SOILS AND GEOMORPHOLOGY OF A LOWLAND RIMU FOREST MANAGED FOR SUSTAINABLE TIMBER PRODUCTION

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Abstract

Saltwater Forest is a *Dacrydium cupressinum*-dominated lowland forest covering 9000 ha in south Westland, South Island, New Zealand. Four thousand hectares is managed for sustainable production of indigenous timber. The aim of this study was to provide an integrated analysis of soils, soil-landform relationships, and soil-vegetation relationships at broad and detailed scales. The broad scale understandings provide a framework in which existing or future studies can be placed and the detailed studies elucidate sources of soil and forest variability.

Glacial landforms dominate. They include late Pleistocene lateral, terminal and ablation moraines, and outwash aggradation and degradation terraces. Deposits and landforms from six glacial advances have been recognised ranging from latest Last (Otira) Glaciation to Penultimate (Waimea) Glaciation. The absolute ages of landforms were established by analysis of the thickness and soil stratigraphy of loess coverbeds, augmented with radiocarbon dating and phytolith and pollen analysis.

In the prevailing high rainfall of Westland soil formation is rapid. The rate of loess accretion in Saltwater Forest (ca. 30 mm ka^{-1}) has been low enough that soil formation and loess accretion took place contemporaneously. Soils formed in this manner are known as upbuilding soils. The significant difference between upbuilding pedogenesis and pedogenesis in a topdown sense into an existing sediment body is that each subsoil increment of an upbuilding soil has experienced processes of all horizons above. In Saltwater Forest subsoils of upbuilding soils are strongly altered because they have experienced the extremely acid environment of the soil surface at some earlier time. Some soil chronosequence studies in Westland have included upbuilding soils formed in loess as the older members of the sequence. Rates and types of processes inferred from these soils should be reviewed because upbuilding is a different pedogenic pathway to topdown pedogenesis.

Landform age and morphology were used as a primary stratification for a study of the soil pattern and nature of soil variability in the 4000 ha production area of Saltwater Forest. The age of landforms (> 14 ka) and rapid soil formation mean that soils are uniformly strongly weathered and leached. Soils include Humic Organic Soils, Perch-gley Podzols, Acid Gley Soils, Allophanic Brown Soils, and Orthic or Pan Podzols. The major influence on the nature of soils is site hydrology which is determined by macroscale features of landforms (slope, relief, drainage density), mesoscale effects related to position on landforms, and microscale influences determined by microtopography and individual tree effects. Much of the soil variability arises at microscales so that it is not possible to map areas of uniform soils at practical map scales. The distribution of soil variability across spatial scales, in relation to the intensity of forest management, dictates that it is most appropriate to map soil complexes with boundaries coinciding with landforms.

Disturbance of canopy trees is an important agent in forest dynamics. The frequency of forest disturbance in the production area of Saltwater Forest varies in a systematic way among landforms in accord with changes in abundance of different soils. The frequency of forest turnover is highest on landforms with the greatest abundance of extremely poorly-drained Organic Soils. As the abundance of better-drained soils increases the frequency of forest turnover declines. Changes in turnover frequency are reflected in the mean size and density of canopy trees (*Dacrydium cupressinum)* among landforms. Terrace and ablation moraine landforms with the greatest abundance of extremely poorly-drained soils have on average the smallest trees growing most densely. The steep lateral moraines, characterised by well drained soils, have fewer, larger trees. The changes manifested at the landform scale are an integration of processes operating over much shorter range as a result of short-range soil variability. The systematic changes in forest structure and turnover frequency among landforms and soils have important implications for sustainable forest management.

Keywords: glacial stratigraphy; loess; soil stratigraphy; Aokautere Ash; oxygen isotope record; phytoliths; thermoluminescence dating; pedogenesis; soil chronosequences; loess; soil variability; soil survey; soil-landform models; forest soils; forest disturbance; disturbance frequency; soil-vegetation relationships; windthrow; earthquakes

DECLARATION OF ORIGINALITY

This thesis reports the original work of the author except where otherwise stated.

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CERTIFICATE OF SUPERVISION

I certify that the work described in this thesis was conducted under my supervision.

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Chapter 1

General Introduction

Background to the study

This study forms part of a growing body of knowledge of the environment and ecosystems of Westland, on the west coast of New Zealand's South Island. It is principally a soil study, but by the very nature of soil science, involves amongst other disciplines, aspects of geology, geomorphology, ecology and botany.

Scientific interest in Westland has stemmed partly from economic motives involving exploitation of mineral and timber resources but also from the unique opportunities to study relatively unmodified environments. Westland is a narrow region, extending over 400 km in length, between the main divide of the rugged Southern Alps in the east and the Tasman Sea to the west. It is linked to Canterbury to the east by only three alpine passes throughout its length. The southern-most, the Haast Pass, was opened for road traffic as recently as 1965.

Accessibility has always been a major limitation to settlement and economic activity in Westland. The coastline is rugged and affords few natural harbours for ships. The historically important ports of Hokitika, Greymouth and Westport (Fig. 1) all require crossing treacherous river bars which accounted for many shipwrecks. The abundance of large and swift rivers, high rainfall, dense rainforest, rugged terrain and aggressive insects combined to ensure that before the latter half of this century travel was dangerous and life arduous. The most concentrated European impact on the Westland landscape spread from the port town of Hokitika, fuelled by the first gold rush in 1865 and the subsequent developing timber trade (Millar, 1959; Peat, 1987). South of Ross in south Westland the landscape remained relatively untouched other than around isolated mining settlements.

The lowlands of Westland to the west of the Southern Alps are a piedmont complex of mostly glacial landforms with scattered glacially modified outcrops of Paleozoic sedimentary and igneous rock. Since the end of the Last Glaciation rivers have infilled formerly ice-filled valleys so that now Pleistocene glacial moraines and outwash terraces are separated by elongate, Late Holocene floodplains. Where the Southern Alps rise abruptly above the lowland piedmont at the Alpine Fault, Holocene age fan deposits drape older landforms or interfinger with the alluvial fill of the valley floors (Mew, 1980b).

The dichotomy in ages of landforms is reflected in the soils, vegetation patterns and land use. Soils on Pleistocene age landforms which are under high rainfall have become very strongly leached, often podzolised, and of very low nutrient status. Their low nutrient status is accompanied on all but the steepest slopes by very poor drainage and aeration due to the presence of poorly structured, silty upper horizons and/or the occurrence of pans at depth

(Jackson, 1984; Mew and Leamy, 1977). Soils of this type became known as *pakihi soils* because they were often associated with swampland vegetation known locally as pakihi (Mew, 1983). In contrast, soils on Holocene floodplains or fans are poorly drained to well drained recent soils of moderate fertility (Mew and Leamy, 1977).

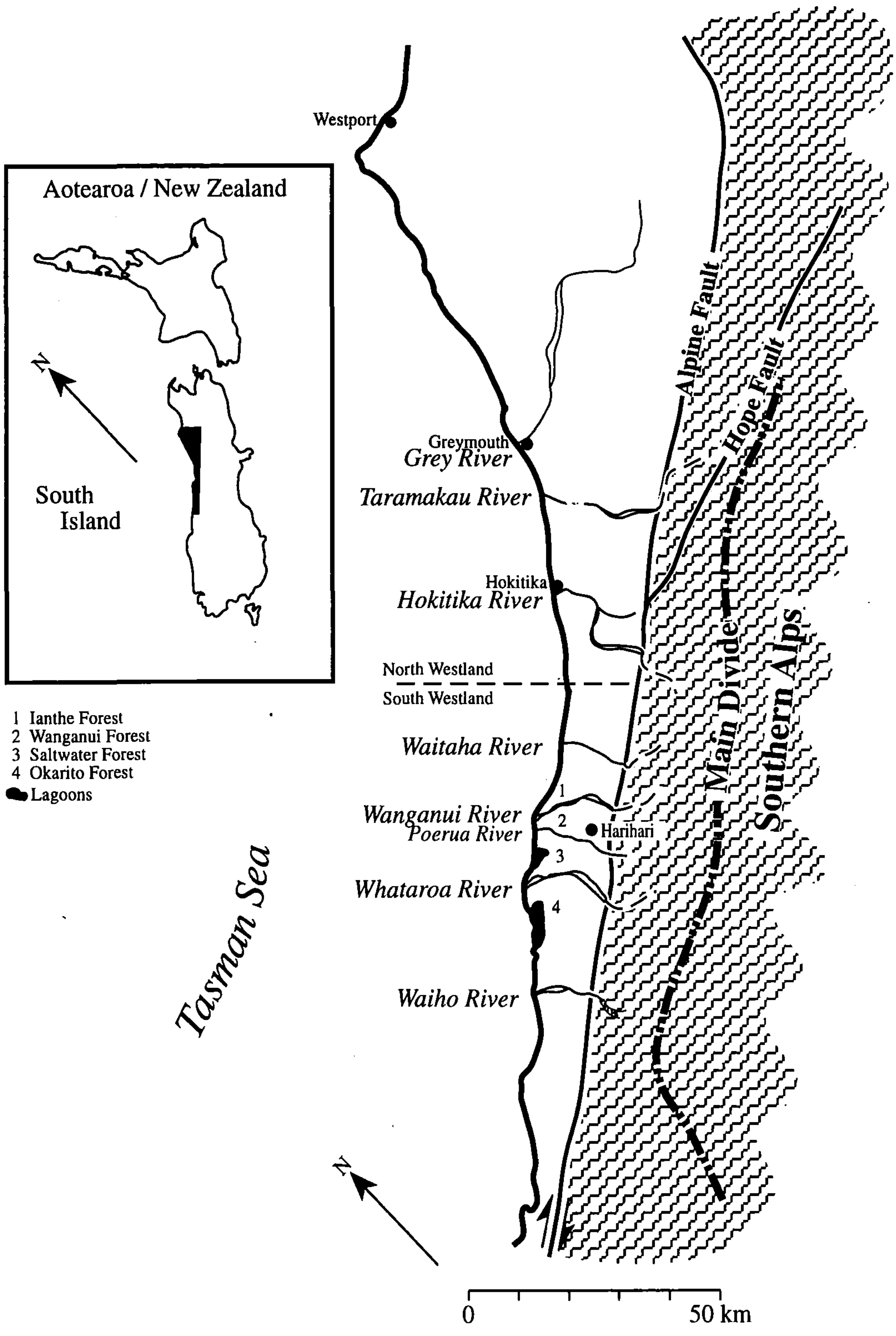

Fig. 1. Part of Westland showing features referred to in the text.

Forest types show a strong association with soils and landforms (Chavasse, 1971). Within the region described as the beech gap between the Grey River in the north and the Paringa River in the south, canopy species in forests are dominantly podocarps. On Pleistocene landforms rimu (*Dacrydium cupressinum*), miro (*Prumnopytis ferruginea*) and silver pine (*Lagarostrobos colensoi)* are dominant. The podocarps are long-lived, slow-growing trees. Rimu routinely lives to 400 years and some reach 800 or even 1000 years (Norton et al., 1988). Volume increment in dense stands (ca. 200 stems/ha) ranges between 0.5 and 2 m^3/ha/annum. On terraces rimu forms dense stands often with a continuous, 30-40 m high canopy above subdominant and less common miro in the subcanopy. In very poorly drained areas silver pine can become codominant. A broadleaf understorey is dominated by kamahi (*Weinmannia racemosa)* and quintinia (*Quintinia acutifolia)* with *Phyllocladus alpinus.* Some taller individuals may reach the subcanopy. The lower tiers include a shrub and tree fern storey and a ground layer of bryophytes and ferns.

Rimu forest on rolling to steep moraines has a different character. Scattered, large diameter, tall rimu are emergent over a broadleaf canopy comprising mainly kamahi and quintinia. Large southern rata (*Metrosideros umbellata)* or Hall's totara (*Podocarpus hallii*) also occur as emergents but are rare in comparison to rimu. Epiphytes are more common and as slopes become steeper so does the abundance of lianes including supplejack (*Rhipogonum scandens),* climbing rata (*Metrosideros sp.)* and kie kie (*Freycinetia baueriana*) until they form almost impenetrable tangles in places.

On the more fertile recent soils the podocarps kahikatea (*Dacrycarpus dacrydioides*), matai (*Prumnopytis taxifolia)* and totara (*Podocarpus totara* var. *waihoensis)* are more common. Kahikatea forms dense stands on poorly drained soils, while matai and totara favour the better drained gravelly soils. Across flights of river terraces from young to old, forest composition progressively shifts to dominance by rimu as soils become more infertile and more poorly drained (Duncan, 1993; Norton, 1990; Sowden, 1986).

The gold boom and associated development created the first demand for timber and forest clearance. Timber was required for buildings, flumes, tunnel shorings, bush tramways, firewood and later fence posts, telegraph posts and sleepers for railways. As the gold boom waned the milling of the forest became a primary activity and impetus to the local economy. From 1867 a local oversupply and falling prices threatened the livelihoods of many mill owners and workers, and a local coastal trade to other parts of New Zealand began to develop. By 1876 export to Australia was under way. However competition remained cut-throat and large tracts of forest were cut and sold for prices which afforded sawmillers only a subsistence living. Moreover, cutting was haphazard and very inefficient. Most of the sawmilling concentrated on the rimu (red pine) forests on Pleistocene landforms but the kahikatea, matai and totara forest were cleared for farmland to feed the population. Much of the kahikatea forest was burned until the mid-1890s when its utility as a butter box timber was realised.

Widespread forest clearance for timber continued southwards as far as Ross following the Midland Railway opening in 1906. During the period from 1865 to 1906 there had been a recognition from many people including a group of naturalist politicians in Wellington that a timber famine in New Zealand was imminent. By the mid 1890s it was apparent that Westland had the last major tracts of native forest in New Zealand and was to provide the bulk of timber into the future. The first piece of legislation intended to foster better use of remaining forests and to prevent environmental and 'climatic' degradation was introduced to Parliament by Sir Julius Vogel, premier of New Zealand, in 1874. However it was not until 1919 that major areas of crown land were gazetted or provisionally gazetted as State Forest; these were later transferred to the control of the State Forest Service which was established in 1921. Four hundred and sixty nine thousand hectares of State Forest were created in Westland. The aim was to shift from the prevailing exploitive forest 'mining' to sustainable scientific forestry in the European tradition (Roche, 1987, p. 84).

Research on the composition, life history, growth rates and yields of Westland rimu forest began in the mid-1920s; much of it originating from the Canterbury College School of Forestry (Foweraker, 1925; Hutchinson, 1928; Hutchinson, 1929; Hutchinson, 1931; Hutchinson, 1932). This early research was summarised by Foweraker and Hutchinson (1935) who concluded that rimu terrace forest was a mosaic of essentially even-age patches that should, for silvicultural, practical and economic reasons, be logged accordingly. They advocated clear-cutting of mature and overmature stands in 70-75 year cycles. However this work was ignored, and it was not until 1956 that sustained yield management began (James and Franklin, 1977), although the silvicultural regime was based on hypotheses of Holloway (1954b). Holloway (1954a; 1954b) concluded that Westland rimu forests are a temporary phase in a succession from open mire to hardwood-dominated forest initiated by a change to a drier climate in the 13 th century. He advocated strip felling such that logged strips were no wider than the common seed dispersal range so that they could regenerate from seed sources in intervening unlogged strips. This should only take place, he argued, in early mature and sub-mature stands which stood a chance of regenerating - more or less mature stands would be encouraged along a progressive or regressive succession to hardwood dominated forest or pakihi respectively.

By 1967 strip-felling was replaced by selection logging following the work of Chavasse and Travers (1966). After a reassessment of the earlier work by Foweraker and Hutchinson (Foweraker, 1925; Hutchinson, 1928; Hutchinson, 1929; Hutchinson, 1931; Hutchinson, 1932), they concluded that rimu forest could be managed as uneven-aged and recommended selection logging. The selection logging techniques held sway into the 1980s although a variety of harvesting methods including crawler tractors, rubber-tyred skidders and cable hauler logging employing a variety of extraction patterns were trialled. By the late 1970s it became clear however that mortality induced by windthrow and damage during logging reduced the net volume increment of selectively-logged stands to very low or even negative values (James, unpublished ; James and Franklin, 1977). In response to these problems and also the practicalities of dealing with local logging crews poorly equipped for selection logging, sustainable management shifted in the 1980s to coupe felling. Small areas (up to 20 ha) were clearfelled in the manner similar to that advocated by Hutchinson and Foweraker (1931). This was seen to be more in accord with the natural dynamics and replacement pattern of terrace forests indicated by earlier research (Foweraker, 1925; Hutchinson, 1928; Hutchinson, 1929; Hutchinson, 1931; Hutchinson, 1932; Poole, 1937) and more recent research (Six Dijkstra et al., 1985). Mature and over-mature stands were targeted with 100% harvest and the site replanted with nursery-raised seedlings. In this way problems of post-logging mortality due to logging damage were avoided and an intact forest margin at coupe boundaries reduced windthrow losses relative to selection logging. Furthermore, restocking was not left to the vagaries of natural regeneration.

Parallel to the development of technology of sustained yield forestry in the 1970s was the growth of environmental lobby groups. By the time Saltwater and Okarito Forests in south Westland (Fig. 1) were gazetted for sustained yield management in 1984 all management had effectively stopped. Extension of Westland National Park to include part of North Okarito Forest as a result of environmental lobby group pressure had forced the government to sacrifice the other State Forests in south Westland (Ianthe Forest and Wanganui Forest, Fig. 1) to honour long term contracts between the crown and the logging companies. Ianthe and Wanganui Forests, partially logged in the selective logging era, were systematically clearfelled. Expensive sustained yield management could not produce timber at a price that could compete with cheap clearfelled timber, and it was not until 1992, after forest destined for clearfelling was exhausted, that management in Saltwater and Okarito Forests began. By this stage the New Zealand Forest Service had been disestablished and management of the sustained yield management forest had shifted to Timberlands West Coast Ltd, a state owned enterprise. Management is ongoing and based on an assumption of the forest comprising a mosaic of even-age patches. However, all harvesting except roadside salvage is done by aerial extraction by helicopter to minimise site and stand damage (Hammond and Richards, 1995).

Soil studies in Westland began early this century (Ashton, 1910), and most focussed on the rehabilitation of derelict land left after large scale clearance of rimu forest on Pleistocene landforms. After initial clearance the land was burned, oversown and grazed with cattle which roamed in a semi-wild state grazing what grass was available and browsing the bush margins. The saturated soil conditions, acidity and low fertility were antagonistic to grass growth and the rough state of the land with debris left after logging hampered access. As a result the land often reverted to regenerating scrub, gorse or pakihi vegetation. The Cawthron Institute began work on rehabilitation of pakihi land in the 1920s concentrating on land drainage, fertility and pasture maintenance. From the 1950s the Department of Lands and Survey took up the challenge of land improvement on large tracts of state owned land.

The first reconnaissance soil survey of Westland at four miles to the inch was published in 1950 (Gibbs et al., 1950) with the information to be later recompiled, with some additional field survey, as part of the General Soil Survey of the South Island at the same scale (New Zealand Soil Bureau, 1968). Later, detailed soil surveys were carried out in north Westland principally with the aim of identifying land suitable for agriculture and exotic forestry (Mew, 1980a; Mew, 1980b; Mew and Adams, 1975). As a complement to the soil surveys were a number of detailed studies on the properties, genesis and classification of soils representative of the important soil series (Barratt, 1983; Farmer et al., 1984; Jackson, 1984; Mew and Lee, 1981; Mew et al., 1983; Ross et al., 1977; Thomas and Lee, 1984).

The very dynamic landscape has afforded many opportunities for soil chronosequence studies which have added to our understanding of soil development, particularly in a strong leaching environment. Furthermore many of the chronosequences occur on landforms with unmodified vegetation thereby providing opportunities for studies of primary succession of vegetation from freshly deposited material to landforms and soils older than 22 000 years (Basher, 1986; Birkeland, 1993; Smith and Lee, 1984; Sowden, 1986; Stevens, 1963; Stevens, 1968; Wardle, 1980). Other studies have focussed on the variation of vegetation across soil and other environmental gradients (Collins, 1986; Cornere, 1992; Duncan et al., 1990; Lieffering, 1989; Norton, 1990; Rogers, 1995; Simmons, 1982; Stewart and Harrison, 1987) but apart from the study of Stewart and Harrison (1987) there has been no integrated study of soils and landforms at a landscape scale and none in the lowland rimu-dominated forests. The studies cited were either short range studies along transects or in a limited number of plots (Collins, 1986; Lieffering, 1989; Rogers, 1995; Simmons, 1982) or lacked any rigorous analysis of soils (Duncan et al., 1990; Norton, 1990).

Study aim

The aim of this study was to provide an integrated analysis of soils, soil-landform relationships and soil-vegetation relationships in Saltwater Forest, one of the rimu dominated lowland forests currently under sustained yield management. As well as adding to the current understanding of soil formation, properties and patterns in Westland, it was hoped that information relevant to forest management would be forthcoming. The only existing soil information for Saltwater Forest is provided by the General Soil Survey (New Zealand Soil Bureau, 1968) and a more detailed soil survey over a limited area (975 ha) (Almond, 1986).

Delineating the pattern of soils in the landscape relies on an understanding of the spatial and temporal variability of the soil forming factors (climate, organisms, relief, parent material, time, Jenny, 1941). To gain this understanding in the study area an integrated pedologic and geomorphic approach, known as soil geomorphology (McFadden and Knuepfer, 1990; Olson, 1989), was adopted because the factors of time, parent material and relief (topography) are intimately linked with the evolution of the landscape.

From other soil work in Westland, it was hypothesised that over the limited area studied the factors of relief (topography) and time with its co-related influence on parent material (see below), would be the major controls on soil pattern at least at a landscape scale. At the beginning of this study the only available information on landform age was from a 1:250 000 geological map (Warren, 1967) which grouped glacial deposits into two chronostratigraphic units generalised from the stratigraphy developed in north Westland (Suggate, 1965). Glacial deposits and landforms flanking major valleys were mapped as Moana Formation produced by the K3 advance (14-16.5 ka, Suggate, 1990) and Okarito formation for all older glacial deposits. The latter was thought to comprise deposits older than the Last (Otira) Glaciation included in Waimea and Tansey Formations in north Westland (Suggate, 1990). This age stratification was too broad to be of value for soil studies so the initial part of the study focussed on establishing the temporal array of landforms. This involved:

1. ordering landforms by relative age according to stratigraphic principles;
2. assessing the magnitude of differences of landform age by examining representative soils in light of the established soil chronosequence framework.

Early in this part of the study it was obvious that soils across the range of landforms did not follow a developmental sequence because loess had mantled the older landforms. However the loess coverbeds presented alternative opportunities for dating. Where loess production has been continual (though not necessarily constant) over time and in the absence of significant erosion, loess coverbeds thicken on progressively older landforms and often show a vertical sequence of weakly altered material separated by zones of strong soil development. This is traditionally represented as sequences of loess layers and interbedded paleosols (see for example Kukla, 1987). For loess of Pleistocene age, changes in loess deposition and soil development rates are widely accepted as being climatically controlled (Fink and Kukla, 1977; Hardcastle, 1889; Kukla, 1987; Palmer and Pillans, 1996; Raeside, 1964). If this model is valid, loess layers and paleosols can be correlated in a 'counting-back' fashion to dated climatic events established by the marine oxygen isotope record. Identifiable, dated tephras in loess establish isochronous horizons which greatly enhance the confidence of correlation.

The stratigraphy of buried soils in loess is one aspect of the discipline of soil stratigraphy which, more generally, refers to the chronological analysis of the vertical and lateral relationships between buried soil mantles (or the former land surfaces they underlay) (Brewer et al., 1970; Butler, 1959; Finkl, 1980; Morrison, 1978; Ruhe, 1956). The principles of soil stratigraphy (Finkl, 1980) were applied wherever exposure permitted to establish a relative age chronology between deposits and the landforms they comprise. The chronology was supported by absolute dating techniques (radiocarbon and thermoluminescence dating) as well as pollen analysis, phytolith analysis and tephrochronology.

Loess deposits on landforms of different type and age also presented an opportunity to study the influence of loess deposition on soil properties, soil evolution and ecosystems. Loess deposition rates in Westland are low (Mew et al., 1988a; Mew et al., 1988b) and soil formation rates are high (Tonkin and Basher, 1990). Potentially some of the properties of the older soils of Westland may be better understood by considering soil formation and loess accumulation to be contemporaneous. Chronosequence studies in Westland spanning an age range of > 18 ka had included soils formed at least partly in loess (Campbell, 1975; Ross et al., 1977; Stevens, 1963) yet its influence on soil evolution has not been seriously considered. Loess inputs are a flux of fresh material to soils with high leaching losses and low fertility and as such could retard or redirect soil development.

A stratification of the study area by landform and landform age provided the basis for analysis of soil variability within and between landforms. This was done according to a nested sample design in order to determine the sources of soil variability and the scales over which they operated.

The soil pattern so determined then allowed an analysis of the relationships between vegetation, soils and landforms. Vegetation data comprising about 311 0.2 ha plots distributed across the area of production forest were used for analysis of soil-vegetation relationships at a landscape scale while comprehensive vegetation data along two transects provided a window into processes and soil-vegetation relationships at short range.

Objectives and organisation of the thesis

This thesis is written in paper style so that although each paper forms part of the whole, they are also capable of standing alone. Each paper addresses a specific objective as listed below.

1. To elucidate the geomorphic history of Saltwater Forest to provide a basis for a time subdivision of the landscape.

   Chapter 2: Loess, soil stratigraphy and Aokautere Ash on late Pleistocene surfaces in south Westland, New Zealand: Interpretation and correlation with the glacial stratigraphy.

   Chapter 3: A reinterpretation of loess, soil stratigraphy and Aokautere ash on late Pleistocene surfaces in Saltwater forest, and a revised correlation with the glacial stratigraphy.

2. To compare properties of soils formed in slowly accreting loess with comparable soils formed in a single sediment body to determine the influence of loess accretion during the late Otiran on soil properties and soil genesis.

   Chapter 4: Pedogenesis by upbuilding through slow loess accretion in an extreme leaching and weathering environment, south Westland, New Zealand

3. To determine the nature and sources of soil variability in the area of production forestry in Saltwater Forest.

   Chapter 5: A multiscale approach to soil variability in Saltwater Forest, a podocarp/hardwood forest, south Westland, New Zealand.

4. To determine the relationships between soil and vegetation in the area of production forestry in Saltwater Forest

   Chapter 6: The structure of *Dacrydium cupressinum* (Podocarpaceae) dominated forest along landform and soil gradients, south Westland, New Zealand.

Chapters 7 is a synthesis and summary.

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Chapter 2

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Loess, soil stratigraphy and Aokautere Ash on late Pleistocene surfaces in south Westland: interpretation and correlation with the glacial stratigraphy

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Abstract

The detailed stratigraphy and chronology for Late Pleistocene glaciations in north Westland have not previously been applied to glacial deposits and landforms of south Westland, in part because of the lack of exposure. A stratigraphy for loess coverbeds offers the potential for discriminating and correlating landforms in this region. This study, in Saltwater Forest and surrounding areas in south Westland, confirms the presence of loess and of the incorporated 22 600 BP (radiocarbon years) Aokautere Ash beneath Late Pleistocene moraine and terrace surfaces. Loess sheets are thin (ca. 40 cm thick) and accretion rates have been low (ca. 30 mm ka^{-1} during the last stadial of the last glaciation). Boundaries between loess sheets were placed at the surface of morphologically recognised buried soils. Criteria for recognition of buried soils differ between the moderately well drained moraine sites and poorly to extremely poorly drained terrace sites. Five loess sheets have been recognised and correlated to the oxygen isotope record on the basis of the position of maximum counts of Aokautere Ash shards, pollen analysis and radiocarbon dates. Loess sheets 1 and 2 are of Last (Otira) Glacial age; loess sheet 3 accumulated during the Last (Kaihinu) Interglacial and loess sheet 4 accumulated during the penultimate (Waimea) glaciation. Loess sheet 5 is of at least Waimean age.

Keywords: glacial stratigraphy; loess; soil stratigraphy; Aokautere Ash; oxygen isotope record

1. Introduction

Westland, on the west coast of the South Island, is a key area for the study of the late Quaternary in New Zealand. The region extends from the main divide of the Southern Alps across the coastal piedmont to the Tasman Sea (Fig. 1). The Southern Alps rise abruptly above the piedmont at the Alpine Fault, which is a major element of the Australian/Pacific convergent

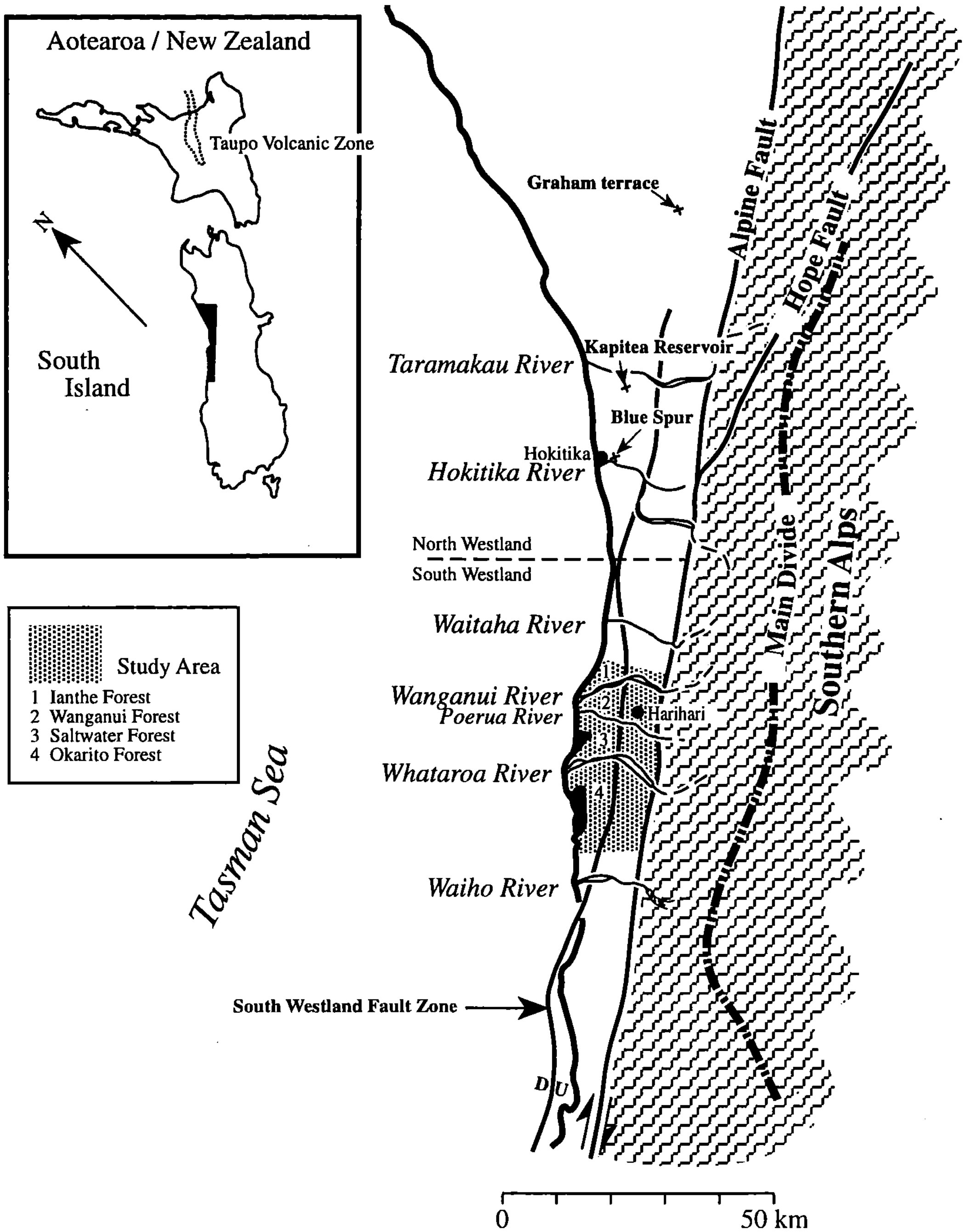

Fig. 1. Westland, South Island, New Zealand.

plate boundary. Precipitation is very high, ranging from 2400 mmyr^{-1} at the coast to 12 000 mmyr^{-1} between the mountain front and the main divide (Griffiths and McSaveney, 1983). Much of the precipitation in the mountains falls as snow which feeds glaciers; two of these, the Franz Josef and Fox glaciers, flow to less than 300 m above sea level. During the Pleistocene, glaciers extended beyond the alpine valleys out on to the piedmont plain, at times and in some areas as far as the Tasman Sea. Near the coast, glacial deposits commonly overlie marine sediments deposited during interglacials. Between the Taramakau and Hokitika rivers (Fig. 1), glacial moraines, outwash surfaces, wave cut platforms and beach cliffs have been preserved by moderate rates of tectonic uplift, and in general, older landforms are at higher elevations. The stratigraphy of glacial and interglacial deposits has been exposed by dissection in response to uplift and in this region a glacial stratigraphy has been established and refined over the last 25 years with recent correlation to the oxygen isotope record (Suggate, 1990)(Table 1).

Table 1. Glacial stratigraphy for north Westland (after Suggate, 1990).

Oxygen Isotope Stage	Glaciation	Glacial Advance		Interglacial
1				**Aranui**
		Kumara-3		
2	**Otira**		*Minor interval*	
		Kumara-2_2		
3			*Major interval*	
4		Kumara-2_1		
5				**Kaihinu**
6	**Waimea**	Kumara-1		
7				**Karoro**
8	**Waimaunga**	Hohonu		
9				Unnamed
10	**Nemona**	Cockeye		

In south Westland, many of the moraines and outwash surfaces remain uncorrelated to the stratigraphy developed in north Westland (Suggate, 1990). A stratigraphy for aeolian coverbeds would be a powerful correlation tool within the region. Assuming that loess sheet accumulation and intervening breaks in sedimentation are climatically controlled, estimated ages can be assigned to loess sheets and the surfaces they drape by correlation of the loess stratigraphy to the dated proxy climate record established from the deep sea oxygen isotope curve.

The presence of loess has been confidently established in north Westland (Young, 1967; Mew *et al.*, 1988) and although loess sheets are thin (less than ca. 60 cm on a ca. 50 ka surface), loess thickness increases on surfaces of increasing age. Westland loess is a massive, impermeable, silt to fine-sandy loam-textured sediment ranging in colour from blue-grey to yellowish-brown depending on soil drainage. Quartz dominates the silt and sand fraction with clay content normally less than 30% (Mew *et al.*, 1983; Thomas and Lee, 1983). In south Westland, loess has been reported as far south as the Waiho River (Fig. 1)(Young, 1967; Bruce *et al.*, 1973) although Stevens (1968a) argued that the silty surficial soil horizons identified as loess can form by pedogenic processes and that little or no loess was produced because of the extreme rainfall, low winds and limited source areas within the narrow valleys in south Westland. Silt loam-textured soil horizons can form from weathering of a sandy and gravelly parent material within 20 ka in Westland (Stevens, 1968b; Campbell, 1975; Basher, 1986) but Stevens' (1968a) assertion that loess is not produced in south Westland is incorrect. Firstly, from my own observations, loess still blows from the channel zones of braided rivers during south-easterly weather, after a few days without rain, which are needed before silty sediments are dry enough to be mobilised by the wind. Secondly, as discussed below, the narrow moraine-bounded valleys have not been the only loess sources.

The occurrence of Aokautere Ash, a 22 600 BP (radiocarbon years) tephra (Wilson *et al.*, 1988) from the Taupo Volcanic Zone (Fig. 1 inset), confirms the origin of fine-textured coverbeds in which it occurs (Robertson and Mew, 1982; Mew *et al.*, 1983; Mew *et al.*, 1986). Although Mew *et al.* (1988) recognised layers with different sand/silt ratios in loess on terraces between the Taramakau and Hokitika Rivers, a loess stratigraphy for Westland has proved elusive. This is because:

1. loess is poorly preserved in Westland's high rainfall climate as a result of erosion and acid dissolution under extreme leaching. Campbell (1975) showed that soil material making up the E horizon of an Alaquod in Westland was a residuum representing as little as 70% of the original parent material.
2. ambiguities exist in the recognition of buried soils within the loess. In aquic sites, organic rich subsurface horizons may form through burial of former A/O horizons or through illuviation of organic compounds by processes of podzolisation (Mew *et al.*, 1988). In moderately well drained sites, loess sections show alternating pale (E horizon) and olive-yellow (Bs horizon) zones. Young (1967) discounted these as evidence of buried soils because of a lack of lateral continuity along exposures. In thin loess sheets, buried soils become subsumed by developing surface soils.
3. radiocarbon dates of peat and wood are unreliable owing to contamination from leached modern carbon (Hammond *et al.*, 1991).
4. Aokautere Ash within loess normally occurs as disseminated glass grains rather than as a macroscopic layer and hence it has limited value as a chronohorizon (Mew *et al.*, 1988).

2. Soils and landforms of Saltwater Forest

This study was centred on Saltwater Forest, 10 km to the west of Harihari in south Westland but also included Ianthe and Wanganui forests to the north and Okarito Forest to the south (Fig. 1). All four forests cover Pleistocene and Holocene landforms. Rainfall increases from about 2 400 mm yr^{-1} at the coast to 7 000 mm yr^{-1} at the Alpine Fault (Griffiths and McSaveney, 1983). According to the USDA Soil Taxonomy (Soil Survey Staff, 1992), by which soils are provisionally classified throughout this paper, soils on late Holocene landforms are Entisols or Histosols whereas on older surfaces Entisols have evolved to Inceptisols or

Spodosols. Soils commonly have perched water tables; the depth of perching is dependent upon land surface slope.

Saltwater Forest is a lowland podocarp forest covering 9 500 ha bordering the Tasman Sea between the Poerua and Whataroa Rivers (Fig. 2). The principle landforms are glacial in origin. Moraines are designated M6, M5 etc, in order of increasing age and their associated outwash terraces are correspondingly designated T6, T5 etc. Relative ages were assigned to moraines using the principle that moraines become progressively younger toward the Southern

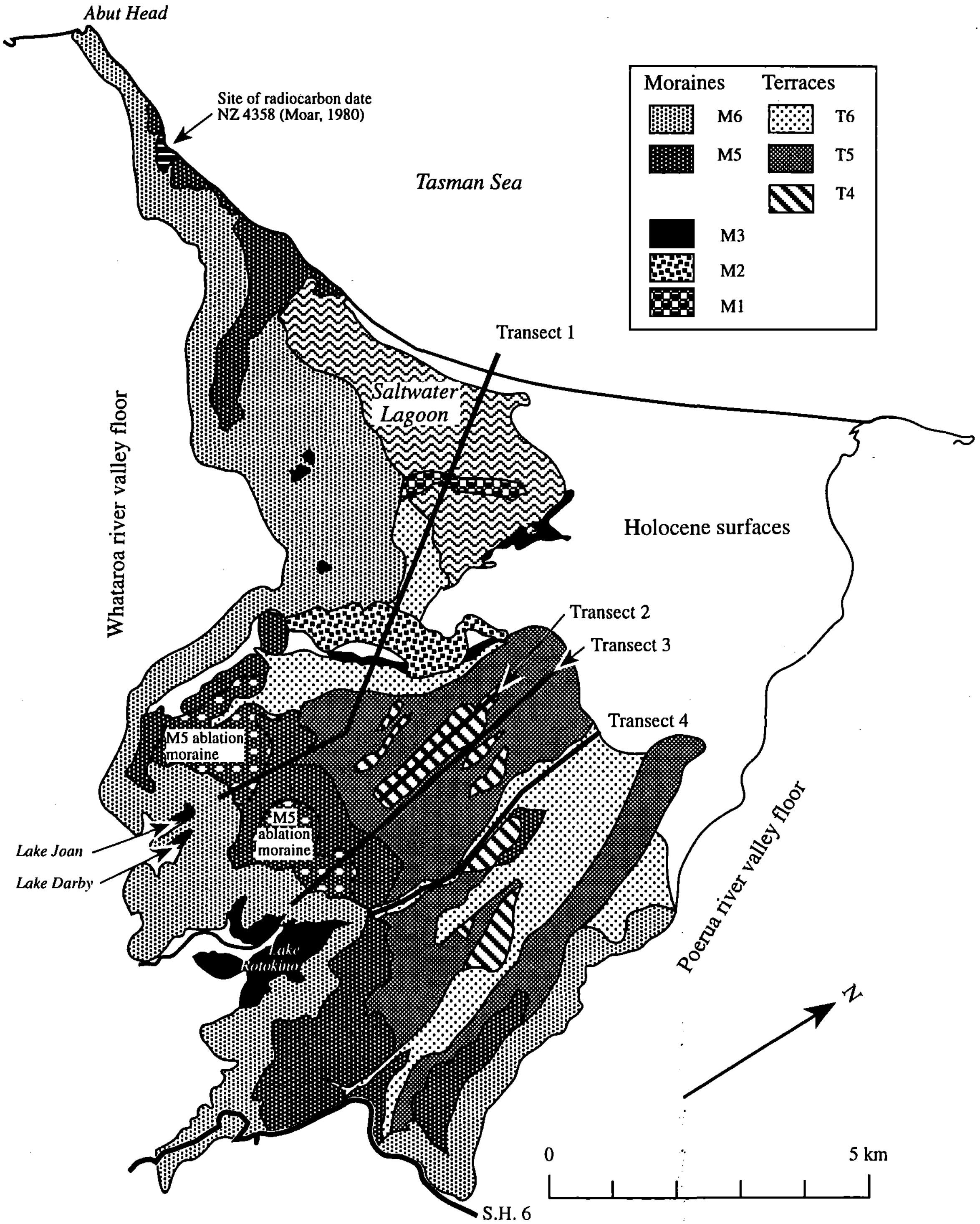

Fig. 2. Landforms of Saltwater Forest.

Alps since early moraines within the range of a later glacial advance are obliterated. The three oldest moraines (M1, M2 and M3) lie within Saltwater Forest, subparallel to the coastline. M1 is a low relief moraine (maximum elevation ≤50 m ASL) in Saltwater Lagoon. M2 is a large, steep moraine reaching a maximum elevation of 190 m. The glacier that formed M3 flowed around M2 to leave sections of moraine at the base of M2 and along the eastern shore of Saltwater Lagoon. Henceforth the codes M1, M2, etc are used to name a moraine and the advance that produced it. Outwash terraces associated with M1, M2 and M3 were not found and are presumed to be buried. The remaining area of Saltwater Forest comprises three Pleistocene outwash terraces (T4, T5 and T6), younger moraines (M5 and M6) and the Holocene swamps, floodplains and dunes surrounding Saltwater lagoon (Fig. 2). T4 and T5 grade from a terminal moraine complex within Saltwater Forest. The younger terrace (T5) and the moraine (M5) must have formed at the same time. No moraine correlative of the older terrace (T4) is preserved in Saltwater Forest but it may be buried beneath M5. T4 exists as finger-like remnants of a previously extensive terrace that was dissected by meltwater from the M5 advance. The remnants are truncated to the north, along a line south-east to north-west, presumably by channel widening of an ancestral Poerua River which flowed through what is now Saltwater Lagoon. To the south, they are buried by the M5 terminal moraine. M5 and younger M6 lateral moraines form the boundary of Saltwater Forest along the margins of the Whataroa and Poerua valleys. This pair of moraines extend to, and are truncated at, the coast in the Whataroa valley but they do not extend so far down the Poerua valley. Ice lobes of the Whataroa glacier extended into Saltwater Forest in the south, during the slightly larger M5 advance to form the M5 terminal moraine complex. M6 terminal moraines rise abruptly above Lakes Rotokino, Joan and Darby on the southern boundary of Saltwater Forest. M5 ablation moraine lies between the M5 and M6 terminal moraines in some areas. T6 grades from M6 and comprises degradational terraces and relict channels formed when meltwater rivers from the M6 glacier terminus incised into older till and outwash gravels. To the north-west, T5 and T6 are buried by Holocene fine-textured floodplain deposits and peat.

Fine-textured coverbeds occur on M1, M2, M3 and T4 and have been identified as loess on the following criteria:

1. they are silt or silt loam in texture
2. the older the surface, the thicker the cover
3. the older the surface, the greater the number of buried soils within the cover
4. they occur on surfaces above the range of alluvial deposition
5. they include volcanic glass shards
6. loess production in the past has modern analogues.

In general, loess is absent or very thin on M5, M6, T5 and T6, except where it is over-thickened on the slopes of moraines bounding the river valleys.

3. Aokautere Ash

Volcanic glass shards have been found in all analysed soils formed in loess. Their identification as Aokautere Ash is supported by elemental glass chemistry in this study and others from north Westland (Mew *et al.*, 1986). Soils were sampled in consecutive 5 or 10 cm increments and glass grains were counted under a polarising microscope from sub-samples of sand (>63μm) suspended in clove oil. Grains per slide were counted because extremely low glass concentrations made grains per thousand counts impractical. In loess profiles, shards are usually restricted to the upper 50 cm with maximum counts between 30 and 40 cm although in some profiles glass grains were found as deep as 80 cm. No glass grains have been found in soils on M5, M6, T5, or T6. At a site immediately in front of the M5 terminal moraine, a soil has been buried sequentially by lacustrine sediments then outwash gravels from the M5 advance. Glass shards are concentrated in the upper 10 cm of the buried A horizon. Where the same geomorphic surface is not buried, glass is mixed through 40 cm of overlying loess indicating that progressively increasing dispersal of glass has taken place during the last ca. 22 ka of loess deposition and soil development. Glass shards can be mixed upward in an aggrading soil profile by soil faunal activity or tree overturn, downward by soil movement into root voids or eluviation with macro-pore flow. At Blue Spur in north Westland (Fig. 1), glass grains were found at a depth of 1.8 m in a buried soil identified from pollen analysis as having formed during the Last Interglacial (N.T. Moar, pers. comm., 1994)(Fig. 3)[1]. Although the grains were not analysed in any way, they are presumed to be Aokautere Ash since no other tephra dating from 200 ka (the upper age limit of the Waimea Formation) have been reported in the South Island. Maximum glass grain counts occurred in the loess above this soil at 60 cm depth. The glass grains probably moved downward by macropore flow since the grains were found in a coating of clean sand grains around a macropore. In light of these observations, I used the maximum concentration datum in preference to the first appearance datum when establishing a chronohorizon by glass counting. Using this criterion, between 30 and 40 cm of loess has accumulated in the last ca. 22 ka in Saltwater Forest compared to 1 m or more in many North Island or east coast South Island sites. Assuming that loess production is climatically controlled and hence most of the loess fell prior to 10 ka (Eden and Froggatt, 1988), respective average loess accretion rates are ca. 30 mm ka^{-1} and 80 mm ka^{-1} or more. Most valleys in south Westland were often filled with ice during the Pleistocene, so loess sources must have included the coastal plain exposed by sea level lowering, drift on the surface

[1] A section 750 m to the south on the same terrace at Blue Spur was described by Moar and Suggate (1973). The stratigraphy of the two sections are difficult to correlate and Aokautere Ash was not identified in the original section. After subsequent reinterpretations (Moar, 1984; N.T. Moar and R.P. Suggate, pers. comm., 1995) the lower part of the section from 95 to 170 cm is, like the lower part of the section described here, regarded to be of Last (Kaihinu) Interglacial age.

of the glaciers themselves or ice free areas between glaciers. With approximately 120 m sea level lowering at the Last Glacial Maximum (Bloom *et al.*, 1974), as much as 20 km of now submerged coastal plain was exposed near the study area (New Zealand Navy, 1981). Periods of strong onshore (westerly) winds, particularly in times of lower rainfall would have favoured the mobilisation and transport of silt from the coastal plain.

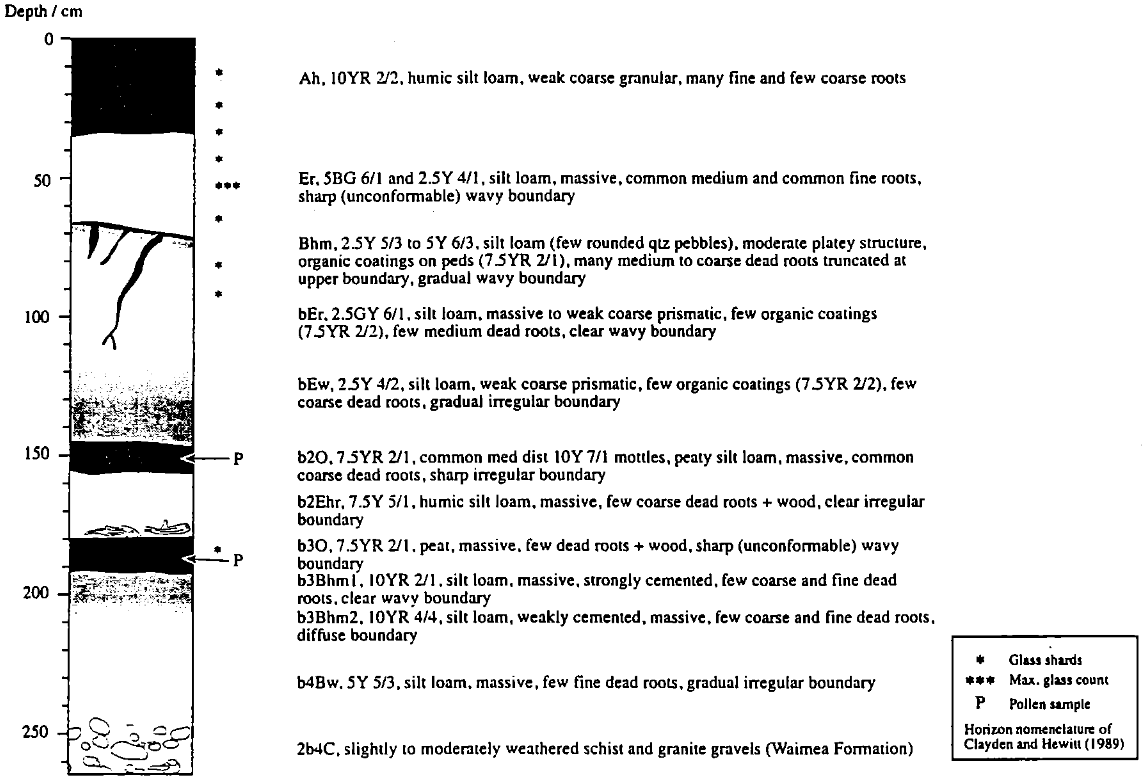

Fig. 3. Section near the edge of Waimean age terrace at Blue Spur, north Westland.

4. Recognition of buried soils and loess sheets

In south Westland where loess accretion rates are low and soil formation rates are high (Stevens, 1968b; Smith and Lee, 1984; Sowden, 1986), the whole of a loess column is strongly altered by pedogenesis so that soil stratigraphic units and loess sheets are coincident in depth and time. The sediment is most likely to be mineralogically and texturally different from the primary loess (Campbell, 1975) but the degree of alteration differs between soil environments (Mew *et al.*, 1983). Therefore, in the following discussion I have classified loess according to the type of soil now formed in it.

4.1. Terraces

Soils on terraces are Alaquods or Hemists that range from poorly to extremely poorly drained. Recognition of buried soils in loess beneath terraces is aided by the preservation of buried A or O horizons in the prevailing aquic moisture regime. These horizons were distinguished from pedogenic Bh horizons by the presence of pollen from species rare in the modern podocarp forest, and the stratigraphic consistency of the horizons. Over time these horizons have developed chemical properties typical of Bh horizons. On terraces in Saltwater Forest, organic matter accumulation rates can equal or exceed loess accretion rates; one metre of peat above the loess beneath T4 is common. This depth of peat equates to an average rate of peat accumulation exceeding 40 mm ka^{-1} and consequently the stratigraphy must be interpreted as much in terms of changes in rates of organic matter accumulation as in changes in rates of loess accretion. Rates of organic matter accumulation are determined by the balance between rates of organic matter input and organic matter decay. The former is linked with plant production and has a climatic control. The latter is influenced very strongly both by site drainage conditions and climate. Changes in the balance between the rates of loess accretion and net organic matter accumulation are reflected by changes in the colour and texture of the loess. Pollen analysis of loess sections has shown that peaty loess forms in relatively warm climates and low organic matter content loess accumulates during periods of cool climate (N.T. Moar, unpublished information). Figure 4 is a model which shows the effect of different rates of organic matter accumulation on loess colour and texture relevant to sites T5/1, T4/2 and T4/3 (Fig. 5), assuming that at all sites, the rate of loess accretion at any given time is the same. Climatically controlled trends in organic matter accumulation rate are similar between sites but absolute rates vary according to site conditions. As the average rate of organic matter accumulation increases from T4/3 through T4/2 to T5/1, the resolution of the stratigraphy within the loess declines. The erosion break beneath the upper blue-grey loess at T4/3 was observed in other sites in Saltwater Forest and can be seen at a depth of 65 cm in the Blue Spur section (Fig. 3). It is not clear whether the olive-grey colour in the loess beneath the erosion break is caused by illuviation from above or is a relict feature from the eroded soil.

4.2. Moraines

On moraine crests, soils are moderately well drained Placorthods or Haplorthods. In this soil environment organic matter in buried soil surface horizons is mineralised resulting in a horizon indistinguishable from the immediately underlying E horizon. Thus buried soils were recognised by repeating E/Bs horizon pairs. Ambiguity in the recognition of buried soils arises from the possibility of the pale subsurface horizons, recognised as buried A/E horizons, forming by subsurface gleying. Major element trends for loess profiles do not resolve this dilemma. Marked trends of enrichment or depletion from A and E to B horizons are obvious in surface soils but are much subdued or absent in buried soils. This contrasts to major element trends in multiple loess sheets of the eastern South Island where repeating depth profile trends of P, K, Ca and Mn correspond to morphologically identified buried soils (Runge *et al.*, 1974; Childs, 1975). The lack of correlation between major element trends and buried soils in south Westland may be due to diagenesis of buried soils. Evidence from soils in Wanganui Forest, undoubtedly buried in the Holocene by debris flows (K. Hewson, unpublished data), shows that buried A/E horizons within 2 m of the soil surface receive illuvial Al and Si compounds soluble in ammonium oxalate. In slowly accreting loess, illuviation may act to resaturate horizons depleted during their time as surface soils.

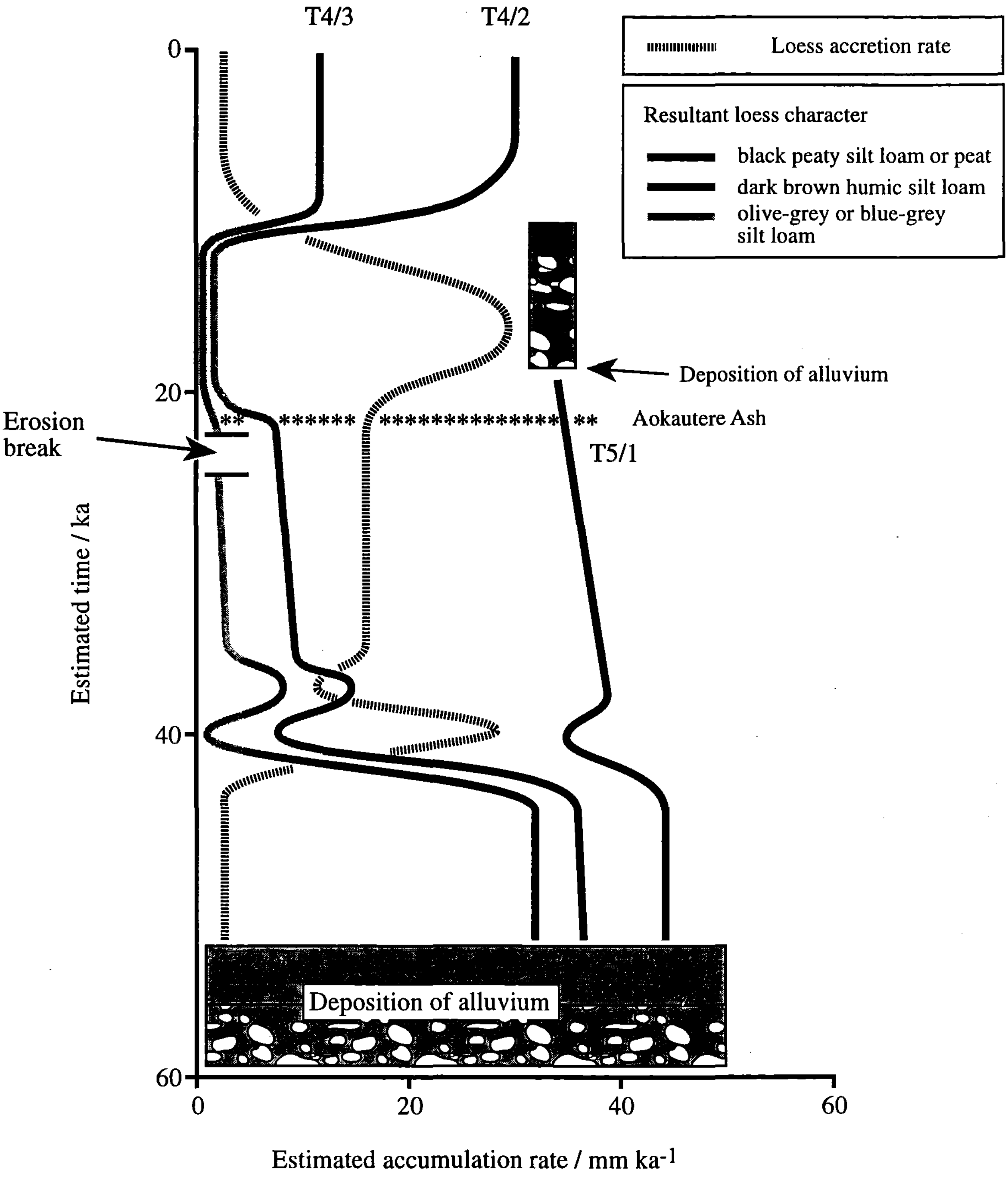

Fig. 4. Schematic model showing the relationship between loess accretion, net organic matter accumulation and loess texture and colour for sites T5/1, T4/2 and T4/3.

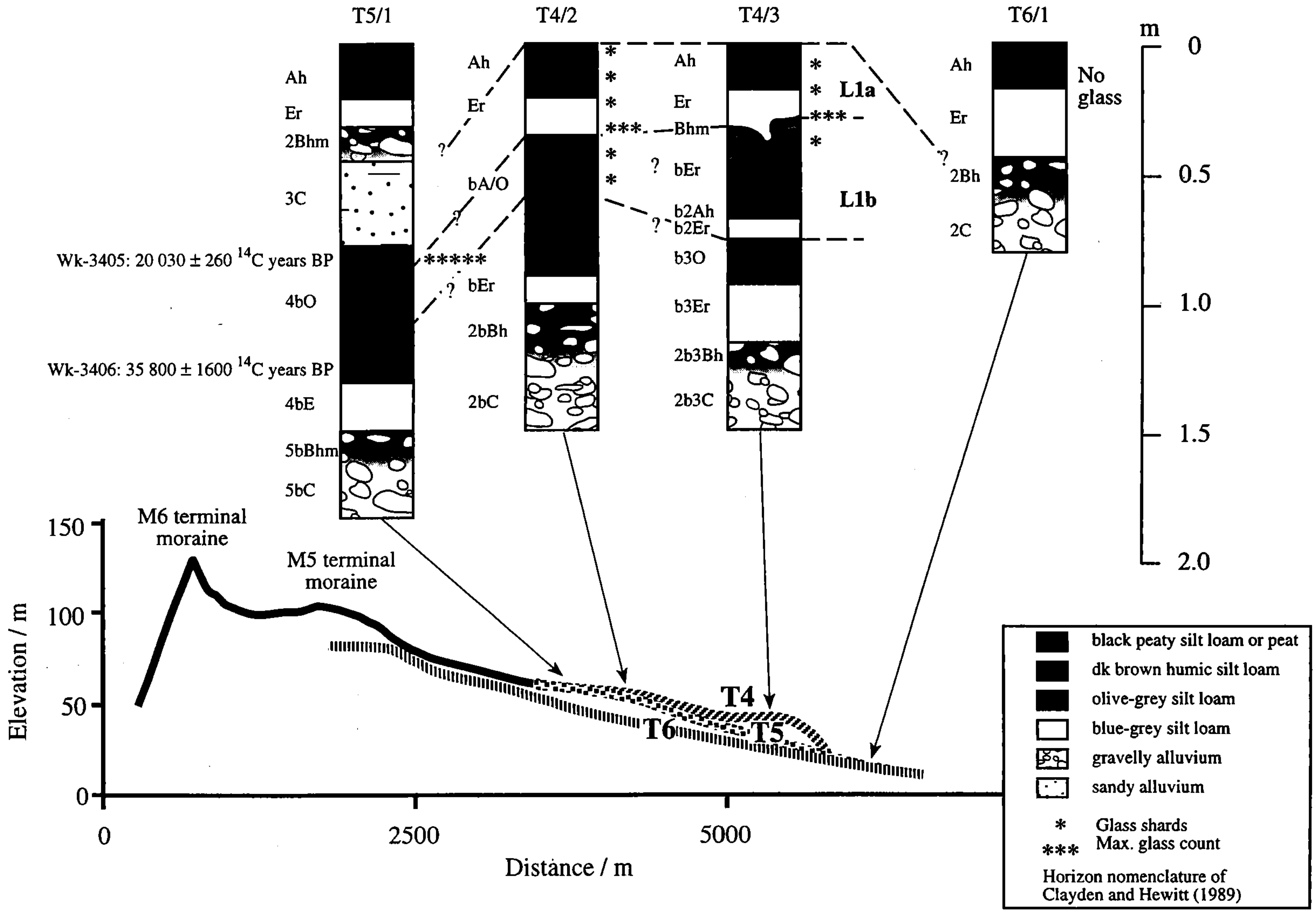

Fig. 5. Soil and loess stratigraphy for terraces along transects 2, 3 and 4 in Saltwater Forest (see Fig. 2).

Stratigraphic detail and interpretation

5.1. Loess and soil stratigraphy on terraces

Beneath T4 up to 80 cm of loess covers an obvious buried soil (cf. T4/3, b3O and b3Er horizons, Fig. 5). The loess varies from a blue-grey silt loam through an olive-grey silt loam to a brown humic silt loam to a black peaty silt loam or peat. The particle size distribution of the horizons above the buried soil at T4/3 fall within a narrow envelope (Alaquod loess, Fig. 6a). The upper part of the buried soil (T4/3, b3O 73-84 cm) also falls within this envelope but beneath this, particle size distribution curves show a progressive increase in sand content and fall outside the Alaquod loess envelope. From this I infer that the buried soil formed in fine-textured alluvium grading upwards into loess. On T5 and T6, silty textured soils have a particle size distribution which generally falls outside the Alaquod loess envelope, having a higher clay and sand content (Fig. 6b) and hence these soils are also interpreted to have formed in fine-textured alluvium with or without a loess component. Maximum numbers of glass grains were counted at about 35 cm depth in T4/2 and T4/3 within a blue-grey loess layer. At T4/3 and other sites, this blue-grey loess layer rests on an erosional surface cut into an underlying olive-grey loess layer. At T5/1, which is near the M5 terminal moraine, T4 is buried by alluvium deposited in the T4/T5 interval. Here, Aokautere Ash is concentrated in the upper 10 cm of the buried soil (4bO).

The deposition of the gravels beneath T4 was followed by a period of declining loess accretion rate and increasing organic matter accumulation culminating in the formation of a peaty soil across much of T4. After this period until 22.6 ka loess accretion rate was greater, while organic matter production continued at moderate levels, punctuated by short periods when loess accretion slowed and organic matter accumulation was relatively high. Around 22.6 ka loess accretion rate increased and organic matter accumulation declined markedly. In north Westland, Aokautere Ash appears macroscopically in an O horizon buried by sands at Graham Terrace (Mew *et al.*, 1986; Hammond, 1989) and in an A horizon buried by loess at Kapitea Reservoir (Fig. 1). In Westland it appears that a major pulse of loess production began after 22.6 ka. This contrasts with eastern South Island loess, where Aokautere Ash occurs 50 cm to 1 m above the base of the upper loess sheet (Eden and Froggatt, 1988). I infer that the difference is due to differences in loess source area over time, related to the size of the piedmont and the proximity of glacier termini to the coastline. In south Westland, loess source areas would have increased markedly after 22 ka, during the time of maximum sea-level depression (the Last Glacial Maximum) whereas in eastern South Island, the loess source areas increased before 25 ka B.P. (Eden and Froggatt, 1988) when aggradation began on large piedmont fans.

Most of the loess beneath T4 is included in one loess sheet (L1), which is subdivided into 2

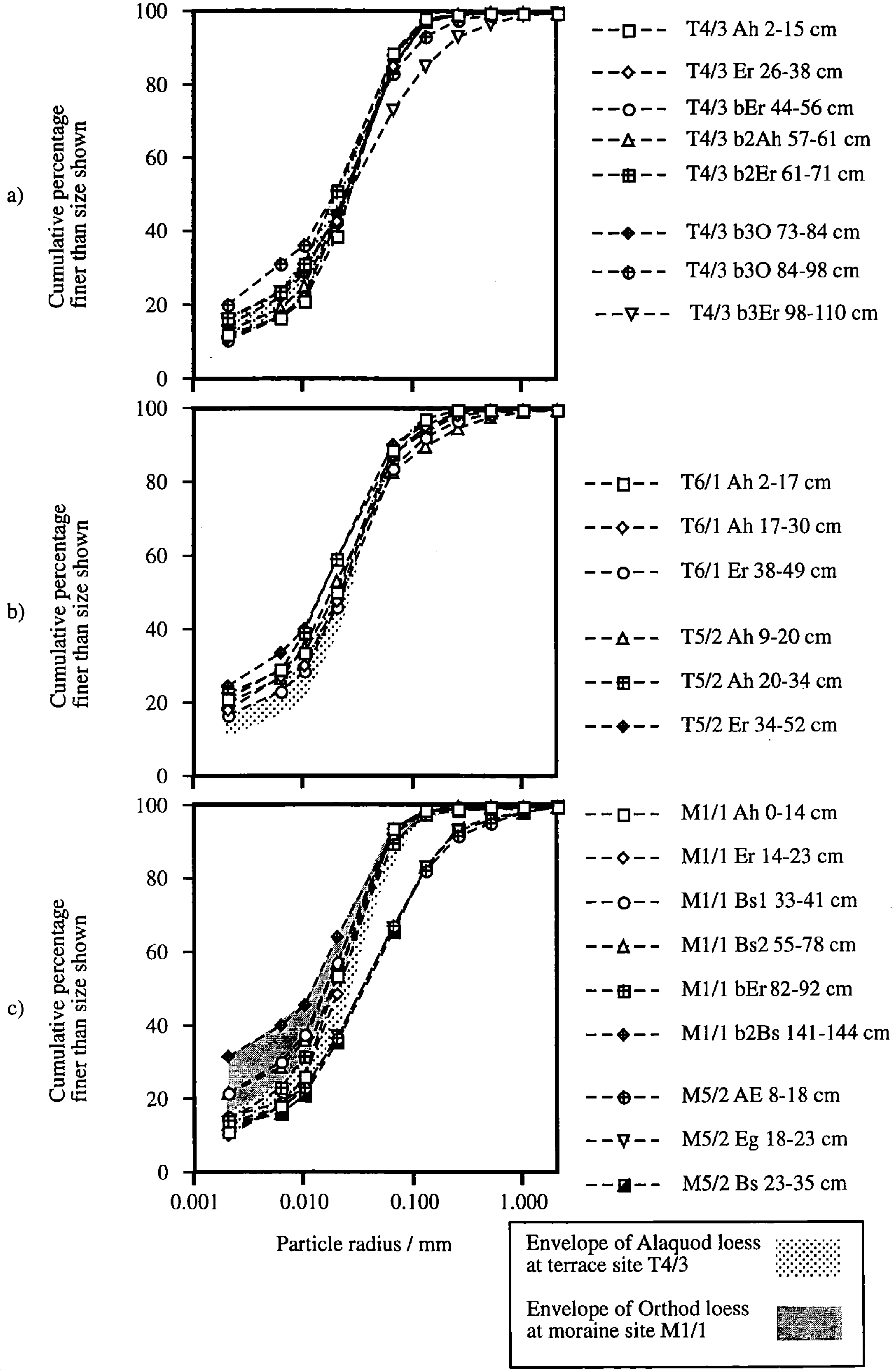

Fig. 6. Cumulative particle size distribution for soils in Saltwater Forest: (a) T4/3. (b) T6/1 and T5/2. (c) M1/1 and M5/2.

units, 1a and 1b (Fig. 5). Unit 1a post-dates the Aokautere Ash and accumulated during the last stadial of the Last Glaciation (oxygen isotope stage 2). Unit 1b accumulated in a longer period of slow loess accretion presumably in a milder climate during oxygen isotope stage 3. The loess which contributes to the soil beneath L1b is included in loess sheet 2 (L2). No moraine in Saltwater Forest was found with only loess sheet 1, and an M4 is inferred to be missing, hence the gap in numbering of moraines from M3 to M5; the former has more loess than T4 and the latter has none.

5.2. Loess and soil stratigraphy on moraines

Pits were dug and described on moraine crests because these are relatively stable sites free from loess over-thickening by colluvial additions. The greatest thickness of loess occurs on M1 where 3 m of loess incorporating 4 buried soils has been found (Fig. 7). In all loess sections on moraines, the lowermost buried soil formed partly in loess and partly in the underlying till. On M2 the maximum loess thickness found was 1.5 m with 2 buried soils. This is a very steep moraine and it is likely that some loess has been lost even from ridge crest positions. M3 is mantled by up to 2 m of loess containing 3 buried soils. The particle size envelope for Orthod loess from M1/1 is wider than the Alaquod loess envelope (Fig. 6c) because of higher clay contents particularly in buried Bs horizons. On M5 and M6, soils have formed directly in till although loess may contribute to the fine earth fraction of the surface horizons (Fig. 6c).

On M1 and M3, loess sheet 1 extends downwards to the top of the first obvious buried soil between 70 and 80 cm depth (Fig. 7). In this soil environment, the distinction between loess units 1a and 1b is extremely subtle but by analogy with the loess beneath T4, the base of loess unit 1a and the zone of maximum glass counts should be coincident at about 40 cm depth. Glass counts are in accord with this. On both moraines, the top of a second buried soil is at 110 cm making loess sheet 2 about 40 cm thick. Some of this loess probably blew from the aggrading T4 terrace and its regional correlatives. Loess sheet 3 extends from 110 cm to the top of a third buried soil at 145 cm depth, a thickness once again of about 40 cm. Loess sheet 4 on M3 lies on the moraine and the soil formed in it extends into the till. On M1 this loess sheet extends to 210 cm and is underlain by a fourth buried soil in loess sheet 5 which grades into till. On M2 the loess record is incomplete and interpretation is difficult.

5.3. Summary of the loess stratigraphy

A maximum of 5 loess sheets has been recognised, all on the oldest moraine, M1. Loess sheets 1 to 4 occur on M3 but part of this record is missing on the older and much steeper M2. Loess sheet 1, subdivided into units 1a and 1b, lies beneath T4 above a buried soil formed in

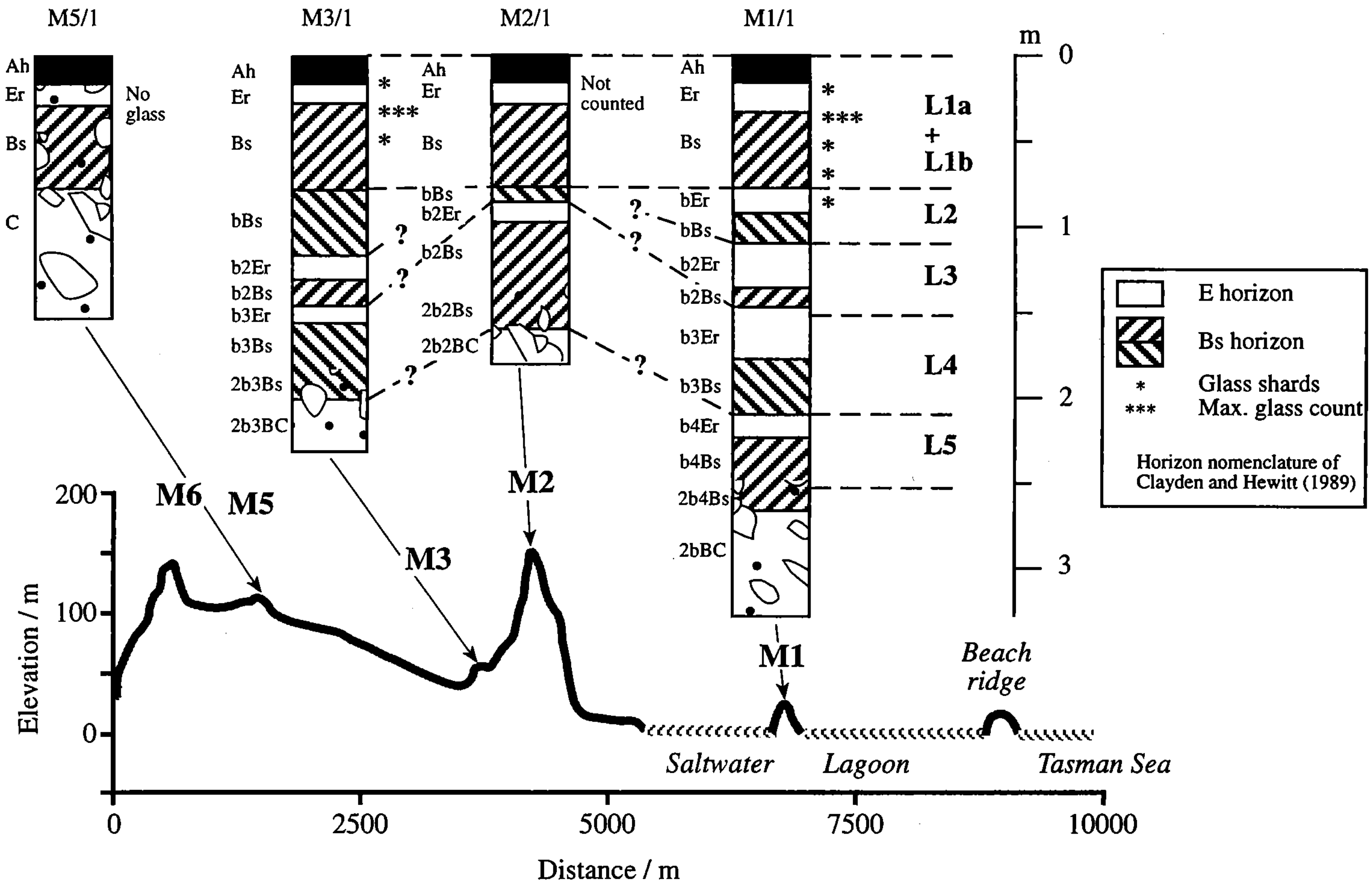

Fig. 7. Soil and loess stratigraphy for moraines along transect 1 in Saltwater Forest (see Fig. 2).

fine-textured alluvium grading up into loess, above aggradation gravels.

6. Sequences in adjoining catchments

Figure 8 shows the loess stratigraphy on moraines in Ianthe, Wanganui and Okarito Forests with correlation to the moraines of Saltwater Forest where appropriate. Lateral moraine pairs bounding the river valleys that have little or no loess mantle were correlated with M5 or M6 depending on their proximity to the river valley. Based on the loess stratigraphy, a correlative of the "missing" moraine in Saltwater Forest (M4), occurs in Ianthe, Wanganui and Okarito Forests. Large, steep moraines in Wanganui Forest have, in places, a loess cover which includes 4 loess sheets, like M3 in Saltwater Forest. Despite this, these moraines have been correlated to the older M2 on the basis of similarity of morphometry, although I suspect that there is little difference in the ages of M2 and M3. Where examined, the depths of maximum glass counts in all loess profiles were consistent with those in Saltwater Forest.

7. Correlation to glacial stratigraphy and oxygen isotope record

Figure 9 shows correlation of periods of time represented by loess, soils and glacial deposits in south Westland to the glacial chronology for north Westland (Suggate, 1990). M5 and T5 are correlated to the Kumara-2_2 advance during oxygen isotope stage 2. According to Suggate (1990), the Kumara-2_2 advance began "somewhat before 22.3 ka and culminated at ca. 18 ka BP or a little later". This corresponds well to the timing of the M5 advance which post-dates the eruption of Aokautere Ash. A radiocarbon date of the upper 10 cm of the 4bO horizon at T5/1 (Fig. 5) indicates that the overlying outwash gravels may have begun to accumulate after 20 ka, although the radiocarbon age may still be too low owing to contamination, despite a pretreatment of the peat by conc. HNO_3 hydrolysis (Hammond *et al.*, 1991). M6 and T6 are correlated with the Kumara-3 advance of north Westland between about 16 ka and 14 ka (Suggate, 1990). In fact the upper age limit for the Kumara-3 advance is constrained by a radiocarbon date from near Abut Head in Saltwater Forest (Moar, 1980) (Fig. 2). At this site, peaty sediments on till are buried by lacustrine sediments in a lake that was impounded between M6 and M5. Coastal erosion has removed parts of M5 and exposed the section. The radiocarbon age of (NZ 4358) 16 450 ± 200 years BP (untreated sample) is a maximum age for the M6 advance and a minimum age for the M5 advance. Warren (1967) mapped M5 and M6 together as Moana Formation, which is the glacial formation associated with the Kumara-3 advance. He probably did this because the Kumara-3 moraine in north Westland is a double moraine (Suggate, 1965) similar to the M5 and M6 pair. However, his interpretation can no longer be supported.

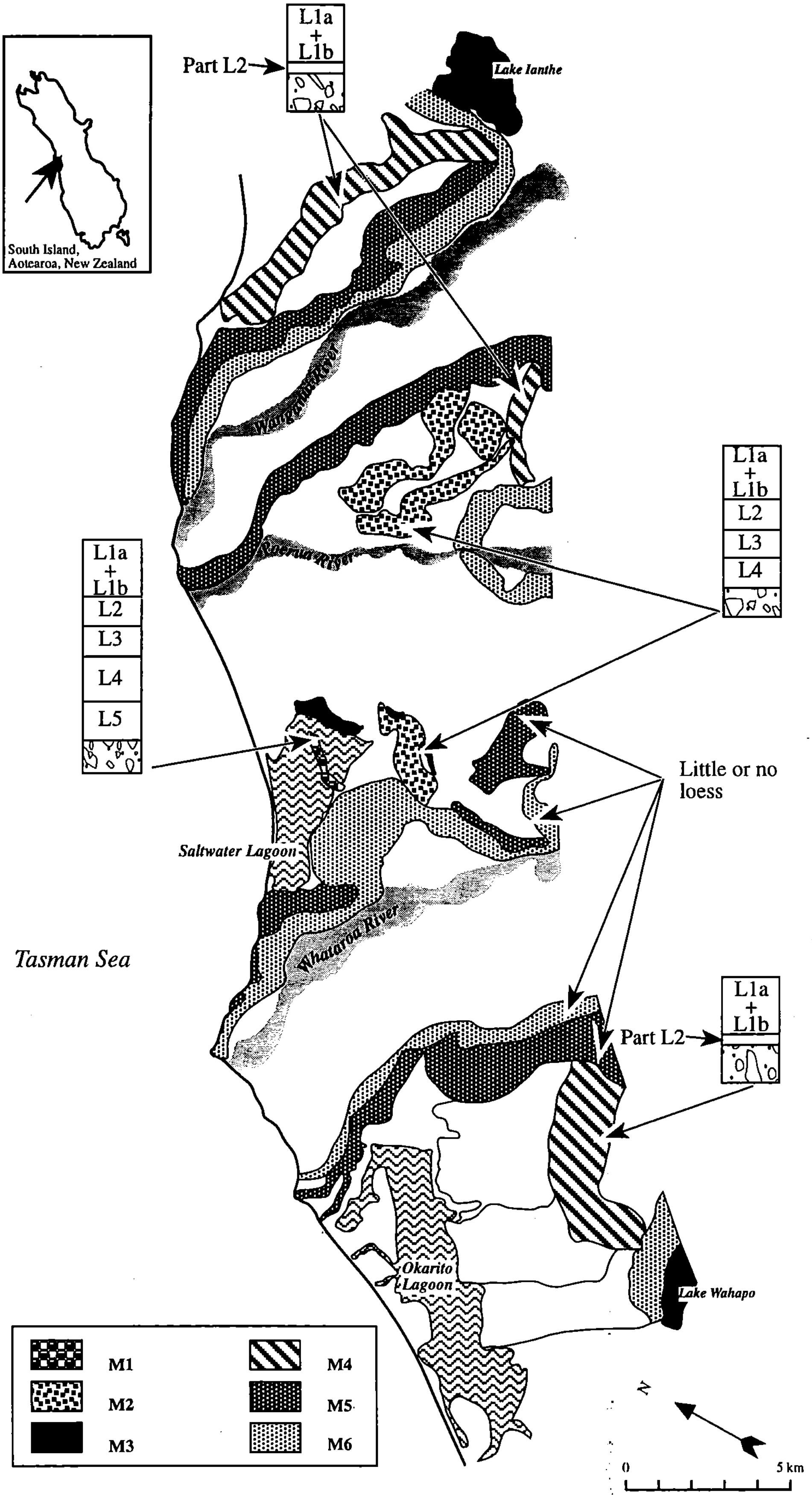

Fig. 8. Correlation of moraines from Saltwater Forest to Okarito, Wanganui and Ianthe forests based on soil and loess stratigraphy.

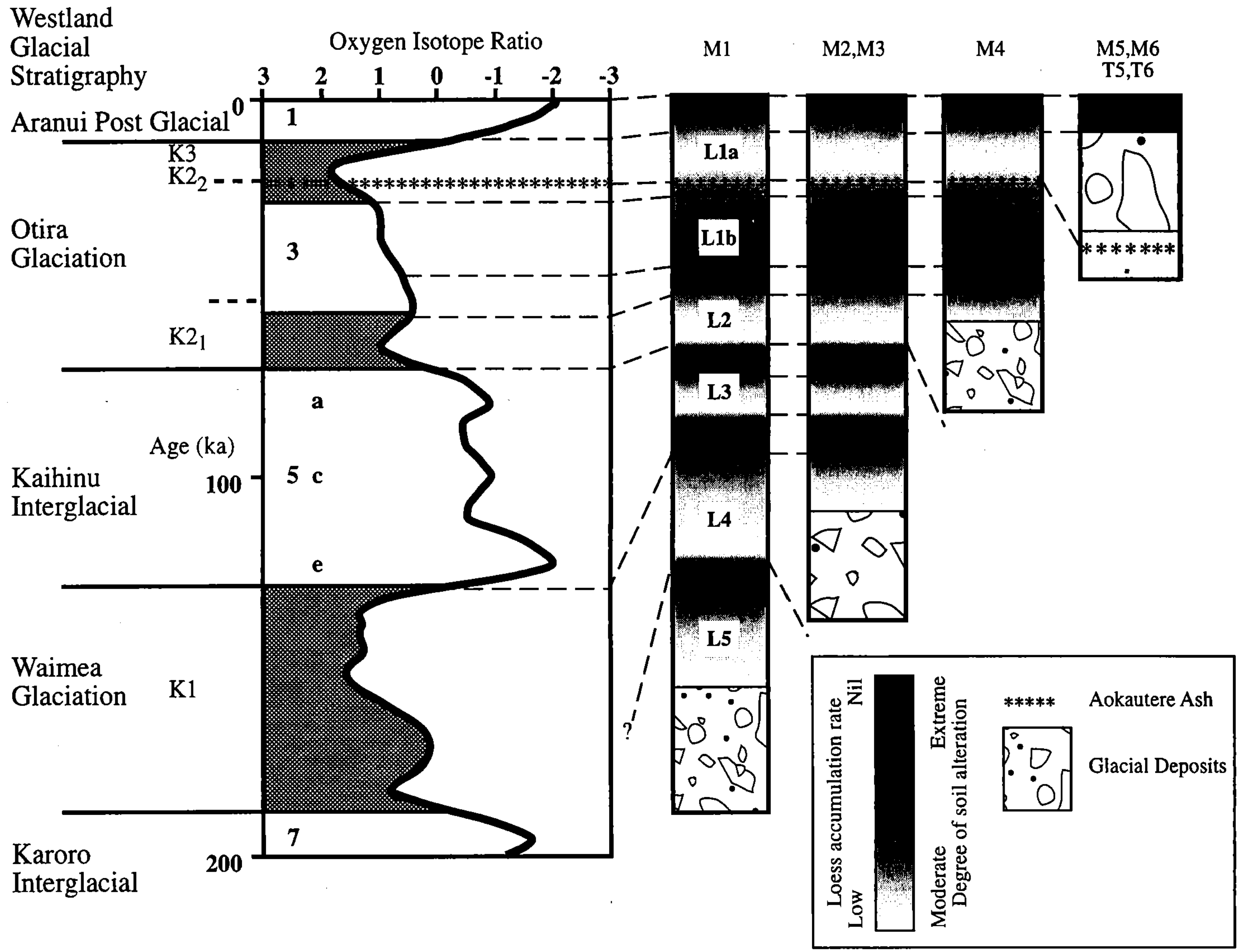

Fig. 9. Correlation of periods of time represented by soils, loess and glacial deposits in south Westland to the Westland glacial chronology and the oxygen isotope record (after Imbrie *et al.*, 1984).

M4 and associated outwash terraces are correlated to the Kumara-2_1 advance of north Westland, in the first stadial of the Last Glaciation (oxygen isotope stage 4). The buried soil beneath loess sheet 1b formed in glacial deposits covered by part of loess sheet 2 on M4/T4 and in loess sheet 2 on older surfaces, during the last interstadial (oxygen isotope stage 3). Pollen analyses of samples from T5/1 show a warming from the basal gravels (5bBhm) to the base of the buried peat (4bO) with cooling thereafter to the top of the 4bO horizon immediately above the zone of maximum counts of Aokautere Ash (N.T. Moar pers. comm., 1994). A treated sample from the lower 10 cm of the 4bO returned an age of (Wk-3406) 35 800 ± 1600 BP (Fig. 5). I interpret this as effectively a "beyond radiocarbon" age since even the slightest contamination by young carbon will significantly alter radiocarbon ages of samples which are older than 40 000 years (Hammond *et al.*, 1991).

M3 is assigned to the Waimea glaciation on the basis of the thickness of loess and the number of buried soils. M2 is considered to be only slightly older and is also assigned to the Waimean. As shown in figure 9, loess sheet 4 accumulated in the Waimean and loess sheet 3 accumulated during the Last Interglacial. By this correlation, the soils in loess sheets 3 and 4 formed in the Last Interglacial. Evidence for this interpretation includes:

1. an organic-rich horizon above M2 tills in Wanganui Forest has a pollen spectrum consistent with an interglacial climate (N.T. Moar pers. comm., 1994) and;
2. at the Blue Spur road section (Fig. 3), a 35 cm loess sheet separates 2 organic rich horizons (b2O and b3O) both of which yielded a pollen spectrum typical of the Last Interglacial (N.T. Moar pers. comm., 1994). Beneath the b3O, an eroded 50 cm thick loess sheet (b4Bw horizon) rests on Waimean outwash gravels.

It is not clear yet, as to the correlation of loess sheet 5 and M1 on which it rests, nor is it known whether the soil in loess sheet 5 formed in an interglacial or interstadial; however all are of at least Waimean age.

8. Tectonics

Nathan *et al.* (1986 p. 74) concluded that the block to the west of the South Westland Fault Zone (the Western Platform) (Fig. 1) which includes the study area, has been subsiding since the Middle Miocene. The absence of wave-cut platforms on pre-Last Interglacial landforms at the coast is consistent with this conclusion. The nearest elevated Late Pleistocene marine sediments are mapped at Pug Creek and Cement Hill in Waikukupa Forest (Warren, 1967). These sites are about 4 km south of the Waiho River, on the eastern side of the South Westland Fault Zone which has been intermittently uplifted since the Pliocene (Nathan, 1978).

9. Conclusions

Loess is present on Late Pleistocene surfaces near Saltwater Forest in south Westland and continues to be produced from valley floors despite the high rainfall (ca. 2 500 to 7 000 mm yr^{-1} MAR). Aokautere Ash, a 22.6 ka, rhyolitic ash from the Taupo Volcanic Zone, occurs in trace quantities in the upper part of the loess. Counts of glass grains from the sand fraction of loess, consistently show a peak at about 40 cm depth although glass grains are commonly mixed to 80 cm. The depth of maximum glass counts is preferred as a chronohorizon to their first appearance because glass grains can be moved downward by soil movement *en masse* or by eluviation of sand-sized grains. From these data, average loess accretion rate was about 30 mm ka^{-1} during the last stadial of the Last (Otira) Glaciation. The oldest moraine in Saltwater Forest is mantled by 3 m of loess that is divided into 5 loess sheets on the basis of the soil stratigraphy. L1 and L2 are of Last Glacial age, L3 accumulated during the Last (Kaihinu) Interglacial and L4 accumulated during the Penultimate (Waimea) glaciation. L5 is at least of Waimean age. The youngest pair of moraines and associated outwash terraces are not mantled by L1 and post-date the Aokautere Ash. Landforms of intermediate age have progressively more loess sheets as they increase in age except where affected by erosion. The soil and loess stratigraphy provides a means of correlating glacial landforms locally and to the established glacial chronology of north Westland. Correlatives of the Kumara-3, Kumara-2_2, Kumara-2_1 and Kumara-1 advances of north Westland, as well as potentially older landforms, are in south Westland.

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Chapter 3

A reinterpretation of loess, soil stratigraphy and Aokautere ash on late Pleistocene surfaces in Saltwater forest, and a revised correlation with the glacial stratigraphy

Abstract

The published stratigraphy of buried soils and loess sheets on late Pleistocene surfaces in south Westland, which was based on limited chronological data and a "counting-back approach" (Almond, 1996), has been refined by the application of total element, mineralogical and phytolith analysis. Total element and mineralogical analysis were used to indicate the degree of soil alteration of different parts of loess columns. The results were consistent with the existing morphological identification of buried soils. Phytolith analysis allowed reconstruction of broad vegetation changes during loess accretion. However, significant dissolution of phytoliths may have skewed representation of taxa in phytolith spectra. Thermoluminescence dating was also used in an attempt to provide absolute time control but the method gave imprecise dates and inexplicable age reversals. Loess sheets 1, 2, 3 and 4 are now thought to be of Last (Otira) Glacial age. Loess sheet 1 accumulated in latest OIS 3 through OIS 2 (L1a from < 22.6 ka to 16 ka, L1b from < 37 ka to 22.6 ka). L2 accumulated in OIS 3 from 50 to 37 ka. Soils in loess sheets 3 and 4 do not show phytolith spectra, or weathering sufficiently extreme, for them to have formed in the Last Interglacial as previously thought. Loess sheets 3 and 4 are now correlated to OIS 4 and the soils in them to early OIS 3. Only loess sheet 5 is altered to a degree consistent with it having passed through the Last (Kaihinu) Interglacial. Unfortunately, phytoliths, which might have provided corroborating evidence, are extremely rare and in poor condition. L5 is correlated to OIS 6. The correlation of the glacial advances in Saltwater Forest to the central Westland stratigraphy, based on loess and soil stratigraphy is: M6 = K_3, M5 = K_{22}, M4 - no central Westland correlative, M2 and M3 = K_{21}, M1 = K_1.

Keywords: glacial stratigraphy; loess; soil stratigraphy; phytoliths; thermoluminescence dating

1. Introduction

1.1. Background to the study

The soil and loess stratigraphy of late Pleistocene surfaces in south Westland is reported in Almond (1996). However, the correlations and stratigraphic interpretations in that study suffer from a paucity of time control beyond 22.6 ka, the datum provided by Aokautere Ash. Suspicion of radiocarbon dates in this environment (Gillespie et al., 1992; Hammond et al., 1991; Mew et al., 1986; Mew et al., 1988a; Moar and Suggate, 1996) exacerbates the problem.

Beyond 22.6 ka, correlation relied on a 'counting back' approach whereby loess sheets and strongly developed soil features recognised on morphological grounds, were assigned in a sequential fashion from youngest to oldest, to climatic episodes established in the local glacial stratigraphy and to the global oxygen isotope record (Suggate, 1990). Correlations were supported by some pollen data but these came from poorly drained terrace sites which spanned only the latest part of the total loess record. Deeper sections that include the older part of the loess and soil record are found on well drained moraine-crest sites, and here pollens are not preserved.

Subsequent work on the deep loess sections in the study area has necessitated a reassessment of the existing correlations (Almond, 1996). Total element, mineral and phytolith analysis together with thermoluminescence dating, were used in an attempt to test whether :1) the degree of soil alteration; 2) paleovegetation evidence and 3) depositional age were consistent with the existing assignment of buried soils to climatic events.

1.2. General Introduction

1.2.1. Total element and mineral analysis

The extent of alteration of soil horizons can be assessed by the degree of divergence of elemental chemistry or soil mineralogy from some *a priori* initial state. Mobile elements or weatherable minerals are depleted in eluvial horizons and the former may be relatively concentrated in illuvial horizons. Conversely, immobile elements or resistant minerals are relatively concentrated in eluvial horizons and decreased in concentration (diluted) in illuvial horizons. If the elemental composition of the parent material of a soil is known it is possible to calculate losses and gains of elements relative to the initial state or in absolute terms, if the bulk density of soil horizons and parent material are known. These calculations require an internal standard i.e. a soil component which is not altered, lost or gained from soil horizons during pedogenesis. A number of internal standards have been used for mass balance calculations including the minerals zircon, rutile, tourmaline, garnet (Marshall, 1940), quartz, albite and orthoclase (Barshad, 1964) or as a more easily measured surrogate, elements that occur exclusively in a resistant mineral e.g. Zr in zircon, Ti in rutile or anatase or B in tourmaline (Evans, 1978). Evans and Adams (1975) and Akhtyrtsev (1968) used a tri-acid residue of the whole soil.

Quantitative or qualitative assessments of profile development are contingent on pedological alteration exceeding the inherent variability of the parent material. It is necessary therefore to confirm original parent material uniformity. Parent material anisotropy may originate from lithological or textural stratification inherited from bedrock or sedimentary layering and may be assessed by a number of techniques including mineralogical analysis, ratios of resistant minerals or elements in restricted particle size ranges, or ratios of non-clay particle size fractions (Barshad, 1964; Brewer, 1964; Chittleborough and Oades, 1980; Evans, 1978; Evans and Adams, 1975; Rabenhorst and Wilding, 1986). In this study elemental and particle-size ratios have been used.

1.2.2. Phytolith analysis

Phytoliths are particles or grains of biogenic hydrated silica ($SiO_2.nH_2O$), also known as plant opal or opal phytoliths, that develop in the tissue of plants and accumulate in soils (Kondo, 1977) and other depositional environments. Most phytoliths are in the range of 2 to 200 μm, and may be colourless, brown, or black, depending on the amount of occluded carbon (Wilding and Drees, 1974). Under a polarising microscope they appear as high relief grains in common mounting media, as a result of their low refractive index (1.14-1.47) (Piperno, 1988), and are mostly isotropic. They have a low specific gravity in relation to other common soil minerals (1.5-2.3) and this is exploited in their extraction. The silica comprising phytoliths is dominantly amorphous but a continuum of forms from amorphous to crystalline (α quartz and cristobalite) exists (Wilding and Drees, 1974). The more crystalline forms usually occur as biosynthesised, as opposed to diagenetic, inclusions within a groundmass of amorphous opaline silica.

Phytoliths arise from deposition of silica as encrustations on cell walls or as solid plugs in cell lumen and in intercellular cortex. Silica uptake may occur as a passive accession of soluble monosilicic acid in the transpiration stream or as an active metabolic process involving an energy cost (Piperno, 1988). Active uptake implies some metabolic or physiologic utility to silica; this is thought to include enhanced resistance to predation and fungal attack, mechanical support or even metabolic function (Piperno, 1988; Rovner, 1971; Wynn Parry and Smithson, 1964). The variation in phytolith production in plants is influenced by environment but genetic factors are the prime determinant. Monocotyledons, particularly grasses, tend to be more prolific producers of phytoliths than dicotyledons or gymnosperms to the extent that, in the 1960s and 1970s, it was considered that only the monocotyledons warranted any attention in phytolith research (Piperno, 1988, p. 21). However, it has become clear that phytolith production within the monocotyledonae is not uniformly high nor is it uniformly low in other plants; some families of dicotyledons and pteridophytes are now classified as common or abundant phytolith producers (Piperno, 1988, p. 21).

Phytolith morphology is determined by the kind of cell that accumulates silica and its position in the plant. Cells within, or associated with, the leaf epidermis (epidermal, hair and hair base cells) usually completely silicify to give mineralised pseudomorphs of the original cell, whereas cell lumen are usually only partly silicified so that phytolith shape does not correspond to that of the cell. Classification schemes for phytoliths have followed two schools of thought; non-taxonomic classification, which focusses on phytolith shape alone, is based on the premise that phytoliths have little taxonomic significance; taxonomic classification, in contrast, stresses "the correspondence between phytolith shape, the species that produced it, and the evolutionary relationships of that taxon with other plants" (Piperno, 1988). Phytolith nomenclature includes taxonomic terms (panicoid, festucoid, chloridoid, bambusoid, Kondo et al., 1994), morphologic terms (hat-shaped, dumbbell, cross, elongate, Kondo et al., 1994; Piperno, 1988; Twiss et al., 1969; Weatherhead, 1988; Wilding and Drees, 1971; Wilding and Drees, 1974) and generic terms indicating origin in plant tissue (tracheid, anticlinal epidermis, schlerid, stomata, Kondo et al., 1994; Piperno, 1988).

Rovner (1971) lists the following criteria that must be met for a fossil system to be of value for environmental reconstruction. The material must:

1. withstand decomposition
2. exhibit sufficient morphological differences to be of taxonomic significance
3. provide sufficient quantities to reflect the nature of the entire assemblage from which it is derived.

Phytoliths satisfy these criteria to a greater or lesser extent and have been used in archeological reconstruction[1], palaeoecology and botany, paleoclimatology[2], radiocarbon dating[3] and studies of soil development[4]. Generally speaking, phytoliths are a very resilient component of soils and other deposits and have been found in sedimentary rocks of Cenozoic age (Rovner, 1971). The inorganic nature of phytoliths makes them a potentially valuable microfossil in oxidising environments where organic microfossils such as pollen are not preserved. The solubility of silica is low and independent of pH between pH 3 and 9 and therefore most soil and sedimentary environments preserve phytoliths. Other factors which influence phytolith preservation include: phytolith morphology, high surface area phytoliths such as cell wall encrustations, hair cell phytoliths and prickle hairs tend to be less stable than solid silica plugs; content of tannins in soil solution, higher concentrations increase silica solubility; amount of occluded carbon and adsorbed Fe and Al, the more of these impurities the lower the solubility and; the intensity of biological activity and silica cycling, some micro-organisms have been implicated in phytolith dissolution (Piperno, 1988). The intensity of leaching (Piperno, 1988) and perhaps soil solution silica concentration also have a bearing on phytolith preservation. Kondo *et al.* (1994) noted the relative scarcity of phytoliths in loess of South Island of New Zealand compared to that of North Island. This may be due to the

[1] Pearsall, 1978; Pearsall and Trimble, 1984; Piperno, 1988

[2] Jones et al., 1963; Kondo et al., 1994; Piperno, 1988; Reider et al., 1988; Sase, 1981; Sase et al., 1988

[3] Wilding, 1967

[4] Bakeman and Nimlos, 1985; Kondo and Iwasa, 1981; Sase et al., 1990; Takahashi et al., 1994

abundance of siliceous tephra, sourced from the Taupo Volcanic Zone, in much of North Island loess.

The greatest limitation of phytoliths for reconstructive work is the lack of taxonomic significance of phytolith morphology. One species can include many different phytoliths and phytolith shapes are shared across species, genera, families and higher taxonomic categories. The latter problem, known as redundancy, is more severe than for pollens because, unlike pollens, phytoliths do not have distinct physiological functions (Rovner, 1971). This notwithstanding, there are phytolith forms with taxonomic significance. Commonly there are unique associations of phytolith forms with families, subfamilies or tribes, sometimes with genera but rarely with species. A degree of caution is warranted with any taxonomic associations however because some "diagnostic" phytoliths may yet be proved to be more catholic in their relations to taxa (Hart, 1990).

Even for taxa that do not produce unique phytoliths, assemblages and relative abundance of phytolith forms may distinguish some taxa from others. Pearsall (1978) used relative abundance of different short cell phytoliths to distinguish maize (*Zea* sp.) from wild grasses and thus was able to infer the onset of cultivation of maize in lower Central America by phytolith analysis of an archaeological soil. Phytoliths that have no taxonomic discrimination at family level may still be useful if they are restricted to plants with particular growth habits. Schlerid phytoliths for example, which originate in schlerenchyma cells, are generally restricted to arboreal species and are rarely found in herbs or vines.

The last of Rovner's criteria underlines the requirement that fossil phytoliths adequately represent the local or regional vegetation. The quantity of phytoliths produced by plants is obviously a consideration but not the only one. Phytolith content of soils varies from between 0.1 to 0.3% in surface horizons to <0.03% in subsurface horizons (Wilding and Drees, 1971) but generally quantities of soil in the order of grams is adequate to yield enough phytoliths for analysis. More serious problems arise from over and under-representation of taxa. Taxa that do not produce phytoliths will always be unrepresented and their presence or absence will be indeterminate. Systematic errors in vegetation reconstruction introduced by variation in phytolith production between taxa, can be reduced by calibration of phytolith abundance in surface soil horizons with species composition of the overlying vegetation (Piperno, 1988). Similar exercises have been undertaken in pollen analysis. Under-representation can also arise from selective dissolution of different phytolith forms. Wilding and Drees (1974) found that phytoliths of tree origin in the clay fraction of soils in Ohio were 10-15 times more soluble than grass phytoliths of equivalent size. They attributed this to the greater surface area of the tree phytoliths (see also Hart, 1988). Piperno (1988) noted that partially silicified hair cells and epidermis, while being abundant in vegetation, were at best rarely found and were usually absent in tropical soils.

Other considerations in the reconstruction of paleovegetation from phytolith assemblage are the lateral and vertical mobility of phytoliths. Phytoliths accumulating in a soil or sediment may be of local, regional, or inter-regional source. Piperno (1988) showed that phytoliths in soils on flat terrain beneath forest or grassland in Panama moved less than 20 m from their source. In accretionary sediments or accumulative (upbuilding) soils however, the source area of phytoliths depends on the type of depositional environment. Phytoliths in lacustrine or other water laid sediment usually reflect more regional vegetation throughout a catchment. Phytoliths in loess could represent local sources from vegetation growing at the site of deposition or regional sources reflecting the vegetation in the loess source area, or even an inter-regional phytolith flux. Some of the earliest documentation of phytoliths by Darwin (1909, cited in Piperno, 1988) recounts the accumulation of siliceous plant tissue in the sails of *HMS Beagle* while she lay anchored hundreds of miles off the North African coast. The dilemma of the source of phytoliths in loess has not been addressed but it would seem reasonable to expect that the contribution from loess source areas, which by their nature are relatively devoid of vegetation, is less than that of the plants at the site of deposition.

Vertical mobility of phytoliths in a soil or sediment, once deposited, has major significance for phytolith interpretation. If phytoliths were a relatively mobile soil component, assemblages at any given stratigraphic position could be a composite of phytoliths derived from all overlying material and inferences on vegetation composition through time would be meaningless. Piperno (1988) presents evidence and cites other work in support of the relative vertical immobility of phytoliths and concurs with Rovner (1986) that "Vertical movement cannot be ignored, but it is a non-issue warranting no special attention. It is certainly no invalidation of phytolith analysis in archaeology".

The limited history of phytolith studies in New Zealand has been reviewed by Kondo et al. (1994). The early studies documented the morphological and optical properties, and vertical distribution of phytoliths in soils and loess sections (Claridge and Weatherhead, 1966; Fieldes and Weatherhead, 1966; Raeside, 1964; Weatherhead, 1988), although Raeside (1970) studied phytoliths separated from three species of tussock grasses and three species of sedges. He noted phytoliths specific to individual sedge species but concluded that there was little in phytolith morphological variety to separate the species of tussock. It was not until Japanese workers studied Andisols and deep tephra sections in central North Island that phytoliths were used for paleoenvironmental reconstruction and investigations of pedogenesis in New Zealand (see citations in Kondo et al., 1994, p10, most are published in Japanese). These studies compared phytoliths extracted from surface and buried soils with phytoliths extracted from native trees and herbaceous plants but it was not until Kondo et al. (1994) published their book that there was a comprehensive catalogue and classification of phytoliths of New Zealand vegetation. These workers created an extensive reference collection of phytoliths extracted from New Zealand plants with which they reconstructed climate and vegetation history for a number of North Island tephra and loess sites. In a novel application they also documented the diet of the extinct Moa (*Megalapteryx*) from phytoliths in preserved faeces and made other suggestions for the application of phytolith research.

The general taxonomic associations Kondo et al. (1994) reported are as follows:

1. Spherical smooth, verrucose and nodular phytoliths of 5-20 µm in diameter originate from beeches (*Nothofagus* sp.) (other than silver beech (*Nothofagus menziesii)*, tawa (*Beilschmiedia tawa*), rata (*Metrosideros* sp.), and rewarewa (*Knightia excelsa).*
2. Spherical nodular or rosette phytoliths of 25-60 µm are typically found under broad-leaved forest
3. Spherical spinulose phytoliths of about 10 µm originate from nikau palms (*Rhopalostylis sapida)*
4. Chionochloid and chloridoid phytoliths originate from snow tussock (*Chionochloa* sp.), red tussock (*Chionochloa rubra)* and *Cortaderia*
5. Boat-shape, hat-shape and truncated-cone phytoliths originate from *Poa*, *Festuca,* and *Chionochloa,* and panicoid phytoliths originate from Panicoideae, *Rytidosperma, Cortaderia,* and some *Chionochloa*
6. Fan-shape and rectangular tablet-like phytoliths from bulliform cells are derived from *Rytidosperma* sp. of short tussock grasses.

For the purpose of the present study a detailed characterisation of species composition of paleovegetation was not necessary. A key correlation, and one with high uncertainty in the existing stratigraphy, is that of the soils of Last Interglacial age. Pollen evidence has clearly established that Last Interglacial vegetation in Westland was essentially similar to that of today, disregarding some differences in the distribution of beech species (Moar and Suggate, 1996). The few pollen diagrams from south Westland for that period indicate rimu dominated forest (N. Moar unpublished information). A phytolith assemblage indicative of a grassland in a buried soil would, therefore, exclude the possibility of that soil having formed in the Last Interglacial. However the corollary is less informative. The existence of an arboreal-species-dominated phytolith assemblage would exclude full glacial conditions during deposition, but is not sufficient to discriminate between interglacial or interstadial because the nature of vegetation in Westland during the last interstadial (OIS 3) is unknown.

1.2.3. Thermoluminescence dating

Thermoluminescence (TL) dating exploits geochronometers provided by sediment grains acting as natural radiation dosimeters. A portion of the energy from low-level background ionising radiation is stored in charge traps caused by defects and impurities in crystal lattices. The amount of energy stored is proportional to the time since the last emptying (or partial emptying) of the traps. Ionising radiation includes ambient and exogenous alpha, beta, gamma and cosmic radiation. This energy excites electrons from the valence band into the conduction band whereby trapping may occur into energy levels intermediate between the two bands. Heating provides the thermal energy necessary for electrons to escape from traps whence they can recombine with holes at luminescent centres to produce light (Aitken, 1985). Luminescence centres are a particular type of defect usually associated with impurities such as Ag^{2+}, Mn^{2+} or Fe^{2+}; the impurity determines the colour of the luminescence (Aitken, 1985). The life-time of traps at normal soil or earth-surface temperatures varies from less than hours to millions of years (Aitken, 1985). Feldspars and quartz in bulk samples are the most important minerals that contribute to a TL.

Thermoluminescence dating was first used in archaeology for dating pottery. Firing of pottery drains the thermoluminescence of the constituent minerals and "zeros" the thermoluminescence clock. The age or time since firing can be calculated from the quotient of the equivalent dose (or paleodose) responsible for the acquired thermoluminescence and the dose rate (eqn. 1).

$$\text{age} = \frac{\text{equivalent dose}}{\text{dose rate}} \qquad (1)$$

Units are usually grays (1 Gy = 1 Joule/Kg) for the equivalent dose (D_E) and Gy/ka for dose rate (D_R) giving age in kiloannums (ka). Later, thermoluminescence was applied to Quaternary sediments or materials which fall into two broad classes (Berger, 1988a):(1) those for which the clock is zeroed, i.e. heated, authigenic, biogenic, or precipitated material and (2) those for which the TL is only partially reduced to some unknown, non-zero value. The latter includes aeolian or other terrestrial sediment for which partial zeroing of the TL clock is effected by optical bleaching in sunlight (Wintle and Huntley, 1979; Wintle and Huntley, 1980) or, in rare circumstances, shock.

TL dating is performed on fine grain (4-11 μm) or coarse grain (100-300 μm) fractions because they provide convenient relationships between particle size and ionising radiation path-length and therefore simplify the calculation of effective dose-rate. Samples may be poly- or mono-mineralic. The quartz and feldspar mineral groups dominate the thermoluminescence of most pottery or geological polymineral samples although in the latter zircons can be important. Feldspars are by far the most intense emitters being 10-50 times brighter than quartz (Aitken, 1985, p. 181). The measurement of TL involves heating a sample in an oxygen-free environment (to avoid spurious TL, Aitken, 1985) at a constant rate, typically 5-20 °C/s. The TL is measured in units of photon counts $°C^{-1}$ or photon counts s^{-1}, with the use of a photomultiplier tube. TL emission occurs from about 200°C through to about 500°C in the range of 200-800 nm (Berger, 1995). TL emissions are filtered before reaching the photomultiplier to remove wavelengths produced by black-body radiation. The useful wavelengths are in the blue/violet, violet and ultra-violet ranges. The output, graphically presented, is known as a glow curve. The equivalent dose can be determined by a number of methods (discussed below) but all involve artificially radiating a sample and measuring TL dose-response over a range of applied doses.

Of the two components of the TL age equation the dose rate is the least problematic. It can be estimated by: (1) calculation from concentrations of radionucleides in the sample and surroundings under the assumption that parent and daughter nucleides in the decay series are in secular equilibrium and an allowance is made for cosmic ray flux and sediment water content (water is an effective alpha absorber), or (2) *in situ* measurements with synthetic TL dosimeters or gamma spectrometers (Berger, 1995). Ambient ionising radiation is derived from the decay of ^{238}U, ^{235}U, ^{232}Th and ^{40}K which produce alpha, beta and gamma radiation. Beta, gamma and exogenous cosmic radiation are lightly ionising and almost all the energy incident on a mineral is absorbed and converted to TL. Alpha particles, in contrast, are heavily ionising and only a relatively small fraction of particle energy is converted to TL. Alpha fluxes are therefore reduced by a factor (k, a, or b factor, Aitken, 1985; Berger, 1988a) in dose-rate calculations. *In situ* measurement of dose-rate have some advantages over calculations from nucleide concentration, not least that they take account of the particular geometry of a sample which may have a bearing on radiation flux depending on stratigraphic setting. The disadvantages of *in situ* techniques are that TL dosimeters need to be left in place for up to one year and gamma spectrometers are very expensive and not widely available (Berger, 1995). The major disadvantage of calculating dose-rate from element analysis is the reliance on secular equilibrium. This assumption may be breached where there is leaching (Wintle, 1990) or accumulation of radionucleides, or in strongly radiometrically heterogeneous material (Berger, 1995).

Water content is measured gravimetrically and though it may well have varied over the period of burial the error introduced is not usually large; a 10% change in water content introduces a 2-5% error in TL age (Berger et al., 1994).

Measurement of D_E potentially has more systematic errors, the most significant arising from the variability of effectiveness of "zeroing", and anomalous fading. Berger (1988b) states that the former, discussed further below, can contribute to 50-100% error in estimates of D_E. Anomalous fading refers to a type of "malign" TL manifested as a short-lived (hours to months) unstable component induced in traps over a wide-range of thermal stabilities by artificial laboratory irradiation of a sample. It is called anomalous because its decay is temperature independent and not governed by kinetic theory. This component has decayed from natural samples and if not accounted for or removed from irradiated samples, produces D_E and therefore age underestimates.

For sediments zeroed by sunlight the most appropriate methods for determining D_E are the partial bleach and regeneration methods (Wintle and Huntley, 1980), and a modification of the partial bleach procedure known as the total bleach method (Singhvi et al., 1982). With the partial bleach method irradiated and natural subsamples are divided into two groups. One group is given a short optical bleach under natural sunlight or, more commonly, artificial light. A range of filters is used to simulate natural insolation in the appropriate depositional environment (Berger, 1988b). The optical bleach is intended to remove a proportion of the accumulated TL which is less than the proportion lost by optical bleaching prior to deposition. To measure D_E, TL in unbleached and partially bleached subsamples is plotted against applied

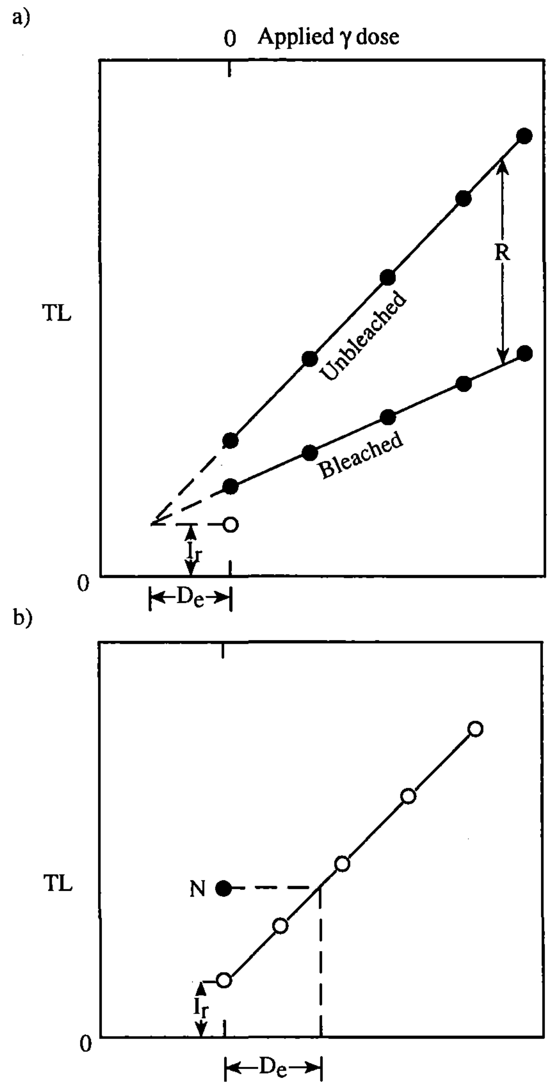

Fig. 1. Partial-bleach, total-bleach (Fig. 1a) and total-bleach regeneration (Fig. 1b) procedures for measuring equivalent dose (D_E) values (after Berger, 1988).

dose; D_E being the intercept of the bleached and unbleached dose-response curves (Fig. 1a). With the total bleach method a subsample is given a long optical bleach and this value extrapolated to the dose response curve to give the equivalent dose (Fig. 1a). With the regeneration method a number of subsamples are exposed to a strong optical bleaching and the dose-response curve is regenerated by applying a range of added doses across the subsamples (Fig. 1b). The equivalent dose is equal to the applied dose corresponding to the TL in the natural sample.

Calculated D_E varies with temperature across a glow curve owing to the presence of short-lifetime (low temperature) TL induced by laboratory irradiation which has decayed in the natural sample. This is thermally controlled decay not to be confused with anomalous fading. The effect is to give reduced D_E values at lower glow-curve temperatures. When D_E is plotted against T the curve often rises to a plateau at higher temperatures (ca. >300°C) indicating the temperature range corresponding to long-lifetime TL. The D_E plateau value is used for age calculation. The absence of a plateau (the plateau test) can indicate over-bleaching or the presence of anomalous fading, although not all anomalous fading can be detected this way. The total bleach and regeneration methods suffer from the serious flaw that the level of bleaching may exceed that of the sediment prior to deposition and consequently D_E and age will be overestimated. In the partial bleach method this is less likely and can be tested (Aitken, 1985, p. 226). The regeneration method also suffers from altered mineral sensitivity (slope of dose-response curve) after the intense bleaching used (Aitken, 1985).

Anomalous fading, despite being a non-thermally controlled TL decay, has been shown to be reduced by storage at elevated but modest temperatures (Berger, 1988a). A range of recipes from short-duration, relatively high temperature storage (e.g. 150°C for 16 h) to long-duration relatively low temperature storage (e.g. 75°C for 8 d) has been used with success (Wintle, 1990). Another alternative is to avoid minerals with a strong tendency to exhibit anomalous fading. Principal amongst the fading minerals are the volcanic (mid-to-high anorthite) feldspars (Wintle, 1973) although other feldspars are also implicated (Berger, 1995). Quartz and feldspar-free volcanic glass have been touted as being free of anomalous fading but in aeolian sediments glass is only occasionally present and quartz has two limitations for TL dating: (1) the TL saturates at relatively low doses giving a limited dating range (Berger, 1995) and (2) it shows resistance to optical bleaching for wavelengths longer than 450 nm and yet paradoxically is also susceptible to laboratory "over-bleaching" (Berger, 1988a). The latter poses difficulties in establishing the level of un-zeroed TL prior to burial.

There has been some resistance amongst Quaternary scientists to TL dating particularly for sediments with expected ages greater than ca. 100 ka. A tendency for TL dating to underestimate ages in this range has been attributed to a lifetime for the dominant TL signals much less (ca. 150 ka, Wintle, 1990) than the 1 Ma suggested by kinetics (Aitken, 1985). Berger (1994) suggested however that the ambiguities in TL dating of sediments > 100 ka arose from technical and methodological inadequacies rather than intrinsic limitations of the method. He listed the following as being the most significant: (1) poor age-control on sediments dated; (2) use of regeneration methods for calculating D_E; (3) use of peak integration to construct dose-response curves; (4) use of only one preheating schedule; and (5) use of UV TL emissions. Berger *et al.* (1992; 1994) demonstrated that TL dating could produce expected ages for loess up to 800 ka. They used a range of experimental parameters, including different filters for TL analysis, preheat recipes and optical bleaching and a number methods of determining D_E. They found that no one preheat option could be prescribed for all samples but a system of trial and error should be used. Blue TL gave dates in accord with those expected whereas UV TL, for unknown reasons, produced underestimates. They concluded that the partial and total bleach methods were most reliable but over different age ranges (partial bleach up to 300 ka and total bleach from 100-800 ka).

2. Methods

The soils analysed include M1/1, M3/1, as described in Almond (1996) and a second profile on M3 (M3/2) located on the crest of the moraine where it borders the eastern shore of Saltwater Lagoon. Soil stratigraphy is broadly consistent with M3/1 although loess is thinner (1.6 m vs 2.0 m) and there is less apparent detail in the soil stratigraphy. The position of maximum glass counts of Aokautere Ash is similar to M3/1 and other soils formed in loess (ca. 40 cm), but the boundary between L1b and L2 is not identifiable. Also, L3 between the second and third buried soils in M3/1, cannot be recognised in M3/2 because either part of the loess has been eroded, or less loess was deposited at the M3/2 site. Thus the third and fourth buried soils (b3 and b4 prefixes respectively) in M3/1 may be a subdivided geosol (Morrison, 1978), represented by a single composite geosol (b prefix) in M3/2. Soil horizons, stratigraphy and location of soils are presented in Fig. 2 (c.f. Fig. 7 of Almond, 1996).

M3/1 and M3/2 were sampled in contiguous increments, up to 20 cm thick, by soil horizon. M1/1 was sampled in the same manner to a depth of 2 m then by grab samples in all horizons below. In addition, 3-6 cm, near contiguous, core samples of M3/1 were available from a previous bulk density sampling. All samples were air dried, lightly crushed and sieved to 2 mm before analysis.

Total elements were measured on channel samples from M3/1 and M3/2 by x-ray fluorescence using a Philips PW2400 spectrometer (3 kw rhodium tube and vacuum path). The machine was calibrated against NIST mineral and sediment standards. Analyses were carried out on pellets formed by compressing (30 tonnes for 10 s) equal quantities of oven dry

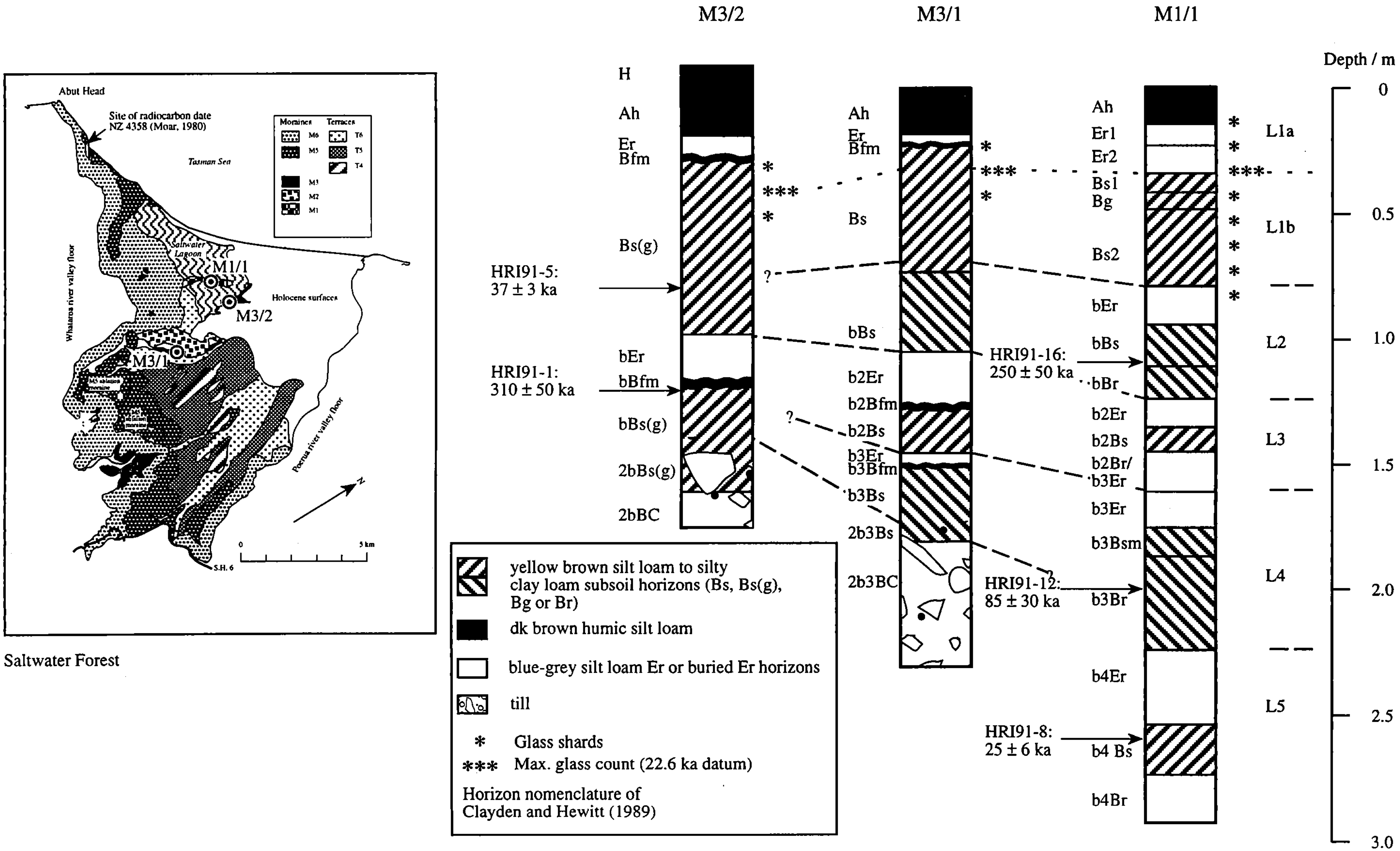

Fig. 2. Location, morphological features and stratigraphic relations of soils.

(60 °C) < 2 mm soil material and cellulose, ground to pass through a 75 μm sieve.

Particle-size distribution was measured in all (channel) samples of all soils as part of an assessment of original mineral material uniformity. Boiling hydrogen peroxide (30%) and acid ammonium oxalate were used separately to remove organic matter and secondary Fe minerals respectively. The >63 μm fraction was separated from treated material by wet sieving and oven dried, to be later separated into 63-125 μm, 125-250 μm, 250-500 μm and 500 μm - 2 mm size fractions by dry sieving. The <63 μm fraction was oven dried and then dispersed with an ultrasonic probe in about 500 ml of water treated with 10 ml of 10% Calgon solution, in a measuring cylinder. The proportion of material <2, 2-10, 10-20, and 20-63 μm was determined by pipette analysis at 20 °C with the equipment and method used by Grewal (1991). From the results the total proportion of material in the above size fractions was recalculated for the whole sample.

Mineral analyses were carried out on channel samples from profiles M3/1 and M1/1. Samples were pretreated as for particle-size analysis and a 53-500 μm fraction separated by wet sieving. Oven dry material was then separated into 5 mineral fractions with the use of a Frantz Isodynamic Separator. The settings employed for the separation were determined by trial and error by checking resultant mineral fractions for purity with a binocular microscope. Current, sideslope and forward slope settings and composition of the resultant mineral assemblages are given in Table 1. The yield of each assemblage was calculated as a percentage of the mass of the original 53-500 μm sample.

Table 1. Forward slope, side slope, current settings and resultant mineral assemblages separated by Frantz Isodynamic Separator.

Forward slope (deg)	Sideslope (deg)	Current / A	Mineral Assemblage
20	20	0.3	ilmenite
20	20	0.7	biotite + chlorite micas
20	20	1.2	quartz with ferrous inclusions, minor chlorite and tourmaline
20	5	1.2	muscovite mica
non-magnetic			quartz and feldspars

Phytoliths were extracted from channel samples of M1/1 and core samples of M3/1. Soil material was removed of organic material and deferrated as for particle-size analysis. A 5-200 μm fraction was separated by wet sieving to remove the >200 μm fraction and sedimentation in a centrifuge, followed by decanting, to remove the < 5 μm fraction (Jackson, 1956). Phytoliths were concentrated in Na polytungstate adjusted to 2.3-2.4 s.g. About 20 g of oven-dry 5-200 μm material was slowly added, with stirring, to 50 ml of heavy liquid in a 100 ml centrifuge bottle cut in half. The centrifuge bottles were sat in an ultra-sonic bath while the soil was added to aid dispersion. Upper soil horizons presented problems for dispersal because particles retained waxy organic coatings even after peroxide treatment which prevented grains from wetting up and sinking. Consequently some A horizon samples were unsuitable for counting and have been excluded from the results.

Heavy liquid/soil suspensions were stirred then centrifuged at 1000 rpm for 10 mins or until there was a clear separation between light and heavy fractions. This procedure was repeated 3 times. Because the yield of light material containing phytoliths was very small it could not be separated by decanting as most of the light fraction would have stuck to the vessel walls. Instead an eye-dropper was used to suck the material from the surface of the heavy liquid. Each eye-dropper-full was emptied into a syringe then passed through a 5 μm millipore filter when the syringe was full. When the scum of phytoliths was no longer apparent on the surface of the heavy liquid the filter was given a final wash and dried in an oven at 60 °C. The dry material on the filter was gently scraped off and mounted on a microscope slide in Canada Balsam and covered with a glass cover-slip. Between 180 and 600 phytolith grains were counted on each slide and classified according to the scheme of Kondo *et al.* (1994). The number of grains in each phytolith class were expressed as a percentage of the total.

To aid identification of phytoliths, a reference collection of phytoliths from common plants in the study area was established. Plant samples were ashed in a low-temperature plasma-induction furnace. Ashed material was washed in distilled water and sieved through a 5 μm millipore filter and mounted as above. The low temperature plasma induction furnace was effective in ashing grasses, sedges and flaxes but was less successful on woody material. Nonetheless, it produced examples of phytoliths useful in aiding identification of those extracted from soils.

TL dating was performed by G.W. Berger at Quaternary Sciences Centre, Desert Research Institute, Reno, Nevada. Samples were collected in February 1991 at which time profile M3/1 had not been excavated. Consequently only selected horizons of M3/2 and M1/1 were dated. A fine-silt (4-11 μm diameter) polymineralic fraction was used for TL dating, after removal of carbonates (with 1N HCl acid) and organics (with 30% H_2O_2). Other procedures were similar to those of Berger *et al.* (1994) and Berger *et al.* (1996). Blue-wavelength light was selected for TL measurements by use of appropriate glass filters (see above citations). Preheating prior to TL measurement was for 4 days at 130-140°C, depending on the sample.

3. Results

3.1. Profile uniformity

An expectation of initial uniform elemental and mineralogical composition for the loess in the study area seems reasonable given the provenance of sediments from which loess has been sourced. Loess sources for the study area would have included floodplains or perhaps till on the surface of glaciers that occupied valleys to the southwest (up prevailing wind direction).

The catchment of these rivers are within the Haast Schist and Torlesse Terrains of the Southern Alps. The changing exposure of Paleozoic igneous and sedimentary rocks west of the Alpine Fault, as a result of uplift, burial or right lateral motion, may have brought about some localised variation in loess mineralogy but these exposures are of limited extent (Warren, 1967) and the effect is likely to be negligible.

Textural variability of loess potentially is another source of non-uniformity, particularly if mineral abundances vary across particle-size fractions. Mew *et al.* (1988b) used sand (200-20 μm):silt (20-2 μm) ratio to test for lithological discontinuities in loess on 3 terraces in the Lamplough area, central Westland. They suggested there had been changes "in the pattern of deposition" or reworking of loess on the basis of breaks in slope of sand:silt ratio depth plots. Ratios were usually in the range 1.0 to 1.8. They noted and ascribed some significance to the fact that breaks in slope coincided with 'brown layers' which are possibly buried A horizons. Sand:silt ratio depth plots for M1/1, M3/1 and M3/2 are more consistent within profiles than those reported by Mew *et al.* (1988b) and are in a similar range (0.7-1.7) (Fig. 3a). Error bars plotted about selected points in M3/1 and M1/1 show the difference of the ratio between pairs of duplicate samples - error bars are plotted about the mean. Their magnitudes indicate that differences among horizons are real and not an artifact of analytical procedure.

Differences in sand:silt ratio within and between profiles may reflect temporal and spatial variability of loess deposition or just as likely, variation in degree of soil alteration. Chronosequence studies in Westland show that as soils become more weathered with time the proportion of silt increases at the expense of sand and coarser fractions (Basher, 1986; Campbell, 1975; Mokma et al., 1973). Similarly, other techniques for assessing parent material uniformity, such as immobile element ratios, suffer from the same problem. Depth plots of Ti:Zr ratio for M3/1 and M3/2 (Fig. 3b) show very similar trends between profiles; Ti:Zr ratio is low in A, E and buried E horizons and higher and very consistent elsewhere (Fig. 3b).

Sand:silt and Ti:Zr ratio depth plots do not challenge the assumption of an originally mineralogically and texturally uniform loess except perhaps for the Ah horizon (0-18 cm) of M3/1 and the b3Er horizon (160-174 cm) of M1/1. Both horizons have a very high sand:silt ratio. This is contrary to what might be expected of strongly weathered horizons and indicates deposition of a different material or perhaps local reworking with concentration of sand or loss of silt. In M3/1 and M3/2 the variation of Ti:Zr ratio could reflect differences in original loess mineralogy but accordance of minima with horizons of maximum alteration suggests it is more likely to be due to differences in degree of weathering and differential element mobility. Limitations to the applicability of Ti:Zr ratio for assessing parent material uniformity have been highlighted in studies of strongly weathered soils in Australia, particularly where biotite is a common soil mineral (Chartres et al., 1988). Other soils in the study area show a consistent drop in Ti:Zr ratio in surface horizons indicating that Ti is more mobile than Zr (unpublished

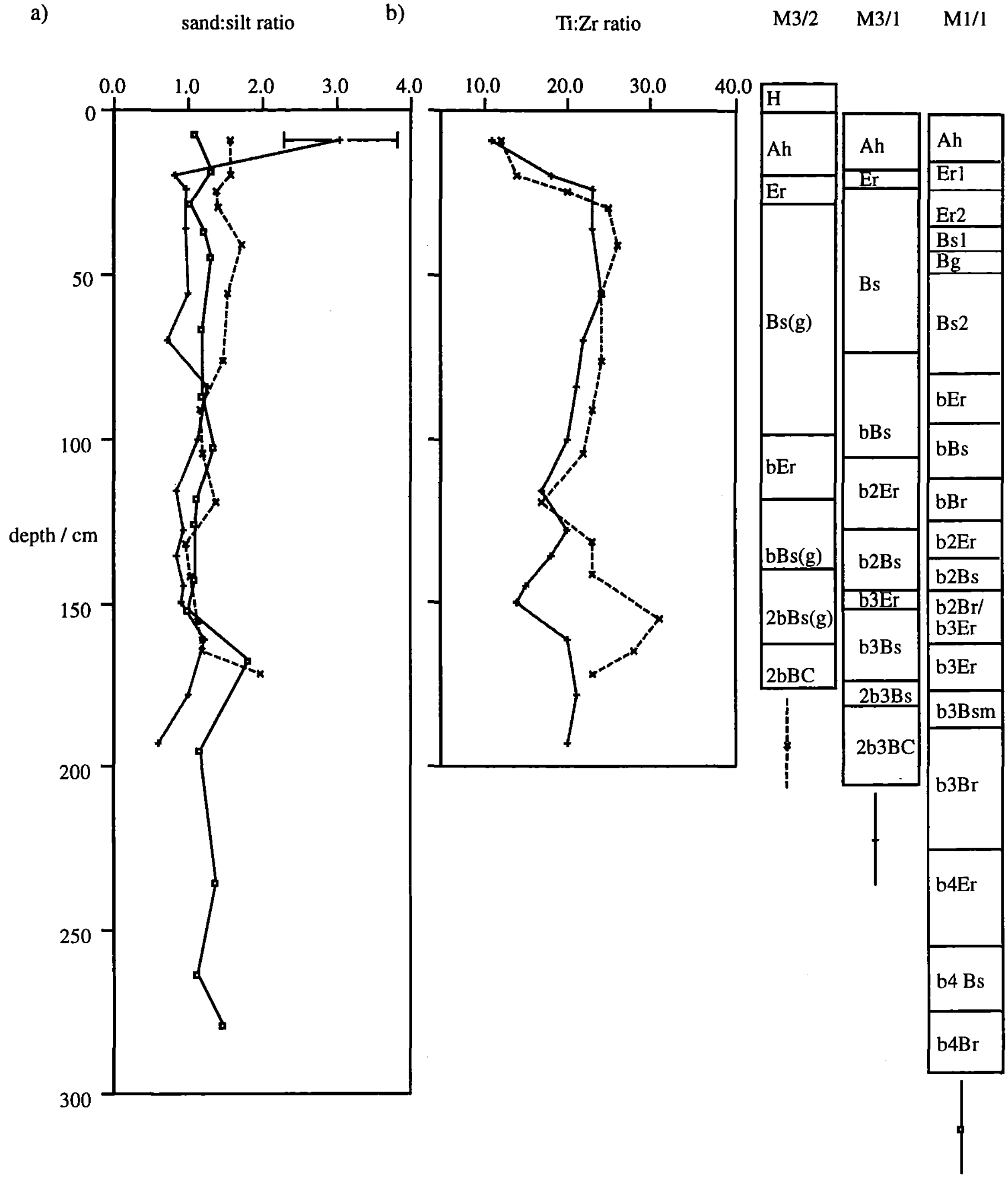

Fig. 3. Depth profile trends of a) sand:silt ratio for profiles M1/1, M3/1 and M3/2 and b) Ti:Zr ratio for M3/1 and M3/2.

data). If the minima in Ti:Zr depth plots are mostly a consequence of soil processes the otherwise consistent ratio suggests a uniform original loess mineralogy for these two soils.

3.2. Total element analysis

Total element data are presented for 3 groups of elements for profiles M3/1 and M3/2 (Fig. 4). The 3 groups of elements are distinguished by their relative mobility in soils. Group 1 includes highly mobile alkali earths, K, Na, Mg and Ca (Campbell, 1975; Loughnan, 1969;

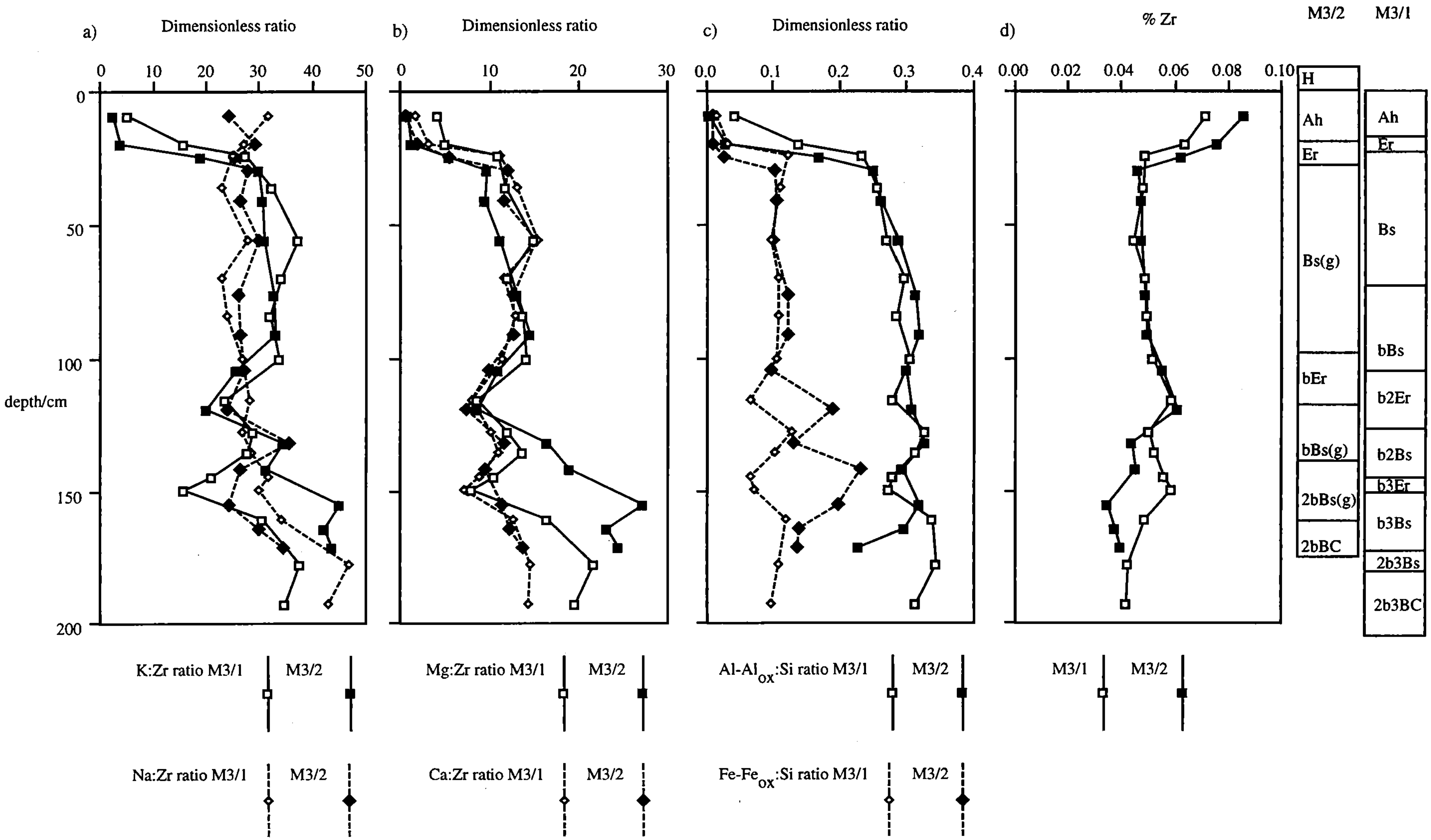

Fig. 4. Depth profiles of total element data for M3/1 and M3/2. a) Na:Zr and K:Zr ratios, b) Mg:Zr and Ca:Zr ratios, c) Al-Alox:Si and Fe-Feox:Si ratios, d) Total Zr (%) (ignition free).

Young, 1988); Group II includes Fe, Al whose mobilities are vary depending on soil Eh and pH conditions as well as the concentration of dissolved organic carbon (Campbell, 1975; Funakawa et al., 1993; Gustafsson et al., 1995; Loughnan, 1969; Lundstrom, 1993; Young, 1988); Group 3 includes the immobile element Zr (Barshad, 1964; Brewer, 1964).

Results for Group I are expressed as element: Zr ratio (Fig. 4a and b). Assuming Zr is highly immobile, the total mass of Zr within an horizon remains constant over time. If element:Zr ratio was originally constant throughout the profile, present variation within a profile reflects relative gains and losses of that element due to soil processes. All elements but Na show major depletion in surface A and E horizons with less depletion throughout the subsoil. Secondary minima coincide with buried E horizons but the magnitude of depletion, using the element:Zr ratio of the majority of the subsoil as a baseline, is about half of that in the surface A and E horizons. There appears to be no significant differential enrichment or depletion of Na in the loess of M3/1 or M3/2. This is somewhat surprising but it is possible that losses of Na from surface horizons may be compensated by inputs of cyclic salts from the sea, less than 5 km away.

Podzolisation, with a greater or lesser degree of gleying (depending on site conditions), dominates soil forming processes in the study area (Almond, submitted). In a study of a soil chronosequence in a similar soil forming environment in north Westland Campbell (1975) found that podzolisation resulted in the progressive loss of Fe and Al and concentration of Si. Furthermore, his study and that of Young (1988), also in a high rainfall environment, showed that under reducing conditions Fe had a mobility similar to that of K. In light of these findings Group II element depth plots have been expressed as Fe:Si and Al:Si ratios to highlight variation in intensity of podzolisation and gleying throughout the profiles (Fig. 4c). However to achieve a valid comparison between buried soil horizons and surface soil horizons it is desirable to remove any effects of mobile secondary Fe, Al or Si bearing compounds that have partially resaturated buried soil horizons since burial (Hewson, 1994). To this end % oxalate extractable Fe and Al values (Fe_{OX} and Al_{OX} respectively) (Almond, submitted) were subtracted from totals before dividing by Si percentage. Fe_{OX} ranged from 10 to 36% of total Fe and Al_{OX} ranged from 5 to 12% of total Al. The effect of mobile Si was ignored because oxalate extractable Si was extremely low (<0.4%) and negligible in the context of total Si percentage (> 50%).

Fe-Fe_{OX} and Al-Al_{OX} are most depleted, and Si most concentrated in surface A and E horizons. This implies that podzolisation is active in the upper horizons of the surface soil, with or without gleying. In subsoils Fe-Fe_{OX}:Si and Al-Al_{OX}:Si ratios are relatively constant. Small minima occur in buried E horizons but only for Fe are the minima significant with respect to the degree of depletion of Fe in the A and E horizons of the surface soils. These data indicate that podzolisation was not of sufficient intensity or duration to significantly mobilise and remove Al from the now-buried A and E horizons of the soils formed in L3 (b2Er and b3Er of M3/1: bEr of M3/2). I suggest that gleying was the dominant process in these horizons since it can mobilise Fe but not Al; the enhanced loss of Fe in the surface A and E horizons may be attributable to the combined effect of gleying and podzolisation.

Zr is most concentrated in the A horizons at > 0.07% compared to a baseline of 0.05% throughout most of the subsoil (Fig. 4d). Secondary maxima coinciding with buried E horizons reach 0.06%. Although there is a possibility that the high concentrations in the A horizon of M3/1 are due to parent material inhomogeneity (see earlier) a similar peak in M3/2 indicates that the effect is likely to be of pedogenic origin. On the basis of Zr enrichment the buried E horizons are altered to a lesser degree than surface horizons.

3.3. Mineral analysis

The dominant mineral assemblages separated by the Frantz Isodynamic Separator (Table 1) ranked in order of increasing resistance to weathering are: biotite + chlorite < muscovite < quartz + feldspar (Campbell, 1975; Goldich, 1938). The fact that these are assemblages of minerals rather than pure separates causes some difficulties in ranking by resistance to weathering particularly for the quartz and feldspar assemblage. Quartz is the most resistant mineral but albite, the dominant feldspar in the Haast Schists (Reed and Watters, 1978), ranks approximately alongside biotite. However, the quartz and feldspar assemblage is likely to be dominated by the former (Mew et al., 1983) and increasing abundance of this assemblage is therefore indicative of more intense weathering.

Depth plots of abundance of mineral assemblages for M3/1 and M1/1 (Fig. 5) show a marked concentration of the quartz + feldspar assemblage at the expense of all other assemblages in the A and E horizons of both soils. Here quartz and feldspar make up over 97% of the 53-500 μm size fraction. The quartz and feldspar assemblage abundance progressively declines with depth with a concomitant increase in the biotite + chlorite and muscovite assemblages, to the bBs horizon in L2. The quartz + feldspar assemblage increases again in the b2Er horizon in L3, remains roughly constant through the b2Bs then increases again into the b3Er horizon in L4. Lower horizons in L4 (b3Bs or b3Br) show a decrease in quartz + feldspar abundance. The peaks in the quartz + feldspar assemblage abundance are very close to 87% and 94% for the b2Er and b3Er horizons respectively, in both soils. This indicates firstly that the b3Er horizon in L4 is more intensely altered than the b2Er horizon in L3, and secondly, that the degree of alteration is not as great as in the A and E horizon of the surface soil. It is not until the b4Er horizon in L5 is reached that the abundance of the quartz + feldspar assemblage (96.3%) approaches that of the surface soil A or E horizon. Samples in

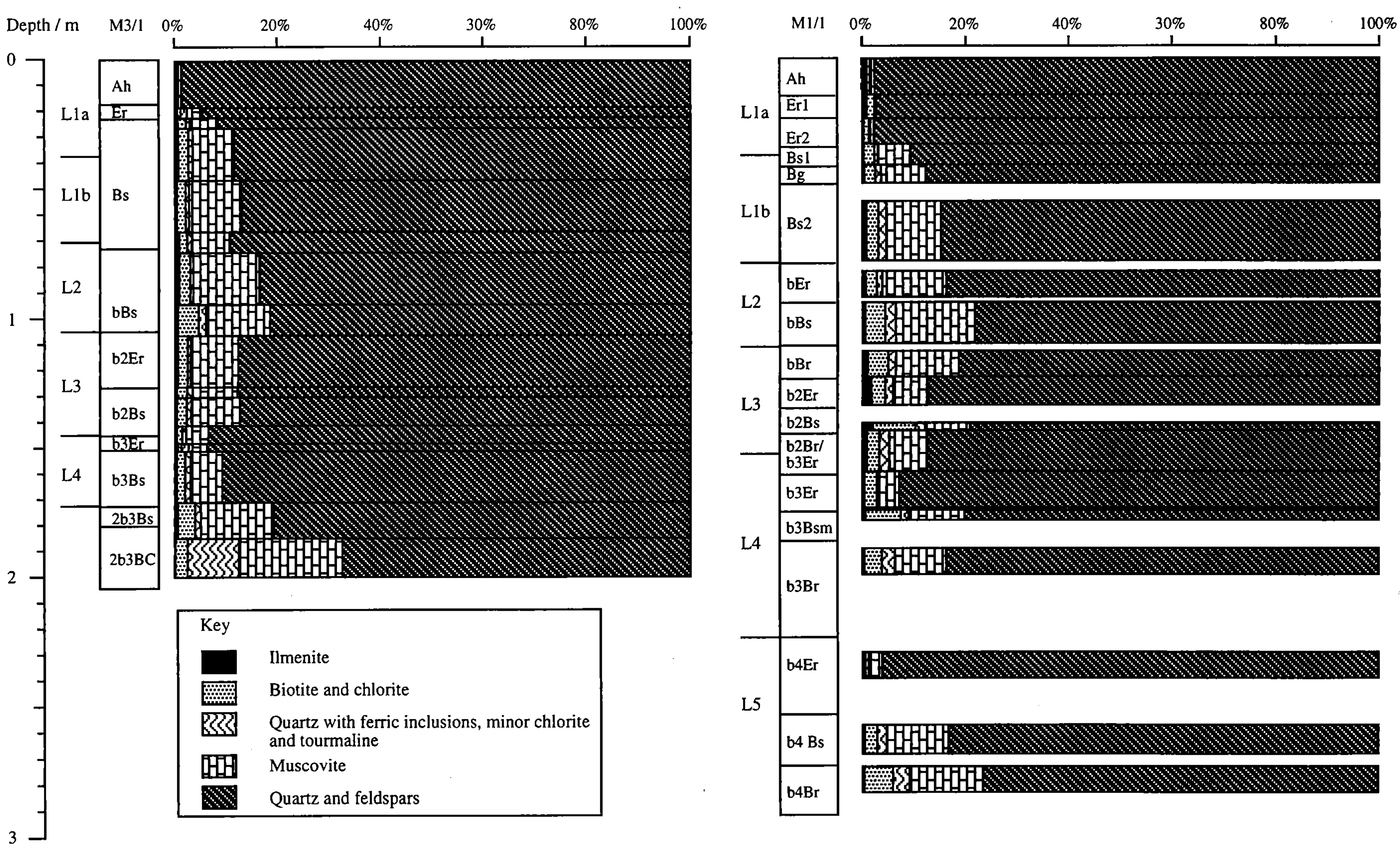

Fig. 5. Abundance of mineral assemblages within 53-200 µm fraction for M3/1 and M1/1.

horizons below, once again, have a lower quartz + feldspar abundance.

Depth profile trends of weathering indices calculated from biotite+chlorite:quartz+feldspar and muscovite:quartz+feldspar ratios are similar between the two profiles, as are absolute values in horizons occupying the same stratigraphic position (Fig. 6). This is strong support for the morphological identification of buried soils and suggests that both profiles are essentially intact (uneroded). Trends are more apparent for muscovite:quartz + feldspar ratio (Fig. 6a)

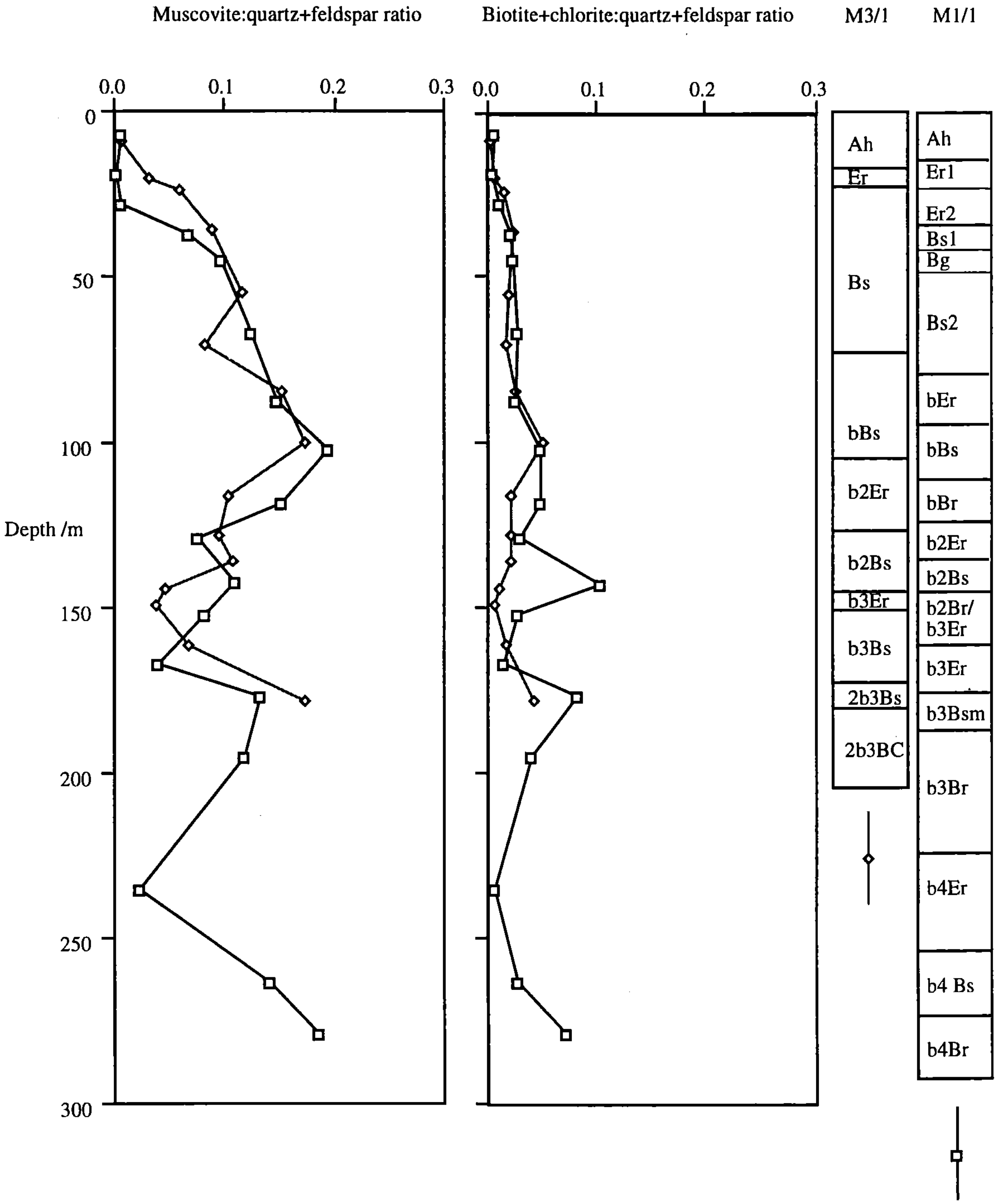

Fig. 6. Weathering index depth profiles for M3/1 and M1/1. a) Muscovite:quartz+feldspar, b) Biotite+chlorite:quartz+feldspar.

than biotite + chlorite:quartz + feldspar ratio (Fig. 6b) because of the greater absolute amounts of muscovite compared to biotite and chlorite. A somewhat poorer correspondence between the two profiles for the biotite + chlorite:quartz + feldspar ratio is probably due to greater relative errors generated by lower absolute amounts. These data also indicate that the soils in L3 and L4 (b2Er, b2Bs and b3Er, b3Bs(m)) are not as altered as the surface soils and that only the b4Er horizon in L5 approaches the degree of weathering of the surface soils. Had the soil in L5 been sampled more comprehensively, horizons with weathering as extreme as that in the surface soils may have been identified.

3.4. Phytolith analysis

Phytolith yields from M3/1 and M1/1 were very small and impossible to quantify. Furthermore many phytoliths showed evidence of dissolution in the form of solution pits and there were many unidentifiable fragments, particularly in deeper horizons. My inexperience with a wide spectrum of phytolith forms and the poor preservation of phytoliths in the soils studied has meant that phytoliths of unknown origin make up between 23 to 63% of total counts. A similar or greater proportion of unidentified phytoliths was found by Kondo *et al.* (1994) in the deepest and most strongly weathered sections of a ca. 18 m loess and tephra section at Rangitatau East Road in the North Island. Nevertheless, some broad trends are discernible and phytolith spectra provide some insight into vegetation changes in the time represented by loess in M3/1 and M1/1.

The phytolith spectra for both profiles have been divided into 3 zones although the zones are not coincident between profiles (Figs. 7 and 8). Phytolith zones M3/1-A and M1/1-A, which correspond closely with L1a, are dominated by spherical nodular and aggregate forms derived from post-glacial forest. In M3/1, phytolith zone M3/1-B (Fig. 7) which includes the upper half of L1b and the zone of maximum counts of Aokautere Ash, is characterised by an abrupt decrease of tree origin phytoliths and dominance of phytoliths from short cells of grasses. Here chloridoid and phytoliths from other short cells of grasses dominate. Their dominance, the occurrence, albeit at low abundance, of chionochloid phytoliths and a lack of festucoid phytoliths point to a tall tussock grassland containing *Chionochloa* (Kondo et al., 1994). There is no zone of comparable phytolith assemblage and stratigraphic position in M1/1 (Fig. 8). In the upper part of M171-B there is only a slight increase in abundance of short cell grass-origin phytoliths between 23 and 41 cm depth (Fig. 8), which does not justify the establishment of a separate zone.

In M3/1-C (L1b (lower half)-L4) and M1/1-B (L1b-L4) fan-shape and rectangular tablet, and elongate phytoliths are the dominant forms in an assemblage of low diversity. Phytoliths of tree origin persist but do not exceed 14% in M3/1 or 20% in M1/1. The peaks coincide with the b3Er horizons in both profiles.

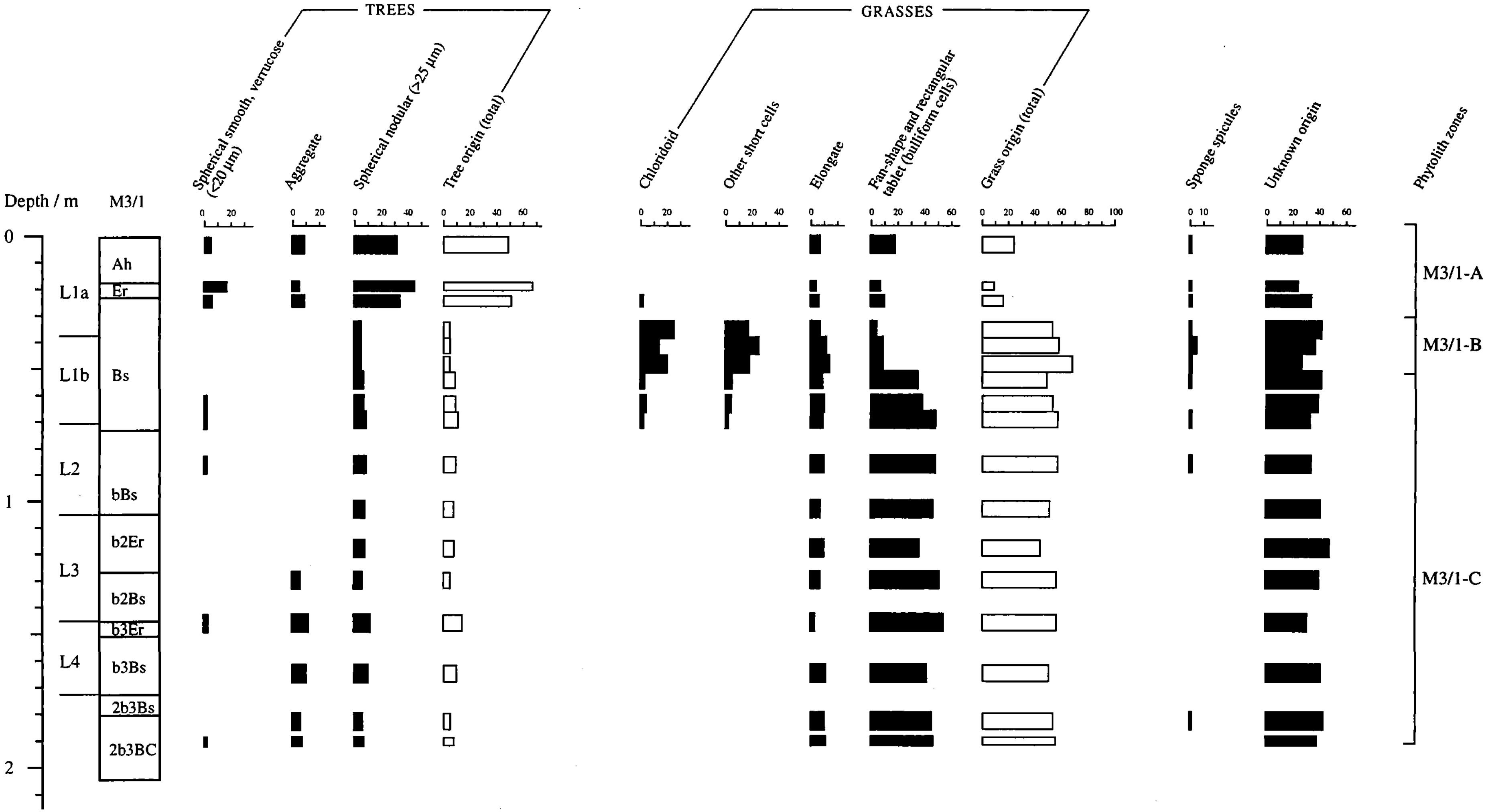

Fig. 7. Phytolith diagram for M3/1 (percentages for each sample) Note: some phytolith classes for which abundances are consistently low have not been shown individually but are included in the totals for trees or grasses.

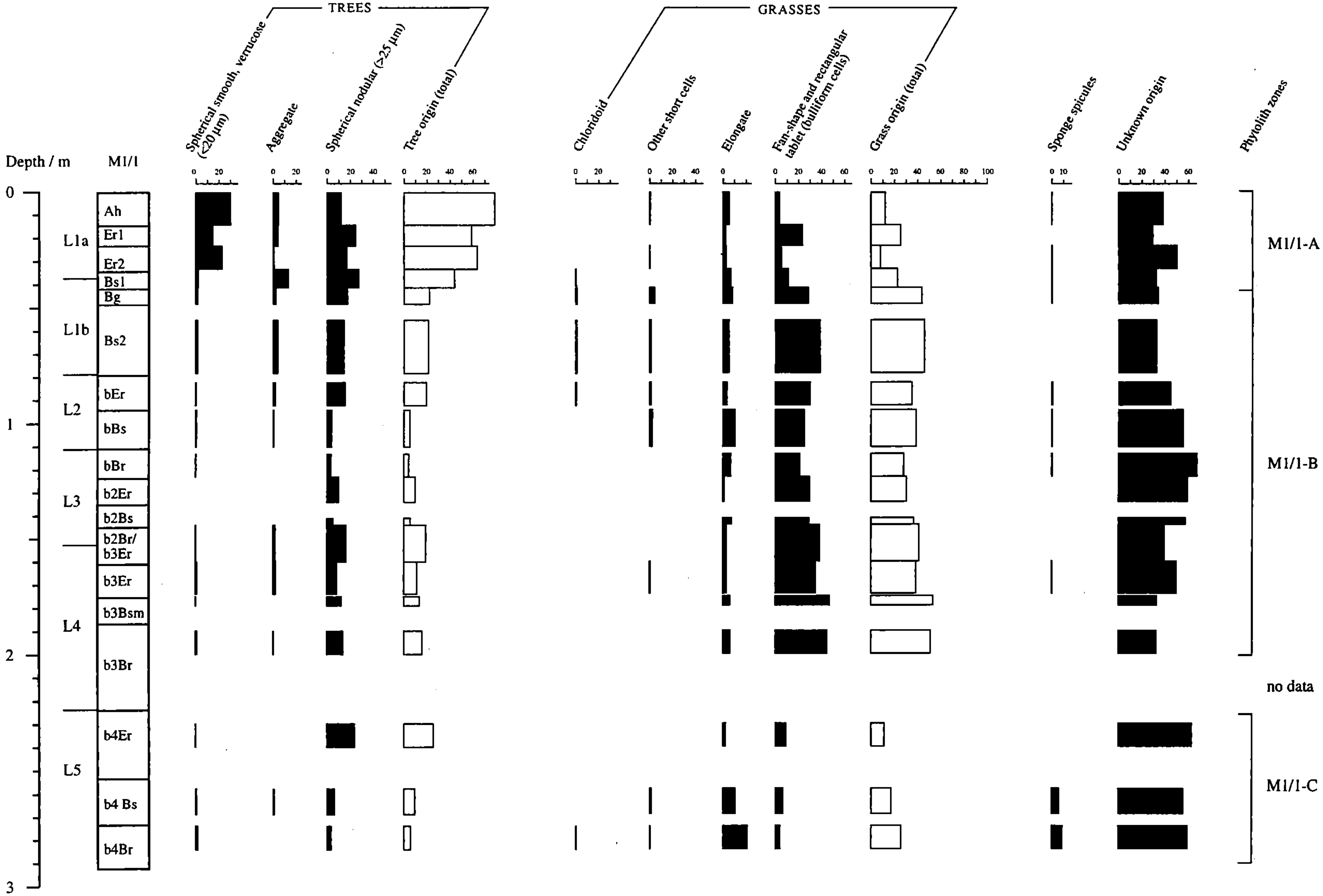

Fig. 8. Phytolith diagram for M1/1 (percentages for each sample). Note: some phytolith classes for which abundances are consistently low have not been shown individually but are included in the totals for trees or grasses.

Although the phytolith assemblage in M1/1-C (L5) differs little in composition from the overlying zone it has been separated on account of its very low phytolith content. Results from this zone are difficult to interpret because not only were phytoliths very rare, they were also poorly preserved. The slide prepared for counting for the b4Er horizon gave a total of only 180 phytoliths and 63% of these were of unknown origin. Throughout L5 fan-shape and rectangular tablet phytoliths are relatively rare but in the b4Br horizon elongate phytoliths increase in abundance as do small (< 15 μm) smooth spherical phytoliths characteristic of wire rush (*Empoedisma* sp.) (Kondo et al., 1994). This horizon also shows the greatest abundance (9%) of sponge spicules of M3/1 and M1/1; the rest of profiles M3/1 and M1/1 have counts of sponge spicules <1%. The increase may merely reflect an increased representation of sponge spicules due to the diminished phytolith flux.

3.5. Thermoluminescence dating

Thermoluminescence ages are presented in Table 2 and alongside the stratigraphic position of samples in Fig. 2. The date for HRI91-5 (Bs(g) horizon of M3/2, 80 cm depth) distinguishes itself from the others because of its greater precision (± 10%). This sample was dated earlier in a different laboratory than the other four samples. The precision of the later dates (± 20-30%) is lower because the beta source used for laboratory irradiation had not been calibrated but was only roughly known. Furthermore none of these samples permitted measurement of the alpha efficiency factor (b-value) and all evinced unusual high-dose response, indicating interference to the desired feldspar signal from an undesired quartz signal. In addition to the greater precision, the quartz interference effects were also least severe in HRI91-5.

Table 2. Loess samples and TL ages.

Sample	Profile and horizon	Depth (m)	Dose-rate (Gy/ka)	Equivalent dose (Gy)	TL age (ka)
HRI91-5	M3/2 Bs(g)	0.80	2.92±0.12	107±9	37±3
HRI91-1	M3/2 bBs(g)	1.20	3.05±0.17	950±150	310±50
HRI91-16	M1/1 bBs	1.10	est 3.0	750±120	250±50
HRI91-12	M1/1 b3Br	2.00	3.31±0.18	280±90	85±30
HRI91-8	M1/1 b4Bs	2.60	3.08±0.20	77±19	25±6

Given the lack of precision, methodological problems, and the more serious reversals of age in M1/1, the TL dates, other than HRI91-5, are of little value. Apparent age over-estimation of samples HRI91-1 (310 ka) and HRI91-16 (250 ka) may be due to un-zeroed TL. This is more likely for HRI91-1 which is close to the underlying till and could contain grains that have not experienced substantial optical bleaching. The age HRI91-8 (25 ka) is certainly an underestimate in view of its position relative to the zone of maximum counts of Aokautere Ash. The usual anomalous fading tests (several months post-irradiation storage plus preheating) failed to show any measurable anomalous fading thereby removing this as a source of age underestimation. The most likely explanation is that the TL has rapidly saturated because of the dominance of quartz. The age of HRI91-12 (85 ka) is believable but I place little weight on this date because of its precision (± 30 ka) and the unreliability of the other ages in the later suite of samples to which it belonged.

4. Discussion

Soil profile trends of total elements (Fig. 4), mineral assemblages (Fig. 5) and weathering indices (Fig. 6) collectively confirm the field identification of buried soils. Buried E horizons generally show a greater degree of soil alteration, be it loss of mobile elements, concentration of resistant elements or changes in ratio of resistant to less resistant minerals, than underlying or overlying horizons. Buried E horizons represent still-stands of loess accretion when the effects of weathering and leaching were not mitigated by inputs of unweathered material. Consequently they are more altered than other parts of the loess columns. However the data show that, apart from the b4Er of L5 in M1/1, no buried soil horizons are as altered as the A and E horizons of the surface soils and the buried soils appear to have experienced different pedogenic processes. The differences in Fe:Si and Al:Si ratios of the buried E horizons to the surface A and E horizons indicate that conditions were less favourable to podzolisation than at present, although gleying was still important. In view of the evidence for similar rainfall during the Last Glaciation (Soons, 1978) changes in soil conditions (pH, concentrations of DOC) due to different vegetation seem to be the most likely explanation for reduced mobility of Al.

The evidence does not support the interpretation of Almond (1996) whereby the soils in L3 and L4 of M3/1 were correlated to the Last Interglacial (Oxygen Isotope Stage 5). Oxygen isotope stage (OIS) 5 covered a period of at least 50 ka when climate and vegetation in Westland were similar to that of the Holocene (Moar and Suggate, 1996). Given that the duration of OIS 5 is in excess of 35 ka longer than the present interglacial (Holocene) we should expect buried paleosols formed during OIS 5 to show a greater degree of soil alteration than soils formed during and since the Otiran, even disregarding post-burial soil alteration. Since this is not so with the soils in L3 and L4, it seems likely that they formed sometime during the Otiran, perhaps OIS 3, under a non-forest vegetation.

Phytolith data support the interpretation above. Overall trends in phytolith assemblage indicate that the current forest vegetation replaced a grassland growing during the deposition of L1b around the time of the Last Glacial Maximum (phytolith zone M3/1-B). The phytolith assemblage of M3/1-B suggests that, at least in this location, it was a tall tussock grassland. The lack of corroboration from M1/1 could reflect inter-site differences or limited resolution in M1/1 for which sampling was much coarser. A pollen spectrum from a peat lens at Abut Head (Fig. 2 inset) radiocarbon dated at 16 450 ± 140 yrs BP, (untreated) confirms the dominance of grassland communities, at least on a regional scale, around this time (Moar, 1980). The different, yet complementary, information provided by phytolith and pollen analysis highlights the advantage of carrying out combined studies. More localised information provided by phytolith analysis complements the regional record of pollens. Furthermore, one technique may offer taxonomic discrimination where the other does not. For example, this study indicates that *Chionochloa* sp. were present in the grassland around the Last Glacial Maximum where pollen analysis had no discrimination below family level.

Phytolith zones M3/1-C and M1/1-B indicate that the grassland around the Last Glacial Maximum represented by M3/1-B, replaced a grassland of different character. The dominant fan-shape and rectangular tablet phytoliths of M3/1-C and M1/1-B are are relatively large (20-60 μm) solid silica plugs derived from the bulliform (motor cells) of grasses, suggested by Kondo *et al.* (1994) to be *Rytidosperma.* The presence of some phytoliths of tree origin suggests a sparse woody overstory.

Kondo *et al.* (1994) did not observe fan-shape and rectangular tablet phytoliths in *Rytidosperma* but based their suggestion on descriptions of the anatomy (Zotov, 1963). According to my reading of Zotov (1963), although the author describes dumbbell shaped silica cells (panicoid phytoliths), and notes the presence of motor (bulliform) cells in *Notodanthonia* (now *Rytidosperma)*, there is no explicit mention of the latter being silicified. The work of Wynn Parry and Smithson (1964) on British grasses justifies some caution in this respect. They observed silicified bulliform cells in a number of grasses but found that the phenomenon was common only in 3 species; 2 species of *Danthonieae,* and 1 species of *Agrostideae.*

In an effort to confirm the presence of fan-shape and rectangular tablet (hereafter referred to as bulliform) phytoliths in *Rytidosperma,* I ashed leaves of 1 specimen each of 3 species of *Rytidosperma* found in Westland (*R. gracile, R. unarede and R. setifolium*[5]) (Connor and Edgar, 1979). None of them produced any bulliform phytoliths. In contrast, *Microlaena avenacea* (rice grass) which is common in the study area and throughout Westland, produced abundant bulliform phytoliths (Plate 1). *M. avenacea* grows in a range of environments including low altitude conifer/ broad-leaf forest or beech forest, subalpine conifer/broadleaf forest, beneath kanuka tree heath, amongst a herbaceous lower tier in grey scrub in areas subject to severe frosts, and in oligotrophic mire-forest ecotones (Wardle, 1991). In Saltwater

[5] Specimens were provided by the Landcare Herbarium at Lincoln, Canterbury.

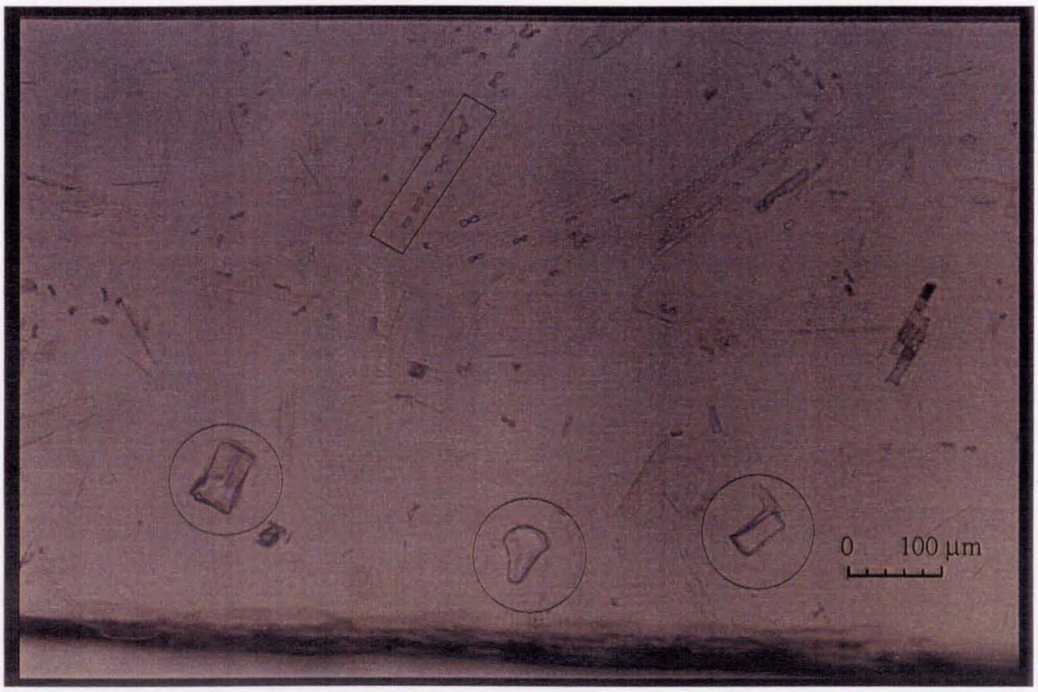

Plate 1. Phytoliths in *Microlaena avenacea* (bulliform phytoliths circled, panicoid phytoliths in the rectangle).

Forest and other rimu dominated conifer/broadleaf forests in Westland it tends to be most common near streams and in other areas where light penetrates the canopy but generally does not grow in full sunlight (pers. obs.) (Plate 2).

The degree of silicification of bulliform cells can vary within individuals of a species depending

Plate 2. *Microlaena avenacea* on the edge of Tunnel Creek, Saltwater Forest.

on soil moisture conditions, location of cells on the leaf blade, and tissue age (Wynn Parry and Smithson, 1964). The samples of *Rytidosperma* I ashed may have, for one or more of these reasons, been atypical. Nevertheless, I have assumed *M. avenacea* is the source of bulliform phytoliths preserved in the loess of M3/1 and M1/1 - *M. avenacea* is a prolific bulliform phytolith producer and its environmental tolerance suggests it could have persisted throughout the late Pleistocene. Phytolith zones M3/1-C and M1/1-B in which bulliform phytoliths are the most abundant class, are therefore inferred to represent a grassland dominated by, or at least containing, *M. avenacea* with a woody overstory. The climatic interpretations are that conditions were probably not as severe as during the Last Glacial Maximum, still sufficiently wet to provide suitable habitat for *M. avenacea,* but certainly not full interglacial.

The monotony of the phytolith spectra and the obvious evidence of dissolution (Plate 3) highlight the limitations of this technique for vegetation reconstruction. Bulliform phytoliths, whether sourced from *Microlaena* or *Rytidosperma,* would have been associated with smaller panicoid phytoliths (Plate 1) and yet none of these are to be found in the soils. In the deeper parts of the loess sections, the phytolith spectra are skewed toward large, solid morphologs with low surface area to volume ratio which resist dissolution, and any plants that produced only small phytoliths are completely unrepresented.

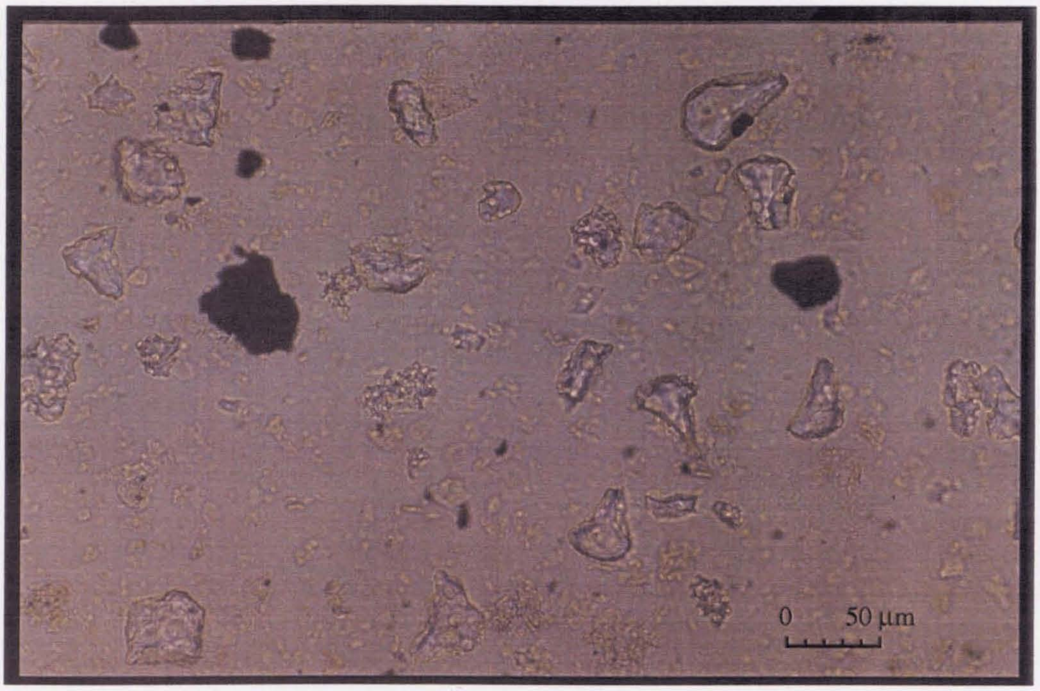

Plate 3. Pitted and weathered (mostly bulliform) phytoliths from b3Er horizon, M1/1.

The extremely low abundance of phytoliths in phytolith zone M1/1-C, corresponding to L5, requires some explanation. Either the material is not loess and was deposited rapidly enough for very little vegetation to be growing on the site, or it provides an extreme example of phytolith dissolution. Particle size distribution data do not offer any strong evidence for L5 being different to overlying material. Particle size grading curves for horizons in L5 lie within the envelope defined by all overlying loess sheets except for the sample increment in the b4Br horizon (lowermost increment) which is slightly more sandy (Fig. 9). A similar lack of phytoliths has been documented in loess on the east coast of South Island (Kondo et al., 1994; Raeside, 1964) under a much lower leaching regime. From my own observations, below 2 m depth in loess from the Timaru area, phytoliths are extremely rare even within the upper,

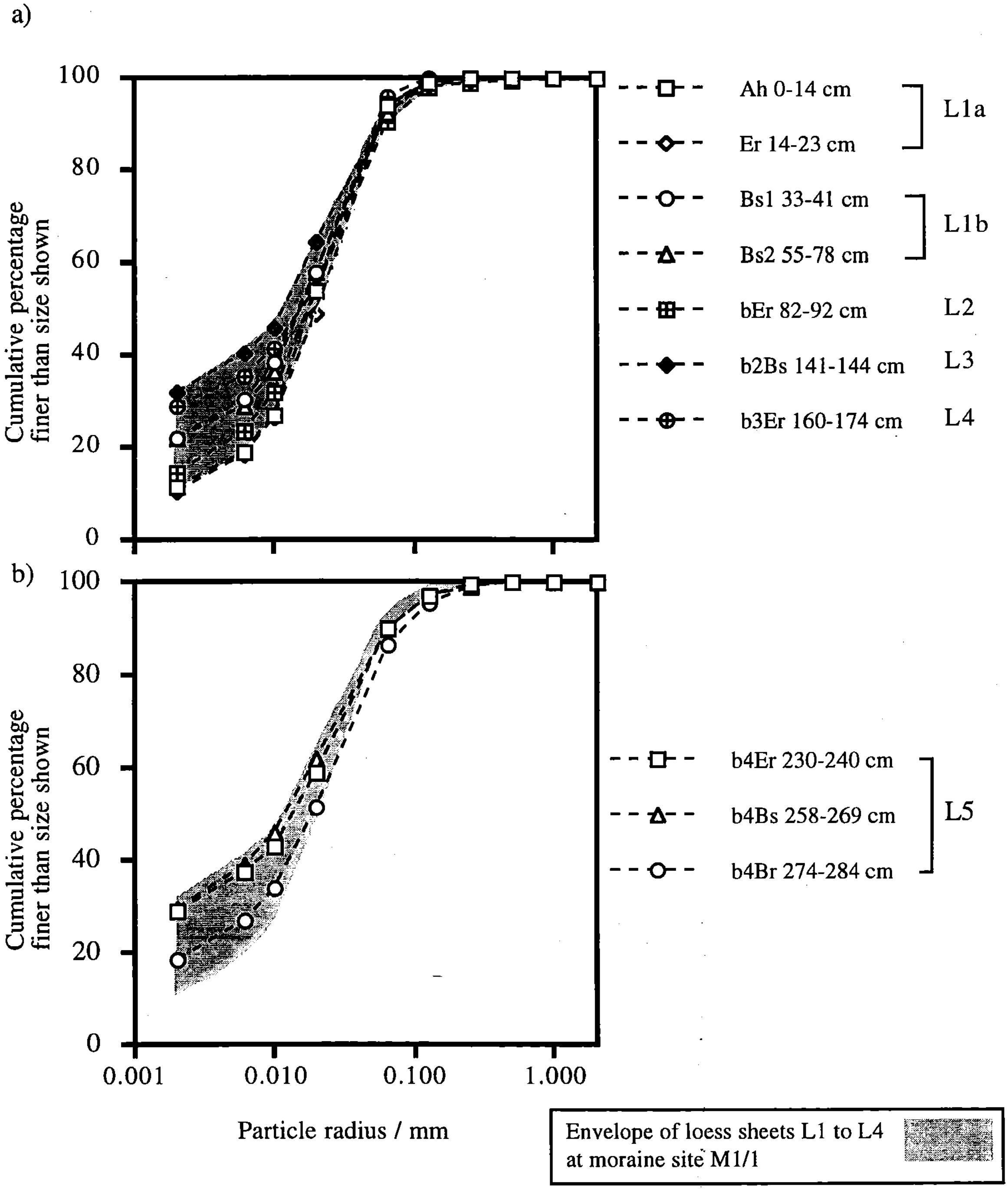

Fig. 9. Cumulative particle size distribution for selected horizons and loess sheets in M1/1.
a) L1 to L4 b) L5.

formerly near-surface, parts of loess sheets. The absence of phytoliths from a deposit does not preclude it from being loess.

The interpretation of the increased abundance of sponge spicules in the lower part of M1/1-C is problematic. The spicules are fragments varying from <10 to 100 μm and it was not possible to determine whether they originated from freshwater or marine sponges. Given the setting and the fact that sponge spicules are readily transported by wind (Jones and Beavers, 1963; Smithson, 1959), both sources are possible. Any environmental interpretation of the increased sponge spicule flux is tenuous but, assuming the mode of transport to the site was aeolian, I infer that there were bodies of water in closer proximity during the initial, compared to later stages of loess deposition at M1/1.

In summary, phytolith data provide little information about vegetation or climate during deposition of L5. L4 to L2 were deposited, and soils in them formed, under a grassland with woody overstory. Climate was most likely to be cool and moist but not full glacial. L1b and part of L1a accumulated under a tall tussock grassland in a severe climate around the time of the Last Glacial Maximum.

The TL ages measured in this study are of little value because of a lack of precision due to laboratory limitations (beta source calibration) and, more seriously, age reversals. The strongly leached and weathered loess in Westland presents a significant challenge to application of TL dating because of the dominance of quartz in silt and sand fractions, and potentially significant violation of the closed-system assumptions made in the procedure. Calculations of dose-rate in this study were made on the basis of radionucleide concentrations under the assumption of secular equilibrium. The mobility and significant depletion of K from some horizons is signalled by total element ratio plots (Fig. 4). Furthermore U and Th, if they are associated with heavy minerals such as zircon, might well be concentrated over time. Future studies should use *in situ* determination of dose-rate although this still does not avoid potential depletion or concentration of radionucleides over time.

The most reliable date, from the Bs(g) horizon in M3/2 (80 cm), indicates that about 80 cm of loess has accumulated since 37 ka. Assuming that the upper parts of the loess section (i.e. L1a, L1b) are intact, the sample is in the upper part of L2 and thereby provides a maximum age for the onset of deposition of L1b. Being close to the surface of L2, the date may more closely represent the onset of deposition of L1b than the depositional age because TL clocks in the sample may have been partially re-zeroed by pedoturbation (Berger and Busacca, 1995). This effect would have stopped when the sample had been buried deeply enough.

The beginning of accumulation of L1b on gravels beneath T4 was dated by radiocarbon at 35.8±1.8 ka radiocarbon years (Almond, 1996) (ca. 38 ka calendar years, Mazaud et al., 1991) but this was interpreted to be effectively an infinite age. In light of the consistency between the TL and radiocarbon ages I now suggest that L1b started accumulating sometime after ca. 37

ka. Assuming that the soils in L3 and L4 are not Last Interglacial soils, they, or the composite soil representing them in M3/2, probably formed during the early part of OIS 3 in loess sheets accumulated during OIS 4. The conformable contact between L4 and the underlying tills on M3 suggests that the M3 advance also occurred in OIS 4 and is the equivalent of the $K2_1$ advance of central Westland. This correlation places the accumulation of L2 in OIS 3, sometime between 50 and 37 ka and implies that the M4 advance and the associated aggradation event (gravels beneath T4) also occurred in this time. No equivalent advance is documented for central Westland (Suggate, 1990). Either this advance did not occur, has not been recognised, or has left no trace in that region. Given the proximity of the study area to the Southern Alps it is conceivable that glacial advances extended into the study area where they did not in the Hokitika and Kumara regions where the central Westland stratigraphy was established. The problems of correlation discussed above notwithstanding, the soil in L5 is correlated to the Last Interglacial, and L5 and the M1 advance to OIS 6. The M1 advance is equivalent to the K1 advance of central Westland. The revised correlation is presented in Fig. 10 (c.f. Fig. 9 of Almond, 1996).

These correlations imply M1 is of Penultimate Glaciation age and hence the conclusions of Almond (1996) regarding the tectonic setting of the coastal part of the study area are still valid. If this part of the south Westland piedmont were actively uplifting we would expect to find raised beaches or cliffs on M1.

5. Conclusions

The published stratigraphy of buried soils and loess sheets, and the late Pleistocene glacial landforms in south Westland (Almond, 1996) needs revision. Total element and mineralogical data support the existing morphological definition of soil and loess stratigraphic units but together with phytolith data, contradict the existing correlation of prominent soils to climatic events. Both L1a and L1b are now correlated with OIS 2 (perhaps including late OIS 3) with the onset of accumulation of L1b beginning some time after 37 ka. L2 accumulated during OIS 3 between 37 ka and ca. 50 ka. L2 is less strongly pedogenically altered than upper parts of other loess sheets which indicates the period between the cessation of deposition of L2 and the onset of accumulation of L1b was not long. L3 is a thin loess sheet that is not consistently present. It probably represents a short period of localised deposition. L4 is conformable with the till beneath M3 and is inferred to have accumulated during OIS 4. The prominent soils in L3 and L4 are strongly altered although not to the degree, or in the same way as the surface soil. This evidence and phytolith data support the correlation of the period of soil development to the beginning of OIS 3 when climate was less severe than the rest of the Last Glaciation but not as mild as the Last Interglacial (OIS 5). The soil in L5 is correlated with the

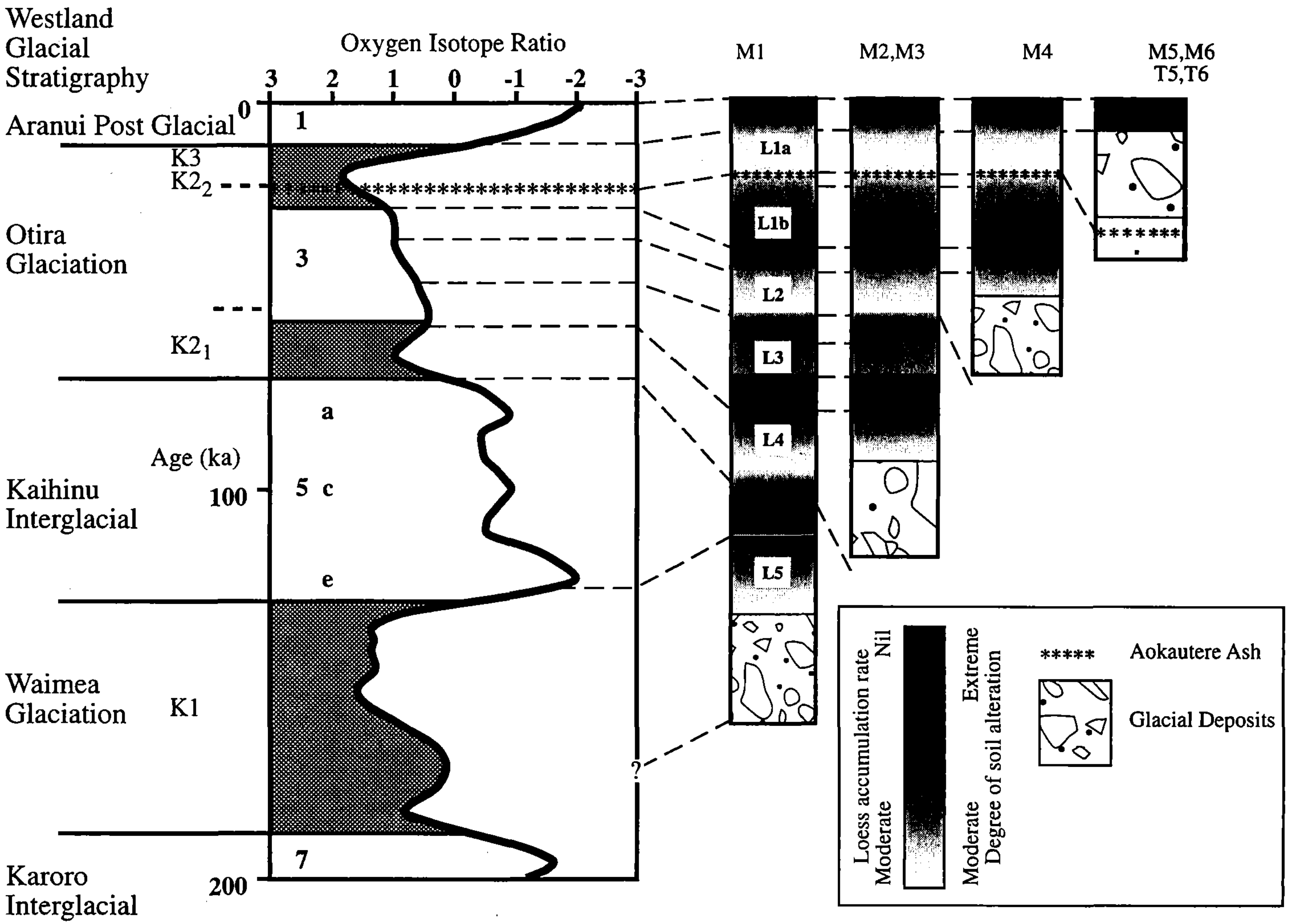

Fig. 10. Revised correlation of periods of time represented by buried soils, loess and glacial deposits in the study area to the Westland glacial chronology and the oxygen isotope record (after Imbrie *et al.*, 1984).

Last Interglacial on the grounds of its extreme modification. L5 and the moraine beneath M1 are correlated to OIS 6.

The revised correlation of glacial advances in Saltwater forest and neighbouring catchments to the central Westland glacial stratigraphy is as follows:

1. the correlation of glacial advances younger than 22.6 ka is unchanged; the M6 and M5 advances remain correlated with the K_3 and K_{22} advances respectively;
2. the M4 advance took place between 37 and 50 ka in OIS 3. No correlative exists in central Westland;
3. the M3 and M2 advances are correlated with the K_{21} advance;
4. the M1 advance is correlated to the K_1 advance.

Thermoluminescence dating was used in an attempt to provide absolute time control but there appear to be serious limitations of the method in the highly weathered loess of Westland.

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Chapter 4

Pedogenesis by upbuilding in an extreme leaching and weathering environment, and slow loess accretion, south Westland, New Zealand

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Abstract

Soil formation is usually envisaged as a topdown process whereby weathering and other soil process act downwards into an existing parent material. However, in many terrestrial environments the accumulation of sediment and soil formation are contemporaneous. It is only where the rate of one process far exceeds the rate of the other that assumptions of topdown pedogenesis are appropriate. The distinctive feature of upbuilding soil formation is that each increment of soil below the A horizon has experienced processes characteristic of all horizons above it. This study examines the influence of upbuilding soil formation in a high rainfall environment of rapid soil formation and slow loess accretion - an environment where upbuilding could have a noticeable influence on soil morphological and chemical properties. Soils formed by topdown soil formation (simple soils) are compared with soils formed in loess that accumulated from early Last Glaciation to the Holocene (upbuilding soils). Soils are compared in two environments: Humose or Silt-mantled Perch-gley Podzols on poorly-drained terraces and Placic Perch-gley Podzols or Acidic Allophanic Brown Soils on well-drained moraine crests. I have assumed that properties of the upbuilding soils inherited during the current topdown phase of pedogenesis beginning in the Holocene are analogous to those in the simple soils. In this manner soil properties evolved during upbuilding are distinguished.

In well-drained upbuilding soils on moraines, subsoil organic carbon has mineralised so that profile trends are very similar to corresponding simple soils. Upbuilding has enhanced weathering in subsoils which is reflected in lower pH, lower total P and lower primary inorganic P than in simple soil counterparts. Total P concentrations in upbuilding soils are much lower than unweathered loess and constant throughout the subsoil, and there is almost no P in primary forms. In contrast, total P in C horizons of simple soils reaches concentrations typical of the parent material and more than 30% of this is in a primary inorganic form. The differences are attributed to each subsoil increment of upbuilding soils having been part of acid A and E horizons. Oxalate extractable Fe and Al data indicate that podzolisation was active during upbuilding.

Poorly-drained upbuilding soils on terraces have higher subsoil organic carbon (inherited from former A or O horizons) than comparable simple soils. Buried A and O horizons inherited during upbuilding and preserved

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in the reducing conditions have acted as sinks for mobile Al and now resemble podzolic illuvial Bh horizons. They are also pools of relatively immobile organic P.

The rate of P input from loess has been sufficient to maintain total P levels in >35 ka upbuilding soils at levels typical of 5 to 12 ka simple soils. This may have significance for understanding plant primary succession in the wider study area. Upbuilding through addition of loess to soils promotes a pedogenic pathway distinct from topdown pedogenesis. Consequently, the soil chronosequence studies in Westland that included soils on old, loess-mantled surfaces do not strictly satisfy the assumptions of soil chronosequences. The nature and rate of soil processes inferred from these studies need revisiting. The findings of this study are transferable to other environments where the rate of soil formation and sediment accumulation are comparable.

Keywords: Pedogenesis; soil chronosequences; loess; soil stratigraphy

1. Introduction

Theories of soil formation provide a framework for understanding the combinations of soil properties in a single pedon or, in conjunction with an appreciation of landscape dynamics, the spatial variability of soil in the landscape. Theories of soil formation reflect the state of knowledge at any time but also establish the perspective from which new knowledge is viewed. Inferences based on new information are not independent of the paradigms under which we operate.

Soil formation is most often presented as occurring in a topdown sense i.e. a set of processes act on a preexisting body of sediment or rock (the parent material) such that the degree and depth of soil alteration increases with time (Duchaufour, 1977; Simonson, 1959). This is a simplification of reality which holds for many situations but in aggrading landscapes it is inappropriate (Nikiforoff, 1949; Raeside, 1964). Soils on surfaces incrementally mantled by loess, overbank deposits, tephra or other extrinsic inputs form contemporaneously with sediment addition to a greater or lesser extent depending on sedimentation rates.

Johnson (1987) presents a model of soil evolution which explicitly includes the potential for soil formation and sediment accumulation to occur together. In this model soils evolve along two coacting pathways; one progressive, the other regressive. The progressive pathway brings about development of horizons, soil deepening and chemical stability while homogenisation, chemical instability or soil thinning result from the regressive pathway. Soils on aggrading landforms may undergo developmental upbuilding if rates of addition permit assimilative pedogenesis and profile deepening, or regressive upbuilding if additions are so rapid as to prevent horizonation. A degree of subjectivity is inevitable as additions may lead to profile deepening (progressive pedogenesis) but also to chemical instability (regressive pedogenesis). The final judgment rests on the weight given to different soil properties in determining the degree of soil development.

The significant feature of upbuilding soils is that below the current A horizon all the soil has experienced processes, in sequence, characteristic of A (or E) horizons followed by processes characteristic of B horizons. Consider soil formation to be a depth function encompassing the range of intensities of additions, losses, transformations and translocations in a profile in a particular environment (Simonson, 1959). The collective history of soil formation for a soil increment of an upbuilding soil is equivalent to that increment passing down through the depth function at a rate determined by sedimentation rate. Sedimentation rate and soil formation rate are unlikely to remain constant over time and this is exploited in Quaternary soil stratigraphy (particularly in loess) where it is usually assumed that the two are 180° out of phase. If soil formation only takes place in warm climate periods (Morrison, 1978) and loess deposition in cold climate periods, buried soils are intercalated with little-altered sediment. This model was adopted by Palmer and Pillans (1996, pp155, 157) for their interpretation of the soil and loess stratigraphy of the west coast, North Island. They implied loess accretion and soil formation operate according to a climatically controlled on/off switch.

The superimposed effects of soil formation during sediment accumulation have received little attention. McDonald and Busacca's (1990) study of the Palouse loess in USA highlighted the fact that development of subsurface horizons had been preconditioned by features inherited during upbuilding. Intensely cicada-burrowed buried A horizons became the focus for precipitation of pedogenic carbonate leached from the overlying upbuilding soil. These horizons developed into densely cemented petrocalcic horizons. Raeside (1964) argued that the Pallic Soils (Fragiaquepts) of New Zealand inherited their dense subsoil (fragipan) from the period of loess accumulation in a cold, dry climate. The dense subsoil and seasonally perched water table above it in the soil now are therefore a result of a superposition of two distinctly different phases of soil formation. He suggests that current processes of soil formation are destroying features of the earlier phase.

Soils in Westland, on the west coast of South Island, form under a superhumid and mild temperate climate. They experience high rates of soil formation dominated by processes of podzolisation and gleying. Rapid landscape evolution due to high rainfall and tectonic uplift rates has produced arrays of geomorphic surfaces that have presented many opportunities for soil chronosequence studies. Studies by Ross et al. (1977); Smith and Lee (1984); Stevens (1968); Campbell (1975); Sowden (1986); Basher (1986) all show a consistent trend of soil evolution. On well-drained sites, Recent Soils evolve to Acid or Allophanic Brown Soils to Orthic Podzols and finally Perch-gley Podzols (New Zealand Soil Classification; Hewitt, 1992). The sequence is marked by rapid and progressive loss of mobile followed by more immobile primary and secondary soil constituents as a result of intense leaching and acidification in association with coniferous or beech forest.

It is now widely recognised that loess has accumulated on late Pleistocene and older surfaces in Westland (Almond, 1996; Bruce et al., 1973; Mew and Lee, 1981; Mew et al., 1988a; Mew et al., 1988b) albeit at comparatively low rates. As yet there has been no serious consideration as to how progressive upbuilding may have influenced soil formation partly because some of the chronosequence studies were on surfaces too young to have loess and also because some studies predate the recognition of loess.

Campbell (1975) in his study near Reefton (MAR of 2000 mm), concluded that a pedogenic threshold is reached sometime between 20 and 50 ka whereby soil development shifts from a lateritic pathway (concentration of iron) to a podzolic pathway (concentration of Si). The change in direction was suggested to be strongly influenced by the local southern beech vegetation (*Nothofagus sp.*) which has a high potential to acidify soils once their reserves of weatherable minerals are depleted. However the soils either side of the inferred threshold are formed in sandy and gravelly alluvium (Ahaura Soil) on the one hand, and loess on the other (Campbell's Kumara Soil). The threshold may not be intrinsic as thought but extrinsic, determined by the change in style of pedogenesis from topdown to upbuilding.

Walker and Syers (1976) developed a model of the fate of P during pedogenesis based partly on the study of Stevens (1968) near Franz Josef in south Westland (Fig. 1). Their interpretations were that total P declines monotonically from soil inception with little change after soils reach the persistent Perch-gley Podzol form. The initial rapid decline of total P principally reflects the leaching of acid extractable P (primary apatite P). Occluded, non-occluded and organic P fractions increase while primary P declines. After about 10 000 years of soil formation primary P is still in decline and organic and non-occluded P fractions have peaked. Beyond this point to 22 000 yrs (the age ascribed to the oldest soil in the sequence) primary and non-occluded P decline to zero while occluded P gradually increases. Although their treatise is widely accepted and supported in the literature (Birkeland, 1984; Cross and Schlesiger, 1995; Lajtha and Harrison, 1995; McLaren and Cameron, 1996), it is has become clear there are limitations to its general application even in the environment where the data were collected. Walker and Syers (1976) oldest soil (Okarito Soil) is formed in loess which may potentially predate the Last Interglacial (ca. 125 ka) (P. Almond unpublished information). Landsurfaces intermediate in age between that of the second oldest soil (Mapourika site ca. 12 ka) and the Okarito Soil have received loess from early to late Last Glaciation (Almond, 1996). Extrinsic inputs of loess may have been sufficient to offset the losses documented by Walker and Syers (1976) particularly since P is tightly cycled (Wood et al., 1984). The potential for an upbuilding style of pedogenesis could have significant implications for the understanding of ecosystem dynamics and plant succession.

In this study I investigate the influence progressive upbuilding has had on soil development in a region of relatively slow loess accretion and rapid soil formation. These circumstances are most likely to highlight any differences engendered from the two styles of soil formation. Morphological and chemical properties of soils formed by topdown pedogenesis (simple soils) are compared with soils formed by upbuilding. The latter formed with the accumulation of

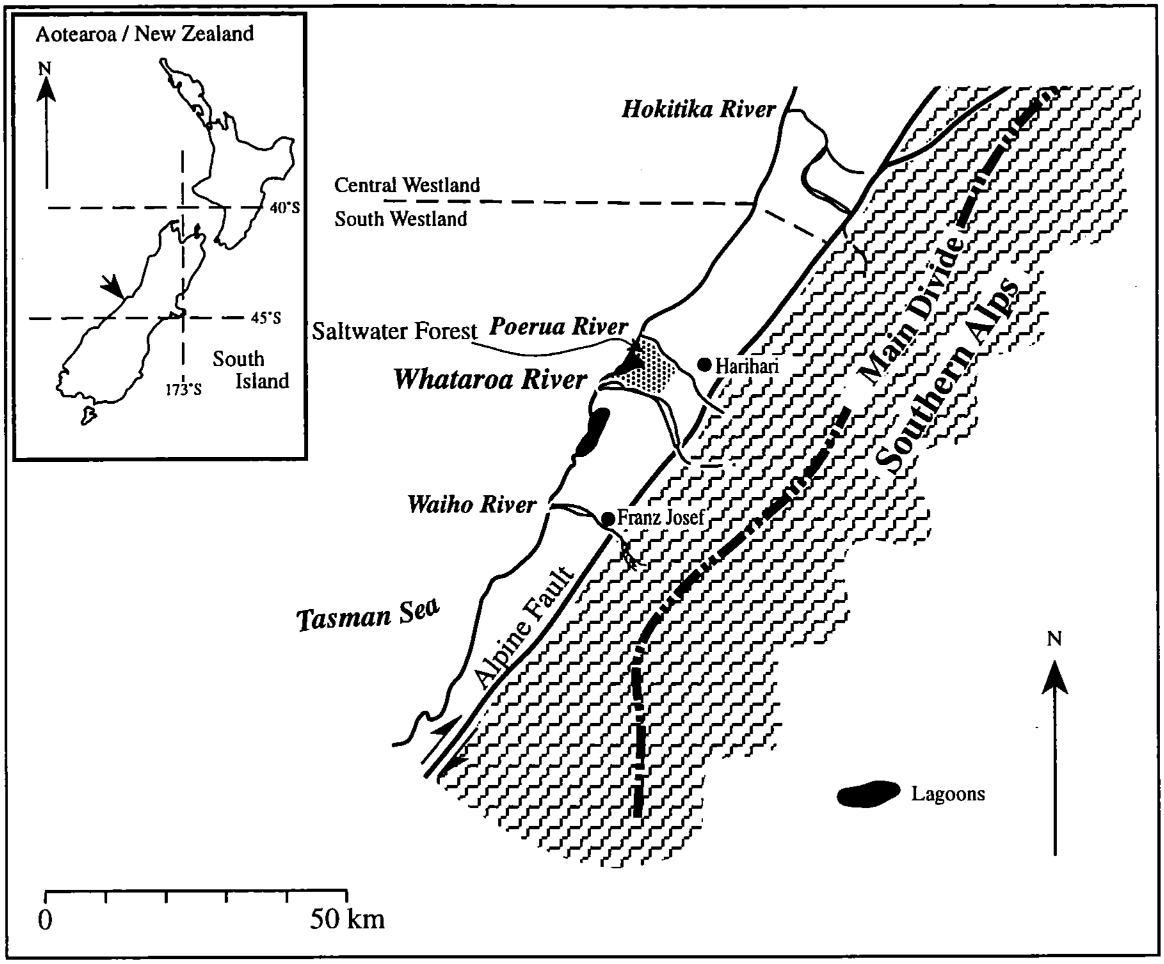


Fig.1. Location map showing Saltwater Forest.

loess in the late Pleistocene but have been partially overprinted by subsequent topdown pedogenesis during the Holocene. Comparisons are made in two pedogenic settings.

2. Study area and selected soils

The study was carried out in Saltwater Forest which occupies the area between the Whataroa and Poerua Rivers in south Westland on the west coast of South Island (Fig, 1). Saltwater Forest is a conifer/broadleaf forest dominated by rimu (*Dacrydium cupressinum*), a podocarp (Norton and Leathwick, 1990; Six Dijkstra et al., 1985). The climate is mild temperate with a high rainfall distributed evenly throughout the year. Mean annual rainfall at Harihari, 10 km to the east, is 3789 mm (New Zealand Meteorological Service, 1980). However a mean annual rainfall range of 2500 to 5000 mm is likely over the study area because it straddles part of a steep rainfall gradient from the coast to the main divide of the Southern Alps in the southeast (Griffiths and McSaveny, 1983). Mean monthly temperatures range from 16.3° C in January to 6.9° C in July (New Zealand Meteorological Service, 1980). The climate in the Last Glaciation, when upbuilding soils were forming, would have been colder than present, reaching a maximum temperature depression of ca. 4.5 °C in the Last Glacial Maximum (Soons, 1979). Annual temperatures were within 2°C of present by 12 ka, and at least equivalent to those of today by 10 ka (McGlone et al., 1993). Annual rainfall is likely to have been as high during the Last Glaciation as it is today (Soons, 1979). Since rainfall is a major determinate of the rate of pedogenesis (see Basher, 1986), rates of pedogenesis during the Last Glaciation are likely to have been high.

Landforms are principally glacial in origin and include late Pleistocene moraines and outwash terraces (Almond, 1996). The coastline has prograded in the Holocene as sediment has been deposited on floodplains and in swamps and dunes surrounding Saltwater Lagoon (Fig. 2). The oldest landforms are a Penultimate Glaciation terminal moraine (M1) and early Last Glaciation (Otiran) terminal moraines (M2 and M3) lying parallel to the present coastline in a nested fashion. Behind these moraines, the next oldest landform is a mid-Otiran aggradation terrace (T4). The moraine associated with the T4 terrace has been buried by a late Otiran terminal moraine (M5) formed by a lobe of the Whataroa glacier. Glaciers during this advance extended in narrow tongues down the Whataroa and Poerua valleys leaving steep-sided sinuous lateral moraine ridges on the valley margins. The most recent advance followed soon after M5 and its moraines (M6) are nested within M5 lateral moraine ridges in the valleys. It was not large enough to extend ice lobes on to the area of Saltwater Forest, as the M5 advance had, but meltwater rivers incised into T5 to form mostly degradational alluvial surfaces (T6) (Fig. 2).

Sediment transported by glaciers and rivers is almost entirely derived from the cirques and valley walls of the Southern Alps. Lithologies grade from garnet and biotite schists adjacent to the Alpine Fault (Fig. 1) on the Alp's western margin into chlorite schists and eventually to greywacke sandstone and argillite at the main divide (Warren, 1967). Glacial till is a very compact, matrix supported diamicton comprising mostly angular to subangular, schist clasts in a sandy loam to silt loam matrix. Terrace alluvium is characteristically a clast supported, coarse sandy gravel overlain with up to 80 cm (but usually less than 50 cm) of silty overbank deposits. This sort of material is presumed to have been the major source of loess in the Pleistocene .

Loess is absent on M6, T6 and M5, T5 then progressively increases in maximum thickness from T4 to M1 except for M2. The M2 moraine is steep with a very narrow crest and loess is thin or absent. Loess thickness is as much as 1 m on T4, 2 m on M3 and 3 m on M1.

For this study I selected two soils from each of the following landforms:

1. loess free terraces, one from T6 and one from T5
2. loess mantled terrace, two from T4
3. loess free moraines, one from M6 (moraine ridge crest) and one from M5 (shoulder slope)
4. loess mantled moraine, two from M3 (moraine ridge crests)

to give a total of eight analysed profiles. Numbers 1 and 3 are examples of simple soils on

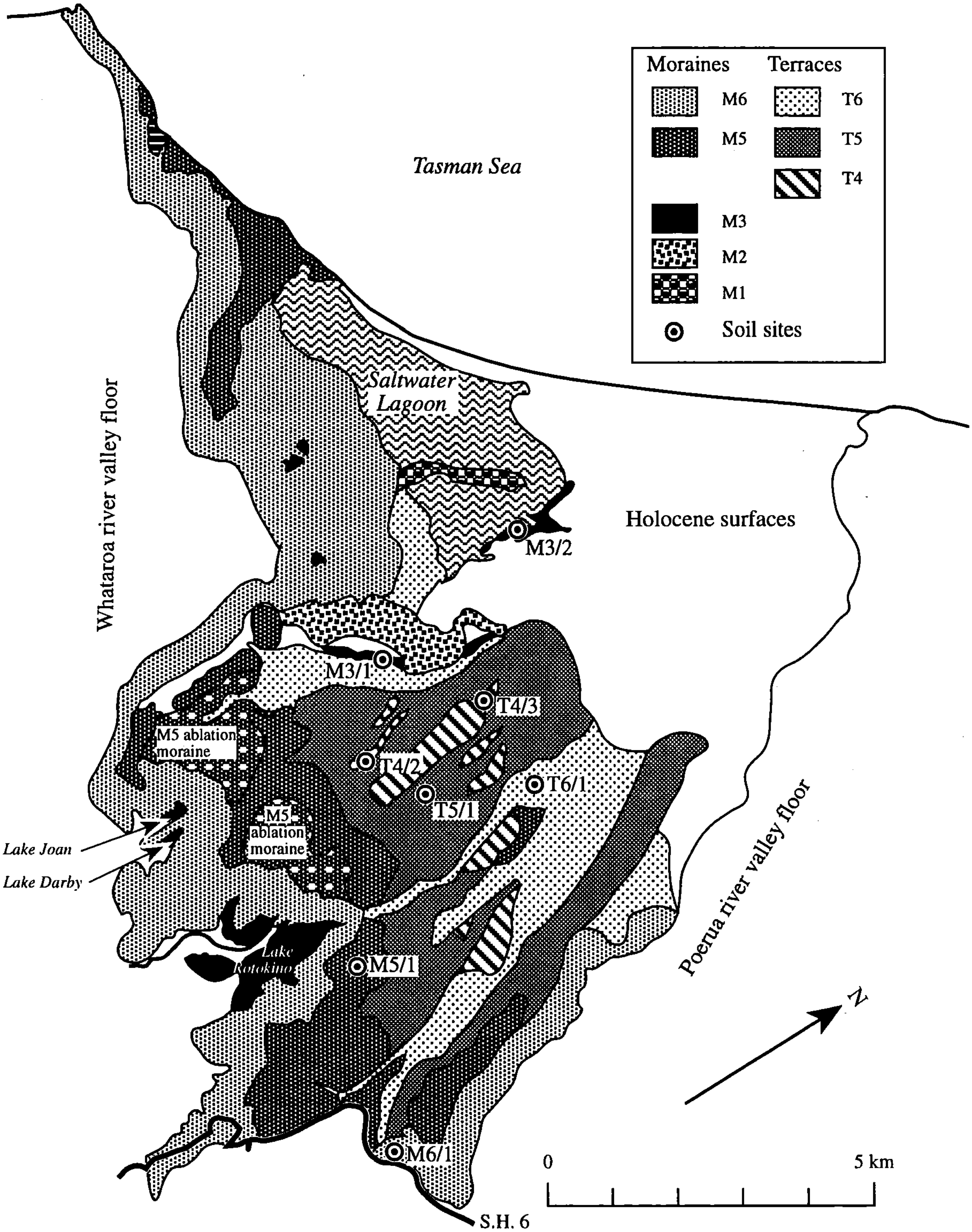


Fig. 2. Landforms and soil profile site locations within Saltwater Forest.

terraces and moraines respectively while 2 and 4 are examples of upbuilding soils on terraces and moraines respectively. Descriptions and samples were taken from fresh exposures from soil pits except in the case of T4/3 which was obtained by excavating a track cutting. I used the horizon nomenclature of Clayden and Hewitt (1989) and a description methodology based mostly on Hodgson (1976).

Figure 3 depicts soil morphological features and stratigraphic relationships between soils. Soils on terraces are very poorly-drained Perch-gley Podzols (Alaquods); Humose Perch-gley Podzols on T5 and T6 and Silt-mantled Perch-gley Podzols on T4. Soils on T6 began forming sometime after 14 ka whereas soils on T5 began forming sometime between 16.5 and 18 ka (Almond, 1996). The difference in soil age has no apparent effect on soil morphology. On T4 soil formation into silty alluvium on gravels began sometime before 36 ka producing amongst other soil features a well expressed, peaty O horizon (b3O Fig. 3). After 36 ka but before 22.6 ka about 35 cm of loess (L1b) accumulated (Almond, 1996). The 22.6 ka datum is provided by microscopic occurrences of Aokautere Ash erupted from central North Island. The soil morphological features developed in L1b depended on site drainage (Almond, 1996). In very wet sites the loess is a peaty or humic silt loam (see T4/2) with a subtle soil stratigraphy whereas on slightly better drained sites peaty or humic A or O horizons are separated by distinct olive grey or blue grey, silt loam Er horizons (see T4/3). This difference reflects a change, between sites, of the balance between loess input and organic matter decomposition rates (Almond, 1996). From the time of the deposition of Aokautere ash through to about 16 ka, 35 to 40 cm of loess (L1a) accumulated relatively rapidly. Surface A horizons and underlying blue-grey Er horizons have developed in L1a.

Soils on M3 are Placic Perch-Gley Podzols (Placorthods) comprising an (H)/Ah/Er/Bfm/Bs horizon sequence while soils on M6 and M5 are Acidic Allophanic Brown Soils (Dystrochrepts) with an AE/Bw/C horizon sequence. Soils on M5 and M6 are presumed to have begun forming at about the same time as their equivalents on T5 and T6. In contrast to the soils on terraces, soils on moraines have oxidised lower subsoils, and consequently the loess and soil stratigraphy apparent on T4 is manifested differently on M3. Organic matter in buried soils mineralises in nearly all soils so soil stratigraphy is more subtle. Whereas the boundary between L1a and L1b is defined soil-stratigraphically on terraces, the same boundary can only be approximated by counting glass shards in soil increments - the boundary is placed at the zone of maximum glass counts. The lower boundary of L1b, while coinciding with the top of a distinctly peaty horizon on T4 (see b3O in T4/3), is also difficult to locate. In M3/1 it is placed immediately above a bright, buried Bs horizon (bBs), assuming this horizon formed not far below a relatively stable soil surface. In M3/2 the boundary is impossible to locate, although on the evidence of a TL date of 36 ± 3.4 ka (G.L. Berger pers. comm.) at 80 cm below the mineral soil surface, the same boundary is placed at 70 cm depth.

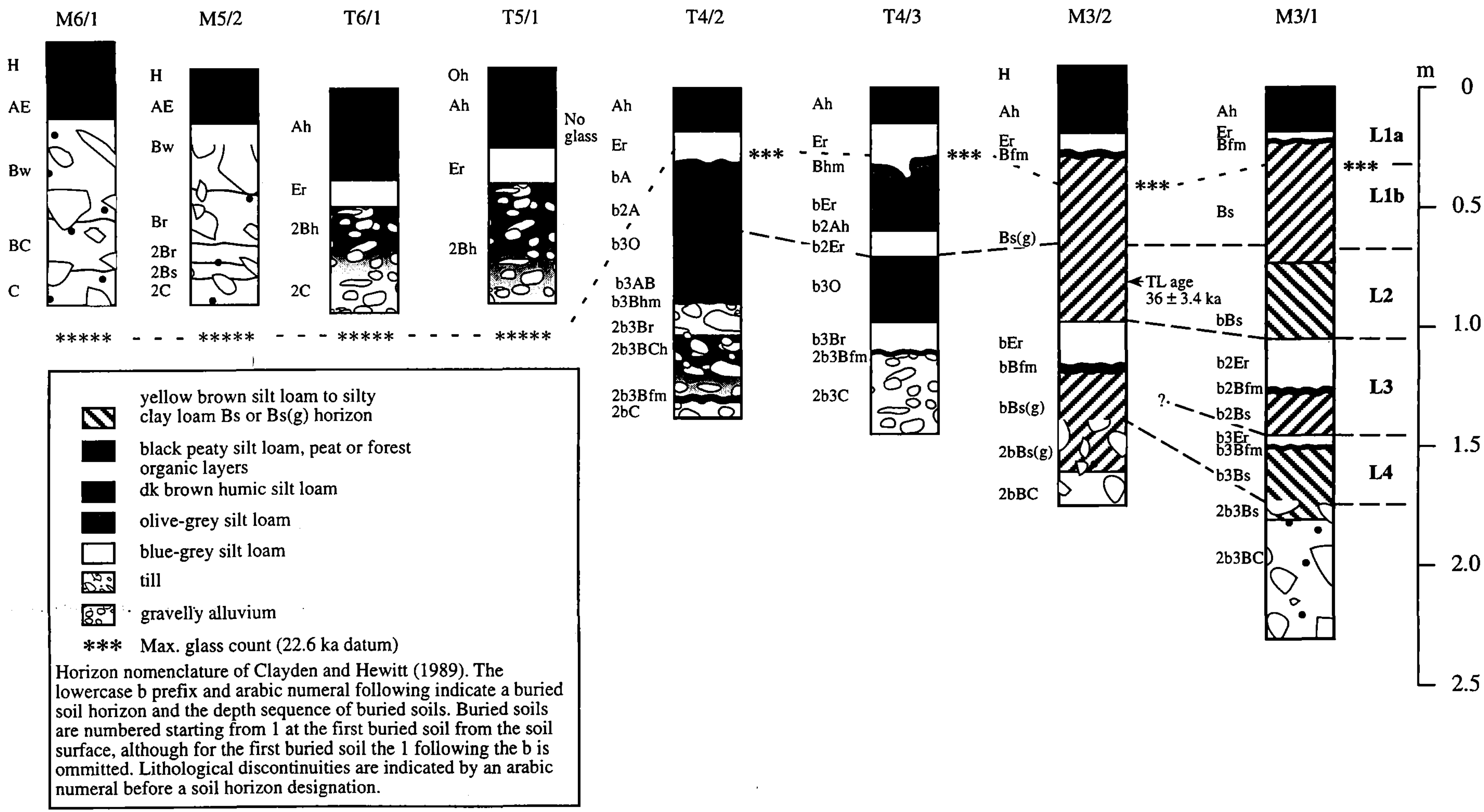


Fig. 3. Soil morphological properties and stratigraphic relationships between soils.

The lower boundary of L2 at about 1 m coincides with the top of a distinct buried soil (b2 prefix) with the same horizon sequence as the surface soil, apart from the absence of an A or H horizon. It should be understood from a buried E horizon designation (e.g. bE or b2E) that this includes the former A horizon which is no longer apparent as such. Horizons are classified according to how they appear now.

The loess beneath L2 is divided into two loess sheets, L3 and L4, because of the presence of a buried soil (b3 prefixes) at 145 cm in M3/1. The tills of M3 are correlated with glacial advances of Oxygen Isotope Stage 4 and it is likely that L3 and L4 accumulated towards the end of this stage. The obvious buried soil in L3 probably formed from the beginning of Oxygen Isotope Stage 3 at ca. 60 ka and was buried by L2 sometime afterward, through to about 36 ka.

In each pedogenic setting the soils formed by topdown pedogenesis (simple soils) are an analogue for the morphological and chemical features developed in the upbuilding loess soils in their Holocene phase of topdown pedogenesis. Thus, it should be possible to isolate the effects of topdown from upbuilding pedogenesis.

3. Methods

Soils were channel sampled for chemical analysis in contiguous increments of up to 20 cm thickness, by soil horizon. I estimated bulk density in all soils except M3/2 by core sampling (M3/1) and gamma probe (the rest). Gamma probe (Erbach, 1987) was favoured because of problems fluidity, roots and stones presented for core sampling. In M3/1 the remoteness ruled out using the gamma probe, but fortunately the uniformly silty soil was suitable for core sampling. Samples for chemical analysis were air dried, sieved to 2 mm and analysed for pH, organic carbon, oxalate extractable Fe and Al, P fractions and total Zr.

Soil pH was measured with a combination electrode from a 1:2.5 soil to distilled water suspension after it was shaken and left overnight to equilibrate. Soil organic carbon was measured colorimetrically after oxidation of 1 g of finely ground soil by the method given by Blakemore et al. (1987) (modified from Walkely and Black, 1934). Fe and Al were extracted by shaking 2 g of soil for 4 hours (in the dark) with 50 ml of pH 3 ammonium oxalate reagent. Fe and Al concentrations were measured by atomic absorption spectrometry (McKeague and Day, 1966 as described in Blakemore et al., 1987).

Total soil phosphorus (P_{tot}) was measured by NaOH fusion (Smith and Bain, 1982) and organic P (P_o) by acid and alkali extraction according to the method of Bowman (1989). Inorganic P (P_i) was measured and divided into an Fe/Al associated form ($P_{Fe/Al}$) and a primary apatite form (P_{Ca}) by sequentially extracting with NaOH followed by dilute HCl. The method is an adaptation of that of Tiessen and Moir (1993). Recalcitrant forms of P (P_r), including occluded P and non-extractable organic P, were calculated by difference of the sum of P_o and P_i from P_{tot}.

Samples were analysed by X-ray fluorescence for total Zr. These data are used for assessment of relative enrichment or depletion of P in profiles.

4. Results and Discussion

4.1. Soil pH and Organic Carbon

4.1.1. Moraines (Fig. 4a, b)

Soil pH and percentage organic carbon (%OC) profile trends in simple soils (M6/1 and M5/2) are consistent with a decreasing intensity of weathering and quantity of organic inputs, with increasing depth. Soil pH rises from around 4.0 in AE horizons to 5.0 in Bw horizons and remains essentially constant into the parent material. Organic carbon decreases monotonically from 5.0% in AE horizons to less than 1.5% by 50 cm. Depth trends in upbuilding soils are very similar although Ah horizons are more organic rich than the AE horizons of simple soils. %OC depth curves for the B horizon are superimposed on those of the simple soils and in deeper parts of profiles %OC remains almost uniformly less than 1%. Soil pH in upbuilding soils is systematically lower than simple soils. Soil pH rises from a minimum near the soil surface (3.4 in M3/1, 3.8 in M3/2) in a similar manner to simple soils but only reaches a value of 4.1 in M3/1 or 4.7 in M3/2 (depth weighted averages). Values below the average correspond to buried E horizons.

Subsoils of upbuilding soils have not retained carbon accumulated during the upbuilding phase presumably because the oxidising environment has permitted organic matter mineralisation. The relationship between soil pH and %OC in simple and upbuilding soils on moraines is demonstrated in Fig. 5a. The trend is for horizons with higher %OC to have lower pH. This is likely to be due to the acidity contributed by the organic matter and more intense weathering in surface horizons where organic matter accumulates. Subsoil pH in the upbuilding soils plots below subsoil pH in simple soils even though they have the same %OC. I attribute this to the enhanced weathering each subsoil increment of the upbuilding soils experienced when it was part of surface A and E horizons. Weathering has reduced the buffering capacity of subsoil and even after the removal of organic matter the pH has not risen to levels comparable to simple soils. The more weathered buried E horizons remain the most acid.

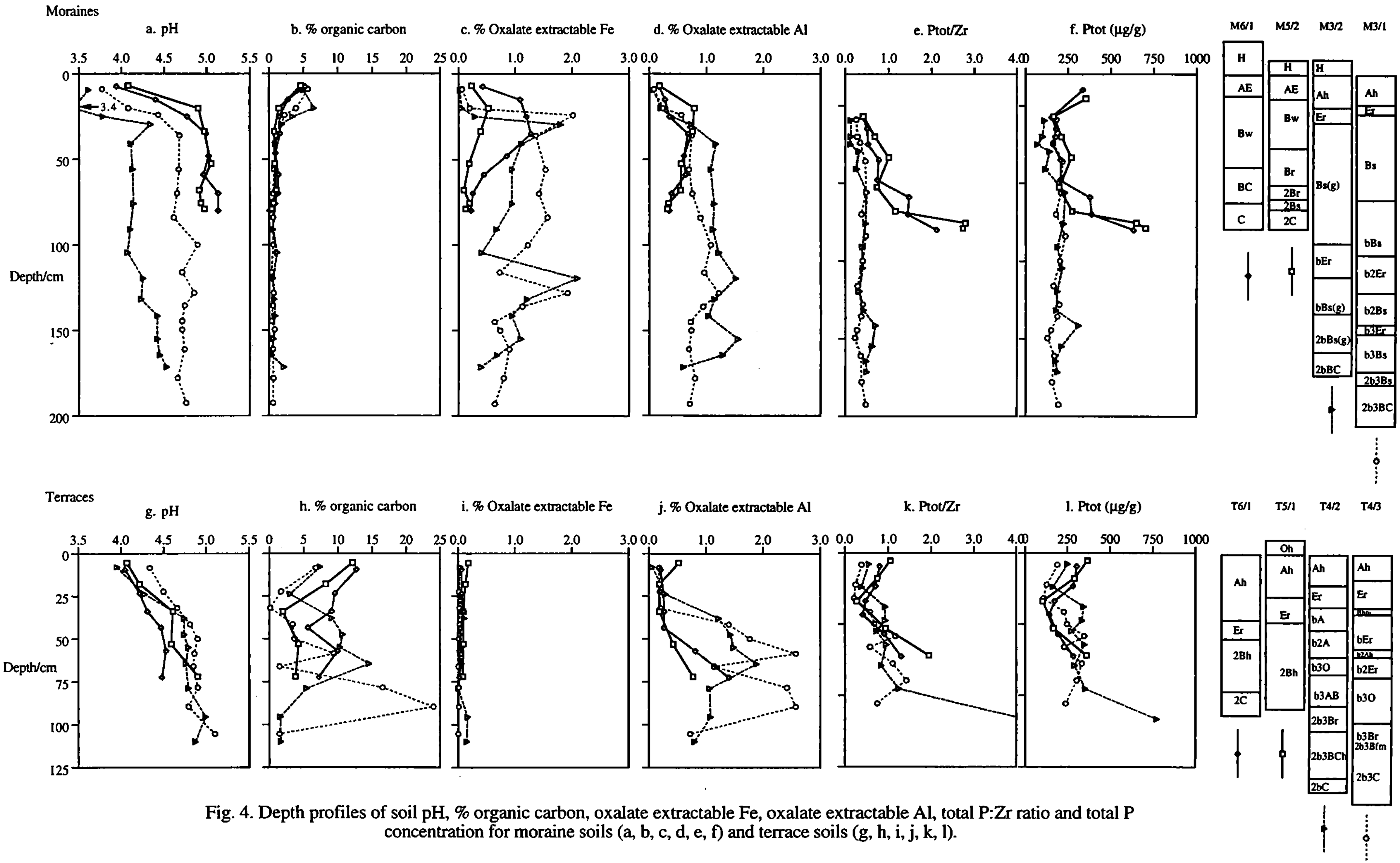


Fig. 4. Depth profiles of soil pH, % organic carbon, oxalate extractable Fe, oxalate extractable Al, total P:Zr ratio and total P concentration for moraine soils (a, b, c, d, e, f) and terrace soils (g, h, i, j, k, l).

4.1.2. Terraces (Fig. 4g, h)

Soil pH is 4.0 in the A horizons of all soils except T4/3 and rises with depth more slowly than in moraine soils. In simple soils pH rises to 4.6 in the subsoil and there is little difference in subsoil pH of upbuilding soils (4.8). Overall, %OC is higher in terrace soils than moraine soils. Soils show a trend of decreasing %OC from Ah to Er horizons (13% to <6% in simple soils, 7% to <3% upbuilding soils), then a subsequent increase in subsoils. In simple soils the increase occurs in the organic illuvial 2Bh horizon whereas in upbuilding soils it corresponds to buried A or O horizons. Highest %OC carbon for all soils and all depths is in the b3O horizons of T4/2 (14.5%) and T4/3 (24.0%).

Terrace soils are acidified to greater depth than their moraine soil counterparts which probably reflects the topographic setting. On the terraces there is deeper percolation of acidic dissolved or colloidal organics than on moraines where, because of the slope, there is greater lateral water movement. This is confirmed by the absence of subsurface illuvial Bh horizons in moraine soils. The relationship between pH and %OC has the same trend as in moraine soils; the more OC the lower the pH. However, there is more scatter than for the moraine soils and there is no appreciable difference between the upbuilding soils and the simple soils. It appears that acidification in the topdown phase of pedogenesis has matched or exceeded that which occurred during the upbuilding phase so that the pH profiles of upbuilding and simple soils are very similar. The buried A or O horizons in the upbuilding soils stand apart in the pH vs %OC plot (Fig. 5b); they have high %OC and yet relatively high pH (4.7-4.9). The organic matter comprising these horizons is not as acid as that from the current podocarp vegetation, perhaps reflecting the degree of decomposition of the organic matter, or the fact that it is derived from herbaceous plants and ferns that grew on the site in the mid-Otiran (ca. 36 ka, Almond, 1996).

4.2. Oxalate extractable Fe and Al

4.2.1. Moraines (Fig. 4c, d)

Profile depth trends of oxalate extractable Fe (Fe_{OX}) and oxalate extractable Al (Al_{OX}) in simple soils and the upper part of upbuilding soils are consistent with 'topdown' podzolisation processes (Childs et al., 1983; McKeague and Day, 1966). Fe_{OX} in all soils shows a rapid increase from A horizons (<0.5%) to B horizons. In simple soils, Fe_{OX} reaches a maximum in the Bw horizon, immediately beneath the AE horizon in M5/2 (0.53%), and towards the base of the Bw horizon in M6/1 (1.28%). Below the maxima Fe_{OX} declines rapidly into the C horizon. Al_{OX} depth trends are similar to Fe_{OX}. Maxima in Al_{OX} coincide with those of Fe_{OX} but the decrease with depth below maxima is more gradual indicating that Al is more mobile than Fe. Al_{ox} is lowest in the most acid upper horizons and accumulates where soil pH exceeds 4.5. At normal soil Al^{3+} concentrations, solubility of inorganic monomeric Al rapidly declines above pH 4.7 whereupon amorphous Al may precipitate (Funakawa et al., 1993) or, if Si concentrations are high enough, above pH 4.8-5.0 imogolite and proto-imogolite allophane may form (Farmer et al., 1980).

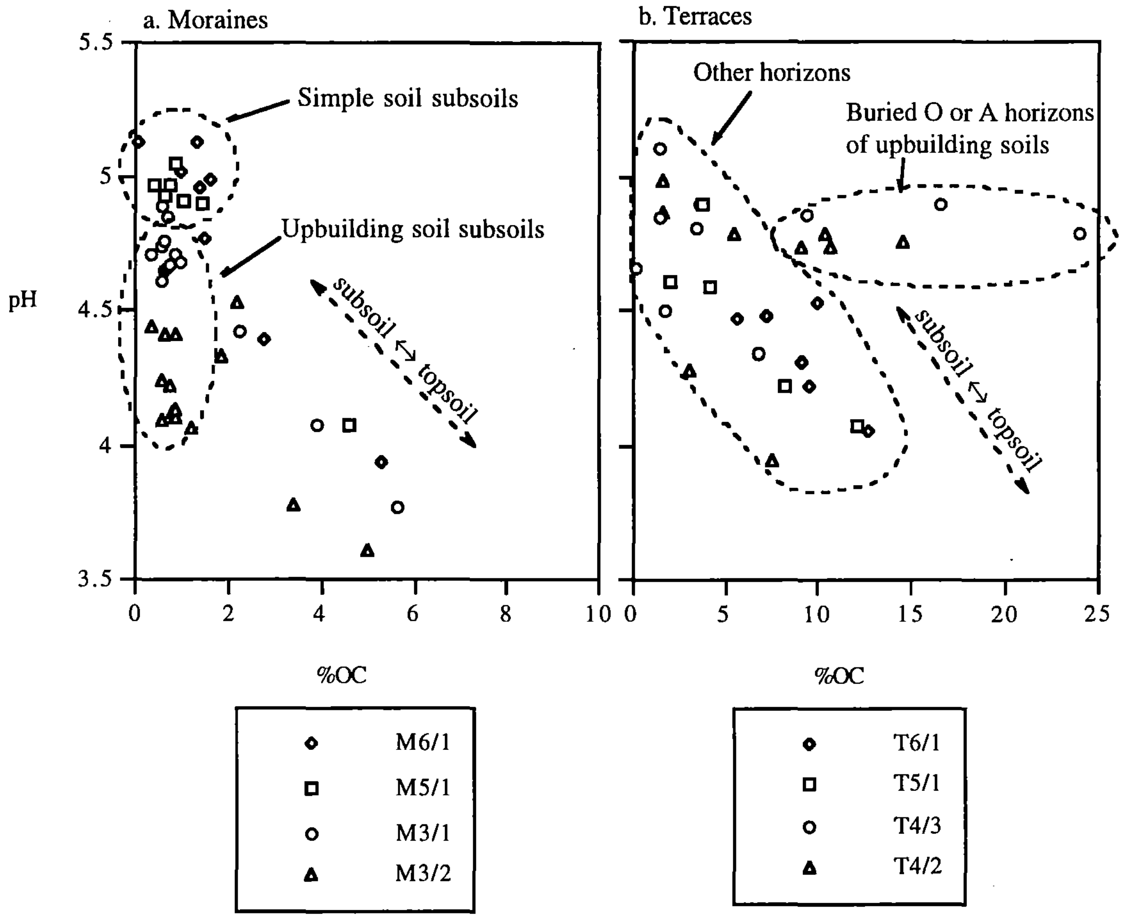


Fig. 5. Soil pH vs percentage organic carbon (%OC) for simple and upbuilding soils on a. moraines and b. terraces.

Fe_{ox} is lower in A horizons of upbuilding soils than simple soils and rises to greater concentrations (1.5-1.8%) in the Bfm horizons. Below this horizon, rather than a steady decline, Fe_{ox} plateaus at > 1% until about 80 cm where it declines into a buried E horizon (b2E in M3/1 and bE in M3/2). The discrepancy between the depth trends for the simple and upbuilding soils stems from accumulation of Fe_{ox} during the upbuilding phase. Each increment of soil loses primary Fe by weathering and downward or lateral translocation when it is at the soil surface. After that increment is buried it becomes the illuvial increment for poorly ordered secondary Fe compounds precipitated from Fe mobilised from above. In other words, the upbuilding profile leaves poorly ordered Fe compounds as a trail immediately beneath the upbuilding A/E horizons.

Al_{OX} depth trends in upbuilding soils are similar to the simple soils to about 70 cm depth but below this depth Al_{OX} remains above 0.7% to the base of the loess. Similar process of weathering and translocation during upbuilding, described above for Fe, are responsible for Al depth trends. Variations in subsoil Al_{OX} include peaks in buried Bfm horizons and minima in buried E horizons, but the amplitude of variation into the buried E horizons is much less than in surface soils or for Fe_{OX} variation. This could be due to resaturation of buried E horizons from Al_{OX} translocated from horizons above, or a low intensity of Al translocation at the time the buried E horizons formed. The latter explanation is supported by total Al:Si ratios (P. Almond unpublished information) which indicate that the buried E horizons were not as severely podzolised as the surface E horizon.

4.2.2. Terraces (Fig. 4i, j)

Fe_{OX} is extremely low in all terrace soils. The highest levels (0.2%) are in the A horizon of T5/1 and in the lower-most subsoil of T4/2 (2b3Br horizon). Secondary Fe has remained mobile because of reducing conditions and subsequently lost to vertical and lateral leaching. Al_{OX} is low, generally < 0.2%, in the Ah and Er horizons of all profiles and then increases in underlying horizons where pH and %OC increase. In simple soils this coincides with the organic illuvial 2Bh horizons. In upbuilding soils Al_{OX} increases sharply in the buried A and O horizons of L1b and varies in step with %OC. Importantly, in all horizons of accumulation for all profiles 100% of Al_{OX} is also extractable by Na pyrophosphate (P. Almond unpublished data) indicating that it is organically complexed.

Together the data above show that current topdown pedogenic processes involve the complete loss of Fe in the aquic moisture regime, and the translocation of Al from acid upper parts of soil profiles to organic-rich lower soil horizons. This is a common finding for poorly-drained soils in Westland and other high rainfall regions (see for example Young, 1988; Mew and Lee, 1988; Mew and Lee, 1981). The simple soils show that the topdown processes of podzolisation on the poorly-drained terraces remove Al to a greater depth than on the well-drained moraines. This may be due to either a greater depth of acidification or deeper percolation on the flat terraces of organic leachates capable of chelating with aluminium, or both. The depth of loess on T4 combined with the depth of translocation of Al has meant that any profile trends inherited during upbuilding cannot be distinguished from those formed by topdown pedogenesis. The fact that the peaks in Al_{ox} correspond to subsoil organic-rich horizons and that all Al_{ox} is organically complexed indicate that Al may not have been retained at all had these horizons been absent. The buried A and O horizons of the upbuilding soils have been transformed during upbuilding and in the subsequent topdown phase into horizons with the chemistry indistinguishable from pedogenic Bh horizons. That these horizons are indeed buried A or O horizons is confirmed by their pollen assemblages which are characteristic of a swamp vegetation association very different to the current podocarp forest (Almond, 1996). So despite the different stratigraphy and modes of formation of the simple and upbuilding soils their evolution has effectively converged so that, at least with respect to oxalate chemistry, they are very similar.

4.3. Phosphorus chemistry

4.3.1. Moraines (Fig. 4e, f, Table 1)

P_{tot} depth trends are presented as concentrations (μg/g) and ratios of P_{tot} to the stable index element Zr. Ratios of P_{tot} to Zr make it possible to assess enrichment or depletion of an element independently of changes in total soil mass (Barshad, 1964). Total P to Zr ratios of unweathered tills and silty river alluvium are about 2 (Table 2). Sampling fresh loess to use as a standard presented many problems but a single sample was trapped on a windy day from the channel zone of the Hokitika River (Fig. 1). Although P and Zr concentrations were greater than in till or alluvium, the P:Zr ratio is again about 2 (Table 2). The ratios of P_{tot}/Zr from fresh samples indicate that all horizons except for C horizons of simple soils have lost significant amounts of P. Plots for both P_{tot} concentration and P_{tot}/Zr have similar shapes which means that, assuming initial profile uniformity of parent material, zones of higher P_{tot} concentrations are also zones of lower net P loss.

In simple soils the depth trend of P_{tot} shows a decline from a relative maximum in surface H horizons (ca. 340 μg/g) to a minimum in AE and upper Bw horizons (ca. 200 μg/g or less) increasing to a maximum in C horizons (600-700 μg/g). Organic P (P_o) and recalcitrant P (P_r) are the dominant P fractions except in: (1) the H horizons where P_{tot} is presumably wholly P_o, and (2) in BC and C horizons where primary P (P_{Ca}) becomes an increasingly important fraction with depth (up to 50% of P_{tot} in M5/2). Depth trends are consistent with downward soil formation into unaltered sediment. Organic forms of P accumulate in surface organic horizons by upward translocation in the biological cycle. P losses are greatest in the upper part of the mineral soil due to plant uptake and the combined effects of weathering and leaching

(Letkeman et al., 1996). As would be expected almost all of the P in this part of the profile is in secondary organic and inorganic forms.

In upbuilding soils P_{tot} rises from <200 μg/g in M3/1 or < 110 μg/g in M3/2 in the Ah and Er horizons to around 200 μg/g throughout the subsoil. The entire depth of the profiles is characterised by an almost complete absence of P_{Ca}. P_o is the dominant P fraction to a depth

Table 1. Bulk density and selected chemical properties of soils at all sites.

Soil	Depth	Horizon	Bulk density (g cm-3)	pH	OC (%)	Feox (%)	Alox (%)	PCa-P (μg/g)	PFe/Al (μg/g)	Pr (μg/g)	Po (μg/g)	Ptot (μg/g)	Zr (μg/g)
M6/1	-20-0	H	0.44	-	-	-	-	-	-	-	-	335	-
	0-11	AE	0.69	3.94	5.25	0.43	0.17	4	21	49	79	152	388
	11-20	Bw	1.09	4.40	2.74	1.09	0.27	2	12	68	98	179	364
	20-30	Bw	1.28	4.77	1.50	1.20	0.35	1	9	58	93	161	314
	30-40	Bw	1.22	4.99	1.60	1.28	0.67	1	14	80	111	206	269
	40-53	Bw	1.32	5.02	0.94	0.85	0.60	4	22	77	101	204	280
	53-65	BC	1.31	4.96	1.34	0.45	0.63	74	73	118	112	377	257
	65-75	BC	1.35	5.13	1.30	0.25	0.38	82	179	101	25	387	269
	75-85	C	-	5.13	0.04	0.22	0.34	222	350	51	9	631	299
M5/2	-9-0	H	0.44	-	-	0.06	0.07	-	-	-	-	354	-
	0-14	AE	1.07	4.08	4.59	0.23	0.18	2	15	95	53	165	406
	14-26	Bw	1.54	4.90	1.45	0.53	0.78	2	14	120	75	211	314
	26-41	Bw	1.60	4.97	0.76	0.39	0.75	2	15	182	70	269	270
	41-64	Br	1.66	5.05	0.82	0.18	0.55	1	11	118	67	197	274
	64-72	Br	1.75	4.91	1.02	0.09	0.54	32	24	162	56	273	238
	72-79	2Bs	1.80	4.93	0.61	0.19	0.33	364	104	168	13	648	233
	79-90+	2C	-	4.97	0.39	0.13	0.31	385	156	161	0	702	258
M3/1	0-18	Ah	0.42	3.77	5.65	0.06	0.08	1	28	86	66	180	718
	18-22	Er	1.05	4.08	3.91	0.19	0.23	1	11	76	87	175	634
	22-26	Bf	0.93	4.43	2.25	2.02	0.56	1	11	50	101	162	486
	26-46	Bs	1.13	4.68	0.97	1.36	0.74	1	10	97	109	217	479
	46-66	Bs	1.21	4.67	0.71	1.54	0.69	0	13	82	113	208	444
	66-74	Bs	1.12	4.65	0.59	1.42	0.75	0	14	70	96	179	491
	74-94	bBs	1.03	4.61	0.57	1.57	0.89	1	18	115	100	233	497
	94-106	bBs	1.18	4.89	0.58	1.22	1.07	2	29	81	90	202	518
	106-126	b2E	1.29	4.71	0.33	0.72	0.95	1	23	68	72	164	586
	126-130	b2Bf	1.12	4.85	0.67	1.92	1.21	1	30	83	84	199	505
	130-141	b2Bs	1.30	4.74	0.54	1.13	0.94	1	28	76	82	187	521
	141-148	b3E	1.32	4.71	0.34	0.64	0.72	1	27	74	47	150	562
	148-151	b3Bf	1.25	4.71	0.83	0.73	0.73	1	30	56	41	128	588
	151-171	b3Bs	1.32	4.74	0.58	0.90	0.69	1	35	87	45	169	488
	171-185	b3BC	1.32	4.66	0.66	0.80	0.80	2	41	61	52	155	428
	185-200	2b3BC	1.36	4.76	0.61	0.64	0.71	2	54	94	43	193	419

Table 1 contd.

Soil	Depth	Horizon	Bulk density (g cm-3)	pH	OC (%)	Feox (%)	Alox (%)	PCa-P (μg/g)	PFe/Al (μg/g)	Pr (μg/g)	Po (μg/g)	Ptot (μg/g)	Zr (μg/g)
M3/2	-9-0	H	-	-	-	-	-					-	-
	9-28	Ah	-	3.61	4.97	0.02	0.06	2	31	0	88	109	854
	27-31	Er1	-	3.43	6.51	0.05	0.20	1	15	0	85	100	756
	31-37	Er2	-	3.78	3.39	0.29	0.38	0	9	0	91	69	619
	37-40	Bfm	-	4.34	1.81	1.80	0.73	1	14	31	96	142	464
	40-55	Bs(g)	-	4.11	0.84	1.11	1.16	1	20	7	89	117	472
	55-75	Bs(g)	-	4.13	0.81	0.94	1.07	1	21	115	96	233	478
	75-95	Bs(g)	-	4.14	0.87	0.94	1.13	0	33	89	100	223	487
	95-105	Bs(g)	-	4.10	0.55	0.67	1.11	0	36	79	71	187	493
	105-124	bEr	-	4.07	1.18	0.41	1.21	1	32	117	67	215	549
	124-133	bBfm	-	4.25	0.59	2.09	1.51	1	40	98	47	186	605
	133-148	bBs(g)	-	4.23	0.75	1.21	1.14	0	48	90	41	179	442
	148-153	2bBs(g)	-	4.42	0.85	0.94	1.03	1	72	171	65	309	454
	153-170	2bBs(g)	-	4.42	0.62	1.10	1.55	1	79	60	70	211	348
	170-177	2bBC	-	4.45	0.35	0.68	1.28	2	79	54	41	176	378
	177-184	2bBC	-	4.53	2.14	0.40	0.59	0	10	68	108	187	394
T6/1	-2-0	L,H	-	-	-	-	-	-	-	-	-	-	-
	0-15	Ah	0.89	4.05	12.7	0.06	0.20	5	29	135	137	305	384
	15-28	Ah	0.97	4.22	9.57	0.06	0.25	3	21	120	139	283	407
	29-36	Ah	1.04	4.31	9.09	0.10	0.27	0	17	49	109	175	372
	36-47	Er	1.13	4.47	5.62	0.04	0.26	1	6	50	85	143	354
	47-63	2Bh	1.34	4.53	9.98	0.02	0.81	2	20	53	119	194	216
	63-80	2Bh	-	4.48	7.21	0.02	1.40	16	68	78	121	283	220
T5/1	-9-0	Oh	0.55	-	-	-	-	-	-	-	-	-	-
	0-11	Ah	0.67	4.07	12.1	0.18	0.52	2	22	128	216	368	352
	11-25	Ah	0.90	4.22	8.26	0.13	0.18	1	27	90	174	291	393
	25-43	Er	1.41	4.61	1.98	0.09	0.18	1	3	20	81	105	385
	43-63	2Bh	1.80	4.59	4.23	0.09	0.42	2	23	82	60	168	180
	63-81	2Bh	-	4.90	3.81	0.08	0.77	48	143	109	60	360	185
T4/2	0-16	Ah	0.91	3.95	7.42	0.03	0.06	22	3	130	97	251	463
	16-32	Er	1.36	4.28	3.03	0.03	0.28	18	1	59	86	165	444
	32-44	bA	1.17	4.74	9.12	0.10	1.22	18	1	90	234	343	371
	44-51	b2A	1.42	4.74	10.7	0.03	1.43	21	1	87	225	334	362
	51-59	b2A	1.29	4.79	10.3	0.05	1.48	23	1	47	203	274	380
	59-70	b3O	1.14	4.76	14.5	0.05	1.88	25	3	78	241	347	366
	70-88	b3AB	1.01	4.79	5.41	0.00	1.06	40	1	52	195	289	351
	88-103	2b3Br	1.23	4.99	1.61	0.16	1.08	202	5	84	60	351	283
	103-130	3b3BCh	-	4.87	1.66	0.15	0.80	596	46	107	23	771	184

Table 1 contd.

Soil	Depth	Horizon	Bulk density (g cm-3)	pH	OC (%)	Feox (%)	Alox (%)	PCa-P (μg/g)	PFe/Al (μg/g)	Pr (μg/g)	Po (μg/g)	Ptot (μg/g)	Zr (μg/g)
T4/3	0-15	Ah	0.86	4.34	6.76	0.04	0.18	3.89	14	100	73	191	507
	15-26	Er	1.20	4.50	1.74	0.01	0.19	9.38	5	40	73	128	516
	26-38	Er	1.25	4.66	0.16	0.02	0.20	1.35	11	35	62	110	500
	38-45	Bh	1.13	4.81	3.42	0.03	1.40	2.85	14	0	217	229	402
	45-56	bEr	1.13	4.90	3.68	0.01	1.77	2.69	25	67	153	248	369
	56-60	b2A	0.96	4.86	9.38	0.04	2.57	6.90	26	84	231	347	299
	60-71	b2Er	1.09	4.85	1.50	0.00	1.14	3.30	31	78	118	230	405
	71-84	b3O	0.83	4.90	16.5	0.01	2.42	5.99	20	81	225	332	303
	84-95	b3O	0.75	4.79	24.0	0.01	2.57	6.00	13	97	186	302	214
	95-113	b3Br	-	5.10	1.46	0.00	0.72	1.65	50	95	92	239	325

of about 60 cm below which adsorbed P ($P_{Fe/Al}$) gradually increases so that at the base of the profiles $P_{Fe/Al}$, P_r and P_o are of roughly equal proportions. The amounts and proportions of P_r remain relatively consistent except in the upper 45 cm of M3/2 where it is absent or of low concentration.

An obvious difference between the simple soil and upbuilding profile P trends is the persistence of P_o at depth in the latter. By 90 cm in the simple soils P_o is negligible. Simple soil P_{tot} profile trends show that the depth of strong alteration during the Holocene phase of topdown pedogenesis reached a depth of between 80 cm to 1 m. Low P_{tot} values extend beyond this depth in the upbuilding soils and at no depth are there P_{tot} or P_{Ca} values indicative of unaltered material (Table 2). There is no reason to believe the front of weathering would have extended significantly deeper in the upbuilding soils, thus it appears that the depth profile P trends in upbuilding soils owe much to the upbuilding phase of pedogenesis. The same observation was made with regard to soil pH.

Table 2. Total P and Zr measured by XRF for unweathered alluvium, till and loess samples.

Sample	P	SD	Zr	SD	P/Zr
Fresh silty and sandy alluvium (n=14)	480	46	245	28	2
Fresh till (n=4)	462	59	251	9	2
Loess (n=1)	624		332		2

Average loess accretion rates calculated from M3/1 are: L1a 5.7 x 10^4 g m^{-2} ka^{-1}, L1b 3.6

x 10^4 g m^{-2} ka^{-1}, L2 2.9 x 10^4 g m^{-2} ka^{-1}, L3+ L4 10 x 10^4 g m^{-2} ka^{-1} which equate to 36, 22, 18 and 62 g of P m^{-2} ka^{-1} respectively, assuming a P_{tot} concentration in fresh loess of 624 µg/g (Table 2). These rates, calculated from present soil mass, are minimum values since mass has been lost as a result of weathering and leaching. When compared to P data and rates of P loss from 75 cm of soil modelled from the Franz Josef chronosequence (Stevens, 1968) (Fig. 6), they demonstrate that inputs of P from loess in the study area are only of the magnitude of losses from strongly developed soils. The comparison should be valid since mean annual

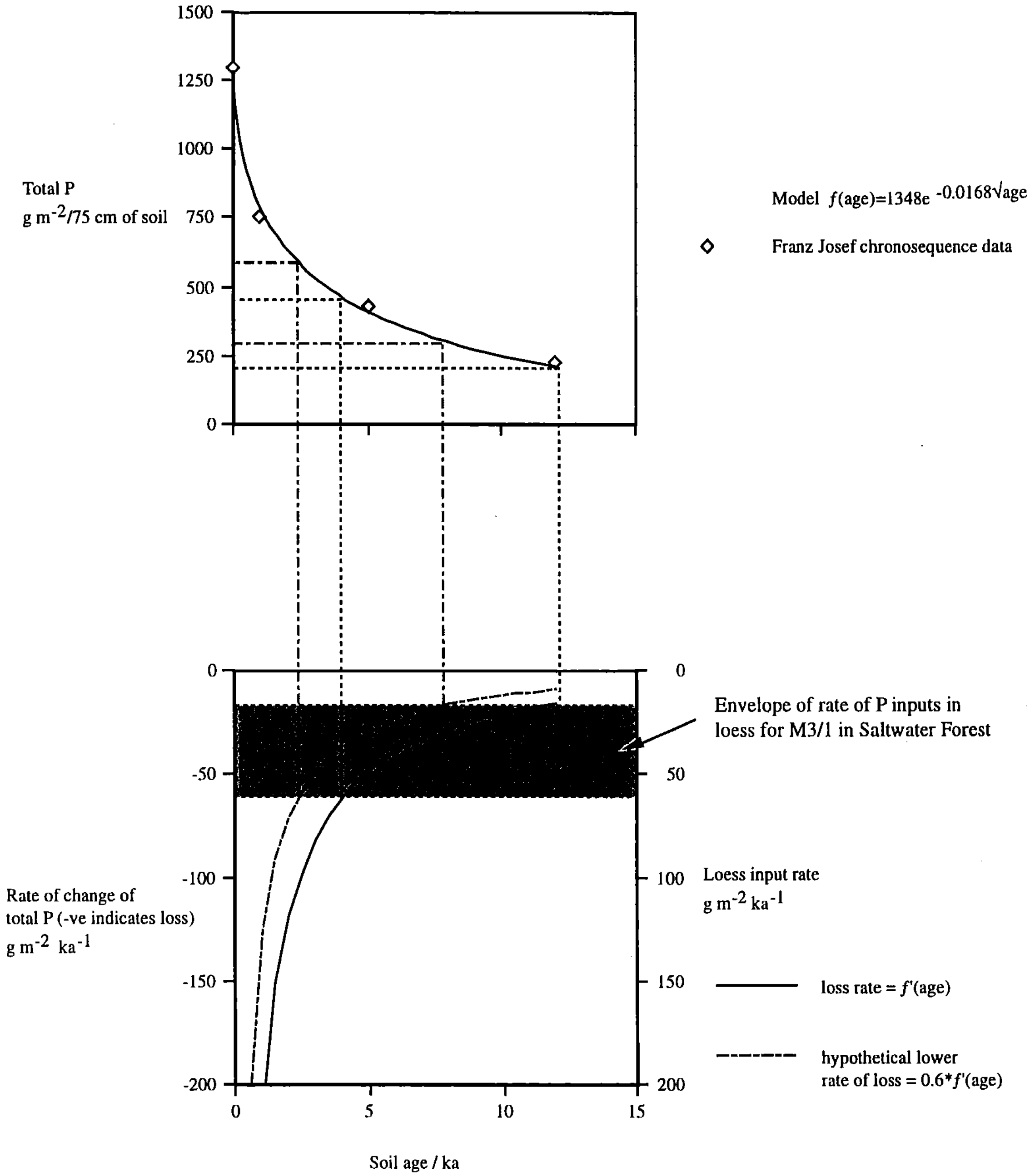


Fig. 6. Comparison of losses of total P with time in the Franz Josef chronosequence (Stevens, 1968) with rates of inputs of P in loess in Saltwater Forest. a) Total P to 75 cm depth in soils of the Franz Josef chronosequence vs soil age. b) Rate of P loss from chronosequence soils vs soil age.

rainfall and other soil forming factors are similar between the two areas.

If rates of P loss during the upbuilding phase of pedogenesis were comparable to rates in the Holocene (the period within which all soils except the Okarito Soil in the FJ chronosequence formed), inputs of P from loess could only maintain P_{tot} levels typical of the Perch-gley Podzols found on the 5 ka and 12 ka surfaces (Fig. 6). Total P (g m^{-2}/75 cm of soil) in the lower 110 cm of M3/1 calculated as an average of a moving 75 cm window from 74 cm to the base of the loess at 185 cm is ca. 170 g m^{-2}. This increment of the profile presumably is unaffected by topdown pedogenesis in the Holocene and yet total P is similar to that of the 12 ka Perch-gley podzol at Franz Josef. If rates of loss during the upbuilding phase had been lower e.g. 60% of the rate in the Holocene (Fig. 6), it might be expected that total P in the lower 75 cm of the loess would be somewhere between 300 and 575 g m^{-2}/75 cm of soil. These inferences rely on an assumption of a simple relationship between rates of P loss and total P status, but they suggest that P losses in the upbuilding phase of pedogenesis were similar to that occurring in the Holocene phase of topdown pedogenesis. Thus the P distribution in upbuilding soils on moraines can only be understood if losses and transformations during the upbuilding phase are considered. This contrasts with soils formed in loess in the subhumid climate of the east coast of South Island where loess accretion rates are about three to six times greater than in Saltwater Forest (Eden and Froggatt, 1988). In this environment depth profile trends of P_{tot} are most strongly influenced by redistribution during topdown pedogenesis phases in periods of no or very low rates of loess accretion. Indeed Runge et al. (1974) used P_{tot} and other P fractions to identify the presence and stratigraphic position of paleosols in east coast South Island loess assuming that P losses and redistribution had only occurred in a topdown sense i.e. after loess accretion stopped. However, other work from dry areas of mid-western United States has shown that significant soil modification during loess accretion is not unique to high rainfall environments. Beavers et al. (1963) showed using CaO/ZrO_2 and other element ratios in a sequence of soils progressively more distant from loess source in southern Illinois, that in distal (more weathered) soils the degree of weathering attributable to the period of loess accretion was similar to that in the period since loess accretion stopped. Their partitioning of weathering between the two phases relied on the assumption that weathering in the proximal soils most strongly reflected topdown pedogenesis which is reasonable if loess accretion rates were sufficiently high. Further by assuming, although not explicitly stated, that rates of weathering were the same in the period of loess accretion to that post-loess accretion, they calculated a period for loess accumulation in good agreement with the radiocarbon chronology for withdrawal of ice from northern Illinois.

4.3.2. Terraces (Fig. 4k, l, Table 1)

In simple soils P_{tot} is highest in Ah horizons at around 300 μg/g, steadily declines into Er horizons to <140 μg/g and rises again in 2Bh horizons to similar concentrations as those of Ah horizons. About 50% or more of P_{tot} is P_o in Ah horizons, and this proportion increases to over 60% in Er horizons. However in 2Bh horizons P_o decreases in importance with depth as $P_{Fe/Al}$ becomes a more significant component. In all other horizons P_r is the only other significant form of P making up between 20 and 50% of P_{tot}.

In upbuilding soils P_{tot} in Ah horizons is <250 μg/g, lower than in the simple soils. P_{tot} minima for upbuilding occur in Er horizons and are similar to the equivalent horizons in simple soils. Maximum P_{tot} concentrations, equivalent to those of the Ah horizons of simple soils, occur in buried A or O horizons in or below L1b. The proportion of P_{tot} made up by P_o increases from the Ah (40%) through Er horizons into the buried A or O horizons (>60%). As for the simple soils P_r is the next most important fraction (usually >30%), except in the gravels at the base of T4/2 where a very high P_{tot} corresponds to an abundance of primary inorganic P (P_{Ca}).

Because P_o is the dominant P fraction, the lower P_{tot} in A horizons of upbuilding soils principally reflects a lower %OC content. Whether this is a response to differences in ecosystem productivity or merely differences in local factors between sites is impossible to gauge with the available data.

Root distributions in simple soils show two peaks; one in the A horizons and a second in 2Bh horizons with very few between (Almond, submitted). Roots in the 2Bh are structural and feeding roots and often completely envelop stones. The source of primary inorganic P provided by the relatively fresh gravels in 2Bh horizons is not readily accessible in the upbuilding soils. Very few roots extend to the gravels beneath T4 because of their depth and the presence of a strongly cemented pan in the top of the gravels (Fig. 3). In these soils root distribution is similar to the simple soils, but the secondary maximum is in the organic-rich buried soil horizons where there is a relatively large pool of P in organic and recalcitrant forms.

The greater P_{tot} values in terrace soils versus moraine soils generally reflects a greater P_o and organic carbon content. Presumably a lower rate of P mineralisation in the wetter terrace soils preserves P in organic forms and decreases leaching losses. The decrease in P_o below 60 cm in upbuilding soils on moraines and the concomitant increase in $P_{Fe/Al}$ may represent the mineralisation of P_o and its conversion to adsorbed forms.

The P data from terrace and moraine sites indicate that additions of loess have been sufficient to maintain P_{tot} in soils (< 2 mm fraction) on >35 ka landforms at levels similar to soils on young <18 ka landforms (Table 3). Furthermore there are no significant differences in the distribution amongst P fractions between simple and upbuilding soils. Wardle (1980) presented a primary successional sequence for south Westland forest. He proposed that pioneering vegetation on fresh surfaces is replaced by shrubland then tall forest and finally heathland or bog on surfaces ca. 50 ka and older. He suggested that the succession is a response to deteriorating soil physical and chemical properties. In areas of south Westland that have received loess in the Otiran, additions of fresh material may have been sufficient to mitigate soil fertility decline. Vegetation patterns perhaps should be analysed with regard to landsurface age and loess distribution patterns.

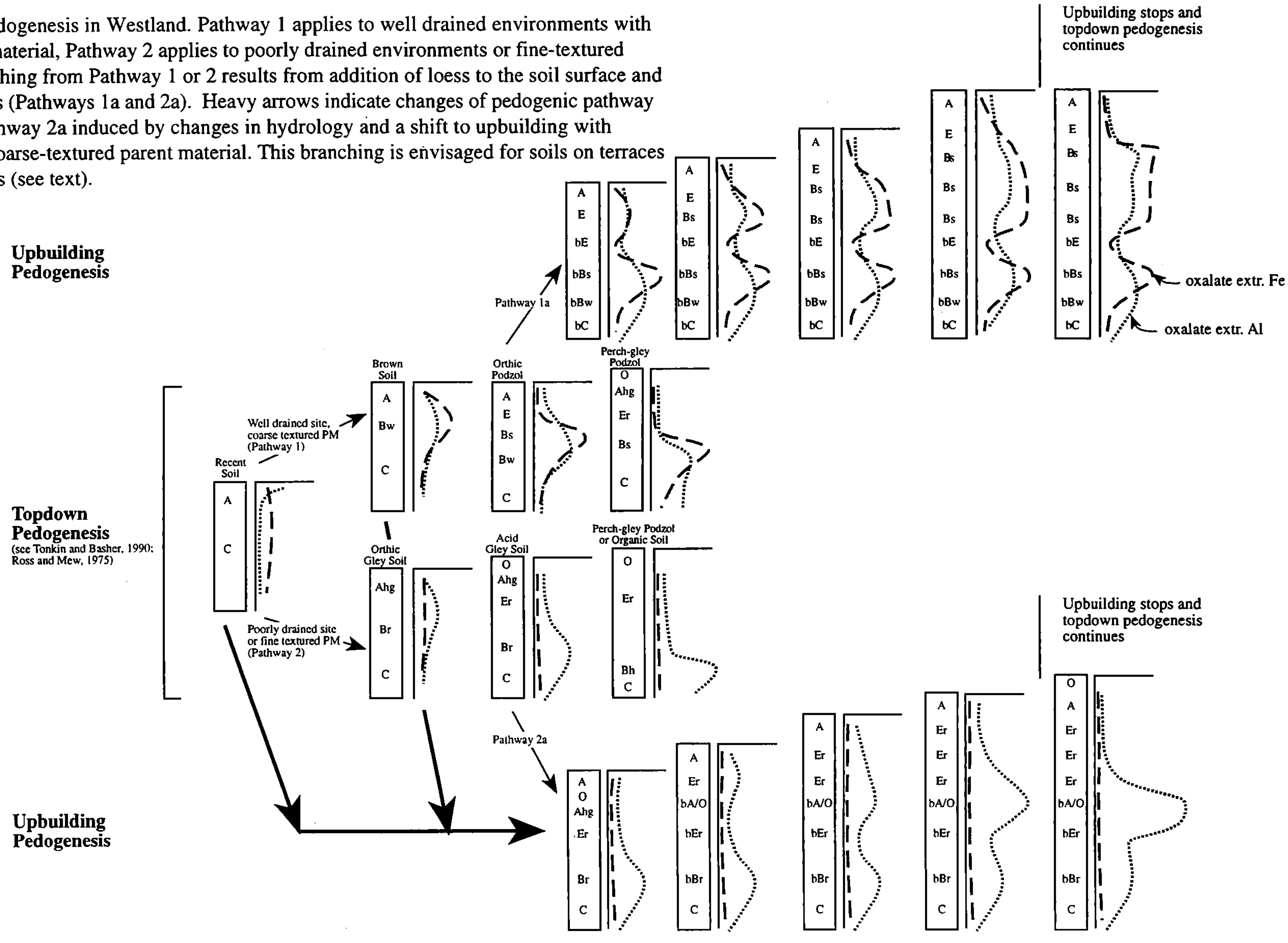
Table 3. Profile mass (g m^{-2}) of total P to 75 cm depth (< 2 mm fraction) for simple and upbuilding soil arranged according to landform and landform age.

Landform age	<14 ka	16.5-18 ka	>35 ka	ca. 70 ka
Terrace	1250	1177	2402	
			1600	
Moraine	969	1947		1636

5. The influence of upbuilding on pedogenic pathways and implications for chronosequence studies

Two pathways of pedogenesis in Westland have been inferred from soil and soil chronosequence studies in Westland (Fig. 7). Pathway 1 applies to well-drained sites and coarse-textured parent material; soils evolve from Recent soils through Brown Soils to Podzols and finally Perch-gley Podzols (see particularly Basher,1986; Sowden, 1986; Smith and Lee,1984). Pathway 2 applies to poorly-drained sites or fine-textured parent material; soils evolve from Recent or Recent Gley Soils through Orthic Gley Soils, Acid Gley Soils to Perch-gley Podzols or Organic Soils (see Mew and Lee, 1988; Ross and Mew, 1975). Generalised hypothetical oxalate Fe and Al depth trends are shown (Fig. 7) for each profile to summarise the effects of the podzolisation and gleying. The trend in Pathway 1 is for Fe_{ox} to increase immediately below the A horizon as soils weather (Brown Soils). After podzolisation begins the peak in Fe_{ox} progressively increases and moves down the profile. In the latter stages of the pathway, reducing conditions enhance the loss of iron from A and Er horizons as the porosity of the E horizon declines. Al_{ox} changes are similar although the peak is slightly deeper than for Fe_{ox} and it is unaffected directly by reducing conditions. In Pathway 2 Fe_{ox} never accumulates because Fe is leached in the reducing conditions. Al_{ox} builds up as in Pathway 1

Fig. 7. Pathways of pedogenesis in Westland. Pathway 1 applies to well drained environments with coarse-texture parent material, Pathway 2 applies to poorly drained environments or fine-textured parent material. Branching from Pathway 1 or 2 results from addition of loess to the soil surface and upbuilding pedogenesis (Pathways 1a and 2a). Heavy arrows indicate changes of pedogenic pathway from Pathway 1 to Pathway 2a induced by changes in hydrology and a shift to upbuilding with addition of loess to a coarse-textured parent material. This branching is envisaged for soils on terraces that have received loess (see text).



but eluvial horizons are much deeper. Ultimately, advanced soils of Pathway 1 approach those of Pathway 2 (Ross and Mew, 1975) as structural collapse in the upper horizons causes perching of water and the depth of perching progressively deepens (Farmer et al., 1984).

Additions of loess cause a branching from the topdown pedogenic pathways to upbuilding pathways. Branching could occur at any point along the two pathways but I have indicated branches most applicable to the upbuilding soils in the study area. Pathway 1a (well drained) depicts the gradual addition of loess to a Brown Soil or podzolised Brown Soil. Mineralisation of organic matter from the A horizon as it becomes buried reveals a bleached horizon subsequently identified as a buried E horizon (see bE horizons M3/1 or M3/2). Further addition of loess results in weathering and translocation of Fe and Al from the upbuilding A and E horizon to lower increments (described above). The result is an upbuilding Bs horizon. When loess addition stops, development of the upbuilding soil continues on the topdown pathway (1). Pathway 2a (poorly drained) depicts gradual loess addition to an Orthic or Acid Gley Soil. Organic matter in the A or O horizons does not mineralise as it becomes buried because of the reducing conditions. Fe released from primary minerals by weathering never reprecipitates and Fe_{OX} does not accumulate as a result. Al_{OX} is mobilised at low pH and in the presence of organic ligands, and precipitates at depth, particularly where it can further complex with buried organic matter. When loess addition stops, development of the upbuilding soil continues on Pathway 2 whereby the translocation of Al is intensified and the peak in Al_{OX} in buried A and/or O horizons increases.

Figure 7 highlights a need for a reinterpretation of the chronosequence studies in Westland that included soils on older loess-mantled terraces (see Campbell, 1975; Stevens, 1968; Tan, 1971; Ross et al., 1977). The requirement of soil chronosequences that all soil forming factors, other than time, remain effectively constant (Jenny, 1941) is to ensure each soil represents a point (in time) along the same pedogenic pathway. Where loess has accumulated on some landsurfaces this can no longer be sustained (Fig. 7). Additions of loess and subsequent upbuilding can change the type and rate of processes, and the balance between progressive and regressive pedogenesis (Johnson, 1987). Furthermore, the final pathway may be dependent on the point at which the branching occurred, i.e. the sort of soil loess accumulated on. Indeed, a change in pedogenic pathway brought about by upbuilding is perhaps responsible for the pedogenic threshold identified by Campbell (1975) between the 18 ka Ahaura soil and the 30-50 ka Kumara soil. The soils in Campell's chronosequence have formed on river terraces underlain by coarse-textured felsic alluvium. Loess addition, principally through its influence on soil hydrology, may have caused the soil on the 30-50 ka surface to switch from Pathway 1 to Pathway 2a early on in its formation (heavy arrows in Fig. 7). This could account for the marked and disproportionate difference in total Fe, Al and Si between these two soils (Fig. 8). The soils in Stevens' (1968) chronosequence have formed on moraines comprised of felsic till dominated by micaceous schists, and soil development passes rapidly along Pathway 1 (Fig. 7). However, the oldest soil which is formed in loess, (Okarito

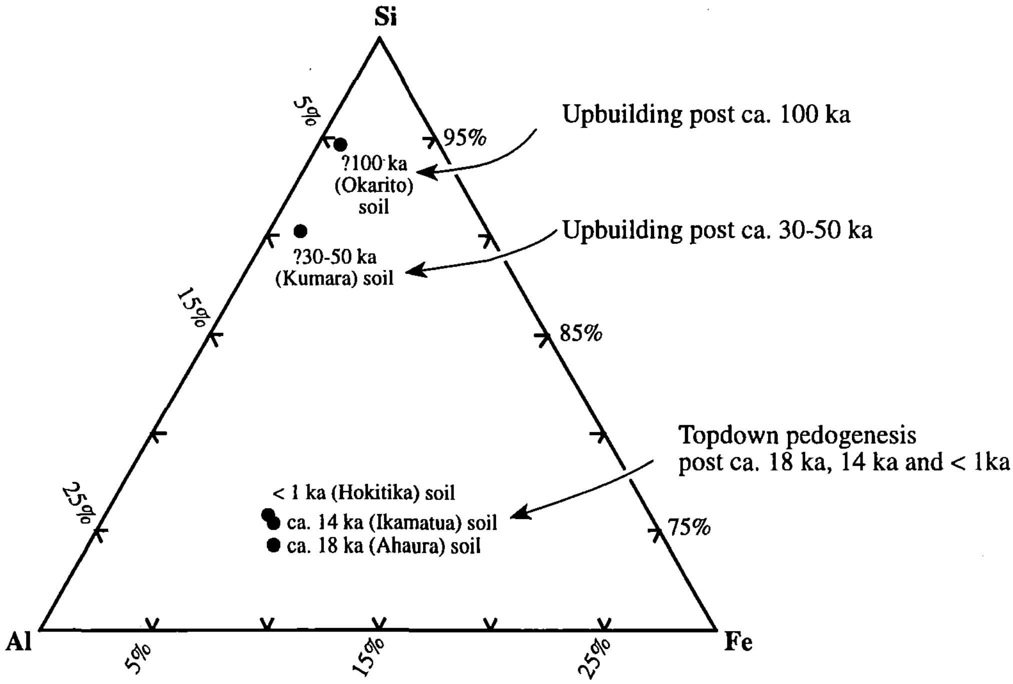


Fig. 8. Triangular plot showing changes in total Fe and Al relative to Si in the upper 70 cm of soils from the chronosequence of Campbell (1975) near Reefton in Westland (Fig. 1).

soil) presents many problems for interpretation because, not only is its age underestimated (P. Almond unpublished information), but it is likely to have had a complicated history of episodes of upbuilding, topdown pedogenesis and erosion. The Okarito soils studied by Stevens (1968) are formed in loess about 50 cm thick, which is around 1.5 m less than the maximum thickness of loess found on the moraine from which the soils were sampled (P. Almond unpublished information). Conclusions on rates and types of soil processes drawn from this soil should be viewed with caution.

6. Conclusions

In Saltwater Forest, and presumably surrounding areas with similar rainfall and loess accumulation rates, soil formation is rapid enough and loess accumulation slow enough that the two processes must be considered to have acted contemporaneously. Loess has accumulated on landforms dating from the middle to early Otiran (Last Glaciation). During loess accumulation, soils formed by developmental upbuilding (Johnson, 1987) until the Late Otiran when loess accumulation stopped. A phase of topdown pedogenesis followed and continues today.

The impact of the upbuilding phase of pedogenesis is demonstrated by comparisons of soil pH, %OC, oxalate extractable Fe and Al and P chemistry between upbuilding soils and soils formed by topdown pedogenesis into a single sediment body (simple soils). Differences are attributable to each soil increment of upbuilding soils having experienced near surface and subsoil pedogenic processes. Poorly-drained upbuilding soils on terraces have higher subsoil organic carbon than simple soils, but on moraines, where soils are moderately well drained, organic carbon accumulated during upbuilding has mineralised. On moraines there is no systematic difference of %OC between upbuilding and simple soils.

Weathering has been enhanced in subsoils of upbuilding soils. This is reflected in total P concentrations (ca. 200 µg/g) much lower than fresh loess (ca. 600 µg/g) and the almost complete conversion of primary mineral P to secondary mineral and organic forms. In C horizons of simple soils P_{tot} rises to concentrations typical of unweathered material with more than 30% of this in a primary form. Enhanced subsoil weathering in upbuilding soils on moraines is also reflected in lower soil pH than in simple soils. The effect is not apparent on terraces where deeper acidification and the thin loess obscure any differences.

Podzolising processes acting contemporaneously with loess accumulation have produced elevated levels of oxalate extractable Fe and Al in subsoil of soils on moraines. Each increment of soil has in turn been part of eluvial then illuvial horizons. On terraces the current phase of topdown pedogenesis dominates distribution of oxalate extractable Fe and Al, although soil accessions inherited during upbuilding have preconditioned soil response. Buried A and O horizons have acted as sinks for mobile Al and now appear similar to pedogenic Bh horizons of simple soils. Permanent anaerobic conditions have prevented accumulation of Fe in a similar manner.

The phenomenon of upbuilding pedogenesis has some significance for chronosequence studies and theories of soil evolution in Westland. In a soil chronosequence the ergodic principle (substitution of space for time) is applied, and the fundamental assumption is that a soil has passed through stages represented by all younger soils in the sequence (Rode, 1961 cited in Stevens and Walker, 1970). Clearly soils on surfaces mantled by loess should not be considered as part of chronosequences of soils formed on geomorphic surfaces that have lacked significant loess inputs. The findings of this study are applicable to any environment where soil formation rates and sediment accumulation rates are comparable.

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Chapter 5

A multiscale approach to soil variability in Saltwater Forest, a podocarp/hardwood forest, south Westland, New Zealand

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Abstract

The relative contribution to soil variability at different spatial scales and the intensity of land management determines the most cost-effective soil survey intensity. This study examines the nature and sources of soil variability at a range of spatial scales in a podocarp rainforest managed for sustainable timber production. The aim was to determine the most effective intensity of soil survey relevant to the current management intensity. The forest covers Pleistocene glacial landforms dominated by strongly developed Podzols and Organic Soils. Soil variability at a macroscale (100s of metres), mesoscale (10-100 m) and a microscale (1-10 m) was examined by a range of soil survey techniques.

Macrovariability is manifest as changes in abundance of soil profile classes (SPCs) among landforms and is due to changes in gross landform morphology responsible for differences in hydrology. Steep lateral moraines are dominated by well drained Orthic Podzols and Allophanic Brown soils; rolling terminal moraines are dominated by poorly drained Acid Gley Soils and Organic Soils while ablation moraines are dominated by extremely poorly drained Organic Soils. Terrace soils are dominantly extremely poorly drained Organic Soils or very poorly drained Perch-gley Podzols. Increases in relative abundance of the former between terraces is related to increasing terrace age and decreasing slope and relief. Mesoscale variability on moraines is often apparently catenary but after examination at microscales it is clear that soil development on slopes is not integrated but occurs across series of independent slope segments. Microvariability of the same order of magnitude as mesovariabilty arises from microtopographic influences on soil hydrology and the effects of individual trees on bioturbation, local hydrology and leaching regimes. Meso- and microvariability were difficult to separate on terraces because of the lack of relief.

Much of the soil variability is concentrated at microscales and it is not until scales characteristic of landforms are reached that a significant systematic component to the soil pattern is obvious. The current forest management prescribes minimum management units averaging 260 ha (1600 m characteristic dimension). A soil survey intensity and soil map scale sufficient to delineate the macroscale variability and major features of forest infrastructure (1:50 000) is most appropriate. More intensive surveys would add little

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to resolution of soil variability, complicate the depiction of the soil pattern and be economically unjustifiable.

Keywords: Soil variability; soil survey; soil-landform models; forest soils

1. Introduction

Soil survey for soil mapping aims to divide the soil continuum into discrete areas such that soil variability within areas is less than that between areas. Schelling (1970) argues the legitimacy of this as follows: soils are a product of five soil forming factors (Jenny, 1941); there are areas over which soil forming factors are relatively constant separated by zones of change; therefore there are areas of homogenous soil (soil landscape bodies) defined by boundary zones of maximum lateral rate of change of soil properties. He argues further that for soil mapping a natural classification of soil individuals, in which individuals of a class share the greatest number of soil properties in common, will when mapped produce the most homogenous delineated soil bodies and mapping units.

This is a formalisation of the precepts by which soil surveyors operate but which have been challenged by quantitative analysis of soil variability. McBratney (1981) and Webster (1981) produced optimal boundaries along a transect by using a moving window analysis to identify zones of maximum rate of change of soil properties. They then showed that the resulting spatial groupings of soils crossed classes formed by optimal (natural), non-spatial classification. This equates to the familiar problem of the field soil surveyor trying to identify spatially coherent regions of soil while also trying to control the variability within.

The source of the phenomenon and departure from the logical arguments of Schelling arise because the judgment of homogeneity depends on the scale of observation. What is apparent uniformity at one scale has variability at another. Each soil forming factor "may act over a different spatial scale, and that within each soil forming factor there can be many spatial scales of interaction" (Burrough, 1981). This gives rise to complexity apparent as noise and possibly irrelevant at one scale but which may be resolved into a structural (deterministically understood) component and a residual random (noise) component at larger scales. The fact that many different soil maps for a given region can be derived depending on map scale is direct evidence of this character of soil variability.

How soil variability is partitioned at different spatial scales has implications for soil survey (Nortcliffe, 1978). The optimal intensity of survey (map scale) depends on the spatial structure of soil variability, the intensity of land management and minimum management unit (MMU), and cost of survey. If most variability is concentrated at ranges less than the MMU (Fig. 1: scenario 1), then little benefit will arise from increasing map scale from that of reconnaissance level. Compound mapping units such as soil associations or complexes are then employed. However, where compound mapping units are a necessary expedient, it is still

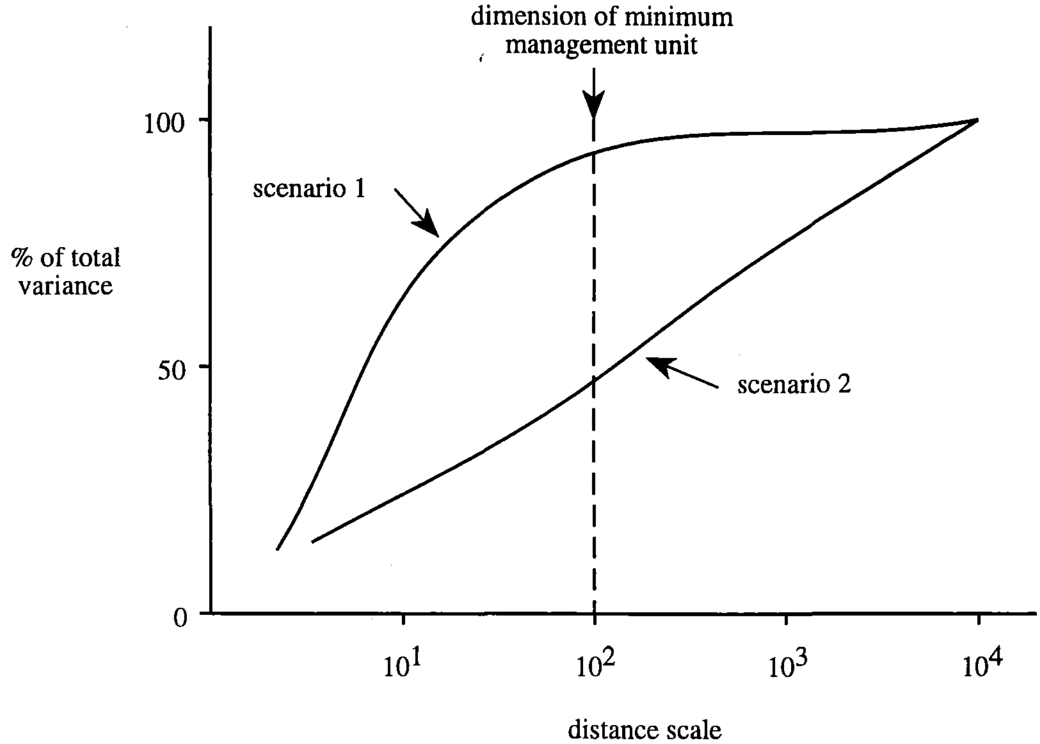

Fig. 1. Hypothetical distribution of soil variability across scales in relation to an arbitrary minimum management unit.

important to appreciate the nature and sources of soil variability to assess its impact on land use.

If variability is distributed more evenly across scales (Fig. 1: scenario 2) then benefits of increased predictive ability will accrue as survey intensity increases until dimensions of delineations are in the order of the MMU. Beyond this, further resolution is irrelevant since management cannot account for it. Optimal map scale depends on the benefits gained by increasing resolution of soil variability balanced against exponentially rising costs with increasing map scale (Dent and Young, 1981).

This study examines the soil pattern in a largely unmodified indigenous forest in New Zealand managed for sustainable timber harvesting. Current management involves helicopter extraction of senescent or mature trees. Single trees or small groups are extracted at a rate considered to be sustained by natural regeneration, over 15 year cycles in compartments averaging 260 ha. Forest structure and dynamics are influenced by soils and landforms and so a subdivision of the landscape by soils would provide an ecological basis for forest management.

The aim of this study was to evaluate the degree and sources of soil variability at different scales in order to identify the most effective scale for soil survey given the intensity of forest management. This paper also serves to document the soil pattern in the study area as a basis for a following paper which examines the interactions between forest and soils.

2. Study Area

The study area includes 4000 ha of partly modified indigenous podocarp/broadleaf forest in Saltwater Forest (Fig. 2) on the coastal piedmont of South Island's west coast. Rimu (*Dacrydium cupressinum)*, a podocarp, dominates. The podocarps are long-lived trees commonly reaching an age of 500 years. On terraces rimu forms a ca. 30 m high closed canopy with miro *(Prumnopytis ferruginea)* and silver pine (*Lagarostrobos colensoi)* (Six Dijkstra et al., 1985). On rolling to hilly land large rimu are emergent along with miro, totara (*Podocarpus hallii)* and the broadleaf rata *(Metrosideros umbellata)* above a lower broadleaf canopy of quintinia *(Quintinia acutifolia)* and kamahi (*Weinmannia racemosa)* (Norton et al.,

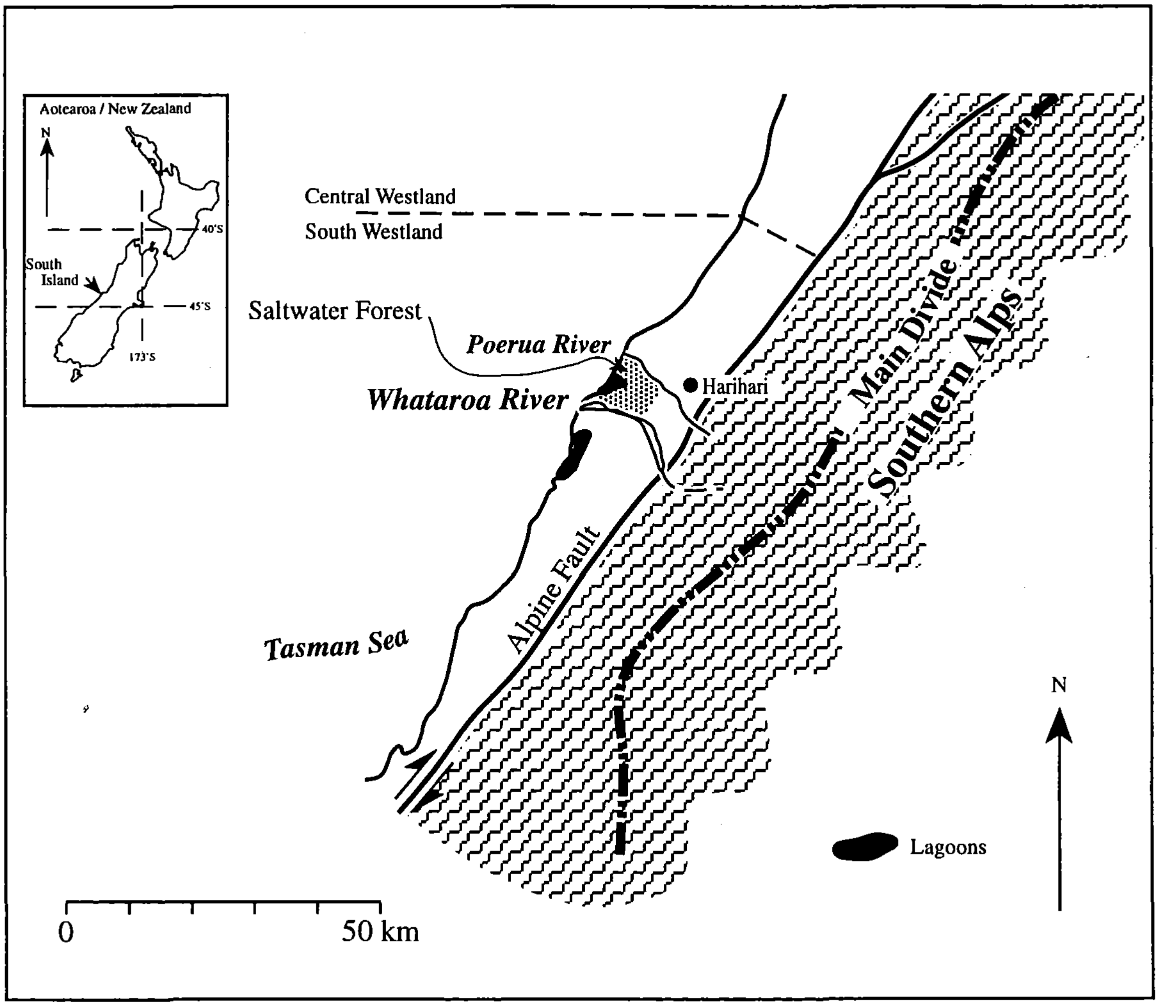

Fig. 2. Location map showing Saltwater Forest.

1988).

The climate is mild temperate and super-humid (Hessell, 1982; New Zealand Soil Bureau, 1968). There is a distinct rainfall gradient from the coast into the Southern Alps (Griffiths and

McSaveney, 1983) and so the rainfall over the study area may range from 3000 to 4000 mm. Mean annual rainfall at Harihari (Fig. 2) is 3790 mm and rainfall exceeds evaporation at all times during the year. Mean annual temperature is about 12°C, peaking in January and being lowest in July (mean monthly temperatures of 16.3 and 6.9°C respectively)(Norton, 1990).

Landforms are glacial in origin and include late Last Glacial (Otiran) moraines and outwash terraces formed during glacial advances from alpine valleys in the Southern Alps immediately to the east (Fig. 2) (Almond, 1996). Clastic sedimentary rocks and metamorphosed equivalents from the Southern Alps dominate glacial deposits in the study area (Pers. obs.). Quartzo-feldspathic sandstones and siltstones of the Torlesse Supergroup of the Main Divide grade into their metamorphosed equivalent, the Haast Schist toward the Alpine Fault. The schists include chlorite, biotite and minor garnet grades. In general, schist is more abundant than the sandstones and siltstones in tills and outwash gravels. These materials also include rare clasts of Paleozoic granite and gneiss of the Tuhua Group which outcrop west of the Alpine Fault.

A pair of lateral moraines (M5 and M6) on the margins of the Poerua and Whataroa Rivers were formed by the most recent glacial advances in the late Otiran (Almond, 1996)(Fig. 3). During the earlier, M5, advance (18-22 ka), ice lobes from the Whataroa glacier extended into Saltwater forest from the south on to an older (>36 ka), early to mid-Otiran outwash terrace (T4). Age discrimination between M5 terminal moraines and the older T4 terrace is based on loess and soil stratigraphy. T4 has about 1 m of loess which includes the 22.6 ka Aokautere Ash in the upper 40 cm and a buried soil, radiocarbon dated at 35.8 ka, formed in fine-textured alluvium immediately above the gravels. Loess and Aokautere Ash are absent on M5 (Almond, 1996). A moraine associated with T4 was not found, and it is presumed to be buried beneath the M5 terminal moraine.

M5 ice lobes produced a belt of terminal moraines and behind this a zone of ablation moraines after the ice melted. Meltwater exiting from channels in M5 terminal moraine incised into T4 leaving a set of narrow (< 500 m wide), northerly aligned terrace remnants. These are truncated to the north along a north-west to south-east trending line probably as a result of channel widening of the meltwater river from the main Poerua valley glacier at the same time or soon afterwards. The outwash surface formed by the meltwater rivers is labelled T5.

No ice lobes from the later M6 advance (14-16 ka) extended onto the study area, but termino-lateral moraines rise abruptly above Lakes Rotokino, Joan and Darby in the south (Fig. 3). Meltwater from this advance formed two narrow 10 m deep gorges which widen to the north into a broad outwash surfaces (T6) inset within T5. In other areas erosional surfaces formed as streams graded to the lower local base level established at this time and these are also included with T6.

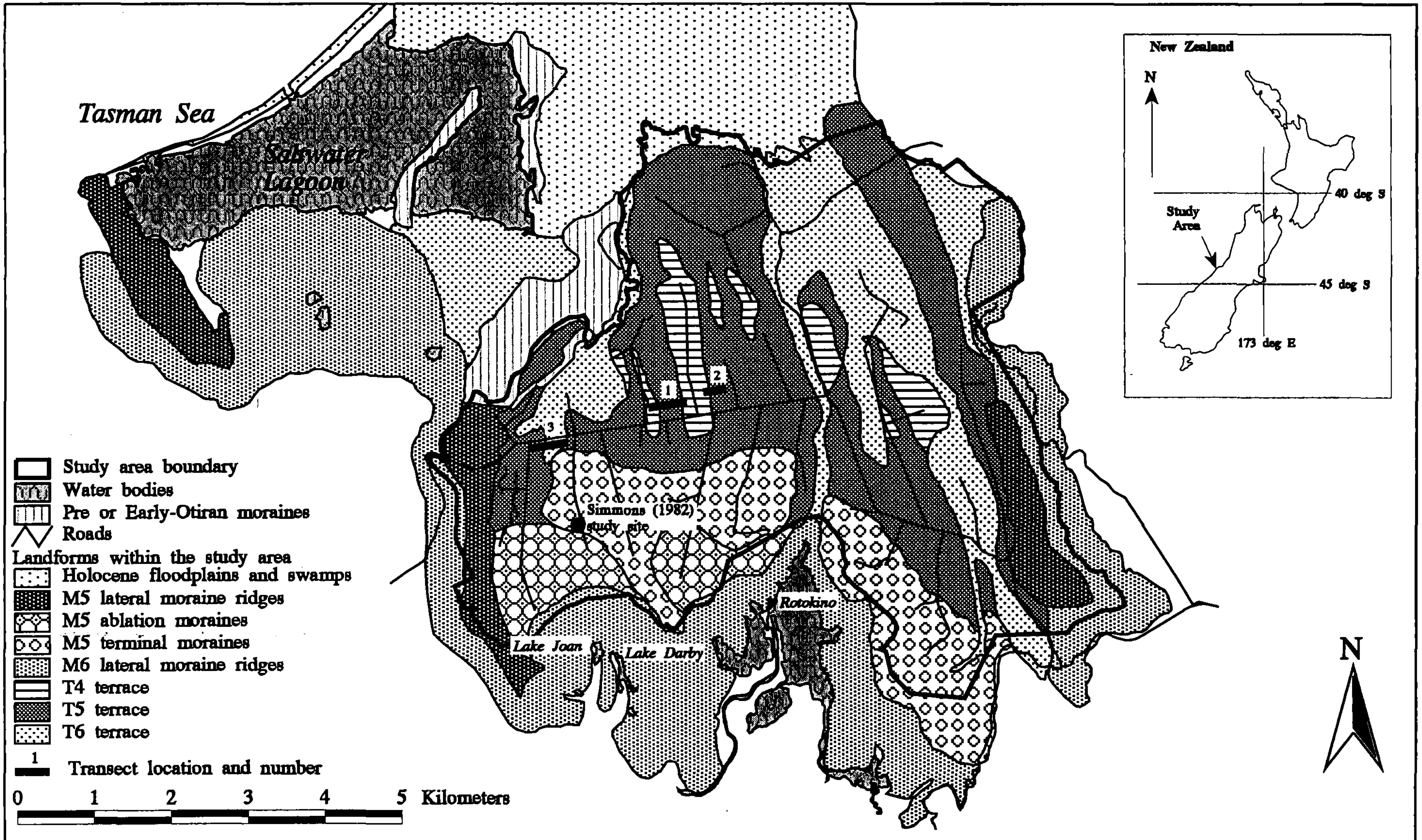

Fig. 3. Landforms of Saltwater Forest and the study area boundary. The locations of the transects described in the text and illustrated in Fig. 7 are also shown.

Soil evolution in Westland is driven by intense leaching under high rainfall and the acidifying podocarp or beech vegetation. Gleying is a common phenomenon which may persist from soil inception or develop as soils evolve. Soil chronosequence studies on well drained sites show a progression from Recent Soils, through Brown Soils, Acid Brown Soils, Podzols and eventually Perch-gley Podzols at a rate dependent on mean annual rainfall (Tonkin and Basher, 1990). Under ca. 2000 mm mean annual rainfall Brown Soils form sometime before 14 ka and persist beyond 22 ka after which Perch-gley Podzols form. In a rainfall of 6000 mm Brown Soils form within 1 ka and Perch-gley Podzols within 3 ka.

Because of the age of land surfaces (>14 ka) and the high rainfall, soils in the study area are strongly developed. Previous work in the study area (Almond, 1986; New Zealand Soil Bureau, 1968; Simmons, 1982) indicates that Organic soils, Perch-gley Podzols and Acid Brown Soils are the dominant soils, but the soil pattern has a high degree of short range variability. (Almond, 1986) mapped part of the study area at a scale of 1: 50 000 and employed only compound mapping units.

3. Methods

The soil pattern was analysed at three nested spatial scales; a macroscale, mesoscale and microscale. The macroscale corresponds to a physiographic division of the landscape into genetically and morphologically defined landforms as described above. The characteristic dimension of units was in the range of 10^2-10^3 m. The mesoscale corresponds to a subdivision of landform units into component landforms with a characteristic dimension in the range 10-10^2 m. The microscale corresponds to an intra-component landform scale with characteristic dimension 1-10 m.

I analysed aerial photographs to delineate moraine landforms and some boundaries between terraces, but generally differences in relief across terraces are so slight that, with the tall forest cover, many terrace edges had to be surveyed by pace and compass. Where there was no relief expression of boundaries between terraces, pace and compass survey was augmented with auger borings.

I carried out a reconnaissance survey of the study area at the macroscale using landforms as a stratification. Natural exposures or soil pits on each landform were sampled by free and transect survey and described according to the horizon nomenclature of (Clayden and Hewitt, 1989) and the soil description methodology of (Hodgson, 1976). Groupings of similar soil profiles into soil profile classes (SPCs) according to presence or absence of horizons, horizon sequence and depth, formed the basis of a preliminary classification. At a mesoscale two approaches were adopted. The first involved sampling representative landform components along transects. On moraines these were aligned downslope to identify any catenary relationships. Terrace soils were sampled along transects for which detailed elevation data were available (Almond, 1986; Cornere, 1992). This allowed recognition of component landforms too subtle to delineate otherwise.

The second approach employed grid surveys. Two 200 x 200 m areas were randomly located in each landform unit and sampled by auger at the intersections of a 50 m grid. At each observation site in moraine units I recorded landform component and measured slope by abney level across a slope segment up to 15 m length depending on visibility. I used a grid survey technique for practical reasons of navigation, and a sample spacing of 50 m was chosen because it was found to be generally greater than the range of mesoscale variability. Hence observations should provide an unbiased sample from which associations of soil class and landform component could be established. Assuming the grid surveys were a representative sample from each landform unit they also provided a means of estimating soil abundance within.

Analysis of soil pattern at the microscale draws upon information from other studies in the study area complemented with additional short-range sampling along selected soil transects established in the mesoscale study.

Final definition of soil profile classes involved analysis of soil profiles from the macro-, meso- and microscale studies.

4. Results

4.1. Soil profile classes

At the prevailing stage of soil development differences in soils arise principally from drainage gradients so soils are divided firstly into 10 soil profile forms (labelled 1 to 10) by soil drainage. Soil profile forms are then divided into soil profile classes according to soil age/parent material. There are two age groups of soils. The older (indicated by .1 after the soil profile form) includes soils on T4 which began forming sometime before 35 ka. Soil formation and loess accretion were contemporaneous through to about 14 ka after which loess accretion stopped but soil formation continued. The younger (indicated by .2 after the soil profile form) includes soils on moraines and associated terraces formed between 22 and 16.5 ka. Soils on M5, T5 and M6, T6 are not separated, despite their age difference, because their suite of soils appear to have reached a persistent soil state and are not distinguishable. Three soil profile forms cross age groups so 13 SPCs are recognised in total (Fig. 4). Soil profile classes are described below in order of decreasing drainage status.

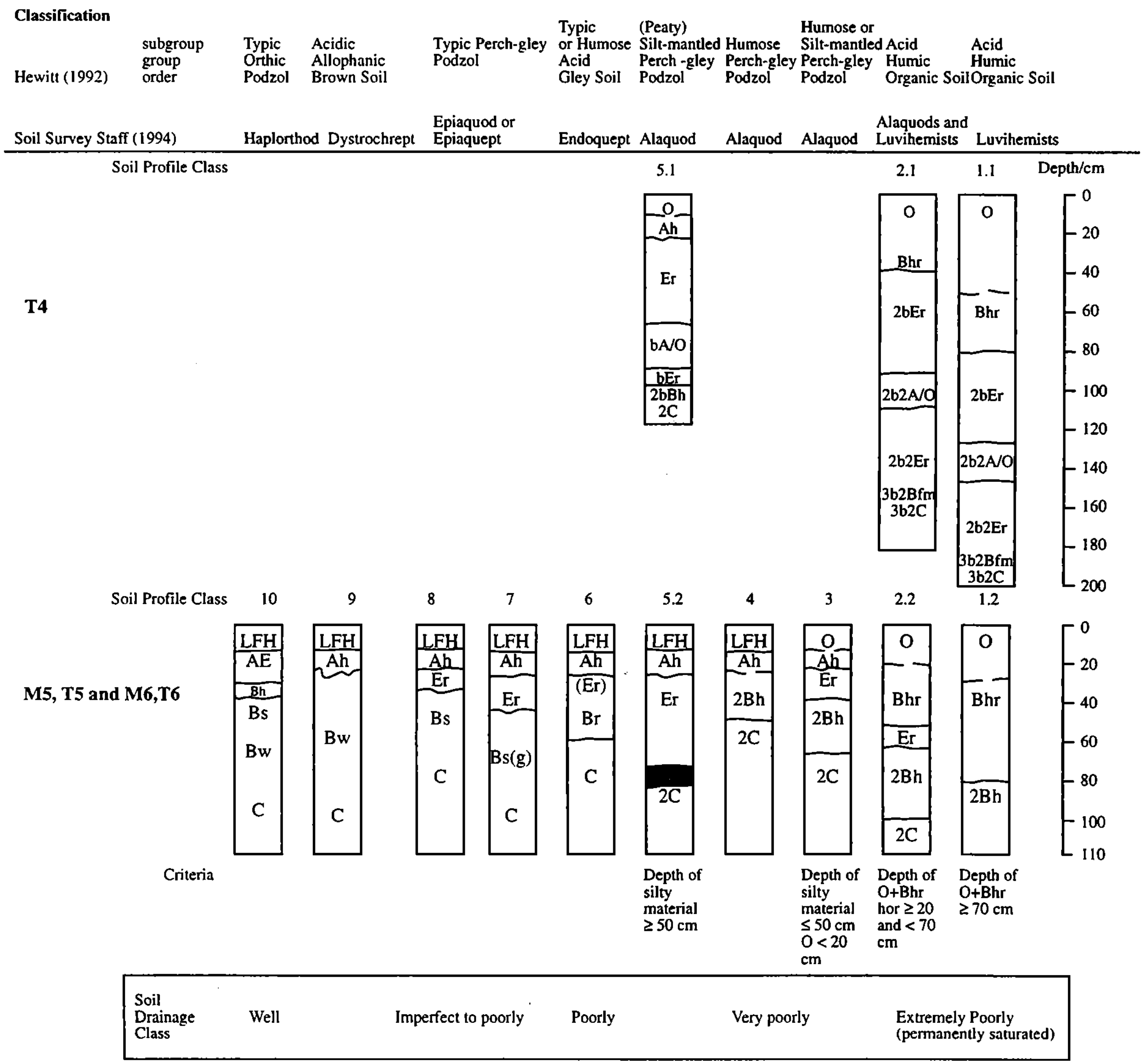

Fig. 4. Soil profile forms and soil profile classes arranged according to drainage status.

4.1.1. Soil profile classes 9 and 10

SPCs 9 and 10 form in till or colluvium derived from till and include soils with litter organic horizons (L,F,H) shallow, greyish-brown Ah or AE horizons and bright yellowish-brown (2.5Y-10YR 5/4-6), gravelly, greasy silt-loam subsoils with a weak to moderate fine blocky structure. SPC 10 (Orthic Podzol) is distinguished from SPC 9 (Allophanic Brown Soil) on the presence of a thin greyish-brown E horizon beneath the A horizon and either a Bh horizon or bright yellowish-brown (7.5YR-10YR 5/6-8) Bs horizon above the Bw horizon.

4.1.2. Soil profile classes 6, 7 and 8

These SPCs form from till and share the same upper soil horizons. All have litter organic horizons (L,F,H) and lack peaty O horizons, above shallow, often gley-mottled, silt loam, Ahg horizons. These horizons in turn overlie greenish or bluish-grey, gravel-free, silt loam, massive Er horizons to a depth of about 30 cm in SPC 8 and 40 cm in SPC 7. In SPC 6 (Acid Gley Soil) it is difficult to distinguish this horizon from the B horizon which is also low chroma. It is in the nature of the B horizon that the three SPCs are distinguished. In SPC 6 it is a stony, silt loam, massive Bg or Br which is 50-85% or >85% low chroma colours respectively. These may be slightly more olive-coloured than the overlying Er horizons. The B horizon of SPC 7 (Typic Perch-gley Podzol) is designated Bs(g) and is an olive-yellow (5Y 6/3), massive, stony silt-loam with 2-49% low chroma colours as above. For SPC 8 (Typic Perch-gley Podzol) the B horizon is yellowish-brown (2.5Y 5/4-6) and lacks low chroma colours (Bs horizon). It is a stony silt loam with weak to moderate fine and medium blocky structure. Some variants show a brightening of the upper Bs horizon and the development of incipient iron pans. SPCs 6, 7 and 8 may include a thin Bh horizon (1-3 cm thick) with or without an iron pan at the contact between the solum and underlying till.

4.1.3. Soil profile classes 3, 4, 5.1, 5.2

All these SPCs are Perch-gley Podzols which have between 0 and 20 cm of mostly humified peat over a peaty or humic silt loam Ah horizon. SPC 4 is a Humose Perch-gley Podzol in which the A horizon rests directly on a brownish-black (7.5YR 3/2) 2Bh horizon formed in gravels or sometimes colluvium derived from till. SPC 3 is a Humose or Silt-mantled Perch-gley Podzol with a massive, light bluish-grey or greenish-grey, silt loam Er horizon above a 2Bh horizon in gravels whose upper boundary is within 50 cm of the soil surface. Soil profile form 5 includes soils where depth to gravels (2Bh horizon) is greater than 50 cm. On T4 these soils (SPC 5.1) are composite soils formed in 80 cm to 1 m of loess resting on gravels. The soil stratigraphy of the loess is complicated and varies significantly spatially (Almond, 1996), however a common feature is the presence of organic rich horizons toward the base. These are former A or O horizons formed after deposition of terrace alluvium and during a period of episodic loess deposition until sometime before 22 ka. On younger land surfaces SPC 5.2 is formed in uniform, deep silty alluvium. The surface O and Ah horizons are underlain by a bluish-grey, massive silt loam Er horizon on a 2Bh horizon in gravels at between 50 to 90 cm from the soil surface.

4.1.4. Soil profile classes 1.1, 1.2, 2.1, 2.2

These SPCs are Organic Soils which typically have very fluid humified peat (O) over very fluid peaty silt loam or silty peat muck (Bhr). Soil profile forms 1 and 2 are separated on the thickness of fluid material; for soil profile form 1 it is > 70 cm thick and for soil profile form 2 it is between 21 and 69 cm thick. Twenty centimetres was chosen as the critical thickness for peaty horizons rather than the usual 40 cm cutoff because this formed a more natural grouping of soils whose distinct properties are reflected in forest structure (Almond and Duncan, submitted). A 40 cm class limit was rejected because it would have divided soils which were spatially contiguous and to which the forest responds in a similar manner. SPCs 1.1 and 2.1 are found only on T4, and the fluid peaty material is underlain by loess which usually has the stratigraphy described for SPC 5.1. SPCs 1.2 and 2.2 are found on all younger landforms. In these SPCs the fluid peaty material is commonly underlain by a silty Er horizon in SPC 2.2 or a gravelly, thick 2Bh horizon in SPC 2.1. The latter also underlies the Er horizon in SPC 2.2. All soil profile classes have water tables permanently within 10 cm of the soil surface yet roots are heavily ramified to gravels or loess. From personal observation rimu roots do not have arenchyma, and there are gilled fish living permanently in the deeper organic profiles (SPCs 1.1 and 1.2). From this we can presume that although these soils are permanently saturated there is sufficient recharge of soil water by rainfall to maintain an oxygenated subsoil.

4.2. Soil Pattern

4.2.1. Macroscale

Reconnaissance survey suggested that the stratification of the study area by landform was meaningful. One exception was the separation of M5 and M6 lateral moraine ridges. After the reconnaissance survey these and M6 termino-lateral moraines above Lakes Rotokino, Joan and Darby, were combined into a single unit (steep lateral moraines) because of their similarity in morphology and the fact that their age difference was not reflected in soil pattern or properties.

At the macroscale the soil pattern is reflected by changes in abundance of SPCs among landforms (Table 1). In steep lateral moraines well drained soils (SPCs 9 and 10) are most abundant together making up nearly 50% of soils. Poorly to imperfectly drained soils (SPCs 6, 7 and 8) make up a further 30%. Very to extremely poorly drained soils make up less than 25% of all soils.

The M5 terminal moraine complex has a 50:50 split between poorly to imperfectly drained soils (SPCs 6-10) and very poorly to extremely poorly drained soils (SPCs 1.2, 2.2, 3 and 4). Well drained soils are rare. The ablation moraines are dominated by extremely poorly drained soils (SPC 1.2, 2.2). Together these SPCs make up 57% of all soils. A further 29% is very poorly drained soils (SPCs 3 and 5). Poorly to imperfectly drained soils make up only 14%, and well drained soils were not observed in the grid surveys.

Table 1. Frequency (%) of soil profile classes across landforms; arranged by component landforms for moraines.

Landform	Component landform SPC	1.1	1.2	2.1	2.2	3	4	5.1	5.2	6	7	8	9	10
Steep lateral moraines	ridge toeslope		9%			5%	2%							
	ridge footslope									2%			7%	
	ridge sideslope					2%				4%	2%	2%	7%	4%
	ridge crest										2%		7%	
	gully				2%		2%			4%	2%			
	spur footslope									2%	2%		7%	
	spur sideslope									2%		2%	4%	2%
	spur crest									2%			4%	7%
	total		9%		2%	7%	4%			17%	9%	4%	35%	13%
Rolling terminal moraines	inter-mound hollow		12%		4%									
	toeslope		2%		2%	2%	2%							
	footslope					2%				6%				
	sideslope		4%		4%		10%			22%	6%	4%		2%
	mound crest						2%				6%	6%		
	total		18%		10%	6%	14%			28%	12%	10%		2%
Ablation moraines	valley floor		40%		17%	15%			4%		4%			
	toeslope					4%			2%					
	footslope										2%			
	sideslope					2%			2%		2%	2%		
	crest										2%	2%		
	total		40%		17%	21%			8%		10%	4%		
T6			6%		16%	54%	2%		16%		2%	4%		
T5			16%		33%	22%	2%		22%			4%		
T4		56%		22%		2%		16%	4%					

Shading indicates percentage class: 0-2%, 3-4%, 5-10%, 11-20%, ≥21%

Nearly all soils on terraces are very poorly drained or worse, although there is a shift in relative abundance of extremely poorly drained and very poorly drained soils from T4 to T6. T4 has the greatest abundance of extremely poorly drained, deep Organic Soils. This unit has a unique assemblage of soils in that soils for the most part have formed in or, in the case of Organic Soils, on loess. Extremely poorly drained soils (SPCs 1.1 and 2.1) make up over 75% of soils. T5 has roughly equal abundance of extremely poorly drained soils (SPCs 1.2 + 2.2 = 49%) and very poorly drained soils (SPCs 3+4 + 5.2 = 46%). On T6 very poorly drained soils make up 62% of soils whereas extremely poorly drained soils make up less than 25%.

4.2.2. Meso- and microscale variability

At the mesoscale soil variability has a systematic and a random component. The systematic component is reflected in the broad associations of SPCs and component landform (Table 1). The random component arises from microscale variability reflected in the spread of SPCs across any one landform component (rows in Table 1). This type of variability was confirmed in transect surveys.

Steep lateral moraines comprise a pair (M5 and M6) of narrow-crested, sinuous ridges whose sideslopes are strongly dissected into a hollow and nose morphology. Much of the topography is obscured by the forest cover, but one of the grid surveys on this unit showed the stream density to be as high as 15 per km length of slope. About 20% of the sample sites on the grid surveys were on slopes greater than 28°, and the slope at 50% of the sites was greater than 13° (Fig. 5). The steepest slopes are generally found on the sideslopes of noses. Soils are

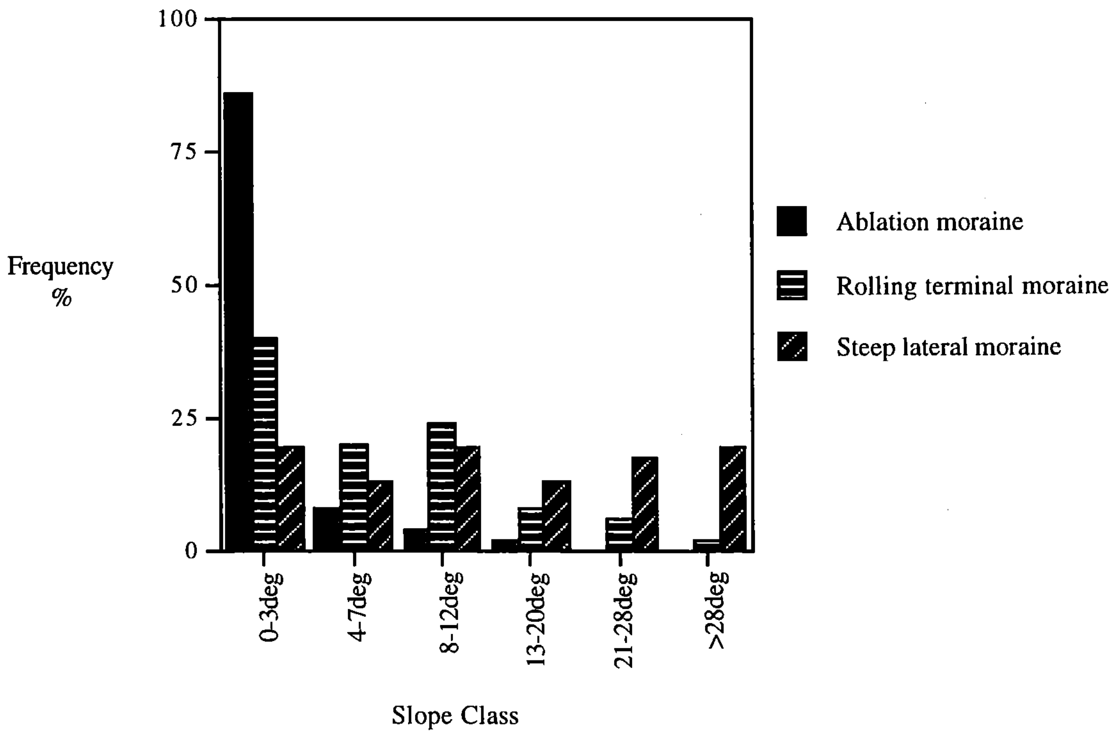

Fig. 5. Slope class distribution by landform.

formed directly in till on stable landform components such as ridge and nose crests or colluvium derived from till on other landform components.

Soils on moraine ridge crests, sideslopes and footslopes, and nose crests, sideslopes and footslopes are most commonly SPCs 9 and 10. SPC 10 appears to be more strongly associated with nose crests. These SPCs are well drained oxidised soils with yellow-brown subsoils. They are well aerated because water incident on the soil surface is conducted rapidly downslope or shed from narrow crest positions by subsurface flow through porous organic surface horizons and does not percolate. Soils in hollows, where flowlines converge, are more poorly drained with SPC 6 most common. Soils on the very wet toeslopes of moraine ridges are dominantly poorly drained Humose Perch-gley Podzols (SPC 3) or extremely poorly drained, deep Organic Soils (SPC 1.2). Most microvariability within component landforms principally stems from short-range hydrological changes. Along with the dominant soil profile class there is a range of soils of different drainage status. The occurrence of SPC 4 in hollows and toeslopes is thought to result from deposition of eroded material from surrounding slopes.

The M5 terminal moraine complex (rolling terminal moraines) is for the most part a relatively low relief mound and hollow landform but with repeating patterns of arcuate recessional moraines separated by depressions on the inner (southern) margin. Modal slope class is 0-3° (40% of sampled slopes) with a further ca. 40% in the range 3 - 12° (Fig. 5). Slope aspects are highly variable in accord with the prevailing deranged drainage pattern, and there are many undrained inter-mound hollows. Soils have formed in often compact, matrix-supported till with angular and subangular clasts. A relatively stone-free silty material overlies till in most soils, and it thickens toward the base of slopes in accord with other studies of similar terminal moraine landscapes (Gerrard, 1992; Ruhe, 1969). This material may be partly loess, but based on its particle-size distribution, which is more clay-rich, it is most likely to have formed by *in situ* weathering (Almond, 1996).

The pattern of soils at mesoscale is usually Perch-gley Podzols on upper slope positions (SPC 8 on crests and SPC 7 on shoulders and upper midslopes), Gley Soils (SPC 6) on midslopes to upper footslopes and Silt-mantled or Humose Perch-gley Podzols (SPC 3 and 4) on lower midslopes to toeslopes. Organic Soils (SPCs 1.1 and 1.2) occur in toeslopes and concave inter-mound hollows associated with Silt-mantled or Humose Perch-gley Podzols. Where summits of mounds or ridges are broad, SPC 6 may be dominant (Fig. 6). Sampling by component landform can lead to the impression that soil pattern is controlled by catenary gradients. However the distribution of SPCs within component landforms highlighted by the grid surveys suggests there are sources of variability at shorter range, independent of these gradients. Slope profiles are highly irregular and are usually broken by benches or swales. Figure 6 (inset) shows soil variability on a 5 m slope segment from a secondary ridge to swale near the shoulder of a moraine slope. Soils vary from Perch-gley Podzols (SPC 8) on the crest

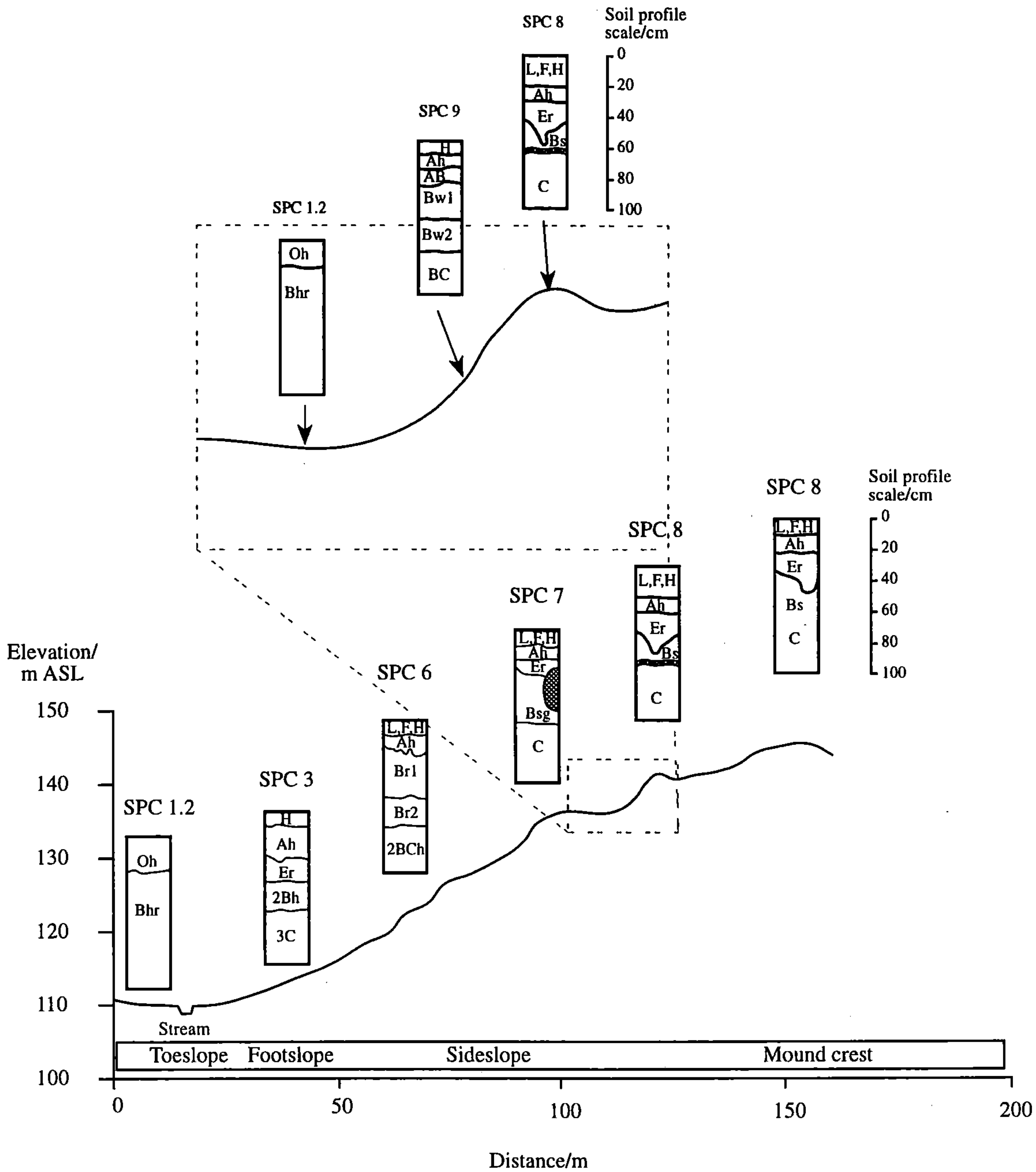

Fig. 6. Soil variation at meso- and microscale in rolling terminal moraines.

to Acid Brown Soils (SPC 9) on the steep sideslope to Acid Gley Soils (SPC 6) on the footslope to Organic Soils (SPC 1.2) in the swale. It appears that within any section of a longer slope, shorter range soil variation can be of the same order of magnitude as at the longer range mesoscale. Although in this example soil variation can be linked to microtopography, this is not always so suggesting further sources of soil variability at a microscale.

The ablation moraines include the low-lying areas with impounded drainage behind (upstream of) the rolling terminal moraines. The depression left after the disappearance of ice is now infilled with fine-textured alluvium and peat through which chaotically arranged piles of melt-out till (ice disintegration features) protrude. The great majority of slopes are less than

3° although on till mounds a few slopes are in the 4-7° class and fewer still in the 6-12° class (Fig. 5).

Meso and microscale soil variability on till mounds is similar to the rolling terminal moraines. However their comparative rarity means that they were poorly represented in the grid surveys (Table 1). On valley floors the soil pattern is highly unpredictable, and five SPCs are represented. Patterns appear to be controlled to some extent by surface relief and alluvial deposition of silt, operating over mesoscales (10-100 m). In low-lying areas where water ponds, thick, fluid peat has accumulated (SPCs 1.2 and 2.2). On raised stream levees soils are more mineral and formed in silt over gravel; SPC 5.2 occurs where depth to gravel ≥ 50 cm and SPC 3 where it is less than 50 cm. A good deal of variability is short-range and unpredictable and probably arises from similar sources operating at shorter range.

The very subtle relief on terraces meant that it was impossible to recognise component landforms during the grid surveys. For this reason there is no subdivision of soils by component landform for T4, T5 and T6 in Table 1. Lack of stratification at a mesoscale confounded the recognition of patterns at this scale, and meso- and microscale variability were difficult to separate. From transect surveys and exposures in drains some patterns were recognised.

On T4 variability was due to:

1. Stream dissection. T4 emerges from T5 immediately downstream of the rolling terminal moraines and in this area small meltwater streams dating to the M5 advance have removed the loess. The former channels are now infilled with peat (SPC 1.2 and 2.2) (Fig. 7, transect 1: 48-60 m).
2. Proximity to terrace edge. Close to terrace edges drainage is improved, and soils more commonly belong to SPC 5.1 with some variants displaying an olive subsoil. Toward the centre of wide (>100 m) terrace remnants these soils are sub-dominant to Organic Soils overlying loess (SPCs 1.1 and 2.1). Without accurate surface levelling it is not clear whether the differences in thickness of peat reflect changes in surface elevation or relief in the surface of the underlying loess. However since the loess beneath the peat usually shows an intact soil stratigraphy, it is likely at least some of the variations are propagated through the loess from relief on the underlying gravels.
3. Alluvial deposition. Soils found alongside the ephemeral streams which flow over T4 are often deep and silty but lack the soil stratigraphy characteristic of the loess (SPC 5.2). These soils are presumed to have formed in reworked material.

On T5 soil meso- and microscale soil variability arises in part from the surface hydrology inherited from the original channel and bar pattern of the depositional surface and immediately post-depositional dissection. Some of the patterns are demonstrated along the detailed transects for which elevation data are available. Over much of the surface there is a silt

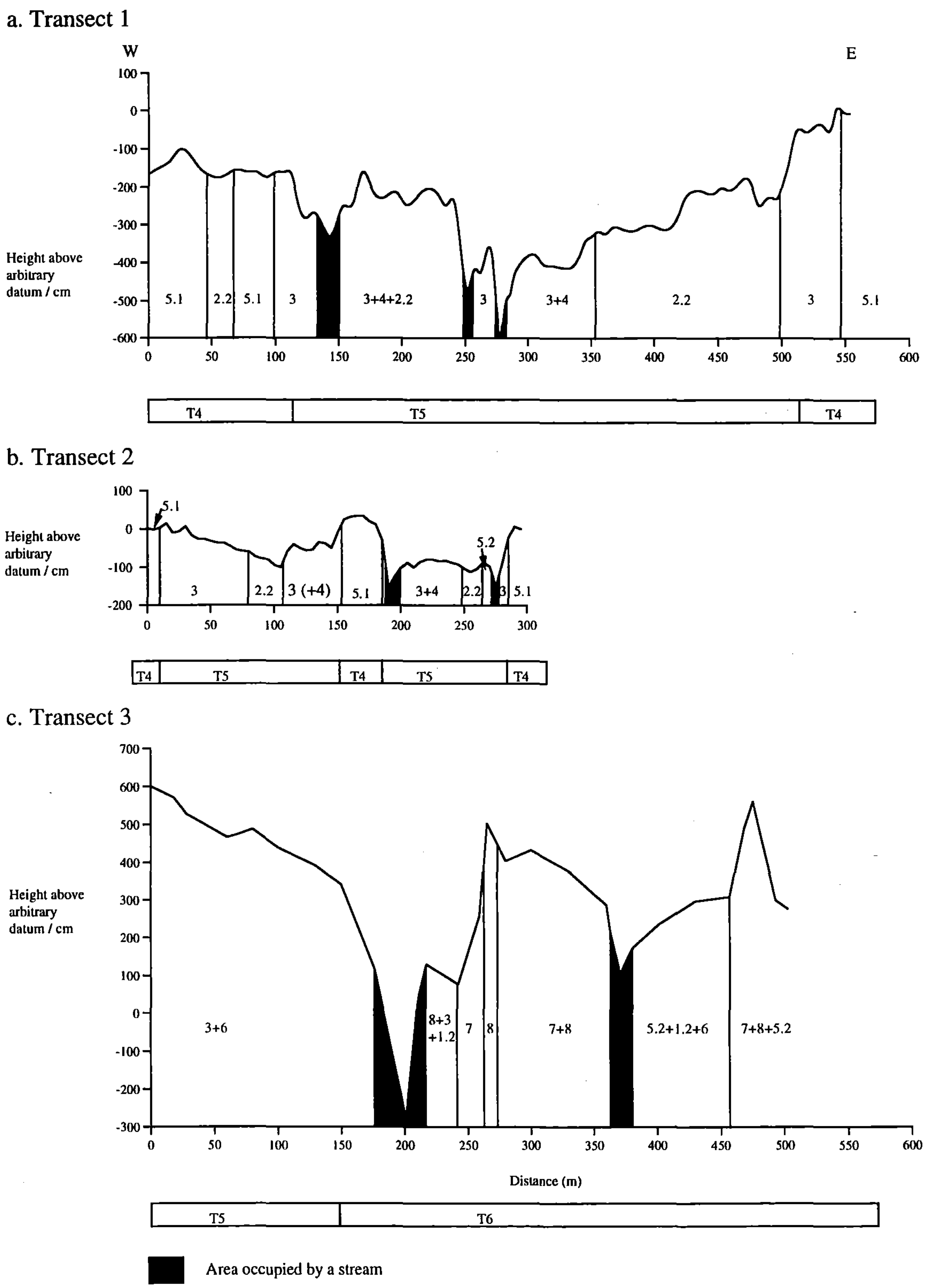

Fig. 7. Distribution of soil profile classes along transects across terraces in the study area. Transects 1 and 2 (Fig. 7a and b) cross T4 and T5 (surface elevation data from Cornere, 1992). Transect 3 crosses T5 and T6 (surface elevation data from Almond, 1986). For the location of transects see Fig. 3.

loam veneer on the gravels obvious as a blue-grey silty horizon in the upper part of soils (SPC 3). Where it is absent, soils form directly in gravels (SPC 4). In former channels fluid peaty O horizons overlie moderately to very fluid humic silt loam on a 2Bh horizon in gravels (e.g. Fig. 7, transect 2: 80-110 m and 250-260 m). SPC 2.2 is recognised where the fluid horizons total between 20 and 70 cm and SPC 1.2 where they are greater than 70 cm. Some inset terraces and former backswamps have extensive areas of Organic Soils (e.g. Fig. 7, transect 1: SPC 2.2 from 350-500 m) although usually Organic Soils and Perch-gley Podzols are intimately associated (e.g. Fig. 7, transect 1: 150-250 m). Levees near streams often have greater thicknesses of silty material over gravels and SPC 5.2 is recognised (Fig. 7, transect 2: 270-275 m).

The surface relief of T6 can be as much as 9 m but usually less than ca. 3 m and is due to erosional remnants of higher terraces, degradational terraces and incised channels. As on T5 there is a veneer of silt loam material over gravels, and the suite of soils is similar to T5. However, there is a more integrated drainage because of the greater surface relief and the incised channels, and for this reason Organic Soils are less common. Over most of the unit, like T5, the soil pattern is determined by micro-topographical features and patterns of sediment deposition inherited from the braided rivers which formed the surface. However, in strongly dissected terrain soil associations similar to ablation or rolling terminal moraines occur on erosional remnants i.e. SPCs 6-8 from sideslopes to summits. Patterns on terraces are similar to T5 (Fig. 7, transect 3).

4. Discussion

At the macroscale changes in "mean" soil properties, reflected in shifts in abundance of SPCs are a result of differences in gross landform morphology and age. Landform morphology (slope, relative relief, dissection, drainage pattern) influences hydrology. The transmission time of water through the soil has a large bearing on the nature of soils. The high drainage density and steep slopes on steep lateral moraines mean that water is efficiently conducted and most soils are well drained. As relative relief decreases and the proportion of low angle or concave slopes increases from steep lateral moraines through rolling terminal moraines to ablation moraines, the proportion of poorly drained soils increases. The effect is enhanced in ablation moraines because drainage is impounded by the terminal moraines.

A similar sequence occurs from T6 to T4 as a result of declining terrace slope and surface relief. Landform age may also be significant. Because of its age T4 is mantled with 1 m of loess. Similar, massive loess has been implicated in the low hydraulic conductivity of other soils in Westland (Young, 1967). Even if this material were not limiting, the gravels beneath T4 are ubiquitously iron cemented in the upper 10 cm and would be a barrier to drainage. When the iron pan is punctured, gas escapes from the underlying gravels (pers. obs.) indicating that they are well sealed.

Variability within landforms, particularly moraines, is partly systematic when considered at the mesoscale. At this scale variation arises from gradients of water content and fluxes which in turn determine soil aeration, leaching and organic matter content. Sequences of soils down slopes show some order when considered in relation to component landform and slope position, but they cannot be considered to be soil catenas in the strict sense (Gerrard, 1993) because soil development is not linked from top to bottom of slopes. Soil development on slopes apparently proceeds as a series of spatially independent soil catenas over small slope segments within component landforms. The nature of soils within each slope segment may be influenced by hydrological regime within landform component (crest vs sideslope vs toeslope) giving rise to the trends seen at a mesoscale. A similar lack of integration of soils on slopes is described by Gerrard (1990) in a study of soil variation on hillslopes in humid temperate environments. Mesoscale variability on terraces is partly systematic only where detailed landform cross-sections allow delineation of subtle component landforms. In grid or free survey variability appears essentially random at this scale.

Microscale soil variability is clearly a major component of soil variability at meso- and macroscale (Fig. 6, 7 and Table 1). Short range changes in soil moisture regime appears to be a major source of variability. The type of variability demonstrated in Fig. 6 (inset) is also reported by Simmons (1982) who studied a 180 m long soil section from the summit of an mound through a inter-mound hollow to the summit of an adjacent mound in the rolling terminal moraine (Fig. 3). Within similar broad trends from mound to hollow as reported here, he found on a 35 m slope a repetitive pattern of SPC 6 and SPC 8 (my classification) coinciding with changes within the range of meters, from sloping to flat slope segments. The changes in morphology were also mirrored in behaviour of water tables monitored from dipwells.

Microhydrological gradients may occur independent of microtopography. Derks (1996) investigated the effects of tree canopy on organic horizon depth and pH in two plots (17 x 14 m and 13 x 8 m) in the rolling terminal moraines in the study area. In both plots large rimu or rata trees were emergent over a lower kamahi/quintinia canopy. Organic horizon pH beneath canopies varied significantly between species, and pH varied in a systematic way within large (ca. 10 m diameter) canopies. Organic horizon pH variation within canopies showed a consistent pattern from tree trunk to canopy edge for both species; pH was high at the trunk, reached a minimum immediately off the tree bole, increased to a maximum at mid-canopy and declined again to canopy edge. The variations in acidification were thought to be caused by concentration of intercepted rainfall toward the trunk as stem flow and toward the canopy edge as drip. Semi-variograms of soil pH for the two plots (Fig. 8) reach a sill between 10 and 14 m indicating spatial dependence out to this range. The rise of semi-variograms from the nugget

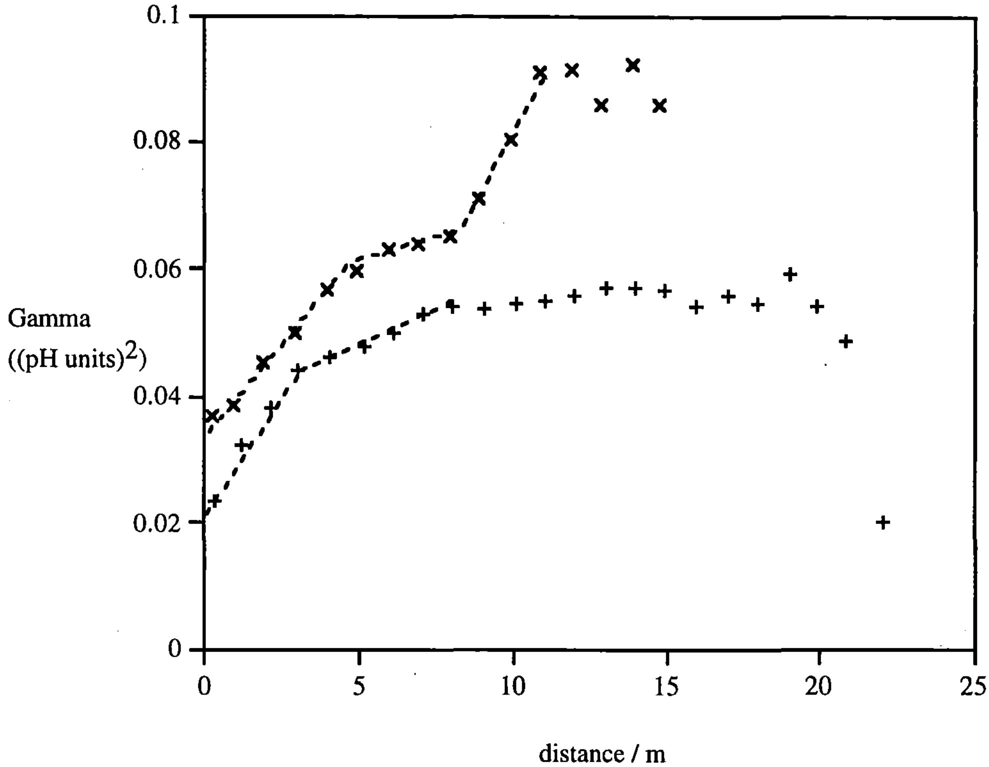

Fig. 8. Non-directional semi-variograms for a 17x14 m plot (+) and a 13x8 m plot (x) in rolling terminal moraines, Saltwater Forest (after Derks, 1996). (+ samples from a 1x1 m grid, x samples from a 0.5x0.5 m grid, dashed lines fitted by eye to linear segments).

variance to the sill can be resolved into linear segments for both plots indicating nested scales of variability (Burrough, 1983) related to intra-canopy and inter-canopy variation. The scales of the nested structures are dependent on the size and geometry of trees in the plots.

Given that many of the soil processes dominant in this environment are driven by micro-site hydrology and thresholds of soil pH (Campbell, 1975), I believe that a significant proportion of short range variability in soil morphology, but also soil chemistry and mineralogy, relates to vegetation history of a site. A possible example of this phenomenon was encountered on the shoulder of a slope in the rolling terminal moraines. Here soil profiles at either end of a 1.5 m long pit belonged to SPC 8 and SPC 6. Along with the morphological changes, indicating locally enhanced gleying, soil pH and acid-oxalate extractable Fe and Al were lower in the soil belonging to SPC 6. These differences can only be attributed to enhanced leaching, acidification and translocation of Fe and Al. Gleying could account for the loss of Fe in the low chroma Br horizon although no micro-topographical features, which might be the cause were apparent, but the loss of aluminium could not. Aluminium is not mobilised by redox reactions, and therefore its removal from one soil profile can only be attributed to enhanced leaching, acidification and concentration of dissolved organic acids (Funakawa et al., 1992; Funakawa et al., 1993; Gustafsson et al., 1995; Lundstrom, 1993) such that might exist beneath different parts of a tree canopy (Gersper and Holowaychuk, 1970b; Johnson, 1990; Stout and

McMahon, 1961). This phenomenon could also account for short-range variation in soil moisture regime (Gersper and Holowaychuk, 1970a; Gersper and Holowaychuk, 1970b).

The effect of tree windthrow on soils and microtopography has been widely documented (Schaetzl et al., 1989) and specifically studied in Saltwater Forest by Adams and Norton (1991). The latter authors found that trees disturbed areas of soil with average dimension 3.2 m x 1.9 m and commonly produced a pit where roots uplifted soil and a mound where soil material was deposited after roots rotted. They studied two soil profiles in detail, a Perch-Gley Podzol (SPC 7) and an Organic Soil (SPC 2.2). The effect of windthrow was to produce complicated composite soil profiles on mounds and Organic Soils in the water filled pits. Conditions on mound sites appear to favour a redirection of pedogenesis toward increased podzolisation rather than a rejuvenation as reported in other studies. Norton (1989) modelled soil turnover with time due to tree windthrow with a first order rate (exponential decay) function in order to account for the probability of trees reestablishing on previously disturbed sites. Using data from Westland podocarp forest he calculated a half-life of soil turnover (the time after which half of the soil has been disturbed) of 2960 years. According to the model and assuming forest has been in this landscape 10 ka (McGlone et al., 1993; Moar, 1980), presently only 10% of the soil area remains undisturbed and 65% has been disturbed twice or more.

The impact that tree windthrow has on soil depends on the proportion of trees that die by toppling, the depth of rooting and stand density. The calculations above assumed a stand density of 185 stems/ha, typical of terrace rimu forest, and that 70% of trees die by toppling. Simmons (1982) presented evidence for a difference in style of wind disturbance of trees on relatively well drained, mineral soils against those on Organic Soils in this area. The former were more prone to snapping while the latter tended to fail at the root plate.

Profile distribution of roots in the poorly drained soils which dominate the terraces have a peak in the upper soil horizons and a secondary peak of fine and coarse roots at depth (Fig. 9). In better drained soils characteristic of the moraines the maximum root density is in the soil surface and declines with depth. Furthermore there are few coarse roots and no very coarse roots at depth. Together with the fact that tree density is on average greater on terraces than moraines (Almond and Duncan, submitted), I infer a difference in the importance of tree windthrow to soil variability on different soils and landforms; it would be expected that a far greater proportion of soils had been influenced and to a greater depth on terraces than moraines.

This highlights an inconsistency however: 90% of the soil should be disturbed at least once in terrace forest and yet on T5 and T6 SPC 3 is abundant. Occurrences of SPC 4 may represent windthrow-disturbed sites but unless soil differentiation is very rapid, the dominance of SPC 3 indicates that tree uprooting is not as important a bioturbatory phenomenon as

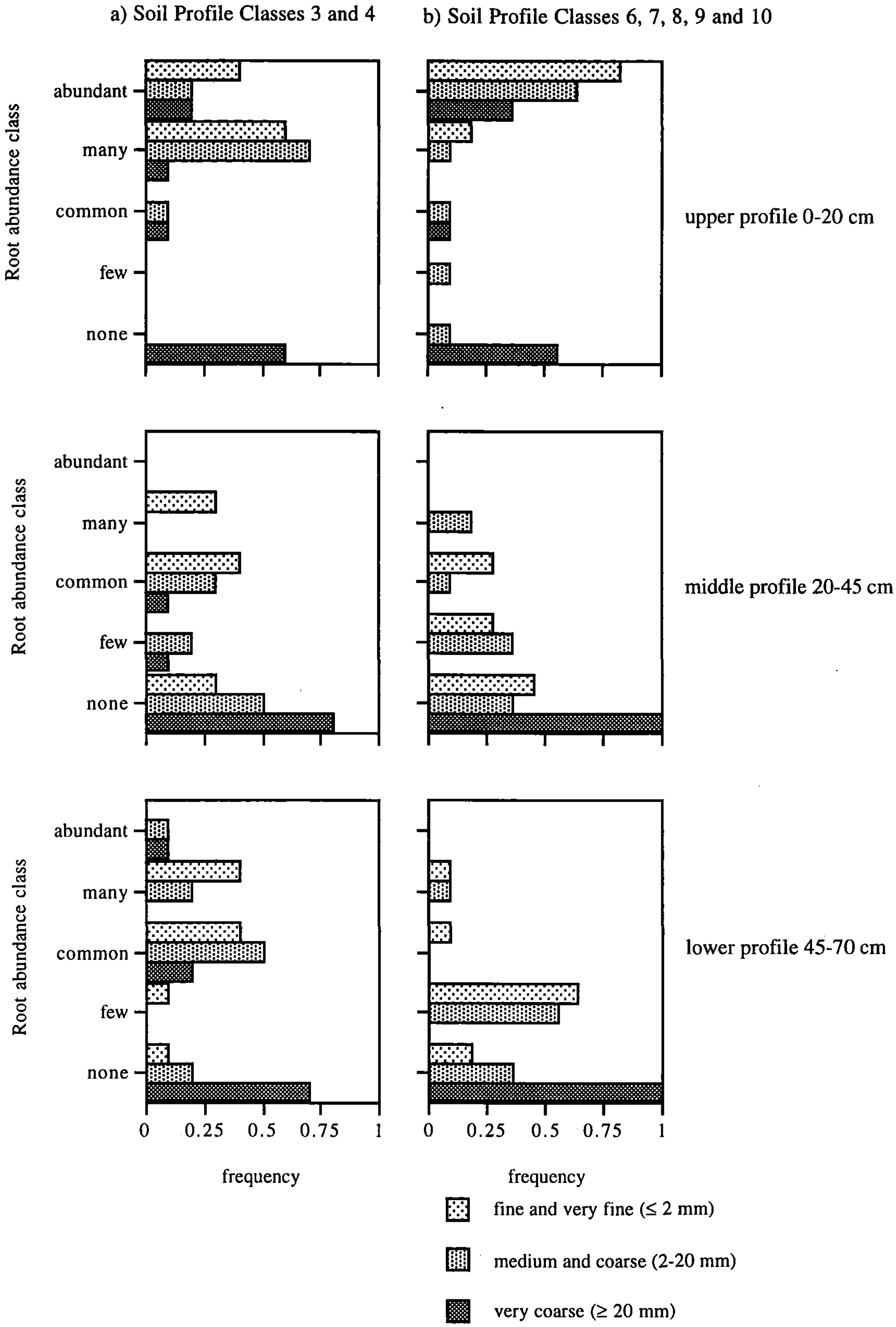

Fig. 9. Frequency distribution of root class with profile depth in a) terrace soils and b) moraine soils.

Norton's (1989) windthrow model would suggest. Departure from model predictions could be accounted for by a strong tendency for trees, particularly rimu since it is dominant, to reestablish on disturbed sites. This has been observed (Norton et al., 1988). Reoccupation rate would need to be in the order of 85% to result in 30% of soil having been disturbed after 10 ka using other parameters above. Another explanation is that fewer trees die by toppling than has been assumed. For 30% of the soil to be disturbed at least once only 10% of trees must die by toppling, the other 90% die standing. Further work is required to resolve this conundrum, but it is worthy of attention since there are significant implications for soil evolution and forest dynamics.

Changes in soil profile class derived from soil microvariability can be as diverse as that of the mesoscale. Some potential sources of soil microvariability have been discussed, but there is no doubt a component which is unpredictable and incomprehensible. This noise, apparent even at a microscale, may be a result of stochastic complexity i.e. short range stochastic variability in soil processes and environmental conditions, or it may arise from inherent features of the soil system. Soils are nonlinear dynamical systems that are likely to display deterministic chaos (Phillips, 1993), a phenomenon which is characterised by sensitive dependence on initial conditions. Very small differences in any of the state factors, climate, relief, parent material or vegetation over time will persist or cause increasing divergence in soil properties rather than being offset or damped. Phillips (1993) shows that for the soil system to behave in a stable manner (i.e. lack sensitive dependence on initial conditions) the interactions between soil development and parent material must outweigh the interactions between soil development and vegetation cover, soil development and relief, and climate and vegetation cover. Climate-vegetation linkages are presumably small over Saltwater forest, but the evidence presented so far indicates that soil-vegetation and soil-relief interactions are likely to outweigh soil-parent material linkages on most landforms.

Despite the nature of the origin of soil microvariability (systematic, stochastic, chaotic), in the context of soil survey, it appears as a random component of sufficient magnitude to thwart delineation of taxonomically pure mapping units. Even at scales of 1:5000 which would allow delineation of areas with dimension of 50 m, mapping units would still be soil complexes. Currently the minimum management unit has spatial dimension in the order of 1600 m which normally would include more than one landform unless compartments are designed specifically to coincide with landform boundaries. The intensity of management does not warrant mapping of soils at scales any larger than that which allows reasonable delineation of landforms and forest infrastructure (about 1:50 000). Maps at this scale would produce coherent mapping units highlighting the systematic variation between different landforms whereas more detailed maps would obscure the macroscale gradients with detail irrelevant to the intensity of land use. Should intensity of management change in future, there would be no significant gain in remapping areas at larger scales unless management units were in the order of 0.25 ha, although it is unlikely that the high cost of soil survey at this scale would be economically justifiable. However with some increases in management intensity there would be value in applying predictive soil-landform models appropriate to the mesoscale for field based management decisions. A model would predict soil profile classes with an assigned probability. On terraces where there is extremely subtle relief, soil-landform models would have nothing to offer but on moraines the soil-landform component relationships presented here (Table 1) would form the basis of a soil-landform model. Rigorous evaluation of predictive ability (reliability) would be necessary.

6. Conclusions

Soil variation in the study area arises from sources operating over different spatial scales. Macrovariability over the range of 100s of meters corresponds to changes in landform and results in significant, sometimes abrupt changes in 'mean' soil properties reflected by the abundance of different classes of soil. Variation arises principally from changes in the combination of parent material, time and landform hydrology factors. Mesoscale variability over the range of 10s to 100 meters is to some extent predictable and arises from hydrological variation determined in part by component landform and slope position. Soil variation on slopes cannot be considered to be catenary because soil-slope sequences are not integrated. Variability arises at short range (microvariability) from microtopography and individual tree effects. The latter include soil disturbance by tree uprooting and changes in leaching and soil moisture within and between tree canopies. The separation of micro- and mesovariability is to some extent arbitrary as they can arise from similar processes acting over overlapping spatial ranges.

The distribution of soil variability over different scales and the present intensity of forest management dictates soil map scale. A soil map scale and soil survey intensity which allows delineation of macroscale soil variability (landform boundaries) and accurate placement of management units (compartments) and infrastructure should be used. A map scale of 1:50 000 is sufficiently large to do this. More intense soil survey and a larger map scale than is required to achieve the above are not justifiable. Larger scale maps would not provide extra useful information at the current intensity of land management and the further detail would only serve to obscure soil patterns of relevance.

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Chapter 6

The structure of *Dacrydium cupressinum* (Podocarpaceae) dominated forest along landform and soil gradients, South Westland, New Zealand

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R. P. Duncan[2]

Abstract

The frequency at which natural disturbances form canopy gaps, and initiate the establishment of canopy trees, may vary across the landscape. Where susceptibility to disturbance varies in a systematic way, forest-size and age-structure may vary accordingly. Regions of high disturbance frequency will be dominated by young cohorts of small diameter trees whereas areas disturbed less frequently will have have older, larger trees. We examined diameter size-class variation in a *Dacrydium cupressinum* dominated forest on glacial landforms on the west coast of South Island, New Zealand, and found that forest size structure varied systematically with changes in soil drainage. The soil pattern in the study area is characterised by large short-range variability, but there are broad differences among landforms. Dominance of soils on moraine landforms progressively shifts from well drained Podzols or Brown Soils on steep lateral moraines to Gley Soils on rolling terminal moraines to extremely poorly-drained Organic Soils on ablation moraines. Across terrace landforms, dominance shifts from poorly drained Perch-gley Podzols to extremely poorly drained Organic Soils from the youngest to the oldest terraces. Diameter data from 311 randomly located 0.2 ha plots showed that mean tree diameter declined and average stem density increased as the abundance of poorly-drained soils on each landform increased. We suggest this predictable relationship between landform and forest size structure reflects the greater susceptibility of trees growing on poorly drained soils to natural disturbances such as windstorms or earthquakes. The frequently disturbed areas of poorly drained soil are dominated by cohorts of young, small diameter trees. Additionally, we examined the relationship between *Dacrydium cupressinum* age structure and soils in more detail along two 20 m wide transects (300 and 600 m long). We found that spatially discrete cohorts of trees with restricted ages coincided with abrupt changes in soils and landforms. The youngest trees grew in patches of extremely poorly-drained Organic Soils, adjacent to older cohorts of trees on better-drained mineral soils. This relationship between soils and cohort age supports our interpretation of the landform-scale study; trees on poorly-drained soils are more susceptible to disturbance and are turning over more often than trees on better drained soils. This fine-scale variation in turnover rate manifests itself as the systematic variation in forest size-structure apparent at the landform scale. The patterns

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of forest structure and turnover-frequency we have highlighted have important implications for the sustainable management of these forests.

Keywords: Forest disturbance; disturbance frequency; soil-vegetation relationships; windthrow; earthquakes

1. Introduction

Most forest trees reach the canopy by growing into openings formed by the death of existing canopy trees. In natural forests, the death of canopy trees is often associated with natural disturbances such as fires, floods, or windstorms, and the resulting canopy gaps are colonised by cohorts of potential new canopy recruits. Consequently, the frequency at which canopy gaps are formed by disturbance is a fundamental driving force in the dynamics of forest systems as it influences the rates of canopy tree mortality and recruitment, and such processes as the rates of biomass turnover and nutrient flux in forest systems (Borman and Likens, 1979; Clark, 1991; Vitousek and Reiners, 1975; Waring and Schlesinger, 1985).

The frequency at which canopy gaps are formed by disturbance can vary in different portions of a heterogeneous landscape and may be predictable from site characteristics. Perhaps the best studied example is spatial variation in the frequency of forest fires which can burn at different frequencies in regions that differ in topography and rates of fuel accumulation (Heinselman, 1973; Romme, 1982; Zackrisson, 1977). Sites that burn frequently tend to carry younger cohorts of smaller trees at earlier stages of forest development following fire, than sites that burn less often (Romme and Knight, 1981; Segura and Snook, 1992). Other natural disturbances may similarly interact with landscape features to generate predictable spatial variation in the rates of forest turnover, and this is often manifested as predictable variation in the ages and sizes of trees growing in different parts of the landscape (Bellingham, 1991; Jane, 1986).

Dacrydium cupressinum Lamb. (Podocarpaceae) is the dominant forest canopy tree on the lowland Pleistocene age landforms of the coastal piedmont in Westland, New Zealand. For a long time it has been observed that the structure of *Dacrydium*-dominated forest in Westland varies across different landforms (Chavasse, 1971; Holloway, 1954; Norton et al., 1988; Norton and Leathwick, 1990; Poole, 1937; Wardle, 1977). On undulating or steep moraine surfaces, *Dacrydium* typically occur as large, widely spaced trees over a lower canopy of hardwoods. In contrast, *Dacrydium* on adjacent outwash terraces grow in denser stands and the trees, on average, appear to be smaller in diameter.

Historically, it has been suggested that these differences in forest structure are a consequence of *Dacrydium* having different modes of regeneration on the different landforms. On the poorly-drained terraces, it was hypothesised that *Dacrydium* established episodically in large canopy openings (up to 20 ha in area) formed by catastrophic windthrow of the original canopy trees. The terrace forest therefore comprised a mosaic of large, relatively even-aged cohorts of trees regenerating after major disturbances. In contrast, *Dacrydium* establishment on the better-drained moraines was thought to occur more frequently and in smaller gaps, with trees regenerating in canopy openings formed by the fall of just one or a few trees. The moraines were therefore thought to comprise widely spaced older trees interspersed with small groups of trees of various ages regenerating in scattered treefall gaps.

Three recent studies have investigated the regeneration dynamics of *Dacrydium cupressinum* on terrace and moraine surfaces by ageing trees and reconstructing the temporal and spatial patterns of tree establishment (Cornere, 1992; Rogers, 1995; Stewart et al., submitted). Contrary to expectation, *Dacrydium* appears to regenerate in a similar manner across both the terrace and moraine surfaces. On both landforms, trees typically occur as cohorts with a restricted age range in often large patches, suggesting episodic canopy recruitment after major and often widespread disturbance to the previous canopy. However, at least in the areas sampled in the above studies, the moraine surfaces are dominated by older trees while the terraces have a mix of old and younger trees (Stewart et al., submitted), a pattern that could account for the generally larger size and lower density of *Dacrydium* on the moraines compared to the terraces. This suggests that canopy openings suitable for *Dacrydium* regeneration are formed more frequently on the terraces than on the adjacent moraines, as the predominance of younger trees on the terraces implies more recent cohort initiating disturbances and a higher rate of forest turnover (canopy tree mortality and subsequent cohort recruitment).

Why would trees growing on the terraces suffer higher rates of mortality and be turning over more often than trees growing on the moraines? Windstorms are thought to be the major disturbance agent in *Dacrydium cupressinum* forest (Hutchinson, 1932; Poole, 1937; Six Dijkstra et al., 1985) and one possibility is that trees growing on the poorly-drained terraces are less wind-firm than trees growing on the better-drained moraines, and so are more susceptible to blowing over during windstorms. There is some evidence to support the suggestion that poorer drained sites are disturbed more frequently. Within plots of less than 0.5 ha in area, Rogers (1995) observed that areas of poor drainage characteristically carried young, approximately even-aged groups of *Dacrydium* trees, while neighbouring, better-drained areas supported patches of older trees. The older trees on the well-drained sites must have survived the disturbance events that removed the canopy trees in the poorly-drained areas and initiated establishment of the younger cohorts of trees.

The above studies hint that the frequency of canopy gap formation that drives the process of tree replacement in *Dacrydium* forest is intimately tied to landform and soil variation. Our aim in this study is to test the generality of this hypothesis by investigating how the age, diameter, and density of trees vary across Pleistocene age landforms in lowland Westland. To do this, we take advantage of a detailed landform and soil survey of Saltwater Forest, south Westland (Almond, submitted). If poorly-drained sites are more prone to canopy disturbance and are turning over more frequently than well-drained sites, then poorly-drained areas should carry a greater proportion of younger cohorts of smaller diameter trees compared to well-drained areas. We test this hypothesis at two spatial scales. First, at a broad scale we use tree-diameter and density data from 311 0.2 ha plots in 4000 ha of Saltwater Forest to compare forest structure across macro-landform units that differ in soil drainage. Second, at a smaller scale we use age data from a previous study (Cornere, 1992) to examine how the age of *Dacrydium* cohorts vary in relation to soil drainage along two transects several hundred meters in length.

Dacrydium cupressinum is the most important native timber species in New Zealand. The majority of *Dacrydium* dominated forest used for timber production has been clearfelled and Saltwater Forest is now one of only two forests in New Zealand that have been set aside for the sustainable management of *Dacrydium cupressinum* for timber production. Clearly, an understanding of the way *Dacrydium* regenerates in these forests, and in particular the degree to which patterns of natural canopy mortality and cohort recruitment may vary along landform and soil gradients is important to developing sustainable harvesting strategies, particularly if harvesting aims to mimic the natural processes of tree replacement. We therefore consider the findings of this study in relation to the past and current management of *Dacrydium cupressinum* dominated forest.

2. Study area

2.1. Climate

The study area is a 4000 ha portion of Saltwater Forest in south Westland, South Island, New Zealand (Fig. 1). The climate of the area is superhumid and mild temperate. Mean annual rainfall at Harihari, 10 km to the east of Saltwater Forest is 3789 mm (New Zealand Meteorological Service, 1980) but over the study area we expect a range in annual rainfall from 3000 to 5000 mm since the study area falls along a steep rainfall gradient increasing from the coast to the Southern Alps in the southeast (Griffiths and McSaveny, 1983). At Harihari mean monthly temperatures range from 16.3°C in January to 6.9°C in July (New Zealand Meteorological Service, 1980).

2.2. Geomorphology

We use the system of landform nomenclature in Almond (1996) and refer the reader to that source for a detailed description of the geomorphology and stratigraphy of the study area. We summarise the key points below.

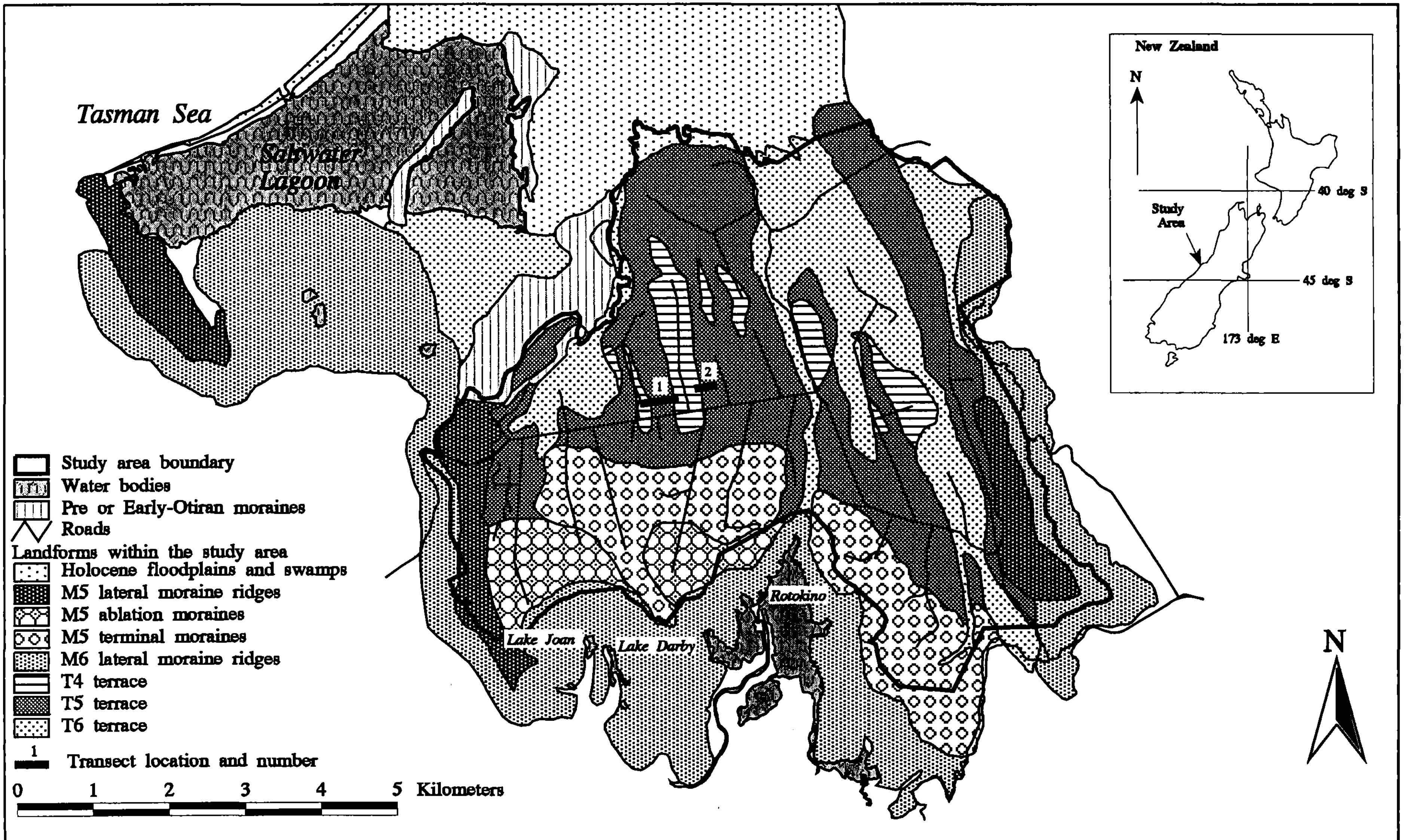

Fig. 1. Location and landforms of the study area showing transect locations.

In the study area, landforms have originated from three glacial advances in the late Pleistocene. The oldest advance, which occurred sometime before 35 ka, produced an extensive outwash aggradation terrace. This terrace was dissected by outwash streams and rivers from a late Last Glacial advance (18-22 ka). Terrace T4 comprises the remnants of this old terrace, now mantled by up to 1 m of loess deposited between 35 ka and ca. 14 ka (Fig. 1). The moraine associated with terrace T4 has not been found and we presume it was buried during the subsequent advance (18-22 ka) which produced moraine M5 and an associated outwash terrace, T5. M5 moraines are subdivided by morphology and genesis into rolling terminal moraines (M5RT), ablation moraines (M5ABL), and lateral moraine ridges (M5LAT - which, for the purposes of this study, also include the lateral moraines from the youngest, M6, advance that are indistinguishable from the M5 lateral moraines on soil properties). Terminal moraines form hummocky topography covering arcuate zones marking the termini of former ice lobes. Behind these, and in the area formerly occupied by ice, are ablation moraines. These formerly ice-filled depressions have infilled with peat and fine textured alluvium through which chaotically arranged piles of melt-out till protrude. M5 lateral moraine ridges flank the major river valleys on the margin of the study area. These are steep-sided, high relief, sinuous ridges formed on the flanks of the major ice-tongues that occupied the Poerua and Whataroa valleys. On the inner valley margin are moraines from the youngest advance (M6 lateral moraine ridges) between 16 and 14 ka. This advance did not produce terminal or ablation moraines in the study area but meltwater rivers flowed across the study area to form the youngest terrace, T6. Meltwater rivers were erosive in character and produced sets of degradational terraces and incised channels, all included in T6.

2.3. Soil Pattern

Organic soils and strongly leached, often gleyed, mineral soils are found in the study area. The latter form from alluvium, till and loess derived from the Mesozoic felsic sandstones and schists of the Southern Alps. The range of soil classes is most simply presented as a subdivision of a soil drainage gradient which includes Organic Soils, Podzols, Gley Soils and Brown Soils (Hewitt, 1992).

The soil pattern is extremely complex. A high proportion of soil variability arises from hydrological gradients, erosional and depositional processes and individual tree effects operating on a micro-scale (1-10 m) and a mesoscale (10-100 m) (Almond, submitted; Derks, 1996). At a macro-scale broad changes in landform morphology are reflected in differences in the abundance of the various soil groups recognised in the study area (Table 1).

Terraces are dominated by extremely poorly-drained and very poorly-drained soils. T4 has the highest proportion of extremely poorly-drained soils because of the prevalence of deep Organic Soils which have accumulated above a ca. 1 m thick mantle of loess on the aggradation gravels. T4 has the lowest slope of all the terraces which has likely compounded the poor drainage of this surface. T5 and T6 have progressively fewer extremely poorly-drained Organic Soils and a higher proportion of very poorly-drained mineral Humic or Silt-mantled Perch-gley Podzols. This is due to the lack of loess on these younger surfaces and their greater slope or relief. T6 comprises flights of degradational terraces and incised channel elements which give the surface greater relief and accordingly a more integrated and efficient drainage network.

Table 1. The relative abundance (%) of soil groups and constituent soil profile classes on different landform units. Abundance estimates are based on 50 soil observations from each landform made at the intersection of a 50 m grid on two randomly located 200x200 m areas. For details, including descriptions of soil profile classes see Almond (submitted). For each landform unit the number in bold highlights the most abundant soil group.

				Terraces		Moraines		
Soil Group (New Zealand Soil Classification, Hewitt, 1992)	Constituent soil profile classes	Drainage class	T4	T5	T6	Abl	RT	Lat
Deep Humic Organic Soils	1.1, 1.2	extr. poor	**56**	16	6	**40**	18	9
Humic Organic Soils	2.1, 2.2	extr. poor	22	34	17	17	10	2
Humic and Silt-mantled Perch-gley Podzols	3, 4, 5.1, 5.2	v. poor	22	**46**	**71**	29	20	11
Acid Gley Soils	6	poor					**28**	17
Typic Perch-gley Podzols	7, 8	imperfect-poor		4	6	14	22	13
Allophanic Brown Soils	9	well						**35**
Orthic Podzols	10	well					2	13

Moraine landforms, except the ablation moraines, have a greater proportion of moderately to well-drained soils than the terraces. The ablation moraines includes low lying areas behind terminal moraines with impounded drainage where deep Organic Soils dominate. As slope increases from the terminal moraines to the steep lateral moraines the dominant soil groups change from poorly-drained Acid Gley Soils and moderately well-drained Typic Perch-gley Podzols to well-drained Allophanic Brown Soils and Orthic Podzols.

2.4. Forest Pattern

The forests of the study area are dominated by conifers in the family Podocarpaceae, mainly *Dacrydium cupressinum* with some *Prumnopitys ferruginea* (D. Don) del Laub., *Lagarostrobos colensoi* (Hook.) Quinn, and on young river terraces *Dacrycarpus dacrydioides* (A. Rich) de Laub. All of these podocarps are long-lived trees, with *Dacrydium* being aged to over 1000 years in the North Island (Lusk and Ogden, 1992) and routinely living to greater than 500 years. *Dacrydium* is the dominant canopy tree, typically comprising > 75% of stems > 30 cm dbh and often forming a continuous upper canopy at 25-40 m (Norton and Leathwick, 1990). Below this there is a tier of broadleaf trees at 10-25 m, often with *Prumnopitys* and *Lagarostrobos*. The composition and structure of the forest varies along landform and drainage gradients (Norton and Leathwick, 1990). *Dacrydium* forms dense stands on the terrace surfaces, particularly on poorly-drained sites where there is usually only a sparse lower canopy of broadleaf species. On the poorest drained sites *Dacrydium* often shares canopy dominance with *Lagarostrobos*. On the moraine surfaces, where drainage improves the density of *Dacrydium* declines and it often occurs as an emergent over a dense canopy of broadleaf trees. *Prumnopitys* is common in the lower canopy on all but very poorly-drained sites.

3. Methods

Data from three independent sources were used to investigate the relationship between forest structure (the age, diameter, and density of the dominant canopy tree, *Dacrydium cupressinum*), landform, and soil properties at two spatial scales. At a broad-scale, the relationship between forest structure, landform, and soil pattern across 4000 ha of Saltwater Forest was investigated using data from an extensive forest survey of the area, combined with an independent landscape stratification and associated soil pattern analysis (Almond, submitted). At a fine-scale, the relationship between tree age-structure and small-scale topographic and soil variation was investigated along two transects, each several hundred meters in length.

3.1. Broad-Scale Study

A survey of the podocarp component of Saltwater Forest was carried out in the summers of 1985/86 and 1986/87. Saltwater Forest was stratified first by human modification (logged or unlogged forest), second by slope (flat or sloping), and finally by four forest canopy types identified from aerial photographs (G. Uhlig pers comm). An equal number of 0.2 ha circular plots were then randomly located between forest roads in each stratified unit and the location of each plot on the New Zealand standard map grid was recorded. In each plot, the diameter at 1.3 m (dbh) of all podocarp trees was measured and assigned to one of twelve size-classes (0-9.9 cm, 10-19.9 cm, 20-29.9 cm,...., > 110 cm). Trees of the three dominant podocarp species, *Dacrydium cupressinum*, *Prumnopitys ferruginea* and *Lagarostrobos colensoi*, were identified to species and all other podocarps were grouped in a category "Other". For our analyses, we excluded from our data set all plots in logged forest and all trees except *Dacrydium cupressinum*. We also confined our analysis to trees in the >10 cm diameter classes because in plots where there were many individuals in the 0-9.9 cm class a complete count was not attempted.

The locations of all plots were digitised into a Geographic Information System (ARC/INFO®) from their location on a 1:10 000 scale topographic map. In addition, the six landform units identified in the study area (Almond, 1996, see Study Area) were mapped onto aerial photographs and the unit boundaries digitised into ARC/INFO® with a separate coverage to the plot locations. Each plot was then assigned to a landform unit by overlaying the plot locations and the landform unit boundaries. We removed from our data set five plots that did not occur on the mapped landform units and two plots that contained no *Dacrydium* trees, leaving 311 plots for analysis. In each plot, we calculated the number of *Dacrydium* stems/ha in each of the eleven diameter-classes, the total number of *Dacrydium* stems/ha, and mean *Dacrydium* diameter, calculated as the average of the midpoint of each diameter-class weighted by the number of stems in that class. We then pooled the plot data by landform and compared mean *Dacrydium* diameter and density among landforms using analysis of variance.

3.2. *Fine-Scale Study*

We used data on the age of *Dacrydium* trees along two transects in Saltwater Forest, the first 560 m x 20 m and the second 300 m x 20 m, that were collected by B. Cornere in 1990 (Cornere, 1992). The transects were located in unlogged forest between three parallel roads, such that the first transect finished and the second transect started near the centre road. The transects began at random points along each road, and the start and finish of each transect was ca. 100 m from the road edges. Both transects traversed parts of the terrace landforms T4 and T5 (Fig. 1).

Data on the spatial location and the age from ring counts on increment cores were collected from each *Dacrydium* tree ≥ 5 cm dbh along both transects (Cornere, 1992). Cornere used the tree age and spatial location data to identify approximately even-aged patches of trees, representing cohorts established in past disturbance openings (see Duncan and Stewart, 1991). In addition, he surveyed with a dumpy level, the soil surface elevation at 5 m intervals along the middle of each transect.

We returned to Cornere's transects in 1992/3 and described the soil pattern by auguring holes at 10 to 50 m intervals (the distance between auger holes depended on the amount of local soil variability) along the middle of each transect. Five soil profile classes out of a total of 13 (Table 1) encompassed the range of soils on the two transects We visually examined the relationship between the age of *Dacrydium* cohorts and the soil pattern along each transect by overlaying the two on the same diagram.

4. Results

4.1. Broad-Scale Study

The diameter-distribution of *Dacrydium* trees in individual plots usually shows a lack of stems in the smaller size-classes. This was evident when we pooled the data from plots on the same landform units; on all units the *Dacrydium* diameter-distribution peaks in size-classes greater than 20 cm (Table 2). The lack of small stems in individual plots is consistent with our understanding of the way *Dacrydium* predominantly regenerates in these forests. Localised and episodic disturbances form canopy openings that initiate cohorts of a restricted age and consequently a restricted size range (Cornere, 1992; Rogers, 1995; Six Dijkstra et al., 1985; Stewart et al., submitted).

Table 2. Mean number of stems/ha by size-class, total number of stems/ha and mean tree diameter at 1.3 m (dbh) by landform unit for *Dacrydium cupressinum* in Saltwater Forest. For total stems/ha and mean dbh, values with the same letter indicate no significant difference among those landform units (Duncan's multiple range test, $P < 0.05$). For each landform the number(s) in bold highlights the size class with the greatest average stem density.

	Terrace			Moraine		
Landform unit	4	5	6	Abl	RT	Lat
Number of 0.2 ha plots	24	96	71	35	39	46
Dbh (cm)						
10-19.9	42	25.8	18	**45.7**	13	3.4
20-29.9	**50.2**	34.4	16.4	**45.7**	13.9	3.9
30-39.9	43.5	38.1	23.9	35.3	15.7	6.6
40-49.9	43	**40.2**	23	35.7	**16.1**	6.3
50-59.9	31.7	32.6	**24.2**	21	12.4	6.4
60-69.9	18.3	24.1	18.5	12.6	12.1	**9.8**
70-79.9	9.1	11.9	11.4	12.8	12.2	7.8
80-89.9	2.6	6.3	6.1	6	7.1	7.3
90-99.9	0.7	2.8	2.6	2.2	5.9	4.5
100-109.9	0.2	1.1	1	2.1	1.6	2.4
>110	.	0.5	0.6	1.6	2.7	2.5
Total stems/ha	241a	218a	146b	221a	113bc	61c
Mean dbh (cm)	42a	48ab	50b	49ab	61c	76d

There are marked differences in the structure of forests on different landform units (Table

2). Forest on terraces and on the ablation moraines has a higher density of *Dacrydium* and smaller diameter trees than those growing on the steeper terminal and lateral moraines. This variation in tree size and density parallels variation in soil drainage. The terraces and ablation moraines, dominated by denser, smaller trees, almost exclusively have very poor to extremely poorly-drained soils (Table 1). In contrast, the terminal and lateral moraines, with more widely spaced, larger trees, have a high proportion of poor to well-drained soils.

Among the terrace units, variation in forest structure similarly parallels differences in soil properties. *Dacrydium* stem density decreases and mean tree diameter increases going from the oldest, loess covered T4 through T5 to the younger T6. This increase in mean *Dacrydium* diameter parallels an improvement in soil drainage from T4, dominated by extremely poorly-drained often deep organic soils, through T5 with an equal mix of extremely poorly-drained Organic Soils and very poorly-drained mineral soils, to T6 dominated by very poorly-drained mineral soils.

The moraine units show the same pattern of increasing mean tree diameter and reducing density with improvement in soil drainage across landforms. The extremely poorly to very poorly-drained ablation moraines have on average the smallest diameter *Dacrydium*, the poorly to well-drained steep lateral moraines have low density, large diameter *Dacrydium*, while the terminal moraines are intermediate, comprising a mix of extremely poorly to imperfect-poorly-drained soils, and having trees of a mean density and diameter intermediate between those on the ablation and lateral moraines.

4.2. Fine-Scale Study

4.2.1. Transect 1

Cornere (1992) recognised five *Dacrydium* cohorts of restricted age-ranges along his first transect (Transect 1, Fig. 2a) that could have been initiated by four separate disturbance events. The oldest cohort includes trees 400-650 years old, occurring as a discrete clump at 0-150 m along the transect (cohort 1 in Fig. 2a), and as scattered individuals in the last 250 m. Cohorts 2, 3, 4 and 5 are spatially discrete and, judging by the age of the oldest tree in each cohort, were probably initiated by disturbances approximately 500, 375, and 250 years ago.

The position of the *Dacrydium* cohorts along transect 1 appears closely tied to the topographic and soil pattern, with younger cohorts of trees generally found in the areas of more poorly-drained soil. This is most obvious with the youngest cohort 5 that is confined to the lowermost degradational terrace (part of T5) dominated by extremely poorly-drained Humic Organic Soils. The boundaries of the remaining cohorts also broadly coincide with topographic or soil boundaries. Cohort 2 aligns with a terrace between two stream channels on

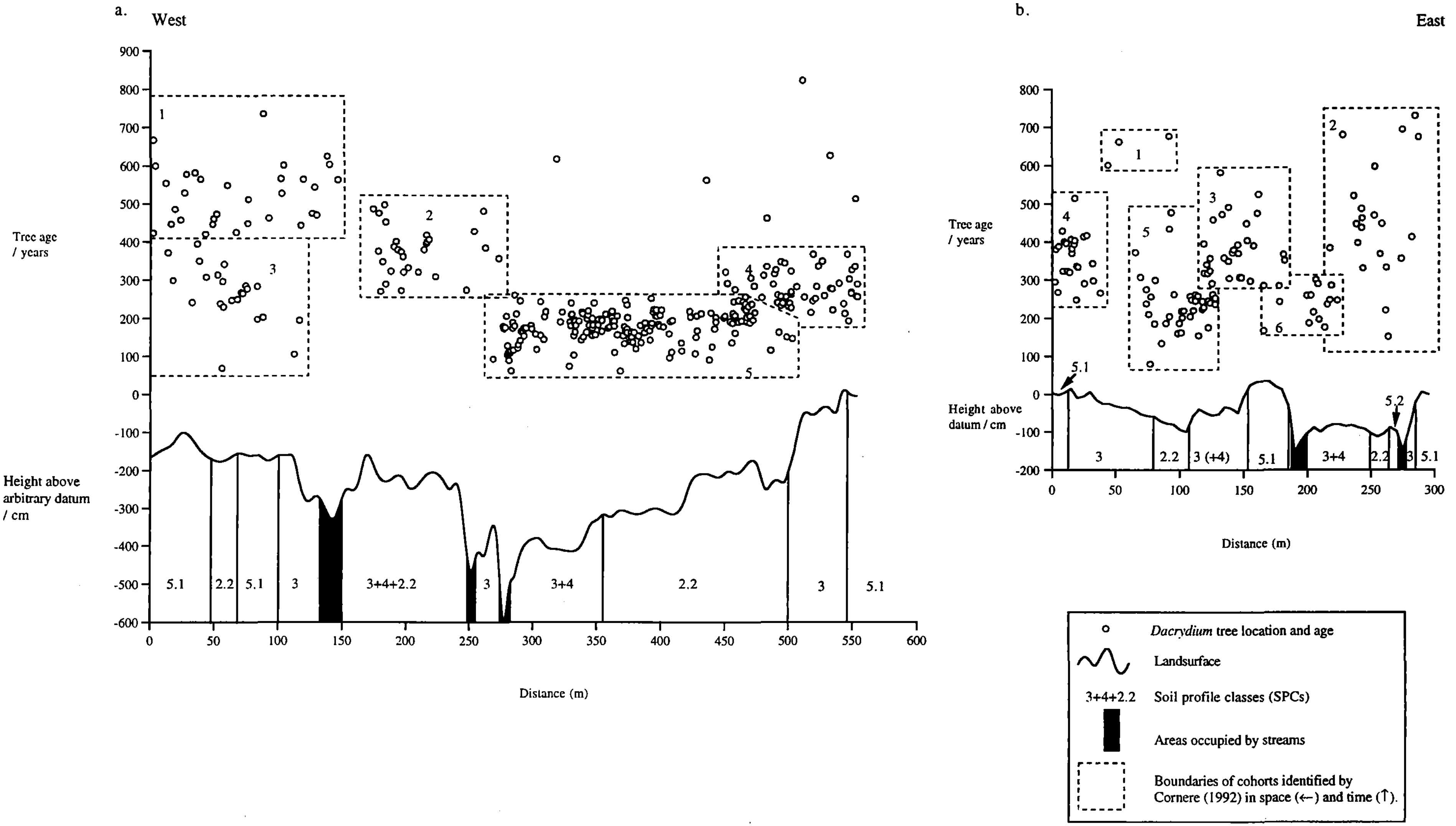

Fig. 2. Dacrydium tree age, cohort boundaries and topography along transects 1 and 2; data from Cornere (1992). For transect locations see Fig. 1. Table 1 groups soil profile classes according to drainage and the New Zealand Soil Classification. Detailed descriptions of soil profile classes are given in Almond (submitted).

T5 where meltwater flowed across T4 and removed the loess, leaving a complex of Perch-gley Podzols formed directly in gravels (SPC 4) or in a thin mantle of alluvial silt (SPC 3), or Humic Organic Soils (SPC 2.2). Cohort 3 comprises trees that established in openings beneath an older surviving cohort 1 on a portion of T4 where the soils are predominantly Silt-mantled Perch-Gley Podzols formed in loess (SPC 5.1). Two regions of these soils are separated by ca. 10 m of Humic Organic Soils (SPC 2.2) occupying a former stream channel that cuts across the transect, and this extremely poorly-drained area contains a dense patch of young cohort 3 trees. Cohort 5 and the older cohort 4 overlap but their boundary coincides with a terrace riser separating T5 and T4, where the extremely poorly-drained Humic Organic Soils give way to better drained Silt-mantled Perch-gley Podzols (SPC 3) on the riser and deep Silt-mantled Perch-gley Podzols in loess (SPC 5.1) on the terrace.

4.2.2. Transect 2

As for transect 1, transect 2 crosses T4 and T5 but has less relief and overall resembles the better drained portions of transect 1. Soils through the regions on T4 (0-10 m, 150-180 m, and 285-300 m) are Silt-mantled Perch-gley Podzols formed in loess (SPC 5.1). Between 10 and 150 m, T5 comprises 2 degradation terraces either side of an abandoned stream channel. The abandoned channel has filled with Humic Organic Soils (SPC 2.2) while the degradational terraces have Perch-gley Podzols formed in silty alluvium over gravels (SPC 3) or directly in gravels (SPC 4). The region of T5 between 180 and 285 m includes 2 small streams at the boundaries and between them the soil pattern is similar to the first T5 region. Soils are mostly SPCs 3 and 4 except in a narrow abandoned stream channel (250-260 m) and an adjacent levee where soils are Organic Soils and deep Silt-mantled Perch-gley Podzols (SPC 5.2) respectively.

The cohorts recognised by Cornere (1992) in transect 2 are not as discrete as those in transect 1, often showing considerable spatial and temporal overlap (Fig 2b). It was therefore difficult to recognise distinct pulses of tree establishment that would indicate localised disturbance events. Furthermore, there is a much weaker correspondence of cohort boundaries with soil/landform regions. Nevertheless, the youngest groups of trees within cohorts are found in the small regions of extremely poorly-drained Organic Soils (at ca. 100 m and 250 m).

5. Discussion

Both our broad scale study and in particular the detailed work of Cornere (1992) support the widely accepted hypothesis (Norton et al., 1988) that *Dacrydium*-dominated lowland forest in Westland is made up of a patchwork of cohorts of trees with restricted age-ranges and consequently restricted size-ranges (see also (Rogers, 1995; Six Dijkstra et al., 1985). These cohorts presumably originated as pulses of tree recruitment following episodes of canopy disturbance. The significant finding of this study is that landforms with a high proportion of poorly-drained soils, particularly extremely poorly-drained Organic Soils, are dominated by cohorts of small diameter trees while landforms with better-drained soils are dominated by cohorts of larger diameter trees. If larger diameter trees are generally older than smaller diameter trees then this finding suggests that episodes of canopy disturbance and cohort initiation have occurred more recently and more often on those poorly-drained landforms. This explanation is supported by the fine-scale study showing that, where there are marked soil drainage differences over 10s to 100s of meters, younger cohorts of trees are associated with poorly-drained areas. These results suggest that in lowland *Dacrydium* dominated forest the patterns of tree mortality and subsequent cohort recruitment are predictable from the soil pattern; the forest is turning over more frequently in poorer drained areas.

Nevertheless, there are two alternative explanations for the observed variation in mean tree diameter among landforms in the broad-scale study. First, variation in mean tree diameter could be due to differences among landforms in diameter growth rate rather than differences in tree age. To test this possibility, we collated all the available radial growth data for *Dacrydium* growing in Saltwater forest and compared mean radial growth rates among trees growing on different landforms (Table 3). There were no significant differences in mean radial growth rate among landforms suggesting that, on average, larger diameter trees on better-drained landforms were indeed older.

Table 3. Mean radial growth rate of *Dacrydium cupressinum* on different landforms in Saltwater Forest.

Landform	*n*	Mean radial increment (mm/yr) ± 1 SD	Source
T4 and T5	449	0.77 ± 0.35	Cornere (1992)
T4	191	0.7 ± 0.3	Rogers (1995)
T5	90	0.6 ± 0.3	Rogers (1995)
T6	94	0.7 ± 0.4	Rogers (1995)
RT	110	0.72 ± 0.24	Stewart et al. (submitted)

The second possibility is that patterns in our data could be an artifact of logging history. For example, many stands of large trees on the terraces may have been logged, leaving only stands of small trees for inclusion in our analysis, while the moraines may have escaped extensive logging and consequently include more larger trees with a greater mean diameter. Two lines of evidence counter this hypothesis: (1) the landforms with the greatest preponderance of small trees (T4 and ablation moraines) are among the least affected by logging (Table 4), and (2) the dominant mode of logging (selection logging by radial hauler) lacked the flexibility to selectively remove large trees. Extraction involved removing a portion of the standing volume in a circular area of ca. 400 m radius centred on a landing. The practicalities of this type of logging demanded that the prescribed harvest (25-30% of standing volume) was taken uniformly over the whole area accessible. This, together with the fact that the scale of the logging was much larger than the size of individual cohorts, has meant that selective extraction was quite indiscriminate; cohorts of small trees were affected as much as cohorts of large trees.

Table 4. Area (ha) and percentage of total area of each landform affected by clearcut or selective logging for *Dacrydium cupressinum.*

		Terrace			Moraine	
	4	5	6	Abl	RT	Lat
Area unlogged	179	838	487	217	307	455
Area logged by: clearfelling including logging for *Lagarostrobos*	0	33	163	166	0	46
selection logging by radial hauler	49	515	182	32	247	50
% logged for *Dacrydium*	24	45	41	13	49	10

Our results show a similar pattern at both the fine and broad-scales; groups of young or small diameter trees are consistently associated with poorly-drained soils. We suggest that the patterns observed at the broad scale are essentially an integration of the soil/vegetation patterns evident at the fine scale. The mean diameter and density of trees on a given landform is influenced by the average frequency at which cohort-initiating disturbances affect that landform (see Clark, 1991). In Saltwater Forest this appears to be largely controlled by the distribution and abundance of soils of different drainage classes. Areas of very poorly-drained soil are more susceptible to disturbances that cause canopy mortality and initiate new cohorts, and so poorly-drained areas carry younger cohorts of trees (Fig 2). Compared to landforms dominated by well-drained soils, landforms with a high proportion of poorly-drained soils consequently have more young trees and a smaller mean diameter.

The fine scale study additionally offers some insight into the spatial scale at which the soil/vegetation patterns may break down. Along transect 2 patches of poorly-drained soil are much smaller than along transect 1 and, although young trees occurred in the small patches of Organic Soils, it was not possible to unequivocally identify cohorts of trees associated with different drainage classes. It appears that the very poorly-drained patches of soil on transect 2 are not large enough to create discrete, disturbance-prone groups of trees (Curtis, 1943;

Everham and Brokaw, 1996). Obviously, where soil patch size is of the order of the dimension of tree root systems, trees are integrating soil conditions across a range of soil types. At this scale of soil patchiness canopy disturbance and cohort initiation may show little relationship to soil pattern. Transect 1 and 2 together suggest that the minimum linear dimension of a soil patch that may predispose groups of trees to synchronous mortality and recruitment is somewhere between 10 and 200 m.

Wind is hypothesised to be a major disturbance agent in Westland's *Dacrydium* dominated forest (James and Franklin, 1977; Six Dijkstra et al., 1985). Windfirmness in trees decreases with declining soil drainage because of lower soil cohesion and interaction between soil and roots, increased lubrication at shear planes at the base of root systems, and reduced rootable volume (Behre, 1921; Day, 1950; Everham and Brokaw, 1996; Faulkner and Malcolm, 1972; Gratkowski, 1956). In addition to the mechanical aspects of poorly-drained soils, trees on wet soils are also more prone to fungal attack of roots and butts which can be a secondary cause of windfall (Schaetzl et al., 1989) or even a primary cause of tree death. Fungal attack is more chronic on poorly-drained soils because firstly, trees are more stressed and secondly, fungi can enter root tissue through wounds caused by non-critical damage when trees sway in strong winds (Hubert, 1918). This may be a cause of the standing dieback noted particularly in areas of very wet soils (James, 1987). All of factors above provide plausible mechanisms for a greater frequency of wind-induced mortality on poorly-drained soils.

From our broad-scale study we have inferred a relationship between soil drainage and disturbance frequency although we are not certain of the functional relationship between soil profile class and susceptibility to windfall. We suspect that tree windfirmness varies along the soil drainage gradient in a step-wise rather than a continuous manner. There are marked changes in soil strength from Organic Soils to mineral soils, and marked changes in root morphology from poorly-drained mineral soils to well-drained mineral soils. Root systems in the mineral soils that characterise the terraces (SPCs 3, 4 and 5.1) have most structural roots on the surface but with a secondary maximum at about 60 cm depth in contact with underlying gravels (Almond, submitted). Root systems in the dominant mineral soils within the terminal and ablation moraines have all structural roots in the surface horizons (Almond, submitted). Differences in windfirmness of trees between these landforms may reflect the differences in root morphology but are more likely to be influenced by soil strength determined by soil moisture. In the well-drained soils dominating the steep lateral moraines, not only are soils drier and stronger but extensive lateral root systems are augmented with large diameter sinkers (P. Almond unpublished information). Transect 1 indicates that forest age-structure on terraces, ablation moraines and terminal moraines could be strongly driven by the abundance and spatial continuity of the weakest and wettest (Organic) soils.

A widely noted feature of Westland podocarp forests is an abundance of mature trees and a lack of young, small-diameter trees (Chavasse, 1954; Cockayne, 1928; Holloway, 1954; Hutchinson, 1932; Poole, 1937). This pattern was evident in our data. The diameter distribution of all *Dacrydium* in the broad-scale study peaked in the 40-49.9 cm class, while the age-distribution of *Dacrydium* in the fine-scale study showed few stems less than 100 years old. The paucity of young and small diameter *Dacrydium* in Saltwater Forest suggests that cohort initiating disturbances were more common in the past than in recent times. Wells et al. (submitted) found a similar pattern, and reached the same conclusion, by collating all of the published data on podocarp age structures from throughout Westland. One explanation for the lack of young cohorts is that rather than wind, earthquakes have been a major disturbance agent in lowland *Dacrydium* forests and there have been no major earthquakes in recent times (Bull, 1996a; Bull, 1996b; Stewart et al., submitted; Wells et al., submitted). Bounding the lowland piedmont of Westland in the south-east is the Cook segment of the Alpine Fault which for about 350 km is a single right-lateral, oblique slip fault which takes up part of the transpressional motion between the Pacific and Australian plates. In historical times it has been aseismic (Adams, 1980) but there is geomorphic and other evidence (Adams, 1980; Berryman et al., 1992; Cooper and Norris, 1990; Hull and Berryman, 1986; Smith and Berryman, 1986) for large (>M 7) earthquakes with up to 9 m horizontal displacement during single events. Indeed, Bull (1996b) cited Cornere's (1992) data to support his inference, from lichen dating of rockfalls, of a large magnitude earthquake on the Cook segment in 1748 A.D. (± 10 yrs). This date is very close to the ages of *Dacrydium* in Cornere's extensive cohort 5 of transect 1, although the ages reported by Cornere (1992) are from increment cores extracted at greater than 1 m above the ground and so are underestimates of true age. Bull (1996a) argued that seismic shaking was in fact a more likely mechanism of disturbance in *Dacrydium* forests given the lack of evidence for disturbance events since about 1830. For the same reasons as for windthrow, we presume that earthquake disturbance of trees would be more severe on poorly-drained soils, particularly the saturated Organic Soils that are prone to flowage (Gallart et al., 1994).

6. Management Implications

Six Dijkstra et al. (1985) concluded that much of the spatial variation in *Dacrydium* diameter and density across Saltwater Forest was a consequence of different portions of the forest being at different stages along a sequence of cohort development. They envisaged that cohorts went through five identifiable stages (early competition phase, competition phase, homeostatic phase, late homeostatic phase, senility/regeneration phase), following a sequence from dense cohorts of small diameter trees at an early stage of development following disturbance through to widely spaced trees of large diameter, as trees grew and thinned out. Six Dijkstra et al. (1985) advocated a forest harvesting plan that aimed to mimic this pattern of cohort development. They argued that harvesting should concentrate on clearfelling cohorts in the later stages of development (homeostatic growth phase, late homeostatic phase or with special care for regeneration, the senility/regeneration phase) to simulate the natural collapse or disturbance of mature stands. Clearfelled cohorts (coupes) would then naturally regenerate and begin the process of cohort development again, while cohorts presently in the early stages of development would progress through the sequence to reach a stage suitable for harvesting.

We support Six Dijkstra et al.'s (1985) contention that dense cohorts of small diameter trees are generally younger than groups of widely spaced, large trees. However, we argue that the forest is not a patchwork of cohorts at different stages along the same development sequence. Cohorts on the most poorly-drained, weak soils may never attain the size-structure or density of cohorts on better drained soils because natural disturbance affects them more often and may continually reset the development process. Thus, the present set of cohorts in the early stages of development may not provide the required future supply of cohorts in the later stages of development that would be necessary for a long-term sustainable harvest under the Six Dijkstra et al. (1985) model.

Furthermore, on the better drained soils of the rolling terminal moraines and particularly the steep lateral moraines, soil conditions appear better suited to hardwood species so that the *Dacrydium* seedlings face intense competition for light and space during the colonisation of disturbance gaps, and may naturally occur at low density even in the early stages of cohort development. In contrast, the high density of *Dacrydium* on the terraces may be in part be due to reduced competition from hardwood species in the very wet conditions, in addition to the young age of the cohorts. Thus, the broad differences in forest structure among landform units are most likely a function of the frequency at which disturbance openings of different size form, along with the environmental conditions found within openings, all factors that are influenced by local soil properties.

The current management of *Dacrydium* in Saltwater Forest has moved away from the coupe felling system advocated by Six Dijkstra et al. (1985), largely because the ground based extraction methods that were used in coupe felling caused extensive site damage. Current harvesting now uses helicopter extraction and is conservative and reactive in that only trees that are likely to die within a management cycle (15 years) in defined compartments are removed. The forest is divided into three broad forest types: hill forest, equivalent to forest on steep lateral moraines; upland forest, equivalent to forest on rolling terminal moraines; and terrace forest. The distribution of the cut among size classes is different for each forest type although more than 80% of the volume is taken from diameter size classes ≥ 50-60 cm for all three forest types. The distribution of harvest among size classes is calculated from assumptions of the rate of transfer of volume from smaller to larger diameter classes, and mortality rate among diameter classes. If, as our results suggest, many of the younger trees on poorly-drained soils are disturbed before they reach maturity, then calculating harvestable volume based on the growth of existing trees may overestimate the true volume that will be available in the future. We suggest that a first step to calculating a sustainable volume harvest from *Dacrydium* forest is to stratify the forest by mean natural turnover frequency, a stratification that could be broadly based on the landform units recognised in this study. In this way standing volume and volume increment can be better estimated in more uniform areas of forest that may have a similar risk of mortality through natural disturbance. Currently, selection of trees for harvest takes place in the field with regard to stand structure and tree health. If management is to take account of the likelihood of significant and sudden changes in standing volume, tree selection would need to be based also on underlying soils. This too could only be done in the field because the complexity of the soil pattern precludes precise mapping of soils at practical map scales. Simple soil-landform models could be developed for moraines (see Almond, submitted) to provide the degree of prediction required, but on terraces where there is almost no relief, foresters would need to examine the soil directly.

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Chapter 7

Synthesis and Summary

Saltwater Forest is a rimu (*Dacrydium cupressinum*)-dominated rainforest on the lowland piedmont of the west coast of South Island, New Zealand. Nearly 4000 ha of the total 9000 ha is gazetted for the sustainable production of timber from indigenous trees. The production area is currently being managed by Timberlands West Coast Ltd who use mostly helicopter extraction for log harvesting. There have been a number of vegetation studies in Saltwater Forest, some of which incorporate soil analysis. However, to date there has been no study of Saltwater Forest or any other lowland podocarp forest in south Westland that integrates forest ecology with a rigorous analysis of soils at a landscape scale. The only published soil map of Saltwater Forest is part of the General Soil Survey of South Island at a scale of 1:253 440 (New Zealand Soil Bureau, 1968). This study has aimed to fill that gap and provide a framework in which existing studies can be placed or future studies can be planned.

The landforms of Saltwater Forest are principally glacial in origin and date to the late Pleistocene. They include moraines (lateral, terminal and ablation) and outwash aggradation and degradation terraces. The existing geological map at 1:250 000 groups the materials comprising the glacial landforms into two chronostratigraphic units: the Moana formation (14-23 ka) and Okarito formation (> 23 ka). The first objective of this study (Objective 1, p. 8) was to establish a more refined chronological stratification of the landscape because time is an important influence on soil properties and patterns. A more refined relative age chronology was constructed using stratigraphic principles and by an analysis of the geometrical relationship between landforms with regard to the direction of ice flow. There are six age groupings of landforms corresponding to separate glacial advances that produced moraines (M) or terraces (T). Arranged from oldest to youngest they are (Fig. 2, p. 19):

1. M1 (terminal moraine)
2. M2 (terminal moraine)
3. M3 (terminal moraine)
4. T4 (aggradation terrace)
5. M5, T5 (lateral, terminal and ablation moraines and an aggradation terrace)
6. M6, T6 (lateral moraines and degradational terraces)

The important landforms in the production area are T4, M5, T5, M6 and T6. Loess mantles the older late Pleistocene landforms (T4 and older) and presents an opportunity to estimate absolute ages of land surfaces. It is often assumed that loess accumulation rate is greatest during cold climate periods and least during warm climate periods. Under this assumption loess sheets and the buried soils that bracket them, can be correlated to dated climatic events. A land surface draped with loess must be older than the oldest loess sheet on it and land surfaces with no loess (erosion notwithstanding) must be younger than the youngest loess sheet. A maximum of five loess sheets are found on the oldest moraine (M1) and loess is thinner and loess sheets fewer on younger landforms. Combined evidence from (1) the degree of alteration of buried soils, (2) plant microfossil spectra, (3) the position of microscopic occurrences of Aokautere Ash (22.6 ka) and (4) radiocarbon dating, points to the correlation of loess sheets given in Table 1. The loess in Saltwater Forest and presumably neighbouring areas is unsuitable for thermoluminescence dating because extreme weathering has removed feldspars which have the best thermoluminescence properties for dating. The ages of landforms in Saltwater Forest and their correlation to the Westland glacial stratigraphy and oxygen isotope stages (Table 1) are based on the loess stratigraphy and other stratigraphic evidence.

Table 1. Chronology of glacial landforms and loess sheets in Saltwater Forest with correlation to the Westland glacial stratigraphy and oxygen isotope stages.

Time (ka)	Landform Saltwater Forest	Loess sheet Saltwater Forest	Westland Glacial Stratigraphy (glacial advance; Suggate, 1990)	Oxygen Isotope Stage
0 - 14				1
14 - 16.5	M6, T6		K3	2
		L1a		
18 - 22	M5, T5		K22	
23				
		L1b		
37				3
	T4			
		L2		
50				
		L3	K21	4
		L4		
	M3			
80	M2			
				5
120				
		L5		
	M1		K1	6

Soils on chronological arrays of landforms such as those in Saltwater Forest have been used elsewhere in Westland to study rates and processes of soil formation. Some of these studies included soils on middle Otiran or older landforms which were mantled with loess. Including these soils as part of the chronosequence under investigation breaches the assumptions of a chronosequence study in that not all soil forming factors, other than time, are effectively constant. Soils on landforms mantled with loess form in a different parent material to that of the soils comprising the younger parts of sequences. The younger soils in the chronosequences studied have been formed in coarse-textured till or alluvium. More seriously, with slow loess accretion and rapid soil formation, typical of Westland, pedogenesis in soils formed in loess has taken place contemporaneously with loess accretion. Soils formed in this way are called upbuilding soils. The slow accumulation of loess during the late Pleistocene in the study area has provided an opportunity to study the influence upbuilding pedogenesis has on soil development (Objective 2, p. 9). Upbuilding soils have unique properties because each increment of the subsoil has initially experienced processes characteristic of surface soil environments. Once an increment is buried, acquired properties are modified or the properties themselves may influence ongoing soil development in the subsoil environment.

In the soil environment of the study area, loess is rapidly weathered as it accumulates on acid surface soil horizons. Organic matter is added and Fe and Al are redistributed from upper to lower soil increments. Resulting soil profiles have different properties to those for which soil formation occurred in a topdown sense into an existing parent material. The most striking difference is the depth of strong alteration. Addition of loess over time is also an input of fresh material that can offset soil fertility decline. Inputs of phosphorus in loess have been sufficient to maintain P status in soils on >35 ka landforms at levels similar to soils on <22 ka landforms. Plant communities in Westland pass along a successional sequence from shrubland to tall forest and eventually heathland as soil fertility first improves then declines. The later stages are driven by deterioration in soil physical and chemical properties. The vegetation pattern on older land surfaces in Westland should be analysed with regard to patterns of loess deposition.

Upbuilding soil formation is a separate pedogenic pathway to that of topdown soil formation. Therefore, soil chronosequence studies that have included soils formed by topdown pedogenesis and upbuilding pedogenesis compare soils formed by two distinct pathways. The rates and types of processes inferred from these studies should be reviewed.

The chronostratigraphic framework established in the first part of the study (Objective 1) combined with an analysis of landform morphology and genesis provided the basis for the study of the soil pattern in the production area. The third objective (p. 9) was to identify the sources and spatial scale of soil variability so that the most cost-effective scale of soil mapping in relation to the intensity of forest management could be determined. Soils in Saltwater Forest are uniformly strongly leached and weathered owing to the high rainfall (2500 - 5000 mm) and the age of the landforms. Drainage status varies from extremely poorly drained to well drained depending on landform and topographic position. According to the New Zealand Soil Classification (Hewitt, 1992) the important soil groups are Humic Organic Soils, Perch-gley Podzols, Acid Gley Soils, Allophanic Brown Soils, and Orthic or Pan Podzols. Soil profile classes within groups have been defined in this study to embrace subtle taxonomic variation related to parent material or substrate, and soil age. In general, spatial variability of soils at any scale of observation is the sum of all soil variability contributed from shorter range (larger scales). In the production area of Saltwater Forest a large proportion of total soil variability arises at very short range. This is a result of microtopographic and individual tree effects on local hydrology and leaching regimes. Soil mapping units displaying areas dominated by a single soil profile class or soil group cannot be drawn at any practical map scale. Moreover, the complexity of soil pattern cannot be routinely accounted for at the current intensity of forest management. At present the production area is divided into 15 compartments averaging 260 ha. Therefore the most practical soil delineations correspond to the boundaries of morpho-genetic landforms of different age i.e. T4, M6 and M5 lateral moraines, M5 ablation moraines, M5 terminal moraines, T5, and T6. These boundaries delineate soil complexes at a scale that can be accounted for in forest planning and operations. Furthermore, they have significance for understanding forest structure and dynamics (see below). The complexes differ in the abundance of soil profile classes in accord with gross changes in landform morphology and age that are responsible for changes in hydrology. Some predictability of soil pattern is possible at a landform component scale (10 - 100 m) by applying rules developed from soil landform models, but only on moraine landforms where there is some relief, and only in terms of probabilities of finding a given soil profile class.

The final objective (Objective 4, p. 9) was to determine the relation between soils and forest structure and dynamics. Variation in forest size and age-structure were analysed at a landform scale and at much shorter range (10-100s of metres) in relation to the soil pattern previously established (Objective 3). Although rimu is the dominant tree species across all landforms, forest structure varies between rolling to steep moraines and flat to gently sloping terraces. On moraines, rimu typically are large and widely spaced, rising above a canopy of hardwoods. In contrast, rimu on terraces are smaller and grow in denser stands with a sparse hardwood subcanopy. Within this broad division of forest types it is obvious that the forest is made up of a patchwork of stands with different diameter and age-class distribution. The patchwork is made up of essentially even-age stands at different stages of development. Stand structure is a function of the time since cohort recruitment and site conditions. Synchronous recruitment is initiated in canopy gaps created by disturbance of a previous generation of trees. Wind is thought to be an important agent of disturbance although large earthquakes on the nearby Alpine Fault may also have been responsible. The frequency of forest disturbance varies according to soil conditions. Forest on the wettest, most poorly-drained soils experiences a higher frequency of turnover than forests on better drained soils. Better drained soils also enable hardwoods to compete more effectively with rimu and other canopy species. Differences in turnover frequency and types of regeneration niches between different soils means that stands making up the forest patchwork have different life histories.

Parts of forested landscapes with a high frequency of turnover have cohorts of small, young trees whereas areas of low turnover-frequency have older, larger trees. This generalisation holds true for the production area of Saltwater Forest. Mean diameter and density vary among landforms in accord with the abundance of different soil profile classes. Terrace T4 which has the highest proportion of deep, extremely poorly-drained Organic Soils has on average the smallest diameter trees growing at the greatest density. From T4 to T6 the abundance of Organic Soils declines and the abundance of better-drained mineral soils increases. This is paralleled by an increase in average rimu diameter and decrease in density. A similar trend occurs from ablation moraines to lateral moraines. Ablation moraines have a high proportion of extremely poorly-drained Organic Soils and a low mean rimu diameter and high density. The lateral moraines which have the highest proportion of well drained Allophanic Brown Soils or Orthic Podzols have on average the largest trees growing at the lowest density. The terminal moraines are intermediate in drainage between ablation and lateral moraines and have trees of intermediate average diameter and density.

The systematic variation of forest structure and dynamics among landforms provides a basis for stratifying the production area for management. Forest units defined by landform would permit better estimates of standing volume and volume increment over more uniform areas of forest, with a similar risk of mortality through natural disaster. Current harvesting rate is calculated on assumptions of the rate of transfer of volume from smaller to larger diameter size-classes, and mortality rate among size classes. Stands on the most poorly drained soils appear to turn over at a rate whereby they are unlikely to provide trees in the larger size-classes that are most commonly harvested. If management is to take account of the possibility of sudden changes in standing volume, tree selection should take account of the nature of soils stands are growing on. This can only be done in the field because of the large short-range soil variability.

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