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ASPECTS OF THE GEOMORPHOLOGY OF THE GREYWACKE RANGES BORDERING THE LOWER AND MIDDLE WAIKATO BASINS

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Abstract

The fault-bounded blocks which make up the Greywacke Ranges bordering the Lower and Middle Waikato Basins have a deep red-weathered regolith and are covered by mantles of volcanic ash which can be used for dating ground surfaces. The drainage texture is exceedingly fine compared with that of Dartmoor (U.K.) and Unaka Mountains (U.S.A.) This is attributable to rainfall type, regolith, vegetation cover and soil physical properties. The major types of mass movement are deep fossil slumps on upper slopes where the regolith is deep; debris slides on mid- and lower slopes where the regolith is thin; and seepage heads controlled by ground water conditions. The valley floors show both stream incision, and aggradation resulting from infill with mass movement debris. Deforestation has increased the frequency of mass movement during high intensity rainstorms and the slopes are at present becoming adjusted to changed equilibrium conditions.

INTRODUCTION

The Lower and Middle Waikato Basins are situated in the South Auckland Land District of the North Island of New Zealand. The northern or lower basin contains the estuary of the Waikato River and the middle or southern one is the Hamilton Basin (Figure 1). Both basins are structural depressions bounded by faults which, where exposed, define the boundary between the infilled grabens of the lowlands and the surrounding upland horsts which are composed largely of Jurassic sandstones, siltstones and conglomerates which are commonly referred to as greywacke (Bartrum 1926; Battey 1949; Henderson and Grange 1926; Kear 1960; Kear and Tolley 1957; Laws 1931; Mead 1938).

The bordering greywacke ranges are not continuous and are overlain by more recent rocks in the lowerlying areas—particularly to the east of the Hamilton basin where lowland routes give access to the Hauraki Plains which are structurally another graben.

The ranges are deeply dissected by streams, which in the asymmetrical blocks of the east descend from a main watershed which is closer to the eastern than the western edges of the ranges. The greywacke ranges have maximum altitudes of about 1,700 feet and many ridges are above 1,000 feet.

Considerable areas of the Hapuakohe Range are still covered by native forests, but in the last 70 years extensive land development has converted most of the forest into pasture.

Climatically most of the ranges would fall into the category of AB'_1ra' in Thornthwaite's 1948 classification (Critchfield 1966). In other words the area is perhumid (A), mesothermal (B'), has no seasonal water deficit (r), and less than 48% of the potential evapotranspiration is concentrated into the three summer months (a').

General reconnaissance field work has been undertaken throughout the ranges, but it has been concentrated in the area from the Hunua Ranges in the north to the Whitehall district in the south. Three localities (marked on Figure 1) have been chosen for detailed study.

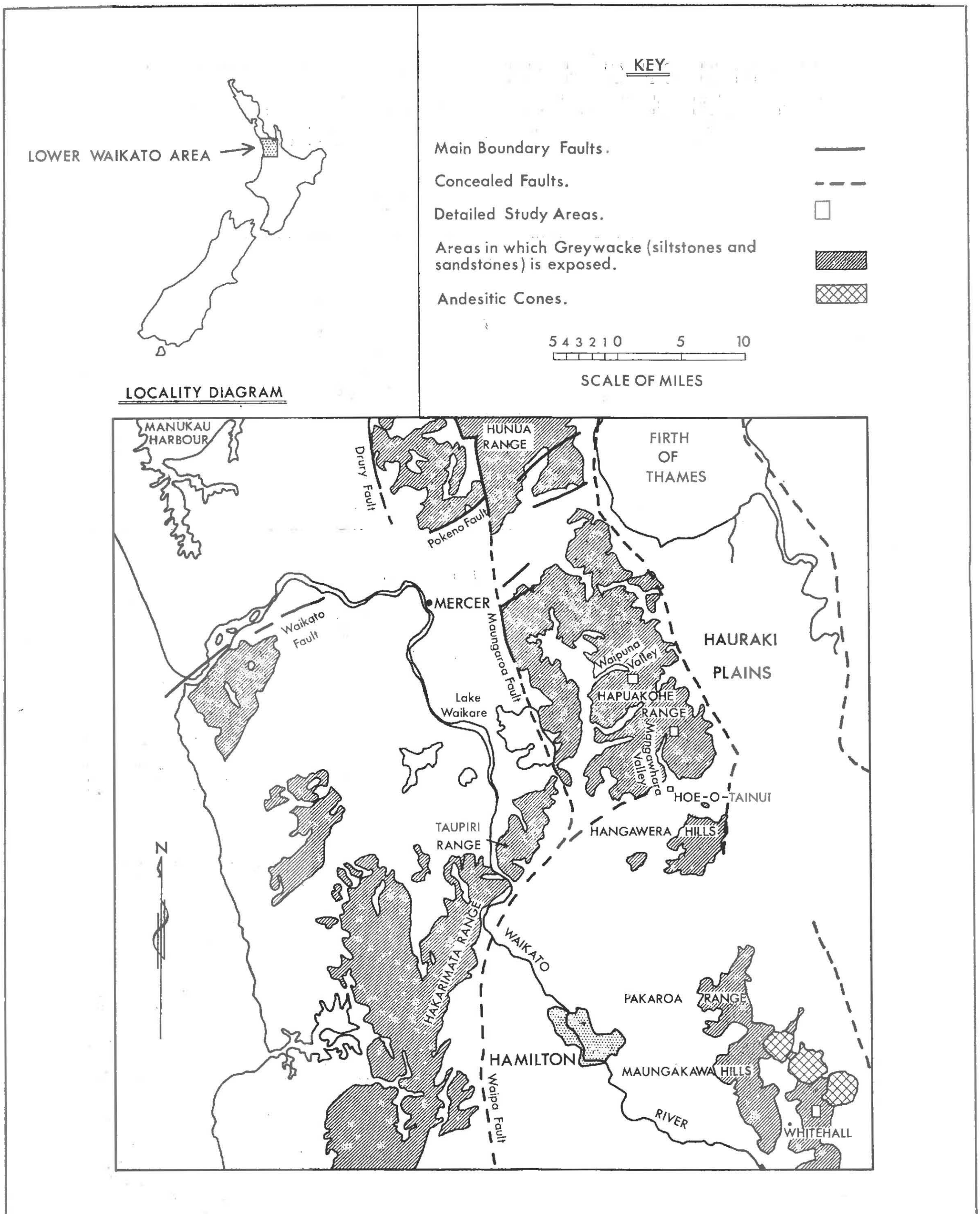


Figure 1: Location of the Middle and Lower Waikato Basins showing the bordering greywacke ranges and main boundary faults.
(Compiled from published maps of N.Z. Geol. Survey.)

THE REGOLITH

The term 'regolith' is used in the sense of being all weathered and superficial material. It thus includes weathered greywacke, the overlying volcanic ashes and the colluvium derived from either or both of these.

Deep Weathering

The weathering mantle on the ranges has a very variable depth. Some exposures, as at the Whitehall quarry (GR. N66/063352) show deep red-weathered material to a depth of over 80 feet. A typical deep section will show a weathering front forming a boundary between unweathered greywacke and slightly weathered material with yellow and brown staining, but with the original rock structures still preserved. Above this the intensity of weathering increases until the rock is entirely replaced by yellow and brown material composed largely of clay minerals. The original rock structure may still be seen and in a few localities (as at GR. N66/114372) core stones of unweathered rock surrounded by weathering skins still exist where the greywacke was once massive. The upper part of a deep section, often with a depth of 20 ft. is typically composed of red-weathered greywacke.

Red beds are derived from red soils developed under a savanna type of climatic regime (Krynine 1949). Red beds in the Wellington area of New Zealand are described as relict from the Mindel-Riss or Riss-Würm interglacial (Te Punga 1964) and those of the South Auckland district may be assumed to have a similar age.

Volcanic Ash Beds

Overlying the red-weathered greywacke are volcanic ash beds. Their thickness, like that of red-weathered regolith is very variable and largely depends upon their position on the slope. Broad interfluves usually have the deepest deposits and the most complete sequence, while the midslopes bear thinner colluvial deposits. In some areas midslopes are bare of ashes and deep regolith material. Footslopes usually have re-deposited ashes and weathered greywacke as a colluvial mantle of varying thickness. On nearly flat sites the lower, and therefore the oldest ash beds, called as a group the Hamilton Ash, can have a thickness of ten to twenty feet, although thicknesses of more than six feet are unusual on the greywacke ranges. This ash is of rhyolitic and andesitic composition (New Zealand Soil Bureau, 1954) and aged 75,000 to 125,000 B.P. (Kear, Schofield and Kermode, 1964). These ashes have weathered to a heavy brown clay.

Overlying the Hamilton in the south of the area, are the Tirau ash beds (N.Z. Soil Bureau, 1954). The beds of the Tirau ash vary in age from about 15,000 B.P. to 35,000 B.P. (Mr W. A. Pullar, N.Z. Soil Bureau pers. comm.). The soils for which these ashes are the parent materials are much more friable, with a more granular structure than those on the older ashes or red-weathered greywacke. Permeability and other physical characteristics of the soil consequently depend upon the nature of the parent material.

In the southern part of the study area younger ashes of the Rotorua sub-group overlie the Tirau and have a depth of up to ten feet. In age they are between 3,270 ± 200 B.P. and 15,000 B.P. (Mr W. A. Pullar pers. comm.) Like those on the Tirau ashes the soils on these younger ashes are friable and permeable.

Extending as a thin film over the hills of the Whitehall area is pumice lapilli derived from the Taupo eruption (Selby 1966), dated at 1900 ± 60 B.P. (Healy, Vucetich and Pullar 1964).

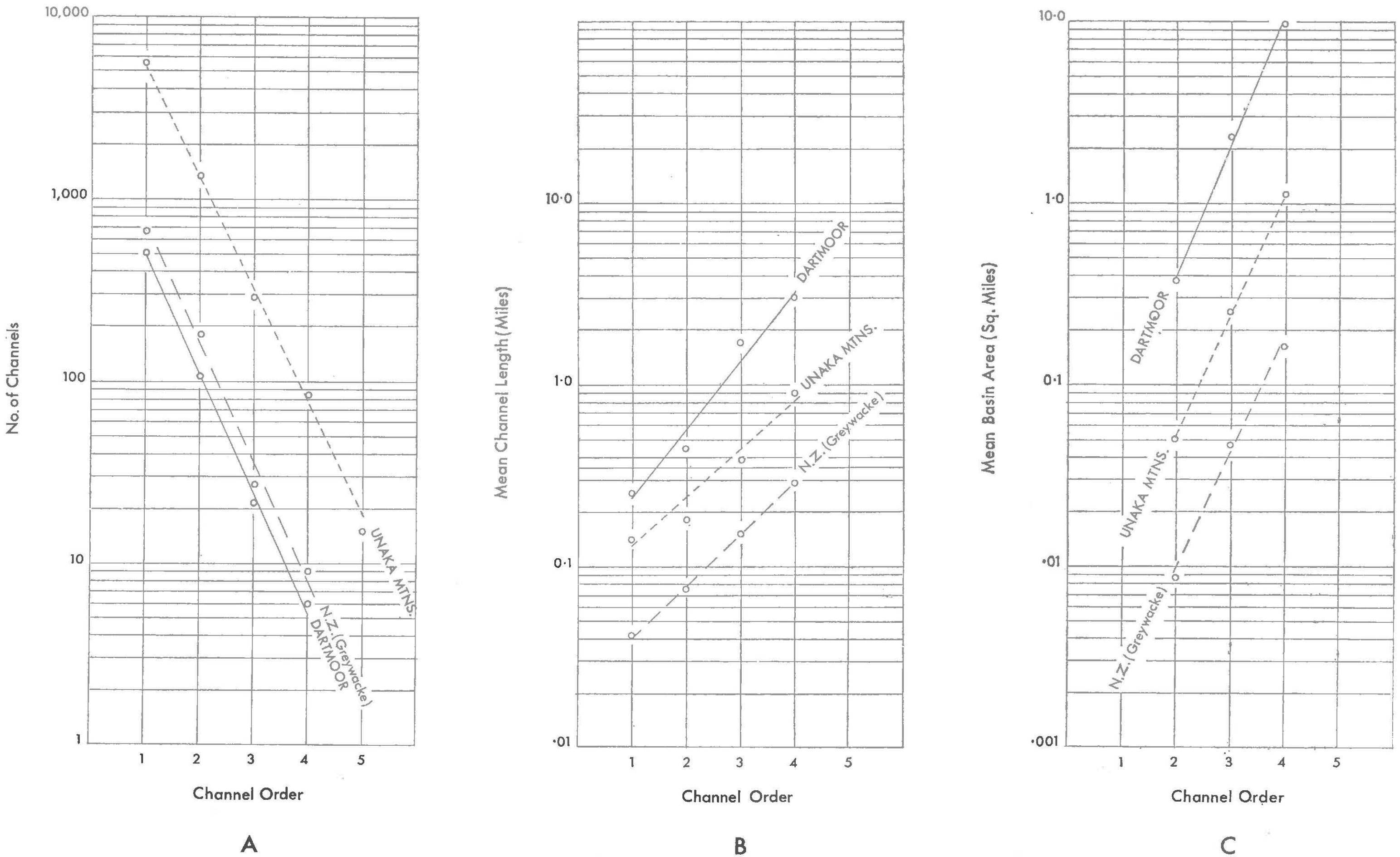


Figure 2 : Plots of channel order against:
A—number of channels in sample basins;
B—mean channel lengths;
C—mean basin areas;
(Data for Unaka Mountains and Dartmoor from Chorley and Morgan, 1962).

These ash beds are of considerable geomorphic significance, partly because they are the parent material for the soils, partly because their various physical characteristics have an influence on the rates and types of slope erosion. To the geomorphologist they have the added value that they can be used to date ground surfaces. The detailed mapping of individual beds is not yet complete, but the type of dating of landforms which was attempted by Selby (1966) in the Whitehall area should eventually be extended over all the ranges.

DRAINAGE BASIN MORPHOMETRY

In a number of recent papers Cotton (1958a; 1958b; 1962; 1963a; 1963b) has contended that New Zealand relief is fine textured. To test this contention morphometric data from the study areas has been plotted against that derived by Chorley and Morgan (1962) from Dartmoor, England, and the Unaka Mountains, U.S.A. The data from New Zealand is possibly not directly comparable with that from the other two areas for the Dartmoor maps were at a scale of 1:25,000 and those of the Unakas at 1:24,000. In the study area the only contour maps available are at a scale of 1:63,360, but there are semi-controlled air-photo mosaics at a scale of 1:15,840. To obtain accurate results stream orders have been plotted on to the mosaics using stereoscopic pairs of photographs as a check.

By Strahler's method (1957, p. 915) the smallest streams with no tributaries are said to be of order one. Those stream segments, with only order one tributaries are said to be of order two, and those segments below the junction of two or more order two segments but with only order one or two tributaries, are said to be of order three, etc. One difficulty with the method is that when maps are used it has to be decided whether to observe only the channels marked in blue as streams, or to extend the channels along a depression denoted by nicks in the contours, or whether to extend all channels to the watershed. Morisawa (1957) has demonstrated that the contour method is the most reliable. In using the mosaics only those channels have been plotted which might be expected to be marked on an accurate map. This factor must be remembered when the data from the three areas are compared. Three order four basins from each of the three study areas have been measured. The order one and two channels were found, during field work, to be scars of mass movement features and even in order three channels there were often no permanent streams. A check demonstrated that the order one and two channels were never shown on the 1:63,360 maps but that consistent results could have been obtained if the smallest channels marked on the maps had been regarded as order three.

It is evident from the graphs in Figure 2 that the same simple relationships between stream order and basin geometry exist in the New Zealand ranges as on Dartmoor or the Unakas. Plot B emphasises the extreme shortness of channel lengths in New Zealand compared with the other areas and also near-perfect geometrical relationships between channel length and order. Plot C further demonstrates the extremely fine texture of erosional relief in New Zealand compared with the other two areas. Cotton's contention is demonstrably correct for the greywacke ranges of the South Auckland area. Chorley and Morgan have calculated ideal drainage basin forms and stream lengths for Dartmoor and the Unakas and these are shown in Figure 3. Using mean values of all the basins studied a similar basin has been drawn for the New Zealand area. The figure gives a visual impression of the differences in texture between the areas.

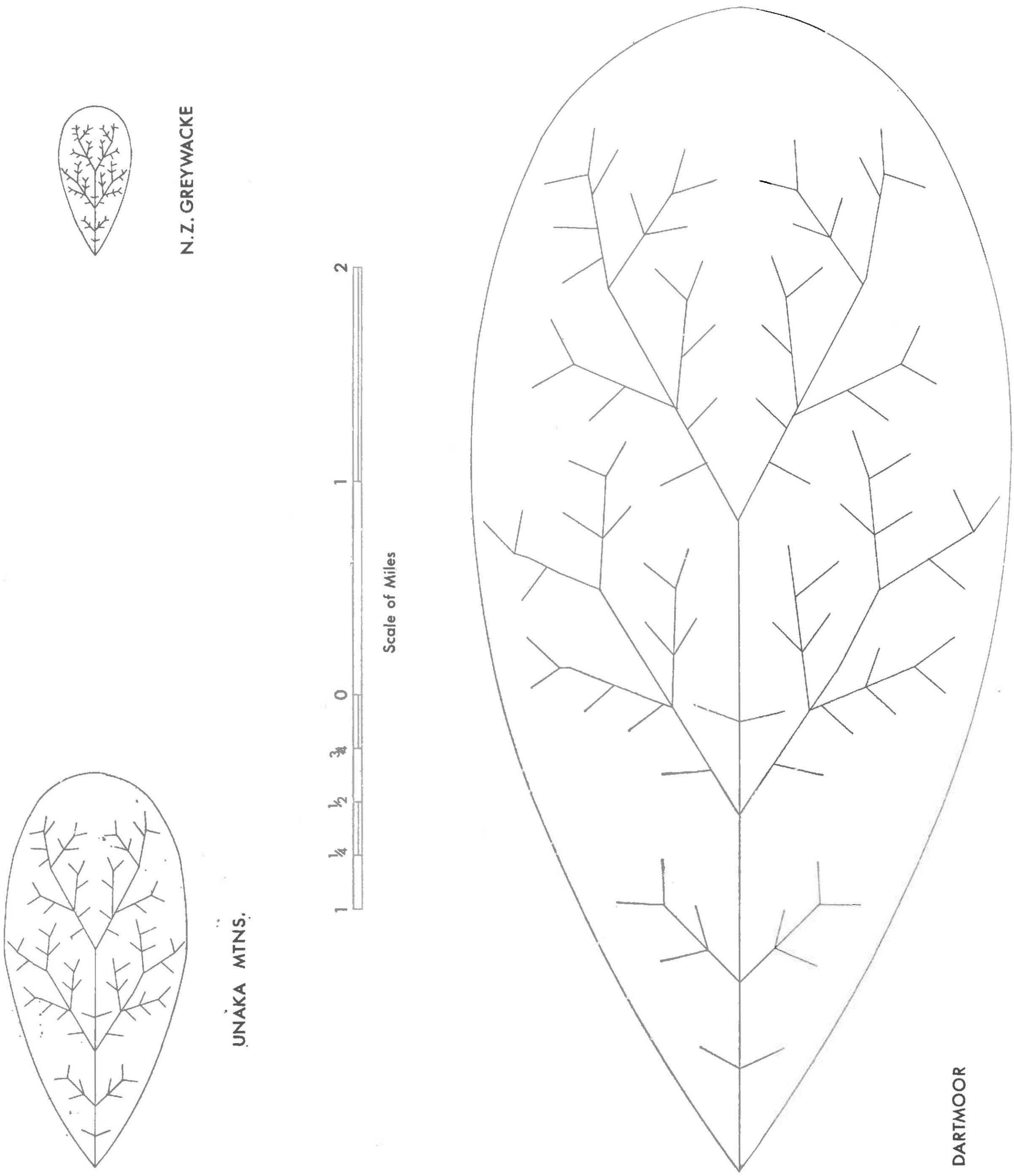


Figure 3 : Idealised fourth order drainage basins.

The causes of the differences in relief texture are not so easily assessed. Dartmoor and the Unaka Mountains are areas of crystalline igneous and metamorphic rocks which may be more resistant to erosion than the greywacke of New Zealand. Dartmoor has a low relative relief compared with the other two areas but this is probably the result rather than the cause of coarse-textured dissection. The channel system in any area appears to be related to conditions of maximum, rather than mean, runoff (Wolman and Miller, 1960). New drainage lines once cut appear to be persistent. Chorley and Morgan assert that Dartmoor and the Unaka Mountains have both been "under the influence of humid climatic conditions since perhaps the beginning of the Tertiary" (p. 32), and that past periglacial activity has not been significant. They suggest, therefore, that the drainage texture in these areas is related to the degree of intensity of the precipitation. Mortensen (1959) has made a similar suggestion for New Zealand. The lack of information about past and present rainfall regimes prevents a valid comparison, but even the limited information presented in Table 1 does not entirely support the suggestion that intense rainfall is the main factor. At Hoe-o-Tainui the precipitation is less intense and of smaller amount than that given for the Unaka Mountains. Field study also suggests that any explanation in terms of one factor is a gross oversimplification. In areas where relief texture is dependent upon stream incision, the fineness of the texture must depend upon the size of the catchment necessary to initiate a stream. In the study areas, however, the fine texture is a result of mass movement, and is finest where the mass movement forms are smallest and closest together. Such features are the result not only of hydrological factors but many factors such as depth of regolith, slope angle, vegetation cover and soil physical properties.

Table 1: PRECIPITATION DATA

	Number of Raindays per Year	Mean Annual Precipitation in Inches	Maximum Precipitation in 24 Hours in Inches
N.Z. (Hoe-o-Tainui)	189 (11 yr. record)	71.24	6.5 (50 yr. estimate for Ruakura)
Unaka Mountains	120	82.47	15.0 (40 yr. record)
Dartmoor	200	67.90	3.0 (23 yr. record)

N.B. The figures given above should merely be regarded as indications of major differences, as the length of records is not uniform and the maximum precipitation figure for New Zealand had to be taken from the nearest lowland station—Ruakura—this figure is therefore probably too low (Robertson 1963).

A—debris slides in the Hunua Ranges after the rainstorm of February 28th, 1966.

B—amphitheatre-shaped depressions at the heads of slopes near Whitehall. Boulders litter the lower slopes and the valleys are infilled with colluvium.

C—type 3 slope in the Waipuna Valley.

Figure 4

D—details of a debris slide in the Hunua Ranges. Terracettes and cracks are evident.

SLOPE FORMS

The valley-side slopes are largely controlled by the deep fine dissection and the large relative relief.

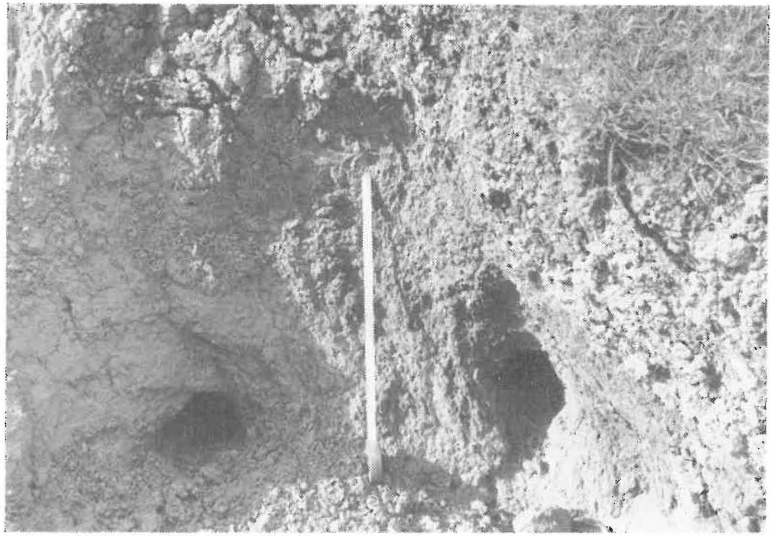
The most common slope profiles fall into one of three main categories (Figure 6A). In areas where streams are actively eroding or valley floors are infilled with fine sediment and interfluves are narrow, the upper slopes are short and convex, midslopes are long and straight with angles of between 20° and 40°, and footslopes are either very short and concave, or they are non-existent (see Figure 4A). The second major type occurs only where interfluves are broad. Here the upper part of the slope is occupied by a deep depression with a relatively low angle of slope on its floor of between 5° and 15°, and a lip above a long straight midslope. The footslope again may be non-existent or short and concave (Figure 4B). The third type is rare. It has a midslope bench or break of slope in which the upper part is gently sloping and the lower part is steeper (Figure 4C). In some areas the upper part appears to be due to the coalescence of upper slope depressions, but elsewhere it seems more probable that it results from a structural control. It is quite clear that the bench is not the result of cyclic erosion for it is irregular, does not necessarily dip down-valley, and is not continuous through a valley and tributary system. For the same reasons it is most unlikely to be caused by stream incision.

The midslopes, although generally straight, have many detailed forms of microrelief. Soil creep and stock trampling have given rise to flights of discontinuous terracettes. Deforestation has left pits up to four feet deep where old tree roots have been removed, and corresponding mounds where the soil around the old roots and stump has been left and grassed over. Common also are shafts, locally called tomos, which are parts of the collapsed roofs or terminal sections of subsurface drainage tunnels or pipes. These pipes and shafts are usually confined to linear depressions on slopes where the regolith has a depth of at least eight feet. Their origin and incidence on the greywacke ranges has not yet been studied in detail. Workers in other areas (Ward, 1966) have concluded that they are the result of the localisation of subsurface water, and the presence within upper soil horizons of clays which swell and disperse on wetting, and crack on drying.

A—seepage heads with infilled valleys below them at Whitehall.

B—a shallow slide with thin, regolith material moving over coherent weathered greywacke. In the foreground cracks mark the site of a potential slide.

C—a slickensided manganese-coated slide-plane. Regolith material is exposed in the headwall of the slide. Hunua Ranges.



D—pipes exposed in the headwall of the debris slide.

Figure 5

MASS MOVEMENT

Mass movement forms fall into three categories (Figures 4B, 4D, 5A). On upper slopes where the interfluves are broad there are deep flat-floored features previously described as amphitheatre-shaped forms (Selby, 1966). The other two types of feature are debris slides and seepage heads; these can occur at any position on the slopes, but are most common on mid- and footslopes.

Amphitheatres

Amphitheatres in the Whitehall area have dimensions of about 600 x 400 feet; they are between 80 and 30 feet deep, and have floors sloping at between 7° and 5°. The floors of many of them are strewn with boulders—a type of surface often called a boulder field. In some amphitheatres the boulders within a matrix of fine ash and colluvium form a mound. The existence of the mounds and the deep backwalls of the amphitheatres suggest that they are the result of slumping. Amphitheatres are no longer being formed in the greywacke ranges and all have smoothed slopes which suggest they are old.

Similar features have been described from elsewhere. Cotton and Te Punga (1955) suggest that in the Wellington area of New Zealand they are similar to the European **Dellen** produced by movement of solifluxion materials in a periglacial environment. Berry and Ruxton (1961) attribute similar features in Hong Kong to slow mass movement in a humid tropical climate, and Waters (1957) found that basins on Dartmoor were the result of excavation of residual materials derived from deep chemical weathering in areas of close jointing.

Borings in and around amphitheatres in the greywacke ranges of the lower Waikato Basin indicate that a section through an amphitheatre has some of the features shown in Figure 6. On the interfluve the deep weathering mantle and the overlying volcanic ashes have their greatest depth. On the steeper slopes the regolith is thin because of continuous erosion. A deep mass movement feature can therefore only be formed at the head of the slope where slope angle and deep regolith are both suitable. In such a position it reveals deeply buried core stones which are the origin of the boulder fields. On the mid- and footslopes the regolith is thinner and therefore only shallow features can form. By using the ashes for dating it has already been demonstrated that the deep slumps in the Whitehall area were formed between 15,000 B.P. and 1,900 B.P. (Selby, 1966). It is possible that the slumps occurred soon after the deposition of the ashes when unstable slope conditions might be expected to be at a maximum. There have been no slumps since 1,900 B.P., and possibly before, because the edges of all the broad interfluves are fully fretted by the old slumps so that there are no potentially unstable deep deposits left, and elsewhere the interfluves are too narrow to have the required deep regolith. It is not necessary to presuppose a change of climate to account for the amphitheatres and in any case it is possible that these ranges did not experience a periglacial climate (Willett, 1950), but such changes cannot be excluded.

In brief, then, it appears that dells or amphitheatres may occur in many different environments, just as boulder fields may be formed in periglacial conditions (Joyce, 1950; King and Hirst, 1964; Smith, 1950) or as a result of deep tropical or subtropical weathering (Demek, 1964; Rother, 1965; Selby, 1966). Neither forms nor deposits alone are unequivocal diagnostic tools for determining past processes.

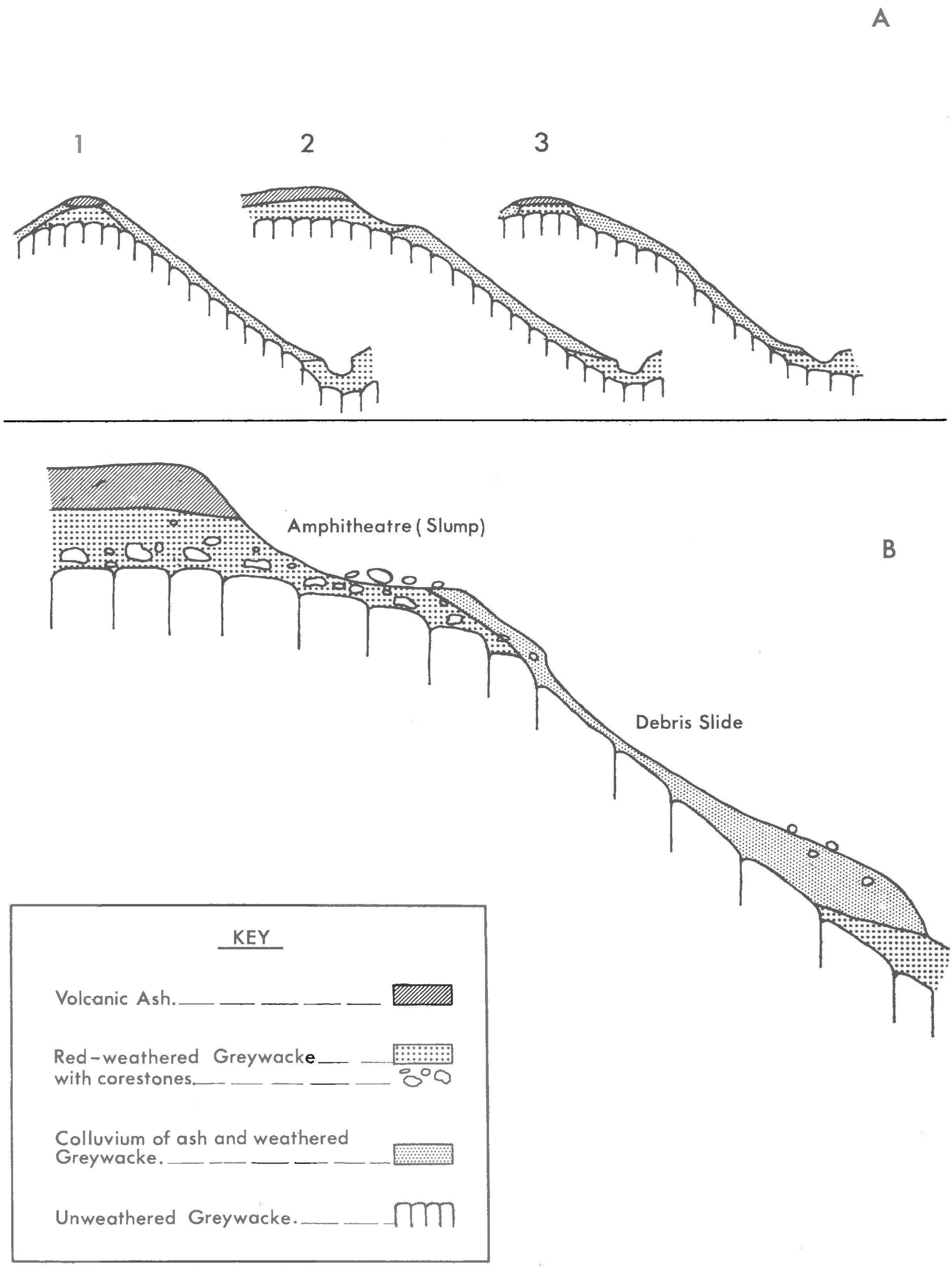


Figure 6 : A—slope types:
(1) long straight slope below narrow interfluve;
(2) straight slope below wide interfluve with marginal amphitheatre;
(3) compound slope with steeper footslope;
B—schematic profile through a slope with wide interfluve, amphitheatre and debris slide.

Debris Slides

Sharpe (1938) defines a debris slide as a ". . . movement of predominantly unconsolidated and incoherent earth and debris in which the mass does not show backward rotation but slides or rolls forward" (p. 74). The movement, according to the definition, has to be rapid. In the study areas there are many features which fit Sharpe's description. The debris slides are found at almost any position on the slopes but it is most usual for the slide to originate at a break of slope either at the upper convexity (Figure 4A), at an inflexion in the midslope, or at the footslope. Compared with the upper-slope slumps the slides are shallow, varying in depth from two to eight feet, and are narrow, seldom having a width exceeding 100 feet. The length of the scar downslope is often uncertain as the lower part of the scar is obscured by the debris from above.

The slides occur entirely within the regolith. Many fresh slides, resulting from movements after an intense rainstorm on 28th February to 1st March, 1966, have been examined in the study areas. The slide plane may occur in any one of a number of positions: on a clay pan or iron pan within the soil; at the base of the colluvium which forms the parent material of the present soil and above the underlying ash bed or weathered greywacke; between ash beds; within an ash bed which has had a soil developed on it and in which a pan has been formed; between an ash bed and the weathered greywacke; and within the weathered greywacke where a structural plane derived from the original rock is nearly parallel to the slope. The common feature of these slide planes is that they are lines of movement for subsurface water. In Figure 5B is shown a slide plane on red-weathered greywacke beneath only a thin colluvial mantle. In Figure 5C the slide plane is a slickensided manganese-coated surface on red-weathered greywacke representing a joint or bedding plane of the original rock. It is evident from both these sites that the plane extends upslope beneath the still-coherent soils. Pipes which are further evidence of subsurface water movement are shown in Figure 5D. These pipes, and others like them, occur at the level of the slide plane. Some appear to be related to seepage along the base of vertical cracks in the soil, in others are remnants of old tree roots.

Debris slides are the only large mass movement forms which are developing at present, as is shown by the evidence of erosion in the 28th February rainstorm. Most commonly they are caused by undercutting of the base of the slope or develop in the scars of older debris slides (Figure 7).

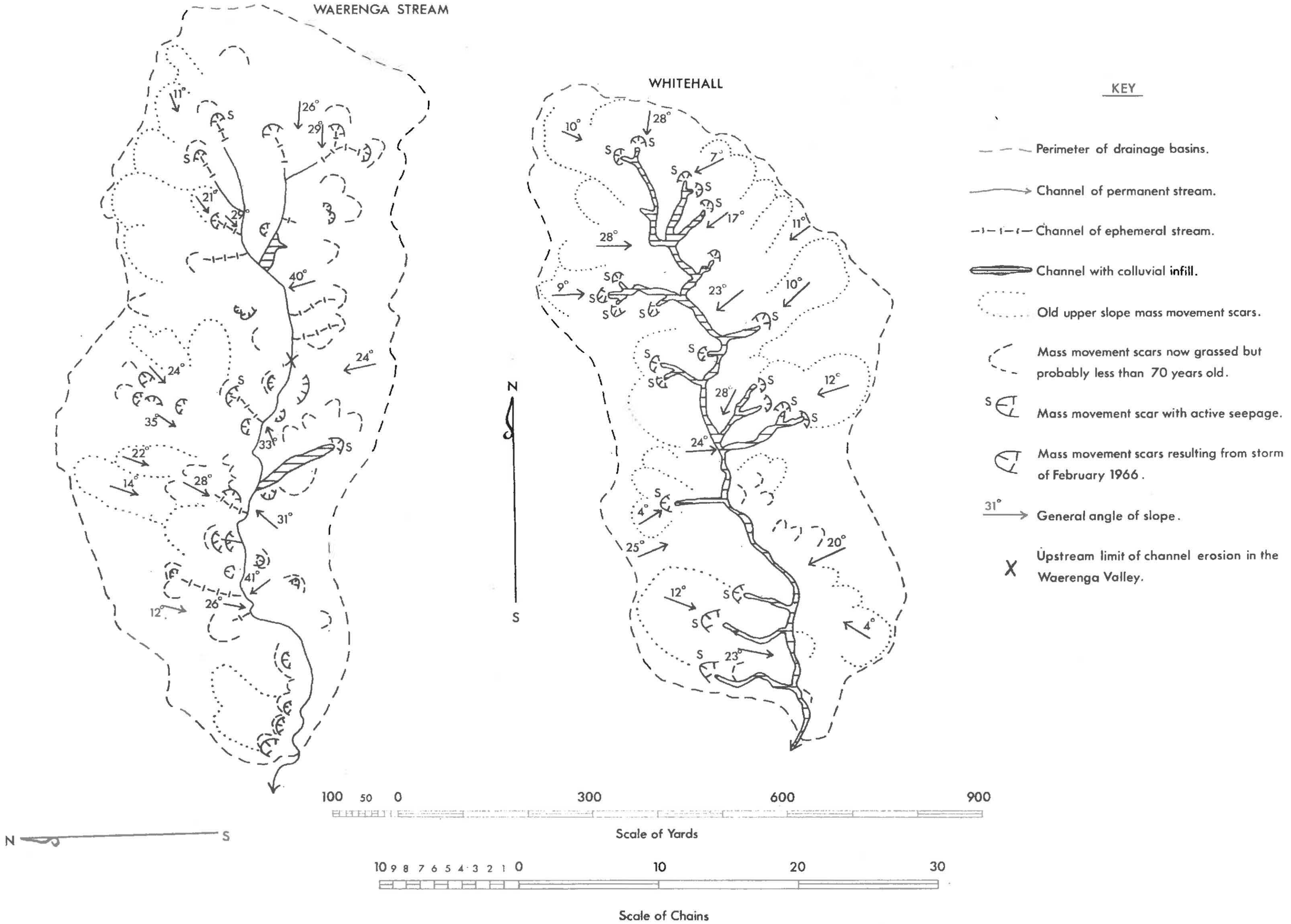


Figure 7: Maps showing mass movement and other geomorphic features in two study areas.

Seepage Heads

Seepage heads are small hollows, usually with terracetted backwalls, from which originate swampy channels infilled with fine sediment. The origin of these features is not entirely clear but it is possible that the head is usually a small debris slide or shallow slump. The colluvial infill was derived from the head and is probably removed by the slow seepage from the head which keeps the infill saturated. The seepage heads are only found on the midslopes and usually all occupy a similar position along any valley wall. This suggests that they result from seepage from a perched water table, which in a few cases may be controlled by a structural plane.

The relationships of the mass movement features with each other can be seen in Figure 7. In the headwaters of the Waerenga stream amphitheatres occupy the heads of the northern slopes where the interfluves are broad and the slope angle is less than about 15°. Below the amphitheatres the slopes are steeper and debris slides, both old and new, occur on slopes of between 20° and 40°. Many of the seepage heads occur within old debris slides as do many of the newer debris slides. In the Whitehall area the upper-slope amphitheatres are even more significant while below them most of the forms are of the seepage type. The storm of 28th February, 1966, did not affect this area and all the scars are grassed over. It can be seen that, at the present, nearly all the slopes are affected by some type of mass movement and that the corrasive action of streams is of little significance in shaping the valleys, or any of the landforms of channels with a lower order than three.

It is interesting to note that the depth to length ratio of the mass movement forms in the study area is about the same as those described by Skempton (1953) for the Castle Eden Dene near Shotton, Durham, England. The slumps for both areas have a depth/length ratio of about 20 and the debris slides one of 5, but whereas the English slopes are being flattened by mass movement, the New Zealand slopes appear to be retreating parallel to themselves because the slides are of uniform depth throughout and the debris is removed from the base of the slope.

Causes of Mass Movement

Mass movements occur because occasionally the intensity of shearing stresses becomes greater than the shearing resistance of the materials on a slope, or the shearing resistance decreases until it becomes less than the shearing stresses (Terzaghi, 1950, p. 87). The causes therefore may be external—that is, they increase shearing stresses—or they may be internal—shearing resistance is decreased.

Slope instability results from failure at sufficient points to define a surface along which movement can take place. Failure is seldom spontaneous for the slope is usually unstable for some time before failure occurs and warnings, such as the formation of cracks and small movements, are often seen. To define **the** cause of a failure is therefore impossible. As Sowers and Sowers (1961, p. 319) put it: "Calling the final factor **the cause** is like calling the match that lit the fuse that detonated the dynamite that destroyed the building **the** cause of the failure". Gravity in the form of the weight of the soil and its contained water is the major shearing force tending to produce failure, while the shearing resistance of the soil is the major resisting force.

The factors that contribute to high shear stress are (Varnes, 1958):

(1) Removal of support by stream undercutting.

(2) Weathering.

(3) Wetting and drying causing volume changes.

(4) Weight of rain.

(5) Removal of soluble material.

(6) Lateral pressure due to hydration of clay.

(7) Lateral pressure due to water in cracks.

(8) Earthquakes.

Factors that contribute to low shear strength are:

(1) Inherently weak materials.

(2) Existence of structural planes which provide lines of movement for ground water.

(3) Weathering processes causing decrease of cohesion and disintegration of particles.

(4) Hydration of clay minerals, and resulting decrease of cohesion of clay soils at high water contents.

(5) Reduction of cohesion as cracks develop in drying soils.

(6) Absence of deep tree roots.

(7) Increases in pore water pressure.

(8) Development of tension cracks.

It should be noted that lubrication by water is not a cause of mass movement (Terzaghi, 1950, p. 91). Water in contact with some common minerals such as quartz actually acts as an anti-lubricant, and furthermore only a very thin film of lubricant is required to produce a full lubricating effect. Most humid areas have soils with permanently higher moisture than is required for full lubrication, hence additional water has no further lubricating effect.

The immediate cause of most slope failures is a sudden rise in pore-water pressure (Terzaghi, 1950, p. 119). In a soil in which the voids between the grains are filled with water the grain structure cannot immediately support an increased load because compression of the water cannot occur. It is the pressure in the water (also known as the neutral stress) which supports the load. As the water seeps away the grain structure assumes the load and the soil compresses. At the same time the neutral stress reduces to zero, but in a potentially unstable situation the rise in pore-water pressure can induce movement of the soil.

It is clear from the history of recent slope failures in the study areas (Selby, in press) that debris slides are the result of both decreased shear strength and increased shearing stresses on potentially unstable soils during exceptionally intense rainstorms. Jackson (1966) and Grant (1965) have also demonstrated that elsewhere in New Zealand severe rainstorms can trigger large-scale mass movements.

VALLEY FLOORS

The valley floors of the study areas are of two types: those in which streams are actively downcutting, and those which are infilled with fine sediment.

Active corrasion is confined to relatively short periods during and after intense rainstorms, while for most of the year stream activity is limited to transport of fine sediment. This point can be illustrated by reference to the headwaters of the Waerenga stream. The stream usually has a depth which does not exceed six inches and a width of five feet in the lower section of the third order channel. The bed is 'armoured' by gravels and boulders which the stream cannot move, and the water is normally clear. During the intense rainstorm of 28th February, 1966, the stream occupied a channel with a maximum depth of seven feet and a width of 32 feet at the lower end of the order three basin. The normal estimated discharge is about five cusecs, but the peak discharge during the rainstorm is estimated at between 3,000 and 5,000 cusecs. Such a peak discharge from a drainage basin of 2.8 square miles is of the same magnitude as that calculated for the forested headwaters of the Mangatangi Stream in the Hunua Ranges, during the same storm, where a catchment of 17 square miles yielded a peak discharge of 12,500 cusecs (Mr G. T. Ridall, Waikato Valley Authority pers. comm.).

The largest boulder which is known to have been moved had an estimated weight of 240 lbs. and was transported 200 yards downstream. Many other boulders of approximately the same size were moved in this flood. The rock cut channel was abraded and chips of resistant greywacke broken off at the walls three feet above the normal stream level. Many of the boulders had fragments chipped off them. Corrasion was limited to about two thirds of the length of the order three channel. The upstream limit of corrasion is marked on the map in Figure 7. In the eleven years during which records have been kept, the 28th February, 1966, precipitation of 5.8 inches in twelve hours is a record, but seven other years out of the eleven have had at least one rainfall of 3.0 inches in 24 hours. It seems then that a significant eroding flood can be expected at least once in two years.

The colluvially infilled valley floors like that at Whitehall (Figure 7) are confined to channels of order three or lower. Cotton (in Cotton and Te Punga, 1955, p. 1019) has described similar forms from the Wellington area and suggests that they are infilled by "... mass movement of a water-saturated stream of debris", and by creep from the valley sides. Berry and Ruxton (1961, p. 628) suggest that similar features in Hong Kong are the result of deforestation and deposition of sediment derived from gullying of the slopes.

In the South Auckland greywacke ranges colluvial infill is not necessarily a permanent feature of the valley floors, for until 1930 when a major flood occurred, the floor of the Waerenga Valley was infilled, but the flood cleared it out and downcutting has continued ever since (pers. comm. Mr E. Cheyne, a local farmer). Infilling of valley floors here seems to be a result of mass movement debris sliding into the valley floor and being smoothed over by surface wash, but not removed, before grass establishes itself. If the precipitation is prolonged the debris may all be entrained and removed, but if it is short it will remain until the next eroding flood. Once grass is established on the infill the erosive force needed to remove it will be great. Surface runoff can be so deep in these areas that it will flatten grass and move as a sheet down slopes and over the grassed infill, but there is no debris for it to entrain and incision cannot occur in order one or two channels and only low down those of order three.

EFFECTS OF DEFORESTATION

The slopes of the study areas were covered with a dense forest of evergreen trees of which the principal genera were **Podocarpus** and **Dacrydium** (Holloway, 1959). Towards the end of the 19th century these were cleared and European pasture grasses and clover were established in many areas. The forest which remains has been affected by introduced animals—deer, goats, pigs and opossums—so that it no longer gives as complete a protection to the slopes as formerly. In areas where pasture and forest are contiguous it is notable that recent intense rainfalls have had far less effect upon forest-covered than upon grassed slopes. Debris slides do occur in the forest both in the study areas and elsewhere (Cunningham and Arnott, 1964), but they occupy a much smaller area than those in the grasslands.

The erosion-inhibiting effects which forests have on soils and slopes can be summarised as:

(1) Tree roots keep the water table lower than do grasses and so prevent loss of cohesion in the soil.

(2) Deep roots give structural strength to the soil and resist soil movement.

(3) Tree roots produce a soil structure with more macropores, and hence greater permeability, than grass roots.

(4) Forest soils have a higher humus content to bind the clay micelles and mineral grains.

(5) Forest soils do not dry out so readily, or suffer as much volume change as grassland soils.

(6) Cracks do not form in the permanently moist soils of forests.

(7) Forest trees intercept precipitation and prevent the rapid rise of pore-water pressures.

(8) Water is detained in the forest litter, so that runoff and stream erosion is reduced.

(9) Stream banks are protected by tree roots.

(10) Tree roots do not compact the soil like grass roots, and the lack of grazing animals further reduces soil compaction.

The changes of soil structure after deforestation become more pronounced with time; soil develops grassland structures; tree roots rot; and animal treading causes terracettes and compaction. The maximum effect of deforestation reaches a peak, then, some years after the initial clearance.

The changes in soil characteristics can be illustrated by a comparison between compaction, infiltration rates and organic matter content (as measured by loss on ignition). Samples were taken from each of the three study areas along the boundary between forest and pasture. One set of samples was taken from the forest and one set from the pasture. The infiltration rates were obtained from six replications on each site after the ground had been pre-wetted by heavy rain. The length of time is that taken for 500 ml. of water to infiltrate the soil within a four-inch ring inserted to a depth of one inch. The unconfined compressive strength was obtained from twenty penetrometer readings at each site, and the percentage loss of weight on ignition was calculated after heating ten dried samples, from each site, of the top two inches of soil, at 600° C. for six hours. Litter was removed from the forest samples and green vegetable matter from the pasture samples. The pasture samples have a high organic matter content because of the presence of roots, so that the percentage of humus in the forest soils is proportionally higher than the figures suggest.

Table 2 : SOIL PHYSICAL CHARACTERISTICS

	Pasture	Forest
Mean Unconfined Compressive Strength, Tons/sq. ft. (Kg./sq. cm.)	3.04	1.46
Mean Infiltration Time for 500 ml. of Water (minutes)	14.60	3.05
Mean Percentage Loss of Weight on Ignition	8.70	10.90

It is clear from Table 2 that the pasture soils are more compacted, are less permeable and have lower humus contents than the forest soils. Each of these factors will increase their liability to erosion.

CONCLUSIONS

The dominant geomorphic process on the slopes of the South Auckland greywacke ranges, at present, is the formation of debris slides. These are largely confined to areas of pasture although they can occur in the forests. Deep slumping from an earlier period has fretted the edges of all wide interfluves but it can no longer occur because the deep regolith materials have already been removed. It is not yet clear whether the phase of slumping occurred during a period of different climate from that of the present or whether it was solely the result of unstable slope conditions soon after the deposition of the ash. Floors of order one and two channels are infilled by colluvium derived from the mass movement, and stream erosion seems to be confined to channels of order three or larger.

The concept of drainage basins whose geometry is adjusted to the forces acting upon them may be expressed in terms of a steady state in an open system. "For a given intensity of erosion processes, acting on a mass of given physical properties, the conditions of surface relief, slope and channel configuration reach a time-independent state in which the morphology is adjusted to transmit through the system just the quantity of sediment and runoff characteristically produced under the controlling regime." (Strahler, 1964). If one factor, such as vegetation, is varied, it is probable that a period of relatively-rapid adjustment of basin geometry must occur until a new steady state is achieved.

A dynamic equilibrium of energy and forms appears to be directly applicable to the landforms of the greywacke ranges. Under forest the slopes were nicely adjusted to precipitation and runoff conditions so that channels could carry the material moved during a flood with a medium return period. Between such events erosion was largely by the slow and minor processes of creep and surface wash with streams transporting the debris. Major events with long return periods—of several hundred years—could do catastrophic damage (Grant, 1965), and resulting from such damage there would be increased runoff on mass movement scars so that erosion as a whole would be accelerated. With the regrowth of vegetation and renewed soil formation there would be a return to earlier equilibrium conditions and slow eradication of the forms left by the catastrophic event.

After deforestation by man equilibrium conditions would be changed permanently. Runoff would increase and channel scouring and deposition would change. Slope forms and drainage basin geometry would not necessarily be seen to respond immediately to the disequilibrium. When next an event with a medium return period occurred, however—such as the intense rainstorm of 28th February, 1966, with a frequency—i.e. return period—of 20 years there would be a rapid readjustment within the drainage basin, but in the forested areas adjustments would be small. This accords with the small number of slides seen in the forested areas and the large number in the pastured ranges. The medium return period event has now caused a change of drainage basin geometry towards a state of finer-textured relief in accordance with the need for more channels to receive the runoff. A further step has occurred towards the attainment of new equilibrium conditions adjusted to pasture grass vegetation. Until that new equilibrium state is fully achieved—and the presence of infilled valleys suggests that it is not—it must be expected that medium return period events will cause damage in grassland areas, while forested land will be largely unaffected.

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