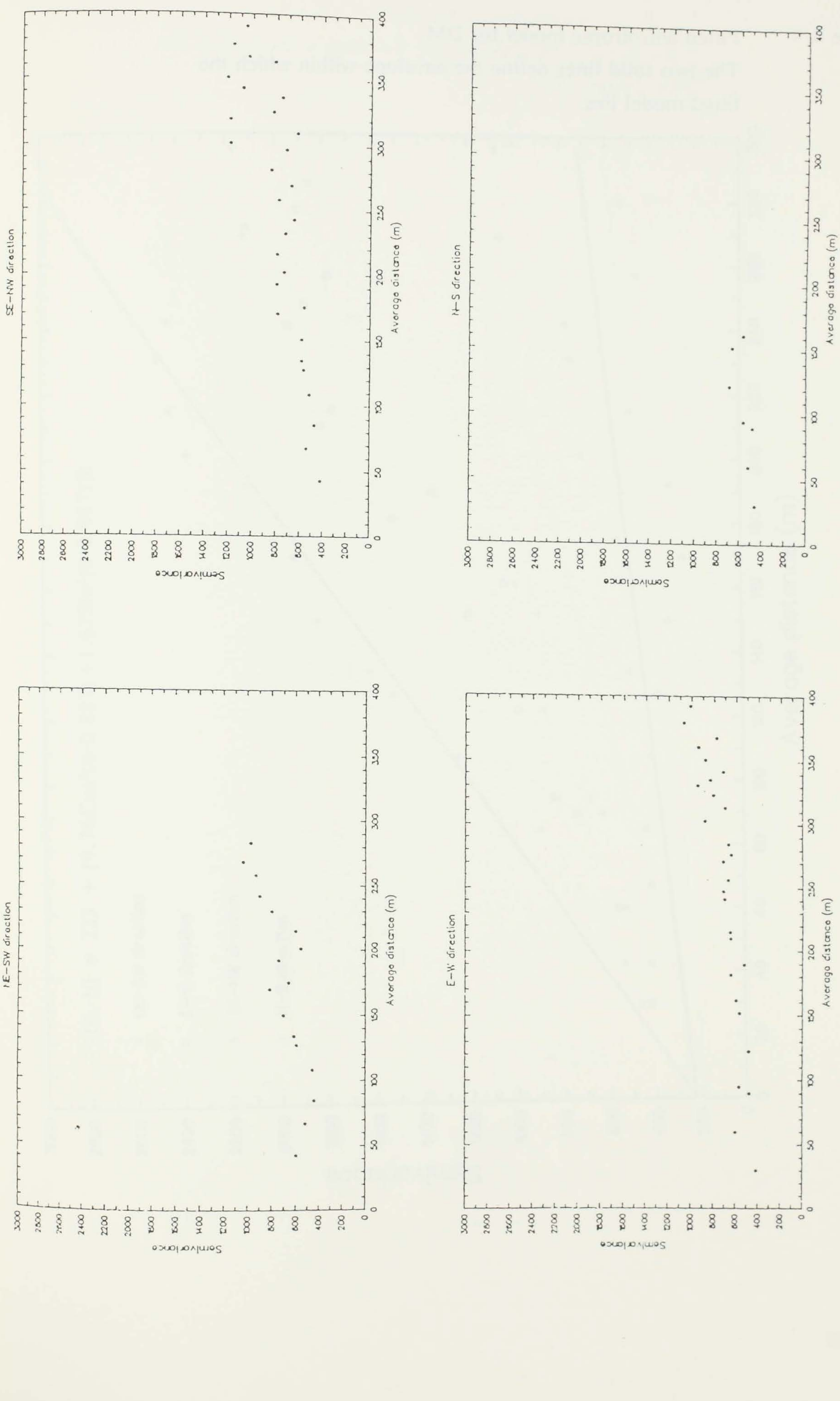

SE-NW direction
N-S direction
NE-SW direction
E-W direction
Average distance (m)
Semivariance

Figure 4.9 Fitted anisotropic model for DM

The two solid lines define the envelope within which the fitted model lies.

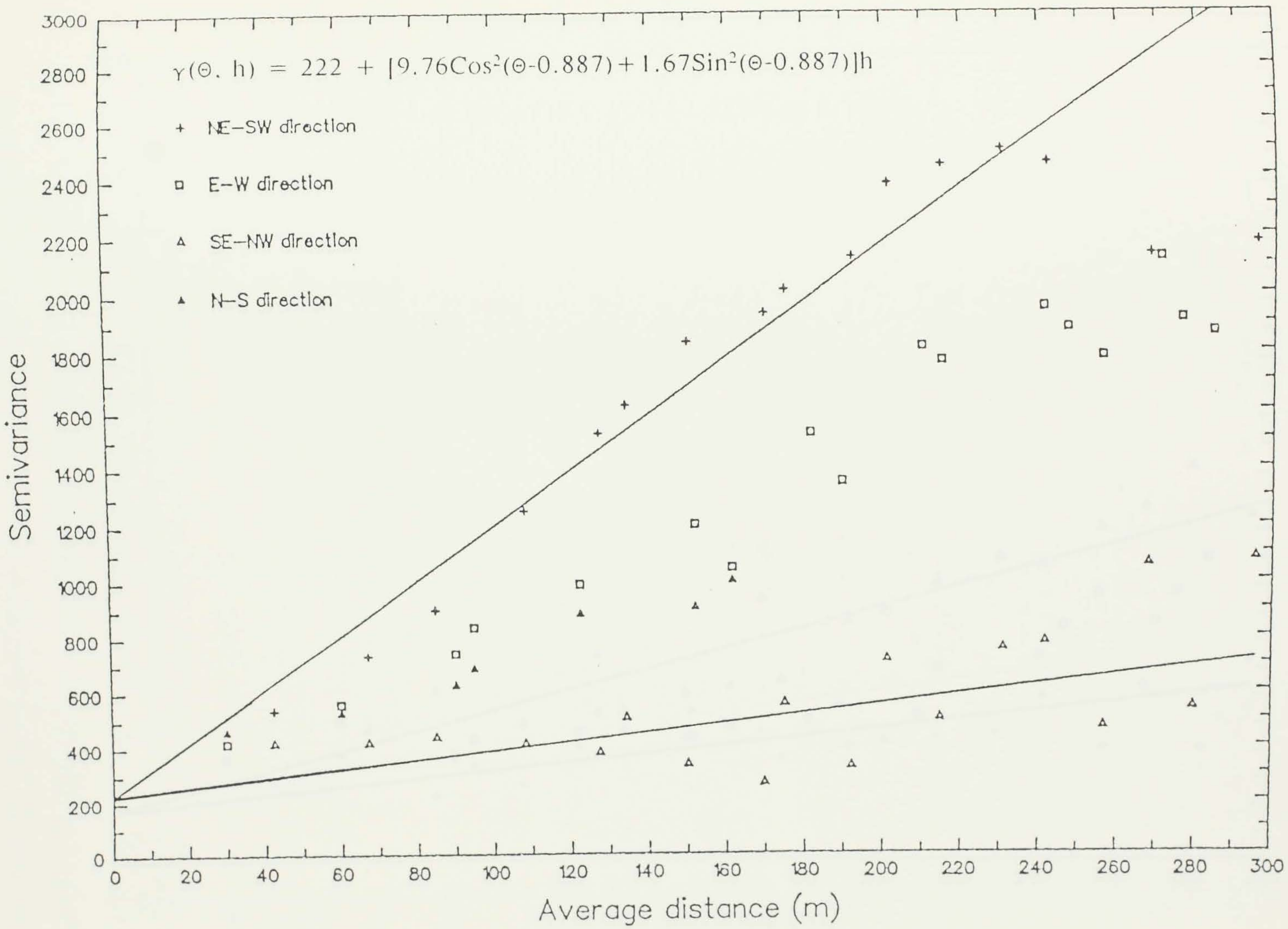
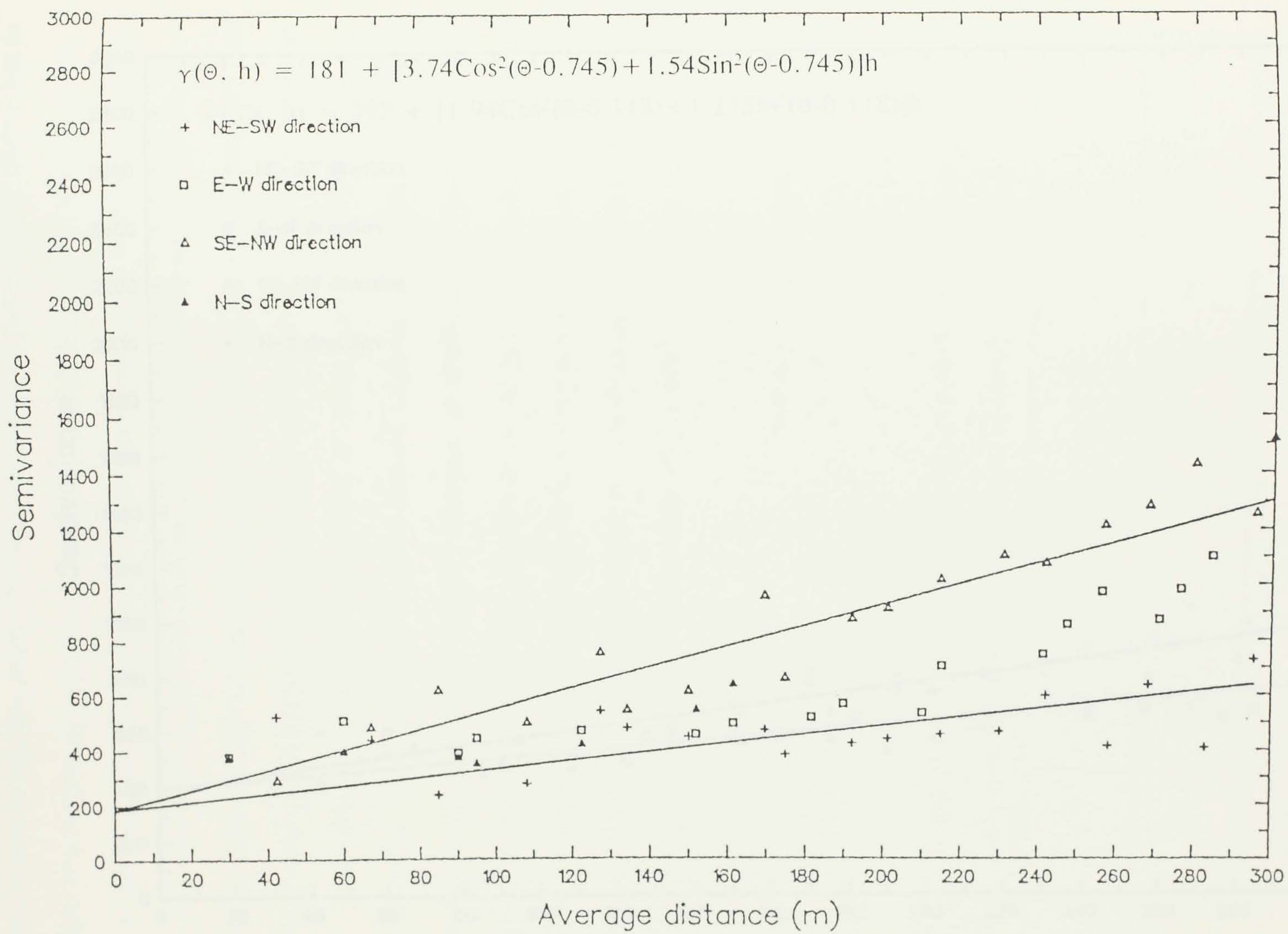

γ(Θ, h) = 222 + [9.76Cos²(Θ-0.887)+1.67Sin²(Θ-0.887)]h
NE-SW direction
E-W direction
SE-NW direction
N-S direction
Semivariance
Average distance (m)
0
200
400
600
800
1000
1200
1400
1600
1800
2000
2200
2400
2600
2800
3000
20
40
60
80
100
120
140
160
180
200
220
240
260
280
300

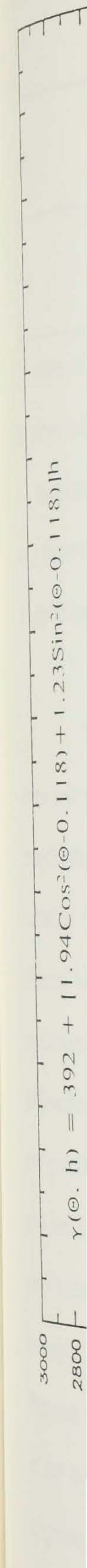
Figure 4.10 Fitted anisotropic model for DG

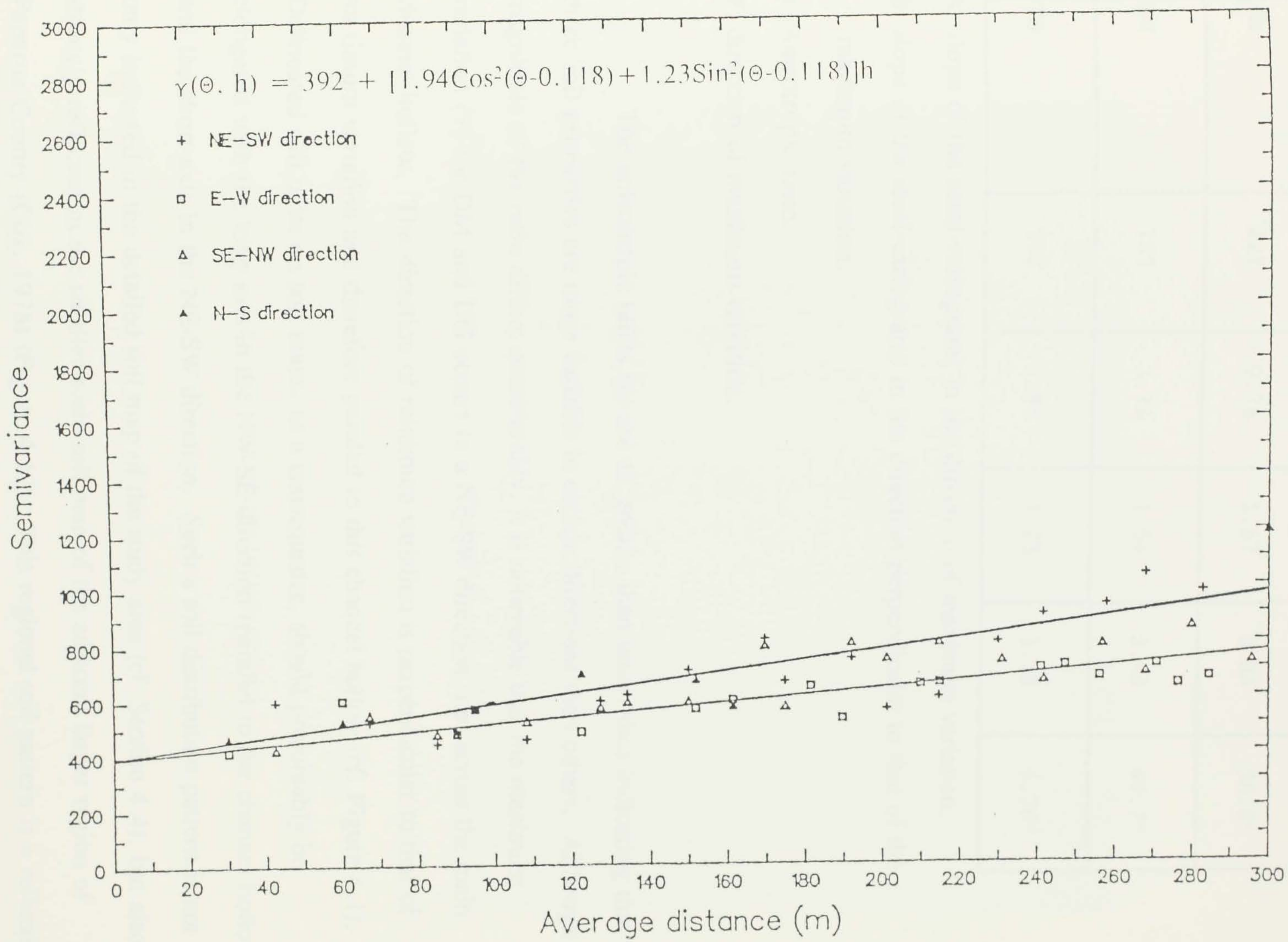
The two solid lines define the envelope within which the fitted model lies.



γ(Θ, h) = 181 + [3.74Cos²(Θ-0.745)+1.54Sin²(Θ-0.745)]h
NE--SW direction
E--W direction
SE--NW direction
N--S direction
Semivariance
Average distance (m)

Figure 4.11 Fitted anisotropic model for TS
The two solid lines define the envelope within which the fitted model lies.





$\gamma(\Theta, h) = 392 + [1.94Cos^2(\Theta\text{-}0.118)+1.23Sin^2(\Theta\text{-}0.118)]h$
NE–SW direction
E–W direction
SE–NW direction
N–S direction
Semivariance
Average distance (m)
0
20
40
60
80
100
120
140
160
180
200
220
240
260
280
300
200
400
600
800
1000
1200
1400
1600
1800
2000
2200
2400
2600
2800
3000

Table 4.2 Parameters of anisotropic semi-variograms

Soil properties	C_0	A	B	k (A/B)	Ψ
DM	222	9.76	1.67	5.84	50.8°
DG	181	3.74	1.54	2.43	42.7°
TS	392	1.94	1.23	1.58	6.76°

A: slope of the semi-variograms in the direction of maximum variation.

B: slope of the semi-variograms in the direction perpendicular to that of the maximum variation.

k: Anisotropic ratio.

Ψ: direction of maximum variation.

The anisotropic ratios (k) are all greater than unity, thus indicating that the three soil properties are more variable in certain directions than others. Although the magnitude of the ratio differs considerably, it is noticeable that the maximum variation (Ψ) for DM and DG occurs in a NE-SW direction, i.e. across the main channel hollow. The direction of minimum variation is perpendicular to that of maximum variation and therefore parallel to this channel hollow (cf. Figure 4.1). Delineated soil units on soil maps, as a consequence, should presumably be elongated with the long axis in the NW-SE direction (parallel to the channel hollow) and the short axis in the NE-SW direction. Such a soil distribution pattern is not only indicated in the detailed soil map of the study area (cf. Section 4.4), but also strongly reflected in the smaller-scale soil map of the adjacent larger region of Paparua County (Cox, 1978) (Figure 4.12). This regional soil pattern is a reflection of the overall fluvial depositional environment associated with the various ancient migrating channels of the NW-SE flowing Waimakariri river.

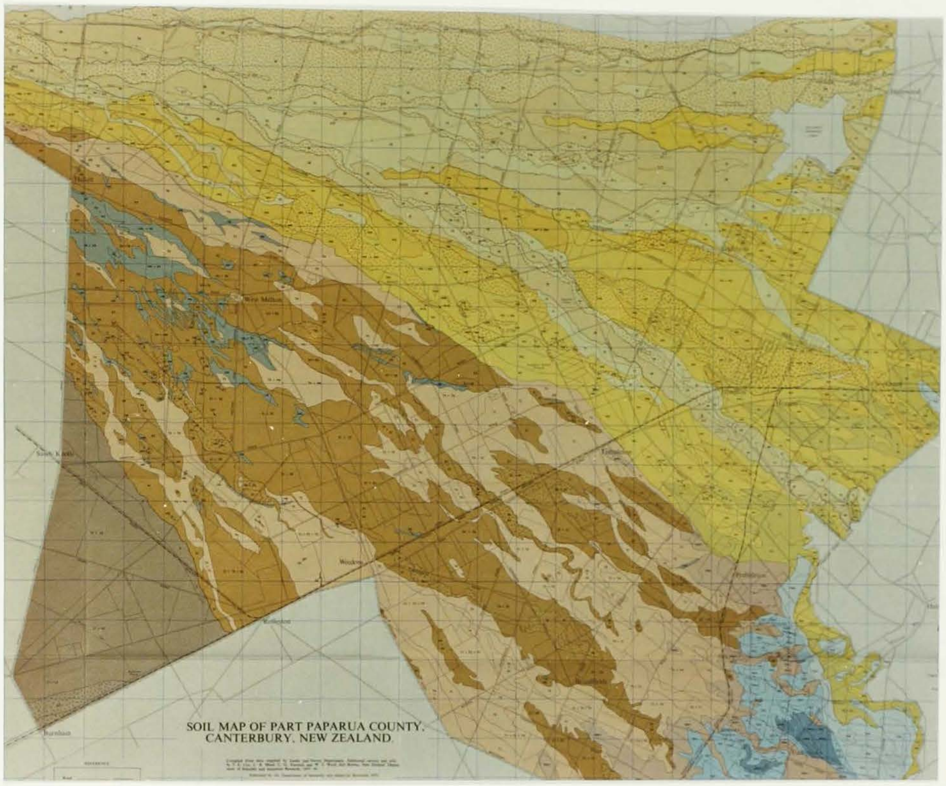


Figure 4.12 Soil map of part Paparua County, Canterbury, New Zealand (from Cox, 1978) in which there is a general NW-SE alignment of delineated soil units

4.3.4 Kriged isarithmic maps

Figures 4.13-4.15 compare the conventional property contour maps derived from the 30 m grid data with the equivalent block-kriged maps.

The patterns within both maps for each property are generally similar. A conspicuous characteristic of kriged property maps, however, is the regularised nature of the contour lines which are not simply interpolated between adjacent data points as with the conventional maps: these kriged contours are estimated on the basis of observations within the whole specified neighbourhood, each of the observations being weighted according to the spatial relationships that are reflected in the semi-variograms. Very sharp changes in soil properties are therefore smoothed on the basis of the spatial dependence in the neighbourhood. If there are only one or two observations that are distinctly different from the surrounding observations they tend to be completely removed from the block kriged map. Particularly noticeable is the way the abandoned channel running NW-SE across the eastern part of the area, which is clearly isolated on the conventional DG and TS contour maps, is not depicted on the block kriged counterparts. The smoothing effect, however, is influenced by the kriging method and the searching radius within which observations are weighted for kriging. If punctual kriging had been used and/or a shorter searching radius adopted, more weights would have been given to the data points inside the isolated small areas and the kriged soil map would probably be even more similar to the manually drawn map.

Estimation can be by extrapolation as well as interpolation. For instance, the two zones (NE and SW corners) outside the paddock boundaries or occupied by houses, which were not included within the original survey, have been extrapolated on the kriged maps. Such estimates, however, would be expected to have high errors attached to them. One advantage of the kriged contour lines is that they are drawn with known estimation errors. Estimates are conducted based on regionalised areas rather than on single specific points. Kriging standard errors are further considered in Section 4.5.

Figure 4.13 (a) Conventional contour maps and (b) equivalent block-kriged maps of DM derived from the same 30 m grid data
ns = areas where no grid data was collected.

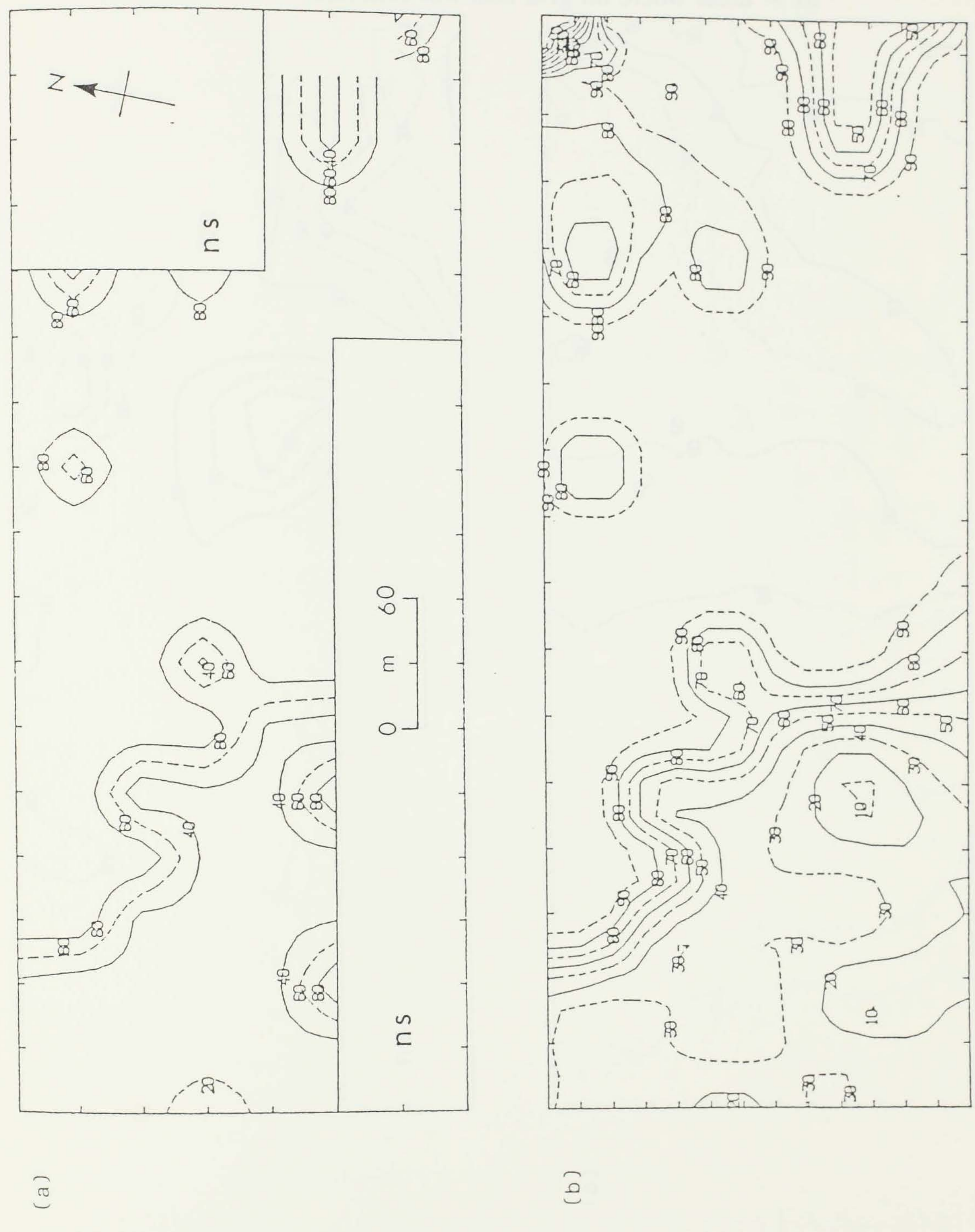

(a)
(b)
N
n s
n s
0 m 60

Figure 4.14 (a) Conventional contour maps and (b) equivalent block-kriged maps of DG derived from the same 30 m grid data

ns = areas where no grid data was collected.

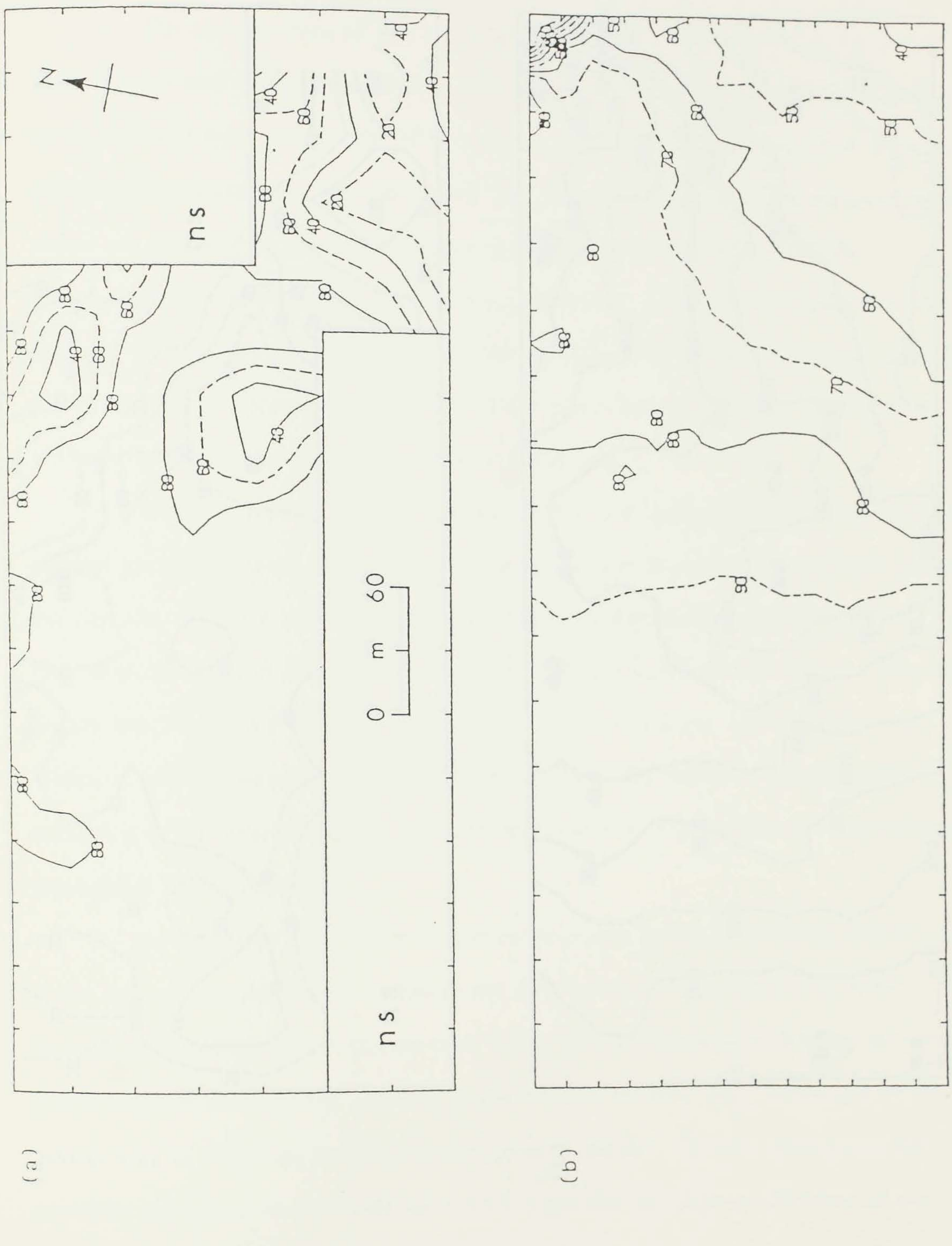
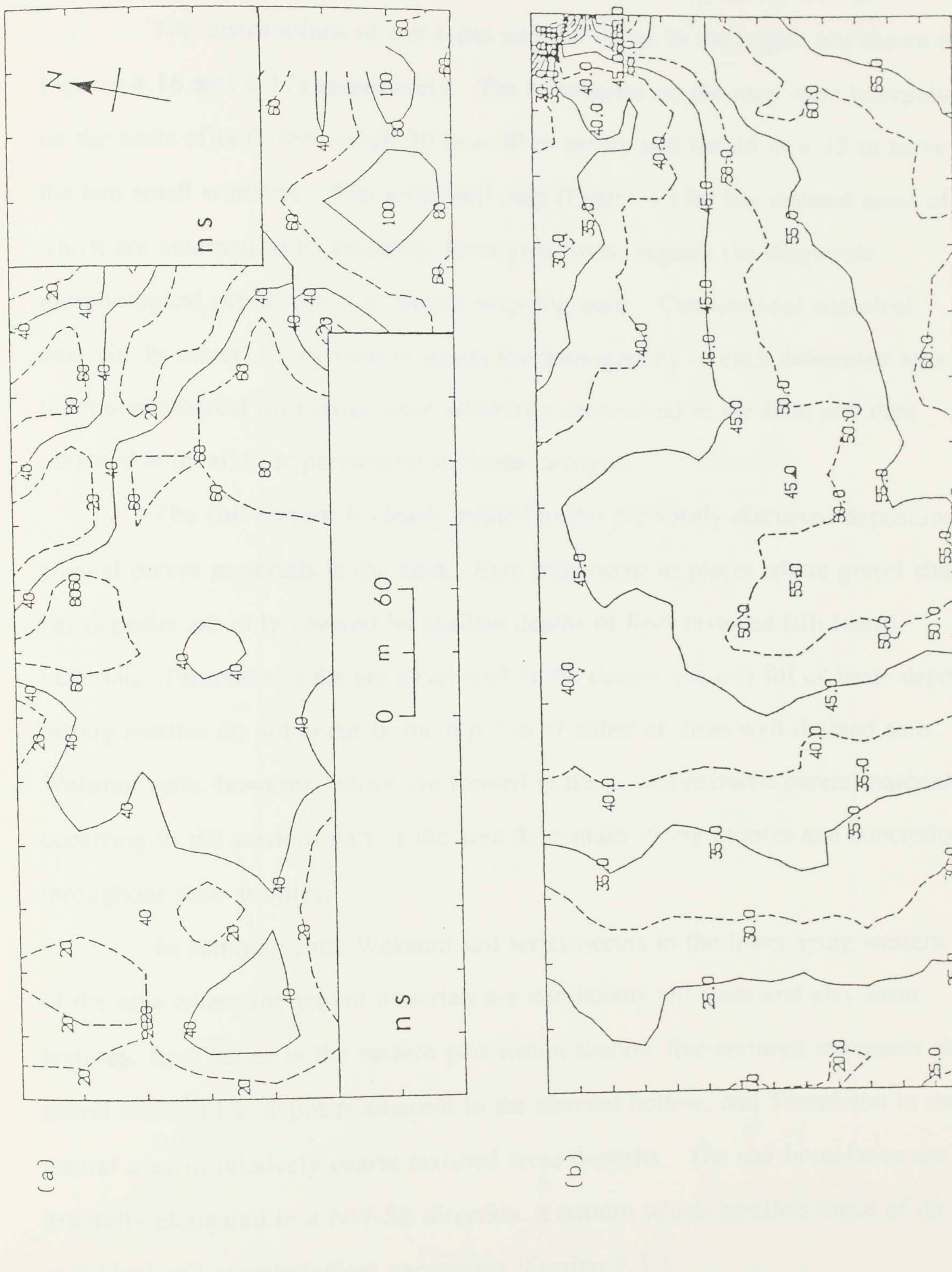

N
n s
0 m 60
n s
(a)
(b)

Figure 4.15 (a) Conventional contour maps and (b) equivalent block-kriged maps of TS derived from the same 30 m grid data ns = areas where no grid data was collected.


(a)
(b)
0 m 60
ns
ns

4.4 Soil classification and mapping

The classification criteria described by Cox (1978). and outlined in Section 3.3. were used to divide the soils of the study area into three soil series. Eyre. Templeton. and Wakanui. Each series was further subdivided into soil types.

The distributions of soil types and soil series in the region are shown in Figures 4.16 and 4.17a respectively. The boundaries on the map were interpolated on the basis of both the overall 30 m x 30 m survey and the 15 m x 15 m survey of the two small windows. The series soil map (Figure 4.17a) has isolated areas of land which are assumed to be relatively homogeneous as regards the diagnostic morphological properties. i.e. simple mapping units. Conventional statistical analysis. however. is not used to assess the homogeneity of each delineated area as the morphological properties were arbitrarily determined in the field and data obtained is invalid for parametric statistical analysis.

The soil pattern is clearly related to the previously-discussed deposition of alluvial parent materials in the area. Eyre soils occur in places where gravel channel-bar deposits are only covered by shallow depths of finer-textured (silt loam) materials. Templeton soils are developed in the deeper channel-fill or levee deposits. Strong mottles do not occur in the top 1 m of either of these well-drained soils. Wakanui soils. however. which are formed in deep. fine-textured parent materials occurring in the western part of the area do contain strong mottles and concretions throughout their profiles.

In summary. the Wakanui soil series occurs in the lower-lying western part of the area where the parent materials are dominantly silt loam and clay loam textures. Eyre series in the eastern part within shallow fine-textured sediments above gravel channel-bar deposits adjacent to the channel hollow. and Templeton in the central area in relatively coarse-textured levee deposits. The soil boundaries are generally elongated in a NW-SE direction. a pattern which parallels those of the individual soil morphological parameters (Section 4.3.1)

Figure 4.16 Conventional soil map of the study area
Boundaries drawn by manual interpretation between observation points.

Key to the soil map

Eyre Series:

(1) E_1: Eyre shallow silt loam
(2) E_2: Eyre very shallow silt loam
(3) E_3: Eyre stony silt loam
(4) E_4: Eyre shallow fine sandy loam
(5) E_5: Eyre very shallow fine sandy loam
(6) E_6: Eyre stony sandy loam
(7) E_7: Eyre very stony sandy loam.

Templeton Series:

(1) T_1: Templeton silt loam
(2) T_2: Templeton silt loam on loamy sand
(3) T_3: Templeton silt loam, moderately deep phase
(4) T_4: Templeton fine sandy loam
(5) T_5: Templeton fine sandy loam on sand
(6) T_6: Templeton fine sandy loam, moderately deep phase

Wakanui Series:

(1) WK_1: Wakanui silt loam
(2) WK_2: Wakanui silt loam on loamy sand

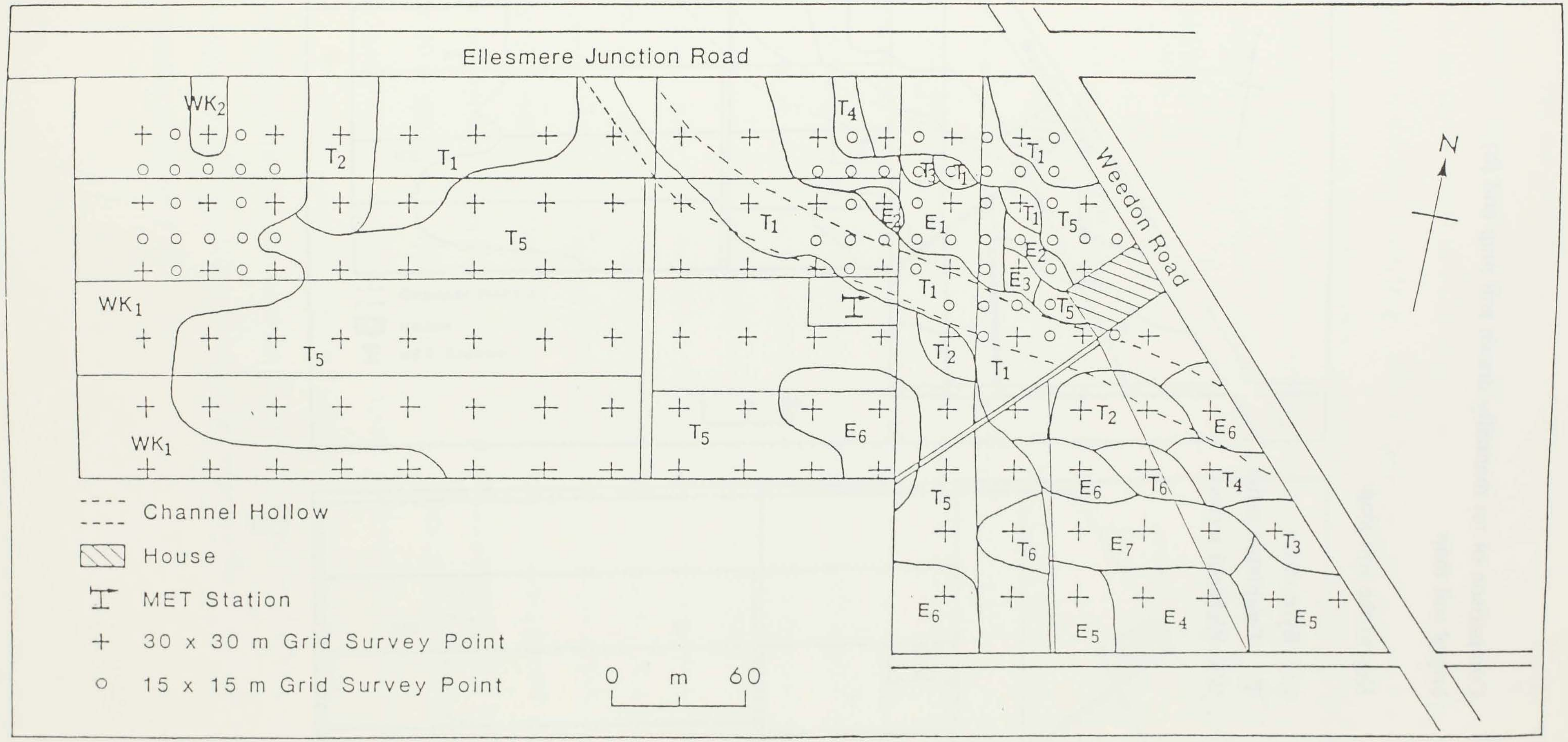

Ellesmere Junction Road
Weedon Road
N
Channel Hollow
House
MET Station
30 x 30 m Grid Survey Point
15 x 15 m Grid Survey Point
0 m 60
WK1
WK2
T1
T2
T3
T4
T5
T6
E1
E2
E3
E4
E5
E6
E7

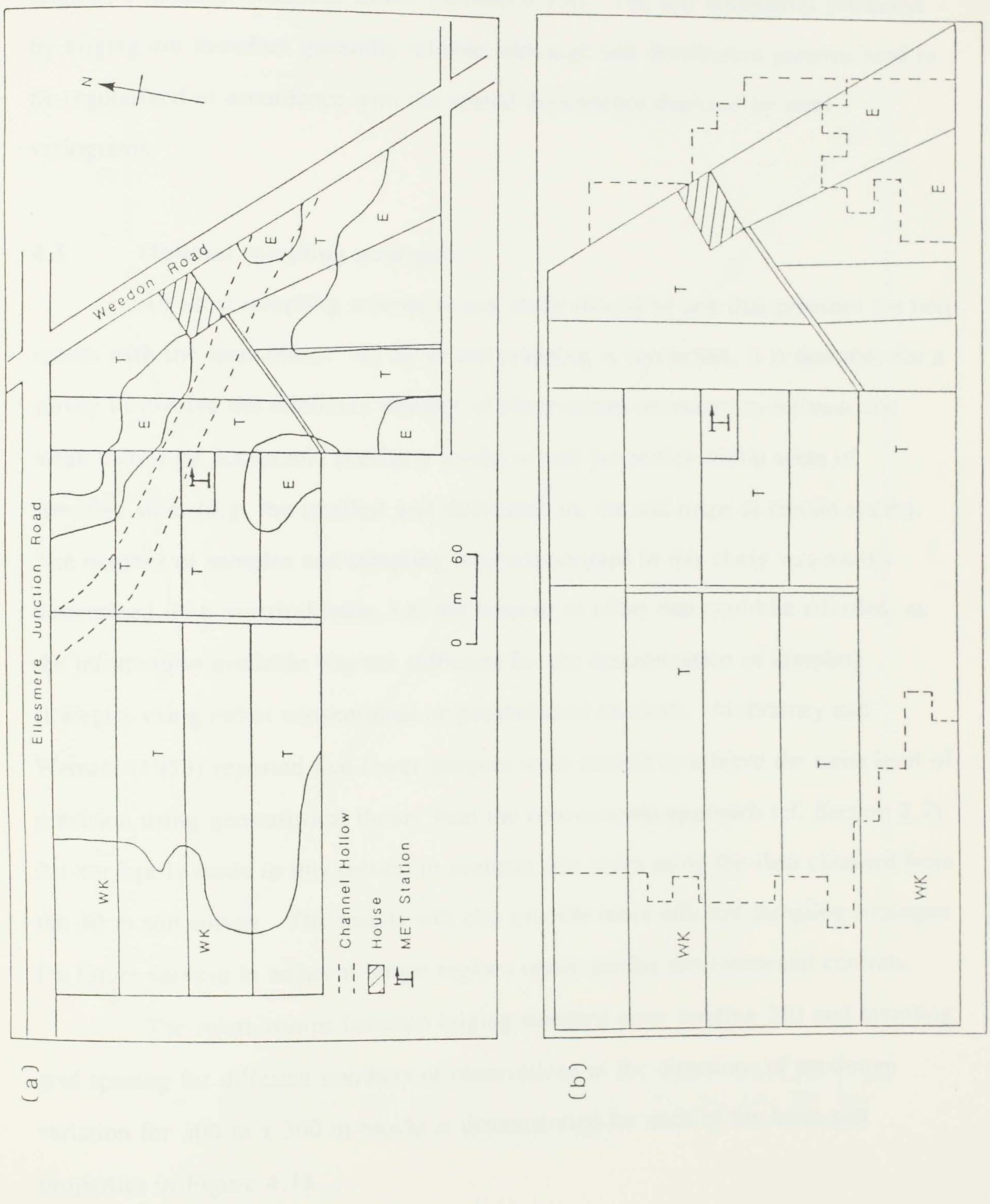
Figure 4.17 Comparison of (a) manually-drawn soil map and (b) kriged soil map

Key to the soil map

E: Eyre series

T: Templeton series

WK: Wakanui series


(a)
(b)
Weedon Road
Ellesmere Junction Road
N
0 m 60
Channel Hollow
House
MET Station
WK
T
E

The soil pattern of the kriged map (Figure 4.17b) is similar to that of its conventional equivalent. The main difference is that some small isolated patches within the conventional soil map (e.g. the Eyre series in the northeast corner) disappear in the kriged soil map. This is due to the smoothing effect of block kriging, a function discussed earlier (Section 4.3.4). The soil boundaries predicted by kriging are therefore generally reliable, although soil distribution patterns tend to be regularized in accordance with the spatial dependence depicted by semi-variograms.

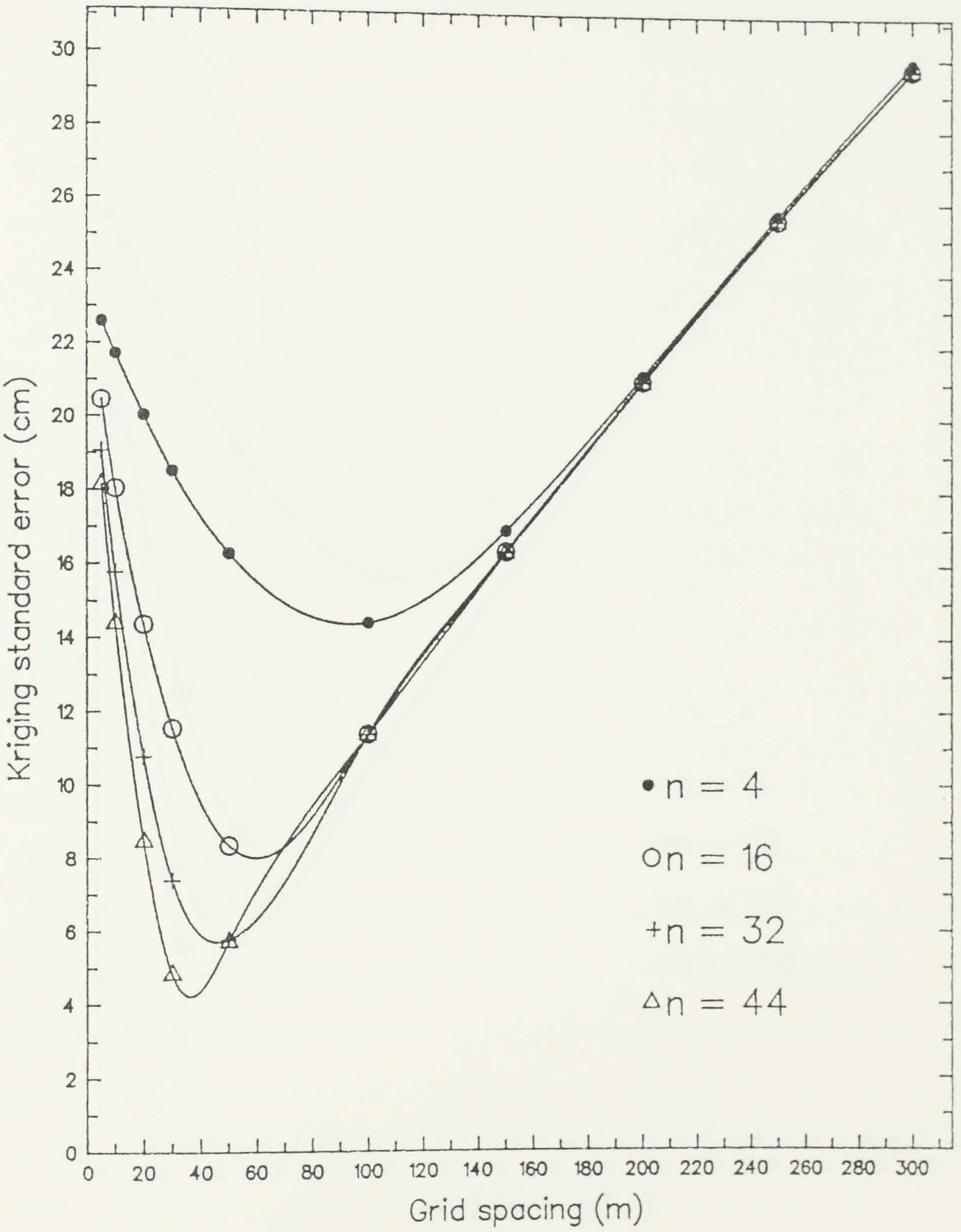
4.5 Optimal sampling strategies

An ideal sampling scheme in any study should be one that provides the best results with the least effort. As far as soil mapping is concerned, it is desirable for a survey to involve the minimum number of observations necessary to estimate the mean values (at acceptable precision levels) of soil properties within areas of specified sizes (e.g. the smallest unit delineated on the soil maps at certain scales). The number of samples and sampling intervals adopted in this study was mainly determined on a practical basis, i.e. the amount of effort that could be afforded, as the information available was not sufficient for the determination of sampling strategies using either conventional or geostatistical methods. McBratney and Webster (1983) reported that fewer samples were needed to achieve the same level of precision using geostatistical theory than the conventional approach (cf. Section 2.7). An attempt is made in this section to evaluate this claim using the data obtained from the 30 m soil survey. The results will also provide more efficient sampling strategies for future surveys in adjacent larger regions under similar environmental controls.

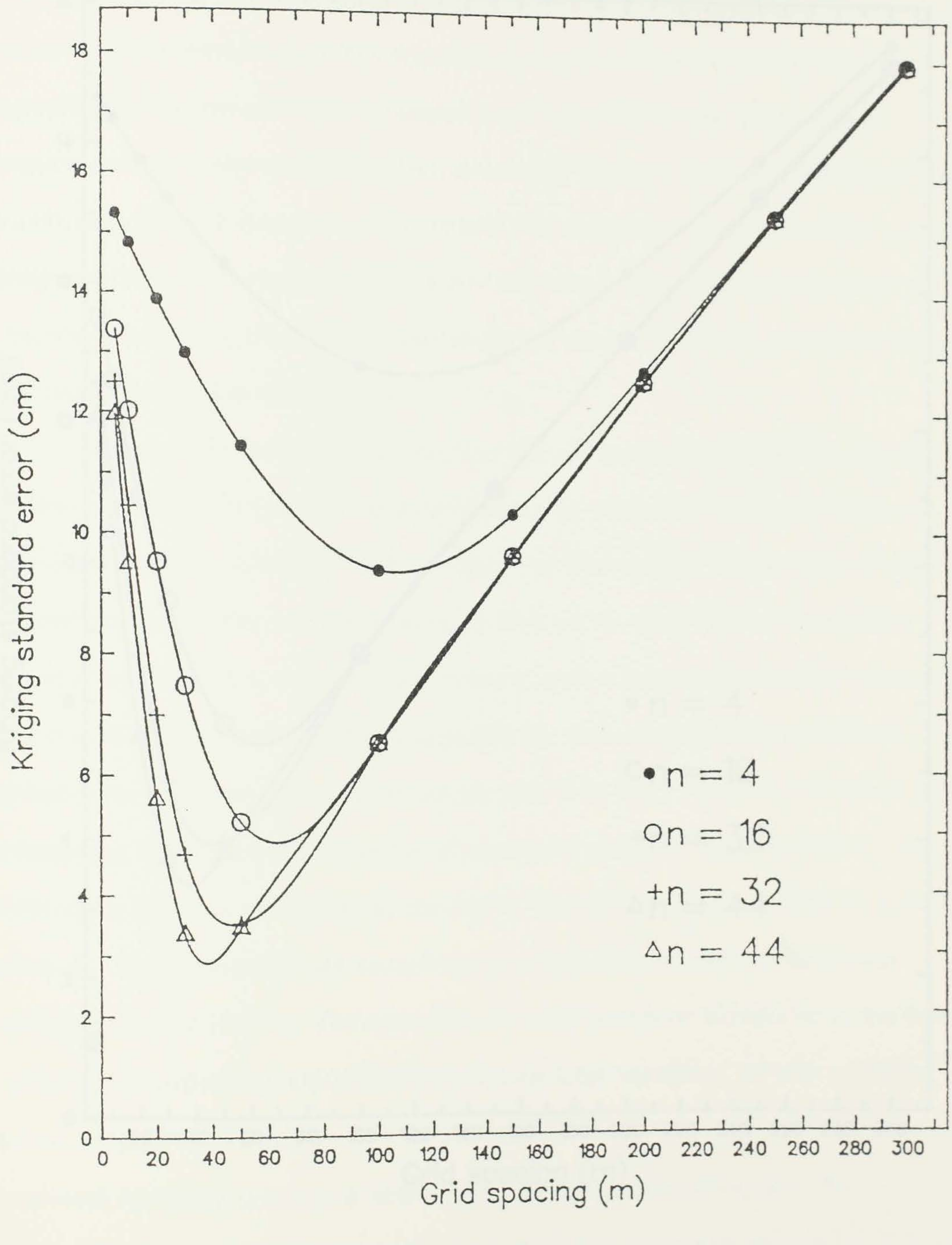
The relationships between kriging standard error (kriging SE) and sampling grid spacing for different numbers of observations in the directions of maximum variation for 300 m x 300 m blocks is demonstrated for each of the three soil properties in Figure 4.18.

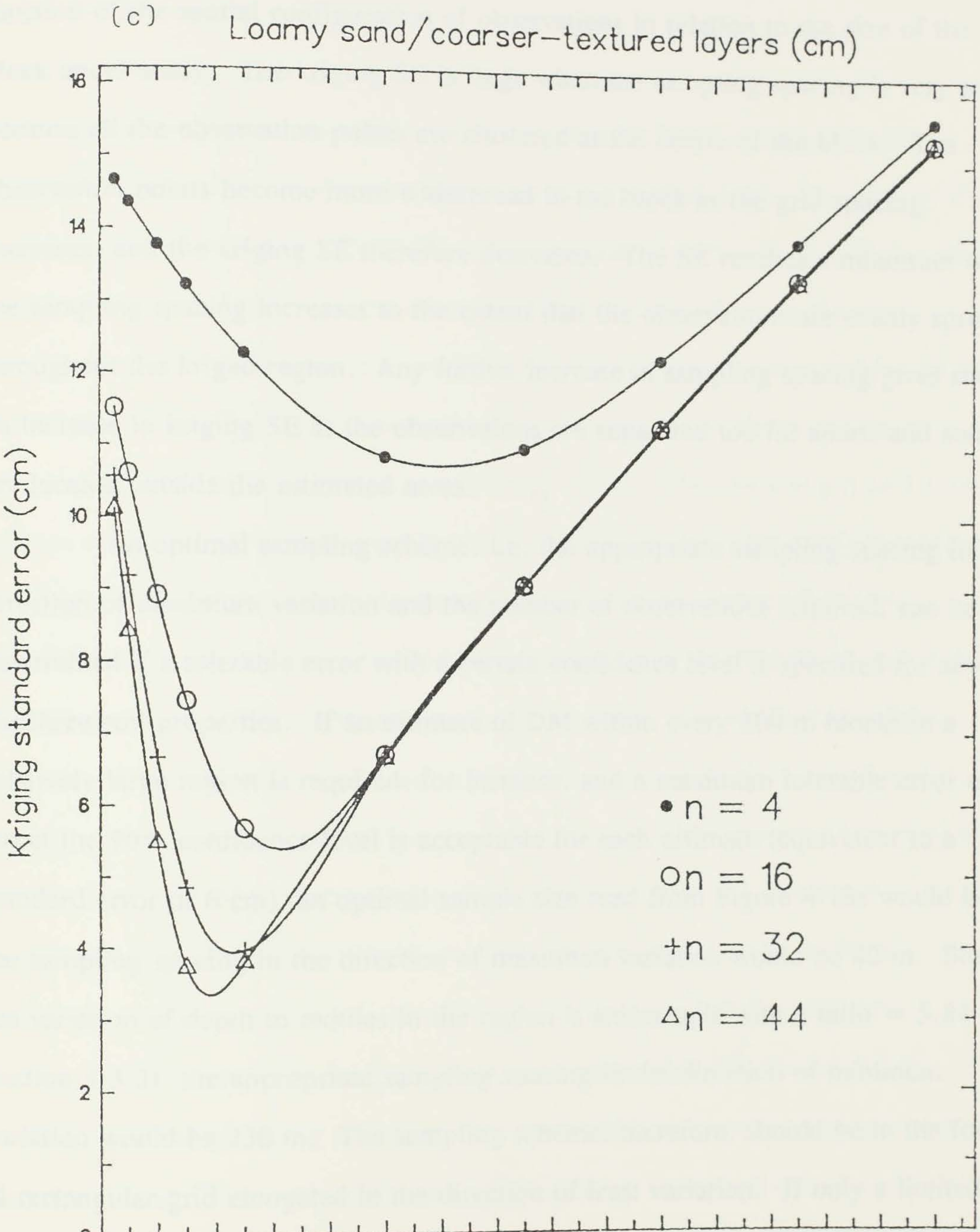
Figure 4.18 Relationships between kriging standard error and sample grid spacing for different numbers of observations in the directions of maximum variation within a 300 m × 300 m block for (a) DM, (b) DG and (c) TS

(a)


Kriging standard error (cm)
Grid spacing (m)
0
2
4
6
8
10
12
14
16
18
20
22
24
26
28
30
0 20 40 60 80 100 120 140 160 180 200 220 240 260 280 300
• n = 4
○ n = 16
+ n = 32
△ n = 44

(b)


18
16
14
12
10
8
6
4
2
0
Kriging standard error (cm)
0 20 40 60 80 100 120 140 160 180 200 220 240 260 280 300
Grid spacing (m)
• n = 4
o n = 16
+ n = 32
△ n = 44


(c)
Loamy sand/coarser-textured layers (cm)
Kriging standard error (cm)
Grid spacing (m)
• n = 4
○ n = 16
+ n = 32
△ n = 44

The kriging SE generally decreases with increase in sample numbers, though beyond a certain sampling spacing it tends to have the same value irrespective of sample number. The kriging SE for uniform sample numbers initially decreases, then increases rapidly, with increase in sample spacing. These relationships are a function of the spatial configuration of observations in relation to the size of the block under study. The kriging SE is large when the sampling spacing is very small, because all the observation points are clustered at the centre of the block. The observation points become more widespread in the block as the grid spacing increases, and the kriging SE therefore decreases. The SE reaches a minimum when the sampling spacing increases to the extent that the observations are evenly spread throughout the kriged region. Any further increase in sampling spacing gives rise to an increase in kriging SE as the observations are separated too far apart, and some are located outside the estimated areas.

An optimal sampling scheme, i.e. the appropriate sampling spacing in the direction of maximum variation and the number of observations required, can be determined if a tolerable error with a certain confidence level is specified for any of the three soil properties. If an estimate of DM within every 300 m blocks in a relatively large region is required, for instance, and a maximum tolerable error of 10 cm at the 90% confidence level is acceptable for each estimate (equivalent to a standard error of 6 cm), an optimal sample size read from Figure 4.18a would be 32; the sampling spacing in the direction of maximum variation would be 40 m. Since the variation of depth to mottles in the region is anisotropic with a ratio = 5.84 (cf. Section 4.3.2), the appropriate sampling spacing in the direction of minimum variation would be 230 m. The sampling scheme, therefore, should be in the form of rectangular grid elongated in the direction of least variation. If only a limited number of observations can be afforded, however, the graphs can be used to indicate the optimal sampling spacing in accordance with the minimum kriging SE.

The minimised SE for each of the three soil properties derived from kriging and conventional estimations were plotted against the number of observations for three different sizes of blocks. The block sizes (50 m × 50 m, 100 m × 100 m, and 300 m × 300 m) were chosen arbitrarily to represent the smallest units likely to be delineated on soil maps of different scales. Figure 4.19 shows the results from the 100 m blocks, whilst graphs for the 50 m and 300 m blocks are presented in Appendix 2.

The kriging SE is considerably lower than that estimated by the conventional method for the same number of observations. Kriging therefore requires less observations than the conventional estimation method to achieve the same amount of precision. If DM is to be estimated for a block of 100 m × 100 m (1 ha), for example, and the maximum tolerable error is 10 cm at the 90% confidence level (equivalent to 6 cm SE), only 14 samples are required for kriging, but 34 are needed for the conventional estimation method. These results are similar to those obtained by McBratney and Webster (1983), who claimed a 3 to 9 times gain in efficiency by kriging over the conventional method. This gain is explained by the fact that kriging takes into account the spatial dependence between observations in the region, whereas random variation is assumed by conventional statistical theory. This account of spatial dependence also means that kriging becomes more advantageous in terms of estimation error as the blocks become smaller. This feature is not very conspicuous for the TS parameter, however, due to the large amount of nugget variance in its semi-variogram (cf. Section 4.3.2). The advantage of geostatistical method over the conventional disappears when observations are separated too far apart and become spatially independent.

This part of the study provides some important information for the determination of sampling strategies for future soil survey and mapping in adjacent larger regions of similar environment. The optimal number of samples and their field configurations can be determined for certain desired precisions using the methods as described above. Soil boundaries can be kriged based on the spatial relationships among the observations as expressed in semi-variograms. Where there are different environmental controls and spatial relationships it is recommended that

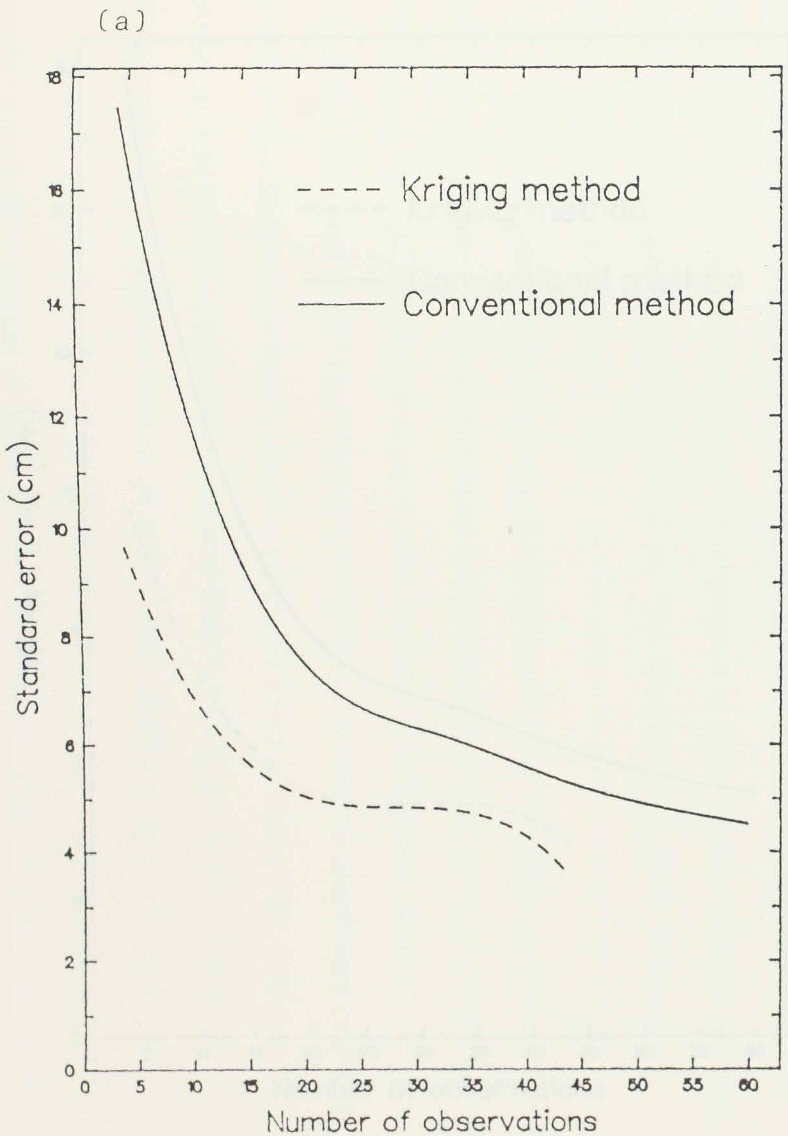
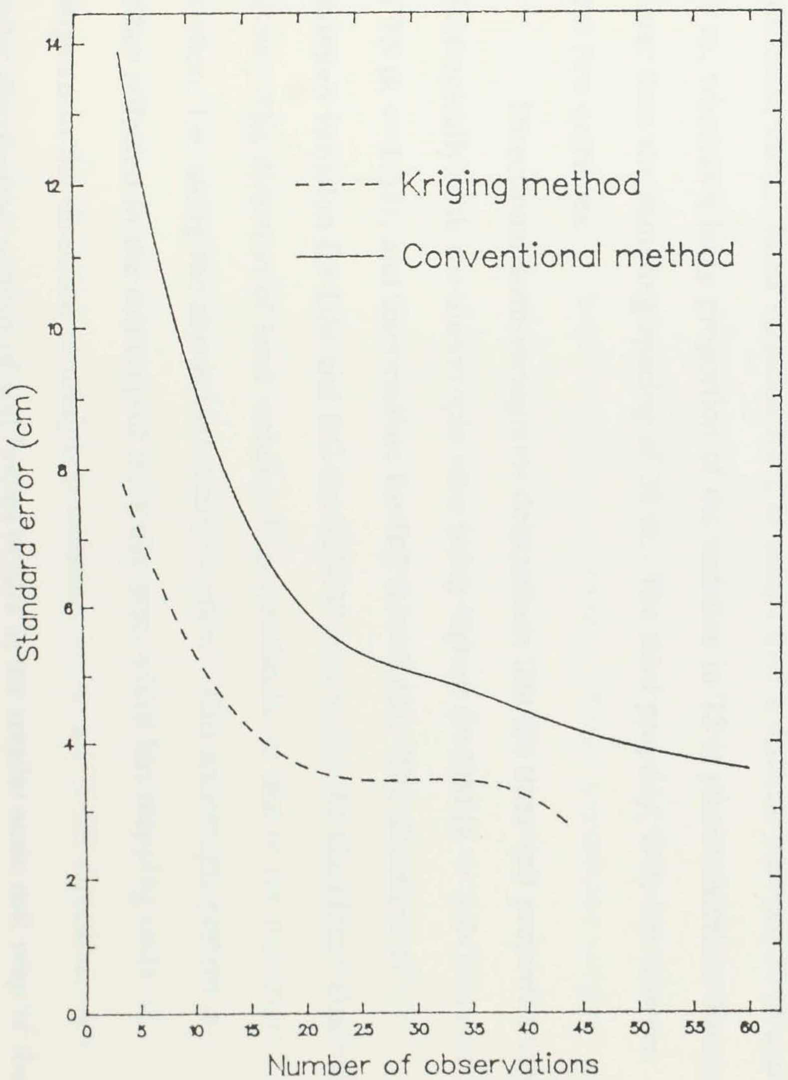


Figure 4.19

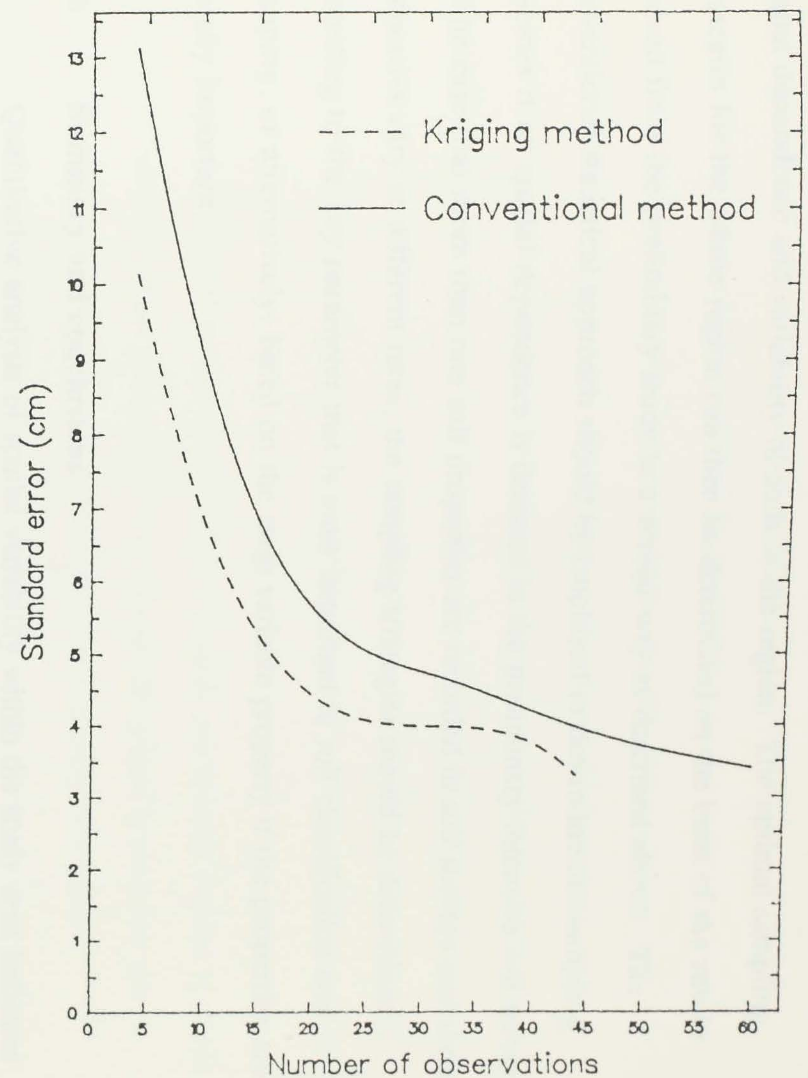
Graphs of kriging and conventional standard error against sample size with 100 m × 100 m blocks for (a) DM, (b) DG and (c) TS

(b)



Kriging method
Conventional method
Standard error (cm)
Number of observations

(c)



Kriging method
Conventional method
Standard error (cm)
Number of observations

an intensive soil survey be conducted first on a small representative area to reveal the spatial dependence and variability of soils in the region. The optimal sampling strategies for the whole region can then be determined on the basis of the results derived from the preliminary study in a similar way as described above. The conventional statistical approach should be employed to determine the sampling schemes if no spatial dependence is detected in the preliminary intensive soil surveys. In the case that more than one soil properties are recorded in soil surveys and these properties vary at different rates, the sampling strategies should be determined according to the key parameter that is most important to soil classification and mapping, or alternatively, based on the most variable property if the properties are equally important.

4.6 Summary and conclusions

Quantitative analysis of spatial variability within the study area indicates that the three soil morphological properties, depth to strong mottles (DM), depth to gravels (DG), and thickness of loamy sand and/or coarser-textured layers (TS), vary at different rates. Most variation for DM occurs over a distance between 30 m and 430 m, whereas a large proportion of the variation in TS is present within a distance shorter than the sampling spacing of 30 m. The third property, DG, lies between these two extremes.

Directional semi-variograms demonstrate that the three soil properties vary anisotropically with the anisotropic ratio being highest for DM ($k = 5.84$), lowest for TS ($k = 1.58$), and intermediate for DG ($k = 2.43$). The directions of maximum variation for DM and DG are NE-SW, i.e. across the abandoned channel hollow. The direction of least variation is perpendicular to that of the maximum variation, i.e. along the abandoned channel hollow. This anisotropic pattern is further reflected in the soil map of the study area, which has mapping units elongated in the direction of minimum variation, i.e. in a NW-SE direction. A similar distribution pattern of soil bodies occurs in the smaller-scale soil map of the adjacent regions. Such variations reflect the general past drainage patterns of channels flowing in a NW-SE direction across the broad region.

The spatial variation of the three soil morphological properties and the related distribution of soil series are closely related to the pattern of alluvial deposition within the study area. Eyre soils are developed in thin layers of finer-textured materials overlying gravel channel-bar deposits. Templeton soils in sandy channel-fill and levee deposits. and Wakanui soils in finer-textured sediments characteristic of an intermediate zone between levee and floodbasin deposits. This distribution is similar to the general soil pattern depicted by Cox (1978) in adjacent regions.

The kriged soil property and soil series maps are broadly similar to those interpolated manually from the survey data. though the kriged boundaries are regularised and some isolated small parcels that differ sharply from their neighbourhoods are removed.

Optimal sampling schemes for soil survey and variability studies can be determined based on kriging standard errors. Kriging SE can be computed for different sampling spacings in the direction of maximum variation and different number of observations given the sizes of blocks. The optimal sampling spacing and sample size to achieve certain specified precision can be read from the graphs showing the relations between the three parameters. If only a restricted numbers of observations can be afforded. however. the optimal sampling spacing in the direction of maximum variation can be obtained to minimise the estimation error. The sampling spacing in the direction of least variation is k (anisotropic ratio) times the spacing in the direction of most variation. i.e. a rectangular scheme elongated in the direction of minimum variation.

Less samples are needed for kriging than for the conventional method to achieve the same level of precision. The amount of gain in efficiency by kriging over the conventional method in this study is similar to that claimed by McBratney and Webster (1983) in their studies. These results provide useful guidelines for sampling strategies for future soil surveys in adjacent larger regions. In dissimilar regions, it is recommended that intensive soil survey be conducted first on small representative areas to reveal the spatial dependence of soil properties. Optimal sampling strategies for the whole region can then be determined on the basis of the results derived from the preliminary study in a similar way as this study.

CHAPTER 5

PHYSICAL PROPERTIES OF A TYPICAL PROFILE FROM EACH SOIL SERIES

5.1 Introduction

Soils in the study area have been delineated into three simple mapping units at series level according to morphological criteria (Figure 5.1). The high individual purity of each mapping unit and the diagnostic differences between mapping units were assessed solely in terms of morphological properties. The fundamental assumption of soil mapping, however, is that other related (but not so easily-measured) accessory properties should display similar spatial patterns as the diagnostic morphological properties. Such mapping units should therefore be individually pure and distinct from other mapping units in terms of these accessory properties.

The diagnostic textural and mottling properties which distinguish the Eyre, Templeton and Wakanui series are normally used to predict related accessory physical properties, particularly those concerned with water movement and storage. Direct assessment of soil physical (and chemical) properties of a simple mapping unit is traditionally achieved by sampling and analysing a single profile considered characteristic of the soil series taxonomic unit (e.g. Joe and Watt, 1983; Joe, 1984). The same approach is followed in this chapter in order to establish and generally compare the physical properties of the Eyre, Templeton and Wakanui series taxonomic units from a relatively less quantitative perspective. The succeeding chapter uses data from a larger number of sampling points to statistically quantify any differences, and to examine the purity of each taxonomic unit from a physical property perspective.

5.2 Methods

The locations of the soil profiles representative of the three soil series taxonomic units were determined from the soil map produced in Chapter 4 (Figure 4.16). Each soil pit was dug in the central part of the appropriate soil body (Figure 5.1). Soil profiles were described using standard terminology (Taylor and Pohlen. 1979). The Templeton and Wakanui profiles were described to a depth of 100 cm; the Eyre profile was restricted to 25 cm depth. the level at which continuous gravels were encountered.

Six soil physical properties were measured or assessed in each soil profile: particle size distribution. particle density. bulk density. equivalent pore-size distribution. pore pattern and "field-saturated" hydraulic conductivity (K_{fs}). These properties were chosen as they relate to. and directly or indirectly determine. water movement and storage characteristics of the soils. Other relevant physical properties are not concerned in this study because of time constraints.

Samples were only taken from the surface horizon of the Eyre profile (at the equivalent depth to those in the Templeton and Wakanui A horizons). No attempt was made to determine the properties at lower depths in the Eyre profile due to the difficulties of sampling within gravels. In order to substantiate the described textural changes within and between the Templeton and Wakanui profiles. a sample for particle-size analysis was collected from every horizon (Table 5.1). Samples for other measurements. however. were taken from selected horizons or specific depths. The derived data was considered sufficient and adequate for the interpretation of the overall hydraulic characteristics of the soil profiles. Triplicate samples were collected for each property measurement. though some undisturbed sand samples collapsed when being processed and could not be used. The aim of triplication was mainly to ensure a reliable mean value rather than to provide data for any stringent statistical analysis. T-tests. however. were applied wherever appropriate to assess the significance of noted differences between horizons or profiles. though it was appreciated that real differences may not be statistically substantiated with so few replicates.

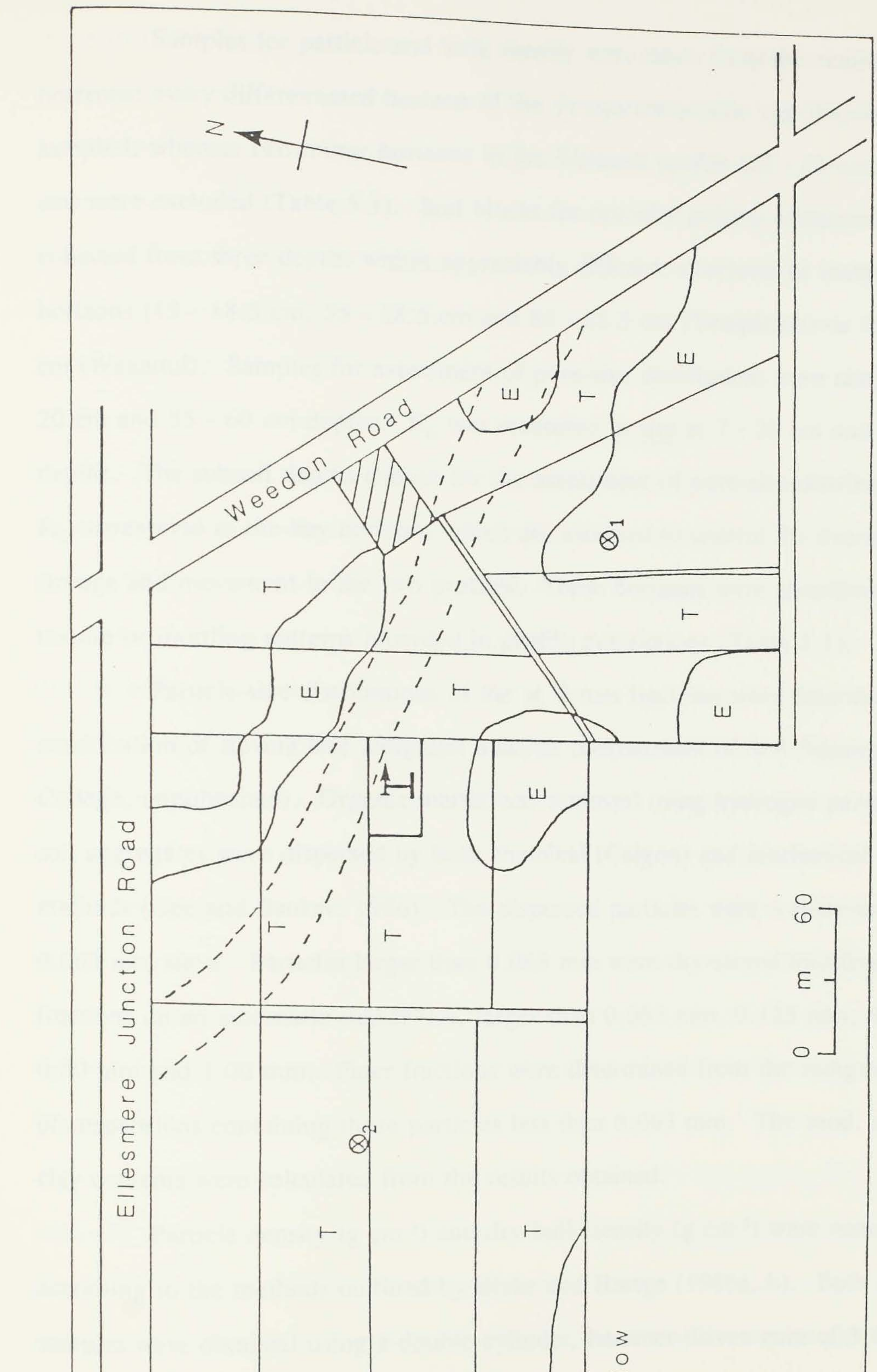
Figure 5.1 Soil map of study area and locations of three profiles

Key to the soil map

E: Eyre series

T: Templeton series

WK: Wakanui series


Ellesmere Junction Road
Weedon Road
N
T
E
WK
0 m 60
Channel Hollow
House
MET Station
Soil Profile

Samples for particle and bulk density were taken from the centres of four horizons: every differentiated horizon of the Templeton profile was therefore sampled, whereas two minor horizons in the Wakanui profile (23 - 31 cm; 60 - 74 cm) were excluded (Table 5.1). Soil blocks for porosity pattern assessment were collected from three depths within appreciably different structural or textural horizons (15 - 18.5 cm, 55 - 58.5 cm and 85 - 88.5 cm (Templeton) or 75 - 78.5 cm (Wakanui). Samples for assessment of pore-size distribution were taken at 15 - 20 cm and 55 - 60 cm depths. K_{fs} was measured in situ at 7 - 25 cm and 42 - 60 cm depths. The subsoil depths chosen for the assessment of pore-size distribution and K_{fs} correspond to the key horizons which are assumed to control the overall water storage and movement in the two profiles. These horizons were identified by the texture or mottling patterns recorded in profile descriptions (Table 5.1).

Particle-size distributions of the < 2 mm fractions were determined using a combination of sieving and sedigraph analysis (Department of Soil Science, Lincoln College, unpublished). Organic matter was removed using hydrogen peroxide, and soil aggregates were dispersed by both chemical (Calgon) and mechanical (shaking) methods (Gee and Bauker, 1986). The dispersed particles were wet-sieved through a 0.063 mm sieve. Particles larger than 0.063 mm were dry-sieved into five different fractions on an automatic shaker, i.e. larger than 0.063 mm, 0.125 mm, 0.25 mm, 0.50 mm and 1.00 mm. Finer fractions were determined from the sedigraph analysis of suspensions containing those particles less than 0.063 mm. The sand, silt and clay contents were calculated from the results obtained.

Particle density (g cm^{-3}) and dry bulk density (g cm^{-3}) were measured according to the methods outlined by Blake and Hartge (1986a, b). Bulk density samples were obtained using a double-cylinder, hammer-driven core of 5.4 cm diameter and 5 cm length. Total porosity was calculated from these two density measurements.

Equivalent-pore size distributions were determined using tension table and pressure plate apparatus. Undisturbed ring samples (10.3 cm in diameter and 5 cm in height) were first equilibrated on tension tables under three suctions: 20 mbar, 50 mbar and 100 mbar. The samples were then subsampled using smaller rings (3.2 cm in diameter and 1 cm of height). The subsamples were equilibrated on the pressure plate under two different pressures: 1500 mbar and 15.000 mbar. The amount of moisture released under each suction or pressure was determined by weighing the samples prior to and after equilibration. The soil-moisture-release characteristic curves were produced and the equivalent pore-size distributions determined (Hillel, 1982).

Macropore shapes, distributions and patterns were assessed from photographs of resin-impregnated soil blocks. The undisturbed soil blocks (7.5 cm × 6 cm × 3.5 cm) were horizontally orientated, i.e. the sample tins were pressed vertically downward into each sampled horizon. Water was removed by submerging the blocks in acetone for six weeks, the acetone being changed weekly. The blocks were then impregnated under vacuum with a polyester resin containing a U.V. dye, and left to harden for a further six weeks. The impregnated blocks were cut into sections and one surface from each block polished on a Logitech automatic grinder. The polished surfaces were then photographed under U.V. light. Pores are indicated on the photographs by white zones with dark areas representing the solid soil particles. Due to problems caused by poor impregnation of certain samples, only selected porosity images from the Wakanui and Templeton profiles are included in this chapter.

"Field-saturated" hydraulic conductivity (m s^{-1}) was measured in situ for each soil profile using a Guelph permeameter (Reynolds et al., 1983; Reynolds and Elrick, 1985, 1986, 1987). Measurements were made before the soil pits were dug and were 40 cm away from the sampled profiles. "Field-saturated" hydraulic conductivity refers to the quasi-saturated hydraulic conductivity of a porous soil medium containing entrapped air. This parameter is believed to be more appropriate than the true saturated hydraulic conductivity (K_s) as complete saturation is rarely achieved in the field situation (Reynolds et al., 1983). The K_{fs} value can be as much as 25% - 50% below K_s (Reynolds et al., 1983).

The Guelph permeameter is an "in hole" Mariotte bottle constructed of concentric, transparent plastic tubes. The inner air-inlet tube provides the air supply to the permeameter, and the outside tube provides the liquid reservoir and outlet into the well. Once the apparatus is installed into an auger hole in the field, the steady-state water recharge Q ($m^3\ s^{-1}$) necessary to maintain a constant depth of water H (m) in an uncased, cylindrical well of radius a (m) is measured. This is done by monitoring the rate of fall of the water surface in the permeameter.

The K_{fs} value is calculated from equations using either a "Laplace" or "Richards" analysis (Reynolds and Elrick, 1986). The equations are based on steady-state solutions for infiltration into unsaturated soil from a well. The "Laplace" analysis was used in this study and K_{fs} was calculated from the following equations:

$$K_{fs} = BQ \tag{5.1}$$

$$B = C/2\pi H^2[1+C/2(a/H)^2] \tag{5.2}$$

where C is a dimensionless proportionality parameter dependent on the H/a ratio. The radius (a) was 0.018 m and the constant water height (H) was 0.18 m in this study. The C value was obtained from the C vs H/a graph presented by Reynolds and Elrick (1986).

5.3 Results

5.3.1 Profile morphology and particle-size distribution

The detailed descriptions of the three soil profiles are recorded in Table 5.1. The Eyre profile has only 25 cm of fine sandy loam A horizon over loose, non-cemented gravels within a sand matrix (2Bw). Gravels are not present in either of the other two profiles. The Templeton and Wakanui A horizons have fine sandy loam and silt loam textures respectively. Both profiles contain sharply banded

Table 5.1 Soil profile descriptions

Eyre series:	
0 - 25 cm (A):	very dark greyish brown (10YR 3/2) slightly stony fine sandy loam: friable: moderately developed fine nutty and granular structure: few fine tubular pores: many fine roots.
25 - cm (2Bw):	gravels with sand matrix.
Templeton series:	
0 - 25 cm (A):	very dark greyish brown (10YR 3/2) fine sandy loam: friable: moderately developed medium nutty and granular structure: many medium tubular pores: many fine roots.
25 - 54 cm (Bw):	dark greyish brown (2.5Y 4/3) sandy loam; friable; moderately developed medium blocky structure: common medium tubular pores: few fine faint brown (7.5YR 4/4) mottles: few fine roots.
54 - 83 cm (2BC):	olive brown (2.5Y 4/4) sand; loose; single grain.
83 - 100 cm (3C):	olive (5Y 4/3) sandy loam; firm; weakly developed medium blocky structure: few medium faint brown (7.5YR 4/4) mottles.
Wakanui series:	
0 - 23 cm (A):	very dark brown (10YR 2/2) silt loam: friable: strongly developed medium granular structure: many medium and coarse tubular pores: few fine distinct dark brown (7.5YR 4/4) mottles: few fine and medium noddules: many fine roots.
23 - 31 cm (Bg):	greyish brown (2.5Y 5/2) sandy loam: friable; weakly developed fine blocky structure: few fine medium tubular

pores; many medium distinct dark yellowish brown (10YR 4/6) mottles; few fine roots.

31 - 60 cm (2BCg): greyish brown (10YR 5/2) silty clay loam; firm; moderately developed medium blocky structure; abundant medium prominent dark yellowish brown (10YR 4/6) mottles.

60 - 74 cm (3BC): greyish brown (10YR 5/2) loamy sand; loose; weakly developed fine and medium blocky structure; few fine faint dark brown (7.5YR 4/4) mottles.

74 - 85 cm (4Cg): greyish brown (10YR 5/2) sandy loam; friable; weakly developed fine and medium blocky structure; many medium prominent dark yellowish brown (10YR 4/6) mottles.

85 - 100 cm (5Cg): greyish brown (10YR 5/2) loamy sand; loose; weakly developed fine and medium blocky structure; common fine faint dark brown (7.5YR 4/4) mottles.

textural layers beneath. The Templeton series has a sequence of fine sandy loam. sandy loam. sand and sandy loam layers; the Wakanui series has a silt loam. sandy loam. silty clay loam. loamy sand. sandy loam and loamy sand sequence.

The depth functions of sand. silt and clay contents for the Templeton and Wakanui series (Figures 5.2 and 5.3) clearly illustrate this textural layering. Additionally, they demonstrate that the Templeton series generally has a higher sand content at equivalent depths throughout its profile (57 - 86% compared to 34 - 70%). particularly between about 25 cm and 85 cm. The correspondingly finer texture of the Wakanui profile is especially marked in the silty clay loam 2BCg horizon between 31 and 60 cm where clay and silt contents are both over 30%.

The field-determined soil texture is generally in agreement with the results obtained from the particle size analysis. The Eyre A horizon, however, provides a noticeable exception. Particle-size analysis determined the percentage sand. silt and clay contents as 50.2. 24.2. and 25.6 respectively. The consequent silt loam texture (Taylor and Pohlen. 1979) contrasts with the field-assessed texture of fine sandy loam. This discrepancy presumably reflects the inclusion of very small stones within the horizon; these make the field textural assessment of the < 2 mm fraction more difficult.

The three soils differ considerably in mottling patterns and structure. The Eyre profile contains no mottles; a few fine faint brownish mottles occur in the subsurface (Bw) horizon above the sand layer (2BC) of the Templeton profile; abundant prominent mottles are present in the Wakanui profile. particularly in the fine-textured 2BCg horizon. Moderately developed nutty and granular structures characterise the A horizons of the Eyre and the Templeton profiles. whereas the Wakanui A horizon has a strongly developed granular structure. The Templeton and Wakanui subsoils have weakly to moderately developed blocky or single-grained (if sand texture) structures.

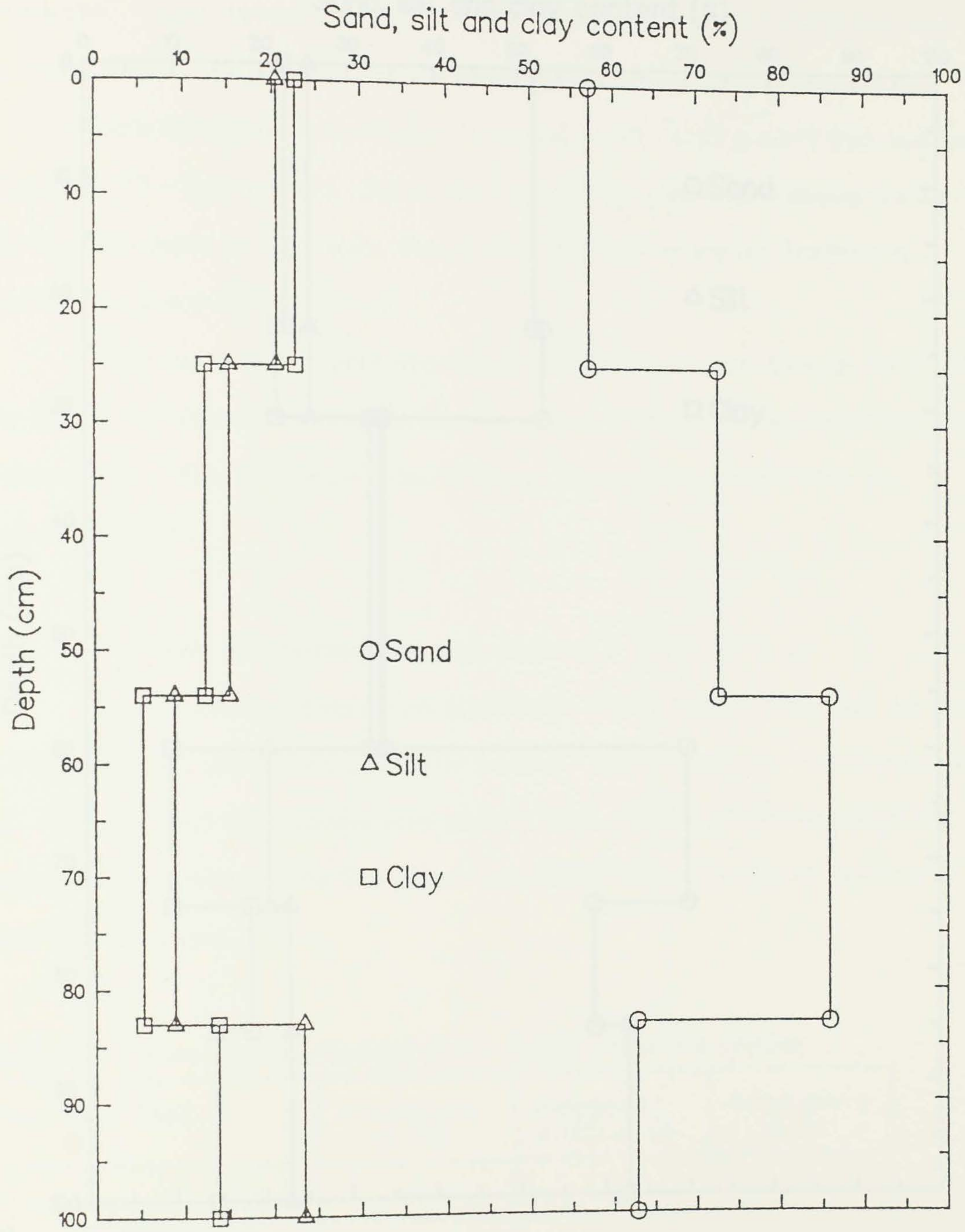


Figure 5.2 Depth function of sand, silt and clay contents for the Templeton series profile

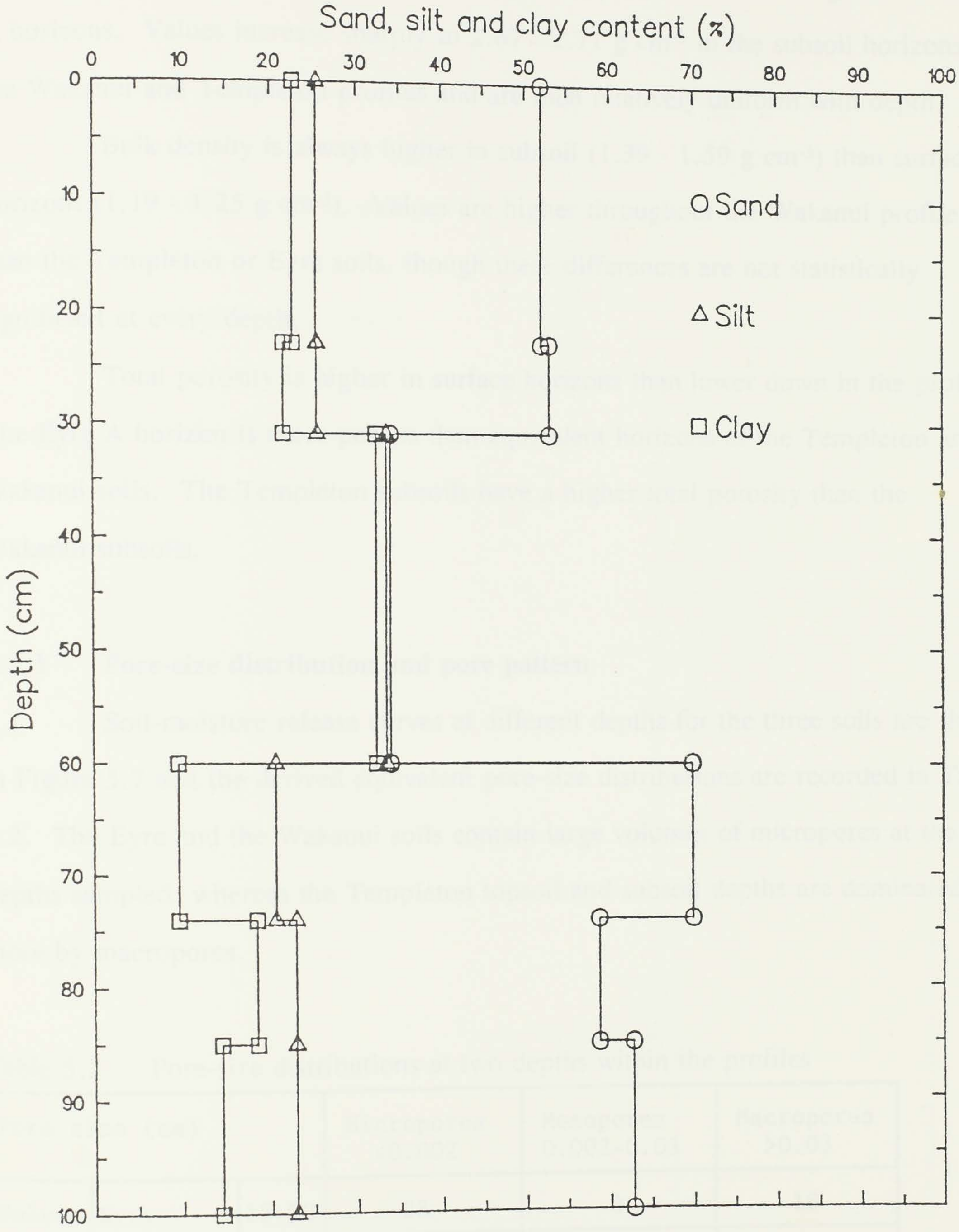


Figure 5.3 Depth function of sand, silt and clay contents for the Wakanui series profile

5.3.2 Particle density, bulk density and total porosity

Depth functions of the mean values of particle density, bulk density and derived total porosity for the three soil profiles are illustrated in Figures 5.4, 5.5 and 5.6 respectively. All three soils have particle densities of 2.56 - 2.57 g cm^{-3} in their A horizons. Values increase sharply to 2.67 - 2.71 g cm^{-3} in the subsoil horizons of the Wakanui and Templeton profiles and are then relatively uniform with depth.

Bulk density is always higher in subsoil (1.39 - 1.59 g cm^{-3}) than surface horizons (1.19 - 1.25 g cm^{-3}). Values are higher throughout the Wakanui profile than the Templeton or Eyre soils, though these differences are not statistically significant at every depth.

Total porosity is higher in surface horizons than lower down in the profiles. The Eyre A horizon is more porous than equivalent horizons of the Templeton and Wakanui soils. The Templeton subsoils have a higher total porosity than the Wakanui subsoils.

5.3.3 Pore-size distribution and pore pattern

Soil-moisture release curves at different depths for the three soils are shown in Figure 5.7 and the derived equivalent pore-size distributions are recorded in Table 5.2. The Eyre and the Wakanui soils contain large volumes of micropores at the depths sampled, whereas the Templeton topsoil and subsoil depths are dominated more by macropores.

Table 5.2 Pore-size distributions at two depths within the profiles

Pore size (mm)			Micropores <0.002	Mesopores 0.002-0.03	Macropores >0.03
Volume of pores (%)	Eyre	15-20	28	8	18
	Templeton	15-20	21	6	25
		55-60	13	12	23
	Wakanui	15-20	28	8	15
		55-60	26	0	18

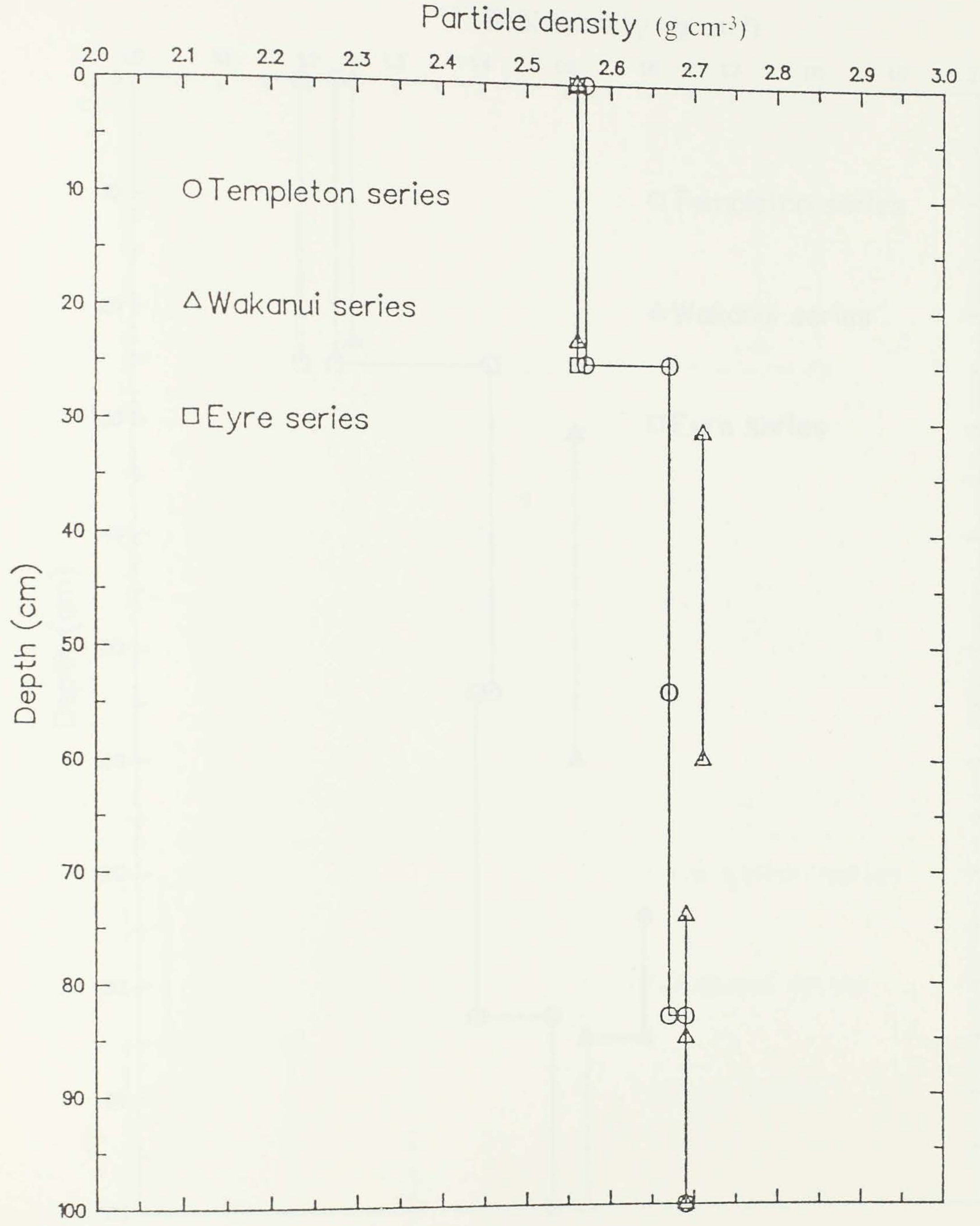


Figure 5.4 Depth function of particle density for the sampled horizons in the three soil profiles

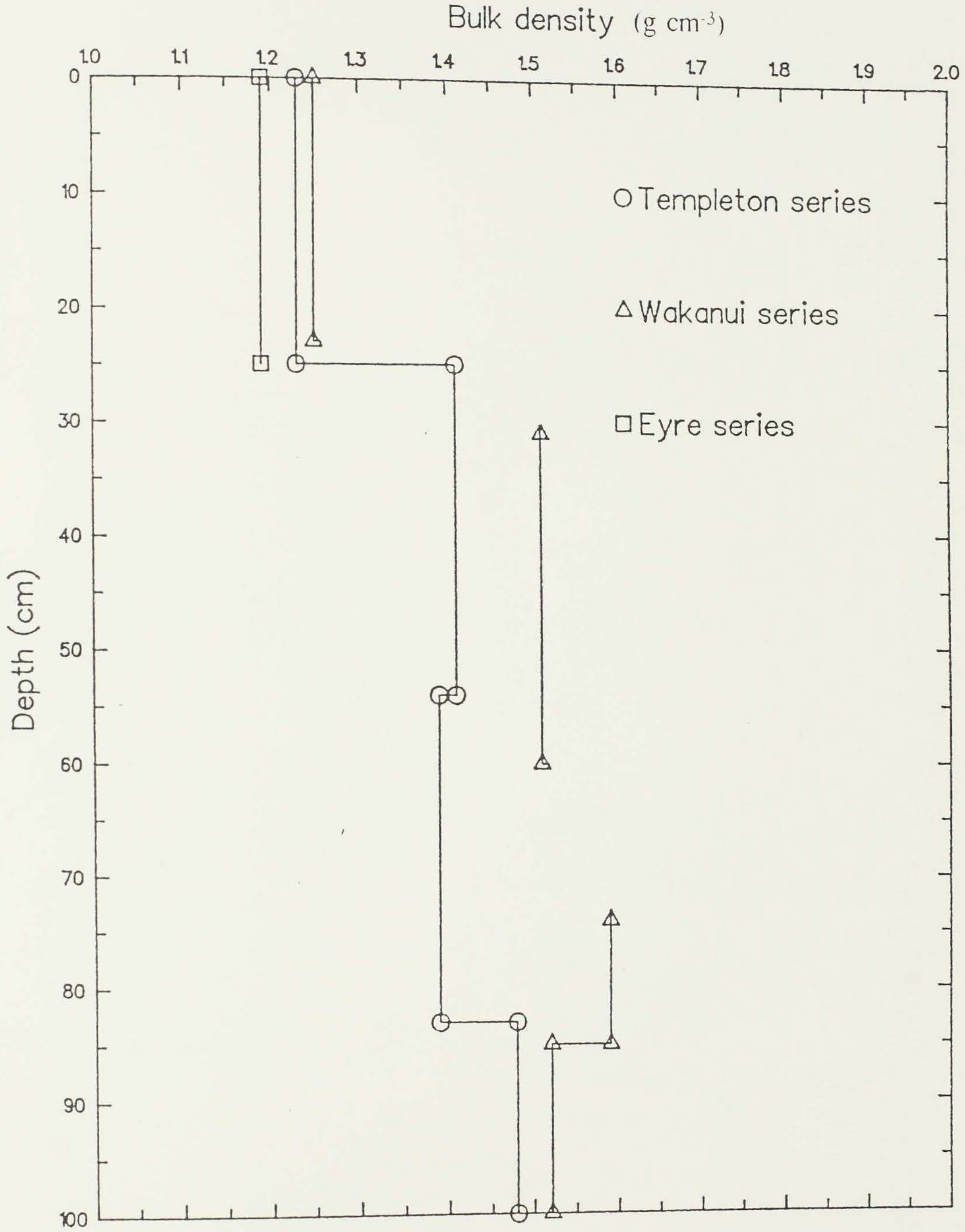


Figure 5.5 Depth function of bulk density for the sampled horizons in the three soil profiles

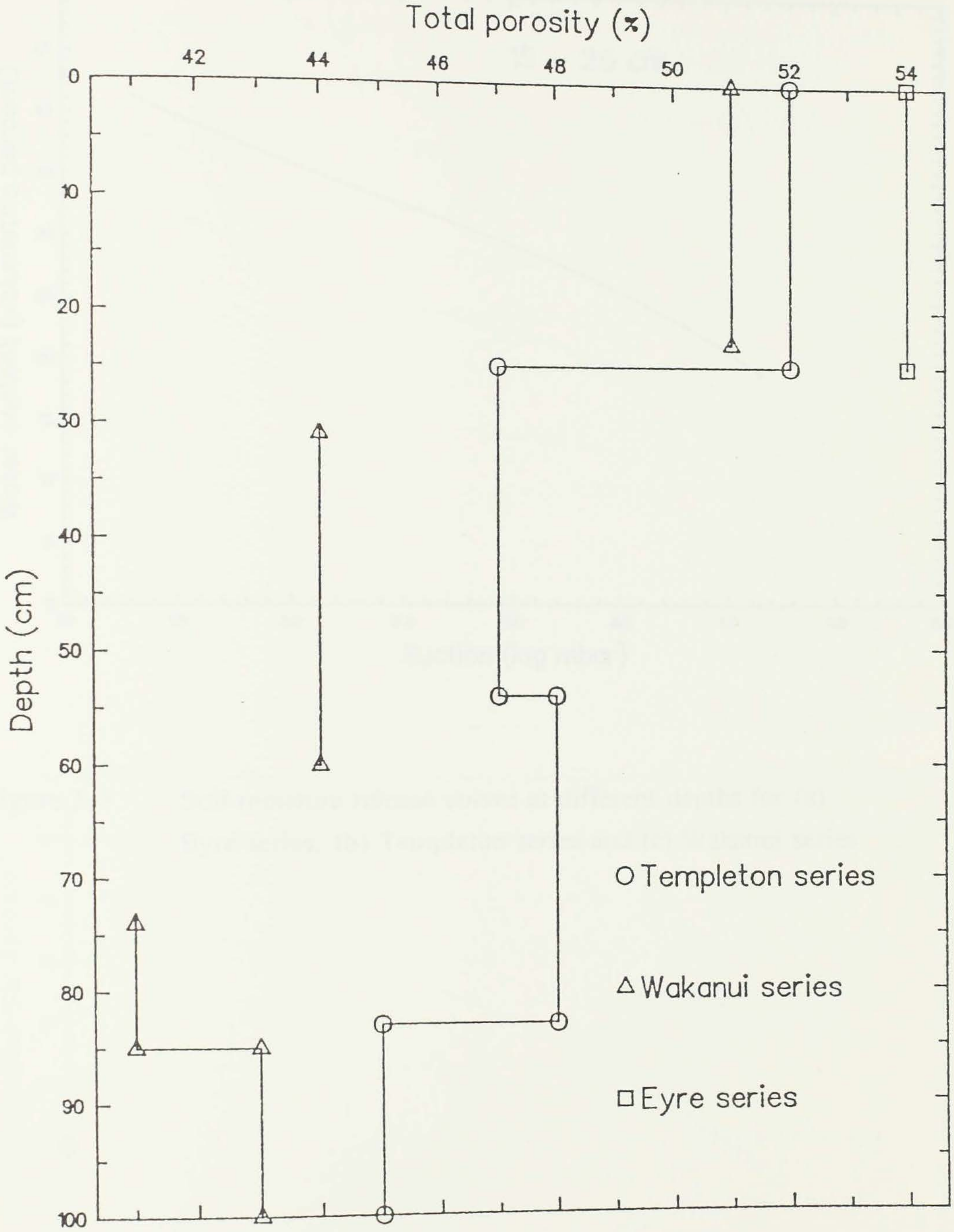


Figure 5.6 Depth function of porosity for the sampled horizons in the three soil profiles

(a)

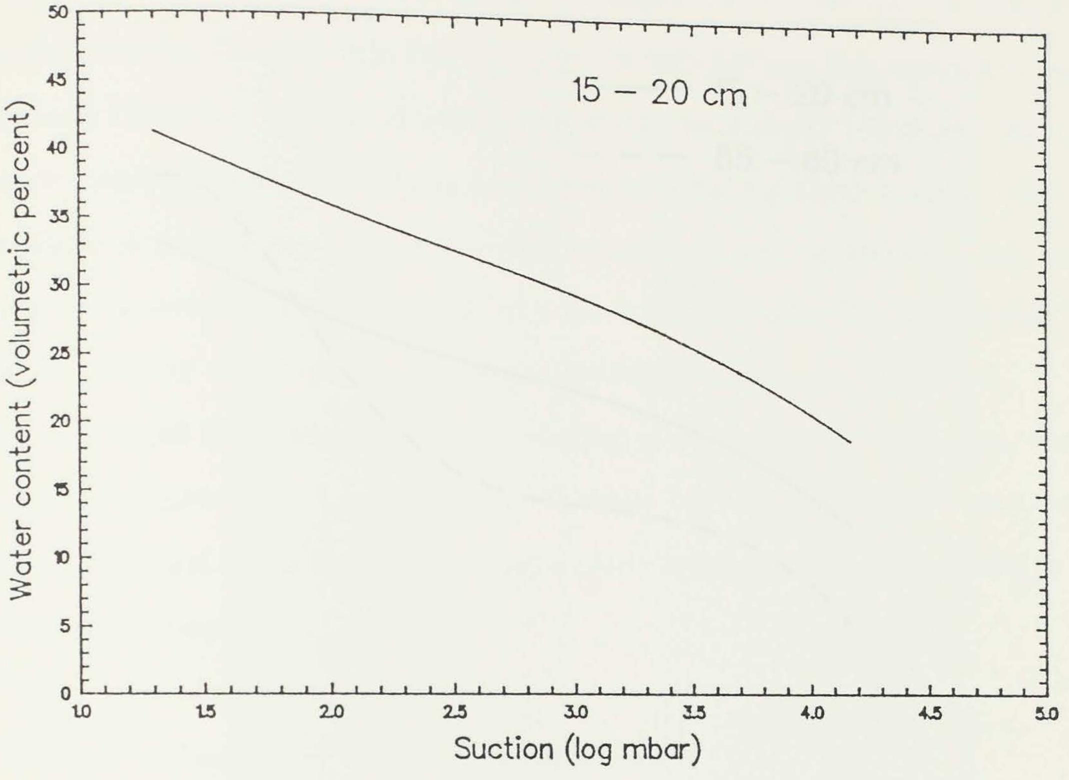
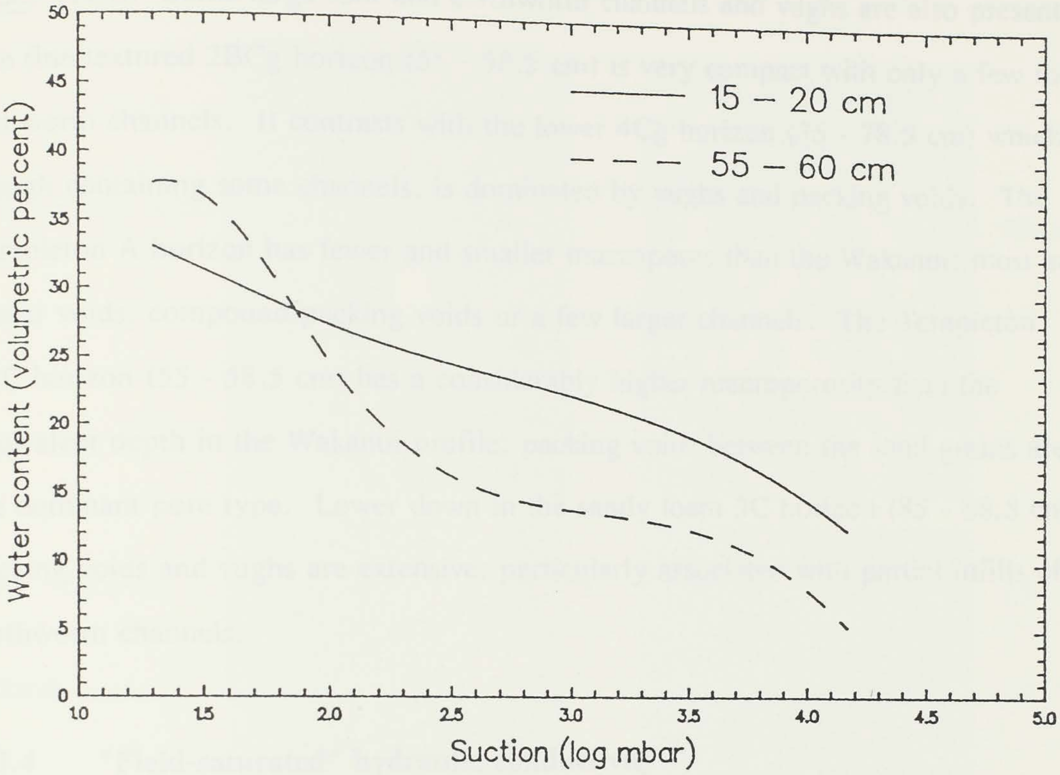
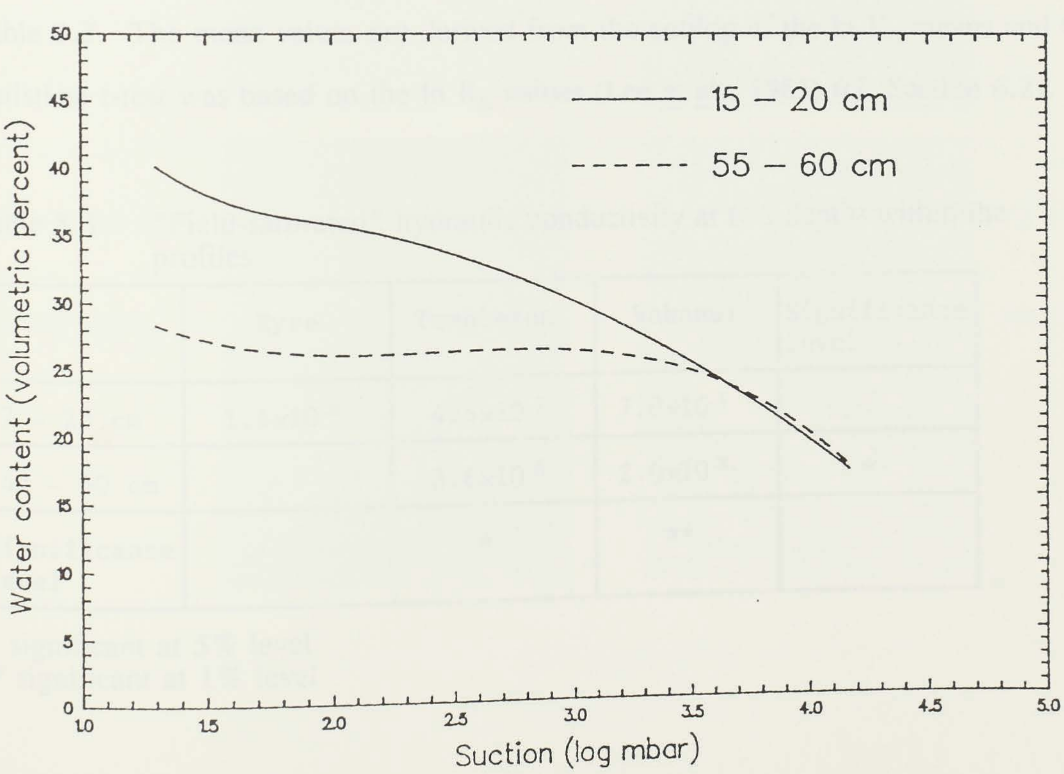


Figure 5.7 Soil-moisture release curves at different depths for (a) Eyre series, (b) Templeton series and (c) Wakanui series

(b)


15 – 20 cm
55 – 60 cm
Water content (volumetric percent)
Suction (log mbar)

(c)


15 – 20 cm
55 – 60 cm
Water content (volumetric percent)
Suction (log mbar)

Changes in macropore patterns down the Wakanui and Templeton profiles are illustrated in Figure 5.8. Granular and nutty aggregates in the Wakanui A horizon (Figure 5.8b) are separated by compound packing and planar voids (Bullock et al., 1985). Some large root and earthworm channels and vughs are also present. The fine-textured 2BCg horizon (55 - 58.5 cm) is very compact with only a few root and worm channels. It contrasts with the lower 4Cg horizon (75 - 78.5 cm) which, though containing some channels, is dominated by vughs and packing voids. The Templeton A horizon has fewer and smaller macropores than the Wakanui; most are planar voids, compound packing voids or a few larger channels. The Templeton 2BC horizon (55 - 58.5 cm) has a considerably higher macroporosity than the equivalent depth in the Wakanui profile: packing voids between the sand grains are the dominant pore type. Lower down in the sandy loam 3C horizon (85 - 88.5 cm) packing voids and vughs are extensive, particularly associated with partial infills of earthworm channels.

5.3.4 "Field-saturated" hydraulic conductivity

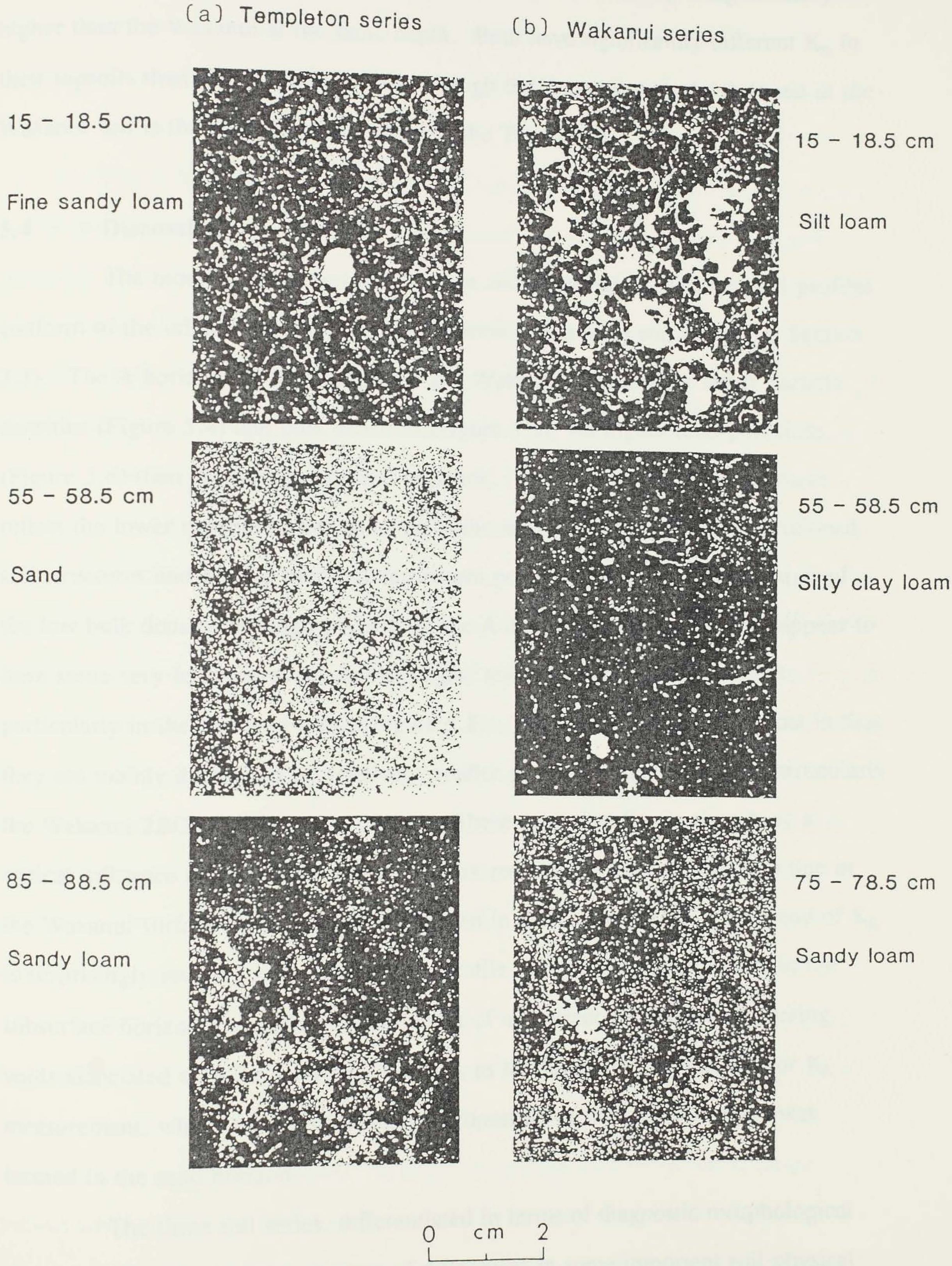
The mean "field-saturated" hydraulic conductivity values are summarised in Table 5.3. The mean values are derived from the antilog of the ln K_{fs} means and the statistical t-test was based on the ln K_{fs} values (Lee et al., 1985) (cf. Section 6.2).

Table 5.3 "Field-saturated" hydraulic conductivity at two depths within the profiles

	Eyre	Templeton	Wakanui	Significance level
7 - 25 cm	1.4×10^{-6}	4.4×10^{-7}	7.0×10^{-7}	
42 - 60 cm		3.6×10^{-6}	2.0×10^{-8}	*
Significance level		*	**	

* significant at 5% level
** significant at 1% level

Figure 5.8 Pore patterns at three depths within the (a) Templeton and (b) Wakanui profiles


(a) Templeton series
(b) Wakanui series
15 - 18.5 cm
Fine sandy loam
55 - 58.5 cm
Sand
85 - 88.5 cm
Sandy loam
15 - 18.5 cm
Silt loam
55 - 58.5 cm
Silty clay loam
75 - 78.5 cm
Sandy loam
0 cm 2

The Eyre A horizon has a higher K_{fs} than the Templeton and Wakanui A horizons, though the difference is not statistically significant at the specified probability levels. The K_{fs} in the Templeton subsoil (42 - 60 cm) is significantly higher than the Wakanui at the same depth. Both have significantly different K_{fs} in their topsoils than at the subsoil depths, though the decrease with depth found in the Wakanui soil is the opposite of the trend in the Templeton.

5.4 Discussion

The morphological features (texture and mottling) of the three soil profiles conform to the criteria used for the classification of the three soil series (cf. Section 3.3). The A horizons of the Templeton and Wakanui profiles have lower particle densities (Figure 5.4) and bulk densities (Figure 5.5), yet higher total porosities (Figure 5.6) than the sampled subsoil horizons. The particle density differences reflect the lower organic matter contents of the subsoil, whilst the better developed soil structures and greater root and earthworm penetrations are the chief causes of the low bulk density and high porosity in the A horizons. The A horizons appear to have some very large pores (channels, vughs, and compound packing voids), particularly in the Wakanui series (Figure 5.8). The subsoil horizons contrast in that they are mainly dominated by relatively smaller packing voids and vughs, particularly the Wakanui 2BCg horizon (31 - 60 cm). These pore-size distributions have a critical influence on water movement, and are responsible for a higher K_{fs} value in the Wakanui surface horizon than lower down in the profile. This depth trend of K_{fs} is surprisingly reversed in the Templeton profile, though the measurement in the subsurface horizon may reflect the influence of a continuous network of packing voids associated with the sand 2BC horizon, as the lower part of the well for K_{fs} measurement, which accounts for a large proportion of the total K_{fs} value, was located in the sand horizon.

The three soil series, differentiated in terms of diagnostic morphological characteristics show varying degrees of differences in some important soil physical properties. The slightly stony Eyre A horizon has the lowest bulk density, yet highest total porosity and hydraulic conductivity among the three soil profiles. These properties are presumably related to the presence of non-fitting gravels. The gravel subsoil of the Eyre profile, though not examined, would also be expected to yield very high hydraulic conductivities. The whole profile therefore is freely-drained as indicated by the absence of mottles.

The A horizon K_{fs} is higher in the Wakanui than Templeton soil, a feature that is attributed to the difference in the type and degree of development in soil structure: The Wakanui A horizon has a more strongly developed granular structure due to its finer texture (silt loam compared to fine sandy loam), and more larger pores (cf. Table 5.1 and Figure 5.8). Soils with water stable granular structures conduct water much more rapidly than those with less strongly developed structures.

The Templeton subsoil has high sand contents (Figure 5.2) and a consequent large proportion of packing meso- and macrovoids, particularly in the 2BC horizon (54 - 83 cm) (Table 5.2). This in turn causes a high subsoil saturated hydraulic conductivity (Table 5.3). The overall soil profile is thus also well-drained and mottles are not extensively developed. The fine faint brown mottles occurring in the Bw horizon (Table 5.1) are caused by the impedance to unsaturated water flow of the underlying sand layer (2BC). This layer has a lower matric suction than the overlying finer-textured horizon (sandy loam). Surface-added water does not drain through immediately and tends to be held up in the overlying finer-textured horizon until the moisture content is high enough, and the water suction low enough, for water to be released into the underlying horizon.

The Wakanui soil is characterised by strong mottling patterns throughout the profile, features which suggest that the soil is imperfectly-drained. Such mottling patterns, particularly in the 2BCg horizon, are associated with the fine textures, high bulk densities, and low porosities. A large proportion of pores in the sampled subsoil are very fine (< 0.002 mm), and thus make little contribution to saturated water flow (Table 5.2 and Figure 5.8). The hydraulic conductivities are therefore very low in the subsoil compared to that at the equivalent depth of the Templeton profile.

The overall water movement down soil profiles is controlled by the horizon that has the lowest saturated hydraulic conductivity. Both Eyre and Templeton soils possess an intermediate hydraulic conductivity (Reynolds and Elrick, 1986) and are therefore freely-drained. The Wakanui subsoil (31 - 60 cm) has a very low hydraulic conductivity; surface-added water cannot always drain freely down through the profile. Water therefore tends to be periodically held up by the low conductivity layer, thus allowing gleying processes to take place with the consequent development of mottling patterns. These drainage differences will presumably lead to corresponding differences in moisture content between the Wakanui and other soils at certain times of the year. This relationship is further considered in Chapter 6.

5.5 Summary and conclusions

The three examined soil profiles are representative of the Eyre, Templeton and Wakanui soil series in terms of the diagnostic morphological criteria used for the classification of soils in the region (Cox, 1978). The examined accessory soil physical properties (related to water storage and movement) do differ between the three soils, though to varying extents and levels of significance. Bulk density increases and porosity decreases from Eyre to Templeton to Wakanui soils due to the differences in soil texture. "Field-saturated" hydraulic conductivity is relatively similar for the three A horizons, though the Eyre soil values are slightly larger presumably because of the presence of some very large pores produced by the non-fitting gravels occurring within and beneath the A horizons. K_{fs} is significantly higher in the Templeton subsoil (42 - 60 cm) than at the equivalent depth in the Wakanui soil. This difference reflects the high macroporosity of the Templeton subsoil and corresponding high microporosity of the Wakanui subsoil.

The Eyre and Templeton soils are freely-drained due to their intermediate hydraulic conductivities, whilst the Wakanui soil is probably close to saturation for certain periods of the year due to the very low hydraulic conductivity of parts of the subsoils. The Wakanui soil should also be capable of storing more moisture than the other two soils during drier phases. The examined soil physical properties seem to be most closely related to textural changes among the three soil profiles. Mottling patterns are a secondary feature governed by effect soil texture has on water movement.

CHAPTER 6

VARIABILITY OF SOIL PHYSICAL PROPERTIES WITHIN AND BETWEEN TAXONOMIC UNITS

6.1 Introduction

The widespread effectiveness of a morphologically-based soil classification was partially assessed in Chapter 5 by characterizing and comparing modal soil profiles of different taxonomic units in terms of various physical properties. This traditional approach, although providing some comparative data which may be used to substantiate the taxonomic differentiation, is limited due to the restricted number of samples used and the overall area (or volume) of soil considered. A greater number of replicate samples taken from identical depths would enable a more thorough statistical differentiation, whilst it would be particularly useful from a practical viewpoint to establish whether an area of morphologically-uniform (i.e. taxonomically-pure) soil is similarly uniform as regards physical properties. One of the main aims of this chapter therefore is to assess and compare the variability of selected physical properties within areas of different mapping units which are morphologically-pure, and thus equivalent to taxonomic units. The data will also be analysed to more stringently test the assumption that there are significant differences in physical properties between the three taxonomic units. These analyses will provide a quantitative test of the efficiency of the morphologically-based classification system to account for spatial variability of physical properties. The significance test of difference, however, is based on specified probability levels; differences shown not significant at such levels in the text may be significant at slightly higher levels.

The three soil physical properties chosen to characterise the soil series taxonomic units in this part of the study were selected for practical and logistical reasons. "Field-saturated" hydraulic conductivity (K_{fs}) can be measured relatively rapidly and provides an indication of how fast water moves through horizons and whole profiles. Moisture content varies considerably throughout the year, yet measurements made at a particular time of the year still provide a general indication of soil-water characteristics. Bulk density can be measured at the same time as moisture content and is closely related to both these parameters and other properties such as texture and porosity.

The intended statistical comparison of results necessitated that the samples or measurements should be taken at the same depths for all three soils, though the presence of gravels again precluded measurements from the lower depths of the Eyre soil. The two sampled depths are referred to in the text as "topsoil depth" and "subsoil depth". The topsoil depth extended to within the A horizons; the subsoil depth corresponded to key horizons which play governing roles in determining the overall water storage and movement within the Wakanui and Templeton soils. The areas sampled for each taxonomic unit was 30 m x 30 m. This size was practically determined by balancing time/effort considerations with the necessity for studying an area of realistic size such as would be useful for experimental trials.

6.2 Methods

Bulk density and K_{fs} were measured using the same procedures as described in Chapter 5 (cf. Section 5.2). The samples for bulk-density measurement were all taken on the same day (August 15, 1987), and also used for the measurement of moisture content. The latter was obtained from the difference in weight between moist- and oven-dried samples, expressed as a volumetric percentage.

Bulk density (and moisture-content) samples were taken at depths of 20 - 25 cm and 55 - 60 cm, whilst "field-saturated" hydraulic conductivity was measured in situ at 7 - 25 cm and 42 - 60 cm. Only samples or measurements from the shallowest depth was obtainable for the Eyre series due to the presence of gravels below 25 cm.

Samples and measurements were taken from three "window" areas which appeared, on the basis of the spatial analysis of soil morphological properties (cf. Chapter 4), to be taxonomically pure and representative of the three taxonomic units. This purity was confirmed by a detailed auger survey conducted on the same grid as for the sampling and measurements.

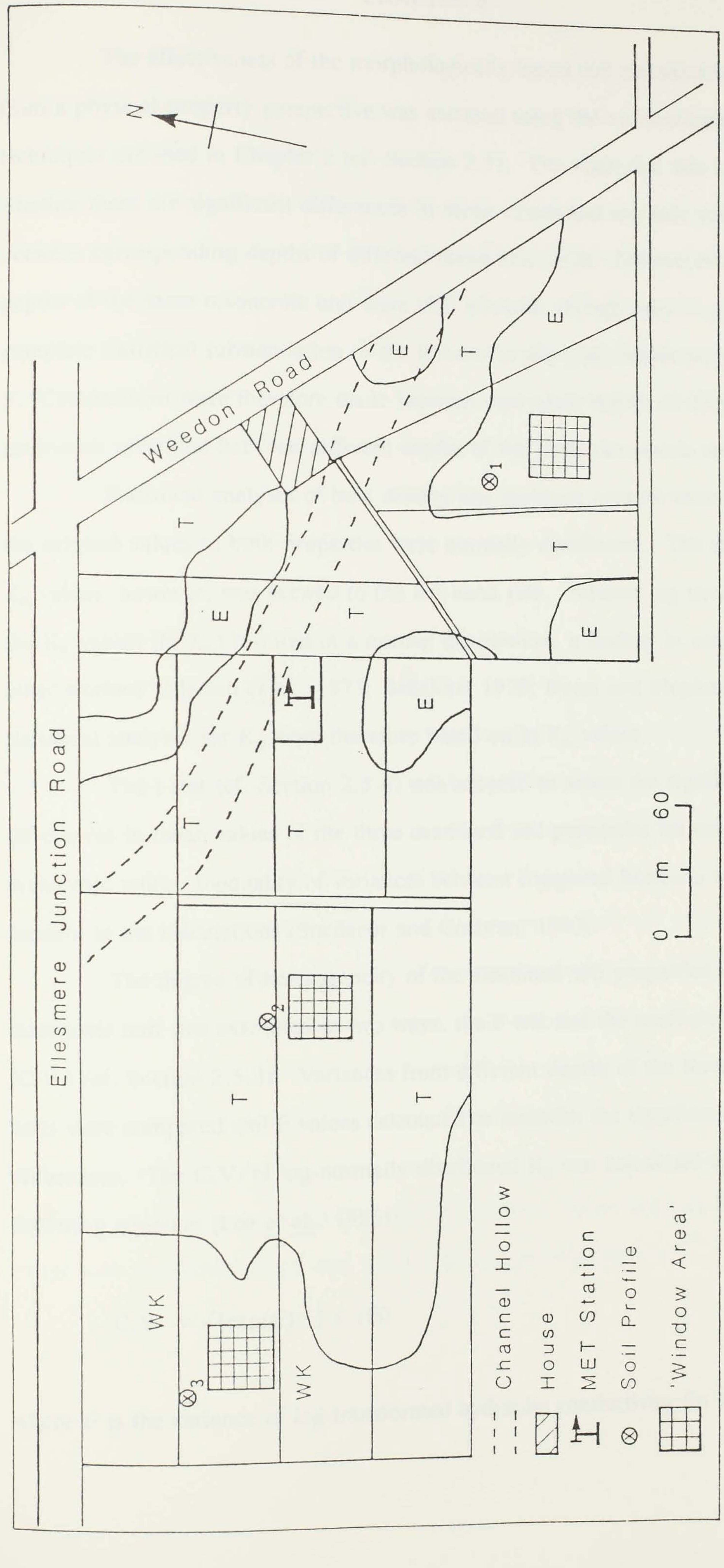
An attempt was made to devise an optimal sampling strategy to provide the mean values of each taxonomic unit by utilising the geostatistical techniques discussed in Section 4.5. A preliminary study carried out on a series of transects with staggered sampling intervals (1 m for K_{fs} and 2 m for bulk density and moisture content) within each taxonomic unit indicated that there was no structural dependence revealed in most of the situations. The conventional approach (Section 2.7.2) was therefore employed to determine the number of samples necessary to characterise the mean values of the examined physical properties with specified maximum tolerable errors (0.03 g cm^{-3} for bulk density, 0.5 units for ln K_{fs} and 1% for moisture content at the 95% confidence level). This approach led to the establishment of different sample sizes for each property at each depth. For comparative purposes, however, the largest number of samples or measurements necessary for any property or depth was assumed to be appropriate as a sample size for all properties from each taxonomic unit. The sample size was 36 and the observations were evenly distributed within each "window" area (30 m x 30 m) on a square grid basis with sampling spacing therefore being 6 m apart. The locations of each sampled area are indicated in Figure 6.1.

Three-dimensional diagrams produced by the C3D programme (Baird, 1986) were used to illustrate the spatial changes in the properties at both depths. Those diagrams concerning the Wakanui soil are included in this thesis as examples of the spatial variability.

Figure 6.1 Soil map of the study area and locations of the three taxonomically-pure "window areas" sampled for soil physical property assessment

Key to the soil map

E: Eyre series
T: Templeton series
WK: Wakanui series


N
Ellesmere Junction Road
Weedon Road
T
E
WK
⊗1
⊗2
⊗3
0 m 60
Channel Hollow
House
MET Station
Soil Profile
'Window Area'

The effectiveness of the morphologically-based soil classification system from a physical-property perspective was assessed using the conventional statistical techniques outlined in Chapter 2 (cf. Section 2.5). The main aim was to establish whether there are significant differences in mean values and intrinsic variability between corresponding depths of different taxonomic units. Differences between depths of the same taxonomic unit were also assessed, though only to provide a more complete statistical substantiation of the inferences and conclusions made in Chapter 5. Comparisons were therefore made between equivalent depths of different taxonomic units and between different depths of the same taxonomic unit.

Statistical analyses of bulk density and moisture content were performed on the original values as both properties were normally distributed. The distribution of K_{fs} values, however, was skewed to the left-hand side. Natural log transformation of the K_{fs} values ($\ln K_{fs}$) resulted in a normal distribution, a finding in accordance with other workers (Nielsen et al., 1973; Babalola, 1978; Byers and Stephens, 1983). All statistical analyses for K_{fs} were therefore based on $\ln K_{fs}$ values.

The t-test (cf. Section 2.5.4) was adopted to assess the significance of differences in mean values of the three examined soil properties between the taxonomic units. Inequality of variances between compared horizons was taken into account in the calculations (Snedecor and Cochran, 1980).

The degree of heterogeneity of the examined soil properties within each taxonomic unit was examined in two ways, the F-test and the coefficient of variation (C.V.) (cf. Section 2.5.3). Variances from different depths of the three taxonomic units were compared and F values calculated to establish the significance of any differences. The C.V. of log-normally distributed K_{fs} was calculated using the following equation (Lee et al., 1985):

$$C.V. = \sqrt{[\exp(s^2)-1]} \times 100 \qquad (6.1)$$

where s^2 is the variance of log-transformed hydraulic conductivity ($\ln K_{fs}$).

Analysis of variance (cf. Section 2.5.4) was used to partition the components of variance from different sources. The intraclass correlation coefficient (r_i) was calculated on the basis of the products derived from this analysis. The F-test indicated that the variances between some of the horizons are significantly different (Tables 6.2 and 6.5), a fact that should theoretically violate one of the assumptions behind the analysis of variance. This is not a serious problem in this study, however, as the F-test for two variances is "sensitive both to the real differences of variance and to departures from normality. Analysis of variance is much more robust, and even when significant differences are found among variances, the investigator is often quite justified in proceeding with analysis of variance on the assumption that the variances are equal" (Webster, 1977).

6.3 Results and discussion

6.3.1 Spatial distribution of data within windows

The raw grid data for each soil property in the three taxonomic units are summarised in Appendix 3. Spatial changes in properties at both depths in the Wakanui taxonomic unit are illustrated by the block diagrams in Figures 6.2 and 6.3. Similar patterns occur within the Templeton and Eyre taxonomic units, though the appropriate diagrams are not included in this chapter. Each soil physical property varies from place to place within the taxonomic unit that is morphologically uniform in terms of the soil series criteria. Figures 6.2b and 6.3b demonstrate that a majority of the K_{fs} values have a relatively small range: they contrast markedly with occasional large values apparent at a few locations. This demonstrates the left-hand skewed nature of the K_{fs} distribution pattern as discussed above. The other properties display a more normal distribution pattern with most values being in the medium range with occasional larger and smaller values occurring in places.

Figure 6.2 Spatial variation of physical properties in the Wakanui topsoil for (a) bulk density (g cm^{-3}). (b) hydraulic conductivity ($\times 10^{-6}$ m s^{-1}) and (c) moisture content (Vol. %)

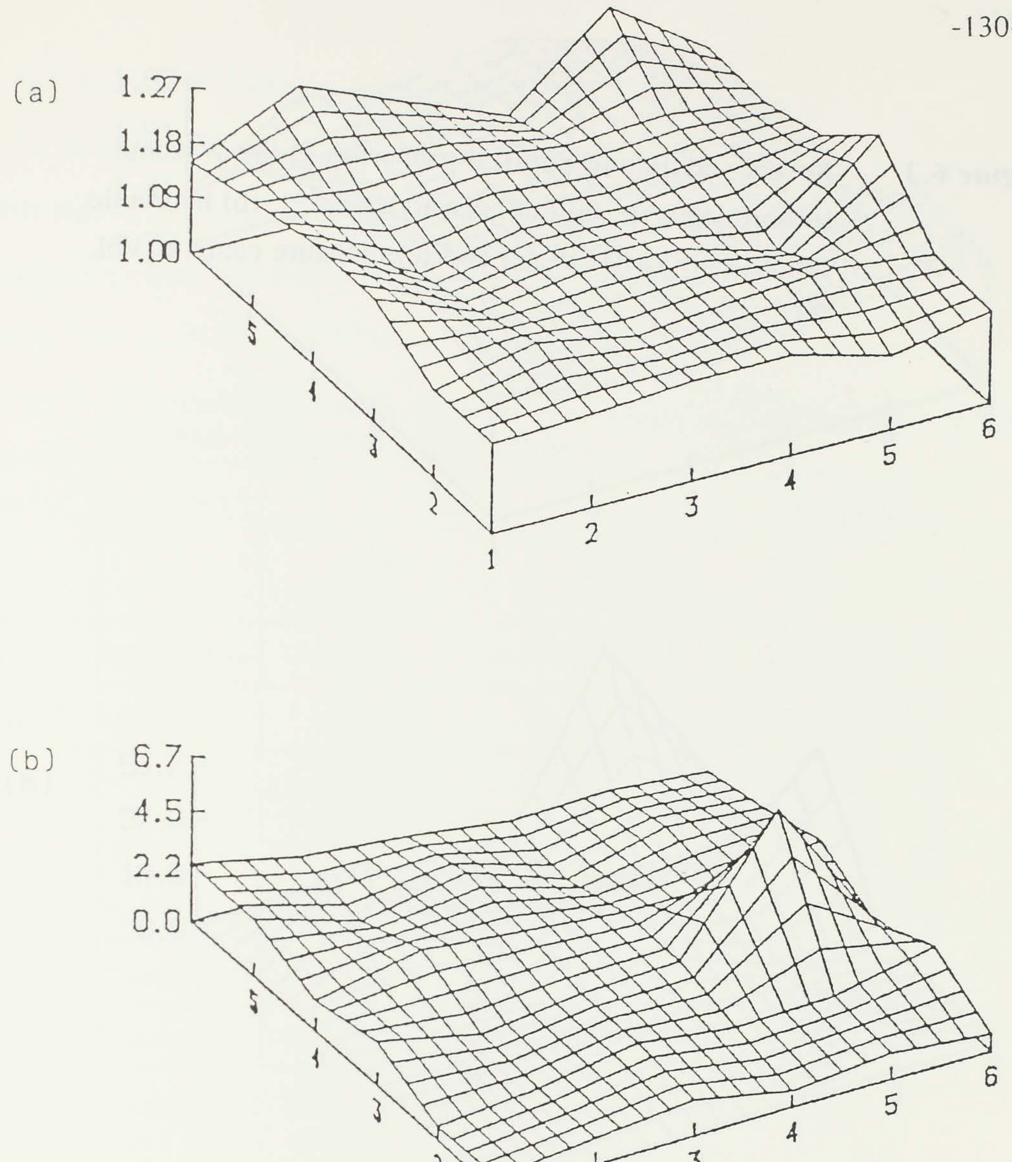
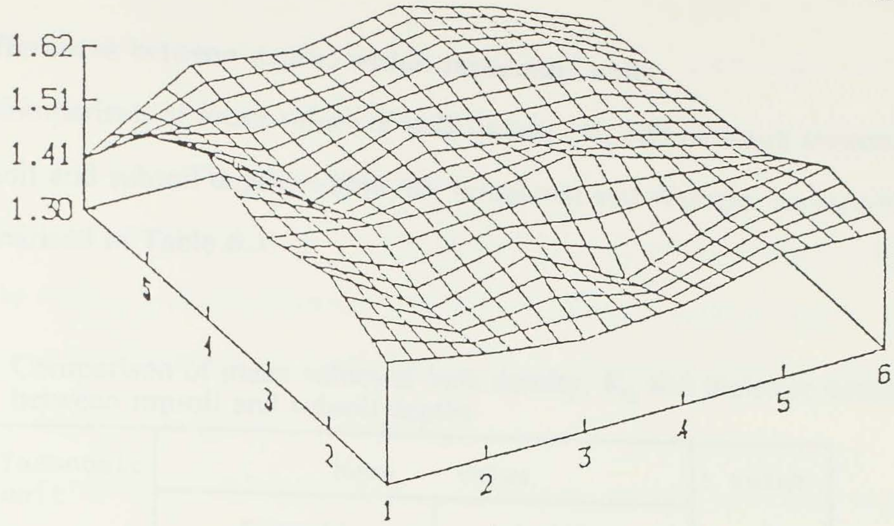
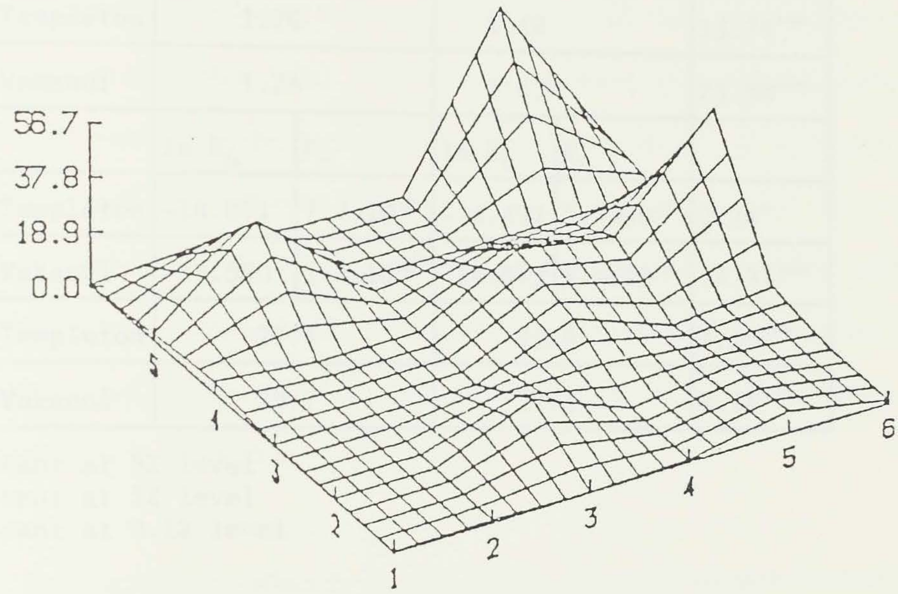

(a)
1.27
1.18
1.09
1.00
1
2
3
4
5
6
(b)
6.7
4.5
2.2
0.0
(c)
30.5
29.0
27.5
26.0

Figure 6.3 Spatial variation of physical properties in the Wakanui subsoil depth for (a) bulk density (g cm^{-3}). (b) hydraulic conductivity ($\times 10^{-8}$ m s^{-1}) and (c) moisture content (Vol. %)

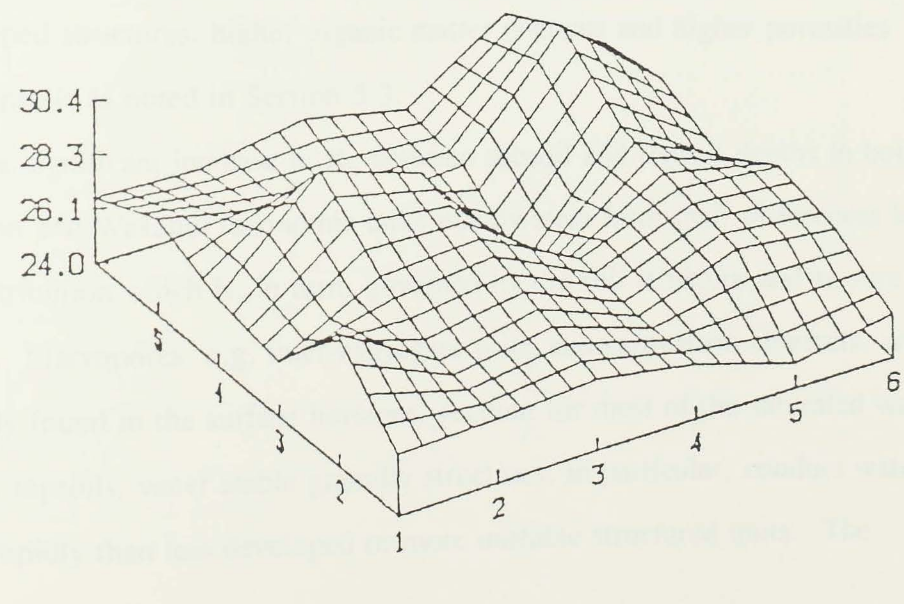
(a)


1.62
1.51
1.41
1.30
5
4
3
2
1
2
3
4
5
6

(b)


56.7
37.8
18.9
0.0
5
4
3
2
1
2
3
4
5
6

(c)


30.4
28.3
26.1
24.0
5
4
3
2
1
2
3
4
5
6

6.3.2 Differences between depths within taxonomic units

A comparison of mean values of bulk density, K_{fs} and moisture content between topsoil and subsoil depths within the Templeton and Wakanui taxonomic units is summarised in Table 6.1.

Table 6.1 Comparison of mean values of bulk density, K_{fs} and moisture content between topsoil and subsoil depths

Property	Taxonomic unit	Mean value				t value
		Topsoil depth		Subsoil depth		
Bulk density ($g\ cm^{-3}$)	Templeton	1.26		1.49		13.71***
	Wakanui	1.26		1.61		23.50***
K_{fs} ($m\ s^{-1}$)		ln K_{fs}	K_{fs}	ln K_{fs}	K_{fs}	
	Templeton	-14.071	7.7×10^{-7}	-14.492	5.1×10^{-7}	2.24*
	Wakanui	-13.580	1.3×10^{-6}	-18.073	1.4×10^{-8}	14.31***
Moisture content (Vol. %)	Templeton	27.4		23.9		7.06***
	Wakanui	29.7		27.2		8.64***

* significant at 5% level
** significant at 1% level
*** significant at 0.1% level

Bulk density increases significantly between topsoil and subsoil depths within both the Templeton and Wakanui taxonomic units. This trend reflects the better-developed structures, higher organic matter contents and higher porosities within the topsoils as noted in Section 5.3.

The significant increase in K_{fs} between subsoil and topsoil depths in both the Templeton and Wakanui taxonomic units mainly stem from their differences in pore-size distribution which is, in turn, governed by the soil structure and texture (cf. Section 5.3). Macropores, e.g. inter-aggregate voids and earthworm channels, which are commonly found in the surface horizons, account for most of the saturated water movement in topsoils; water stable granular structures, in particular, conduct water much more rapidly than less developed or more unstable structured units. The subsoils of both Templeton and Wakanui taxonomic units, have fewer connected large intra-aggregate pores and water-stable aggregates, and thus have significantly lower hydraulic conductivities than their topsoils.

The depth trend in K_{fs} for the Templeton series is opposite to that recorded for the profile study. This emphasises the limitations of using a single profile to characterise a soil taxonomic unit, particularly in terms of certain physical properties which tend to display a large spatial variation.

The significantly higher moisture content of the topsoils compared to the subsoils of the Templeton and Wakanui taxonomic units is partly a function of higher organic matter contents, and larger volumes of fine (< 0.03 mm) pores which store water in the topsoils (cf. Table 5.2). Such variation patterns also probably reflect recent addition of water from the surface. The vertical distribution of moisture content, however, will vary substantially from season to season, and with the length of time since rainfall or irrigation. The information provided here therefore only indicates the soil-water-storage characteristics at the particular time of sampling.

Results from the equality test of variances between topsoil and subsoil depths are presented in Table 6.2.

Table 6.2 Comparison of variances in bulk density, K_{fs} and moisture content between topsoil and subsoil depths

Property	Taxonomic unit	Variance		F Value
		Topsoil depth	Subsoil depth	
Bulk density	Templeton	0.0022	0.0073	3.25**
	Wakanui	0.0032	0.0047	1.45
ln K_{fs}	Templeton	0.43	0.83	1.95
	Wakanui	0.39	3.15	8.05**
Moisture content	Templeton	2.47	6.62	2.68**
	Wakanui	1.08	1.92	1.77

* significant at 5% level
** significant at 1% level

Variance is always greater in subsoil than topsoil depths, though the differences are only significant for bulk density and moisture content in the Templeton, and for K_{fs} in the Wakanui series. The relatively low variability of the topsoil properties probably reflect the smoothing effect of biotic mixing and uniform management over the area. It is also attributable to the fact that the thickness of topsoils is relatively constant and measurements are liable to be taken entirely within the texturally-uniform A horizon. The subsoils, however, remain less modified and the variable attributes inherited from the alluvial parent materials still exist. It was observed from the auger survey, for instance, that the sand particles in the Templeton subsoil are coarser in places than others, and that the degree of compaction varies considerably. The thickness of textural layers in subsoils also changes markedly over short distances in the region: measurements in the subsoil may have been taken within one textural layer at some locations, whereas at others effects of overlying or underlying textural layers may have been incorporated within the measurements and therefore cause markedly different values from those derived from within one textural layer. This is especially true for K_{fs} because the measurements were taken from a relatively large depth range (18 cm).

In summary, there are significant differences in the three physical properties between topsoil and subsoil depths within the Templeton and Wakanui taxonomic units. The topsoils tend to have lower bulk densities, yet higher hydraulic conductivities and moisture contents (Table 6.1). Subsoils, however, are generally more variable than topsoils in terms of the three properties.

6.3.3 Differences in means between taxonomic units

A comparison of property mean values between the three taxonomic units is shown in Table 6.3.

Table 6.3 Comparison of mean values of bulk density, K_{fs} and moisture content between the three taxonomic units

Depth	Taxonomic unit	Bulk density ($g\ cm^{-3}$)		K_{fs} ($m\ s^{-1}$)			Moisture content(Vol%)	
		Mean	t	lnK_{fs}	K_{fs}	t	mean	t
Topsoil depth	Eyre Templeton	1.18 1.26	6.75^{***}	-13.60 -14.07	1.2×10^{-6} 7.7×10^{-7}	2.33^{*}	27.6 27.4	0.43
	Eyre Wakanui	1.18 1.26	6.28^{***}	-13.60 -13.58	1.2×10^{-6} 1.3×10^{-6}	0.08	27.6 29.7	4.31^{***}
	Templeton Wakanui	1.26 1.26	0.05	-14.07 -13.58	7.7×10^{-7} 1.3×10^{-6}	3.25^{**}	27.4 29.7	7.11^{***}
Subsoil depth	Templeton Wakanui	1.49 1.61	6.90^{***}	-14.49 -18.07	5.1×10^{-7} 1.4×10^{-8}	10.8^{***}	23.9 27.2	6.76^{***}

* significant at 5% level
** significant at 1% level
*** significant at 0.1% level

Bulk density in the Eyre topsoil is significantly lower than at the same depth in the Templeton and Wakanui soils, whereas similar values are found in the topsoils of both the Templeton and the Wakanui taxonomic units. The low bulk density of the Eyre topsoil is probably due to the existence of large packing voids associated with included gravels. The considerably higher bulk density of the compact Wakanui subsoil than the Templeton subsoil is obviously a function of the former's finer texture, and is consistent with the difference noted between the profiles in Chapter 5.

Both Eyre and Wakanui topsoils conduct water more rapidly than the Templeton topsoil. The high K_{fs} in the Eyre topsoil is probably associated with the presence of the large packing pores between the gravels; the underlying gravels with presumably high hydraulic conductivity may also partially contribute to the high K_{fs} in the A horizon. The strongly developed granular structure is mainly responsible for the high conductivity in the Wakanui A horizon. The Templeton A horizon was earlier noted to have a less well-developed structure because of its coarser texture, and have fewer large pores than the Wakanui topsoil. The hydraulic conductivity in the Templeton subsoil, however, is significantly higher than at the equivalent depth in the Wakanui subsoil, a contrast which is clearly a function of their differences in texture, bulk density (total porosity), macroporosity and possibly pore continuity.

Moisture content, one of the most important parameters to plant growth, is significantly higher at both depths in the Wakanui series than in the Eyre and Templeton series. These differences are related to their differences in K_{fs} and pore-size distribution. The fine-textured, tightly-compacted Wakanui subsoil horizon has a very low hydraulic conductivity; water added from the surface by rain or irrigation cannot penetrate through this layer immediately. Rapid drainage is impeded and water stored above and within the fine-textured horizon with the result that both topsoils and subsoils contain more moisture than at equivalent depths in the other two soil series. Waterlogged conditions may even occur at certain periods of the year. Lateral movement of water may accentuate these moisture content differences between taxonomic units. The differences in soil moisture contents appear to correlate closely with the texture and mottling parameters. Evidence of this relationship is important in view of the frequent reliance on these easily-examined morphological features to predict soil hydraulic behaviour.

In summary, it appears that the classification of soil according to morphological properties of texture and mottling provides a useful means of separating soils of distinctive physical properties in this region. Although some topsoil properties do not differ significantly between pairs of taxonomic units, all subsoil characteristics are clearly distinguishable between Templeton and Wakanui series.

6.3.4 Differences in variability between taxonomic units

Although the classification system has separated out soils which differ in certain physical properties, there is no guarantee that these properties will display the same level of uniformity within each taxonomic unit. An indication of the variability of properties within units can be obtained from coefficient of variation (C.V.) values and the F-test.

The C.V. values (%) for the three properties in topsoil and subsoil depths of each taxonomic unit are summarised in Table 6.4.

Table 6.4 C.V. values for bulk density, K_{fs} and moisture content

Property	Depth	C.V. value (%)		
		Eyre	Templeton	Wakanui
Bulk density	20-25	5.1	3.8	4.5
	55-60		5.7	4.2
K_{fs}	7-25	137.3	73.3	69.1
	42-60		113.7	472.6
Moisture content	20-25	9.4	5.7	3.5
	55-60		10.8	5.1

The Eyre topsoil has higher C.V. values for all properties than the other topsoils. The Templeton subsoil is more variable than the Wakanui subsoil in bulk density and moisture content, but the reverse holds for K_{fs}. The C.V. values for the three soil properties are in close agreement with other studies reported in literature (Warrick and Nielsen, 1980; Wilding and Drees, 1983; Lee et al., 1985).

The C.V. values give no indication whether the differences in variability are statistically significant. An F-test on the variances is required to achieve this purpose (Table 6.5).

Table 6.5 Comparison of variances of bulk density, K_{fs} and moisture content between the three taxonomic units

Depth	Taxonomic unit	Variance and F value					
		Bulk density		ln K_{fs}		Moisture content	
		s^2	F	s^2	F	s^2	F
Topsoil depth	Eyre Templeton	0.0036 0.0022	1.59	1.06 0.43	2.47*	6.77 2.47	2.74**
	Eyre Wakanui	0.0036 0.0032	1.10	1.06 0.39	2.71**	6.77 1.08	6.25**
	Templeton Wakanui	0.0022 0.0032	1.44	0.43 0.39	1.10	2.47 1.08	2.28*
Subsoil depth	Templeton Wakanui	0.0073 0.0047	1.56	0.83 3.15	3.77**	6.62 1.92	3.46**

* significant at 5% level
** significant at 1% level

The variances in bulk density at equivalent depths between any two of the three taxonomic units are never significantly different, i.e. the three soils are similarly uniform (or variable).

Moisture content and K_{fs} are significantly more variable in the Eyre topsoil than either Templeton or Wakanui topsoils. This heterogeneity is probably due to the spatially-variable gravel content of the Eyre A horizon, at least with regards sampling volumes. It is also attributable to the fact that the sampled window area includes different soil types of the Eyre series (Figure 4.16), whereas the Templeton and Wakanui areas only encompass a single soil type of the appropriate soil series.

K_{fs} is more variable, and moisture content less variable, in the Wakanui subsoil than the Templeton subsoil. The high variability of K_{fs} in the Wakanui subsoil is probably due to the existence of occasional cracks, earthworm channels and the variable depth of the coarse-textured layer (loamy sand) beneath the fine-textured horizon (sharp contrast in texture). The textural layering has a strong effect on K_{fs} as the measurement was made within a relatively large depth range (18 cm); the effect is not so marked with bulk density and moisture content as the samples were small (5 cm length) and are more likely to have been taken from within one textural layer.

The Templeton subsoil is more variable in moisture content than the Wakanui subsoil, presumably because of its heterogeneous pore-size distribution. This difference, however, may alter if samples were to be taken at other times of the year, when different sizes of pores are filled with water.

In summary, the topsoils of the Templeton and Wakanui taxonomic units are similarly variable in terms of the three physical properties due to their textural uniformity and the smoothing effect of uniform management and biotic activities. Properties within their subsoils, however, display different levels of heterogeneity, a reflection of the variability of textural layering and volume of material measured. The Eyre topsoil is more variable than the other two topsoils due to the presence of included gravels.

6.3.5 Analysis of variance

The three taxonomic units differ in terms of the physical properties, yet are not completely uniform themselves in terms of these properties. The question therefore arises as to the effectiveness of the morphologically-based soil classification system in differentiating units which simplify and reduce the overall spatial variability of soil physical properties within a region. The proportion of the total variation in physical properties accounted for by this classification system can be assessed using analysis of variance.

The components of variance in topsoil properties from different sources (between taxonomic units, within taxonomic units and over the three combined window areas), and the proportion of variance accounted for by the classification system are summarised in Table 6.6.

Table 6.6 Analysis of variance for bulk density, K_{fs} and moisture content among the topsoils of the three taxonomic units

Property	Sources	D.F.	S.S.	M.S.	F	r_i
Bulk density	Between series	2	0.17797	0.08898	29.50**	0.44
	Within series	105	0.31678	0.00302		
	Total	107	0.49474			
ln K_{fs}	Between series	2	5.591	2.795	4.45*	0.09
	Within series	105	65.953	0.628		
	Total	107	71.544			
Moisture content	Between series	2	109.10	54.55	15.86**	0.29
	Within series	105	361.21	3.44		
	Total	107	470.31			

* significant at 5% level
** significant at 1% level

The F-tests indicate that there are significant differences in mean values of the examined physical properties between the three taxonomic units. The differences, however, may be only significant between two of the three taxonomic units as shown by the t-test (Table 6.3). The intraclass correlation coefficients (r_i) indicate that nearly half of the variance in topsoil bulk density and about one third in moisture content are accounted for by differentiating soils in the region into the three soil series. The classification, however, made little contribution to the reduction of variability in topsoil hydraulic conductivity over the combined areas due to the very high variability of K_{fs} throughout the region.

The effectiveness of the classification system was further tested by using analysis of variance to compare the physical properties of pairs of taxonomic units at each depth (Table 6.7). Total variance in this case represents the variance of the combined data from each pair of designated window areas.

Table 6.7 Analysis of variance for bulk density, K_{fs} and moisture content for every two of the three taxonomic units

Depth	Taxonomic unit	Bulk density		ln K_{fs}		Moisture content	
		F	r_i	F	r_i	F	r_i
Topsoil depth	Eyre Templeton	45.63**	0.55	5.43*	0.11	0.19	-0.02
	Eyre Wakanui	39.48**	0.52	0.01	-0.03	18.60**	0.33
	Templeton Wakanui	0.00	-0.03	10.53**	0.21	50.54**	0.58
Subsoil depth	Templeton Wakanui	47.65**	0.56	115.61**	0.76	45.69**	0.55

* significant at 5% level
** significant at 1% level

The F-test is equivalent to the t-test shown in Table 6.3, though the inequality of variances between the two compared taxonomic units is considered in the t-test, but neglected in the analysis of variance. The results derived from the t-test and from the analysis of variance, however, are in general agreement with each other.

The r_i values reveal that more than half of the total variance in topsoil bulk density is reduced by separating the Eyre from either the Templeton or Wakanui series. A similar amount of improvement is made in subsoil bulk density when separating Templeton from Wakanui taxonomic units. The differentiation of Templeton from Wakanui taxonomic units, however, does not reduce the variability in topsoil bulk density (r_i = -0.03).

The intraclass correlations for K_{fs} show that the classification achieves very little in the reduction of the overall heterogeneity of topsoil "field-saturated" hydraulic conductivity. In contrast, the classification is very effective in separating Templeton from Wakanui soils in terms of the subsoil hydraulic conductivities (r_i = 0.76). This is particularly significant and important in view of the critical role that subsoils play in governing water movement.

More than half of the variance in moisture content, at both topsoil and subsoil depths, amongst Templeton and Wakanui taxonomic units is accounted for by the classification, and is thus due to differences between the two soils. Each taxonomic unit, therefore, is more homogeneous in moisture content than any two units combined as a whole. Only one third of the variance in topsoil moisture content amongst Eyre and Wakanui soils is related to the differences between the two series.

In summary, the overall spatial variability of bulk density is substantially decreased in topsoils by differentiating Eyre from either Templeton or Wakanui taxonomic units, and in subsoils by separating Templeton from Wakanui. The classification is especially effective and useful in separating Templeton from Wakanui taxonomic units in terms of subsoil K_{fs}. A large amount of variation in K_{fs} among the two subsoils is accounted for by the classification; the two soil series therefore have distinctly different subsoil hydraulic conductivities, and each taxonomic unit is more uniform than the two soil series as a whole. The heterogeneity in moisture content is also significantly reduced by separating Templeton from Wakanui taxonomic units, i.e. the two soil series taxonomic units differ from each other substantially in moisture content, and each series is more homogeneous in terms of moisture content at both topsoil and subsoil depths than the two series combined. The morphologically-based classification system, however, is not very effective in reducing spatial variability in topsoil K_{fs}. Separations of Templeton from Wakanui, and Eyre from Templeton taxonomic units are similarly ineffective in reducing the overall variability of topsoil bulk density and moisture content respectively.

6.4 Summary and conclusions

This chapter has substantiated some of the results obtained from the profile studies considered in the last chapter. General patterns of vertical and lateral differences in certain soil physical properties of taxonomic units may be revealed by the characterisation of typical profiles. In some cases, however, misleading conclusions may be drawn due to the limitations of having only relatively few samples to draw upon.

There are significant differences in mean values of bulk density, K_{fs}, and moisture content between topsoil and subsoil depths in both Wakanui and Templeton taxonomic units. Topsoils, because of the high organic matter content, well-developed structure, and profusion of biotic macropores, tend to have lower bulk densities and higher hydraulic conductivities than subsoils. The variability of these soil physical properties in topsoils is significantly lower than at the lower depths where inherited textural layers have not been modified or homogenized by biotic activity and management.

The subsoils of the Templeton and Wakanui taxonomic units are clearly separated in terms of mean values of the three examined soil physical properties. The classification is especially useful in differentiating the three taxonomic units in terms of mean soil moisture contents: at the time of sampling the Wakanui taxonomic unit contained significantly higher moisture than the Eyre and Templeton taxonomic units at both topsoil and subsoil depths. Significant differences in mean values of the examined properties do not always occur between topsoil depths, a feature that is again probably related to the effect of uniform management, addition of organic matter and biotic activities.

The topsoils of the Templeton and Wakanui taxonomic units are similar in terms of their physical-property variability. The Eyre topsoil, however, is more variable than the other two taxonomic units, probably due to the presence of included gravels and inclusion of different soil types within the Eyre series window area. The homogeneity of K_{fs} and moisture content also differs significantly between the subsoils of the Templeton and Wakanui taxonomic units: K_{fs} is more variable, and moisture content less variable, in the Wakanui subsoil than in the Templeton subsoil.

The morphological differentiation of Eyre from either Templeton or Wakanui taxonomic units accounts for a substantial amount of the overall topsoil variation in bulk density. The homogeneity of bulk density at the subsoil depth is significantly improved by separating Templeton from Wakanui taxonomic units. Substantial gain in homogeneity of moisture content is also obtained by separating Templeton from Wakanui taxonomic units. The classification, however, does not reduce the variability of K_{fs} among the topsoils of the combined window areas of the three taxonomic units. The separation of Wakanui from Templeton soils is particularly justified in terms of subsoil K_{fs}: a large amount of variation amongst the combined areas of the two soils is attributed to the difference between the two series. This is an important conclusion as the hydraulic conductivity of subsoils is a crucial parameter controlling water storage and movement.

CHAPTER 7

CONCLUSIONS

A 30 m × 30 m grid survey, supplemented by more intensive 15 m × 15 m surveys of selected zones, has led to the identification and delineation of Eyre, Templeton and Wakanui soil series simple mapping units within the study area. The complex soil distribution is related to, and largely determined by, the history and pattern of alluvial deposition within the area. The Eyre, Templeton and Wakanui soils are developed in a thin layer of finer-textured materials overlying channel-bar gravel deposits, in sandy channel-fill and levee deposits, and in fine-textured materials associated with intermediate zones between levees and the floodbasin respectively. Such patterns conform to those depicted by Cox (1978) in adjacent regions. Both the Eyre and Templeton soils are well-drained; strong mottles are not found in the profiles. The Wakanui soil, however, is imperfectly-drained and strong mottles are encountered throughout the profiles.

Geostatistical analysis of the 30 m grid data shows that each soil morphological parameter used as classification differentiae is spatially dependent among observations within certain localised regions, though they vary at different rates. Most variation of DM (depth to strong mottles) occurs over distances between 30 m (sampling spacing) and 430 m (range), whereas a large amount of variation in TS (thickness of loamy sand and/or coarser-textured layers) is present within distances shorter than the sampling spacing of 30 m. The third property, DG (depth to gravels) lies in between these two extremes.

The three morphological parameters also vary anisotropically, though to different extents. The anisotropic ratio is highest for DM (k = 5.84), and lowest for

TS (k = 1.58). with DG again being intermediate (k = 2.43). This anisotropic variation is clearly related to the pattern of alluvial deposition in the area. The direction of maximum variation for DM and DG is NE-SW. i.e. across a major abandoned channel hollow; the direction of least variation is perpendicular to that of maximum variation. i.e. parallel to the channel hollow. Soil mapping units. whose classification is based on the DM and DG parameters. are therefore elongated NW-SE in the direction of least variation. This pattern is not locally confined. however. as mapping units on smaller-scale soil maps of adjacent larger regions are similarly aligned. Such variations reflect the general past drainage patterns of channels flowing in a NW-SE direction across the broad region.

Block-kriged soil and single property maps compare favourably with their manually- and computer-drawn counterparts based on the original grid-survey data. The two methods produce generally similar distribution patterns. though the kriged boundaries tend to be regularised according to the spatial relationships expressed in semi-variograms. Some isolated small parcels. which differ sharply from their neighbourhoods. distinguished on the manually-drawn maps are not present on the kriged maps. This is due to the smoothing effect of kriging. whereby observation points are weighted within their neighbourhoods according to the spatial relationships reflected in the semi-variograms.

Geostatistical methods have been adopted to derive sampling strategies for future soil survey and soil variability studies in adjacent regions. Kriging standard errors can be computed for different sampling spacings in the direction of maximum variation. and for different numbers of observations for estimating mean soil property values of certain specified sizes of areas. The sampling spacing and sample size can be read from graphs showing the relationships between the three parameters. Where only a limited number of samples can be afforded. the sampling spacing in the direction of maximum variation that minimises the estimation error can be determined. The sampling spacing in the direction of minimum variation is k (anisotropic ratio) times the spacing in the direction of most variation. i.e. a rectangular scheme elongated in the direction of minimum variation. This geostatistical method is better than the conventional method in determining sampling strategies if structural dependence is present: less samples are required for kriging than for the conventional method to achieve the same level of precision. Only 14 samples, for instance, are required using the kriging method, as compared to 34 for the conventional method, to estimate the mean DM of an area of 100 m × 100 m (1 ha) with a tolerable error of 10 cm at the 90% confidence level. This gain in efficiency by kriging over the conventional method corresponds closely to that claimed by McBratney and Webster (1983) in their studies.

These results are important for the determination of sampling strategies in future soil surveys of adjacent larger regions having similar environmental controls. For soil surveys in other regions, it is recommended that an intensive soil survey be conducted first on a small representative area to reveal the overall spatial dependence and variability of soils. The sampling strategies can then be determined for the whole region using the approach outlined in this study. The conventional method of determining sampling strategies, however, should be applied if no spatial dependence is revealed in the initial intensive survey. Where more than one property is recorded and each varies at different rates, the sampling strategies should be determined according to the most important parameter to soil classification, or if they are of equal significance, based on the most variable property.

The Eyre, Templeton and Wakanui soil series, which are differentiated according to morphological criteria, are generally assumed to have markedly different hydraulic characteristics. This assumption is confirmed by the comparison of related soil physical properties between typical profiles of each soil series (taxonomic unit) from within the study area. Results indicate that bulk density generally increases, and porosity correspondingly decreases, from Eyre to Templeton to Wakanui soils due to differences in texture. "Field-saturated" hydraulic conductivities (K_{fs}) are similar for the three A horizons, though the Eyre soil has a slightly higher value due to the presence of large packing voids associated with gravels within and beneath the A horizon. The Templeton subsoil (42 cm - 60 cm) has significantly higher K_{fs} than the equivalent depth of the Wakanui soil, a difference that reflects its coarser texture and higher macroporosity. The Eyre and Templeton soils are freely-drained because of their higher hydraulic conductivities, whilst parts of the Wakanui soil are probably saturated, or close to saturated, for periods of the year. The soil hydraulic characteristics are most closely related to textural changes: mottling patterns are secondary features governed by soil texture.

These general differences in physical properties between the three soil series are substantiated by the quantitative assessment of soil-water movement and storage variability within, and between, selected morphologically-pure "window areas" (30 m x 30 m) of each taxonomic unit. The collection of 36 replicate samples for each measured soil property (bulk density, K_{fs} and moisture content) at two depths within each area, however, allows a more stringent statistical examination of the relationships. It also demonstrates how misleading conclusions may be potentially drawn from traditional single profile comparisons, where relatively few samples are considered.

The subsoils of the Templeton and Wakanui taxonomic units have significantly different mean values of the three examined physical properties. This is important as the subsoils play an important role in governing the overall water storage and movement within the whole soil profile. The classification is particularly useful in separating the three taxonomic units in terms of moisture contents: the Wakanui soil at the time of sampling contained significantly higher soil moisture than did the Eyre and Templeton soils at both topsoil and subsoil depths. Although moisture content varies substantially throughout the year, such information still provides an indication of the differences in soil water storage characteristics among the three soils. Significant differences in mean values of the examined physical properties, however, do not always occur between the three topsoils, a feature that probably reflects uniform management and biotic activities.

The variability of examined topsoil physical properties is similar within Templeton and Wakanui taxonomic units. The Eyre topsoil is more variable than the other two topsoils due to the existence of gravels and inclusions of different soil types within the Eyre series "window area". K_{fs} is more variable, and moisture content is less variable, in the Wakanui subsoil than the Templeton subsoil. The high variability of K_{fs} in the Wakanui subsoil is related to the existence of occasional cracks, earthworm channels, and the intermittently exposed sharply contrasted textural layers (loamy sand) beneath the fine-textured horizon (silty clay loam) where the K_{fs} measurements were made. The high variability of moisture content in the Templeton subsoil is probably due to the heterogeneous pore-size distribution of the sand layer.

The effectiveness of the morphologically-based soil classification in terms of soil physical properties is further assessed by using analysis of variance to partition the combined physical property variation from all three window areas into different components. This allows the proportions of variation accounted for by the classification to be derived. Analyses were first made for the three topsoils altogether and then for every pair of the three taxonomic units. The differentiation of Eyre from either Templeton or Wakanui taxonomic units substantially reduces the overall variability of bulk density in the topsoils. Homogeneity of subsurface bulk density and moisture content at both depths is significantly improved by separating Templeton from Wakanui taxonomic units. The classification, however, does not substantially reduce the variability of topsoil K_{fs}. The classification is particularly effective in separating Wakanui from Templeton taxonomic units in terms of the subsoil K_{fs}: a large amount of variation amongst the two soils is attributed to the differences between the two series. The fact that these soil series have distinctly different subsoil hydraulic conductivities is particularly important as this property plays a vital role in controlling water movement and storage.

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APPENDIX 1

Results from the 30 m × 30 m grid soil survey: depth to strong mottles (DM), depth to gravels (DG), and thickness of loamy sand and/or coarser-textured layers (TS)

X (W-E)	Y (S-N)	DM (cm)	DG (cm)	TS (cm)
0	210	30	100	0
30	210	35	100	37
60	210	35	100	15
90	210	100	100	38
120	210	100	80	20
150	210	100	75	25
180	210	100	100	0
210	210	100	70	70
240	210	100	75	25
270	210	100	100	0
300	210	100	45	55
330	210	100	65	65
360	210	100	100	40
390	210	100	100	0
0	180	30	100	15
30	180	25	100	25
60	180	25	100	0
90	180	100	100	35
120	180	100	65	35
150	180	100	100	55
180	180	100	100	40
210	180	100	100	70
240	180	100	85	80
270	180	100	100	0
300	180	50	100	0
330	180	100	20	80
360	180	100	35	65
390	180	30	100	10

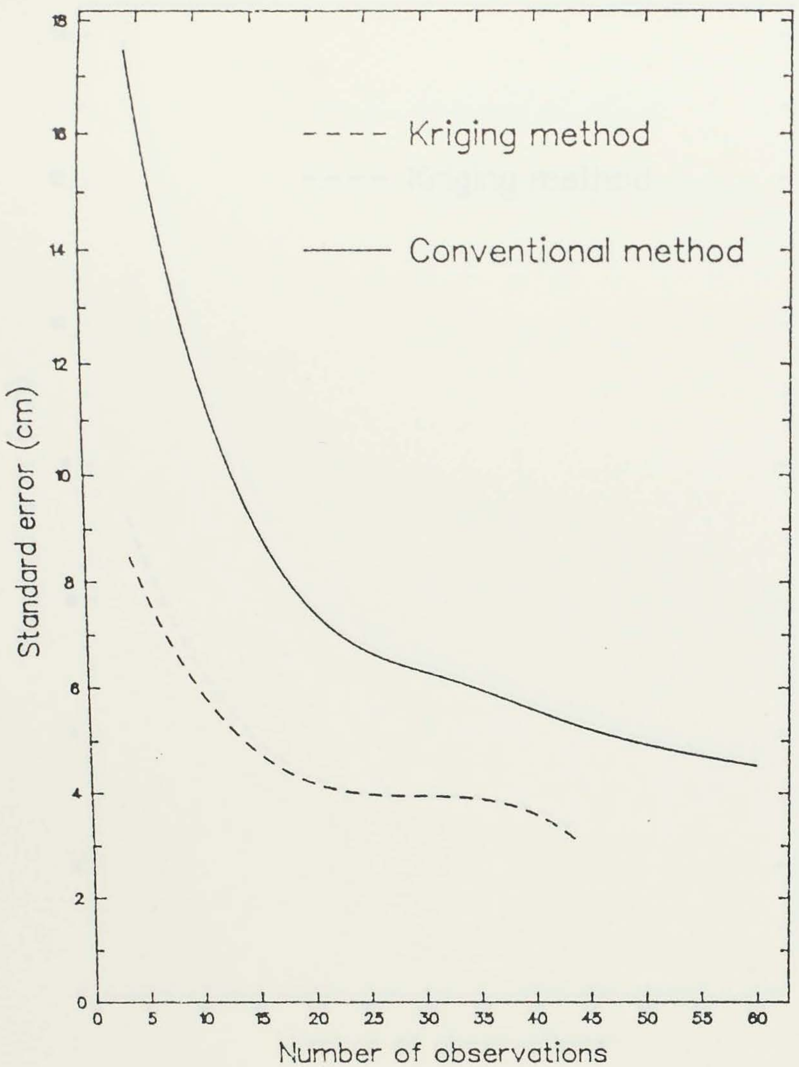
0	150	30	100	0
30	150	30	100	20
60	150	30	100	20
90	150	30	100	40
120	150	100	100	35
150	150	20	100	38
180	150	100	100	50
210	150	100	100	60
240	150	100	100	50
270	150	100	100	60
300	150	100	100	65
330	150	100	100	0
360	150	100	80	20
390	150	100	35	65
0	120	0	100	0
30	120	35	100	55
60	120	35	100	35
90	120	30	100	0
120	120	35	100	45
150	120	35	100	50
180	120	100	100	55
210	120	35	100	30
240	120	100	100	50
270	120	100	75	50
300	120	100	65	60
360	120	100	95	5
390	120	60	100	0
0	90	30	100	10
30	90	30	100	50
60	90	30	95	50
90	90	30	100	35
120	90	30	100	55
150	90	30	100	50
180	90	35	100	50
210	90	100	100	45
240	90	100	100	40
270	90	100	100	55
300	90	100	35	80
330	90	100	25	75
360	90	100	100	70

390	90	100	75	25
420	90	100	90	50
450	90	100	85	45
480	90	100	15	85
0	60	30	100	0
30	60	30	100	25
60	60	0	100	0
90	60	40	100	0
120	60	30	100	0
150	60	0	100	40
180	60	25	100	30
210	60	100	100	45
240	60	100	100	55
270	60	100	100	60
300	60	100	85	70
330	60	100	40	60
360	60	100	100	10
390	60	100	75	55
420	60	100	20	80
450	60	30	60	40
480	60	30	65	35
360	30	100	85	40
390	30	100	50	75
420	30	100	0	100
450	30	100	25	75
480	30	100	0	100
510	30	30	50	50
360	0	100	30	70
390	0	100	100	55
420	0	100	25	75
450	0	100	40	80
480	0	100	55	55
510	0	100	35	65

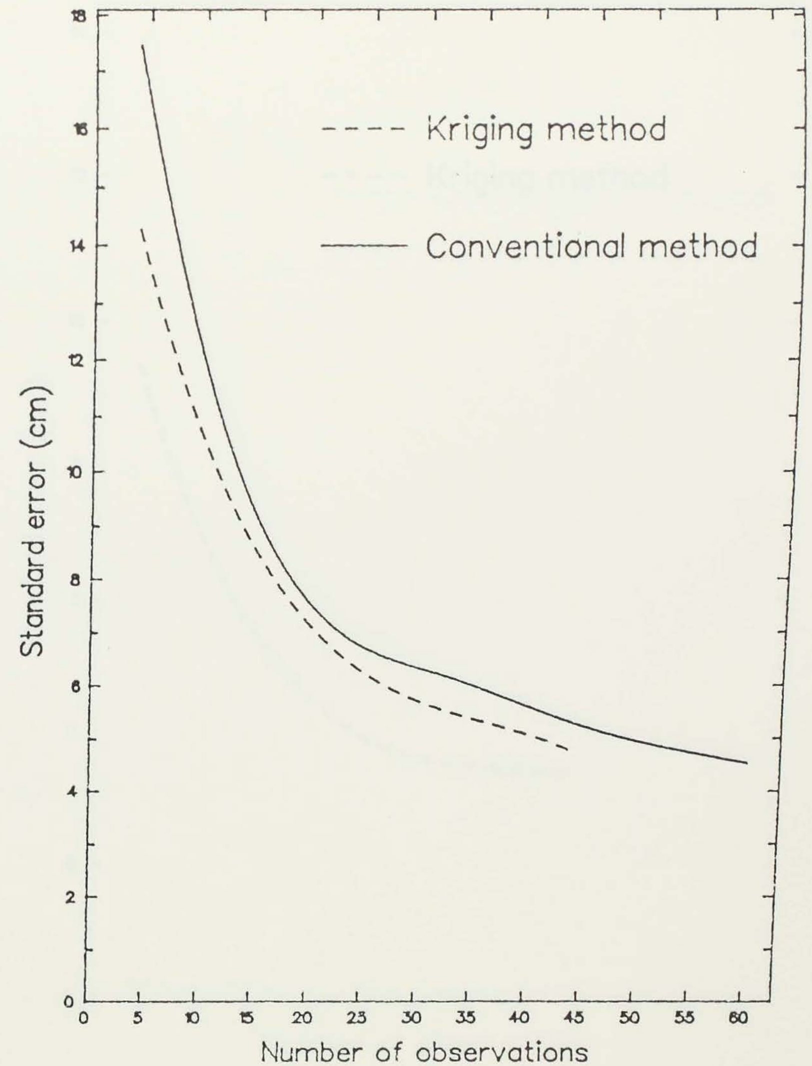
APPENDIX 2

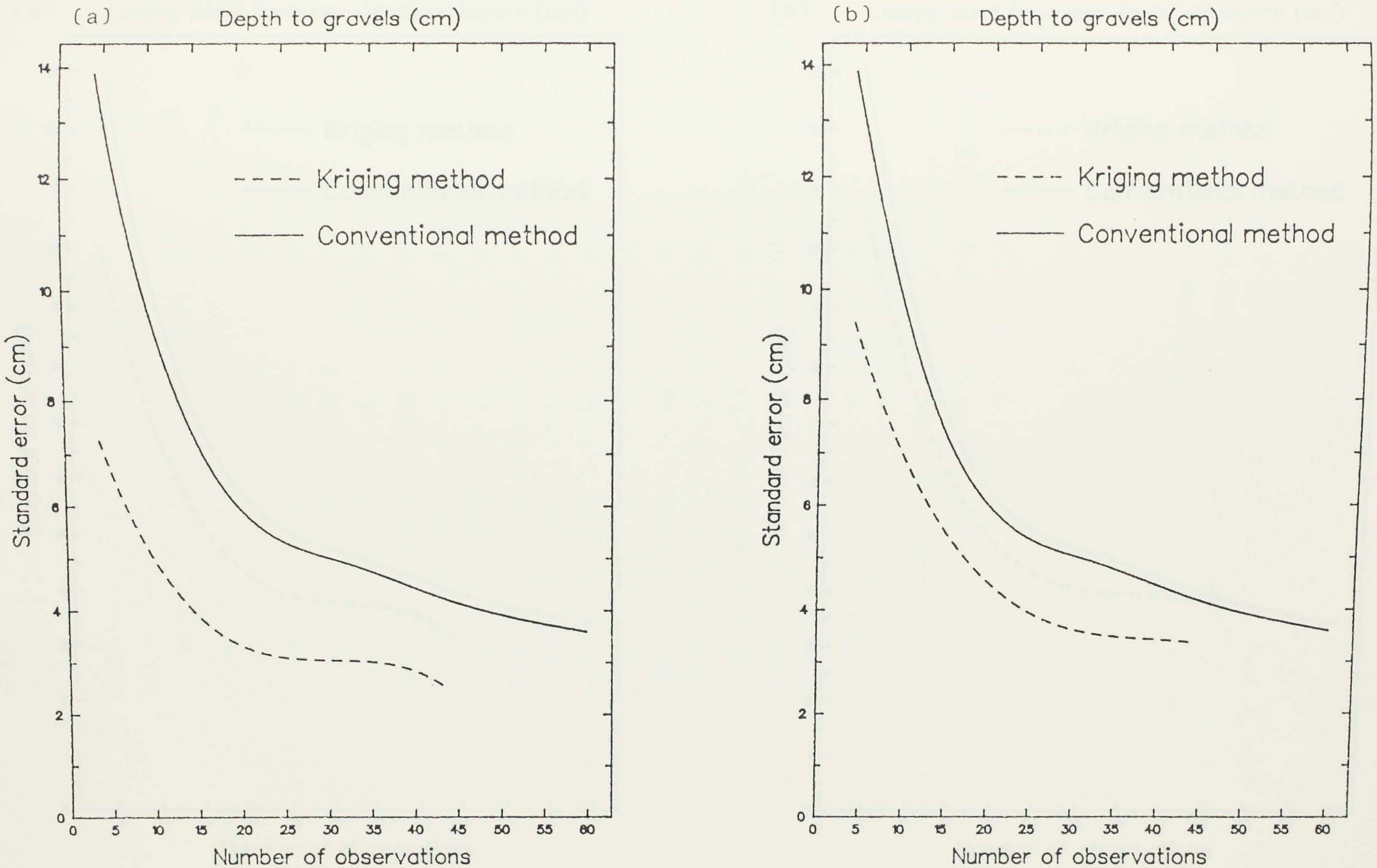
Graphs of standard error against sample size estimated by kriging and conventional methods for (a) 50 m × 50 m and (b) 300 m × 300 m blocks

(a) Depth to mottles (cm)

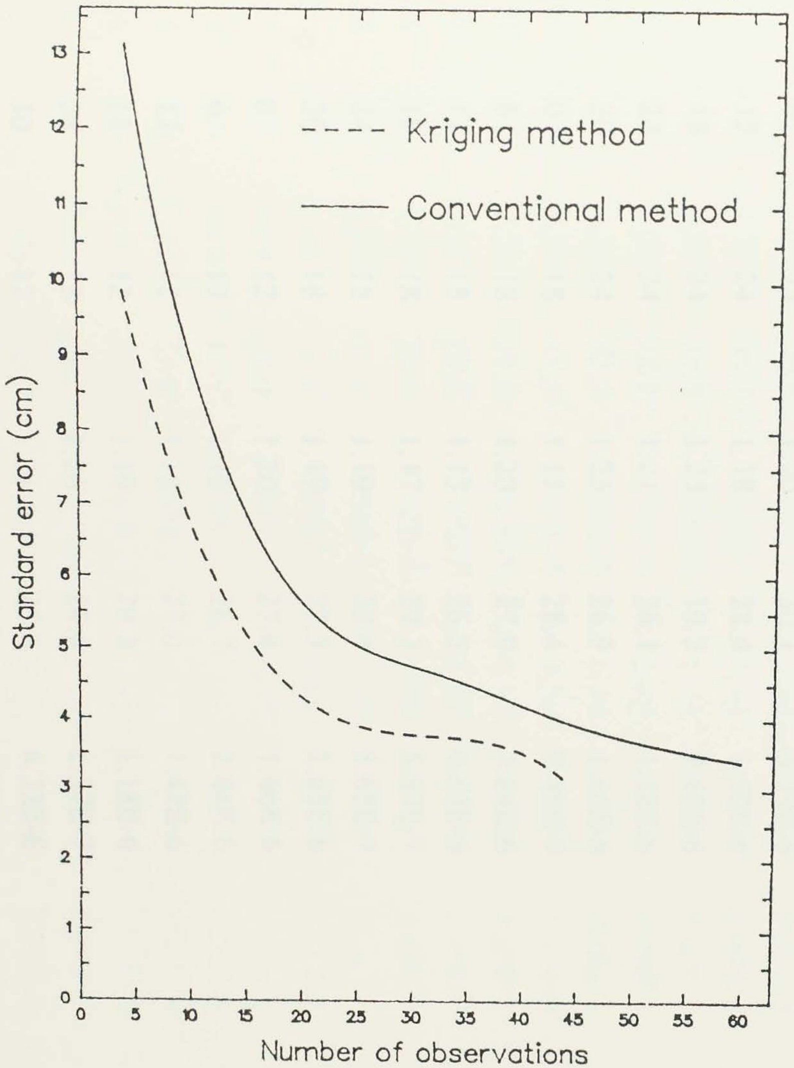


(b) Depth to mottles (cm)

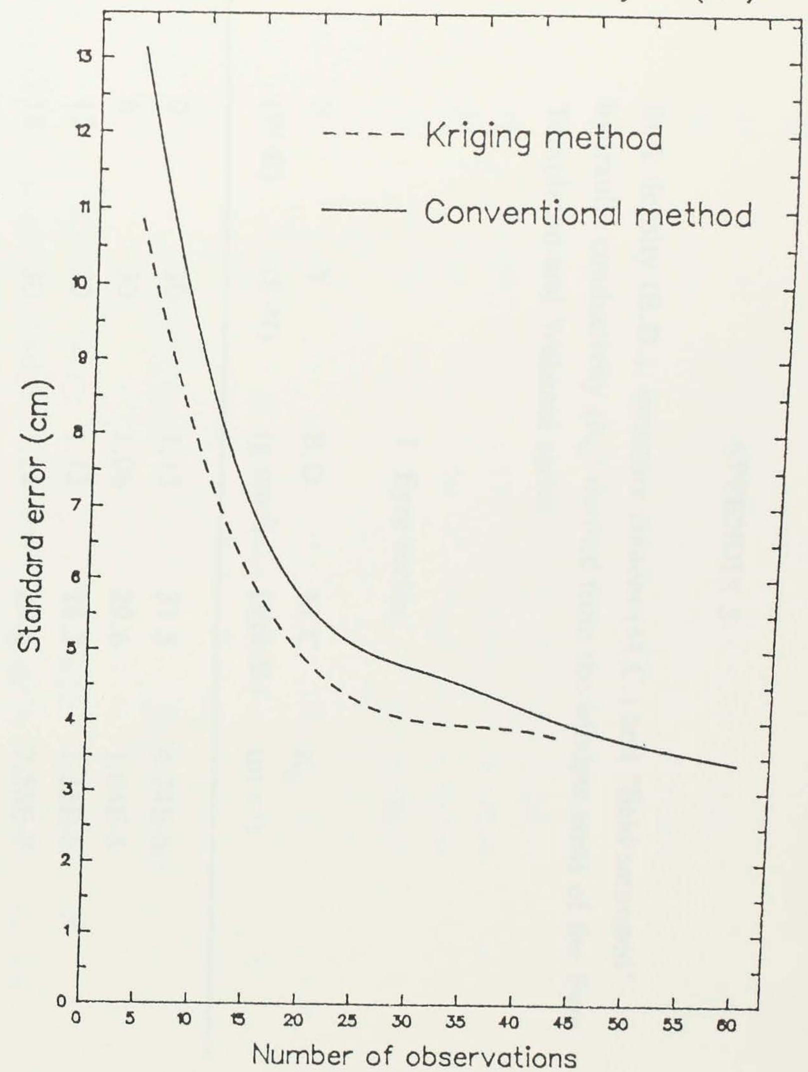



(a)
Depth to gravels (cm)
Kriging method
Conventional method
Standard error (cm)
Number of observations
0
2
4
6
8
10
12
14
0
5
10
15
20
25
30
35
40
45
50
55
60
(b)
Depth to gravels (cm)
Kriging method
Conventional method
Standard error (cm)
Number of observations
0
2
4
6
8
10
12
14
0
5
10
15
20
25
30
35
40
45
50
55
60

(a) Loamy sand/coarser–textured layers (cm)



(b) Loamy sand/coarser–textured layers (cm)



APPENDIX 3

Bulk density (B.D.), moisture content (M.C.) and "field-saturated" hydraulic conductivity (K_{fs}) derived from the window areas of the Eyre, Templeton and Wakanui series

I Eyre series

X (W-E)	Y (S-N)	B.D. (g cm^{-3})	M.C. (Vol.%)	K_{fs} (m s^{-1})
0	30	1.13	31.5	4.74E-6
6	30	1.06	29.6	1.04E-5
12	30	1.12	28.3	1.33E-6
18	30	1.24	29.6	7.58E-7
24	30	1.12	28.8	1.90E-7
30	30	1.21	30.1	7.80E-7
0	24	1.09	28.4	2.71E-7
6	24	1.20	30.1	3.79E-6
12	24	1.18	28.4	1.90E-6
18	24	1.25	19.2	9.48E-6
24	24	1.21	26.1	1.18E-6
30	24	1.25	26.9	1.66E-6
0	18	1.13	28.4	9.48E-7
6	18	1.20	27.9	2.84E-6
12	18	1.13	26.5	6.63E-6
18	18	1.17	29.3	5.69E-7
24	18	1.19	28.4	5.69E-7
30	18	1.19	27.5	2.65E-6
0	12	1.20	27.4	1.66E-6
6	12	1.16	28.7	2.84E-6
12	12	1.13	27.0	1.42E-6
18	12	1.19	29.3	1.18E-6
24	12	1.26	28.6	2.37E-7
30	12	1.21	25.6	6.32E-6
0	6	1.17	26.1	3.79E-7
6	6	1.18	28.8	5.69E-7

12	6	1.07	27.4	7.58E-7
18	6	1.22	18.1	7.58E-7
24	6	1.20	29.1	7.58E-7
30	6	1.15	27.5	1.14E-6
0	0	1.28	26.7	3.79E-7
6	0	1.29	27.3	5.69E-7
12	0	1.16	27.9	7.58E-7
18	0	1.17	29.1	2.46E-6
24	0	1.03	25.6	1.71E-6
30	0	1.21	30.1	7.58E-7

II Templeton series

		Topsoil depth			Subsoil depth		
X (W-E)	Y (S-N)	B.D. (g cm^{-3})	M.C. (Vol.%)	K_{fs} (m s^{-1})	B.D. (g cm^{-3})	M.C. (Vol.%)	K_{fs} (m s^{-1})
0	30	1.21	26.3	1.90E-7	1.32	17.5	7.18E-6
6	30	1.29	31.1	5.69E-7	1.38	23.8	1.26E-7
12	30	1.27	26.6	3.79E-7	1.45	25.4	5.67E-7
18	30	1.27	26.2	5.69E-7	1.61	23.5	1.89E-6
24	30	1.26	25.4	7.58E-7	1.49	30.2	5.67E-7
30	30	1.28	25.8	3.79E-7	1.45	22.3	3.78E-7
0	24	1.24	30.2	5.69E-7	1.49	22.5	7.56E-7
6	24	1.30	31.5	3.79E-7	1.58	24.9	1.32E-6
12	24	1.26	27.8	6.63E-7	1.35	20.4	5.67E-7
18	24	1.23	25.8	9.48E-7	1.67	25.0	2.65E-7
24	24	1.18	26.1	1.42E-6	1.56	26.2	9.45E-7
30	24	1.24	25.4	7.58E-7	1.51	23.7	1.89E-7
0	18	1.22	27.3	7.58E-7	1.48	23.7	3.79E-7
6	18	1.28	28.0	7.58E-7	1.45	23.1	3.79E-7
12	18	1.27	27.9	5.69E-7	1.41	21.9	1.33E-6
18	18	1.25	27.8	7.58E-7	1.43	23.0	5.69E-7
24	18	1.29	27.4	6.44E-7	1.42	25.7	4.74E-7
30	18	1.23	26.6	1.42E-6	1.44	26.7	1.90E-7
0	12	1.19	28.3	3.79E-7	1.61	24.8	1.90E-7
6	12	1.34	30.7	5.69E-7	1.54	21.5	5.69E-7
12	12	1.25	25.5	3.98E-6	1.56	23.7	1.90E-7

18	12	1.25	27.2	1.71E-6	1.58	23.7	1.90E-7
24	12	1.21	26.2	7.58E-7	1.52	24.7	3.79E-7
30	12	1.25	27.6	1.52E-6	1.38	25.1	1.90E-7
0	6	1.33	26.8	5.67E-7	1.50	22.3	3.79E-7
6	6	1.29	28.7	4.74E-6	1.57	23.0	7.58E-7
12	6	1.27	28.4	9.48E-7	1.38	20.1	1.14E-6
18	6	1.26	28.1	9.48E-7	1.55	25.9	3.79E-7
24	6	1.32	28.6	1.33E-6	1.46	27.2	7.58E-8
30	6	1.26	27.4	7.58E-7	1.59	24.3	1.90E-7
0	0	1.26	28.5	5.69E-7	1.43	24.3	1.23E-6
6	0	1.43	26.1	3.79E-7	1.61	26.0	9.48E-7
12	0	1.22	26.5	3.79E-7	1.42	28.1	1.71E-6
18	0	1.31	27.2	5.69E-7	1.40	17.3	1.14E-6
24	0	1.20	26.4	1.71E-6	1.46	24.5	3.79E-7
30	0	1.23	26.0	1.42E-6	1.42	23.6	1.14E-6

III Wakanui series

		Topsoil depth			Subsoil depth		
X (W-E)	Y (S-N)	B.D. (g cm^{-3})	M.C. (Vol.%)	K_{fs} (m s^{-3})	B.D. (g cm^{-3})	M.C. (Vol.%)	K_{fs} (m s^{-1})
0	30	1.24	29.6	2.53E-6	1.44	26.6	5.44E-9
6	30	1.38	28.6	1.58E-6	1.63	26.5	3.64E-8
12	30	1.27	31.3	1.26E-6	1.60	27.2	3.11E-9
18	30	1.17	31.4	7.58E-7	1.66	26.3	1.55E-9
24	30	1.41	30.8	9.48E-7	1.59	28.9	5.68E-7
30	30	1.23	30.5	5.69E-7	1.48	26.5	5.38E-8
0	24	1.28	28.3	2.46E-6	1.66	27.8	2.20E-8
6	24	1.25	31.0	1.14E-6	1.59	26.9	3.06E-7
12	24	1.27	29.9	2.08E-6	1.62	25.9	6.83E-9
18	24	1.24	31.2	5.69E-7	1.68	26.1	2.58E-9
24	24	1.26	28.9	1.71E-6	1.64	30.5	6.23E-9
30	24	1.23	29.6	1.33E-6	1.51	27.1	1.23E-8
0	18	1.32	29.2	1.32E-6	1.76	26.7	4.15E-8
6	18	1.17	27.8	2.27E-6	1.55	30.5	2.54E-9
12	18	1.23	29.8	1.90E-6	1.67	29.1	2.54E-9
18	18	1.22	29.3	1.52E-6	1.60	24.1	1.90E-7

24	18	1.25	30.6	1.52E-6	1.60	28.5	1.35E-7
30	18	1.29	30.5	2.27E-6	1.53	26.0	4.87E-7
0	12	1.32	30.1	1.90E-6	1.68	26.2	5.19E-9
6	12	1.22	29.5	9.48E-7	1.64	28.3	3.27E-9
12	12	1.26	29.4	1.52E-6	1.64	28.5	4.64E-9
18	12	1.26	27.7	1.71E-6	1.70	26.4	1.33E-7
24	12	1.19	29.8	7.31E-6	1.62	28.0	5.98E-9
30	12	1.44	29.6	9.48E-7	1.55	27.7	4.03E-8
0	6	1.23	28.8	5.69E-7	1.73	29.3	3.11E-9
6	6	1.29	30.3	9.48E-7	1.56	26.9	1.56E-9
12	6	1.24	30.5	1.71E-6	1.64	27.7	1.39E-7
18	6	1.27	30.7	1.16E-6	1.50	26.1	3.95E-9
24	6	1.29	30.4	1.33E-6	1.60	27.6	4.93E-8
30	6	1.26	29.0	2.32E-6	1.67	27.2	3.27E-9
0	0	1.24	29.2	7.58E-7	1.64	26.5	2.31E-9
6	0	1.25	29.1	1.18E-6	1.59	26.6	1.02E-8
12	0	1.27	29.4	1.58E-7	1.56	26.9	1.35E-9
18	0	1.27	29.8	7.11E-7	1.58	26.2	9.24E-9
24	0	1.20	26.8	1.18E-6	1.65	25.0	7.52E-8
30	0	1.25	29.4	8.12E-7	1.64	25.8	1.30E-8