GEOLOGY AND ITS RELATIONSHIP TO EROSION IN THE SOUTHERN RUAHINE RANGE, NORTH ISLAND, NEW ZEALAND

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Volume II

P A R T   T W O

RELATIONSHIPS BETWEEN GEOLOGY AND EROSION

- SOUTHERN RUAHINE RANGE

PART TWO

RELATIONSHIPS BETWEEN GEOLOGY AND EROSION

INTRODUCTION

Previous work on erosion in the southern Ruahine Range has been concerned with defining causes for the increased erosion evident during the last 40 years. Much of this increased erosion has occurred since the introduction of domestic and feral browsing animals into New Zealand. Scientific documentation of natural phenomena that might also be responsible for the increased erosion has involved studies of climatic fluctuations, earthquake activity and natural cycles in composition of the forest cover in the Range. These mechanisms are well substantiated in overseas studies but their role in the southern Ruahine Range remains undetermined.

Previous Ruahine studies have largely been concerned with the incidence of shallow-seated slope movements with very little attention being focussed upon the existence of deep-seated slope movements. The failure of these latter movements occurs at depths beyond the influence of previously cited major causal erosion factors except that of earthquake activity.

It is the aim of this section to examine the types of slope movement present in the study area and to evaluate the role of major causal factors in the genesis of each type of movement. Factors of geologic origin are discussed in detail. Physiographic, climatic and seismic factors, together with factors of human origin that interact with these geologic factors, are also outlined.

A chronological account of episodes of Late Quaternary slope movements and subsequent fluvial aggradation within the Range, based on evidence of terrace remnants, radiocarbon dates and tree ring counts, is presented.

On the basis of known geologic structure and interpretation of aerial photographs spanning a 28 year period, an attempt has been made to: (1) predict sites of potential slope movement; (2) assess the type of movement most likely to occur; and (3) predict possible trends in the pattern of future erosion.

CHAPTER 9

SLOPE MOVEMENT TYPES AND PROCESSES

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CHAPTER 9

SLOPE MOVEMENT TYPES AND PROCESSES

9.0 DEFINITIONS

Slope movements may be classified in many ways, each based on features that recognise, avoid, control or correct the nature of slope movements. Factors that have been used as criteria for identification and classification are type of movement, kind of material, rate of movement, geometry of the failure area and the resulting deposit, age, causes, degree of disruption of the displaced mass, relationship of slide geometry to geologic structure, degree of development, geographic location of type examples, and state of activity. The most widely used classifications are those of Sharpe (1938) and Varnes (1958, 1978). The chief criteria used in Varnes' (1978) classification are: (1) type of slope movement; and (2) type of material. These criteria can be easily discerned in the southern Ruahine Range so Varnes' classification has been adopted. An abbreviated version of this classification is shown in Table 9.1.

TABLE 9.1: Classification of slope movements in the southern Ruahine Range *(after Varnes, 1978)*.

Principal type of slope movement			Principal Type of Material		
			Bedrock	Engineering Soils	
				Debris	Earth
Falls			Rock fall	Debris fall	Earth fall
Topples			Rock topple	Debris topple	Earth topple
Slides	Rotational	Few Units	Rock slump	Debris slump	Earth slump
	Translational	Many Units	Rock block slide	Debris block slide	Earth block slide
			Rock slide	Debris slide	Earth slide
Flows			Rock flow (deep creep)	Debris flow (soil creep)	Earth flow (soil creep)

All four principal types of movement have been recognised in the study area.

Materials are divided into two classes; bedrock and engineering soil. Bedrock designates strata comprising Torlesse Supergroup bedrock that was intact and in its natural place before the initiation of movement. Engineering soil includes a loose, unconsolidated or poorly cemented aggregate of solid particles. Engineering soil is divided into debris and earth. Debris refers to surficial colluvium that directly overlies bedrock as well as to alluvium comprising terrace and fan deposits. Debris contains a significant proportion of coarse material with 20-80% of fragments being greater than 0.02m in size. Earth refers to weakly consolidated marine deposits that include sandstone, siltstone, mudstone, conglomerate and limestone of which about 80% or more of the fragments are smaller than 0.02m in size.

This study was primarily concerned with elucidating relationships between the types of slope movement and causal factors of geologic origin, such as lithology and structure of the materials involved.

The more frequent forms of mass movement have been studied in greatest detail because of: (1) the greater opportunity for study; but more importantly: (2) significant volumes of detritus involved in these movements choke drainage channels and result in major flood control problems. Thus slides and flows are here better documented than falls and topples.

9.1 DESCRIPTIONS

9.1.1 FLOWS

In this study the term 'flow' is used for a fast or slow creep of rock, debris and earth. Creep, as used in mechanics, involves deformation under continual stress. Some creep deformation may be recoverable upon relief of the stress, but generally most of it is not. One of the essential attributes of creep, as defined in geomorphology, is that movement is commonly imperceptible. However, the term may also embrace perceptible movements that immediately precede failure. Slip surfaces within the moving mass are usually not visible or are short lived and the boundary between moving mass and material in places may be a sharp surface of differential movement or a zone of distributed shear. Water is an essential component for flow formation. Flow movements within debris and earth are often more readily recognised than in rocks because these materials tend to behave more like fluids.

A. Debris Avalanches

The term 'avalanche' should refer only to slope movements of snow or ice. The term debris avalanche, however, is well entrenched and designates a variety of very rapid to extremely rapid debris flows (Varnes, 1978). In debris avalanches, progressive failure is rapid and the mass either because it is wet or because it is on a steep slope partially liquefies, flowing and tumbling downslope.

Debris avalanches are the dominant form of slope movement upon steep valley sides of the southern Ruahine Range (Maps 5 and 6). Debris avalanches are characteristically long and narrow scars cut through dense vegetation cover (Fig. 9.1). They may be found at any position on a valley slope but it is more usual for them to originate at a break in slope. Many debris avalanches are found within the outline of large-scale slump movements. The biggest ones result from failure high up near ridge crests and may extend the entire length of a valley side to the stream channel. Often at the head of each debris avalanche is an uphill tapering serrate or V-shaped scar which marks the point of failure (Figs 9.1 and 9.2). Their length is largely dependent upon the position of failure on the slope, slope steepness, and the volume of material involved. The latter two factors determine the momentum of the avalanche which becomes restrained by the forest vegetation through which it flows. The depth to which such avalanches develop is largely dependent upon the thickness of colluvium and the lithological and structural features of the underlying bedrock (see Section 10.1.1). At the toe there is usually a fan-shaped accumulation of coarse blocky detritus, here referred to as scree deposits (Fig. 9.11). Scree deposits are of convex outline in transverse cross-section with the highest point near the longitudinal axis. In the axial region is sometimes a channel that is the site of ephemeral surface runoff from the exposed debris avalanche scar. This channel is bordered by small natural levees. The length of the channel is dependent upon the distance that runoff and transported 'fines' can travel as a high concentration surface flow, before major infiltration into the highly porous scree occurs. A longitudinal profile of a representative scree has a constant steep gradient of 30-40 degrees. The toe slope is usually steeper, being 40-60 degrees, due to the high angle of repose maintained by the coarse, blocky detritus. Once formed, scree deposits remain devoid of vegetation for long periods of time. This is largely due to periodic collapse of the over-steepened margins around the debris avalanche scar (Fig. 9.1).

FIGURE 9.1:

Debris avalanche resulting from failure within bedrock. Height of scarp at head of debris avalanche is 5m. Note (1) the position of the point of failure (near ridge crest), (2) V-shaped scar, (3) blocky scree deposit at toe of slope, and (4) the difference in stage of seral development between tussock covered debris avalanche scars and the surrounding shrubby vegetation. Photograph taken in Rokaiwhana catchment.

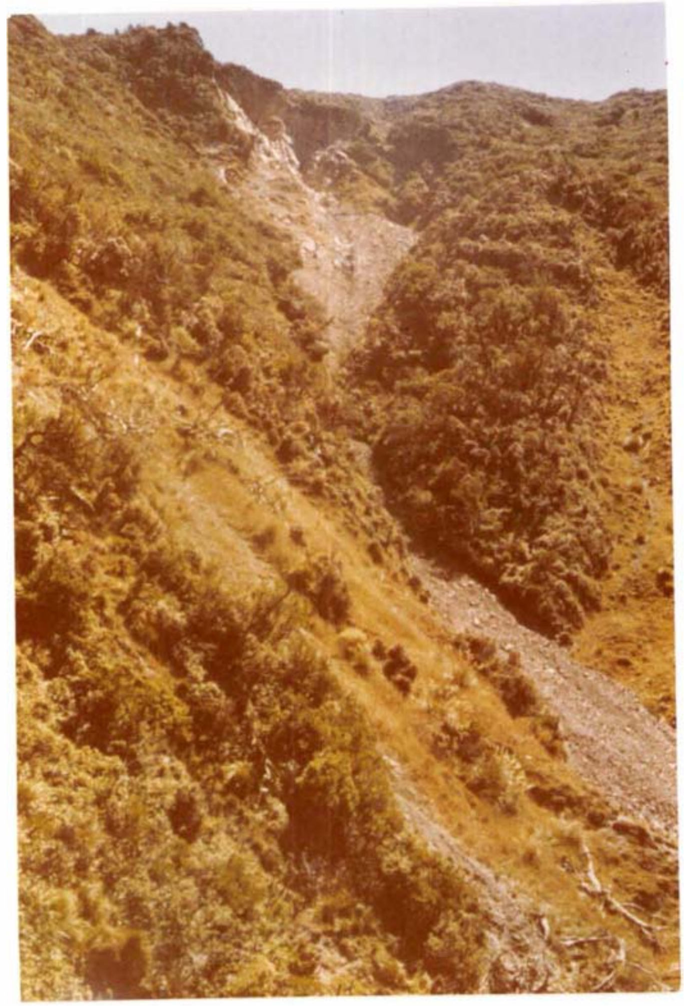

FIGURE 9.2: Debris avalanche scars on steep valley slopes within Mangapuaka catchment. Photograph looks northward across the plane of marine erosion referred to as 'Delaware Ridge'. Photograph taken from track leading up to Maharahara Trigonometrical Station.

B. Debris Flows

Debris flows, called mudflows in some classifications, are here distinguished from the latter on the basis of particle size (Varnes, 1978). The term debris denotes material that contains a relatively high percentage of coarse fragments, whereas the term mudflow is reserved for an earth flow consisting of material that is wet enough to flow rapidly and that contains a high percentage of sand-, silt- and clay-sized particles. Debris flows may originate from any form of slope movement. Once in motion a debris flow, heavily laden with detritus, has transporting power that is disproportionate to its size and as more material is added to the flow its size and power increases (Varnes, 1978). Such flows tend to follow stream channels and are often of such high density, perhaps 60-70% solids by weight, that huge boulders many metres in diameter may be carried along.

In Coppermine Creek at locality 72 (Map 6) a slope movement in 1976 is thought to have begun as a rotational slump with movement occurring along a surface of failure, at depth within the bedrock (Mosley & Blakely, 1977). This movement occurred during a prolonged period of heavy rainfall. The resultant debris filled the valley at the foot of the movement to about 10m depth and then, behaving like wet concrete in a chute, moved as a debris flow for about 600m along the swollen stream channel. The surface of the debris flow was initially impossible to walk on because it remained in a near fluid state for several days. Near its source the flow surface was hummocky but in a downstream direction it became planar. Throughout its length material comprising the flow appeared to be thoroughly mixed, but some of the coarser fragments were heaped along the sides to form natural levees. While this debris flow appears to have moved as a single event, other debris flows may move as a series of surges (Pierson, 1980). Such surges are caused by periodic mobilisation of material in the source area or by periodic damming and release of debris.

In the southern Ruahine Range, because of the narrow and deeply dissected nature of stream channels, much evidence of debris flow activity is removed by stream erosion a short time after. Only the largest of such events produce levees that become stabilised and protected by vegetation to remain as evidence. During historical times debris flows within the southern Ruahine Range have only developed subsequent to large-scale slump movements and not from any other forms of slope movement. Fifty six slump and slump-like features have been documented within the Range (Appendix VI) of which the majority show no sign of subsequent debris flow development. This is probably due to much of the moved mass remaining on the valley slope rather than the valley slope collapsing entirely. Subsequent movements at these localities are likely to result in debris flow activity in the future (see Chapter 12).

C. Earth Flows

Earth flows, involving failure of massive sedimentary deposits of Plio-Pleistocene age, are located on pastured farmland flanking the western side of the Range. They are, in general, small in extent and are thus of minor importance within the study area. A detailed descriptive account is therefore not included. In the past, however, earth flow activity may have been an important landslide process on the flanks of the Range. Earth flows could explain the complete removal of material from many of the chute-like landslide scars at localities 76 and 78-80 (Maps 5 and 6 and Appendix VI).

D. Rock Flows

Rock flows involve deformations that are distributed among many large or small fractures, or even unconnected microfractures, without concentration of displacement along a major fracture. The movements are often extremely slow and are apparently more or less steady in time, although few data are available (Varnes, 1978). Flow movements may result in folding, bending, bulging or other signs of plastic behaviour. These kinds of movements are being recognised world wide in areas of high relief. They are quite varied in character and several kinds have been described, such as gravitational slope deformation (Nemcok, 1972), gravitational faulting (Beck, 1967) and ridge-top depressions, (Tabor, 1971).

A number of geomorphic features described in this study are thought to be the result of plastic deformation of greywacke bedrock under the influence of gravity. They are here referred to as ridge-top depressions, benches and scarps (Appendix VI). Some of these features are closely associated with large-scale slump movements, usually on steep upper catchment slopes near ridge-crests. They are distinguished from similar features associated directly with faulting because they are mass movement features. Scarps of tectonic origin are relatively common in the study area (see Chapter 6 and Map 4).

Many flows in bedrock occur near ridge crests, either parallel or oblique to the axis of the ridge, at altitudes between 1000m and 1300m. At these elevations details of ridge-top features are masked by dense, shrubby leatherwood vegetation of podocarp-hardwood forest. Fewer are located at lower elevations (500m) within pastured foothills adjacent to the Range.

1. Ridge-top Benches

Ridge-top benches range from 100-250m in length with an average width of 10-30m. Each surface is separated from the next by a steep, curved, or straight scarp up to 10m high. The bench surfaces are planar and may be horizontal or slightly inclined downslope. Examples occur at localities 18 and 19 on Maps 5 and 6.

2. Ridge-top Depressions

Ridge-top depressions (Tabor, 1971) consist of a shallow swale or U-shaped trough that may either parallel the Range crest or subsidiary ridge crests, or strike divergently across valley slopes and ridges. The depression is usually less than 3m deep and averages 5m in width. Continuous depressions range from 100-200m in length. Often a series of sub-parallel depressions arranged in en-echelon fashion extend up to 500m in length. Where depressions cross ridges they form low 'cols' analogous to those created by faults. Examples can be seen at localities 2c and 41-44 on Maps 5 and 6.

3. Ridge-top Scarps

Ridge-top scarps (Beck, 1967) may be linear, arcuate, sharp in outline with steep fronts, or subdued in outline due to erosion. Scarps are generally 2-10m high and vary in length from 150-300m. The slope of scarp faces is usually greater than 50^{o}. Examples occur at localities 2b, 9 and 10 on Maps 5 and 6.

All of the ridge-top features described can grade into one another along their length but most commonly depressions and scarps occur together. At four localities in the study area ridge-top features can be shown to be of gravitational origin through association with obvious mass movement features (predominantly slumps). However, ridge-top features at a further 20 localities are of obscure origin. The latter are here interpreted as being the result of precursory movements of gravitational origin, that have occurred at depth, and may indicate potential sites of future large-scale mass movement (see Chapter 12).

9.1.2 SLIDES

In a slide, shear results in displacement along one or several surfaces. The shear failure may be instantaneous or it may be progressive. In progressive failure, movement propagates along a defined surface of rupture from an area of local failure. The displaced mass may slide beyond the original surface of rupture onto the original ground surface, which then becomes a surface of separation. In the present classification (Varnes, 1978) emphasis is placed on a distinction between translational and rotational slides. In translational sliding the mass progresses outward and/or downward along a more or less planar or gently undulatory surface and has little of the rotational movement or backward tilting characteristic of slumps. A translational slide may continue to enlarge if the surface on which it rests is sufficiently inclined and the resistance to shear along this surface remains lower than the shear stress. A translational slide in which the moving mass consists of a single unit or a few closely related units that are not greatly deformed may be called a block slide. If the moving mass consists of many semi-independent units, it is termed a broken or disrupted slide. The movement of translational slides is commonly structurally controlled by surfaces of weakness, such as faults, joints, bedding planes, lithological discontinuities, or by the contact between firm bedrock and overlying detritus (see Section 10.1.1). In rotational sliding the mass (here referred to as a slump) moves along a surface of rupture that is concave upwards (Fig. 9.3). The movement takes place only along internal slip surfaces and is more or less rotational about an axis that is parallel to the slope. Cracks exposed at the surface are concentric in plan and concave toward the direction of movement (Fig. 9.3). According to Varnes (1978) the development of classic, purely rotational slumps on a surface of smooth curvature within bedrock are relatively uncommon. In the study area, however, gravitational failures exhibiting rotational movement, though subordinate in number to other forms of failure, are relatively abundant (see Chapter 10).

A. Translational Slides

1. Debris Slides

This term is used here to describe slides that involve slope detritus, soil, peat and vegetation and occasionally the surficial weathered layers of bedrock. In the study area there is complete gradation from debris slides to debris flows to debris avalanches as deformation and disintegration increase. Increased disintegration is the result of more rapid movement because of lower cohesion, higher water content, steeper slopes and suitable lithologies (see Section 10.1.1).

Debris slides appear to be randomly distributed throughout the mapped area and can originate at any site on a valley slope. They commonly occur along steep-sided stream banks, gully edges, around headwall and lateral scarps and along the toe region of large-scale slump movements (Maps 5 and 6). Most debris slides are initially relatively small in areal extent when compared to other forms of slope movement. However, once sliding has begun the scar tends to increase in area both laterally and upslope as a result of further collapses around the over-steepened margins. Eventually several discrete debris slides may coalesce to form very extensive areas of bare ground.

Where vertical displacement by sliding is small, the moved mass remains relatively undisrupted and the vegetation cover is undisturbed. However, where vertical displacement is large the moved mass disintegrates and accumulates as a scree deposit below the debris slide scar. These scree accumulations may be found at the base of valley slopes, adjacent to gully or stream channels. In such localities scree accumulation is temporary because fluvial erosion soon removes it. Elsewhere, on bush covered slopes, debris slides often appear small in size because the scree accumulates beneath the forest cover and is largely hidden. Debris slide scars and scree deposits are subject to the processes of surface runoff including sheetwash, rilling and gullying (see Section 10.3). Debris sliding is the second most abundant type of slope movement found within the southern Ruahine Range.

2. Rock Slides

Rock slides consist of coherent blocks which fail along planar surfaces of weakness within the bedrock (see Section 10.1.1). Failure usually occurs at between 1-3m depth. With distance travelled a rock slide disintegrates to become a disrupted rock slide. These can develop into debris avalanches comprised of disintegrated bedrock, colluvium and vegetation.

3. Earth Slides

Earth slides involving marine sedimentary deposits of Plio-Pleistocene age, at localities 76, 78, 79 and 80 (Maps 5 and 6) suggest that large-scale displacement has taken place along a translational surface. At each of these localities earth has been totally stripped from this area occupied by the slide to expose a smooth, basal plane. It is likely that these translational surfaces correspond with an inferred very low angle of dip of the poorly exposed Plio-Pleistocene strata at these localities. Very high and steep headwall and lateral scarp faces prove that considerable volumes of material have been displaced. During the latter stages of displacement much of the material may have formed flows of considerable magnitude. Earth slides are an uncommon type of slope failure in the study area (Appendix VI).

B. Rotational Slides (Slumps)

1. Earth and Rock Slumps

Earth and rock slumps can be identified readily from surface features. Maps 5 and 6 in combination with Appendix VI give details of 93 slumps and 24 slump-related features from 109 localities throughout the mapped area. Of these, 56 examples have been formed in bedrock lithologies within the Range and 37 have formed in the Plio-Pleistocene aged deposits flanking the Range. Irrespective of the nature of the underlying material, most of these slumps have many of the following features in common.

In the vicinity of the headwall, vertical movement may be either slight with little apparent rotation (localities 5b, 11 and 36) or substantial with obvious backward rotation (localities 61 and 62). In the latter cases rotation has resulted in the formation of deep depressions between the slumped mass and the headwall (Fig. 9.3).

With two exceptions, ponding of surface runoff does not occur on earth and rock slumps in the mapped area. One exception at locality 2b (Maps 5 and 6) is a rock slump where ponding has occurred at the base of an east-facing scarp of suspected gravitational origin. The other exception is associated with an earth slump at locality 77. This appears to be natural ponding that has been modified by man for water storage purposes.

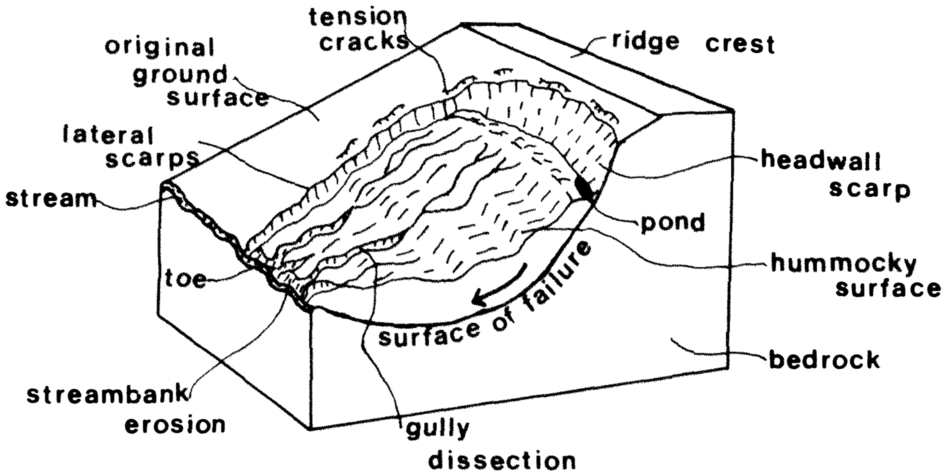

FIGURE 9.3: A schematic block diagram showing the characteristic features of a typical slump in the southern Ruahine Range.

Seeps, springs and marshy conditions commonly mark the foot and toe regions of most slumps. The strata in the toe region are generally very sheared and soft. Consequently, it is susceptible to gully incision by surface runoff and fluvial erosion along stream banks (localities 2, 53 and 61). The instability of these toe slopes, especially in rock slumps, is shown by numerous secondary debris avalanches and debris slides. Minor scarps often develop on the displaced slump mass as a result of differential movement of the sliding mass (localities 2 and 24). Other scarps may develop to the side or above the recognisable outline of the slump (localities 1, 5a and 5b). Such scarp development has taken place at localities within the Range in historical times and appears to have formed between the years 1946 to 1978 (see Chapter 12).

9.1.3 FALLS

In a fall, a mass of any size is detached from a steep slope or cliff along a surface on which little or no shear displacement takes place and descent is mostly through the air by freefall, leaping, bounding or rolling. Movements are very rapid to extremely rapid and may or may not be preceded by minor movements leading to progressive separation of the mass from its source.

A. Rock Falls

Rock falls in the study area originate from the near vertical bluffs of either massive sandstone or bedded bedrock strata. In the latter instance a rock fall may begin with a translational slide component along a bedding plane prior to becoming a fall. Many of the observed rock falls are of very limited areal extent and rarely involve more than a few tonnes of bedrock.

The parent cliff of a rock fall is often marked by an irregular scar that lacks the crescentic shape characteristic of a slump. Instead the irregularity of its surface is controlled largely by joints, bedding planes and faults in the parent material (see Section 10.1.1). Rock fall scars are slow to establish a vegetation cover. Most scars observed support patchy growths of weeds, tussock and occasional shrubby species usually growing from discontinuities in the solid rock. Rock fall activity in many cases appears to have ceased but the scars remain fresh as a result of continuing spalling. Most rock fall locations occur adjacent to waterways. Thus detritus from rock fall activity often falls into stream channels to accumulate at the base of the scar as an irregular pile of angular rock detritus ranging in size from a few centimetres to five metres in diameter (*e.g.* at T23/598184). Accumulation of rock fall detritus has been of sufficient volume to partially infill an ephemeral gully at T23/593091 (Maps 5 and 6). Elsewhere stream activity is generally effective in removing all but the largest detritus.

B. Debris Falls

Debris falls are found in situations where steep banks have been carved into thick accumulations of alluvial or colluvial detritus. Debris falls are generally the result of bank undercutting by stream activity (*e.g.* at T23/697190 and T23/713221; Maps 5 and 6). They are best observed in those catchments where abundant alluvial bedload has formed high terrace and alluvial fan accumulations. The vertical fall component ranges up to 10m in height. In the study area debris falls are very small in extent and rarely involve more than 10 tonnes of material. Although the debris is unconsolidated, the fallen mass may remain intact where the fall component is small (2-5m height). Accumulations of debris fall material are rarely preserved because many falls occur during periods of flooding and are removed by successive periods of high water.

C. Earth Falls

Earth falls in the study area occur in sandstone, siltstone, mudstone, conglomerate and minor limestone strata of Plio-Pleistocene age. They are restricted to the near vertical bluffs along stream channels that have been deeply incised within the marine strata from the western base of the Range westward to the Pohangina River. Bluffs vary considerably in height but on average are 30-50m high where earth fall activity has occurred. Many earth falls appear to be the result of bank undercutting by streams.

Earth fall detritus rarely accumulates for any length of time because it is readily dispersed and transported by fluvial processes. However, the presence of large blocks of material in stream beds is an indication that earth falls or topples have occurred. Hanging fence lines and gaps in shelter belts along bluff edges also indicate recent earth fall activity. Surface cracks along many bluff edges indicate sites of potential earth falls.

9.1.4 TOPPLES

Topples consist of a forward rotation of a unit or units about some pivot point, below or low in the unit, under the action of gravity and forces exerted by adjacent units or by fluids in cracks. It is tilting before collapse (Varnes, 1978).

A. Rock Topples

Rock topples are to be found only at the base of valley slopes where bed-rock exposures border stream channels. Examples of rock topples occur where large masses of bedrock are detached from valley walls and are separated by a gap 0.1 - 1m wide. Often the planar-sided detached blocks and parent wall surfaces are fresh in appearance and devoid of vegetation indicating relatively recent movement. Separation surfaces vary in attitude from vertical to very steeply inclined and may in places correspond with joints or faults. As the component of topple is difficult to substantiate and as so few examples could be located in this area, rock topples are not documented in detail in this study.

B. Earth Topples

Earth topples often occur in association with earth falls along vertical bluffs in marine strata of Plio-Pleistocene age bordering westerly draining streams. Earth topples are also associated with the headwall, lateral margins and the toe region of major slumps, for example at T23/553146 (Maps 5 and 6). Here the strata are often carved into blocks that are separated by transverse open tension cracks. These blocks are often re-ferred to as slump-topples (Varnes, 1978).

CHAPTER 10

FACTORS THAT INFLUENCE SLOPE STABILITY

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CHAPTER 10

FACTORS THAT INFLUENCE SLOPE STABILITY

10.0 INTRODUCTION

Slope movement can seldom be attributed to a single definitive cause. Overall terrain must therefore be analysed in detail so that individual but interrelated factors are identified before areas of potential slope movement can be recognised. These factors may be broadly grouped into: (1) geologic; (2) physiographic; (3) climatic; (4) seismic; and (5) human origin. This chapter outlines how these factors affect slope stability in the study area.

Factors of geologic origin leading to failure are lithology, structure and mineralogy. This includes physical hardness of the constituent materials and their susceptibility to weathering, structural attitude of strata and presence of structural discontinuities such as joints, bedding planes, faults and folds. Many of the details of geologic origin are described in Part One.

The physiographic factors centre on the southern Ruahine Range being between the youthful and mature stages of development in the common fluvial geomorphic cycle. Most valleys are V-shaped in cross-section, the majority of drainage divides are sharp and the drainage system is well developed. Physiographic features of such a terrain important to slope movement include: angle of slope; length of slope; altitude and aspect.

Climatic factors greatly affect the development of landforms and slope movements. Variations in microclimate, such as differences in altitude, exposure to moisture-bearing winds and exposure to sunlight create different geomorphic processes. Numerous investigations have demonstrated that the most active periods of slope movement occur during times of heaviest precipitation (James, 1973; Stephens, 1975; Mosley, 1977). Undercutting by stream action and the erosion of slopes by alternate wetting and drying and frost action are also well known factors related to climate that result in slope movement.

Slope movements may also be triggered by fault rupture or seismic shaking.

Factors of human origin have been largely through settlement practices such as felling and burning of native forest and the introduction of domestic and feral animals into the remaining forests. A resultant change in the forest structure and general deterioration of the forest cover have contributed to an increase in slope movements during the time of human occupation.

10.1 GEOLOGIC FACTORS INFLUENCING SLOPE STABILITY

10.1.1 LITHOLOGICAL AND STRUCTURAL FACTORS

It is here intended to group together those slope movement types that are thought to be controlled by similar structural and/or lithological discontinuities during their initial stages of movement in order to emphasise the influence such discontinuities have upon the stability of materials. For example, debris slides and debris avalanches are generally the result of slope failures that are initiated in a similar manner but are classified as different slope movement types in their latter stages as deformation and disintegration increases in debris avalanches. The grouping and order in which slope movements are discussed in this chapter differs from that in Chapter 9. Firstly, slides and flows that are initiated by translational movement along lithological discontinuities are discussed together. Such discontinuities include the contact between bedrock and colluvium and bedding planes within bedded sequences of Torlesse bedrock. Secondly, falls and topples are described together. These result from failure along discontinuities, largely of structural origin, such as fault and joint planes. Some of these failures are preceded by translational sliding. Thirdly, earth and rock slumps are described. The surface of failure along which earth and rock slump movements take place is largely unknown but in some cases is thought to coincide with a fault plane. In other cases a combination of both lithological and structural discontinuities are thought to control the type, amount and size of these movements. Still other earth and rock slumps form independently of any known discontinuity.

The relationship between the occurrence of slumps and major fault zones is discussed in detail later in this chapter.

Fourthly, there are a number of ridge-top features of large size, where failure is thought to be of similar origin to that of slumps. However, they lack an ovate outline and rotational sliding cannot be demonstrated, so they are described separately.

A. Shallow Translational Slides and Flows (debris avalanche, debris slide, rock slide and earth slide)

Within the study area each of these slope movements is initiated by translational sliding at shallow depths. Movement begins as a result of failure at some discrete surface or zone of shear that is commonly of lithological origin but may have been subjected to subsequent tectonic disruption. Once sliding begins the moved mass becomes progressively more deformed and disintegrated. If the vertical displacement is small, the moved mass remains relatively undisturbed and the movement is referred to as a slide. On the other hand, if the vertical displacement is large and the moved mass disintegrates, the process becomes one of flowage (see Chapter 9). Debris slides and rock slides generally remain translational slides throughout their development. However, debris avalanches and some earth slides become flows in their latter stages of development.

There are exceptions, where gradations between debris slides and debris avalanches become particularly difficult to classify. In general, shallow movements that involve small volumes of predominantly soil and colluvium develop as slides. In contrast, shallow movements that involve large volumes of material begin as debris slides and develop into debris avalanches with characteristic V-shaped scars of considerable depth (Figs 9.1 and 9.2). Intergrades between debris avalanche and debris slide movements show complete disintegration of the moved mass but lack the scar shape and depth characteristics of other debris avalanches in the study area. These movements are grouped with debris slides. On the basis of materials involved in these movements, slides and flows can be grouped into two forms. The first includes movements that predominantly comprise soil and colluvium. Translational sliding in these instances results from failure either within the soil or colluvium or at the soil- or colluvium-bedrock contact. The second includes movements that predominantly comprise bedrock. Translational sliding in these instances results from failure either along a zone of shear within the uppermost weathered horizon or along internal bedding plane surfaces at depth.

Slides and flows comprising earth materials occur along the base of the Range. Here translational sliding is thought to result from failure along low dipping bedding plane surfaces. These are discussed in turn.

1. Movements Comprising Soil and Colluvium

These materials are formerly derived from the underlying bedrock. Colluvium is coarse with large grain size variation, is shallow and skeletal, has loose and unconsolidated structure, is porous and consequently is inherently weak. Colluvium thickness varies depending upon the type of underlying bedrock present. In general, colluvium appears to be thinner upon slopes underlain by Wharite Lithotype rather than the Tamaki Lithotype. Chemical weathering and physical disintegration of the regularly bedded lithologies comprising the Tamaki Lithotype, facilitated by the numerous bedding plane and joint surfaces, appears to occur more rapidly and to greater depths than upon the predominantly foliated lithologies comprising the Wharite Lithotype. Colluvium thickness also varies with position on slope being thickest near the base of slopes and thinnest near ridge crests. In general an average of 1-2m of soil and colluvium could be expected on upper valley slopes and between 2-3m on most lower slopes, apart from those subjected to active gully erosion or streambank undercutting.

Slope movement types involving soil and colluvium are debris avalanches and debris slides. These movements vary considerably in depth, size and shape. There is no obvious relationship between variations in the physical appearance of these slope movements and: (1) their position on valley slopes; (2) the depth of colluvium; and (3) the underlying lithologies.

The occurrence of debris avalanches and debris slides particularly at high elevations may be related to the presence of an iron pan in the soil. Hubbard (1978) suggests that a perched water table may develop at the junction between the Ah (or Ha) and Bg (or Br) horizons, within Takapari hill soils, as a result of the development of an iron pan at the base of the Bg (or Br) horizon. Where the soil is shallow, the iron pan offers resistance to both downward movement of water and to root penetration. The perched water table is also adverse to root growth. With a marked decrease in root abundance there is a consequent decrease in the shear strength of these shallow soils and hence slope movement would be facilitated. In addition water movement along the perched water table may have a lubricating effect so that the iron pan becomes a surface of weakness along which sliding may be initiated.

At lower elevations debris avalanches and debris slides are initiated as a result of translational sliding along a zone of shear that is irregular in outline and difficult to trace. Most of these slope movements can be attributed in part to the structural weakness of the soil (Hubbard, 1978) and in part to the lubricating effect of emergent ground water.

Irrespective of their position on valley slopes the majority of debris avalanches and debris slides resulting from failure within soil or colluvium are of very shallow depth. Whilst some are less than 0.5m deep (Figs 10.1 and 10.2) the majority tend to be up to 1m deep.

Approximately 20% of all debris slide and 5% of all debris avalanche movements in the southern Ruahine Range result from failure within soil or colluvium (Table 10.1).

Slides and flows that predominantly comprise soil and colluvium also result from translational sliding at the smoothed and often even-surfaced contact between underlying greywacke bedrock and overlying soil or colluvium (Fig. 10.3). In the majority of these slides translational sliding has occurred where the slope direction of this contact was concordant with discontinuities in the uppermost layers of bedrock (Figs 10.4 and 10.5; diagrams A and B). In these cases the discontinuity was always a lithological contact. Where concordant slides occurred in areas of Wharite Lithotype, the contact coincided with exposed faces of massive sandstone. Where concordant slides occurred in areas of Tamaki Lithotype the contact coincided with the dip direction of the uppermost bedding plane surface.

In the majority of cases observed, translational sliding occurs where the slope direction of the soil- or colluvium-bedrock contact was discordant to lithological or structural discontinuities in the underlying bedrock (Figs 10.4 and 10.5; diagrams C, D and E).

As these slides are ubiquitous it is apparent that lithological and/or structural discontinuities within the underlying bedrock do not influence the formation of translational slides at this contact.

Translational sliding at the soil- or colluvium-bedrock contact is frequently coincidental with areas of emergent ground-water. Close investigation indicated that some of this water was derived from within the bedrock and emerged as seeps through joints. Another source appeared to come from higher slopes where water percolated through porous soil and colluvium to emerge at the head of slips (Figs 10.4 and 10.5) at this contact. Whilst the presence of water at these localities is not

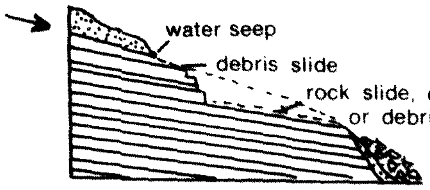
FIGURE 10.1: Extensive debris slide development within South Oruakeretaki catchment. Note the poor condition of the forest vegetation and the large number of dead trees on the surface above the debris slide.

FIGURE 10.2: Close-up of debris slide scar within Mangapuaka catchment. Note: (1) the shallow depth of failure (less than 0.5m) within colluvial slope material; (2) the dense shrubby vegetation (background) adjacent to where the movement occurred; and (3) the shallow root systems of the vegetation on the slide (foreground).

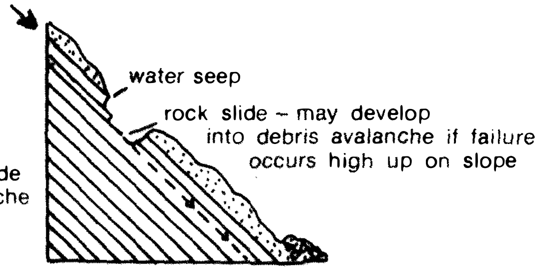
TABLE 10.1: Percentage and type of shallow translational slope movement in the southern Ruahine Range resulting from failure at varying depths. *(Figures are based on field observation and photographic interpretation of approximately 5650 erosion* scars - see Maps 5 and 6).

Position of origin of slope failure	Type of shallow translational slope movement		
	Debris slide (n = 2500)	Debris avalanche (n = 3050)	Rock slide (n < 100)
Within soil and/or colluvium (at less than 1m depth below ground surface)	20	5	-
At bedrock contact (1-3m below ground surface)	70	35	-
At < 1m depth below colluvium - bedrock contact (within bedrock material)	10	40	50
At > 1m depth below colluvium - bedrock contact (within bedrock material)	-	20	50

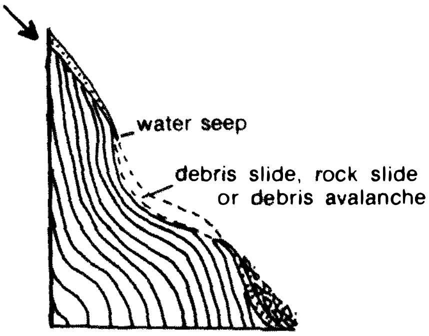
FIGURE 10.3: Torlesse bedrock overlain and in sharp contact with peaty loam. On inclined valley slopes the lubricating affect of water along this contact facilitates slope movement. Debris slides or debris avalanches may occur as a result of failure at this contact.



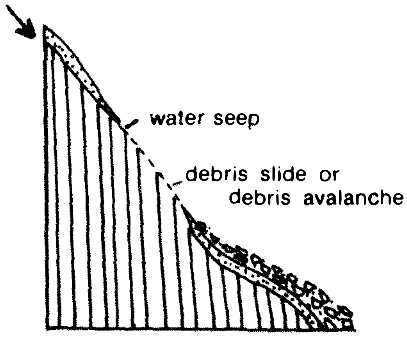
(A) Dip of strata less than steepness of free face.



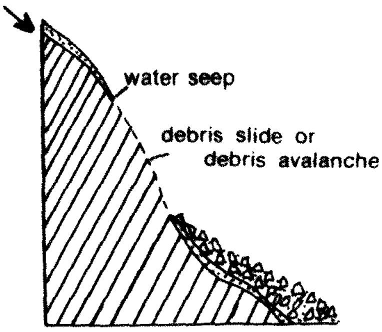
(B) Dip of strata parallels and is of equal steepness to that of the free face.



(C) Flexure in bedrock strata increases or reduces angle of dip to that greater than, equal to or less than steepness of the free face.



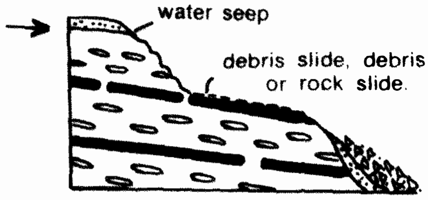
(D) Dip of strata greater than steepness of free face.



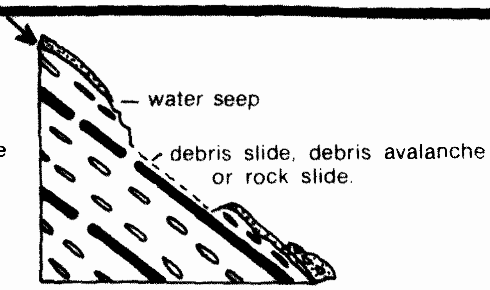
(E) Dip of strata perpendicular to the free face.

LEGEND	
---	Surface of failure.
	Soil and colluvium.
	Landslide debris.
	Graded-bedded Lithozone.
	Soil- and colluvium-bedrock contact.

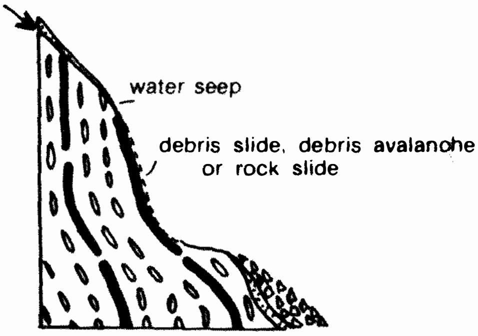
FIGURE 10.4: Variations in the structural attitude of strata with respect to slope angle and direction, that largely determine the type of slope failure likely to occur in catchments underlain by the Tamaki Lithotype. Diagrams A, B and C illustrate the favourable attitudes of bedded strata conducive to debris avalanche and rock slide. Diagrams D and E illustrate bedrock attitudes less favourable for these movements. Both debris slides and debris avalanches occur where the strike of the strata is perpendicular to the free face (see Fig. 10.3). They most commonly result from failure at the colluvium-bedrock contact irrespective of the structural attitude or the lithological composition of the underlying bedrock.



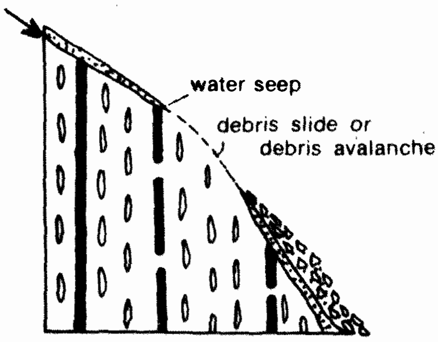
(A) Dip of foliation less than steepness of free face.



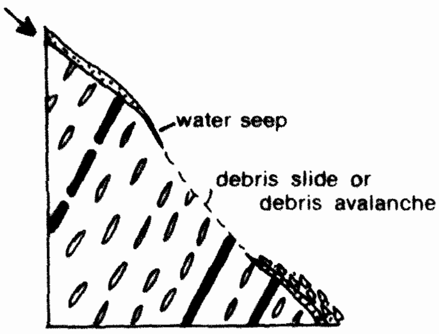
(B) Dip of foliation parallels and is of equal steepness to that of the free face.



(C) Flexure in bedrock strata increases or reduces angle of dip to that greater than, equal to or less than steepness of the free face.



(D) Dip of foliation greater than steepness of free face.



(E) Dip of foliation perpendicular to the free face.

LEGEND
Surface of failure.
Soil and colluvium.
Landslide debris.
Foliated or Diamictite Lithozones.
Sandstone, Chert or Argillite Lithozones.
Soil- and Colluvium-bedrock contact.

FIGURE 10.5: Variations in the structural attitude of foliated bedrock with respect to slope angle and direction, that largely determine the type of slope failure likely to occur in catchments underlain by the Wharite Lithotype. Diagrams A, B and C illustrate the favourable structural attitudes and bedrock compositions most frequently associated with debris avalanche and rock slide movements resulting from failure within the bedrock. Diagrams D and E illustrate bedrock attitudes less favourable for these movements. Both debris slides and debris avalanches occur where the strike of the strata is perpendicular to the free face. They most commonly result from failure at the colluvium-bedrock contact irrespective of the structural attitude or lithological composition of the bedrock.

necessarily a prerequisite for slide movements, it is here considered that the lubricating effect of water along this contact greatly enhances their occurrence.

The physical contrast in materials at this contact in conjunction with a smoothly eroded bedrock surface facilitates the sliding of overlying materials. Such unconsolidated materials are in a delicate balance on most slopes within the study area.

All failures at the soil- or colluvium-bedrock contact result in debris slide and debris avalanche activity. Overall the greatest number of slide and flow movements within the southern Ruahine range are the result of failure at this contact. These movements account for approximately 70% of all debris slides and 35% of all debris avalanches in the study area (Table 10.1). This contact is therefore the single most important discontinuity with which slope movement can be directly related.

2. Movements Comprising Bedrock

These movements result from failure along discontinuities within the bedrock which most commonly coincide with bedding planes, lithological contacts where bedding is absent and less frequently with joints and low angle thrust fault planes. The depth at which translational sliding occurs within the bedrock often determines the resultant slope movement type. Sliding was noted to occur at two distinct depths:

(1) Some slides and flows only involve the uppermost layers of weathered bedrock to a depth of 1m. Here weathering has been essentially concentrated along joint, cleavage and bedding plane surfaces to loosen and fragment the *in situ* bedrock. Failure within such material occurs along a zone of shear that is irregular rather than planar in outline. Where downslope movement is small, the moved mass remains relatively undeformed to form either a debris slide or an undisrupted rock slide (Figs 10.6 and 10.7) but where movement is large the moved mass completely disintegrates to form either a debris avalanche or disrupted rock slide.

    There is no apparent correlation between the occurrence of these slides and the underlying bedrock lithologies. Slides can occur where the direction of sliding is either concordant or discordant with the structural attitude of discontinuities within the bedrock.

FIGURE 10.6: A rock slide resulting from downslope movement along a bedding plane surface within bedrock comprising the Tamaki Lithotype. The direction of bedding plane dip coincides with topographic slope. Strike of the strata parallels valley slope. Downslope displacement of the moved mass is small hence the rock is relatively undisrupted and the vegetation cover has not been disturbed. *Locality T23/591083.*

FIGURE 10.7:

An undisrupted rock slide resulting from downslope movement along a lithological contact within bedrock comprising the Wharite Lithotype. Downslope displacement is small. Note the water seep on the exposed sandstone surface.

Locality T23/657251.

Approximately 10% of all debris slides, 40% of all debris avalanches and 50% of all rock slides (Table 10.1) result from translational sliding within the bedrock at a depth of less than 1m.

(2) Other slides and flows result from translational sliding within the unweathered bedrock at depths greater than 1m. Irrespective of the depth of failure the surfaces along which movement occurs are bedding planes. Exposed slide surfaces are step-like in outline, indicating that sliding occurred along several bedding plane surfaces at differing levels both down the length of, and across, the slide scar. These slides, by virtue of the volumes of material involved, are large in size and form deep erosion scars. Approximately 20% of all debris avalanches and 50% of all rock slides result from translational sliding within the bedrock at depths greater than 1m (Table 10.1).

The scale to which these translational slides can develop within bedrock lithologies in the study area is largely governed by the lithologies present, their internal organisation and the attitude of the strata with respect to valley slopes.

Different rock types have inherently different weaknesses and strengths as a result of their origin and composition. Of the sedimentary bedrock lithologies, argillites are in general the most susceptible to failure, especially when interbedded with permeable sandstone. When a slope is underlain by a rock unit consisting of several alternating and interbedded rock types such as sandstone and argillite their combined mechanical strength differs considerably from that of the constituent interbeds themselves. The mechanical strength of a rock unit consisting of interbedded lithologies can only be as strong as its weakest member. Argillite and sandstone are the commonest interbedded lithologies where argillite is less competent than sandstone and on exposure breaks down more readily. Failure by translational sliding along any of the multitude of bedding plane surfaces is possible but appears to occur more frequently where the depositional contacts are sharp rather than graded, *i.e.* where coherence is weakest. The sharpest depositional contact occurs between the T_e interval of one Bouma sequence and the T_a interval of the succeeding Bouma sequence (see Chapter 3). This contact separates two lithologies of greatly differing composition and mechanical strength, that is, argillite and sandstone. In addition this contact is often tectonically disrupted. Since the strata are largely overturned (see Chapter 2), most newly exposed slide surfaces are basal bedding planes of sandstone composition, indicating that failure occurs either along the sandstone-argillite contact or within argillite beds.

Two types of translational slide occur at depth within bedrock. Rock slides are more frequently associated with bedded strata of the Tamaki Lithotype and less frequently with the foliated lithologies comprising the Wharite Lithotype. On the other hand, the distribution of debris avalanches is more widespread and appears to be associated equally with catchments underlain by either lithotype. This suggests that rock slides are controlled largely by bedding plane surfaces whereas debris avalanche activity is controlled to a greater extent by other discontinuities, such as joints, within the bedrock.

Within the Tamaki Lithotype the majority of rock slides and debris avalanches occur in association with sequences of regularly bedded strata that comprise the Thin-Bedded Association (see Chapter 2). This is largely a function of the widespread occurrence of this Association throughout the Tamaki Lithotype and also because this Association usually underlies the steepest of valley slopes. Elsewhere throughout the Tamaki Lithotype, slopes underlain by regularly bedded strata comprising the Very Thin-Bedded, Thick-Bedded and Very Thick-Bedded Associations of identical structure and similar lithologies but in differing proportions display very little of the translational slide activity that is characteristic of slopes underlain by strata of the Thin-Bedded Association. In the case of the Very Thin-Bedded Association, the absence of slide activity is largely a factor of location where relief is low and stream dissection is absent, such as is found along Delaware Ridge. The paucity of rock slide and debris avalanche activity associated with the Thick-Bedded and Very Thick-Bedded Associations, which comprise thick units of sandstone with few thin interbeds of argillite, may be due to fewer slide surfaces and therefore greater overall mechanical strength. However, this cannot be substantiated in the mapped area as these Associations have not been found in localities where favourable structural relations with valley slopes are conducive to translational sliding. As the bedrock comprising Wharite Lithotype is essentially non-bedded, the incidence of rock sliding and debris avalanching by failure along bedding plane surfaces is rare. However, one plane of weakness along which translational sliding may take place is the lithological contact between competent lithologies comprising either the Sandstone or Chert Lithozones (Fig. 10.5, diagrams A and B) and other lithozones such as the Foliated-, Diamictite- or Argillite Lithozones, the latter three containing a higher percentage of argillaceous material.

Translational sliding in bedrock material is highly dependent upon the attitude of the strata and its relation to slope orientation, the influences of which are discussed next. The most favourable locations for sliding occur where the strike of the strata is parallel or subparallel to slope contours and where the dip of the strata is towards the free face. There is particular danger of translational slides forming if the dip of the strata is less than (Figs 10.4 and 10.5; diagram A) or equal to (Figs 10.4 and 10.5; diagram B) the slope of the free face. Here, the bedding planes emerge at the free face and in the absence of a toe slope, rock slides are likely to occur. In the majority of locations, however, the near vertical dip of the strata is greater than the slope of the free face (Figs 10.4 and 10.5; diagram D). Here the valley slope is relatively stable in that rock sliding and large debris avalanches are unlikely to occur by bedding plane translation. Similarly, these slope movements are unlikely to be found at localities where the strata dip into the hillside (Figs 10.4 and 10.5; diagram E). The least favourable locations occur where the strike of the strata is perpendicular to the free face (Fig. 10.8).

Near surface, localised and open stratal flexures can produce a potentially unstable free face where translational sliding could result in rock slide or debris avalanche activity (Figs 10.4 and 10.5; diagram C).

Slope movement by translational sliding is sometimes associated with other discontinuities in the bedrock such as fault planes. Fault planes dipping towards the free face are potentially unstable surfaces. Their instability increases proportionally as the strike of the fault plane approaches that of the slope. Many of these fault planes are high angle faults that correspond with bedding plane surfaces. Fault planes that do not correspond with bedding plane surfaces are generally low-angle thrusts. Sliding along such surfaces in many instances is facilitated by the presence of blue-grey, soft, wet, fault gouge (pug) which varies in thickness from between 0.1 - 0.5m. The amount of translational sliding along these surfaces is difficult to assess.

Joint and cleavage surfaces have not been recognised as major discontinuities along which translational sliding occurs largely because the continuity of these surfaces is of limited extent.

3. Movements Comprising Earth Materials

The origin of the surface along which translational sliding in earth

FIGURE 10.8: Very shallow debris avalanches resulting from failure at the colluvium-bedrock contact. Note the near vertical bedding attitude of the thin-bedded strata comprising the Tamaki Lithotype. In such situations the orientation of the strata, perpendicular to the free-face of the slope, is not conducive to those slope movement types involving failure along planes of weakness within the bedrock. These debris avalanches are thus very shallow and are of small areal extent in comparison to those debris avalanches that involve bedrock material.

material takes place in this area is uncertain. A number of earth slides in the lower reaches of No. 1 and No. 2 Lines (localities 76 and 78-80; Maps 5 and 6; see also Fig. 10.9) illustrate downslope movement towards the SW, W and NW. At each locality the surface of translational sliding is planar in outline, gently-dipping and approximates what is thought to be a bedding plane.

The removal of all the material within the outline of each earth slide is thought to indicate that movement was rapid. Upon reaching a drainage channel, slide material appears to develop into earth flows.

B. Falls and Topples

1. Rock Falls

Rock Falls are predominantly the result of failure along joint surfaces within stratigraphically thick units of unbedded sandstone. Jointing on a regional scale is poorly developed but locally may be strongly developed. Two dominant joint orientations have been mapped; one strikes from northwest to southeast and the other strikes in an east-west direction at near right angles to the former set of joints. The angle of dip of both sets of joints is highly variable but in general is steep. No consistent pattern of dip direction could be discerned for either set of joints. Both sets are open joints.

Rock falls associated with strata comprising the Tamaki Lithotype are predominantly restricted to localities where very steeply dipping bedded strata are exposed in stream channels. Here rock falls are the result of failure along bedding planes which in many instances have been tectonically disrupted to form open joints. There is no correlation between bed thickness and rock fall activity. Rock falls associated with strata comprising the Wharite Lithotype occur along near vertical bluffs of thick sandstone. Here rock falls are particularly evident in eastward and westward draining catchments where the strike of the east-west set of joints parallels valley slopes and stream channels within which the sandstones occur. Some rock falls occur irrespective of the dip direction of joint surfaces (Fig. 10.10, diagrams A and B) whilst others occur where joint surfaces dip towards the stream channels (Fig. 10.10, diagram C).

Most rock falls are preceded either by toppling or translational sliding. Where rock fall activity has occurred the resultant parent rock-face is often blocky in appearance. It is at these localities that it is readily apparent that both sets of intersecting joints have largely controlled

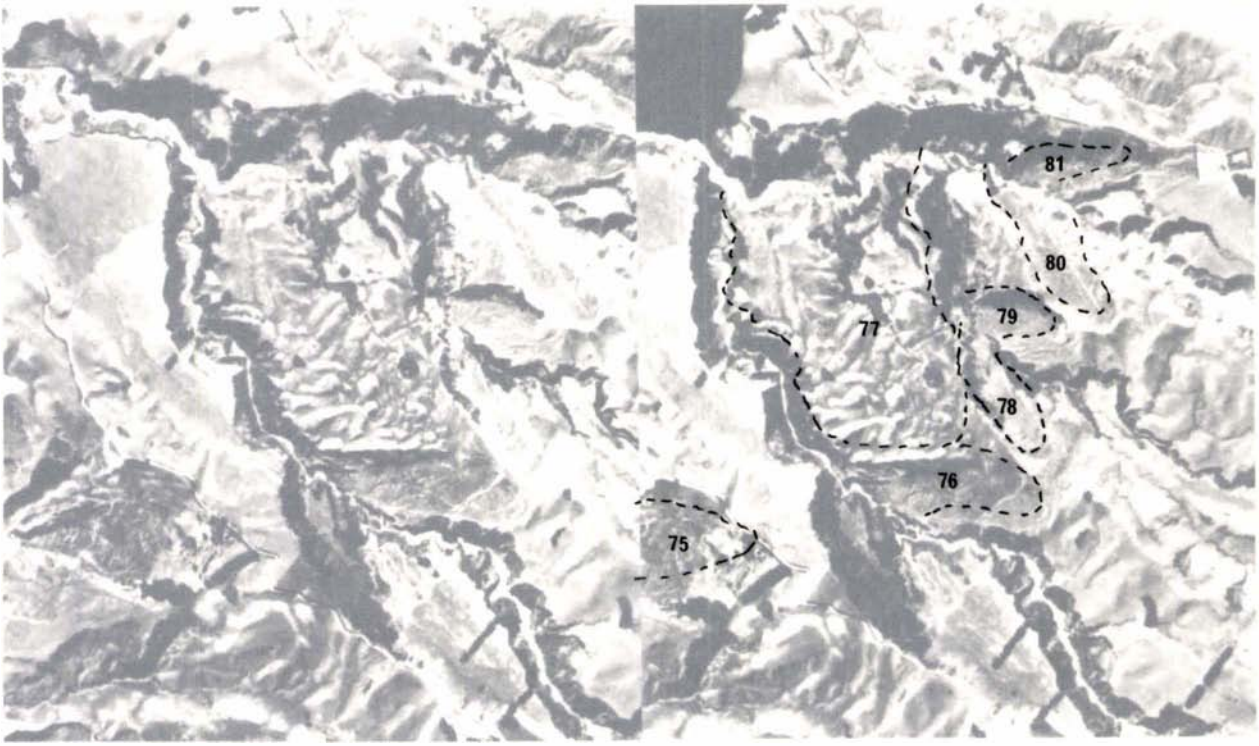


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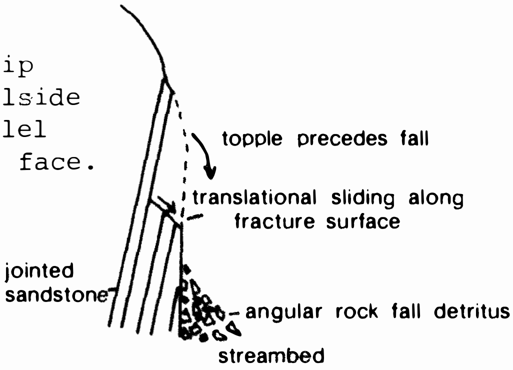
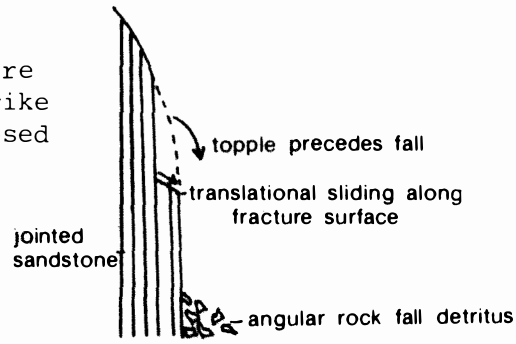
FIGURE 10.9: Stereoscopic view of large-scale earth movements in the lower reaches of No. 1 Line and No. 2 Line streams. Five mass movement features (localities 76 and 78-81) have been classified as earth slides and two (localities 75 and 77) have been classified as earth slumps (Appendix VI). Localities 76-81 are thought to be the result of translational sliding along a westward dipping ?bedding plane. The sliding of earth material from localities 76 and 78-81 is thought to have resulted in the formation of earth flows down the channels of No. 1 Line and No. 2 Line streams. For locality details of mass movement features, refer to Maps 5 and 6 and Appendix VI.

(A) Joint surfaces dip steeply into hillside and strike parallel to exposed bluff face.



(B) Joint surfaces are vertical and strike parallel to exposed bluff face.



(C) Joint surfaces dip towards the free face and strike parallel to exposed bluff face.

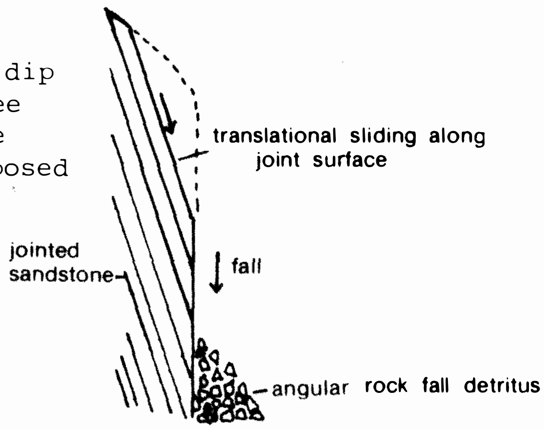


FIGURE 10.10: Schematic diagram illustrating variations in the structural attitude of joint surfaces, with respect to the free face of sandstone bluff, along which failure results in rock fall activity. Sometimes topple precedes rockfall activity (diagrams A and B) and in other instances translational sliding precedes fall activity (diagram C).

rock fall activity. Where the resultant rock-face is planar in appearance, it appears that rock fall activity has been controlled by one dominant set of joints (Fig. 10.11). Sets of filled joints were not found in association with rock fall activity. Consistent relationships between joint size, spacing, open or filled joints and the incidence of rock falls could not be established. Rock fall locations are shown on Maps 5, 6 and 7. Slope movement as a result of rock fall activity is of a small scale and of minor importance in this area.

2. Rock Topples

The plane of separation associated with some rock topples is clearly of fault origin. It is apparent that the horizontal separation between the parent wall and the toppled block results from the erosion of fault gouge from along a fault plane. However, the orientation of these features is not always consistent with the dominant northeasterly striking fault pattern often tending to follow the orientation of adjacent valley slopes. Here, topples coincided with open joint surfaces of limited extent. Examples of bedding plane failure resulting in topple could not be verified. Examples of topple culminating as rock falls were recognised at strongly jointed bluffs of sandstone lithologies (Fig. 10.10, diagrams A and B).

3. Earth Falls and Earth Topples

The origin of discontinuities along which movement results in earth fall and earth topple activity is unknown. The attitude of such discontinuities is too steep to correspond with infrequent, gently dipping bedding surfaces within these materials. Joint patterns are indistinct. Fault planes are present but rarely do they correspond in attitude and strike direction to the surfaces along which movement propagates. Nonetheless, the failure surfaces are of planar outline which always parallel near vertical bluffs and thus it is thought that toppling results from separation of blocks of material from the parent bluff purely as a consequence of gravity; *i.e.* the material behaves isotropically.

Some earth falls and earth topples are found in association with large earth slumps. Here blocks of earth material become dislodged from a headwall scarp of a slump along transverse open tension cracks. Vertical cracks develop in response to downslope movement of the slump mass, and the mass finally falls or topples due to gravity.

FIGURE 10.11: Rock fall resulting from failure along the north-south trending joint system. The planar joint surfaces dip steeply towards the west (left). Channel at base of rock fall is choked with coarse angular rock fall detritus. Photograph taken in North Oruakeretaki catchment at T23/593091.

4. Debris Falls

There is no recognisable surface of failure associated with debris falls. Failure largely results because the shear strength of the free-standing deposit of unconsolidated material has been exceeded by the shear stress applied. Stress is distributed throughout a zone of shear of steep attitude that develops parallel or sub-parallel to the open face of the deposit. Again the material behaves isotropically and falls as a debris fall.

C. Deep-Seated Rotational Slides (Slumps)

1. Rock Slumps

Rock slumps generally form in homogenous materials but within the study area they form within two lithologically and structurally different bed-rock lithotypes (see Chapter 2). Each comprises a number of lithologies of diverse origin, differing in composition, and having very different mechanical strengths. The even distribution of rock slumps throughout the study area indicates that differences in the lithological characteristics of the bedrock (comprising Wharite and Tamaki Lithotypes) bears little relationship to their formation, *i.e.* the bedrock behaves isotropically and fails as a rock slump. Nonetheless a great many rock slumps in this area occur either adjacent to a surface trace of a major fault or within a zone of fault brecciated bedrock. This relationship is examined further with examples of rock slumps along the strike of the Wellington and the Ruahine Fault Zones.

Fault related factors that influence the stability of valley slopes are also detailed. These include examination of fault deformation of bed-rock lithologies, relationships between fault zone width and orientation with respect to structural attitude of bedrock strata and to valley slope, and the location of faults in relation to slope steepness and to rock slump movements.

(a) Characteristic features of major fault zones

Fault zones usually consist of a number of discontinuous, sub-parallel, en-echelon traces that are characteristically straight and are often expressed physiographically as a series of scarps, trenches, saddles, notches, truncated spurs and fault aligned stream reaches (see Chapter 6 and Map 4). The Wellington Fault Zone demarcates the eastern margin of the southern Ruahine Range. Between the Manawatu Gorge and Otamarahu

Stream a number of discontinuous, sub-parallel fault traces are preserved either within Late Quaternary sediments or form the contact between them and Torlesse bedrock. From Otamarahu Stream northwards, a single discontinuous fault trace is predominantly confined within bedrock strata comprising the Tamaki Lithotype. Only here at altitudes of between 300-500m do steep slopes (>30^{o}) border this Fault Zone. Displacement of terrace surfaces of Late Quaternary age indicate that this fault was active in Late Quaternary time.

The Ruahine Fault Zone has also been mapped throughout the study area (see Chapter 6 and Map 4). This Fault Zone strikes across steep upper catchment reaches at considerably higher altitudes (average altitude of between 750-1000m) than the Wellington Fault Zone. Valley slopes adjacent to the Ruahine Fault Zone are steep and have an average gradient of 30^{o}. However, slopes of 40^{o} to 50^{o} are common where this fault crosses stream channels. Throughout part of its length this Fault Zone displaces and brecciates bedrock of the Wharite Lithotype, and in part separates the two bedrock lithotypes (Map 1). There are few surface expressions to indicate that the Ruahine Fault Zone was active during Late Quaternary times, however this fault is considered to be 'Potentially Active' (see Chapter 6).

Rock slumps are also found in association with splay faults and cross faults that abound throughout the Range (see Chapter 6 and Map 4).

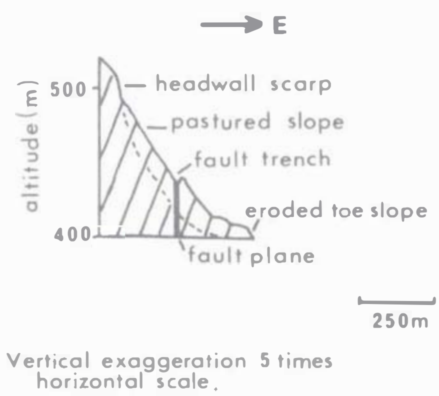
It is often difficult to determine the amount or direction of throw of these faults because the fault planes are rarely exposed; dense vegetation often masks the fault scarps and removal of the scarps by erosion is common, particualrly within the Range itself. Consequently, some of the fault zones are largely identified by the presence of bedrock exposures of fault breccia or strongly disrupted strata in places measuring several hundred metres across. In the study area bedrock within fault zones may exhibit one of two forms of deformation. The first comprises extensive brecciation of the bedrock. Sandstone lithologies are reduced to a grey or white coloured breccia that may show signs of cementation. Blue-grey coloured gouge is often found adjacent to the fault plane along which movement took place. If both sandstone and argillaceous lithologies are present within the crush zone, the resultant breccia is 'blackened' in appearance due to the presence of the argillaceous component. Where argillite is present it tends to be more abundant within the breccia zone than in adjacent unfaulted bedrock.

This apparent increase in the argillaceous component may be due to argillite remobilisation during intense deformation (Reed, 1957). The second form of fault deformation involves loosening of *in situ* bedrock along discontinuities such as bedding planes, joints, cleavages and shear surfaces. The result is reduced cohesion along the discontinuities, thereby decreasing the shear strength of the rock unit. The Tamaki Lithotype is particularly prone to this form of deformation because of the numerous bedding-plane surfaces. Within the Wharite Lithotype (mélange terrane) loosening occurs along numerous cleavage surfaces within the pervasively sheared argillaceous matrix. In addition, loosening occurs along contacts between blocks, pods and lenses of competent lithologies and the surrounding argillaceous matrix. These clasts are then easily removed from the matrix by physical erosion and fall out of near vertical bedrock exposures.

In the field, bedrock disrupted as a result of deep-seated slump movement is distinguishable from fault brecciated bedrock: firstly, by the degree of fragmentation; and secondly, by the absence of veining, gouge and cementing matrix. Slump movement results in fragmentation (without crushing) whereas fault movement generally results in brecciation of all lithologies.

(b) The relationship of rock slumps to major fault zones

This relationship is discussed with respect to: firstly, the Wellington Fault Zone; secondly, the Ruahine Fault Zone; and thirdly, the Piripiri Fault. The first three examples illustrate variations in the position of the trace of Wellington Fault with respect to rock slumps. Each of these slumps involves failure of slopes underlain by bedrock comprising the Thin-Bedded Association of the Tamaki Lithotype. The first example is at locality 36 (Maps 5 and 6) where vertical slump movement has been small but sufficient to produce a distinct headwall scarp. Subsequently, much of the slumped bedrock mass has remained intact as a coherent unit; the grassed surface is not unduly hummocky in appearance. The slope of the slump surface has remained as steep as the adjacent unslumped hillside. The slump covers an area of 18.1 hectares. A Late Quaternary fault trace cuts across the midslope of the slump (Fig. 10.12). Here, fault pug within a fault trench cuts across the slump, and fault brecciated bedrock occurs at two localities 250m and 750m further north along the strike of the fault trace. Outcrops of highly disrupted bed-



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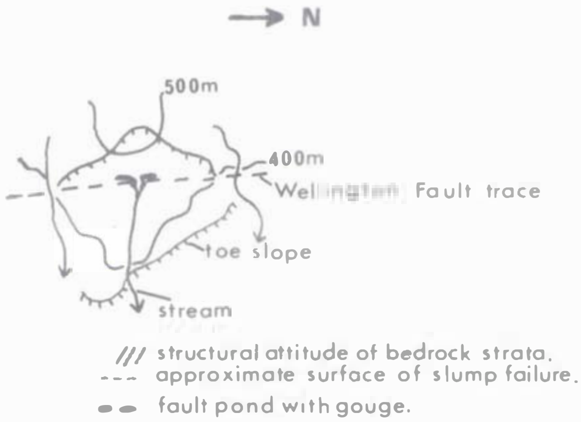

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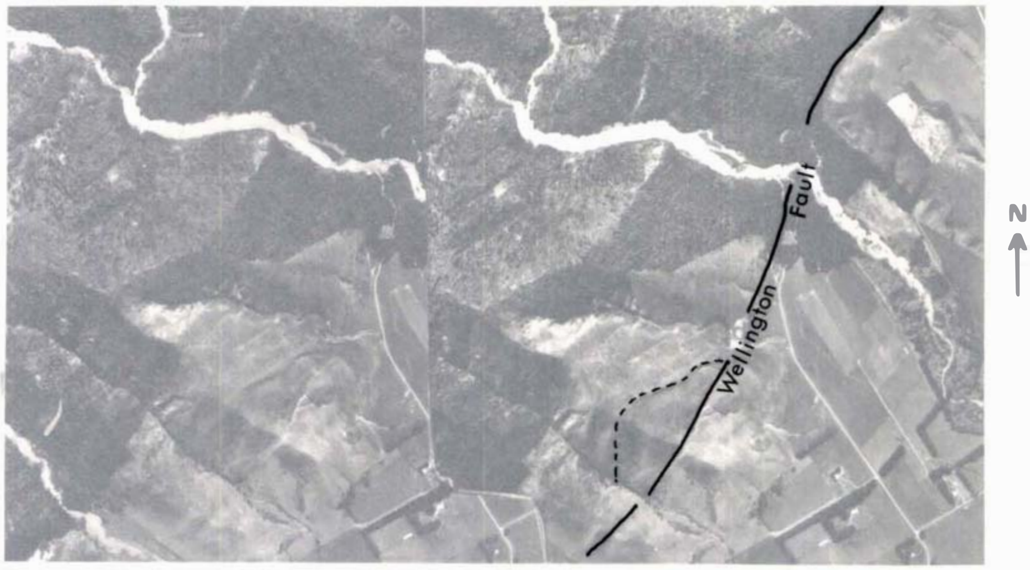

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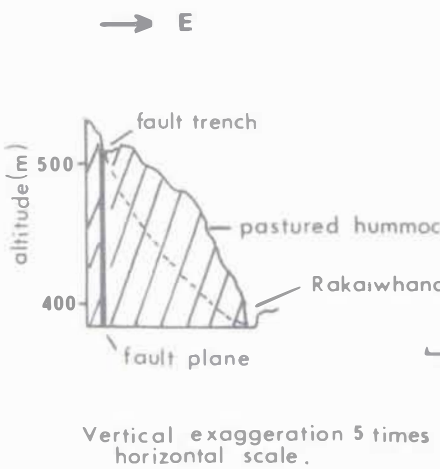
FIGURE 10.12: Position of Wellington Fault trace in relation to rock slump at locality 36. Cross-section (a) illustrates the structural attitude of the bedded bedrock strata and fault plane in the vicinity of this slump. Slump failure involves bedrock lithologies comprising the Tamaki Lithotype. The same locality is shown in plan view (b) and in stereoscopic view (c). For details of slump locality refer to Maps 5 and 6 and Appendix VI. For details of tectonic features along Wellington Fault refer to Map 4 and Appendix Va.

rock are exposed downslope of the fault trace near the base of the rock slump.

The second example at locality 30 (Maps 5 and 6) is less regular in outline but displays greater surface irregularity as a result of substantial vertical displacement. This slump covers an area of 26 hectares. Slump movement has formed a 20m wide, flat-bottomed trench approximately 5m deep at the head of the slump. This trench separates the head of the slump from the steep bedrock slope above (Fig. 10.13) and coincides with the Late Quaternary trace of Wellington Fault (Chapter 6 and Map 4). Brecciated bedrock crops out in stream channels to the north and south along the strike of the fault trace but is not evident within the trench. Highly disrupted bedrock occurs along the right bank of Rokaiwhana Stream where it cuts across the toe of the slump, and along a small stream that follows the southern margin of the slump.

The third example at locality 11 (Maps 5 and 6) involves minimal slump movement of 8.2 hectares of forested slope, at the base of which there is an extensive zone of fault brecciated bedrock. Streambank undercutting parallel to the base of the slump has exposed fault breccia. The Late Quaternary trace of Wellington Fault cuts across the lower portion of this slump (Fig. 10.14).

At each of the above localities bedrock strata strike to the north and northeast parallel to slope contours. As the bedrock strata dip steeply into the hillside towards the west and the slumps have moved downslope towards the east, it is not possible for slumping to have occurred as a result of sliding along bedding plane surfaces. At these localities the trace of Wellington Fault also coincides with the strike of the bedrock strata and lies parallel to slope contours. The attitude of the fault plane is thought to be either near vertical or steeply westward dipping (Ower, 1943). The relatively straight surface trace of this fault and others in the study area supports a high angle fault plane. Although the fault plane is not exposed in proximity to these slumps, the presence of a zone of fault brecciated bedrock and a topographical fault trace can be verified. Neither the surface along which rotational slump movement took place nor the plane of fault displacement are exposed. Thus the origin of the surface of rotational slump movement is currently unknown but it may: (1) coincide in part with the plane of fault displacement; or (2) originate independently of existing litho-



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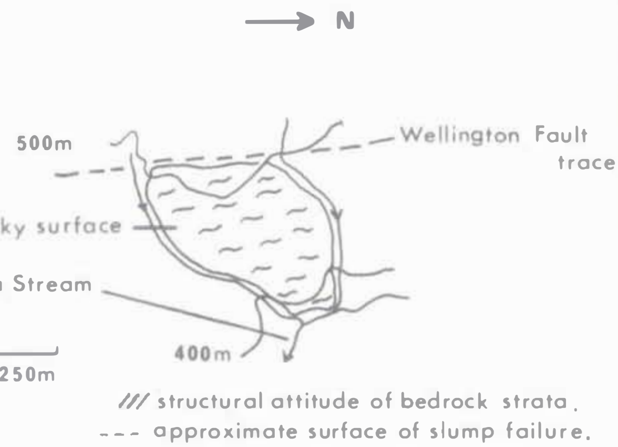

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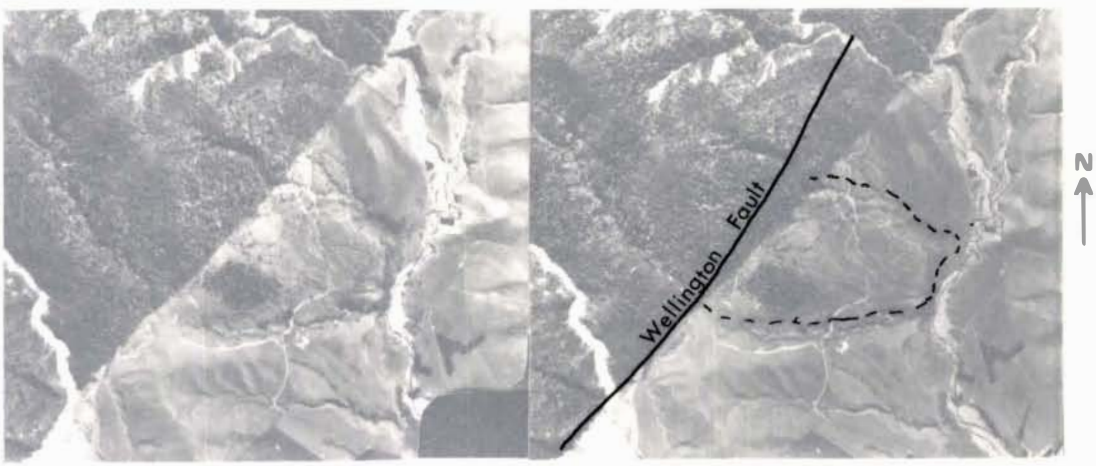

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FIGURE 10.13: Position of Wellington Fault trace in relation to rock slump at locality 30. Cross-section (a) illustrates the structural attitude of the bedded bedrock strata and fault plane in the vicinity of this rock slump. Slump failure involves bedrock lithologies comprising the Tamaki Lithotype. The same locality is shown in plan view (b) and in stereoscopic view (c). For details of slump locality refer to Maps 5 and 6 and Appendix VI. For details of tectonic features along Wellington Fault refer to Map 4 and Appendix Va.

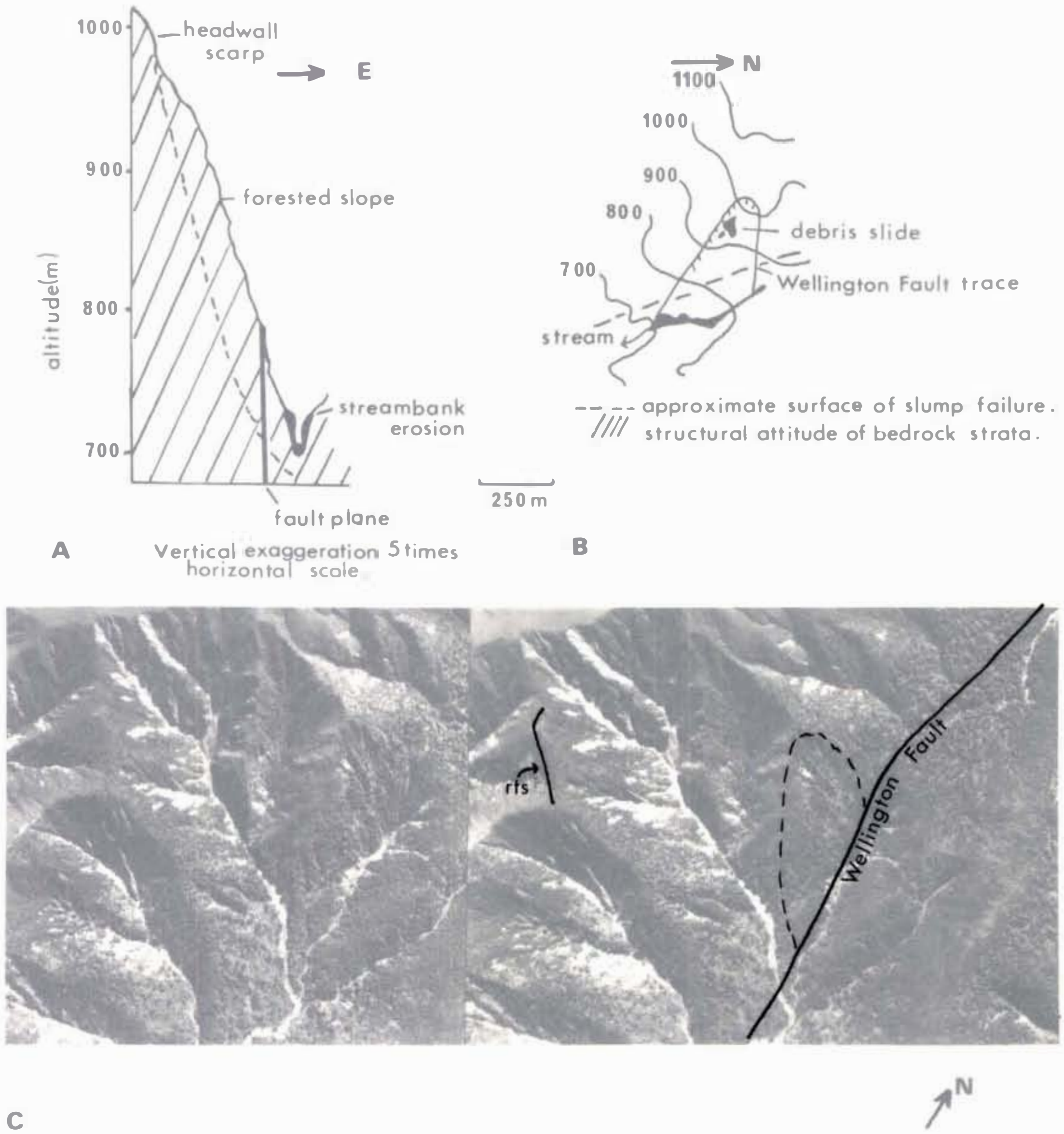

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FIGURE 10.14: Position of Wellington Fault trace in relation to rock slump at locality 11. Cross-section (a) illustrates the structural attitude of the bedded bedrock strata and fault plane in the vicinity of this rock slump. Slump failure involves bedrock lithologies comprising the Tamaki Lithotype. The same locality is shown in plan view (b) and in stereoscopic view (c). Other mass movement features are also illustrated, *e.g.* r.t.s. (ridge-top scarp). For locality details of mass movement features refer to Maps 5 and 6 and Appendix VI. For details of tectonic features along Wellington Fault refer to Map 4 and Appendix Va.

logical or structural discontinuities within the bedrock being exacerbated by the presence of a zone of fault brecciated and/or fault disrupted bedrock.

A similar relationship exists between large-scale slump movements and the Ruahine Fault Zone which displaces bedrock comprising the Wharite Lithotype. At locality 61 (Maps 5 and 6) a rock slump measuring 500m in width at its base extends 200m upslope from an altitude of 700m to 900m a.s.l. (Fig. 10.15). The ovate slump outline is sharply defined by pronounced headwall and lateral scarps that delineate an area of 14.5 hectares (Fig. 10.17). The surface of the slumped mass is hummocky with the major part still retaining its original forest vegetation cover. The foot of the slump has been deeply dissected by gully erosion and the toe slope is severely eroded, owing to present day fluvial undercutting by the eastward draining Raparapawai Stream. The strata comprising Wharite Lithotype in the vicinity of the base of the slump have been severely brecciated and contorted by movement on the Ruahine Fault. The north to northeast striking and steeply eastward dipping bedrock foliation within this fault breccia zone is barely recognisable. The strike of the Ruahine Fault Zone parallels bedrock foliation and the fault plane is likely to have a near vertical attitude. As the surface of failure dips towards the north and is at right angles to both the dip direction of bedrock foliation and the fault plane, it is unlikely that either lithological or structural discontinuities are related to the development of the surface of failure beneath this slump. This slump sits astride a zone of brecciated bedrock which in this vicinity exceeds 200m in width. Signs of fault brecciated and disrupted bedrock extend intermittently over a total width of approximately 500m. Much of the breccia has been carbonate-cemented, thus indicating an early episode of faulting (see Chapter 6). Refaulting, possibly during Late Quaternary times, has produced much fault gouge, recrushed the fault breccia and loosened clasts of competent lithologies within the encompassing matrix. The result of such extensive faulting has been to effectively homogenise the bedrock. The brecciated bedrock lost its original stratal fabric as a result of movements along this Fault Zone and behaved isotropically. Thus the slump is considered to have formed by gravitational collapse of a mass of extensively brecciated bedrock along a northward sloping surface of failure.

The last example of a rock slump occurs on a splay fault, the Piripiri Fault, which strikes southwest of Ruahine Fault Zone into Piripiri Stream

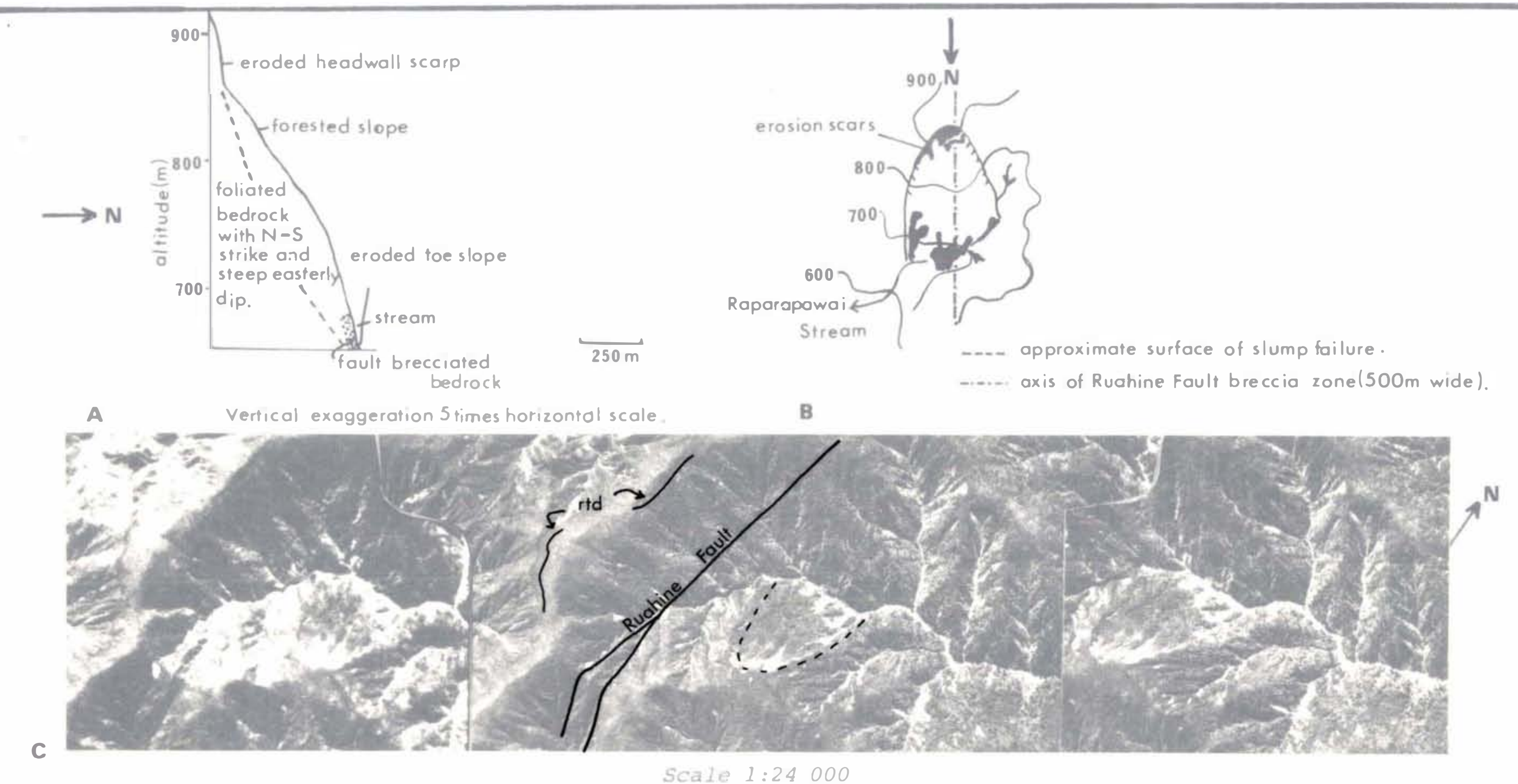

FIGURE 10.15: Position of Ruahine Fault trace in relation to rock slump at locality 61. Cross-section (a) illustrates the structural attitude of the foliated bedrock strata in the vicinity of this slump. Slump failure involves bedrock lithologies comprising the Wharite Lithotype. The same locality is shown in plan view (b) and stereoscopic view (c). Other mass movement features are also illustrated, *e.g.* r.t.d. (ridge-top depression). For locality details of mass movement features refer to Maps 5 and 6 and Appendix VI. See also Figure 10.17. For details of tectonic features along Ruahine Fault refer to Map 4 and Appendix Va.

catchment where a major rock slump has been identified. It is the largest rock slump in the study area covering an area of 39 hectares (locality 2, Maps 5 and 6). From its base at valley floor level, where it is 750m wide, it extends 400m upslope to the crest of an adjoining ridge at 1200m a.s.l. The roughly ovate outline is sharply defined by very steep scarps, the highest of which occurs along its eastern margin (Fig. 10.16). Scrub and forest cover the upper two-thirds of the slump. Recent fluvial bank undercutting at its toe has caused renewed small-scale slumping and stream bank collapse. These active eroding lower slopes have been colonised by tussock grasses. A deeply dissected tributary drains down the centre of the slump (Fig. 10.16). The trace of Piripiri Fault coincides with the toe of this slump. Piripiri Fault can be traced across the upper reaches of several tributary streams of the Pohangina River located to the north and east of Piripiri Stream. Several other large-scale slump features are aligned along this fault trace beyond the northern boundary of the study area. The catchment east of Piripiri Stream contains rock slumps that parallel the ridge crests, indicating similar large-scale slope movements have occurred (localities 2b and 2c, Maps 5 and 6).

The bedrock strata in Piripiri Stream catchment differ from that exposed near the Raparapawai slump (locality 61) in that it is tectonically less deformed. Thus much of the strata consists of regularly bedded lithologies, each separated by a distinct bedding plane surface. However, disrupted strata exhibiting a distinct foliation are also present in adjacent outcrops. Units of strata showing these two distinctly different structural styles appear to be interbedded at this locality. The strike of the foliation and bedding planes varies between 60° and 80° east of north, which approximately parallels the orientation of both Piripiri Stream channel and the fault trace. The strata dip towards the southeast at between 50°-60°. Here (locality 2), failure may have originated in two different ways. First, the strike of the strata parallels Piripiri Stream channel and dips steeply towards the southeast in the direction of the free face of this slump (Fig. 10.16), so discontinuities within the bedrock, coupled with the landscape configuration, may have initiated slope failure. Second, failure may be related to a lowering of shear strength of the fault breccia present along the toe of the slump.

It is not postulated that all large-scale rock slumps are necessarily associated with faults or that the presence of fault brecciation or dis-

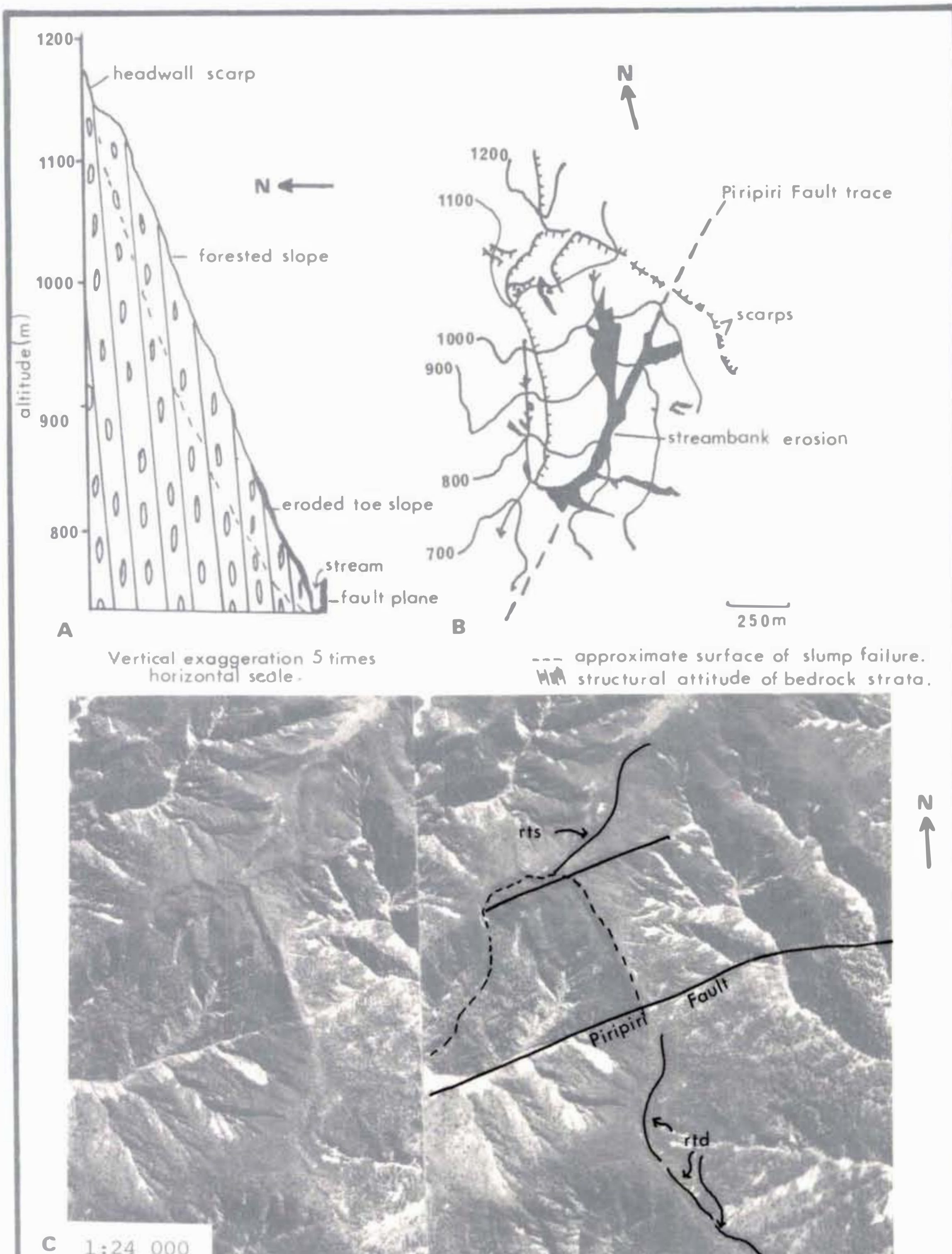

FIGURE 10.16: Position of Piripiri Fault trace in relation to rock slump at locality 2. Cross-section (a) illustrates the structural attitude of the foliated bedrock strata in the vicinity of this slump. Slump failure involves bedrock lithologies comprising the Wharite Lithotype. The same locality is shown in plan view (b) and stereoscopic view (c). Other mass movement features are also illustrated, e.g. r.t.s. (ridge-top scarp) and r.t.d. (ridge-top depression). For locality details of mass movement features refer to Maps 5 and 6 and Appendix VI. For details of tectonic features along Piripiri Fault refer to Map 4 and Appendix Va.

rupted bedrock is a prerequisite for their formation in the study area. Of the 56 rock slumps documented (Appendix VI) the majority (60%) occur in localities where there is a wide zone of fault brecciated bedrock. The greater the width of the fault zone the greater is the potential for the development of rock slumps of large size. The orientation of fault zones with respect to valley slopes appears unrelated to rock slump formation because they occur both where the orientation of the fault zone is parallel (*e.g.* Wellington and Piripiri Faults) and perpendicular to slope contours (*e.g.* Ruahine Fault). Approximately 25% of the rock slumps in this area occur in association with faults of known orientation whilst 35% are found in localities where faults have no physiographic expression and the orientation of fault breccia zones is unknown. Forty percent of all rock slumps in this area do not appear to be related to faults in any known way.

Most rock slump movements in the southern Ruahine Range are the result of gravitational forces acting upon large masses of fault disrupted bedrock. Brecciation and disruption of the bedrock preceded and facilitated subsequent rock slump movement. Failure seems to coincide with periods of heavy rainfall but may also coincide with periods of rejuvenated fault rupture or seismic shaking (see Section 10.4). The temporal relations between fault displacement and slump movement in this area is further discussed in Chapter 12.

Small-scale slump features develop at the toe slope of larger deep-seated slumps. These small-scale slumps often fail as a result of streambank undercutting into disrupted bedrock resulting in oversteepening of the toe slope (Fig. 10.18).

2. Debris Flows

In the study area there are two requirements necessary to form a debris flow. Firstly, there must be a mass movement feature of substantial size that is capable of contributing large amounts of rock detritus to a stream channel. Within the mapped area it appears that debris flows only develop where large-scale, deep-seated rock slumping has occurred. A relationship thus exists between the incidence of rock slumps and the formation of debris flows, given that other conditions, including moisture content and detritus of suitable composition are met. Shallow types of slope movement such as debris slides, debris avalanches, rock slides and rock falls have been of insufficient size to result in debris flows within the period of observation. Although these shallow slope movements

FIGURE 10.17: Large-scale rock slump with well defined lateral and headwall scarps. Slump movement is towards the north (right). This slump sits astride a 500m wide, northeast striking fault breccia zone mapped as the Ruahine Fault. Note the occurrence of debris slides in the headwall area and gully erosion together with debris slides in the toe area of the slump, much of which has developed since 1946 (see Chapter 11).

Photograph taken in Raparapawai Catchment at locality T23/563071.

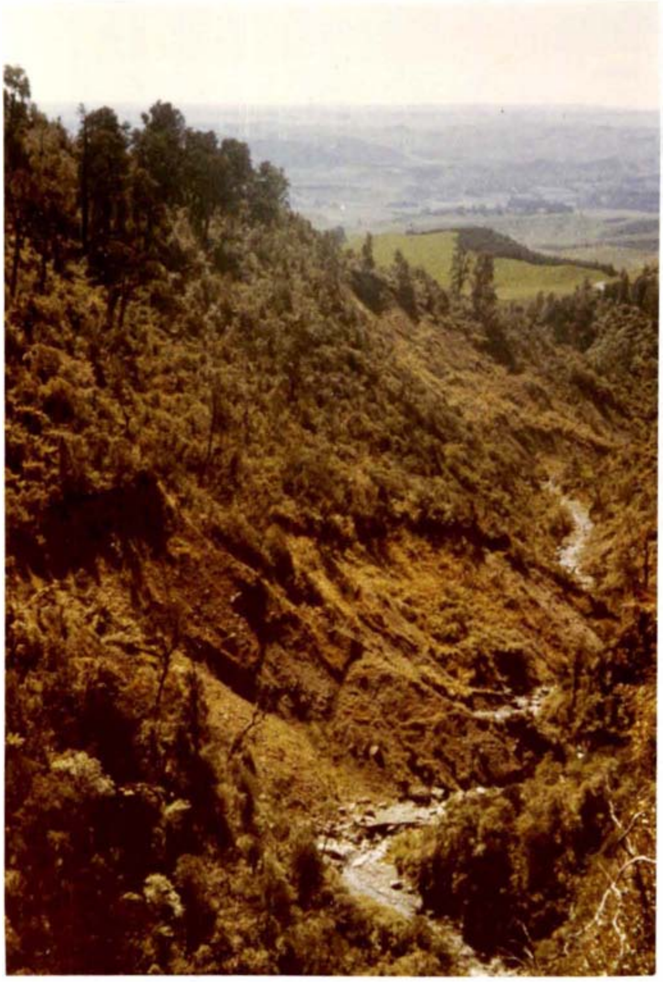

FIGURE 10.18:

Small-scale rock slumps and debris slides involving disrupted bedrock lithologies of the Wharite Lithotype and thick accumulations of colluvium. Failure is mainly the result of streambank undercutting.

Photograph taken in No. 1 Line catchment.

are capable of supplying sufficient volumes of material, the material comprises colluvium which is coarse and lacks sufficient "fines" to lead to debris flow formation.

Second, the detritus must comprise a substantial volume of fine-grained argillaceous material. This is always derived by rock slump activity from bedrock lithologies comprising either the Wharite or the Tamaki Lithotypes. Within the Wharite Lithotype, argillaceous material forms a cementing matrix around all clasts and varies from 0-100% between lithozones. The overall impression of the *in situ* bedrock is its dark colouration due predominantly to the omnipresent argillaceous matrix. Slumping on a large-scale is most likely to occur in the structurally weaker Foliated and Diamictite Lithozones (see Chapter 2) in which the argillaceous content varies from 30% to 100%. Rock slumps within catchments underlain by the Tamaki Lithotype generally involve bedded strata comprising the extensive Thin-Bedded Association (see Chapter 2), with an argillaceous content of approximately 30%.

Evidence for debris flow activity in this area is limited to one historic example (see Chapter 9) which occurred within a catchment underlain by lithologies comprising the Wharite Lithotype. Nonetheless the argillaceous content of both the Tamaki and Wharite Lithotypes is considered sufficient for formation of debris flows. The size and distance travelled by debris flows will be largely determined by other factors such as volume of material involved in the initial rock slump movement and the amount of moisture present within bedrock and stream channels at the time of slump movement. Within the study area all 56 rock slumps (Appendix VI) are potentially capable of developing into debris flows in future movements. This suggests that debris flow activity could be an important mass movement process in the future and discussion of this possibility is included in Chapter 12.

3. Earth Slumps

Earth slump movements in the study area are found at localities where inherent lithological discontinuities, particularly bedding planes or faults, are present. Fifty eight percent of all earth slumps in this area occur on traces of known faults. Some of these earth slumps (localities 90-97 and 103-108, Maps 5 and 6) are aligned along the active Late Quaternary trace of Wellington Fault. At most of these localities the failure surface along which movement occurred shows no known structural relationship with the fault plane. However, at localities 105, 106 and 108 (Fig. 10.19) it is probable that fault scarps coincide with steeply eastward-dipping and relatively straight head-wall scarps of these earth slumps. This may indicate that deep-seated, rotational downslope movement initiated from a steeply eastward-dipping fault plane. Other earth slumps are found in association with active contact faults (*e.g.* Umutoi Fault). These contact faults separate Torlesse bedrock from unconformably overlying Plio-Pleistocene marine deposits. The marine deposits have been dragged as a result of tectonic upwarping into steeply dipping attitudes (see Chapter 6). Slump formation along these contact faults may thus result from failure at: (1) the fault plane; (2) the major unconformable contact between Torlesse bedrock and Plio-Pleistocene marine deposits; or (3) steeply dipping bedding plane surfaces within the marine deposits.

(1) Slumping as a result of movement along the fault plane is considered unlikely in this area because the majority of contact faults are considered to be high angle reverse faults, *i.e.* the fault planes dip towards the Range. Extensive areas of brecciated Plio-Pleistocene marine deposits are not present adjacent to these contact faults. Thus slump movement within breccia zones as a result of gravity collapse, similar to that proposed for rock slumps, is not evident at these localities.

(2) Slumping as a result of movement at major unconformable contacts seems probable because this contact slopes steeply away from the Range towards lower lying areas adjacent to the Range. Such a movement would involve translational sliding of Plio-Pleistocene marine deposits along this contact. The majority of slumps located at contact faults, however, show obvious signs of rotational rather than translational movement and nowhere is the bedrock contact exposed. It is therefore unclear if slump movement at these localities resulted from initial translational sliding along the bedrock contact.

(3) The strike of bedding plane surfaces within Plio-Pleistocene marine deposits parallels the strike of contact faults. The dip of the strata varies between 30° and 60°, although overturned strata have been reported at one locality adjacent to Whareroa Fault (see Chapter 6). Thus the strata predominantly dip away from the Range front. At these localities slump movement may begin with translational sliding along a bedding plane surface

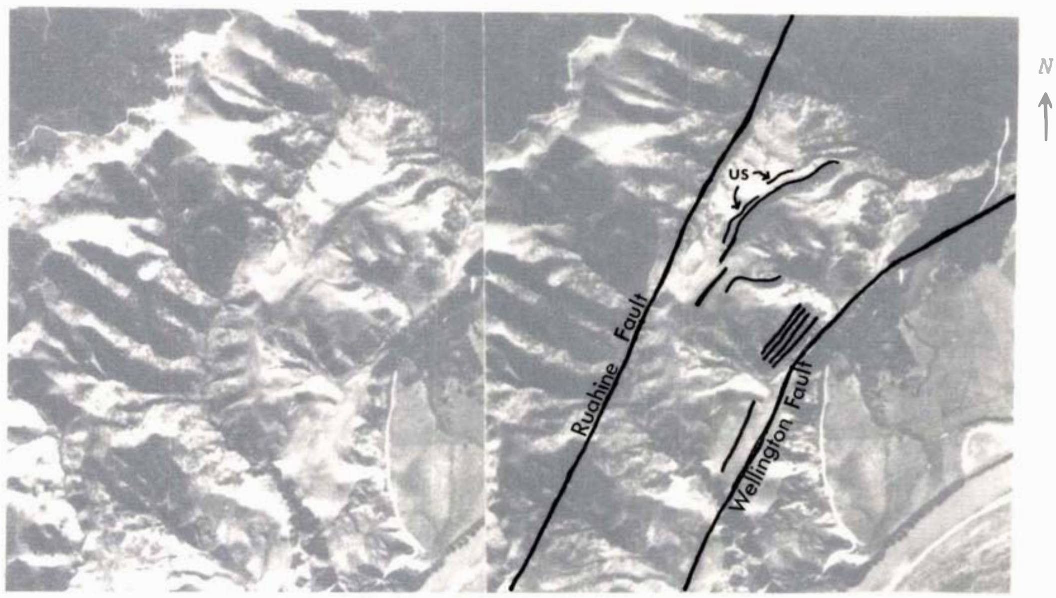

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FIGURE 10.19: Stereoscopic photo-pair of large-scale earth slump at locality 108 (Maps 5 and 6). This slump is wedged between a 'Potentially Active Contact Fault' (Ruahine Fault) and the 'Active Fault' trace of Wellington Fault. The straight headwall scarps of this slump are thought to be the result of deep-seated movement of an earth mass down the steeply eastward dipping fault plane of Wellington Fault. Physiographic features of tectonic origin, in the vicinity of this earth slump, are shown on Map 4 (Inset E) and documented in Appendix Va.

which progresses to a new curved surface of failure upon which rotational movement takes place.

Earth slumps occur adjacent to the Faulted Pohangina Monocline where Plio-Pleistocene marine deposits have been folded into steeply westward dipping attitudes. Earth slumps at localities 84 and 85 are situated immediately to the west of the Monocline where the marine strata are at their steepest attitude between 50° and vertical (Map 4). Here slump movement is considered to develop in a similar way to that described in (3) above.

An earth slump at locality 86 (Maps 5 and 6) has moved northward in a direction at right angles to the westward dip direction of the marine strata. Here the north sloping curved surface of failure appears to have developed at right angles to and independently of the bedding planes. There does not appear to be any relationship between the surface of failure of this slump and an eastward-dipping contact fault plane exposed at T23B/585192.

In summary, movement of the majority of earth slumps appears to show no consistent relationship to lithological or structural discontinuities within weakly indurated Plio-Pleistocene marine deposits in this area. Only at a few localities does the initial downslope movement of a slump mass coincide with bedding planes, bedrock contacts or an active fault plane. It is concluded that the majority of earth slumps in the study area fail when the shear strength of earth materials is exceeded to result in downslope rotational movement along a curved surface of failure. This failure surface develops independently of and is discordant to planar discontinuities within these earth materials. Earth slumps in the southern Ruahine Range are thus thought to be largely of gravitational origin. They occur predominantly in topographically oversteepened localities where climatic and seismic factors (see Sections 10.3 and 10.4, respectively) cause the marine deposits to fail as earth slumps.

4. Earth Flows

Earth flows, like debris flows, are poorly represented within the study area. There is no evidence to suggest that earth flows have been an active process in recent times. However, they may have been locally important in the past. Earth flows undoubtedly formed within stream channels at the base of earth slides at localities 76 and 78-80 (Maps 5 and 6; see also Fig. 10.9). Few earth flows appear to have developed subsequent to earth slumps. This is largely because earth slumps have occurred in areas distant from stream channels or because downslope movement of earth slump material has been minimal with much of the slump mass remaining *in situ*. When saturated, the predominantly fine-grained earth materials in the study area are highly conducive to earth flow activity. Earth flows could be an important mass movement process in the future (see Chapter 12).

D. Ridge-top Features

The uplifted bedrock block comprising the southern Ruahine Range is aligned between SW and NE. The pattern of east-west oriented narrow ridge crests is the result of deep valley dissection of the Range. The following features either form parallel or sub-parallel to ridge-tops and are possibly related to large-scale rock slumps.

1. Ridge-top Benches

Ridge-top benches occur at two localities in the study area. The first is in the headwaters of Car Park catchment, a tributary of the West Tamaki River, where a series of short parallel benches are arranged in step-like fashion, decreasing in altitude from the crest of the Range to the top of a debris avalanche scar known as "Catspaw" (locality 18, Maps 5 and 6). Immediately adjacent and to the north side of "Catspaw" there are four bench surfaces in an area where shallow translational slope failures have not previously occurred and the dense vegetation cover is intact. The second locality is in Rokaiwhana Stream (locality 19, Maps 5 and 6).

The alignment of ridge-top benches along the crest of the Range is paralleled by the northeast-southwest strike of the bedrock strata comprising the Tamaki Lithotype. At both localities bedding planes dip steeply towards the east. The parallelism of bench formation to bedrock strike suggests a genetic relationship. The exposed bedrock near the middle of the slip at "Catspaw" gives the impression of bulging outwards so that the eastward-dipping strata are less steep here than at the top of the slip. This is possibly a minor open flexure that typically characterises the style of fold deformation in the Tamaki Lithotype. However, the discovery of benches above Catspaw indicates that successive gravitational collapses have been initiated by downward movement of large tracts of hillside. The probable bending of bedded strata together with bench formation may therefore indicate that bedrock creep is important in some large-scale movements. There are no bedrock exposures above Catspaw so that deformation of the bedrock beneath the bench surfaces is not observable. Disrupted rock at the base of the valley slope has not been seen so that a deep-seated movement either by translational or rotational sliding along a discontinuity cannot be confirmed. The bench surfaces do not appear to show backward tilting and the separating scarps are linear and not curved. For these reasons rotational sliding at this locality has been ruled out. The absence of displaced surfaces and the limited lateral extent of the benches indicate that these features are not the result of fault displacement. These benches are not formed by differential erosion of argillite between resistant sandstone beds. Hubbard (1978) suggests that bench formation in the area adjacent to "Catspaw" may be the result of either: (1) creep of the surficial soil over the bedrock; or (2) deep-seated slumping within the bedrock in response to active stream undercutting. However, it is considered that the limited thickness of the soil in this area (0.7m) is unlikely to produce bench surfaces of 100m length, 10-30m width and up to 10m vertical separation. Rotational slumping has been discounted (see above) and active streambank undercutting at the base of the slope is not evident. The most likely mechanism is that of gravitational adjustment of oversteepened slopes in upper catchment areas by downslope movement of large masses of bedrock along a surface or surfaces of failure, which are most likely to coincide either in part or wholly with bedding planes.

2. Ridge-top Depressions

Ridge-top depressions are numerous along the axis of the Range where they have developed parallel to ridge crests. Many depressions are associated with bedrock strata comprising the Wharite Lithotype and occur at sites where the strike of the depression parallels bedrock foliation direction. This suggests failure along a discontinuity parallel to bedrock foliation. The direction of dip of bedrock foliation in the vicinity of depressions in some cases appears to control the direction of failure even where structural discontinuities dip into the hillside. This can be illustrated using two adjacent localities situated almost back to back along the same section of ridge crest. Here both the lithologic and structural features of the Wharite Lithotype consistently strike towards the northeast and dip towards the southeast. At locality 60 (Maps 5 and 6) failure towards the east has formed a depression along the eastern side of the ridge crest at the head of

Raparapawai catchment. To the west of this depression at locality 63 (Maps 5 and 6) large-scale, down-dip movement towards the east has formed a depression on the western side of the ridge. At other localities, *e.g.* locality 2c (Maps 5 and 6; see also Fig. 10.16), a ridge-top depression has formed at near right angles to the strike of foliated strata but parallel to the ridge crest. These examples suggest that whilst at some localities neither lithological nor structural discontinuities within the bedrock influence formation of ridge-top depressions, at other localities their formation may be assisted by lithological and/or structural discontinuities at favourable attitudes with respect to valley slope.

Ridge-top depressions are also associated with strata comprising the Tamaki Lithotype. Here, they are not dependent upon failure along bedding plane surfaces because their orientation cross cuts the strike direction of bedrock strata. Ridge-top depressions within the study area are not the result of differential erosion of an incompetent lithology. Because of their sinuous outline and the absence of fault breccia, ridge-top depressions are not considered to be the direct result of vertical fault displacement.

At locality 41 (Maps 5 and 6) it is readily apparent that this ridge-top depression is a result of deep-seated gravitational failure. Here, one large rock slump (locality 40) occurs below a ridge-top depression within Mangapuaka catchment. Two other rock slumps are situated further along the hill slope from the depression. Each of the slump features is delineated by obvious headwall and lateral scarps, indicating rotational movement has taken place. Within Mangapuaka catchment the southern stream bank gives the impression of being highly unstable and the bedrock shows obvious signs of disruption. Signs of faulting are absent. The depression is thought to mark the upper limit of an extensive mass movement feature which has undergone a small initial downslope movement to produce the ridge-top depression and possibly trigger the three smaller rock slumps.

3. Ridge-top Scarps

There is no evidence within the study area to suggest that ridge-top scarps are the result of failure along lithological or structural discontinuities within the bedrock. All the examples occur parallel to ridge crests irrespective of the trend of the ridge. A few scarps are coincidental with stratal trends (localities 2b and 64) but most are not (localities 9, 10 and 12; Maps 5 and 6). The diversity of strike directions of ridge-top scarps distinguishes them from the predominantly northerly striking tectonic scarps. While some of the ridge-top scarps are associated with large-scale rock slumps such as at locality 2b (Maps 5 and 6; see also Fig. 10.16), others appear as isolated features in localities where there is no other sign of slope movement, *e.g.* localities 9, 10 and 11 (Maps 5 and 6; see also Fig. 10.14).

All these ridge-top features are suggestive of downslope movement by creep within bedrock. Their formation is strongly dependent upon gravity because they always form parallel to slope contours. In most instances it is not possible to establish if translational sliding has occurred in association with them. Neither is it possible to ascertain whether a surface of rotational movement has developed independently of existing discontinuities. Where there are rock slumps with no associated ridge-top features further rotational movement is likely to result in ridge-top collapse. Ridge-top features may also be the precursor of rock slumps. Thus future large-scale, deep-seated rock slumps may be expected where ridge-top features display evidence of incipient movement.

E. Summary

Slope movements within the study area are predominantly the result of failure at shallow depths by translational sliding either along recognised bedrock discontinuities (bedding planes and faults) or along zones of shear within soil, colluvium or weathered bedrock. Movement within soil or colluvium may result from sliding along a perched water table or a thin iron pan but in most instances the zone of shear is indeterminable. Translational slides are generally less than 1m deep and the majority have been classified as debris slides. The soil-bedrock or colluvium-bedrock contact is the single most important discontinuity along which translational sliding occurs in the study area. The presence of sub-surface water, the smooth outline of the eroded bedrock surface and the physical contrast between these materials all contribute to the instability of soil or colluvium on the slightest of valley slopes. The depth of these movements varies considerably with position on slope, being shallowest (1-2m) near ridge crests where the colluvium mantle is thinnest and deepest (2-3m) at the base of valley slopes where the colluvium mantle is thickest. Translational sliding at this contact occurs independently of lithological or structural discontinuities within the underlying bedrock. Both debris slides and debris avalanches result from sliding on this contact.

Translational sliding along internal slip surfaces within the bedrock occurs at two distinct levels. At depths of less than 1m, sliding occurs within the uppermost weathered bedrock layers. There is no apparent correlation between the location of these slides and the underlying bedrock structure. Sliding usually takes place along an irregular rather than planar zone of shear. Such sliding produces debris slides, debris avalanches and small rock slides. At depths of greater than 1m within unweathered bedrock, sliding occurs predominantly along bedding plane surfaces or high angle fault planes and less frequently along low angle thrust fault planes. The resultant movements are either debris avalanches or rock slides. The scale to which these translational slides develop is largely governed by the internal organisation and attitude of the strata with respect to physiographic slope. Translational slides resulting from movement along bedding plane surfaces are most frequently associated with interbedded argillite and permeable sandstone sequences of the Tamaki Lithotype. These slides are most common where the bedded sequences consist of thin beds of sandstone with only minor argillite. Failure usually occurs at the sharpest depositional contact between these two lithologies. Here subsequent tectonic deformation has further weakened the coherence between sandstone and argillite at this contact. The paucity of planar surfaces within strata comprising the Wharite Lithotype limits the development of rock slides and debris avalanches to lithological contacts between competent lithologies such as Sandstone and Chert Lithozones and other lithozones such as Foliated, Diamictite and Argillite Lithozones.

Translational sliding within the bedrock is dependent upon the attitude of the strata and its relation to slope orientation. The most favourable locations occur where strata strike parallel or sub-parallel to slope contours and dip in the direction of slope. The least likely locations occur where the strike is perpendicular to the slope and the strata dip vertically or at steep attitudes into the hillside. Shallow stratal flexures and low angle fault planes enhance the probability of translational sliding. Joint and cleavage surfaces are not recognised as major discontinuities associated with translational sliding in this area.

Earth slides are thought to be the result of translational sliding along planar, low-dipping bedding plane surfaces.

Rock falls and rock topples are predominantly the result of failure at open joint surfaces along either east-west or northwest-southeast joint orientations. Some falls result from spalling along steeply inclined bedding plane surfaces. The structural attitude and orientation of joint surfaces with respect to slopes is a key factor in rock fall and rock topple activity. Most rock falls and rock topples predominantly occur where stratigraphically thick units of unbedded sandstone crop out and are therefore more commonly associated with sandstones comprising Wharite Lithotype. The size of these slope movements is limited by the poor horizontal and vertical continuity and spatial distribution of joint surfaces.

The discontinuity along which movement results in earth fall and earth topple activity does not appear to be a joint surface, a bedding plane or a fault plane. Nonetheless, the surfaces of failure are planar which is suggestive of lithological and/or structural control. Such movements are thought to be the result of separation, along surfaces of unknown origin, of blocks of earth material from the parent bluff purely as a consequence of gravity due to the earth materials behaving isotropically.

Debris falls are the result of failure along a zone of shear of steep attitude that develops parallel to an open face of unconsolidated free standing material. Again debris materials behave isotropically.

Large-scale rock slumps are found in catchments underlain by either the Wharite or Tamaki Lithotypes. A genetic relationship between large-scale rock slumps and faults has been established in this area. At each slump locality detailed in the text, the strike of a steeply dipping fault plane parallels the strike of equally steeply dipping lithological discontinuities (*i.e.* bedding or foliation) within the bedrock. As the structural attitude of these discontinuities rarely coincides with that of the failure surface it is apparent that rock slumps form irrespective of the direction of stratal dip or strike. In the majority of rock slumps therefore the bedrock materials behave isotropically. However, in a few cases initial translational sliding at depth either along a bedding plane or a fault plane may initiate downslope movement of a large rock mass. The mass then moves along a concave-upwards surface of failure that forms subsequently. Rotational sliding thus takes place during the latter stages of slump formation. The formation of the curved failure surface is thought to be facilitated by the presence of extensive zones of fault breccia and disrupted bedrock particularly where deformation has been intense and has resulted in destruction of the original stratal fabric. For this reason, it is concluded that rock slumps occur more frequently within zones of fault brecciated and disrupted bedrock than elsewhere within the study area. The wider the zone of fault deformation the greater is the potential for the development of large-scale rock slumps. Rock slump formation occurs both where the fault zone is parallel to or perpendicular to valley slopes. The majority of rock slumps are the result of gravitational downslope movement and not fault displacement. However, those rock slumps that occur on active Late Quaternary faults may have failed as a result of fault rupture (see Section 10.4).

Small-scale rock slumps are not controlled by structural or lithological discontinuities within the bedrock but develop in response to gravitational downslope movement of bedrock that has either been fault brecciated or disrupted as a result of previous large-scale slump movements.

Debris flows in the study area develop from large-scale rock slump movements. The debris essentially comprises large volumes of bedrock lithologies of which argillite and water content are by far the most important components.

Earth slumps are found along active Late Quaternary fault traces, potentially active contact faults and faulted monoclines in Plio-Pleistocene marine strata which dip steeply. At a few localities downslope movement is considered to have been controlled by a fault plane, bedding plane or the steeply inclined planar contact between Torlesse bedrock and overlying Plio-Pleistocene marine deposits. In each case an initial component of translational sliding is suggested. As downslope movement progresses a curved failure surface forms within these marine deposits, along which continued movement is rotational. However, in the majority of localities there is no relationship between existing discontinuities and the slump failure surface. Here, the surface of failure cross cuts existing lithological or structural discontinuties within the earth materials. It is therefore concluded that the majority of earth slumps move when the shear strength of isotropic earth materials is exceeed. They occur predominantly on oversteepened slopes where climatic and seismic factors cause weakly consolidated earth materials to fail.

Earth flows form within stream channels at the base of earth slides and comprise fine-grained earth materials. Few earth flows appear to have developed subsequent to earth slumps. Earth material is conducive to flowage whenever saturated.

Ridge-top features are suggestive of downslope movement as a result of creep within bedrock. In the majority of cases movement is thought to take place along steeply inclined lithological or structural discontinuities. However, in the majority of cases downslope movement of large tracts of hillside occurs irrespective of the structural attitude of discontinuities in the underlying bedrock. Ridge-top features are not the result of vertical fault displacement nor are they the result of differential erosion of an incompetent lithology. The diversity of strike of these features and their orientation parallel to ridge crests indicates that they are largely of gravitational origin. Movement in some instances may be translational and in other instances rotational. Where there are rock slumps with associated ridge-top features, further rotational movement is likely to result in ridge-top collapse. Ridge-top features may also be the precursor of rock slumps. Thus future large-scale, deep-seated rock slumps may be expected where ridge-top features display evidence of incipient movement.

In conclusion, the majority of slope movements in the study area involve translational sliding at shallow depth. Of these, debris avalanches and debris slides are the most important, with rock slides and earth slides being of lesser significance. Falls and topples are few in number, isolated in occurrence, small in extent and are hence of little importance. Large-scale, deep-seated and small-scale surficial slump movements involving rotational sliding are numerically small but incorporate significant volumes of material. Slumping and subsequent flow movements though currently non-active are likely to be major slope movement types in the future. Ridge-top features demonstrating incipient downslope gravitational movement of mountain slopes are likely to be sites of future large-scale, deep-seated rock slumping.

10.1.2 MINERALOGICAL FACTORS

The composition, texture and stratigraphic juxtaposition of materials of differing origins largely determines the strength of rock outcrops. Of the bedrock lithologies present in the study area the fine-grained argillites have the least competence. Slope movements involving argillaceous lithologies are, however, not necessarily associated with extensive outcrops of argillite, for these are few. Nonetheless it has been noted that there is a relationship between the degree of shearing and the stability of argillite outcrops. Sheared argillite is consistently more prone to instability than unsheared argillite, a feature also noted in the southern Ruahine Range by Smale *et al.* (1978). Thus the majority of slope movements that involve argillaceous lithologies are not the result of failure due to the mineralogical composition of argillite but are the result of a loss of coherence between argillite and competent lithologies by tectonic shearing. Slope movement types that result from failure of argillaceous material when interbedded with permeable sandstone as extensive and repetitive beds are discussed in Section 10.1.1. Coherence of these interbedded lithologies is weakest where contacts are sharp rather than graded. Where argillite is the major lithologic component within non-bedded or tectonically disrupted outcrops (Diamictite and Foliated Lithozones, respectively) it is surprisingly coherent. However, Late Quaternary tectonic shearing, much of which is concentrated within the argillaceous matrix, severely reduces the shear strength of these lithozones as a whole. Despite tectonic shearing, slope movements in these lithozones rarely result from failure along tectonic fractures within the argillaceous matrix but rather result from failure at the sheared contact between argillite and competent lithologies. Rotational rock slumping and localised streambank undercutting along stream channels comprising sheared argillite-dominated lithozones are particularly well developed.

Fewer slope movements are associated with Lithozones that comprise lithologies of greater competence, such as massive sandstone and chert. Where slope movements do occur in association with these lithologies, instability is due to the presence of discontinuities rather than to factors of mineralogical origin. The calcareous lithologies are prone to dissolution but as these lithologies are only a minor constituent of the bedrock it is not likely that slope instability is related to their presence.

Zones of non-indurated fault gouge within breccia zones or along bedding plane contacts facilitates slope movement either by acting as a lubricant if wet or resulting in slope movement if removed by fluvial erosion. Current research indicates that the presence of fault gouge along bedding plane surfaces and thrust faults is of major significance in affecting slope stability where their attitude with respect so slope is conducive to translational slide movement (see Section 10.1.1). Slope instability as a result of fluvial erosion of non-indurated fault gouge is evident throughout the study area wherever this material occurs in streambank exposures. Here, streambank undercutting results in debris slide, debris avalanche and slump movements. Surface runoff on exposed outcrops of fault gouge results in rilling and gullying and the consequent rapid removal of large volumes of this material (see Section 10.3).

Zones of indurated fault gouge are cemented predominantly by carbonate, the dissolution of which is thought to facilitate rapid removal of this material by fluvial processes.

The majority of slope movements involving earth materials are associated with outcrops of massive, predominantly fine-grained mudstone lithologies, probably for no other reason than that this lithology predominates. There is no apparent indication that differences in mineralogy are related to the stability of or to the type of slope movements associated with these materials. Nonetheless these weakly indurated marine sediments may be susceptible to instability on account of their fine-grained texture and consequent inability to support high shear stresses. Failure in these materials may be triggered by gravity, saturation during periods of heavy rainfall, seismic shaking or removal of support by streambank undercutting.

The instability of colluvium and alluvium is more closely related to their texture than to their mineralogical composition. As these materials are non indurated, permeable and coarse in texture, they are of low shear strength. Colluvial deposits are particularly susceptible to failure when saturated or shaken by seismic activity. In riparian localities scree deposits and fan or terrace alluvium are highly susceptible to collapse as a result of streambank undercutting (see Section 10.3).

10.2 PHYSIOGRAPHIC FACTORS INFLUENCING SLOPE STABILITY

10.2.1 INTRODUCTION

The geomorphology of the southern Ruahine Range is closely related to the tectonic history of the area. It comprises a topographically high NNE-trending upthrusted mass of Triassic-Jurassic Torlesse bedrock, elevated to 900-1200m a.s.l. during the Late Cenozoic Kaikoura Orogeny. Much of this uplift has taken place since Waitotaran times by upwarping and faulting. This has neither been uniform over time nor constant over the length of the study area, as is indicated by the tectonically depressed area in the vicinity of the Manawatu Gorge. Uplift has principally been confined to major fault lines that bound the eastern and western flanks of the Range (see Chapter 6). It has been greater on the eastern Wellington Fault than on the western contact faults, thereby giving the Range an appearance of having been tilted westwards. The cross-sectional profile of the Range is therefore strongly asymmetric with a smooth, even-surfaced, long western flank dipping at about 10° to the west and a short, steep eastern flank dipping at about 40° to the east. Streams draining the Range generally flow perpendicular to the NE-SW axis of the Range, but stream alignment along the NE-SW trending fault lines is conspicuous. Those streams draining the eastern flank are short and steep, whilst those draining the western flank have longer channels with gentler profiles, reflecting the asymmetric outline. Sharp crested ridges and closely spaced streams are typical of the deeply dissected flanks of the Range, with broad ridge tops being restricted to the northern end of the study area. Valleys were downcut largely in response to tectonic uplift and are characteristically V-shaped and steep sided.

There is no consistent relationship between each specific type of surficial slope movement and physiographic factors such as slope length, angle, aspect or altitude. It was therefore considered more meaningful to compare and contrast the distribution of all surficial slope movements collectively with respect to these physiographic factors. This was attempted by firstly counting the total number of scars within the study area for the time periods 1946-49 and 1974-78 from Maps 5 and 6, respectively, and secondly by grouping scars according to the contour interval in which they occurred (Table 10.2) and according to aspect (Table 10.3). Catchments on the eastern flank of the Range were considered separately from those on the western side. In addition comparisons of the distribution of erosion scars between 1946-49 and 1974-78 were attempted as a result of which several conclusions have emerged.

10.2.2 ALTITUDE

A major proportion of all upper catchment areas within the southern Ruahine Range occur at altitudes between 400 and 1100 metres a.s.l. This is approximately 500m above the surrounding lowland in the south of the area and 850m above the lowland in the north. Throughout the study area the distribution of surficial erosion scars with respect to altitude indicates that during the period 1946-49, 60% of all erosion scars occurred in the 500-900m contour interval, of which the

TABLE 10.2: Distribution of erosion scars with respect to altitude.

Contour Intervals (m)	Number of scars on western side of Range		Number of scars on eastern side of Range		Total number of scars per contour interval		Percentage of total number of scars per contour interval		Cumulative %	
	1946-49	1974-78	1946-49	1974-78	1946-49	1974-78	1946-49	1974-78	1946-49	1974-78
0 - 100	17	28	0	0	17	28	0.45	0.49	0.45	0.49
100 - 200	5	22	0	0	5	22	0.13	0.38	0.58	0.87
200 - 300	10	16	0	0	10	16	0.29	0.28	0.87	1.15
300 - 400	65	68	41	41	106	109	2.85	1.92	3.72	3.07
400 - 500	121	172	246	286	367	458	9.87	8.09	13.59	11.16
500 - 600	178	361	329	494	507	855	13.64	15.11	27.23	26.27
600 - 700	201	401	362	510	563	911	15.15	16.10	42.38	42.37
700 - 800	436	651	407	535	843	1186	22.68	20.96	65.06	63.33
800 - 900	372	710	271	443	643	1153	17.30	20.38	82.36	83.71
900-1000	272	420	175	235	447	655	12.02	11.57	94.38	95.28
1000-1100	138	189	25	40	163	229	4.38	4.04	98.74	99.32
1100-1200	33	28	3	0	36	28	0.96	0.49	99.72	99.81
1200 & over	9	7	0	0	9	7	0.24	0.12	99.96	99.93
Total number of scars	1858	3073	1858	2584	3716	5657	99.96%	99.93%		

TABLE 10.3: Distribution of erosion scars with respect to aspect.

Facing Direction of valley slopes	Number and percent of scars on western side of Range				Number and percent of scars on eastern side of Range				Total number of scars with respect to slope facing direction		Percentage of scars with respect to slope facing direction	
	1946-49		1974-78		1946-49		1974-78		1946-49	1974-78	1946-49	1974-78
North	704	38	1143	37	665	36	1168	45	1369	2311	37	41
South	212	11	250	8	319	17	271	11	531	521	14	9
East	438	24	289	10	426	23	416	16	864	705	24	13
West	504	27	1391	45	448	24	729	28	952	2120	25	37
Total number of scars	1858	100%	3073	100%	1858	100%	2584	100%	3716	5657	100%	100%

largest concentration (40%) occurred between 700-900m (Table 10.2). Less than 18% of all scars were located above the 900m contour and less than 14% occurred below the 500m contour. The same pattern is once again evident in the 1974-78 period, despite a 52% increase in the number of scars during this 28 year interval. In the 1970s, 72% of all scars occurred in the 500-900m contour interval of which 41% were located within the 700-900m contour interval. Less than 17% of all scars were located above the 900m contour and less than 12% occurred below the 500m contour (Table 10.2). A significant number of erosion scars occur within the 400-500m contour interval (Table 10.2) compared to those present at lower altitudes. This contour interval corresponds with the approximate boundary between earth materials flanking the Range and Torlesse bedrock comprising the Range and also with a marked change in gradient between the lowland and the base of the Range.

In a small-scale study James (1973) determined that most erosion scars within Pohangina catchment occurred between 800-1000m a.s.l. Figures obtained in the current study for that part of the Pohangina catchment included within James' studv area support his conclusions. Yet this appears to disagree with the findings above. This apparent disagreement is related to valley slopes within Pohangina catchment being mostly above 900m a.s.l. Thus the distribution of erosion scars within the Pohangina catchment, with respect to altitude, is skewed towards the contour interval within which the largest proportion of this catchment falls. In contrast, the majority of slopes within the current study area occur at altitudes of less than 900m a.s.l. where slope movements are confined largely to upper catchment slopes between 700-900m a.s.l. A typical example is Raparapawai catchment where Stephens (1975, p. 64) also noted that the majority of erosion scars occurred between 700-900m a.s.l.

The largest increases in the number of erosion scars between the period 1946-49 and 1974-78, in the study area, have occurred within the 500-900m contour intervals. The most dramatic increases have occurred in catchments on the western flank of the Range where the total number of scars increased from 1858 in 1946-49 to 3073 in 1974-78 (Table 10.2), an increase of 65%. Erosion scar numbers on the eastern flank of the Range during this time period increased from 1858 (*sic*) to 2584 (Table 10.3), an increase of 39%.

10.2.3 SLOPE ASPECT

As the major tributaries draining the Range flow either towards the east or west, the majority of valley slopes are north- or south-facing. Only where minor tributaries and fault aligned reaches of some major tributaries drain to the north or south are valley slopes east- or west-facing.

The overall distribution of erosion scars with respect to aspect indicates that the greatest number of scars are to be found on north-facing valley slopes (Table 10.3). The second largest number occur on west-facing slopes and the least number of scars occur on south-facing slopes. This pattern was evident in the 1940s and is further enhanced by the greater number of scars present in the 1970s (Table 10.3). In particular, slopes with a northwesterly aspect appear to be the most severely eroded . Scar counts indicate that 62% of all scars present in the 1940s and 78% of scars present in the 1970s occurred on north- and west-facing slopes. This pattern is the same on both sides of the Range (Table 10.3), but along the eastern flank of the Range there is also a secondary mode with substantial numbers of scars occurring on slopes with a northeasterly aspect. This observation was also made in the Raparapawai catchment (Stephens, 1975) and a sub-catchment of West Tamaki River (Hubbard, 1978).

The greater instability of north- and west-facing slopes in the past is apparent on comparing the stage of development of the vegetation on these slopes with that on south-facing slopes. On north-facing slopes, successive episodes of slope movement are apparent. These range from the most recent movements represented by areas of bare ground, to slip surfaces colonised by weeds and grasses, to those colonised by shrubby species and finally to those covered mostly by shrubby species with a few tall emergent tree species beginning to protrude. On south-facing slopes the vegetation cover is dominated by forest trees. The greater instability of north-facing slopes is also indicated by the predominance of valley slopes that are less steep than those on south-facing slopes (Tables 10.4a and b). Research in the northern hemisphere, where the above situation is reversed, has indicated that variations in microclimate, such as differences in altitude, exposure to moisture-bearing winds, and exposure to sunlight, can cause significant differences in geomorphic processes. South-facing slopes of east-west valleys in the northern hemisphere are less steep than adjacent north-facing slopes.

TABLE 10.4a: Results of measurement of slope gradient, determined from unpublished 1:25000 topographic maps, within the study area. Number of readings = 221.

Catchments on western flank	Average slope gradient (degrees)	Catchments on eastern flank	Average slope gradient (degrees)
Piripiri	26	Mangatera foothills	22
Pohangina	30	East Tamaki foothills	30
Konewa	29	West Tamaki	32
Makawakawa	29	Rokaiwhana	33
Te Ekaou	27	Otamaraho	33
Porewa	30	Mangapuaka	33
Opawe	29	Otamarahu	32
Ohinetapu	27	Mangapukakakahu	32
Dundas Creek	26	North Oruakeretaki	30
No. 1 Line	29	South Oruakeretaki	31
Mangatuatou	26	Raparapawai	27
No. 2 Line	30	Manga-a-tua	28
Te Awaoteatua	22	Coppermine	31
Tokeawa	23	Mangapapa	23
Makohine	24		
Whareroa	25		
Maungatukurangi	30		
Manawatu Gorge	30		
Mean	27	Mean	30

TABLE 10.4b: Measurements of slope gradient with respect to aspect.

Facing direction of valley slope	Average slope gradient (degrees)	
	Western flank	Eastern flank
North	26	29
South	29	31
East	27	31
West	26	29

There, north-facing slopes have snow cover longer, experience fewer days of freeze and thaw, retain their soil moisture longer, and have a better vegetation cover, all of which result in less active erosion and steeper slopes (Rib & Liang, 1978; p 35).

In the southern Ruahine Range the most significant factor contributing to greater slope instability on north-facing slopes is considered to be the dominant north-westerly direction of major climatic elements experience by this area. Between the period 1946-49 to 1974-78 there was a 52% increase in the number of erosion scars. On the western flank of the Range the greatest increase in number of new erosion scars occurred upon north- and west-facing slopes. On the eastern flank of the Range the greatest increase in number of erosion scars during this period occurred on north-facing slopes and to a lesser degree on west-facing slopes (Table 10.3). There is therefore a close relationship between the highest incidence of erosion scars and northwesterly slope aspect.

Episodes of increased instability within the study area are known to be influenced by climate. Of particular importance is the periodic occurrence of tropical cyclones that approach the Range from the north-west. This relationship is discussed further in Section 10.3.

10.2.4 SLOPE GRADIENT

A total of 221 slope gradients were measured throughout the study area. The average slope gradient within catchments on the western flank of the Range is 27°, and for catchments on the eastern flank of the Range 30° (Table 10.4a). Steeper slopes of between 40° and 50° were found adjacent to active and ephemeral stream channels in most catchments throughout the Range (Stephens, 1975; Mosley, 1977).

Valley slopes in headwater areas of catchments show considerable variation in steepness. In the northeast of the study area where valley dissection is greatest and has in places reduced the dividing ridge between east- and west-draining catchments to a sharp crest, slope gradients of between 35° to 40° may be found. Less steep headwater valley slopes occur in catchments in the south. In contrast, some catchments in the northwest of the study area are little dissected in their headwater reaches and slopes of between $4\text{-}20^{\circ}$ are characteristic.

Catchments on the western flank with valley slope gradients consistently steeper than the average include West Tamaki, Rokaiwhana, Otamaraho, Mangapuaka, Otamarahu and Mangapukakakahu. The majority of slopes in these catchments average 32-33^{o} (Table 10.4a).

Irrespective of altitude, slopes of less than 25^{o} appear to be considerably more stable than steeper slopes. On such slopes there are fewer surficial erosion scars, either at low altitudes, *e.g.* foothill areas at the base of the Range or at high altitudes, *e.g.* the crest of the Range in Konewa catchment, than on steeper slopes at similar altitudes. The majority of erosion scars occur on slopes of 25-50^{o}, predominantly in steeper than average localities, *e.g.* adjacent to drainage channels and at all altitudes.

10.2.5 SLOPE LENGTH AND DISSECTION

The length of valley slopes is related to aspect and to depth of valley dissection; consequently the longest valley slopes are found in the largest catchments and predominantly on north- and west-facing slopes. Conversely, many of the shorter valley slopes are found in catchments that are weakly dissected and with a predominant southerly or easterly slope aspect. The incidence of erosion scars is proportional to slope length, with the longer valley slopes having better developed drainage networks, along which erosion scars are concentrated, than short valley slopes. That is, the denser the drainage network on valley slopes the greater the incidence of slope movement.

It is particularly noticeable that catchments, including Konewa and Te Ekaou, within which valley dissection along a small number of major tributaries is minimal, are among the least eroded within the study area. The major tributaries in these catchments have short, steep valley slopes immediately adjacent to the stream channels where the majority of erosion scars occur, above which undissected and uneroded slopes of gentler gradient ascend towards the ridge crest.

10.2.6 COMBINED AFFECTS OF PHYSIOGRAPHIC FACTORS, PARENT MATERIALS AND VULNERABLE LOCATIONS

The combined affects of unstable parent materials and/or certain vulnerable locations together with physiographic factors is considered to have greater influence upon the occurrence and distribution of surficial erosion scars than physiographic factors alone.

The majority of surficial erosion scars are the result of failure of thin deposits of unconsolidated, coarse slope colluvium (see Section 10.1.1). Colluvium is in sharp contact with underlying bedrock lithologies, with which there is little cohesion, and is consequently unstable on slopes in excess of 25^{o} steepness. The same material on lesser slopes is relatively stable.

Colluvium and bedrock materials disrupted either by faulting or by large-scale slump movement are less stable than undisrupted materials. Where disrupted materials coincide with either oversteepened fault scarps or headwall and lateral scarps of slumps further instability occurs in the form of surficial erosion scars. At sites of fault rupture and/or slump movement the combination of steep slopes and weak materials is highly conducive to instability. This is apparent in the study area where surficial erosion scars at some localities appear to be concentrated and indeed delineate fault traces and slump outlines.

10.3 CLIMATIC FACTORS INFLUENCING SLOPE STABILITY

10.3.1 INTRODUCTION

The climate of an area is one of the most important factors influencing slope movement. The effects of rainfall, temperature, wind, snowfalls, relative humidity, and barometric pressure can seldom be evaluated individually because their relationships are too complex. At best one has to make empirical correlations of one or more climatic factors with episodes of slope movement. Many factors involve the presence of water (Varnes, 1978), the most accessible information being that of rainfall. Rainfall may directly influence slope stability both in the short term during periods of high intensity rainfall or in the long term by contributing to storage.

Although slope instability is closely related to rainfall amount, duration and maximum intensity, equally important is the condition of the colluvium, soil and vegetation prior to a period of rainfall. For example: (1) the presence of shrinkage cracks not only reduces soil strength but also greatly influences the rate of infiltration; (2) the presence of a high water content is often accompanied by a decline in slope stability; (3) decaying root systems reduce soil strength with a resulting decrease in slope stability; and (4) an already saturated soil may promote increased runoff.

Whilst many climatic factors have been recorded as seriously damaging forest vegetation (see Section 10.5.3) only wind and heavy rainfall can be directly associated with incidences of slope instability in the study area.

10.3.2 WIND AND HEAVY RAINFALL

Examples of windthrow involving the uprooting of large trees in the study area are rare and the resulting debris slide is small and shallow. Further instability at these sites is not common as the exposed debris slide scar heals quickly.

The predominance of patches of sheet erosion at the heads of west-facing debris slides and avalanches suggests prevailing winds may be a prime factor in causing sheet patches to retreat headward beyond slide outlines. Evidence of sheet erosion is most prominent in upper catchment areas above the forest line (900m altitude). Some of the better examples occur at the head of Whareroa catchment immediately below Wharite Trig., reputed to be one of the windiest places in New Zealand. Sheet erosion became extensive during the 1940s in the pastured foothill area surrounding East Tamaki and Mangatera catchments where the original forest vegetation was cleared in the early 1900s. The maximum effects of deforestation were therefore in evidence some years after the initial clearance. By the late 1970s most of the earlier sheet erosion scars had healed so that today the incidence of sheet erosion is very much reduced. Sheet erosion scars covered approximately 23.5 hectares (2.2% of upper catchment area) of Mangatera catchment in the mid- to late 1940s. By the 1970s this area had reduced by 78% to 5.1 hectares (0.48% of upper catchment area) (see Chapter 12). Here, increased stability probably resulted from better grass cover and improved pasture management.

Grant (1965, 1966, 1969, 1971) has noted the importance of high-intensity rain storms as factors triggering slope movement. He has been a leading advocate of the hypothesis that periods of increased erosion are closely related to periods of increased storminess. Based on depositional terraces in the Tukituki River, just 27 km to the north of the study area, Grant (1965) recorded several broad erosion periods the earliest of which dates back to 1650 AD. The latest of these periods is thought to have begun prior to European settlement probably a little before 1800 AD. In historic times, the maximum daily rainfalls recorded at 16 stations were used by Grant (1966) to determine the relative storminess of each decade since 1900. On this basis the periods of greatest potential for mass movement activity on a regional scale were ranked in descending order of magnitude as: (1) 1931-40; (2) 1911-20; (3) 1951-60; and (4) 1961-65. From this data Grant concluded that the 1931-60 period was one of greater potential for mass movement activity than the 1901-30 period. He also postulated that for the Tukituki River region, small area rainstorms had increased in intensity during the 1931-60 period (Grant, 1965).

One of the most important features of the climate is the passage of cyclonic storms across the Range. These storms come from a northwesterly direction, bringing heavy rains that may last from two to four days. They are often locally very intense and may produce daily rainfalls of over 300 mm. Of interest is the observation made by Cunningham & Arnott (1964) of the effect of a heavy rainfall in the Rimutaka Range during 1962. They concluded that during rainfall intensities in excess of 200 mm/day and 25 mm per hour, "one of the most alarming features . . is the discovery that such rains can cause erosion under forest conditions which we at present regard as reasonably satisfactory. Serious damage occurred in three catchments, in all of which the forest cover and soils were in much better condition than many we are familiar with in the Aorangi, Tararua, Ruahine and Kaweka Ranges".

Further evidence to support a direct relationship between high intensity rainfall and the incidence of slope movement comes from the south Auckland - Waikato Basin area (Selby, 1967, 1976), the eastern Hunua Range (Pain, 1968) and the Ruahine Range (James, 1973; Stephens, 1975).

From historical records of climatic and river gauging data, together with eye witness accounts, it is apparent that periods of intense rainfall in the study area are associated with episodes of slope movement and subsequent transportation of debris from the Range to the lowlands during periods of peak flood flow. Between the years 1930-60 at a time when erosion within the Range began to be noticed and problems encountered with excess streambed detritus, the Manawatu River recorded twenty flood flows (>50,000 cusecs) (Cowie & Osborn, 1977). In using this information it is stressed that very heavy local storms within the Range are likely to do considerable damage there, without causing a significant rise in the Manawatu River level at Palmerston North.

Conversely, high flood flows in lower catchment reaches do not necessarily imply heavy rainfall in upper catchment reaches of the Ruahine Range as intense rainfall may have occurred in adjoining lowlands. Figure 10.20

shows flood flows (>200 cumecs) recorded in the upper Manawatu River catchment at Weber Road, less than 20 km from the eastern flank of the southern Ruahine Range. Here, a flood flow of only 287 cumecs was recorded in March 1975 during the passage of Cyclone Alison, yet considerable slope movement occurred during this storm event within the Ruahine Range. In contrast, a storm in 1976 produced a higher flood flow of 392 cumecs at Weber Road yet slope movement within the Range was markedly less severe than that which occurred during the 1975 storm. In the years 1976-83 no fewer than 14 flood events with flows greater than that measured during Cyclone Alison have been recorded. During this time period slope movement within the Range has been negligible.

It is apparent, therefore, that since 1975 there has been little correlation between flood flow levels at Weber Road and the incidence of slope movement within the Range. In the absence of other confirmatory evidence flood flow levels, measured downstream of the Range front, should therefore not be used to infer the relative potential of storm events to trigger slope instability within the Range. Of greater relevance to determining slope stability threshold values are measures of rainfall duration and intensity, information on soil moisture content prior to storm events and the location of storms. However, such information is generally unavailable in steepland terrain.

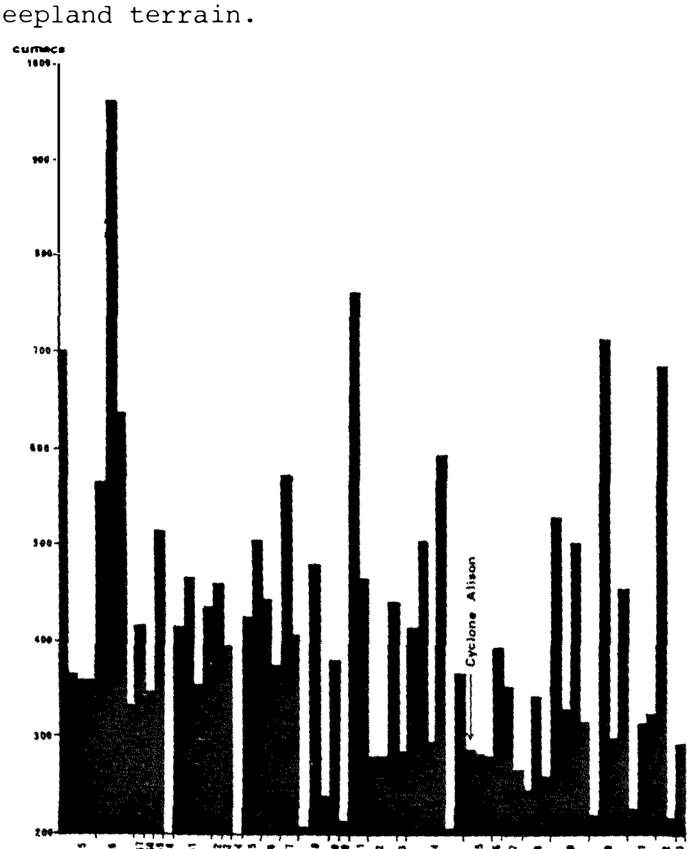


FIGURE 10.20:

Frequency and magnitude of flood events (>200 cumecs) per year (1955-83) recorded in the upper Manawatu River catchment at Weber Road (U23/750027). Data provided by Manawatu Catchment Board.

The effects of several episodes of heavy rainfall in the southern Ruahine Range upon a man-made 'drain' designed to divert flood water from Mangapuaka (Kumeti) Stream into an adjacent stream, have been documented by Hubbard (1976). This drain was initially dug in 1920 to 1m depth and 3m width. Exceptionally heavy rainfalls in this area during 1920 and 1922 resulted in high flood flows during which detritus was transported from the Range to adjacent pastured lowland and the drain was scoured to 8m depth and 20m width. Heavy rainfall and consequent flooding in 1936 resulted in the infilling of the drain to a depth of 4m (Hubbard, 1976). Subsequent episodes of flooding during 1942 and 1975 caused infilling of the drain with resultant overbank deposition of detritus upon surrounding farmland. The source of this detritus was the upper reaches of Mangapuaka and Otamarahu catchments where episodes of valley slope collapse were triggered by these heavy rainfalls. An eye witness account by a local resident, Mr G Miller, indicates that the cloudburst in February 1936 caused substantial damage in the upper Mangapuaka and Otamarahu catchments. He '. . . observed massive erosion, with "whole mountain sides having slipped into the stream." Twelve metre high debris flows had buried trees along the sides of the upper valley; and debris was dumped in the channel right down to the picnic area. All the headwater streams were choked with debris . . ' (Mosley, 1977).

The most recent storm events with which slope movements can be directly correlated occurred in 1975 during the passage of Cyclone Alison across the Range and in 1976 following several days of continuous rain of moderate intensity (Mosley & Blakely, 1977).

It appears that there was little sign of erosion in the southern Ruahine Range prior to the 1930s (see Section 10.4). However, from this time on huge slips began appearing in what had been well bush-covered ranges. The effect of the increased detritus, as bedload, on the stream channels soon became of major concern. Damage, resulting from transportation of this detritus onto lowland farms, escalated during periods of increased storm activity. An indication of the severity of the problem is the cost (past, present and future) of efforts to curb and control both the incidence of slope movement and the movement of detritus down stream channels (Appendices Ia, Ib, IIa and IIb).

Because north-westerly slope aspect coincides with the greatest erosion it is considered that the frequency, duration and intensity of periods of heavy rainfall from the northwest (often cyclonic) are the major contributory factors in promoting slope instability and in triggering slope movements in this area. Considerable supportive evidence has been documented from neighbouring ranges with similar physiographic characteristics and more importantly from the southern Ruahine Range itself.

10.3.3 WEATHERING ON EROSION SCARS

The climate of New Zealand is renowned for creating some of the most rapid weathering in the world (Birkeland, 1974). The present climatic conditions within the study area are highly conducive to rapid weathering of bedrock. Fine-grained moderately permeable rocks such as argillites and some sandstones are the most susceptible to physical weathering. Repeated cyclic wetting-drying, freeze-thaw and heating-cooling causes expansion and contraction to produce tensional fatigue which eventually results in rock failure. These cyclic processes are largely dependent upon the duration and amount of moisture, the frequency with which temperatures fluctuate through freezing point, relative humidity and exposure to prevailing winds which ultimately will determine the rate at which physical weathering takes place. Variations in the frequency of these processes is largely a function of aspect.

Fragmentation of outcrops at high altitudes is likely to occur at a faster rate due to a more severe climate than outcrops on sheltered slopes at lower altitudes. Some fragmentation of argillaceous materials may be predetermined by numerous cleavage surfaces. However, it is not always cleavage controlled, as evidenced by many fracture surfaces that are conchoidal. The rate at which these processes occur to cause fragmentation is largely unknown. Quantitative measurement of material removed from exposed bedrock was beyond the scope of this project.

Water on freezing undergoes about a nine percent volume increase and exerts tremendous pressure when in a confined space. Freezing of water in joints and along bedding planes is thought to account for many rock falls. Freezing of pore water in soil and colluvium on eroded surfaces produces needle-ice, the effect of which promotes downslope movement of particles of soil and rock (Zotov, 1940; Soons, 1967). This process was observed to be most effective in headwater areas of catchments where sheetwash and wind erosion processes are also active.

Weathering by dissolution of soluble minerals may be an important causitive factor in some slope movements. Petrological analyses indicate a high proportion of soluble carbonate in some lithologies. However, slope movements involving these lithologies have not been recognised. Dissolution of carbonate cement from fault breccia zones may weaken outcrops sufficiently to initiate slope movement. Rock falls in lithologies with calcite filled joints were observed, but failure is not proven to be due to dissolution of calcite.

10.3.4 SURFACE WATER

Next to gravity, water is the most important single contributing factor in most slope instabilities. Identificaton of the source, movement, amount of water, and water pressure is as important as identification of the soil and rock strata. Water conditions are, however, subject to fluctuations controlled by the weather and on the cumulative effects of rainfall, snowmelt, surface runoff, infiltration, evaporation and transpiration throughout the year and during long-term climatic cycles.

A. Surface Runoff and Rilling

Subsequent to an inital slope movement, rainsplash and surface runoff processes scour the bared soil, colluvium or bedrock of an exposed scar. Their importance is a function of rainfall amount, intensity and duration, water falling directly on the scar plus sub-surface water from upslope flowing down the scar to concentrate in rills. Here the term rill is used to describe sub-parallel channels of relatively shallow dissection which in the study area occur in colluvial materials. Many rills drain the upper parts of scars, which downslope usually merge into a master rill of greater depth and width (Fig. 10.21). Rills are indicative of rapid incision, transportation and deposition. Continued incision results in deeper dissection that cuts into the underlying bedrock to form gullies.

Debris slides, debris avalanches and slumps are particularly susceptible to rilling as the initial scar is often of sufficient areal extent to involve a considerable amount of surface runoff. Consequently, surface runoff processes remove large quantities of slope detritus from exposed scars and inhibit the establishment of a vegetation cover, thus perpetuating scar instability.

FIGURE 10.21: Rilling within colluvium at the head of a debris slide scar. Note how the smaller rills near the top of the scar merge further downslope into a single master rill. Photograph taken in Otamarahu catchment at locality T23/612107.

B. Gully Development

1. Description

Gullies within the study area are seldom more than 5m wide and vary in depth from 4m to 10m. They are characteristically steep-sided, of restricted length (20-100m) and chute-like in appearance. Water flow in gullies is predominantly ephemeral with episodes of dissection and transportation of detritus occurring mainly during the wet winter months and less frequently during periodic storm events in the drier summer.

2. Origin of Gullies

Gullies are rapidly developed erosional features that often result from changes in the environment such as faulting, climatic changes affecting vegetation, extreme storms or other factors that cause a break in the vegetation cover which exposes the underlying unconsolidated colluvium.

Two main erosional processes appear to generate gullies; slope movement followed by surface runoff. Debris slides and debris avalanches usually precede gully development. These slope movements and subsequent gully development within the study area occur concurrently during periods of intense or prolonged rainfall. It has also been noted that headward erosion within some existing gullies may cause further gullying and slope movement.

3. Factors Influencing Gully Development

Gully development is largely influenced by: (1) attitude of the strata in relation to slope direction; (2) slope steepness and length; and (3) the presence of fault zones.

The best examples of gully development are found where bedrock consists of thin-bedded alternating sandstones, siltstones and argillites, particularly where the strata strikes at right angles to the contours of a slope and dip is near vertical. Such is the case in the majority of catchments to the north of Mangapukakakahu Stream where gullies have incised within bedrock of the Tamaki Lithotype. Here gully development is aligned approximately north-south parallel to strike and perpendicular to the principal west- to east-draining catchments. As a result of this structural control, gullies tend to be remarkably straight throughout their length. Here, gully length is restricted because valley slopes are steep and short. The steepness of the near vertical gully walls is maintained by the spalling of vertically inclined bedrock (Fig. 10.22). Gully development within bedrock comprising Wharite Lithotype is restricted to less resistant, indurated Foliated and Diamictite Lithozones. Here, gullies often develop in an approximate northeast-southwest direction parallel to the strike of bedrock foliation. However, their development is not always structurally controlled because some gullies cross-cut the foliation strike direction. Where gully walls coincide with and parallel steeply inclined faces of large blocks of resistant lithologies they may be near vertical and planar in outline. In other localities where mass movements or streambank undercutting predominate within less resistant lithologies, gully walls tend to be less steep and non-planar in outline (Fig. 10.23).

Gully development is often found in association with faults. At some localities gullies develop within narrow (1-5m width) zones of fault gouge. Most of the gouge zones are steeply dipping, hence fault gullies tend to be straight, narrow, deep and confined between near vertical walls of non-faulted lithologies. Elsewhere, extensive zones of fault brecciated bedrock exceeding several hundred metres in width and within which the bedrock has been homogenised are also sites of gully development. Many of these zones are also prone to slumping (see Section 10.1.1) at the toe of which gullying is often developed. These gullies are deeply incised, irregular in outline, short in length and steep sided. Good examples are found in the Manga-a-tua (T23/569040), Raparapawai (T23/564072) and Piripiri (T23/679256) catchments.

C. Streambank Undercutting

Removal of material from stream banks results in widening of a channel and sometimes removal of toe slopes that adjoin drainage channels. This process is particularly evident in the southern Ruahine Range. Here, the large number of slope movements that occur adjacent to stream channels can be attributed to undercutting by fluvial erosion. Riparian slope movements usually involve loose, unconsolidated materials that are easily removed by fluvial erosion. These include: (1) alluvial gravel terrace or fan deposits; (2) colluvium from previous slope movements; (3) fault crushed and fault loosened bedrock; and (4) disrupted bedrock previously involved in large-scale mass movements, particularly slumps. Successive failures account for the slow regeneration and stabilisation of riparian strips. Undercutting of riparian slopes comprising coherent bedrock is rare. Undercutting may occur where a major dis-

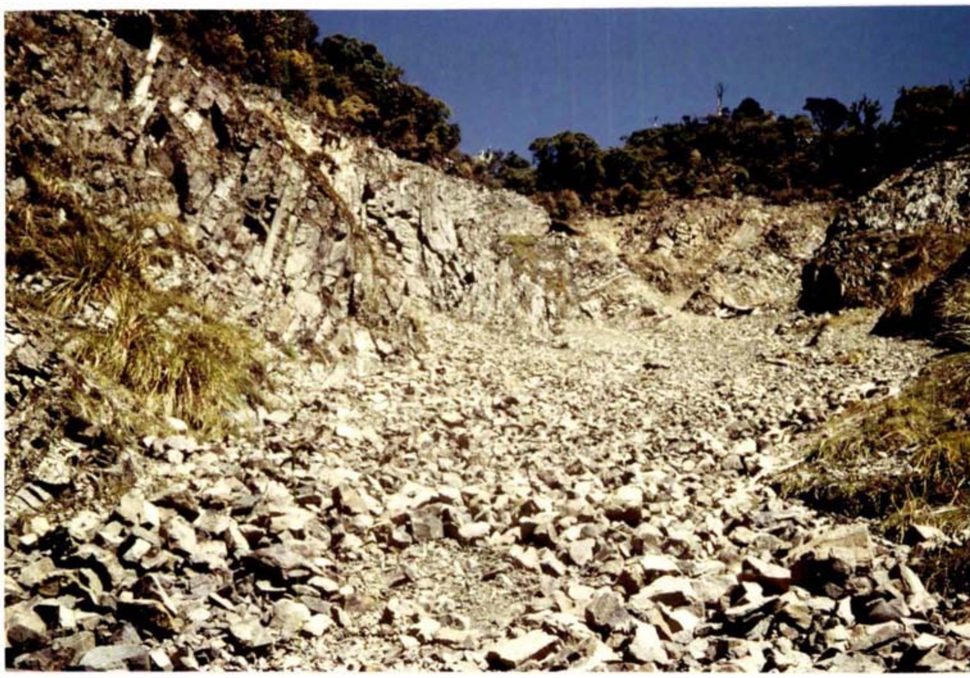


FIGURE 10.22: A typical example of structurally controlled gully erosion developing parallel to the NE-SW strike of vertically inclined thin-bedded strata comprising the Tamaki Lithotype. This gully has developed downslope of a debris avalanche scar from which a water seep emerges. Drainage is by sub-surface flow beneath coarse, angular detritus lining the floor of the gully. Photograph taken in West Tamaki catchment at locality T23/679184.

FIGURE 10.23:

An example of gully erosion within bedrock strata comprising the Wharite Lithotype. This ephemeral gully is bordered by steep sided slopes that are susceptible to mass movement and are consequently bare of vegetation. Photograph taken in Coppermine catchment at locality T23/551041.

continuity such as a bedding plane, fault or joint strikes parallel to the stream channel.

The dominant types of slope movement resulting from streambank undercutting within the Range are debris slides, debris avalanches, small-scale slumps, debris falls and very rarely rock falls. While streambank undercutting is known to rejuvenate slope instability, in many instances it is not the cause of initial failure. For example, the initial failure of most large-scale slumps is not thought to be the result of streambank undercutting but subsequent rejuvenation of slope instability in the toe region of these slopes is.

Many riparian slope movements that involve Plio-Pleistocene marine deposits along the flanks of the Range are largely the result of streambank undercutting. Here, earth falls, earth topples and earth slumps predominate.

10.3.5 GROUNDWATER

Groundwater can be defined broadly as all water below the ground surface. However, the term may be restricted to water that is not restrained in soil by capillary tension, or linked to soil or rock minerals. Although soil capillary moisture, absorbed water and water of hydration may not be considered true groundwater, they are part of the total ground moisture system. Groundwater has been more narrowly defined in Sowers & Royster (1978) as that part of the soil-rock-water system that is free to move from point to point under the influence of gravity. This definition is used here.

An evaluation of the effect of soil capillary moisture, absorbed water and water of hydration upon slope stability was not attempted. The importance of such water in promoting disintegration of bedrock outcrops is, however, recognised in the study area, some aspects of which have been outlined briefly in Section 10.3.3.

Free groundwater is regarded as a major causative factor in most slope movements within the study area. Evidence for this includes the occurrence of slope movements at localities where groundwater seeps and springs emerge at the surface. These occur in areas where structural discontinuities within the bedrock facilitate movement of groundwater to the surface. In the study area emergent groundwater and slope instability have been observed in conjunction with: (1) joint surfaces in massive or bedded sandstone; (2) bedding plane surfaces in sequences of alternating sandstone, siltstone and argillite; (3) within fault breccia zones; and (4) at the base of upthrown fault scarps.

(1) Jointed sandstone bluffs are frequently sites of rock fall activity, due to failure along one or more prominent joint surfaces (see Section 10.1.1). These joint surfaces are sometimes sites of emergent groundwater, usually highlighted by hydrophilic vegetation on an otherwise barren outcrop. Considerable volumes of pressurised groundwater may occur in these joints particularly at depth, such as that encountered in open jointed greywacke sandstone in the Kaimanawa Range 100 km to the northwest of the study area by Hegan (1980). Whilst groundwater pressure is likely to be substantially greater at depth than at the surface it is possible that surface pressures may be sufficient to overcome the frictional resistance of joint surfaces to result in rock fall activity. Within the Manawatu Gorge drill holes have been bored into sandstone outcrops at localities, where rock falls are currently active, in an attempt to reduce groundwater pressures and thereby minimise the incidence of rock falls. Here, groundwater flow is seen to be climatically influenced, with periods of highest groundwater flow occurring during wet winter months, the time of greatest rock fall activity.

(2) It has been noted that where a water-carrying stratum crops out on a valley slope, the incidence of slope movement is notably greater and the size of movement larger than at localities where groundwater seeps are absent. The best examples occur in association with bedded sequences of strata comprising the Tamaki Lithotype where groundwater penetrates along one or more steeply dipping bedding planes. The intersection of an aquifer with the ground surface produces either a concentrated flow in the form of a spring or a diffuse flow in the form of a seep. Flow is highly variable though generally greatest during wet winter months. The high incidence of slope movements, such as debris slides and debris avalanches, that are coincidental with groundwater seeps and springs and which predominantly fail during periods of greatest groundwater flow has led to the conclusion that emergent groundwater plays an important role in the initiation of many shallow translational movements.

(3) In the study area substantial volumes of groundwater emerge along

major fault breccia zones. No measurement of the volume of water involved was attempted during this study. Nonetheless, it is of interest to note that during the excavation of the Rimutaka Deviation through Torlesse bedrock, substantial volumes of groundwater were encountered at depth within fault breccia zones (Keller, 1954). Similarly, during the excavation of the Moawhango to Tongariro tunnel, in Torlesse bedrock that in part comprises mélange, groundwater flowed into the tunnel at an average rate of 105 litres/sec/km (Hegan, 1980).

Emergent groundwater in fault breccia zones is known to promote slope instability in the study area. For example, within the Ruahine Fault breccia zone at locality T24/656072 emergent groundwater has resulted in rilling and gullying. Where gully incision has been greatest shallow slope movements have occurred adjacent to the gully. Whether or not the presence of emergent groundwater facilitated the initial development of large-scale slumps within these breccia zones or has promoted subsequent slump movement is unknown. It is, however, considered possible that high groundwater pressures at depth and substantial groundwater flow at the surface in conjunction with the greater porosity of fault brecciated bedrock may be major causal factors of some large-scale slumps. This would explain the association of these movements with extensive areas of brecciated bedrock along major fault zones.

(4) The emergence of groundwater as springs at the base of fault scarps is a particularly common feature associated with the Late Quaternary trace of the Wellington Fault (Map 4). Groundwater escapes to the surface presumably by seepage up the high angle fault plane. These springs generally flow all year around and where drainage is impeded they form fault ponds or areas of swampy ground on the downthrown side of the fault trace. Where the emergent water flow is substantial, slumping of the scarp may occur adjacent to the spring.

Within the study area, slope movements involving colluvium and bedrock materials frequently occur in areas where groundwater is seen to emerge at the surface via a structural discontinuity. The importance of these structural discontinuities upon slope stability has been stressed in Section 10.1.1. Groundwater seeps and springs along lines of structural discontinuity coinciding with a high incidence of slope movements is here considered clear evidence that bedrock structure and groundwater together are major causal factors of many slope movements in the study area.

10.4 SEISMIC FACTORS INFLUENCING SLOPE STABILITY

10.4.1 INTRODUCTION

New Zealand lies across the boundary between the Indian and Pacific plates. The boundary is a 200 km wide zone that cuts diagonally across New Zealand at an azimuth of 050° from the Hikurangi active margin in the northeast to the Fiordland margin in the southwest. Within this boundary zone considerable deformation is being accommodated along a 70-100 km wide 'axial tectonic belt' This deformation is a result of strain generated by compressive forces from the eastward movement of the Pacific plate relative to the Indian plate at a rate of 50 mm/yr (Walcott, 1978). Two types of deformation in crustal rocks occur within this boundary zone: a seismic, elastic deformation resulting in fault displacements; and an aseismic ductile flow. The relative proportions of these varies along the length of the boundary zone.

The most abundant and detailed information on the nature of the boundary zone and its internal deformation comes from present-day observations of its vertical movements, patterns of horizontal shear and its seismicity. Large and rapid amounts of vertical Quaternary uplift of the central mountain system of the North Island, of which the Ruahine Range is a part, are inferred to be a response to the compression across the boundary zone. Measured rates of compression are considered to be sufficient to induce observed rates of vertical movement (Walcott, 1978).

Internal deformation is indicated by horizontal shearing, much of which occurs along slip surfaces within the boundary zone that closely parallel the direction of plate movement (Walcott, 1978; p 148).

There is however a clear discrepancy in the amount of horizontal movement observed to have taken place on major faults and the amount of movement deduced from plate tectonics and from strain measurements. The rate of horizontal displacement across faults has been estimated by dividing the offset of river terraces across the fault trace by the inferred age of the terrace. Considerable disagreement centres around the age of specific offset aggradational terrace surfaces. Consequently estimates of the rate of horizontal displacement for parts of the southern North Island vary between 24 mm/yr (Wellman, 1972) and 13 mm/yr

(Suggate & Lensen, 1973). The currently accepted plate tectonic rate and the rate calculated from strain measurements indicate horizontal displacement rates of 45-50 mm/yr (Walcott, 1978), that is three times the rate of Suggate & Lensen (1973) and twice that of Wellman (1972). This anomaly is explained by some form of deformation other than by displacements on major faults, and is presumably distributed over the width of the axial tectonic belt on numerous minor faults, by rotation of blocks between the major faults or by ductile flow in the crust (Walcott, 1978).

Seismicity is indicative of active deformation and the distribution of crustal earthquakes in New Zealand indicates that deformation is widespread throughout the boundary zone. The scattered nature of seismic activity is evident on both regional and local scales for both large- and small-magnitude events. The historically large earthquakes in New Zealand have been concentrated near the axis of the boundary zone (Eiby, 1975). Since most of the relative displacement in an earthquake zone occurs during these earthquakes (Brune, 1968), then the most rapid seismic deformation is probably occurring along this axis.

Aseismic deformation along this axis within closely spaced belts of shear may be greater still. The location of these belts of shear is largely unknown but may correspond to areas of mélange terrane where a large component of the deformation, as horizontal shear, results in rotation, bending and flowage of lithologies comprising Torlesse bedrock (see Chapter 7).

The following section outlines the influence of seismic activity within this zone upon the southern Ruahine Range and examines its role in relation to slope stability. Slope stability can be greatly altered during seismic activity, either through ground displacement by earthquake-generating faults, or through secondary movements such as slope failures due to seismic shaking.

10.4.2 UPWARPING AND FAULT UPLIFT

The present day relief of the southern Ruahine Range is largely the result of a combination of tectonic upwarping and successive episodes of vertical fault uplift, much of which has taken place in Quaternary times. It is unlikely that tectonic uplift has been uniform over the whole Range, but rather is considered to be greater in the north of the study area than at the Manawatu Gorge where the 'Manawatu Saddle' represents minimal uplift rates. Tectonic uplift along much of the axial ranges of the North Island is thought to exceed 7 mm/yr (Wellman, 1967).

The cross-sectional asymmetry of the Range is suggestive of greater fault uplift along the steeper eastern than the western side of the Range. The south and westward dip of the surface of marine planation in the north of the area indicate a general tilting of the block westward. The asymmetry of the Range matches the asymmetry of the Pohangina Anticline lying immediately to the west of and parallel to the Range (Te Punga, 1957), uplift on which is still occurring at a suggested rate of 1.1 mm/yr during the last 0.8 million years (Boellstorf & Te Punga, 1977).

In the southeast of the study area at T24/517945 an Ohakean aged surface has been vertically displaced 12m giving an average rate of displacement since this time of around 0.9 mm/yr. In the northeast of the study area successive generations of truncated spurs indicate that vertical fault displacement along the eastern flank of the Range in pre-Late Quaternary times has been considerable, recurrent and episodic. However, the rate at which vertical fault displacement occurred during these times in this area is unknown (see Chapter 6).

Several authors have indicated that there is less active faulting in this area than is suggested by Kingma (1962) or might be expected from a tectonically active region and have thus come to the conclusion that perhaps much of the uplift of the Range may have to be accounted for by upwarping rather than faulting (Lillie, 1953; Smale *et al.*, 1978). Whatever the mechanism it would appear that tectonic uplift has not yet ceased.

Previous reports discussing slope stability and erosion have either emphasised <u>rapid uplift</u> of the Range as a significant factor in maintaining the youthfulness of streams and in continuing the flow of gravel out of the Range (Mosley, 1977), or suggest that present day active erosion may be a consequence of <u>recent increases in the rate of uplift</u> (Cunningham & Stribling, 1978; p 6). Uplift of the Range has occurred in geologically recent times. Opinions of the duration of uplift vary between 0.5 million and 2 million years; consequently, estimates of rates of uplift vary considerably. In geological terms uplift of the southern Ruahine Range has been rapid and in response streams draining the Range have cut into the bedrock to produce steep sided valleys. Continual but episodic uplift of the Range has led to the youthful appearance of deeply-dissected, steep-sided valleys being perpetuated through Quaternary time. Presumably stream downcutting and slope instability were greatest during periods of rapid uplift.

There is no doubt that tectonic uplift and subsequent stream downcutting in the past have produced valley slopes that are at present susceptible to failure, irrespective of the factor that finally results in slope movement. Whether present day tectonic uplift rates may be considered to be average or "accelerated", is debatable. It is here considered very doubtful that the amount of uplift of the Range within historic times has contributed significantly to the observed increase in slope instability in this area.

10.4.3 FAULT RUPTURE AND SEISMIC SHAKING

As the study area lies within the main seismic region of New Zealand (Eiby, 1971), active fault traces are not uncommon (see Chapter 6) and earthquake activity is relatively frequent.

Rupture of the ground surface along a fault trace is associated with fault movement that may be instantaneous or creeping, and which may or may not be accompanied by seismic shaking.

A. Fault Rupture and Slope Stability

Within New Zealand mass movement resulting directly from fault rupture along active faults is well documented (Fyfe, 1929; Henderson, 1933; Ongley, 1943). In 1931 a destructive earthquake (M = 7.9) rocked Napier (approximately 100 km to the northeast of the study area) during which there was considerable ground rupture and mass movement (Henderson, 1933). In 1942 an earthquake (M = 7.0) in the Wairarapa district (70 km to the southeast of the study area) was also accompanied by ground rupture and mass movement (Ongley, 1943a).

Within the study area there are faults of known pre-Quaternary age along which rejuvenated movements in Early and Late Quaternary times resulted in dextral transcurrent and vertical displacements to uplift much of the Torlesse bedrock and Plio-Pleistocene marine deposits. There is no written evidence of fault rupture having occurred in historic times in the study area. Thus, evidence suggestive of fault rupture being accompanied by instances of slope stability during this time is non-existent. However, many deep-seated slope movements including ridge-top features and slumps in this area are aligned along faults, especially the Ruahine and Wellington Faults (Maps 5 and 6). Such alignments are suggestive of slope movements having resulted from fault rupture in prehistoric times. Nearly all of the documented active faults in this area have moved more than once and in many cases repeated movements have taken place along the same trace. Consequently, there may be several generations of fault related slope movements present. It is equally likely that some of these deep-seated slope movements post-date fault rupture and are the result of seismic shaking or other non-seismic influences within early historic times (see Chapter 12). Repeated movements of faults along the same Fault Zone have resulted in zones of extensive fault brecciation. As previously mentioned (Section 10.1.1), the majority of large-scale rock slumps occur where large tracts of brecciated bedrock are mapped along major fault traces. The presence of these breccia zones is here considered a major causal factor in the development of large-scale rock slumps in the study area.

Fault rupture of earth materials, particularly along contact faults separating Torlesse bedrock and Plio-Pleistocene marine deposits, has produced areas of physiographically oversteepened terrain where earth slump activity is common.

Fault rupture of colluvial and alluvial materials has produced fault scarps of variable height (see Chapter 6). At some localities subsequent collapse of these scarps is here considered to be the result of rejuvenated fault activity. However, the majority of fault-scarp collapses are triggered by non-seismic processes. Fault scarp collapses include slumps and debris slides of small extent.

Slope instability along lines of fault rupture can be clearly demonstrated within the southern Ruahine Range. It is, however, difficult in the majority of cases to verify whether episodes of instability coincided with fault rupture or occurred at a later date due to seismic shaking or non-tectonic events such as rainstorms. Nonetheless, fault rupture is clearly an important influence upon slope stability as: (1) a triggering mechanism; and (2) a means of producing both physiographically oversteepened slopes and wide zones of brecciated material conducive to slope instability.

B. Seismic Shaking and Slope Stability

Deep earthquake shocks occur as commonly as shallow shocks, particularly in the North Island, but there has rarely been more than minor damage associated with them. Consequently much of the following discussion centres on the effect of shallow shocks on slope stability.

Earthquakes have not been witnessed as having triggered large-scale slope movements in the study area. However, it is important to note that elsewhere within the North Island and particularly in the South Island in areas of similar physiographic terrain, very large-scale slope movements have been triggered by earthquakes within historic times. Many such movements have been known to block rivers and form 'earthquake dammed lakes'. Adams (1981a) has recorded the existence of 15 historic landslide-dammed lakes that have resulted directly from earthquake shaking, eleven of which were formed in an area of more than 5000 km^2 during the Buller earthquake ($M = 7.6$) in 1929. At least nine other New Zealand lakes dammed by landslides were probably formed by prehistoric earthquakes. Two large landslide-dammed lakes to the north of the study area - Waikaremoana and Colenso - were formed in prehistoric times but in both cases it is uncertain that damming resulted from an earthquake triggered landslide (Adams, 1981a). Thus, although a seismic origin for many of the prehistoric lakes seems probable it is possible that some lakes have formed by damming due to storm triggered landslides or other non-seismic events.

Only one type of landslide within the study area involves sufficient volumes of material capable of forming landslide dammed lakes. These are rock slumps (Maps 5 and 6) most of which occur in situations favourable for the blocking of stream channels. The steep sided, narrow river valleys are ideal for blockage of drainage leading to the formation of lakes but a dam will only result if slump movement is total. Partial slump movements into stream channels form debris flows that rapidly move downstream so damming does not eventuate. Small-sized lakes have most probably formed in the past, as a result of slump-damming, in the southern Ruahine Range. However, little evidence of their existence remains today because even very large slumps are unlikely to dam streams permanently, the lakes soon filling with water and the outflow rapidly incising through the dams.

To date, formation and failure of landslide-dammed lakes have caused little damage in New Zealand. However, they are realised as being a significant geologic hazard in mountainland within a seismically active zone like the southern Ruahine Range.

To the south of the study area, in the Rimutaka Range, Robins (1958) examined the direct effect of the 1855 earthquake (M = 8.1) on the forest vegetation and slopes of the Orongorongo Valley, Wellington, where numerous shallow landslides caused by the earthquake formed scars 400m in length and 200m in width. The debris slide and avalanche scars resulted from sudden slope movement as a result of seismic shaking rather than by slow erosional processes. Debris accumulated as extensive screes at the base of valley slopes. Many of the scars and screes, although long since revegetated, remained visible at the time of Robins' (1958) report. The susceptibility of these mountainland valley slopes to shallow slope movements during seismic shaking is due to the combined influences of a thin layer of loose, porous colluvium, overlying and in sharp contact with greywacke bedrock and steep valley slopes (see Section 10.2.6).

Seismic shaking of the soil and colluvium mantle is likely to cause debris sliding and debris avalanching and where the underlying bedrock is involved slumping may occur. Seismic shaking is also likely to trigger other forms of movement, including topples and falls, particularly where the shear strength of the material has already been severely reduced by discontinuities. The effects of seismic shaking of earth materials flanking the Range can vary, depending on the water content of the material, from relatively dry earth slides and slumps to wet earth flows. In poorly consolidated alluvium standing at the angle of repose seismic shaking will create open fissures and bring about collapse.

It cannot be demonstrated that any shocks have originated from active faults within the study area during historic times, the time period during which many of the shallow slope movements in the southern Ruahine Range were triggered. Neither shallow nor deep-seated slope movements suspected of having been triggered by seismic shaking occur on slopes facing a particular direction. Slope movements may have been triggered by earthquake shocks generated some distance from the Range, or alternatively they may have been triggered by movement at depth along faults within the Range but without obvious surface rupture. The Wellington and Ruahine Faults cut through bedrock of high shear strength which can store large amounts of elastic strain energy. Sudden release of this energy as a result of fault displacement would produce an earthquake with considerable seismic shaking potential. Many of the fault traces in the study area within alluvium and Plio-Pleistocene marine deposits are thought to be splinter faults originating from movements along major fault zones. Consequently considerable seismic shaking may be associated with fault movements not only along major fault zones but along all active faults in the study area.

Dr W D Smith of Geophysics Division, DSIR, considers that only earthquakes of magnitude 6 or greater in the epicentral area would generally trigger landslides (cited in Cunningham & Stribling, 1978). Since 1940 there have been six such earthquakes; two occurred in 1942 and the others in 1947, 1951 (Stephens, 1975; p 88), 1958 and 1975 (Mosley, 1977; p 31). There is no direct evidence to suggest that slope movement in the study area occurred during any of these earthquakes. Nonetheless, it has been postulated that an increase in seismic activity during historic times may have resulted in increased slope movement during the 1940-1970 period. Stephens (1975) suggested that the large number of historical earthquakes (M $\geqslant$5) centred within 100 km of Raparapawai catchment, had a profound influence on mass movement occurrence and erosion rates in this area. Records of shallow earthquakes (defined as those with focal depths of 40 km or less) known to have occurred in New Zealand between 1840 and 1975, with intensities greater than MM 6.5 (Modified Mercalli Scale of Eiby, 1966), indicate that seismicity has varied over this time period. However, there has been rather less seismic activity than on average since 1940 (Smith, 1978a).

Hatherton's (1970) charts of North Island earthquake epicentres show no particular concentration in the vicinity of the southern Ruahine Range and no other records could be found that attribute a particularly high frequency of earthquakes to the study area. No statistically significant increase in earthquake frequency between 1939 and 1975 within the study area has been recorded. It therefore seems unlikely that the incidence of increased slope instability in the study area during historic times can be attributed to seismic activity alone. However, it is here considered likely that seismic shaking in both prehistoric and historic times has been an important factor in triggering all forms of slope movement.

10.4.4 CONCLUSIONS

The absence of documented evidence of slope instability resulting directly from fault rupture or seismic shaking suggests that the increase in observed slope instability within historic times in the southern Ruahine Range has not been greatly influenced by active faulting. However, in view of the frequent occurrence of seismic events of damaging intensity within historic times, both throughout New Zealand and particularly in areas adjacent to the study area, the potential of seismicity in initiating slope instability within the southern Ruahine Range should not be underestimated. Slope instability is likely to be most severe should seismic activity coincide with saturated ground conditions.

10.5 HUMAN FACTORS INFLUENCING SLOPE STABILITY

10.5.1 INTRODUCTION

In studies dealing with problems of erosion and slope stability it is generally accepted that a dense forest cover in healthy condition substantially enhances the stability of slopes and that deterioration of the forest cover by natural or induced causes results in progressive instability and eventual slope movement. Such a premise can only be substantiated when and where the original condition of the vegetation cover can be accurately assessed and where observed changes in the vegetation cover are closely monitored through time. It cannot be substantiated that the vegetation cover of the southern Ruahine Range was ever in perfect condition but from observations dating from the late nineteenth century to the present day it is readily apparent that the vegetation has undergone major changes. This chapter outlines these changes to the forest composition and structure, many of which were in evidence prior to European settlement but have undergone an accelerated rate of change during historic times (*c* 1870 A.D.). Causal factors responsible for this change to forest structure and composition commonly referred to in the literature as "forest deterioration", are outlined. As forest deterioration has accelerated during historic times much of the blame for this deterioration has been attributed to causes of human origin. It is, however, impossible to divorce natural causal factors, operative during historic times, from a discussion on forest deterioration. Consequently natural causal factors currently cited as dominant causes of forest deterioration are included.

The effect of forest deterioration upon slope stability is examined and an opinion as to whether human-related factors or factors of natural origin have been dominant in promoting slope movement is expressed.

10.5.2 THE NATURE OF CHANGES TO THE FOREST VEGETATION

In an ideal protection forest the treetops are usually continuous enough to form a closed canopy; there is generally a lower tier of secondary trees and shrubs, a ground tier of ferns, mosses and seedlings, and a forest floor cover of dead or decaying plant material. Bare soil, stones and exposed roots are rare. Death or windthrow of trees will occasionally occur, and the consequent canopy gap will be replaced within a few decades. Landslides will also occur, and on the resulting exposed surfaces a succession of grasses and shrubs will eventually give way to forest, perhaps within a century (Cunningham & Stribling, 1978).

Little is known about the condition of the forest vegetation cover in the southern Ruahine Range prior to European settlement. An early report by Colenso (1884) contains vivid descriptions of severe forest damage and extensive slope movements that he observed during crossings of the Ruahine Range in 1845 and again in 1847. "Here and there an immense mass of earth had slipped quietly down the upright cliffs bringing the large trees with it. . .; in two or three spots during the day I noticed a double slip or subsidence of this nature . . ." (Colenso, 1884; p 10). Many of the observations were from the Waipawa and Makaroro river catchments just to the north of the mapped area. During the early period of European settlement, observations of the mapped area by local residents indicate that the forest undergrowth was extremely dense and that there was little or no sign of damage to the forest cover until late in the 1930s. "From the farming country practically <u>no</u> slips could be seen on the Ranges" (Hubbard, 1976). However, the latter statement is here considered to be misleading because it is equally applicable to many views of the Range, even today. This point is made in order to emphasise that most of the erosion, both in the past and at present, is not located along the Range front within view but occurs deep within the Range at sites largely hidden from view. It is therefore considered that the magnitude of erosion, within that part of the southern Ruahine Range comprising the study area, during the period of early European settlement has been grossly underestimated. Consequently, it would be incorrect to assume that the vegetation cover of the Range within this area was complete prior to 1930. Rather, it is likely that the vegetation cover was in good condition but did show signs of deterioration, particularly in areas where slope movement had already occurred.

During years of European settlement the forest vegetation underwent a rapid change, the result of which is very much in evidence today. "The canopy is open and in places has completely collapsed, exposing secondary trees and shrubs to wind and temperature extremes they are not able to withstand (*sic*). Replacement of the canopy has often been retarded by heavy browsing of seedlings and shrubs. Debris avalanches are abundant;

by comparison, signs of a return to forest are uncommon" (Cunningham & Stribling, 1978; p 18).

10.5.3 FACTORS RESPONSIBLE FOR CHANGES TO THE FOREST VEGETATION COVER

This section discusses some of the factors often quoted as reasons for changes to the forest cover in the southern Ruahine Range.

A. Fires

Evidence of pre-European fires in the southern Ruahine and northern Tararua Ranges has been found by Esler (1963). In addition, changes to the vegetation resulting from European fires during the 1880s and again in 1915 and 1946 have been recorded in Pohangina and Cattle Creek catchments (Elder, 1965). These fires are thought to be responsible for anomalous patches of scrub evident today in the Pohangina Saddle area. The area affected by these fires is relatively small. There is no other evidence of any extensive fires in the southern Ruahine Range, although it is probable that during clearance of the flanks of the Range fires may have become uncontrolled and spread into the lower forest margins and up some of the outer spurs (Elder, 1965). During the summer of 1978 fire broke out on the right bank of the western end of the Manawatu Gorge, above the roadway. The vegetation cover was completely stripped to expose the bedrock. Sheetwash and weathering processes have perpetuated the instability of this burnt-over area to the present day.

B. Man and Introduced Animals

There is no doubt that the forest composition has further changed since the time of European settlement. The forest was axed, sawn and burnt on the lowlands adjacent to the Range fronts and up to the base of the Range from as early as 1870. Selective milling of large trees from within the Range occurred well into the 1920s.

Some of the first introduced animals to be brought into the area included domestic cattle and sheep, many of which penetrated considerable distances into the Range, particularly along stream beds and ridge crests, to forage within the forest. At this time the forest boundary was unfenced and common farming practices entailed the overwintering of stock within the forest (Elder, 1965). Opossums were first liberated in the Pohangina Valley in 1893 (Pracy, 1966). This was followed by many other liberations on both sides of the Range during the 1920s and by 1953 opossum colonisation was considered complete throughout the forested area. Opossums caused severe damage to the mid-valley slopes of the southern Ruahine Range by the repeated defoliation of preferred species such as *Metrosideros robusta* and *Weinmannia racemosa*. By the late 1950s the mid-valley slopes contained an abundance of dead trees, which have since collapsed, interspersed with scattered *N. fusca, P. ferrugineus* and other less palatable trees. Red deer were released in the upper Pohangina (George, 1979) and at Delaware in 1922 (Elder, 1965; Logan & Harris, 1965) and by the mid-1940s had spread southward throughout the forested area. Deer population numbers peaked in the 1960s but significant effects upon plant viability due to browsing were evident by the 1950s (James & Beaumont, 1971). Deer and other ground browsing animals (goats, sheep, cattle) eat foliage up to 2m above the ground and can remove the current year's growth or eliminate certain palatable plants altogether. Intensive animal browsing can prevent tree seedling development that would otherwise replace openings in the forest canopy. This is most readily seen on grassed slip faces where even the development of shrub species may be inhibited (Cunningham & Stribling, 1978).

Goats were first seen in the Raparapawai catchment in 1919. They were later liberated in the Manga-a-tua catchment in the 1930s, and spread northwards to reach the Tamaki catchment by 1955 and the upper Pohangina River by 1959. The goat population is considered to have peaked around 1956 but now goats have almost disappeared from the forested areas. Pigs were abundant in the Oruakeretaki and Manga-a-tua catchments before 1920. In 1955 their presence in the Mangapukakakahu catchment was thought to be the major cause of erosion there. As a consequence of increased hunting pressures, pig numbers were reduced considerably in these catchments by 1958 (Elder, 1965).

Bark eating by goats and deer, tree rubbing by goats, deer, pigs, sheep and cattle and scratching by opossums damage the bark and promote tissue decay and serious injury to trees. Hares, rabbits, rodents and other introduced mammals are not present in high numbers and their influence upon the vegetation is probably slight by comparison with opossums, deer and domestic browsing animals. However, all are capable of damaging the vegetation.

As a result of the high density of browsing animals within the forest throughout the 1940s - 1960s it is considered likely that soil compaction by trampling could have been an insidious but important factor influencing plant health. Compaction by animals reduces soil aeration, impedes sub-surface drainage, and greatly alters the habitat of soil fauna which, in turn, affects plant growth and paves the way for disease (Cunningham & Stribling, 1978).

The overall impression from the literature is that the Ruahine forests, up until the time of European settlement, evolved in the absence of browsing animals and reached a delicate state of balance with the steep slopes and climatic conditions of the time. Thus the ensuing introduction of mammals had a profound influence on this ecosystem, altering the forest structure, exposing sub-canopy plants to storm damage, influencing the water cycle and encouraging the proliferation of fungal and insect pathogens. In the words of McKelvey (1960) "Introduced animals have weakened the precipitation-absorbing and soil-stabilizing indigenous protection forests, causing accelerated erosion and increased flooding".

C. Native Animals

Not all damage is done by introduced animals. Certain native birds such as wood pigeons and kakas can be responsible for tree defoliation or bark damage. Insects too may play an important role in defoliation (Meads, 1976). Insects become important as pathogens in vegetation initially damaged by browsing mammals (Batcheler, 1967). Logan (1971) and Pracy (1971) report that white scale insect has been responsible for defoliation of *Metrosideros robusta* and *Weinmannia racemosa*, and was first noticed in the West Tamaki catchment in 1955.

D. Climatic Factors

Change in the forest cover has also been attributed to climatic factors such as wind, precipitation and temperature. In the southern Ruahine Range windthrow appears to result in the toppling or breaking of emergent living trees rather than wholesale felling of large tracts of forest. Sites of windthrown trees were most commonly observed along ridges and steep sided valley slopes adjacent to stream channels. Windthrown trees in the latter locations are largely due to the funnelling effect of strong wind gusts up or down the valleys. Forest damage by very strong winds in historic times has been documented by Elder (1965), Esler (1969) and Logan (1971), most of which is thought to have occurred during a period of gale force winds in 1936.

Mountain vegetation may be severely damaged by sudden increases in precipitation. Cunningham (1966) describes extensive damage to forests dominated by *N. fusca* following the heavy snowfall of 1961. He also notes that damage from snow avalanches can occur in sub-alpine scrubland. Intense rainstorms of short duration result in slope instabilities of various types that often completely remove the vegetation cover. Elder in 1956 spoke in terms of "a retreat of forest due to climatic changes". In 1965 he was more cautious about attributing the deterioration of *Libocedrus* and *Dacrydium biforme*, and the retreat of the mountain beech timberline, to climatic change. It has been suggested that climatic changes could be the primary factor causing the death of old even-aged and exposed *Weinmannia racemosa* (Stephens, 1975). There is still little clear evidence of the effect of climatic changes on the forest vegetation in the southern Ruahine Range. For example, influence of the 1946-47 drought on the vegetation of the Range is little known. Elder (1958) concluded "It may be expected that a vegetation assumed to be adapted to high average humidities with a low exposure to sunlight would be vulnerable to any prolonged and sunny period". Drought may significantly influence vegetation vulnerability (Mosley, 1977).

There is no clear manifestation of the effects of temperature fluctuations on the Ruahine vegetation (Cunningham & Stribling, 1978).

E. Factors of Unknown Origin

The possibility of some unknown factor damaging the southern Ruahine forest has often been raised. Logan (1955) describing forest deterioration in the southern Ruahine Range, wrote "This upper level forest is not able to regenerate itself and has not been able to do so since before animals were ever introduced into the forest. What has caused the recession of the upper forest margin can only be guessed at, but it is not animal life". Pracy (1971) commented that "The profusion of dead trees suggests other factors have been involved in the decline of the vegetation". James (1973) considers that some forest mortality cannot be linked with mammals. "There appears to be a natural downslope movement of some forest communities; for example, the widespread mortality of cedar and pink pine at the upper timberline. Similar trends are shown by mountain totara (*Podocarpus hallii*) and red beech. The mortality is particularly noticeable at the upper limit of each species, and is often accompanied by a lack of effective regeneration" (James, 1973; p 97). According to Robins (1958) the general decline of the podocarp element in New Zealand mixed forests is merely a phase in the trend towards a replacement of podocarp forest by broadleaf forests that is evident in the New Zealand forest pattern.

F. Other Factors

Mosley (1977) points out that slope instability and mass movement, once initiated, may in itself inhibit redevelopment of a continuous forest cover (p 25). Fault rupture and seismic shaking may affect the vegetation either directly through loosening root systems or indirectly by promoting slope failure with consequent removal of the vegetation.

The unhealthy condition of some tree species, for example *Weinmannia racemosa*, may be explained if the trees' normal cycle includes widespread collapse of the old even-aged trees followed by a new even-aged canopy (Strand, 1981).

10.5.4 DISCUSSION

The vegetation of the southern Ruahine Range is complex, is undergoing change, and is in many areas in ill health. Changes to the forest composition and structure seem to be widespread and are, in the opinion of the author, the result of a combination of factors as obviously no single factor is responsible for this situation. The previously published contributory factors are listed in order of importance: (1) browsing animals, especially opossums, have certainly damaged selective canopy tree species and deer have undoubtedly inhibited regeneration in the understorey; (2) climatic factors have brought about changes in forest composition and resulted in localised damage to living canopy trees; (3) insect and pathogen attack may also be an important factor; (4) the effect of fires and milling on the overall pattern of forest deterioration has been localised and of minor importance; and (5) the effect of other factors is indeterminate. It is not possible to firmly state to what extent each of these factors has promoted forest deterioration.

Largely overlooked but briefly mentioned by Cunningham (1966) and Mosley (1977) is the possible damaging effect of continual slope movement upon forest composition and structure. Observations of many slope movements in this area indicate that their effect upon the vegetation is more widespread than is at first apparent. These effects are twofold: (1) movement of loose and angular scree downslope of an initial scar, buries or damages the understorey and inhibits regeneration for long periods but has little effect upon the canopy. This insidious creeping of scree may therefore account for the paucity of understorey vegetation downslope of some erosion scars; (2) Most slope movements result in the removal of all vegetation from the site of failure thereby creating gaps in the canopy. Opening of the canopy exposes the remaining tree species to windthrow and recolonisation of slope movement scars is slow. In general, a few species of grasses are present, but for the most part the plants are herbs, *e.g. Raoulia, Taraxacum, Gnaphalium, Acaena, Epilobium, Wahlenbergia, Euphrasia, Pratia,* the tall pampas grass *Arundo,* the ferns *Polystichum* and *Pteris,* together with small woody shrubs such as *Coprosma, Cassinia* and *Helichrysum* (Robins, 1958).

The paucity of revegetated scars with a forest tree cover is the result of numerous episodes of intermittent movement at the same site, the vegetation cover not having sufficient time to mature. It follows that slope movement and subsequent revegetation of slipped areas is bringing about a change in the composition and structure of the forest vegetation in this area and is therefore part of the process of forest deterioration largely unrelated to browsing animals.

The majority of slope movements fail during or following periods of heavy rainfall (see Section 10.3.2) as a consequence of soil and colluvium saturation. A good vegetation cover may reduce the frequency of saturation by: (1) intercepting moisture; and (2) reducing the amount of water that is retained as groundwater through increased rates of transpiration. On the other hand a depleted vegetation cover promotes saturation and eventual slope movement. As well, rainwash, surface runoff, local rilling and gullying are enhanced.

Periods of heavy rainfall accompany major storm events that are derived predominantly from the northwest. Hence soil and colluvium upon slopes with a northwesterly aspect are likely to become saturated more frequently than other slopes. A relationship between soil and colluvium saturation and the incidence of slope movement is indicated by the distribution of slope movement scars throughout the study area, the greatest number of erosion scars occurring on north- to west-facing slopes (see Section 10.2).

While it is accepted that depletion of the forest vegetation as a result of overbrowsing has increased the susceptibility of soil and colluvium to saturation which in turn has increased slope instability, it is not considered likely that slope movements are the direct result of forest deterioration by browsing animals because:

(1) Changes in both the structure and composition of the forest vegetation and extensive slope movements were in evidence prior to the introduction of browsing animals;

(2) Browsing animals cannot account for extensive slope areas devoid of canopy trees;

(3) Slopes upon which the understorey has been eaten out and upon which there are large numbers of dead and dying forest trees are not necessarily sites of slope movement;

(4) Browsing cannot account for slope movements where the forest is considered to be in good health;

(5) Browsing is unlikely to explain the northwesterly aspect of the worst eroded slopes.

In addition, Mosley (1977) noted that in areas of stream bank undercutting repeated failure of riparian slopes prevented the establishment of a forest cover whereas on adjacent slopes not affected by streambank undercutting there existed a good forest cover. He concluded that animal browsing could not have caused the changes in vegetation cover necessary to produce this pattern.

A generalised picture of the forest vegetation today consists of a dying and much thinned out secondary canopy beneath which a sparse understorey essentially comprises grasses and shrubby species. There are extensive areas where the canopy is absent and where grasses, ferns and herbaceous species form a dense ground cover. Areas of bare ground within the confines of the forest are largely the result of slope movements which are currently at varying stages of revegetation.

The capability of this vegetation cover to stabilise valley slopes is now discussed in relation to root depth. Forest trees are generally regarded as having better slope stabilising capabilities than shrubby species or grasses. Some of the erosion-inhibiting effects which forests have on soils are summarized in Selby (1976, p 54). In particular, the reinforcing effect of a strong root mat in many instances is thought to be the difference between stability and failure with deep root systems giving coherence and strength to the soil to resist downslope movement. Root systems of even the largest trees within the study area rarely exceed depths of more than one metre. Numerous mature podocarps were observed to have maximum root depths of only 0.5m by Mosley (1977). He also indicates that one of the most common tree species, *Weinmannia racemosa,* is regarded as a shallow-rooting tree and that there is little evidence that the forest in any part of the study area has ever had rooting depths in excess of one metre (p 32). Root systems, in all but one of the soils mapped in the West Tamaki catchment, penetrated to the base of the soils to a maximum thickness of 0.7m (Hubbard, 1978; Table 15, p 171). In the remaining soil a waterlogged horizon at 0.5m depth provided adverse conditions for root growth; consequently there was a marked decrease in root abundance and a decrease in shear strength in this horizon.

Comparison of rooting depths with depth of slope movements suggests that many slope movements, in particular debris slides, result from failure within the root-zone at depths of less than one metre (Table 10.1). This concurs with observations by James (1963), Stephens (1975), Mosley (1977) and Hubbard (1978) which suggests that the forest vegetation on steep valley slopes within the Range is unable to prevent shallow movements resulting from failure within colluvial materials.

However, the majority of slope movements are deeper seated and result from failure below the level of most tree root systems. These movements may result from failure within colluvium, at the colluvium-bedrock contact or within bedrock (see Section 10.1.1) and fail irrespective of the type or condition of the vegetation. It is therefore concluded that in general the instability of steep valley slopes, particularly where slope movement has occurred, may be more closely related to the physical characteristics of the colluvium, the presence of a sharp colluvium-bedrock contact and to a lesser extent the nature of the bedrock, than to the composition and structure of the forest vegetation cover. There is, however, evidence that the stability of some valley slopes has declined as a consequence of the removal of the vegetation cover by slope movement particularly within historic times. For example, many of the erosion scars visible in the 1940s appear to have become partially revegetated by the 1960s but have again become sites of slope movement in the 1970s (see Chapter 12).

The susceptibility of these sites to further failure is likely to reflect the combined influences of unstable colluvium, a slow rate of vegetative recolonisation of eroded areas and susceptibility of these areas to surface runoff processes. Also to be considered is the probability that grasses, ferns and shrubby species are less likely to achieve the same degree of stability as that of the original forest cover. The progressive enlargement of these sites during successive episodes of failure indicates that the surrounding forest vegetation may have been weakened as often it is unable to prevent upslope retrogression of the slide.

Enlargement of old erosion scars and increased incidence of new scars during historic times (see Chapter 12) is bringing about a gradual but progressive decline in slope stability. The threshold of slope stability may therefore be declining also to the extent that a storm with a lesser return period will trigger slope movements whereas previously a storm of greater return period was necessary to trigger similar slope movement types. In addition, the trend towards increased numbers of erosion scars of larger size during historic times may be a result of the increased frequency of storm events for which supportive evidence is considerable (see Section 10.3.2).

10.5.5 CONCLUSIONS

Changes in the stucture and composition of the forest vegetation were in evidence prior to European settlement. These changes were largely the combined result of natural forest evolution and slope movements under specific climatic, geomorphic and tectonic conditions in existence at that time. The introduction of browsing animals early in the twentieth century greatly accelerated the rate of change to the vegetation cover largely through the influence of additional pressures which included depredation of the understorey and selected depletion of tree species by preventing their regeneration.

During this period of forest deterioration there occurred an increase in slope movement activity which further opened the forest canopy and destroyed large tracts of forest downslope of initial failure sites. Recolonisation of these sites by grasses and shrubby species has been incomplete as further failures have occurred in more recent times to lay bare substantial areas of valley slope. Continual slope movement activity has brought about a rapid change to the composition and structure of the forest vegetation and is therefore considered to be part of the process of forest deterioration. It is here considered that forest deterioration in this area has primarily been the result of progressive vegetation cover removal by continual slope movement processes and to a lesser extent due to depredation of the understorey and selected depletion of tree species by browsing animals. The combined processes of overbrowsing and increased slope movement activity have effectively reduced the stability of valley slopes by increasing the susceptibility of soil and colluvium to failure. There is considerable evidence to suggest that failure of these materials is brought about as a result of saturation both within the root zone and at deeper levels below most tree root systems. The distribution of erosion scars throughout the southern Ruahine Range shows that the majority occur on slopes with a northwesterly aspect, the direction from which major storm events approach this area. Hence soil and colluvium are likely to become saturated and fail more frequently upon these slopes than upon slopes facing in other directions. Successive episodes of slope movement severely reduce the stability of these slopes. As the majority of slope movements occur during or shortly after periods of heavy rainfall it is considered probable that slope stability in this area is in general primarily governed by climatic factors and to a lesser extent by the condition of the forest vegetation. Indeed, similar slope movement types have been triggered by storm events in neighbouring mountainland areas where forest and soil conditions are at present regarded to be in better condition than those in the southern Ruahine Range. It follows that the observed increase in the number of new erosion scars and enlargement of older scars, within historic times, is predominantly the result of increased frequency of storms as proposed by Grant (1965, 1966) combined with and exacerbated by the progressive deterioration of the forest vegetation.

Other factors contributing to slope instability include the nature and thickness of slope materials (see Section 10.1.1), slope steepness, length, aspect and degree of dissection (see Section 10.2) and seismicity (see Section 10.4).

The sum total of these factors has increased the probability that an event of given magnitude will now trigger slope movement in this area when previously it would not.

CHAPTER 11

LATE QUATERNARY EROSION EVENTS

CONTENTS

CHAPTER 11

LATE QUATERNARY EROSION EVENTS

11.0 INTRODUCTION

Within upper catchment reaches of the southern Ruahine Range there are a number of discontinuous depositional surfaces of alluvial origin, each recording at least one episode of erosion. Chronological accounts of past erosional events in the Ruahine Range (Grant, 1978) and, more specifically, in the West Tamaki River catchment (Hubbard, 1978; Grant, 1981a) have been largely based upon soil profile development and vegetation maturity on alluvial terrace and fan surfaces. This chronology is aided by ages determined by: (1) a tephra, the Aokautere Ash; and (2) radiocarbon dating of wood. There is evidence for two early erosional events dated at about 20 000 years B.P. and 12 000 years B.P., respectively (Hubbard & Neall, 1980), followed by a succession of erosional events (Table 11.1) dating from the 13th century to present day (Grant, 1978; 1981a).

TABLE 11.1: Periods of erosion in the southern Ruahine Range and West Tamaki Basin.

Erosion periods in the Ruahine Range (after Grant, 1978)		Erosion periods in the West Tamaki Basin (after Grant, 1981a)	
Erosion period	Tentative date A.D.	Erosion period	Tentative date A.D.
Waihirere	closed *c.* 1400-1450	Waihirere	*c.* A.D. 1270 - *c.* A.D. 1370
Matawhero	closed *c.* 1600	Matawhero	*c.* A.D. 1530 - *c.* A.D. 1620
Wakarara	*c.* 1780 - 1830s	Wakarara	*c.* A.D. 1780 - *c.* A.D. 1830?
Early Modern	1880s - 1890s	Tamaki	*c.* A.D. 1870 - *c.* A.D. 1900
Modern	mid 1930s to present	Waipawa	late 1940s to present

The following account combines information from these sources and provides additional evidence to support the recognition of these erosion events in other catchments elsewhere throughout the study area. Also, evidence is presented to suggest that erosional events within upper catchment reaches were coeval with episodes of extensive aggradation in lowland areas adjacent to the Range front.

11.1 EROSIONAL HISTORY

11.1.1 DEPOSITS OF THE OTIRAN STAGE (see Table 6.1)

The oldest dated deposit found in the Range consists of colluvial gravels within West Tamaki River catchment at locality T23/681162. Here, a 0.25-0.4m thick tephra bed, identified as Aokautere Ash (Hubbard & Neall, 1980), occurs within and is overlain by at least 10m of colluvial gravel. The age of the ash approximates 20 000 years B.P. and its presence within these gravels suggests that: (1) the Aokautere Ash was deposited in an environment of active erosion and deposition; and (2) gravels were no higher than this point (10.1m above the level of the present river channel) when Aokautere Ash was deposited (Hubbard & Neall, 1980). The colluvium is overlain by 1.9m of loess that presumably accumulated towards the end of the Ohakean Substage, and during Aranuian time (Table 6.1).

11.1.2 WHITEYWOOD CREEK FAN

This comprises an areally extensive fan-shaped deposit of alluvial material up to 15m thick within which there is evidence of at least three levels of aggradation. The apex of the fan can be traced into the lower reaches of Whiteywood Creek (T23/698192), a tributary of the West Tamaki River, thereby suggesting that much of the material comprising this fan was derived from this tributary. The absence of lateral terrace remnants within the middle and upper reaches of Whiteywood Creek make it difficult to define the exact source area of the large volume of material comprising the fan. Similarly, the absence of correlative terrace remnants in the principal channel of the West Tamaki River, upstream of Whiteywood Creek junction, indicates the localised nature of this aggradational deposit.

At the base of this alluvial fan at locality T23/698191, on the left bank of West Tamaki River, is a buried soil which overlies an undated gravel deposit which in turn overlies greywacke bedrock (Fig. 11.1). Above the buried soil is a second gravel horizon from which wood identified as *Griselinia littoralis* (broadleaf) has been radiocarbon dated (NZ 4314B) at 12 150 ± 150 years B.P. (Appendix Vb), suggesting that active deposition was occurring on the fan about 12 000 years ago. A 2m thick layer of fine gravel and silt mark the end of this second period of gravel aggradation. A seepage zone, highlighted by the line of *Cortaderia fulvida* vegetation (Fig. 11.1), indicates the position of this layer of fine gravel. The third period of aggradation is represented by the upper gravel unit upon which a mature stand of rimu has grown (Fig. 11.1). Tree ring dating of

FIGURE 11.1: Whiteywood Creek fan deposit in which at least three levels of gravel aggradation are recorded. The basal gravel (blue-grey colour) and uppermost gravel deposit are undated. The middle level has been dated at 12 150 ± 150 years B.P. Each level of gravel aggradation is separated by a thin accumulation of fine gravel and silt. Here, emergent water seeps and hydrophilic vegetation highlights the boundaries between successive gravel deposits. The terrace surface supports a thick stand of *Dacrydium cupressinum* dated by tree ring dating as being approximately 450 years old. Figure is 2m high. Locality T23/698191 on the left bank of West Tamaki River.

seven old rimu indicates an age of about 450 years (Grant, cited in Stephens, 1977). This gives a minimum age for the last period of aggradation which is thought to have been responsible for blocking the entrance of a small tributary on the left bank of the West Tamaki River at T23/690190 within which carbonaceous silts accumulated. A sample of wood found at the base of the 3m thick carbonaceous silts has been radiocarbon dated (NZ 4547c) at 770 ± 60 years B.P. (Appendix Vb). This date is thus considered to represent the commencement of the last period of fan aggradation at this locality, approximating to early Waihirere time.

On the basis of soil profile development, Hubbard (1978) has identified another aggradational fan deposit at T23/690180 and a buried soil at T23/684170 of comparable age to that developed on the surface of Whiteywood Creek fan. She suggests that the surfaces upon which these soils have developed may represent lateral down valley continuations of a single depositional episode.

11.1.3 WEST TAMAKI TERRACES

On the left bank of the West Tamaki River at localities T23/688175 and T23/686172 there are remnant alluvial terrace surfaces at about 2.5m above present river level. The surficial soil suggests that these terrace surfaces predate the upper surface of the Whiteywood fan (Hubbard, 1978). The terraces support a mature podocarp-hardwood forest. The bases of some of these trees have been buried by later alluvium of between 0.3-1m thickness, which supports a less mature vegetation dominated by pepperwood and mahoe. Tree ring dating of several large mahoe trees indicates that the uppermost alluvium was deposited during the 1880s, that is, during the Tamaki erosion period (Grant, 1981a). Tree ring dating of large podocarps were inconclusive and Grant (1981a) was unable to determine with certainty in which erosion period the lower portions of these terraces were deposited. He concludes that deposition of these terrace deposits occurred either during the Matawhero or the Waihirere periods.

11.1.4 HUT CREEK FAN

At the junction of Hut Creek and the main channel of West Tamaki River (T23/699197) there is a small fan deposit with a weakly developed soil that supports a stand of red beech (*Nothofagus fusca*). Tree ring dating indicates that trees first became established on this fan during the early 1880s. This gives a minimum age for the uppermost 2m thick veneer of alluvial gravel at this site of deposition in the Tamaki erosion period (Grant, 1981a). Beneath this veneer of alluvial gravel stumps of *Nothofagus fusca*, with approximately the same diameter as the largest now growing on the surface of the fan, are buried. The size of the stumps indicate that the dead trees were 90-100 years old at the time of the Tamaki deposition, thus an earlier depositional period preceded the Tamaki erosion period of 90-100 years, which approximates the dating for the Wakarara erosion period of Grant (1981a). A soil within surface gravels of a fan deposit at T23/684170 is at a similar stage of development to that of the uppermost veneer of gravels on Hut Creek fan and is considered to be of similar age (Hubbard, 1978).

11.1.5 DRY CREEK FAN

An extensive terrace surface is preserved in the lower reaches of West Tamaki River catchment. Gravels comprising this terrace have buried a well developed soil to depths varying between 3m at T23/682166 to 2m at T23/682162. The gravels are angular and very fresh-looking, and there has been little time for soil development (Hubbard, 1978). Tree ring dating suggests that the vegetation became established around A.D. 1900 indicating that depositional activity probably closed on the fan surface during the 1890s which was a very active phase of the Tamaki erosion period elsewhere (Grant, 1981a). Since this time further deposition of alluvium on the terrace surfaces during A.D. 1915-1920 covered the surface formed in the 1880s and buried the bases of numerous trees growing on the 1880 surface (Grant, 1981a).

11.1.6 RAPARAPAWAI TERRACE

On the left bank of Raparapawai Stream at T23/580056 Stephens (1975) located a 3.5m high aggradational terrace, much of which has since been destroyed. Wood identified as *Podocarpus spicatus* extracted from a 0.2m thick paleosol at 1.8m depth has been radiocarbon dated (NZ 3879c) at 680 ± 40 years B.P. (Appendix Vb). The period of aggradation represented by the gravel horizon beneath the paleosol is undated. The gravel horizon overlying the paleosol may be interpreted as being the product of a period of erosion that corresponds in time with the Waihirere erosion period of Grant (1981a).

11.1.7 MISCELLANEOUS TERRACE SURFACES

Throughout the study area there are a number of localities at which terrace surfaces are preserved. These surfaces were observed to fit into two general groups: (A) high-level terraces with tree-sized vegetation; and (B) low-level terraces of varying heights with a range of vegetation types including trees, shrubs and grasses to those which are non-vegetated.

A. High-Level Terraces

High-level terrace remnants occur in Andersons Stream (T23/605178), in Cattle Creek (U23/713221) and in Coppermine Stream (T23/554034). Each surface is approximately 10m above present day stream level and is mantled by a weakly developed soil of similar thickness and appearance to that on Hut Creek fan. Remnant tree stumps indicate that trees of approximately 100 years age grew on these surfaces. These terraces are considered to be of similar age and were probably deposited during the Tamaki erosion period of Grant (1981a). Each terrace deposit represents deposition, of a considerable volume of unweathered alluvial gravel, during a single erosional event. In each case the terrace has formed at a stream junction where the tributary stream that provided the gravel joins the main stream channel. Downstream extension of these high-level terraces is either very short or non-existent so that correlation relative to other surfaces is difficult.

B. Low-Level Terraces

Low-level terraces occur in most catchments; the better examples are presented here. The height of these surfaces is variable and there is no consistent relationship between terrace height and maturity of vegetation. These terrace surfaces are underlain by a very weakly structured soil that overlies unweathered gravels. In the West Tamaki catchment, low level terraces covered with *Melicytus ramiflorus* occur at T23/697195 and T23/684169. They are in general no more than 0.5m above stream level and are thought to be less than 98 years old (Hubbard, 1978). Other vegetated surfaces of probable equivalent age are 0.5-2m in height and occur in Mangapuaka Stream (T23/627111), in Pohangina River (T23/690234) and in Makawakawa Stream (T23/620186). At locality T23/584052 in the Raparapawai Stream catchment, Stephens (1975) dated a 2m high terrace surface on the basis of age of its vegetation cover. Using an increment borer he determined the minimum age of rimu trees to be approximately 200 years old. In allowing for a reasonable time for the vegetation to establish on a freshly deposited aggradational gravel surface, Stephens estimated the age of this surface to be at least 250-300 years old (p 63). Insufficient evidence precludes the assigning of this period of terrace formation to a specific erosion period. Most low level terraces are subject to streambank overflow and indeed are either veneered with recent alluvium or have been dissected by channel scour. It is unclear to which erosion period the formation of many can be attributed but most of them were probably deposited either during the Tamaki or Waipawa erosion periods of Grant (1981a).

In historic times the formation of alluvial terraces has coincided with periods of slope instability during storm events. Severe slope instability, sediment transport and terrace formation occurred in 1975 during the passage of Cyclone Alison across the area. As a result of this storm many catchments contain terrace surfaces of bare gravel that are still in evidence today, for example, in the West Tamaki River catchment at T23/699199. In 1978 a localised storm caused severe damage to valley slopes in the North Oruakeretaki Stream catchment, as a result of which the main stream channel became choked with alluvial gravel. The formerly boulder-armoured stream bed was transformed overnight into an even-surfaced bed comprising fine alluvial gravels up to 1m thick. Since this storm event the stream has incised into the fine gravel to form 1m high lateral terrace deposits along its banks and the former boulder-armoured streambed is once again exposed. Many such terrace systems are to be found in upper catchment reaches where they are formed by a build-up of gravel behind log dams. When the dam breaks the stream rapidly incises into the gravel as a result of which the former streambed is abandoned and is preserved in part as lateral terrace remnants. Terraces formed in such a way are generally short-lived as subsequent storm events resulting in high flood flows generally destroy these unconsolidated deposits. Still other surfaces are preserved for sufficient periods of time for the establishment of a meagre vegetative cover. The majority of depositional terrace surfaces of this nature have been formed during the Waipawa erosion period of Grant (1981a).

11.2 DISCUSSION

Evidence of major erosional periods dating back to Late Quaternary times is preserved within upper catchment reaches of the southern Ruahine Range. The gravel deposits in which Aokautere Ash is buried preserve evidence of a major period of erosion that occurred during the Ohakean Substage of the Otiran (last glacial) Stage. At this time the Ruahine Range was subjected to a sequence of alternating climatic fluctuations. During the cold glacial periods solifluction is known to have been widespread in the Wellington district (200 km to the south of the study area) (Cotton & Te Punga, 1955) and similar periglacial conditions probably existed in the

Ruahine Range (Hubbard & Neall, 1980). Resultant increased erosion rates in mountainland areas produced gravels that now form extensive terraces in lowland areas adjacent to the Range. The presence of Aokautere Ash within loess upon aggradational terrace surfaces of Ratan age in the south of the study area (see Section 6.6.1C) suggests that here loess accumulation was approximately coeval with a major erosion period in the Ranges during which this ash was preserved within aggradational gravels. This erosion period is thought to have begun at about 20 000 years ago and based upon a radiocarbon date from the West Tamaki catchment (NZ 4314B, Appendix Vb), is thought to have ceased about 12 000 years ago. These dates suggest that a period of mountainland erosion in the southern Ruahine Range spanned most of Ohakean time.

At least five erosional periods within Holocene times have been recorded in upper catchment reaches of the southern Ruahine Range. Grant (1965, 1966, 1978, 1981a, 1981b, 1983) attributes each episode of erosion and sedimentation to periods of sustained increase in the frequency of major rainstorms (see Section 10.3). Some erosion is the result of a single storm such as that which occurred during Cyclone Alison in March 1975.

As previously outlined (see Chapter 10) climate is regarded as being the most important causal factor of erosion in the southern Ruahine Range during historical times. Presumably this relationship has existed throughout the Holocene as one of the most important factors responsible for triggering erosion periods. Older erosion periods in Ohakean times coincided with a known period of climatically induced increased erosion rates. Thus it is likely that the majority of erosion periods recorded in this area have been triggered by long term climatic oscillations and/or short-term heavy rainstorms. Evidence from the southern Ruahine Range suggests that erosional periods in Ohakean (25 000-12 000 years B.P.) times were of considerably greater magnitude than in Holocene times.

Since the 13th century each successive erosion period, apart from the present day Waipawa period, has been of lesser magnitude than previous periods and all but the Tamaki period exceed in magnitude that of the present Waipawa period (Grant, 1983, fig. 24). The Waipawa erosion period has been characterised by increased erosion rates in headwater valley slope areas, increased supply of coarse sediment to stream channels and increased rates of bedload transportation from upper catchment areas onto adjacent lowland areas. A quantitative measure of erosion rates in all upper catchment reaches within the study area during the present Waipawa erosion period has been attempted; the results and consequences of which are outlined in Chapter 12.

CHAPTER 12

PAST, PRESENT AND POTENTIAL PATTERNS OF SLOPE STABILITY

CONTENTS

CHAPTER 12

PAST, PRESENT AND POTENTIAL PATTERNS OF SLOPE FAILURE

12.0 INTRODUCTION

The few historical accounts which refer specifically to erosion in the southern Ruahine Range record the presence of slope movement during the early twentieth century but unfortunately no estimate of the magnitude of erosion can be determined from these accounts. An attempt to quantify the magnitude of erosion in the study area and to establish a trend in the erosion rate, within historical times, is outlined in this chapter. The scope of this study was, however, largely governed by the availability of aerial photographic coverage. In an area the size of the southern Ruahine Range there are problems in obtaining sequential photographic coverage that: (1) spans the entire area; (2) is devoid of cloud cover; and (3) is of an appropriate scale for use in such a study. These constraints limited this study to two periods of time. The earliest available coverage is provided by two photographic series (SN 230 and SN 181) flown between 1946 and 1949. The second period for which entire aerial photographic coverage is available (SN 3721 and SN 5163) spans the years 1974 to 1978. Thus from these two time periods measurements of areas of actively eroding slips about 28 years apart is possible. From differences or similarities in the distribution and amount of eroded areas, current trends in the rate of erosion can be ascertained. Reference to additional but incomplete aerial photographic coverage during the intermediary years provides valuable clues to the timing of an accelerated erosion period that occurred within this time interval.

In this chapter large-scale, deep-seated mass movements are discussed separately from areas of actively eroding slips, the latter being largely of shallow translational origin (see Section 10.1.1). Falls and topples are few in number and too small to measure but where good examples have been located, they are plotted on Maps 5 and 6.

12.2 METHODS

Measurement of upper catchment areas and of large-scale, deep-seated slumps with an obvious ovate outline was achieved using a Hewlett Packard digitiser. Areal measurement of ridge-top features of indeterminable shape was not attempted. Areal measurements, in hectares, were repeated until figures accurate to two decimal places were obtained.

Areas of actively eroding slips (excluding old erosion scars that have since revegetated) that were present during the period 1946-49 were traced off aerial photographs at a scale of 1 : 13 600 (series SN 230 and SN 181). With the aid of an epidiascope, these images were transferred onto a base map drawn at 1 : 25 000 scale (Map 5). Areas of actively eroding slips present in the period 1974-78 were traced off aerial photographs at 1 : 24 000 scale and 1 : 10 000 scale (series SN 3721 and SN 5163, respectively) and projected to 1 : 25 000 scale in the same manner (Map 6). Photographic distortion was kept to a minimum by tracing only those areas of actively eroding slips within the middle portion of each contact print. Measurement of the areas of actively eroding slips was carried out in upper catchment areas only using a dot grid of 100 dots/cm^2. These results are presented in Tables 12.1 and 12.2). These areal measurements were not adjusted for the slope angle of the ground.

12.2 SHALLOW TRANSLATIONAL ACTIVELY ERODING SLIPS

12.2.1 DISTRIBUTION

During the 1940s the total area of actively eroding slips occurring on both the western (180 ha) and eastern (160 ha) flanks of the Range was essentially the same (Tables 12.1 and 12.2). However, the distribution of these areas of active erosion along each flank is uneven. On the western flank 61% of the total eroded area was located within three deeply dissected catchments - Piripiri, Pohangina and Makawakawa catchments (Table 12,1, column 6) at the northern end of the Range. In addition, large areas of actively eroding slips occurred within three moderately dissected catchments further to the south, *e.g.* Mangatuatou, Opawe and No. 1 Line catchments. Only small areas of actively eroding slips occurred within the least dissected catchments both in the north (Konewa and Te Ekaou) and in the south (Te Awaoteatua and Maungatukurangi) on the western flank of the Range.

In contrast, areas of actively eroding slips on the eastern flank of the Range during the 1940s were more evenly distributed. Nonetheless, more than half (55.7%) of the total eroded area on this flank was located within catchments northward of Mangapuaka catchment (Table 12.2, column 6). This included a substantial area of actively eroding slips within the Mangatera foothills (to the northeast of the Range) at this time (Table 12.2, column 6).

During the 1970s the location of actively eroding slips remained unchanged despite an average increase of 91% in eroded area since 1946

TABLE 12.1: Calculation and comparison of the area in actively eroded slips within upper catchments along the western flank of the southern Ruahine Range for the years 1946-49 and 1974-78.

Column number	1	2	3	4	5	6	7	8	9	10
Catchment	Upper catchment area (ha)	Area in actively eroding slips				% of total eroded area within each catchment		% of total eroded area in the southern Ruahine Range		% increase/ decrease in eroded area between 1946-74
		1946-49		1974-78						
		ha	%	ha	%	1946-49	1974-78	1946-49	1974-78	
Maungatukurangi	445	0	0	2	0.4	0	0.5			+200
Whareroa	723	2	0.3	5	0.7	1.1	1.4			+150
Makohine	484	2	0.5	3	0.6	1.1	0.8			+ 50
Tokeawa	519	5	1.0	10	1.9	2.8	2.7			+100
Te Awaoteatua	493	1	0.2	0	0	0.6	0	not determined		-100
No. 2 Line	292	5	1.7	8	2.7	2.8	2.2			+ 60
No. 1 Line	477	14	2.9	19	4.0	7.8	5.2			+ 36
Dundas Creek	291	2	0.7	3	1.0	1.1	0.8			+ 50
Ohinetapu	155	3	1.9	6	3.9	1.7	1.6			+100
Mangatuatou	368	12	3.3	13	3.5	6.7	3.6			+ 8
Opawe	798	19	2.4	20	2.4	10.6	5.5			+ 5
Porewa	498	3	0.6	13	2.6	1.7	3.6			+333
Te Ekaou	661	0	0	3	0.5	0	0.8			+300
Makawakawa	3 982	34	0.9	105	2.6	18.9	28.7			+209
Konewa	1 334	2	0.1	2	0.1	1.1	0.5			0
Pohangina	3 320	58	1.7	120	3.6	32.2	32.8			+109
Piripiri	639	18	2.8	34	5.3	10.0	9.3			+ 89
Western flank	15 479	180	1.2	366	2.4	100%	100%	53	56	+103
Western and eastern flanks	27 132	340	1.3	648	2.4	100	100	100	100	+ 91

TABLE 12.2: Calculation and comparison of the area in actively eroding slips within upper catchments along the eastern flank of the southern Ruahine Range for the years 1946-49 and 1974-78.

Column number	1	2	3	4	5	6	7	8	9	10
Catchment	Upper catchment area (ha)	Area in actively eroding slips				% of total eroded area within each catchment		% of total eroded area in the southern Ruahine Range		% increase/ decrease in eroded area between 1946-74
		1946-49		1974-78						
		ha	%	ha	%	1946-49	1974-79	1946-49	1974-78	
West Tamaki	1 250	20	1.6	48	3.8	12.5	17.0	not determined		+140
East Tamaki foothills	1 445	11	0.8	7	0.5	6.9	2.5			- 36
Mangatera foothills	1 061	24	2.3	5	0.5	15.0	1.8			- 79
Rokaiwhana	1 042	10	1.0	34	3.3	6.3	12.0			+240
Rokaiwhana foothills	614	1	0.2	2	0.3	0.6	0.7			+100
Otamaraho	172	2	1.2	10	5.8	1.3	3.5			+400
Mangapuaka	777	21	2.7	40	5.1	13.1	14.2			+ 95
Otamarahu	346	3	0.9	10	2.9	1.9	3.5			+233
Mangapukakakahu	438	8	1.8	13	3.0	5.0	4.6			+ 63
North Oruakeretaki	855	19	2.2	21	2.5	11.9	7.4			+ 11
South Oruakeretaki	275	4	1.5	7	2.5	2.5	2.5			+ 75
Raparapawai	1 045	16	1.5	35	3.3	10.0	12.4			+119
Manga-a-tua	895	14	1.6	27	3.0	8.8	9.6			+ 93
Coppermine	630	5	0.8	16	2.5	3.1	5.7			+220
Mangapapa	808	2	0.2	8	1.0	1.3	2.8			+300
Eastern flank	11 653	160	1.4	282	2.4	100	100	47	44	+76
Western and eastern flank	27 132	340	1.3	648	2.4	100	100	100	100	+ 91

throughout the Range (Tables 12.1 and 12.2). Much of this increase was located within Piripiri, Pohangina and Makawakawa catchments, which at this time contained 70.8% of the total eroded area in the southern Ruahine Range (Table 12.1, column 7). Further to the south No. 1 Line, Opawe and Mangatuatou catchments contained substantial eroded areas, though by the 1970s they had a lesser percentage than in the 1940s. Konewa, Te Ekaou, Te Awaoteatua and Maungatukurangi catchments continued to be sites of least actively eroding slips (Table 12.1, column 7).

The even distribution of actively eroding slips on the eastern flank of the Range was still apparent in the 1970s with 51.7% of the total eroded area on this flank occurring within catchments northward of Mangapuaka catchment (Table 12.2, column 7). Here, a reduction in the percentage of the total eroded area since the 1940s is due to stabilisation of farmed foothill areas within East Tamaki and Mangatera catchments (Table 12.2, column 7).

It could be interpreted that greater areas of actively eroding slips in the north of the study area are a consequence of larger catchment size. Indeed streams in the north of the area are longer and catchments are of larger size where the Range is widest (13 km). However, the presence of substantial areas of actively eroding slips in the southeastern sector of the Range indicates that factors other than catchment size are important in determining the location of actively eroding slips. In the south, where the range is narrow (7 km) it can be shown that eastern catchments are shorter and steeper so that valley dissection is greater and valley slopes are steeper (Table 10.3a) than in western catchments. These differences account for the greater area of actively eroding slips within southeastern catchments compared with southwestern catchments.

12.2.2 TRENDS

Of the 32 catchments surveyed in this study, all but three show an increase in the area of actively eroding slips between the years 1946-49 and 1974-78; a period of about 28 years. The three exceptions have one factor in common. Within each catchment a substantial area of original forest vegetation was cleared during the early 1900s and replaced with pasture. As a result of forest clearance, the slopes in these areas became unstable and the effects were still apparent on the 1946-49 aerial photographs. Much of the erosion at that time is thought to have been the result of shallow slope movements, such as debris slides and debris avalanches, sheetwash and wind erosion. By the 1970s the area of actively eroding slips within these catchments had declined by 100% in Te Awaoteatua catchment (Table 12.1, column 10), by 79% in the Mangatera foothill area and by 36% in East Tamaki catchment (Table 12.2, column 10).

No attempt was made to quantify the area of actively eroding slips on pastured slopes associated with earth materials flanking the Range. However, visual comparison of Maps 5 and 6 indicates that areas of actively eroding slips within these lower catchment reaches have declined since the 1946-49 period.

Of those upper catchments to show an increase in the percentage of actively eroding slips during the 28 year period, the most dramatic increases have occurred in Porewa, Te Ekaou, Maungatukurangi and Makawakawa catchments on the western flank of the Range; and in Rokaiwhana, Otamaraho, Otamarahu, Coppermine and Mangapapa catchments on the eastern flank of the Range. In each catchment the percentage of actively eroding slips has increased by between 200% and 400% (Tables 12.1 and 12.2, column 10). Catchments with minimal increases in area of actively eroding slips (between 0% and 50%) during this same time period include Makohine, No. 1 Line, Dundas Creek, Mangatuatou, Konewa and Opawe catchments, all on the western flank of the Range and North Oruakeretaki catchment on the eastern flank. Most of the remaining catchments show an increase in percentage of actively eroding slips of between 50% and 150% (Tables 12.1 and 12.2, column 10).

Results of this work, in general, agree with findings of previous authors. James (1973) established that 1.7% of the slope area within a small portion of Pohangina catchment was actively eroding in 1946. By 1963 this area had increased to 2.7% of catchment slopes, an increase of 60% in 17 years (p 98). Table 12.1, column 3 of this study indicates that in 1946, 1.7% of slopes within part of the Pohangina catchment surveyed in this study were actively eroding. By 1974 this area had increased to 3.6% of catchment slopes, an increase of 109% in 28 years (Table 12.1, column 10).

Between 1946 and 1974 there was a 120% increase in the actively eroding slips in No. 1 Line and Raparapawai catchments (Stephens, 1975). This present study indicates that during this same period there was a 119% increase in the area of actively eroding slopes within Raparapawai catchment (Table 12.2, column 10) but only a 36% increase in No. 1 Line catchment (Table 12.1, column 10). There is little correlation between the results of this study and those of Mosley (1977, Table 7).

All but two catchments contain a significant area of actively eroding slips during the 1946-49 period. These two catchments (Te Ekaou and Maungatukurangi) were essentially devoid of actively eroding slips at this time. Therefore, the appearance of an area of actively eroding slips within each catchment by 1974, though small in areal extent, has resulted in a greatly exaggerated percent increase in actively eroding slip areas (Table 12.1, column 10).

The sum total of all catchment areas on the western flank of the Range is greater than that on the eastern flank of the Range (Tables 12.1 and 12.2, column 1). It is therefore not surprising that the total area of actively eroding slips on the western flank is greater both during the 1946-49 and 1974-78 periods. It is, however, interesting to note that the area of actively eroding slips expressed as a percentage of the total area on each flank of the Range indicates that in the period 1946-49 a higher percentage (1.4%) of slopes on the eastern flank were actively eroding compared to 1.2% on the western flank (Tables 12.1 and 12.2, column 3). However, during the interval 1946-1974, there occurred a greater increase (103%) in the area of actively eroding slips on the western flank of the Range compared with a lesser 76% increase on the eastern flank (Tables 12.1 and 12.2, column 10). By the 1970s the percentage of actively eroding slips on each flank of the Range was identical and measured 2.4% (Tables 12.1 and 12.2, column 5).

The higher percent increase on the western flank of the Range is due to the greater rate at which actively eroding slips have expanded in areal extent during the years 1946 to 1974 (Table 12.4). An increase from 53% in 1946 to 56% in 1974 is matched by a corresponding decrease on the eastern flank from 47% in 1946 to 44% in 1974 (Tables 12.1 and 12.2, columns 8 and 9).

The sum total of upper catchment areas within the southern Ruahine Range approximates 27 132 hectares (Tables 12.1 and 12.2, column 1) of which approximately 2.4% (Tables 12.1 and 12.2, column 5) is considered to have been actively eroding during the years 1974-78. This figure compares favourably with an assessment of eroded areas in the southeastern Ruahine Range by Cunningham & Stribling (1978, p38) of 2.7%. These are, however, mean values and do not reflect the true condition of individual catchments within which areas of actively eroding slips vary between less than 1% and up to 6% of total catchment areas. On the western flank of the Range, catchments within which a high percentage of the total catchment area is actively eroding include No. 1 Line (4%), Ohinetapu (3.9%), Mangatuatou (3.5%), Pohangina (3.6%) and Piripiri (5.3%) (Table 12.1, column 5) and on the eastern flank, the West Tamaki (3.8%), Otamaraho (5.8%) and Mangapuaka (5.1%) catchments (Table 12.2, column 5).

During the 28 year interval between 1946 and 1974 the area of actively eroding slips on the western flank of the southern Ruahine Range increased by 103% (Table 12.1, column 10) increasing in area from 180 ha to 366 ha (Table 12.1, columns 2 and 4). On the eastern flank, the area of actively eroding slips increased by 76% (Table 12.2, column 10) from 160 ha to 282 ha (Table 12.2, columns 2 and 4). This totals an overall increase in area of actively eroding slips of 91% for the southern Ruahine Range in 28 years from 340 ha in 1946 to 648 ha by 1974.

12.2.3 TIMING OF INCREASED SLOPE INSTABILITY

The majority of actively eroding slips present on the 1946-49 aerial photographs are still clearly visible on aerial photographs taken 15 years later in 1961. It appears that this intervening period was not one of marked increase in slope movement because many of the original slips had not enlarged greatly as a result of subsequent movements. Also, very few new sites of active slope movement were initiated during this period. Within this 15 year interval some revegetation of erosion scars had taken place but in spite of this the majority of scars appear to have been sites of continuing movement up to 1961.

Comparison of aerial photographs taken in 1961 and 1966 indicates a substantial increase in the area of actively eroding slips in almost all catchments over this five year period. The same trend was noted by Stephens (1975) in Raparapawai catchment. This increase resulted predominantly from enlargement and amalgamation of old erosion scars identifiable on earlier photographs but many new areas of active erosion were also initiated. A trend of enlargement of old erosion sites and initiation of new ones was again very apparent in 1975. This was the last major episode of extensive slope movement in the southern Ruahine Range.

For reasons stated previously, a measure of the rate of increase in areas of actively eroding slips at these times was not possible.

The relationship between the above episodes of increased slope instability and periods of heavy rainfall is discussed in Section 10.3.

12.2.4 EROSION RATES

Erosion rates have been calculated using a method employed by James (1973) in which he measured areas of actively eroding slips from aerial photographs taken at different times. This method was also used by Stephens (1975) and Mosley (1977) for other parts of the Range and by this author for all catchments within the study area (see Section 12.2).

A. Annual Increase in Eroded Area

The annual increase in eroded area is greater on the western flank (429 $m^2/km^2/yr$) than on the eastern flank ($374m^2/km^2/yr$) of the Range (Table 12.3). It is, however, noticeable that the highest increase in eroded areas on the western flank occurred within a few catchments in the northwestern part of the study area. Here, the high rates of increase occurred within the largest catchments, *e.g.* Makawakawa ($637m^2/km^2/yr$) and Pohangina ($667m^2/km^2/yr$), within which there have been extensive areas of actively eroding slips since the 1940s, most of these having enlarged substantially in later years. Alternatively, some of the smaller catchments, *e.g.* Piripiri, show high rates of increase ($894m^2/km^2/yr$) because of the presence of large-scale, deep-seated slope movements that have triggered shallow debris slides and debris avalanches.

In contrast, high rates of increase in eroded area on the eastern flank are more evenly distributed along the length of the Range but are highest north of Otamarahu catchment (Table 12.3). The higher rates of increase in eroded area throughout the north of the study area are thought to be due to the presence of steeper valley slopes, and deeper valley and slope dissection (see Section 10.2) in an area subject to higher, more intense rainfall.

The highest rate of increase in eroded area ($1661m^2/km^2/yr$) occurs within Otamaraho catchment (Table 12.3). This rate is approximately twice that of adjacent catchments, within which the strata are composed of identical lithologies with similar structural attitude, valley slopes are of similar steepness, aspect and altitude, the condition of forest vegetation cover is not noticeably different to that in neighbouring catchments and the dominant types of slope movement are the same. The reasons for the increased instability of slopes within Otamarahu catchment are not clear.

TABLE 12.3: Calculation of annual erosion rates 1946-74.

Western catchments	Annual increase in eroded area ($m^2/km^2/yr$)*	Annual erosion rate ($m^3/km^2/yr$)†	Eastern catchments	Annual increase in eroded area ($m^2/km^2/yr$)*	Annual erosion rate ($m^3/km^2/yr$)†
Maungatukurangi	161	483	West Tamaki	800	2400
Whareroa	148	444	East Tamaki foothills	-99↓	-297
Makohine	74	222	Mangatera foothills	-640↓	-1920
Tokeawa	344	1032	Rokaiwhana	823	2469
Te Awaoteatua	-72↓	-216	Rokaiwhana foothills	58	174
No. 2 Line	367	1101	Otamaraho	1661	4983
No. 1 Line	374	1122	Mangapuaka	873	2619
Dundas Creek	123	369	Otamarahu	723	2169
Ohinetapu	691	2073	Mangapukakakahu	408	1224
Mangatuatou	97	291	North Oruakeretaki	84	252
Opawe	45	135	South Oruakeretaki	390	1170
Porewa	717	2151	Raparapawai	649	1947
Te Ekaou	162	486	Manga-a-tua	519	1557
Makawakawa	637	1911	Coppermine	624	1872
Konewa	0	0	Mangapapa	265	795
Pohangina	667	2001			
Piripiri	894	2682		374	1122
	429	1287	TOTAL CATCHMENTS	405	1215

* Calculations based on figures from Tables 12.1 and 12.2
↓ Negative values indicate a reduction in area of actively eroding slips in these catchments between the years 1946-74
† Calculations based on observation that average depth of erosion scars is 3m.

The negative values in table 12.3 indicate a reduction in the area of actively eroding slips between the years 1946-74 within foothill catchments that are predominantly pastured farmland (discussed in Section 10.2).

B. Annual Erosion Rate (volume)

While many of the translational slope movements are of less than 2m depth (see Section 10.1.1), other translational debris avalanches, rock slides and rotational slumps often fail at depths between 2m to several tens of metres. Here, the calculation of volumes of material derived from slope movements is based on the average depth of the most common form of slope movement.

On the basis of extensive field observation, a mean depth of 3m has been assumed. As this is a constant, then volume variation will be the same as area variation. As expected, the results (Table 12.3) indicate that the highest annual erosion rate occurs in those catchments in the northern sector of the study area where the annual increase in eroded area is greatest. On the western flank of the Range these rates range from $1911m^3/km^2/yr$ in Makawakawa catchment to $2682m^3/km^2/yr$ in Piripiri catchment. On the eastern flank of the Range high erosion rates occur in catchments throughout the length of the Range from Coppermine ($1872m^3/km^2/yr$) in the south to West Tamaki ($2400m^3/km^2/yr$) in the north. The highest erosion rate of any catchment throughout the Range was determined for the Otamaraho catchment at $4983m^3/km^2/yr$ (Table 12.3).

Comparison of annual erosion rates obtained in this study generally concur with those of Stephens (1977) but differ from those of Mosley (1977). For example, on a per catchment basis, Stephens calculated a rate of $2000\text{-}3000m^3/km^2/yr$ (Table 5.1, p 30) for West Tamaki catchment which compares favourably with a rate of $2400m^3/km^2/yr$ (Table 12.3) obtained in this study. Both results, however, differ markedly from a rate of $5850m^3/km^2/yr$ determined by Mosley (1977, Table 7). Similarly, Stephens (1977, Table 5.1) calculated a rate of $1800\text{-}2700m^3/km^2/yr$ for Raparapawai catchment; in this study $1947m^3/km^2/yr$ (Table 12.3) was obtained but this differs from $5200m^3/km^2/yr$ determined by Mosley (1977, Table 7). However, disturbingly, no correlation exists between calculations of the annual erosion rate for No. 1 Line catchment. Stephens (1977, Table 5.1) obtained a rate of between $2260\text{-}3390m^3/km^2/yr$ whereas $1122m^3/km^2/yr$ (Table 12.3) was calculated for the same catchment in this study.

TABLE 12.4: Annual erosion rates, 1946-74, southern Ruahine Range.

Catchment	Upper catchment area (ha)	Area of actively eroding slips (ha) 1946	Area of actively eroding slips (ha) 1974	Annual increase in eroded area ($m^2/km^2/yr$)	Annual erosion rate ($m^3/km^2/yr$)
Western flank	15 479	180	366	429	1287
Eastern flank	11 653	160	282	374	1122
TOTAL CATCHMENTS	27 132	340	648	405	1215

TABLE 12.5: Annual erosion and sediment transport rates, New Zealand and overseas.

Locality	Annual erosion/ transport rate ($m^3/km^2/yr$)	Comments	Author
NEW ZEALAND EROSION RATES:			
Tongawai catchment Opihi Opuho	74 84 66	South Island high country, tussock vegetation. Greywacke and schist.	Cuff, 1974
Hapuakohe Range	250-500 1000	Under forest) greywacke Under pasture) range	Selby, 1976
Tangoio, Napier	1500	Under pasture, 5 yr return period storm. Tertiary rocks.	Eyles, 1971
OVERSEAS EROSION RATES:			
Areas of steep relief	Mean value 500	Includes mountainous areas.	Young, 1969
Mountain areas	915		Schumm, 1963
Papua-New Guinea	800-1000		Pain & Bowler, 1973
Japan	1000	Tanzawa mountains.	Tanaka, 1976
Swiss Alps	400-1000		Clark & Jager, 1969
Himalayas	700		Curry & Moore, 1971
NEW ZEALAND TRANSPORT RATES:			
Folgers Lake	1925	Tukituki Stream, Ruahines.	de Leon, *pers com*
Mangahoa Roxburgh Lake Frazer Dam	1580 259 32	Tararua Range, greywacke) South Island schist)	Thompson, 1976
Torlesse Stream	125	Sth Island, greywacke	Hayward & Blakely, 1976
Shotover River catchment	820-2340	Sth Island, schist	Min. of Works, 1975

(modified from Stephens, 1977)

The average annual erosion rate for the western flank of the Range is about $1287m^3/km^2/yr$ compared with $1122m^3/km^2/yr$ for the eastern flank of the Range. This gives an average annual erosion rate for the entire southern Ruahine Range of $1215m^3/km^2/yr$ (Table 12.3) over an area of 27 132 hectares. The annual erosion rate was calculated for ten catchments along the eastern flank of the southern Ruahine Range by Mosley (1977). Eight of these catchments occur in the present study area and two lie to the north of the study area. His calculations were based on the assumption that the mean depth of erosion scars in these ten catchments is 3m from which he calculated an annual erosion rate of $2870m^3/km^2/yr$ (p 28). This figure is approximately twice the annual erosion rate calculated in the present study.

The above quoted annual erosion rate obtained by Mosley is considered to have been inflated by the inclusion of exceptionally high erosion rates of $5200m^3/km^2/yr$ and $5850m^3/km^2/yr$ obtained for Raparapawai and West Tamaki catchments, respectively. Evidence to suggest that these figures are not representative of the magnitude of erosion currently taking place in these two catchments is the close correlation between annual erosion rates calculated by Stephens (1977) and during this study. If it is accepted that the annual erosion rates determined for Raparapawai and West Tamaki catchments by Mosley (1977) are incorrect, and they are then excluded from calculations of erosion rates then the annual erosion rate for the remaining six catchments studied by Mosley (1977, Table 7) along the eastern flank of the Range is $2200m^3/km^2/yr$. Figures obtained in the present study for the same six catchments indicate an annual erosion rate of $1900m^3/km^2/yr$. This higher correlation illustrates the point that the annual erosion rate for catchments along the eastern flank of the southern Ruahine Range is not as substantial as was previously thought. For this reason it is considered likely that the average annual erosion rate for the southeastern Ruahine Range approximates to $1122m^3/km^2/yr$ rather than to a rate of $2870m^3/km^2/yr$ as proposed by Mosley (1977) and that higher annual erosion rates of around $2000m^3/km^2/yr$ are locally restricted to the northern sector of the study area.

Periods during which a higher than average erosion rate existed in the southern Ruahine Range is not restricted to present times. It can be inferred, from the distribution of aggradational alluvial terrace surfaces and fan deposits within the Range, that erosion rates in the past were at times considerably higher than at present. One such period of valley slope erosion and subsequent terrace aggradation, recorded in the northeastern sector of the study area, has been dated at around 12 000 years B.P. (see Chapter 11). This date corresponds with a period of extensive terrace aggradation in the southeast of the study area dated at around 13 000 years B.P. (see Chapter 6). It is here suggested that the overall pattern during post-glacial times is one of declining erosion rates (see Chapter 11).

It is clear that contemporary erosion rates in the southern Ruahine Range are not exceptional despite increases in erosion rates within historical times and in particular since the 1940s. Periods of erosion are episodic in nature and differ in intensity and duration through time. Periods of instability in which erosion rates are often accelerated are separated by periods of less active erosion. Periods of instability may be caused by many factors such as increased tectonic activity (including uplift, fault displacement and earthquake shaking), changes in weather and climatic patterns (*e.g.* onset of glaciations or the duration and frequency of intense rainstorms). The past and/or present effects of any one or combination of these factors will vary considerably within relatively small areas so that erosion rates may vary significantly over short distances. Thus areas with high erosion rates can be found within short distances of areas with low erosion rates.

C. Long Term Erosion Rate

In order to compare present-day erosion rates with erosion rates of the past Mosley (1977) estimated a long-term minimum erosion rate by calculating the volume of material removed from four valleys along the eastern flank of the southern Ruahine Range. As he pointed out, this method is based upon some very restrictive assumptions, the most important of which is the time period during which uplift of the Range occurred. On the assumption that uplift of the Range has occurred in the last 1.5M years, Mosley attained an erosion rate of $868m^3/km^2/yr$. On the basis of a suggestion that uplift of the Range has in fact occurred within the last 0.8M years, Mosley recalculated the erosion rate at $1638m^3/km^2/yr$ (p 27). A minimum long term erosion rate of $1000m^3/km^2/yr$ was considered representative. Mosley then deduced that since about 1940 A.D. the annual erosion rate of $2780m^3/km^2/yr$ in this area has approximately doubled from that of the long term minimum erosion rate.

The annual increase in eroded area within catchments in the northeast of the study area is higher and atypical of erosion rates in other catchments further to the south. It therefore follows that a mean long term erosion rate, based upon these catchments, will be an overestimate.

D. Comparative Erosion Rates

When erosion rates from the southern Ruahine Range (Table 12.4) are compared with erosion rates determined from elsewhere (Table 12.5), they more closely approximate those determined from overseas mountainous areas and in New Zealand from the Hapuakohe Range than those rates determined from South Island high country areas.

The erosion rate of $1215 m^3/ha/yr$ for the southern Ruahine Range is comparable to most other areas listed in Table 12.5. Although short term (30 yrs or less) bedload transport rates do not necessarily indicate erosion rates, they have been included to give an indication of their comparative magnitude. Differences may be explained by any number of factors, the most significant of which include absence of measurement of suspended loads, size of the catchment(s) under consideration, the proportion of lowland areas as opposed to upper catchment area, the nature of the bedrock, the amount of valley dissection, the climatic regime of the area, aspect of the catchment(s) and the type of vegetative cover.

The significance of the erosion rate determined for the southern Ruahine Range is that it is equally as high as for those areas where erosion rates of a similar magnitude are considered a serious problem.

12.3 LARGE-SCALE, DEEP-SEATED MASS MOVEMENTS

12.3.1 STATE OF ACTIVITY

Large-scale, deep-seated mass movements within this area are of gravitational origin and involve very large quantities of either earth or bedrock materials. Their classification is discussed in Chapters 9 and 10. An assessment of changes in their state of activity between the years 1946 and 1978 has been attempted by comparing aerial photographs spanning this time period. Details of location (Maps 5 and 6) and classification (Appendix VI) are outlined.

Photographic interpretation of these mass movements indicates that all have moved downslope at least once and some have undergone several episodes of movement. Further, whilst some movements clearly pre-date historical times, others have failed in recent times. These differences have been used as a basis for grouping mass movements with similar failure history.

A. Group A

This group comprises all the ridge-top features, earth slumps and earth slides and the majority of rock slumps. The age of this group of mass movements is largely unknown. All failed prior to 1946 and none show signs of subsequent downslope movement. At localities where complete removal of the slumped mass has occurred, *e.g.* locality 58 in No. 1 Line catchment, the re-establishment of a dense forest vegetation at these sites is interpreted as evidence of a prehistoric age. At the majority of localities downslope movement of the slumped mass was minimal and did not severely affect the original forest vegetation. An attempt by Stephens (1975) to date the movement of one slump within Raparapawai catchment, using tree ring dating, proved unsuccessful. Earth slumps and earth slides to the west of the Range have been cleared of their original forest cover so no age estimates can be made. Where earth sliding has occurred, the entire mass of material has been removed. Such removal would require a considerable period of time.

B. Group B

These mass movements are not present on the 1946-49 aerial photographs but have failed at some time prior to 1974-78. These include localities 4, 5a and 5b in Pohangina catchment and localities 20 and 21 in Makawakawa catchment.

C. Group C

These mass movements are indistinct on the 1946-49 aerial photographs but show an obvious slump outline by 1974-78. These include localities 32a (Rokaiwhana catchment), 35 (Otamaraho catchment), 49 (Mangapukakakahu catchment), 53 (Raparapawai catchment) and 67, 68, 69, 70 (Manga-a-tua catchment). It cannot be proven that further episodes of slump movement since 1949 have occurred at each locality. However, shallow translational slides associated with all of these mass movements have noticeably increased in area.

D. Group D

These include mass movements that show obvious slump outlines on the 1946-49 aerial photographs and have failed prior to 1974-78. An example is locality 73 in Coppermine catchment. Analysis of aerial photographs taken in 1961 indicate that this slump failed prior to 1961.

E. Group E

These include mass movements that are not visible on the 1946-49 photographs but are known to have failed in recent years. The slump at locality 72 in Coppermine catchment failed in August 1976 (Mosley & Blakely, 1977).

Slumps are often sites of obvious debris slide and debris avalanche activity. In particular the headwall, lateral scarps and toe slopes are prone to shallow translational slide activity. Here various stages in seral development of the regenerated vegetation not only highlight the outline of the slump but also indicate that there have been several episodes of instability during which shallow translational sliding occurred. It is not certain whether discrete episodes of translational sliding are due to periodic slump movement or whether other influencing factors cause the headwall and lateral scarps to periodically fail. In Raparapawai catchment the increased incidence of shallow translational slope movement on the toe slope and around lateral and headwall scarps of the rock slump at locality 61 has been attributed to gradual downslope movement of this slump since 1946 (Stephens, 1975, p 68). Whilst shallow translational slope movements associated with the slump at locality 61 are clearly increasing in areal extent, no evidence was found in the current study to suggest that this was due to downslope movement of the slump.

At localities where recent slump movement cannot be demonstrated but where shallow translational movements have occurred nonetheless, *e.g.* Group C slumps, it is more likely that translational sliding is not due to renewed slump movement but to processes such as streambank undercutting, or the result of seismic shaking and storm events.

Shallow translational slides are not a feature of areas where ridge-top features have formed. Their state of activity has remained unchanged since 1946.

Some earth slides and earth slumps show increased shallow translational sliding in headwall and toe slope areas since 1946, *e.g.* at locality 86. This is not thought to be due to recent downslope movement of the slump but instead is considered to be the result of climatic influences. Other earth slumps, however, show distinct signs of increased stability, particularly where the grass covering on headwall and lateral scarps and in toe slope areas has reverted to a shrubby vegetation, *e.g.* localities 77-79.

These earth slumps show no change in their state of activity since 1946.

12.3.2 STABILITY

Movement of slumps and ridge-top features may relieve stress within earth or rock materials and therefore temporarily enhance the stability of slopes at these sites. However, movement of a mass also results in a lowering in material strength. Once movement has occurred the probability of repeated movements occurring at these localities is greatly increased. At many of the localities where mass movements have been documented, the strength of the materials has been severely reduced as a result of either extensive fault brecciation or by disruption of the material during former large-scale, deep-seated mass movements. Hence fault zones and sites of former mass movement determine to a large extent the location of more recent slump movements.

It is here shown that at each of the localities recorded in Appendix VI at least one phase of downslope movement has occurred and at some locations repeated movement can be demonstrated. However, it is not possible to establish whether slopes at these localities are of greater or lesser stability than adjacent slopes because it can be shown that similar mass movements are just as likely to occur at localities where former mass movement had not taken place, for example locality 72. The rock slump at locality 61 is considered to be in a precarious position and may slide into Raparapawai Stream channel (Stephens, 1975, p 70). It is highly probable that the ridge-top benches at locality 18 will respond to gravity collapse during, for example, an intense rainstorm (Hubbard, 1978, p 179). It is suggested therefore that large-scale, deep-seated mass movements in this environment are inevitable and may occur on any slope. Many such movements in the future will be unpredictable. On the other hand, some future movements can be predicted at those localities where downslope movement is presently in evidence. Because collapse of these large-scale mass movements may be hazardous to man, it is important to identify and document their presence.

12.3.3 VOLUME

An estimate of the volume of material involved in individual mass movements has been determined. The depth of the concave plane of failure along which rotational slump movement takes place is highly variable both within a single slump and between slumps. Much of this variation is determined by valley-slope height, steepness, length and size of the slump. This, in turn, affects the degree of curvature of the plane of failure. A measure of the volumes of material involved in slumps is possible by assuming a constant depth for all slumps. Then, as for all translational slides, the volume equals area times average depth. At localities where slumped masses have been totally removed from sites of failure, it has been determined that the average depth of slumps in this area approximates 6m depth.

The single largest mass movement involving earth materials (locality 90) is estimated to contain approximately 4.6 x 10^6 m^3 of material (Table 12.6). Within Te Awaoteatua and No. 1 Line catchments several earth movements (slides and slumps) occur within small localised areas in the lower reaches of each catchment. Six earth slumps within Te Awaoteatua catchment collectively contain 3.4 x 10^6 m^3 of earth materials and within No. 1 Line catchment three earth slumps total 3.7 x 10^6 m^3 (Table 12.6).

The majority of slumps in the southern Ruahine Range are rock slumps and occur within upper catchment reaches. The single largest rock slump incorporates approximately 2.4 x 10^6 m^3 of bedrock (Table 12.6) within Piripiri catchment (locality 2a). Other rock slumps of substantial size include localities 30 (Rokaiwhana), 98 and 99 (West Tamaki), 36 and 38 (Mangapuaka), 53 and 61 (Raparapawai), 4 (Pohangina) and 23 and 24 (Makawakawa). Volumes of bedrock material at these localities and at others throughout the Range are indicated in Table 12.6.

12.3.4 CONSEQUENCES OF LARGE-SCALE, DEEP-SEATED MASS MOVEMENTS

The consequences, both within and beyond the Range front, of failure of large-scale, deep-seated mass movements depend on several factors, including: (1) the amount of downslope displacement of each mass movement during a period of instability; (2) volume of material involved in each movement; (3) the number and location of mass movements within upper catchment reaches; (4) the number of mass movements that fail simultaneously during a period of instability; (5) the groundwater conditions prior to and during a period of instability; and (6) the flow conditions of streams at the time of a period of instability.

Most of the potential sizes of large-scale, deep-seated mass movement have been identified by the presence of either an ovate slump outline or a linear ridge-top feature. Each represents a phase of downslope movement which at most localities has resulted in only minor amounts of vertical displacement. When further downslope movement occurs at these localities their potential as a hazard will be dependent upon how much downslope displacement takes place. Should displacement be minor, the effect, particularly at those localities within the Range, will be minimal. Then further headwall, lateral scarp and toe slope instability will be evidenced. A potential hazard may arise at localities such as 84 and 85 where minor vertical displacement of earth slumps could result in partial or total closure of a road.

Should a major vertical displacement occur at any of the documented localities (Appendix VI) where downslope movement has occurred previously, resultant damage will largely be a function of the size and volume of material involved in the movement. Within upper catchment reaches many of the rock slumps contain considerable volumes of bedrock (Table 12.6), displacement of which could result in the damming of streams to form a sizeable landslide - dammed lake (see Section 10.3.2).

Such debris dams may constitute a hazard of considerable magnitude depending upon whether the dammed water was released gradually through downcutting of the slump debris or whether the dammed water was released suddenly due to collapse of the debris dam. In the event of collapse of a debris dam, a debris flow is likely to form that could be capable of travelling a substantial distance to reach the Range front and beyond. The probability of landslide-dammed lakes forming as a result of rock slumping is high for all catchments within which rock slumps and ridge-top features occur. The narrow, deeply-dissected valleys in this area are conducive to damming. The greatest potential for the formation of landslide-dammed lakes occurs in those catchments documented in Table 12.6.

Damming will not occur at localities where mass movement features are small. However, debris flows are likely to result from such movements. The size and distance travelled by a debris flow will depend upon how many mass movements fail at any one time. If more than one slump fails simultaneously the greater is the chance of a resultant debris flow reaching the range front. Location within catchments is also important because a debris flow resulting from mass movement close to the range front is likely to flow beyond the confines of the Range onto the adjacent lowland. Conversely, a debris flow originating from a mass movement in the headwaters of catchments is unlikely to reach the range front on account of the sinuous stream channels where the speed of a flow would be slowed and eventually the kinetic energy dissipated. However, accelerated streambed aggradation would occur in these reaches.

TABLE 12.6: Estimated volumes of material contained in large-scale, deep-seated mass movements.

Catchment	Classification† location number	Classification† type of movement	Area (ha)	Volume ($m^3 \times 10^6$)	Total volume per catchment
Te Ano Whiro	90	e.sl	76.5	4.59	4.59
Piripiri	1	r.sl	12.3	0.74	
	2a	r.sl	39.1	2.35	
	101	e.sl	11.6	0.70	
	102	e.sl	4.3	0.26	4.05
Pohangina	3	r.sl	4.9	0.29	
	4	r.sl	20.7	1.24	
	5a	r.sl	7.9	0.47	
	5b	r.sl	8.5	0.51	
	6	r.sl	10.1	0.61	
	7	r.sl	10.4	0.62	
	8	r.sl	5.7	0.34	
	13	r.sl	3.2	0.19	
	14	r.sl	1.7	0.10	
	84	e.sl	16.0	0.96	
	85	e.sl	7.9	0.47	5.80
Konewa	87	e.sl	2.7	0.16	
	88	e.sl	6.2	0.37	
	89	e.sl	3.3	0.20	0.73
Makawakawa	20	r.sl	10.6	0.64	
	21	r.sl	6.5	0.39	
	22	r.sl	6.4	0.38	
	23	r.sl	16.1	0.97	
	24	r.sl	19.5	1.17	
	25	r.sl	2.1	0.13	
	26	r.sl	4.7	0.28	
	27	r.sl	4.6	0.28	
	86	e.sl	34.0	2.04	6.28
Opawe	83	e.sl	6.6	0.40	0.40
Ohinetapu	52	r.sl	2.9	0.17	0.17

† explanation of abbreviations is given in Appendix VI.

TABLE 12.6: (cont)

No. 1 Line	57	r.sl	5.2	0.31	
	58	r.sl	5.8	*	
	59	r.sl	5.5	0.33	
	77	e.sl	45.4	2.72	
	78	e.sld	5.6	*	
	79	e.sld	5.1	*	
	80	e.sld	10.4	*	
	81	e.sl	8.2	0.49	
	82	e.sl	7.7	0.46	4.31
No. 2 Line	76	e.sld	10.3	*	
Te Awaoteatua	74a	e.sl	14.9	0.89	
	74b	e.sl	7.7	0.46	
	74c	e.sl	4.2	0.25	
	74d	e.sl	8.4	0.50	
	74e	e.sl	4.5	0.27	
	75	e.sl	16.9	1.01	3.38
West Tamaki	11	r.sl	8.2	0.49	
	98	r.sl	15.3	0.92	
	99	r.sl	16.9	1.01	
	100	r.sl	5.3	0.32	2.74
Rokaiwhana	28	r.sl	4.5	0.27	
	30	r.sl	26.0	1.56	
	32a	r.sl	2.9	0.17	
	32b	r.sl	4.2	0.25	
	33	r.sl	3.3	0.20	
	34	r.sl	8.8	0.53	
	35	r.sl	4.5	0.27	3.25
Mangapuaka	36	r.sl	18.1	1.09	
	37	r.sl	6.3	0.38	
	38	r.sl	21.5	1.29	
	39	r.sl	11.4	0.68	
	40	r.sl	12.2	0.73	4.13
Otamarahu	46	r.sl	13.0	0.78	0.78
Mangapukakakahu	48	r.sl	9.7	0.58	
	49	r.sl	3.3	0.20	0.78

* material has been removed from site of mass movement.

TABLE 12.6: (cont)

South Oruakeretaki	54	r.sl	5.3	0.32	
	55	r.sl	3.6	0.22	0.54
Raparapawai	53	r.sl	22.1	1.33	
	61	r.sl	14.4	0.86	
	62	r.sl	9.7	0.58	2.77
Manawatu	108	e.sl	14.7	0.88	0.88

Other important factors include the ground water conditions prior to and during an episode of mass movement instability. Under conditions of saturation the greater is the chance of mass movement failure and the greater is the potential for flowage than when the moisture content of the bedrock is low. Further, if an episode of mass movement coincides with a period of high stream flow the resultant debris flow will travel considerably greater distances than during a period of normal stream flow.

Damming of a stream bed could occur if considerable downslope displacement of some earth slumps were to occur, *e.g.* at localities 74a-e, (Te Awaoteatua Stream), 77, 81 and 82 (No. 1 Line Stream), 83 (Opawe Stream), 86 (Makawakawa Stream), 87-89 (Konewa Stream) and 90 (Te Ano Whiro Stream). Where damming does not occur earth flows will form. The greatest potential for earth flow development is within lower catchment reaches of streams draining westward from the range front. Here the majority of earth flows would be confined to stream channels incised within Plio-Pleistocene marine deposits. Siltation of streambeds and possible structural damage to highway bridges would be caused by earth flows in this area.

In the event of a major episode of mass movement occurring in this area it is considered likely that debris flows would have least impact on lowland farming areas bordering the western flank of the Range and greatest impact on lowland farming areas bordering the eastern flank of the Range. This is because stream channels within the Range are in general longer and less steep on the western flank than on the eastern flank, hence there is little chance of a debris flow maintaining sufficient momentum to reach the Range front. Also, the greater is the potential of westward draining streams to store mass movement detritus. In addition, debris flows upon emerging from the western range front would be channeled along deeply entrenched stream courses that have incised into Plio-Pleistocene deposits to depths of 30m or more. Here, a debris flow would have little or no effect upon farmland adjacent to stream channels, although aggradation could be considerable. In contrast, a debris flow upon emerging from the eastern range front would infill shallowly entrenched streambeds, thereby resulting in overbank flow, especially adjacent to the Coppermine, Manga-a-tua, Raparapawai and Rokaiwhana Streams and West Tamaki River. In some streams a debris flow, on emerging from the eastern range front, would travel upon an alluvial fan surface that is perched above the level of surrounding farmland. Eastward of the range front (valley throats of Mosley, 1977) such debris flows would be unrestricted by channel banks and would fan outwards over adjacent farmland causing much destruction. Such a process could occur at the valley throats of North and South Oruakeretaki Streams, Mangapukakakahu, Otamarahu, Mangapuaka and Otamaraho Streams.

The potentially most dangerous locality, to human life, occurs downstream of Mangapuaka and Otamarahu valley throats. Within Mangapuaka catchment there are three large-scale, deep-seated rock slumps (localities 38-40) located along the main channel, and within Otamarahu catchment a large rock slump occurs at locality 46. All of these rock slumps are found at less than 2 km distance from the Range front. Both stream channels are relatively straight and lack barriers that may restrict or slow down debris flow movement. Should any one of these rock slumps collapse, huge volumes of rock debris will be added to the already choked streambed where further aggradation and downstream fluvial transportation of detritus would cause serious management problems. Should simultaneous collapse of rock slumps within these catchments occur, approximately 2.7 x 10^6 m^3 of detritus would be released into Mangapuaka Stream from localities 38-40 and approximately 0.8 x 10^6 m^3 would be released at locality 46 into Otamarahu Stream (Table 12.6). The resultant debris flows could endanger the lives of people living along Kumeti Road. An additional danger lies in the possibility that if a triggering mechanism created simultaneous collapse of several rock slumps then it would be likely to trigger other mass movement features within these catchments, such as rock slumps at localities 36 and 37 and ridge-top features at localities 41, 42, 44 and 45. Although such an event is speculative, the consequences would be calamatous, widespread and long lasting.

12.4 PREDICTING AREAS OF FUTURE SLOPE INSTABILITY

Natural slopes undergo progressive failure in time. Most of the forces involved in such failures are dependent upon the effects of regional tectonic stresses, amounts and forms of weathering products, physical and chemical action of groundwater and seasonal variations of rainfall. The time required for deep-seated movements to develop is almost impossible to evaluate. In contrast, near-surface failures such as debris slides and debris avalanches tend to be more sensitive and are triggered almost immediately, either during or shortly after the triggering conditions have occurred.

There is no sure way to evaluate the stability of an existing slope, or to predict whether or not a landslide will occur. However, analyses of slope movements can provide information on the mechanics and patterns of their formation. By using information from existing or former slope movements, one can extrapolate and make predictions concerning possible sites of future slope movement. Predictions are somewhat easier to make in areas where slope movement has already occurred. It is not so easy to define, in advance, sites where there is potential for movement that has not yet occurred. However, by using known physiographic and geologic criteria together with density of known episodes of slope movement, an attempt has been made to predict areas of future instability (Map 7). Much of this prediction is based on the premise that areas involved in slope movement in the future will be larger than those comprising the suspected activity or known movement because most slope failures enlarge with the passage of time. Moreover, many unstable areas are much larger than first suspected from the obvious overt indications of activity. Particularly in steep terrain like the southern Ruahine Range, failure may progress to the top of a slope, or to a major change in lithology or slope angle. The extent to which slope movements develop is dependent largely upon factors outlined in Sections 10.1 to 10.5.

In making predictions over a given time range, there is a difficulty that predictions based upon observations during a period in which the climatic conditions are less severe than the average will prove too optimistic. Those made during a period in which the climatic conditions are more severe than the average may appear too pessimistic. For this reason, a time scale has not been included.

12.4.1 AREAS SUSCEPTIBLE TO INSTABILITY

Map 7 is designed as a guide for use in the evaluation of slope movement potential. It is not intended to predict precise boundaries of movements but rather to identify and rank on a broad scale areas of different potential for movement. Subdivision is based largely upon physiographic factors (*e.g.* slope steepness and aspect) and on drainage patterns (*i.e.* density and degree of dissection). Within each broad subdivision, influences relevant to slope movement potential such as rainfall regime, materials involved and structural attitude of discontinuities are indicated. Where possible the predominant form of slope movement most likely to occur under these conditions is indicated.

Within these subdivisions there are certain vulnerable locations that are conducive to slope movement. Typical locations include: (A) areas of steep slope; (B) banks undercut by stream action; (C) areas of drainage concentration and seepage zones; (D) areas of former slope movement; (E) areas of fracture concentration; and (F) fault zones.

A. Areas of Steep Slope

The majority of valley slopes within the Range are between 25-33°. However, steeper slopes of between 35-50° are particularly common along major drainage channels in areas where valley incision is deepest. The steepest slopes are, however, not necessarily the most vulnerable because aspect is particularly important in this area (see Section 10.2). North-, northwest- and northeast-facing slopes are in general more susceptible to instability than other slopes. Hence the amount, duration and intensity of rainfall, together with other climatic influences originating from a northwesterly direction greatly influence the stability of these slopes. The greatest potential for slope movement occurs on such valley slopes. Other valley slopes within the Range are less susceptible to instability predominantly because they are less steep and little dissected. Such slopes occur in the northwest of the area within Konewa and Te Ekaou catchments that comprise an area of marine planation. Here, valleys are not as deeply incised as elsewhere and steep valley slopes are restricted to a narrow strip immediately adjacent to drainage channels. The remainder of the valley slope to the ridge crest is little dissected as there are few tributary channels upon this part of the slope. Similarly, in the southwest of the area, between No. 2 Line and Wharite Trig. and including the northern end of Tararua Range, valley slopes are in general considerably less susceptible to slope movement. The greater stability of this area is primarily the result of less well developed valley incision hence valley slopes are not as steep as elsewhere. There are fewer tributary channels on valley slopes hence valley slope dissection is considerably less than elsewhere. Also of considerable importance is that rainfall amount is susbtantially less in this area than further to the north. The most common cause of landslides on slopes within the Range is colluvium and soil sliding over the underlying material. This permeable, loose, unconsolidated material of low shear strength cannot maintain as steep a slope as the underlying material and is, consequently, in delicate balance. The most common forms of slope movement on steep slopes are debris slides and debris avalanches. Any of several factors, such as sudden heavy rainfalls, earthquake shaking or the excavation of the toe of a slope may result in movement.

Valley slopes, within a foothill area lying immediately to the east of the Range are, in part, equally as steep as those within the Range; however, the stability of these slopes appears to be considerably greater. These slopes are less susceptible to movement than those within the Range largely because they are 500m lower than the height of the Range and due to a rainshadow effect, the amount of rainfall is considerably less.

Lowland areas immediately adjacent to the Range consist of moderate to steep slopes (10-25°) that have developed on weakly indurated marine deposits of low inherent shear strength. Although slope movement at present is not particularly well developed, there is evidence to suggest that these areas in the past have been, and in the future are likely to be, susceptible to instability of considerable magnitude.

B. Banks Undercut by Stream Activity

Banks along stream reaches within the Range are particularly susceptible to instability. Here numerous slope movements are the result of streambank undercutting into unconsolidated materials principally riparian terrace and fan alluvium or colluvium. Debris falls, debris slides and sometimes small-scale slumps form as a result of streambank undercutting at these localities. Streambank undercutting into Torlesse bedrock is relatively uncommon but does occur at the points of maximum curvature of streams. Generally undercutting of bedrock slopes is restricted to areas where the bedrock has either been disrupted as a result of large-scale slope movements or has been brecciated as a result of fault movement. Rarely will streambank undercutting occur where a discontinuity strikes parallel with the stream channel. Streambank undercutting into steep bluffs of weakly indurated marine deposits creates instability along the lower reaches of westward draining streams. Here earth falls, earth topples, earth slides and earth slumps are likely to occur as a result.

C. Areas of Drainage Concentration and Seepage

The majority of slope movements are concentrated along existing stream channels. The denser the concentration of streams, whether ephemeral or perennial, the greater the incidence of slope movement. The deeper the dissection of the stream, the steeper and longer are the slopes adjacent to the stream channel, and hence the greater the instability of such slopes.

Groundwater seepage in association with areas of slope instability has been documented in fault zones, upon slopes where soils are prone to saturation and in areas where structural discontinuities such as fault planes, bedding planes and open joints permit seepage of groundwater to the surface. The dominant forms of slope movement in these areas include slumps, debris slides, debris avalanches and rock slides.

D. Areas of Former Slope Movement

The presence of hummocky ground with characteristics inconsistent with the general regional slope or the presence of scarps, benches and depressions are often an indication of large-scale slope movement. Once such movements have been identified, they serve as a warning that the general area has been unstable in the past and that further instability is likely to occur in the future. To date, 109 such localities have been documented (Appendix VI), both within the Range in bedrock materials, and in areas adjacent to the Range in earth materials. These movements have been identified as slumps, and ridge-top benches, scarps and depressions.

E. Areas of Fracture Concentration and Bedding Planes

Slope movements involving Torlesse bedrock are primarily controlled by the presence of lithological and structural discontinuities that may be: (1) inherent within the bedrock, *e.g.* bedding planes; or (2) the result of subsequent deformation, *e.g.* cleavage, joints and faults. The horizontal and vertical continuity together with spatial distribution and width of these discontinuities largely determine the size, shape and depth of failure within bedrock materials. Bedding planes are the most common discontinuity associated with slope movements that involve Torlesse bedrock. The continuity and frequency of bedding plane surfaces is particularly well developed in the regularly bedded strata comprising the Tamaki Lithotype and less well developed in the tectonically disrupted foliated strata of the Wharite Lithotype. Slope movements resulting from bedding plane failure are therefore most commonly found in association with valley slopes underlain by strata comprising the Tamaki Lithotype. These slope movements include some debris avalanches and all rock slides. Valley slopes susceptible to such failure occur where bedding planes parallel the slope. The most favourable locations occur where slopes are steeper than the dip of the bedding planes (see Section 10.1.1). Within the study area these slope movements are generally of small size. Their development is limited because:

(1) The majority of valley slopes are east-west aligned and therefore lie across the northeast-southwest strike of the strata and other prominent discontinuities such as faults and one joint set. Valley

slopes susceptible to bedding plane movements are therefore predominantly restricted to small tributaries and gullies that are aligned parallel with the northeast-southwest strike of the strata;

(2) Bedding plane dip is predominantly steeper than valley slopes, thus bedding planes seldom crop out on slopes at inclinations conducive to sliding. However, small-scale flexures in the bedrock sometimes lessen bedding plane dip to angles less than that of valley slopes and thereby increases slope instability;

(3) Frequent localised reversals in dip direction of bedding planes further limit those sites where dip direction coincides with valley slope (see Section 10.1.1).

Slope movements resulting from failure along joint surfaces are most commonly associated with outcrops of very thick sandstone and outcrops of alternately bedded sandstone and argillite. The two dominant joint systems with which rock fall activity has been recognised have poor horizontal and vertical continuity. Hence rock falls tend to be of small size and localised in occurrence. The structural attitude and orientation of joint surfaces with respect to valley slope controls rock fall activity to a large extent (see Section 10.1.1). Cleavage and low angle thrust faults are not recognised fracture surfaces with which slope movements can be directly related.

Fracture surfaces within earth materials are poorly developed and difficult to recognise. Bedding planes in earth materials are few because of their massive nature, and are rarely found in situations conducive to sliding. However, limited slope movements resulting from sliding along bedding planes have occurred in the past so must be considered areas susceptible to future instability.

F. Fault Zones

The major faults in this area displace Torlesse bedrock and are considered to be active (see Chapter 6). Nearly all active faults are known to have moved more than once and repeated movements have taken place along many of the fault lines. Therefore future movements can also be expected to occur on existing fault lines. Movement of the fault will cause slope instability either by ground displacement or as a result of seismic shaking (see Section 10.4.3).

Fault zones comprise extensive areas of fault brecciated and disrupted bedrock. In many cases brecciation has destroyed previously existing structural elements within the bedrock, thereby reducing the shear strength of this material. Slope instability within areas of fault brecciated bedrock arises when such material comprises steep valley slopes and is thus susceptible to gravitational failure. In this study a direct relationship between fault zones and the incidence of large-scale, deep-seated rotational slumps has been established. However, not all slopes crossed by these fault traces are of equal susceptiblity to slope movement because extensive fault brecciated bedrock is not present at all localities along the entire length of a fault trace. Also, it has been noted that the wider the zone of fault deformation the greater is the potential for slump development. However, fault zone width is highly variable, being up to 300m in width at some localities and less than 1m in width in others. In addition, the susceptibility of slopes within fault zones to movement is greatly dependent upon their steepness. For example, where a fault trace crosses areas of low relief, valley slopes are less susceptible to movement than in localities where a trace crosses steeper valley slopes in areas of high relief.

Slope instability within fault zones also arises through rapid stream dissection and gully development into the non-cemented or weakly cemented fault gouge. Many shallow forms of slope instability are associated with these areas. Fault zones are particularly susceptible to slope movement and are considered sites of high potential for further slope instability.

12.4.2 TIMING OF FUTURE SLOPE MOVEMENTS

In the future, as in the past, many shallow and deep-seated slope movements will occur during the wetter winter months when precipitation of moderate intensity during several consecutive days is common. The majority of slope movements, however, are associated with less frequent periods of intense rainfall of short duration during the passage of cyclonic storms. Cyclones of the intensity of Cyclone Alison in March 1975, during which extensive slope movement occurred, are considered to have a return period of 40 years (cited in Hubbard, 1978). Both shallow translational slides and deep-seated rotational slumps are likely to be triggered during such storms.

Not to be overlooked is the probable high incidence of slope movement that could result from the effects of either fault rupture or seismic shaking greater than MM VI (see Section 10.4.3). Earthquake activity during historical times provides an estimate of the return periods for earthquake intensities that are reliable if they are of the order of the observation period or less. On this basis, Smith (1978a) calculated a return period of ten years for earthquake intensities of MM VI; 50 years for an intensity of MM VII; 100 years for an intensity of MM VIII; and 200 years for an intensity of MM IX for this area. Although the frequency of high intensity earthquakes is not great in this area, it does occur within the main seismic region of New Zealand and corresponds with the area of highest earthquake risk. This area must be expected to be periodically subjected to earthquake shaking of the highest intensities which will probably result from fault movements within the study area along the major active Wellington and Ruahine Faults. Supportive evidence includes the observation that all active faults are known to have moved more than once along the same fault line. Thus future movements can be expected to occur along existing fault lines. In the event of an earthquake of MM VI or greater, slope movements will be extensive and dramatic. Shallow translational debris slides and debris avalanches, and deep-seated rotational slumping will be the dominant forms of slope movement to occur within the Range. However, the most destructive effects of such movements are likely to be felt during wet weather beyond the range front as a result of debris flow activity (see Section 12.4.4).

12.5 PREVENTIVE AND CORRECTIVE MEANS OF SLOPE STABILISATION

There are essentially two governing bodies concerned with erosion control in the vicinity of the southern Ruahine Range.

The New Zealand Forest Service carries out animal control and revegetation schemes (Hathaway, 1977; Cunningham & Stribling, 1978) within the Range in an endeavour to control valley slope erosion and prevent the products of slope movements reaching stream channels. The animal control programme involves the trapping and poisoning of opossums and aerial and ground shooting of deer. Revegetation schemes within the Range comprise: (1) pole planting of species of poplar or willow along undercut stream banks and on extensive terrace flats to prevent gravel remobilisation and streambank undercutting; and (2) aerial seeding and fertilising of valley slopes. To control surface runoff in steep terrain slope treatment can only be effected by vegetative planting. At present only slopes that have failed are being revegetated as the forested slopes are considered to be sufficiently vegetated to control surface runoff. Sites below 1200m which are not severely eroding can generally be effectively treated by aerial sowing with pine or lupin seed (at a cost of about $50/ha) or with selected grasses and clover plus fertiliser (at about $275/ha) (Cunningham & Stribling, 1978). The latter is a favoured method because it can initiate plant succession back to indigenous vegetation. Above about 1200m, and on more severe sites, intensive planting (at a cost of about $500/ha) may be effective in controlling erosion. More severe sites still may require special measures which, because of their very high cost, have generally been avoided in New Zealand. All erosion control measures, particularly those involving herbaceous plants, must be accompanied by intensive animal control (Cunningham & Stribling, 1978). Ground planting pines has largely been abandoned because of cost, although this practice was once widespread; examples are readily apparent throughout the southern Ruahine Range. The main thrust of revegetative work in the next few years is to be concentrated in catchments along the eastern flank of the Range (Appendix IIb) with particular emphasis on the West Tamaki, Rokaiwhana, Mangapuaka and Manga-a-tua catchments. Streams on the western flank of the Range to receive particular attention include Pohangina, Makawakawa and No. 1 Line catchments.

The Manawatu Catchment Board is largely concerned with stabilising stream reaches at the base of the Range. It is considered that sediment transport rates have been accelerated by deforestation of former, natural depositional areas where valleys broaden and become less steep (the valley "throats" of Mosley, 1977) (Blakely, 1978; Mosley & Blakely, 1977). These workers suggest that the most obvious course of action is to "enhance the natural tendency of the streams to store soil and rock eroded from the valley sides in the valley bottom, by judicious use of structural and vegetative techniques". An example of a successfully constricted natural fan designed to curb the flow of streambed detritus is the Kumeti Gravel Reserve (Mangapuaka Stream) (Blakely, 1978). The cost of these works to date and for the next few years is shown in Appendices Ia and Ib.

12.6 EFFECTIVENESS OF THESE STABILISATION TECHNIQUES

As many of the programmes concerned with stabilising valley slopes have only been in operation for a relatively short period of time, it is difficult to assess their chances of success on a long term basis. Nonetheless during the last six years, a time in which the revegetation of valley slopes and culling of noxious animals was increased, regeneration of native tree species has been observed. This regeneration can presumably be attributed to the effective reduction in noxious animal numbers to a controllable level. Also during this short period of time there has been considerable regeneration of both native and artificially seeded grasses and to a lesser extent woody species, upon some erosion scars. The value of this initial vetetation cover lies in its ability to restrict surface runoff processes and to foster the growth of native plant species in an endeavour to re-establish a forest vegetative cover on these steep valley slopes. The effectiveness of the vegetation cover, irrespective of its stage of seral development, in stabilising these valley slopes is however governed to a large extent by other factors.

In a mountainland environment like the southern Ruahine range where geologic, environmental (tectonic and climatic) and physiographic influences contribute significantly to promoting slope instability, the presence of a healthy and complete vegetation cover can only but ameliorate the magnitude of erosion.

The dominant types of slope movement in the southern Ruahine Range include debris slides and debris avalanches, largely of translational origin, that result from failure within colluvial or surficial bedrock materials at shallow depths. The depth of failure of the majority of these slope movements rarely exceeds the maximum penetration depth of the deepest rooted tree species thereby indicating that slope movement is influenced to a greater extent by factors other than the vegetation cover. Slope movement is frequently initiated by environmental influences particularly of climatic origin and less frequently of tectonic origin. The effectiveness and success of the vegetation cover in stabilising shallow slope movements is therefore greatly dependent upon the frequency and magnitude of rainstorms and earthquakes.

In considering revegetation of mass movements it is necessary to establish whether the cost of preventive measures offsets the consequences of the movement and allowance must be made for the uncertainties in our ability to evaluate the possibilities of slope movement. It will often be the case that we cannot prevent slope movement. Mosley & Blakely (1977) in discussing possible erosion control measures with respect to a landslide in Coppermine Stream, state that no manageable techniques could have prevented the landslide.

In this study at least 109 other mass movement features have been identified. Although some appear at present to be stable, further movements at these localities are inevitable, unpredictable in time and beyond control. Others have been major source areas of sediment in the past and are likely to continue to be a major source area of sediment for many years to come. Even if they could be physically controlled, it is not economic to do so. It is therefore better to ascertain the possible detrimental effects of possible future mass movements and to take measures to minimise their effects.

12.7 CONCLUSIONS AND RECOMMENDATIONS

This investivaton, together with other investigations pertaining to the area (Mosley, 1977; Blakely, 1978), clearly show that there are two main areas from which streambed sediment is derived. Instability of upper catchment slopes, in the form of translational debris slides, debris avalanches and to a lesser extent rock slides of shallow depth, together with occasional large-scale, deep-seated rotational slumps and ridge-top features, provide a substantial volume of colluvium and bedrock detritus to stream channels. Much of this detritus remains within the confines of the Range as bedload, gravel waves, alluvial fan and terrace deposits and as colluvium adjacent to stream channels. The remainder of detritus is transported downstream to lower channel reaches. During periods of flood when stream flow is high, stored detritus within upper catchment reaches is often remobilised, some of which may include material derived from older deposits that have long been stabilised and vegetated. This reworking of older stored materials constitutes the second source area of considerable volumes of bedload detritus. Both source areas contribute significantly to the problems of downstream management of transported fluvial detritus, solutions to which include the stabilisation of upper catchment slopes and channel source areas within the Range and channel source areas downstream of the Range.

It is important that areas of shallow slope instability on steep valley slopes within the Range should be revegetated at the earliest opportunity because areas devoid of vegetation undoubtedly increase in areal extent with time. We cannot afford to leave these areas to natural regeneration because it is not known how long such regeneration will take, or what erosion could occur in that time. If corrective measures are not undertaken at the earliest possible time, the cost of revegetative work becomes increasingly more prohibitive.

It is recommended that future revegetative work be carried out by the cheapest means available for which a reasonable level of success seems attainable. The costing of the various methods of revegetative work currently employed in this environment have been outlined in Section 12.6. The most economical method is that of aerial seeding. This work should be concentrated upon areas where shallow translational sliding has occurred but only where sufficient soil or colluvium remains to provide a medium in which seeds can establish. Where translational sliding has exposed the underlying bedrock the chances of revegetating such sites is minimal as much of the seed will be washed off during periods of rainfall.

Where the erosion is too severe to respond to aerial seeding, such as on scars that are currently active source areas of detritus, other means of stabilisation should only be applied if the cost of this work can be justified. Attempts to stabilise these sites by tree planting in the past have too often been to no avail. However, failure in stabilising such sites has not always been due to the severity of the erosion but to the result of inexperience in the use of appropriate techniques. For example, some attempts to stabilise areas of active erosion have been approached by endeavouring to revegetate unstable scree slopes at the base of large scars. Layering of willow poles or planting of pines across such slopes has sometimes met with little success because the continued supply of detritus, as a result of further collapse around the margins of the initial scar, has either uprooted or buried these plants. A more judicious approach in these circumstances may have been to plant the margins of the initial scar in an attempt to curtail further supplies of detritus being added to the scree slope at the base. Once the source of detritus has been slowed or eliminated the remainder of the slope will revegetate naturally. Given sufficient time to stabilise, the rate of revegetation could then be accelerated by seeding.

Pole planting of willow stakes along streambanks within the Range has in places successfully curtailed further lateral bank undercutting. However, their effectiveness in other places where stream bedload mobility is high has been of little value. It is considered impractical to try to stabilise many of the channel source areas of detritus within the Range because of their high susceptibility to further streambank erosion. In addition, many of these deposits are actively removed during periods of peak flow despite the presence of a mature vegetation cover. It is here recommended that future revegetation programmes be approached in a more scientific way. Perhaps certain areas with varying degrees of severity of erosion could be used as experimental plots upon which different revegetation techniques using a variety of plant species could be attempted. In this way, experience in the use of appropriate techniques for specific types of slope instability could be developed. An awareness of the capabilities of the various vegetative techniques would go a long way towards eliminating the often ineffectual choosing of sites (often the most severely eroded) to be stabilised. This in turn would minimise the time, materials and effort at sites where the chances of success are negligible.

In association with continued revegetation programmes, it is imperative that the New Zealand Forest Service maintain sufficient control over the noxious animal population to enable the growth of a strong vegetation cover throughout the Range. This recommendation was emphasised by Cunningham & Stribling (1978).

It is suggested that certain localities where further slope instability is inevitable should be left to run their course. These include localities where extensive outcrops of bedrock form steep-sided valley slopes adjacent to stream channels, areas of rapid gully dissection and areas where active streambank undercutting of the toe slopes of large-scale mass movements is likely to promote shallow and/or deep-seated movements.

Stabilisation of the sediment source area downstream of the Range front is at present achieved by temporarily storing gravel in one reserve in Mangapuaka Stream referred to locally as the 'Kumeti Gravel Reserve'. The purpose of such a reserve is to slow the rate of gravel movement from the upper catchment storage areas in the Range to those reaches that drain through dairy farm land at the base of the Range. This is achieved by spreading flood flows through an area of regenerating bush which slows the velocity of the flow and encourages deposition of coarse and medium sized gravel. Planting of exotics within the reserve, together with construction of diversion groynes and retards, has also be implemented. This means of stabilisation has proven to be successful.

A works programme outlined in Blakely (1978) recommends the establishment of similar gravel reserves for each of the eastward draining catchments along the southern Ruahine Range. The establishment of such reserve areas for the purpose of gravel storage is of key importance in the long term stabilisation programme for this area and should be given high priority. Such reserves are not required for westward draining catchments as these streams upon leaving the confines of the Range are deeply entrenched within Plio-Pleistocene marine deposits, hence the excessive quantities of bedload are of no threat to adjacent farmland.

There is general concensus that as the final perimeter of the Forest Park becomes established it should be fenced (Cunningham & Stribling, 1978), primarily to exclude domestic stock from upper catchment reaches. On the eastern flank of the Range it has been suggested (Blakely, 1978) that this fence line should form an extension of the proposed gravel reserves and be designed so as to include within the reserves all steep and eroding spurs along the foot of the Range. Its construction will greatly assist the process of natural regeneration of the somewhat depleted forest within upper catchment areas.

Finally, much could be achieved by promoting awareness of the problems of erosion through increased public use of the park. Perhaps through schools or societies a planting campaign of designated areas could be undertaken with the object of establishing gravel reserves similar to the one on Mangapuaka Stream. Recommended sites would be those suggested in Blakely (1978).

REFERENCES

Adams, J. (1981a): Earthquake-dammed lakes in New Zealand. *Geology*, 9, 215-219.

Adams, J. (1981b): Contemporary uplift and erosion of the Southern Alps New Zealand: Summary. *Geological Society of America Bulletin*, Part 1, 91, 2-4.

Adkin, G.L. (1930): The Origin of the Manawatu Gorge. *New Zealand Journal of Science and Technology*, 11(6), 353-356.

Ager, D.U. (1965): Mesozoic and Cenozoic Rhynchonellacea. In Moore, R.C. (ed.). Treatise on invertebrate paleontology, Part H (Brachiopoda), Vol. 2. *Geological Society of America and University of Kansas Press*. H597-H632.

Allen, J.R.L. (1966): On bed forms and paleo-currents. *Sedimentology*, 6, 153-190.

Andrews, P.B. (1974): Deltaic sediments, Upper Triassic, Torlesse Supergroup, Broken River. *New Zealand Journal of Geology and Geophysics*, 17(4), 881-905.

Andrews, P.B. (1976): Guide to recording of field observations in sedimentary sequences. New Zealand Geological Survey Report, 79. 68p.

Andrews, P.B., Bishop, D.G., Bradshaw, J.D. & Warren, G. (1974): Geology of the Lord range, central Southern Alps, New Zealand. *New Zealand Journal of Geology and Geophysics*, 17(2), 271-299.

Andrews, P.B., Speden, I.G.& Bradshaw, J.D. (1976): Lithological and paleontological content of the Carboniferous-Jurassic Canterbury Suite, South Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, 19(6), 791-819.

Angelucci, A., De Rosa, E., Fierro, G., Gnaccolini, M., La Monica, G.B., Martinis, B., Parea, G.C., Pescatore, T., Rizzini, A., Wezel, F.C. (1967): Sedimentological characteristics of some Italian turbidites. *Geology Romana*, 6, 345-420.

Arabasz, W.J. & Lowry, M.A. (1980): Microseismicity in the Tararua-Wairarapa area: depth-varying stresses and shallow seismicity in the southern North Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, 23, 141-154.

Aumento, F. (1969): Diorites from the mid-Atlantic ridge at 45°N. *Science*, 165, 1112.

Ballance, P.F. (1964): The sedimentology of the Waitemata Group in the Takapuna section, Auckland. *New Zealand Journal of Geology and Geophysics*, 7(3), 466-499.

Batcheler, C.L. (1967): Preliminary observations of alpine grasshoppers in a habitat modified by deer and chamois. *Proceedings of the New Zealand Ecological Society*, 14, 15-26.

Beck, A.C. (1968): Gravity faulting as a mechanism of topographic adjustment. *New Zealand Journal of Geology and Geophysics*, 11, 191-199.

Birkeland, P.W. (1974): Pedology, weathering and geomorphological research. New York. 285 p.

Blake, M.C. & Landis, C.A. (1973): The Dun Mountain ultramafic belt - Permian oceanic crust and upper mantle in New Zealand. *United States Geological Survey Journal of Research*, 1, 529-534.

Blake, M.C., Jones, D.L. & Landis, C.A. (1974): Active continental margins: contrasts between California and New Zealand. In Burk, C.A. & Drake, C.L. (eds). The geology of continental margins: New York, Springer-Verlag, 853-872.

Blakely, R.J. (1978): South Eastern Ruahine Management Scheme 1980-2000, including works programme for 1980-1985. Unpublished Manawatu Catchment Board and Regional Water Board Report, 57.

Blatt, H., Middleton, G. & Murray, R. (1972): Origin of Sedimentary Rocks. Englewood Cliffs, N.J. Prentice-Hall, 634 p.

Blick, G.H. (1976): Woodville 1975 Survey report (Wellington Fault). New Zealand Geological Survey, Earth Deformation Studies (EDS) Report, 37.

Blick, G.H. (1977): Woodville site report (Wellington Fault). New Zealand Geological Society, Earth Deformation Studies (EDS) Report, 38.

Bishop, D.G. (1972): Progressive metamorphism from prehnite-pumpellyite to greenschist facies in the Dansey Pass area, Otago, New Zealand. *Geological Society of America Bulletin*, 83, 3177-3197.

Boellstorff, J.D. & Te Punga, M.T. (1977): Fission-track ages and correlation of Middle and Lower Pleistocene sequences from Nebraska and New Zealand. *New Zealand Journal of Geology and Geophysics*, 20(1), 47-58.

Bouma, A.H. (1962). Sedimentology of some flysch deposits: A graphic approach to facies interpretation. Elsevier, Amsterdam, 168 p.

Bradshaw, J.D. (1972): Stratigraphy and structure of the Torlesse Supergroup (Triassic-Jurassic) in the foothills of the Southern Alps near Hawarden (560-61), Canterbury. *New Zealand Journal of Geology and Geophysics*, 15(1), 71-87.

Bradshaw, J.D. (1973): Allochthonous Mesozoic fossil localities in melange within the Torlesse rocks of north Canterbury. *Journal of the Royal Society of New Zealand*, 3(2), 161-167.

Bradshaw, J.D. & Andrews, P.B. (1973): Geotectonics and the New Zealand Geosyncline. *Nature, Physical Sciences*, 241(105), 14-16.

Bradshaw, J.D., Adams, C.J. & Andrews, P.B. (1981): Carboniferous to Cretaceous on the Pacific margin of Gondwana. In Vella, P. & Cresswell,M. (eds). *Gondwana V*, Amsterdam, Balkema, 217-221.

Brougham, G.G. (1977): South-eastern Ruahines. Progress Report on Investigation and Channel Improvements. Unpublished Manawatu Catchment Board Report, 41p plus appendices.

Brune, J. (1968): Seismic moment, seismicity and rate of slip along major fault zones. *Journal of Geophysical Research*, 73, 777-784.

Campbell, H.J. (1982): *Halobia* (Bivalvia, Triassic) and a gastropod from Torlesse Supergroup rocks of Wellington, New Zealand. *New Zealand Journal of Geology and Geophysics*, 25, 487-492.

Campbell, J.D. (1955): The Oretian Stage of the New Zealand Triassic System. *Transactions of the Royal Society of New Zealand*, 82(5), 1033-1047.

Campbell, J.D. (1974): *Heterastridium* (Hydrozoa) from Norian sequences in New Caledonia and New Zealand. *Journal of the Royal Society of New Zealand*, 4, 447-453.

Campbell, J.D. & McKellar, I.C. (1960): The Otamitan Stage (Triassic): Definition and Type Locality. *New Zealand Journal of Geology and Geophysics*, 3, 643-659.

Campbell, J.D. & Warren, G. (1965): Fossil localities of the Torlesse Group in the South Island. *Transactions of the Royal Society of New Zealand, Geology*, 3(8), 99-137.

Campbell, J.D. & Coombs, D.S. (1966): Murihiku Supergroup (Triassic-Jurassic) and Southland and south Otago. *New Zealand Journal of Geology and Geophysics*, 9(4), 393-398.

Cann, J.R. (1969): Spilites from the Carlsberg Ridge, Indian Ocean. *Journal of Petrology*, 10, 1-19.

Cann, J.R. (1970): Rb, Sr, Y, Zr and Nb in some ocean-floor basaltic rocks. *Earth and Planetary Science Letters*, 19, 7-11.

Carson, B., Yuan, J., Myers, P.B. Jr. & Barnard, W.D. (1974): Initial deep sea sediment deformation at the base of the Washington continental slope: A response to subduction. *Geology*, 2 (11), 561-564.

Carter, R.M. (1963): Komako, Pohangina Valley. BSc(Hons) thesis, Otago University, Dunedin.

Carter, R.M. (1972): Wanganui strata of Komako district, Pohangina Valley, Ruahine Range, Manawatu. *Journal of the Royal Society of New Zealand*, 2(3), 293-324.

Carter, R.M. (1975): A discussion and classification of subaqueous mass-transport with particular application to grain-flow, slurry-flow and fluxoturbidites. *Earth Science Reviews*, 11, 145-177.

Carter, R.M., Landis, C.A. & Norris, R.J. (1974): Suggestions towards a high-level nomenclature for New Zealand rocks. *Journal of the Royal Society of New Zealand*, 4(1), 5-18.

Carter, R.M., Hicks, M.D., Norris, R.J. & Turnbull, I.M. (1978): Sedimentation patterns in an ancient arc-trench-ocean basin complex: Carboniferous to Jurassic Rangitata Orogen, New Zealand. In Stanley, D.J. & Kelling, G. (eds), Sedimentation in submarine canyons, fans and trenches: Stroudsburg, Pennsylvania, Dowden, Hutchinson and Ross, 340-361.

Clark, R.H., Dibble, R.R., Fyfe, H.E., Lensen, G.J., & Suggate, R.P. (1965): Tectonic and earthquake risk zoning. *Transactions of the Royal Society of New Zealand*, 1(10) General, 113-125.

Clark, S.P. & Jager, E. (1969). Denudational rate in the Alps from geochronologic and heat flow data. *American Journal of Science*, 267, 1143-1160.

Coish, R.A. (1977): Ocean floor metamorphism in the Belts Cove ophiolite, Newfoundland. *Contributions to Mineralogy and Petrology*, 60, 255-270.

Cole, J.W. & Lewis, K.B. (1981): Evolution of the Taupo-Hikurangi subduction system. *Tectonophysics*, 72, 1-21.

Colenso, W. (1884): In Memorium. An account of visits to and crossings over the Ruahine mountain range, Hawkes Bay, New Zealand; and of the natural history of that region; performed in 1845-1847. Hawkes Bay Philosophical Institute, 72p.

Collinson, D.J. & Thompson, D.B. (1982): Sedimentary Structures. George Allen and Unwin Publishers, London.

Cook, H.E., Field, M.E. & Gardner, J.V. (1982): Characteristics of Sediments on Modern and Ancient Continental Slopes. In Sandstone Depositional Environments, Scholle, P.A. & Spearing, D. (eds). *American Association of Petroleum Geologists Memoir* 31, 1, 329-364.

Cook, H.E. & Taylor, M.E. (1977): Comparison of continental slope and shelf environments in the Upper Cambrian and Lowest Ordovician of Nevada. *Society of Economic Paleontologists and Mineralogists Special Publication*, 25, 51-82.

Conley, G. (1980): The shock of '31. The Hawke's Bay Earthquake. A.H. and A.W. Reed Ltd, Wellington, New Zealand.

Coombs, D.S., Ellis, A.J., Fyfe, W.S. & Taylor, A.M. (1959): The zeolite facies, with comments on the interpretation of hydrothermal syntheses. *Geochimica et Cosmochimica Acta*, 17, 53-107.

Coombs, D.S., Landis, C.A., Norris, R.J., Sinton, J.M., Borns, D.J. & Craw, D. (1976): The Dun Mountain ophiolite belt, New Zealand: Its tectonic setting, constitution, and origin, with special reference to the southern portion. *American Journal of Science*, 276, 561-603.

Cooper, R.A. (1975): New Zealand and southeast Australia in the early Paleozoic. *New Zealand Journal of Geology and Geophysics*, 18, 1-20.

Cotton, C.A. (1922): Geomorphology of New Zealand: Part I - Systematic. Dominion Museum, Wellington, 245 p.

Cotton, C.A. (1942): Geomorphology. Whitcombe and Tombs, Christchurch.

Cotton, C.A. (1956): Geomechanics of New Zealand mountain-building. *New Zealand Journal of Science and Technology*, 38B, 187-200.

Coulter, J.D. (1975): The Climate. In Biogeography and Ecology in New Zealand, Kuschel, G. (ed). The Hague.

Cowan, D.S. (1978): Origin of blueschist-bearing chaotic rocks in the Franciscan complex, San Simeon, California. *Geological Society of America Bulletin*, 89, 1415-1423.

Cowie, J.D. (1964a): Aokautere Ash in the Manawatu District, New Zealand. *New Zealand Journal of Geology and Geophysics*, 7, 67-77.

Cowie, J.D. (1964b): Loess in the Manawatu District, New Zealand. *New Zealand Journal of Geology and Geophysics*, 7, 389-396.

Cowie, J.D. & Wellman, H.W. (1962): Age of Ohakea Terrace, Rangitikei River. *New Zealand Journal of Geology and Geophysics*, 5(4), 617-619.

Cowie, J.D. & Milne, J.D.G. (1973): Maps and sections showing the distribution and stratigraphy of North Island loess and associated deposits, New Zealand. New Zealand Soil Survey Report, 6.

Cowie, J.D. & Osborn, W.L. (1977): Soil resources of the Manawatu and the expansion of Palmerston North City. Published by Advisory Services Division, Ministry of Agriculture and Fisheries, Palmerston North.

Craddock, C. (1975): Tectonic evolution of the Pacific margin of Gondwanaland. In Campbell, K.S.W. (ed), Gondwanaland geology: Canberra. Australian National University Press, 609-618.

Crippen, T.F. (1977): Geology of part of the Kaweka Range, Hawke's Bay. Unpublished MSc (Hons) thesis, University of Auckland, Auckland.

Crowell, J.C. (1957): Origin of pebbly mudstones. *Geological Society of America Bulletin*, 68, 993-1010.

Cuff, J.R.I. (1974): Erosion in the Upper Opihi Catchment. South Canterbury Catchment Board and Regional Water Board Report, 66p.

Cunningham, A. (1966): Catchment condition in the Ruahine Range, 1962. New Zealand Forest Service, Forest Research Institute. Protection Forestry Report, 23 (unpublished).

Cunningham, A. (1977): Introduction. In Field Excursion to Southern Ruahines, Neall, V.E. (ed), Handbook for 25th Jubilee Conference of New Zealand Society of Soil Science, 6-8.

Cunningham, A. & Arnott, W.B. (1964): Observations following a heavy rainfall on the Rimutaka Range. *Journal of Hydrology (N.Z.)*, 3(2), 15-24.

Cunningham, A. & Stribling, P.W. (1977): A situation review and proposals for integrated management of the Ruahine Range and the rivers affected by it. Unpublished report prepared for the Ruahine Range Control Scheme Committee, 151 p.

Cunningham, A. & Stribling, P.W. (1978): The Ruahine Range - A situation review and proposals for integrated management of the Ruahine Range and the rivers affected by it. Water and Soil Technical Publication, 13, 60p.

Curry, J.R. & Moore, D.G. (1971): Growth of the Bengal deep sea fan and denudation in the Himalayas. *Geological Society of America Bulletin*, 82, 563-572.

Davis, E.F. (1918): The radiolarian cherts of the Franciscan Group. University of California. *Department of Geological Sciences Bulletin,* 11(3), 235-432.

De Raaf, J.F.M. (1964): The occurrence of flute casts and pseudomorphs after salt crystals in the Oligocene "Gres à ripple marks" of the Southern Pyrenees. In Bouma, A.H. & Brouwer, A. (eds): "Turbidites. Developments in Sedimentology, No. 3". Elsevier, Amsterdam, 192-198.

Dickinson, C. (1953): Geological aspects of abnormal reservoir pressures in Gulf Coast Louisiana. *Bulletin of the American Association of Petroleum Geologists,* 37, 410-432.

Dott, R.H. & Howard, J.K. (1962): Convolute lamination in non-graded sequences. *Journal of Geology,* 70, 114-121.

Dunbar, C.O. & Rogers, J. (1957): Principles of Stratigraphy. Wiley, New York, 356 p.

Dzulynski, S. & Walton, E.K. (1965): Sedimentary Features of Flysch and Greywackes. In Developments in Sedimentology 7. Elsevier Publishing Company, 274 p.

Eiby, G.A. (1966): The Modified Mercalli scale of earthquake intensity and its use in New Zealand. *New Zealand Journal of Geology and Geophysics,* 9, 122-129.

Eiby, G.A. (1968): An annotated list of New Zealand earthquakes 1460-1965. *New Zealand Journal of Geology and Geophysics,* 11, 630-634.

Eiby, G.A. (1971): Seismic Regions of New Zealand. Recent Crustal Movements. *Royal Society of New Zealand Bulletin,* 9, 153-160.

Eiby, G.A. (1975): Seismology in New Zealand. *Geophysical Survey,* 2, 55-72.

Elder, N.L. (1958): Southern Ruahine Range: Ecological report. Unpublished New Zealand Forest Service report.

Elder, N.L. (1965): Vegetation of the Ruahine Range: An introduction. *Transactions of the Royal Society of New Zealand (Botany),* 3(3), 13-66.

Epstein, A.G., Epstein, J.B. & Harris, L.D. (1977): Conodont colour alteration - An index to organic metamorphism. *Geological Survey Professional Paper* 995. United States Department of the Interior, 1-27.

Esler, A.E. (1963): The influence of the Pre-European fires in the Tiritea catchment, northern Tararuas. *Proceedings of the New Zealand Ecological Society,* 10, 8-12.

Eyles, R.J. (1971): Mass Movement in Tangoio Conservation Reserve, Northern Hawkes Bay. *Earth Science Journal,* 5, 79-91.

Fagan, J.J. (1962): Carboniferous cherts, turbidites and volcanic rocks in northern Independence Range, Nevada. *Geological Society of America Bulletin,* 73, 595-612.

Fair, E.E. (1968): Some structural features of the Manawatu River valley. New Zealand Geographical Society Record, 46, 18-19.

Feary, D.A. (1974): Geology of the Mesozoic "basement" in the Waioeka Gorge, Raukumara Peninsula. Unpublished MSc thesis, Auckland University, Auckland.

Feary, D.A. (1979): Geology of the Urewera Greywacke in Waioeka Gorge, Raukumara Peninsula, New Zealand. *New Zealand Journal of Geology and Geophysics,* 22(6), 693-708.

Feary, D.A. & Pessagno, E.A. Jr. (1980): An Early Jurassic age for chert within the Early Cretaceous Oponae Melange (Torlesse Supergroup), Raukumara Peninsula, New Zealand. *New Zealand Journal of Geology and Geophysics,* 23, 623-628.

Feldmeyer, A.E., Jones, B.C., Firth, C.W. & Knight, J.R. (1943): Geology of the Palmerston-Wanganui Basin, "West Side", North Island, New Zealand. Superior Oil Co. of N.Z. Ltd, Oil Report, 171, New Zealand Geological Survey Library.

Finlow-Bates, T. (1970): Petrology and structure of the metagreywacke facies rocks east of Otorohanga. Unpublished MSc thesis, Auckland University, Auckland.

Firth, G.H. & Feldmeyer, A.E. (1943): The geology of the Pahiatua-Dannevirke Basin "East Side", North Island. Unpublished report held at the New Zealand Geological Survey, Lower Hutt.

Fisher, T.V. (1971): Features of coarse-grained, high concentration fluids and their deposits. *Journal of Sedimentary Petrology,* 41, 916-927.

Fleming, C.A. (1962): New Zealand biogeography: A paleontologist's approach. *Tuatara,* 10, 53-108.

Fleming, C.A. (1970): The Mesozoic of New Zealand: Chapters in the history of the Circum-Pacific mobile belt. *Quarterly Journal of the Geological Society of London,* 125, 125-170.

Fleming, C.A. (1974): The geological history of New Zealand and its life. Auckland, New Zealand, Auckland University Press, 141 p.

Fleming, C.A. (1975): The geological history of New Zealand and its biota. In Kuschel, G. (ed), Biogeography and ecology of New Zealand. The Hague.

Folk, R.L., Andrews, P.B. & Lewis, D.W. (1970): Detrital sedimentary rock classification and nomenclature for use in New Zealand. *New Zealand Journal of Geology and Geophysics,* 13(4), 937-968.

Force, E.R. (1974): A comparison of some Triassic rocks in the Hokonui and Alpine belts of South Island, New Zealand. *Journal of Geology,* 82, 34-49.

Friedman, G.M. & Sanders, J.E. (1978): Principles of Sedimentology. John Wiley and Sons, 792 p.

Fyfe, H.E. (1929): Movement on White Creek Fault, New Zealand, during the Murchison earthquake of 17th June, 1929. *New Zealand Journal of Science and Technology,* 11, 192-197.

Gary, M., McAfee, R. Jr. & Wolf, C.L. (1977): Glossary of Geology. American Geological Institute, Washington DC, 805p.

Gay, N.C. & Jaeger, J.C. (1957): Cataclastic deformation of geological materials in matrices of differing composition: II. Boudinage. *Tectonophysics,* 27, 323-331.

George, D.I. (1977): "Apiti! Where's that?" Dudley Rabone & Co. Ltd, Palmerston North.

Gillot, P.Y. & Nativel, P. (1982): K-Ar chronology of the ultimate activity of Pinot Des Neiges Volcano, Reunion Island, Indian Ocean. *Journal of Volcanology and Geothermal Research,* 13, 131-146.

Graff-Petersen, P. (1967): Intraformational deformations and porewater hydrodynamics. *Abstract of the International Congress on Sedimentology,* England.

Grammer, T.R. (1971): The geology of the Eastern Saddle Road area, Manawatu, North Island, New Zealand. Unpublished BSc(Hons) thesis, University of Otago, Dunedin.

Grant, P.J. (1965): Major regime changes of the Tukituki River, Hawkes Bay, since about 1650 A.D. *New Zealand Journal of Hydrology,* 4(1), 17-30.

Grant, P.J. (1966): Variations of rainfall frequency in relation to erosion in Eastern Hawkes Bay. *New Zealand Journal of Hydrology,* 5(2), 73-86.

Grant, P.J. (1975): Introduction to a research programme for the Ruahine Range, North Island. *Proceedings of the Eastcofert Seminar*, 28-36.

Grant, P.J. (1977): Recorded channel changes of the Upper Waipawa River, Ruahine Ranges, New Zealand Water and Soil Division, Ministry of Works and Development. Water and Soil Technical Publications, 6, 22p.

Grant, P.J. (1978): Major erosion periods in the Ruahine Ranges since the 13th Century and their impact on the vegetation. Unpublished paper presented at the New Zealand Ecological Society (Inc.) Conference 1978, Massey University, New Zealand.

Grant, P.J. (1981a): Major periods of erosion and sedimentation in the Ruahine Range since the 13th century. In Erosion in the Ruahines, Neall, V.E. (ed), International Conference on Soils with Variable Charge.

Grant, P.J. (1981b): Major periods of erosion and sedimentation in the North Island, New Zealand since the 13th century. In Erosion and Sediment Transport in Pacific Rim Steeplands. *IAHS Publication*, 132, (Christchurch), 288-304.

Grant, P.J. (1983): Recently increased erosion and sediment transport rates in the upper Waipawa River Basin, Ruahine Range, New Zealand. Published for the National Water and Soil Conservation Organisation by the Soil Conservation Centre, Aokautere, Publication 5, 127 p.

Grant-Mackie, J.A. (1975): The stratigraphy and taxonomy of the Upper Triassic bivalve *Monotis* in New Zealand. Unpublished PhD thesis, University of Auckland, Auckland.

Grant-Taylor, T.L. (1964): Stable angles in Wellington greywacke. *New Zealand Engineering*, 19, 129-130.

Grindley, G.W. (1960): Sheet 8, Taupo. Geological Map of New Zealand, 1:250 000. Department of Science and Industrial Research, Wellington.

Grindley, G.W. (1961): Mesozoic orogenies in New Zealand. *Proceedings of the 9th Pacific Science Congress*, 12, 71-75.

Grindley, G.W., Harrington, H.J. & Wood, B.L. (1959): The Geological Map of New Zealand. *New Zealand Geological Survey Bulletin*, 66.

Grow, J.A. (1973): Crustal and upper mantle structure of the central Aleutian Arc. *Geological Society of America Bulletin*, 84, 2168-2229.

Haast, J. (1865): Report on geological exploration of the West Coast. *Journal of Proceedings of the Provincial Council, Province of Canterbury, New Zealand Session* XXIII, 13-21.

Hamblin, W.K. (1976): Patterns of displacement along the Wasatch Fault. *Geology*, 4(10), 619-622.

Hampton, M.A. (1970): Subaqueous debris flow and generation of turbidity currents. Unpublished PhD thesis, Stanford University, Stanford, California.

Hampton, M.A. (1972): The role of subaqueous debris flow in generating turbidity currents. *Journal of Sedimentary Petrology*, 42, 775-793.

Hathaway, R.L. (1977): Revegetation Research. In Field Excursion to Southern Ruahines, Neall, V.E. (ed), Handbook for 25th Jubilee Conference of New Zealand Society of Soil Science.

Hatherton, T. (1969): Geophysical anomalies over the eu-and miogeosynclinal systems of California and New Zealand. *Geological Society of America Bulletin*, 80, 213-230.

Hayward, J.A. & Blakely, R.J. (1976): Sediment yields and transport rates in a mountain stream. *Proceedings of New Zealand Hydrological Society, Rotorua Conference*, 1976,

Hector, J. (1890): Maharahara Copper-mine. *Report of Geological Explorations*, 20, xxiv-xxvii.

Hector, J. (1892): Maharahara Copper. *Report of Geological Explorations*, 20, xxvi-xxviii.

Hector, J. (1894): Maharahara Copper-lode. *Report of Geological Explorations*, 22, xxxiv-xxxv.

Hegan, B.D. (1980): Engineering Geology of the Moawhango to Tongariro Tunnel. Unpublished Geological Survey Report, EG343.

Heine, R.W. (1962): An interpretation of the tectonic features of the Tararua and Rimutaka Ranges. *Transactions of the Royal Society of New Zealand*, 1(13), 201-205.

Henderson, J. (1933): The geological aspects of the Hawkes Bay earthquakes. *New Zealand Journal of Science and Technology*, 15, 38-75.

Hill, H. (1893): On Geology of the Country between Dannevirke and Wainui, Hawkes Bay. *Transactions of the New Zealand Institute*, 26, 392.

Hill, H. (1911): A letter to Adkin, G.L., dated 4th August, 1911. Cited in Adkin, G.L. (1930).

Holland, M.K. (1975): Southeastern Ruahines: the erosion threat on the Range and proposal for treatment. *Proceedings of the Eastcofert Seminar*, 39-44.

Hollister, C.D. & Heezen, B.C. (1966): Ocean Bottom Currents. In Fairbridge, R.W. (ed), The Encyclopedia of Oceanography. Reinhold, New York, 567-583.

Hopgood, A.M. (1961): The geology of the Cape Rodney-Kawau district, Auckland. *New Zealand Journal of Geophysics*, 4, 205-230.

Howell, D.G. (1980): Mesozoic accretion of exotic terranes along the New Zealand segment of Gondwanaland. *Geology*, 8, 487-491.

Howell, D.G. & Normack, W.R. (1982): Sedimentology of Submarine Fans. In Sandstone Depositional Environments, Scholle, P.A. & Spearing, D. (eds). *American Association of Petroleum Geologists Memoir*, 31, 365-404.

Hubbard, C.B. (1978): The distribution and properties of soils in relation to erosion in a selected catchment of the southern Ruahine Range, North Island, New Zealand. Unpublished MSc thesis, Massey University, Palmerston North.

Hubbard, C.B. & Neall, V.E. (1980): A Reconstruction of late quaternary erosional events in the West Tamaki River catchment, southern Ruahine Range, North Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, 23, 587-593.

Hubbard, W.W. (1976): Historical record of the Kumeti Stream. Unpublished Manawatu Catchment Board Report.

Hsu, K.J. (1968): Principles of mélanges and their bearing on the Franciscan-Knoxville paradox. *Geological Society of America Bulletin*, 79, 1063-1074.

Hsu, K.J. & Ohrbom, R. (1969): Mélanges of San Francisco Peninsula - a geological reinterpretation of type Franciscan. *American Association of Petroleum Geologists Bulletin*, 53, 1348-1367.

Ingle, J.C. Jr. et al. (1973): Western Pacific floor, Leg 31, Deep Sea Drilling Project. *Geotimes*, 18(10), 22-25.

Isaac, M.J. (1972): The geology of the Oponae and Waiata Valleys, Raukumara Peninsula. Unpublished BSc(Hons) thesis, Auckland University, Auckland.

James, I.L. (1973): Mass movements in the Upper Pohangina Catchment, Ruahine Range. *New Zealand Journal of Hydrology*, 12(2), 92-102.

James, I.L. & Beaumont, P.E. (1971): Report on a survey of the vegetation of the southern Ruahine Range. New Zealand Forest Service, Forest Research Institute Report, 80.

Jenkins, T.B.H. & Jenkins, D.G. (1971): Conodonts of the Haast Schist and Torlesse Groups of New Zealand. Part 1 - Biostratigraphic significance of the Triassic conodonts from the Mount Mason and Okuku limestones. *New Zealand Journal of Geology and Geophysics*, 14(4), 782-794.

Jones, D.L., Silberling, N.J., Berg, H.C. & Plafker, G. (1981): Tectonostratigraphic terrane map of Alaska. *United States Geological Survey Open-File Report*, 81-792.

Kaewyana, W. (1980): Late Quaternary alluvial terraces and their cover bed stratigraphy, Eketahuna and Pahiatua Districts, New Zealand. MSc thesis, Victoria University, Wellington.

Kear, D. (1971): Basement rock facies - northern North Island. *New Zealand Journal of Geology and Geophysics*, 14, 275-283.

Keller, G.P. (1954): The Rimutaka Deviation. *New Zealand Engineering*, 9, 399-420.

Kennett, J.P. (1982): Marine Geology. Prentice-Hall Inc., New Jersey, 752 p.

Kingma, J.T. (1957a): The North Island geanticline in the Hawkes Bay sector. *New Zealand Journal of Science and Technology*, B38(6), 496-499.

Kingma, J.T. (1957b): Tectonic setting of the Ruahine-Rimutaka Range. *New Zealand Journal of Science and Technology*, B38(8), 858-861,

Kingma, J.T. (1958a): Geology of the Wakarara Range, Central Hawkes Bay. *New Zealand Journal of Geology and Geophysics*, 1, 76-91.

Kingma, J.T. (1958b): The Tongaporutuan sedimentation in Central Hawkes Bay. *New Zealand Journal of Geology and Geophysics*, 1(1), 1-30.

Kingma, J.T. (1959): The tectonic history of New Zealand. *New Zealand Journal of Geology and Geophysics*, 2(1), 1-55.

Kingma, J.T. (1962): Sheet 11 Dannevirke (1st ed.) Geological Map of New Zealand 1:250 000. New Zealand Department of Scientific and Industrial Research, Wellington.

Klein, G. de V., (1967): Paleocurrent analysis in relation to modern sediment dispersal patterns. *American Association of Petroleum Geologists Bulletin*, 51, 366-382.

Korsch, R.J. & Wellman, H.W. (in press): Pocket geology of the New Zealand Region. In Nairn, A.E.M. *et al.* (eds). The Ocean Basins and Their Margins, Volume 7, New York, Plenum.

Krumbein, W.C. & Sloss, L.L. (1963): Stratigraphy and Sedimentation. Freeman, San Fransisco, 660 p.

Kuenen, P.H. (1953): Significant features of graded bedding. *American Association of Petroleum Geologists Bulletin*, 37, 1044-1066.

Kuenen, P.H. (1957): Sole markings of graded greywacke beds. *Journal of Geology*, 65, 231-258.

Kuenen, P.H. (1964): Deep-sea sands and ancient turbidites. In Bouma, A.H. & Brouwer, A. (eds). Turbidites. Developments in Sedimentology No. 3. Elsevier, Amsterdam, 3-33.

Kuenen, P.H. (1967): Emplacement of flysch-type sandbeds. *Sedimentology*, 9, 203-243.

Kuenen, P.H. & Migliorini, C.I. (1950): Turbidity currents as cause of graded bedding. *Journal of Geology*, 58, 91-127.

Landis, C.A. & Coombs, D.S. (1967): Metamorphic belts and orogenesis in southern New Zealand. *Tectonophysics*, 4, 501-518.

Landis, C.A. & Bishop, D.G. (1972): Plate tectonics and regional stratigraphic-metamorphic relations in the southern part of the New Zealand Geosyncline. *Geological Society of America Bulletin*, 83(8), 2267-2284.

Landslides Analysis and Control, Special Report 176 by the Transport Research Board. Schuster, R.L. & Krizek, R.J. (eds) 1978. 234 p.

Le Roex, A.P. & Erlank, A.J. (1982): Quantitative evaluation of fractional crystallization in Bouvet Island lavas. *Journal of Volcanology and Geothermal Research*, 13, 309-338.

Lensen, G.J. (1958a): The Wellington Fault from Cook Strait to Manawatu Gorge. *New Zealand Journal of Geology and Geophysics*, 1, 178-197.

Lensen, G.J. (1958b): A method of horst and graben formation. *Journal of Geology*, 66, 579-587,

Lensen, G.J. (1962): Notes on the regional consistency of PHS directions. *New Zealand Journal of Geology and Geophysics*, 5, 175-177.

Lensen, G.J. (1964): The faulted terrace sequence of Grey River, Awatere Valley, South Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, 7, 871-876.

Lensen, G.J. (1968a): Analysis of progressive fault displacement during downcutting at the Branch River, South Island, New Zealand. *Geological Society of America Bulletin*, 79, 545-556.

Lensen, G.J. (1968b): Sheet N158 Masterton (1st ed.) Late Quaternary tectonic map of New Zealand, 1:63 360. Department of Scientific and Industrial Research, Wellington, New Zealand.

Lensen, G.J. (1969): Sheet 153 Eketahuna (1st ed.) Late Quaternary tectonic map of New Zealand, 1:63 360. Department of Scientific and Industrial Research, Wellington, New Zealand.

Lensen, G.J. (1976): Sheets N28D, O28C and N29B, Hillersden (1st ed.). Late Quaternary tectonic map of New Zealand, 1:50 000. Department of Scientific and Industrial Research, Wellington, New Zealand.

Lensen, G.J. & Otway, P.M. (1971): Earthshift and post-earthshift deformation associated with the May 1968 Inangahua earthquake, New Zealand. *Royal Society of New Zealand Bulletin*, 9, 106-117.

Leslie, W.C. & Hollingsworth, R.J.S. (1972): Exploration in the East Coast Basin, New Zealand. *The A.P.E.A. Journal*, 12(1), 39-44.

Lillie, A.R. (1950): Two New Zealand rivers following Tertiary transverse furrows. *Transactions of the Royal Society of New Zealand*, 78, 329-341.

Lillie, A.R. (1953): Geology of the Dannevirke Subdivision. *New Zealand Geological Survey Bulletin* 46, 156p.

Logan, I.R. (1971): A Review of Management Problems in a High Country Protection Forest, Catchment Control Seminar, New Zealand Association of Soil Conservators, Massey University, 1971.

Logan, P.C. (1955): Inspection report - District 2, zone D. Unpublished Internal Affairs Department Report.

Logan, P.C. & Harris, L.H. (1967): Introduction and establishment of red deer in New Zealand. New Zealand Forest Service Information Series, 55.

Lowe, D.R. (1976): Subaqueous liquefied and fluidized sediment flows and their deposits. *Sedimentology*, 23, 285-308.

MacKinnon, T.C. (1980): Geology of *Monotis*-bearing Torlesse rocks in Temple Basin near Arthur's Pass South Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, 23(1), 63-81.

MacKinnon, T.C. (1983): Origin of the Torlesse terrane and coeval rocks, South Island, New Zealand. *Geological Society of America Bulletin*, 94, 967-985.

Macpherson, E.O. (1946): An outline of Late Cretaceous and Tertiary diastrophism in New Zealand. *New Zealand Geological Survey Memoir*, 6.

Manawatu Catchment Board (1972): Eastern Ruahine Catchment Control Scheme, preliminary report. Unpublished Manawatu Catchment Board report.

Manawatu Catchment Board (1976): Southeastern Ruahine scheme, 1975-80.

Mangin, J.P. (1962): Traces de pattes d'oiseaux et flute - casts associés dans un "facies flysch" due tertiaire Pyreneen. *Sedimentology*, 1, 163-166.

Marden, M. (1981): The relationship between fault zones and mass movements in the southern Ruahine Range, North Island, New Zealand. Geological Society of N.Z. Inc, Hamilton Conference Programme and Abstracts.

Martin, R.S. (1978): S.E. Ruahine investigation report on hydrology. Unpublished report to Manawatu Catchment Board.

Marshall, P. (1905): Geography of New Zealand, 183 p.

Marshall, P. (1912): Geology of New Zealand, 38 p.

Marwick, J. (1953): Divisions and Faunas of the Hokonui System (Triassic and Jurassic). *New Zealand Geological Survey Palaeontological Bulletin*, 21, 1-141.

McKay, A. (1877): Report on the Country between Masterton and Napier. *Report of Geological Explorations*, 10, 67-94.

McKay, A. (1888a): On the Copper Ore at Maharahara, near Woodville. *Report of Geological Explorations*, 19, 6-8.

McKay, A. (1888b): On Mineral Deposits in the Tararua and Ruahine Mountains. *Report of Geological Explorations*, 19, 1-6.

McKay, A. (1888c): On the Tauherenikau and Waiohine Valleys, Tararua Range. *Report of Geological Explorations*, 19, 58-67.

McKay, A. (1894): On the Maharahara Copper-mine, Woodville, Hawkes Bay. *Report of Geological Explorations*, 22, 2-6.

McKay, A. (1901): Copper at Maharahara, near Woodville, Hawkes Bay. *Parliamentary Paper*, C-10, 3.

McKee, E.D. & Weir, G.W. (1953): Terminology for stratification and cross-stratification in sedimentary rocks. *Geological Society of America Bulletin*, 64, 381-389.

McKelvey, P.J. (1960): Condition assessment of protection forests. *New Zealand Journal of Forestry*, 8(2),

Meads, M.J. (1976): Effects of opossum browsing on northern rata trees in the Orongorongo Valley, Wellington, New Zealand. *New Zealand Journal of Zoology*, 3, 127-139.

Middleton, G.V. (1967): Experiments on density and turbidity currents, III, Deposition of sediments. *Canadian Journal of Earth Science*, 4, 475-505.

Middleton, G.V. (1970): Experimental studies related to problems of flysch sedimentation. In Lajoie, J. (ed). Flysch Sedimentology in North America. *Geological Association of Canada, Special Paper*, 7, 253-272.

Middleton, G.V. & Hampton, M.A. (1973): Sediment gravity flows: Mechanisms of flow and deposition. In Turbidites and Deep-water Sedimentation. Society of Economic Paleontologists and Mineralogists, Pacific Section, 1-32.

Middleton, G.V. & Hampton, M.A. (1976): Subaqueous sediment transport and deposition by sediment gravity flows. In Stanley, D.J. & Swift, D.J.P. (eds). Marine Sediment transport and environmental management. John Wiley and Sons Inc., 197-218.

Milne, J.D.G. (1973a): Upper Quaternary geology of the Rangitikei drainage basin, North Island, New Zealand. Unpublished PhD thesis, Victoria University, Wellington.

Milne, J.D.G. (1973b): Maps and sections of river terraces in Rangitikei Basin, North Island, New Zealand. 4 sheets. New Zealand Soil Survey Report, 4.

Milne; J.D.G. & Campbell, J.D. (1969): Upper Triassic fossils from the Oroua Valley, Ruahine Range, New Zealand. *Transactions of the Royal Society of New Zealand (Geology)*, 6(18), 247-250.

Milne, J.D.G. & Smalley, I.J. (1979): Loess deposits in the southern part of the North Island of New Zealand: an outline stratigraphy. Proceedings of the West European working group of the INQUA Loess Commission, Budapest.

Ministry of Works and Development (1975): Unpublished report on the Shotover River Catchment, two volumes.

Minster, J.B., Jordon, T.H., Molnar, P. & Harris, E. (1974): Numerical modelling of instantaneous plate tectonics. *Geophysical Journal of the Royal Astronomical Society*, 36, 541-576.

Molnar, P. & Atwater, T. (1978): Interarc-spreading and cordilleran tectonics as alternatives related to the age of subducted oceanic lithosphere. *Earth and Planetary Science Letters*, 41, 330-340.

Moore, J.C. (1973): Complex deformation of Cretaceous trench deposits, southwestern Alaska. *Geological Society of America Bulletin*, 84, 2005-2020.

Moore, P.R. (1975): Tertiary Sediments in the Ruahine Range. Unpublished New Zealand Geological Survey Report.

Morgan, P.G. (1914): Petroleum prospects in southern Hawke's Bay and Eastern Wellington. *New Zealand Geological Survey 8th Annual Report*, 135.

Morgan, P.G. (1920): Maharahara Copper Mine. *New Zealand Geological Survey 14th Annual Report*, 9.

Mosley, M.P. (1977): Southeastern Ruahine investigation report on erosion and sedimentation. Manawatu Catchment Board and Regional Water Board Report, Palmerston North. 105 p and appendices, corrections.

Mosley, M.P. & Blakely, R.J. (1977): The Coppermine landslide, southeastern Ruahine Range. *Soil and Water*, 13(5), 16-17 and 32-33.

Muir, I.D. & Tilley, C.E. (1964): Basalts from the northern part of the rift zone of the Mid Atlantic Ridge. *Journal of Petrology*, 5, 409-434.

Munday, P.M. (1977): The geology of the Oroua Valley. Unpublished MSc(Hons) thesis, Auckland University, Auckland.

Nemcok, A. (1972): Gravitational Slope Deformation in High Mountains. Proceedings, 24th International Geological Congress, Montreal, Section 13. *Engineering Geology*, 132-141.

Officers of the New Zealand Geological Survey (1979): Active earth deformation. New Zealand Geological Survey Report, 89.

Ongley, M. (1935): Manawatu Gorge. *New Zealand Journal of Science and Technology*, 16(5), 249-260

Ongley, M. (1943a): Wairarapa earthquake of 24th June 1942, together with map showing surface traces of faults recently active. *New Zealand Journal of Science and Technology*, 25(2), 67-78.

Ongley, M. (1943b): Surface trace of the 1855 earthquake. *Transactions of the Royal Society of New Zealand*, 73(2), 84-89.

Ongley, M. & Williamson, J.H. (1931a): A note on the north end of the Tararua Range. *Parliamentary Paper*, H-34, 55.

Ongley, M. & Williamson, J.H. (1931b): Maharahara district. *Parliamentary Paper*, H-34, 55.

Ower, J.R. (1943): The geology of the Manawatu Saddle and adjacent fronts of the Ruahine Range, North Island, New Zealand. Superior Oil Co. Report, New Zealand Geological Survey Library, Lower Hutt, 1-17.

Pain, C.F. (1968): Geomorphic events of floods in the Orere river catchment, eastern Hunua ranges. *New Zealand Journal of Hydrology*, 7(2), 62-74.

Pain, C.F. & Bowler, J.M. (1973): Denudation following the November 1970 earthquake at Modang, Papua New Guinea. *Zeitschrift fur Geomorphologie Supplements*, 18, 92-104.

Palmer, A.S. (1982): Kawakawa Tephra in Wairarapa, New Zealand, and its uses for correlating Ohakea loess. *New Zealand Journal of Geology and Geoghysics*, 25, 305-315.

Park, J. (1887): On the Geology of the western part of Wellington Provincial District, and part of Taranaki. *Report of Geological Explorations*, 18, 24-73.

Pearce, J.A. (1975): Basalt geochemistry used to investigate past tectonic environments on Cyprus. *Tectonophysics*, 25, 41-67.

Pearce, J.A. & Cann, J.R. (1971): Ophiolite origin investigated by discriminant analysis using Ti, Zr, and Y. *Earth and Planetary Science Letters*, 12, 339.

Pearce, J.A. & Cann, J.R. (1973): Tectonic setting of basic volcanic rocks determined using trace element analyses. *Earth and Planetary Science Letters*, 19, 290-300.

Pearce, J.A. & Norry, M.G. (1979): Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks. *Contributions to Mineralogy and Petrology*, 69, 33-47.

Peterson, G.L. (1965): Implications of two Cretaceous mass transport deposits, Sacramento Valley, California. *Journal of Sedimentary Petrology*, 35, 401-407.

Petrie, D. (1908): Account of a visit to Mount Hector, a high peak of the Tararuas. *Transactions of the New Zealand Institute*, 40, 290.

Pettijohn, F.J. (1975): Sedimentary Rocks. Third Edition. Harper and Row Publishers Inc, New York.

Pierson, T.C. (1980): Erosion and deposition by debris flows at Mt Thomas, North Canterbury, New Zealand. *Earth Surface Processes*, 5, 227-247.

Pierson, T.C. (1981): Dominant particle support mechanisms in debris flows at Mt Thomas, New Zealand, and implicatons for flow mobility. *Sedimentology*, 28, 49-60.

Piyasin, S. (1966): Plio-Pleistocene geology of the Woodville area. Unpublished MSc thesis, Victoria University, Wellington.

Poole, A.L. (1973): Protecting Ruahines more important than saving Manapouri. *Soil and Water*, 9(4), 27.

Pracy, L.T. (1966): Introduction and Liberation of the Opossum (Trichosurus velpecula) into New Zealand. *New Zealand Forest Information Series*, 45.

Pracy, L.T. (1971): Inspection southern Ruahine Range. New Zealand Forest Service file note.

Pringle, I.J. (1980): Petrology, geochemistry and igneous and metamorphic mineralogy of low-grade metamorphosed basalts of the Torlesse terrane, South Island, New Zealand. Unpublished PhD thesis, University of Otago, Dunedin.

Pryor, W.A. & Barr, J.L. (1968): Sole marks in siltstones of non-turbidite origin (Abstract). *American Association of Petroleum Geologists Bulletin*, 52(3), 546.

Ramberg, H. (1955): Natural and experimental boudinage and pinch-and-swell structures. *Journal of Geology*, 63, 512-526.

Ramsay, J.G. (1967): Folding and fracturing of rocks. New York, McGraw-Hill Book Co., 568p.

Rattenbury, M.S. (1982): Geology of Otaki Forks, Tararua Range. Unpublished MSc thesis, Auckland University, Auckland.

Reed, J.J. (1957a): Fault zones in part of the Rimutaka Range. *New Zealand Journal of Science and Technology*, 38B, 686-687.

Reed, J.J. (1957b): Petrology of the lower Mesozoic rocks of the Wellington district. *New Zealand Geological Survey Bulletin*, 57, 60p.

Reineck, H.E. & Wanderlich, F. (1968): Classification and origin of flaser and lenticular bedding. *Sedimentology*, 11, 99-105.

Reineck, H.E. & Singh, I.B. (1975): Depositional Sedimentary Environments. Springer-Verlag, New York, 439p.

Rhea, K.P. (1968): Aokautere Ash, loess and river terraces in the Dannevirke District. *New Zealand Journal of Geology and Geophysics*, 11, 685-692.

Rib, H.T. & Liang, T. (1978): Recognition and Identification Chapter 3. In Schuster, R.L. & Krizek, R.J. (eds). Landslides: Analysis and Control, Special Report, 176, 34-80.

Rich, C.C. (1959): Late Cenozoic geology of the lower Manawatu valley. Unpublished PhD thesis, Victoria University, Wellington.

Riedel, W.R. (1959): Siliceous organic remains in pelagic sediments. In Ireland, H.A. (ed). Silica in sediments - a symposium. *Society of Economic Geologists and Mineralogists Special Publication*, 7, 80-91.

Robins, R.G. (1958): Direct effect of the 1855 earthquake on the vegetation of the Orongorongo Valley, Wellington. *Transactions of the Royal Society of New Zealand*, 85(2), 205-212.

Roser, B.P. (1983): Comparative studies of Cu and Mn mineralisation in the Torlesse, Waipapa and Haast Schist terranes, New Zealand. Unpublished PhD thesis, Victoria University, Wellington.

Rowe, G.H. (1980): Applied geology of Wellington rocks for aggregate and concrete. Unpublished PhD thesis, Victoria University, Wellington.

Saunders, A.D., Tarney, J., Marsh, N.G. & Wood, D.A. (1979): Ophiolites as ocean crust or marginal basin crust: a geochemical approach. Proceedings of the International Ophiolite Symposium, Cyprus, 193-204.

Schumm, S.A. (1963): The disparity between present rates of denudation and orogeny. *United States Geological Survey Professional Papers*, 454H, 13p.

Schumm, S.A. (1975): Episodic Erosion: A modification of the geomorphic cycle. A proceedings volume of the 6th Annual Geomorphology symposium series, Birmingham and New York, 69-85.

Schumm, S.A. (1977): South-eastern Ruahine Investigation - Consultant's Appraisal of Erosion Processes. Manawatu Catchment Board and Regional Water Board Report, Palmerston North, 7p plus attachments.

Scott, K.M. (1966): Sedimentology and dispersal patterns of a Cretaceous flysch sequence, Patagonian Andes, southern Chile. *American Association of Petroleum Geologists Bulletin*, 50, 72-107.

Selby, M.J. (1967): Aspects of the geomorphology of the greywacke ranges bordering the lower and middle Waikato basins. *Earth Science Journal*, 1(1), 37-56.

Selby, M.J. (1976): Slope failure due to extreme rainfall; a case study from New Zealand. *Geografiska Annaler*, 58A, 131-138.

Sharpe, C.F.S. (1938): Landslides and related phonomena, Columbia University Press.

Shervais, J.W. (1982): Ti-V plots and the petrogenesis of modern and ophiolitic lavas. *Earth and Planetary Science Letters*, 59, 101-118.

Silver, E.A. & Beutner, E.C. (1980): Mélanges. Conveners, Penrose Conference Report. *Geology*, 8, 32-34.

Smale, D., Houghton, B.F., McKeller, I.C., Mansergh, G.D., Moore, P.R., Grant-Taylor, T.L. & Te Punga, M.T. (1978): Geology and Erosion in the Ruahine Range. New Zealand Geological Survey Report, G20

Smith, W.D. (1978a): Earthquake risk in New Zealand: statistical estimates. *New Zealand Journal of Geology and Geophysics*, 21(3), 313-327.

Smith, W.D. (1978b): Spatial distribution of felt intensities for New Zealand earthquakes. *New Zealand Journal of Geology and Geophysics*, 21(3), 293-311.

Soons, J.M. (1967): Erosion by needle ice in the Southern Alps, New Zealand. In Arctic and alpine environments, 217-227. Indiana University Press.

Speden, I.G. (1975): An association of the late Triassic bivalves *Halobia*, *Manticula*, and *Monotis* in the Torlesse rocks of north Canterbury. *New Zealand Journal of Geology and Geophysics*, 18(2), 279-284.

Speden, I.G. (1976): Fossil localities in Torlesse rocks of the North Island, New Zealand. *Journal of the Royal Society of New Zealand*, 6(1), 73-91.

Sporli, K.B. (1978): Mesozoic tectonics, North Island, New Zealand. *Geological Society of America Bulletin*, 89, 415-425.

Sporli, K.B. & Barter, T.P. (1973): Geological Reconnaissance in the Torlesse Super-group of the Kaimanawa Ranges along the Lower Reaches of the Waipakihi River, North Island, New Zealand. *Journal of the Royal Society of New Zealand*, 3(3), 363-380.

Sporli, K.B. & Lillie, A.R. (1974): Geology of the Torlesse Supergroup in the northern Ben Ohau Range, Canterbury. *New Zealand Journal of Geology and Geophysics*, 17(1), 115-141.

Sporli, K.B., Stanaway, K.J. & Ramsay, W.R.H. (1974): Geology of the Torlesse Supergroup in the southern Liebig and Burnett Ranges, Canterbury, New Zealand. *New Zealand Journal of Geology and Geophysics*, 4(2), 177-192.

Sporli, K.B. & Bell, A.B. (1976): Torlesse mélange and coherent sequences, eastern Ruahine Range, North Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, 19(4), 427-447.

Stanley, D.J. (1969): Sedimentation in slope and base-of-slope environments. In Stanley, D.J. (ed). The New Concept of Continental Margin Sedimentation, DJS-8, 1-25. Washington American Geological Institute.

Stauffer, P.H. (1967): Grain-flow deposits and their implications, Santa Ynes Mountains, California. *Journal of Sedimentary Petrology*, 37(2), 487-508.

Stephens, P.R. (1975): Determination of procedures to establish priorities for erosion control as determined in the southern Ruahine Ranges, New Zealand. Unpublished MAg Sci thesis, Massey University, Palmerston North.

Stephens, P.R. (1977): Erosion in the upper West Tamaki Catchment. In Field Excursion to Southern Ruahines. Neall, V.E. (ed). Handbook for 25th Jubilee Conference of the New Zealand Society of Soil Science, 27-36.

Stevens, G.R. (1963): Jurassic belemnites in the Torlesse Group of the North Island. *New Zealand Journal of Geology and Geophysics*, 6, 707-710.

Stevens, G.R. & Speden, I.G. (1978): In The Mesozoic A. The Phanerozoic Geology of the World, II. Moullade, M. & Nairn, A.E.M. (eds). Chapter 8, New Zealand. Elsevier Scientific Publishing Co., Amsterdam, 251-328.

Strand, S. (1981): The vegetation and management of the West Tamaki catchment. In Neall, V.E. (ed). Erosion in the Ruahines. International conference on soils with variable charge, Massey University, Palmerston North, 24-33.

Suggate, R.P. (1961): Rock-stratigraphic names for South Island schists and undifferentiated sediments of the New Zealand Geosyncline. *New Zealand Journal of Geology and Geophysics*, 4(4), 392-399.

Suggate, R.P. (1978): Torlesse Supergroup. In Suggate, R.P., Stevens, G.R. & Te Punga, M.T. (eds). The Geology of New Zealand. Wellington, Government Printer, Vols 1 and 2.

Suggate, R.P. & Lensen, G.J. (1973): Rate of horizontal fault displacement in New Zealand. *Nature*, 242, 518.

Suggate, R.P., Stevens, G.R. & Te Punga, M.T. (1980): The Geology of New Zealand. Government Printing Office, Wellington, New Zealand. Vol. 1, 343p; Vol. 2, 477p.

Tabor, R.W. (1971): Origin of ridge-top depressions by large-scale creep in the Olympic Mountains, Washington. *Geological Society of America Bulletin*, 82, 1811-1822.

Taliafero, N.L. (1943): Franciscan-Knoxville problem. *American Association of Petroleum Geologists Bulletin*, 27, 109-219.

Tanaka, M. (1976): Rate of erosion in the Tanzawa Mountains, central Japan. *Geografiska Annaler*, 58A, 155-163.

Ten Haaf, E. (1956): Significance of convolute lamination. *Geologie en Mijnbouw*, 18(6), 188-194.

Te Punga, M.T. (1952): The geology of Rangitikei Valley. *New Zealand Geological Survey Memoir*, 8.

Te Punga, M.T. (1957): Live Anticlines in Western Wellington. *New Zealand Journal of Science and Technology*, B38(5), 328-341.

Te Punga, M.T. (1978): Upper Jurassic fossils from the Ruahine Range (Note). *New Zealand Journal of Geology and Geophysics*, 21(6), 773-774.

The Soil Conservation and Rivers Control Council of New Zealand (1957): Floods in New Zealand 1920-53 with notes on some earlier floods, 239p.

Thompson, S.M. (1976): Siltation of Clutha hydro-electric lakes. *Proceedings of New Zealand Hydrological Society Rotorua Conference*.

Thomson, J.A. (1914): Mineral prospects of the Maharahara district - Hawke's Bay. *Parliamentary Paper*, C-2, 62-70.

Van der Lingen, G.J. (1969): The turbidite problem. *New Zealand Journal of Geology and Geophysics*, 12, 7-50.

Varnes, D.J. (1958): Landslide Types and Processes. In Eckel, E.B. (ed). Landslides and Engineering Practice. Highway Research Board Special Report 29.

Varnes, D.J. (1978): Slope Movement Types and Processes. In Schuster, R.L. & Krizek, R.J. (eds). Landslides: Analysis and Control. Special Report 176.

Vella, P. (1962): Determining Depths of New Zealand Tertiary Seas. An introduction to depth paleo-ecology. *Tuatara*, X(1), 19-40.

Vucetich, C.G. & Howarth, R. (1976): Proposed definition of the Kawakawa Tephra, the *c.* 20 000-years-B.P. marker horizon in the New Zealand region. *New Zealand Journal of Geology and Geophysics*, 19, 43-50.

Waghorn, R.J. (1927): "Earthquake rents" as evidence of recent surface faulting in Hawkes Bay. *New Zealand Journal of Science and Technology*, 9, 22-26.

Walcott, R.I. (1978a): Present tectonics and late Cenozoic evolution of New Zealand. *Geophysical Journal of the Royal Astronomical Society*, 52, 137-164.

Walcott, R.I. (1978b): Geodetic strains and large earthquakes in the axial tectonic belt of North Island, New Zealand. *Journal of Geophysical Research*, 83, 4419-4429.

Walker, R.G. (1965): The origin and significance of the internal sedimentary structures of turbidites. *Proceedings of the Yorkshire Geological Society*, 35, 1-32.

Walker, R.G. (1966): Shale Grit and Gindslow Shales: transition from turbidite to shallow water sediments in the Upper Carboniferous of northern England. *Journal of Sedimentary Petrology,* 36(1), 90-114.

Walker, R.G. (1967): Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. *Journal of Sedimentary Petrology,* 37(1), 25-43.

Walker, R.G. (1973): Mopping-up the turbidite mess. In Ginsburg, R.N. (ed). Evolving Concepts in Sedimentology, 1-37. John Hopkins Press, Baltimore.

Walker, R.G. (1976): Facies Model 2. Turbidites and associated coarse clastic deposits. *Geoscience Canada,* 3(1),

Walker, R.G. (1981): Facies Models 8. Turbidites and Associated Coarse Clastic Deposits. In Walker, R.G. (ed). Facies Models. *Geoscience Canada Reprint Series,* 1, 91-103.

Walton, E.K. (1956): Limitations of graded bedding and alternative criteria of upward sequence in the rocks of Southern Uplands. *Edinburgh Geological Society Transactions,* 16, 262-271.

Warren, G. (1967): Hokitika (1st ed). Geological Map of New Zealand, 1:250 000. New Zealand Department of Scientific and Industrial Research, Wellington.

Waterhouse, J.B. (1975): The Rangitata Orogen. *Pacific Geology,* 9, 35-73.

Webby, B.D. (1959): Sedimentation of the alternating greywacke and argillite strata in the Porirua district. *New Zealand Journal of Geology and Geophysics,* 2, 461-478.

Wellman, H.W. (1948): Tararua Range summit height accordance. *New Zealand Journal of Science and Technology,* 30 (Sect. B), 123-127.

Wellman, H.W. (1952): The Permian-Jurassic stratified rocks. Symposium sur les Series de Gondwana. *Proceedings of the 19th International Geological Congress,* 13-24.

Wellman, H.W. (1956): Structural outline of New Zealand. *Department of Scientific and Industrial Research Bulletin,* 121, 1-36.

Wellman, H.W. (1967): Report on studies relating to Quaternary diastrophism in New Zealand. Minutes of the working group meeting for the Neotectonic study of the Pacific Regions, Appendix 5. *Quaternary Research (Japan),* 6, 34-36.

Wellman, H.W. (1972): Rate of horizontal displacement in New Zealand. *Nature,* 237, 275-277.

Wentworth, C.K. (1922): A scale of grade and class terms for clastic sediments. *Journal of Geology,* 30, 377-392.

Wentworth, C.M. (1967): Dish structure, a primary sedimentary structure in coarse turbidites (abstract). *American Association of Petroleum Geologists Bulletin,* 51, 485.

Wheeler, H.E. *et al.* (1950). Stratigraphic classification. *American Association of Petroleum Geologists Bulletin,* 34, 2361-2365.

Wilcox, R.E., Harding, T.P. & Seely, D.R. (1973): Basic wrench tectonics. *American Association of Petroleum Geologists Bulletin,* 57, 74-97.

Willett, R.W. (1948): Undifferentiated Jurassic-Triassic-Permian. In Marwick, J. (ed). Outline of the Geology of New Zealand. New Zealand Geological Survey, Wellington.

Williams, E. (1960): Intra-stratal flow and convolute folding. *Geological Magazine,* 97(3), 208-214.

Wood, D.A., Gibson, I.L. & Thompson, R.N. (1976): Elemental mobility during zeolite facies metamorphism of the Tertiary basalts of eastern Iceland. *Contributions to Mineralogy and Petrology,* 55, 241-254.

Young, A. (1969): Present rate of land erosion. *Nature,* 224, 851-852.

Zotov, V.D. (1940): Certain types of soil erosion and resultant relief features on the higher mountains of New Zealand. *New Zealand Journal of Science and Technology,* 5(2) (Sect B), 256-262.

Zutelija, B. (1974): The basement structures at the Manawatu Gorge. Unpublished BSc(Hons) thesis, Auckland University, Auckland.

APPENDICES

Appendix Ia: Manawatu Catchment Board expenditure on erosion control and investigations into the erosion problem along the southeastern Ruahine Range front.

Financial Year	Investigations	Status Quo Work	Maintenance	Yearly Total
1974 - 1975	$ 23,288			$ 23,288
1975 - 1976	$ 51,625	$ 58,485		$110,110
1976 - 1977	$ 65,830	$ 75,938		$141,768
1977 - 1978	$ 41,873	$ 54,542		$ 96,415
1978 - 1979	$ 1,040	$ 72,751	$ 2,083	$ 75,874
1979 - 1980		$ 74,604	$ 3,144	$ 77,748
1980 - 1981		$ 73,217	$11,950	$ 85,167
	$183,656	$409,537	$17,177	$610,370

Appendix Ib: Estimated government subsidies (on a basis of 3 : 1) required by the Board during the next five year period.

1981 - 1982	$ 94,000	1982 - 1983	$130,000
1983 - 1984	$150,000	1984 - 1985	$111,500

Source: Manawatu Catchment Board, Palmerston North

Appendix IIa: Forest Service expenditure on slope and streambed stabilization in the southern Ruahine Range.

Financial Year	Southwest Ruahines	Southeast Ruahines	Yearly Total
1975 - 1976	$ 2,004	$ 220	$ 2,244
1976 - 1977	$ 5,006	$ 1,453	$ 6,459
1977 - 1978	$ 9,389	$27,761	$ 37,150
1978 - 1979	$ 4,451	$15,843	$ 20,294
1979 - 1980	$ 9.066	$17,369	$ 26,435
1980 - 1981	$ 3,419	$11,899	$ 15,318
	$33,335	$74,545	$107,880

Appendix IIb: Estimated future expenditure by the Forest Service on slope and streambed stabilisation.

Financial Year	Southwest Ruahines	Southeast Ruahines	Yearly Total
1983 - 1984	$15,000+ -	$16,000+ $13,000*	$44,000
1984 - 1985	$18,000	$32,000	$50,000
1985 - 1986	$23,000	$40,000	$63,000

+ Estimated expenditure on aerial seeding and fertilizing of valley slopes.
* Estimated expenditure on pole planting and streambed stabilization.

Source : New Zealand Forest Service, Palmerston North.

APPENDIX IIIa: Faunal list for limestone block T23/f7530

Macrofossil identification by H J Campbell:

- Clionidae
- Porifera
- Foraminifera
- *Lingula* sp.
- Disciniscinae gen.
- ? *Athyris* sp.
- *Sulcirostra*
- *Pisirhynchia*
- *Aulacothyris*
- Brachiopoda indet.
- Gastropoda indet.
- ? *Arcestes* sp. indet.
- Ammonoidea indet.
- *Parallelodon* sp.
- *Halobia lilliei* Marwick
- *Halobia hochstetteri* Mojsisovics
- *Halobia* sp.
- ? *Parahalobia* sp.
- *Oxytoma* sp.
- Pectinacea indet.
- Ostreacea indet.
- aff. *Plicatula* sp.
- ? *Pseudolimea* sp.
- Bivalvia indet.
- Crinoidea (2 forms)
- Echinoidea (2 forms)
- ? *Serpula* sp.
- Ostracoda
- Conodonta
- Vertebrata (teeth)

Macrofossil identification by I W Keyes:

- Hybodus sp. (Elasmobranch tooth)

Macrofossil identification by J G Begg:

- ? Dicyclocidaris denticulata Fell (Interambulacral plate)

Conodont identification by J R Simes:

- *Gondolella navicula*
- *Gondolella Polygnathiformis*
- *Gondolella* sp.
- *Enantiognathus ziegleri*
- *Prioniodina excavata*
- *Prioniodina* sp.
- ? *Hindeodella suevica*

Foraminifera identification by C P Strong:

Nodosarid Foraminifera

- *Lenticulina*
- *Pseudonodosaria*
- *Astacolus*
- *Lingulina*
- *Nodosaria*
- *Vaginulina*
- *Frondicularia*

Polymorphinids

- *Eoguttulina*
- *Pyrulinoides*

Spirillinids

- ? *Spirillina*

Agglutinated foraminfera

- *Trochammina*
- *Ammobaculites*

Miliolids

- ? *Hemigordius*

APPENDIX IIIb: Comments on the identification, interpretation and significance of the faunal content of T23/f7530 (extracts from unpublished manuscript and/or pers. coms.

Macrofauna (H J Campbell):

The limestone may be regarded as a bivalve-brachiopod-echinoderm shell limestone. The most abundant forms are halobiid bivalves and terebratulid and rhynchonellid brachiopods. The fauna (Appendix IIIa) includes two inarticulate and at least four articulate brachiopods. At least one terebratulid form is present. No spiriferinid forms were recognised. At least two rhynchonellides are present, both of which are regarded as cosmopolitan Upper Triassic forms (Ager, 1965). Neither has been recorded from New Zealand before now.

A single gastropod columella was identified from acid residue and several indeterminate gastropod cross-sections were observed in hand specimen.

Two small ammonoids were identified of which one specimen may be referrable to the genus *Arcestes*.

Several well-preserved specimens of a small *Parallelodon* sp. are present and one small ostreacean and one form of *Oxytoma* were identified. Crinoid and echinoid elements are abundant.

Halobiid bivalves are particularly abundant of which *Halobia lilliei* (Marwick, 1953) is the most common. This form is regarded as being restricted to the Oretian stage (Marwick, 1953; Campbell 1955) as defined for sequences of the Murihiku Supergroup. Where faunal similarities can be recognised, this molluscan-based local stage scheme is used for Torlesse strata. In this case, the macrofauna of T23/f7530 may be regarded as Oretian. Traditionally, correlation between the Oretian stage and world Triassic has been poor but in the absence of strong direct lines of correlation the stage has logically been attributed to Late Karnian-Early Norian time (Campbell, 1974).

Halobia lilliei has been described from Torlesse strata once before (Campbell, 1982). Rocks recognised as being specifically Oretian in age within the Torlesse are rare but *Halobia* specimens and faunas attributed to the *Halobia* Zone (Campbell & Warren, 1965; Andrews *et al.*, 1976; Speden, 1976) may be shown to be Oretian in some cases.

Halobia hochstetteri Mojsisovics is present but in fewer numbers than *H. lilliei*. It is regarded as ranging from Oretian to Otamitan time in New Zealand (Marwick, 1953; Campbell & McKellar, 1960).

Overall, this macrofauna is markedly different from any Oretian fauna yet described from either Murihiku or Torlesse terranes of New Zealand. It is the first Oretian limestone to be recognised from either New Zealand or New Caledonia, it is remarkably diverse and it contains forms hitherto unrecorded from New Zealand.

An additional item that has been identified from the macrofauna is an echinoid plate ? *Dicyclocidaris denticulata* Fell (Dr J G Begg).

Conodonts (J E Simes):

The conodont fauna (Appendix IIIa) is rich and well preserved, particularly the more robust platform conodonts. Ramiform conodonts are also abundant but are far more prone to breakage. The most striking feature of this conodont fauna is the abundance of the platform conodont *Gondolella navicula* Huckriede (Appendix IIIc, photo A). This species has been previously recognised from New Zealand at Mount Mason (Jenkins & Jenkins, 1971). While the specimens from T23/f7530 are Oretian in age, the Mount Mason specimens are of Warepan age.

A second diagnostic conodont *Gondolella polygnathiformis* Burdurov and Stefanov is recorded from New Zealand for the first time (Appendix IIIc, photo B).

Correlation of the New Zealand Triassic with the world Triassic is aided by this fauna. Of particular significance is the joint occurrence of conodonts and halobiid bivalves, both of which have a restrictive time range. The New Zealand Oretian Stage has been correlated with the lower-most Norian of the world Triassic.

Conodont colour is a preservational feature that has been demonstrated to be a valuable indicator of the thermal metamorphism of the rock (Epstein & Epstein, 1977). These conodonts are a light amber colour and have a CAI (colour alteration index) of 1.5 based on the recommended reading of the lightest elements only. A reading of CAI 1 indicates a maximum temperature of 90°C which, in turn, indicates a maximum depth of burial of 9000 ft (2743m) (assuming 1°C = 100 ft (30m).

Foraminifera (C P Strong):

This sample is dominated by nodosarid foraminifera. Other calcareous taxa include the polymorphinids, a single spirillinid genus and rare agglutinated foraminifera (Appendix IIIa).

Triassic foraminifera in New Zealand are poorly documented and comparison with overseas literature is proving to be difficult on account of the poor preservation of the specimens and because of the probable endemic nature of much of the fauna.

The age and correlation of this sample based upon foraminiferal evidence is at present unknown.

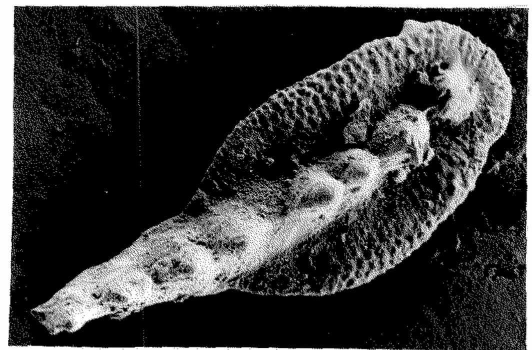
Other fauna

Sample T23/f7530 contains abundant ostracods, many of which are intact but unfortunately preservation is by a crude silicification process. To date little is known about the ostracods from this sample.

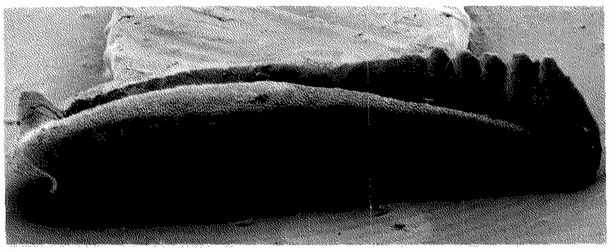
Teeth and dermal ossicles of elasmobranchs are well preserved. The elasmobranch remains include a tooth of *Hybodus* sp. (I W Keyes).

APPENDIX IIIc: Conodont fauna from fossiliferous limestone block T23/ f7530. Fossil identification by J E Simes.

A. top view

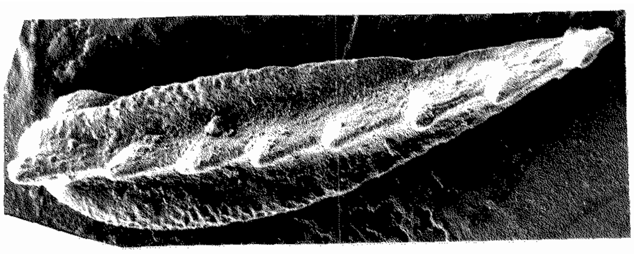


B. top view



A. lateral view

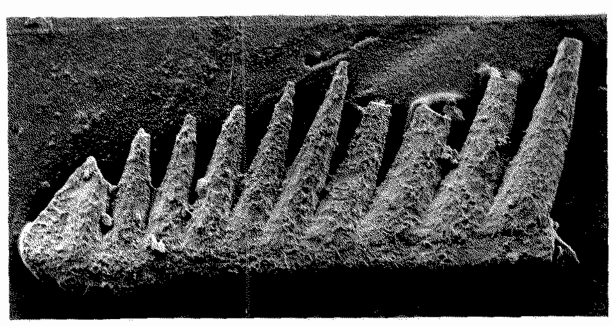
B. lateral view



C.

D.

E.



F.

A. Gondolella navicula. Top and lateral views, x 100.

B. Gondolella polygnathiformis. Top and lateral views, x 140.

C. Gondolella navicula juvenile, x 160.

D. Prioniodina excavata, x 110.

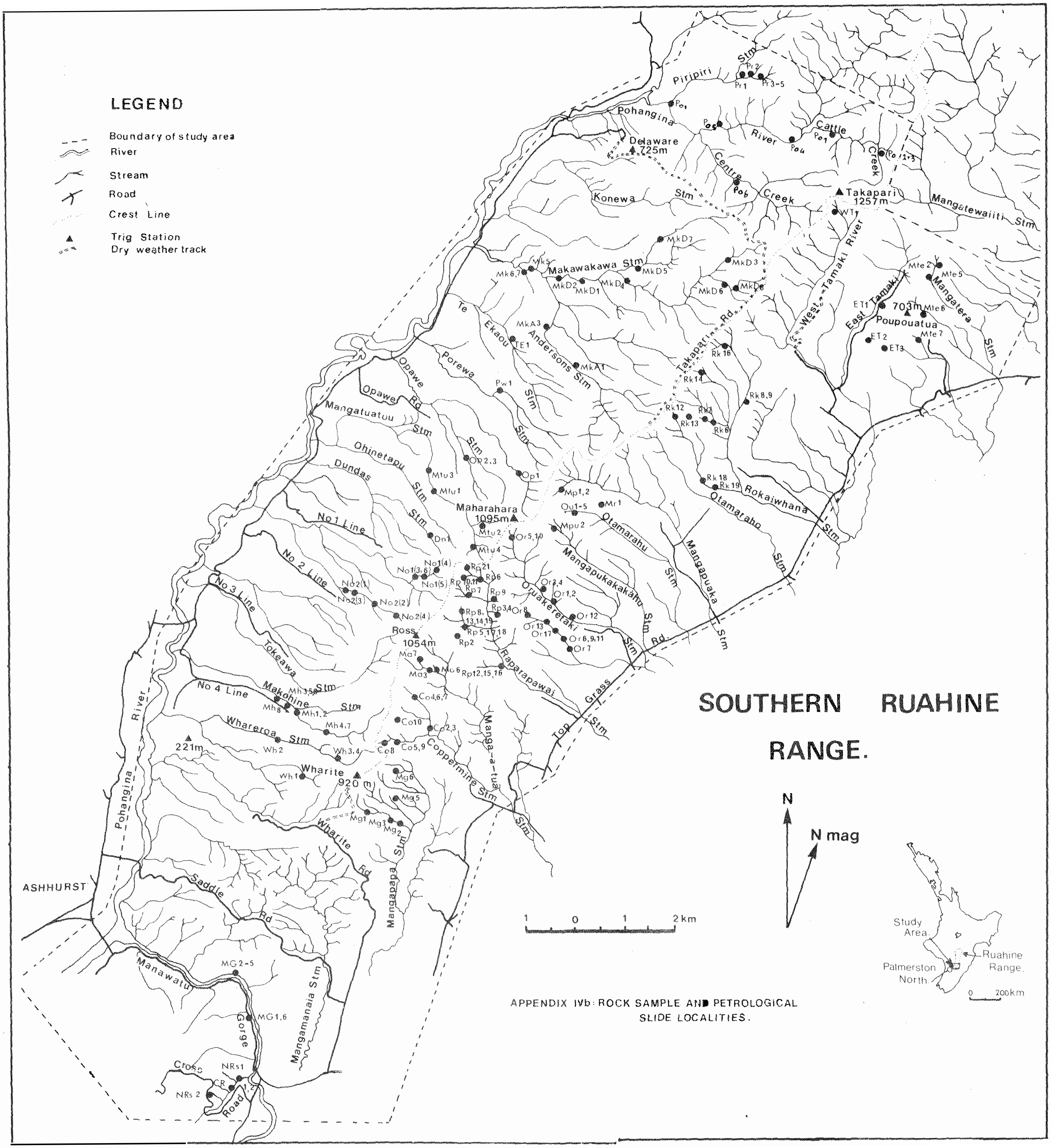
E. Enantiognathus ziegleri, x 100.

F. ? Hindeodella suevica.

APPENDIX IVa: Abbreviations used to denote locality of rock samples and corresponding petrological slides.

Coppermine Stream	Co	No. 1 Line Stream	No. 1
Cross Road	CR	No. 2 Line Stream	No. 2
		North Range Road Stream	NRs
Dundas Stream	Dn		
		Ohinetapu Stream	Oh
East Tamaki Stream	ET	Opawe Stream	Op
		Oruakeretaki Stream	Or
Konewa Stream	Ko	Otamarahu Stream	Ou
Maharahara Track	Mr	Piripiri Stream	Pr
Makawakawa Stream	Mk	Pohangina River	Po
Andersons Stream	MkA	Porewa Stream	Pw
Diggers Stream	MkD		
Makohine Stream (No. 4 Line)	Mh	Raparapawai Stream	Rp
Manawatu Gorge	MG	Rokaiwhana Stream	Rk
Manga-a-tua Stream	Ma		
Mangapapa Stream	Mg	Te Ekaou Stream	TE
Mangapuaka (Kumeti) Stream	Mp	Tokeawa Stream (No. 3 Line)	Tk
Mangapukakakahu Stream	Mpu		
Mangatera Stream and Foothills	Mte	West Tamaki River	WT
Mangatuatou Stream	Mtu	Whareroa Stream	Wh

Each rock sample and petrological slide is prefixed by letters (as above) followed by a number. The letters denote the catchment from which the rock sample was collected. The number denotes the location within the catchment. Each location from which samples were collected is shown in Appendix IVb.



APPENDIX IVb: ROCK SAMPLE AND PETROLOGICAL SLIDE LOCALITIES.

APPENDIX IVc: Major trace element analyses of spilitic rocks from the southern Ruahine Range.
(Analyses carried out by Dr B Roser, Geology Department, University of Wellington.)

	OUTCROP												
	MG3	MG5	Cr2	Or9	Pw1	Rk2	MkD4	Mk5	Mk6	Mh3	No. 1 (5)	No. 2 (4)	Mtu4
SiO_2	48.80	41.38	51.44	48.11	50.80	67.24	47.47	43.60	38.77	46.67	30.65	51.19	46.39
TiO_2	0.79	1.64	2.33	3.14	2.48	0.56	0.81	1.39	2.45	1.86	2.09	3.39	1.08
Al_2O_3	13.36	13.05	12.83	16.91	15.76	14.43	12.49	18.21	13.86	13.15	11.23	12.21	14.40
Fe_2O_3	2.24	9.25	3.30	4.29	2.17	1.37	2.92	4.51	4.56	0.85	2.98	1.47	5.20
FeO	6.17	3.99	11.48	5.92	5.39	2.86	3.93	2.82	7.09	0.51	4.59	6.50	6.18
MnO	0.19	0.57	0.26	0.22	0.13	0.08	0.12	0.23	0.20	0.06	0.22	0.17	0.23
MgO	6.95	4.61	3.12	3.99	3.12	2.01	2.26	2.88	2.46	0.52	2.75	1.74	3.71
CaO	13.95	15.18	4.66	4.54	8.84	1.94	17.69	8.94	11.98	15.62	21.97	9.48	12.24
Na_2O	1.96	2.75	4.03	2.59	5.84	3.67	1.85	3.61	4.44	6.89	2.69	4.73	2.48
K_2O	0.03	0.51	0.06	2.61	0.09	2.18	0.01	3.34	0.89	0.26	1.33	0.19	0.22
P_2O_5	0.04	0.16	1.00	1.03	0.23	0.10	0.23	0.30	0.61	0.94	0.71	1.69	0.11
Loss	5.28	6.81	5.08	6.31	4.66	3.31	10.04	9.80	12.16	12.21	18.60	7.00	7.54
TOTAL	99.76	99.90	99.59	99.66	99.51	99.75	99.81	99.64	99.47	99.53	99.81	99.76	99.78
Fe_2O_3T	9.10	13.69	16.06	10.87	8.16	4.55	7.28	7.65	12.44	1.41	8.08	8.69	12.07
V	240	373	42	205	325	100	234	215	317	120	286	254	305
Ba	28	215	108	465	91	505	38	5850	151	95	243	147	71
Cr	256	128	5	176	61	110	408	78	126	272	52	16	138
Cu	80	85	27	46	44	24	43	46	53	15	30	59	78
Ga	16	16	28	22	22	15	18	18	19	10	16	20	16
Nb	< 1	5	41	60	19	8	3	11	29	53	23	50	2
Ni	58	73	6	132	45	24	114	37	58	8	26	27	78
Pb	< 2	3	2	3	< 2	17	2	< 2	< 2	2	< 2	6	< 2
Rb	< 2	16	< 2	102	< 2	83	< 2	106	42	11	56	6	8
Sr	75	155	349	360	248	286	153	1187	322	234	290	182	145
Th	1.7	2.7	5.1	6.6	2.5	10.5	2.2	1.8	3.8	5.1	3.0	4.8	1.3
Y	19	37	80	54	37	21	19	28	53	75	34	77	31
Zn	64	124	185	146	97	60	70	74	205	43	94	186	115
Zr	40	96	486	483	174	163	44	109	270	360	169	356	57

	FLOAT							
	Mk1	Mk2	Mk3	MkA2	Ma1	Rp20	No. 1 (1)	Tk1
SiO_2	40.36	45.93	64.53	34.43	49.89	45.43	42.51	49.68
TiO_2	2.51	3.03	0.28	2.84	0.20	2.05	2.04	1.36
Al_2O_3	11.39	17.03	17.74	12.19	16.27	14.34	19.50	19.06
Fe_2O_3	7.85	1.50	1.81	10.46	1.05	1.88	2.23	3.17
FeO	3.47	8.75	3.05	1.13	0.58	7.65	5.84	4.38
MnO	0.21	0.16	0.10	0.13	0.22	0.17	0.17	0.15
MgO	1.20	1.92	0.60	1.51	0.56	1.64	2.68	3.10
CaO	14.79	6.13	0.54	17.37	13.82	10.20	10.38	7.66
Na_2O	4.67	5.55	3.95	3.11	1.81	5.73	3.35	4.92
K_2O	0.68	1.09	4.10	1.79	0.95	0.14	1.81	0.71
P_2O_5	0.43	0.86	0.00	0.65	0.22	0.68	0.30	0.13
Loss	12.55	7.55	2.93	14.63	14.80	10.20	9.33	5.65
TOTAL	100.10	99.50	99.63	100.23	100.36	100.12	100.14	99.97
Fe_2O_3T	11.70	11.22	5.20	11.71	1.70	10.39	8.72	8.04
Ba	649	208	872	172	531	159	667	320
Cr	7	249	< 2	41	5	23	82	115
Cu	27	36	3	20	7	38	21	47
Ga	12	19	55	18	7	16	22	15
Nb	26	46	257	21	2	31	21	12
Ni	25	103	< 2	58	7	39	29	42
Pb	3	3	8	< 2	7	4	< 2	< 2
Rb	25	52	157	71	31	4	65	11
Sr	536	283	442	261	656	236	408	466
Th	2.7	2.6	25.4	1.7	<0.5	2.8	1.7	1.3
V	227	227	< 2	262	23	306	383	206
Y	37	44	147	43	5	43	27	21
Zn	125	152	284	190	16	210	80	74
Zr	205	286	1176	216	31	147	153	94

APPENDIX Va: Late Quaternary Tectonic Data

ABBREVIATIONS - TABLE HEADINGS			
NZMS	New Zealand Map Series	Lk	Length in kilometres
260	1:50,000 scale (in quarter sheets)	FPLANE	Faultplane
		Dir	Direction of dip
Loc	Locality number plotted on upthrown side	Am	Amount of dip in degrees
Gr ref	Grid reference to nearest 100 metres	VERT	Vertical displacement
		Up	Upthrown side
Na	Nature of tectonic feature	Amm	Amount in metres
Dp	Displaced reference feature	HORZ	Horizontal displacement
Lm	Length in metres	S	Sense of horizontal displacement
HDm	Height or depth in metres		
TrFOLD	Trace or fold	Fm age	Age of formation displaced
Str	Strike of feature	Rf	Reference cited and remarks

ABBREVIATIONS - TECTONIC DATA			
ag	Aggradational	os	Offset stream
dg	Degradational	pr	Pressure ridge
ds	Diverted stream	rd	Riser downstream side
E	East	ru	Riser upstream side
fg	Fault guided stream	rg	Ridge
gg	Fault gouge	S	South
gr	Grid reference	sc	Scarp
H	Hawera Series terrace surface (see Table 6.1 forexplanation)	sd	Saddle
		Sh	Shutter ridge
ic	Intermittently continuous	sl	Slope
K-O	Kawhia-Oteke age of strata comprising Torlesse Supergroup bedrock	tc	Trace
		te	Terrace
lk	Linear Depression	tr	Trench
lf	Local fan	TS	Torlesse Supergroup bedrock
lin	Lineation	us	Uphill-facing scarp
md	Mound	W	West
N	North	Wc	Castlecliffian aged marine deposits
op	Offset spur	Wn	Nukumaruan aged marine deposits
or	Offset riser	Ww	Waitotaran aged marine deposits

WELLINGTON FAULT

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		Fm age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T24A	020	479907	238	40	us	sl	30	1.0	005	ic			W				Wn-c	Inset E.grades into gravity sc. fault pond.
T24A	021	484914	238	40	sc	sl	45	2.0	030	ic			W				Wn-c	Inset E.two scarps at head of slump.
T24A	022	484914	238	40	sc	sl	60	2.5	010	ic			W				Wn-c	Inset E.grades into gravity sc.
T24A	023	485916	238	40	us	sl	30	1.0	010	ic			W				Wn-c	Inset E.tr 1m wide sc faces E.
T24A	024	485916	238	40	us	sl	60	1.0	020	ic			W				Wn-c	Inset E.sc faces E.
T24A	025	486917	238	40	us	sl	45	0.5	020	ic			W				Wn-c	Inset E.040° at N end.tr 1m wide
T24A	026	471892	238	40	sc	ag	930	16	005	ic			E				Wn-c	Inset E.gr at S end
T24A	027	473893	238	40	tr	ag	14	3.0	025	ic							Wn-c	Inset E.
T24A	028	473894	238	40	tr	ag	5.0	0.5	010	ic			E				Wn-c	Inset E.
T24A	029	473895	238	40	sc	ag	15	1.0	020	ic			E				Wn-c	Inset E.
T24A	030	474896	238	40	us	ag	35	0.5	000	ic							Wn-c	Inset E.
T24A	031	476897	238	40	sc	ag	900	5.0	005	ic			W				Wn-c	Inset E.
T24A	032	478901	238	40	sc	sl	20	4.0	030	ic			W				Wn-c	Inset E. Spring at S end. Sc. of variable height.
T24A	033	479902	238	40	sc	sl	45	4.0	030	ic			W				Wn-c	Inset E.
T24A	034	479902	238	40	sc	sl	45	4.0	030	ic			W				Wn-c	Inset E.
T24A	035	480902	238	40	sc	sl	100	3.0	030	ic			W				Wn-c	Inset E.
T24A	036	480902	238	40	sc	sl	30	0.5	045	ic			NW				Wn-c	Inset E.
T24A	037	481903	238	40	sc	sl	15	0.5	035	ic			W				Wn-c	Inset E.
T24A	038	481903	238	40	sc	sl	30	0.5	030	ic			W				Wn-c	Inset E.
T24A	039	481903	238	40	sc	sl	45	3.0	050	ic			NW				Wn-c	Inset E.
T24A	040	478900	238	40	sc	sl	45	3.0	025	ic			E				Wn-c	Inset E.
T24A	041	479900	238	40	us	sl	70	1.5	025	ic			W				Wn-c	Inset E.
T24A	042	480901	238	40	sc	sl	75	8.0	010	ic			E				Wn-c	Inset E.
T24A	043	481902	238	40	sc	sl	50	8.0	010	ic			E				Wn-c	Inset E.
T24A	044	481903	238	40	sc	sl	50		010	ic			E				Wn-c	Inset E.
T24A	045	481903	238	40	us	sl	15	1.0	010	ic			E				Wn-c	Inset E.
T24A	046	482904	238	40	sc	sl	15	3.0	005	ic			W				Wn-c	Inset E. Spring at N end.
T24A	047	482904	238	40	sc	sl	70	4.0	005	ic			W				Wn-c	Inset E.
T24A	048	483907	238	40	ds	sl	25	2.0	015	ic			E				Wn-c	Inset E. ds 20m.
T24A	049	483907	238	40	sc	sl	30	3.0	015	ic			E				Wn-c	Inset E.
T24A	050	484910	238	40	pr	sl	180	3.0	015	ic			E				Wn-c	Inset E.
T24A	051	485911	238	40	pr	sl	330	2.0	005	ic			E				Wn-c	Inset E. 5m high at N end.
T24A	052	485912	238	40	us	sl	75	0.5	020	ic			E				Wn-c	Inset E. tr 1m wide.
T24A	052	485912	238	40	us	sl	75	1.0	020	ic			E				Wn-c	Inset E. tr 1m wide.
T24A	052	485912	238	40	us	sl	75		020	ic			E				Wn-c	Inset E. See on photo. Now a farm track.
T24A	053	486912	238	40	us	sl	75	0.3	020	ic			E				Wn-c	Inset E. tr 1m wide.
T24A	053	486912	238	40	us	sl	75	1.0	020	ic			E				Wn-c	Inset E. tr 1.5m wide.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T24A	053	486912	238	40	us	sl	75	0.5	020	ic			E				Wn-c	Inset E. tr 1m wide.
T24A	053	486912	238	40	us	sl	75		020	ic			E				Wn-c	Inset E. tr destroyed by farm track.
T24A	054	486913	238	40	us	sl	75	0.3	020	ic			E				Wn-c	Inset E.
T24A	054	486913	238	40	us	sl	80	1.0	020	ic			E				Wn-c	Inset E.
T24A	054	486913	238	40	us	sl	100	0.5	020	ic			E				Wn-c	Inset E.
T24A	054	486913	238	40	us	sl	100	0.5	020	ic			E				Wn-c	Inset E.
T24A	055	487914	238	40	us	sl	25	0.5	020	ic			E				Wn-c	Inset E. Cut by farm track.
T24A	055	487914	238	40	us	sl	45	0.5	020	ic			E				Wn-c	Inset E.
T24A	055	487914	238	40	us	sl	15	0.5	020	ic			E				Wn-c	Inset E. tr destroyed by farm track.
T24A	056	489916	238	40	sc	sl	30	1.0	005	ic			E				Wn-c	Inset E.possible shr across stream.
T24B	201	504930	237	40	us	sl	50	1.0	015	ic			E				Wn	Cut by stream. Swampy ground at S end of tr.
T24B	201a	506932	237	40	op	sl			015	ic			W		d	20	Wn	sd at point of offset 20m wide, 6m deep.
T24B	201b	509933	237	40	sc	sl	150	2.0	015	ic			E				Wn	Modified scarp along base of slope.
T24B	201c	510934	237	40	sc	sl	150		015	ic							Wn	Collapsed scarp face.
T24B	202	509936	237	40	sc	lf	25	15	015	ic			W				H	tc truncates Wn spur at S end. gr taken at N end of tc. lf capped with 1-2m ag gravels.
T24B	203	511939	236	40	sc	lf	250	15	015	ic			W				H	Swampy ground at base of scarp. lf capped with 1-2m ag gravels.
T24B	204	513941	236	40	sc	lf	275	15	020	ic			W				H	Inset B. lf capped with 1-2m ag gravels.
T24B	205	514942	236	40	sc	lf	75	4.0	020	ic			W				H	Inset B.
T24B	205a	515943	236	40	sc	lf	25	0.5	020	ic			W				H	Inset B.
T24B	206	517947	236	40	sc	ag	385	10	010	ic			W				H	Inset B. gr taken at N end. Surface dg at S end.
T24B	207	516946	236	40	sc	ag	250	2.0	035	ic			W				H	Inset B. Rejuvenated scarp 30 cm high?
T24B	208	521951	236	40	sc	ag	650	15	010	ic			W				H	Inset B.
T24B	209	520952	236	40	sc	ag	80	1.0	015	ic			W				H	Inset B.
T24B	210	521953	236	40	ld	sl	150	1.0	015	ic			W				Ww	Inset B. Swampy ground in depression 5m wide.
T248	211	522954	236	40	ld	sl	175	4.0	015	ic			W				Ww	Inset B. Swampy ground in depression.
T24B	212	522952	236	40	tc	ag				ic							H	Inset B.unable to locate this feature in field.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T24B	213	523953	236	40	us	sc	12	1.0	005	ic			E				Wn	Inset B.tr 1m wide.
T24B	214	524953	236	40	md	lf	50	4.0	020	ic			W				H	Inset B. Steep E side cut by old stream channel.
T24B	215	524954	236	40	sc	lf	185	5.0	015	ic			W				H	Inset B.displacement highest in middle of fan tapers to 2m height at periphery of fan.
T24B	216	523954	236	40	sc	te	90	15	010	ic			W				H	Inset B. gr taken at S end. Two bench levels 6m below.
T24B	217	526958	236	40	sc	te	625	15	010	ic			W				H	Inset B. Base of scarp wet and swampy.
T24B	218	529962	236	40	us	sl	60	2.0	015	ic			E		d	20	Wn	4m wide. Spring offset stream 20m.
T24B	219	529962	236	40	sc	sl	35	6.0	015	ic			E				Wn	Sc runs into headwall of slump.
T24B	220	532966	236	40	us	sl	55	6.0	015	ic			E				Wn	tr 6m wide. Shallower tr at S end 25m long, 2m wide. Spring.
T24B	221	533967	236	40	us	sl	80	0.5	015	ic			E				Wn	sc west facing on E side of tr.
T24B	222	533967	236	40	us	sl	40	2.0	015	ic			E				Wn	tr decreases in depth northward. 5m wide.
T24B	223	534968	236	40	us	sl	95	3.0	015	ic			E				Wn	tr 2m deep at S end, 2.5m long, 5m wide.
T24B	223a	535968	236	40	us	sl	8.0	1.0	015	ic			E				Wn	tr 1m wide. Gas pipeline separates tr at locality 223 from that at 223A.
T24B	224	536969	235	52	sh	sl	125	15	015	ic			E		d	35	Wn	os 35m.
T24B	225	537970	235	52	sc	sl	35	2	015	ic			W				Wn	
T24B	226	537971	235	52	sc	sl	50	1.5	015	ic			E				Wn	sc height decreases upslope.
T24B	226a	537973	235	52	us	sl	20	0.5	005	ic			W				Wn	tr 2m wide.
T24B	226b	537973	235	52	us	sl	40	0.5	005	ic			W				Wn	tr 3m wide.
T24B	227	539972	235	52	us	sl	23	2.0	015	ic			E				Wn	tr 6m wide curves around hillside.
T24B	228	541977	235	52	sc	ag	320	15	005	ic			W				H	Inset C.
T24B	229	542979	235	52	sc	ag	90	13	005	ic			W				H	Inset C.rejuvenated. Collapse features along scarp change in strike.
T24B	230	543980	235	52	sc	ag	60	2	010	ic			W	6.0			H	Inset C.
T24B	231	544981	235	52	sc	ag	80	4	010	ic			W	2.0			H	Inset C.
T24B	232	544982	235	52	sc	ag	9.0	4	010	ic			W		d	9.0	H	Inset C.distance between 2 channels.
T24B	233	544982	235	52	sc	ag	5.0	4	010	ic			W				H	Inset C.
T24B	234	545982	235	52	sc	ag	70	10	010	ic			W				H	Inset C.
T24B	235	544981	235	52	sc	ag	175	2.0	010	ic			E		d	40	H	Inset C. 2m sc on E side of tr.
T24B	236	545983	235	52	sc	ag	45	8.0	010	ic			E				Wn	Inset C.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		Fm age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T24B	237	545983	235	52	us	sl	25	2	010	ic			W				Wn	Inset C. Lies to E of location 236
T24B	238	547985	235	52	tr	te	60	2	010	ic			W	10			Wn	Inset C. Fault pond.
T24B	239	548986	235	52	us	sl	30		010	ic			E				Wn	Inset C. Unstable hillside.
T24B	240	548986	235	52	sc	sl	60	3	010	ic			E				Wn	Artificial pond on W side of W facing sc.
T24B	241	548987	235	52	sc	sl	40	4	010	ic			E				Wn	ds along base of W facing sc.
T24B	242	548988	235	52	sc	sl	90	0.5	010	ic			E				Wn	Subdued W facing sc.
T24B	243	550990	235	52	sc	sl	60	6	010	ic			E				Wn-c	Two parallel scarps. Sc to W is 20m long by 2m high. Flat surface between scarps
T24B	244	551992	235	52	sh	rg	100	6	010	ic			E		d	90	Wn-c	Stream offset southward.
T24B	245	551992	235	52	sc	sl	100	6	010	ic			W				Wn-c	Sc. faces east.
T24B	246	552994	235	52	sh	sl	60	4	010	ic			E		d	60	Wn-c	50m long W facing sc. Offset stream southwards. S end of sh eroded.
T24B	247	552994	235	52	us	sl	40	2	010	ic			E				Wn-c	W facing sc.
T24B	248	553996	235	52	sh	sl	60	4	010	ic			E		d	67	Wn-c	S end of sh eroded. Abandoned stream channel at N end of sh.
T24B	249	553996	234	16	sc	sl	20	2.0	010	ic			E				Wn-c	
T24B	250	553996	234	16	us	sl	60	6.0	010	ic			E				Wn-c	
T24B	251	554997	234	16	sc	sl	100	4.0	010	ic			E				Wn-c	
T24B	252	556999	234	16	sh	rg	60	10	010	ic					d	60	Wn-c	Stream offset southwards.
T24B	253	556999	234	16	sc	sl	25	10	010	ic			W				Wn-c	East facing sc dies out northwards.
T24B	254	556000	234	16	sh	sl	20	4.0	010	ic			E		d	20	Wn-c	Stream offset southwards.
T24B	261	547995	234	16	sc	ag	125	1.0	000	ic			W				H	Sc partly modified by small drainage channel. Displaces topmost te surface.
T24B	262	547996	234	16	sc	ag	125	0.5	000	ic			W				H	E facing subdued sc. Swampy ground on E side of sc.
T24B	263	549999	234	16	us	sl	400	1.0	020	ic			W				Wn-c	Swampy ground tr 8m wide. E facing sc on W side of tr.
T24D	608	555008	234	17	sc	sl	70	0.5	020	ic			E				Wn-c	S end disappears into forest.
T24D	609	556008	234	17	us	sl	10	1.0	010	ic			E				Wn-c	tr 5m wide.
T23D	610	556009	234	17	sd	rg		2.0	010				W	0.5			Wn-c	
T23D	611	557010	234	17	sc	sl	30	1.0	010	ic			E				Wn-c	
T23D	612	557010	234	17	sc	sl	90	6.0	010	ic			E				Wn-c	
T23D	613	558011	234	17	sc	sl	90	5.0	010	ic			E				Wn-c	Swamp on W side spring.
T23D	614	558011	234	17	sc	ag	10	1.0	010	ic			E	2.0	d	15	H	

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		Fm age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	615	562015	233	18	tc	sl	15	2.0	010	ic							Wn-c	Two tc on photo not located in field.
T23D	616	563016	233	18	sc	sl	45	2.0	055	ic			N					
T23D	617	564017	233	18	ld	sl	90	0.5	025	ic							Wn-c	Two parallel depressions with spring at head of each.
T23D	618	565018	233	18	sc	sl	75	2.0	090	ic			S				Wn-c	Borders S edge of slump.
T23D	618a	565018	233	18	sd	rg			025								Wn-c	Lies between two parallel tc.
T23D	619	566019	233	18	sc	sl	46	2.0	025	ic			NW				Wn-c	Lies between farm track and Coppermine Rd. sc on line with other traces parallel to E side of slump.
T23D	620	571021	233	18	sc	rg	100	4.0	025	ic			NW				Wn-c	ld at base of sc. 1m high ridge along E side of ld.
T23D	621	562014	233	18	sc	sl	30	2.0	045	ic			SE				Wn-c	Less regular in outline further up slope.
T23D	622	564016	233	18	us	sl	10	1.0	035	ic			SE				Wn-c	tr at base of E facing 4m high sc
T23D	623	557001	234	17	sc	sl	60	4.0	010	ic			E				Wn-c	sc reduces in height northward.
T23D	624	558003	234	17	sc	sl	10	4.0	010	ic			E				Wn-c	
T23D	625	558003	234	17	sc	sl	30	6.0	010	ic			E				Wn-c	
T23D	626	559004	234	17	sc	sl	150	15	010	ic			E				Wn-c	8m wide.
T23D	627	561006	234	17	sc	sl	225	10	010	ic			E				Wn-c	8m wide. tr cut by streams at S end.
T23D	628	562007	234	17	us	sl	75	4.0	010	ic			W				Wn-c	4m wide. Swamp at S end.
T23D	629	562008	234	17	us	sl	50	4.0	010	ic			W				Wn-c	8m wide. Contains pond.
T23D	630	562008	234	17	us	sl	20	0.5	010	ic			W				Wn-c	1m wide - is extension of tr at 629.
T23D	631	562008	234	17	sc	sl	40	3.0	010	ic			E				Wn-c	sc faces W with flat bench on E side.
T23D	632	563009	234	17	us	sl	25	0.5	010	ic			W				Wn-c	tr 1m wide.
T23D	633	563009	234	17	md	sl	65	3.0	010	ic			E				Wn-c	Subdued E facing sc along E wide of md.
T23D	634	564010	234	17	sc	sl	15	3.0	010	ic			E				Wn-c	sd at N end of sc and to the west.
T23D	635	564010	234	17	sc	sl	34	5.0	010	ic			W				Wn-c	md at loc 633 converges with N end of sc.
T23D	636	565011	234	17	us	sl	35	2.0	010	ic			E				Wn-c	tr 3m wide.
T23D	637	565011	234	17	us	sl	18	1.5	010	ic			E				Wn-c	tr 5m wide.
T23D	638	566012	234	17	sh	sl	125	4.5	010	ic			E		d	80	Wn-c	Swamp pond on W side.
T23D	639	566013	234	17	sc	sl	200	5.0	010	ic			W				Wn-c	sc faces east.
T23D	640	567014	234	17	tr	sl	30	3.0	010	ic			E				Wn-c	5m wide.
T23D	641	568015	234	17	tr	sl	41	2.0	010	ic			E				Wn-c	8m wide.
T23D	642	568016	234	17	sh	sl	30	4.0	010	ic			E		d	30	Wn-c	Swamp on western side of tc.
T23D	643	570018	233	19	ds	ag	75		010	ic			E				H	

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	644	571018	233	19	tc	te	100		015	ic			E				Wn-c	Unstable slope. ld to W of tc.
T23D	645	572019	233	19	sc	te	65	2.0	015	ic			E				Wn-c	
T23D	646	572019	233	19	us	sl	35	1.0	015	ic			W				Wn-c	tr 2m wide. Spring and swamp at s end
T23D	647	572020	233	19	ld	te	70	0.5	015	ic			E				Wn-c	5m wide
T23D	648	572020	233	19	sc	te	30	2.0	015	ic			W				Wn-c	sc faces E.
T23D	649	573021	233	19	us	sl	100	4.0	020	ic			NW				Wn-c	8m wide tr.
T23D	650	576023	233	19	tc	sl	175		035	ic			SE				Wn-c	
T23D	651	577024	233	19	tr	sl	90	3.0	035	ic			SE				Wn-c	10m wide. Spring at N end.
T23D	652	577024	233	19	sc	rg	50	4.0	035	ic			SE				Wn-c	
T23D	653	577024	233	19	ld	rg	140	0.5	020	ic							Wn-c	10m wide. Sd across rg
T23D	654	578025	233	19	ld	rg	45	0.5	020	ic							Wn-c	10m wide.
T23D	655	578025	233	19	sh	sl		1.5	015	ic			E		d	14	Wn-c	8m wide swamp on upstream side.
T23D	656	578025	233	19	us	sl	10	1.0	015	ic			E				Wn-c	1m wide tr.
T23D	657	579026	233	19	sh	sl	60	1.5	015	ic			E		d	25	Wn-c	Swamp and pond on upstream side Drainage at S end blocked by slump.
T23D	658	580028	233	19	sh	sl	70	6.0	025	ic			E		d	50	Wn-c	Swamp and pond on upstream side.
T23D	659	581028	233	19	sh	sl	15	1.0	025	ic			E				Wn-c	Stream cuts through N end.
T23D	660	581028	233	19	us	sl	10	1.0	025	ic			E				Wn-c	
T23D	661	582029	233	19	sh	sl	138	6.0	025	ic			E		d	40	Wn-c	Large swamp on upstream side.
T23D	662	583030	233	19	us	sl	65	2.0	025	ic			E				Wn-c	Artificially dammed 5m wide tr.
T23D	663	583030	233	19	us	sl	20	3.0	025	ic			E				Wn-c	Drains southward
T23D	664	584030	233	19	us	sl	14	3.0	025	ic			E				Wn-c	Drains northward.
T23D	665	584030	233	19	us	sl	7.0	3.0	025	ic			E				Wn-c	Drains southward.
T23D	666	584031	233	19	sh	sl	75	3.0	025	ic			E		d	5	Wn-c	Large swamp on upstream side.
T23D	667	584031	233	19	us	sl	25	2.0	020	ic			E				Wn-c	3m wide tr.
T23D	668	585032	233	20	us	sl	10	1.0	015	ic			E				Wn-c	5m wide tr.
T23D	669	585032	233	20	us	sl	25	1.0	015	ic			E				Wn-c	2m wide.
T23D	670	585032	233	20	us	sl	40	3.0	015	ic			E				Wn-c	5m wide swamp on upstream side. Spring.
T23D	671	586033	233	20	us	sl	20	3.0	015	ic			E				Wn-c	
T23D	672	586033	233	20	sh	sl	15	3.0	015	ic			E				Wn-c	Strata dip 65°E.
T23D	673	586033	233	20	sh	sl	70	4.0	015	ic			E		d	37	Wn-c	
T23D	674	586034	233	20	sc	sl	15	1.0	020	ic			E				Wn-c	Swamp on up-stream side.
T23D	675	587035	233	20	sc	ag	45	2.0	020	ic			E				H	Stream destroyed S end of sc. ag surface tilted 3° to E.
T23D	676	588035	233	20	sh	sl	46	5.0	010	ic			E		d	25	Wn-c	
T23D	677	587035	233	20	sc	ag	12	2.0	015	ic			E				H	Stream drains west side of sc.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	678	587035	233	20	sc	ag	6.0	0.5	015	ic			E				H	Stream drains along western side of sc.
T23D	679	589037	233	20	us	sl	17	1.0	045	ic			W				Wn-c	2m wide - 3 similar tr's in same locality.
T23D	680	590037	233	20	us	sl	25	1.0	055	ic			E				Wn-c	2m wide tr.
T23D	681	590037	233	20	us	sl	8.0	0.5	050	ic			E				Wn-c	1m wide tr.
T23D	682	590037	233	20	us	sl	15	0.5	035	ic			W				Wn-c	1m wide tr.
T23D	683	591037	233	20	sc	sl	45	2.0	010	ic			E				Wn-c	Cut by east draining stream.
T23D	684	592038	233	20	sc	sl	70	2.0	010	ic			E				Wn-c	Cut by east draining stream.
T23D	685	593039	233	20	pr	sl	225	8.0	010	ic			E				Wn-c	Swamp fed from springs.
T23D	686	594040	233	20	sc	sl	100	9.0	010	ic			E				Wn-c	Swamp on west side. Large pond.
T23D	687	594041	233	20	sc	sl	75	4.0	010	ic			E				Wn-c	Swamp on west side.
T23D	688	595042	233	20	sc	ag	225	1.0	020	ic			E				H	Cut in two places by streams. Grid taken at S end.
T23D	689	600048	528	64	fp		225											Inset A. Drainage restricted by alluvial deposits at S end. Pond 80m wide.
T23D	690	602050	528	64	tr	sl	800		015	ic			E				Wn-c	Inset A. Swamps and ponds.
T23D	691	603051	528	64	pr	sl	600		005	ic							Wn-c	Inset A.
T23D	691a	604052	528	64	sc	ag	115	10	015	ic			E				H	Inset A. 12-15m wide swamp on west side.
T23D	691b	600053	528	64	sc	sl	100		070	ic			W				Wn-c	Inset A. Artificially dammed to form pond. Spring and tc further downslope.
T23D	692	604054	528	64	sc	ag	140	1.0	010	ic			E				H	Inset A. sc cut by stream draining along W side. Rejuvenated at base of 691a.
T23D	693	605054	528	64	sc	ag	100	3.0	010	ic			E				H	Inset A.
T23D	694	605054	528	64	sc	ag	50	0.5	010	ic			E				H	Inset A. Subdued scarp modified at N end by erosion. ds on W side.
T23D	695	609059	528	64	sc	ag	250	6.0	040	ic			W				H	Inset A.
T23D	696	600045	528	64	ld	ag	50	0.5	025	ic			E				H	Inset A. 2m wide.
T23D	697	600045	528	64	sc	sl	40	2.0	025	ic			E				Wn-c	Inset A. Spring on W side. Unstable slope.
T23D	698	601046	528	64	sc	sl	60		025	ic			E				Wn-c	Inset A.
T23D	699	602047	528	64	pr	sl	275			ic			E				Wn-c	Inset A. Artificially dammed pond in depression between two md.
T23D	700	603048	528	64	sc	sl	50	2.0	015	ic			E				Wn-c	Inset A. On line with 2 springs. Stream cuts thru N end of sc.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	701	604050	528	64	sc	sl	100	0.5	015	ic			E				H	Inset A. Swamp and drainage along west side of sc.
T23D	702	605051	528	64	sc	ag	210	2.0	005	ic			E				H	Inset A. sc borders E side of pond in part.
T23D	703	606056	528	64	sc	ag	150	2.0	010	ic			E				H	Inset A. ds cuts through sc at N end.
T23D	704	607054	528	64	sc	ag	75	1.0	010	ic			E				H	Inset A. Subdued sc.
T23D	705	607054	528	64	sc	ag	90	1.0	015	ic			E				H	Inset A. Spring at S end.
T23D	706	608055	528	64	sc	ag	135	5.0	005	ic			E				H	Inset A. Stream flows N along W side of sc. gg in stream bank.
T23D	707	610058	528	64	sc	ag	300	5.0	010	ic			E				H	Inset A. Swamp on W side of sc is 100m wide at N end.
T23D	708	611060	528	64	us	sl	40	1.5	340	ic			W				Wn-c	tr 6m tr.
T23D	709	612061	528	64	us	sl	20	1.0	000	ic			W				Wn-c	tr 2m wide.
T23D	710	612063	528	64	pr	ag	75	8.0	000	ic			E				Wn-c	sc on W side, tr on E side at loc 711.
T23D	711	612063	528	64	ld	ag	75	4.0	000	ic			W				H	ld 7m wide.
T23D	712	613064	528	64	sc	ag	200	1.5	000	ic			E				H	
T23D	713	613064	528	64	sc	ag	100	1.0	000	ic			E					
T23D	714	614064	528	64	sc	ag	75	4.0	045	ic			SE				H	Modified by farm track.
T23D	715	615065	527	67	sc	ag	350	6.0	010	ic			E				H	Stream follows base of sc on W side.
T23D	716	618069	527	67	sc	ag	250	5.0	010	ic			E				H	Old channel 8m wide, 2m deep cuts sc.
T23D	717	619071	527	67	sc	ag	250	6.0	010	ic			E				H	
T23D	718	621073	527	67	sc	ag	50	4.0	010	ic			W				H	Modified by fluvial erosion.
T23D	719	624077	527	67	sc	sl	325		010	ic			E				Wn-c	Inset I. gr at N end.
T23D	720	622072	527	67	sc	sl	10	0.5	010	ic			E				Wn-c	sc faces W with bench on E side.
T23D	721	622072	527	67	us	sl	12	1.0	010	ic			W				Wn-c	tr 1m wide. Unstable slope.
T23D	722	622072	517	67	us	sl	30	1.0	010	ic			W				Wn-c	tr 3m wide. Unstable slope.
T23D	723	623076	527	67	us	sl	70	1.0	010	ic			W				Wn-c	tr 8m wide. Swamp and spring.
T23D	724	624077	527	67	us	sl	11	1.0	010	ic			W				Wn-c	tr 2m wide.
T23D	725	624077	527	67	us	sl	30	1.0	010	ic			W				Wn-c	tr 2m wide. sd at N end.
T23D	726	624078	527	67	or	te	125		010	ic			W		d	150	Wn-c	Inset I. Unstable slope with springs.
T23D	727	623079	527	67	sc	ag	125		013	ic			E				Wn-c	
T23D	728	625079	527	67	sc	ag	75	3.0	000	ic			W				H	Inset I. Spring at N end. sc cut by farm track.
T23D	729	619082	527	67	ld	sl	125		000	ic							K-O	sd at top of ld across rg.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	730	620083	527	67	sc	sl	25	2.0	000	ic			E				H	Swamp and spring.
T23D	731	620084	527	67	sc	sl	25	2.0	000	ic			E				H	Swamp and spring on W side.
T23D	732	627082	527	67	sc	sl	40	1.0	015	ic			W				Wn-c	
T23D	733	628083	527	67	tc	sl	75		010	ic							Wn-c	
T23D	734	629084	527	67	us	sl	7.0	0.5	010	ic			E				Wn-c	tr 1m wide.
T23D	735	625085	527	67	tr	rg	15	0.5	010	ic			W				Wn-c	tr 1m wide, sd across rg at S end.
T23D	736	627085	527	67	tr	sl	200	6.0	010	ic			W				Wn-c	tr 7m wide. Slopes S.fg drainage.
T23D	737	628087	527	67	sc	sl	300	6.0	010	ic			W				K-O	sc faces E. Parallels W side of tr at 736.
T23D	738	629088	527	67	sc	sl	100	3.0	010	ic			W				K-O	
T23D	739	630089	527	67	tc	sl	175		010	ic							K-O	Unstable hillside. Many springs.
T23D	740	628088	527	67	us	sl	15	0.5	345	ic			SW				K-O	tr 1m wide.
T23D	741	628089	527	67	us	sl	25	1.0	325	ic			SW				K-O	tr 3m wide.
T23D	742	628089	527	67	sd	rg											K-O	30m wide.
T23D	743	630094	526	1	sc	sl	150	1.0	345	ic			SW				K-O	
T23D	744	630094	526	1	sd	rg			010	ic			W	2.0			K-O	sd 12m wide.
T23D	745	631096	526	1	sc	sl	75	2.0	045	ic			W				K-O	
T23D	746	630097	526	1	sc	sl	20	3.0	045	ic			E				K-O	
T23D	747	632097	526	1	sc	sl	75	1.0	045	ic			E				K-O	
T23D	748	632092	526	1	sc	sl	50	1.5	005	ic			E				Wn-c	
T23D	749	632092	526	1	us	sl	25	1.0	000	ic			W				Wn-c	tr 2m wide.
T23D	750	633094	526	1	us	sl	50	2.0	000	ic			E				Wn-c	tr 1-3m deep cut by farm track.
T23D	751	633094	526	1	sc	ag	45	1.0	010	ic			E				H	Swamp on W side.
T23D	751a	634096	526	1	us	sl	15	1.0	020	ic			E				Wn-c	tr 1.5m wide.
T23D	752	633098	526	1	sc	sl	60	2.0	005	ic			E				K-O	Spring and swamp on W side.
T23D	753	633098	526	1	sd	rg	10	5.0	005				E				K-O	sd 10m wide.
T23D	753a	633098	526	1	us	sl	30	1.0	005	ic			E				K-O	Unstable hillside - slump.
T23D	754	633099	526	1	us	sl	50	5.0	000	ic			E				K-O	Swamp and spring on W side.
T23D	755	633099	526	1	us	sl	30	4.0	000	ic			E				K-O	Old stream gap between 754 and 755 rejuvenated?
T23D	756	634099	526	1	us	sl	50	2.0	000	ic			E				K-O	Present stream gap at S end of 756.
T23D	757	635101	526	1	us	sl	23	1.0	000	ic			E				K-O	tr 3m wide.
T23D	758	635103	526	1	us	sl	30	1.0	005	ic			E				K-O	tr 2m wide.
T23D	759	636106	526	1	sc	ag	150	6.0	005	ic			W				H	2m ag gravels deposited on TS.
T23D	760	638110	525	74	sc	ag	20	2.0	020	ic			W					
T23D	761	638110	525	74	us	ag	5.0	0.5	015	ic			W				H	tr 2m wide.
T23D	761a	639111	525	74	us	sl	25	1.5	015	ic			W				H	sc faces E.
T23D	762	640112	525	74	sc	sl	90	1.0	020	ic			E				H	Swamp with channel cut through sc.
T23D	763	641112	525	74	pr	sl	55	2.0	035	ic			E				K-O	Inset F.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	763	641112	525	74	us	sl	20	0.5	035	ic			E				K-O	Inset F.
T23D	763	642113	525	74	us	sl	35	1.0	035	ic			E				K-O	Inset F.
T23D	763	641113	525	74	us	sl	55	1.0	040	ic			E				K-O	Inset F.
T23D	763	642113	535	74	us	sl	20	0.5	035	ic			E				K-O	Inset F.
T23D	763	642113	525	74	us	sl	40	1.0	035	ic			E				K-O	Inset F.
T23D	763	642114	525	74	us	sl	33	0.5	035	ic			E				K-O	Inset F. Strike swings to 010 at S end.
T23D	764	643115	525	74	us	sl	6.0	1.0	015	ic			E				K-O	tr 2m wide.
T23D	765	644117	525	74	sc	ag	100	4.0	010	ic			W	4.0			H	
T23D	766	644117	525	74	us	ag	10	2.0	010	ic			E				H	Swamp on W side.
T23D	767	645118	525	74	us	sl	50	2.0	040	ic			E				K-O	Inset G.
T23D	767	645118	525	74	us	sl	25	8.0	020	ic			E				K-O	Inset G.
T23D	767	645118	525	74	us	sl	50	1.0	025	ic			E				K-O	Inset G.
T23D	767	645118	525	74	us	sl	45	0.5	020	ic			E				K-O	Inset G.
T23D	767	646119	525	74	us	sl	140	1.0	030	ic			E				K-O	Inset G.
T23D	767	646119	525	74	us	sl	90	1.0	030	ic			E				K-O	Inset G.
T23D	768	647121	525	74	us	sl	15	1.0	030	ic			E				K-O	
T23D	769	649124	524	69	sc	ag	40	1.0	010	ic			E				H	Eroded at S end.
T23D	770	649124	524	69	sc	ag	90	3.0	010	ic			W	3.0			H	Unstable slope with springs at N end.
T23D	771	650126	524	69	us	sl	120	2.0	010	ic			E				K-O	tr 4m wide.
T23D	772	651127	524	69	os				010	ic			E		d	30	K-O	
T23D	773	651127	524	69	sc	ag	15	2.0	010	ic			E	2.0			H	
T23D	774	652127	524	69	us	sl	15	0.5	030	ic			E				K-O	Inset H.
T23D	774	652127	524	69	us	sl	22	0.5	035	ic			E				K-O	Inset H.
T23D	774	652127	524	69	us	sl	25	1.5	040	ic			E				K-O	Inset H.
T23D	774	652127	524	69	us	sl	7.0	1.5	035	ic			E				K-O	Inset H.
T23D	774	652127	524	69	us	sl	40	1.0	020	ic			E				K-O	Inset H.
T23D	775	652127	524	69	us	sl	50	2.0	030	ic			E				K-O	sc faces W.
T23D	776	653129	524	69	us	sl	100	2.0	000	ic			E				K-O	sc faces W.
T23D	777	654129	524	69	us	sl	40	1.0	020	ic			E				K-O	sc faces W.
T23D	778	654132	524	69	sc	sl	675		005	ic			W				K-O	
T23D	779	654131	524	69	us	sl	50		005	ic			E				K-O	
T23D	780	658138	524	69	sh	sl	250	20	005	ic			E				K-O	Possible offset stream.
T23D	781	661143	523	74	sh	sl	125	20	005	ic			E				K-O	Partially eroded by stream.
T23B	296	670155	523	74	sh	sl	250	10	025	ic			E				K-O	Partially eroded by stream.
T23B	297	673159	523	74	sh	sl	125	10	025	ic			E				K-O	
T23B	298	677155	523	76	us	sl	125		030	ic			E				K-O	
T23B	299	681170	523	76	us	sl	125		030	ic			E				K-O	
T23B	300	684175	522	70	us	sl	125		030	ic			E				K-O	
T23B	301	691185	522	70	us	sl	250		020	ic			E				K-O	Unstable hillside.
T23B	302	698198	521	71	us	sl	60	10	010	ic			E				K-O	tr 5m wide - shallows
																	K-O	towards north.

BEAGLEY ROAD FAULT

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T24B	271	553989	235	53	us	sl	75	2.0	060	ic			NW				Wn	Swamp within tr
T24B	272	559991	235	53	sc	sl	100		050	ic			NW				Wn	Subdued sc faces SE
T24B	273	560991	235	53	sc	sl	75		050	ic			NW				Wn	Subdued sc faces SE
T24B	274	561991	235	53	sc	sl	50	2.0	050	ic			NW				Wn	Subdued sc faces SE
T24B	275	553989	235	53	us	sl	75	2.0	060	ic			NW				Wn	Swamp within tr on SE side
T24B	276	553989	235	53	sc	sl	35	2.0	060	ic			SE?				Wn	Flat bench 8m wide on SE side of NW facing sc.
T24B	277	554989	235	53	us	sl	35	6.0	060	ic			NW				Wn	Sc on NW side of tr faces SE. Swampy ground.
T24B	278	554989	235	53	us	sl	55	1.0	060	ic			NW				Wn	Sc on NW side of tr faces SE. Swampy ground.
T24B	279	555989	235	53	us	sl	50	4.0	060	ic			NW				Wn	Sc on NW wide of tr faces SE. Swampy ground.
T24B	280	557989	235	53	sc	sl	35	2.0	050	ic			NW				Wn	Subdued sc faces SE
T24B	281	560990	235	53	sc	sl	45	4.0	050	ic			SE?				Wn	Sc may be sh as stream is diverted at right angles to its original course. Possible dextral displacement towards SW.
T24B	282	562991	235	53	sc	sl	60	6.0	050	ic			NW				Wn	ds.
JAMES HILL ROAD FAULT																		
T24B	264	565998	234	17	sc	sl	40	3.0	045	ic			NW				Wn	Sc faces SE
T24B	265	566998	234	17	sc	sl	130	6.0	045	ic			NW				Wn	Sc faces SE
T24B	266	568999	234	17	sc	ag	10	2.0	045	ic			NW	2.0			H	Sc faces SE. Displaces te surface.
T24B	267	568999	234	17	sc	ag	85	2.0	045	ic			NW	2.0			H	Sc faces SE. Displaces te surface.
T24B	268	569999	234	17	sc	ag	15	2.0	045	ic			NW	2.0			H	Sc faces SE. Displaces te surface.
T24B	269	571999	234	17	us	sl	100	4.0	045	ic			NW				Wn	Sc faces SE. Swampy ground.
T24B	270	571000	234	17	us	sl	40	0.5	045	ic			NW				Wn	Subdued tr peters out towards the NE.
COPPERMINE ROAD FAULT																		
T24B	294	562999	234	17	sc	sl	60	2.0	060	ic			NW				Wn	Sc broken by stream.
T24B	295	563999	234	17	sc	sl	90	2.0	060	ic			NW				Wn	Sc broken by stream.
T23D	791	565000	234	17	sc	sl	80	2.0	020	ic			NW				Wn	Sc faces east. 2m wide bench on upthrown side.
T23D	792	565001	234	17	us	sl	3.0	0.3	020	ic			SE				Wn	Tr 0.5m wide on left bank of small stream.
T23D	793	566002	234	17	sc	sl	130	2.0	020	ic			NW				Wn	Subdued in outline.
T23D	794	566003	234	17	us	sl	40	0.5	350	ic			SE				Wn	Curves around hillside. 2m wide tr.
T23D	795	566003	234	17	us	sl	25	1.0	030	ic			SE				Wn	Tr 2m wide.
T23D	796	567004	234	17	sc	sl	40	3.0	015	ic			NW				Wn	Sc faces E. Subdued in outline.
T23D	797	568005	234	17	sc	sl	120	2.0	015	ic			NW				Wn	Sc faces E. Subdued in outline.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	798	568006	234	17	sc	sl	25	1.0	015	ic			NW				Wn	Subdued E facing sc.
T23D	799	569007	234	17	sc	sl	120	3.0	015	ic			NW				Wn	2m wide bench on upthrown side of sc.
T23D	800	570007	234	17	sc	sl	70	3.0	015	ic			NW				Wn	Spring at N end.
T23D	801	571008	234	17	sc	ag	180	2.0	035	ic			NW				H	ds at S end of sc.
T23D	802	572009	234	17	sc	ag	175	1.0	035	ic			NW				H	
T23D	803	573009	234	17	sc	ag	45	1.0	025	ic			NW				H	
T23D	804	575011	234	17	sc	ag	200	1.0	025	ic			NW				H	
T23D	805	576011	234	11	sc	ag	200	1.0	040	ic			NW				H	Sc cut by stream.
MANGARAWA FAULT																		
T24B	283	558966	235	54	sc	ag	200	2.0	040	ic			NW	3.5			H	Inset D. Sc peters out towards SW end.
T24B	284	560966	235	54	sc	ag	30	1.0	055	ic			NW				H	Inset D. Sc modified by recent channel realignment scheme.
T24B	285	561967	235	54	sc	sl	175	2.0	055	ic			NW				Wn	Inset D. Springs on line of fault tc.
T24B	286	562967	235	54	sc	sl	80	2.0	055	ic			NW	7.2			Wn	Inset D. Line of springs at base of sc.
T24B	287	564967	235	54	sc	sl	130	3.0	055	ic			NW	7.2			Wn	Inset D. Subdued SE facing sc.
T24B	288	565967	235	54	us	sl	45	1.0	055	ic			SE?				Wn	Inset D. ds issuing from spring.
T24B	289	567967	235	54	sc	rg	75	6.0	055	ic			NW	15.6			Wn	Inset D. Two parallel scarps each showing similar amounts of vertical displacement.
T24B	290	568968	235	54	sc	sl	250	8.0	055	ic			NW				Wn	Inset D. Sc faces SE.
T24B	291	570968	235	54	sc	sl	150	8.0	055	ic			NW	15.6			Wn	Inset D. Probable 75m stream offset at N end of sc.
T24B	292	573968	235	54	sc	rg	175	8.0	055	ic			NW				Wn	Inset D. Sc faces SE. Faint tc to NW of sc.
T24B	293	578971	235	54	md		300	8.0	030	ic			NW				H	Inset D. Steep SE facing sc along E side of md.
WHARITE FAULT																		
T24B	255	539998	234	15	us	sl	5.0	0.5	035	ic			SE				Wn	W facing sc on E side of tr.
T24B	256	539998	234	15	us	sl	5.0	0.5	035	ic			SE				Wn	W facing sc on E side of tr. Stream cuts through sc between localities 256 and 257.
T24B	257	540998	234	15	us	sl	20	1.0	035	ic			SE				Wn	W facing sc on E side of tr.
T24B	258	541998	234	15	us	sl	25	2.0	020	ic			SE				Wn	W facing sc on E side of tr.
T24B	259	541999	234	15	us	sl	20	4.0	035	ic			SE				Wn	W facing sc on E side of tr.
T24B	260	541999	234	15	us	rg	2	0.5	035	ic			SE				Wn	tr 10m wide.
T23D	601	540000	234	15	us	sl	20	1.0	010	ic			E				Wn	Stream issues from spring.

NZMS 260	Loc	Gr Ref	Air Photo		Feature				Tr Fold		F Plane		Vert		Horz		FM age	Remarks and Ref
			Run	No	Na	Dp	Lm	HDm	Str	Lk	Dir	Am	Up	Amm	S	Amm		
T23D	602	540000	234	15	us	sl	15	1.0	010	ic			E				Wn	Stream issues from spring.
T23D	603	541001	234	15	us	sl	25	4.0	010	ic			E				Wn	
T23D	604	541002	234	15	sc	sl	70	0.5	350	ic			E				Wn	Swamp on W side of sc.
T23D	605	541004	234	15	sc	sl	60	0.5	350	ic			E				Wn	Swamp on W side of sc.
T23D	606	542006	234	15	sc	sl	10	4.0	350	ic			W				Wn	4m wide flat bench on W side of sc.
T23D	607	542007	234	15	us	sl	15	4.0	350	ic			E				Wn	W facing sc on E side of tr.
TOTARA FAULT																		
T24A	001	452940	236	34	sd	rg	30	10	010	ic			E				Wc	Contact between Wc and K-O.
T24A	002	453942	236	34	sd	rg	15	10	010	ic			E				Wc	Contact between Wc and K-O.
T24A	003	475946	236	34	sd	rg	30	10	010	ic			E				K-O	
T24A	004	465956	236	34	fg		500		010	ic							K-O	gr at northern end.
T24A	005	459944	236	35	sd	rg	20	10	050	ic			SE				K-O	
T24A	006	461945	236	35	sd	rg	20	5	050	ic			SE				K-O	
T24A	007	462945	236	35	sd	rg	20	5	050	ic			SE				K-O	
T24A	008	464945	236	35	us	sl	105	1	050	ic			NW?				K-O	Swampy 1m wide.
T24A	009	465946	236	35	sc	rg	25	6	050	ic			SE				K-O	Subdued scarp
CROSS ROAD FAULT																		
T24A	010	462917	237	36	sc	sl	100	6	025	ic			W				K-O	Inset E. Subdued scarp faces east.
T24A	011	463918	237	36	tr	sl	40	2	025	ic			W				K-O	Inset E. Spring and swamp.
T24A	012	464918	237	36	sc	sl	45	10	025	ic			W				K-O	Inset E. Sc faces E.
T24A	013	464918	237	36	sc	sl	30	6	025	ic			W				K-O	Inset E. Springs, local subsidence.
T24A	014	468922	237	36	sc	sl	200	10	025	ic			W				K-O	Inset E. Sc height varies from 6m to 15m. Local subsidence.
T24A	015	472925	237	37	sc	sl	200	4	050	ic			NW				K-O	Inset E. Line of springs.
T24A	016	475928	237	37	sc	sl	75	8	050	ic			SE				K-O	Inset E. Face NW.
T24A	017	477928	237	37	sc	sl	200	8	060	ic			SE				K-O	Inset E. Face NW.
T24A	018	478928	237	37	tc	sl	125		035	ic			NW				K-O	Inset E. Flat surface between 018 and 019. 8m wide.
T24A	019	479929	237	37	sc	sl	125	6	025	ic			NW				K-O	Inset E. gr at northern end.
MISCELLANEOUS FAULT TRACES																		
T23D	782	578031	233	19	us	sl	50	2.0	035	ic			NE				Wc	
T23D	783	579031	233	19	us	sl	45	2.0	035	ic			NE				Wc	Tr 10m wide.
T23D	784	580031	233	19	us	sl	50	1.5	035	ic			NE				Wc	
T23D	785	580032	233	19	us	sl	45	1.5	035	ic			NE				Wc	Tr 7m wide.
T23D	786	582033	233	19	us	sl	5.0	0.5	035	ic			NE				Wc	Tr 4m wide.
T23D	787	583034	233	19	us	sl	30	1.0	020	ic			NE				Wc	
T23D	788	593025	233	20	sh	sl	65	3.0	030	ic			NE		d	50	Wn	6m wide swamp on upstream side.
T23D	789	594026	233	20	tc	sl	200		030	ic							Wn	Destroyed by cultivation.

APPENDIX Vb: Radiocarbon dates collected from study area

Fossil record number	Metric grid reference	NZ radiocarbon laboratory sample no.	Old $T^{1/2}$	New $T^{1/2}$	New $T^{1/2}$ corrected for secular affect	Wood identification	Collectors
T23/f1	T23/698191	NZ 4314	11 800 ± 150	12 150 ± 150	-	*Griselinia*, most probably *G. littoralis*	M Marden P R Stephens V E Neall
T23/f2	T23/697189	NZ 4547	790 ± 60	810 ± 70	770 ± 60	*Pittosporum* sp.	C Hubbard M Marden
T24/f12	T24/517946	NZ 4651	12 900 ± 200	13 300 ± 200	-	*Metrosideros robusta*	V E Neall M Marden
T24/f15	T24/476958	NZ 5231	3 430 ± 80	3 530 ± 80	3 700 ± 90	*Dacrydium cupressinum*	C Lees
T24/f17	T24/517945	NZ 5320	10 350 ± 100	10 650 ± 150	-	*Leptospermum* sp.	M Marden V E Neall
T23/f19	T23/658168	NZ 5274	8 160 ± 120	8 400 ± 130	-	Compositae wood	C Lees
T24/f19	T24/517946	NZ 5591	12 650 ± 150	13 000 ± 150	-	-	V E Neall G Lensen M Marden
T23/f20	T23/607045	NZ 5592	376 ± 56	387 ± 57	442 ± 57	-	G Lensen M Marden
N144/f589	N144/386516 T23D/580057	NZ 3879	680 ± 60	700 ± 60	680 ± 40	*Podocarpus spicatus*	P R Stephens

APPENDIX VI

Changes in the State of Activity of Large-Scale, Deep-Seated Mass Movement Features in the southern Ruahine Range between the years 1946-49 (Map 5) and 1974-78 (Map 6).

Symbols:

e sl	-	earth slump
e sld	-	earth slide
r sl	-	rock slump
r t b	-	ridge-top bench
r t d	-	ridge-top depression
r t s	-	ridge-top scarp

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	Aerial Photograph 1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	State of Activity 1974 - 1978
T23B	1	675265	Piripiri	Wharite	r sl	SN230	519	65	SN3721	C	12	Slump outline visible. Total forest cover except for small areas of bare ground at toe of slump. Ridge top depression and scarp above slump outline.	Not moved since 1946. Debris slides around lateral scarps. Increase in debris slide activity at toe of slump. Scarp and depression unchanged.
T23B	2a	678256	Piripiri	Wharite	r sl	SN230	519	65	SN3721	C	12	Slump outline visible. Thick forest cover over much of slump area. 3 small debris slides near headwall. Many slides along toe slope are unvegetated. Former small slump features and debris slides are tussock covered.	Not moved since 1946. Debris slide scars still unvegetated. Marked increase in area of bared ground along toe slope largely due to stream bank undercutting.
T23B	2b	680260	Piripiri	Wharite	r t s	SN230	519	65	SN3721	C	12	N.E. striking scarp with pond at base of eastern dowhthrown side. Very small area of bared ground on scarp face.	No change.
T23B	2c	682248	Piripiri	Wharite	r t d	SN230	519	65	SN3721	C	12	Scarp and depression visible. No bared ground.	Few small debris slide scars on NE side of scarp and depression.
T23B	3	695240	Pohangina	Wharite	r sl	SN230	520	61	SN3721	D	13	Slump outline visible. Moved mass is forested. Scrubby vegetation around headwall scarp. No bare ground in headwall area or along toe slope.	Not moved since 1946. One small debris slide on toe slope.
T23B	4	678238	Pohangina	Wharite	r sl	SN230	520	61	SN3721	D	12	No slump outline. Ridge top scarp visible. Total area forest covered.	Slump outline visible. Debris slide scars around head wall and along toe slope. Headwall scars colonised by shrubby vegetation. Ridge top scarp shows no additional vertical movement.
T23B	5a	680233	Pohangina	Wharite	r sl	SN230	520	61	SN3721	D	12	No slump outline. No bare ground. Total area forest covered. Ridge top scarps not noticeable.	Slump outline visible. No toe erosion. Headwall area marked by small debris slides. Ridge top scarps show noticeable vertical displacement.
T23B	5b	682231	Pohangina	Wharite	r sl	SN230	520	61	SN3721	D	12	Same as 5a.	Slump outline visible. Vegetation undusturbed. No headwall or toe erosion. Ridge top scarps show noticeable vertical displacement.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	Aerial Photograph 1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	State of Activity 1974 - 1978
T23B	6	681225	Pohangina	Wharite	r sl	SN230	520	61	SN3721	D	12	Slump outline visible. Slumped mass is forested. Headwall and lateral scarps are covered in shrubby species. Headwall scarp outlined by recent debris slide scars.	Not moved since 1946. Some debris slide scars revegetated. Others remain as bare ground.
T23B	7	648234	Pohangina	Wharite	r sl	SN230	520	58	SN3721	D	10	Slump outline visible. Forest covered. No debris slide scars in headwall or along toe slope.	Not moved since 1946. Increase in debris slide scars in headwall and lateral scarp areas and also along toe slope.
T23B	8	650233	Pohangina	Wharite	r sl	SN230	520	58	SN3721	D	12	Slump outline visible. Forest covered. Large debris slide scar along toe slope. Small scar near headwall.	Not moved since 1946. Increase in debris slide scar area at toe of slope to include most of the area of the slump.
T23B	9	693217	Pohangina	West Tamaki	r t s	SN230	521	70	SN3721	E	13	Short linear scarp facing south. No slump structure on valley slope below. Densely vegetated with leatherwood.	No change.
T23B	10	695208	West Tamaki	West Tamaki	r t s	SN230	521	70	SN3721	E	13	Short linear scarp facing to NE. Densely vegetated with leatherwood. No slump structure on valley slope below.	No change.
U23A	11	703207	West Tamaki	West Tamaki	r sl	SN230	521	71	SN3721	E	13	Outline difficult to discern beneath forest vegetation. Scrub covered debris avalanche scars extend down length of valley slope. Small debris slides around headwall scarp. Large area of bared ground along stream bank at toe of slope.	Not moved since 1946. No further slope activity in headwall area. Increase in bared ground along toe slope.
U23A	12	708202	West Tamaki	West Tamaki	rtd/rts	SN230	521	71	SN3721	E	13	Scarp and depression visible but are heavily vegetated with forest.	No change.
T23B	13	674216	Pohangina	Wharite	r sl	SN230	521	68	SN3721	E	12	Outline of slump visible. Forest cover largely destroyed - vegetated by shrubby species. Small debris slide scars at toe to slump adjacent to stream bed.	Not moved since 1946. Shrubby vegetation predominates. Small areas of toe erosion. No bare ground in headwall area of slump.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph						State of Activity	
						1946 - 1949			1974 - 1978			1946 - 1949	1974 - 1978
						Series	Run	Photo No.	Series	Run	Photo No.		
T23B	14	673215	Pohangina	Wharite	r sl	SN230	521	68	SN3721	E	12	Slump outline visible. Most of slump surface clothed in shrubby vegetation. Few large emergenttrees in central part of slump. No bare ground. Small toe erosion debris slide scars.	Not moved since 1946. Difference in vegetation cover between slump scar and surrounding slope very obvious. No bare ground. No additional toe erosion.
U23A	15	704180	East Tamaki	West Tamaki	r t d	SN230	522	72	SN3721	F	13	Depression visible. Pastured. No bare ground.	No change.
U23A	16	700180	East Tamaki	West Tamaki	r t d	SN230	522	71	SN3721	F	13	Depression visible. Partially in pasture and partially in bush. No bare ground.	No change. May of of tectonic origin if trace of Wellington Fault passes along this valley.
T23B	17	678188	West Tamaki	West Tamaki	r sl	SN230	522	68	SN3721	F	11	Slump outline visible. Densely clothed in leatherwood vegetation. No old erosion scars, no bared ground. Headwall scarp is suggestive of two slump features side by side.	Not moved since 1946. No bare ground on east facing slump. Bare ground opened up on north facing slump. Considerable debris slide activity in stream channels immediately below slump features.
T23B	18	675181	West Tamaki	West Tamaki	r t b	SN230	522	67	SN3721	F	11	Bench scarps visible below very steep scarp parallel to ridge. Benches may be part of large slump structure at the top of this catchment. No headwall erosion scars. Large debris avalanche known as 'Catspaw' lies next to bench features.	Not moved since 1946. No bare ground. No old erosion scars. Slight increase in debris slide activity on slopes adjacent to benches. Scars confined to waterways.
T23B	19	661168	Rokaiwhana	West Tamaki	r t b	SN230	523	74	SN3721	F	11	Benches difficult to pick out beneath dense leatherwood vegetation. No bare ground.	Not moved since 1946. No bare ground. Increase in debris slide activity along waterways beneath bench features.
T23B	20	643179	Makawakawa	Wharite	r sl	SN230	522	63	SN3721	F	9	No slump outline present. Heavy forest vegetation. No bare debris slide scars.	Headwall scarp discernible - no bare ground. Gully and stream dissection have caused extensive debris slide activity that has since revegetated. Small debris slide scars of recent origin adjacent to stream channels.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	1974 - 1978
T23B	21	640178	Makawakawa	Wharite	r sl	SN230	522	63	SN3721	F	9	No slump outline visible. Heavily forested. No toe erosion.	Headwall scarp visible. Much debris slide activity around head-wall scarp and along toe of slope.
T23B	22	629185	Makawakawa	Wharite	r sl	SN230	522	62	SN3721	F	9	Slump outline visible. Heavily forested. Small debris slides at toe of slump.	Not moved since 1946. Increase in debris slide activity around head-wall scarp and along toe of slope.
T23B	23	622185	Makawakawa	Wharite	r sl	SN230	522	62	SN3721	F	9	Slump outline visible. Heavily forested. Headwall scarp has mix-ture of tree and scrub vegetation. No bare ground. Small debris slides at toe of slump.	Not moved since 1946. Headwall scarp is largely bared ground. Large debris slides around three sides of slump along toe slope adjacent to streams.
T23B	24	596193	Makawakawa	Wharite	r sl	SN230	522	60	SN3721	F	7	Slump outline visble. Upper half grassed; lower half in bush. Numerous small lateral scarps within slump outline. No bare ground. Small area of toe slope lost bush covering due to erosion - scrub covered.	Not moved since 1946. No bare ground in headwall area. Scarps still present. Fully erosion at toe of slump has caused much debris slide activity.
T23B	25	600184	Makawakawa	Wharite	r sl	SN230	522	60	SN3721	F	7	Slump outline visible. Heavily forested. No bared ground in head-wall area or along toe slope.	No change.
T23B	26	606175	Makawakawa	Wharite	r sl	SN230	522	60	SN3721	F	7	Slump outline visible. Total area is heavily forested apart from narrow debris avalanche scars - grassed. No toe erosion.	Not moved since 1946. Large areas of exposed ground in middle of slump. Large recent debris slide at eastern end.
T23B	27	604174	Makawakawa	Wharite	r sl	SN230	522	60	SN3721	F	7	Near vertical linear headwall scarp - could be a rockfall. Scarp is tussock covered. Slumped mass vegetated with trees and scrub. Little bare ground at toe of slump.	Not moved since 1946. Greater area of bared ground on toe slope at western end. Headwall scarp un-changed.
T23B	28	655152	Rokaiwhana	West Tamaki	r sl	SN230	523	72	SN3721	G	10	Slump outline visible. Dense forest vegetation down centre of slump. Margins delineated by shrubby species. No bare scars - no toe erosion. Very steep headwall scarp.	Not moved since 1946. Forest vege-tation and shrubby cover appear to have thinned-out. No bare ground in headwall area. Small debris slide scar along toe slope.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949			Aerial Photograph 1974 - 1978			State of Activity 1946 - 1949	State of Activity 1974 - 1978
						Series	Run	Photo No.	Series	Run	Photo No.		
T23B	29	650150	Rokaiwhana	Wharite	r t d	SN230	523	72	SN3721	G	10	Depression not obvious beneath dense leatherwood vegetation. Parallels ridge above suspected but unnumbered slump feature.	No change.
T23D	30	656128	Rokaiwhana	West Tamaki	r sl	SN230	524	69	SN3721	G	10	Slump outline visible. Pastured hummocky ground surface. No bare ground. No toe erosion.	No change.
T23D	31	641139	Rokaiwhana	West Tamaki	r t d	SN230	524	61	SN3721	G	9	Depression visible despite dense leatherwood vegetation.	No change.
T23D	32a	641134	Rokaiwhana	West Tamaki	r sl	SN230	524	67	SN3721	G	9	Outline not obvious. Shrubby vegetation dominant. Grassed debris avalanche trails cut through scrub. No bare ground.	Recent scouring at toe of slump.
T23D	32b	639137	Rokaiwhana	West Tamaki	r sl	SN230	524	67	SN3721	G	9	Slump outline visible. Upper part is forested. Toe slope covered in shrubby vegetation. One large area of bare ground on toe slope.	Not moved since 1946. Debris slide scar on toe slope increased in area.
T23D	33	639132	Rokaiwhana	West Tamaki	r sl	SN230	524	67	SN3721	G	9	Outline visible. Forested central portion with shrubby vegetation along lateral and headwall scarp. Considerable bared ground along toe slope.	Not moved since 1946. Increase in area of debris slide activity across width of slump along toe slope.
T23D	34	643128	Rokaiwhana	West Tamaki	r sl	SN230	524	67	SN3721	G	9	Slump outline visible. Mostly covered in shrubby vegetation - very few trees. Very small areas of bared ground along lateral margin of slump. No headwall or toe slope erosion.	Not moved since 1946. Two large debris slide scars on mid slope and 3 large scars across toe of slump.
T23D	35	635129	Otamaraho	West Tamaki	r sl	SN230	524	67	SN3721	G	9	Outline not obvious. Slump is covered by shrubby vegetation along old debris avalanche scars with forest vegetation in between. No bare ground on slump. No toe erosion.	Not moved since 1946. Marked increase in debris slide activity along toe of slump.
T23D	36	633099	Mangapuaka	West Tamaki	r sl	SN230	526	1	SN3721	G	9	Outline visible. Pastured. No bare ground. Fault trace across centre of slump.	No change.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	1974 - 1978
T23D	37	621121	Mangapuaka	West Tamaki	r sl	SN230	525	71	SN3721	H	8	Slump outline visible. Headwall area forested. Shrubby vegetation along toe slope. No headwall or toe erosion. Gully of centre of slump has large area of bare ground at its head.	Not moved since 1946. Extensive debris slide activity in scrub covered toe slope area. Slides extensively developed along sides of gully.
T23D	38	617112	Mangapuaka	West Tamaki	r sl	SN230	525	71	SN3721	H	8	Slump outline visible - enhanced by areas of shrubby vegetation along margins of slump. Central part heavily forested. Small debris slides at toe of slump.	Not moved since 1946. Debris slide activity along toe slope has increased area of bare ground.
T23D	39	613112	Mangapuaka	West Tamaki	r sl	SN230	525	71	SN3721	H	8	Slump outline visible. Much of slump covered in shrubby vegetation - few large trees remain. No bare ground.	Not moved since 1946. Debris slide activity along toe slope has increased area of bare ground.
T23D	40	610112	Mangapuaka	West Tamaki	r sl	SN230	525	71	SN3721	H	8	Slump outline visible. Small revegetated debris slide scars. No bare ground.	Not moved since 1946. Increase in debris slide activity along margin of slump.
T23D	41	606110	Mangapuaka	West Tamaki	r t d	SN230	525	69	SN3721	H	8	Depression visible. Densely vegetated. No bare ground.	No change.
T23D	42	606127	Mangapuaka	Wharite	r t d	SN230	525	69	SN3721	H	8	Depression visible. Densely vegetated. No bare ground.	No change.
T23D	43	601122	Mangapuaka	Wharite	r t d	SN230	525	69	SN3721	H	8	Depression visible. Densely vegetated. No bare ground.	No change.
T23D	44	599119	Mangapuaka	Wharite	r t s	SN230	525	69	SN3721	H	8	Scarp visible. Densely vegetated. No bare ground.	No change.
T23D	45	598116	Mangapuaka	Wharite	r t s	SN230	525	69	SN3721	H	8	Scarp visible. Densely vegetated. No bare ground.	No change.
T23D	46	612105	Otamarahu	West Tamaki	r sl	SN230	526	66	SN3721	H	9	Slump outline visible. Central portion is forested. Headwall scarp and lateral scarp dominated by shrubby vegetation. Small areas bare ground along lateral margin and toe slope.	Not moved since 1946. Increased debris slide activity along toe slope.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph						State of Activity	
						1946 - 1949			1974 - 1978			1946 - 1949	1974 - 1978
						Series	Run	Photo No.	Series	Run	Photo No.		
T23D	47	600100	Mangapukakakahu	West Tamaki	r t d	SN230	526	65	SN3721	I	6	Depression heavily vegetated. No bare ground.	No change.
T23D	48	603099	Mangapukakakahu	West Tamaki	r sl	SN230	526	65	SN3721	I	7	Slump outline visible. Mostly shrubby vegetation. No headwall erosion. Small debris slide scars along toe slope.	Not moved since 1946. Slight increase in debris slide activity on toe slope.
T23D	49	601096	Mangapukakakahu	West Tamaki	r sl	SN230	526	65	SN3721	I	7	Outline not obvious. Mostly shrubby vegetation. Large areas of bare ground at base of slump.	Not moved since 1946. Total width of toe slope is bare ground - extends for considerable distance upslope.
T23D	50	582101	Oruakeretaki	Wharite	rtd/rts	SN230	526	62	SN3721	H	6	Depression and scarp visible but covered in heavy vegetation. No bare ground.	No change.
T23D	51	569094	Mangatuatou	Wharite	r t d	SN230	526	61	SN3721	I	4	Depression visible. Heavily vegetated. No bare ground.	No change.
T23D	52	557109	Ohinetapu	Wharite	r sl	SN230	526	60	SN3721	H	5	Slump outline visible. Heavily forested. Two large debris slides at toe of slump.	Not moved since 1946. Debris slides partially revetated but still much bare ground. Small slide in headwall area.
T23D	53	574087	Raparapawai	Wharite	r sl	SN230	527	61	SN3721	I	4	Slump outline not obvious. Ridge-top-depression around head of catchment. Stream draining this small catchment is lined with re-vegetated and bared debris slides. Much of catchment heavily forested. No headwall erosion. Debris avalanches and slides along toe slope.	Not moved since 1946. Depression still visible. Greater stream dissection with increased debris slide activity up centre of slump. Increased toe erosion with much bared ground.
T23D	54	592070	Sth Oruakeretaki	Wharite	r sl	SN230	527	63	SN3721	I	6	Outline of slump visible - obvious signs of rotation. Moved mass heavily forested. Headwall area clothed in shrubby vegetation. No headwall erosion. Few small areas of bared ground along lateral margin. Debris slides along toe of slump.	Not moved since 1946. No headwall or lateral scars. Increased debris slide activity along toe slope.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	Aerial Photograph 1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	State of Activity 1974 - 1978
T23D	55	590072	Sth Oruakeretaki	Wharite	r sl	SN230	527	63	SN3721	I	6	Slump outline visible. Dense forest vegetation all around slump but slumped mass is clothed in shrubby vegetation with only a few tall trees. No headwall erosion. Debris slide scars along toe slope.	Not moved since 1946. Increased debris slide activity along toe of slump.
T23D	56	570076	Raparapawai	Wharite	r t d	SN230	527	61	SN3721	I	4	Depression visible. Heavily vegetated. No bare ground. At base of slope are revegetated erosion	No change in the nature of the depression, however there is considerably more debris slide and avalanche activity at the base of this slope.
T23D	57	555083	No. 1 Line	Wharite	r sl	SN230	527	58	SN3721	I	3	Outline visible but doesn't give appearance of a slump feature. Moved mass heavily forested. Bare ground around margin and along toe slope.	Not moved since 1946. Marked increase in amount of bare ground within the slump particularly in headwall and along toe slope areas.
T23D	58	546089	No. 1 Line	Wharite	r sl	SN230	527	58	SN3721	I	3	Slump outline visible. Open canopy forest on slumped area contrasts with dense forest cover on adjacent slopes. No bare ground.	No change.
T23D	59	544088	No. 1 Line	Wharite	r sl	SN230	527	57	SN3721	I	3	Slump outline visible. Forest cover down centre of slump, shrubby vegetation around margins. Small headwall erosion scar. Unstable toe slope with extensive areas of bare ground.	Not moved since 1946. Debris slide in headwall area enlarged. Large areas of bared ground in toe slope area but doesn't appear to have enlarged.
T23D	60	560075	Raparapawai	Wharite	r t d	SN230	527	59	SN3721	I	4	Depression visible beneath dense leatherwood vegetation. No bare ground, or old erosion scars.	No change.
T23D	61	563071	Raparapawai	Wharite	r sl	SN230	527	60	SN3721	I	4	Outline of slump visible. Central part of slump heavily forested. Debris slide scars in headwall and lateral scarp areas. Small scar on toe slope at eastern end.	Not moved since 1946. Marked increase in bared ground around headwall region and on toe slope at eastern end. Central forested part of slump unchanged.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	Aerial Photograph 1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	State of Activity 1974 - 1978
T23D	62	570068	Raparapawai	Wharite	r sl	SN230	527	61	SN3721	I	4	Outline of slump visible. Mostly dense forest cover. Revegetated debris avalanche scars in toe area. Very little toe or headwall erosion.	Not moved since 1946. Debris avalanche scars show signs of opening up again, in the higher parts of the slump but are still tussock vegetated in lower parts. Headwall scarp opened up again.
T23D	63	555071	Raparapawai	Wharite	r t d	SN230	527	59	SN3721	I	4	Depression visible. No erosion scars. Densely vegetated.	No change.
T23D	64	553067	No. 2 Line	Wharite	r t d	SN230	528	58	SN3721	J	4	Depression not obvious through dense leatherwood vegetation.	No change.
T23D	65	558064	Manga-a-tua	Wharite	r sl	SN230	528	59	SN3721	J	4	Slump outline visible. Totally forest covered apart from narrow debris avalanche scars. No bare ground.	Not moved since 1946. Definite change in vegetation cover in head-wall area to shrubby species. No current headwall erosion. Gully erosion with debris slides at base of slump.
T23D	66	563062	Manga-a-tua	Wharite	r sl	SN230	528	59	SN3721	J	4	Slump outline visible. Predominantly shrubby vegetation. Little bare ground.	Not moved since 1946. Vegetation cover looks thinner. Bare ground in headwall and along toe slope.
T23D	67	560058	Manga-a-tua	Wharite	r sl	SN230	528	59	SN3721	J	4	Outline of slump vague. Forest cover on slumped mass. Shrubby vegetation around margins. Re-vegetated debris avalanche scars in headwall area. One patch of bared ground in headwall area. Large area of bare ground on toe slope.	Outline of slump more obvious but further movement of the slump is difficult to prove. Thinned out forest cover in centre of slump. Debris avalanche scars opened up. Toe erosion across width of slump.
T23D	68	565056	Manga-a-tua	Wharite	r sl	SN230	528	59	SN3721	J	4	Slump not obvious in outline. Forest vegetation on slumped mass. Shrubby vegetation around margins. Few areas of bared ground along headwall and margins of slump.	Outline visible. Forest in centre of slump still intact. Shrubby species dominate headwall and marginal areas. Bare ground in headwall region.
T23D	69	567042	Manga-a-tua	Wharite	r sl	SN230	528	59	SN3721	J	6	Slump outline not obvious. Pastured. Long narrow debris avalanche scars.	Outline vague but small headwall scarp has developed. Increased debris avalanche activity on higher slopes and slide scars on toe slope. Area within slump has been site of former large sized failure most of which have now grassed over.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	Aerial Photograph 1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	State of Activity 1974 - 1978
T23D	70	570040	Manga-a-tua	Wharite	r sl	SN230	528	59	SN3721	J	6	Slump outline not obvious. Pastured. Long narrow debris avalanche scars.	Same as 69 but with a large very deep gully up centre of slump lined with extensive debris slide scars.
T23D	71	578038	Manga-a-tua	Wharite	r t s	SN181	233	19	SN3721	J	6	Outline of scarp visible around head of catchment. Pastured.	No change.
T23D	72	546035	Coppermine	Wharite	r sl	SN181	233	16	SN5163	Q	5	No indication of slump outline or of any bare ground in this area. Thick forest vegetation.	Slump failed totally in August 1976. Large bare slip face with small area of trees near centre.
T23D	73	548032	Coppermine	Wharite	r sl	SN181	233	16	SN5163	Q	5	Slump outline visible. Covered in shrubby vegetation. One small area bare ground on toe slope. Forest vegetation above slump may be part of a larger slump feature that extends upslope from the failed area.	Total area covered in 1946 with shrubby vegetation collapsed - all bare ground. Forested upper part of slope still intact.
T23C	74a	500800	Te Awaoteatua	Plio-Pleist	e sl	SN230	527	53	SN3721	I	1	Headwall scarp visible with few small areas of bare ground around steepest part. Hummocky grass surface. Small area of toe erosion.	Slump not moved since 1946. Headwall scars partially revegetated. Areas of toe erosion visible in 1946 have healed. Signs of regeneration by shrubby species.
T23C	74b	495081	Te Awaoteatua	Plio-Pleist	e sl	SN230	527	53	SN3721	I	1	Very steep headwall scarp with areas of bare ground. Predominantly grassed surface with some shrubby regeneration. Slumped mass is in place at base of slope. Steep toe slope is unstable due to stream bank undercutting.	Not moved since 1946. Headwall scarp covered with shrubby vegetation. Some areas of bare ground visible in 1946 have healed. Bare ground present at eastern end of slump outline. Greater area of toe erosion than in 1946.
T23D	74c	504082	Te Awaoteatua	Plio-Pleist	e sl	SN230	527	53	SN3721	I	1	Two headwall scarps visible indicating differential displacement of slumped mass. Small area of bare ground along headwall scarp. Little toe slope erosion along stream bank. slumped mass largely in place at base of slope. Grassed surface.	Not moved since 1946. Much of slumped area is covered in regenerated shrubby vegetation. Very little evidence of bare ground.
T23D	74d	508082	Te Awaoteatua	Plio-Pleist	e sl	SN230	527	53	SN3721	I	1	Short, steep headwall scarp with few small areas of bare ground. Long, steep toe slope with erosion scars along stream bank due to stream undercutting. Slump mass in place at base of slope. Small	Not moved since 1946. Toe slope is largely covered in regenerated shrubby vegetation. No areas of bare ground either along the headwall scarp or along the toe slope.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	Run	Photo No.	Aerial Photograph 1974 - 1978 Series	Run	Photo No.	State of Activity 1946 - 1949	State of Activity 1974 - 1978
												scarp and depression across slumped mass indicate differential movement of the slumped mass. Surface is grassed.	
T23D	74e	506084	Te Awaoteatua	Plio-Pleist	e sl	SN230	527	53	SN3721	I	1	Steep lateral scarp along western side of slump outline. Smaller vertical displacement in headwall area. Grassed surface with no areas of bare ground visible. Slumped material totally removed.	Not moved since 1946. Some of slump area reverting to shrubby vegetation. Very little evidence of bare ground.
T23D	75	508092	Te Awaoteatua	Plio-Pleist	e sl	SN230	527	53	SN3721	I	1	Slump outline visible. Hummocky pastured surface. Multiple head-wall and lateral scarps may indi-cate several episodes of slumping.	Not moved since 1946. Few small headwall scars.
T23D	76	518093	No. 2 Line	Plio-Pleist	e sld	SN230	526	55	SN3721	I	1	Slide outline visible. Pastured surface. Very steep headwall and lateral scarp along western side. No toe erosion. Small debris slide scar on headwall.	Not moved since 1946. No bare ground - former scars healed.
T23D	77	515099	No. 2 Line	Plio-Pleist	e sl	SN230	526	55	SN3721	I	1	Slump outline visible. Very hummocky pastured surface. Toe slopes on two sides are very steep adjacent to stream channels - few small areas of bare ground on these steep slopes.	Not moved since 1946. Steep sided toe slopes revegetated with shrubby vegetation. No bare ground at all.
T23D	78	520096	No. 1 Line	Plio-Pleist	e sld	SN230	526	56	SN3721	I	1	Slide outline visible. Pastured. Bare ground along steep lateral scarp and along toe slope adjacent to stream channel.	Not moved since 1946. Steep sided lateral scarp reverted to shrubby vegetation. Similarly stream bank now bush covered. No bare ground on either of these slopes.
T23D	79	520099	No. 1 Line	Plio-Pleist	e sld	SN230	526	56	SN3721	I	1	Outline visible. Part of steep scarp is covered with shrubby vege-tation. Very small areas of bare ground.	Not moved since 1946. Steep scarp covered with shrubby vegetation. No bare ground.
T23D	80	521101	No. 1 Line	Plio-Pleist	e sld	SN230	526	56	SN3721	I	1	Slide outline visible. Steep head-wall and lateral scarp. Pastured. Small areas bared ground along lateral scarp.	Not moved since 1946. Former areas of bare ground have grassed over. Small amount of toe erosion.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph 1946 - 1949 Series	1946 - 1949 Run	1946 - 1949 Photo No.	1974 - 1978 Series	1974 - 1978 Run	1974 - 1978 Photo No.	State of Activity 1946 - 1949	State of Activity 1974 - 1978
T23D	81	523104	No. 1 Line	Plio-Pleist	e sl	SN230	526	56	SN3721	I	1	Slump outline visible. Pastured. Steep headwall scarp. Very little bare ground.	Not moved since 1946. Steep scarp reverted to shrubby vegetation cover. No increase in bare ground.
T23D	82	530096	No. 1 Line	Plio-Pleist	e sl	SN230	526	57	SN3721	I	1	Outline visible. Pastured. Much bare ground along steep scarps.	Not moved since 1946. Much bare ground along steep scarp - has not healed since 1946. Slumped mass at base of slope has narrowed the stream channel.
T23D	83	553147	Opawe	Plio-Pleist	e sl	SN230	524	58	SN3721	G	4	Slump outline visible. Collapsed high level terrace edge. Pastured. No bare ground. No toe erosion.	No change.
T23B	84	552162	Opawe	Plio-Pleist	e sl	SN230	523	61	SN3721	G	4	Slump outline visible. Pastured. Three large debris slide scars in headwall area and at base of slump.	Not moved since 1946. All previous debris slide scars healed. No bare ground. Planted recently in pines.
T23B	85	568185	Te Ekaou	Plio-Pleist	e sl	SN230	522	56	SN3721	F	5	Slump outline visible. Pastured. Small debris slide scars around headwall scarp.	Not moved since 1946. No bare ground. Previous scars all healed.
T23B	86	583190	Makawakawa	Plio-Pleist	e sl	SN230	522	58	SN3721	F	5	Slump outline visible. Pastured. Large proportion of bare ground around headwall scarp and in toe area.	Not moved since 1946. Many areas of bare ground grassed over. Fewer, smaller areas of bare ground still present.
T23B	87	581207	Konewa	Plio-Pleist	e sl	SN230	521	58	SN3721	E	6	Slump outline visible. Pastured. No bare ground.	No change.
T23B	88	584209	Konewa	Plio-Pleist	e sl	SN230	521	58	SN3721	E	6	Slump outline visible. Pastured. No bare ground.	No change.
T23B	89	588213	Konewa	Plio-Pleist	e sl	SN230	521	58	SN3721	E	6	Slump outline visible. Pastured. Bare ground around headwall scarp. Severe toe erosion along steep face.	Not moved since 1946. Bare ground in headwall scarp area. Steep toe reverted to a shrubby vegetation cover with only one small area of bare ground.
T23B	90	652264	Te Ano Whiro	Plio-Pleist	e sl	SN230	519	61	SN3721	C	10	Slump outline visible. Much of headwall scarp and slumped mass has established forest vegetation. Small area of bare ground in headwall area.	Not moved since 1946. Most of bush on slump has been cleared by man. No bare ground. Former headwall scar has grassed over.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph						State of Activity	
						1946 - 1949			1974 - 1978				
						Series	Run	Photo No.	Series	Run	Photo No.	1946 - 1949	1974 - 1978
T23D	91	567018	Coppermine	Plio-Pleist	e sl	SN181	233	18	SN5163	R	6	Slump outline visible (two slump features side by side). Pastured. No bare ground.	Not moved since 1946. Areas of bare ground have grassed over. Recently planted with trees.
T23D	92	573027	Manga-a-tua	Plio-Pleist	e sl	SN181	233	19	SN5163	R	7	Slump outline visible. Steep headwall with large vertical displacement. Slumped mass is at base of slope. No bare ground. Slump largely covered in grass but headwall has regenerated shrubby vegetation on it.	Not moved since 1946. Small bare areas of ground. Shrubby vegetation cover removed since 1946. Two slump features opened up on adjacent slope. They are present on 1946 photographs but are stable and covered with regenerated shrubby vegetation. Date of recent movement unknown.
T23D	93	579025	Manga-a-tua	Plio-Pleist	e sl	SN181	233	20	SN5163	R	7	Obvious signs of slope instability along line of fault trace. Irregular outline around headwall. Small vertical displacement has taken place. Slump is grass covered. No bare ground.	No change.
T23D	94	580027	Manga-a-tua	Plio-Pleist	e sl	SN181	233	20	SN5163	R	7	Outline visible with small vertical displacement in headwall area. Grassed surface. No bare ground.	No change.
T23D	95	581029	Manga-a-tua	Plio-Pleist	e sl	SN181	233	20	SN5163	R	7	Long narrow chute-like outline with small vertical displacement in headwall area. Grassed. No bare ground.	No change.
T23D	96	589029	Manga-a-tua	Plio-Pleist	e sl	SN181	233	20	Not covered by SN3721 or SN5163			Large slump feature with steep lateral scarp around western side and subdued shallow depression around headwall and eastern side. grassed surface. No bare ground.	No change.
T23D	97	590038	Manga-atua	Plio-Pleist	e sl	SN181	233	20	SN5163	R	9	Two semi-circular headwall scarps. Vertical displacement small. Displaced mass essentially in place. Grassed surface. No bare ground.	No change.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph						State of Activity	
						1946 - 1949			1974 - 1978			1946 - 1949	1974 - 1978
						Series	Run	Photo No.	Series	Run	Photo No.		
T23B	98	673162	West Tamaki	Tamaki	r sl	SN230	523	76	SN3721	F	12	Slump outline visible. No areas of bare ground. Total area of slump densely forested.	No change.
T23B	99	686177	West Tamaki	Tamaki	r sl	SN230	522	70	SN3721	F	12	Slump outline visible. Densely forested. One small area of bare ground in headwall area.	Increase in area of bare ground along headwall.
T23B	100	692186	West Tamaki	Tamaki	r sl	SN230	522	70	SN3721	F	12	Slump outline visible. Densely forested. No bare ground.	No change.
T23B	101	651243	Piripiri	Plio-Pleist	e sl	SN230	519	62	SN3721	D	10	Slump outline visible. Small areas of bare ground around headwall and toe slope. Felled bush vegetation lying on ground. Grassed.	Marked increase in amount of bare ground along the toe slope. No indication of downslope movement of slump.
T23B	102	647242	Piripiri	Plio-Pleist	e sl	SN230	519	62	SN3721	D	10	Slump outline visible. Narrow areas of bare ground around head-wall. Felled bush vegetation lying on ground. Grassed.	Increase in the slope erosion but otherwise there is little change. No indication of downslope movement of slump.
T24B	103	555994	Mangapapa	Plio-Pleist	e sl	SN181	235	52	NOT COVERED BY SN3721 OR SN5163			Slump outline visible. Slumped mass essentially in-situ on hill slope. Deep depression below head-wall outline. Grassed.	No further downslope movement of slumped mass has taken place. Ponded water accumulates in de-pression in wet winter months.
T24B	104	535969	Mangapapa	Plio-Pleist	e sl	SN181	235	52				Series of headwall scarps up hill-side indicate that downslope move-ment has taken place. Grassed.	No change. Trace of Wellington Fault strikes across the base of this slump. Gas pipeline runs down the length of this slump.
I24B	105	531965	Mangapapa	Plio-Pleist	e sl	SN181	235	52				Series of headwall scarps mark position of at least two slump features. Grassed.	No change. Wellington Fault trace strikes across these slumps.
T24B	106	513943	Mangamanaia	Plio-Pleist	e sl	SN181	237	40				Slump outline visible. Small pond in headwall scarp area. Grassed.	No change.
T24B	107	504935	Mangamanaia	Plio-Pleist	e sl	SN181	237	40				Several linear depressions indicate downslope movement has taken place at more than one locality within this large mass movement feature. Grassed.	No change.

NZMS 270 No.	Locality No.	Grid Ref.	Catchment	Lithotype	Feature	Aerial Photograph						State of Activity	
						1946 - 1949			1974 - 1978				
						Series	Run	Photo No.	Series	Run	Photo No.	1946 - 1949	1974 - 1978
T24A	108	485915	Manawatu River	Plio-Pleist	e sl	SN181	238	41	) NOT COVERED BY SN3721 or SN5163			Slump outline visible. Discontinuous fault traces aligned along head of slump. Grassed.	No change.
T24A	109	451441	Manawatu River	Plio-Pleist	e sl	SN181	236	34				Slump outline visible. Very small areas of bare ground around headwall.	No change.

APPENDIX VII: Reference numbers of rock samples in the Massey University Reference Collection sampled from the southern Ruahine Range

Field number	M.U. number	Metric grid reference	Field number	M.U. number	Metric grid reference	Field number	M.U. number	Metric grid reference
Po_5	MU 3	T23/660235	Mk_5	MU 19	T23/593190	MG_3	MU 20	T24/485950
MG_5	MU 21	T24/486951	CR_2	MU 22	T24/487928	Pw_1	MU 24	T23/581149
Rk_2	MU 25	T23/645141	MkD_4	MU 26	T23/628185	Mk_6	MU 27	T23/591189
Mh_3	MU 28	T23/506045	No. 1 (5)	MU 29	T23/557087	No. 2 (4)	MU 30	T23/550072
Mk_1	MU 32	T23/593190	Mk_2	MU 33	T23/591189	Mk_3	MU 34	T23/587190
MkA_2	MU 35	T23/605163	Rp_{20}	MU 37	T23/571088	No. 1 (1)	MU 38	T23/553085
Tk_1	MU 39	T23/512051	Op_1	MU 40	T23/589121	Mk_4	MU 41	T23/635181
Mh_5	MU 42	T23/506046	Rk_{10}	MU 44	T23/649143	Rk_{11}	MU 45	T23/656161
Rp_2	MU 46	T23/564069	Rp_7	MU 48	T23/570080	Rp_{19}	MU 49	T23/568074
Rp_8	MU 51	T23/568074	MG_1	MU 52	T24/492944	Co_4	MU 63	T23/547047
Wh_3	MU 67	T23/525025	No. 1 (2)	MU 68	T23/553085	Rp_{18}	MU 71	T23/569070
Mtu_1	MU 76	T23/562113	Rk_1	MU 77	T23/656161	Rk_3	MU 78	T23/649143
Mh_1	MU 79	T23/512041	OP_3	MU 80	T23/571129	Ko_1	MU 81	T23/608208
Po_2	MU 82	U23/712226	CR_1	MU 83	T24/488928	Or_2	MU 87	T23/594080
Mp_1	MU 91	T23/601113	Rk_6	MU 92	T23/655136	Mte_2	MU 93	U23/731182
Mte_3	MU 94	U23/734183	Mte_4	MU 95	U23/711161	Mte_8	MU 97	U23/710168
ET_1	MU 98	U23/712175	Rp_1	MU 99	T23/577076	NRs_1	MU 101	T24/488928
NRs_3	MU 102	T24/480922	MG_4	MU 103	T24/486951	MG_6	MU 104	T24/492944
Rk_{15}	MU 105	T23/646145	Mp_2	MU 106	T23/515037	Mh_4	MU 108	T23/516037
Wh_4	MU 111	T23/525025	No. 1 (4)	MU 113	T23/560088	No. 1 (6)	MU 114	T23/555087

APPENDIX VII: (cont)

Op$_2$	MU 117	T23/571129	Un$_2$	MU 121	unknown	Mte$_6$	MU 122	U23/727173
Mtc$_7$	MU 123	U23/720168	ET$_3$	MU 124	U23/713162	Rk$_9$	MU 125	T23/661141
Pr$_3$	MU 128	T23/673251	Mg$_3$	MU 129	T24/538992	Dn$_1$	MU 130	T23/562096
MkD$_7$	MU 131	T23/643199	MkD$_9$	MU 132	T23/636193	Rk$_4$	MU 133	T23/657164
Rp$_{12}$	MU 135	T23/581056	Co$_8$	MU 145	T23/540032	Ou$_2$	MU 150	T23/605107
MkA$_1$	MU 154	T23/613155	Co$_{10}$	MU 158	T23/546041	Or$_7$	MU 159	T23/601064
Or$_{13}$	MU 161	T23/592072	Rk$_{13}$	MU 162	T23/646141	MkD$_8$	MU 163	T23/664180
Rk$_{16}$	MU 169	T23/660163	WT$_1$	MU 171	T23/698208	Po$_7$	MU 173	T23/699232
Rk$_{18}$	MU 178	T23/652114	Or$_4$	MU 185	T23/593082	Rp$_{17}$	MU 186	T23/569070
WT$_2$	MU 189	T23/682162	ET$_2$	MU 190	U23/710165			