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Quantifying Sediment Production in Steepland Environments

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Abstract

Five published contributions to our understanding of the impacts of erosion processes on sustainable land management are reviewed and discussed. These focus on rapid shallow landsliding and gully erosion which are among the most prevalent forms of environmental degradation in New Zealand's hill country. The over-arching goal of this research has been to quantify the on-site (e.g., soil erosion, land productivity) impacts of these processes. Rather than measure erosion rates over long periods of time, geomorphic techniques such as 'space-for-time substitution' and paired catchment approaches have been employed to overcome the naturally high spatial and temporal variability in erosion processes. Digital elevation models (DEMs) have proven invaluable as a means of measuring rates of gully erosion and of quantifying landform properties at small and large catchment scales.

The spatial variability in shallow landslide erosion, recovery of soil on scar surfaces, and long-term evolution of hillslopes is investigated in an area of sandstone hill country in the North Island of New Zealand, to help elucidate the role that vegetation plays in maintaining slope stability. Hillslope evolution is primarily by landsliding on steep slopes and by diffuse creep processes on gentle slopes, punctuated by periods of slope instability related to climatic and vegetative variability. Variation in slope form near to channels, with over-steepened sideslopes and higher benches, implies a history of fluctuating erosion rates, driven by changes in stream base level. Systematic variation in soil depth and slope angle measured at the hillslope scale implies spatial variability in erosion rates and a landscape that is not in morphologic equilibrium. There is about an order of magnitude difference in long-term erosion rates between relatively steep (> 30º) and gentle (< 30º) hillslopes. Steep slopes are located at the head of 1st-order drainage basins and are more closely coupled to base level changes in the drainage network and erode at about the rate of tectonic uplift.

A novel 'paired hillslope' approach is used to directly measure the net loss of soil caused by post-deforestation landslide erosion and conclusively show that, had the forest remained, the landslides would not have occurred. Over an 85-year period since deforestation, average net soil loss per unit hillslope area was 0.15 ± 0.04 m, equivalent to a depletion rate of 1.8 ± 0.5 mm yr^{-1}. The rate was higher at 2.7 ± 0.8 mm yr^{-1} on slopes above 28° where most landslides are concentrated. Contemporary erosion rates are at least 3 to 10 times, or more, the long-term rates of erosion depending on hillslope location.

The distribution and average depths of soils to bedrock are measured and used to derive a logarithmic function of increasing soil depth with landslide scar age. The rate of soil recovery is found to diminish with time: from 3.5 mm yr^{-1} over the first 40 years after slipping to 1.2 mm yr^{-1} over the following 50 years. The functional form of the relationship implies continued decrease in the accumulation rate with time. When averaged at the scale of hillslopes, accumulation rates are much less than erosion rates implying that there are semipermanent losses in soil depth and associated soil properties, something that has important consequences for sustainable land management of this landslide prone hill country.

A further study investigates the methodology and errors involved in determining the amount of sediment produced from gully complexes in catchments at the headwaters of the Waipaoa River basin. Multi-date DEMs constructed from historical aerial photography are used to determine the volume and mass of sediment produced from gullies of varying size over timescales of decades. Results show that the average denudation rate of gullies is proportional to the square root of gully area: large gullies are producing a disproportionately large amount of sediment, not only because of their larger area, but also because of their higher denudation rates. This simple power-law relationship provides an efficient and quick means of estimating the total amount of sediment producted from gullies (of similar morphology) at the catchment scale, from the measurement of gully area alone, and this has direct application to catchment scale sediment budgets.

Many of the smaller gullies were found to have stabilized, became smaller, and reduced in sediment yield after tree plantations were established, demonstrating that reforestation should be a major

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conservation strategy wherever gully erosion is encountered. Large gullies produce a disproportionately large amount of sediment and are less affected by reforestation. Measured sediment yields from the two largest gullies when compared to the basin yield, demonstrate that these gullies have neither individually nor collectively dominated the sediment budget of the Waipaoa River. This is because greater contributions are made by numerous contemporary gullies and/or by diverse sources in other parts of the Waipaoa River basin.

Research findings presented in this paper provide information that is valuable for the sustainable land management of areas subject to accelerated erosion processes. Being able to identify and map areas of land that are sensitive to disturbance, and evaluate the on-site and off-site impacts, is an important part of assessing soil sustainability in these steep topographic settings.

Key words: Erosion; Landform evolution; Landslide; Gully; Sediment budgets

1. GENERAL INTRODUCTION

Erosion and associated soil degradation are among the most pervasive environmental problems in many areas of the world, causing loss of soil productivity, pollution of rivers, sedimentation in water storage and lagoons, and regularly causing loss of life and millions of dollars of damage, particularly in areas affected by mass movement processes. Both on-site (productivity losses, damage to buildings, roads and farm infrastructure) and off-site (water quality, flooding, sedimentation) are the main consequences of erosion processes. Permanent losses in land productivity associated with depleted soils, together with poor water quality from increased levels of sediment and nutrients in rivers, are major concerns of catchment and land management agencies worldwide. Yet erosion is a also key component of the natural functioning of land environments. Without it there can be no down-wearing of landscapes, construction of floodplains, replenishment of soil on hillslopes and floodplains, bedload transport and bed-armoring of rivers, long-shore transport, coastal accretion and natural protection of coastlines. Understanding the relative size and impacts of contemporary erosion processes to their background or natural rates is critical in evaluating the role of man induced (anthropogenic) land use changes on the natural environment.

Unfortunately, the measurement of erosion rates is not an easy task owing to the naturally high spatial and temporal variability in these processes. Added to this is the complex shape of many erosion features (e.g., landslides), which limits the accuracy with which erosion measurements can be made, and means that particularly innovative methods are often required when trying to assess contemporary rates of erosion relative to background rates. There has been a need to better develop methodologies for directly measuring erosion rates and in particular quantifying the impact that woody or forest vegetation has on reducing rates compared with grasslands. There has also been a need to be able to integrate research finding into sediment budget applications (Reid and Trustrum, 2002; Reid and Dunne, 2003) and scale the results of field trials, which are generally undertaken at the hillslope scale, to the whole of catchments. This is necessary in order to determine the relative contributions that different erosion processes make to river sediment loads and sediment export from river basins.

Empirical models which relate erosion intensity to terrain, geology and soils, vegetation, land management and climatic characteristics are generally required to make predictions about erosion rates over large spatial scales. Perhaps the best known and most widely used model of this type is the Universal Soil Loss Equation (USLE) for predicting rates of sheetwash and rill erosion from arable fields (Wischmeier and Smith, 1978). Model development was based on thousands of observations from runoff plots. Such broadly applicable models have generally not been developed for other erosion processes, mostly because of the limited availability of the large datasets necessary to develop and validate such models.

The main goal of the research undertaken here was primarily to develop and improve methods for quantifying landslide and gully mass movement erosion processes in New Zealand steeplands. Important aspects of the work were to: (1) directly measure rates of erosion; (2) determine the affect of vegetation cover in mitigating erosion rates; and (3) develop methods for extrapolating research findings to larger spatial scales. To determine the impact that deforestation has on erosion requires that the natural or background rates of erosion are known. Since it is not possible to directly measure erosion from the past, studies of hillslope stability and evolution are necessary in order to make inferences about long-term erosion rates. The production and accumulation of soil materials from bedrock is an important process that replenishes soil materials lost through erosion and controls the long-term mass balance of soil materials sitting on hillslopes. This paper therefore also investigates soil weathering and production on hillslopes subject to recurrent landslide erosion.

Research was conducted in two regions of the North Island of New Zealand, these being: (1) eastern Taranaki hill country where shallow landsliding is the dominant erosion process (Figure 2); and (2) East Coast hill country in the Raukumara Range where spectacular gully complexes are one of a number of processes contributing to elevated river sediment loads (Figure 20). Both regions have experienced major anthropogenic disturbances in historical times with widespread clearance of the native forest vegetation communities and conversion to grasslands for grazing animals.

1.1 Research background

Mass movement erosion is particularly prevalent on hillslopes in tectonically active regions of the world underlain by weak, crushed and sheared or poorly consolidated lithologies, and typically contributes high sediment loads to coastal margins. Countries which are located along the plate margins of the Pacific Ocean, or so called Pacific Ring of Fire, such as New Zealand, Japan, and the Pacific Northwest of America are particularly prone to these forms of erosion due to the predominance of steeply sloping and mountainous physiographies. In these regions, the combination of high relief, steep slopes, frequent heavy rainfall and large earthquakes, often produce potentially hazardous mass movement phenomena.

Principle types of erosion common to New Zealand hill country and steeplands are shallow landslides (commonly referred to as soil slips), deeper seated landslides, rotational slumps, earthflows, soil creep, tunnel gullying along ephemeral watercourses, and much larger gully complexes incorporating slumping and sheet processes. There can be considerable variability in the type and intensity of erosion processes depending on local physiography. Landslides are particularly common on steep slopes underlain by uplifted Tertiary marine sediments, whereas gully erosion is much more common on hillslopes underlain by the older, generally more crushed and sheared, Cretaceous sediments. Earthflows are prevalent on the more gently sloping terrain underlain by clay rich rock types. Landslide and gully erosion, account for much of the sediment delivered to streams in mountainous regions and contribute to the high sediment yields of coastal rivers. Some rivers, such as the Waipaoa, have among the highest specific sediment yields documented worldwide (Walling and Webb, 1996; Hicks *et al*., 2000). Although such high yields are undoubtedly sustained by high tectonic uplift rates over geologic timescales, a significant component of contemporary loads can be attributed to relatively recent catchment disturbance and inappropriate land management practices.

Inappropriate land use activities, such as removal of forest cover, can result in accelerated erosion above their natural background rates. In New Zealand, mass movement processes predominate on steep hillslopes that have been cleared of the native forest cover and converted to grasslands for livestock production. These lowland steeplands occupy more than 6 million hectares of the North Island and more than 40% of New Zealand's land area (Blaschke *et al*., 1992). Evergreen rain forests covered much of New Zealand's land surface during the Holocene. Although removal of this forest began over 1000 years ago with the arrival of the first Polynesians (McGlone, 1983) a new era of forest removal began when large numbers of European settlers arrived in the 1840s. Between 1847 and 1909 approximately 4.5 million hectares of forest or 20% of the total area of New Zealand was converted from forest to grassland in order to provide feed for large numbers of introduced grazing animals (Masters *et al*., 1957).

In the North Island much of this conversion to grassland took place on hill country comprised of poorly consolidated Tertiary and Cretaceous marine sedimentary rocks (predominantly mudstones and sandstones) which have been tectonically uplifted many hundreds of metres during the Quaternary. These sediments have become deeply dissected by erosion since uplift began in the late-Tertiary (Neogene Period). Erosion continues to the present day and it is estimated that some 36% of the total area of New Zealand is affected by mass movement processes (Blaschke *et al*., 1992). Because of the generally low population densities in New Zealand, mass movement processes mainly affect rural farming communities and transport links (road and rail), although deep-seated mass movements can on occasion cause considerable infrastructural damage in urban environments.

Due to the prevalence of mass movement erosion in New Zealand and the problems it can cause, in the past there has been considerable efforts by research and catchment management agencies at recognizing and mitigating the impacts of erosion. These had tended to focus on qualitative broad-scale mapping exercises such as the New Zealand Land Resource Inventory (Eyles, 1985; Eyles and Newsome, 1990) or localized on-farm conservation efforts which generally involve space planting with broad leaf conservation species (Populus spp., Salix spp.) or afforestation, principally with Pinus Radiata. Despite on-going debate over the effectiveness of conservation programs, very few studies were undertaken to address this issue (Hawley and Dymond, 1988). There was also an increasing recognition over the link between on-site erosion and off-site impacts, particularly as during large storm damage events such as Cyclone Bola in 1988, the scale of flooding and sedimentation in lowlands was considered a consequence of the lack of forest cover and widespread erosion in upland catchments. These issues provided the primary impetus for the research undertaken in this paper. Although the stabilizing influence that forest or woody vegetation has on hillslope erosion is now well understood (Marden and Rowan, 1993; Marden *et al*., 2005), at the time the research presented in this review paper was undertaken, there was a general lack of quantitative information about the relative importance of vegetation at controlling erosion rates and in controlling sediment related off-site impacts. In East Coast rivers for example, the erosion rates from gullies were largely unknown and consequently the contribution that gullies made to river sediment loads could only be guessed. Similarly, although conservation plantations were shown to decrease the size and extent of some gullies, the long-term impacts of afforestation at reducing river sediment loads could not be predicted with any degree of certainty.

Similarly, with respect to landslide erosion, although various studies had shown that landslide densities were lower in forested terrain compared with grasslands (O'Loughlin and Pearce, 1976; Pain and Stephens, 1990; Hicks, 1991; Blaschke *et al*., 1992), the results of these studies were often inconclusive, because other

factors such as terrain characteristics and rainfall patterns were also known to influence the distribution and density of landslides. Furthermore, most of these studies were undertaken following individual storm events and did not consider the cumulative impacts over longer periods of time. Although differences in river load (Hicks, 1988; Quinn and Stroud, 2002) and lake sedimentation (Page and Trustrum, 1997) generally support the results from landslide studies, indicating similar relative increases in the magnitude of erosion rates, it is often unclear which of the erosion process occurring within catchments is primarily responsible for these changes. Time lags between source and delivery and variability in climate or tectonic activity can further complicate the assessment of erosion rates. In general, sediment deposits cannot distinguish between specific erosion processes or source areas, unless the eroding sediments and subsequent deposits have distinctive chemical or mineralogical signatures. The link between hillslope erosion processes and off-site sedimentation in lowlands and coastal regions continues to be the subject of on-going research (Gomez *et al*., 1999; Orpin, 2004; Gomez *et al*., 2004; Marden *et al*., 2008).

1.2 Paper structure

The research summarized here is based on five publications (Trustrum and De Rose, 1988; De Rose *et al*., 1991; De Rose *et al*., 1995; De Rose *et al*., 1998; Betts and De Rose, 1999). Sections 2 to 5 focus on shallow landslide erosion while Section 6 in concerned with the measurement of gully erosion.

The overall manuscript layout is illustrated below in Figure 1. The second section describes methods used in the measurement of the soil recovery process on landslide scars within eastern Taranaki hill country. Results are then used to assess the contemporary and long-term impacts of landslide erosion on the mass balance of soil materials remaining on hillslopes. Section 3 investigates the long-term evolution of hillslopes as a means of determining the background erosion rates for this region. Measurements of the spatial variability in soil depth and slope form, coupled with stratigraphic dating, are used to demonstrate morphologic disequilibrium for hillslopes, and significant spatial and temporal variability in erosion. Sections 2 and 3 have been presented prior to the study on measurement of contemporary landslide erosion rates in Section 4 as they provide the necessary background information and context for this research, which was undertaken several years after the initial work on soil recovery and slope evolution.

Section 4 conclusively demonstrates the importance of forest vegetation in maintaining slope stability and provides a direct measure of the rates of erosion caused by landsliding. Section 5 investigates the potential of both empirically derived landslide susceptibility models and limit-equilibrium slope stability models as a means of predicting landslide erosion at larger spatial scales. The existence of threshold slopes for landslide failure provides a means of predicting landslide susceptibility using high resolution digital elevation models (DEMs).

Section 6 describes the methods used and results of measurement of gully erosion in East Coast catchments from multi-date DEMs. Results are discussed within context of the relative contribution that gullies make to river sediment loads and catchment sediment export. The role of vegetation in stabilizing gullies is also investigated. Gully erosion rate - area relations are developed as a means of extrapolating research findings, and calculating the cumulative impact of gully erosion, at larger spatial scales. The research results are summarized within the concluding Section 7.

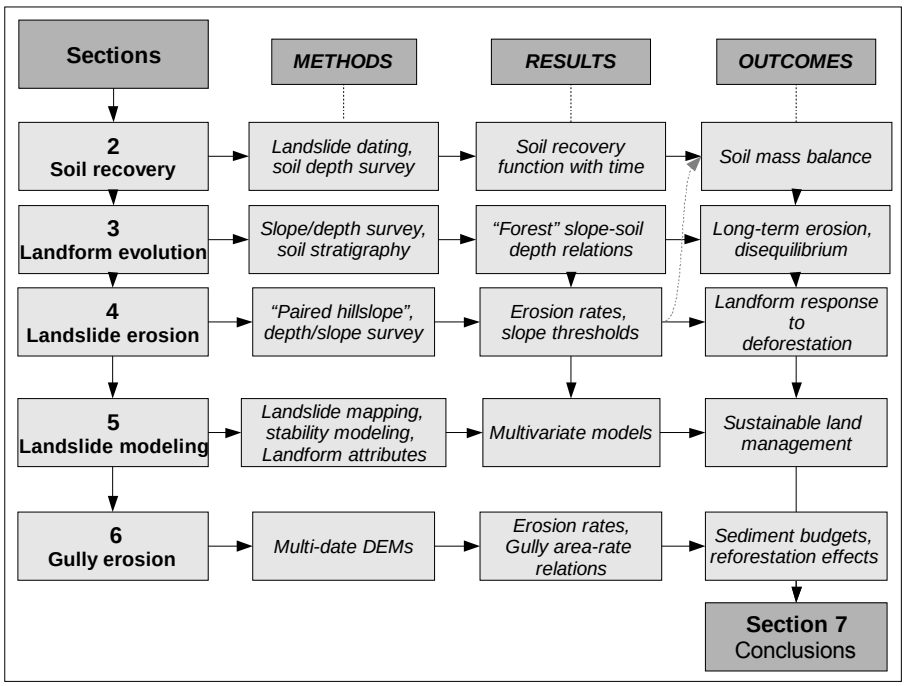

Fig. 1. Structure of study.

2. THE SOIL RECOVERY PROCESS

2.1 Introduction

The weathering, formation, and accumulation of soil material from bedrock is one of the most fundamental processes operating on soil-mantled hillslopes, yet it is one of the most poorly understood or quantified. A knowledge of soil accumulation rates is important for understanding the long-term consequences of soil erosion, yet there are few published examples (Minasny and McBratney, 1999). In New Zealand, most investigations of hillslope erosion have tended to emphasize rates of soil denudation and their implications (Selby, 1982; Crozier, 1986; Pearce, 1986) while far less attention has been paid to rates of soil recovery. Research was therefore required to address the paucity in information about the soil recovery process following landslide erosion.

The soil chronosequence (Birkeland, 1984) in which soil properties show consistent trends with age, is used as a conceptual framework to examine soil recovery on landslides. In particular, soil depth provides an index of pedogenic development in which rates of soil formation are calculated from changes in mean soil depth over time. Study sites within eastern Taranaki hill country were used to explore the nature of the soil recovery process on landslides and how this is likely to influence landslide immunity and recurrence intervals. This site was selected because of a well established landslide chronology and the ease and accuracy with which measurement of soil depth to the underlying consolidated sandstone bedrock can be made. Furthermore, landslides usually remove soil to bedrock, providing a common base line (i.e., time zero) for the onset of soil formation.

2.2 Study sites

Two locations at Tutatawa and Makahu within eastern Taranaki hill country were selected for study because of the availability of aerial photography and history of landslide erosion (Figure 2). Both locations have similar slope morphology, climate, soils, vegetation and history of land management.

Eastern Taranaki hill country is characterized by fluvially dissected Upper Tertiary marine silty sandstones with interbedded mudstones and conglomerate shellbeds. Dissection of these sediments over more than 600,000 years, at a regional uplift of 0.5 – 1.0 mm yr^{-1} (Pillans 1986), has led to a fine dendritic drainage pattern and the development of steep, largely rectilinear hillslopes. Small alluvial terraces of variable height above current stream levels are common along major stream and river courses. Gentle tephra mantled surfaces, that may be remnants of an earlier more mature landscape, occur furthermost from streams along major ridge systems at elevations of about 300 m (130 m above current stream elevation).

Most hillslopes have a crenelated microtopography of alternating spurs and swales (Figure 3). Spurs (convex contour) parallel swales and are often continuous from ridge to valley. Valley sideslopes vary in length from 50 to 250 m and extend from narrow (1 – 2 m) ridges down to similarly narrow valleys. Swales (concave contour) equate with zero-order drainage basins. Swales tend to be pear-shaped and converge towards the valley bottom where they are at their narrowest. In general soils are free draining, have silt loam to loam textures, are friable, of medium to low bulk density, moderate to strongly acid and have high phosphate retention. However, soil properties are extremely variable due to hillslope position and erosion, and additions of airfall andesitic ashes erupted from Mt Taranaki during the late Pleistocene and Holocene (Geddes and Neall, 1982; Alloway, 1989). There is commonly little obvious lithological variation on hillslopes. In general, soils offer little resistance to penetration compared with the underlying more consolidated sandstone bedrock, and it is a relatively easy task to measure soil depths using cone-tipped impact penetrometers.

Removal of the native podocarp forest communities and conversion to grasslands has occurred sporadically since initial European settlement in the late 19th century. Widespread forest removal occurred in the early 20th century, but much of this land reverted during the Great Depression in the 1930's to a shrubland comprising

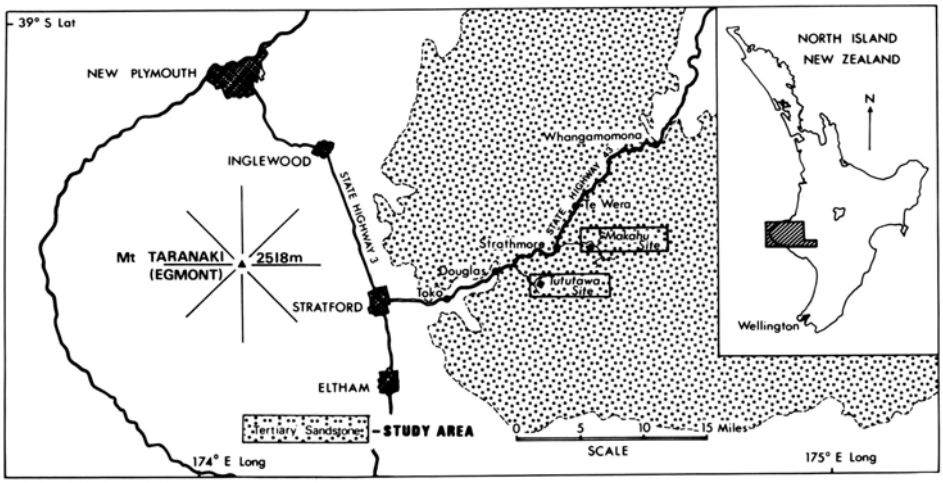

Fig. 2. Eastern Taranaki hill country (from, Trustrum and De Rose, 1988).

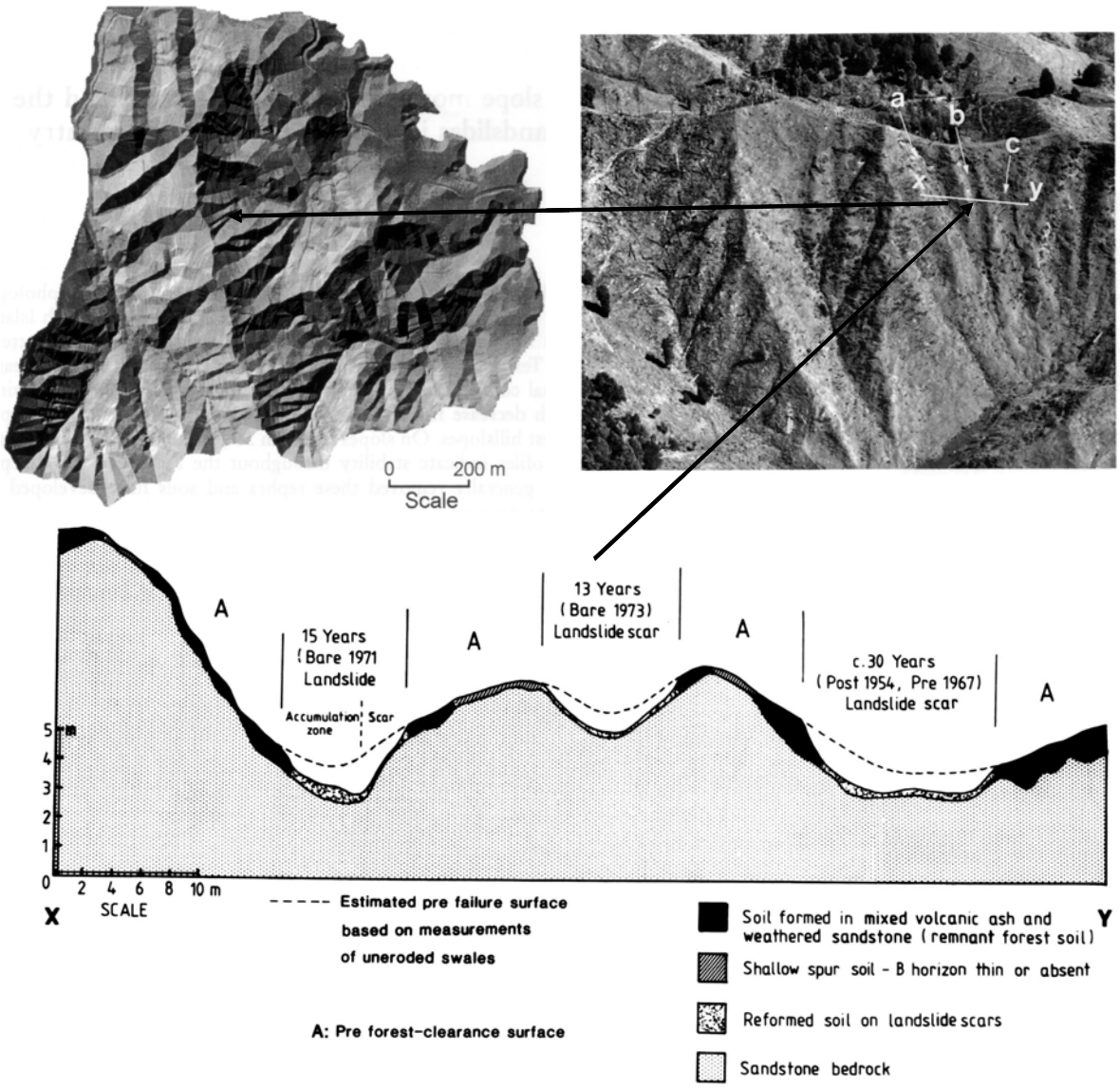

Fig. 3. Survey location (top right), shaded DEM (top left), and cross-section (bottom) across a hillslope at Makahu (from, Trustrum and De Rose, 1988; De Rose, 1996).

regenerating native species. Government land development grants in the 1970's saw a new wave of forest clearance with much of this land being cleared again.

Following clearance of native forest and conversion to grassland, numerous shallow landslides were triggered by high intensity rainfall events. Landslides of different age occur on mid and upper slopes where they have clearly discernible scar margins. They occur most frequently in swales and have an elongate form with the long axis parallel to spurs. The return period of meteorological events causing wide-spread landsliding on deforested hillslopes is about 10-20 years. Contemporary reports indicate that landslides are triggered by rainfalls in excess of about 120 mm over 24 hours. Annual rainfall is about 1800 mm.

2.3 Methods

Landslide scars were first identified and mapped onto 1:7000 scale aerial photographs. Analysis of photographs indicated that, since forest removal began in the late 19th century, there had been at least eight landslide events in Taranaki hill country. Sequential aerial photographs, historical terrestrial photographs and documented erosion and rainfall events were used to locate and date 14 post-deforestation landslide scars. These ranged in age from 13 to 82 years and provide age control for onset of soil development on scar surfaces.

For each of the dated landslide scars, soil depths to bedrock were measured over a 1 m interval grid spacing using an 18 mm cone-tipped metal probe inserted normal to slope by hand. The survey grid had previously been laid out with flexible survey tapes. Depths were verified by occasional soil excavation. For each landslide site, between 40 and 350 measurements provided a reliable estimate of the mean and standard deviation of soil depth. For each scar, the mean soil depth was calculated and plotted against landslide scar age.

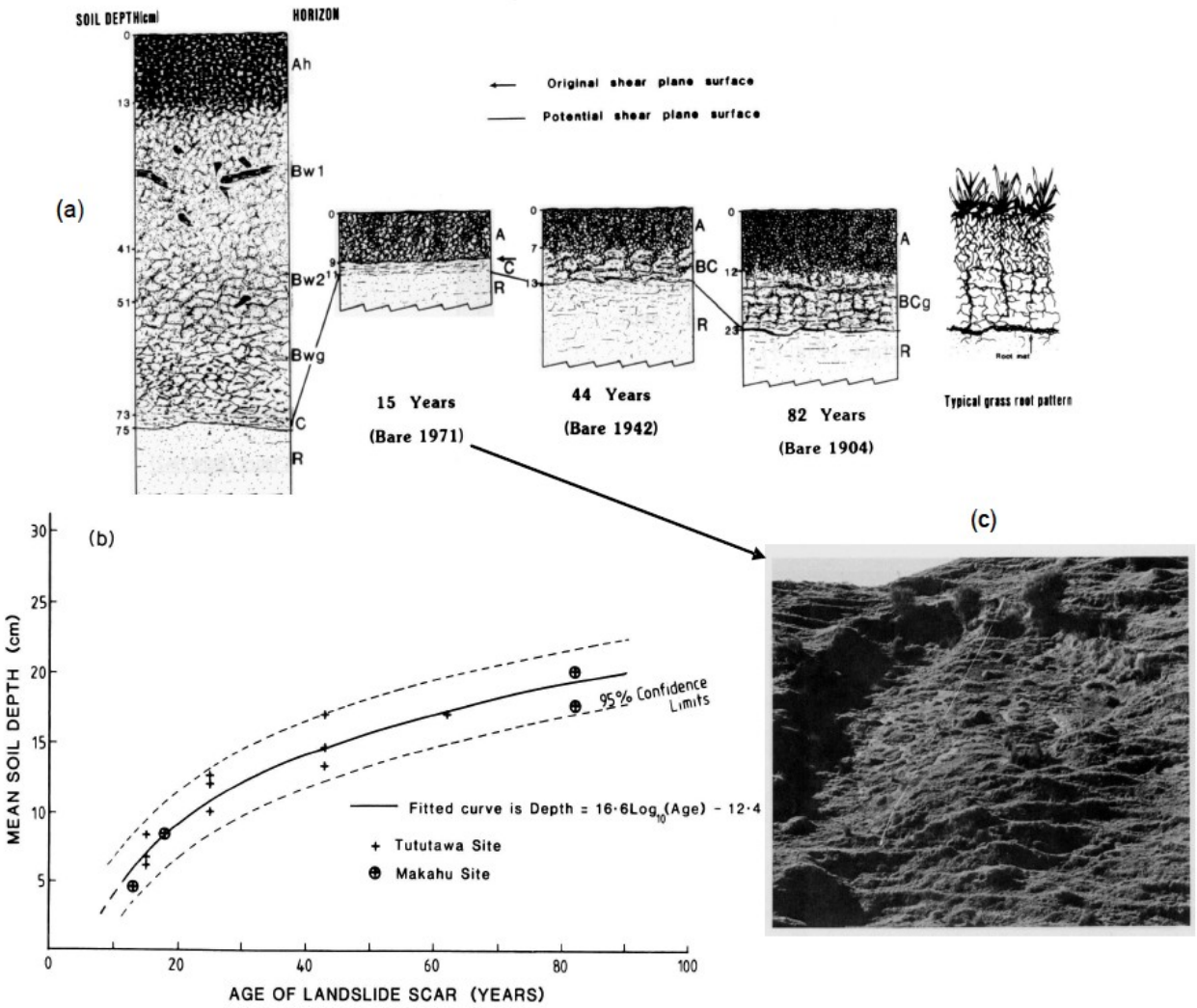

Fig. 4. Soil recovery on landslide scars: a) Soil formation; b) Soil depth chronofunction; c) 1971 landslide scar in 1986 (photo: N.A. Trustrum) (from, Trustrum and De Rose, 1988; reproduced in Soons and Selby, 1992).

2.4 Results

Systematic changes in soil morphology and vegetation were observed with landslide age and these provided some important insights into the process of soil formation (Figure 4). Soil formation is primarily a result of both bedrock weathering and accumulation of colluvium derived from surface fragmentation of exposed bedrock and crumbling scar margins. In early stages, increase in soil depth is due largely to colluvial transport. On bare areas a characteristic platy structured rind forms by chemical and physical weathering of bedrock. Wetting and drying cycles and stock trampling cause the rind to flake and the resultant sediment is washed downslope, together with material derived from scar headwalls, contributing to colluvial accumulation on "island" areas (Figure 4c). After about 5-10 years lichens and mosses colonize these bare sites and help trap colluvial sediments. Deposition of colluvium further decreases as the head and side-walls of scars revegetate, first with flatweeds and then grasses. Further increase in soil depth is related to both the weathering of bedrock and colluvial accumulation.

Soil depth was found to increase with landslide age (Figure 4b). Mean soil depth increased from 5 cm on 15-year old scars to 20 cm on 82-year old scars. Measured soil depths were attributed to "rafted" (translocated/residual) soil, colluvium, and bedrock weathering. Since areas of "rafted" soil do not represent development from bare rock, the mean depths were calculated by excluding the deeper (upper quartile) measurements. A soil chronofunction was derived by regressing mean soil depth against the logarithm of known scar age. The rate of soil accumulation averaged 3.5 mm yr^{-1} over the first 40 years after slipping, but dropped to 1.2 mm yr^{-1} over the ensuing 50 years. This trend, which can be described by a reciprocal function, suggests that the rate of soil formation further decreases beyond 90 years. This chronofunction can give an estimate of the age of other landslide scars to within ± 27 % for ages up to 90 years.

2.5 Discussion

The research has shown that by placing landslide scars in chronological order, and regressing soil depth

against scar age, it is possible to derive a soil chronofunction of increasing mean soil depth with scar age. The results importantly show that, although rates of soil recovery are initially rapid, there is a sharp decline after about 40 years once scars become revegetated. Repeated landsliding on previously failed surfaces has not been observed, suggesting a minimum landslide recurrence interval of 82 years. Although soils will continue to accumulate beyond 82 years, it will take a considerable length of time to replace the original forest soils which average about 1 metre in depth.

To compare soil accumulation rates with erosion rates, requires scaling to the whole of hillslope areas. Landslides occupy at most 30% by area of the steeper hillslopes (section 5). The average accumulation rate of soil for a hillslope will therefore be under one third that observed for individual landslide scars, and will depend on the age distribution of scars. For example, assuming landslides average 40 years in age and have a mean soil accumulation rate of 1.9 mm yr^{-1}, then the average accumulation rate for a hillslope covered with 30% of landslides would be 0.57 mm yr^{-1}. This represents a maximum rate since most hillslopes contain fewer landslides. The average landslide erosion rate for steep slopes (2.7 mm yr^{-1}) is nearly 5 times the rate of soil accumulation demonstrating that landslide erosion is leading to semi-permanent reductions in soil depth over time. Obviously, average hillslope soil depths will continue to decline as more landslides scars occupy hillslope areas.

Similar soil recovery processes and functional forms have been observed elsewhere. For example, Shimokawa (1984) found the soil depth to bedrock to increase from an average 0.19 m at the time of landsliding to around 0.78 m after 250 years, on steep (30 – 40°) forested granodiorite slopes in southern Japan. Increase in soil depth was most rapid at about 2.7 mm yr^{-1} in the first 150 years, declining to about 1.8 mm yr^{-1} thereafter. Similarly, Smale *et al*., (1997) found that soil depth increased from an average 0.20 m at 10 years to 0.58 m at 70 years after landsliding under regenerating Kanuka (Kunzea spp.) forest in East Cape, New Zealand, equivalent to a soil accumulation rate of 6.3 mm yr^{-1}. Blaschke (1989) found an increase in average soil depth to bedrock with increasing land surface age to about 1000 years on forested slopes in eastern Taranaki hill country (summarized in Trustrum *et al*., 1990). Although additional mechanisms were responsible for forest disturbance other than landsliding (e.g., wind-throw), the general trend and functional form of soil recovery was consistent with what had been observed on recent landslides (Figure 4). A soil depth of 150 mm on 25 year old landslide scars, however, suggested a soil accumulation rate about 50% faster than on equivalent landslides on pasture hillslopes. This was attributed to greater residual depths of soil remaining on scars (which were not excluded from the analysis as with pasture scars), greater rates of colluvial transport of soil and organic materials, and possibly a greater ability for the roots of regenerating tree species to penetrate into the sandstone bedrock and promote weathering (Trustrum *et al*., 1990).

The soil recovery studies on landslides (Shimokawa, 1984; Trustrum and De Rose, 1988; Blaschke, 1989; Smale *et al*., 1997) all indicate relatively rapid rates of initial soil recovery, but that these rates decrease with time. The results from Taranaki hill country suggest that increases in depth are due to both surface transport (diffuse creep and headwall collapse) and to weathering of the underlying sandstone. The rapid decrease in soil accumulation rates is coincident with revegetation of both the scar surface and headwalls. Significant depths of soil material can also remain on scars at the time of landsliding, and this may help accelerate soil development. Smale *et al*., (1997) attributed the higher rates of soil accumulation on the landslide scars they investigated to the softer mudstone bedrock which tends to fritter rapidly into 1-2 cm blocks upon exposure to the atmosphere and to greater rates of headwall collapse of the strongly pedal mudstone soils. Furthermore, their soil recovery function was more linear in form when compared with sandstone lithologies, suggesting that a significant depth of soil material had remained on slip scars at the time of landsliding. The high initial rates after bedrock exposure are therefore most likely due to surface transport of preexisting soil materials from exposed headwalls and from residual material remaining on the scars at the time of landsliding. Pillans (1997), who compiled data on rock weathering and soil development rates from numerous sources, suggested that rates of soil accumulation > 0.1 mm yr^{-1} occur on either weakly consolidated (e.g., dune sands) or transported (alluvial/colluvial) parent materials.

Since a significant component of the accumulating soil materials is derived from surface transport, then rates of soil accumulation on landslide scars will overestimate the rate at which soil is being replaced by the weathering of underlying bedrock materials. Recently, Heimsath *et al*., (1999, 2000) developed a method to estimate the rates of soil production from bedrock, based on measurements of cosmogenic nuclide ^{10}Be and ^{26}Al concentrations in quartz grains, at the base of soil columns. Convex (nose) slope morphologies were selected for sampling as they were assumed to have equilibrium soils depths and were free of colluvial deposition: losing soil primarily through diffuse creep processes. Their results suggest an order of magnitude reduction in the rate of soil production as soils deepen. They have termed this a "soil production function". For example, on greywacke sandstones in Tennessee Valley, Marin County, California (Heimsath *et al*., 1999), soil production was found to decline exponentially from 0.077 mm yr^{-1} with no soil mantle to 0.0077 mm yr^{-1} under 1 m of soil. Heimsath *et al*., (2000) similarly found an exponential decrease in soil production rate from 0.053 to 0.008 mm yr^{-1} with increasing depth from 0 to 1 m depth, on granitic derived soils at the base of the eastern Australian escarpment. These rates of soil production are much less than the soil accumulation rates derived from soil chronosequences. Although the regions examined by Heimsath (1999, 2000) have lower annual rainfall (760

and 910 mm, respectively) and are somewhat different in terms of bedrock lithologies compared with Taranaki hill country, their results support the notion that the high initial rates of soil accumulation observed on landslide scars are due mostly to surface transport of soil materials from surrounding land (scar margins).

Although a logarithmic function has been used to describe the recovery process, other functions provide similarly good fits to the spread of data points (Figure 5). Both negative exponential (Equation 2) and power series on the square root of time functions (Equation 3) (Minasny and McBratney, 1999) can be also used to predict soil depths from landslide scar age with a similar level of accuracy.

$$Depth\ (mm) = 166\mathrm{Log}_{10}t - 124 \quad (1)$$

$$Depth\ (mm) = 232(0.83 - \mathrm{Exp}(-0.042\ t)) \quad (2)$$

$$Depth\ (mm) = 19.8\sqrt{t} + 0.23t \quad (3)$$

While it matters little which of the equation is used within the observed range of scar ages, each equation makes different future predictions about likely soil depths beyond 82 years. For example, at 1000 years, soil depths of 374 mm, 193 mm, and 866 mm are predicted by the logarithmic, exponential and square root functions, respectively. Clearly other methods are needed to establish soil accumulation rates over longer periods of time.

Land surface ages can be estimated using dendrochronology and studies of forest succession (Blaschke, 1989). Average soil depths measured under mid-aged forest communities in Taranaki hill country (250 to 650 years B.P.) have been found to increase by 0.15 m, equivalent to an accumulation rate of about 0.4 mm yr^{-1} (Trustrum *et al.*, 1990). A further increase in average soil depth from 0.9 to 1.10 m for the oldest aged forest community at 1000 yrs B.P., suggests that soils continue to accumulate, but at a rate less than 0.4 mm yr^{-1}. Since these sites are well vegetated, and colluvial transport onto and off monitoring sites can be assumed similar, then soil accumulation is most likely the result of weathering from underlying bedrock. This places an upper limit on the rate of soil production for this depth of soil at about 0.4 mm yr^{-1} (equivalent to a denudation rate of 0.2 mm yr^{-1}, assuming a 2:1 ratio of dry bulk density between soil and bedrock).

Tephrochronological dating of colluvial hollows in Taranaki hill country (section 4) shows that the deeper soils (1 – 1.5 m) have been accumulating soil material throughout the Holocene period at an average rate of about 0.1 mm yr^{-1}.

Combining the rates of accumulation observed on recent landslide scars with those from mid-aged forest communities (Blascke, 1989) and tephrochronological dating of colluvial hollows, yields a reciprocal function of declining accumulation rate with time (Figure 6). This function suggests that there is at least an order of magnitude decrease in the rate of soil accumulation over the first 500 years after bedrock exposure by landsliding, consistent with the general model of decrease in the rate of mechanical weathering as soils deepen (Ahnert, 1977; Heimsath *et al.*, 1999, 2000). Both the rate of soil accumulation and production from bedrock is seen to decrease with time. Although the relative contribution from both processes remains unknown in Taranaki hill country, bedrock weathering is likely to make up the greater proportion of total accumulation as time passes, because a stable vegetative cover will tend to limit the amount of surface transport on older slopes.

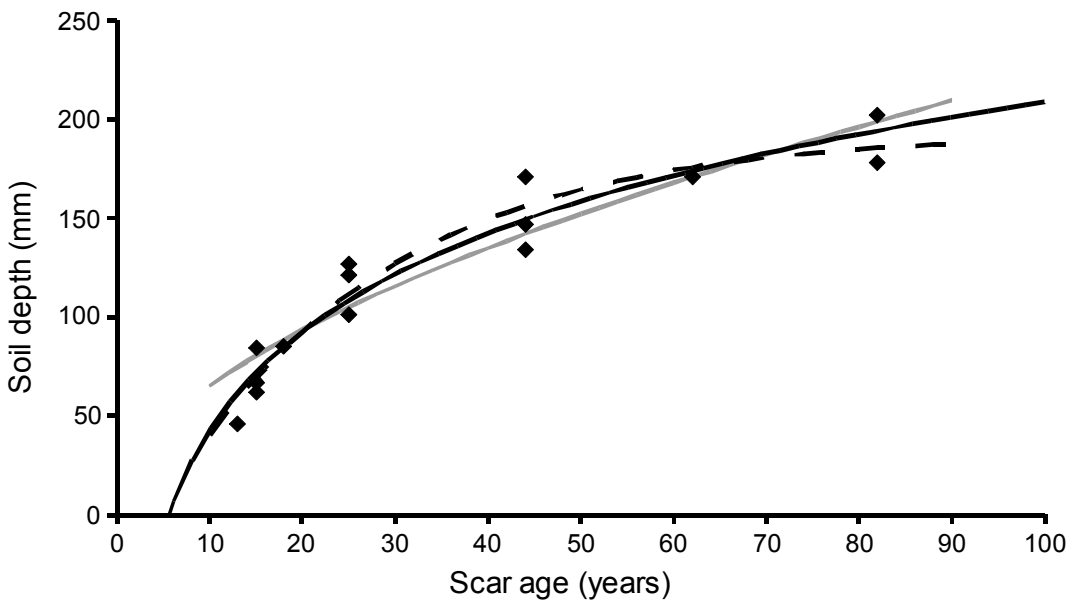

Fig. 5. Relation between scar age and soil depth fitted using logarithmic (black line), negative exponential (dashed line), and square root (grey line) functions.

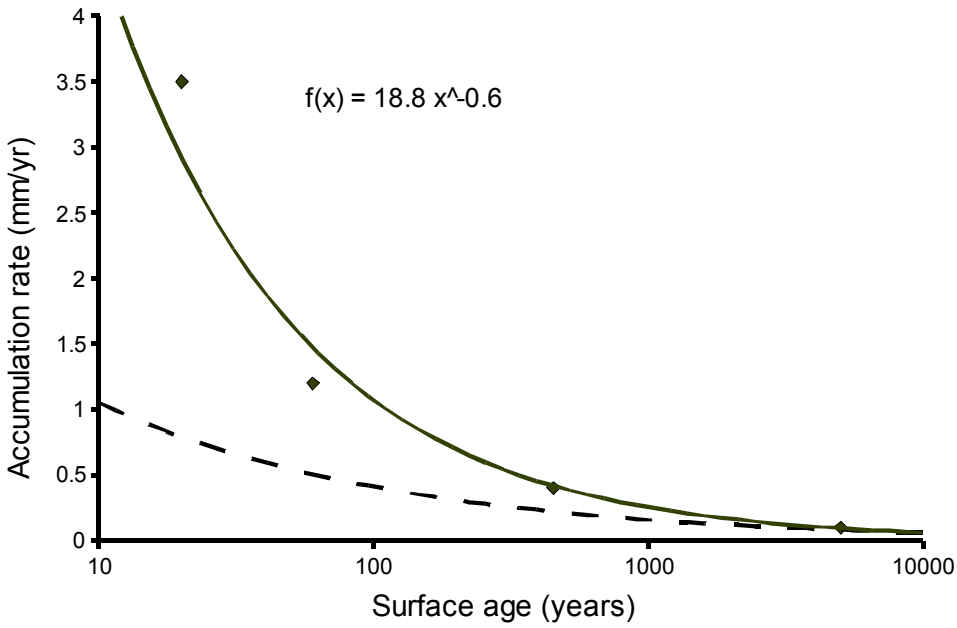

Fig. 6. Relation between soil accumulation (solid line) and probable bedrock production (dashed line) rates on consolidated sandstone.

These results are important in interpreting the long-term evolution of hillslopes, which is the subject of the next section. The relationships between surface age, soil depth and soil production and accumulation rates place important controls on the long-term erosion rates in weathering limited situations.

3. LANDFORM EVOLUTION

3.1 Introduction

Knowledge of the long-term rates of erosion for any region is essential for evaluating the relative magnitude of contemporary erosion processes and anthropogenic disturbances. In theory, the evolution of soil-mantled hillslopes is considered a weathering limited process (Carson and Kirkby, 1972), that is, the rate of denudation is controlled by the rate at which bedrock is converted to soil, rather than by erosion of soil from slopes. For soil to persist it must be replenished at a rate greater than or equal to that of soil erosion. Rates of soil production therefore place an upper limit on the long-term rates of erosion that hillslopes can sustain. Hence, if weathering and soil production rates are known, then this should also provide a measure of the long-term rates of erosion from hillslopes.

In general, the methodology presented in section 2 cannot be applied to periods of time much older than 100-years because landslide scars become difficult to identify and date with any certainty. Instead, approximate estimates of long-term erosion rates rely on studies of slope evolution where soil coverbeds and slope deposits are dated and hillslopes are placed in some sort of assumed evolutionary sequence. Under conditions of dynamic equilibrium it is often assumed that all parts of a landscape should be downwasting at a uniform rate, denudation is balanced by vertical uplift and there is no change in basin wide relief over time (Strahler, 1950; Hack, 1960; Reneau and Dietrich, 1991). Long-term rates of landform denudation can therefore be inferred from rates of tectonic uplift.

This section presents a study of slope evolution within Eastern Taranaki hill country and takes a detailed look at the spatial variability in slope morphology and related distribution of soils. Methods are designed to test assumptions about the concepts of uniform downwaisting and dynamic equilibrium and to see whether or not these assumptions hold for the shorter timescales associated with soil development. It is an extension of the work presented in section 2 and examines the process of soil recovery over much longer timescales than considered by contemporary landsliding.

3.2 Study site

Research was undertaken at the Makahu site in eastern Taranaki hill country, the general characteristics of which are described in section 2 (Figure 2). Uplift rates in this region are in the range of 0.5 – 1.0 mm yr^{-1} (Pillans, 1986). This site proved ideal for studying the process of landform evolution because the consolidated sandstone lithology is generally not subject to deep-seated slope failures and subsidence, and the form of hillslopes is a consequence of shallow erosion processes operating over long time-scales. Furthermore, the generally sharp rock contact between friable soil and underlying compact massive sandstone, means that soils depths to bedrock can be easily measured using impact penetrometers.

Some limited stratigraphic dating of coverbeds sequences and colluvial slope deposits has been undertaken in the past (Trustrum *et al*., 1989). Although radio-carbon (^{14}C) dating of woody material is the common technique employed in dating slope deposits

(Reneau *et al*., 1991), dateable woody material has not survived at the base of colluvial hollows in Taranaki hill country. Age control was instead provided by tephrochronological dating.

In regions peripheral to volcanic centres, tephra deposited on the surrounding topography from past eruptions, can provide a valuable set of chronohorizons with which to date cover bed sequences (Yanai, 1989; Eden *et al*., 1993; Trustrum *et al*., 1989). In New Zealand, the silicic Aokautere Ash (Kawakawa Tephra), dated at 22,590 radiocarbon yrs B.P., (c. 26,500 calender years), is the most widely recognized of these (Wilson *et al*., 1988). However, eruptions of this magnitude are rare and provide only limited resolution within relatively recent coverbed sequences. In contrast, tephra have been erupted with much greater frequency from andesitic stratovolcanoes such as Mt's Ruapehu and Taranaki, and although generally not forming distinctive macroscopic layers, can be identified through microscopic analysis of hornblende, pyroxene, titano-magnetite, and weathered glass mineral constituents (De Rose, 1984; Trustrum *et al*., 1989; Cronin *et al*., 1996).

In eastern Taranaki hill country, Holocene andesitic tephra erupted from Mt Taranaki (Egmont Volcanic Centre) form an accretionary sequence in soil profiles that can be differentiated from the underlying older slope deposits by differences in texture, mineralogy, colour and soil structure (Figure 7). A comparison of deeper colluvial fills on slopes to about 30º with stable terrace surfaces, suggests they have been stable throughout the Holocene. With the exception of all but a few colluvial hollows, the general absence of the c. 26.5 ka Aokautere Ash from hillslopes, suggests widespread stripping of soils during the Last Glacial period.

3.3 Methods

The study methodology at the Makahu site was as follows. Slope profiles form the basic unit of study and were used to quantify variability in soil depths and slope morphology. Spurs (convex across slope) and swales (concave across slope) were the two main morphologic units identified on hillslopes and the survey was designed to replicate measurements on these sites (Figure 8). A total of 22 slope profiles were established in the field, 11 each on spur and swale morphologic units. This number was chosen to provide a reasonable statistical sample.

Both slope and soil depths to bedrock were simultaneously measured at regular five metre intervals along profile lines, from the top of ridges, to the bottom of valleys. This measurement interval was selected to help smooth localized variability in depth and slope, while retaining sufficient detail of the slope shape. A Suunto inclinometer affixed to the top of 1 m high ranging poles was used to measure slope angle to within ± 0.5°. Although less precise than total-station surveying equipment or Abney levels, this technique permitted rapid collection of a large number of slope observations. Soil depths to bedrock were measured using specially designed 18 mm diameter cone tipped steel penetrometers, inserted normal to slope by hand, until the bedrock interface was encountered. The steel shafts were graduated in 0.5 cm intervals to enable ease of measurement from the cones tip to the soil surface. Occasional soil profile pits were excavated to verify soil depth measurements and to collect samples for mineralogical analysis. The maximum reliable depth that could be measured in this way was three metres, though few sites exceeded this depth.

The slope angle and soil depth data were analyzed by standard linear-least-squares regression. The mean of slope angle and soil depth were also calculated for each profile line (the equivalent of a hillslope) and for regions of relatively uniform slope, bounded by breaks in slope. These slope segments averaged 20 to 30 m in length, each comprising 4 to 6 observations.

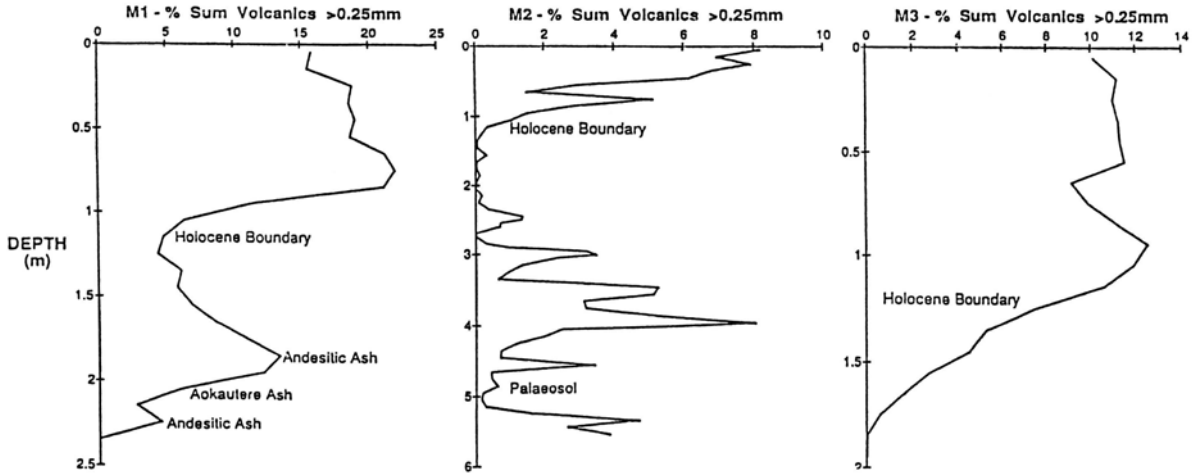

Fig. 7. Distribution of volcanic minerals in coverbed sequences, eastern Taranaki hill country: M1 - alluvial terrace; M2 - deep colluvial fill on 24° slope, and; M3 - colluvial fill on 30° slope (from, Trustrum *et al.,* 1989).

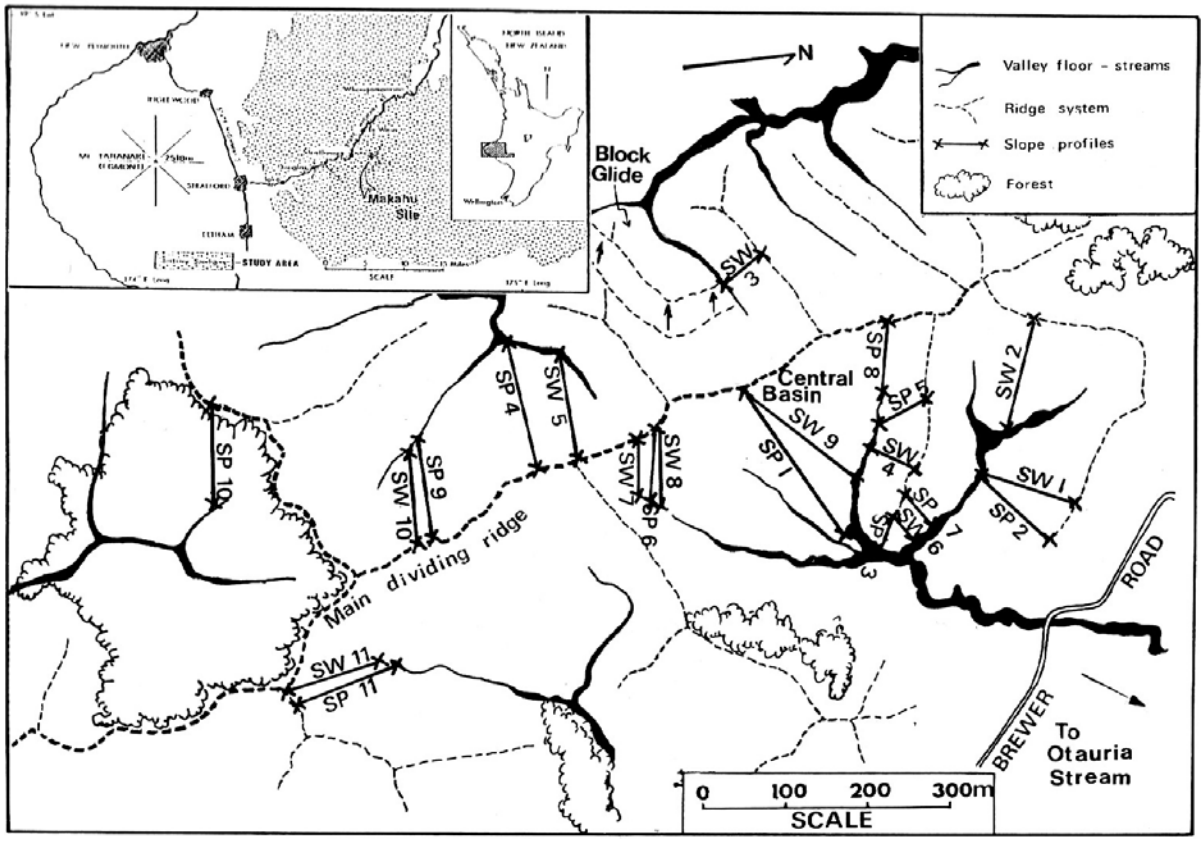

Fig. 8. Location of profile line surveys at the Makahu study site.

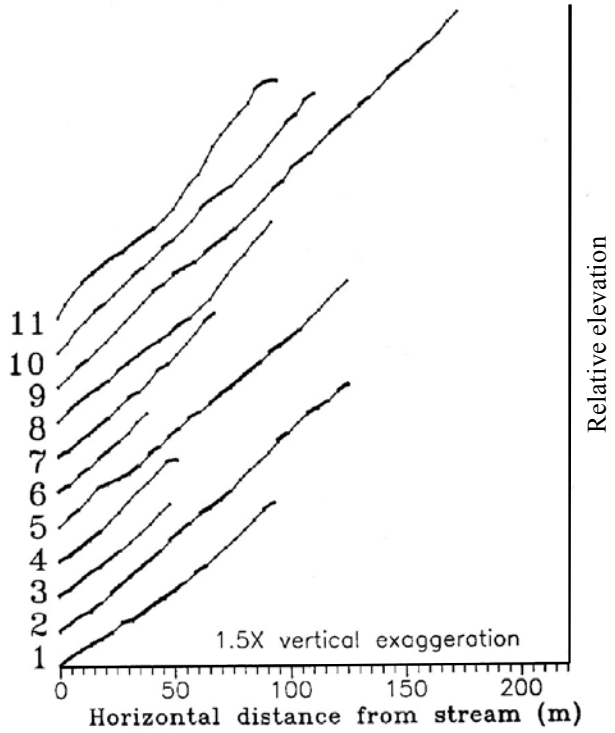

Fig. 9. Slope profiles arranged in order of increasing slope (bottom to top).

3.4 Results

In general, slope measurements vary in a cyclic manner from 10 to 65°, along the length of broadly rectilinear hillslopes (Figure 9). Mean profile slopes were in the range 25 to 36° while modal profile slopes were in the range 28° - 32°. The gentler profiles were found to occur adjacent to 2nd- and 3rd- order stream channels while the steeper profiles were adjacent to 1st-order stream channels, located at the head of drainage basins. Most 1st- order streams are bedrock channels with steep valley sideslopes of 32° – 35°, or more.

The profile line surveys show considerable localized variability in soil depth, which can range from bare rock to depths of 3 metres or more in some locations (Figure 10, Figure 11). Microtopographic variation in regolith depth was attributed mainly to vegetation effects, recurrent landsliding, hillslope position, and slope curvature and angle. Despite this high localized variability in soil depth there are clear relationships between average soils depths, slope position, slope curvature and angle. Shallow soils (< 0.5 m) are associated with convex spur sites, landslide scars or over steepened valley sidewalls near to streams. Exposures of the sandstone bedrock are common along streams and areas immediately adjacent to streams. Deep soils are associated with gentler slopes, concave sites, and lower slope positions.

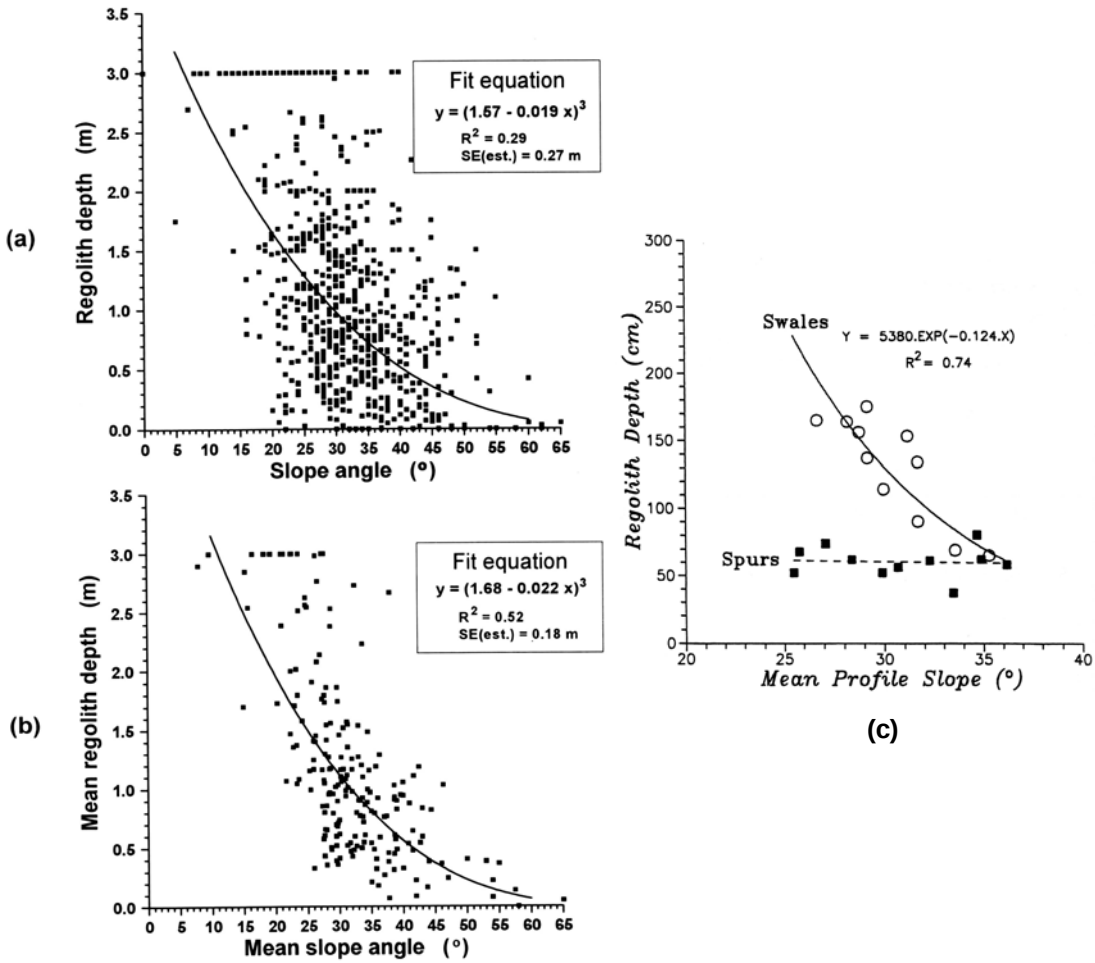

Fig. 10. Relationship between soil depths and slope for: a) point observations; b) profile segments, and; c) spur and swales profiles means (from, De Rose *et al.,* 1991; De Rose, 1996).

Soil depths are shown plotted in relation to slope angle at various scales of observation in Figure 10 and to relative hillslope position in Figure 11. The relationship between soil depth and slope can be described by inverse power or negative exponential functions, showing that soil depth decreases on average with increasing slope (Figure 10). Average depths are about 1.5 m or deeper for gentler slopes (< 28°), decreasing to about 1 m for the modal class of slopes (28 – 32°), and to 0.5 m or less for the steepest slopes (> 32°). This is apparent at all scales of observation, from individual measurements, to the mean for slope segments and profile lines, though importantly the correlation coefficient increases with increasing scale of observation. This shows that the influence of slope on average soil depth becomes more important at larger scales of observation: something that has important implications for the statistical sampling and analysis of this type of data.

Hillslope shape also influences the distribution of soil depths. For the same slope, soil depths are on average deepest in concave swale sites (Figure 10c). This is to be expected as these are sites of convergence of overland flow and therefore also soil materials. The slope-depth relationship for swales is similar to that for combined sites, except that average depths tend to be somewhat greater. Spurs in contrast, show no such relationship, and soil depths average approximately 0.6 m across the full range of slopes. This shows that slope does not control variation in soil depth on convex sites.

Hillslope position further influences average soil depths, but this depends on the average slope of hillslopes (Figure 11). There is a progressive increase in soil depth with increasing distance from the ridge along the gentler swale profiles (<31º), though occasional shallow soils may occur near channels. There is no similar trend on steep slopes > 31º. In contrast, soil depth decreases with distance from the ridge line on spur profiles, and is shallowest near to streams. Soil depths are shallowest, averaging about 0.3 m, on the lowerslopes of the steeper spur profiles.

3.5 Discussion

The general rectilinear form of hillslopes and fact that many cluster in a narrow range between 28 and 32°, would tend to support the notion of dynamic equilibrium and uniform down-wasting and a balance between denudation and vertical uplift (Strahler, 1950). If this is the case, then long-term bedrock denudation rates are in the range of 0.5 – 1 mm yr^{-1}. There is

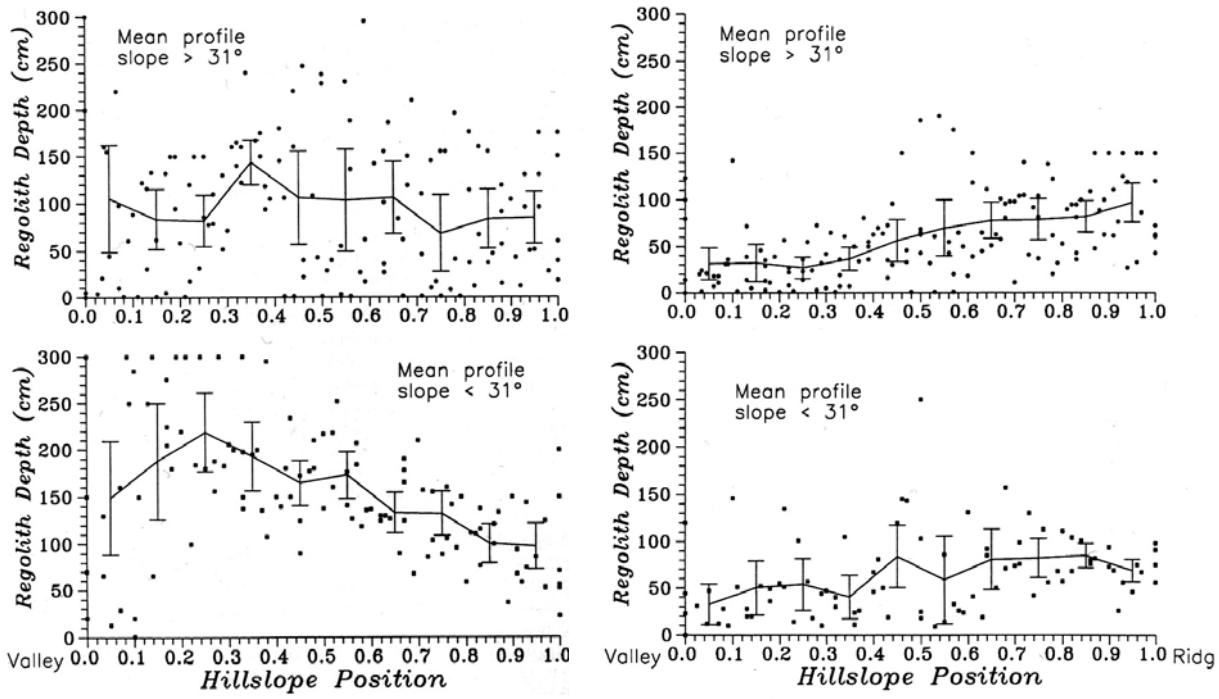

Fig. 11. Variation in soil depth in relation to position from the channel for swale (left) and spur (right) profiles, for slopes greater (top) or less (bottom) than 30° (from, De Rose *et al.,* 1991).

however enough variation in hillslope gradients, and together with systematic (rather than random) variability in soil depths, to suggest that denudation and erosion rates are not spatially uniform within basins over shorter timescales.

The generally shallow soils on hillslopes (averaging c. 1m) indicate weathering limited slope processes. Although shallow soils are associated with a number of morphological units, the trend of decreasing soil depth with increasing slope angle is interpreted to result from the increased prevalence of landslides and higher erosion rates on steeper slopes (> 31°). Figure 10b shows that very shallow soils (<0.5 m) occur mostly on slopes steeper than 35°, suggesting that in the past, landslides have tended to concentrate on these slopes. The high variability in soil depths is a reflection of the patchwork of soils at different stages of recovery from previous landslide events.

On steep (30 – 40°) forested slopes, a mean soil depth of 0.66 m and average land surface age of 450 years (Blaschke, 1989) indicates that erosion rates for landslide sites are in the order of 1.5 mm yr^{-1}. These are consistent with inferred soil erosion rates of 1 mm yr^{-1} or more, in shallow landslide dominant, steep and forested, hillslopes elsewhere in the world (Shimokawa, 1984; 1989). However, as with landslide scars on pasture slopes (section 2), these rates require scaling to the whole of hillslope areas, because a significant component of the accumulated soil will have come from downslope colluvial transport of soil from surrounding areas (e.g., convex spur sites). Soil production rates of 0.4 mm yr^{-1} for mid-aged forest communities suggests that as much as two thirds of the accumulated soil may have come from redistribution of soil on slopes. Since soil production rates from bedrock are assumed to be higher immediately after landsliding (Figure 6), then the relative contribution from bedrock production is likely to be somewhat greater. Bedrock denudation rates are about half soil production rates because of differences in the bulk density between soils and consolidated sandstone bedrock. Figure 12 shows the likely mass balance of soil materials on slopes of different average gradient at the study site.

The shallower soils on convex spur profiles are indicative of colluvial transport (rather than landsliding) from divergent sites. Spurs tend to shed soil materials and their shallow depths are a reflection of insitu weathering from the underlying sandstone in the absence of accumulations of colluvial slope deposits. The trend of increasing soil depth with distance down gentle (< 31°) swale profiles, where soil depths may reach 2 m or more, is strongly indicative of diffuse transport processes in these locations. Tephrochronological dating (Figure 7) has shown that these deeper colluvial fills have been stable for the past 10,000 – 15,000 years. The age and average depths of soil that have accumulated on gentler slopes (1 – 1.5 m), imply average long-term soil accumulation rates of about 0.1 mm yr^{-1}, assuming negligible loss of soil from hillslopes (Figure 12). In contrast, soils on the steeper slopes generally contain little material of volcanic origin and have formed from the underlying bedrock in much more recent times.

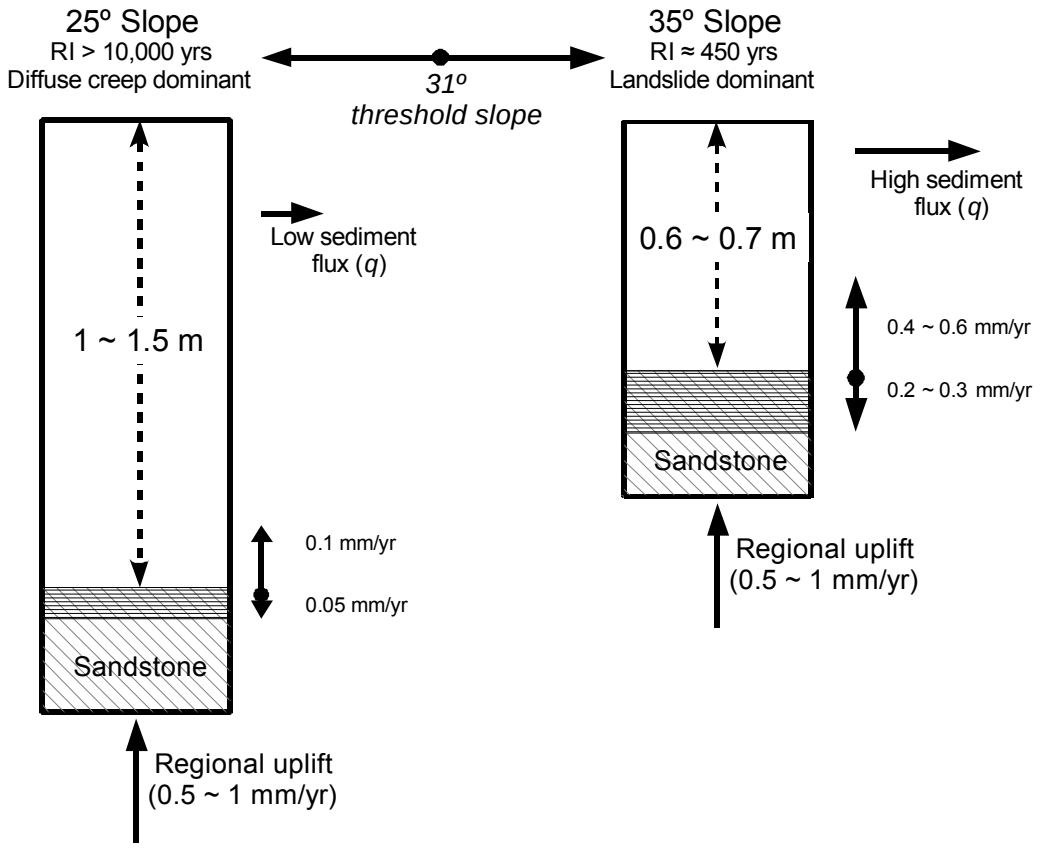

Fig. 12. Equilibrium depths and material budget for hillslopes of different average slope.

The evolution of hillslopes in Taranaki hill country appears to be controlled by a process of colluvial infilling (diffuse transport) of hollows and infrequent evacuation by landsliding, in a manner consistent with the Pacific North West model of Deitrich *et al*., (1986). This region is comparable to Taranaki hill country because of the similarity in bedrock lithology, stability and age of bedrock hollows, climate and prevalence of shallow landsliding. In the Oregan Coast Range, Reneau and Dietrich (1991) calculated from the age and depth of colluvial fills that the average bedrock denudation rates in this region were 0.066 ± 0.027 mm yr^{-1} (equivalent to a soil production rate of 0.13 mm yr^{-1} based on the difference in average bulk density between bedrock and soil). These are similar to long-term erosion rates estimated for the gentler slopes in Taranaki hill country. Reneau and Dietrich (1991) suggest that the similarity between bedrock denudation rates, bedrock exfoliation rates (0.091 ± 0.025 mm yr^{-1}), sediment yields (0.05 – 0.08 mm yr^{-1}) and tectonic uplift rates (0.1 – 0.2 mm yr^{-1}) provide evidence for approximate landscape equilibrium in their region. However, more recent research findings (Dietrich *et al*., 1995; Heimsath *et al*., 1999; Roering *et al*., 1999) suggest that because of spatial variability in soils depths, topographic curvature, and local soil production rates, hillslopes in this region are not in morphologic equilibrium.

When considering together the age of soils (Figure 7), distribution of soil depths (Figure 10) and accumulation rates on landslide scars (Figure 4), it is clear that, depending on slope gradient and position within the drainage network, hillslopes are subject to a different tempo and frequency of disturbance. Steep slopes have shallower, more youthful soils; are subject to more frequent landsliding and exposure of bedrock; and have higher erosion and soil production rates. They tend to occur at the head of 1st- order drainage basins, implying that these are also regions of most active erosion. Steep slopes are more closely coupled to the fluvial drainage network, and more likely to respond to changes in rates of channel incision. By contrast, gentler slopes have deeper, much older soils; are subject to only very infrequent landsliding and exposure of bedrock; and have lower erosion and soil production rates. Gentler slopes tend to occur adjacent to higher order streams and are somewhat decoupled from the channel network, often with intervening benches or terraces, and are less likely to respond to changes in stream base level.

Such localized variability in soil production and erosion rates is seemingly at odds with the apparent dynamic equilibrium of landscapes over longer timescales. In Taranaki, only the steeper landslide prone slopes appear to be in approximate equilibrium with the long-term rates of denudation and regional uplift of 0.5 – 1.0 mm yr^{-1} (Pillans, 1986). Gentler slopes are eroding at a slower rate (at least during the Holocene) and if this trend were to persist through time, these slopes would eventually become elevated above streams and surrounding steeper slopes. In fact, the

presence of 'benches' of relatively gentle slope that can be traced along the side of some hillslopes at the Taranaki study sites, provides morphologic evidence for this having happened at times in the past. Eventually however, these sites become unstable and subject to higher rates or erosion. Many of the hillslopes at the Taranaki study sites show evidence of periodic slope rejuvenation, starting initially at the stream channel and subsequently migrating upslope through parallel retreat. In particular, the over-steepened sidewalls adjacent to streams, suggest that a contemporary phase of rejuvenation is presently active within the catchments (De Rose *et al*., 1991). These have been likened to 'waves of aggression' by Brunsden (2001) which periodically sweep through basins reshaping and re-plaining hillslopes and removing any deep pockets of soils and gentler parts of slopes.

Fluvial incision rates since the Last Glacial maximum have been shown to be 1.5–5 times uplift rates in many regions of New Zealand (Pillans, 1986; Litchfield and Berryman, 2006; Marden *et al*., 2008). Slopes that are strongly connected to the fluvial drainage network should also have experienced similar responses. In Taranaki hill country, over-steepened sidewalls and gentler local slope gradients further upslope, provide strong evidence for high contemporary rates of fluvial incision, preceded by periods of relative stability. Although recent fluvial incision coincides with the wetter Holocene Epoch, the absence of soils and colluvial fills which predate the late Pleistocene and Holocene, suggests widespread stripping of soils on hillslopes during previous colder climates (Dietrich *et al*., 1986). It would appear that, the relative intensities of erosion in different parts of the landscape depend on connectivity with base level responses in the drainage network and external controls on hillslope stability, imposed by synchronous fluctuations in climate and vegetation (McGlone, 1983; Newnham *et al*., 1999).

Erosion and denudation cannot be viewed as being either constant in space or time (Figure 13), and it is unlikely that the assumption of uniform downwasting (Hack, 1960), applies to periods of time when changes in climate or spatial variation in soil depth and slope morphology enforce variability is soil production and erosion rates (thousands to tens of thousands of years). Erosion rates are upwards of an order on magnitude greater on the steeper landslide dominant slopes compared with the relatively gentle, creep dominated slopes, and this needs to be taken into consideration when evaluating the impact of catchment disturbance on contemporary erosion rates.

4. SHALLOW LANDSLIDE EROSION

4.1 Introduction

Shallow landslide erosion is the most frequent and widespread form of mass-wasting recognized within New Zealand (Soons and Selby, 1992) and is the principal natural hazard (Crozier, 2005). Shallow regolith landslides, seldom greater than 2 – 3 metres in depth, are the most common of the various forms of landslide morphology. They typically occur as regional clusters of “debris slides” triggered during high intensity rainfalls or following periods of prolonged soil saturation and can simultaneously involve many thousands to tens of thousands of landslides over areas extending up to 20,000 km^2 (Crozier, 2005; Hancox and Wright, 2005). Landsliding events occur somewhere in New Zealand two to three times a year during highly localized storm events, though rainfall events capable of causing widespread landslide damage occur much less frequently (Glade, 1998). Threshold rainfalls

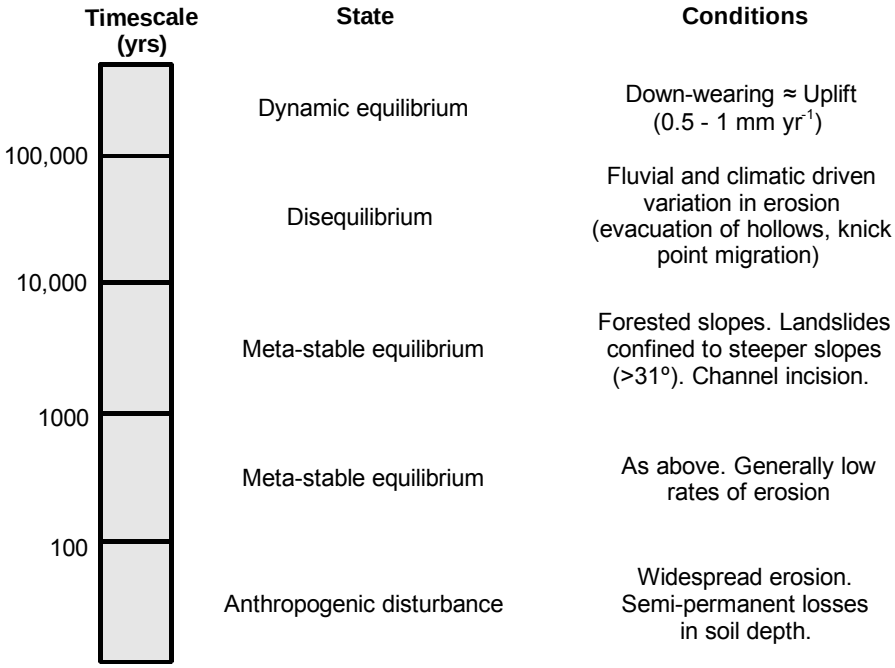

Fig. 13. Variation in the geomorphic state and conditions for landform evolution in relation to timescale.

intensities are usually in excess of 100 – 180 mm per 24 hour depending on location (Crozier *et al*., 1980; Page *et al*., 1999). Rates of denudation in areas of active landsliding, usually where the vegetation has been severely disturbed, vary from 0.01 to 5 mm yr^{-1}, equivalent to 1.2 – 60 t ha^{-1} yr^{-1} (Selby 1982).

Landslides are a particularly common sight on steep hillslopes where the native forest cover has been removed and replaced by grassland for farming. Landslides are commonly interpreted to result from forest removal because measured landslide densities are typically 3 – 20 times greater on pasture covered than forested hillslopes (O'Loughlin and Pearce, 1976; Pain and Stephens, 1990; Hicks, 1991; Blaschke *et al*., 1992; Marden and Rowen, 1993; Reid and Page, 2002; Hancox and Wright, 2005; Dymond *et al*., 2006). River sediment loads from paired catchments (Hicks, 1988; Quinn and Stroud, 2002), or changes in lake sedimentation over time (Page and Trustrum, 1997; Eden and Page, 1998) also imply similar increases in landslide erosion due to forest removal.

There are problems however with such simple comparisons of landslide densities because most landslide surveys employ aerial photographic or remote sensing analyses and landslide densities are often underestimated in forested areas because scars can be obscured by tree canopies. Most of these studies have considered only a single snapshot in time and not the cumulative impact of landsliding over longer time periods. Furthermore, regions of land are rarely identical and the type of vegetative cover may be only one of a number of factors influencing variation in landslide densities (e.g., soils, geology, slope, aspect, rainfall intensity can influence landslide densities). Added to this, the fact that landsliding is considered the dominant disturbance mechanism on steep forested slopes (Shimokawa, 1984; Blaschke *et al*., 1992), suggests that an alternative method is required to more precisely determine the effect of vegetation cover on landsliding.

There was thus a need to develop a methodology which could conclusively demonstrate the effectiveness of vegetation on controlling shallow landslide erosion in New Zealand hill country. As a result, this section presents a method for the comparison of landslide densities and erosion rates between forested and pasture covered areas of land, based on a novel 'paired hillslope' approach (analogous to a paired catchment study). In this study, neighboring hillslopes are identified which vary only in terms of the elapsed time since deforestation. In theory, any difference in landslide densities should be a direct consequence of the different vegetation histories. Furthermore, as landsliding results in reduced soil depths (section 2) any difference in average soil depth should relate to differences in landslide activity, and provides a direct measure of the rate of soil erosion caused by forest removal.

4.2 Study site

In most regions of New Zealand widespread removal of the native forest has resulted in few areas where catchments vary only in terms of their vegetation. In the Taranaki region, however, because land development was more sporadic, there is a patchwork of vegetation types where forested hillslopes alternate with grassland hillslopes. Conversion to grassland has occurred at different times in the past, and adjacent forested and grassland hillslopes can be found that have the same slope morphology and history of storm activity. As in section 2 and 3 the Makahu site in eastern Taranaki hill country (Figure 2) was selected for study.

4.3 Methods

Neighboring drainage basins and hillslopes were located that had been deforested for 10- and 85- years, respectively. To detect statistically significant differences in mean depth, a survey strategy was required that was randomly distributed and provided a large enough number of soil depth observations within each area. The survey also needed to be relatively easy to implement in the field by considering ease of access to hillslopes and sampling time (i.e., days rather than months or years). For these reasons, the profile line technique was adopted as the means of surveying and collecting soil depth measurements. A systematic random sample of profiles lines was selected by overlaying a regularly spaced point grid onto the 1:7,000 scale aerial photographs. A slope profile line running through each point, from ridge top to valley bottom, was located in the field and soil depths were measured at 5 m regularly spaced intervals along its length. Measurements of the slope angle between each 5 m sample location were also made to ensure that the two areas had similar slope morphology. In this way, 49 profile lines provided a total of 826 point observations. Although similar data had been collected during the previous survey at the Makahu site (section 3), these were considered unsuitable for inclusion into this study because they were collected on a systematic rather than random basis.

The profile lines were further subdivided into a total of 236 profile segments of relatively linear slope angle and uniform regolith depth, that averaged 17.5 m in length. Breaks in slope and scar margins demarcated profile segments. Landslides were recognized by the presence of clearly distinguishable scar margins. The profile segments were further grouped into those representing recent landslide scars and those representing undisturbed forest soils.

Two weeks after completion of the survey, a large-magnitude, low-frequency storm occurred in March 1990 (Cyclone Hilda). The survey was repeated to permit a direct comparison of pre- and post-storm event soil depths, and provide an accurate measure of soil loss due to this one event. It was only necessary to repeat survey profile lines where new landslides were observed.

The slope angle and soil depth data were analyzed by standard linear-least-squares (LLS) regression. A cubed root transformation was applied to soil depth data, to normalize the distribution of soil depths prior to regression analysis.

4.4 Results

In all, 5% and 20% respectively of the areas deforested for 10- and 85-years were affected by contemporary landslide erosion. This demonstrates that it is the removal of forest vegetation that is the primary cause for the majority of landslides to have occurred in the area deforested for 85-years: such that landslides would not have occurred, had the forest remained.

Average soil depths were 1.105 ± 0.09 m and 0.953 ± 0.08 m for the 10- and 85-year deforested areas, equating to a difference of 0.15 ± 0.11 m. This difference will only relate to landslide erosion if both areas had the same mean depth prior to deforestation, because as shown in section 3, average soil depths also vary in relation to slope. To test the assumption of uniformity in initial soil depths, regression models between depth and slope for the non-landslide profile segments, and hence forest soils, were computed for the two regions. The regression models (Table 1) were statistically indistinguishable, demonstrating spatial uniformity in depth across the range of slopes, and that the same regression model (using pooled data, Table 1) can be used to estimate the preexisting soil depths prior to landslide occurrence. The spatial uniformity in average depth, importantly suggests that an equilibrium condition may have existed between rates of soil accumulation and erosion under the indigenous forest cover.

The two areas have slight, but significantly different average slopes of 31.0° and 32.8°. The 1.8° difference in mean slope, suggests that the area deforested for 85-years would have a 0.03 – 0.1 m shallower average soil depth in the absence of any recent landslide erosion, and therefore a portion of the 0.15 m difference in average depth may be due to differences in average slope angle.

An alternate way to analyze this same data, is to calculate the net loss in soil for landslide sites alone, then spread this loss across the area as a whole, to separately derive erosion rates for each of the two areas. Twenty percent (20%) of slope segments were landslide scars in the area deforested for 85-years. These had an average depth of 0.153 m and average slope of 33.3°. For these sites, the average regolith depth that existed prior to landslide occurrence is estimated from Table 1 at 0.934 m, giving a net soil loss of 0.781 m for landslide sites. This equates to an average surface lowering by post-deforestation landslides of 0.15 ± 0.04 m. In the area deforested for 10-years, the average surface lowering due the 5% of landslide scar segments was similarly estimated at 0.02 ± 0.04 m. Hence, erosion over the 75-year period between times of deforestation amounted to 0.13 ± 0.06 m, demonstrating that most of the difference in average depth of 0.15 ± 0.11 m between the two areas was due to landslide erosion, and not to differences in average slope angle.

A comparison of the frequency distributions of measured depths (Figure 14), shows that in the area deforested for 85-years, the proportion of samples in the range 0.7 – 2 m has reduced, while the proportion in the range 0 – 0.5 m has increased, i.e., landslides have preferentially removed soils with mid-range depths. An increase in the proportion of observations greater than 2 meters in the area deforested for 85-years is due either, to debris accumulation from landslide deposits or, to a slightly higher proportion of gentler slopes or concave hollows being sampled.

The 0.15 m average surface lowering represents a regolith depletion rate of 1.8 ± 0.5 mm yr^{-1}. However, landslides are not evenly distributed across hillslopes. For hillslopes steeper than 28°, where most post-deforestation landslides occur, average surface lowering is higher at 0.20 ± 0.05 m, and the soil depletion rate is 2.4 ± 0.6 mm yr^{-1}. Average surface lowering is greatest at 0.23 ± 0.07 m on hillslopes steeper than 32° where the majority of post-deforestation are concentrated. Here, the soil depletion rate is 2.7 ± 0.8 mm yr^{-1}. These results demonstrate an increase in the rate of erosion with increasing slope angle.

During Cyclone Hilda, twenty landslides were triggered on or adjacent to the survey lines, affecting 12% and 4% of the length of profile lines in the areas deforested for 10- and 85-years, respectively. These landslides produced an average surface lowering of 0.041 m over the study site. This one, extreme rainfall event, produced about a quarter as much erosion as had all previous landslides over the 85 years since deforestation, and demonstrates the importance of large magnitude, infrequent storm events to hillslope erosion. There were proportionately more landslides in the area deforested for 10 years, illustrating the importance of previous erosion history of hillslopes on the spatial distribution of landslides. There were also comparatively few landslides on steeper hillslopes because previous lower magnitude storms had already removed much of the unstable regolith.

Table 1. LLS regression models of soil depth on slope for two hillslope areas. Two standard errors of coefficients are shown in brackets.

Hillslope area	$D = (a + bS)^3$, D = depth in metres, S = slope in degrees		
	a	b	R^2
10-year	1.62 (0.16)	-0.020 (0.005)	0.44
85-year	1.71 (0.12)	-0.022 (0.004)	0.55
combined	1.68 (0.10)	-0.021 (0.003)	0.52

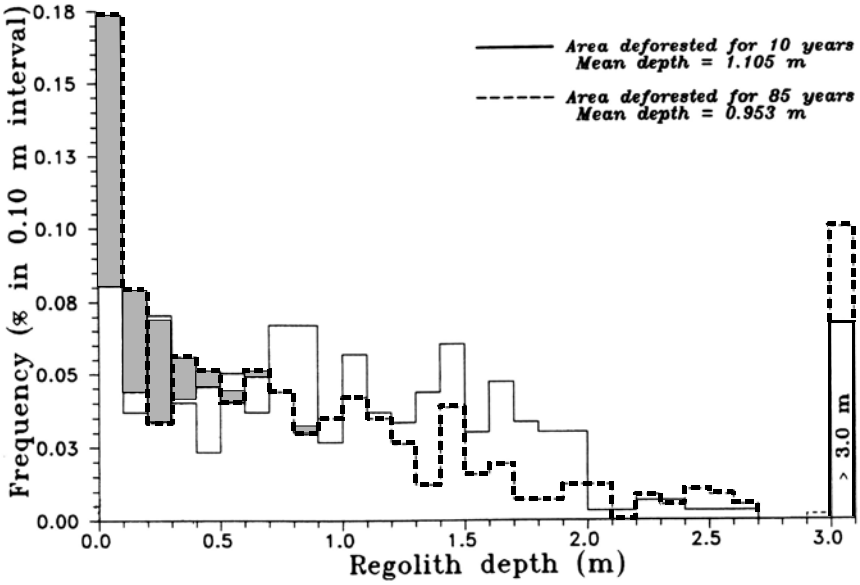

Fig. 14. Frequency distribution of soil depth. Depth measurements on landslide scars are shown in black (from, De Rose *et al.,* 1993).

4.5 Discussion

Results of this study demonstrate that in regions which have clearly defined boundaries between soil and bedrock, it is possible to quantify soil depletion caused by landslide erosion, by simply comparing the mean depth of regolith between 'paired hillslopes' that have had their forest removed for different periods of time. The survey would have benefited from a greater density of sampling. For example, the 95 % confidence error of ± 0.11 m would reduce to ± 0.05 m if 6 times as many samples had been taken.

The results also conclusively demonstrate, that had the forest remained on hillslopes, the majority of landslides would not have occurred. Furthermore, they show that landslide erosion rates increase with increasing slope angle and there is a threshold slope at about 28º above which there is large increase in landslide density. A comparison of the contemporary erosion rates calculated from landslide area densities and mean hillslope soil depths under pasture vegetation, and estimates of the long-term erosion rates derived from soil accumulation rates (section 3), suggests that there is about a 2 – 10 fold increase in erosion rate following deforestation (Figure 15). The relative increase tends to be greatest on the modal class of slopes (28º – 32º) where soils are deeper and have been stable throughout much of the Holocene. Although there are typically fewer landslides on the modal class of slopes, they tend to be larger and erode a greater volume of soil.

The results from Taranaki hill country, though generally consistent with findings from landslide prone regions elsewhere in New Zealand, importantly demonstrate that slope morphology needs to be taken into consideration when comparing landslide densities or frequencies between forested and pasture areas of land. Regions with a greater proportion of steep slopes above the threshold for landsliding, will have a greater overall response in terms of erosion rates and hillslope sediment yields.

The storm event of March 1990 (Cyclone Hilda) was also very fortuitous, because the results of resurveying profile lines, provide numerical evidence for the notion that hillslopes can acquire event resistance. Previous erosion events appear to have removed much of the unstable regolith and remaining soils, because of their greater shear strength or location away from sites of saturation, are resistance to failure in subsequent rainfall events. The probability of landslides occurring in subsequent events is therefore reduced. This suggests that the initial rates of landslide erosion that follow

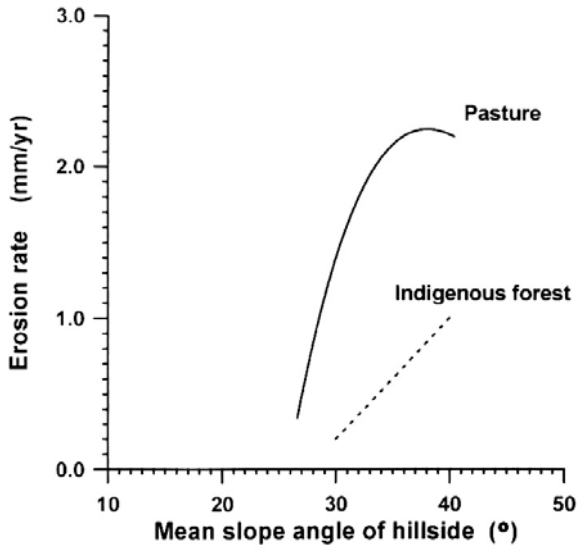

Fig. 15. Comparison of contemporary landslide erosion rates under pasture and long-term erosion rates under forest (from, De Rose 1996).

forest removal (O'Loughlin and Pearce, 1976; Pearce, 1986; Crozier, 1986; Trustrum and Hawley, 1985) are unlikely to be sustained in the long-term.

A key outcome of this survey was to show that spatial variability in landslide occurrence was controlled by a number of factors: slope angle, recency of forest clearance and previous landslide activity. These factors must therefore be included in modeling applications which predict spatial variability in landslide occurrence.

5. MODELING AND PREDICTION OF LANDSLIDE EROSION

5.1 Introduction

The ability to predict sites of future landslide failure is one of the most important aspects of hazard mitigation. Most detailed surveys of landslide occurrence are generally restricted to small areas of land and results require upscaling to larger areas of land using one of a number of modeling approaches. Both physically-based (geotechnical) slope stability models and empirically-based (statistical) landslide susceptibility models have generally been used, though often with limited success (van Westen *et al*., 2006). Each has their own advantages and disadvantages.

Empirical (statistical/multivariate) models that relate the spatial occurrence of landslides to landform attributes (vegetation, slope, geology, soil type, etc) are more applicable at larger spatial scales and although much easier to implement in GIS systems, can have poor model performance because of high within and between event variability of landsliding and due to poor resolution and limited availability of spatial data sets (Lineback Gritzner *et al*., 2001; van Western *et al*., 2006; Miller and Burnett, 2007). Moreover, specific models are often not transferable to other regions and maybe of limited usefulness in future hazard assessments because they are constrained by historical patterns of failure (Montgomery *et al*., 1998). These models however can shed light on the key driving variables for landslide initiation.

Physically-based stability models rely on a higher degree of parameterization and are most applicable at the hillslope scale where relevant soil and hydrological data maybe available. Although less applicable at larger spatial scales, stability models can be used to make future predictions about the instability of hillslopes under varying soil, rainfall and slope seepage conditions (Brooks *et al*., 2002). Some recent applications have derived slope thresholds for landslide initiation under varying soil, geological or hydrological conditions and then applied these to DEM derived slope and soil/geology maps to produce maps of landslide susceptibility (e.g., Liener and Kienholz, 1996; Hennrich and Crozier, 2004). Correlating topographic wetness indices (which are readily derived from DEMs) with field pieziometric measurements of the degree of soil saturation during rainfall events, maybe the most tractable way of incorporating hydrological data at larger spatial scales (Hennrich and Crozier, 2004). Ultimately, however, it is the short-scale variability in soil depth, strength (including tree root size and densities) and hydrological properties that limit the accuracy of slope stability models (Hawke and McConchie, 2003).

This section presents the results of modeling of landslide erosion at the Makahu study site in Taranaki hill country. This work was undertaken as a means of investigating the potential modeling approaches that might be used to extrapolate and map landslide hazard over much larger areas of hill country.

5.2 Methods

5.2.1 Empirical model

An empirical model (Table 2) was formulated by correlating the density of post-deforestation landslides within 1st-order drainage basins to landform attributes that included slope, curvature and flow accumulation. Landslides sites were located using a combination of field surveys and historical aerial photography. The density of landslides was on an area basis and was determined by mapping the spatial extent of landslide scar boundaries onto a high resolution (1:7000 scale) orthophotograph of the study site. Once digitized, the planimetric area of landslide scars and the 1st-order drainage basins that contain them, could be calculated.

Landform properties and basin areas were derived from a 2 m resolution DEM of the study site. The DEM was captured by an analytical stereoplotter from aerial photographs. Points recorded at the corners of tessellated triangles and break-lines were interpolated to form the 2 m grid of elevations (Dymond *et al*., 1992). An index of the degree of flow accumulation was given by the 50th percentile on the cumulative frequency distribution of log transformed (40Log10) flow accumulation values (i.e., upslope contributing area) within each basin area. Divergent slopes had on average a lower median 40Log10 value than convergent slopes.

5.2.2 Physical model

Instability in soil masses depends on the balance between shear stresses induced by gravity and resisting forces provided by the shear strength of slope materials. Limit-equilibrium analysis is the common method used to predict sites of instability (Terzaghi, 1942). The infinite slope stability model (Equation 4), which assumes a slip surface parallel to the ground surface, is considered most applicable to shallow landslide failures of the type common in New Zealand hill country (Preston and Crozier, 1999; Hennrich and Crozier, 2004).

$$\mathrm{FoS} = \frac{c' + (\gamma - m\gamma_w) z \cos^2 \beta \tan \phi'}{\gamma z \sin \beta \cos \beta} \qquad (4)$$

where FoS = factor of safety, c' = effective cohesion (kg m^{-2}), γ = regolith density (kg m^{-3}), γ_w = water density (c.1000 kg m^{-3}), m = ratio of water table height to regolith thickness, z = regolith thickness (m), β = angle of shear plane, (º), ϕ' = effective friction angle of shearing resistance (º). Failure is predicted when FoS < 1. The friction angle typically varies from 25 – 35º for

naturally occurring soils, while effective cohesion can vary from 70 – 2000 kg m^{-2} depending on soil texture and vegetation (O'Loughlin and Pearce, 1976; Sidle, 1984; Tsukamoto, 1987; Fannin *et al*., 1996; Preston and Crozier, 1999).

The assumption of a planar slip surface may not always be valid since hillslopes can be curvilinear in form and there maybe an element of rotation at the time of landslide failure. Additional soil strength may derive from internal stresses within the landslide body at the time of failure. Hence, alternate methods such as in Bishop's simplified method can be used to model slope stability (Bishop, 1955). Drainage conditions are also important for stability and consequently recent model development has incorporated 2D numerical simulations of seepage, which is usually based on Richard's (1931) equation and Darcy's law extended to unsaturated conditions (Sammori and Tsuboyama, 1991; Brooks *et al*., 2002). For Bishop's simplified method, the Factor of Safety is given by:

$$\mathrm{FoS} = \frac{1}{\sum W \sin\beta} \sum \frac{c' l \cos\beta + (W - ul\cos\beta)\tan\phi'}{\cos\beta + \sin\beta \tan\phi' / FoS} \quad (5)$$

where W and l are the slice weight and base length, respectively, u is the poor pressure at the base of the slice, and β, c' and ϕ' are defined as before. Unsaturated soil strength is determined as: $\sigma' = \sigma - x.u_w$, where σ' is the effective stress, σ is the total stress, u_w is pore water pressure and x is a fraction depending on water content (1 at saturation and 0 when dry). Van Genuchten (1980) equations are used to derive the unsaturated hydraulic conductivity at varying pressure potential.

For the study site at Makahu, critical depths of failure (FoS = 1) were calculated for the range of local slopes typically occurring at the field site (15 – 40º). Uniform soil properties were assumed. As no field shear tests were available, soil strength properties (c' = 200 kg m^{-2}, ϕ' = 29º) were obtained from a region with similar soils and slope morphology (O'Loughlin and Pearce, 1976) to the Makahu field site. Given that water bypass occurs during storms events and that relative water heights often do not exceed m = 0.6 – 0.8 (Crozier *et al*., 2003; Hennrich and Crozier, 2004), the critical depths of failure were also recalculated by assuming m = 0.8.

For the 2D seepage-stability coupled model, a 50 m

Table 2. LLS regression models (y = a+bS+cF) for predicting landslide density (as % of hillslope area) for hillslopes above 28º at the Makahu study site. Standard error of coefficient is in brackets.

Parameters	Coefficients			
	a	b	c	R^2
Slope (º)	-35 (4)	1.4 (0.1)	-	0.32
Flow accumulation (50th percentile)	-6 (3)	-	0.9 (0.2)	0.11
Slope + flow accumulation	-0.43 (4)	1.3 (0.1)	0.7 (0.1)	0.38

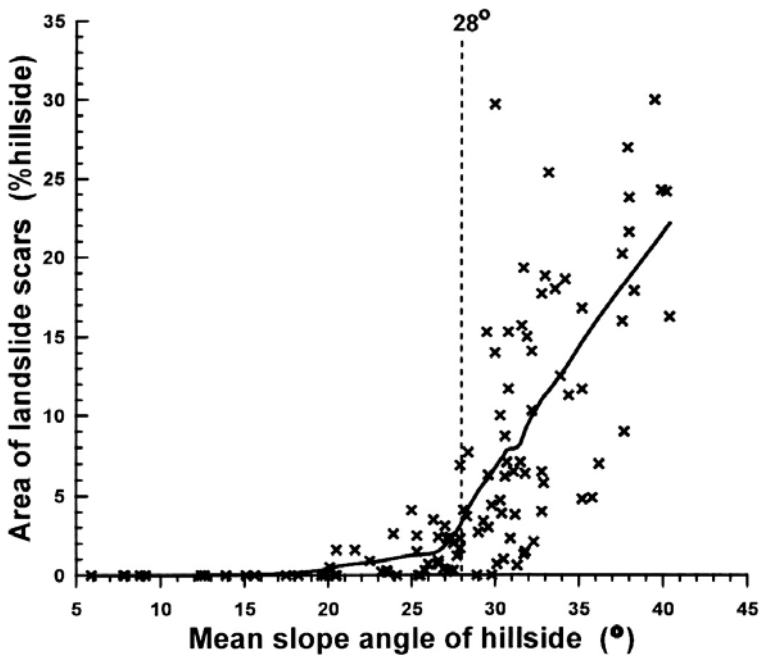

Fig. 16. Area density of landslide scars in relation to mean slope of hillside. Fitted line is the local area weighted average (from, De Rose 1996).

long, uniformly inclined slope was used to simulate depths of failure under conditions of constant rainfall intensity over a defined period. For the Makahu site, rainfall intensities equivalent to 120 mm and 240 mm per 24 hours were used to simulate 10- and 50-year recurrence interval storm events. Soil density (DBD = 800 kg m^{-3}, γ_{sat} = 1450 kg m^{-3}) and hydrological data (K_{sat} = 7 x 10^{-5} m s^{-1}) were from the Egmont black silt loam (NZ National Soils Database), the closest soil type with available soil moisture retention data (NZ National Soils Database), to soils at the Makahu study site. Volumetric desorption curves were used to derive the Van Genuchten parameters (Ksat = 7 x 10^{-5} m s^{-1}, saturated soil water content θ_s = 0.75, residual soil water content θ_r = 0.27, α = -4.47, n = 1.35). The low DBD reflects the high volcanic ash and allophanic clay content within these soils.

5.3 Results

5.3.1 Empirical model

Slope explains most of the variation in landslide density: there are no landslides on slopes < 20º few < 28º and above 28º there is a more or less linear increase in landslide density with increasing slope (Figure 16). Neither plan nor profile curvature was related to landslide density. Flow accumulation explained some additional variation in landslide density as landslides were generally more common on hillslopes with greater flow accumulation. A piecewise model was found best suited to predicting landslide density, ie., no landslides < 20º, 20º < average 2 % < 28º, and > 28º use the multivariate model (Table 2). This model was used to predict the area density of landsliding on pasture hillslopes at the Makahu site (Figure 17). The multivariate model shown in Table 2, applies only to slopes above 28º, where there is a more or less linear relationship between slope and landslide density.

5.3.2 Physical model

Critical depths of failure calculated from the various stability models when compared with the average distribution of soil depths (Figure 18) would suggest that the majority of soils are unstable using the assumed soil strength parameters. A back calculation of critical depths of failure to match the observed average soil depths, shows that larger effective cohesion, lower effective friction angles, and relative water table depths (c' = 391 kg m^{-2}, ϕ' = 26º, m = 0.8) are required to maintain stability in these soils (Figure 18). For most model runs there is a large decrease in critical depths of failure for

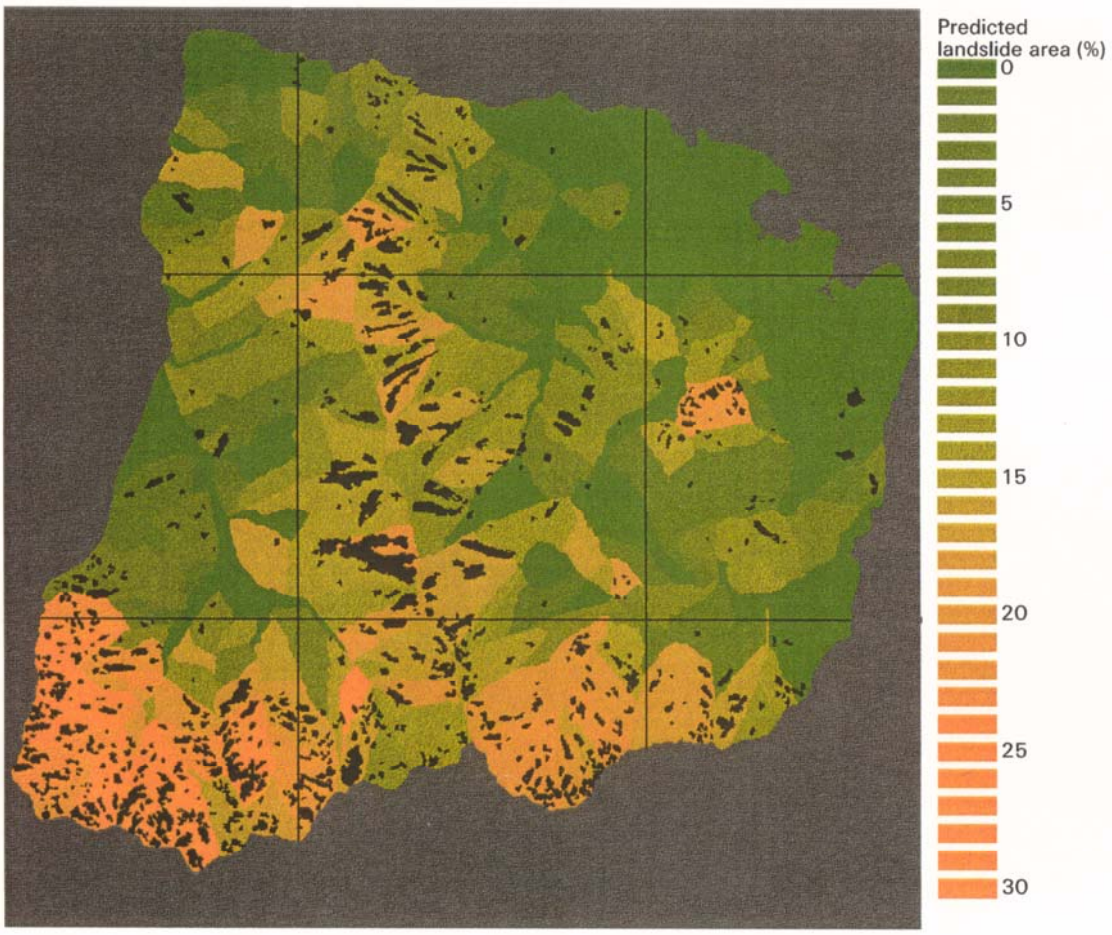

Fig. 17. Predicted area-density of shallow landslide scars, based on an empirical model with slope and flow accumulation, at the 120 ha Makahu research site (NZMS 260 Q20 498128). Actual landslide scars are in black. Landslides are concentrated onto slopes above 28º and hillslope facets with high flow accumulation. Grid lines are 500 m apart.

slopes above about 20º.

In general, Bishop's model suggests greater relative stability, with critical depths of failure on average 20 – 40 cm deeper, than in the case of the infinite slope stability model (m = 1). Critical depths of failure increase with decreasing relative water table heights such that both the infinite slope stability model (m = 0.8) and Bishops model give similar results. For the two simulated rainfall events, there is a decrease in the threshold slope for landslide initiation from about 28 to 21º, with increasing total rainfall.

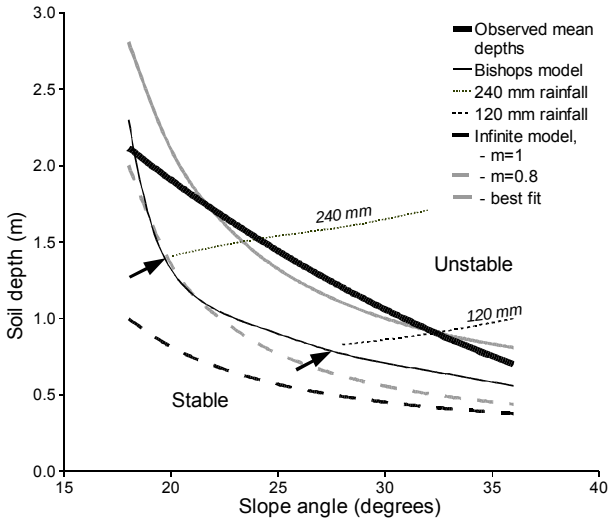

Fig. 18. Critical depths of slope stability/failure derived from slope stability models compared with observed mean soil depths.

5.4 Discussion

Both modeling approaches show that slope morphology and soil hydrological and strength properties are important in governing slope stability at the study site. Soils are relatively more stable on gentler slopes and in regions which are likely to experience lower relative water table heights during storm events, such as sites of flow divergence. Conversely soils are more unstable on steeper slopes and in sites of flow convergence. The distribution of landslides at the study sites largely supports the results from stability modeling. In particular, both modeling approaches indicate that landslides are unlikely to occur on slopes less than 20º. However, while the stability model suggests that the majority of hillslopes should be unstable above this slope, the empirical model shows that a relatively small percentage of their area is affected by landsliding, and landslides only become common on slopes above about 28º. Despite the occurrence of a significant number of high intensity rainstorm events capable of causing landsliding over the last 85-years since forest clearance, many sites remain stable. Clearly the assumption of uniform soil properties has limited the accuracy of stability models. Physically based stability models will probably only be useful in predicting sites of failure once techniques are developed for acquiring detailed spatial information about variation in soil strength and hydrological properties (Hawke and McConchie, 2003). The present stability model, for example, had employed hydrological properties from a soil located some distance from the study site and this could explain why it tends to over predict the extent of landsliding, particularly if the hydraulic conductivities were less and there were a greater tendency for overland flow to occur in these steepland soils.

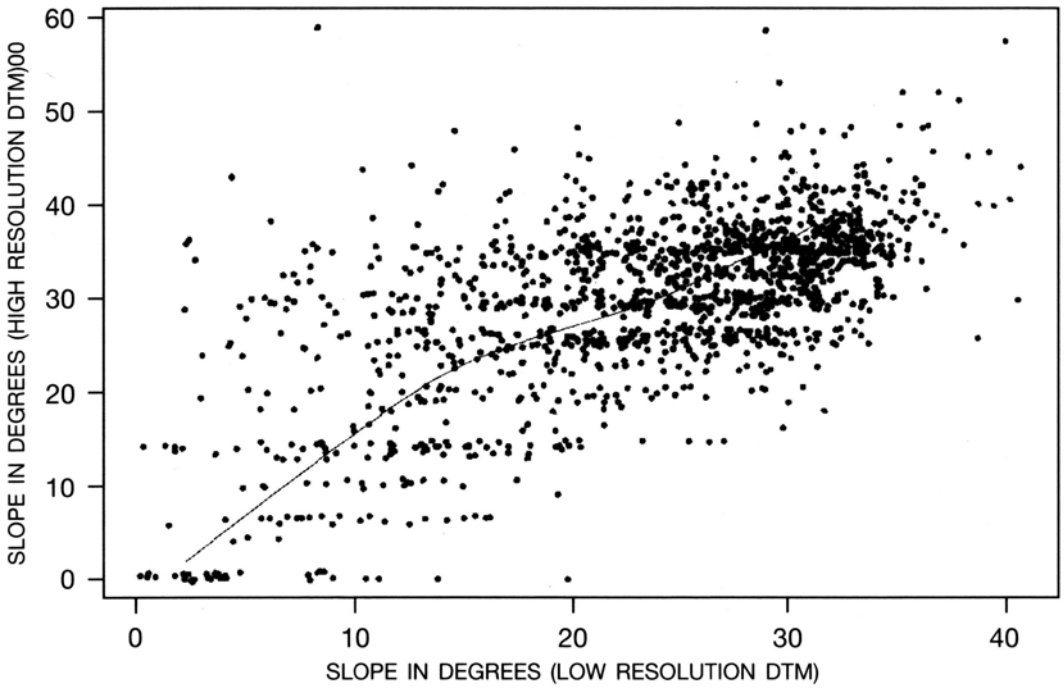

Fig. 19. Comparison of slope derived from high (2 m) and low resolution (12.5 m) DEMs (from, Dymond and Harmsworth, 1994).

Currently, the empirically based approach provides the best method for mapping zones of landslide susceptibility as it relies only on the availability of a high resolution DEM to derive slope and flow convergence. It is, however, currently restricted in application to the consolidated sandstone hill country of inland Taranaki hill country. Extrapolation to other areas of hill country will rely on recalibration, as landslide density is also affected by geology and soil type (De Rose, 1995; Dymond *et al*., 2006). Probabilities for landsliding are generally greater, and threshold slope angles for failure lower, on hillslopes underlain by soft sediments such as Tertiary mudstones, crushed Cretaceous lithologies or deeply weathered Jurassic Greywacke. Mean slope angles for landsliding are in the range 27 – 36° and serious slipping generally starts on slopes between 20 and 24°, depending partly on storm size (Page *et al*., 1994; Hancox and Wright, 2005). In contrast, slope angles are in the range 45 – 49° and serious slipping starts above 33° where hillslopes are underlain by consolidated, relatively unweathered Jurassic Greywacke.

Recently, Dymond *et al*., (2006) suggested that the importance of threshold slopes has been overstated and all slopes are susceptible to landsliding, though to different degrees. Their conclusion was based on correlations between landslide density and slopes for the February 2004 storm event, which showed that landslide probability on slopes from 10 to 20° was one third that of slopes above 30°. Their results are however, partly an outcome of the DEM-derived slope. Slope derived from DEMs is particularly prone to error because of inaccuracies in elevations and the tendency for larger grid spacings to "smooth" computed topographic surfaces (Desmet, 1997; Walker and Willgoose, 1999), such that slope varies inversely with DEM grid size (Montgomery *et al*., 1998; Zhang *et al*., 1999). As illustrated by Figure 19, there will be a significant proportion of steep areas above 30° in which the corresponding DEM derived slopes will appear to be between 10 and 20°: landslides will appear to occur on slopes < 20° with a greater frequency, than actually occurs in the field.

6. GULLY EROSION

6.1 Introduction

The contribution that gullies make to the sediment yield of river basins is an important issue in many areas of the world. Poesen *et al*., (2003) note that the relative contribution of gully erosion to total sediment production within catchments is variable, typically ranging from 10 to 94%. In New Zealand for example, gully complexes rank as the largest point sources of sediment, particularly within the North Island where they can be locally widespread. Gully erosion is thought to dominate the 15,000 t yr^{-1} and 20,520 t yr^{-1} sediment yields of the Waipaoa and Waiapu Rivers (Hicks *et al*., 2000), who's yields are among the highest in the world (Walling and Webb, 1996). It is estimated that gully erosion affects some 10 % of New Zealand catchments (Eyles, 1985).

In the past, only qualitative statements could be made about the contribution of gullies to basin sediment yields because of the lack of adequate methodologies for quantifying the surface erosion rates of gullies. There was thus an identified need to develop a reliable methodology for the measurement of gully erosion. This would then enable a much more precise evaluation of the relative role that gully erosion makes to sediment yield at the catchment scale. This section demonstrates the development of a methodology for measuring gully erosion by constructing high resolution digital elevation models (DEMs) from historical aerial photography using digital photogrammetry techniques. It further shows how the measurement of gully erosion at the hillslope scale can be easily upscaled to whole of catchment areas through the development of gully area - erosion rate relations.

Although various definitions of gullies exist, that of Schumm *et al*., (1984) where "gullies are relatively deep, unstable, eroding channels that form at the head, side or floor of valleys where no well-defined channel previously existed", provides a straightforward and unambiguous definition. Gullies are often thought of as long sinuous features that form at the head of streams and are particularly common in arid and semi-arid environments (Prosser and Winchester, 1996). However, there can be considerable morphological variability in gullies depending on local substrate conditions (geology, soils). Gullies in the East Coast of the North Island of New Zealand are generally associated with intensely crushed and sheared bedrock lithologies of Cretaceous age. Although generally consistent with the definition of Schumm *et al*., (1984), these gullies are characterized by a combination of slump, sheet, rill and bank erosion processes and are best thought of as 'mass movement gully complexes'. Because of significant variation in gully morphology in East Coast catchments, gullies of varying size and morphology were considered in the assessment of gully erosion.

6.2 Study sites

Gully erosion is investigated at Mangatu and Haunui Forests in the eastern Raukumara Range. The Mangatu Forest site lies within the headwaters of the Waipaoa River basin (Figure 20), straddling the boundary between the Te Weraroa and Upper Mangatu subcatchments. The Haunui Forest site is located in the Waingaromia subcatchment and lies 26 km to the southwest of Mangatu Forest. Both locations have a similarly high annual rainfall of 1000 – 2500 mm and history of vegetation disturbance with clearance of the indigenous forest cover for pastoral farming in the late 19th and early 20th centuries and reforestation for soil conservation purposes, predominantly with Pinus radiata, beginning in the 1960's.

The two survey locations differ in terms of their geology and landforms. Land systems are used to distinguish principal landforms. Mangatu Forest consists of the Mangatu land system while Haunui Forest consists of two land systems: Reporua and Waingaromia. The Mangatu land system is underlain by Whangai Formation, crushed late Cretaceous and Paleocene argillites (Moore *et al*., 1989). These argillite

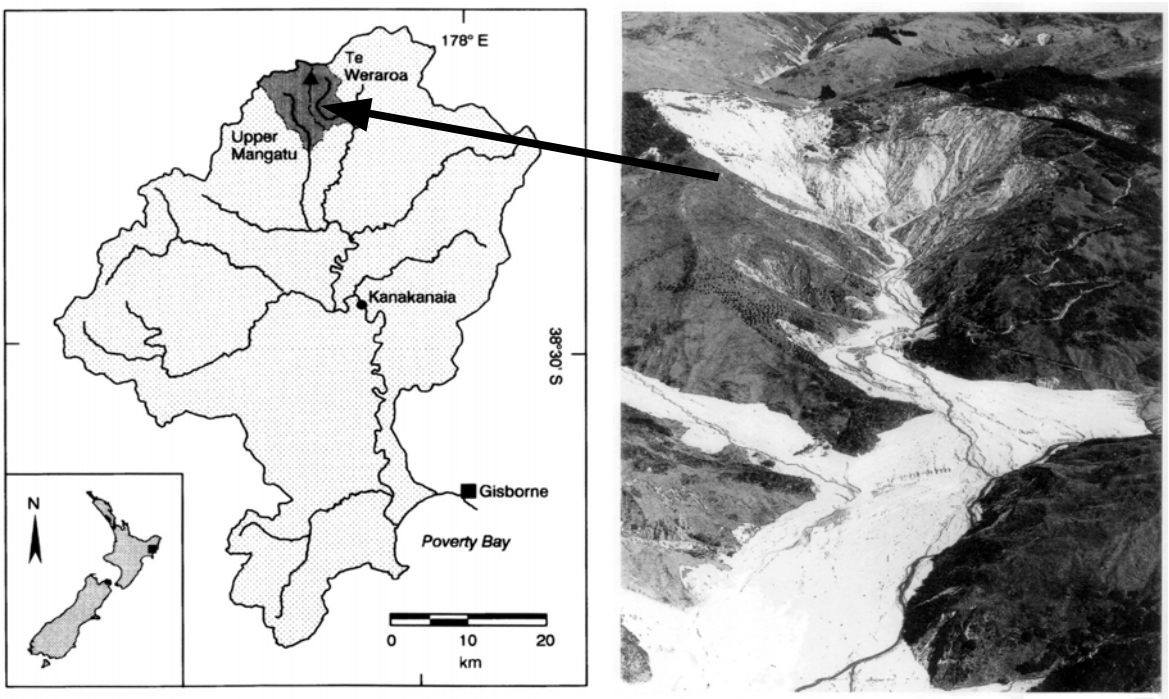

Fig. 20. Waipaoa River basin, Mangatu Forest site and Tarndale gully complex in 1961 (photo: J. H. Johns). The gully is c. 300 m deep and c. 500 m wide, and the width of the feeder channel into Te Weraroa Stream is c. 100 m (from, De Rose *et al.,* 1998).

rocks, which were crushed during emplacement of the East Coast Allochthon (Mazengarb *et al*., 1991), are particularly susceptible to acid sulphate weathering (Pearce *et al*., 1981) and the development of large, amphitheater shaped gully complexes. Hillslopes are long (c. 800 m) with an average slope angle of 25° and elevation range of between 250 and 630 m or more. Drainage basins vary from a few thousand square metres to 0.45 km^2 in area.

The Waingaromia land system in contrast, is underlain by Miocene to Pliocene, loosely jointed mudstones that are particularly prone to mechanical weathering processes. Here, deep seated earthflows are common, slope gradients average < 19° and hillslopes are long and range in elevation from 190 to 490 m. The Reporua land system has similar slope morphology, but is underlain by Paleocene to Oligocene bentonitic mudstones. Gullies tend to be more linear in form in the Waingaromia compared with the Reporua land system. Gullies at both these localities are generally smaller in size than at Mangatu Forest.

Among the most extensive and best known gully complexes are the Tarndale and Mangatu gullies, which are each about 0.2 km^2 in area (Figure 20) and engross much of the 1st- order drainage basin which contain them. Large quantities of sediment are moving out of these gullies into tributary river systems.

Under the native forest cover, capacious mass movements and gullying were sustained by tectonic uplift (c. 0.01 m yr^{-1} over the past 10^4 years), but it is thought that the gully complexes examined here were initiated early this century as a result of increased runoff that occurred after the native forest cover was converted to pasture (Allsop, 1973). Anecdotal evidence suggests that the Tarndale gully was initiated on the site of a mass movement that formerly had occurred under the native forest cover, some 20 years after the forest had been cleared, in the winter of 1915 (Gage and Black, 1979). Early attempts to control gully erosion in the headwaters of the Waipaoa catchment by check dams were largely ineffective, but post-1960 reforestation of the most severely eroded land in the critical headwater regions, helped stabilize many active gullies and reduced the rate of gully formation (Allsop, 1973; Water and Soil Directorate, 1987). On the largest amphitheater-like gully complexes, where the bare head- and side-walls can extend from the channel to the ridge top, erosion was too far advanced to be mitigated by reforestation. It was speculated that these gullies were the principal source of sediment entering the river system (Water and Soil Directorate, 1987).

6.3 Methods

Gullies deepen and enlarge over time. If the topography of a gully is known at two or more points in time, it should be possible to ascertain the volume of displaced sediment and estimate the amount of sediment generated. The contribution that gully erosion makes to sediment production has been previously evaluated using sequential aerial photography and orthodox photogrammetric techniques (Seginer, 1966; Dymond and Hicks, 1986; Poesen *et al*., 1996). Here, for the first time, high resolution multi-date DEMs are used to derive area and elevation differences on the rapidly eroding bare gully surfaces. DEMs permit calculation of the volumetric and mass rates of gully

erosion between specified time intervals.

At Mangatu Forest, a 4 km^2 area, that encompassed the Tarndale and Mangatu gully complexes and nine smaller gullies, was selected for study. Large format sequential aerial photography, suitable for DEM construction, had been captured on the 2nd June 1939 (1:12000), 23rd November 1958 (1:18000), and 20th February 1992 (1:15000). This gave time intervals of 19.5 (1939–1958) and 33.2 years (1958–1992) respectively, for estimation of erosion rates. At Huanui Forest, 11 gullies were investigated in the Reporua land system and 4 gullies were investigated in the Waingaromia land system. A different availability of dates of aerial photography (1955, 1969 and 1988) meant that DEMs were constructed for different periods (14 and 19 years) at Huanui Forest. The scale of photography was also somewhat greater (1:18,000 – 1:25,000) compared with the Mangatu site.

DEM construction was as follows. For each survey date, contact diapositives of stereo-pairs of photography covering the study site were obtained from aerial photographic companies and scanned at between 25 and 52 um, depending on scale. The necessary camera lens calibration data and fiducial mark measurements were obtained from relevant companies. GPS measurement to within ± 1 m on stable terrain features (survey markers, fence, buildings, tree stumps and trig stations), provided ground control for orientating stereo-pairs.

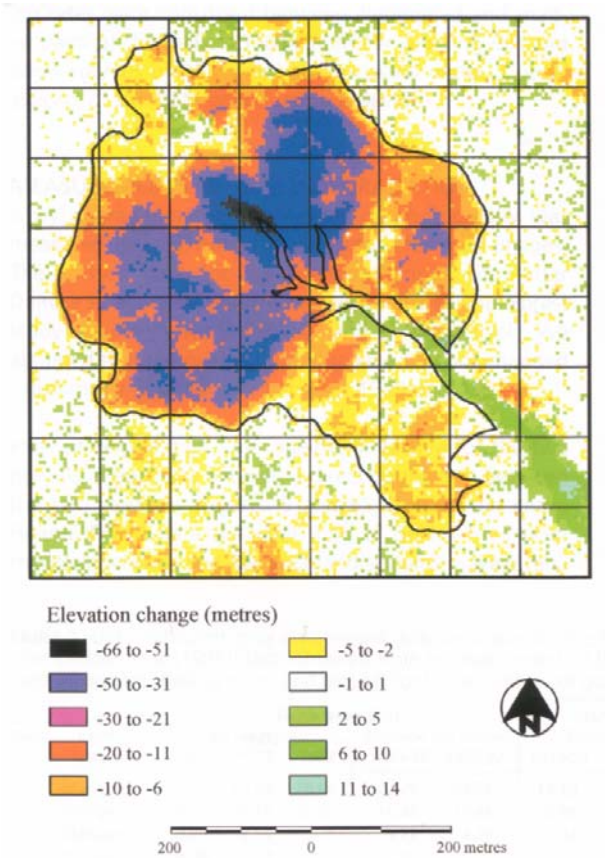

Fig. 21. Denudation and fan aggradation, Tarndale Gully (1939 – 1958).

DEMs at 5 m resolution were then constructed through block triangulation and parallax measurement during pixel matching (area correlation) of the scanned diapositives using IMAGINE Orthomax photogrammetric software. This computer software uses a textural recognition algorithm to match successive surface features on adjacent stereo images (analogous to landing the floating point in conventional photogrammetry). Elevation at regular grid spacings is then interpolated from the measured parallax. The accuracy of elevation measurements depend on the textural definition of imagery. Where definition is poor, such as in areas of shadow or near forest margins, then errors in elevation can result due to interpolation from surrounding areas. On average, between 20% and 40% of pixels were interpolated in each of the DEMs. Gully faces were generally well illuminated resulting in precise pixel matching in these locations. Interpolation occurred mainly in surrounding grassed or forested hillslopes, or in areas of shadow.

The relative error in elevation between DEMs was evaluated by comparing one hectare areas of stable land (usually interfluves or terraces). A total of 48 control areas were used at the Mangatu site while 30 were used at that Huanui site. In the absence of random and systematic errors, the mean and variance of differences in elevation at the same location on DEMs, should be negligible. However, random errors occur as a result of imprecise pixel matching, and systematic errors are generated by differences in triangulation and ground control. Both contribute to error in elevation estimates made by the software. Stochastic (random) errors due to inexact pixel matching within control sites were slightly larger for the Huanui DEMs (St.dev ± 2.4 to 2.9 m) than for the Mangatu DEMs (St.dev ± 2.0). Systematic errors were similar at both sites: -1.0 to +4.0 m at the Huanui site and -6.0 to -0.8 m at the Mangatu site. The DEMs were subsequently adjusted to correct for these systematic errors.

Shadowing was a particular problem in the 1939 scenes due to the low sun angle at the time of photography. Systematic errors from shadowing in the 1939 imagery at the Mangatu site required special treatment. Errors due to shading within individual gullies were estimated by applying the average elevation difference for only the unshaded areas to the gully as a whole. This assumes that average surface lowering is no different between shaded and unshaded parts of a gully. The interpolation process was found to "fill" areas in shadow, producing higher average elevation and overestimation of surface degradation between 1939 and 1958.

For each time period, elevation differences and erosion volumes were calculated by subtracting the DEM elevations of the earlier date of photography from the latter. The mass of sediment eroded during each time period was calculated by multiplying the sediment volume by an average dry bulk density of 2.0 ± 0.1 t m^{-3}. This measurement was determined from a number of cores of relatively unweathered argillite bedrock sampled at the base of the Tarndale gully.

6.4 Results

6.4.1 Gully sediment production rates

Volumetric production of sediment from the Tarndale and Mangatu gully complexes during the period 1939 – 1958 was initially estimated at 3.96×10^6 m^3 and 3.12×10^6 m^3, respectively. The estimated volumes are less at 3.48×10^6 m^3 and 2.58×10^6 m^3, respectively, if areas of shadowing are excluded from the calculation of mean elevation differences. During the latter period (1958 – 1992) volumetric production was greater at 4.83×10^6 m^3 and 5.03×10^6 m^3, respectively, due to the longer time period involved. The volumetric yields are equivalent to sediment production rates of $3.6 \pm 0.2 \times 10^5$ t yr^{-1} and $3.0 \pm 0.2 \times 10^5$ t yr^{-1} for the Tarndale gully, and $2.7 \pm 0.2 \times 10^5$ t yr^{-1} and $3.0 \pm 0.2 \times 10^5$ t yr^{-1} for the Mangatu gully, for the former and latter periods, respectively.

Total sediment production from all 11 gullies at Mangatu Forest from 1939 – 1958 was 7.88×10^6 m^3, equivalent to 8.1×10^5 t yr^{-1}. Sediment production declined to $6.2 \pm 0.2 \times 10^5$ t yr^{-1} during the period from 1958 to 1992, when many of the smaller gullies were stabilized by reforestation. The two largest gullies collectively produced 77% and 95% of total sediment production from all gullies during the former and latter time periods, demonstrating the importance of large gullies to sediment delivery to tributary rivers.

At Huanui Forest, the areal extent of individual gullies was typically smaller (0.1 ~ 12 ha) compared with those at Mangatu Forest (0.2 ~ 27 ha). Volumetric changes between dates of photography were correspondingly less. In most cases, the larger gullies individually produced 2×10^4 to 13×10^4 m^3 of sediment from 1955 – 1969 and 0.7×10^4 to 7.0×10^4 m^3 of sediment from 1969 – 1988. Total erosion at Huanui Forest from the 15 gullies amounted to 4.52×10^5 and 2.75×10^5 m^3, equivalent to sediment production rates of 6.5×10^4 and 2.9×10^4 t yr^{-1}. These are less than $^1/_{10}$ the rates of sediment production from gullies at Mangatu Forest: i.e., gullies at Huanui Forest are producing significantly less sediment because of their smaller size. The decrease in sediment production from gullies during the later period at both Mangatu and Huanui Forests reflects the stabilizing influence of trees after reforestation of hillslopes commenced in the 1960's. Four out of the 11 gullies at Mangatu Forest and seven out of the 15 gullies at Huanui Forest were completely stabilized by reforestation, reducing their sediment yields to almost zero.

6.4.2 Gully erosion rate - area relations

Plots of surface denudation rates versus gully area demonstrate, that with the exception of a few outliers, there is a power law relation between denudation rate and surface area of gullies: the bigger the gully, the greater the rate of denudation (Figure 22). This non-linear relation shows that larger gullies will produce a proportionately greater volume of sediment per unit area, such that:

$$D = kA^{0.5} \qquad (6)$$

where D is the average denudation rate in m yr^{-1}, A is the average gully area in hectares and k is the proportionality constant, which for the Mangatu and Reporoa gullies is similar at 0.12 ~ 0.26. Although gullies were too few in number to construct a significant relationship between gully denudation rates and area for the Waingaromia land system, the generally lower surface denudation rates for these gullies, suggest that the proportionality constant is also likely to be smaller in this land system.

Gully development over time is therefore similar for the Mangatu and Reporoa land systems, but is different in the more elongate gully morphologies of the Waingaromia land system. The greater degradation rate with increasing gully area presumably reflects the greater runoff volumes generated during rainfall events. Concentration of overland flow, particularly at the base of small ephemeral channels which extend onto gully faces, is producing greater relative channel incision on larger gullies. Results from the Waingaromia land system suggest that the more elongate gullies (shown as crosses in Figure 22) will tend to have lower average surface denudation rates for the same gully area, and therefore produce proportionality less sediment. This may, however, be a product of the higher proportion of shadows and/or vegetation in and around these types of gully.

6.4.3 Sediment contribution to export

The mean annual suspended sediment load of the Waipaoa River at Kanakanaia (Figure 20), where measurements of suspended sediment discharge have been made since 1960, is c. 10.7×10^6 t yr^{-1} (Hicks *et al*., 2000). The average bedload yield (estimated using Wilcock's (1998) modified Parker–Einstein formula) amounts to a small fraction (<1%) of the Waipaoa's suspended load. Much of the bed material released from gullies breaks down rapidly into finer particle sizes upon exposure to air. Sediment generated within the Tarndale gully complex has a D_{50} of 1·4 mm, and 60% is finer than 2 mm in diameter (Phillips, 1988). Consequently only a small portion of the sediment load comprises coarse bedload material and most sediment produced from gullies can be assumed to contribute to the suspended sediment yield of the Waipaoa River.

The Tarndale and Mangatu gully complexes therefore each generate the equivalent of 2 – 3 per cent of the total annual suspended sediment load of the Waiapaoa River at Kanakania, assuming no net change in channel sediment storage over time. The actual amount of sediment delivered to downstream reaches, however, depends on in-channel storage elements such as fans and terraces. Debri fans, which had formed at the base of the Tarndale and Mangatu gullies, increased in elevation during the measurement periods showing that some of the eroding sediment had gone into storage. Net aggradation on the Tarndale fan was computed to be 0.52×10^5 and 0.46×10^5 t yr^{-1}, amounting to 13% and 16% of sediment production for the periods from 1939 to 1958 and from 1958 to 1992, respectively. Aggradation on the Mangatu fan was 0.07×10^5 t yr^{-1} and 0.19×10^5 t yr^{-1}, amounting to 2% and 6% of

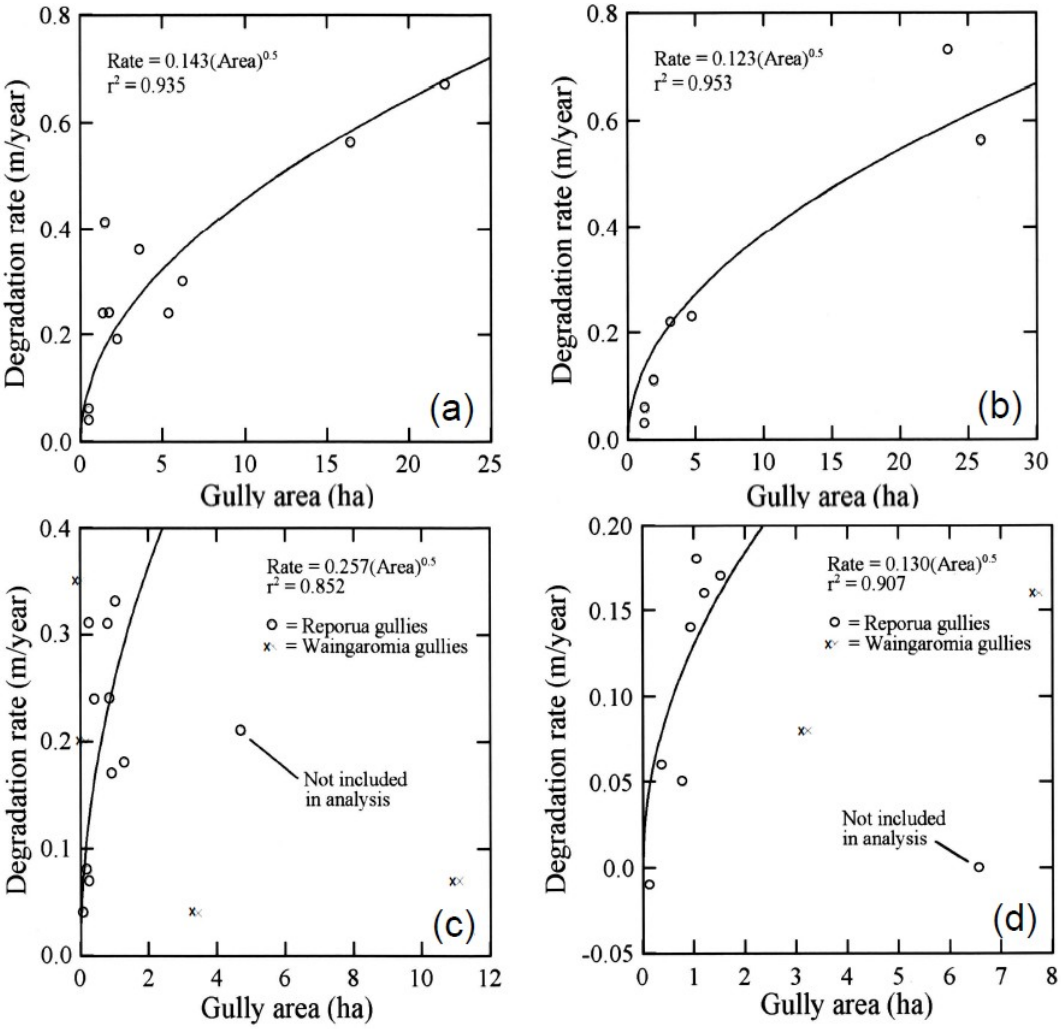

Fig. 22. Denudation rate versus gully area at Mangatu Forest site; a) 1939 – 1958, b) 1958 – 1992; and at Huanui Forest site; c) 1955 – 1969, d) 1969 – 1988 (from, Betts and De Rose, 1999).

sediment production, for these same periods. Clearly, the majority of sediment eroding from gully faces is reaching tributary streams.

6.5 Discussion

6.5.1 DEM construction

The results presented here show that the construction of high resolution multi-date DEMs from historical aerial photography can be used to quantify rates of gully erosion over periods of several decades or more. The method provides an efficient and cost effective way of deriving volumetric production rates from gullies. The generally small-scale of historical photography available in most regions (typically > 1:25,000) will however tend to limit the accuracy with which elevations can be determined.

Random errors, which are apparent as small pits and mounds on shaded DEMS and produce localized errors in elevation of ± 2 - 3 m, tend to average out and become relatively unimportant over larger areas of land. This high frequency noise can be effectively removed by using local neighborhood filters (Walker and Willgoose, 1999). Systematic errors that arise from errors in triangulation and ground control, shadows on photographs, scanning equipment used and camera lens calibration data have proven to be of much more concern. In this study, the use of one hectare training areas in regions of stable land, demonstrate that there were systematic errors in elevation of several meters or more across DEM difference images between dates of photography. In contrast to random errors, systematic errors do not average out over larger areas of land, and corrections are necessary to DEM elevations, before volumetric production rates from gullies can be calculated.

Shadows, resulting from low sun angles at the time of photography or trees surrounding gullies, are of much greater concern because the errors they generate are more localized in extent and difficult to evaluate. The research results presented here suggest that the interpolation algorithms used in digital photogrammetry software, overestimate elevations on shaded hillslopes. Shadows of the two largest gullies produced systematic errors in estimated sediment production volumes of 13 – 17% and these are significantly larger that the 3% error attributable to

random noise in DEM elevations.

More recent investigations (Betts *et al*., 2003; Martínez-Casasnovas *et al*., 2003) have further extended the methodology developed here, to producing more accurate measurements of gully erosion at larger scales (<1:10,000). Martínez-Casasnovas *et al*., (2003) achieved a high level of vertical accuracy with maximum random errors in elevation of ± 0.07 m and a systematic error of 0.35 m, using 1:5000 – 1:7000 scale photography. Betts *et al*., (2003) similarly produced DEMs with low systematic errors in elevation of −0.13 ± 0·01 m and random errors of ± 1 m or less, using 1:8000 – 1:10,000 scale photography. Such levels of accuracies should in future permit a study of the impact of individual storm events on sediment production from gullies over much shorter periods than investigated here.

6.5.2 Sediment budget applications

Perhaps the single most important research outcome from this study is to show that large gullies complexes (such as the Tarndale and Mangatu gullies) within the Waipaoa River basin do not individually dominate the sediment budget of this catchment, and produce at most 3% of the basins sediment yield at Kanakania. The catchment area above Kanakania (1582 km^2) comprises 72% of the Waipaoa River basin and the estimated contribution from large gullies has been recently revised by Marden *et al*., (2005) to be individually only 1 ~ 2% of the 2205 km^2 Waipaoa River basins sediment yield of 15 ~ 17 × 10^6 t yr^{-1}.

Furthermore, small gullies may be relatively unimportant sediment contributors to river loads, because based on results from the Mangatu Forest site, the largest gullies produce 77% - 95% of the total sediment yield from all gullies. However, the study sites at Huanui and Mangatu Forests cover relatively small areas and clearly the sediment contribution from numerous gullies elsewhere means that gullies may collectively dominate the sediment budget of the Waipaoa catchment. Determining the overall contribution that gullies make to the sediment budget of the Waipaoa River requires examination of a much larger catchment area and number of gullies. It is generally not feasible to construct multi-date high resolution DEMs over such large areas (the cost would be prohibitive) and alternative methods are required.

The finding that gully surface denudation rates are closely related to their size is an important outcome of this research work (Figure 22). The power law relationship (Equation 6) that exists between denudation rate and gully area provides an efficient means of estimating gully volumetric and sediment yields across much larger areas of land without the need to construct high resolution DEMs. This general approach was recently implemented by Gomez *et al*., (2003) and Marden *et al*., (2005) to determine the total sediment yield from gullies within headwater subcatchments of the Waipaoa basin. Using research findings presented in this paper, they adopted a modified relation between volumetric production rates (V in m^3 yr^{-1}) and gully area (A in ha) due to the slightly higher correlation coefficient when combining measurements from both time periods (Equation 7).

$$V = 460 + 2750A + 160A^2 \qquad (7)$$

Within the 29 km^2 Te Weraroa catchment (which contains the Tarndale gully), Gomez *et al*., (2003) found that the number of active gullies declined from 55 (occupying c. 1.7 km^2) at the peak of activity in 1960 to 12 (occupying c. 0.18 km^2) by 1988. Using Equation 7, gullies were estimated to have produced 28.7 Mt of sediment over the 38 year period, equivalent to an average sediment production rate of 7.6 × 10^5 t yr^{-1}. This is only about 2.5 times the average sediment production estimated from the Tarndale gully complex. From an analysis of channel surveys taken along the lower 8 km of the Te Weraroa Stream between 1948 and 1996, Banbury (1996) found that nearly half (48%) of sediment produced by gullies went into channel storage, causing bed elevations to rise by around 8 m in most places. A further ~10% of sediment from gullies is stored in lower-order feeder channels and fan deposits immediately downstream from gullies (Marutani *et al*., 1999; Kasai *et al*., 2001; De Rose *et al*., 1998). Consequently, when channel storage elements are taken into consideration, gullies within Te Weraroa catchment have collectively contributed on average, only about 2% of sediment to the Waipaoa River. Although the level of contribution was about twice this amount at the peak of gully activity in 1960, the relative contribution has since declined to about 1.9% by 1988, owing to a decline in sediment production from gullies (4 × 10^5 t yr^{-1}) and proportion of sediment going into channel storage (24%).

Within a larger 107 km^2 areas of Cretaceous terrain at Mangatu Forest (including the Te Weraroa Stream catchment), Marden *et al*., (2005) investigated sediment production from all gullies for 3 periods between 1939 and 1988. Total sediment production amounted to 2.9, 3.2 and 1.2 Mt yr^{-1} for 1939 – 1960, 1960 – 1970 and 1970 – 1988, respectively. This represents 17%, 19% and 8% of the estimated 16.7, 17 and 15 Mt yr^{-1} suspended sediment yield of the Waipaoa River for the same periods (Marden *et al*., 2005). Since about ½ of the sediment went into storage along channels in the two early periods (Gomez *et al*., 2003), then the net percentage contributions to the Waipaoa River was at most about 10% from this area. Today, 37 gullies remain active within Cretaceous terrain at Mangatu Forest and sediment contribution from these gullies makes up a small component of the sediment yield of the Waipaoa River.

These studies, which cover comparatively small catchment areas, tend to suggest that sediment produced from gullies is relatively unimportant at the whole of basin scale. There are however numerous gullies within Tertiary terrain, and other catchments within the Waipaoa River basin. Griffiths (1982) estimated that c.33% of the Waipaoa River's suspended yield came from gullies within the larger Waingaromia and Waihora sub-catchments. Today, there are about 420 active gullies remaining within the Waipaoa

catchment, of which only about 25% have been reforested (Marden *et al*., 2005). This is over 10 times the number of currently active gullies within Cretaceous terrain at Mangatu Forest and hence it is probable that total sediment contribution from all gullies is > 50%, and perhaps as much as 80%, of the Waipaoa River's suspended sediment yield. Without treatment, the remaining 289 gullies on pastoral hillslopes are likely to expand in size, and continue to deliver large volumes of sediment to the Waipaoa River and its tributaries.

There are other erosion processes operating within the Waipaoa River basin that contribute sediment to river loads and no sediment budget is complete without considering these processes. River bank, channel incision, earthflow and landslide erosion are probably the most important processes other than gully erosion. Reid and Page (2002) analyzed landslide density – rainfall relationships and estimated that shallow landslides contribute 15 ± 5% of the suspended sediment load in the Waipaoa River above the Kanakanaia gauging station. Furthermore, 75% of the sediment produced by landslides occurs during storms with recurrence intervals of less than 27 years. Further downstream at the Matawhero gauge (1910 km^2) the maximum proportional contribution of shallow landslides is estimated at about 13% (Reid and Page, 2002). Hence, gullies contribute significantly more sediment than do landslides to the sediment budget of the Waipaoa River. The contribution from other erosion sources, remains largely unknown.

Clearly, channels provide a considerable buffer for sediment delivery, having adsorbed much of the sediment eroded from gullies prior to 1970, only to be later re-released as channels degrade. In Te Weraroa Stream, the rate at which sediment accumulated in the channel was at a maximum in the period from 1948 to 1960, but declined throughout the period from 1960 to 1975, and showed little change after 1975 (Banbury, 1996). This trend suggests that the relative contribution of sediment from the Tarndale gully complex to the sediment load of the Waipaoa River will have increased over this period. Today, degradation along many smaller tributary channels (Marutani *et al*., 1999; Kasai *et al*., 2001) continues to supply sediment to the main channel and offset declines in sediment production from gullies. Thus, although the total amount of sediment generated by gully erosion has declined with time, the decline in the amount of sediment that goes into storage along Te Weraroa Stream, suggests that the proportion of the sediment generated by the Tarndale gully complex that exits the catchment is probably greater now than it was in the past. Gomez *et al*., (2003) estimate that it will take ~90 years for the Te Weraroa channel to degrade to its 1948 level, and much longer to reach the paleo-channel that existed prior to deforestation.

6.5.3 Influence of vegetation

Gully measurements demonstrate that the program of reforestation which commenced in the 1960's has proven effective at stabilizing many gullies. Within the 140 km^2 Mangatu Forest in the headwaters of the Waipaoa River, Marden *et al*., (2005) mapped 300 gullies of varying size from 0.02 to 26 ha. Together, they occupied about 4% of the catchment area at the peak of gully activity in 1960, but their combined extent had declined to c. 1.5% by 1988, following a programme of conservation plantings which began in the 1960's. Most of the smaller gullies were stabilized by reforestation while the largest gullies were largely unaffected. Marden *et al*., (2005) estimate that there are about 420 active gullies remaining in the Waipaoa Catchment. Similarly, Marden (2003) suggest there are about 900 active gullies in the 1734 km^2 Waiapu catchment. Twenty-four of these were examined in detail by Parkner *et al*., (2006) and found to occupy 12.5% of the 5 km^2 Weraamaia subcatchment. Due to the higher temporal resolution of their study, the authors were able to establish that the extent of gully activity corresponded not only to changes in vegetative cover (gradual reversion from pasture to scrub), but also to the frequency of storm events: phases of gully expansion were synchronous with periods of increased storminess and the occurrence of extreme rainfall events. Thus the effectiveness of woody vegetation in stabilizing gullies seems to depend not only on gully size, but also on the frequency of large magnitude storm events capable of causing instability.

6.5.4 Future trends

Multi-date DEMs derived from historical photography permit not only an accurate measure of the volumes and rates of surface degradation, but can also show on what parts of the gully surface, erosion is most active.

An important aspect of the management of gullies is to know how long these features will remain active. While many of the smaller gullies at Mangatu Forest have been effectively stabilized by afforestation (Marden *et, al*., 2005), both the Tarndale and Mangatu gully and numerous other large gully complexes remain active 100 years after initial development. The high average surface denudation rates (~ 0.8 m yr^{-1}) cannot be sustained for long periods of time, else hillslopes would eventually erode to below current stream elevations. It is thought that riverbed incision makes the active stage of the gully evolution much longer than with a stable base level (Sidorchuk, 2006). Stream base levels therefore exert a significant control over gully development and it is likely that the gullies will eventually adopt a much more stable morphology, particularly if the channel bed elevations remain stable. How long this process takes is largely a matter for speculation, though observations of gully development on the Tarndale and Mangatu gullies over the two time periods from 1939 to 1992, suggest that the gullies will eventually exhaust themselves.

Both the surface area of active erosion and rate of erosion on the Tarndale gully complex has decreased by 25% from the peak of activity in 1959 to the latest date of aerial photography taken in 1992. Over the same period both the size and elevation of the fan formed at the base of the gully has increased. The decrease in

gully area has largely resulted from stabilization and reforestation of valley sideslopes towards the lower end of the gully complex. In contrast, the headwall of the gully has advanced at a rate of about 2 – 3 metres per year, while the locus of gully activity has similarly migrated towards the catchment boundary. The area of the catchment containing the Tarndale has grown in size somewhat because the gully headwall has migrated past preexisting ridge lines and now occupies a small portion of what were once adjacent catchment areas. Although ridges have reduced in elevation, and fans at the base of the gully have increased in elevation, steep gully walls and high rates of erosion have been sustained by parallel retreat and migration of the gully headwall. Eventually, however, the area of steep headwall will diminish as the gully eats into lower elevation hillslopes in the adjacent catchments, but this may take at least as long again as the gully has been currently active.

In comparison, the neighboring Mangatu gully complex is at a earlier stage in gully development. In 1939 the gully complex occupied ½ the area of the Tarndale gully and an incipient fan had just began to form along the feeder channel. The Mangatu gully, although producing a similar volume of sediment annually, has continued to grow in area since 1959 and by 1992 it was 35 % larger than the Tarndale gully. Consequently, peak erosion rates on the Mangatu gully have occurred at a later time than on the Tarndale gully complex, where rates are now declining.

These gully complexes are clearly following an evolutionary path, beginning with an incipient gully forming at the head of 1st-order tributary channels, which then rapidly expands in area due to increased runoff and slope instability from forest removal. The steep gully headwall rapidly encroaches on the surrounding ridge line by the combined action of slumping and sheetwash/rill erosion processes. Parallel retreat of the gully headwall results in migration of the locus of gully activity towards the catchment boundary and into surrounding catchments. At the same time, lower valley sideslopes stabilize from the combined effects of slope decline and reforestation. Eventually, the gully contracts as the steep headwall reduces in area, until such time that gully slope and rate of denudation diminish and allow for reestablishment of forest vegetation. Subsequent incision into tributary fan deposits may reinitiate erosion at the base of the gully, but this is unlikely to be as active as in the first phase of gully development, since the overall relief of surrounding hillslopes has been greatly reduced in elevation. Since smaller gullies occupy a smaller proportion of their respective catchment areas, and have much lower rates of denudation (Figure 22), they will not be incising as actively into the valley floor, and reforestation can help stabilize these gullies before the onset of rapid gully expansion onto surrounding slopes. Marden *et al*., (2005) have shown, for example, that for gullies < 1 ha in area, there is close to a 100% probability of stabilization and forest closure 20 years after reforestation, while for gullies 5 ha in area the probability is only 60%. Poesen *et al*., (2003) similarly note that gully erosion is particularly important above a threshold catchment size of 1 to 10 ha.

Sidorchuck (1999, 2006) suggests that there are two main stages in gully development: a dynamic phase during which the morphological characteristics of a gully are far from stable, followed by a static phase when the gully has largely stabilized in area and adopted morphometric equilibrium. The initial phase lasts a small portion (5%) of a gullies lifetime, but it is during this time that most of the length and area of the gully forms and erosion is most intense at the gully bottom: more than 90% of gully length, 60% of gully area and 35% of gully volume form at this time (Sidorchuk, 2006). For most of a gullies lifetime it is stable and near to its maximum size. Sediment transport and sedimentation are the main processes at the gully bottom, while gully width increases due to lateral erosion processes. Sidorchuk (2006) further considers that the initial phase of gully evolution has characteristics of a self organizing system that is close to crisis. The gully will remain is this state until such time as the gully becomes relatively more stable and loses these characteristics. Sidorchuck (1999, 2006) considers a relatively linear-unidirectional trajectory of change for gullies evolving from an initial state, through rapid expansion, stability and to final quiescence and revegetation. This may, however, be only true of large gullies that can sustain relatively high erosion rates, even when rainfall intensities are comparatively low. Parkner *et al*., (2006) suggest that for many gullies the trajectory of change is much more non-linear and dynamic in character, with phases of expansion inter-dispersed with periods of inactivity. Periods of activity relate to changes in vegetation and occurrence of rainfall events with sufficient intensity and duration to overcome innate stability thresholds of the geological substrate in which gullies are formed.

Gully initiation appears to relate to some critical threshold condition. Erosion begins when the critical combination of slope and catchment area produces characteristics of flow (e.g., stream power) that are sufficient to result in incision into, and destabilization, of the underlying substrate material. Clearly the strength properties of substrate materials are important, with gullies tending to form in regions where the underlying lithologies are weakened by tectonic forces or where the soil materials are deeply weathered and easily dispersed (as with Sodic soil types). Research elsewhere, has tended to focus on slope-area threshold relations in defining the most probable location or sites within stream networks for gully initiation. Slope is usually inversely correlated with catchment area in stream networks and the relation for gullies, although following a similar trend to the overall catchment, occupies relativity steeper channel slopes for the same area. However, such slope-area thresholds may not exist in all catchments. Hancock and Evans (2006), for example, found that there was no threshold slope-area relation for gullies in the catchments of the Northern Territory of Australia they investigated, and gullying occurred throughout catchments. At best, most gullies occur at catchment areas of less than 2 ha.

7. SUMMARY and CONCLUSIONS

This paper has presented and discussed research findings from two regions of New Zealand where mass movement erosion processes are of major concern to land management agencies. The focus of this research has been to quantify volumetric and mass rates of erosion, and to place these within the context of long-term rates of erosion and sediment yields at the basin-wide scale. Shallow landslide erosion is investigated in eastern Taranaki hill county, while mass movement gully complexes are investigated in East Coast catchments.

A novel paired-hillslope approach was used to measure landslide erosion rates on pastural hillslopes in Taranaki hill country and to conclusively demonstrate that, had the forest vegetation not been removed, the landslides would not have occurred. On hillslopes deforested for 85-years, average surface lowering was 0.15 m, equivalent to an average erosion rate of 1.8 ± 0.5 mm yr^{-1}. Erosion rates increase with increasing average gradient of hillslopes, because most landslides are located on slopes above 28° and are concentrated onto slopes steeper than 32°. Erosion rates are higher at 2.4 ± 0.6 mm yr^{-1} and 2.7 ± 0.8 mm yr^{-1} for these slopes. A single high intensity rainfall event (Cyclone Hilda) in 1990 produced an average surface lowering of 0.041 m over the study site, demonstrating the importance of large magnitude, infrequent storm events to hillslope erosion. Moreover, there were proportionately more landslides on slopes deforested for 10 years (12%) than deforested for 85 years (4%), demonstrating that hillslopes can acquire event resistance to subsequent storm events and erosion rates can be expected to decrease with time.

On weathering limited slopes, the rate of soil production from bedrock defines the long-term rates of erosion. Measurement of soil depths on landslides scars of varying age in Taranaki hill country, has demonstrated that soils recover with time due to bedrock weathering and surface transport of soil materials. The process of soil accumulation on landslide scars can be defined by a logarithmic function of increasing soil depth with time. The rate of accumulation diminishes, from 3.5 mm yr^{-1} in the first 40 years after slipping to 1.2 mm yr^{-1} for the ensuing 50 years, due to a decrease in the contribution of soil material from the revegetating scar surface and sidewalls. When scaled to include non-landslide areas, the average rate of soil accumulation is lower at 0.57 mm yr^{-1}. The rate of soil accumulation is expected to further decrease with time and comprise mainly bedrock weathering. Long-term rates of soil production from the consolidated sandstone bedrock are probably not much greater than 0.4 mm yr^{-1}. Hence, the contemporary rates of landslide erosion are over 5 times higher than rates of soil accumulation, implying semi-permanent losses in soil from hillslopes. Relatively deep soils that accumulated over long periods of time under the native forest are gradually being replaced by shallower, less productive soils.

Systematic measurement of slope angle and soil depth at sites in Taranaki hill country, together with stratigraphic dating of soil coverbeds, has demonstrated a technique for assessing the long-term rates of erosion due to landsliding. The pattern of variation in slope and soil depths, imply considerable variability in erosion rates, both in space and time, and a landscape in morphologic disequilibrium. Diffuse creep of soil during periods of relative stability and infrequent landsliding on the steeper slopes are the main erosion processes influencing variation in average soil depths. Soil depths decrease on average with increasing slope angle, reflecting an increase in erosion intensity. Gentle slopes (<31º) have been stable throughout the Holocene and diffuse transport processes have produced deep accumulations of colluvial materials, which increase in depth with distance downslope, and are thinnest along convex ridges and spurs. Here, long-term erosion rates are about 0.1 mm yr^{-1}. In contrast, on steep slopes (>31º), past landsliding has produced a patchwork of soils, who's depths bear little relationship to slope form, but rather reflect different stages in soil recovery with time. Here long-term rates of erosion are about 1 mm yr^{-1}. Hence, there is a threshold slope (31º), near the modal angle for hillslopes, above which there is about an order of magnitude increase in erosion rate under the native forest.

Assessing the relative increase in erosion rates following forest removal therefore depends on which part of the landscape is being compared: for gentle slopes (<28º) there is little or no change; for slopes from 28 to 32º there is about a 10 to 20 fold increase; and on slopes steeper than 32º there is about a 3 to 5 fold increase in erosion rates. Consequently the modal class of slopes appear most affected by deforestation since these slopes were largely immune to landsliding during the Holocene.

The research has demonstrated that the spatial variability in landslide density was controlled by a number of factors: vegetative cover, slope angle, and recency of forest clearance and previous landslide activity. Empirical (multivariate) and slope stability (limit equilibrium) models have been investigated as a means of predicting landslide occurrence and extrapolating the research findings from the small survey sites to larger regions of land. The physically based slope stability model suggests greater relative instability and a threshold slope angle less than is actually observed. Because of the requirements for detailed spatial information about soil strength and hydrological properties at the hillslopes scale, this type of model has limited usefulness. The empirically based landslide susceptibility model currently provides the most suitable model for extrapolating research findings to greater areas of land. DEM derived spatial maps of slope and flow accumulation, provide a way for mapping landslide susceptibility over large areas of land. Scale and resolution are important factors affecting DEM derived attributes and it is unlikely that the coarse resolution, contour derived DEMs, will adequately resolve the short, steep slopes, typically of many areas of hill country in New Zealand. The empirical model is also largely restricted in use to the consolidated sandstone hill country, and will require

re-parameterization before application to other areas of landslide prone hill country.

In East Coast hill country of the Raukumara Peninsula, spectacular large mass movement gully complexes are contributing large volumes of sediment to the Waipaoa River. Outcomes of the research have demonstrated that it is possible to generate multi-date DEMs of gully surface elevations from historical aerial photography, along with adequate ground control provided by GPS. Multi-date DEMs of the same area, once corrected for any systematic errors in elevation, provide a direct measure of change in surface elevation and volumetric rates of erosion. An estimate of sediment yield over two time periods between 1939 and 1992, conclusively demonstrated that large gully complexes do not individually dominate the sediment budget of the Waipaoa catchment, because they produce no more than 1 – 2% of the 6800 t km^{-2} yr^{-1} of sediment exported by this catchment. Prior to the research being undertaken at Mangatu and Haunui Forests, sediment yields and relative contributions to the Waipaoa river system were largely unknown.

Change in surface elevation also provides a measure of fan aggradation at the base of gullies and of the proportion of eroding sediment delivered to tributary channels. Aggradation on fans at the base of the Tarndale and Mangatu gully complexes, over the period from 1939 to 1992, was equivalent to 13 – 16% and 2 – 6% of gully sediment yield. These results show that much of the sediment produced from gullies finds its way to tributary channels and only a relatively small portion is deposited on fans.

An important outcome of this research was to demonstrate that the rate of surface denudation of a gully is proportional to the square root of its planimetric surface area. For amphitheater shaped mass movement gully complexes, erosion rates increase from about 0.2 m yr^{-1} for the smaller gullies of several hectares, to 0.8 m yr^{-1} or more, for the larger (20 – 30 ha) sized gullies. The power law relationship between surface denudation rate and area can be used to calculate the total sediment yield from all gullies within a catchment, where it is not feasible to construct multi-date DEMS because of financial or time constraints. This approach was applied to the 29 km^2 Te Weraroa catchment, to show that gullies had produced 28.7 Mt of sediment between 1950 and 1988, equivalent to 5% of the Waipaoa River's average sediment export (though about 60% of this is deposited in channels and on fans). Similarly, gullies within the larger 140 km^2 Mangatu Forest catchment, produced 3.2 Mt yr^{-1} or 19% of the Waipaoa River's sediment export at the peak of activity between 1960 and 1970 (though this has since declined due to reforestation). Without the initial work undertaken on constructing multi-date DEMs from historical aerial photography, this type of analysis would not have been possible.

Not all gullies have the same morphology and this can influence erosion rate per unit area. Multi-date DEM analysis of the smaller and more elongate gullies at Haunui Forest, showed that these gullies tend to have lower denudations rates per unit area compared with amphitheater shaped gully complexes. This demonstrates the importance of gully morphology in estimating rates of denudation: the same power law relationship cannot be applied to all gullies within a catchment. Catchment sediment budgets need to take account of variation in gully morphology, else significant error in estimated sediment yields will result.

The research into gully erosion has also demonstrated that reforestation with conservation trees is effective at stabilizing smaller gullies of several hectares in size or less. By 1992, total sediment yield from the 9 smaller gullies at the Mangatu Forest study site, had declined to one fifth the level that existed prior to reforestation in the early 1960's. In contrast, large gully complexes were mostly unaffected by reforestation. Because large gullies produce a disproportionately high component of total yield, their presence limits the effectiveness of catchment reforestation programmes. Furthermore, changes in sediment storage along tributary streams, means there will considerable phase lags in response between catchment headwaters and basin outlets. At the peak of gully activity in the 1960's, around 60% of sediment produced from gullies aggraded on fans and in tributary channels. Channel beds have now stabilized and adsorb little of the sediment released from gullies. Declines in channel storage have largely offset declines in sediment produced by gullies, such that sediment exported to the Waipaoa River has changed little in recent decades. Conservation management of this type of land is clearly a long-term strategy and it will take many decades before the declines in gully activity in headwater basins will be felt in terms of reduced sediment loads along lowland rivers.

The research brought together in this manuscript and previously separately published in peer-reviewed journal papers provides important contributions to our understanding of erosion processes at the local and catchment scale. Research findings are of particular significance in understanding the consequences of changes in land use (deforestation/afforestation) and providing quantitative information for sediment budget applications at the catchment scale.

Outcomes of the research are invaluable for providing tools for: the mapping and monitoring of erosion prone lands; providing input to catchment scale sediment budgets; and for providing informed debate about the various land use strategies required for improving catchment condition and water quality. The sustainable management of regions affected by anthropogenic erosion also requires methods for the identification and mapping of areas sensitive to land use change and this is best undertaken using empirically-based land susceptibility modeling.

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