Soil Science

Sustainable Production and Environmental Protection

New Edition

R. G. McLaren and K. C. Cameron

Auckland
OXFORD UNIVERSITY PRESS
Melbourne   Oxford   New York

CONTENTS

PLATES

Colour Plates 1–8 may be found between pages 184 and 185.

FOREWORD

The first edition of Drs McLaren and Cameron's *Soil Science*, published in 1990, achieved immediate and justified success. It received excellent reviews in the local and international soil journals, and demand was such that it was reprinted in 1993 and 1994. It is now a standard text at all the New Zealand universities where soil and earth sciences are taught.

There have, however, been significant developments in soil science since the first edition was published.

Attitudes to the ecological role of soils are changing. While the soil's key function of producing biomass for energy, food, fibre and raw materials remains, increasing attention is being paid to the soil as a medium capable of protecting the environment through its ability to filter, buffer and transform pollutants. We are also increasingly concerned with the impact of soil management on the wider environment, including surface and groundwater and the atmosphere.

A new system of classifying New Zealand soils has been introduced. While an advance on earlier systems, this system, like any new system, will take time to become established: soils literature using earlier systems, from research papers to advisory leaflets, will continue to be used, and practitioners must become familiar with new names and relationships while still depending on the older literature.

Systems for estimating fertiliser requirements for pastoral farming have been further refined in the last few years, with increasing emphasis being placed on the dynamics of soil fertility and the need to include an economic component with advice.

These new concepts and systems require that everyone involved with the soil and its management regularly revise their beliefs and knowledge of this complex part of our environment. Drs McLaren and Cameron have recognised that advances in the ways in which we conceive, study and manage soils will be accepted and applied most rapidly if they are incorporated into basic soil science education. They have used the new classification system in the revised edition of their book and have provided a key for translating from the previous system to the new, which will be of value to all users of soil information. They have also endeavoured to demonstrate the key role of the soil in the wider environment and to show how our treatment of the soil impacts on air and water quality.

By incorporating these and numerous other recent developments in their book, the authors have produced a concise and comprehensive text that will help to ensure that New Zealand students will be equipped to promote the efficient, effective and, above all, sustainable use of one of our most valuable resources.

Lincoln University
Canterbury, New Zealand
January 1996

I.S. Cornforth
Professor of Soil Science

INTRODUCTION TO NEW EDITION

During the past few years, we have become increasingly aware of the need for a new edition of this textbook. The major issues of sustainable production and environmental protection are having a significant influence on the ways in which soil science is developing and being taught. While many soil scientists remain heavily involved in soil fertility and productivity issues, a much greater emphasis is now being placed on the sustainability of land use. In addition, soil science has an increasing involvement in environmental issues, such as the land disposal of wastes and the evolution of 'greenhouse' gases. More specific changes that have occurred since the publication of the first edition of *Soil Science* include the introduction of the new New Zealand Soil Classification (Hewitt, 1992) and changes in the ways in which fertilizer recommendations are made.

This new edition attempts to incorporate these trends and changes, while retaining the general philosophy and layout of the original book. Indeed, many of the basic soil chemistry, soil physics and plant nutrient chapters have been changed very little. The most significant changes occur in Chapter 3 (Classification and characteristics of New Zealand soils) and Chapter 21 (Soils and the quality of the environment), and to a lesser extent in Chapter 1 (The soil profile), Chapter 11 (Inorganic soil colloids) and Chapter 19 (Soil fertility and fertilizer management).

In Chapter 3 we have included a description of the new soil classification system and have used this system elsewhere throughout the book. However, since most of the existing soil resource information available in New Zealand uses the older Genetic Soil Classification, we have also retained a brief description of that system. To assist the reader in 'translating' between systems, correlations between the two systems are also included. Chapter 21 has been expanded to include a discussion of the influence of soils on the atmosphere—for example, on greenhouse gases and stratospheric ozone—and also a more complete description of the influence of land disposal of wastes on soil and water quality.

Many of the other changes to the book have been made as a result of comments from our colleagues both at Lincoln and other tertiary institutions. We hope that this new edition will continue to meet the needs of those students studying soil science in New Zealand.

Lincoln University   *Ronald G. McLaren*
Canterbury, New Zealand   *Keith C. Cameron*
January 1996

THE SOIL PROFILE

variations with depth in colour, texture, structure, and other soil properties, make it possible to recognize a number of distinct horizontal zones, positioned approximately parallel with the soil surface. Some of these zones may be the result of layering within the mineral parent material in which the soil has developed (see Section 1.2.4). Other zones however, known as **soil horizons**, have been produced by the various soil-forming processes described in Chapter 2. Figure 1.1 shows the relationship between the soil continuum and individual soil profiles and horizons. Photographs of a range of different soil profiles and horizons can be seen in Colour Plates 1 to 6.

The study of soils in the field depends on the ability to recognize and describe the morphological features of the soil profile and thereby make comparisons between one soil and another. The individual horizons within a soil

Most people usually regard soil as the top 20 cm or so of 'earth' that a farmer or gardener cultivates before sowing their crops or pastures. However, to a soil scientist, the complete soil includes the topsoil and any subsoil layers extending downward to the mineral rock material (parent material) from which the soil has developed. Soil extends laterally in all directions over the surface of the Earth forming a three-dimensional continuum; the exact nature of the soil and its depth vary considerably across the landscape. In New Zealand, soil depth varies from just a few centimetres at some locations, to depths greater than a metre at others.

Although when mapping soils an understanding of lateral variation is extremely important (see Chapter 4), the study of soils in three dimensions raises considerable problems. In the field, soils are usually studied at individual locations by digging a pit down to the parent material and examining one of the vertical faces of the pit. This vertical, two-dimensional section of the soil, extending from the surface down to the parent material is known as the **soil profile** (Figure 1.1). In most soil profiles,
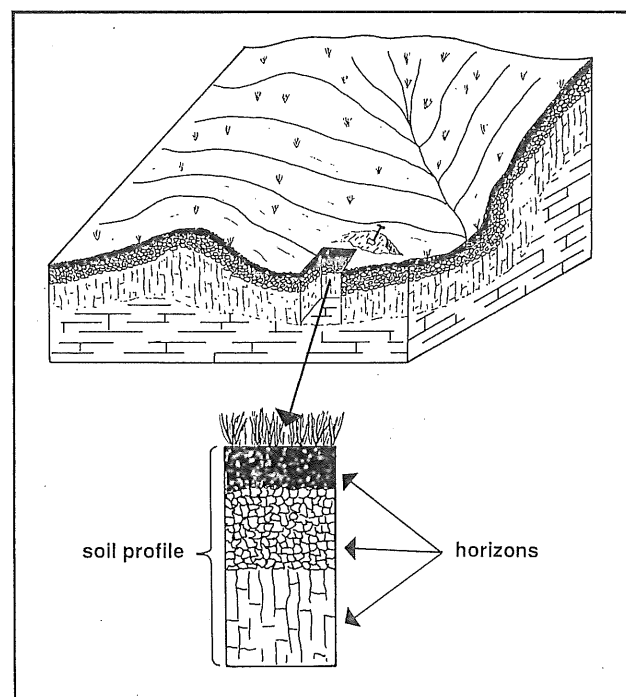

Figure 1.1: The soil continuum and the soil profile.

profile, of which many different types are recognized, have resulted from particular combinations of soil forming processes as will be described in Chapter 2. Examination of soil profiles for the morphological properties described below can help to identify which particular types of horizon are present. Having recognized their presence, a complete **soil profile description** can be assembled, involving a description of the morphological features of each individual horizon.

1.1 SOIL MORPHOLOGICAL PROPERTIES

The following sections describe the methods used to record the morphological features of the soil profile.

1.1.1 Soil colour

Soil colours are recorded by comparison with the standard Munsell Soil Colour Chart. The arrangement of colours on the chart is by hue, value and chroma.
Hue is the spectral colour; it is related to the dominant wavelength of light. Its notation is given at the top of each page, for example 10YR, 5Y, etc. **Value** refers to the relative lightness of colour and is the vertical component of the colour chart. **Chroma** is the relative purity of the colour and increases with decreasing greyness. This is the horizontal component of the colour chart. The Munsell notation for colour consists of a combination of the notations for hue, value and chroma, in that order, for example 10YR 5/6 (Hue10YR, value5/chroma6). In recording soil colour, both the colour name and the Munsell notation is given. The use of colour names other than those on the colour chart is avoided.

It should be noted that the colours of most soils vary with their moisture content and the basic colour(s) of a soil horizon is normally recorded from the surface of a freshly broken soil aggregate in a field moist state. Some horizons show **mottling**, areas of colour different from the overall background colour of the horizon. The most common forms of mottles are the rusty coloured ones associated with poorly drained conditions (see Section 2.4.7 and Colour Plate 5c), however other types of mottles do occur. Since mottling is significant in relation to soil formation and soil moisture regime, it should be carefully described. The abundance, colours, size and contrast of mottles and the sharpness of mottle boundaries are normally recorded (Table 1.1).

1.1.2 Soil texture

Soil texture is defined as the particle size distribution of the solid inorganic constituents of the soil. It normally refers to the solid particles less than 2 mm in diameter,

known as the fine earth fraction. Larger inorganic particles, such as gravel and boulders, and humified organic matter are recognized by qualifying textural terms. Particle size classes for both the fine earth fraction and larger particles are given in Table 1.2.

A quantitative assessment of soil texture can be made by a particle-size analysis determination conducted in the laboratory (see Chapter 5). However, in the field, soil texture is determined qualitatively by manipulating a moistened soil sample between the thumb and fingers, and relies on the different properties of sand, silt and clay under this condition. Sand has a gritty feel, silt has a smooth silky feel, and clay is sticky and adheres strongly

Table 1.1: Terms used for the description of mottles.

Abundance (% of described surface occupied by mottles)		Size (mm, measured along greatest dimension)	
very few	<2	extremely fine	<2
few	2–10	very fine	2–6
common	10–25	fine	6–10
many	25–50	medium	10–20
abundant	50–75	coarse	>20
profuse	>75		

Contrast	
faint	indistinct mottles evident only on close examination.
distinct	although not striking, the mottles are readily seen.
prominent	the mottles are conspicuous, and mottling is one of the outstanding features of the horizon.

Sharpness of Mottle Boundaries	
sharp	knife edge boundaries between colours
clear	colour transition less than 2 mm wide.
diffuse	colour transition greater than 2 mm wide.

Table 1.2: Soil particle size classes (diameters, mm).

(a) Classes for fine earth fraction (ISSS system)	
coarse sand	0.2–2
fine sand	0.02*–0.2
silt	0.002–0.02*
clay	<0.002

(b) Classes for large particles (Milne et al. 1991)	
boulders	>200
very coarse gravel	60–200
coarse gravel	20–60
medium gravel	6–20
fine gravel	2–6

*Other systems use 0.05 or 0.06 mm to separate the fine sand and silt fractions.

to the fingers. With practice, an assessment of the relative proportions of these particles can be easily determined for any sample of soil. However the 'field soil texture' determined in this way may also be influenced by the presence of humified organic matter, stable microaggregates, calcium carbonate and by clay type. The characteristic behaviours of six of the more commonly occurring soil textural classes are summarized in Table 1.3.

The names of texture classes are qualified where required by appropriate terms for organic matter and for mineral particles larger than 2 mm diameter. For example, coarse gravelly silt loam, fine gravelly sand, humic clay loam, peaty clay, etc.

1.1.3 Soil consistence

Soil consistence refers to the inherent qualities of soil that are expressed by the way in which the soil material holds together, deforms or ruptures when put under pressure. The consistence of a soil depends on the kind of cohesive and adhesive forces which hold the soil material together and can vary greatly depending on the soil water content at the time when consistence is assessed.

There are several techniques for assessing consistence in the field and they usually involve taking a sample of soil material in the hand (a natural aggregate or aggregates), subjecting it to pressure between the fingers and assessing its strength or recording the way it deforms or fragments. Some tests require the addition of water to the sample to assess its degree of stickiness or plasticity. One way of classifying the consistence of field moist soils, according to their mode of failure under pressure, is shown in Table 1.4.

A more detailed explanation of consistence and its importance, and more quantitative methods of determining consistence will be discussed in Chapter 5.

1.1.4 Soil structure

The term soil structure refers to the shape, size and degree of development of the aggregation, if any, of the primary soil particles into naturally or artificially formed structural units (peds, clods and fragments).

Table 1.4: A scheme for determining the consistence of moist soil samples.

Consistence category	Behaviour of soil material in response to pressure between fingers
loose	non-coherent
very friable	soil material crushes easily under gentle to moderate pressure between thumb and forefinger, and coheres when pressed together
friable	soil material crushes under very gentle pressure but coheres when pressed together
firm	soil material crushes under moderate pressure but resistance is distinctly noticeable
very firm	crushes under strong pressure, barely crushable
extremely firm	crushes only under very strong pressure, cannot be crushed between thumb and forefinger

Peds are natural, relatively permanent aggregates, separated from each other by voids or natural surfaces of weakness. They are formed in the various ways described in Chapter 5. Less permanent aggregates, termed 'fragments' or 'clods', formed at or near the surface by cultivation or frost action are not true peds. Peds are, however, not always easy to distinguish from fragments. Peds are of two main kinds.

(i) Those in A horizons are often spheroidal, some have distinctly porous surfaces, and they are generally even coloured. They arise particularly from the interaction of organisms, roots and mineral particles; some peds originate as faunal excreta, such as earthworm casts.

(ii) Peds in lower soil horizons often have well formed faces, which can be distinguished from artificially broken surfaces by their smoothness or by colours distinct from the interior of the ped. The ped faces may be coated, for example, by organic matter or clay, and some peds have very smooth shiny faces. The surfaces between peds often form paths for root penetration and ped faces often have abundant live or dead root impressions.

Table 1.3: Texture determination of moistened soil.

Feel and sound	Cohesion and plasticity	Field soil texture class
gritty and rasping sound	cannot be moulded into a ball	**sand**
	will almost mould into a ball but disintegrates when pressed flat	**loamy sand**
slight grittiness, faint rasping sound	moulds into a cohesive ball which fissures when pressed flat	**sandy loam**
smooth soapy feel, no grittiness	moulds into a cohesive ball which fissures when pressed flat	**silt loam**
very smooth, slightly sticky to sticky	plastic, moulds into a cohesive ball which deforms without fissuring	**clay loam**
very smooth, sticky to very sticky	very plastic, moulds into a cohesive ball which deforms without fissuring	**clay**

Peds are described according to:

(i)   **Grade of soil structure**—the proportion of the soil mass that is in the form of peds, see Table 1.5.

(ii) **Class of soil structure**—the size range of peds within each structure type, see Table 1.6.

(iii) **Type of soil structure**—the shape and arrangement of peds, see Figure 1.2.

Although, in detail, peds may be irregular in shape, they can be grouped according to their general form; plate-like, prism-like, block-like and spheroidal (Figure 1.2). Plate-like, prism-like and block-like are fitting structures in that the sides of the peds are the casts of adjacent

Table 1.5: Grades of soil structure.

Apedal
No observable aggregation; without a definite orderly arrangement of natural planes of weakness.

single grain	with more than 85% by weight of individual coarse particles ranging in size from sand to very coarse gravel.
massive	without peds, clods or fragments, and having no partings or fissures at spacings of less than 200 mm.

Pedal
Observable natural planes of weakness that define the surface of peds in some part of the soil.

weakly developed	poorly formed, indistinct, weakly coherent peds that are barely observable in places. Much unaggregated material (only 15–25% of soil consisting of peds greater than or equal to 6 mm).
moderately developed	well-formed, moderately durable peds that are evident but not distinct in undisturbed soil (25–75% of soil consisting of peds greater than or equal to 6 mm).
strongly developed	durable peds that are quite evident in undisplaced soil, adhere weakly to one another and separate cleanly when the soil is disturbed. Little or no unaggregated material (more than 75% of soil consisting of peds greater than or equal to 6 mm).

Table 1.6: Soil structure size classes (mm).

Class	Size (mm)*
very fine	<6
fine	6–10
medium	10–20
coarse	20–60
very coarse	60–100
extremely coarse	>100

* Size classes refer to the dimensions of the shortest axes.

peds indicating that when the soil is wet the swollen peds fit closely together (closing up the macropore space in the soil along ped surfaces). Plate-like peds have a predominance of horizontal planes between peds. Prism-like peds have a predominance of vertical planes over horizontal planes between peds. Block-like peds are bounded by equally developed horizontal and vertical planes between peds.

Spheroidal peds are non-fitting; they differ from the others in that the surfaces of individual peds are not the casts of adjacent peds. They commonly consist of spheroids or polyhedrons. Many of these peds are so irregular in form that they are difficult to distinguish from irregular block-like aggregates, and in such cases the distinguishing criterion is whether they are non-fitting or fitting.

Large peds can sometimes be broken down into sets of smaller peds and are referred to as compound peds. The relationship between such sets of peds is usually described, e.g. medium prismatic structure breaking to medium and fine subangular blocks.

1.1.5 Pores, nodules, coatings, etc.

Pores

Although estimation of the total volume of soil pores cannot be reliably determined in the field, by using a hand lens, some estimate of the size and abundance of the larger pores can be made. Sizes of these large pores can be classified as shown in Table 1.7 and the percentage of such pores per unit area of ped interior can be estimated by comparison to diagrams as shown in Figure 1.3.

Nodules

These are spherical, tubular or irregular coherent bodies, with sharp boundaries sufficiently discrete to be separated from the soil matrix. Most nodules are formed from *in situ* pedogenic precipitates of iron, manganese or aluminium oxides or from calcium carbonate (see Chapter 2). Most are cemented and hard, but soft bodies that can be readily separated are also included. Although most nodules are formed by *in situ* accumulations, in some soils they are derived by transport from older soil or rock materials or inherited from the mineral parent material. In the description of nodules the following features are noted; abundance, size, shape, hardness, colour, composition, kind and location in the horizon.

Coatings

These are accumulations of certain soil constituents, for example, clay, organic matter and oxides of iron and aluminium, at the natural interfaces in soils, such as the faces of peds or the soil pores. In describing coatings the following details are recorded: abundance, continuity, distinctness, thickness, kind and location.
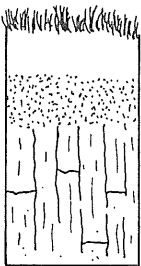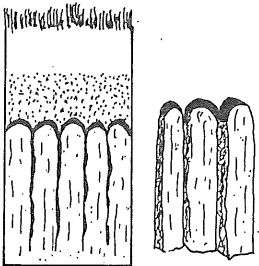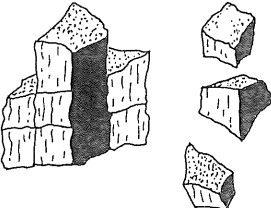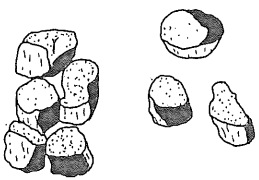
soil structure types		notes
plate-like	platey	may occur in any part of the profile
prism-like	prismatic columnar	The tops of prismatic aggregates are essentially flat, whereas those of columnar aggregates are rounded. Both types of aggregate are found in subsoils of soils of subhumid to arid regions, however, columnar aggregates are less common than prismatic ones.
block-like (fitting)	blocky subangular blocky (nutty)	Blocky aggregates have flattened faces and most vertices are sharply angular - common in clay to clay loam textured subsoils of humid regions. Subangular blocky aggregates have mixed rounded and flattened faces with many rounded vertices - common in topsoils and subsoils of humid regions
spheroidal (non-fitting)	granular crumb	Characteristic of A horizons - commonly found together under grass in humid and sub-humid regions Crumbs are porous peds similar in appearance to crumbs of bread. Granular aggregates are non-porous.

Figure 1.2: Soil structure types.

Other features

Any other noticeable feature in the soil profile is also usually recorded including the presence of earthworms or other fauna.

1.1.6 Roots

The size, abundance and kind of roots present in the horizon are described. Table 1.8 shows one way of classifying root abundance according to their number and size. The relationship of roots to other features in the profile, e.g. faunal channels, ped faces, horizon boundaries, is also noted.

1.1.7 Horizon boundaries

Within the soil profile, soil horizons are distinguished by changes with depth of the above morphological properties. The boundaries between horizons vary in distinctness and in shape. The distinctness of boundaries depends partly on the contrast, as shown by comparison of properties of adjacent horizons, and partly on the thickness of the transition between horizons. Boundary distinctness and shape are described by the terms shown in Table 1.9.

1.2 SOIL HORIZON NOMENCLATURE

Horizons of different kinds are identified by symbols composed of capital letters (master horizon designation) and lower case letters (to indicate subordinate horizon characteristics). The symbols indicate the relationships between horizons within a soil profile as interpreted by the person describing the soil profile.
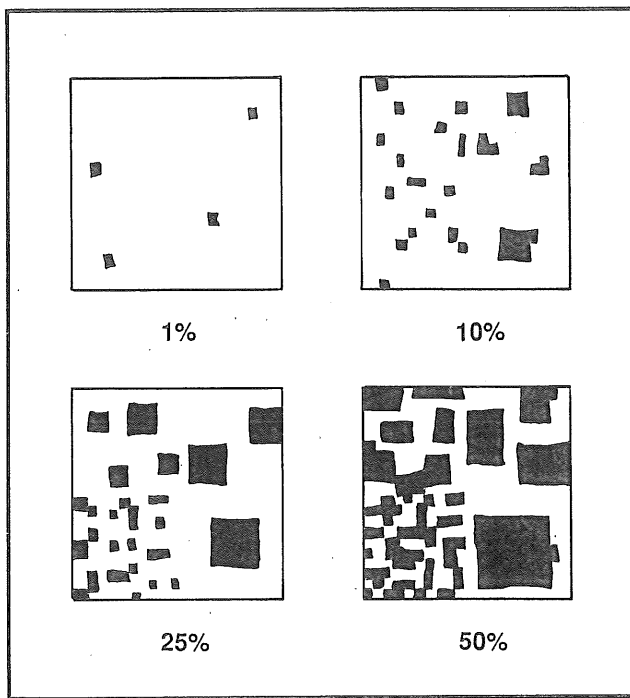
Figure 1.3: Diagrams for estimating macropores, mottles, stones, etc. Each quarter of any one square has the same area of black (adapted from Hodgson, 1976).

Table 1.7: Size classes of large pores, channels or fissures.

Class	Size (diameter, mm)
very fine	<0.5
fine	0.5–2
medium	2–5
coarse	>5

Table 1.8: Abundance classes of roots in relation to their number and size.

Frequency class	number of roots per 100 cm²	
	very fine and fine roots less than 2 mm diameter	medium and coarse roots greater than 2 mm diameter
few	1–10	1 or 2
common	10–25	2–5
many	25–200	>5
abundant	>200	

Table 1.9: Horizon boundary conditions.

Boundary distinctness (thickness of transition zone in mm)		Boundary shape	
sharp	<5	smooth	the boundary is plain with few or no irregularities and usually occurs at the same depth across the profile face.
abrupt	5–20		
distinct	20–50		
indistinct	50–100		
diffuse	>100		
		wavy	the boundary has broad, shallow, relatively regular pockets, wider than they are deep.
		irregular	the boundary has pockets that are deeper than they are wide but not curved back on themselves.
		convolute	the boundary has pockets that are deeper than they are wide and, in parts, curved back on themselves.

The problem of designations for soil horizons is an ever changing one, with different soil survey organizations favouring their own horizon nomenclature. The designations presented here have been found useful by the authors for teaching purposes and are compatible with the *Horizon Notation for New Zealand Soils* published by B. Clayden and A.E. Hewitt (1989). In this text only general horizon descriptions are given, however the reader should be aware that some systems require detailed analytical data in order to accurately categorize horizons.

1.2.1 Master horizons

(a) Organic horizons

These are defined as containing more than 18% organic carbon (30% organic matter).

O horizon—These are organic horizons accumulated under wet conditions. They are saturated with water for at least 30 days in most years, or have been artificially drained. O horizons are subdivided according to the amount of plant fibre present and its degree of decomposition.

L, F and H horizons—These are organic horizons occurring in the upper part of the soil profile, consisting of fresh and/or partly decomposed organic matter accumulated under predominantly aerobic conditions.

L horizon—Fresh litter deposited during the previous annual cycle. It is normally loose and the original plant structures are little altered.

F horizon—Partly decomposed or comminuted litter remaining from earlier years, in which some of the original plant structures are visible to the naked eye.

H horizon—Well decomposed litter, often mixed with mineral matter, in which the original plant structures cannot be seen.

(b) Mineral horizons

These are defined as containing less than 18% organic carbon.

A horizons—Mineral horizons formed at the surface, or below an organic horizon. They are characterized by incorporation of humified organic matter, or properties resulting from disturbance by cultivation. The organic matter has been incorporated by biological activity or by mixing through cultivation rather than by translocation.

E horizons—Subsurface mineral horizons that contain less organic matter and extractable iron and/or silicate clay than the horizon immediately below, as a result of downward or lateral movement. The colour of these horizons is often pale grey to whitish, which is the colour of uncoated sand and silt grains.

B horizons—Subsurface mineral horizons without rock structure, normally underlying A or E horizons and characterized by one or more of the following: (i) illuvial concentration of silicate clay, iron, aluminium or organic matter, alone or in some combination, (ii) coatings of sesquioxides that give the horizon a lower colour value, a higher chroma or redder hue than the horizons above and below, without apparent illuviation of iron, (iii) alteration that has formed silicate clay or liberated oxides or both, and that leads to ped formation, or (iv) evidence of solution and removal of carbonates.

C horizons—Unconsolidated or weakly consolidated mineral horizons that are little affected by soil-forming processes and lack properties of A, E or B horizons. The material may be either like or unlike that from which the overlying horizons were presumably formed. C horizons may have been modified by (i) saturation with water or (ii) accumulation of carbonates or more soluble salts. These horizons are often regarded as the 'parent material' of the soil.

R horizons—Hard or very hard bedrock that is impracticable to dig with a spade. The rock may contain cracks, but these are generally too few and too small for significant root development. However, roots of trees and shrubs may be able to grow into fissures in R horizons.

1.2.2 Transitional horizons

Two main kinds of transitional horizons are recognized:
(i) Horizons transitional between two master horizons, such as gradational horizons in vertical sequence, are indicated by compound horizon designations, e.g. AB, BC.
(ii) Transitional horizons in which discrete parts have recognizable properties of two kinds of master horizon, e.g. where pieces of A horizon are incorporated in a B horizon, or where horizons interfinger. In this case, the transitional horizon is indicated by master horizon designations separated by a solidus (/), e.g. A/B, B/C.

1.2.3 Subordinate horizons

A lower case letter, as listed in Table 1.10, may be added as a suffix to qualify the Master Horizon designation. Suffix letters can be combined to indicate properties which occur concurrently in the same master horizon (e.g. Btg, Cck). Normally no more than two suffixes should be used in combination. In transitional horizons either the suffixes appropriate to each component master horizon are used (e.g. Ah/Bw) or where only one suffix is used it applies to the transition horizon as a whole (e.g. B/Cg). Some suffixes are mutually exclusive (e.g. Ah or Ap, Bw or Bx, Bs or Bk). Suffix letters should only be used if there is sufficient evidence for the implied interpretation.

1.2.4 Buried horizons

Pedogenic horizons that have been buried by deposition of additional material and that have been little modified subsequently are called buried horizons and are denoted by the prefix 'b', e.g. bAh, bBw.

1.2.5 Lithological discontinuities and division of horizons

Where distinct texturally or mineralogically contrasting layers occur within the A, B or C horizons which are not due to processes of soil formation, Arabic numerals are prefixed to the horizon designations concerned. The different layers are numbered from the top downwards, but the numeral 1 is conventionally omitted as unnecessary and numbering begins with 2 (e.g. Ah Bw Bt 2C).

Horizons designated by a single combination of letter symbols can be vertically subdivided by numbering each subdivision consecutively, starting at the top with a suffix number (e.g. Bt1 Bt2 Bt3). The suffix number always follows all of the letter symbols. The number sequence applies to one horizon type only so that the sequence is resumed in the case of a change of horizon type (e.g. Bt1 Bt2 Bx1 Bx2). A sequence is not interrupted, however, by a lithological discontinuity (e.g. Bt1 Bt2 2Bt3) unless this discontinuity also coincides with a change in soil horizon type (e.g. Bt1 Bt2 2Bx1).

1.3 SOIL PROFILE DESCRIPTION

A complete soil profile description involves recording the morphological properties for each horizon recognized within the soil profile. Soil profile descriptions are usually presented in a standard format, an example of which is shown in Table 1.11. The format presents the morphological data for each horizon in the following order: depth (cm); colour (both in words and colour code); texture; consistence; structure (grade, class and type); mottles

Table 1.10: Letter suffixes for subordinate horizons.

c:	localized accumulation, as hard nodules or concretions, of compounds formed by chemical deposition; usually used in conjunction with another suffix to indicate the nature of the deposited material, e.g. Bkc, Csc.
d:	an uncemented E horizon with very high soil strength and bulk density. Sometimes referred to as a 'densipan'.
g:	mottled colour patterns of oxidized and reduced colours (see gleying in Section 2.4.7), with the former predominating.
fm:	sharply defined black to reddish brown, brittle or cemented horizon less than 10 mm thick, enriched in iron and organic compounds (iron pan).
h:	accumulation of humified organic matter in a mineral horizon, e.g. Ah, Bh. Note that, for A horizons, the h suffix is applied only where disturbance and mixing by cultivation is absent.
k:	accumulation of secondary carbonate.
m:	continuously cemented horizon, other than a Bfm horizon, that is strong enough to resist root penetration, e.g. Bkm-cemented by carbonate.
n:	accumulation of sodium.
o:	strongly weathered B horizon (Bo) formed in mixed crystalline iron and aluminium oxides and kaolin minerals.
p:	surface horizon in which incorporation of organic matter has involved mixing as a result of cultivation or of increased biological activity associated with topdressing or manuring.
q:	accumulation of secondary silica.
r:	intensely gleyed horizon (see Section 2.4.7) with predominantly greyish colours and usually few mottles. Associated with prolonged saturation with water.
s:	brightly coloured ochreous B horizon containing iron and aluminium minerals (sesquioxides) present predominantly as a result of downward translocation from horizons above. The iron and aluminium is mainly present as minerals with short-range order, e.g. allophane and ferrihydrite (see Chapter 11).
t:	horizon containing translocated clay, normally shown by the presence of clay coatings on the surface of soil aggregates and in soil pores. Generally with at least 5% more clay than the overlying horizon. A Bt horizon is sometimes referred to as an 'argillic' horizon.
w:	B horizon that shows evidence of alteration (by weathering or structural reorganization *in situ*) under well-aerated conditions and does not qualify as Bh, Bs or Bt. Sometimes referred to as the 'cambic' horizon.
x:	horizon with fragipan properties (see yellow-grey earths/pallic soils, Chapter 3). Fragipans are compact, massive or near massive horizons, rich in silt and generally low in clay. When dry, they are brittle and have the appearance of being strongly cemented, but the cementation, if present, disappears on moistening and the soil material slakes.
(x):	a brittle horizon that does not meet all the requirements of a fragipan, e.g. Bw(x)
y:	accumulation of gypsum.
z:	accumulation of salts more soluble than gypsum, e.g. halides.

Table 1.11: Soil profile description of a Timaru silt loam soil.

Ah 0–20 cm	Very dark greyish brown (2.5Y3/2) silt loam; friable; moderately developed medium and fine crumb with fine nut structure; many roots; gradual boundary.
Bw 20–38 cm	Light olive-brown (2.5Y5/4) silt loam; firm; weakly to moderately developed medium nut and blocky structure; common macropores; few worm casts and channels; few roots; gradual boundary.
Bg 38–52 cm	Light olive brown (2.5Y5/4) silt loam; firm; weakly to moderately developed medium blocky structure; few fine faint light brownish grey (2.5Y6/2) and strong brown (7.5YR5/6) mottles; few fine roots; few macropores; gradual boundary.
Bx 52–101 cm	Light olive brown (2.5Y5/4) silt loam; very firm; weakly developed; coarse blocky with weak, coarse prismatic structure; few distinct grey (5Y6/1) and strong brown (7.5YR5/6) mottles; few thin coatings line some pore spaces and some ped surfaces; few macropores; diffuse boundary.
C 101+ cm	Light olive brown (2.5Y5/4) silt loam; firm; weakly developed coarse prismatic structure; few macropores.

and nodules, etc. (if present); roots; pores and finally the boundary conditions between the next horizon.

The reader should be aware, however, that the profile description shown in Table 1.11 is a relatively simple one. Other formats may require much more detail and involve more comprehensive description systems than the ones described in this chapter. In some systems, quantitative methods are used for recording horizon attributes, enabling storage in computer data bases.

1.4 FURTHER READING

Clarke, G.R. and Beckett, P.H.T. 1971, *The Study of Soil in The Field*, Oxford University Press, Oxford.

Clayden, B. and Hewitt, A.E. 1989, *Horizon Notation for New Zealand Soils*, Division of Land and Soil Sciences Scientific Report 1, DSIR, Lower Hutt.

Hodgson, J.M. 1975, *Soil Survey Field Handbook*, Soil Survey of England and Wales Technical Monograph No. 5, Rothamsted Experimental Station, Harpenden.

Hodgson, J.M. 1978, *Soil Sampling and Soil Description*, Oxford University Press, Oxford.

McRae, S.G. 1988, *Practical Pedology: Studying Soils in the Field*, Ellis Horwood Ltd, Chichester, UK.

Milne, J.D.G., Clayden, B., Singleton, P.L. and Wilson, A.D. 1991, *Soil Description Handbook*, DSIR Land Resources, Lower Hutt.

SOIL FORMATION

Soils are formed ultimately from the rock materials that make up the Earth's crust. Over considerable periods of time, as a result of various erosion and weathering processes, the solid bedrock exposed at the surface of the Earth is broken up and its mineralogical and chemical composition altered substantially. Some of this fragmented rock material may remain *in situ* as a relatively unconsolidated deposit above the bedrock from which it has been derived. However, much of the fragmented rock material has been eroded and transported by water, wind,

ice or gravity to other sites within the landscape, forming a range of deposits collectively referred to as **debris mantles**. It is these weathered, unconsolidated surface deposits, rather than the solid bedrock, which are generally regarded as the parent materials for soils.

The soil, together with any underlying weathered debris and/or weathered bedrock, is referred to as the **regolith**. The regolith overlies either unweathered bedrock or debris which shows no significant chemical weathering. In the case of relatively shallow regoliths, the soil profile may extend right to the base of the regolith. In other situations, for example the deep loess deposits in parts of Canterbury or the volcanic deposits on the central plateau of the North Island, the soil profile may represent just a fraction of the total depth of the regolith. The pathways of soil formation, showing the relationships between the various types of deposit described above, are outlined in Figure 2.1. The terms chemical and physical weathering will be explained in detail in Section 2.3.

The formation of a soil in these mineral deposits is a complex process and it may take many years for a developing soil to acquire distinct profile characteristics. One extremely important aspect of soil formation is the incorporation of organic matter into the mineral fabric of the soil. Indeed, the recognition of a soil, as distinct from a purely geological deposit, is generally agreed to depend on the appearance of a significant biological component of weathering within the surface layer of the regolith. The next four sections in this chapter examine the various stages and processes involved in soil development.

2.1 THE NATURE AND COMPOSITION OF BEDROCKS AND MINERALS

2.1.1 Minerals
Rocks consist of mixtures of various minerals which are themselves chemical compounds formed from the elements present in the Earth's crust. The two most
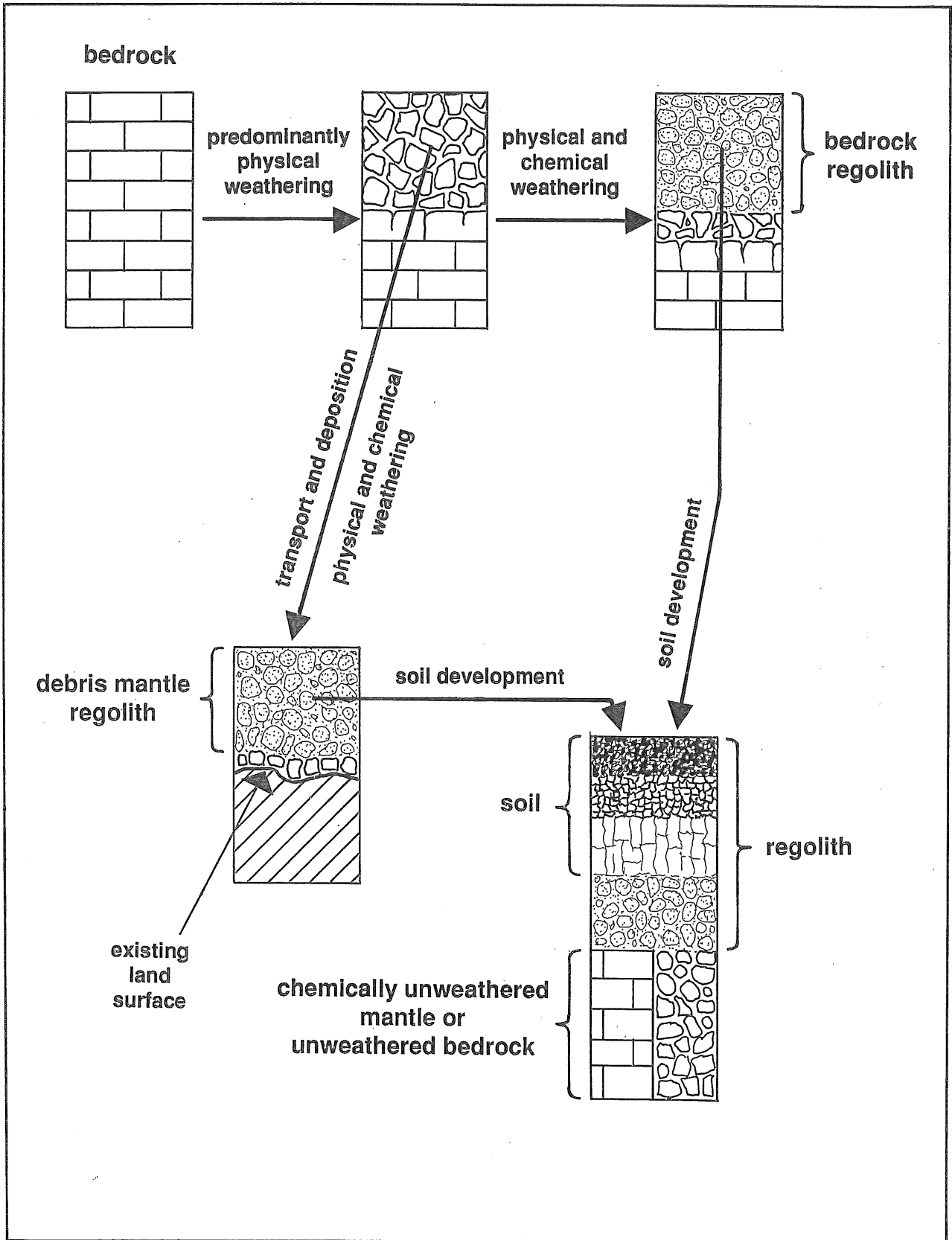
Figure 2.1: Pathways of soil formation showing the relationship between bedrock, regolith and soil.

abundant elements in the Earth's crust are silicon and oxygen, and combinations of these elements form the basis for all the types of silicate minerals that occur in rocks and soils. Silicon and oxygen occur together mainly in a tetrahedral arrangement in which the small silicon ion is surrounded by, and bonded to, four oxygen ions (see Figure 2.2). Depending on the type of mineral, these tetrahedra occur singly or, by sharing oxygen ions, are arranged in chains, sheets or three-dimensional structures (e.g. see Chapter 11). Minerals consist of these structures held together and stabilized by the presence of various cations such as potassium, calcium, magnesium and iron. In many of the minerals some of the silica is substituted in the tetrahedra by aluminium. The structures and elemental contents of the most common rock-forming minerals are shown in Table 2.1. Those minerals which have a high content of cations such as calcium, iron and magnesium, are referred to as ferromagnesian minerals.
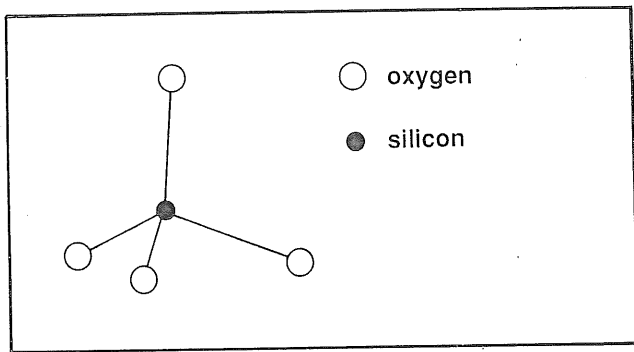
Figure 2.2: Silicon–oxygen tetrahedra.

The minerals listed in Table 2.1 are regarded as examples of primary minerals, that is they occur originally in igneous rocks (see below). However, during rock weathering and soil development, other secondary minerals may be formed. The most important groups of these are the phyllosilicate clay minerals and the non-silicate oxides and hydrous oxides of iron and aluminium. The structures and properties of these types of mineral will be discussed in detail in Chapter 11.

2.1.2 Bedrocks

Bedrocks are divided into three main classes: igneous, sedimentary and metamorphic, depending on how they have been formed. Rocks from all three classes are important contributors to soil parent materials in New Zealand.

Igneous rocks

Igneous rocks are formed from molten silicate material or **magma**, which crystallizes on cooling to form an aggregate of predominantly silicate minerals. Some igneous rocks are formed by the cooling of molten magma below the earth's surface and are referred to as **plutonic** igneous rocks. Others are formed by the eruption of magma at the earth's surface, i.e. **volcanic** igneous rocks. Volcanic rocks include the **lavas**, which are erupted in molten form, and material known as **ignimbrite**, which is pushed over the surface of the land as a flow of solid particles (pyroclastic flow). Volcanic material erupted into the atmosphere and deposited on the surface of the land as airfall material is referred to as **tephra** (see also Section 2.2.4).

Table 2.1: Common primary rock-forming minerals.

Mineral	Structural arrangement of Si-O tetrahedra	Si content	Content of Ca, Mg, and Fe
quartz	continuous framework of tetrahedra each sharing all four oxygens		
feldspars (e.g. orthoclase, plagioclase)			
micas (e.g. biotite, muscovite)	continuous sheets of tetrahedra each sharing three oxygens		
amphiboles (e.g. hornblende)	continuous double chains of tetrahedra sharing alternately two and three oxygens	decreasing Si content	increasing Ca, Mg and Fe content
pyroxenes (e.g. augite)	continuous single chains of tetrahedra sharing two corner oxygens		
olivine	isolated tetrahedra		

All igneous rocks consist of assemblages of individual crystals of various minerals, and different rock types are recognized on the basis of their crystal grain size and mineralogical composition. A simplified classification of the major igneous rocks found in New Zealand is shown in Table 2.2.

The terms felsic, intermediate and mafic refer to the proportions of the different silicate minerals found in these rocks. **Felsic** rocks contain relatively large amounts of silica-rich minerals, particularly quartz, potassium-rich feldspars and micas. **Mafic** rocks contain a higher proportion of ferromagnesian minerals such as augite, olivine, and hornblende. The feldspars in mafic rocks also tend to contain more calcium and less potassium than those in felsic rocks. The term **ultramafic** is used sometimes to describe rocks with extremely high proportions of ferromagnesian minerals.

Sedimentary rocks

These rocks are formed by compaction and cementation of sediments laid down mainly in marine environments. They contain material that has already undergone one or more cycles of weathering, transportation, deposition and consolidation. For this reason, the most easily weathered minerals found in igneous rocks are often absent from sedimentary rocks.

There are two major groups of sedimentary rocks:
(i) **Clastic rocks**. These rocks are composed of materials inherited from previously existing rock formations. These contain predominantly minerals resistant to weathering, mainly quartz and feldspars, together with some clay minerals formed as a result of the weathering of other silicate minerals (see section on chemical weathering). Classification of this type of rock is based initially on particle size (Table 2.3). Further subdivision is based on mineral composition, degree of induration and the nature of the cements holding the particles together.
(ii) **Non-clastic rocks**. These rocks are formed by the accumulation of biological remains (e.g. shells, organic matter) or by direct chemical precipitation. The dominant minerals involved in such processes are calcium carbonate, giving rise to limestone rocks, and magnesium carbonate, giving rise to dolomite.

Metamorphic rocks

These rocks are formed by the action of heat or pressure, or both, on pre-existing igneous or sedimentary rocks. In New Zealand the greatest areas of metamorphic rocks occur in the Otago land district and the western Southern Alps where they have resulted from the immense temperatures and pressures generated by deformation of rocks within the crust, and are now exposed as a result of uplifting and folding of the landscape. During metamorphosis, existing minerals may be altered, new ones formed, and

Table 2.2: Simplified classification of major New Zealand igneous rocks.

Crystal grain size	Mineralogical composition		
	felsic	*intermediate*	*mafic*
coarse	**granite**	**diorite**	**gabbro**
fine	**rhyolite**	**andesite**	**basalt**

Table 2.3: Types of clastic sedimentary rocks.

Rock type	Properties
conglomerate	coarse particle size, cemented stones and gravels
sandstone	dominant particle size 2–0.05 mm (includes **greywacke**, a poorly sorted silica-cemented sandstone)
mudstone	dominant particle size less than 0.05 mm includes **siltstone** and **argillite** (silica-cemented mudstone)

the fabric of the rock transformed with alternating bands of minerals.

Classification of metamorphic rocks is complex and is based on a combination of factors, including mineral composition, rock structure and mineral grain size. The two most important metamorphic rocks from the point of view of soil formation in New Zealand are the **schists** and **gneisses**. Schists are formed mainly from sedimentary rocks and show a well defined subparallel arrangement of cleavage planes (**foliation** see Plate 2.1). They range from the finer textured perlitic schists formed from argillaceous sediments to the coarser psammitic schists formed from sandy sediments. Sub-types of schists are named after the dominant minerals they contain, which indicate the degree of metamorphism the rock has undergone: metamorphism increasing in the order chlorite-biotite-garnet schist. With increasing metamorphism, the crystal size of minerals also increases, eventually resulting in a distortion of the subparallel arrangement of mineral crystals. At this stage the rock is regarded as a gneiss and although it may still show some banding of minerals, unlike the schists, gneiss is not easily split along these layers. Gneiss is formed from both sedimentary and igneous rocks.

2.2 SOIL PARENT MATERIALS

The general distribution of different types of soil parent material (i.e. bedrock and debris mantle regoliths) in New Zealand is quite complex. Soils in the hill and mountain regions of both Islands are generally formed in regolith derived directly from the underlying bedrock or in material moved relatively small distances downslope under the influence of gravity (colluvium, see below). In the
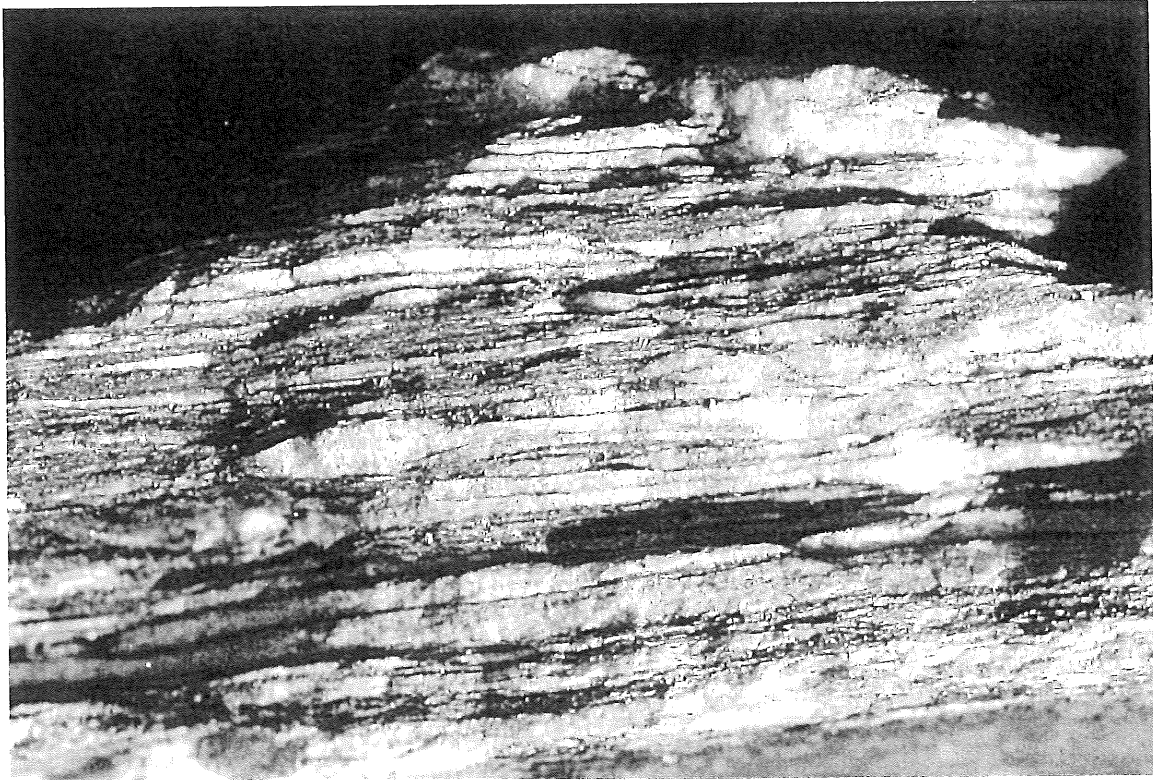
Plate 2.1: Schist rock sample showing parallel banding (foliation) of light and dark coloured minerals (sample approximately 20 cm wide). (Photo: D.P. Hollander)

South Island, the rocks involved are mainly igneous (granite), metamorphic (gneisses and schists) or hard sedimentary rocks such as greywacke, whereas in the North Island mainly softer sedimentary rocks are present. In both Islands there are also significant areas of hard volcanic rocks such as basalt and andesite.

However, many soils in New Zealand are developed in relatively unconsolidated debris mantle deposits that have been transported considerable distances from where the bedrock was originally exposed to the environment. For example, large areas of soil in the North Island are developed in volcanic tephra, and in the South Island substantial areas of soil are developed in wind-blown deposits. Soil formation can be strongly influenced by the nature of such deposits. Debris mantle deposits are classified according to their mode of transport and can usually be identified by the properties of the deposits and their position in the landscape.

2.2.1 Deposits transported by gravity

Colluvium

Colluvium refers to deposits that accumulate on and at the base of slopes as a result of movement by gravity. Such deposits result from a build-up of unconsolidated material which eventually becomes unstable and moves downslope. Often movement is triggered by saturation of

the material with water which decreases frictional forces. The original accumulation of material may be due to weathering of rock *in situ* (i.e. bedrock mantle) or to accumulation as a debris mantle by one of the other processes described below. Movement of the deposits may be a gradual process or may take place in short catastrophic events.

Colluvium often shows evidence of successive periods of deposition (layering, including buried soils) and rock fragments in the deposits often have sharp edges indicating that they have moved relatively small distances.

2.2.2 Water transported deposits

Alluvium

Alluvium refers to sediments deposited by streams and rivers. The particle size of alluvial deposits can vary greatly from large stones and boulders to fine silts and clays. Stones in the deposits are usually rounded or sub-rounded. Alluvial deposits often consist of layers of different textured material, for example, sand and silt layers overlying coarse gravels. Such deposits are common soil parent materials in many parts of New Zealand including the Canterbury, Heretaunga and Manawatu Plains.

Also included under the heading of alluvium, because of its similarity, is the material deposited by streams and rivers of meltwater draining from glaciers; sometimes

these are referred to as **fluvio-glacial deposits**. Such deposits and associated landforms are extremely varied, ranging from narrow ridges to huge outwash plains such as those located at mouths of formerly glaciated valleys along the Southern Alps. For instance the Mackenzie Basin contains the debris discharged by meltwater from the glaciers originally occupying the sites of Lakes Tekapo, Pukaki and Ohau.

2.2.3 Ice transported deposits

Till

Till refers to compacted deposits of poorly sorted material, consisting usually of large stones in a silty to gravelly sand matrix. These deposits occur in areas which were formerly overridden by a valley or piedmont glacier. In New Zealand such deposits are of relatively limited extent being mainly restricted to the previously glaciated areas of the Southern Alps. The most common landforms associated with till are called 'moraines'.

2.2.4 Wind transported deposits

Loess

Loess refers to deposits, usually dominated by silt-size particles, which have been blown by wind from mountains, glacial outwash plains and the floodplains of rivers. These deposits, which blanket and smooth pre-existing topography, vary in depth from a few centimetres to several metres and are common on the plains and downland areas in the eastern half of the South Island from Southland to eastern Marlborough. Substantial areas of loess are also found in the central and southern part of the North Island, for instance in the Manawatu and in southern Hawke's Bay. In places where the loess cover is thin, it may be partially mixed with underlying deposits, for instance alluvium, or the upper part of the soil profile only may be developed in the loess material.

Sand dunes and plains

Accumulations of wind-blown sand occupy narrow areas bordering many parts of the coast of New Zealand. The most extensive areas of these deposits are found on the west coast of the Manawatu lowland and the west coast of North Auckland. The Manawatu sand country consists of alternating dunes (sand hills) and sand plains, in places extending several kilometres inland. The dunes can reach 30 m in height and often feature upwind tails which give them a 'hairpin' or 'canoe' shape. Dune faces have gentle windward and steeper leeward slopes. In places the dune pattern can be more complex with subsidiary apices along the wings. Elsewhere the dunes may be crescent shaped and arranged in chains.

Tephra

Airfall material erupted from volcanoes into the atmosphere and re-deposited on the surface of the earth. Layers of this type of material are common in the central regions of the North Island. The term tephra describes volcanic ash, which consists of fragments or particles less than 2 mm in diameter, lapilli which are particles greater than 2 mm in diameter and pumice consisting of fragments of light-weight 'porous' material.

2.3 WEATHERING

Weathering refers to the many processes which cause alterations in the nature of rocks and minerals to form the regolith and also the further development of regolith material to form soil (see Figure 2.1). The effects of weathering result in the disintegration of mineral matter into increasingly smaller fragments or particles (referred to as **physical weathering**), together with a change in the mineralogical and chemical composition of the material (referred to as **chemical weathering**). These two weathering processes occur simultaneously and are interdependent on each other. Increased physical weathering tends to stimulate further chemical weathering and vice versa. In the early stages of bedrock weathering however, it is probably physical weathering that often appears to be the dominant process taking place.

Most bedrock exposed at the surface of the earth contains stress fractures resulting from rock deformation or unloading as a result of surface erosion. These fractures or fissures in the rock act as sites where various reactions can take place, causing additional stresses in the rock and resulting in further disintegration. For example, if water enters a fissure and then freezes, the formation of ice causes considerable pressure to be exerted on the rock. This may then result in the fissure being widened or even in the rock being split. Crystallization of certain salts from saline solutions in rock fissures can set up similar destructive forces. Disintegration may also be aided by raindrop impact and wind blasting. Other processes such as alternate heating and cooling of rocks, and plant root activity have also been implicated in rock fragmentation, but the effects of such processes overall is likely to be relatively small.

Not all physical weathering of rock materials takes place *in situ*. Once the original bedrock starts to fragment and is eroded and transported to other sites within the landscape, further physical weathering of the rock will take place by simple mechanical abrasion.

Although some of the processes described above have a purely mechanical effect on the rock, in the presence of water it is probably the hydrolysis of minerals such as micas and feldspars (see below), with subsequent volume

increase, that has the major effect on the fragmentation of the rock. Once significant amounts of rock debris start to accumulate, chemical weathering processes, such as hydrolysis and others described below, become increasingly active. Indeed the physical fragmentation of rock into increasingly smaller particles provides a greater exposed surface area of rock at which chemical reactions can take place. As described earlier in this chapter, surface deposits (debris mantles) which show significant chemical weathering are referred to as 'regolith'.
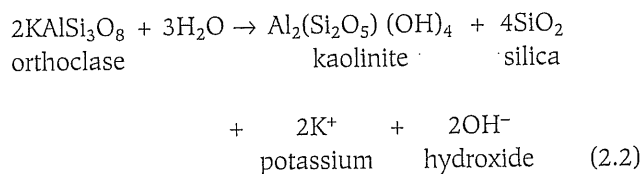
The process of chemical weathering consists of a number of different types of chemical reaction:

Hydrolysis

The chemical reaction between minerals and water, or more precisely between the H^+ and OH^- ions of water and the ions of the mineral, is called hydrolysis. This reaction is probably the most important of the chemical weathering reactions. The result of hydrolysis is the replacement of basic ions in the mineral structure by hydrogen, with consequent collapse and disintegration of the structure. For example the mineral olivine disintegrates into magnesium ions and silicic acid, as illustrated by the following equation.

$$Mg_2SiO_4 + 4H^+ \rightarrow$$
olivine   hydrogen

$$2Mg^{2+} + H_4SiO_4$$
magnesium   silicic acid   (2.1)
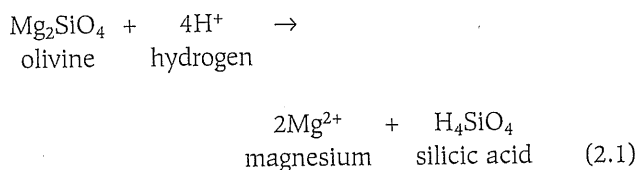
In many cases, where the silicate mineral contains aluminium, complete disintegration does not occur but transformation to a secondary clay mineral takes place. For example orthoclase feldspar can be transformed into the clay mineral kaolinite.

$$2KAlSi_3O_8 + 3H_2O \rightarrow Al_2(Si_2O_5)(OH)_4 + 4SiO_2$$
orthoclase   kaolinite   silica

$$+ 2K^+ + 2OH^-$$
potassium   hydroxide   (2.2)

Another example is the replacement of interlayer potassium in micas to produce the clay minerals illite and vermiculite (see Chapter 11). In addition, where silica and aluminium are released into solution by hydrolysis they may recombine to form the amorphous mineral allophane.

Oxidation/reduction

Many silicate minerals contain cations that can exist in more than one oxidation state. The most important of these are iron (Fe^{2+}/Fe^{3+}) and manganese ($Mn^{2+}/Mn^{3+}/Mn^{4+}$). These ions are generally present in their reduced forms (Fe^{2+} and Mn^{2+}) in mineral structures and become oxidized when exposed to the atmosphere. Such transformations (e.g. Fe^{2+} to Fe^{3+}) disrupt the overall electrostatic charge of the crystal structure and force other ions to leave the lattice to maintain electrostatic neutrality. This causes the mineral structure to become unstable and more susceptible to other reactions such as hydrolysis.

In cases where the reduced forms of elements are released from minerals by other weathering reactions, oxidation of the elements may then follow. In this way oxides of iron and manganese may be formed.

In reducing environments, e.g. under conditions of waterlogging and oxygen deficiency, elements will remain, or be transformed back into their reduced forms.

Solution

Although all minerals are soluble to some degree, the rate of dissolution varies greatly between minerals. The relative ease of dissolution of the silicate minerals is shown in Figure 2.3, the solubility of a mineral like quartz being extremely low indeed. More soluble than the silicate minerals are the simple minerals such as calcium sulphate (gypsum) and calcium and magnesium carbonates (limestones). These minerals are relatively easily dissolved in water or dilute solutions of carbonic acid (i.e. either rainwater or soil water containing dissolved CO_2). On dissolving, the constituents of such minerals are often
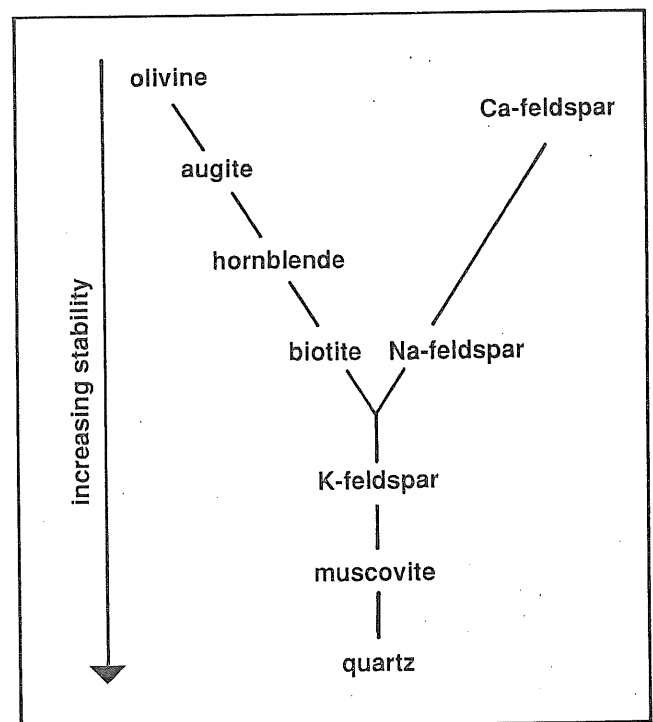

Figure 2.3: Stability of minerals to dissolution and weathering (adapted from Goldrich, 1938).

leached completely from the developing soil, leaving very little in the way of a mineral residue.

Hydration/dehydration

These are processes in which water molecules are added to or removed from a mineral resulting in the formation of a new mineral, usually accompanied by a change in volume. For example:

$$CaSO_4 + 2H_2O \rightarrow CaSO_4 \cdot 2H_2O$$
$$\text{anhydrite} \qquad \qquad \text{gypsum} \qquad (2.3)$$
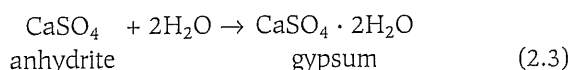
The mica 'biotite' is also able to absorb water between its layers, the resulting expansion causing the mineral to split apart. More commonly however, water molecules are adsorbed at the broken edges of minerals, acting as the first step in hydrolysis. Once decomposition products have been released by hydrolysis reactions, they may then be hydrated to varying degrees. The principal compounds affected in this way are iron and aluminium oxides.

Chelation

Chelation (or complexing) involves the binding of a metallic ion within a chemical ring structure usually of organic origin. Ions can be extracted from otherwise insoluble minerals by chelating agents produced by plants and micro-organisms and decaying organic matter.

Essentially, ions in mineral structures are replaced by H^+ ions from the organic chelates, hence chelation acts in an identical way to hydrolysis. However in some situations the amount of weathering by chelation will far exceed that brought about by hydrolysis alone.

Biological activity and chemical weathering

Chelation is not the only process involved in chemical weathering that can be affected by biological activity in the soil. The respiration of CO_2 by plant roots and micro-organisms increases the acidity of the soil solution through the formation of carbonic acid. The increase in H^+ ion concentration can have a marked effect on hydrolysis reactions and will increase the solubility of carbonate minerals where present. In addition there are also many bacteria which are able to carry out various chemical transformations involving minerals and other inorganic compounds. Probably the best known examples are the various microbiological oxidation/reduction reactions involving iron, sulphur and manganese compounds.

2.3.1 The products of weathering

As a result of the weathering processes described above, the chemical and mineralogical composition of the soil or regolith can be substantially different from that of the bedrock from which they are derived. The products of

weathering can be categorized into three main groups: resistant primary minerals, secondary minerals and soluble materials (Figure 2.4).

Resistant primary minerals

There are considerable differences in the ease with which individual rock-forming minerals are weathered. Quartz, for instance, is extremely stable and very resistant to weathering, whereas the ferromagnesian minerals are relatively easily weathered (Figure 2.3). Most soils contain some primary minerals inherited directly from the bedrock, and in many soils it is the minerals which are relatively resistant to weathering, mainly quartz and to a lesser extent the feldspars, that are the dominant components of the sand and silt-sized mineral fractions and of any rock fragments in the soil.
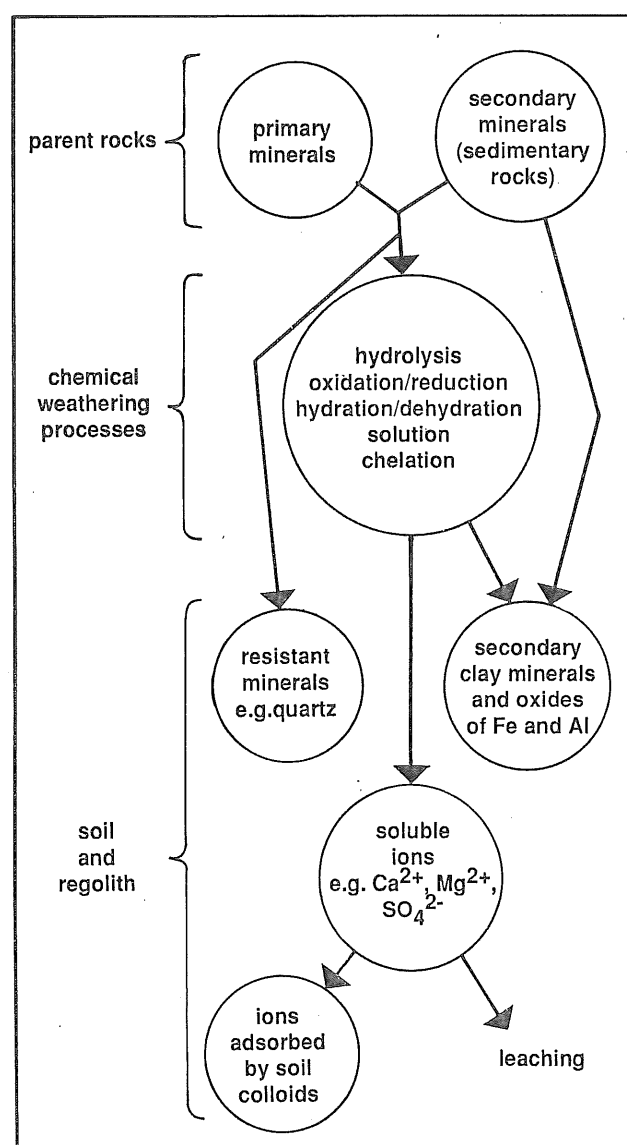

Figure 2.4: Pathways and products of weathering.

Secondary minerals

Secondary minerals are those formed from materials released during rock weathering and soil development. Some secondary minerals may be inherited by soils directly from bedrocks where these are of sedimentary types. Others will be formed during the weathering of the regolith and soil. As mentioned in Section 2.1.1, the most important groups of secondary minerals are the phyllosilicate clay minerals and the oxides and hydrous-oxides of iron and aluminium. In most soils these types of minerals are the dominant constituents of the clay-sized mineral fraction, which, together with soil organic matter, has a major influence on soil chemical properties.

In addition, the iron minerals have a dominant influence on soil colour. The various oxides and hydrous oxides of iron (see Section 11.5) have strong red, yellow or brown colours and, when dispersed through the soil as minute particles or as coatings on other mineral particles, are responsible for the reddish-brown or yellowish-brown colours of freely drained soils. In soils with no other distinguishing morphological features, the colours resulting from the weathering and oxidation of iron may be the only obvious way of distinguishing subsoil horizons from unweathered parent material. Such horizons are usually categorized as Bw horizons and are sometimes referred to as 'cambic' horizons (see Section 1.2.3).

Soluble materials

During weathering and soil development, only some of the ions released from minerals become incorporated into newly formed secondary minerals. Substantial quantities of the more soluble ions (for example Ca^{2+}, Mg^{2+}, Na^+, K^+, Cl^-, NO_3^- and SO_4^{2-}) will be lost from the soil by leaching or may be reprecipitated as new compounds in lower horizons of the soil profile (see Section 2.4).

The relative amounts of material lost by leaching or remaining in the soil as primary or secondary minerals or adsorbed by soil colloids, and the actual types of minerals occurring, will depend to a large extent on the intensity of weathering and leaching processes. These are in turn controlled by the various environmental factors discussed in Section 2.5 below. In most temperate region soils, weathering has resulted in soil mineral fractions dominated by the resistant primary minerals and the secondary phyllosilicate clay minerals, although the relative proportions and types of mineral can vary greatly. The secondary oxides and hydrous oxides of iron and aluminium, although still of great importance for certain soil properties, usually account for only a few per cent of the total bulk of the soil.

In contrast, in extremely highly weathered soils, for example, some very old soils or soils developed under conditions of intensive weathering and leaching, it is the oxides of iron and aluminium and certain aluminium-rich clay minerals, such as kaolinite, that dominate the soil mineral fraction. In New Zealand such soils are found in small areas in Nelson and Marlborough and more extensively in the north of the North Island.

2.4 PEDOGENIC PROCESSES

The chemical and physical weathering processes described in the previous section are important in both the transformation of solid bedrock to form regolith and the continued development of regolith to form soil. The development of mature soil profiles however is not due to weathering alone. There are several other types of process involved, sometimes referred to as 'pedogenic processes', which are responsible in the main for the development of distinct horizons in the soil profile. These processes involve the redistribution of materials within the soil profile and the addition to, and loss of materials from, the developing soil.

Although, for convenience, the various processes are discussed individually below, the reader should be aware that in any one soil, many of the different processes will be taking place simultaneously. The morphological characteristics and properties of a soil will depend on the balance between the various processes, a balance controlled by the environmental factors discussed in Section 2.5.

2.4.1 Accumulation of organic matter

When dead plant or animal material is added to the developing soil it is decomposed by the many organisms present and a dark-coloured, amorphous material known as humus is formed. Humus is the main organic component of the soil and its formation, properties and importance will be described in Chapter 10. Although humus itself is also slowly decomposed, during soil development humus levels will gradually increase until the rate of humus decomposition equals the rate of its formation. The final humus (organic matter) level in the soil will depend on the various environmental factors discussed in Section 2.5.

In well-drained, nutrient-rich soils, earthworms and other organisms will thoroughly mix the humus with the surface mineral matter so forming a dark-coloured Ah horizon (a process referred to as **melanization**). However under acid, nutrient-poor conditions the decomposer and mixing organisms will be few in number and organic matter tends to accumulate on the surface with little mixing with underlying mineral matter. In such a situation organic materials may form separate L (litter), F (fermentation) and H (humus) horizons at the soil surface. Partially decomposed organic matter also accumulates at

the surface under very wet conditions so forming a distinct layer of peat (O horizon).

2.4.2 Leaching

In all areas of New Zealand at some period of the year, water, having fallen as rain, will move downwards through the soil profile and may leach with it any soluble materials present in the soil. The principles of water flow and solute leaching are described in detail in Chapter 7. Over time, some soils, particularly those in high rainfall areas, may lose substantial quantities of the more soluble ions, such as NO_3^-, SO_4^{2-}, Ca^{2+}, Mg^{2+} and Na^+, in water draining from the base of the soil profile. In some soils even less soluble ions such as Si^{4+} and Fe^{3+} may also be lost by leaching. In other soils, certain materials may not necessarily be leached completely from the soil but accumulate in horizons further down the profile.

Clearly, leaching is very closely inter-related with weathering. Weathering of minerals produces the constituents that are leached from the soil and leaching, by the removal of basic cations, causes acidification of the soil (see Chapter 12), which in turn enhances weathering.

The combination of leaching and weathering has marked effects on the sequence of soil profile development. In the early stages of development, apart from the loss of the most soluble constituents from the upper layer of the regolith, the effects of leaching are unlikely to be pronounced. The main process occurring will be the development of a distinct organic-rich Ah horizon as described above (Section 2.4.1). Eventually the effects of weathering and leaching extend below the developing A horizon, producing a horizon distinguished in colour and structural organization from the regolith below. Such horizons are referred to as Bw, or **cambic**, horizons.

Further weathering and leaching result in the development of a pale-coloured horizon (E horizon), consisting mainly of residual quartz grains, from which most other constituents have been weathered and leached downwards. Below the E horizon, some of the products of weathering, particularly iron and aluminium minerals (sometimes collectively referred to as **sesquioxides**), accumulate in the form of a Bs horizon. The presence of iron oxides and hydrous oxides in this layer usually give it a relatively bright ochreous-brown colour. At this stage of soil development, i.e. with the formation of E and Bs horizons, the soil is said to be **podzolized**. By the time the features of podzolization are apparent in the soil profile, continual leaching of the soil has resulted in the development of fairly acidic conditions at the surface of the soil. As described in Section 2.4.1, under such conditions, organic matter tends to accumulate on the surface as L, F and H horizons with very little mixing with underlying mineral matter. Along with the development of

Bs horizons during podzolization, in some soils the iron minerals are concentrated in a narrow, hard cemented layer referred to as an **iron pan** or Bfm horizon. Organic matter may also be leached down through the soil and accumulate as a dark-coloured Bh horizon.

The general sequence of soil development described above is outlined in Figure 2.5. The rate and extent to which such events happen will be determined by the various factors discussed in Section 2.5. In addition, other processes as described below may also take place simultaneously, resulting in the development of other distinct morphological features within the soil profile.
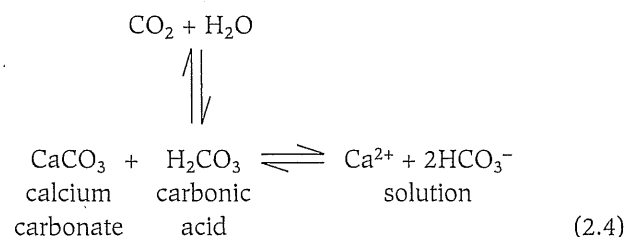
2.4.3 Accumulation of soluble salts

Soils in low-lying areas may receive sub-surface inputs of water that have drained from surrounding higher regions. This water sometimes contains soluble salts that have been leached from the soils and regoliths through which it has passed. In this way, the most soluble products of chemical weathering (for example, chlorides and sulphates of magnesium, calcium and sodium) may accumulate in the low-lying soils. However, such accumulations are only of major significance in semiarid environments, where, as a result of loss of water from the soil surface due to high evapotranspiration rates, the salts become concentrated and precipitate out. Soluble salts may also accumulate in soils of coastal regions due to the presence of shallow saline water tables originating from the sea.

The term salinization is used to describe the accumulation of salts in soils and in some parts of the world salinization creates a major problem in the utilization of soils. Luckily, although there are small areas affected by salinity in the semiarid environment of Central Otago and in coastal areas like the surrounds of Lake Ellesmere in Canterbury, it is by no means a major problem in this country.

2.4.4 Accumulation of calcium carbonate

In some soils calcium carbonate (and possibly other calcium compounds) accumulates at depth in the soil profile (a Bk horizon, see Chapter 1). Calcium carbonate is solubilized in soils mainly by its conversion to calcium bicarbonate through reaction with dissolved CO_2.

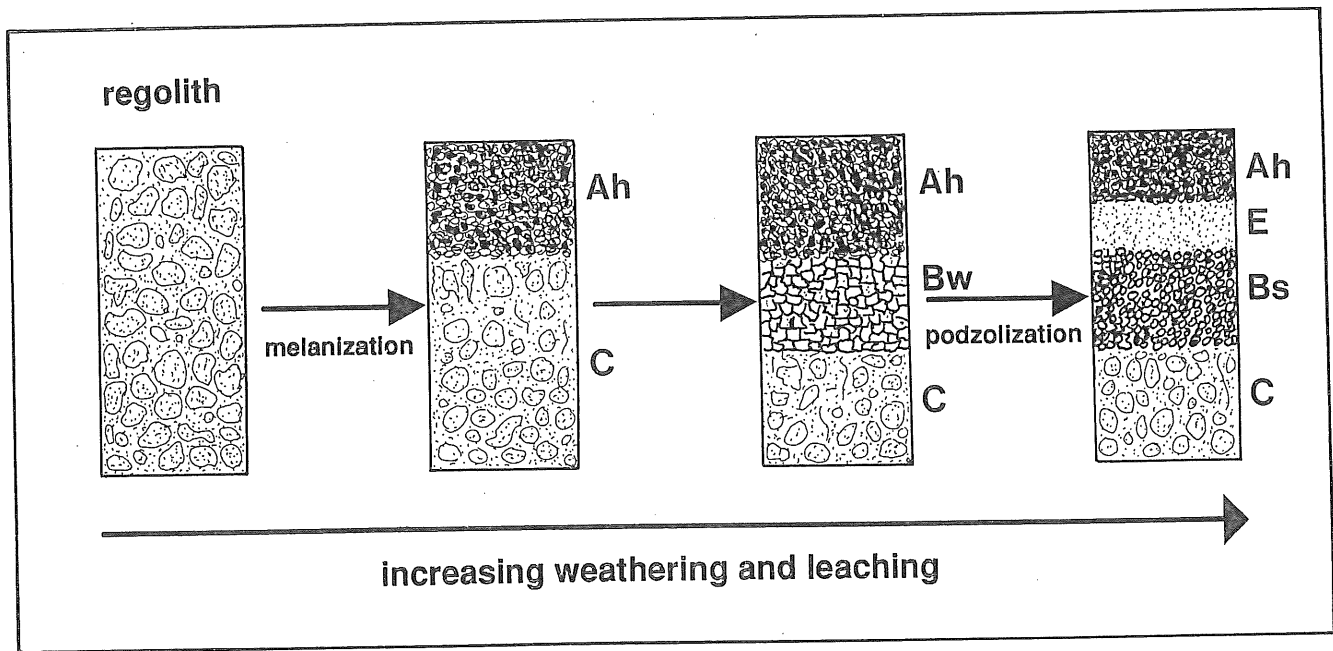$$CO_2 + H_2O$$

$$\Updownarrow$$

$$\underset{\substack{\text{calcium} \\ \text{carbonate}}}{CaCO_3} + \underset{\substack{\text{carbonic} \\ \text{acid}}}{H_2CO_3} \rightleftharpoons \underset{\text{solution}}{Ca^{2+} + 2HCO_3^-} \tag{2.4}$$

Figure 2.5: General sequence of soil development resulting from weathering and leaching.

The level of CO_2 in the soil will, of course, be greatly influenced by the intensity of biological activity of both plant roots and soil micro-organisms.

In humid climates the soluble calcium bicarbonate is removed from the profile by drainage to depth, but in semiarid environments with a marked dry season, re-precipitation of the calcium carbonate can take place further down the soil profile. The re-precipitation is thought to result from a combination of two main factors. Firstly, biological activity and hence CO_2 levels decrease with depth in the soil and the above reaction is reversed. Secondly, the concentration of soil solution, as a result of evapotranspiration and drying out of the profile, also causes conversion of the bicarbonate back to carbonate. The depth of the calcium carbonate horizon in the profile is considered to mark the extent of soil moisture penetration in the soil. In New Zealand, distinct Bk horizons are only found in some Semiarid Soils of Central Otago (see Section 3.3.1).

2.4.5 Accumulation of colloids

Leaching generally refers to the movement of substances in solution, however some solid materials can also be moved downwards through the soil profile. Solids are usually transported in the form of colloidal suspensions, and probably the best example is the movement of colloidal clay particles. In some soils, the transported clay accumulates further down the profile to form a clay enriched horizon referred to as a Bt horizon (see Chapter 1).

The first stage of clay movement is the dispersion by water of fine clay material from unstable aggregates at the soil surface. The clay may remain in suspension as water moves downward through the larger soil pores. If the subsoil is dry, the downward moving water will be absorbed by the soil matrix and the suspended clay will be filtered out and deposited on the walls of the pores. Factors that favour clay movement are: alternating wet and dry conditions, a good system of macropores in the soil, an unstable soil structure, pH between about 4.5 and 6.5, or a high pH associated with exchangeable sodium. Soluble organic matter may help to keep the clay dispersed and leaching of salts or carbonates may produce pores that favour a more rapid flow of soil water. The deposition of clay is probably favoured by high pH, perhaps due to the presence of carbonates, and to less porous material that slows the rate of flow.

2.4.6 Mixing and nutrient cycling

The processes of mixing and nutrient cycling tend to counteract some of the general downward movement of materials through the soil profile caused by leaching, podzolization, etc. Mixing of soil material, sometimes referred to as **pedoturbation**, can be carried out by various organisms in the soil especially earthworms, although ants, termites and larger burrowing animals may also be important in some countries. Soils in forests can be mixed thoroughly when large trees topple over as a result of death or wind throw. Soil associated with the tree root mass is inverted and, in the time span covering

several generations of trees, all of the soil in a forest may undergo this process. In some areas soil mixing may also be aided by purely physical processes such as the alternate freezing and thawing of ground, or the alternate swelling and shrinking of clays.

Nutrient cycling is the process in which nutrients taken up by plants (sometimes from considerable depths) are eventually returned to the soil surface when the plant dies or sheds its leaves. As the plant litter decays some nutrients which are released can be recycled through the vegetation and so escape being leached from the soil.

2.4.7 Gleying

The process of gleying occurs when soils become anaerobic due to waterlogging, brought about by a high water table or impeded drainage within the soil profile (a perched water table). Under these conditions ferric ions (Fe^{3+}) present in iron oxides are reduced to the more soluble ferrous (Fe^{2+}) ion, a transformation carried out by certain anaerobic bacteria and requiring the presence of organic matter. This results in soils dominated by grey/olive/blue-green colours (Br or Cr horizons) compared to the yellowish-brown to reddish-browns of freely drained soils. The colour change is partly due to the formation of ferrous iron compounds which tend to be blue/green in colour and partly due to the removal of ferric oxide coatings (yellowish-brown to reddish-brown) from other mineral particles so revealing their natural greyish colours.

Many wet soils do not remain completely anaerobic and some air (containing oxygen) is able to move through root channels, worm holes or cracks between aggregates, etc. The soluble ferrous ions diffuse towards these regions where they are re-oxidized to form deposits of ferric oxide. These appear as small rust-coloured mottles (a feature of Bg horizons, see Table 1.10). This phenomenon is extremely common in soils with fluctuating groundwater or perched water tables. In some of these soils iron becomes concentrated in small hard nodules. Manganese can act in a similar way to iron, and in some waterlogged soils small black manganese oxide nodules are also visible.

2.4.8 Addition and erosion of materials

In some situations, even where soil development is well established, material may be added to or eroded from the surface of the developing soil. For instance a shower of volcanic ash may add a substantial depth of material in a relatively short time or there may be a more gradual accumulation from wind blown loess. On steep slopes the surface horizons of soil profiles may be eroded by running water or mass movement and the eroded material re-

deposited on soils further down slope. Soils on steep slopes in particular may never develop mature soil profiles (see Section 2.5.1) due to the constant slow movement of material down slope. Conversely, in accumulation sites complex soil profiles, including buried horizons, may develop.

2.5 FACTORS AFFECTING SOIL FORMATION

Although weathering in some form is obviously involved in the development of all soils, the importance of individual weathering processes and reactions will vary greatly between soils. In addition, the relative importance of the different pedogenic processes will vary from site to site. This section attempts to explain the reasons why, even in a relatively small country like New Zealand, there is so much variation in the nature of soil properties and soil profile morphology.

There are five basic factors recognized as controlling the state of soil development. These 'soil-forming' or 'state' factors are parent material (p), climate (cl), organisms (o), relief (r) and time (t). The effect on the soil, or an individual soil property (s), of these factors can be expressed by the following equation:

$$\text{soil (s)} = f(\text{p,cl,o,r,t}) \qquad (2.5)$$

That is, the nature of the soil profile or an individual soil property is a function of the interaction of the above five factors.

Although each of the state factors, for convenience, is usually considered as an independent variable, in practice there will be some inter-dependence between factors, for example, between organisms and climate. In order to study the effect of a single factor on soil formation, soil scientists have examined the variation in soil properties and morphology in sequences of soils across which only one of the factors is known to change, the others remaining more or less constant. Such sequences are referred to by the terms listed in Table 2.4. The occurrence of these types of sequence is by no means common; usually two or more of the state factors will vary across the land surface simultaneously. However, New Zealand is one country in which several good examples have been identified and studied, some of which will be described in the following sections.

2.5.1 Time

It should be clear that the weathering and pedogenic processes discussed above, with few exceptions, do not occur instantaneously but take place gradually over

Table 2.4: Types of soil sequence.

Variable factor	Sequence type
parent material	**lithosequence**
climate	**climosequence**
organisms	**biosequence**
relief	**toposequence**
time	**chronosequence**

substantial periods of time. The term soil development refers to the changes with time that take place in the soil profile as a result of these various processes. In the initial stages of soil development, changes are related mainly to the accumulation of organic matter in and on the surface mineral parent material, however, other chemical and mineralogical changes will also be occurring. With further time, changes in the chemistry, mineralogy and physical characteristics of the developing soil become more marked and extend deeper into the regolith. In addition, the development of distinct horizons as a result of the pedogenic processes becomes apparent. However, as the soil becomes increasingly altered as a result of the combined effects of weathering and leaching, further changes become slower and slower. Eventually a point is reached when the soil and its individual properties change very little over long periods of time. At this stage the soil is often referred to as having reached a 'steady state'.

The length of time a soil takes to reach this 'steady state' will depend, to a large extent, on the state factors other than time; for instance, as will be discussed in the following section, climate can have a large influence on the **rate** of soil development.

The age of a soil is generally considered to be the length of time since the land surface became relatively stable, thus enabling soil development to proceed. In New Zealand, soils vary in age from several thousand years to less than a hundred.

In the South Island most land surfaces have been influenced directly or indirectly by glaciations during the Pleistocene period (2 000 000–10 000 years before present). Substantial alpine regions were actually covered by glacial ice, and large amounts of glacial outwash, alluvial material and loess were deposited on lowland areas during and between glaciations. Any existing soil mantle would have been removed, mixed or buried by the new accumulations. Consequently, few soils in the South Island are older than 20 000 years, and many are less than 10–14 000 years (the end of the last major glaciation). Exceptions are some soils in Central Otago, Nelson and Marlborough formed on highly weathered surfaces of Tertiary age (2–65 million years old), which were relatively unaffected by the glaciations, and some soils formed on glacial moraines and outwash terraces pre-

served from older glacial advances. Since the end of the last glaciation, further erosion, transport and deposition of rock materials has taken place so that many soils on alluvial deposits, sand dunes and slopes, etc., may be just a few hundreds or thousands of years old.

In the North Island there is also a wide range of soil ages. Probably the most highly weathered soils being found in stable regions such as the Coromandel and particularly in Northland and North Auckland, areas unaffected even indirectly by glaciations. Many soils in the North Island are formed on volcanic lava or tephra, which can vary in age from a few hundred years to greater than 40 000 years. As in the South Island, some of these deposits will have undergone subsequent erosion, transport and re-deposition, producing an even greater range of ages for the start of soil development.

New Zealand is in fact an ideal place to study the effect of time on soil formation. Within the country there are several good examples of **chronosequences** which have been used for detailed examinations of the rates of soil development. For example, Figure 2.6 shows the accumulation with time of organic carbon, nitrogen and phosphorus for a chronosequence of soils developed on wind-blown sand in the Manawatu region. After initial rapid rates of accumulation of C, N and organic P during the first 1000 years, subsequent rates of gain were slower, but steady states have still not been reached after 10 000 years. Studies of some of the other New Zealand chronosequences have examined changes with time in the amounts and forms of other nutrients in the soil, and
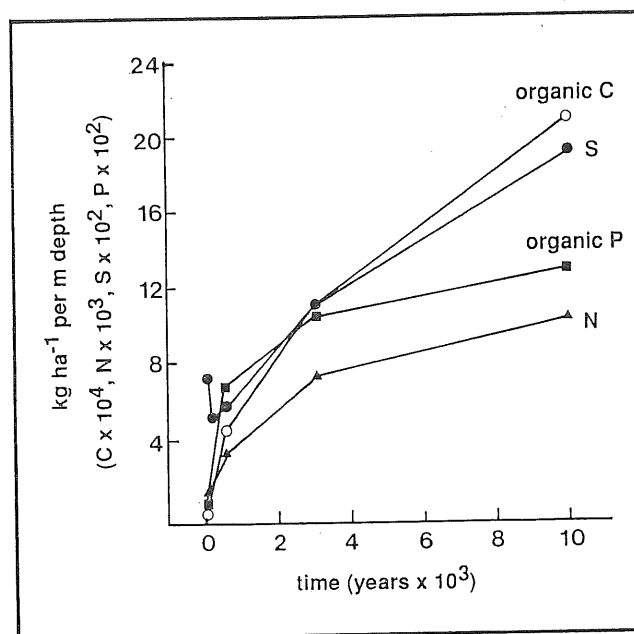
Figure 2.6: Profile weights of organic C, N, S and organic P for a chronosequence of soils developed on wind-blown sand (adapted from Syers et al., 1970).

changes in soil mineralogy. Changes in the forms of soil phosphorus with time are discussed in Chapter 15 (Figure 15.4).

In addition to changes in individual soil constituents, there are also often dramatic changes in soil profile morphology within a chronosequence.

2.5.2 Parent material

Soil parent materials and the types of bedrock from which they are derived have been described in Sections 2.1 and 2.2. It has been noted (Section 2.3.1) that the different rock forming minerals vary considerably in their susceptibility to weathering processes, hence the mineralogy of the parent material will have a significant effect on the rate of soil development. In addition, although the intensity and duration of weathering will play their part, the mineralogy of the parent material will influence the types of weathering products and therefore the mineral composition of the final soil. This is especially true in relation to the formation of the different types of clay minerals.

The overall elemental composition of the parent material will have a strong influence on the likely chemical fertility of the soil. Table 2.5 shows the elemental composition of the major types of rock contributing to parent materials in New Zealand. In this country, even though during soil development there can be considerable loss of some elements by leaching, in many situations the intensity or duration of weathering has not been great enough to override original parent material differences.

Table 2.5: Approximate elemental composition % of some common rock types*.

	Granite	Basalt	Limestone	Sandstone (including greywacke)
P	0.04	0.24	0.2–0.6	0.05
K	3.4	1.1	0.2	0.4
Ca	0.7	5.9	30.2	0.1
Mg	0.1	3.0	4.9	0.17
S	<0.01	0.3	0.1	0.01
Al	7	8	0.5	1.5
Si	35	24	2	35
Na	2.5	2.7	<0.1	0.1

* for micronutrients see Table 18.3.

It is not only the chemical and mineral composition of the parent material that can affect soil development: its physical nature is also important. The state of consolidation and particle size of the parent material will affect its permeability to water, an essential ingredient for virtually all weathering processes. Hence the rate of soil development will be affected by these properties. In addition, the drainage characteristics of the parent material may well influence the types of pedogenic process involved in soil development. For instance, processes involving downward translocation of material are likely to be enhanced in coarse textured freely drained deposits. On the other hand, some forms of gleying are more likely to take place in fine textured materials where the flow of water is restricted.

The effects of parent material on the resulting soil can be seen clearly where different parent materials exist in close proximity to one another. In such a situation the other four factors which can affect soil formation will be constant and the resulting sequence of soils, dependent solely on differences in parent material, is referred to as a **lithosequence**.

2.5.3 Climate

Climate is often considered to be the most important soil forming factor. As well as the direct influence of rainfall and temperature on soil formation, climate will of course affect vegetation (organisms) and interact with the landscape (relief) in influencing soil–water relationships.

Direct effects of rainfall and temperature

Water is an extremely important component of all chemical and many physical weathering processes. Thus the input of rainfall to a soil has a large effect on soil development both through weathering and by the leaching of weathering products through and from the soil. The rate of weathering is also strongly dependent on temperature, increasing with an increase in temperature. For every 10°C rise in temperature, the individual chemical reactions involved in weathering increase in speed by a factor of two or three.

In addition to its effect on weathering, temperature, because of its involvement in evapotranspiration, also influences leaching. As shown in Figure 2.7, the amount of surplus water available for leaching a soil will depend on the balance between rainfall and evapotranspiration. The figure shows the water balance at five contrasting sites in New Zealand ranging from a very dry area (Alexandra) with minimal leaching to an area with extreme leaching (Hokitika on the west coast of the South Island). The distribution of some of the major soil orders in New Zealand is very much related to the water balance in the various climatic regions.

Indirect climatic effects

For soils in their natural state there is usually a strong relationship between climate and the amount of organic matter (humus) in the soil. This results from the effects of climate on both biomass production (growth of vegetation) and on the rate of decomposition of plant litter and
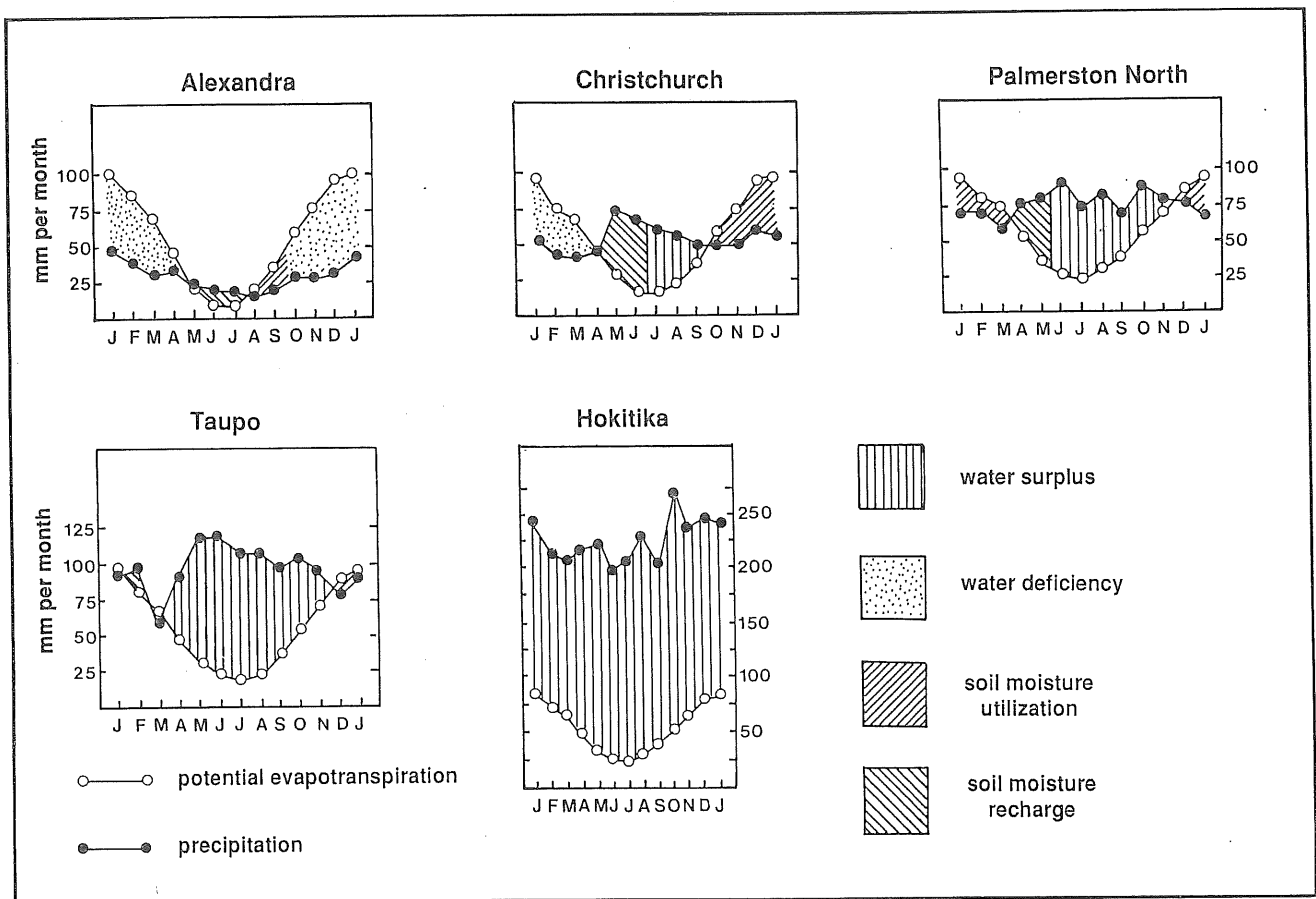
Figure 2.7: Moisture balances for some contrasting areas of New Zealand (adapted from Taylor and Pohlen, 1968).

other organic materials in the soil. As discussed in Section 2.4, soil humus levels depend on the balance between these two processes. As a result of the effects of rainfall on the productivity of vegetation, except where limited by low temperature (e.g. alpine conditions), organic matter inputs to the soil generally increase with increasing rainfall. The rate of decomposition of organic matter, on the other hand, increases with temperature, thus the highest soil humus levels are generally found in cool moist climates. Hot climates, whether dry (arid) or moist (tropical), tend to have low humus levels.

Figure 2.8 shows the effect of increasing rainfall on the organic carbon content of a sequence of soils formed on greywacke in New Zealand. The mean annual temperature is constant across the sequence (8°C). Soil sequences in which climate is the only varying factor are known as **climosequences**. Temperature and rainfall usually both vary across a climosequence, hence it is often impossible to separate the effects of these two climatic components.

2.5.4 Organisms

The many ways in which organisms, both plants and animals, can affect soil development through their influence on both weathering and individual pedogenic processes have been described in detail in Sections 2.3 and 2.4. It should therefore be obvious that the various species of flora and fauna present in an area will have a significant effect on the type of soil formed. The distribution of flora and fauna will, in turn, depend to a large extent on climatic, topographic and parent material influences. Hence it is often difficult to completely separate the effects of organisms from the effect of these other factors.

However, there are some effects of vegetation on soil formation that can be seen quite clearly. For instance the input of organic matter can vary greatly between forest and grassland soils. In forests the major input is as plant litter on the surface of the soil, whereas with grasses, because of their fine, intensively ramifying root systems, there is considerable underground addition of organic
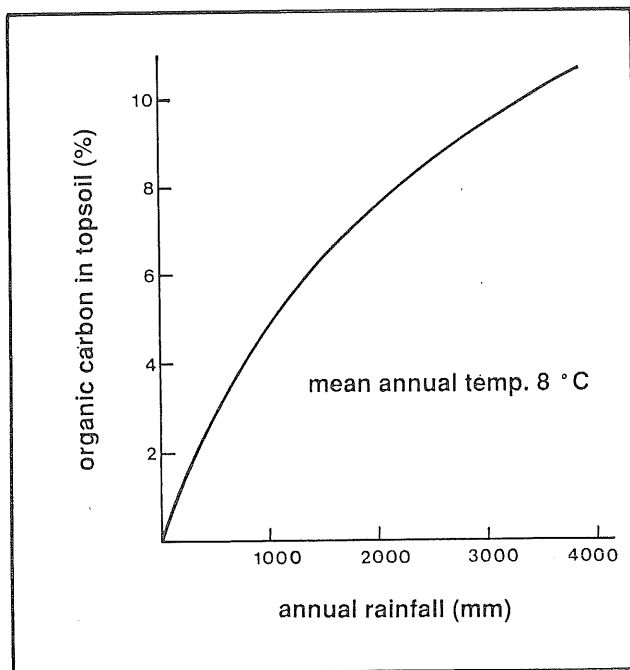
Figure 2.8: Effect of rainfall on the accumulation of organic carbon in soils on parent materials derived from greywacke (adapted from Walker and Adams, 1959).

matter as the roots decay. As a result, although other factors such as differences in pH and soil organism populations are also involved, organic matter in forest soils often occurs as distinct layers concentrated at the surface of the soil (**mor humus**) whereas in grassland soils it is thoroughly mixed by faunal activity in with the mineral matter throughout the upper layers of soil (**mull humus**). In addition, as described in Section 2.4, the decomposition of litter from some forest species releases chelates which encourage the process of podzolization.

A final factor to be considered in this section is the impact of human activity on soil formation. There can be no doubt that once the natural vegetation is removed, and the soil converted to agricultural use, the pathway of soil development is altered drastically. For instance, inputs of lime and fertilizer and removal of farm products will alter the nutrient status of the soil. Similarly the change in vegetation and cultivation of the soil will significantly affect soil humus levels and may well destroy surface pedogenic horizons. Drainage and irrigation schemes will likewise alter soil moisture relationships. It should always be remembered that most of the productive land in New Zealand, with the exception of some areas still under native forest, has been affected to some extent by human activities.

2.5.5 Relief

There are three main ways in which the shape of the landscape (topography) can affect soil formation: (i) through the influence of slope on soil depth, (ii) by modification of the effects of climate, and (iii) by influencing moisture relationships.

Relief and soil depth

Soil depth is influenced strongly by the shape of the land. Soils and regoliths developing on slopes often undergo periods of instability, during which time material moves downslope under the influence of gravity (see Sections 2.2.1 and 4.1.1). The redistribution of material over the surface of the landscape in this way can result in a considerable variation in soil depth with surface topography. In hilly areas where such movement occurs, the overall depth of soil plus regolith is likely to be shallower at sites from which material has been lost (eroded) than at accumulation sites such as in hollows.

Modification of climate by relief

In hilly and mountainous areas the orientation of a hill slope or valley side can have a large influence on soil temperature and moisture regimes. For example, as shown by the data in Figure 2.9 from a hill pasture area in north Canterbury, slopes with sunny aspects can be significantly warmer and drier than shady ones. This can result in substantial differences in soil properties between sunny and shady slopes due to differences in the intensity of weathering and leaching between slopes. However, any differences in soil development between sunny (north aspect) and shady (south aspect) sites are strongly influenced by rainfall.

Studies in the South Island have shown that in sub-humid environments (approximately 600–700 mm annual rainfall), as a result of insufficient rain, there is no difference in soil development between north and south aspects. The low and seasonally variable rainfall limits plant growth and leaching on both aspects. In subhumid to humid environments (approximately 700–1200 mm annual rainfall), north aspects remain drier and therefore soils are less leached than those on the south aspect. The result is a more advanced stage of soil development (and lower fertility) on the moister south aspect. The more favourable moisture regime on the south aspect results in a greater potential for plant growth. In contrast, in humid environments (1200–1400 mm annual rainfall), there is sufficient moisture to promote leaching on the warmer north aspect as well as on the generally cooler south aspect. The result is a more advanced stage of soil development (and lower soil fertility) on the north compared to the south aspect.
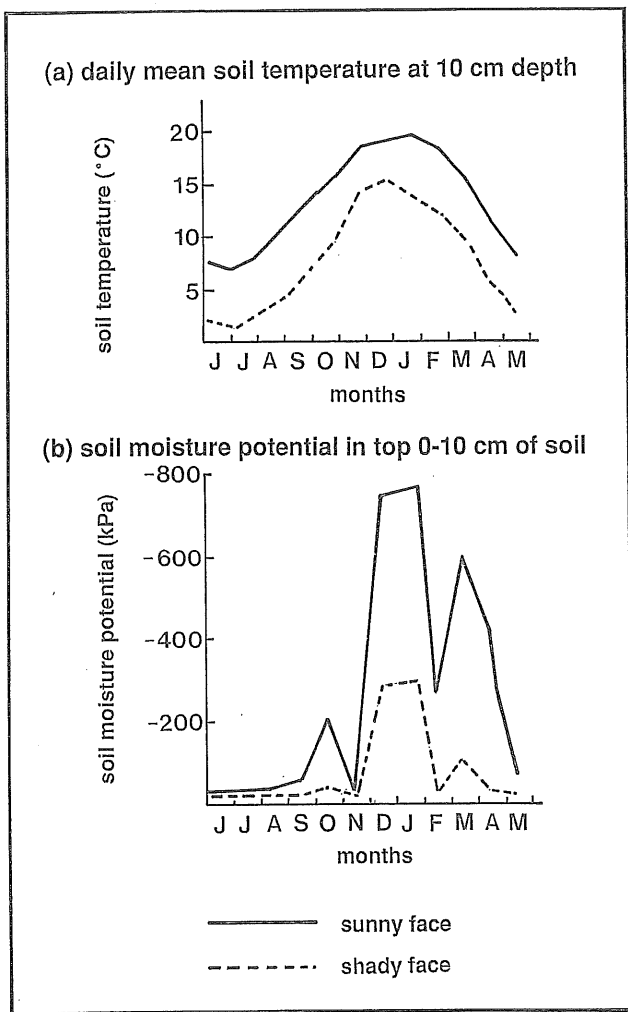
(a) daily mean soil temperature at 10 cm depth

(b) soil moisture potential in top 0-10 cm of soil

——— sunny face

- - - - - - shady face

Figure 2.9: Effect of aspect on soil temperature and moisture regimes (adapted from Radcliffe and Lefever, 1981).

Relief–moisture relationships

As we have already seen, water in one way or another has a considerable influence on soil formation. The flow of water through, or across the surface of a soil, or the position of the soil in relation to the groundwater table is largely controlled by relief. Surface run-off or lateral subsurface flow is much more likely to occur on sloping rather than horizontal land surfaces (Figure 2.10). Conversely, downward flow of water and associated leaching will be greatest for soils developing on horizontal surfaces. At low or depression sites in the landscape the accumulation of excess water can cause waterlogging and give rise to gleying and peat formation.

A good example of the interaction between relief and groundwater table on soil formation can be found in the sand country of the Manawatu region. Figure 2.11 shows the change in soil profile morphology across a sand dune-
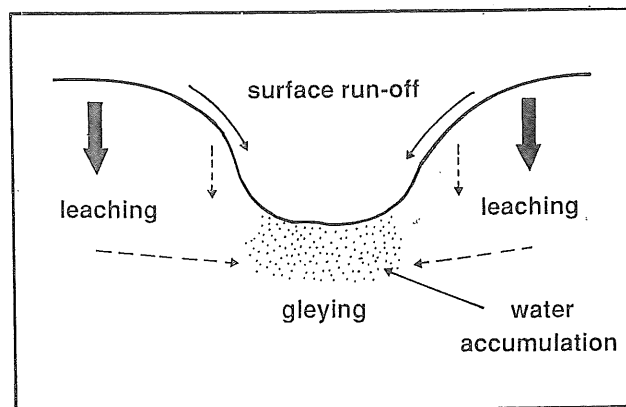

Figure 2.10: Relief–moisture relationships.

sand plain sequence. Moving across the landscape from the dunes onto the plains the soils become increasingly poorly drained as the water table comes nearer to the soil surface.

Sequences of soils in which relief is the major varying soil forming factor are known as **toposequences**. However, in toposequences covering a wide altitudinal range it may be impossible to separate climatic from purely topographical effects.

2.6 SUMMARY OF SOIL FORMATION

It is clear from the preceding sections that soil formation is an extremely complex process, but an attempt has been made to summarize in diagrammatic form the processes and controlling factors involved (Figure 2.12). The five 'soil forming' or 'state' factors control both the rate and nature of the weathering process, and the type of pedogenic processes that will occur in any particular location. The interaction of these factors in New Zealand has produced a wide diversity and complex pattern of soils throughout the country.

However, in spite of this variation, it is still possible to categorize the soils of New Zealand into a relatively small number of major groups. The description and classification of these major soil groups is covered in the next chapter.

2.7 FURTHER READING

Amundson, R., Harden, J. and Singer, M. (eds) 1994, *Factors of Soil Formation: A Fiftieth Anniversary Retrospective*, Soil Science Society of America Inc., Madison, Wisconsin.
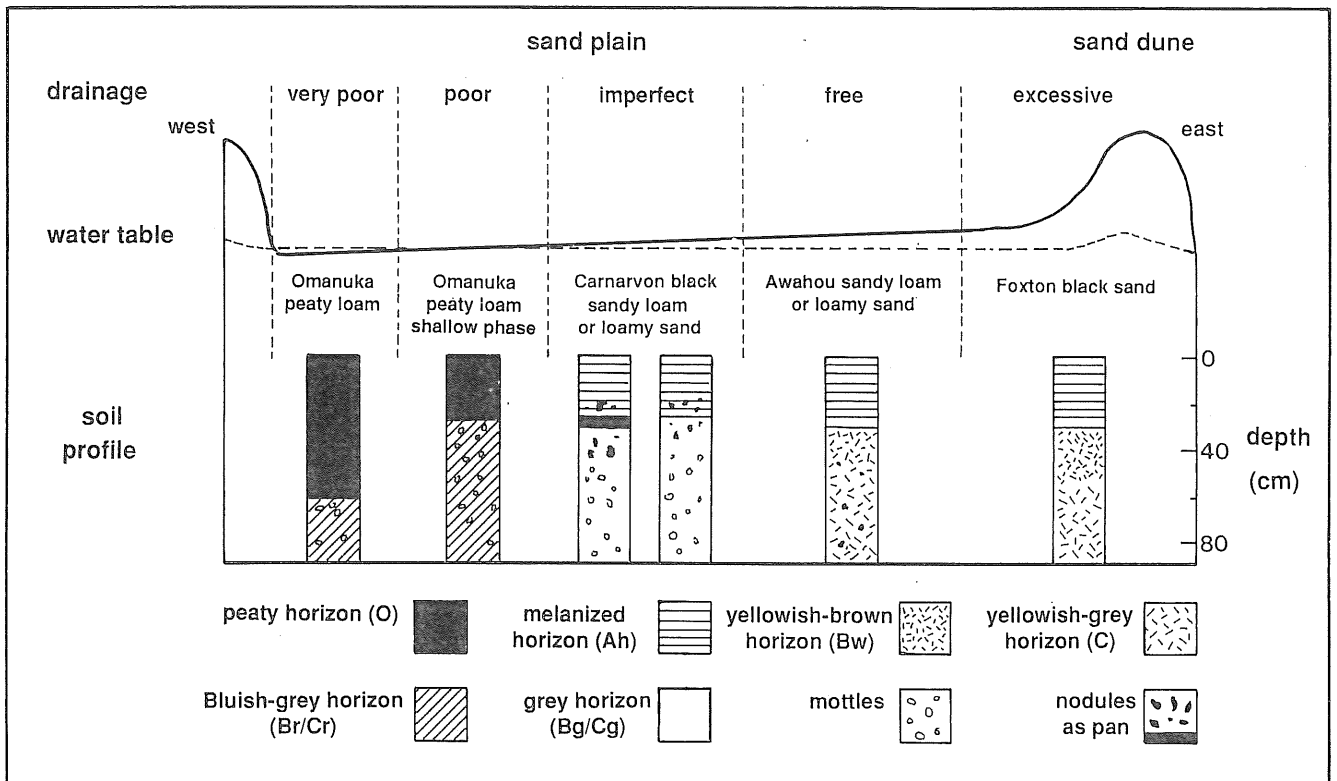
Figure 2.11: Soil profile features in relation to relief and soil moisture status in Manawatu sand country (adapted from Cowie et al., 1967).
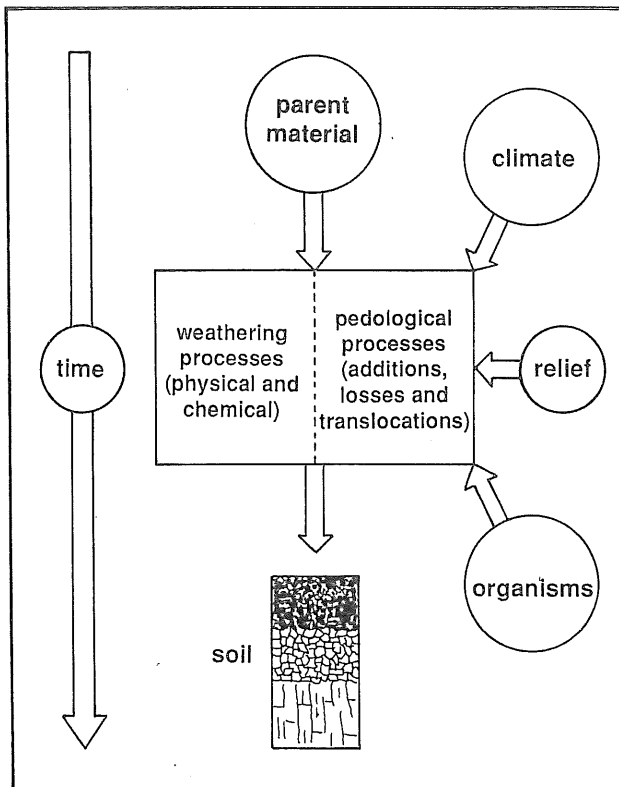

Figure 2.12: Summary of soil formation.

Birkland, P.W. 1984, *Soil and Geomorphology*, Oxford University Press, New York.

Buol, S.W., Hole, F.D. and McCracken, R.J. 1980, *Soil Genesis and Classification*, 2nd edn. The Iowa State University Press, Ames, Iowa.

Duchafour, P. 1982, *Pedology*, trans. by T.R. Paton, Allen & Unwin, London.

Fitzpatrick, E.A. 1980, *Soils: their Formation, Classification and Description*, Longman, London.

Jenny, H. 1988, *The Soil Resource: Origin and Behavior*, Springer-Verlag, New York.

Molloy, L. 1988, *Soils in the New Zealand Landscape: The Living Mantle*, New Zealand Society of Soil Science, Mallinson Rendel Publishers Ltd, Wellington.

Paton, T.R. 1978, *The Formation of Soil Material*, Allen & Unwin, London.