THE MORPHOLOGY AND SEDIMENTOLOGY OF TWO UNCONSOLIDATED QUATERNARY DEBRIS SLOPE DEPOSITS IN THE ALEXANDRIA DISTRICT, CAPE PROVINCE.

By

P.M. Illgner

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"All science is concerned with the relationship of cause and effect. Each scientific discovery increases man's ability to control future events" - Laurence J. Peter, 1978.

"Surficial deposits contain a discontinuous, transient record of depositional geomorphic events. The history contained in these deposits is essential for understanding the spatial distribution and long-term recurrence of rare catastrophic events." - R.B. Jacobson *et al.*, 1989

Dedicated to:

Those heavenly bodies into
whose scintillating orbit I
meandered, only to remain
thrillingly amused!

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ABSTRACT

Research on hillslope surface processes and hillslope stratigraphy has been neglected in southern Africa. The amount of published literature on hillslope stratigraphy in southern Africa is very limited. Hillslope sediments provide a record of past environmental conditions and may be especially useful in calculating the recurrence interval of extreme environmental conditions such as earthquakes and intense rainfall events. The characteristics of hillslope sediments provide information as to their origin, transport and mechanisms of deposition. No published work could be found that had been undertaken on hillslope surface processes or stratigraphy in the eastern Cape coastal region. This study attempted to fill this gap in the geomorphic literature for southern Africa. The surface processes acting on hillslopes at Burchleigh and Spring Grove in the Alexandria district of the eastern Cape were examined in terms of slope morphology, surface sediment characteristics and the internal geometry of the hillslope sedimentary deposits. The late Quaternary hillslope sedimentary deposits at the two study sites are composed of fine grained colluvial sediments intercalated with highly lenticular diamicts. The fine grained colluvial sediments were emplaced by overland flow processes while the diamicts were deposited by debris flows. The sedimentary sequences at both study sites have a basal conglomerate interpreted as a channel lag deposit. Most slope failures preceding debris flow events were probably triggered by intense or extended periods of rainfall associated with cold fronts or cut-off lows. Seismic events may also have triggered slope failure, with or without the hillslope sediments being saturated. The results of this study indicate that a continuum exists between the slopewash dominated processes of the presently summer rainfall regions of Natal to the present winter rainfall regions of the western Cape where mass movement processes are significant. Hillslope deposits, therefore, provide a record of environmental conditions which may greatly facilitate proper management of the landscape.

CHAPTER 1

1.1. INTRODUCTION

"In contrast to glacial, fluvial, and aeolian deposits,..., colluvium has received little study, despite its obvious importance for the study of hillslope evolution." - H.H. Mills, 1987

This thesis examines colluvial sediments of late Quaternary age mantling two slopes near Alexandria, eastern Cape Province, South Africa. It attempts to ascertain the processes by which these sediments were deposited. Figure 1.1 shows the location of both slopes. The processes were identified by the sedimentological characteristics displayed by each deposit.

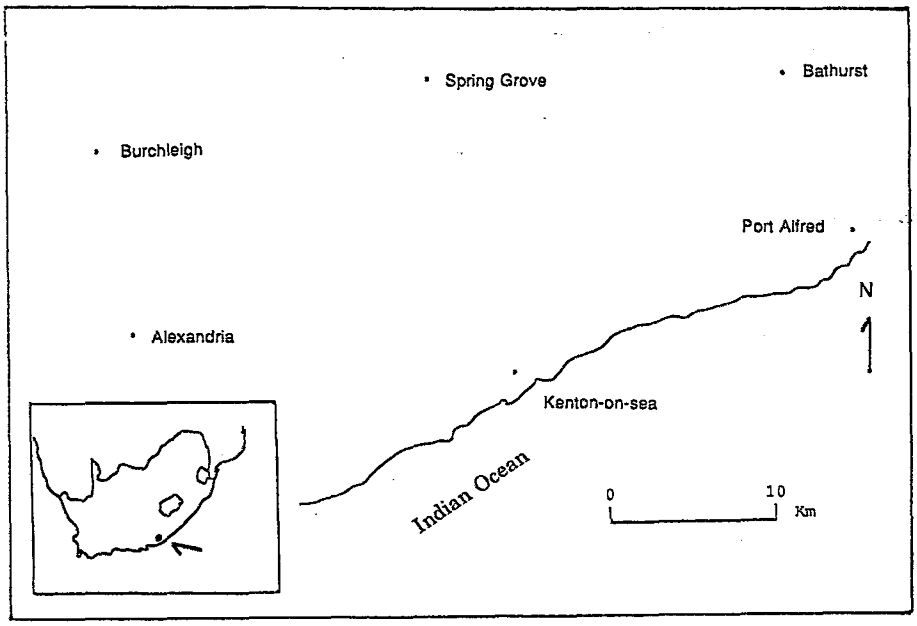

Figure 1.1. The location of the study sites, Burchleigh and Spring Grove in the eastern Cape coastal hinterland.

There is a voluminous literature on the individual processes active on hillslopes. However at present "there is a great paucity of work on the environmental conditions responsible for colluvial deposits and for the processes involved in their formation" (A.S. Goudie -

pers.comm., 1993). Furthermore, knowledge of hillslope palaeohydrology is rudimentary (J.Costa pers.comm., 1993). Govers (pers.comm., 1993) has pointed out that the characteristics of colluvial deposits may change over very short distances making it very difficult to connect them to average hydraulic conditions on hillslopes. However, by using simple flow and sediment transport relations one can attempt to model the character of colluvial sediments (J.R.L.Allen pers.comm, 1993).

Sedimentary systems are process-response systems in which the rate of energy supply is reflected by the texture and configuration of a sedimentary body (Allen, 1974). Conceptually, debris slopes have been viewed as a cascade; slope form reflecting a balance between the rates of input, throughput and output by weathering, hydrological and gravitational processes.

Debris slope deposits are often considered to be colluvia. "Colluvia" are poorly sorted mixtures of clay-, silt-, sand-, and gravel-sized particles, which accumulate on lower hillslopes (Watson *et al.*, 1984). The sediment is derived from upslope, where weathered bedrock has been stripped off and transported to the lower slopes, mantling pediments and choking former stream channels (Watson *et al.*, 1984). The above may be contrasted with alluvium, which is well sorted and bedded and with eluvium, *in situ* weathered bedrock still in "primary" position (Watson *et al.*, 1984).

Research into debris slope deposits is of more than academic interest, for it is widely recognised that an understanding of their processes greatly facilitates effective slope management (Scott, 1971; Carson and Kirkby, 1972; Young, 1978; Schumm, 1979; Nilsen, 1982). For example the slopes within the Alexandria region have been classified as unstable (Paige-Green, 1989). The total social and economic impact of landslides alone in southern Africa during the 1987 fiscal year amounted to close on 20 million dollars (US) (Paige-Green, 1989). Furthermore, analyses of slope deposits may help increase our knowledge of Quaternary climate fluctuations and facilitate prediction of future climate trends. It may also facilitate assessment of slope failure risk in areas marked for development.

This study examines *surface* processes acting on hillslopes in terms of slope morphology, surface sediment characteristics and the internal geometry of debris slope deposits.

Current climate data (approximately 100 year records) was used to predict whether suitable conditions have occurred in the recent past to produce the sedimentary features identified in each deposit and where relevant, their recurrence interval.

1.2. OBJECTIVE

"An approximate answer to the right question is worth a great deal more than a precise answer to the wrong question." Anonymous.

The objective of the study was to identify the surface processes which resulted in the deposition of two unconsolidated late Quaternary deposits near Alexandria, Eastern Cape Province.

1.3. SPECIFIC AIMS

1. To undertake a literature review within which the results of this study could be contextualised.
2. To produce a geomorphic map of each study site.
3. To measure the slope gradients at each site.
4. To take photographs which illustrate the internal structure/relationships of the sediments at each study site.
5. To produce stratigraphic sections of the exposed sediments at each study site.
6. To undertake a textural analysis of the sedimentary units identified above.
7. To undertake a textural analysis of the surface sediments overlying each study slope.
8. To gather data on present seismic activity and meteorological conditions in the study areas, so as to identify day to day conditions and potential trigger mechanisms for catastrophic events.
9. To relate this information on trigger mechanisms and slope sedimentology to known changes in the southern African palaeoenvironment.

CHAPTER 2

THEORETICAL BACKGROUND - HILLSLOPE SYSTEMS

"...Rigorous stratigraphical analysis is a prerequisite to extension of contemporary process studies into the past" - W.J. Vreeken and H.J. Mucher, 1981.

2.1. INTRODUCTION

The stratigraphy of debris slope deposits helps determine which parts of the episodic, stratigraphic record can be attributed to extreme events under present climatic conditions as opposed to major changes in Quaternary climate (Jacobson *et al.*, 1989). Hillslope sediments provide a more sensitive record of extreme climatic events than alluvial sediments, because debris and colluvium have longer residence times on slopes than alluvium (Jacobson *et al.*, 1989). Jacobson *et al.* (1989) suggested that the stratigraphic record may be used to gain better insight into the long-term importance of catastrophic events for which the historical record is too short. The elucidation of process-response relationships on a slope is dependent on the assumption that slope form is in equilibrium with the types of processes acting on a slope, their resultant deposits, and the geological time span within which the results of the study are to be discussed.

Colluvial sediments on footslopes have the potential to record aspects of slope palaeohydrology. Hayward and Fenwick (1983) suggested that "the stratigraphy, morphology, and subfossils of colluvial deposits frequently..." provide "...evidence of changes in local slope movements...related to hydrological conditions."

The role of extreme meteorological and seismic events in the development of slopes has been widely discussed in the literature (e.g. Wolman and Gerson, 1978; Starkel, 1979; Selby, 1982; Keefer, 1984). Extreme events can be regarded as intense short-term events with a recurrence interval of more than a year; mostly once every few hundred or tens of years (Starkel, 1979).

Starkel (1976) suggested that extreme events associated with the processes of slope failure and slope wash may be triggered by the following extreme meteorological phenomena:

1. High intensity rainfall, such as that during thunderstorms.
2. Prolonged rains of lower intensity and covering much wider areas than the above, mostly the result of landward advection of oceanic air-masses. The advection of humid air-masses is often associated with continuous rainfalls (1-5 days) of relatively low intensity. The light rain and high humidity favouring infiltration and saturation of unconsolidated debris slope material.
3. Rapid warming, which causes snow and ice to melt, often in association with precipitation (Starkel, 1976).

Weathering and the removal of rock and soil on hillslopes is not a uniform process in a spatiotemporal sense, but episodic; dependent on the availability of energy, a transporting medium and the incipient instability of a landform (Schumm, 1979; Selby, 1982). Two of the more important influences on hillslope processes are climate and slope form.

2.2. RELATIONSHIP BETWEEN CLIMATE AND HILLSLOPE PROCESSES

Meadows (1988) suggested that geomorphological features are usually more sensitive to changes in precipitation than temperature. One obvious exception is the mean annual air temperature associated with periglacial activity (Lewis, 1988). A very wide range of climates have occurred within southern Africa during the Quaternary, from periglacial through to humid sub-tropical conditions (Tyson, 1986).

Tricart (1970), French (1976) and Karte (1983) have tried to relate climatic conditions to geomorphic processes, specifically periglacial activity. French (1976) defined the periglacial domain empirically as all areas with a Mean Annual Air Temperature (MAAT) of <+3°C. Frost action dominates at MAAT's <-2°C, while frost action processes are present but do not dominate when MAAT's are >-2°C but <+3°C.

Short-lived heavy rains are the main motive power in the transformation of relief in arid areas of the subtropical zone and in less elevated regions of the tropical and temperate zones. Starkel (1976) suggested that short-lived heavy falls cause slope wash, linear erosion and soil flows, while continuous rainfall events initiate mudflows, debris flows, landslides and sijels (Caucasian: "debris-muddy streams" carrying large amounts of material) in valleys.

Starkel (1976) believes that short-lived downpours play an important role in arid/semi-arid climates, whereas continuous rainfall is more important in tropical and temperate humid climates. Mediterranean climates are particularly susceptible to extreme events, where up to 25% of slope surfaces may be mobile at any one time. Starkel (1979) suggested that wetter years with longer-lasting rainy spells initiate deep landslides.

Unlike very arid or humid areas, semi-arid areas are sensitive to very small changes in external conditions such as small changes in rainfall or vegetative cover (Carson and Kirkby, 1972). Statham (1976) studied gully erosion and the associated debris flows on vegetated scree. He related gully formation to a change in environmental conditions. He noted that there appears to be very little climatic control on debris flow activity. Debris flows have been reported from semi-arid, cool and warm humid temperate, periglacial, high arctic and high alpine climates. He related debris flow activity to the availability of coarse material with sufficient fines "...to ensure that high water contents..." are "...maintained, and to situations where water flow is concentrated..." rather than to climatic regimes (Statham, 1976). A minimum volume of material is needed, even in the presence of very heavy rainfall, before slope failure can take place.

2.3. RELATIONSHIP BETWEEN SLOPE FORM AND HILLSLOPE PROCESSES

"Process and form are indivisible. Neither can exist without the other" - Charles S. Denny, 1967.

The direct and indirect action of water forms the basis of most slope processes. The recognition of the modes of water transfer downslope both at and beneath the slope surface is fundamental to the understanding of slope systems (White *et al.*, 1984).

A complex feedback exists between slope form and slope processes. The form of a slope is a function of its initial form, geological structure, vegetative cover, climate and subsequent modification by denudational processes. Everard (1963) has suggested that the "...concavoconvex form and its subsequent destruction are closely related to fluctuations of base-level, climate and vegetation in the late Quaternary." Selby (1982) has argued that the dominant geomorphic processes on hillslopes may vary downslope as the slope angle decreases. He reported that in New Guinea, the threshold angles for certain processes were as follows:

1. Mudflows: 2°
2. Rotational slumps: 8°
3. Debris and complex landslides: 15°
4. Debris avalanches: 25°-30°.

Simple deductions based on the relationship between slope form and hillslope processes are seldom possible. These relationships are often affected by characteristics peculiar to a particular rockface or debris slope, such as its morphology, aspect, geology or gradient (Luckman, 1976; Statham and Francis, 1986) in addition to local perturbations in the regional climatic regime. The net result may be an unconsolidated sedimentary deposit showing a range of failures, degrees of organisation and bedding.

In addition, hillslopes may be viewed as weathering-limited or transport-limited depending on the nature of interaction between the rates of weathering and transport operating on each slope. A weathering-limited environment tends to produce a straight slope profile and a transport-limited environment a concave slope profile (White *et al.*, 1984).

Hogg (1982) suggested that upper slopes are weathering limited, whereas lower slopes are transport limited. The rate of weathering varies over the length of each slope, with more debris on the lower slopes than the upper ones, largely as a result of gravitational processes. This induces a feedback mechanism, with slope form playing the lead role. The concave sections of a slope may be expected to be more thickly covered in debris than their convex counterparts. Concave slopes have a greater tendency to concentrate flows and hence deposition. However, as Emmett (1978) has pointed out, studies can sometimes give contradictory results, as the data collected for the U.S. Geological Survey at Badger Wash, Western Colorado has shown. Research indicated straight segments of slope to have the greatest erosion, with no significant difference in amount of erosion between convex and concave segments.

White *et al.* (1984) have suggested that the magnitude-frequency of various processes may be gleaned from slope profile morphologies. Rapid mass movement processes tend to produce dissected slopes with an irregular profile. The lower frequency hydraulic processes produce a smooth slope profile with few irregularities. Their relative effectiveness: the ratio of soil creep to surface wash in several environments, is shown in Fig.2.1.

	Creep	:	Wash
humid temperate (UK)	10–20	:	1
savanna	1	:	5
warm temperate (Australia)	1	:	7
semi-arid (New Mexico)	1	:	98

Figure 2.1. The ratio of soil creep:slope wash in different climatic regimes (after White *et al.* 1984).

Abrahams *et al.* (1985) have developed a theoretical model which shows that variation in the nature of the hillslope gradient-mean particle size relationship with plan form is consistent with debris slope formation by and adjusted to hydraulic processes. The mean particle size decreases with slope gradient and increased distance downslope. This corroborates earlier work by Hooke (1968) who showed experimentally that the steady state slope is determined by debris size, depositional process, and water discharge. In addition, White *et al.* (1984) have suggested that the laws of hydraulic geometry ensure rills develop concave longitudinal profiles, which causes the lower parts of a hillslope profile to adopt a concave tendency.

Abrahams *et al.* (1985) identified a continuum of debris slope types, with debris slopes underlain by closely jointed and/or mechanically weak rocks at one end and those underlain by widely jointed, mechanically strong rocks at the other. The former are transport-limited, well adjusted to current hydraulic processes, and exhibit a strong relationship between hillslope gradient and mean particle size. The latter are weathering-limited, poorly adjusted to hydraulic processes and display little or no relationship between hillslope gradient and mean particle size. Hooke (1968) ascribed such differences on alluvial fans to their source-area lithologies. The lithology influences fan slope by:

1. Controlling debris size
2. Controlling depositional processes
3. Controlling sediment concentration in flows reaching a fan.

The specific nature of hydrological and gravitational processes are reviewed in the following chapter.

CHAPTER 3

HILLSLOPE PROCESSES

"Yet no matter how sophisticated an open system is devised based on measured contemporary energy fluxes, it cannot hope to do credit to the true evolutionary complexity of landforms." - K.W. Butzer and D.M. Helgren (1972)

3.1. HYDRAULIC PROCESSES

The hydraulic processes operating on the surface of a slope fall into one of two categories, depending on whether the applied forces of falling or running water are responsible for particle movement. The kinetic energy of falling raindrops produces rainsplash erosion, while overland flow causes surface wash and occasionally gully formation (Hogg, 1982; White *et al.*, 1984).

3.1.1. Raindrop Impact and Sediment Transport

Falling raindrops impart a proportion of their kinetic energy to the loose sediment on the soil surface, throwing particles out radially from the point of impact (Elwell, 1986).
Slope gradient tends to lengthen the trajectory of soil particles downslope and shorten them upslope, ensuring net downslope movement of surface sediment particles. Sand and silt-size particles are relatively easily moved, whereas clays, with their cohesive characteristics, resist transportation (White *et al.*, 1984). Pitty (1971) has suggested that raindrop splash may carry a higher percentage of large particles and aggregates than surface flow. Particles in the 50-400 micron range are the most susceptible to movement. Rainsplash is dependent on the vegetative cover, since vegetation absorbs the kinetic energy of raindrop impact (White *et al.*, 1984).

Mucher *et al.* (1981) found during rainwash (combined splash and flow) that "...clay size particles are removed and exported...". They also found that deposits resulting from turbulent rainwash had a ratio $[\%(10\text{-}50\mu m)/(\%<30\mu m)]>1$; whereas for flow without splash (e.g. afterflow), the finer fractions were preferentially deposited with a ratio $[\%(10\text{-}50\mu m)/\%<30\mu m]<1$. This was used as an indicator of the significance raindrop impact had on downslope transport of fine grained sediments.

Young (1972) suggested that while raindrop impact may be a powerful agent of soil detachment it is less important than surface wash as an agent of erosion. Detachment by raindrop impact is not dependent on position on a slope, or to any significant degree on slope angle, although grains moved by this process can be expected to move faster downslope on steep slopes than on more gentle ones.

During periods of intense rainfall, raindrop splash forms a crust on the soil surface by compacting the soil and sealing pore spaces with fine particles (Farres, 1978). This changes the environment from one of high detachment/low transportation of sediment in the pre-overland flow condition, to one of low detachment/high transportation in the overland flow state. Air trapped in soils during heavy downpours may reduce the efficiency of infiltration (Starkel, 1976) and enhance overland flow.

3.1.2. Overland Flow and Sediment Transport

The equations used to describe sediment transport by overland flow were originally derived from sediment transport in rivers (Roels, 1984; Guy *et al.*, 1990). However, they only consider turbulent flow conditions, whereas overland flow may be turbulent, laminar or a combination of both (transitional) (Roels, 1984). A short outline of the principles of sediment transport in river systems follows.

3.1.2.1. Sediment Transport and Deposition in a Fluvial Environment

Particles move when the combined forces of lift and drag "...produced by the fluid exceed the gravitational and cohesive forces of the sediment grain" (Reineck and Singh, 1980). Cohesive forces are only important in fine-grained sediments (<0.1mm). Entrainment is determined by water flow velocity and grain size (Reineck and Singh, 1980). Grain shape, grain position, sediment composition, turbulence, and the type of packing all play a part in grain entrainment (Reineck and Singh, 1980). Angular grains move slower over sediment than more rounded ones (Francis, 1973).

The diagrams of Hjulstrom (Fig.3.1.) (Reineck and Singh, 1980) and Shields (Fig.3.2.)

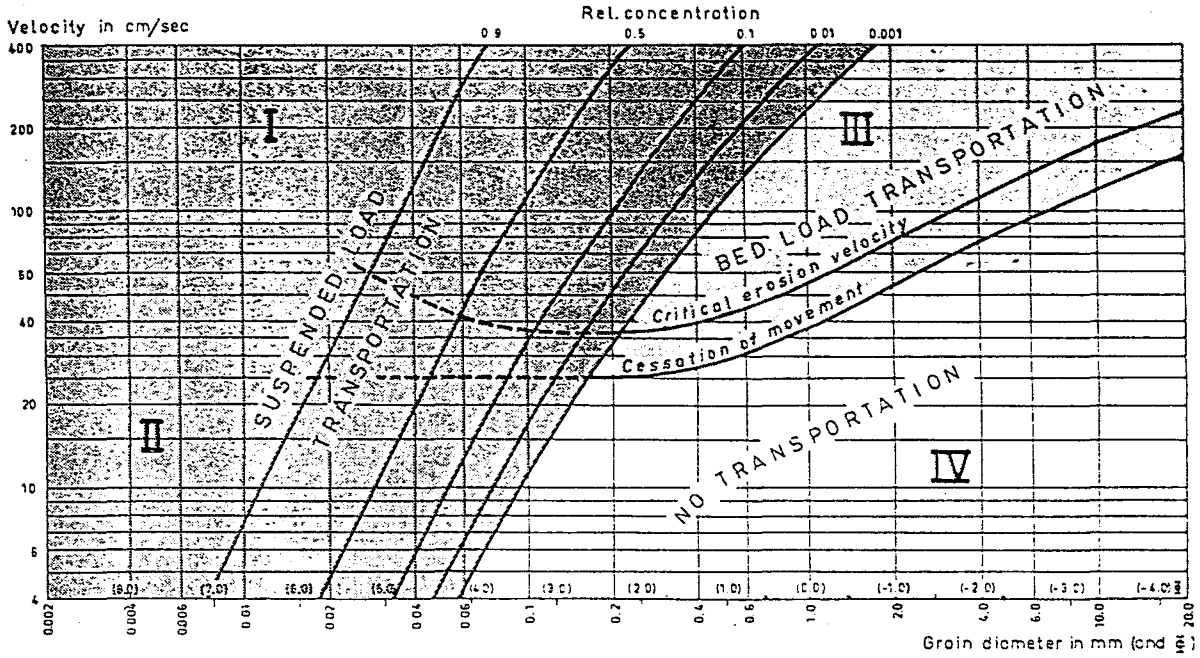

Figure 3.1. The modified Hjulstrom diagram (Sundborg, 1967) illustrating the relationship between grain size, flow velocity and the state of sediment movement for uniform material (density 2.65 e.g. quartz). The velocity is taken as that 100cm above the bed surface (after Reineck and Singh, 1980).

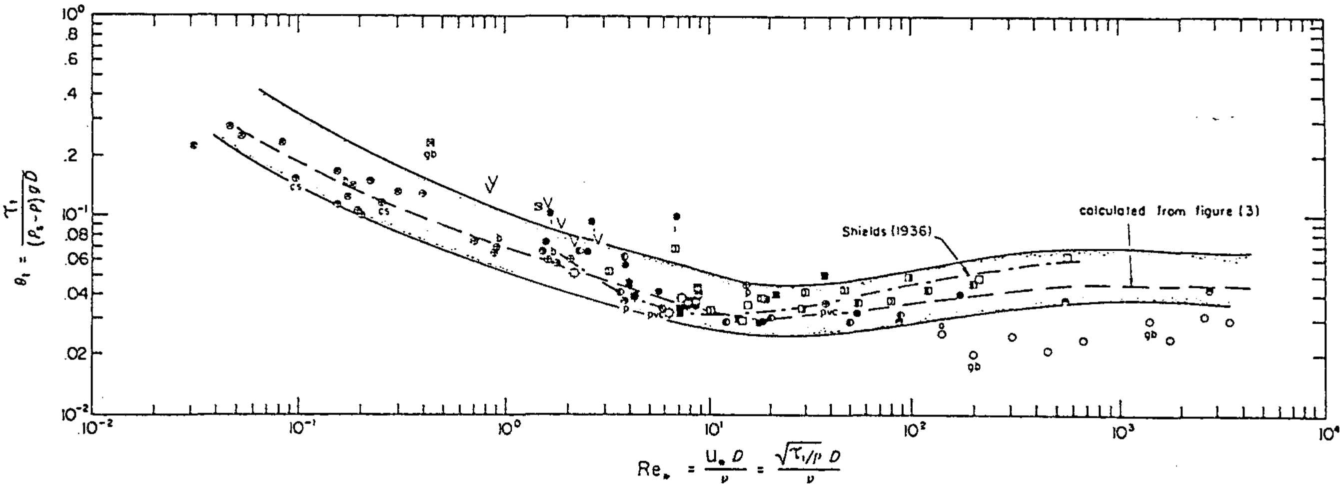

Figure 3.2. The modified Shields curve (after Miller *et al.*, 1977) [θ_t=shields' relative stress, τ_t=shear stress of fluid flow, ρ_s=density of sediment, ρ=density of fluid, g=gravitational acceleration, D=grain diameter, Re_*=dimensionless grain Reynolds number, u_*=friction velocity, ν=kinematic fluid viscosity]

(Miller *et al.*, 1977) are those most frequently used to predict grain entrainment. The former relates mean flow velocity to the grain size of bed particles, but is largely limited to quartz grains under river conditions (Miller *et al.*, 1977). In motion, sediment behaviour becomes a function of settling velocity (Reineck and Singh, 1980). Clay sediments may require more energy for entrainment than some sands, yet will be deposited long after sand grains have settled.

Very few studies appear to have been carried out which relate grain settling velocity to the threshold of motion in unidirectional currents (Komar and Clemens, 1986). This is a problematic relationship as the effects of both grain shape and grain density need to be taken into account.

The boundary between bed load and suspension load in the Hjulstrom diagram, approximately 0.2mm, is diffuse. The diagram is largely qualitative due to the difficulty in obtaining or applying the values used to delimit the four regions indicated (Reineck and Singh, 1980). Bed material is usually non-uniform and flow turbulent. Bed roughness is associated with grain size, while bedforms which strongly influence grain movement are not accounted for in the diagram (Reineck and Singh, 1980).

Shear stress at the bed boundary is a better indicator of the initiation of grain movement than flow velocity, as flow velocity varies with depth (Reineck and Singh, 1980). High concentrations of sediment within flows cause deposition at far lower energy levels than the critical entrainment conditions (Reineck and Singh, 1980). Notwithstanding the problems and shortcomings of the methods outlined above, the "...Shields' criteria or Hjulstrom-type diagrams may provide some information about the minimum energy for flow, especially in gravel sediments" (Reineck and Singh, 1980).

Carson and Griffiths (1985) critically assessed the use of equations in determining the relationship between the critical tractive stress (τ_c) and the onset of bed particle movement. They suggested the τ_c for entrainment of D_{50} bed material in sand bed channels may be represented by:

τ_c=$(D_{84}/D_{16}).(0.91).D_{50}$ (Modification of Shields expression)

D_x=grain size for which percentage "x" is smaller.

In a general review of selective grain entrainment by a current from mixed sediment size beds Komar (1987) concluded that no empirical relationship exists which can provide a good mathematical description of such situations. Komar (1987) found that entrainment measurements from deposits "...of mixed sediment sizes form trends which obliquely cross the threshold curves for uniform grains, the crossing point being roughly at the median diameter (D_{50}) of the size distribution." The coarser size-fractions would thus require lower-flow stresses and the fine fractions higher-flow stresses for their entrainment than would be the case in uniform beds.

The calculated critical shear stresses for grain movement in heterogenous sediments are often different from those for homogenous sediments (Wiberg and Smith, 1987). The difference is "...primarily due to the relative protrusion of the particle into the flow along with differences in the particle angle of repose, or bed pocket geometry, that results from having a mixture of grain sizes on the bed" (Wiberg and Smith, 1987). Grains larger than the mean are moved easier than those smaller than the mean.

3.1.2.2. Overland Flow and Hillslope Sediments

"I can, offhand, think of no reason in principle why one could not reconstruct flow conditions on hillslopes from the deposits. However, one would first have to be certain that one was looking at deposits from flowing water not subsequently modified by other processes" - Michael Church (pers.comm., 1993).

Overland flow has been defined as the flow of water towards a stream channel during the initial phase of surface runoff (Emmett, 1978). Selby (1982) has pointed out that sheetwash rapidly becomes concentrated into rills as it is diverted around objects such as large stones or vegetation.

Knowledge of surface runoff or overland flow is based on Horton's (1945) proposed theory of sheetflow, sheetwash erosion, and rill formation (Selby, 1982; Dunne and Aubry, 1986). Surface runoff as overland flow takes place when rainfall intensity exceeds the infiltration capacity of the soil or as a result of soil saturation (Hillel, 1980; White *et al.*, 1984). Water accumulates in small hollows on the soil surface before spilling over and running downslope as a broad irregular sheet of turbulent flow.

Overland flow only occurs where the "saturated throughflow" is forced to the surface, such as at the base of a slope (Chorley, 1978; White *et al.*, 1984), or when complete soil saturation occurs. "Saturated throughflow" refers to groundwater flow at shallow soil depths (Emmett, 1978; White *et al.*, 1984). Maximum soil erosion in the overland flows takes place when the flow depth approaches raindrop diameter, normally 3-6mm.

Soil particles become entrained when their "critical tractive force" is exceeded by the shear stresses induced by overland flow, which thickens and accelerates downslope (White *et al.*, 1984; Dunne and Aubry, 1986). Long slopes have greater potential for grain entrainment than short slopes (Schumm, 1962). Research (Selby, 1982) has shown that surface wash and soil transport on slopes are minimal except where soils are impermeable, thin, or unprotected by vegetation (e.g. in semi-arid areas). Modest precipitation rates and well structured soils tend to inhibit overland flow in humid environments.

Guy *et al.* (1990) investigated the hydraulics of sediment-laden sheetflow and the influence of simulated rainfall. They found flow to be laminar, while the uniform flows used by them "...were characterised by very shallow depths and large-scale roughness." This supports the assertion by Carson and Kirkby (1972) that surface wash erosion is a special case of fluvial transport, where thin flows occur over a very rough surface. Roels (1984) concluded that overland flow over rough surfaces is laminar at low Reynolds values (<100).

Everaert (1991) found that rainfall had a "negligible" effect on the transport of fine sediments, in contrast to coarser sediment. At higher flow intensities the effect of rainfall decreases and may become negative (Everaert, 1991). Everaert (1991) pointed out that problems with the reliability of grain entrainment formulae still exist even when effective stream power is used which considers depth in trying to account for roughness elements such as irregular bedforms and/or vegetation elements.

These issues influence the efficacy with which palaeo-overland flow conditions may be calculated. Furthermore, the translocation (vertical downward movement) of fines must be considered when reconstructing slope palaeohydrology (Hayward and Fenwick, 1983). Nevertheless the following equations are useful indicators of palaeohydrological conditions on hillslopes. Just as equations developed for overland flow were derived from river studies so palaeohydrological equations developed for palaeofluvial conditions can be useful indicators of palaeo-overland flow conditions.

Andrews (1983) discussed the entrainment of particles from a bed of non-uniform sized material and the implications for palaeohydrological studies. Andrews (1983) found that for particles 0.3 times to 4.2 times the median diameter of the subsurface material, the critical dimensionless shear stress necessary for entrainment (τ_{*ci}) may be expressed as:

$$\tau_{*ci}=0.0834(D_i/D_{50})^{-0.872}$$

D_i=particle diameter
D_{50}=subsurface median grain size

For particles >4.2 times D_{50}, τ_{*ci} approaches a constant value of ~0.020 in non-eroding channels (Andrews, 1983). Reid and Frostick (1984) found that the shear stress required for initial motion of bedload in coarse alluvial channels is about three times that at the time of deposition. The particle size distribution represents the hydraulic conditions at deposition and not entrainment (Reid and Frostick, 1984). They found that sediment behaves as predicted by Shields' curve when D_{90} (grain size for which 90% of sediments are smaller) is taken as the effective particle size. They concluded that a bed behaves "...as though it were composed of its component parts" (Reid and Frostick (1984).

Church *et al.* (1990) have reviewed the palaeovelocity concept and associated problems. They concluded that the modified Keulegen equation of Thompson and Campbell (1979) (as quoted in Church *et al.*, 1990) is the best current representation of palaeohydraulic conditions for very shallow flows.

The Thompson-Campbell equation is given as:

$1/(f^{1/2})=(1-0.1k/R)2.03\log(12.2R/k)$ (Church *et al.*, 1990)

R=hydraulic radius (depth)
k=roughness length scale=$3.5D_{84}$
f=Darcy-Weisbach equation friction coefficient=$[8gRS]/v^2$
D_{84}=grain size for which 84% of sediment is smaller
v=mean velocity of flow
S=slope=sine of slope angle

Savat (1977) assumed that the hydraulic radius (R) was equal to mean depth (d) on slopes as flow depths were always small in relation to width of flow. A channel he assumed to be infinitely wide when the width is >5 times the flow depth.

Rotnicki (1991) has reviewed some of the problems in applying formulae under particular conditions. These include the textural properties of sediment such as particle size, packing and fabric used to estimate various hydraulic parameters (Rotnicki, 1991). However, for the estimation of these parameters there needs to be an unlimited availability of a range of sediment sizes or else flow competence estimated from the largest particles present will be underestimated (Rotnicki, 1991). In addition viscous forces become more important than inertial forces for particles below 5-8mm in diameter. This means that the velocity and depth of palaeoflows "...cannot be estimated on the basis of sand material" (Rotnicki, 1991). Rotnicki (1991) used several different equations to mitigate the effects of using empirical relationships that are only valid under certain prescribed conditions. He used 1:10 000 maps to calculate slope gradients. Rotnicki (1991) concluded that the application of steady uniform flow formulae to palaeo-alluvial channels developed in alluvial sands yields promising results, with standard errors in estimation ranging from 7% to 20%.

Abrahams *et al.* (1988) concluded that the functional relationship between τ_c and D for overland flow on the hillslopes they studied was significantly steeper than predicted by the Shields relationship for river flow and may be represented by:

$$\tau_c=138{,}636D^2$$

The equations may be used to predict τ_c but should not be used to predict D or other relationships between variables (Abrahams *et al.*, 1988). Abrahams *et al.* (1988) emphasized the preliminary nature of their study and the fact that their findings pertain to the transport rather than entrainment of sediment.

There appear to be very few studies which combine descriptive and quantitative data for hillslope sediments. Mucher and De Ploey (1977) investigated experimentally and in the field the effect of slope angle, rainfall intensity, raindrop impact, overland flow, and concentrations of suspended material on erosion and redeposition of loess. They found that the following conditions give rise to the following sedimentary characteristics:

1. "Pluvial" rainfall, by which Mucher and De Ploey (1977) meant overland flow and raindrop impact, results in gravel to laminated deposits which are tightly packed and moderately sorted.
2. After flow (overland flow after rainfall cessation) sediments are very well laminated, well sorted and loosely packed.
3. Raindrop impact they found results in no lamination and no sorting.

Mucher and De Ploey (1984) found that "...artificially created afterflow deposits of calcareous silt loam...were well layered, parallel to the surface, without cross-bedding, while the individual laminae displayed excellent sorting."

3.2. GRAVITATIONAL PROCESSES

Gravitational processes are the downslope movement of unconsolidated sediments or rock under the influence of gravity, but without the direct aid of media such as water, air, or ice. Water and/or ice may however be indirectly involved in mass wasting by reducing the yield strength of the slope materials.

Translational flows occur when coarse debris, fine-grained sediment or clay are liquified. These three conditions are termed "debris flow", "earthflow", and "mudflow" by Selby (1982). For the purposes of this review they have been grouped together under one heading, "debris flows", as a combination of two or all three often takes place. For example, a landslide upslope may become a debris flow downslope, where greater soil pore water pressures are likely to be present. Debris flow development is facilitated by soil remoulding during landslides, the presence of clays with high liquid limits in high rainfall areas, the presence of soils with low liquid limits in areas of low rainfall (as little water is required to liquify the sediment), the presence of soils with open fabrics or where thawing of soil ice has taken place (Selby, 1982). This review is largely confined to debris flows as they are the only gravitational processes evident in the studied stratigraphic sequences.

Enos (1977) cited inclined clast fabrics parallel to flow, preservation of delicate clasts (e.g. shale), blocks projecting from the top of flows and the absence of flute marks at the base of debris flows as evidence for laminar flow. A similar view was expressed by Gloppen and Steel (1981) who ascribed the projection of clasts above the top of the bed and the lack of any significant erosion between beds to high matrix strength and laminar flow, "...at least during the final stages of deposition." Scouring and channel incision and lack of inverse grading are attributed to turbulent flow (Shultz, 1984).

Debris flows occur as single events on slopes normally in excess of 25°, but occasionally on those <10°. Flow activity is dependent on three basic requirements:

1. Steep slopes

2. Available regolith
3. High pore pressures

(Innes, 1983)

A debris flow normally occurs when non-indurated materials, often with a high clay content, absorb a lot of water (White *et al.*, 1984). Loss of material strength is caused by high pore water pressures and a resultant loss in cohesion of the sediment prior to failure. Many scientists (Starkel, 1979; Young, 1978; Caine, 1980; Rogers and Selby, 1980; White *et al.*, 1984; Crozier, 1986; Gallart and Clotet-Pervarnau, 1988; Suwa and Okuda, 1988; Jacobson *et al.*, 1989) have thus associated these processes with high rainfall events. Landslides often occur in clusters, related to the "...isohyetal pattern of the associated weather systems..." (Crozier, 1986). Throughout the world the precipitation associated with low-pressure systems is a major trigger mechanism. These systems include frontal convergence and deep troughs of low pressure.

The material within a debris flow is highly mobile, reaching velocities of up to several metres per second (White *et al.*, 1984). Morphologically the debris flow may be divided into three components in plan view, namely a source area (often a hollow where moisture can accumulate), a narrow flow track and a broad lobate toe at the base of a slope, where the material has fanned out and come to rest. As slope failure often precedes a debris flow on hillslopes it is reviewed briefly below.

3.2.1. Preconditions for Slope Failure

Young (1972) suggested that the following passive slope conditions favour slope failure:

1. Lithological: weak or unconsolidated rocks, particularly clays.
2. Structural: permeable strata overlying impermeable clays and downslope dip.
3. Topographic: cliffs or steep slopes initiated through basal erosion.
4. Hydrological: surface or subsurface concentration of water and seepage within rock or regolith.
5. Climatic: Exposure to high intensity storms.

Intrinsic factors which affect slope instability include regolith shear strength, the shear strength provided by roots, regolith permeability, regolith thickness and slope morphology (Jacobson *et al.*, 1989). Extrinsic factors include the antecedent soil moisture, and the intensity and duration of rainfall (Jacobson *et al.*, 1989). After failure and regolith removal, bedrock weathering and/or accumulation of colluvium within the landslide scar occurs before the critical thickness for subsequent failure is reached (Jacobson et al, 1989). Shallow slides are probably the most common form of slope failure and instability on natural slopes (Carson and Kirkby, 1972).

Karpuz (pers.comm., 1993) suggested that slope failure initiation is mainly controlled by:

1. Gradient
2. Sediment supply
3. Increased rainfall and runoff
4. Rise or fall in base level
5. Earthquakes and/or liquefaction

Shear stresses on a slope are a function of overburden weight, slope angle (which in non-cohesive materials is more important than slope height) and the geotechnical properties of slope sediments. The infinite slope approach to stability analysis for shallow slides can be used to predict the threshold angle for slope failure (Carson, 1976) and hence conditions prior to debris flow development on lower slopes.

3.2.1.1. Infinite slope stability analysis

End and side effects are ignored in the analysis which is expressed in terms of an infinite layer of sediment on an inclined plane, the failure surface being parallel to the slope surface (Fig.3.3.). Carson and Kirkby (1972) justified this approach by citing the great length of the slope relative to sediment depth. The Factor of Safety (F) is used as an indication of slope stability. Carson (1976) cited Skempton and Delory (1957) who demonstrated the effectiveness of the equation in representing long-term stability on clay-mantled slopes, which are the most cohesive sediments.

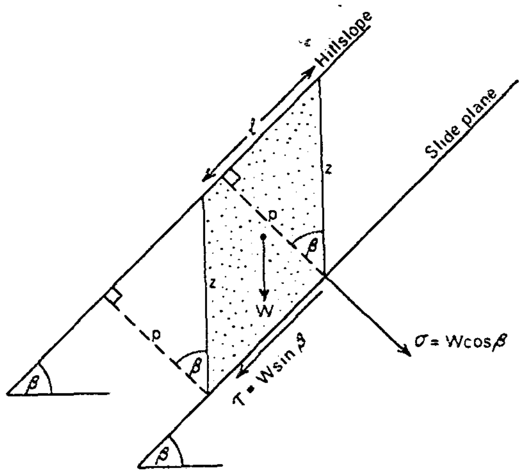

Figure 3.3. Diagram illustrating the components used in the infinite slope stability equation (after Selby, 1982).

The equation for slope stability analysis (p.23) is a simplification of slope failure conditions, omitting the interaction of slopes with various other processes and/or events such as earthquakes, erosion, flooding etc. An F value equal to 1 does not signify real failure, but only "presumed failure" within the context of an adopted model (Chowdhury, 1978). Chowdhury (1978) suggested the following failure probabilities:

F=0.88 - 100% [probability that slope failure will occur]
F=1.00 - 50%
F=1.20 - 10%

Chandler (1986) has considered the relationship between laboratory measured shear strengths and those inferred from slope stability analysis. He concluded that for low plasticity clays (Plasticity Index <20%) the two strengths are in close agreement, while for plastic clays slope processes reduce the "field strength" below that measured in the laboratory. He speculated that "progressive failure" may be responsible for the latter. That is, non-uniform prefailure movements that occur along a potential sliding surface on a slope composed of brittle material.

However, the values obtained for this type of analysis come from samples which represent only a fraction of the material involved and thus only represent a point in space and time on each study slope (Crozier, 1986). The accuracy and appropriateness of the values is thus dependent on homogeneity within slope sediments. Homogeneity is seldom reached in nature. The more detailed the sampling the greater the range of values there is likely to be (Crozier, 1986). Nevertheless, Crozier (1986) suggested that limiting equilibrium analysis' "...representation of strength and shear stress is sound and can be used to explain how many natural and human factors can cause movement."

On scree slopes the translocation of fines may reduce permeability, increase positive porewater pressures and consequently facilitate initiation of shallow debris slides during wet conditions (Crozier, 1986). Shallow instability (e.g. subaerial debris flow) can be achieved on stable bedrock slopes, unlike deep-seated landslides where critical slope height needs to be surpassed (Crozier, 1986). Particle-size, particle sorting and Atteberg limits all play a part in determining material strength. Atteberg limits are particularly useful in determining the characteristics of sediment, as they reflect both particle-size and mineralogy (Crozier, 1986).

Long-term slope stability is assessed by invoking extreme transient conditions, normally the water table at the slope surface. Crozier (1986) suggested that by making these assumptions and assuming porewater pressures are hydrostatic and not artesian, slope failure in terms of (F) may be represented by:

$$F=S/T=[c'+z.\cos^2.(\gamma-\gamma_w)\tan\phi']/[\gamma.z.\sin\beta.\cos\beta]=1.0$$

where S=shear strength

T=shear stress

c'=effective cohesion

ϕ'=angle of internal friction

z=depth of slide (vertical)

γ=total unit weight of material

γ_w=unit weight of water

β=sine of slope angle

Regolith landslides are the most common form of slope failure and are almost always related to extremely wet conditions (Crozier, 1986). Variation in soil saturation is thus an important predictive approach to mass movement. Regolith landslides are associated with converging percolines and hence increased relative porewater pressures.

Iverson and Major (1986) feel that the "seepage force vector" provides a better indication of instability than the "...misleading notion..." that high porewater pressures cause slope instability. The threshold angle for slope failure of water-saturated cohesionless soil is taken as equal to the angle of internal friction if no seepage is taking place or equal to approximately half this angle with uniform slope-parallel seepage (Iverson and Major, 1986).

3.2.1.2. The link between Climate and Slope Failure

Young (1978) found that the distribution of landslides near Wollongong, Australia was related to mantles of ancient mass movement debris deposits. The average excess of precipitation over evaporation was the most important climatic variable controlling long-term instability through its affect on soil moisture status. This determined the amount of water available for runoff and seepage. Young (1978) found that landslides may lag heavy rainfall by 3-4 days and that slope failure sometimes occurred at stream heads where contours converged. Young (1978) pointed out that rainfalls associated with mass movement showed no consistent pattern, citing the inclusion of several intense storms of short duration within a long rainfall event. This is typical of a passing front associated with an incursion of subpolar maritime air.

Caine (1980) reviewed published records of rainfall intensities and duration associated with landsliding and debris flow activity.

He suggested that slope instability related to such processes has a limiting threshold represented by the following equation:

$I=14.82D^{-0.39}$ I=mm/hr and D=duration of rainfall (hr)

or

$d=14.82D^{0.61}$ d=rainfall depth (mm)

The relationship between rainfall and slope instability is best defined for rainfall durations between 10 minutes and 10 days. Rainfall influences slope stability indirectly through its effect on pore water conditions within the slope mantle material. The relationship is not dependent on material factors. The equation is based on the *minimum* rainfall input required for slope failure. Local factors may thus raise local thresholds. However the equation facilitates the calculation of minimum recurrence intervals when linked to local rainfall intensity-duration-frequency data.

No probability value can be assigned to Caines' (1980) data because there is no knowledge of how often the threshold is exceeded without landslides taking place (Crozier, 1986). The geological environment and antecedent conditions are also not taken into account (Crozier, 1986). However in the absence of better similar equations linking rainfall and slope failure Caine's (1980) empirical data will have to suffice as part of the current hillslope model until better information becomes available.

The only similar work has been carried out by Innes (1983).
Innes (1983) calculated a limiting curve for debris flows, plotting total rainfall over a given period of time ("Intensity"=T[mm]) against rainfall duration (D[hrs]). Innes (1983) used the work of Takahashi (1978) to quantify the upper and lower slope threshold angles for debris flows. He was able to define regoliths' susceptible to debris flow activity by two limiting curves:

$S \geq 1.16M + 2.40$ S=Particle size (phi standard deviation)

$S \geq 0.14M + 2.81$ M=Phi mean

The importance of antecedent soil moisture conditions to slope failure led Crozier (1986) to suggest that the Antecedent Precipitation Index (Pa) can be used as an index of antecedent soil water content. Crozier and Eyles (1980) used this as an indicator of slope failure, using the 10 days prior to the landslide. In combination with daily rainfall the index may be used to delimit the approximate threshold at which landslides occur. However, the margin of probability is still wide. The success of these models is dependent on an accurate record of landslide activity, which is not available for the slopes investigated in this study.

3.2.2. Properties of Debris Flow Sediment Transport

At moisture contents greater than the liquid limits of a sediment debris flows behave like a fluid, whereas at moisture contents below the plastic limit a sediment behaves like a brittle solid. Between these two moisture contents the sediment will undergo plastic deformation (Carson and Kirkby, 1972). Most debris flows have at least 2 phases, a hyperconcentrated flow (shear strengths 10-20N/m^2) which approximates Newtonian rheology and a true debris flow phase which displays visco-plastic rheology (shear strengths >20N/m^2)(Selby, pers.comm., 1993). Figure 3.4. is a schematic representation of a typical single surge debris flow.

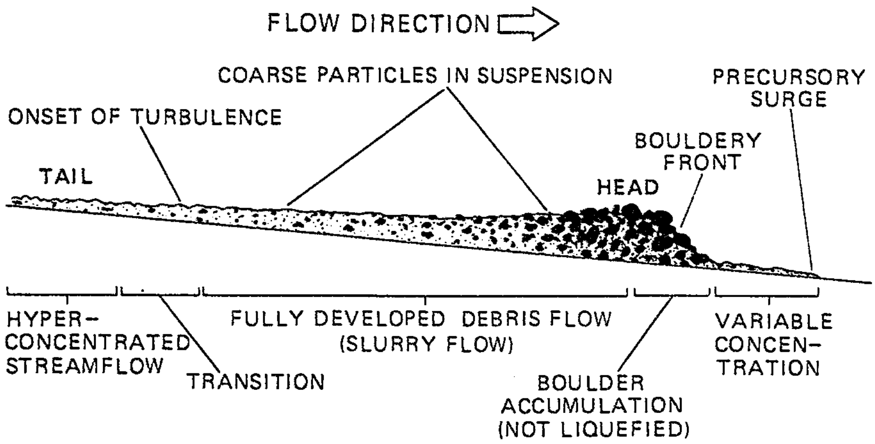

Figure 3.4. Schematic representation of a typical single surge debris flow (after Pierson, 1986).

Sediment flows cease when there is an increase in the resistance to shear or decrease in the applied shear stress. Normally these are a result of:

1. A decrease in bed slope
2. A decrease in thickness of the flow mass
3. A loss of interstitial fluids
4. A combination of the above
   (Lawson, 1982)

Electrochemical surface forces control particle interactions for grain sizes within the fine-silt/clay range. For the larger grains, coarse silt and above, surface forces are small compared to gravitational forces and are thus cohesionless. Frictional and inertial forces control particle interactions. Therefore cohesive forces only dominate flow behaviour when fine-grained particles are most abundant. Plasticity in flows is indicated by the failure of clasts to settle after deposition and by their matrix-supported textures (Lowe, 1976).

The mixture of fine cohesionless sediment and interstitial fluids has two important effects:

1. It maintains the dispersive pressure required for the movement and suspension of the larger clasts by reducing the "immersed weight" of the larger clasts.
2. It promotes high "...flow velocities by increasing the density contrast between the flow and the ambient fluid" (Lowe, 1976) which, in this study, is the density contrast between the air above the slope and the water-saturated sediments.

Rodine and Johnson (1976) concluded that debris flows move "...as a virtually frictionless mass on low slopes" when little interlocking of clasts takes place. Hampton (1979) emphasized that only "...the proportion of clay to water and the total volume percentage of coarse material,..." reflect the competence or mobility of a debris flow deposit. These ratios "...cannot be determined accurately from deposits if dewatering has..." taken place.

Poor sorting is a key factor in the low strength and high mobility of debris flows (89% to 95% of mass by volume can be solids before any significant interlocking will take place -

Pierson, 1980). This and the presence of clay gives debris flow fluids a high density and cohesive strength, allowing boulders to be rafted by the flow.

3.2.3. Sedimentary Character of Subaerial Debris Flow Deposits

Debris flow deposits range in thickness from decimeters to metres. Clast sizes range from boulder to granule within a fine grained matrix of clays and silt present in concentrations as low as 1% by volume. Grading and sorting are poorly developed, apparently with no preferred grain orientations. Although debris flows can not cut channels themselves they may sometimes occupy water-cut channels. Erosion will only occur if the debris flow is turbulent. Hyperconcentrated flow units are poorly sorted and weakly stratified whereas debris flow units are coarse grained with a finer matrix (Selby, pers.comm. 1993). Debris flow units may preserve evidence of lobes and levees with the largest clasts on top and at the face of lobes (Selby, pers.comm. 1993).

Turbulent flows with low sediment concentrations produce "...graded bedding, antidunes, parallel lamination, and ripple cross-bedding" (Postma, 1986). Deposits which result from highly concentrated flows (laminar flow dominant during deposition), are typically (thick) massive and nongraded. Even a small amount of clay (e.g. 5%) can give sediment cohesive strength, although cohesiveness is sometimes significantly reduced during flow as porewater pressures increase and the flow dilates (Postma, 1986). Sharp contacts between layers, straight laminae and inverse grading are the result of high shear rates. At low shear rates laminae are poorly developed and often associated with dish structures.

Scott (1971) investigated the debris flows which occurred near Glendora, California in 1969. He found that the poorly sorted debris flow deposits had strong positively skewed grain size distributions toward the fines. Debris flows, in contrast to the strong bimodal nature of the mudflows, were weakly bimodal or unimodal. Scott (1971) also identified levees and berms on the sides of debris flow channels. Innes (1983) pointed out that levees are characteristic of hillslope as opposed to channel located debris flows. Pierson (1986) found that natural levees form as boulders are continually moved aside by the "slurry" pushing from behind in channelised debris flows.

Large boulders are pushed ahead of the flow (Fig.3.4.) while cobbles and smaller boulders are "overridden" and reincorporated into the flow (Pierson, 1986). The larger the clast in relation to mean particle size, the faster its velocity upwards in the flow and hence movement towards the levees and front of the flow (Innes, 1983). A cohesive debris flow not fully sheared results in a poorly sorted bed with a matrix of fines and clay (Postma, 1986). Large boulders may be rafted within the rigid plug of the flows. These debris flows are typically (diffusely)laminated and sometimes have an inversely graded basal shear zone (Postma, 1986).

Conflicting evidence on the orientation of clasts in debris flows has been obtained (Innes, 1983). Flow lobes show no orientation, but levees do. Clasts dip in the direction of true slope or that of the levee side (*a*-axes). Gloppen and Steel (1981) found that clasts were unoriented or that elongated clasts were oriented subparallel to bed boundaries, with some vertical clast orientation.

The position and orientation of the larger clasts within a debris flow can give an indication of the viscosity of the flow. Tabular gravel fragments in fluid debris flows exhibit a horizontal or imbricated orientation within graded bedding. Larger clasts within the more viscous flows exhibit uniform distribution throughout the thickness of the flow, with clasts in the most viscous flows having a vertical preferred orientation normal to the flow direction (Bull, 1972).

By using the grain size of the coarsest one percentile plotted against median grain size Bull (1972) was able to distinguish between various depositionary environments (Fig.3.5.). However, sedimentary fabrics or structures present within deposits must be interpreted with caution when palaeoflow conditions are inferred as "...observed configurations may have been imported during the final stages of flow (or even post-flow) under conditions atypical of the duration of the flow" (Enos, 1977).

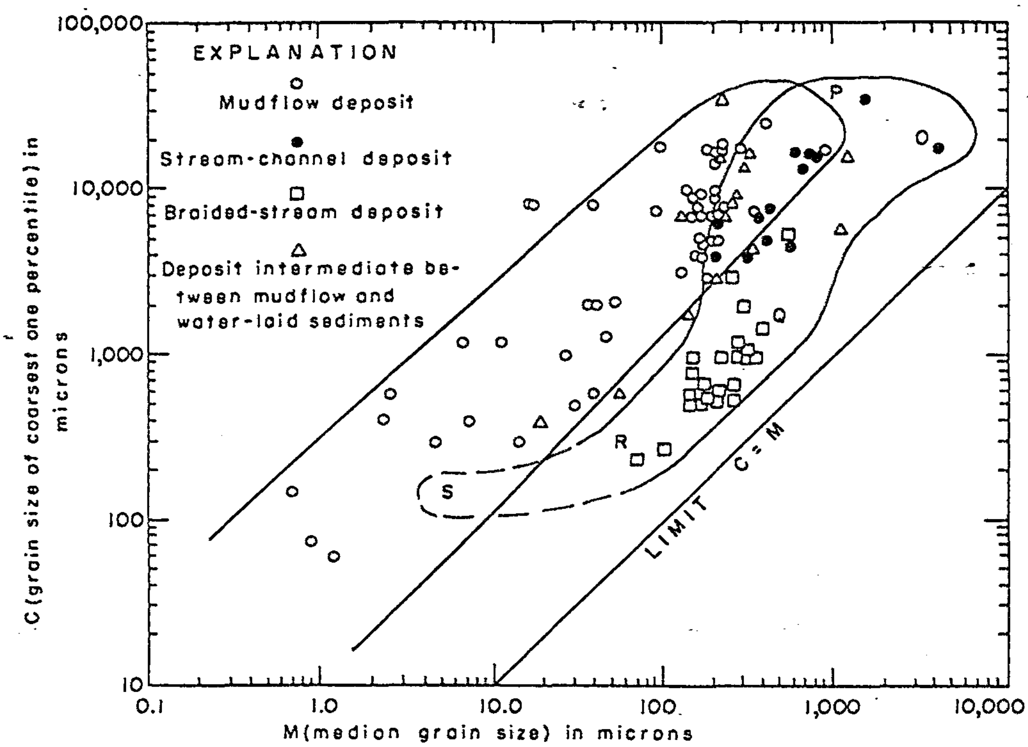

Figure 3.5. The relationship between the coarsest one percentile and median grain size within different environments (after Bull, 1972).

Nemec *et al.* (1980) suggested that if clast-supported conglomerates (diamictons) display a significant positive correlation between bed thickness and particle size, irrespective of grading type, each bed by implication was deposited during a single event. Any "...post-depositional modification of the original bed thickness..." was thought by them to be minimal (Nemec *et al.*, 1980).

Pierson (1980) investigated debris flows at Mt. Thomas, New Zealand. He found that solid material within debris flows ranged in size from boulders to clay and was extremely poorly sorted. Gravel (>2mm) made up 70% by weight of the solids, sand 20%, silt 6% and clay 4% during surges. Gravel comprised only 20%, sand 54%, silt 15% and clay 11% between surges where gravel moved as bedload rather than as suspended load in a watery fluid. Composite curves drawn for data collected from surges indicated an average diameter (D_{50}) of 16mm and a mean diameter of 7mm (i.e. graphic mean-Folk, 1965). The grain size distribution was unimodal with the largest weight fraction falling in the gravel sector and the graph was strongly skewed towards the fines (Sk_1=0.53). This therefore corroborates Scott's (1971) conclusion that debris flow sediments have unimodal grain size distributions. The material was extremely poorly sorted with an inclusive graphic standard deviation of 4.1ϕ (Pierson, 1980).

Pierson (1986) found in active channelised debris flows that their sediment distribution curves indicated bimodality with peaks in the gravel and sand size classes. All samples were poor in fines, with clay particle contents ranging from 0.5 to 2.5 weight %. The total content of fines (silt+clay) ranged "...from 8 to 24 weight % for debris flows and up to 30 weight % for hyperconcentrated flows" (Pierson, 1986). Mean grain size tended to decrease towards the tail of a flow as the sediment concentration decreased.

Gloppen and Steel (1981) described the internal structure of subaerial debris fans from a Devonian alluvial fan/fan delta sequence in Norway. The subaerial debris flows were "...coarse and usually clast-supported...", with a MPS/BTh ratio of three or less (MPS=Maximum Particle Size; BTh=Bed thickness). The deposits were ungraded except for the lowermost few cm's which occasionally displayed inverse grading. Naylor (1980) suggested that inverse grading develops when flows originate from a "ready-mixed" debris source, such as subaerial debris flows developed from soil or talus.

Gloppen and Steel (1981) found that in some cases the coarse debris flow deposit was overlain by an irregular deposit of "...fine, massive conglomerate or laminated granule sandstone...". This thinner overlying bed was generally better sorted, and consisted mostly of grain sizes very poorly represented in the uppermost part of the underlying bed (Gloppen and Steel, 1981). This capping was interpreted as "...the product of waning-stage flow and winnowing of the debris top by waterflow..." (Gloppen and Steel, 1981). Sheetflood deposits in contrast consist of alternating sheets of fine-grained and coarse-grained deposits. These sheetflood deposits consist of sand, granule and very small pebble grain-sizes (Gloppen and Steel, 1981). Figure 3.6. illustrates some typical features of subaerial debris flows.

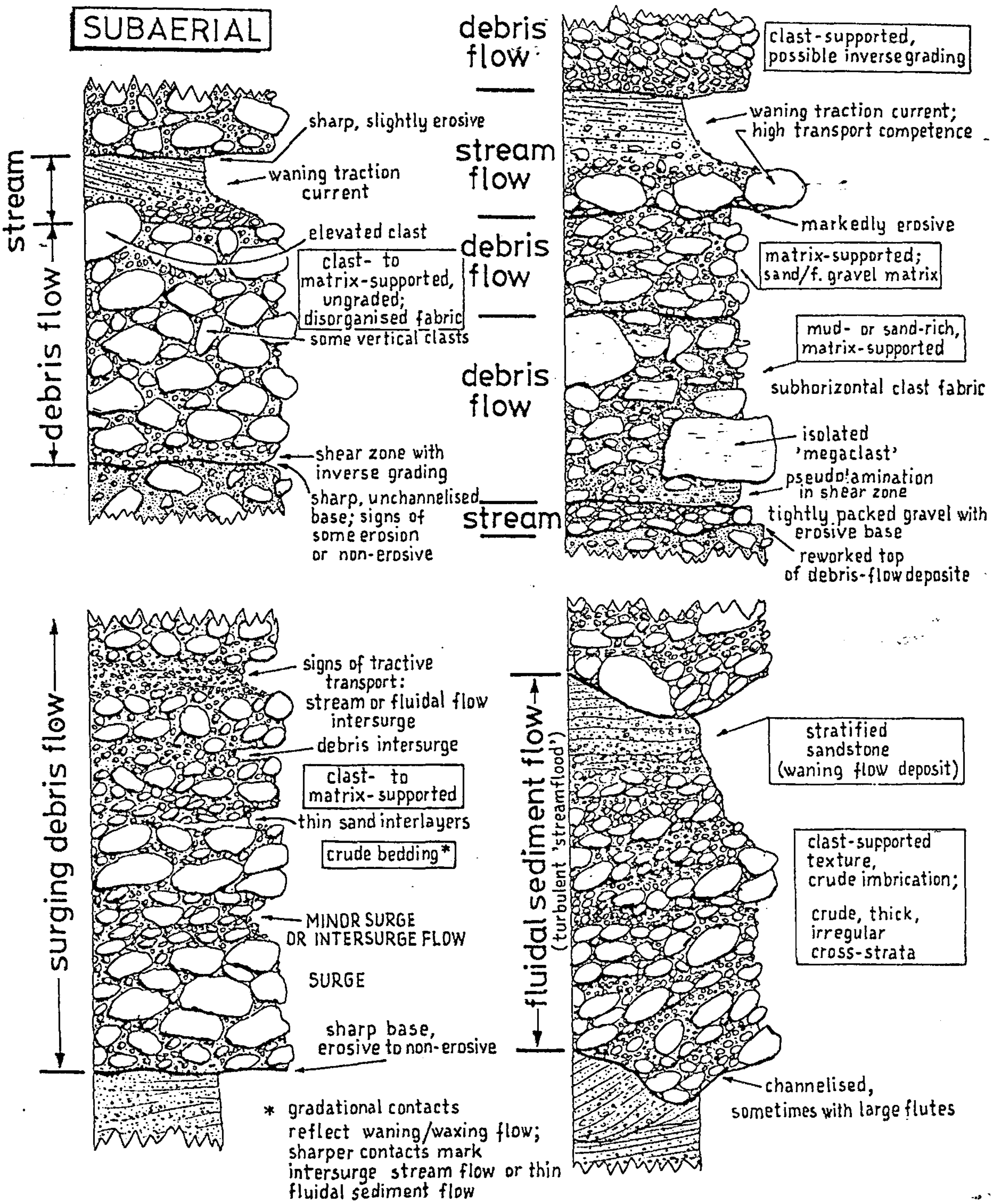

Figure 3.6. Typical features of subaerial debris flow deposits (after Nemec and Steel, 1984).

Nemec and Steel (1984) suggested that the following features are generally characteristic of debris flow deposits:

1. Beds are normally sheetlike, with insignificant or limited basal erosion, although overall geometry is often highly lenticular.
2. Beds are ungraded or well-graded depending on the mode of clast support. The type of grade often changes downslope.
3. While beds show no obvious stratification they can be crudely layered when deposited by surging flows. Successive deposits may show distinct bed boundaries. Nemec and Steel (1984) suggested that "composite units with thick, crude to distinct internal layering or with the presence of thin, discontinuous sandy zones may result from surging flows."
4. Deposits may be polymodal or bimodal, clast-supported or matrix-supported, sometimes containing large "outsize" boulders.
5. Beds often display a significant statistical positive correlation between their thickness and maximum clast size.

Disorganized clast fabric is thought to represent one of three things:

1. Short travel distance
2. Non-sheared (high strength) "plug" flow
3. Weakly sheared (high viscosity) flow

In contrast, "preferred clast orientation, often subhorizontal and parallel-to-flow, may originate from strongly sheared, laminar flow" (Nemec and Steel, 1984). Some significant features of sub-aerial mass-flow deposits are summarized in Fig..
Watery flows may be indicated by an abundance of units variably channelized, with a "...clast-supported texture, crude (subhorizontal to inclined) strata, some imbrication and well-stratified sandstone cappings" (Nemec and Steel, 1984).

Nemec and Steel (1984) developed a conceptual model relating flow competence (D=grain size) to flow thickness. Competence was taken as "...the size of the largest particle that can be completely supported above the bed (i.e. suspended in the debris mixture) over the life of the flow" (Nemec and Steel, 1984). The Maximum Particle Size (MPS, normally the mean of the ten largest clasts, although the size of the largest clast provides a better estimate of flow competence) and thickness of a conglomerate bed (BTh) have often been used to estimate the competence of a flow at the point of measurement (Nemec and steel, 1984). Furthermore, the positive linear correlation between MPS and BTh can be used as evidence for mass flow within a conglomerate as fluvial deposits show no such relationship (Nemec and Steel, 1984). These relationships must be verified statistically. Even where no significant correlation can be found, it does not discount the possibility that the conglomerates may have been deposited by mass flow (Nemec and Steel, 1984). The latter (no correlation) may arise when bed boundaries are wrongly identified (e.g. Amalgamated beds - BTh overestimated) and/or contemporaneous interstratal erosion has taken place (underestimated BTh) (Nemec and Steel, 1984). The above method therefore applies to an assemblage of debris flow deposits and all, rather than a single bed. It thus reveals tendencies rather than exact relationships (Nemec and Steel, 1984). A statement can also be made as to the dominance of cohesive or cohesionless flows. The regression line for cohesionless debris flows intersects the origin, while that for cohesive flows intersects the vertical axis. The fact that the nature of the flow competence may change at the end of its lifetime, also complicates deposit interpretation.

Although fluid mass movements (e.g. mudflows, subaerial debris flows, debris avalanches) and fluvial processes grade into one another, the following field criteria proposed by Crozier (1986) were used to distinguish between the two sets of processes:

1. "Mass movement deposits usually come to rest with convex-upward surfaces or lobes, whereas fluvial deposits have more concave-upward longitudinal profiles" (Crozier, 1986).
2. In fluid mass movements boulders may be carried on the surface or pushed in front of flows (Crozier, 1986).
3. Small cavities, "air pockets", may form between blocks of mass movement debris. Cavities are not normally found in fluvial deposits (Crozier, 1986).
4. Mass movement deposits are poorly sorted whereas woody debris tends to float into segregated lenses in fluvial systems (Crozier, 1986).

3.2.4. Debris Flow Classification Schemes and their Implications for the Interpretation of Stratigraphic Sequences

Numerous classification schemes have been proposed over the last century, many of which have been reviewed by Selby (1982) and Crozier (1986). However, as very little evidence of past debris flow activity may remain on the surface of a hillslope, the classification schemes discussed below are largely based on the nature of the resultant deposit. This review is a summary of the two most useful classification schemes for the purposes of this study, namely those proposed by Shultz (1984) and Postma (1986).

Shultz (1984) suggested that various intermediate flow types occur between two extremes; namely great yield strength, laminar flow, and viscous clast interactions at one end of a continuum and high density turbidity flow at the other end with "...negligible yield strength, turbulent flow, and inertial or collisional clast interactions." Shultz (1984) postulated the following debris flow types assuming constant slope and mineralogical composition. The flow state is dependent on the relative water content and clast concentration. The four flow regimes are based on yield strength, flow turbulence, and intergranular interactions:

I. Plastic debris flow: high-strength, laminar flow, viscous interaction (clasts affect each others' motion by close passage). These flows can easily transport large boulders.

II. Clast-rich debris flow: high-strength, laminar flow, with intergranular collisions. Greater clast concentration than in type I flows. Friction between clasts is more important than cohesion in determining yield strength. The high clast concentrations ensure that these flows fall in the "inertial realm", where inverse grading may be well developed.

III. Pseudoplastic debris flows: low-strength, laminar or turbulent flow, viscous interaction. Matrix strength can only support the smallest clasts. Large clasts are sometimes suspended by turbulence during the high-velocity flow stages, but may be deposited or moved as bedload on flow deceleration.

IV. Pseudoplastic debris flow with inertial bedload: low-strength, laminar or turbulent flow, collisions.

The four main diamictite subfacies which represent these flow characteristics, coded according to the lithofacies scheme of Eyles *et al.* (1983), are (p.73):

1. Dmm: Massive, matrix supported type I flow. Dmm exhibit all the features expected of plastic debris flows.
2. Dmg: Normally graded, matrix supported type III flow. Beds are thinner, finer and better organized than the above (Dmm), with normal grading.
3. Dci: Inversely graded, clast-supported, type II or IV flow.
4. Dcm: Massive, clast-supported, type III flow. Shultz (1984) suggested that the tendency of Dcm's to "...occupy channel-shaped depressions and to grade upward into subfacies Dmg indicates that it comprises the basal sediment or bedload of a relatively mobile mass flow."

A single flow may pass downslope from one flow regime into another (Shultz, 1984). Shultz (1984) found that clast orientations are insufficiently well developed to be of any use within a classification system. The lack of internal stratification and distinct bounding surfaces are characteristic features of debris flows (Shultz, 1984). Mass immobilization rather than selective deposition is the cause of reverse grading (Shultz, 1984).

Postma (1986) developed a mass flow classification scheme based on the character of a flow in its final stage. He based the classification on the following 3 conditions during sedimentation:

1. Flow: laminar or turbulent
2. Flow behaviour: plastic (cohesive) or fluidal (cohesionless)
3. Flow concentration: high or low

The concentration of clasts at the closest possible packing (C_0) was taken as C_0=0.65 (Postma, 1986).

3.2.5. Selected Prominent Studies on Palaeo-debris Flow/Slope Failure

Past research (Chandler, 1976; Hutchinson and Gastelow, 1976; Skempton and Weeks, 1976) has related climate variations to slope stability. Pierson (1980) investigated erosion and deposition by debris flows at Mt Thomas, north Canterbury, New Zealand. He found geological and historical evidence for episodic activity over at least the last 20 000 years. He found that debris flows were the dominant geomorphic process, which could indicate that debris flows may dominate geomorphic development in humid-temperate as well as arid and semi-arid areas (Pierson, 1980).

Shultz (1984) described subaerial debris-flow deposition in the upper Palaeozoic Cutler Formation, western Colorado. He found that the diamictites contained clasts up to 8m in diameter, although some had clasts no larger than pebbles. Maximum clast size and bed thickness were weakly correlated (r=0.66, n=21). The poorly sorted matrix had at least 10% silt and clay. Packing density ranged from 0.1 to 0.77. Long axis orientations were parallel to perpendicular to flow, with clasts dipping in an upflow direction (Shultz, 1984).

Mills *et al.* (1987) ascribed diamicton deposits in colluvial sediments in West Virginia to palaeoclimatic influences. They also cited similar thick colluvial deposits from the upper Ohio valley dating back 8940B.P., 9750B.P., 20 400B.P. and 40 000B.P. years. The flows were often 600m to 900m long, 200m-500m wide and occurred on gradients of 11° to 27°. They suggested that "thick accumulations of old colluvium indicate that rate of production of colluvium has exceeded the rate of removal by fluvial processes during intervals in the late Pleistocene and early Holocene" (Mills *et al.*, 1987). The recurrence intervals of debris slides/flows in the Appalachian mountains were estimated to be thousands of years (Mills et al, 1987).

Mills *et al.* (1987) found that clast roundness in colluvial sediments of the Appalachian Valley and Ridge province of SW Virginia depends mainly on distance of transport rather than on the age of the deposits. Mills (1987) emphasized the importance of hillslope microtopography where material tends to concentrate in hollows perpendicular to a hillslope which is periodically (10^3 to 10^4 years) flushed by intense mass wasting events after intense precipitation e.g. debris flows. Dietrich and Dorn (1984) found that a basal stone layer frequently traces the base of an old landslide along the Pacific coast and mountain system of California, U.S.A. (Dietrich and Dorn, 1984). The stones are transported into the landslide scar and have subsequently buried (Dietrich and Dorn, 1984).

Jacobson *et al.* (1989) investigated the role of catastrophic geomorphic events (e.g. debris flows) in central Appalachian landscape evolution. This is a non-glaciated area of low-seismicity, similar in both respects to the hillslopes investigated in this study. They defined catastrophic geomorphic events as large events which occurred once or twice "per generation" or several times during the Holocene. Stratigraphic evidence showed two scales of temporal variation. One related to Quaternary climate changes and another to "...more-recent, higher-frequency variation due to rare events during the Holocene." Although late-Holocene catastrophic events modified the landscape to a certain degree, landforms related to Quaternary climate changes persisted as the most prominent features (Jacobson *et al.*, 1989). Jacobson *et al.* (1989) assumed a stationary climate over the Holocene and extrapolated historic rainfall data over 1000-10 000 years, in order to obtain the probability distribution of maximum point rainfall. They found three to four debris flows occurred over the past 11 000 years, with a recurrence interval of 3000-4000 years (Jacobson *et al.*, 1989). Relict prehistoric debris flows were attributed to late Pleistocene climates (Jacobson *et al.*, 1989). Most historic catastrophic geomorphic events in the central Appalachians were associated with extreme rainfall events caused by tropical storms or frontal systems and extra-tropical cyclones (Jacobson *et al.*, 1989).

3.2.6. Talus Slopes

True talus slopes do not occur at the two study sites investigated in this thesis. However a short summary of the pertinent characteristics of talus slopes is given below for the sake of completeness.

Kirkby and Statham (1975) found that fines accumulated at the top of a scree slope and coarser fragments lower down. This is the reverse of that expected for sediments moved by fluvial transport. Kirkby and Statham (1975) speculated that the formation of steps at the top of the scree profile below vertical and overhanging cliffs, and randomly elsewhere, were responsible for trapping sediment, which in turn led to small local slip failures. If the rate of input of sediment at the top of the scree slope is greater than a few stones at a time, repeated mass failures occur (Kirkby and Statham, 1975). This produces stratification parallel to the scree surface.

Tanner and Hubert (1991) have investigated the differentiation of Mesozoic talus and debris-flow deposits in Nova Scotia, Canada. They found that the cumulative grain-size distributions of talus breccias are extremely poorly sorted, bimodal, and skewed towards the fines, due to the interstices being filled with a finer-grained matrix. The matrix infiltrated the spaces between the clast-supported framework created by the boulders once they had come to rest. Modern talus is skewed towards the coarser sediments as it lacks the matrix filling. Tanner and Hubert (1991) used the "...orientation of the sub-planar lower surfaces of the discoidal and plate-shaped clasts..." to determine the slope of the palaeosurface (34°), which is consistent with modern rockfall talus gradients between 36°-40°. Platy clasts on modern talus cones are aligned almost parallel to the depositional surface (Tanner and Hubert, 1991).

Tanner and Hubert (1991) interpreted bedded polymodal conglomerates with "...very poor sorting, subhorizontal average fabric of the discoidal clasts, matrix-to-clast-supported fabrics, projection of large clasts above the tops of beds, and flow-parallel alignment of elongate clasts" as debris flow deposits. The conglomerates were described as "...elongate to lenticular deposits of boulders...up to 4m wide and 1m thick measured perpendicular to flow...". They locally truncated other conglomerate beds. Tanner and Hubert (1991) suggested that the debris flows occurred by water-induced mobilization of material in talus at the foot of cliffs after torrential rains in an arid to semi-arid environment.

CHAPTER 4

REGIONAL ENVIRONMENT

"...Owing to the complexity of Quaternary climatic and tectonic histories, topographic and stratigraphic variability can be conveniently explained as a result of climatic and tectonic events." - S.A. Schumm (1979)

Two sites were examined in detail in the area between Alexandria and Grahamstown, eastern Cape. These were Burchleigh and Spring Grove, the location of which is shown in Fig.1.1.

4.1. GEOLOGY

4.1.1. Pre-Quaternary Geology at Study Sites

Burchleigh:

The hillslopes are underlain by quartzites of the late Devonian Witteberg Group and by shales of the early Devonian Bokkeveld Group (Toerien and Hill, 1989). The beds dip steeply to the northeast.

Spring Grove:

The study site is underlain by tillites and sandy shales of the Carboniferous Dwyka Formation at the base of the Karoo Sequence (Toerien and Hill, 1989). The beds dip gently to the southwest.

4.2. REGIONAL GEOMORPHOLOGY

Quaternary climatic and eustatic changes have influenced the development of landforms in southern Africa (Meadows, 1988). These are intimately linked to the erosion cycles experienced by the region after the break up of Gondwanaland in the Mesozoic (Table 4.1.). Partridge and Maud (1987) have considered erosion cycles in analysing topographic sections. They linked key coastal deposits to planation remnants in the interior.

Table 4.1. Summary of the principal geomorphic events in southern Africa since the Mesozoic (after Partridge and Maud, 1987).

Event	Geomorphic manifestation	Offshore sedimentation	Age
Climatic oscillations and glacio-eustatic sea-level changes (most pronounced during middle and late Pleistocene)	Low-level marine benches, coastal dune deposits, river terraces, Kalahari sands.	Accumulation of cones off mouths of major rivers. Widespread erosion elsewhere following development of nearshore current circulations. Renewed sedimentation in deep ocean basins.	Late Pliocene to Holocene
Post-African II cycle of major valley incision, especially in southeastern coastal hinterland.	Incision of coastal gorges, downcutting and formation of higher terraces along interior rivers, formation of Post-African II erosion surface (with planation restricted to the eastern Lowveld region).		
Major uplift (up to 900m in eastern marginal areas).	Asymmetrical uplift of the subcontinent and major westward tilting of previous landsurfaces of interior, with monoclinal warping along southern and eastern coastal margins.		Late Pliocene approximately 2.5Ma
Post-African I cycle of erosion	Development of imperfectly planed Post-African I erosion surface. Major deposition in Kalahari Basin.	Renewed sedimentation giving rise to Uloa Fm. (southeastern coast), upper Alexandria Fm. (southern coast), Bredasdorp Fm. (southern and western coasts) and Elandsfontyn and Varswater Fms. (western coast). Major resurgence of sedimentation in deep ocean basins.	Early mid-Miocene to late Pliocene
Moderate uplift of 150-300m	Slight westward tilting of African surface with limited coastal monoclinal warping. Subsidence of Bushveld Basin.		End of early Miocene approximately 18Ma
African cycle of erosion (polycyclic)	Advanced planation throughout subcontinent. Surface at two levels above and below Great Escarpment. Development of deep-weathered laterite and silcrete profiles. development of Kalahari basin with concomitant onset of sedimentation towards end of Cretaceous.	Widespread epeirogenic sedimentation in several pulses, as exemplified by offshore Alphard Fm., Mzinene and St.Lucia Fms. of southeastern coast and lower Alexandria Fm. of southern coast. General slowing of shelf sedimentation from end-Cretaceous, culminating in major Oliocene unconformity.	Late Jurassic/early Cretaceous to end of early Miocene
Break up of Gondwanaland through rift faulting.	Initiation of Great Escarpment owing to high absolute elevation of southern African portion of Gondwanaland. Deposition of Enon Conglomerate Fm.	Rapid, localised taphrogenic sedimentation producing *inter alia* Uitenhage Group of southern coast.	Late Jurassic/early Cretaceous

Environmental changes and their geomorphological responses in southern Africa have occurred over a wide range of spatio-temporal scales (Meadows, 1988). This section is a short review of landform evolution in relation to tectonic and climatic changes along the southeast African seaboard, with emphasis on the eastern Cape. Table. summarizes the principal geomorphic events in southern Africa since the Mesozoic (Partridge and Maud, 1987).

Pliocene uplift of 600m to 900m occurred along the Ciskei-Swaziland axis of the southeastern coastal hinterland, in contrast to 200m in the southern and 100m in the western sections of the subcontinent (Partridge and Maud, 1987). The concentration of movement in the east accentuated the westward tilt imparted to the subcontinent by the earlier Miocene uplift (Table 4.1.). This caused significant rejuvenation along major inland drainage lines (Partridge and Maud, 1987).

The consequent "Post-African II cycle" is expressed chiefly as "...major dissection of the coastal hinterland" (Partridge and Maud, 1987). Apart from limited areas such as those inland of the Algoa basin, little planation was achieved (Partridge and Maud, 1987). Partridge and Maud (1987) have suggested that except for certain areas along the Natal coastline, the coastal deposits related to this cycle of erosion "...are restricted to the Pleistocene infillings of estuaries."

One result of the incision during the Post-African II cycle of erosion was the development of flights of alluvial terraces along major rivers in response to a variety of structural, geomorphic and climatic factors (Partridge and Maud, 1987). The land of the southeast coastal hinterland underwent monoclinal warping seaward of the Ciskei-Swaziland axis. Movement in the continental interior lagged behind uplift along the Ciskei-Swaziland axis, which caused back-tilting in the headwater zones of sub-escarpment rivers (Partridge and Maud, 1987).

Quantitative data on the effects that this late Pliocene uplift had on erosion rates and sediment supply to the continental shelf are lacking (Dingle *et al.*, 1983; Partridge and Maud, 1987). Evidence for offshore sedimentation rates has been greatly affected by the Agulhas current (Dingle *et al.*, 1983). However, accumulation rates in the cones of major river mouths were high (Partridge and Maud, 1987), with the Limpopo showing increased sedimentation in Pliocene and younger times (Dingle *et al.*, 1983; Partridge and Maud, 1987). This probably indicates increased erosion rates along the southeast coast as a result of river rejuvenation. Glacio-eustatic sea-level changes provided significant pulses of nearshore sediment, resulting in multiple longshore dune cordons (Maud, 1968; Partridge and Maud, 1987). Butzer (1984) suggested that the Cape coastal sector was characterised by either unstable dunes and slopes or by calcification during the early Holocene.

The recognition of isostatic adjustment is not new. Dixey (1942) over fifty years ago correlated the "Pliocene peneplain" of Pondoland with the "Grahamstown peneplains", in conjunction with three major marine-cut terraces of the Pliocene, Pleistocene and Recent time periods respectively. Partridge (1985) suggested that the global sedimentation peak in the upper Pliocene (from 2,5Ma.) owes its origin as much to global tectonics and orogenic activity as to climate change. Partridge (1985) suggested that uplift of 500m could produce a decline in temperature of a magnitude not dissimilar to that experienced during glacial cooling. More recently Hartnady *et al.* (1992) have identified the onset of increased seismic activity in disturbed sediments off the coast of Natal 10 million years ago.

Eastern Cape:

Ruddock (1968) has discussed the Cainozoic sea-levels immediately to the east and west of the Sundays River mouth. He found a total eustatic drop in sea-level of at least 275ft (84m) in the Quaternary period. He related sediments at Birbury, near Bathurst, with a transgression and regression. Davies (1971) studied the Pleistocene shorelines in the southern and eastern Cape, identifying marine terraces at the +60m, 48m, 30m, 18m, 9m and 6m levels.

Marker and Sweeting (1983) have discussed karst development on Tertiary limestones in some detail between the Sundays and Great Fish rivers. They related eroded marine benches to limestone deposits on which the karst developed. Inland the karst area is bounded by remnants of the "silcrete capped Grahamstown planation surface" at an altitude of 300m to 365m (Marker and Sweeting, 1983). The marine benches descend step-like towards the coast at 240m, 200m, 90m, 60m, 30m, 15m, 5m, 2m and -2m (Fig.4.1.). Although caves are not known to occur in the area, dolines, uvalas and poljes have been identified. The planation surface at Burchleigh is associated with the 240 terrace and Spring Grove with the 200m terrace.

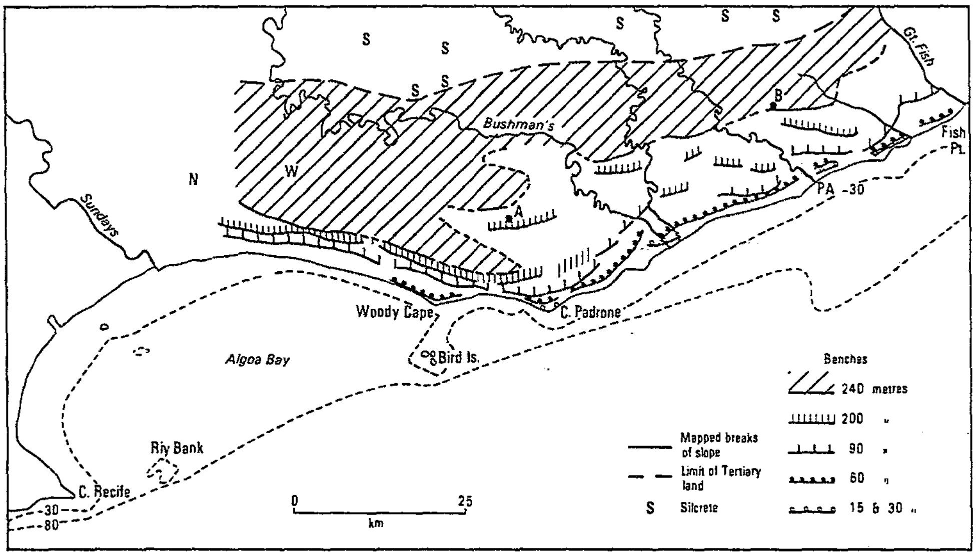

Figure 4.1. Distribution of marine benches in the Alexandria district [A=Alexandria, B=Bathurst, N=Nanaga, W=Wycombe Vale](after Marker and Sweeting, 1983).

Illenberger (1988) has studied the dynamics of the Sundays River and adjacent Alexandria dunefield. He concluded that the Sundays River estuary was a drowned valley during the Last Glacial Maximum (hereafter referred to as LGM). The dunefield developed in conjunction with a number of sand pulses associated with fluctuations in sea level, climatic variation, anthropogenic activity or possibly a combination of all three during the Holocene. Illenberger and Verhagen (1990) associated three sand pulses in the Alexandria dunefield (6500, 3500, 1000 years ago) with climate fluctuations. Each sand pulse is thought to have corresponded with landscape erosion and a fluvial pulse inland.

Both climatic and tectonic factors have therefore played an important role in the development of landforms in the Alexandria district, as Lewis (1995) has recently summarised. Partridge and Maud (1987) have pointed out that tectonism, albeit on a reduced scale has continued to play an active role in the coastal areas of southern Africa throughout the Pleistocene.

4.3. DEBRIS SLOPE DEVELOPMENT IN SOUTHERN AFRICA

While the study of slope form and slope profile development has a distinguished history in southern Africa, with notable contributions by Fair (1947; 1948a; 1948b), King (1953; 1957), Robinson (1966) and lately Russel (1976), Moon (1977; 1984; 1986; 1988; 1990), Botha *et al.* (1990) and de Villiers (1990), the same cannot be said for the sedimentology of the debris slope deposits mantling them. The research work of Botha *et al.* (1990) in northern Natal is a notable exception. Very few studies were carried out on the deposits themselves, most were concerned with other aspects, such as their archaeology (Davies, 1949; 1951), slope evolution (King and Fair, 1944) or mammalian fossil content (Butzer, 1974). This is surprising, since colluvial and hence debris slope deposits in southeastern Africa, "...represent one of the few widespread terrestrial sediments in the region which date from the Late Pleistocene" (Watson *et al.*, 1984).

In Natal King and Fair (1944) suggested that colluvial slopes were the result of sheetflow conditions which originated during torrential downpours. Fair (1947) came to the same basic conclusion, that convectional downpours of high intensity (up to 4 inches per hour) are accompanied by rapid runoff.

Verhoef (1969) observed the widespread occurrence of colluvial deposits in various parts of South Africa, "...often of imposing thickness, reclining on lower slopes of hills and valley-sides and on pediment surfaces." Verhoef (1969) ascribed the deposits to more humid (pluvial) conditions during a prolonged period with saturated soil conditions. He suggested that the mode of downslope transport "sheet-flooding" and "sheet-sludging" (subaerial debris flows?), was possibly aided by "frost action" and diurnal/longer periods of frost.

Botha and Linstrom (1975) found extensive, but isolated hillslope deposits of Quaternary age throughout northwestern Natal and the northeastern Orange Free State, while le Roux and Roos (1982) investigated surface wash on a low-angled slope near Bloemfontein.

Reynardt (1978) has investigated the relationship between colluvial deposits and climate fluctuations near Pretoria. He suggested that they were the result of variations of energy conditions during the Quaternary. He speculated that the 3 colluvial layers he was able to distinguish represented 3 periods when conditions were colder and wetter than at present. Reynhardt (1991) investigated similar deposits near the Hartbeespoort Dam, Transvaal.

Price Williams *et al.* (1982) investigated the archaeology and sedimentology of colluvial deposits in Swaziland and the adjacent regions of Natal, South Africa. They suggested the presence of palaeosols indicated periods of colluvial stabilization and pedogenesis when conditions were moister than present.

Watson *et al.* (1984) believed that the colluvial sediments they investigated of the Late Pleistocene Hypothermal, were deposited under semi-arid conditions, citing the presence of nodular calcrete horizons as evidence. The aridity in southeastern Africa during the LGM they thought to have induced massive colluviation. However, they did identify some ferricrete horizons that they felt attested to periods of pedogenesis under less arid conditions. They suggested that the process of colluvial deposition and re-excavation had been cyclic; some colluvia exhibiting cut-and-fill characteristics. Deposition was a "slow, gentle process" as indicated by the undisturbed nature of prehistoric sites and the micro-textural examination of quartz grains which showed little if any physical or chemical damage (Watson *et al.*, 1984).

Watson *et al.* (1984) found that although differences existed in the colluvium produced from one bedrock type to another, the general characteristics were the same in terms of grain size distribution, particle history and stratigraphy. The mean grain sizes for all sites were 3% gravel, 52% sand, 19% silt, and 26% clay. The lithology was *not* a primary control on colluvia formation. They concluded that the "...broad uniformity in the nature of the sediments over such a large area is a direct result of the similarity in the depositional processes which are themselves the product of very specific environments. A noticeable feature of...the colluvial sites [is]...their similarity in present-day rainfall." This ranged from 600-750mm per annum, most of which falls in the six summer months.

Le Roux and Vrahmis (1987) found a low correlation between slope gradient and the size of rock fragments on debris slopes in the southern Orange Free State. They suggested that large boulders which roll on to the low-angled slopes weakened the relationship. They suggested that debris slopes in semi-arid environments are subject to "slow" mass wasting processes, while debris slopes (screes or talus) in cold environments are formed by "fast" mass wasting processes (le Roux and Vrahmis, 1987).

Dardis and Beckedahl (1988) investigated similar sediments in Swaziland to those examined by Watson *et al.* (1984): a colluvial pediment discordantly overlying a debris fan. The latter consisted of "boulder-sized materials". They (Dardis and Beckedahl, 1988) found that colluvial deposition did not stop during the Late Glacial Period (12 000 - 10 000 years B.P.). They cited the work of Goudie and Bull (1984) who found SEM (Scanning Electron Microscopy) evidence of resedimentation within the upper colluvium layers, to support their assertion that partial resedimentation has taken place within the studied sequences. They (Dardis and Beckedahl, 1988) found that erosion rates averaged 0.07mm/yr from 30 000 to 870 years B.P., 1.0mm/yr between 870-360 years B.P. and up to 20mm/yr from 360 years ago to the present.

Le Roux and Vrahmis (1990) have also investigated the shape of rock particles and their distribution on debris slopes in the semi-arid environment around Bloemfontein (Mean annual rainfall=500mm, 80% October-March; Maximum temperature=January 33°C July 17°C) on Triassic mud-and sandstone with Jurassic intrusions of dolerite. Using the Zingg classification, they found that spherical rock particles increased in number downslope in relation to non-spherical rock particles (very pronounced for spherical particles <500mm in diameter). They ascribed this to:

1. Removal of small non-spherical particles by weathering
2. Production of spherical particles by weathering of non-spherical particles
3. Their unstable equilibrium and relatively small contact area with bedrock and soil inducing faster creep than non-spherical particles.

The last mechanism (3) has also been used to explain the downslope increase in spherical boulders in excess of 500mm in diameter and the momentum these boulders generate once disturbed, which causes them to move a long distance downslope.

Botha *et al.* (1990) investigated the cyclicity of erosion displayed by the Masotcheni Formation, northern Natal, during the Pleistocene and Holocene periods. These sediments were previously linked to pluvial/interpluvial cycles based on the European glacial/interglacial successions (Botha *et al.*, 1990 - based on Gribnitz (1956), van der Merwe (1956), de Villiers (1962). The cyclicity of erosional/depositional/pedogenic events covered a wide area. They identified four such cycles, which entailed "...a swing from relatively stable conditions when pedogenesis was dominant on hilltops and the pediment to unstable conditions during which material stripped from hilltops was deposited on the degraded pediment/footslopes." Botha *et al.* (1990) suggested that a strong relationship exists between *rock type* and *topography*; dependent on the weathering characteristics of Karoo-age sedimentary rocks and post-Karoo intrusives. This feature should not be confused with Watson *et al.*'s (1984) belief that rock type only plays a subordinate role in determining colluvial sediment *composition*. Botha *et al.* (1990) discovered that the pediment thickness *decreases* downslope, the converse of that expected (Dietrich and Dorn, 1984; Embrechts and de Dapper (1987); Botha *et al.*, 1990). Botha *et al.* (1990) found that the bedrock pediment was commonly overlain by "...poorly sorted, angular pebble- and cobble-size gravels...".

As with the sediments studied by Watson *et al.* (1984), cut- and -fill characteristics were common within sediments studied by Botha *et al.* (1990), which featured isolated boulder-sized clasts within fine-textured, laminated sand. The sand grain-size characteristics were very uniform throughout the succession (Botha *et al.*, 1990). Botha *et al.* (1990) believe the bulk of the sedimentary succession accumulated as sheetwash deposits rather than by mass-movement processes, citing the fine-texture of the sediments as evidence. They (Botha *et al.*, 1990) suggested that the "...incised, poorly sorted, diamictic boulder gravels in a clayey matrix exposed..." at the head of dongas represent talus deposits. Botha *et al.* (1992) have also described sediments from a colluvial hollow incised into Ecca Group rocks on the farm Voordrag, upper Mkuze River Valley, northern Natal. The normalized sand distribution is similar to sheetwash deposits from elsewhere in Natal.

In contrast to the limited literature above available on colluvial slopes developed during relatively moderate climates, periglacial slope deposits have been described by a number of workers (e.g. Sparrow, 1964; Harper, 1969; Marker and Whittington, 1971; Hastenrath and Wilkinson, 1972; Nicol, 1973; Fitzpatrick, 1978; Dyer and Marker, 1979; Lewis and Hanvey, 1988; Sanger, 1988; Marker, 1991; Boelhouwers, 1991a, 1991b, 1994; Hanvey and Marker, 1994) from areas at higher altitudes. A selection of these papers is discussed below.

Lewis (1988a) has reviewed the nature of periglacial features in southern Africa, concluding that these phenomena are small-scale, even upland areas are only marginally suited to periglaciation under present climatic conditions. Periglacial features which exist within sediments of lower altitudes are largely an indication of cooler climatic conditions during the Quaternary.

Nicol (1973) tentatively concluded that periglacial processes operated on the south facing slopes of mountains in the Little Caledon Valley, Orange Free State. He identified head deposits and what he believed were protalus ramparts. Lewis (1994) has subsequently argued that the features possess few attributes of protalus ramparts. Further north in the Golden Gate Highlands National Park, Orange Free State, Marker (1989) also noted periglacial features. Le Roux (1990) disputed Markers' (1989) conclusions for the Golden Gate area. He suggested that processes other than periglacial may be used to explain the landforms. In response, Marker (1990) felt that when all the landforms were taken together, they were indicative of periglacial conditions during the LGM.

Dardis and Granger (1986) identified contemporary periglacial phenomena in the Natal Drakensberg, such as stone circles, stone-banked lobes, turf-banked lobes and terracettes. Boelhouwers (1991) and (1994) identified active periglacial features in Natal, concluding that a marginal periglacial environment prevails in the Drakensberg. Grab (1994) identified active thufur in the Mohlesi valley, Lesotho. The thufur are thus reworking existing slope deposits at altitudes +3000m.

Marker (1987) has described "...scree boulder terraces composed of subangular dolerite boulders" in the Amatola mountains, eastern Cape. She attributed these to Pleistocene periglacial conditions and a drop in temperature at the time of around 9°C. The scree are largely restricted to south facing slopes.

Hanvey *et al.*, (1986) concluded that a small debris cone, near Carlisle's Hoek, Rhodes, northeastern Cape was deposited under Quaternary periglacial conditions. The sediments contain a number of debris flow deposits featuring angular to subangular clasts. Lewis and Hanvey (1988) have described the sedimentology of stratified debris slopes near Rhodes, northeastern Cape. The deposits indicate climate fluctuations, some of which may have been periglacial. Periglacial slope wash was used to explain the finer beds, while the coarse angular components were used to invoke emplacement by debris flows following frost shattering. The same authors (Lewis and Hanvey, 1991) identified possible solifluction deposits, formed between 24 000 B.P. and ±12 000 B.P. near Glen Orchy, northeastern Cape. The solifluction deposits are superseded within a sequence of fan deposits by colluvium which accumulated in at least three phases, the last post-dating ±2440 B.P.. Hanvey and Lewis (1991) have also investigated debris slope deposits at Sonskyn, northeastern Cape. They inferred a periglacial origin for the sediments from evidence for freeze-thaw activity, sheetwash and solifluction deposits. Recently Lewis (1994) has identified protalus ramparts at an altitude of 2000m in the eastern Cape Drakensberg. The protalus ramparts date from the Bottlenek stadial (27 000B.P. to 13 000B.P.. This indicates that perennial snowbeds and at least discontinuous permafrost existed in the Drakensberg during that stadial.

The work reviewed above is for the present predominantly summer rainfall regions of the subcontinent. Colluvial slopes in the present winter rainfall region of the southwestern and year-round rainfall regions of the southern Cape are discussed below. They lie further (+5°) south than the debris slopes reviewed earlier and have been exposed to different climatic systems.

Linton (1969) has described in detail a number of features on hillslopes which he believes represent "cryonival phenomena" in southern Africa, including an angular gravel deposit at 445m a.m.s.l. "...a few kilometres southeast of Grahamstown..." on the Port Alfred road. This deposit he identified as "...a characteristic geliflual deposit..." (Linton, 1969). "...A geliflual deposit that must be ascribed to a glacial episode, and most reasonably to the last" (Linton, 1969).

Butzer and Helgren (1972) identified cryoclastic screes while investigating the Late Cenozoic evolution of the Cape coast between Knysna and Cape St.Francis. Slope breccias and other colluvial deposits were widespread "...on the seaward and landward margins of the Coastal Plateau" (Butzer and Helgren, 1972). On coastal slopes, the colluvia were best developed on Bokkeveld shales on gradients from 5° to 35°, sometimes more, projecting below sea-level in places. Shallow dissected talus cones were found on the seaward slopes of the Outeniqua and Tsitsikamma mountains (Butzer and Helgren, 1972). Comparable colluvia rarely form today (Butzer and Helgren, 1972). The formation of these colluvia are not linked to undisturbed modern vegetation (Butzer and Helgren, 1972). The formation of "rubble" was attributed to frost-weathering, even though frost is seldom present today. They (Butzer and Helgren, 1972) therefore believed a lowering of 10°C in winter temperatures would have been needed "...to produce effective frost weathering along the southeastern Cape coast for the type of cryoclastic debris found in these slope breccias...". They suggested that they were "...typical slope deposits transported by sheetwash, creep and other gravitational movements, with or without accessary frost-generated motions." They ascribed these deposits "...to a rupture of slope equilibrium and intensified denudation, presumably related to an opened, incomplete vegetation cover, with marked rainfall periodicity and/or intensity" (Butzer and Helgren, 1972). Butzer (1973) also ascribed "...much or most of the scree and talus of the coastal sector to frost-weathering". Butzer (1973) rejected "...outright Linton's overall interpretation of cryonival and gelifluual activity in the Cape Province." He suggested a 10°C lowering of temperature would be necessary for frost-weathering to occur along the Cape Coast. Butzer (1979) described reworked aeolian sands and talus deposits adjacent to the quartzitic slopes of the Bobbejaansberg at Elands Bay. He concluded that the talus deposits represent periods of frost weathering, large scale mass movement and colluvial deposition. The phases of colluviation were interspersed by aeolian accumulation and pedogenesis. Frost weathering is thought to have led to the formation of the angular debris (Butzer, 1979).

Tankard and Schweitzer (1974) have identified a poorly sorted colluvial deposit containing Middle Stone Age (MSA) artefacts near Die Kelders in the Cape. They suggested that the colluvium was emplaced by "...movement of saturated regolith down a gentle slope."

Deacon *et al.* (1984) investigated angular clasts in Boomplaas Cave, Cango Valley, southern Cape and found no evidence of intense frost shattering phenomena. However, they did find evidence for cold conditions, indicated by cleavage planes and conchoidal fractures on quartz grain surfaces using SEM (Deacon *et al.*, 1984).

The absence of screes and block streams from the Swartberg in the southern Cape and their presence in the mountains of the southwestern Cape indicate different rainfall regimes for the two regions during the LGM (Deacon and Lancaster, 1988). They (Deacon and Lancaster, 1988) suggested that the features were produced by frost wedging of Table Mountain quartzites during the LGM and earlier glacials. The necessity for a coincident cold season and season of precipitation was used to explain the presence/absence of the above slope deposits viz. cold/dry conditions in the Swartberg mountains and cold/wet conditions in the Hottentots Holland and other mountains of the southwestern Cape.

Butzer (1979) also investigated the geomorphology and geo-archaeology at Elandsbaai, western Cape. He suggested that "...several mid-Pleistocene cold intervals led to large-scale frost-weathering of cliff faces and mobilization of 15° block rubbles or *grezes litees* along footslopes" (Butzer, 1979). At Elandsberg he (Butzer, 1979) attributed periodic mobilization of "rubbles" to frost and water-assisted gravity movements or to colluvial transport for sands. Butzer (1979) identified three recurrent environmental anomalies:

1. Frost-weathering and large-scale mass movements or colluvial deposition.
2. Deep pedogenesis
3. Aeolian activation and calcification

Sanger (1988) suggested that Pleistocene glaciation was extant in the mountains of the southwestern Cape; characterised by "...some plateau-, cirque and- valley-glaciers". Sanger (1988) cited periglacial phenomena such as widespread blockfields, glacial pavements and screes as evidence for these glacial features. These features are inactive under present conditions. This built on earlier work by Borchert and Sanger (1981) who claimed to have identified "glacial polishings" and "stria" in the mountains of the southwestern Cape.

The Alexandria area, which presently lies between the summer and winter rainfall regimes, is therefore likely to have been exposed to a wide array of environmental conditions during the Quaternary. This no doubt has led to a change in the suite of geomorphic processes operative at any one time at the study sites investigated for this thesis. The climate and palaeoclimate of southern Africa and the eastern Cape is reviewed below.

4.4. CLIMATE

"The climatic history of southern Africa has been characterised by great change and variability. Seldom have there been prolonged intervals without such change...climates may have changed abruptly over relatively short time intervals...records for climatic changes on the scales of millennia, centuries and decades." - Tyson, 1986.

4.4.1. Climate Classification

The Koppen classification scheme is the most widely used system for geographical purposes (Strahler, 1975). Climate is defined within the system according to fixed values of precipitation and temperature, based on yearly or monthly averages. A more detailed explanation of the scheme appears in Strahler (1975). Strahler (1975) classified the southeastern section of the subcontinent "Cfb", temperate rainy (humid mesothermal) with warm summers. This is a mid-latitude climate controlled by both tropical and polar air masses.

4.4.2. Present Climate in Southern Africa

Apart from the southern Cape the isohyets in southern Africa are aligned N-S, with rainfall decreasing from east to west. The 400mm isohyet divides southern Africa into a wetter eastern half and a drier western half (Tyson, 1986). Rainfall variability increases from east to west. Topography exerts a control on rainfall in many areas. Rainfall is highly seasonal, with the northern and eastern areas experiencing summer rainfall and the southwestern Cape winter rainfall. The southern Cape coast experiences year round rainfall. The western central region exhibits a semi-annual cycle with equinoctial peaks. Winter rainfall is generally initiated by westerly disturbances and the summer rains by convective activity. Evaporation increases from east to west, up to almost 10 times the amount of precipitation received per annum. The greatest rainfall intensities occur in conjunction with convective storms on the highveld or cut-off lows along the coast. A severe storm associated with a cut-off low occurred in Port Elizabeth on 1 September 1968 when rainfall intensities reached 118mm/hr with 552mm falling in 24hrs (Tyson, 1986).

Air temperatures in southern Africa vary a great deal, dependent on altitude, proximity to the coast and the type of associated ocean current. Mean annual temperature ranges from 23,3°C at Goodhouse in the western Orange River valley to 11,5°C at Mokhotlong in the mountains of Lesotho (Tyson, 1986).

No meteorological stations in southern Africa presently experience periglacial conditions (Lewis, 1988b). Although little doubt exists as to their occurrence in the recent past, the identification and interpretation of periglacial phenomena in southern Africa is a contentious issue (Butzer, 1973; Hall, 1991; Hall, 1992). Many researchers have used angular material as proxy evidence for freeze-thaw action (e.g. Butzer and Helgren, 1972), when it could easily form under other conditions, including thermal fatigue or biological weathering (Hall, 1992). Hall (1992) empathising with Butzer (1973), made a plea for greater rigour in the use of cryogenic terminology.

4.4.3. Implications of the Present Climate Model for Palaeoclimate Reconstruction in Southern Africa

Deacon (1983) suggested that glacial/interglacial cycles have left their imprint on colluvial and surficial geological deposits. The most important features of the southern African atmospheric circulation system associated with major changes in climate are summarized below.

Quoting van Zinderen Bakker (1982), Deacon (1983) suggested that climate fluctuations were associated with changes in intensity rather than with shifts of climate zone, cyclonic or winter rains during glacial times being more frequent and penetrating further inland than at present. Citing Morely and Hays (1979), Deacon (1983) suggested that the belt of westerlies could not have been displaced more than 1°N during the LGM (approximately

20 000 B.P.). However, this displacement is in broad disagreement with most other work on the subject. The current ideas and reasons for climate change are discussed below.

During lower temperature conditions, southern Africa, particularly the Cape, would have experienced wetter conditions, with an increase in winter rainfall as perturbations in the temperate westerlies penetrated further north than at present. In contrast, raised temperatures would have enhanced tropical disturbances and increased summer rainfall (Tyson, 1986).

Heaton *et al.* (1986) determined dissolved gas palaeotemperatures for the last 28 000 years using ^{18}O data obtained from groundwater near Uitenhague, South Africa. Mean annual air temperatures at the time of the last glaciation were found to be 5.5°C lower on average than during the Holocene. Similarly, Talma and Vogel (1992) using ^{18}O data from a stalagmite in the Cango Caves, Oudtshoorn, obtained a palaeotemperature record for the last 30 000 years B.P.. They found temperatures were at a minimum 18 500 B.P. to 15 500 years B.P., when they were approximately 5°C to 7°C lower than at present. Essentially constant temperatures have been has been maintained for the last 5000 years, varying between +1°C and -2°C of the present values.

Tyson (1986) quotes Newell *et al.* (1981) who estimated that wind speeds in the southern hemisphere increased 17% during the LGM (25 000 B.P.-15 000 B.P., Butzer, 1984). The "thermal equator" was possibly displaced south of the equator at the time, in contrast to its present annual average position at 6°N (Flohn, 1984; Tyson, 1986).

Palaeo-dune orientations in the Kalahari indicate the equatorward displacement of a strong Continental Subtropical Anticyclone by 2°-3° (Lancaster, 1981, as quoted in Tyson, 1986), and a displacement for the winter rainfall margin by 8° northward (Tyson, 1986). Tyson (1986) suggested that cooling of about 13°C could produce an 8° equatorward shift in the subtropical high pressure systems during the LGM.

During the LGM, storm tracks would have moved further north, precipitation over the winter rainfall region would have increased, and the boundary between the summer and winter rainfall regions would have moved to about 25°S over the western and 30°S over the eastern areas of the subcontinent (Tyson, 1986).

4.4.4. The Affect of Past Oceanographic Conditions on Palaeoclimate

A number of workers have speculated on the palaeo-oceanographic conditions surrounding the subcontinent during the Quaternary, including Be and Duplessy (1976), Hutson (1980), Martin (1981), Butzer (1984), Howard (1985), Tyson (1986), and COHMAP members, (1988). Howard (1985) stated that "...late Pleistocene oceanographic reconstructions place some constraints on terrestrial palaeoclimates, and may provide a framework for the interpretation of the earlier environmental history of southern Africa." Martin (1981) discussed the evolution of the Agulhas Current and its palaeo-ecological implications. He related the Agulhas Current to sea surface temperatures and southern African rainfall receipts.

Be and Duplessy (1976) using oxygen isotope and microfaunal analyses in two ocean cores, one of which (RC 17-69) lies in the southwest Indian Ocean (Location=500km SE of Durban, 31°30'S 32°36'E) identified changes in the meridional position of ocean convergence zones. They showed that the position of the Subtropical Ocean Convergence fluctuated between a northerly limit at 31°S during glacial stages and its current position, the southerly limit 38°S to 40°S. A movement of the Polar Fronts also took place from 48°S-50°S during warm periods to 40°S during colder phases. Be and Duplessy (1976) suggested that "the northward displacement of the Subtropical Convergence to a position off Durban...reflects the general weakness of the Agulhas Current during the glacial stages and parts of interglacial stages...".

Similarly, Hutson (1980) using data from the same core (RC 17-69) concluded that ocean circulation during the LGM in the Agulhas region varied seasonally. The Agulhas current was thus weaker and shallower during cooler phases. The northward displacement of ocean convergence zones was associated with increased meridional temperature gradients, intensification of the westerlies and a northward shift in storm tracks (Tyson, 1986). Others estimated that the Subtropical Ocean Convergence moved 3°-5° north to a position close to but not quite intersecting the present coastline (Howard, 1985; Tyson, 1986), while the Antarctic Ocean Convergence moved 5°-8° north (Tyson, 1986).

Butzer (1984) suggested that the warm Agulhas Current would not have been placed adjacent to the south and east coasts during glacial times when offshore waters were 5°C cooler. A sea temperature drop of 5°C would have caused a 20% to 30% reduction in evaporation. As a result air circulation around the southern Indian Ocean high pressure cell towards southern Africa would become less moisture laden (Martin, 1981). The Benguela Current at the same time was weaker with reduced upwelling and atmospheric stability along the west coast (Butzer, 1984). At 18 000 B.P. enhanced upwelling and lower sea-surface temperatures prevailed off the NW, SW and SE coasts, indicating a strong wind regime over southern Africa (Tyson,1986).

4.4.5. Late Quaternary Climate Fluctuation

This section very briefly summarizes climate fluctuation in southern Africa during the Quaternary, specifically the period following the Last Glacial Maximum (LGM). Partridge *et al.* (1990) have attempted to update the palaeoclimatic models developed by Tyson (1986) and Deacon and Lancaster (1988). They divided the subcontinent into four regions in an effort to diffuse apparent contradictions in precipitation and temperature trends within these models (Fig.4.2.). Their findings are cryptically summarized below:

1. 40-30kyr: Temperatures were low, but warmer than the LGM. All regions were wetter during at least part of this period, except region D.
2. 30-25kyr: Temperatures were low in all regions. The subcontinent was dry, except the Namib and Kalahari which were supposedly uniformly wetter than at present.
3. 25-21kyr: Temperatures were lower than present. Regions A and D were dry, with the Karoo experiencing a modest increase in wetness. The Kalahari was humid but the Namib dry.
4. 21-17kyr: This period coincided with the cold LGM. There was widespread dryness.
5. 17-12kyr: A warming trend, with an increase in change after 16kyr. There was a significant increase in wetness everywhere, the Kalahari remained humid.
6. 12-10kyr: The terminal Pleistocene which experienced continued warming. Regions A and D experienced a degree of humidity similar to today, with the Kalahari

and Namib drier.

7. 10kyr-present: The Holocene. The warming trend continued in the early part of the Holocene in some areas, with a brief cooling period between 8kyr and 9kyr. The warming trend climaxed between 5kyr and 8kyr, the Holocene Maximum. During the early Holocene regions A and D were significantly wetter than at present, with dry conditions in the Kalahari and Namib. During the Holocene Maximum the southern Cape and Karoo were drier than at present, the Kalahari and eastern region of the subcontinent were much wetter. The Namib was dry throughout the interval. The later Holocene experienced fluctuations in precipitation and temperature of small amplitude (e.g. Cango Caves <2°C from current mean for the last 5kyr). However, Region D experienced much wetter conditions between 1kyr and 2kyr. The Namib was dry, except between 3kyr and 4kyr when it was wetter.

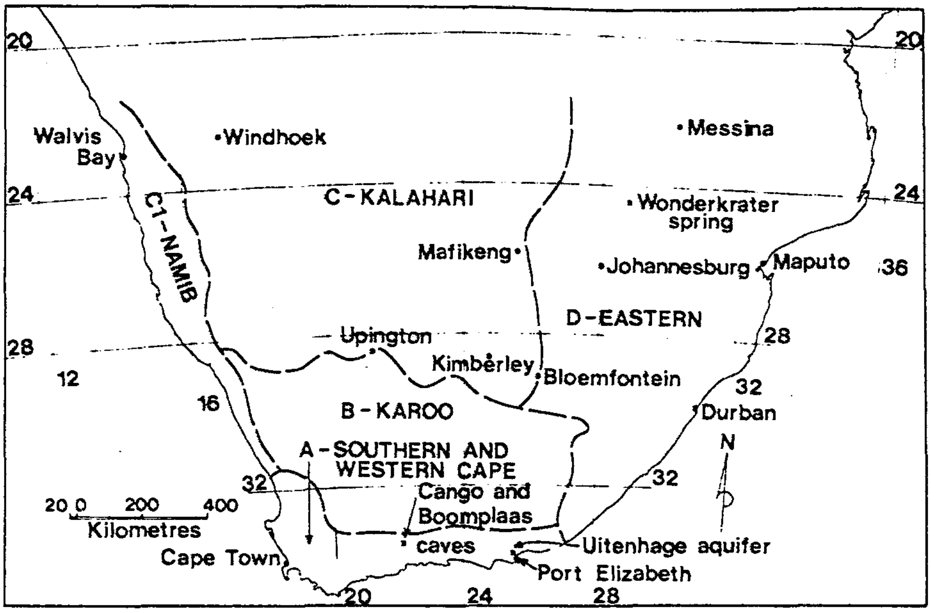

Figure 4.2. The four main climatic zones in southern Africa according to Partridge *et al.* (1990).

Oxygen isotope speleothem data from the Cango Caves indicates that between 4700 B.P. and 4000 B.P. temperatures dropped sharply, with warmer conditions returning around 3500 B.P.. Abrupt cooling set in about 3000 B.P. which reached a minimum at 2000 B.P., associated with the worldwide neoglacial advance of the time (Tyson, 1986). Early evidence appeared to indicate that the local equivalent of the Medieval Warm Epoch was experienced around 1000 B.P. in the southern Cape. Temperatures declined to a minimum during the Little Ice Age (LIA) before increasing again (Tyson, 1986). Dendrochronological data indicates that this cold period (LIA) ended around AD 1570 in the summer rainfall region and around AD 1630 in the southwestern Cape (Tyson, 1986). However, Tyson (1990) citing Talma and Vogel (1988), found no evidence for the Medieval Warm Epoch around AD 1000.

Tyson and Lindesay (1992) have reviewed the climate of the last 2000 years in southern Africa. Cooler conditions over the subcontinent are associated with wetter periods in the winter rainfall region and drier conditions in the summer rainfall regions. They concluded that weather conditions were as follows for the last 2000 years in relation to the present:

1. AD 100 to 200 : Cooler period
2. AD 250 to 600 : Warm period
3. AD 600 to 900 : Variable period of cooling
4. AD 900 to 1300 : Medieval Warm Epoch. Generally warmer period, with highest temperatures between the 10th and 11th centuries.
5. AD 1300 to 1850 : Little Ice Age. Great climatic variability and instability with two major cool phases. The earlier and cooler phase was from 1300 to 1500 and the latter was from 1675 to 1850. The intervening period experienced a period of warming.
6. AD 1850 to present: Post-Little Ice Age. Climatic amelioration exacerbated by the anthropogenically induced greenhouse effect (Tyson and Lindesay, 1992).

Period of the Historical Record:

Weak quasi-periodic oscillations in rainfall occur over the subcontinent, seldom accounting for more than 20% to 30% of the inter-annual variance. Variations are found at 18yr, 10-12yr, 3.5yr and 2.3yr intervals (Tyson, 1986). The 18-year oscillation dominates in the summer rainfall area, the 10-12 year oscillation in the all-season area, and the quasi-biennial oscillation within the area in which the semi-annual rainfall cycle is most evident (Tyson, 1986).

4.5. SEISMIC ACTIVITY

Seismic shaking reduces slope stability by increasing shear stresses, while reducing the resistance of slope material to failure (Crozier, 1986). Earthquake wave propagation has 3 principal effects:

1. The direct mechanical effect of horizontal acceleration, which at high shaking intensities has been known to exceed that of gravity. This provides a temporary increase in shear stress sufficient on occasion to trigger landslides.
2. Seismic loading in saturated material which shifts the weight of particles from granular support onto the porewater. This increases the interstitial porewater pressure, which in turn can lead to liquefaction.
3. The reduction of intergranular bonding imparted by cohesion and internal friction by sudden shock, independent of the degree of saturation. While insufficient to initiate movement, it increases the susceptibility of the slope to future trigger events (Crozier, 1986).

The amount of energy released at the focus of an earthquake is indicated by its "magnitude", while the resultant effects in the surrounding region are estimated in terms of its seismic "intensity".

Unless otherwise stated magnitudes in the following section are based on the "Richter Scale" (M) and intensities on the "Modified Mercalli Scale" (MMS). This discussion is more concerned with intensity than magnitude. Table. lists the characteristics of each level of intensity in the MMS. The effects in each level are not exclusive to that particular intensity. The levels merely indicate particular effects which occur more strongly or more often than in the level below.

Keefer (1984) has been able to discern a number of relationships between slope movement characteristics and seismic parameters. Fourteen types of landslide were identified in the study (Table 4.2.). The most abundant of these were rock falls, disrupted soil slides, and rock slides. Slow earth flows and rock avalanches were among the least abundant. Threshold magnitudes, minimum shaking intensities, and the relationship between magnitude and distance from the earthquake epicentre were used to delineate the relative levels of shaking necessary to trigger landslides in susceptible materials. Amongst these sediments colluvial sand and granular alluvium were prominent. Unsaturated residual or colluvial sand was predominantly associated with disrupted soil slides and soil avalanches, as opposed to rapid soil flows when saturated. Few of the landslides reactivated old landslides, most occurred in material which had not failed before.

Of particular importance to this study were Keefer's (1984) identification of:

1. The smallest earthquake magnitudes at which landslides take place.
2. The minimum shaking intensity at which a particular type of landslide can be expected to take place.
3. The minimum slope angle required for slope failure to take place for each type of landslide.

"Landslides", within this context, encompass all slope mass movements, including "falls" and "flows" but exclude creep and subsidence.

4.5.1. Minimum Magnitude at which Landslides are Initiated

Keefer (1984) found that the smallest earthquake likely to cause a landslide had a magnitude of 4.0. He was also able to estimate the following as minima for the various types of mass movement:

1. M=4.0: rock falls, rock slides, soil falls, and disrupted soil slides.
2. M=4.5: soil slumps and soil block slides.
3. M=5.0: rock slumps, rock block slides, slow earth flows, and subaqueous landslides.

4. M=6.0: rock avalanches.
5. M=6.5: soil avalanches.

[NB. M=5.0 was found to be the minimum magnitude for soil liquefaction (Keefer, 1984).]

Table 4.2. Characteristics of Earthquake-induced landslides (after Keefer, 1984).

Name	Type of movement	Internal disruption*	Water content† D	U	PS	S	Velocity§	Depth**
			LANDSLIDES IN ROCK					
			Disrupted slides and falls					
Rock falls	Bounding, rolling, free fall	High or very high	X	X	X	X	Extremely rapid	Shallow
Rock slides	Translational sliding on basal shear surface	High	X	X	X	X	Rapid to extremely rapid	Shallow
Rock avalanches	Complex, involving sliding and/or flow, as stream of rock fragments	Very high	X	X	X	X	Extremely rapid	Deep
			Coherent slides					
Rock slumps	Sliding on basal shear surface with component of headward rotation	Slight or moderate	?	X	X	X	Slow to rapid	Deep
Rock block slides	Translational sliding on basal shear surface	Slight or moderate	?	X	X	X	Slow to rapid	Deep
			LANDSLIDES IN SOIL					
			Disrupted slides and falls					
Soil falls	Bounding, rolling, free fall	High or very high	X	X	X	X	Extremely rapid	Shallow
Disrupted soil slides	Translational sliding on basal shear surface or zone of weakened, sensitive clay	High	X	X	X	X	Moderate to rapid	Shallow
Soil avalanches	Translational sliding with subsidiary flow	Very high	X	X	X	X	Very rapid to extremely rapid	Shallow
			Coherent slides					
Soil slumps	Sliding on basal shear surface with component of headward rotation	Slight or moderate	?	X	X	X	Slow to rapid	Deep
Soil block slides	Translational sliding on basal shear surface	Slight or moderate	?	?	X	X	Slow to very rapid	Deep
Slow earth flows	Translational sliding on basal shear surface with minor internal flow	Slight			X	X	Very slow to moderate, with very rapid surges	Generally shallow, occasionally deep
			Lateral spreads and flows					
Soil lateral spreads	Translation on basal zone of liquefied gravel, sand, or silt or weakened, sensitive clay	Generally moderate, occasionally slight, occasionally high			X	X	Very rapid	Variable
Rapid soil flows	Flow	Very high	?	?	?	X	Very rapid to extremely rapid	Shallow
Subaqueous landslides	Complex, generally involving lateral spreading, and/or flow; occasionally involving slumping and/or block sliding	Generally high or very high; occasionally moderate or slight			X	X	Generally rapid to extremely rapid; occasionally slow to moderate	Variable

*Internal disruption: "slight" signifies landslide consists of one or a few coherent blocks; "moderate" signifies several coherent blocks; "high" signifies numerous small blocks and individual soil grains and rock fragments; "very high" signifies nearly complete disaggregation into individual soil grains or small rock fragments.
†Water content: D = dry; U = moist but unsaturated; PS = partly saturated; S = saturated.
§Velocity:

extremely slow	0.6 m/yr	very slow	1.5 m/yr	slow	1.5 m/mo	moderate	1.5 m/day	rapid	0.3 m/min	very rapid	3 m/sec	extremely rapid

(velocity terminology from Varnes, 1978).
**Depth: "shallow" signifies thickness generally <3 m; "deep" signifies depth generally >3 m.

The above are not absolute minima as it quite likely that smaller events may trigger landslides if slope failure is imminent (Keefer, 1984). All the earthquake-induced landslides can also be induced by nonseismic causes e.g. intense rainfall. Keefer (1984) mentions two cases reported by Matthews and McTaggert (1978) and Voight (1978) respectively, where rock avalanches occurred in association with earthquakes of M≤3.5. However some uncertainty exists as to whether the earthquakes induced the mass movements or vice versa. Earthquakes must also occur in conjunction with intense rainstorms from time to time, which lower the minimum magnitude for slope failure.

4.5.2. Landslides and Seismic Intensity (Modified Mercalli Scale)

Figure 4.3. clearly indicates the predominant minimum intensity required for slope movement in each slope failure category. Disrupted slides and falls occur on average one intensity level lower than the other types viz. coherent slides and lateral spreads and flows. The lowest intensity reported for any earthquake which causes slope failure was MMI=IV.

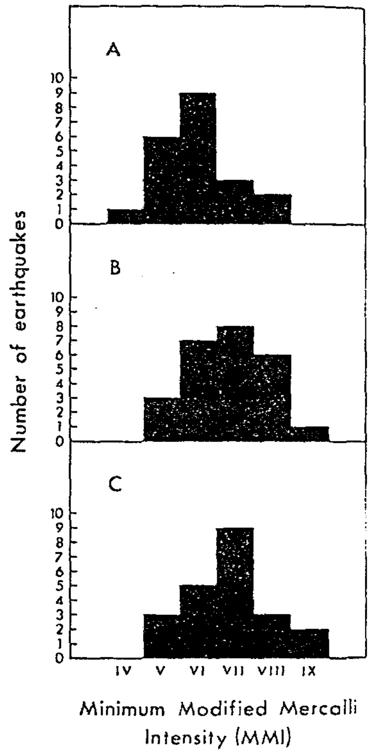

Figure 4.3. The results of Keefers' (1984) survey to determine the minimum Modified Mercalli Scale Intensity for slope failure [A=disrupted slides and falls; B=coherent slides; C=lateral spreads and flows].

4.5.3. Minimum Slope Inclination for Failure

Soil lateral spreads and flows require the lowest angles for slope movement to take place (0.3-10°), whereas rock and soil falls require the highest (40°). Rock avalanches in addition to the minimum slope angle of 25° require a minimum slope height of 150m and need to be actively undercut by fluvial erosion or by active, Holocene, or late Pleistocene glacial erosion (Keefer, 1984).

4.5.4. Seismic or Aseismic Origin of Slope Failure?

If a calculated Factor of Safety indicates that prior to failure a slope had a low probability of failure under saturated soil conditions, the failure can be assumed to have been triggered by an earthquake (Jibson and Keefer, 1989; 1993). This may be tested by calculating the Newark displacement (D_N=cm's).

$\log D_N = 1.460 \log I_a - 6.642 a_c + 1.546$

I_a=Arias Intensity ($m.s^{-1}$)=$0.9T\hat{a}^2$

T=Dohry duration (time needed to build up 90% Arias intensity)

$\log T = 0.432M - 1.83$ (T=seconds, M=unspecified magnitude)

$\hat{a}$=Peak Ground Acceleration (g's)

a_c=critical acceleration (g's)=$(F-1)g.\sin\alpha$ (α=slope angle)

If the D_N>5-10cm, slope failure is likely to have taken place. If the value falls in the 5-10cm range most soils will at least be in a residual state.

Crozier (1992) determined a critical acceleration $[(a_c)_{10}]$ at which a displacement of 10cm would occur, namely:

$\log(A_c)_{10}=0.79\log I_a-1.095$ [$(a_c)_{10}$=g's]

In southern Africa a regression equation (Illgner, in press) defining a line through magnitude and MMS intensity data for the years 1950-1970 is as follows:

$I=-1.15461+1.399118M$ (R^2=0.93)

Seismic intensity (I)(MMS) may be related to horizontal (a_H) and vertical (a_V) ground acceleration by the formulae:

$\log_{10}a_H=0.25I+0.25$

(cm.s^{-2})

$\log_{10}a_V=0.30I-0.54$

(Fernandez and Guzman, 1979b)

However, the values calculated using the above must be interpreted with caution due to the subjectiveness in assigning MMS intensities to tremors and the affects that variable geological conditions may have on earthquakes (Murphy and O'Brien, 1977 cited in Fernandez and Guzman, 1979b). A standard error of 2.9 has been attributed to the formula for horizontal ground acceleration and one of 2.5 for vertical acceleration (Fernandez and Guzman, 1979b).

4.5.5. Seismicity in the Eastern Cape and Southern Africa

Southern Africa and hence the eastern Cape is located in the interior of the large African lithospheric plate. Such intraplate areas are normally subject to moderate earthquakes with shallow foci (Fernandez and Guzman, 1979a). The scatter of these foci in southern Africa are similar to those found in the intraplate areas of the central and eastern United States, and central China (Fernandez and Guzman, 1979a). A comparison of such information indicates that earthquakes of large magnitude can be expected in these low activity areas to be widely scattered in space and time. Earthquakes of magnitude 7.0 can be expected in the Deccan plateau of India with a periodicity of 200 years (Fernandez and Guzman, 1979a). Two such earthquakes have occurred in southern Africa relatively recently, one in the southern Orange Free State in 1912 and the other in the Ceres-Tulbagh district of the Cape in 1969 (Fernandez and Guzman, 1979a) (Fig.4.4). An earthquake of VII caused a block of flats to collapse in Welkom, Orange Free State on 8 December 1976 (Fig.4.5.). Figure 4.6. indicates the location of some of the larger earthquakes which have occurred in southern Africa.

The correlation between seismic foci and geological features in southern Africa remains unclear. Sykes (1978) has ascribed such intraplate activity to the reactivation of pre-existing areas of weakness, such as faults and suture zones (Fernandez and Guzman, 1979b). It has been suggested that the more intense seismic activity in the north and north-eastern part of the region can be attributed to a southern extension of the East African Rift system (Fig.4.6.)(Fernandez and Guzman, 1979b; Hartnady, 1990; Hartnady *et al.*, 1992). The tendency for earthquakes to occur in swarms bears testimony to such a tensional regime. The traditional view presented above has been used in conjunction with isostatic adjustments as a causative mechanism. The isostatic adjustments supposedly occur in response to erosional unloading and offshore deposition (Hartnady, 1990). However, Hartnady (1985) suggested that a hypothetical "Quathlamba hotspot" may have caused tectonic activity. He has subsequently (Hartnady, 1990) revised his ideas to encompass a combination of the above mechanisms, with active rift propagation along a "Quathlamba Seismicity Axis" (QSA) which follows the postulated hotspot trace (Fig.4.7.). Martin (1987) has reviewed the tectonism associated with various hotspot traces in southern Africa, two of which, namely the Shona and Bouvet tracks, follow a very similar path to the QSA.

Little research has been done on seismic activity during the Quaternary in southern Africa, due in large part to the lack of surface evidence (Fernandez and Guzman, 1979b). Korn and Martin (1951) identified a fault scarp 3-4m high running NNW, west of Maltahohe, Namibia, which presented evidence of Pleistocene activity. The fault displaced Tertiary fanglomerate deposits. In northern Botswana, Quaternary neo-tectonics have been identified and used to explain the dynamics of the Okavango-Makgadikgadi Pan complex (Scholz *et al.*, 1976; Fernandez and Guzman, 1979b; Cooke, 1980; Cooke, 1984; Cooke and Verstappen, 1984; Shaw, 1988; McCarthy *et al.*, 1993). At Florisbad in the southern Orange Free State Visser and Joubert (1990 and 1991) identified possible

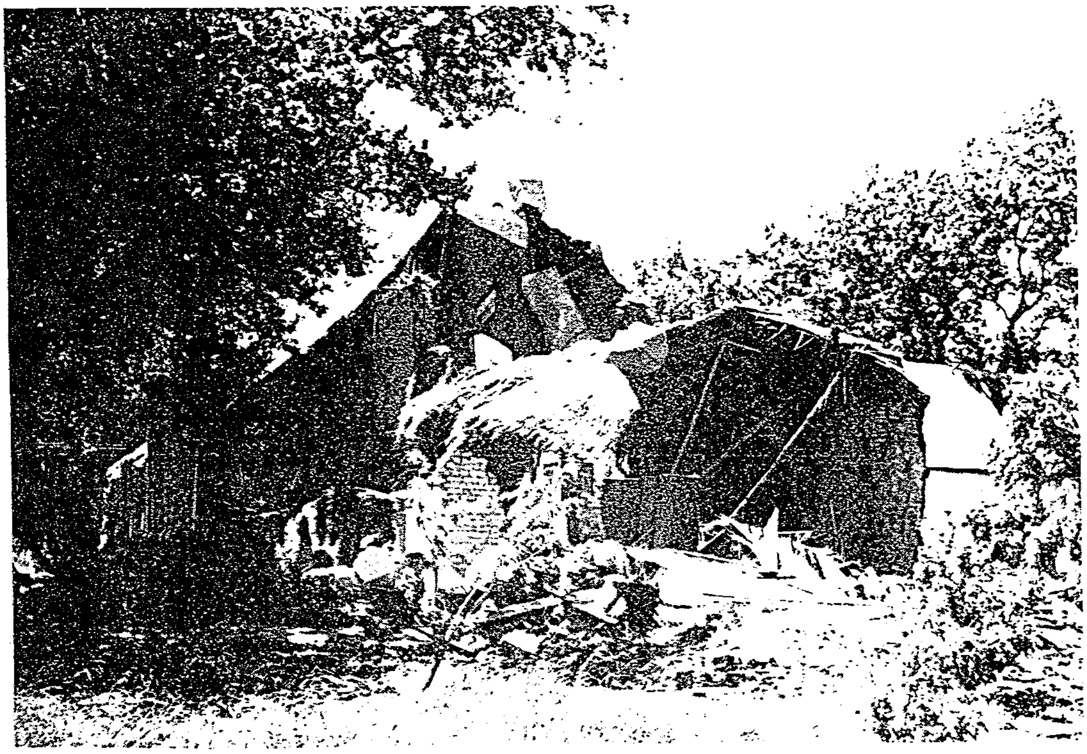

Figure 4.4. Damage to a building caused by the Ceres/Tulbagh earthquake of September 1969 (after Keyser, 1974).

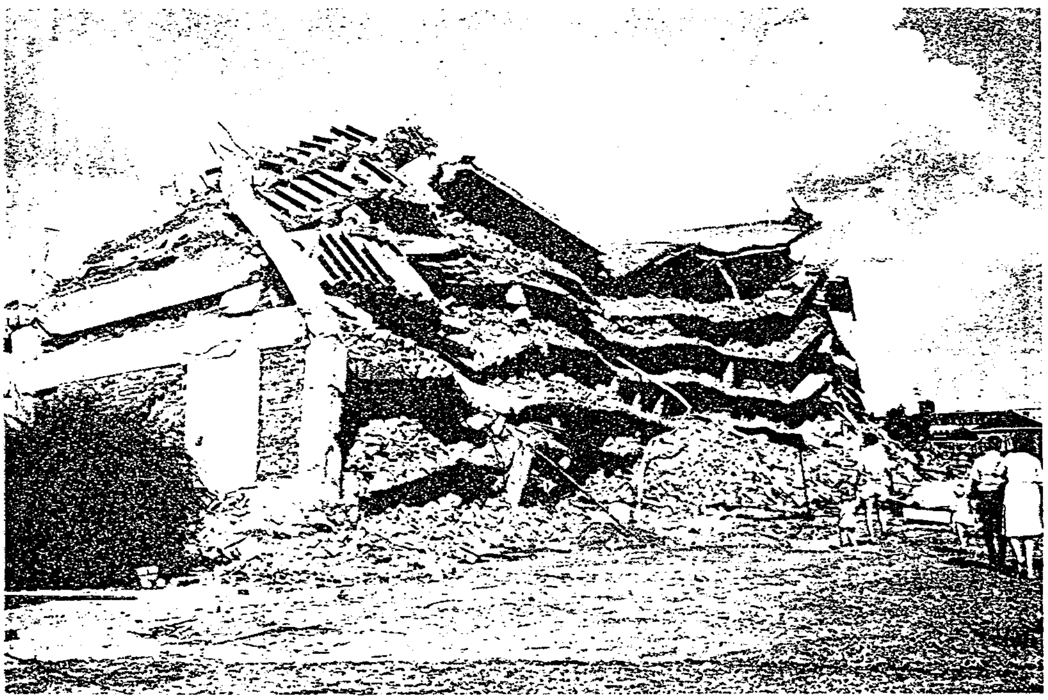

Figure 4.5. "Tempest Hof", the block of flats which collapsed as a result of the earthquake at Welkom during December, 1976 (after Fernandez and Labuschagne, 1979).

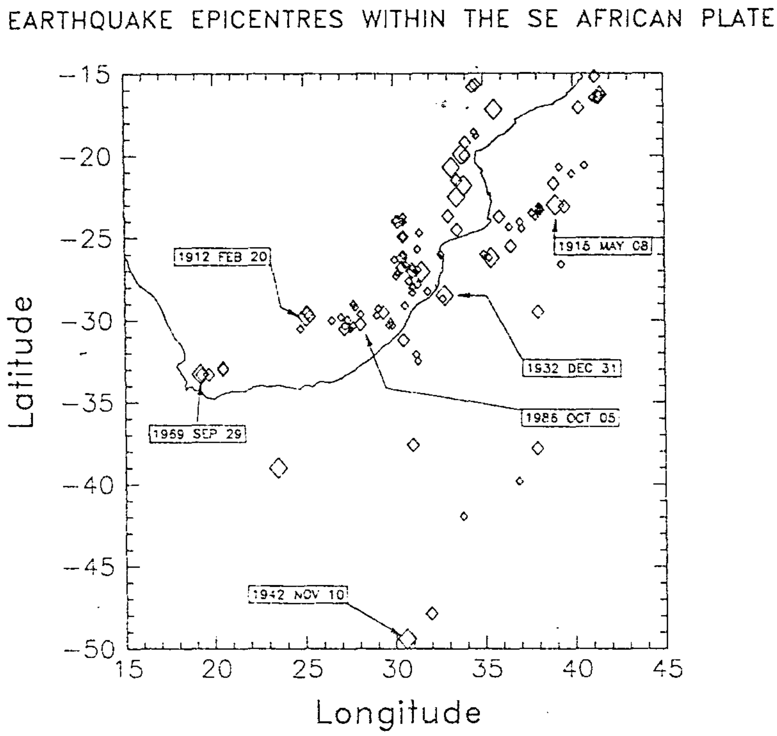

Figure 4.6. The location of some of the larger earthquake epicentres which have been recorded in southern Africa [large open diamond=magnitude≥6, intermediate size diamond=magnitude≥5, small diamonds=magnitude≥4] (after Hartnady, 1990).

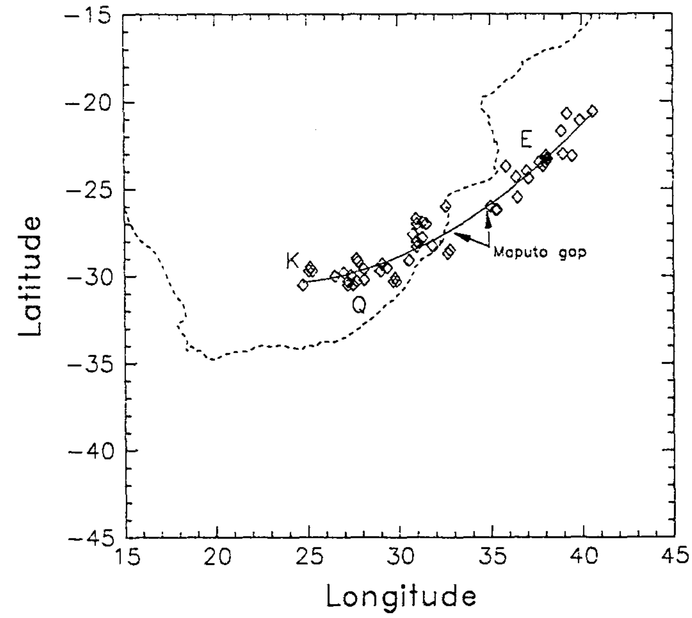

Figure 4.7. The location of the "Quathlamba Seismicity Axis" as proposed by Hartnady (1990).

earthquake-induced sediment liquefaction in thermal spring deposits. They used the periodic migration of the spring head and variation in the water quality to invoke a very incomplete, but lengthy record of probable seismic activity for the past 150 000 years of depositional history for the thermal spring. Hartnady (1990) mentioned geological and geomorphological evidence for late Neogene-Recent surface faulting and deformation along an ESE-WNW axis found in the Cedarville flats area, on the Natal-Transkei-Lesotho border. Krige and Maree (1951) recorded that on 1 November 1942 an earthquake of intensity VI occurred near Port Shepstone, causing "rock falls and small groundslides...". Thompson (1942) suggested that Waterfall Bluff, Transkei, embodied features of a "resurrected fault scarp". The "...upper portion of the Bluff [is]...a remnant of the ancient fault-line (Ergosa fault) scarp lifted above sea-level, probably during the Pleistocene-Recent Period."

Tankard and Schweitzer (1974) attributed the large-scale collapse of a cavern roof at Die Kelders, Cape Province to a catastrophic earthquake approximately 35 000B.P. Hill (1988) found clear evidence for reactivation during the late Quaternary of a 100km long segment of the Kango-Baviaanskloof fault system, with a southerly downthrow not exceeding 5m, east of Oudtshoorn. The reactivated fault truncated an alluvial fan. Avery (1987) studied "rock faults" in caves. He related periods of roof instability to seismicity. He tentatively concluded that seismic activity has had a recurrence interval of 20 000 to 30 000 years over the last 200 000 years. Along the southern Cape coast Anderson and Andreoli (1990) noted the truncation of geomorphological features directly above basement faults and the development of joints in aeolinites directly above them. Most recently, Hartnady *et al.* (1992) have found offshore evidence for late Cainozoic neotectonic deformation of the seafloor off Natal and the Transkei. The tectonism affected the youngest strata in the upper few hundred metres of a >1000m thick seismically profiled sequence. Hartnady *et al.* (1992) have used this as evidence for incipient rifting and the southerly propagation of the east African rift system.

The lack of evidence for Quaternary seismic activity in southern Africa's sedimentary record is probably a reflection of the scarcity of relevant data itself rather than a lack of seismic activity. Nevertheless, while current data may prove inadequate for the above time period, current information can provide some useful indicators of past tectonism and suggests that earthquakes may have been trigger mechanisms for slope movements in the area studied in this thesis, namely the Alexandria district.

Fernandez and Guzman (1979b) have produced two maps and a number of graphs indicating the relationship between Modified Mercalli Scale intensities and a variety of other factors in southern Africa. The maps indicate that earth tremors with an intensity between V and VI are unlikely to occur within the next 100 years, nor one between VI and VII within the next 500 years in the eastern Cape. Nevertheless, if viewed in the context of the Quaternary time period, such activity may prove significant.

To the authors' knowledge only a few earthquakes in southern Africa have been responsible for slope failure. One of these was the Ceres-Tulbagh earthquake in September 1969 (MMS intensity II-III in the eastern Cape), which induced rockfalls and "screeslides" over a wide area in the vicinity of the epicentre (Fig.4.8.). The former

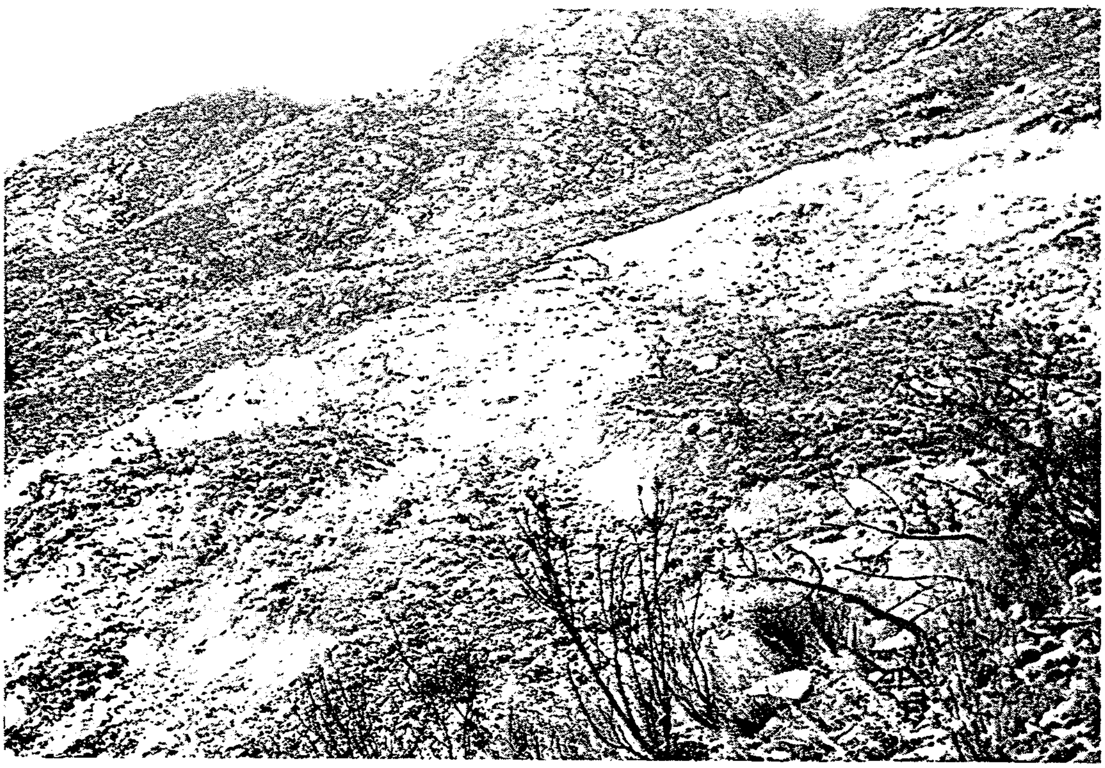

Figure 4.8. A "screeslide" triggered by the Ceres/Tulbagh earthquake of September 1969 (Keyser, 1974).

sparked off bushfires in the mountains (Keyser, 1974). A few rock falls also occurred after the October 5, 1986 Transkeian earthquake (G. Graham, pers.comm., 1993). "Minor secondary failures...triggered at Esikheleni (29° 06'S 32° 02'E - Transkei coast) by an earth tremor which took place in October, 1986" (Beckedahl et al, 1988) (Fig.4.9). These secondary movements occurred within a slope failure complex that Beckedahl *et al.* (1988) speculated was mid- or late Holocene in age. Beckedahl *et al.* (1988) believed that the mass movement complex was not triggered by long term climate fluctuations, but possibly by heavy rainfall events. The gradient of the failed slope was 15°, while the original slope was <20°.

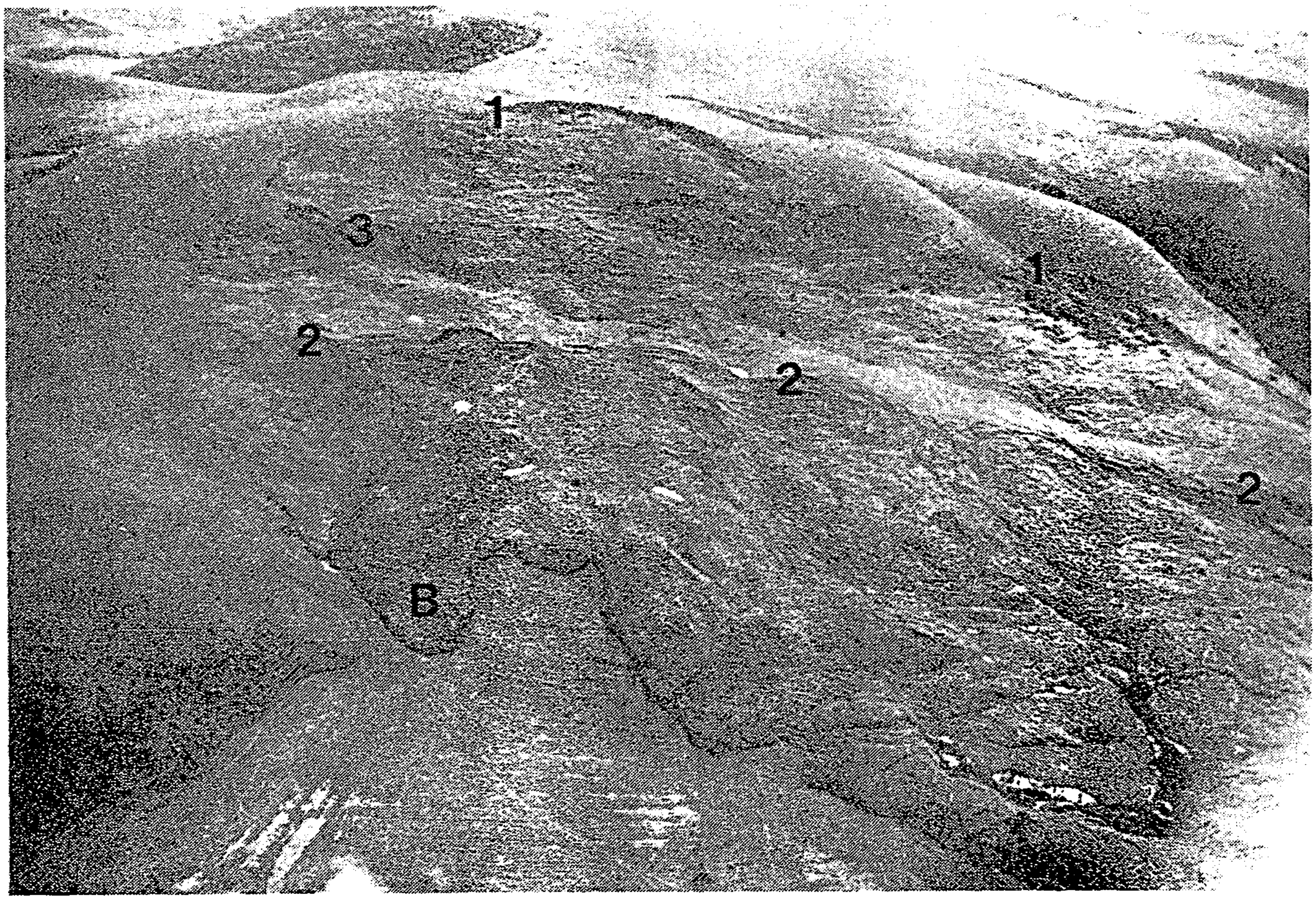

Figure 4.9. The slope failure complex at Esikheleni in the Transkei showing slip scars (1-3) and an earthflow (B) that took place in October, 1986 (after Beckedahl *et al.*, 1988).

CHAPTER 5

MATERIALS AND METHODS

"When you can measure what you are speaking about and express it in numbers, you know something about it; but when you cannot measure it, when you cannot express it in numbers, your knowledge is of a meagre, unsatisfactory kind." - Lord Kelvin (1824-1907)

5.1. CLIMATE

Daily and monthly rainfall, temperature and evaporation data was obtained from the Computing Centre for Water Research (CCWR), Pietermaritzburg, via FTP (File Transfer Protocol) on the internet.

5.2. TOPOGRAPHY

Topography was ascertained by using 1:10 000 orthophoto maps and then checked in the field. These maps were used to draw topographic cross-sections of the hillslopes. The orthophoto maps were also used to draw geomorphic maps, using the scheme proposed by Gardiner and Dackombe (1983). These were checked for inaccuracies in the field.

5.3. SEISMICITY IN THE EASTERN CAPE

A literature search of contemporary media reports (eg. newspapers) of seismic events was used to estimate seismic intensity and to calculate their estimated magnitude. Reports of seismic events by the South African Geological Survey were also utilised. Estimates of the seismic intensity were obtained by using the scheme proposed by Richter (1958).

5.4. STRATIGRAPHY

Lithofacies logs were drawn up for each hillslope exposure based on the scheme developed by Eyles *et al.* (1983). Some of the more important codes, including those used in this study appear in Table 5.1.. Photographs were taken at the study sites to show bed geometry. Access to the cliff sediments was achieved by abseiling off the top of the section.

Table 5.1. Lithofacies codes proposed by Miall (1977) and Eyles *et al.* (1983) relevant to this thesis.

Facies Code	Lithofacies	Interpretation
Dm	diamict, matrix supported	
Dmm	diamict, matrix supported, massive	
Dms	diamict, matrix supported, stratified	
Dc	diamict, clast supported	
Dcm	diamict, clast supported, massive	
Gm	massive or crudely bedded gravel	longitudinal bars, lag deposits, sieve deposits
Gms	massive, matrix supported gravel	debris flow deposits
Sm	sand, massive	

Using terminology developed by Miall (1977) and Eyles *et al.* (1983) Shultz (1984) subdivided the diamictite lithofacies into four subfacies based on fabric and bed geometry (Dmm, Dmg, Dci and Dcm) (see p.36 for interpretation). Using this scheme particles larger than 1cm were termed clasts and those smaller than 1cm, matrix.

5.5. SEDIMENT ANALYSIS

Sediment samples were collected from each of the different lithofacies types and at 10m intervals on a transect line upslope and subjected to grain size analysis. This transect line was parallel to the slope. A 0.5-1kg sediment sample was collected at each sampling site with a trowel and placed in a plastic bag and sealed tightly.

5.5.1. Grain Size Determination

Individual particles were separated from one another for sediment analysis by using a pestle and mortar and later a Calgon solution. The latter was used to prevent flocculation of the clay material. Grinding in the pestle and mortar may have resulted in analyses becoming slightly biased towards the fines. Organic carbon was removed from sediments when neccessary with hydrogen peroxide.

Sediment analyses of grain sizes were based on the techniques suggested by Gale and Hoare (1991). Grain sizes >63µm were separated into phi size class intervals by sieve analysis, whereas those size classes <63µm were separated using a standard pipette analysis technique (Gale and Hoare, 1991).

5.5.2. Interpretation of Grain Size Distribution Curves

The interpretation of grain size distribution curves can be divided into two schools of thought: those who believe that straight lines segments on cumulative frequency curves represent three sub-populations (Fig.5.1.) (Visher, 1969; Middleton, 1976; Bridge, 1981; Viard and Breyer, 1981) and those who believe that they are an artefact of the statistical process (Leroy, 1981; Gale and Hoare, 1991).

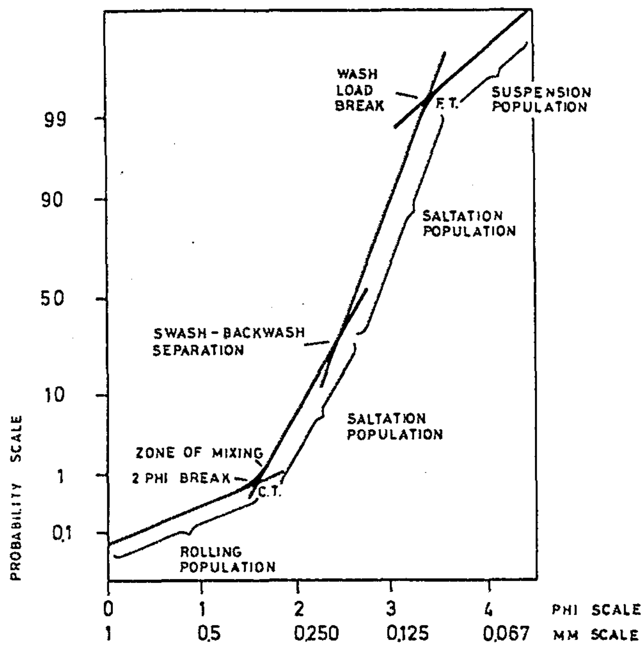

Figure 5.1. An example of a grain size distribution curve for beach sediments plotted on probability scale. The straight-line segments are taken to represent subpopulations associated with a particular mode of transport (indicated on the diagram)(after Reineck and Singh, 1980).

Landim and Frakes (1968) suggested that mean grain size was a record "...of the average expenditure of energy by the depositing agent..." and the available source material (Landim and Frakes, 1968). They suggested that the kurtosis of a curve represents the effectiveness of the velocity of the transporting agent to sort a particular size fraction (Landim and Frakes, 1968). Visher (1969) related grain size distributions in sandy sediments to depositional processes. He identified 3 subpopulations within a log-normal grain size distribution. Each subpopulation related "...to a different mode of sediment transport and deposition." These three modes of transport were suspension, saltation and surface creep/rolling. He believed that the provenance, sedimentary process, and sediment dynamics associated with a sand body cause the systematic variation of "the number, amount, size-range, mixing, and sorting of..." the subpopulations (Visher, 1969).

Middleton (1976) assumed that the "break" between fitted straight line segments in cumulative frequency graphs represented the boundary between truncated normal sub-population distributions of grain sizes. Further unpublished research by Middleton appears to confirm this view (Middleton, pers.comm. 1994). The "breaks" related to:

1. Source (e.g. absence of large grains due to supply rather than hydraulic factors)
2. Mechanical breakage
3. Hydraulic sorting (Middleton, 1976)

Middleton (1976) focused his attention on the position of the "break" between the two coarsest subpopulations. The "break" between the two coarsest fractions he presumed to correspond to the transition between traction and intermittent suspension load (Middleton, 1976). Middleton (1976) suggested that the position of the "break" can be used to estimate the shear velocity of the palaeoflow which deposited the sediment. He believed that the settling velocity of grains at the "break" is nearly equal to the shear velocity of the "dominant" flows moving the bed material. The Shields criterion may be used to check this data (Middleton, 1976).

The fines are more difficult to interpret for they depend more on the supply than on the hydraulics of a river, while the tail of the size distribution is particularly susceptible to diagenetic alteration (Middleton, 1976). Middleton (1976) therefore paid little attention to fine grained material.

For cohesionless sediments, suspension within a flow takes place when:

$w/u^* \leq 1$ (or <0.8 in the case of sorted sediment)

w=settling velocity

u^*=shear velocity

(Middleton, 1975)

When $w/u^*=1$ very little if any sediment is in suspension. Middleton (1976) found this ($w/u^* \leq 1$) to be an "...adequate criterion for separating the traction and intermittent suspension populations...". Middleton (1976) suggested that "the presence of suspended finer grained sediment and large concentrations of wash load may significantly alter the criterion when it is applied to the coarsest fraction of an imperfectly sorted sediment." Likewise, bedforms such as ripples may alter the criterion. Furthermore, the use of whole phi-intervals means it is impossible to determine the precise shape of the cumulative curve and hence there is a large degree of error in locating the position of the "break" between the two coarse components of the distribution (Middleton, 1976). In addition, beds of flow-sediments probably reflect average conditions over a period of time rather than individual flows.

Middleton (1976) suggested that the error in locating the position of the "break" and the uncertainty in converting the size to settling velocity (i.e. unknown water temperature and clay concentration) "...means that the shear velocity can only be predicted within about 2cm/sec". He concluded that "...interpretation must be based on the analysis of many samples from the same sand body, and on averaged rather than on individual size distributions", although the latter may itself introduce uncertainties to an interpretation.

Moore and Burch (1986) pointed out that the strong "primary mode" or "framework population" of most sediments represents the saltating particle population. They noted that "it is this particle size range that controls bed roughness and is therefore probably the size that is important in sediment transport."

Viard and Breyer (1979) investigated the hydraulic interpretation of grain size cumulative curves using log-probability plots. They suggested that the break between the saltating and traction loads could generally be associated with intermediate shear velocities if the subpopulations are assumed to be truncated normal distributions. However, if the subpopulations are believed to be overlapping distributions the "break" is equated with estimated maximum shear velocities. Viard and Beyer (1979) argued that the settling velocity of grains at the "break" between the two line segments that represent the traction and saltating populations is approximately equal to the "dominant" discharge of flow at that point "...if the shear velocity associated with the 'dominant' discharge corresponds to the settling velocity of the grains at the break..."

The shear velocity is represented by:

$u^*=\sqrt{gRs}$ where u^*=average shear velocity
g=gravitational acceleration
R=hydraulic radius (depth)
s=gradient of flow (slope)

"Dominant" discharge was used by Viard and Beyer (1979) as they felt it represented the "average" hydraulic condition at the break. Viard and Beyer (1979) stated that traction load decreases downstream while the "amount of material" saltating increases. However this fining in the direction of transport "...occurs without a major shift in the position of the break between the two subpopulations." This apparent anomaly has been ascribed to the different rates of transport for the two subpopulations (Middleton, 1976; Viard and Beyer, 1979). The sorting in the saltating population increases slightly downstream while the subpopulation in suspension shows no trend.

Grace *et al.* (1978) were sceptical about the significance of dividing grain cumulative frequency distributions into traction, saltation and suspension populations. They examined the size frequency distributions within sand laminae and concluded that these types of curves are highly composite. They found that because "...a uniform distribution or 2 overlapping normal distributions will produce the same pattern..." it is "...difficult to interpret such data in terms of traction, saltation, and suspension processes" (Grace *et al.*, 1978).

Leroy (1981) has pointed out that straight-line segments on cumulative curves are not sufficient evidence for the presence of subpopulations in a grain-size distribution. He showed how a Gaussian distribution plotted as a cumulative percentage using various percentage scales can as a result of mathematical manipulation "develop" "straight-line segments" and hence "subpopulations" as artefacts. Thus dissecting a curve into parts does not necessarily mean that they have any physical reality. Leroy (1981) therefore preferred the use of bivariate plots of such statistical features of grain size distributions as skewness, kurtosis, standard deviation, mean etc to develop an understanding of a sand-body. Nevertheless it is probably inappropriate to discount the usefulness of log-probability cumulative frequency plots as Visher (1969) has pointed out that "an analysis of 1500 samples has shown that these parameters (3 line segment truncation points) vary in a predictable and systematic manner, and that they have significance in terms of transport and deposition."

A reinterpretation of the empirical observations made in the study by Visher (1969) and others is probably part of the solution to the dilemma posed above. Instead of viewing grain size distributions as mixtures of log-normal subpopulations it is probably better to view them in terms of McLaren's (1981) concept of changes in relative skewness of grain size distributions and the mechanisms of grain transport.

McLaren (1981) argued that a single grain-size distribution cannot be used to identify a depositional environment. One sediment sample needs to be compared to another to determine the sedimentary process and transport direction. He found that the "mean grain size, sorting, and skewness of a sedimentary deposit are dependent on the sediment grain size distribution of its source and the sedimentary processes:

1. Winnowing (erosion)
2. Selective deposition of the grain size distribution in transport, and
3. Total deposition of the sediment in transport"

McLaren (1981) identified 3 mechanisms by which a deposit is derived:

Case I. Transported sediment is deposited completely after erosion of source material. The resultant deposit is finer grained, better sorted, and more negatively skewed than the source material.

Case II. The lag left after erosion; that is coarser, better sorted and more positively skewed than the original material.

Case III. The sediment in transport undergoes selective deposition, with the resultant deposits being finer (Type A) or coarser (Type B) than the source material. The sorting for both types will be better and the skewness more positive. This scenario has been associated with waning energy conditions in the transporting process. Coarser sediments have a greater probability that they will be deposited as the energy decreases. The entrapment and shielding of fines in the process of deposition is not thought to affect the relative changes in grain size distribution. The mean grain size will be finer than the source sediment if the energy level of the eroding process is not great enough to transport grains coarser than the mean size at the source. Similarly the mean of the deposited sediment may be coarser if the eroded sediment is coarser than the mean grain size of the original material.

Exceptions to the above cases can occur when:

1. There is more than one sediment source.
2. There is cohesion of fines (e.g. clays)
3. Fines flocculate out of suspension during deposition.
4. Sediment size distribution changes during transport as a result of mechanical or chemical breakdown to cause anomalous trends in Cases I and III.

Therefore, according to McLaren (1981) the skewness displayed by an individual grain-size distribution is a natural feature of a deposit and does not require for its explanation "mixing of modes" or "multiple sediment sources" to justify its existence. From a colluvial point of view an analysis of distribution curves such as the above could help to identify patterns of erosion and deposition and possibly to indicate transport processes.

McLaren and Bowles (1985) later refined McLaren's (1981) model using probabilistic principles. They also found that one conclusion from McLaren's (1981) paper was false, namely that sediments may become finer and more positively skewed in the transport direction. The skewness of such sediments must become more negative. Gao and Collis

(1991) and McLaren and Bowles (1991) have discussed and attempted to fine tune the "McLaren Method" outlined above for sediment transport paths.

Due to the controversial nature of the interpretation of the cumulative frequency curves, flow characteristics were determined for the study sites by using equations developed for overland flow and stream flow. These equations estimated flow shear stress or flow velocity based on grain diameter and slope gradient. They have been discussed in more detail in the chapter on hillslope processes. This data was used in addition to histograms which illustrate the weight percentage particular grain size intervals have for a sediment sample.

5.4.3. Clast Characteristics, Recording and Sampling

Clast characteristics within diamictic facies at both sites studied were recorded using a scheme outlined by Shultz (1984). Shultz (1984) measured clast sizes and orientations in "...cross-sectional views, using line-transects on outcrop faces and photographs." Transects were generally 3m in length, with maximum clast size for a bed taken as the apparent maximum diameter of the largest clast within 10cm of the transect line. Shultz (1984) expressed clast packing density quantitatively as, "...the volumetric fraction of the whole...made up by clasts." Packing density was estimated by measuring the intercepted clast lengths and "...dividing the sum of these lengths by the transect length". He also considered the "apparent imbrication" and "apparent long-axis orientation" of clasts. However, in the present study this approach was greatly complicated by the paucity of suitable outcrops in the study area at which to measure apparent long-axis orientation and the relatively equant shape of the clasts.

Howard (1993) has investigated the statistical aspects of clast counting. The total probable error of a clast count is due to counting and/or sampling errors. The former can occur when some number of clasts less than the actual amount in a population is counted, while a sampling error can occur when there is an uneven distribution of clasts in a population. Howard (1993) suggested that these problems may be minimized by counting 4 closely spaced subsets of 100 randomly chosen clasts each, or by counting all clasts above a certain size. He preferred the latter technique.

The degree of certainty concerning conclusions drawn up in relation to a whole clast population must obey 3 basic requirements:

1. The percentages observed within the sample must be representative of those in the whole population.
2. Each clast counted within the sample must have had an equal chance of being counted within the whole population, based solely on abundance and not on any other factor.
3. Each sample must have an equal and independent chance of being chosen.

Ideally the exact point on an outcrop and the clasts counted at that point are chosen at random. However, in practice clasts may be physically concealed or inaccessible, so not all clasts have an equal chance of being chosen and a truly random sample is not attainable. The nature of the outcrops investigated in this study made it impossible to obtain truly random samples. As a result the degrees of certainty could not be calculated accurately.

5.4.4. Atterberg Limits

Atterberg limits are the water contents at which soils display certain properties. The limits are known as the *liquid limit* and the *plastic limit*. In this study the Casagrande apparatus and cone penetrometer were used to determine the liquid limit of both sections according to methods proposed by Kezdi (1980) and the British Standards Institution (BSI, 1975). The liquid limit is defined by Kezdi (1980) as the "...water content (W_L%) at which the two sides of a groove cut in the soil sample contained in the cup of a Casagrande apparatus would touch over a length of 12mm after 25 impacts". For the cone penetrometer, the water content at a cone penetration of 20mm is taken as the liquid limit.

The plastic limit of sediments from both sites was also determined using a scheme proposed by Kezdi (1980), whereby soil is rolled into threads. He defined the plastic limit as, "...the minimum water content (W_P) of the soil at which threads of 3-4mm thickness can be rolled out without crumbling." The difference between the liquid and plastic limits can be used to calculate a *plasticity index* (I_P):

$$I_P=W_L-W_P$$

These Atterberg limits were used to get an indication of the conditions under which the studied hillslopes could be expected to fail.

CHAPTER 6

RESULTS

6.1 CLIMATE

6.1.1. Temperature

The mean monthly maximum and minimum temperatures were obtained for two coastal weather stations and 6 inland stations in the vicinity (within a 50km radius of Alexandria) of the study sites. The two coastal stations were at:

1. Bathurst NS [33°31'S 26°49'E]
2. Cape Padrone [33°45'S 26°27'E]

The inland stations were located at:

1. Endeavour, Grahamstown [33°28'S 26°22'E]
2. Grahamstown, vet [33°18'S 26°32'E]
3. Midhurst, Grahamstown [33°27'S 26°21'E]
4. Rockhurst, Grahamstown [33°06'S 26°18'E]
5. Southey's Hoek [33°11'S 26°55'E]
6. Vaalkranz, Paterson [33°22'S 26°03'E]

Mean Annual Air Temperature (MAAT) ranges from 16.4°C at Grahamstown-vet to 19.7°C at Southey's Hoek. Half the stations recorded values between 18-19°C. The lowest mean annual maximum temperature occurred at Cape Padrone (21.61°C) and the highest at Southey's Hoek (26.09°C). The lowest mean annual minimum occurred at Grahamstown-vet (10.44°C) and the highest at Cape Padrone and Endeavour (both 14.08°C). The smallest difference between maximum and minimum mean annual temperature occurred at Cape Padrone (7.53°C) and the largest at Rockhurst (14.48°C).

The mean monthly temperatures at all stations are highest in February and lowest in July, except for Grahamstown-vet and Rockhurst where the lowest temperatures occur in June. The mean monthly temperatures range from 24.15°C at Southey's Hoek in February to 12.0°C at Grahamstown-vet in June. The highest values for mean monthly maxima at most stations are in February, except for Rockhurst and Vaalkrantz where they occur in

January and Midhurst where they occur in March. The lowest values for mean monthly maxima occur in June for the inland stations, except for Midhurst where they occur in July. The lowest maxima at the coastal stations are slightly later in the year, namely July at Bathurst and August at Cape Padrone. The lowest values for mean monthly minima occur in July for all stations.

At the coastal stations, Bathurst and Cape Padrone, no months have mean monthly minima below 10°C, although at both stations from June to September these values were between 10°C and 12°C. The inland station at Endeavour shows the same characteristics. At Midhurst, Southey's Hoek and Vaalkrantz mean monthly minima are below 10°C from June to August, while at Grahamstown-vet the same occurs over an additional two months, from May to September. At Rockhurst mean monthly minimum temperatures are below 10°C from May to September, but unlike other stations one month, July, is below 5°C.

The highest recorded temperatures at the coastal stations for a daily maximum was 42.7°C and for a daily minimum, 32.9°C, both for Bathurst. The lowest recorded daily maximum and minimum for these stations was also recorded at Bathurst, 7.9°C and 1.7°C respectively. At the inland stations, the highest daily maxima and minima were both recorded at Vaalkranz namely, 46°C for the former and 28°C for the latter. The lowest daily maximum and minimum were both recorded at Grahamstown-vet, namely 6.4°C and -8°C respectively. The latter, unusually, occurred in November (1 November 1970).

Regression lines were plotted through mean monthly maximum and minimum temperatures for 2 coastal stations (Bathurst, Cape Padrone) and 2 inland stations (Grahamstown-vet, Southey's Hoek) for the months of the solstices (June, December) and the equinoxes (March, September). The length of the temperature record varies between stations, which means the trends are not strictly comparable. Increases or decreases in temperature generally amounted to <1°C. Exceptions are restricted to the two inland stations. Maximum temperatures during the month of the spring equinox have undergone an increase in temperature between 1-2°C, in contrast to a decrease of 1-2°C during the month of the autumn solstice at Grahamstown-vet. At the same station minimum monthly mean temperatures during the summer solstice have decreased between 1-2°C. Maximum mean monthly temperatures at Southey's Hoek have undergone increases of 2-3°C during

June and 3-4°C for September. The minimum mean monthly temperatures at Southey's Hoek during September have decreased between 1-2°C. However most of the trends are of questionable significance as apart from the trends for maximum mean monthly temperatures for June and September at Southey's Hoek interannual variability is greater than net increases or decreases in temperature for the length of the temperature record at each station. This stability in the temperature regime appears to be bourne out both regionally and between stations for the 4 months for which the regression analyses were done. No pattern of general increase or decrease in temperatures is apparent for the solstices or equinoxes inter or intrastationally for maximum and minimum mean monthly temperatures. There is a slight preponderance of trends to show an increase in values rather than a decrease. This is most marked in the mean monthly minimum for the stations. March and December have more decreasing than increasing trends, while the opposite is true for June and September.

6.1.2. Precipitation

6.1.2.1. Rainfall

Mean annual rainfall ranges from 913.3mm per annum at the Alexandria forestry station (33°42'S 26°22'E) to 518.5mm per annum at Langholm (33°26'S 26°45'E). The spatial distribution of mean annual rainfall is indicated in Fig.6.1.. Rainfall decreases from the NW

to the SE. Two stations with long rainfall records (+100 years) were examined in more detail, namely Port Alfred and Alexandria (Police Station). At Port Alfred no rain fell on 79% of the days. On 21 occasions daily rainfall exceeded 100mm. The highest daily rainfall occurred on 27 May 1967 when 167.0mm was recorded. The correspondent for "Grocott's Mail" at the time (Page 3, Tuesday 30 May 1967) reported a "cloudburst" of 4.5 inches (114.3mm) at 4a.m. on the Saturday morning. At Alexandria (Police Station) no rainfall was recorded on 82% of the days. Daily rainfall exceeded 100mm 13 times and 200mm once, on 25 September 1952 when 290.6mm fell.

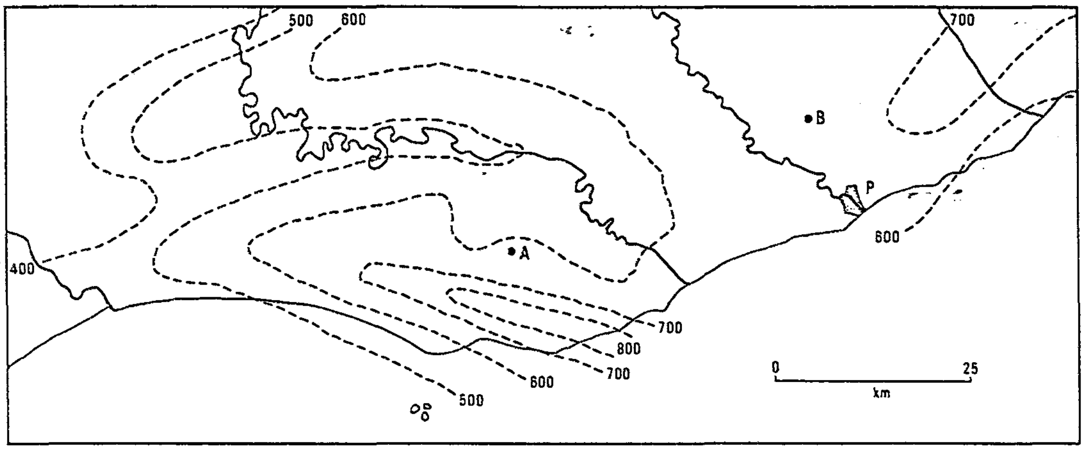

Figure 6.1. Mean annual rainfall (mm) in the Alexandria district [A=Alexandria, B=Bathurst, P=Port Alfred] (after Marker and Sweeting, 1983).

The wettest times of the year are spring and autumn, with peaks in October and March. The autumn peak can often be divided into two subsidiary peaks, namely March and May. March is invariably larger, except for Alexandria (Forestry Station) where May has marginally greater rainfall. The wettest month is in October, except at Clifton and Sydney's Hope where March is the wettest month. The driest times of the year are mid-summer (January) and mid-winter (June/July). January is normally the driest month of the year, but at Bathurst and Sydney's Hope it is June and at Port Alfred July. There is sometimes a drop off in rainfall for April in relation to March and May (i.e. at stations Alexandria [Forestry Station], Alexandria [Police Station], Bushmans River and Port Alfred). All stations have a maximum mean monthly rainfall in October apart from Clifton and Sydney's Hope where it occurs in March. The lowest mean monthly rainfall values occur in January (Alexandria [Forestry Station], Alexandria [Police Station] and Bushmans River), June (Bathurst, Clifton, Sydney's Hope) and July (Port Alfred).

6.1.2.2. Rainfall threshold values for slope failure and debris flow

Caine (1980) suggested that slope failure would occur after a minimum of 103mm in 24hrs. Innes (1983) proposed that if 24.5mm fell in a day debris flows could occur. At Alexandria (Police Station) the Caine threshold value has been exceeded 11 times and the

Innes value 567 times between June 1885 and June 1994. At Port Alfred the corresponding values are 11 and 524 for the period October 1882 to June 1994. At Alexandria (Police Station) this amounts to a recurrence interval of 0.13/year for Caine's threshold value and 5.25/year for Innes' threshold value. The corresponding values at Port Alfred are 0.1/year and 4.71/year. Alexandria (Police Station) and Port Alfred are the only two stations near the research sites with records over 100 years long. A search of newspaper articles which reported the most extreme rainfall figures made no mention of any slope failures or debris flows.

6.1.2.3. Snowfall

Snow has fallen in the Grahamstown district 6 times since the first recorded snowfall on the 9 August 1883 (Table 6.1.). There are 3 other references to snowfall in the Grahamstown district in the local newspaper, the Grocott's Mail. One of these appears to refer to a snowfall at Bedford on 15 July 1917 and is thus not applicable to this study, while the other two for 1927 and 1908/9 could not be verified. On the 7 July 1934 "snow-clad hills" were reported to surround Grahamstown in all directions (Grocott's Mail - 9 July 1934). The heaviest snows appear to have fallen on the 20 June 1976. Snow was reported to have fallen within a wide radius of Grahamstown from the Blaaukrantz river 20km southeast of the city to a point 10km to the northwest and at least as far as the village of Riebeeck East 40km to the west. However, none of the snow falls appear to have ever been to a depth greater than 30cm (Fig.6.2.). Most of the snowfalls occurred in the morning between 8.00am and 10.00am.

Table 6.1. Days on which snowfall has been recorded in Grahamstown.

Date	Comment
9 August 1883	Snow "6 inches deep on the flats"
7 September 1931	
7 July 1934	Snow-clad hills in all directions, particularly those surrounding the city of Grahamstown. Heaviest snowfall ever in the Grahamstown region.
20 June 1976	
2 September 1978	Snowed between 10.00am and 10.45am. Ice on the university library windows.
12 June 1979	Snow started falling at 8.20am. and stopped about 11.15am.

Figure 6.2. Snow on the "Great Field", Rhodes University, Grahamstown, 20 June 1976.

6.1.3. Evaporation

Evaporation values were obtained in order to estimate soil moisture conditions at each study site. This was achieved by comparing the rainfall and evaporation values at stations where both had been recorded. The greater the estimated soil moisture content, the greater the susceptibility of the hillslope to slope failure (Crozier, 1986).

For all the stations for which monthly evaporation data could be obtained the summer months have the highest evaporation values and the winter months the lowest. Evaporation exceeded mean monthly rainfall at Bathurst in January and December, at Cape Padrone from December to February, at Midhurst from March to September and December and at Endeavour in January, March, May-July, November and December.

The driest and only months at Bathurst where evaporation exceeds rainfall [evaporation excess over rainfall in parenthesis, the opposite for the wettest months] is in January

(15.05mm) and December (8.16mm). The wettest months are March (23.68mm), September (20.80mm) and October (25.73mm) when rainfall exceeds evaporation. At Cape Padrone the driest months are December (13.90mm), January (18.55mm) and February (6.11mm). The wettest months at Cape Padrone are May (29.88mm), June (39.41mm), July (41.83mm), August (33.53mm), September (21.47mm) and October (32.24mm).

The driest months at Midhurst are August (37.84mm), September (32.6mm) and December (41.94mm). The wettest months at the same station are February (22.44mm) and October (61.56mm). These two months are also the only two months for which rainfall exceeds evaporation. The driest months at Endeavour are May (22.32mm) and July (26.00mm). The wettest months of the year are February (21.09mm) and April (17.37mm).

The driest month at most of the stations investigated was January, while the wettest month was generally October.

6.2. SEISMICITY IN THE EASTERN CAPE

Fifteen seismic events have been reported for the eastern Cape coastal areas since the first report on the 21 May 1850 (Table 6.2.). This also appears to have been the strongest earthquake recorded in the area (MMS Intensity of V - VI). The minimum MMS intensity for slope failure (IV - Keefer, 1984) has been exceeded 7 times. Seismic events therefore occur at a rate of 10.5/100 years or if only those capable of triggering slope failure are considered 4.9/100 years.

6.3. HILLSLOPE PROFILE AND GEOMORPHOLOGY

6.3.1. Burchleigh

A road cutting has exposed a tangential section of a hillslope just above a dry stream bed at the base of a steep slope (Fig.6.3.). The section is on the inside of a broad sweep of the stream above its confluence with the Bushmans River (Fig.6.4). The tributary stream forms an alluvial fan before entering the Bushmans River. The rivers have cut into a planation surface which Marker and Sweeting (1983) identified as a marine erosion surface. This marine surface, at 240m, is the highest they identified. The Bushmans River flows in straight lines connected by meander segments. The straight lines probably follow master joints in the underlying bedrock.

Table 6.2. The historical incidence of seismic events (modified Mercalli scale intensities) in the eastern Cape.

LOCATION	DATE	EST.MMS	a_H	a_V	EVENT/ 1000yr
GT	1850/05/21	V-VI	42.17	12.88	07
GT	1870/08/03	II	5.62	1.15	21
PE	1883/09/30	III	10.00	2.29	21
GT	1895/04/09	II	5.62	1.15	21
PE	1910/10/21	II	5.62	1.15	21
GT	1912/02/20	III	10.00	2.29	21
PE/PA/GT	1912/12/04	IV	17.78	4.57	21
GT	1932/08/09	V	31.62	9.12	21
GT	1933/02/25	IV	17.78	4.57	21
PEDDIE	1952/09/23	II-III	7.50	1.62	21
GT	1953/01/20	III	10.00	2.29	21
GT	1957/04/13	V	31.62	9.12	21
PE	1969/09/29	V	31.62	9.12	21
PE	1977/11/04	III-IV	13.34	3.24	07
PE	1986/10/05	IV	17.78	4.57	21

-(GT=Grahamstown, PE=Port Elizabeth, PA=Port Alfred.)
-(Estimated MMS intensities are according to the 1956 version formulated by Richter)
-(a_H=horizantal ground acceleration, a_V=vertical ground acceleraton; both in $cm.s^{-2}$)

Overfolded Late Devonian Witteberg Group quartzites (Fig.6.5.) are exposed on the upstream side of the section. These quartzites have supplied the clastic debris strewn across the hillslope and buried within the colluvial sediments.

The hillslope directly above the exposed section is characterised by a straight gently inclined lower and mid-slope and a straight steeper upper slope (Fig.6.6.). The hillslope crest is rounded and terminates on the planation surface (Fig.6.6). The lower slope near the base has a gradient of 26°, the mid-slope 35° and the steep, rocky upper slope 53°. The gradient at the crest of the hill is <10°.

6.3.2. Spring Grove

The section is located at the start of the outer bank of a broad meander of the Assegaai, a tributary of the Kariega River. The river has cut back into the hillslope exposing a deep (+10m) tangential section of the hillslope (Fig.6.7.). The hillslope rises steeply away from the section up on to another greater hillslope which forms the inside of the upstream meander. This latter surface slopes gently up on to a regional planation surface, one of the marine surfaces (200m) identified by Marker and Sweeting (1983) (Fig.6.8.). Shale bedrock of the Carboniferous Dwyka Formation outcrops irregularly on the slope (eg.40m from the base) and has supplied most of the clastic debris. Quartzite clasts are also common.

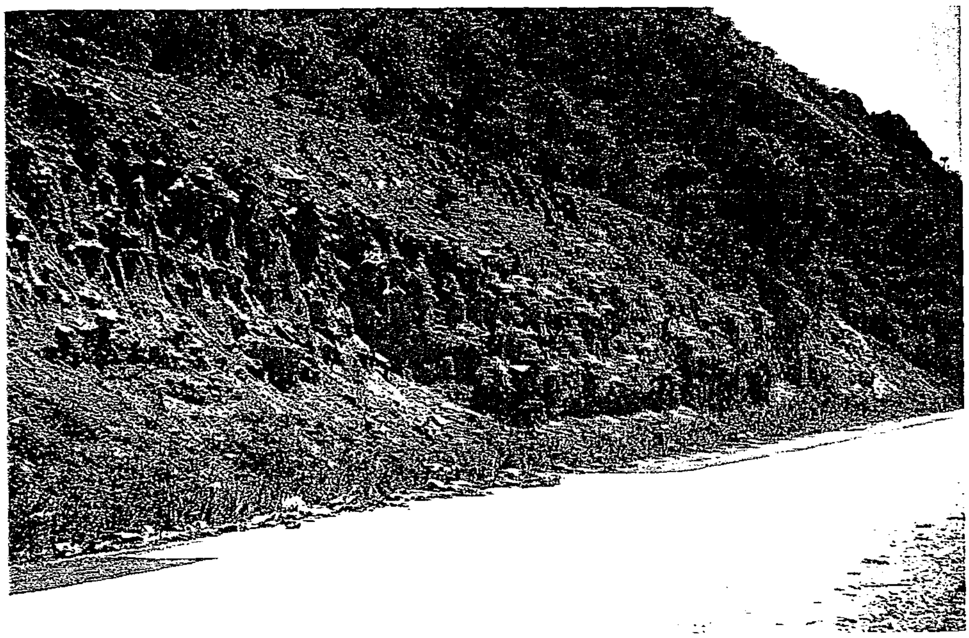

Figure 6.3. Road cutting at Burchleigh exposing a tangential section of the hillslope.

The hillslope profile is displayed in Fig.6.9.. The lower slope directly above the exposed section is straight and inclined at a low angle (22°). The mid slope is straight and slightly more steeply inclined (30°), while the upper slope is also straight but inclined at a low angle (15°). The hillslope crest is sharply rounded.

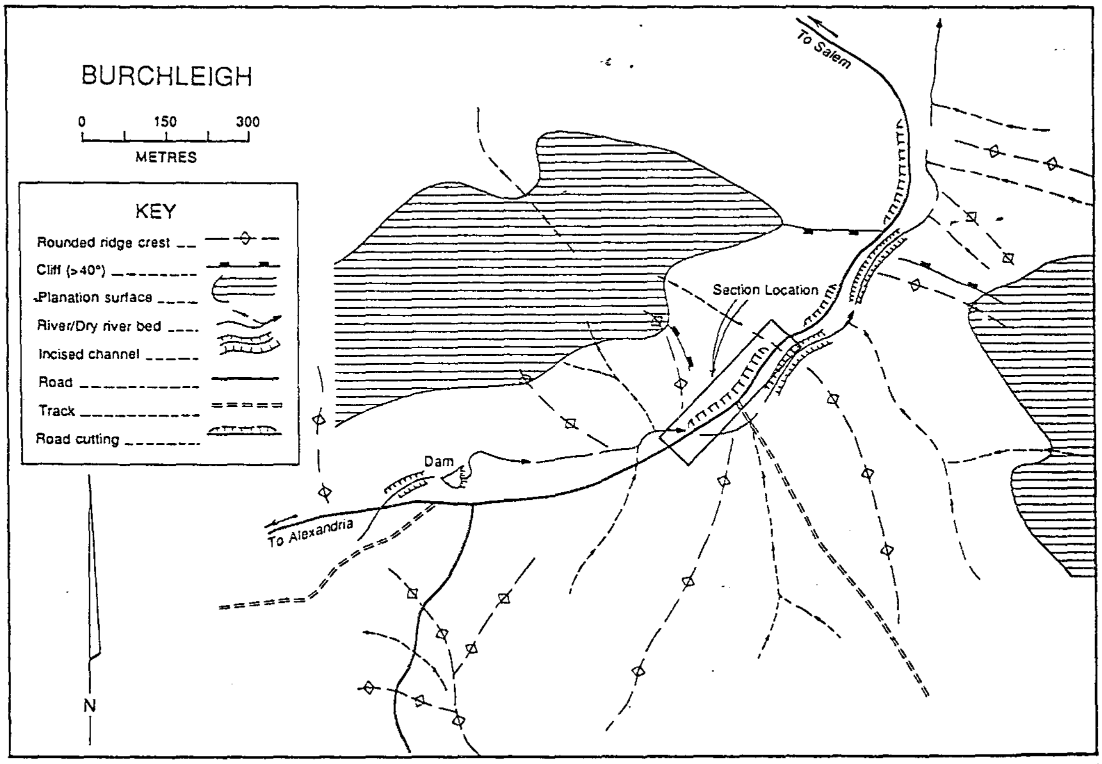

Figure 6.4. A geomorphic map illustrating the dominant features of the Burchleigh landscape.

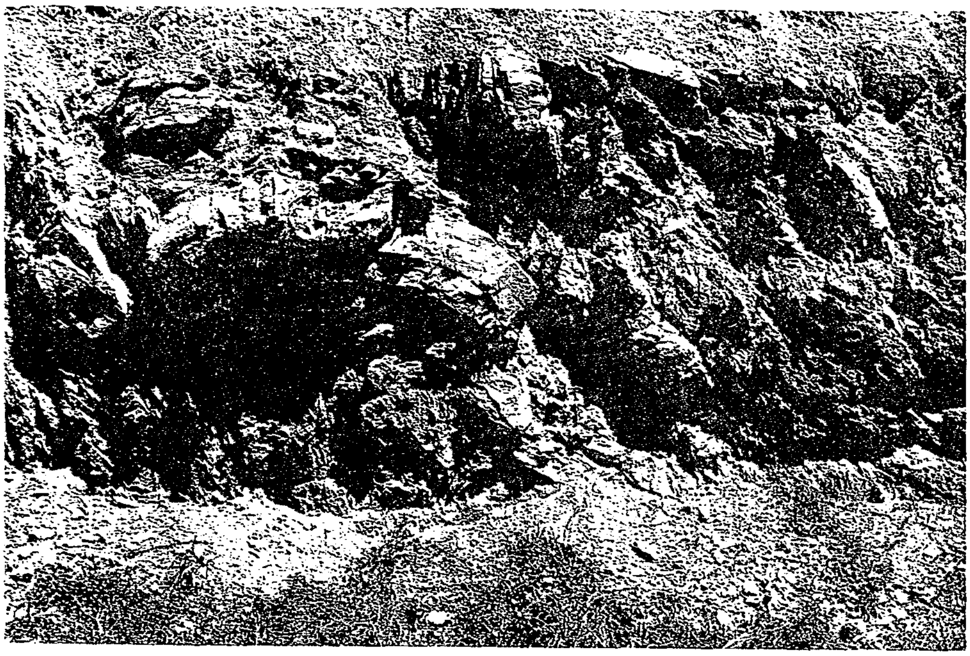

Figure 6.5. Overfolded Late Devonian Wittebrg quartzites exposed within the road cutting at Burchleigh.

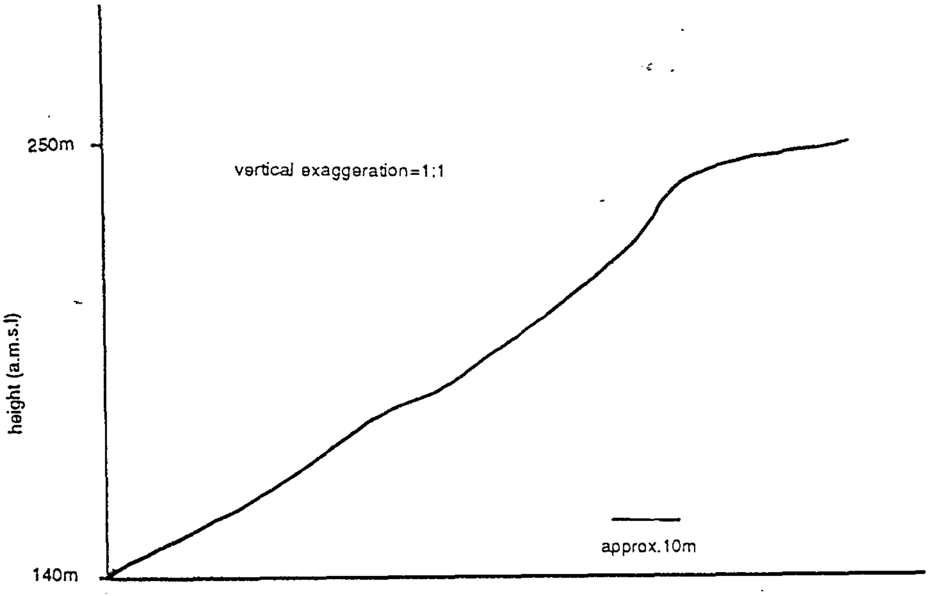

Figure 6.6. Hillslope profile at Burchleigh.

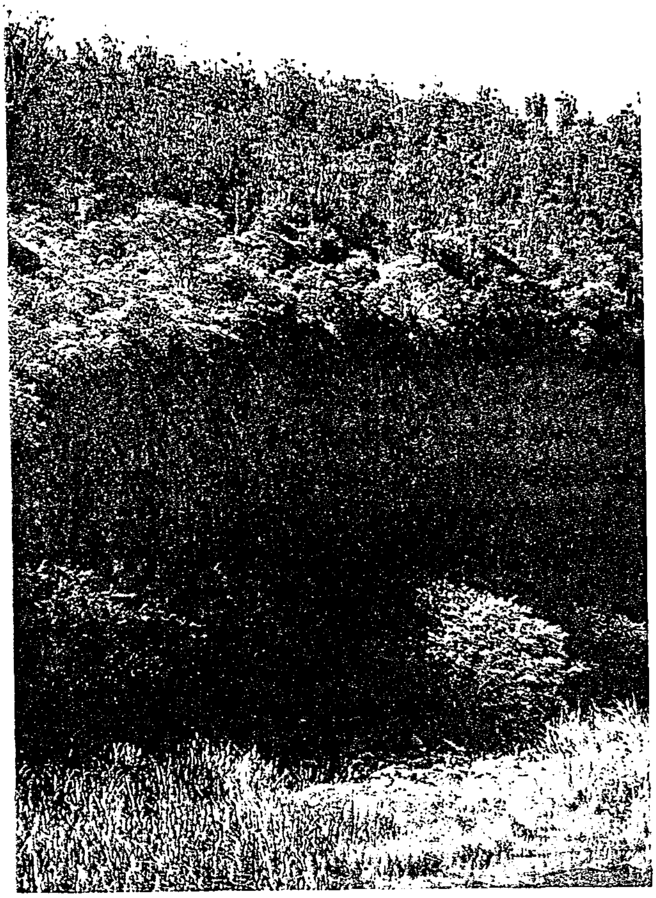

Figure 6.7. The cliff of unconsolidated sediment and the lower hillslope at Spring Grove.

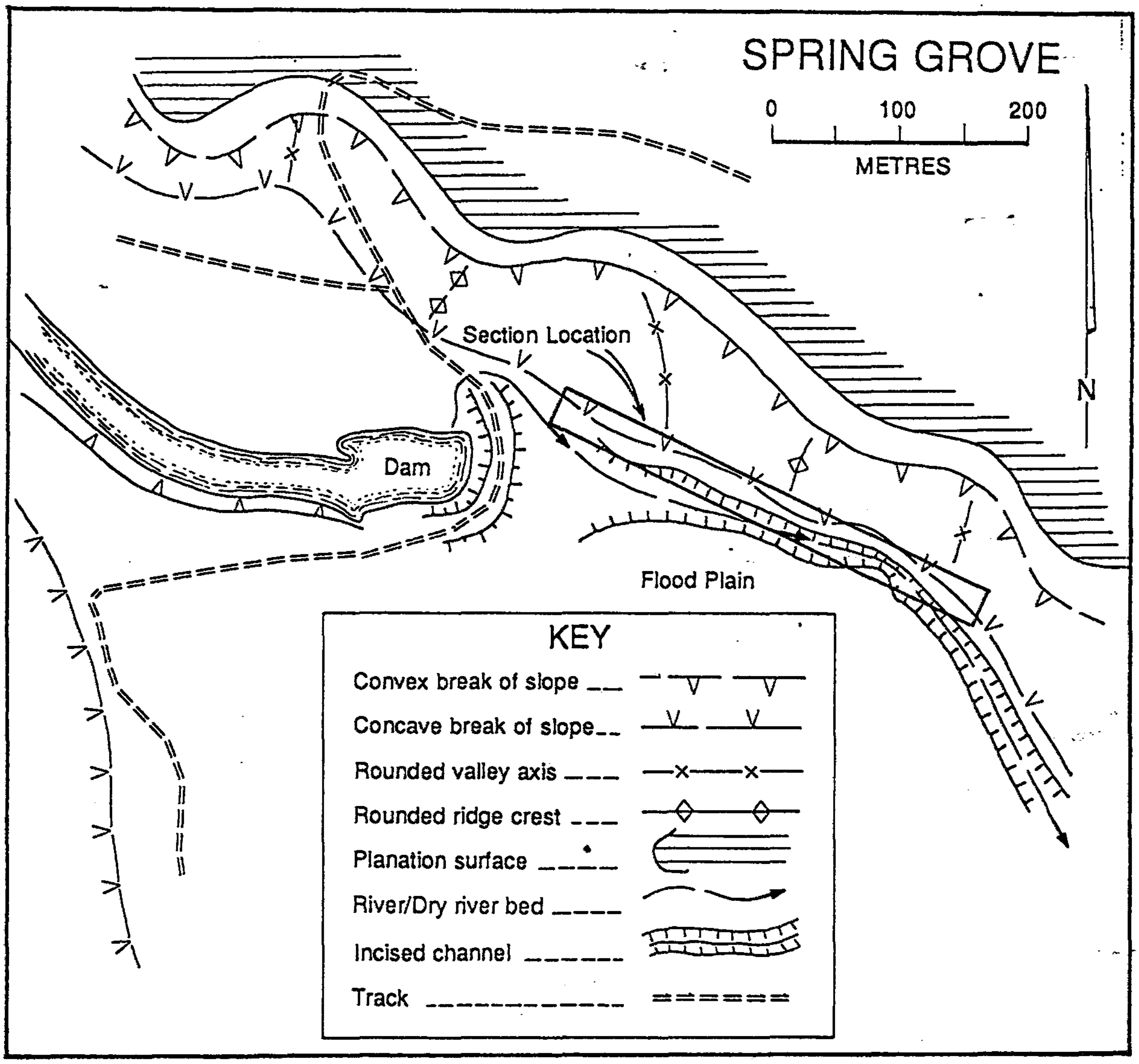

Figure 6.8. A geomorphic map illustrating the dominant features of the landscape at Spring Grove.

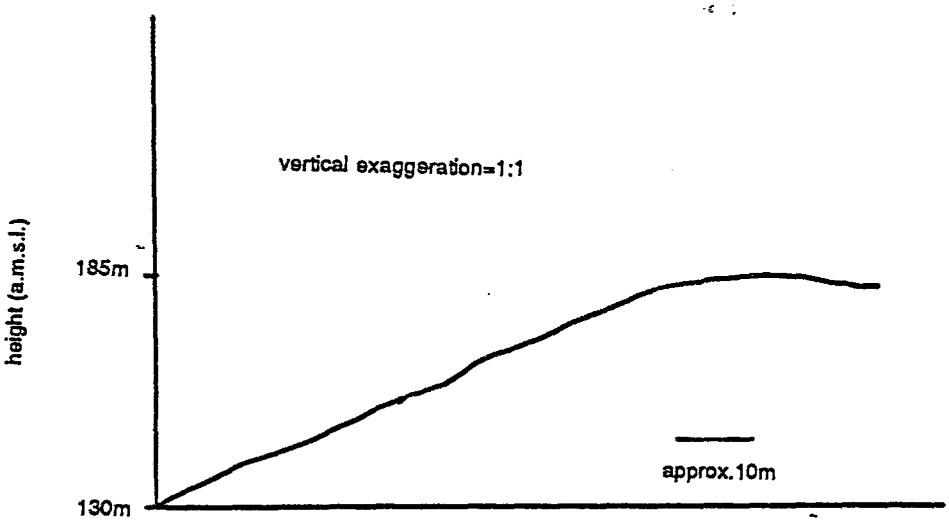

Figure 6.9. The hillslope profile at Spring Grove.

6.4. SEDIMENTOLOGY

6.4.1. Slope Architecture (Bed Geometry)

6.4.1.1. Burchleigh

Tabular quartzite clasts at the contact between the bedrock and colluvial sediments lie parallel to the interface. The sediments within the section (Fig.6.10.) are predominantly fine grained colluvium (Sm) interspersed with stone lines of large clasts (Fig.6.10. & 6.11.) These large (i.e.<30cm) clasts lie flat, parallel to the hillslope surface. They are not imbricated. At irregular intervals within the above sediment and parallel to the stone lines are thin (mostly <50cm thick) lenses of clast supported gravels (Gcm) (Fig.6.12.). They may extend laterally for 5-10 times the bed thickness. The gravels do not occupy palaeochannels. There is no obvious upward or downward fining within each bed. The clasts are angular and show no apparent preferred orientation. They do not lie parallel to the hillslope surface like the stone line clasts. Only quartzite clasts were found, none were shales. Large clasts (20-30cm) are only common in a bed (Dcm) at the base of the sedimentary sequence near the beginning of the upstream side of the section.

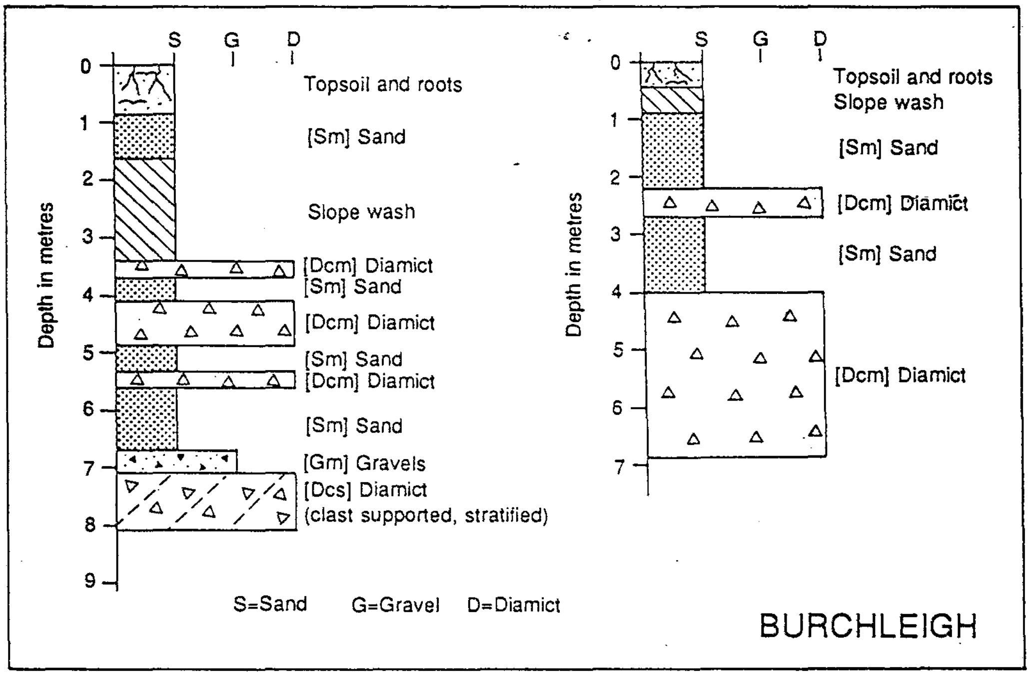

Figure 6.10. Log sections illustrating the typical stratigraphy at Burchleigh.

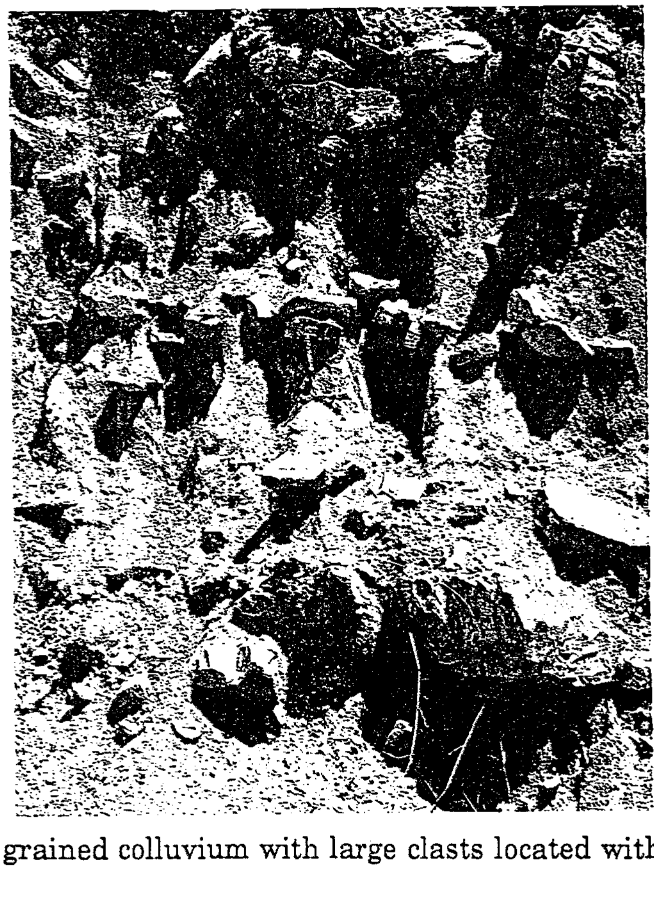

Figure 6.11. Fine grained colluvium with large clasts located within a stoneline.

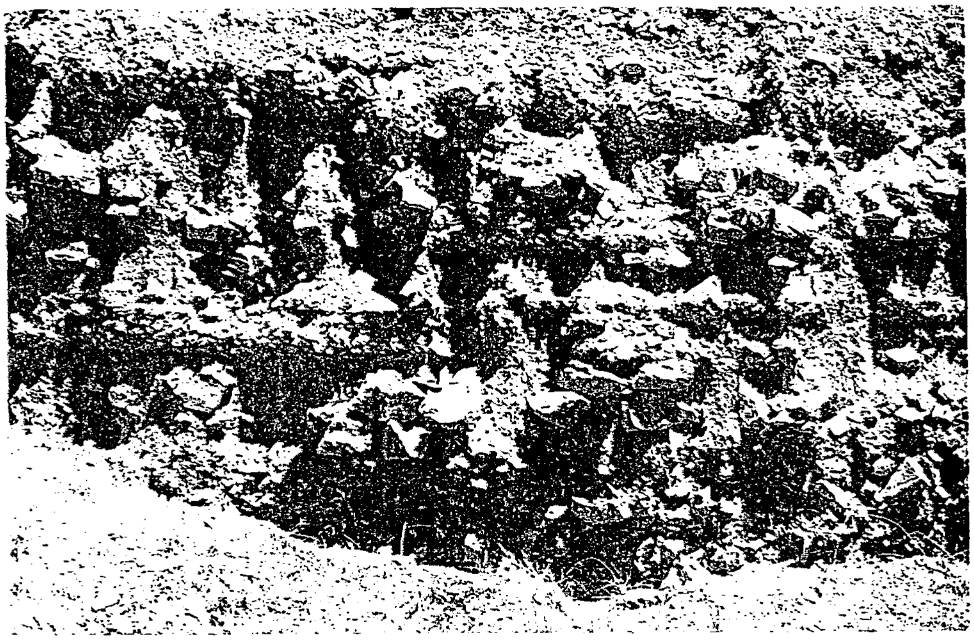

Figure 6.12. Lenses of clast supported gravels within the fine grained colluvium at Burchleigh.

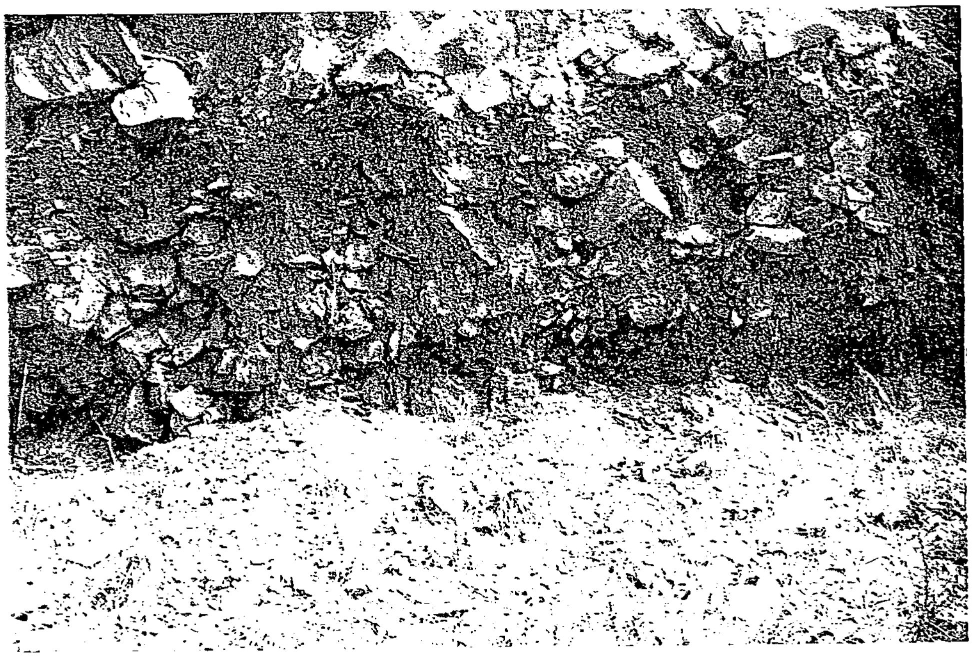

Figure 6.13. Basal diamict with large clasts grading to the right into rounded clast supported gravels.

Figure 6.14. Upper "basal" diamict. Note the the three large imbricated clasts. The diamict is underlain by fine sands.

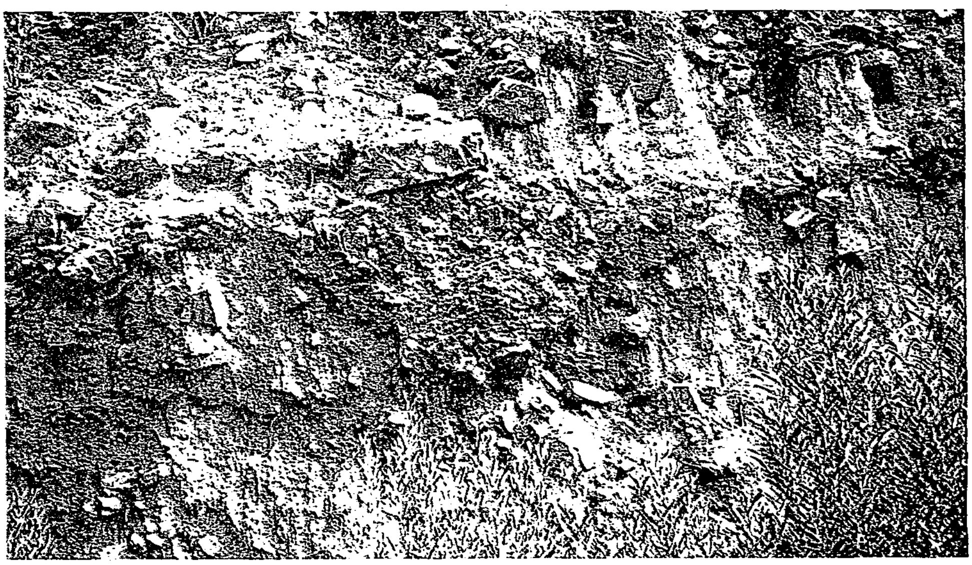

Figure 6.15. The "channelised" gravel/diamict deposit. Note the inverse grading and presence of a few large clasts carried as bedload.

This diamictic bed (Dcm) is different to the gravel beds. It consists of conspicuously larger clasts (Fig.6.12.). It grades laterally downstream into a gravel bed comprised of well rounded pebbles of both shale and quartzite. The clasts within the diamict are very angular. The gravel bed in turn grades downstream laterally into fine grained sands. The diamict shows some signs of imbrication, while there is clear imbrication in the gravel sediments. These sands are overlain in part by another diamict similar to the one described above (Fig.6.13).

A "channelised" gravel/diamict deposit (Fig.6.14.) occurs near the top of the sedimentary sequence halfway across the section. It grades laterally into a gravel deposit very similar to those which form the laterally extensive lenses mentioned earlier above (Fig.6.11). However, in contrast to these tabular beds this deposit has a flat base and convex upper surface. No imbrication is present. The bed is inversely graded (Fig.6.14), although some relatively large clasts (10cm) about the same size as those at the top of the bed were found at its base.

The section is dominated in the central area by a large clast supported diamictic bed (Fig.6.15.). There is no imbrication or preferred orientation of the clasts, which appear very angular and tightly packed. There is no indication of any "levees" or raised portions on the fringes of the bed. The top surface of the downstream side of the diamictic bed is convex. On the far downstream end of the section there is a stratified gravel with clasts that show no imbrication.

6.4.1.2. Spring Grove

The sedimentary sequence is composed of fine grained colluvium (Sm) intercalated with gravel beds (Gm) of approximately the same thickness (Fig.6.16. & 6.17.). Isolated clasts, occasionally arranged as a stoneline, are apparent within the sequence. These clasts lie parallel to the slope surface. In an exceptional case a boulder lies within the fine grained colluvium (Fig.6.18.). The gravel beds are highly lenticular and form a sharp contact with the sediments above and below. They show no apparent imbrication and the clasts appear to have no preferred orientation (field observation) (Fig.6.19.). In some instances the diamicts (Dcm) and gravels (Gm) display inverse grading (Fig.6.20.). A diamict (Dcm) lies at the base of the sequence accompanied by some river gravels (Fig.6.21.). It appears to

Figure 6.16. The thick, centrally situated diamict dominated by large clasts.

occupy a palaeochannel. The clasts appear well rounded (Fig.6.23.) and much larger than those which occur in the gravel beds. The lithology of the clasts within this basal diamict are quartzites whereas shales are common in the gravel deposits higher up in the sequence. Some of the clasts within the this basal diamict display percussion marks (Fig.6.24.).

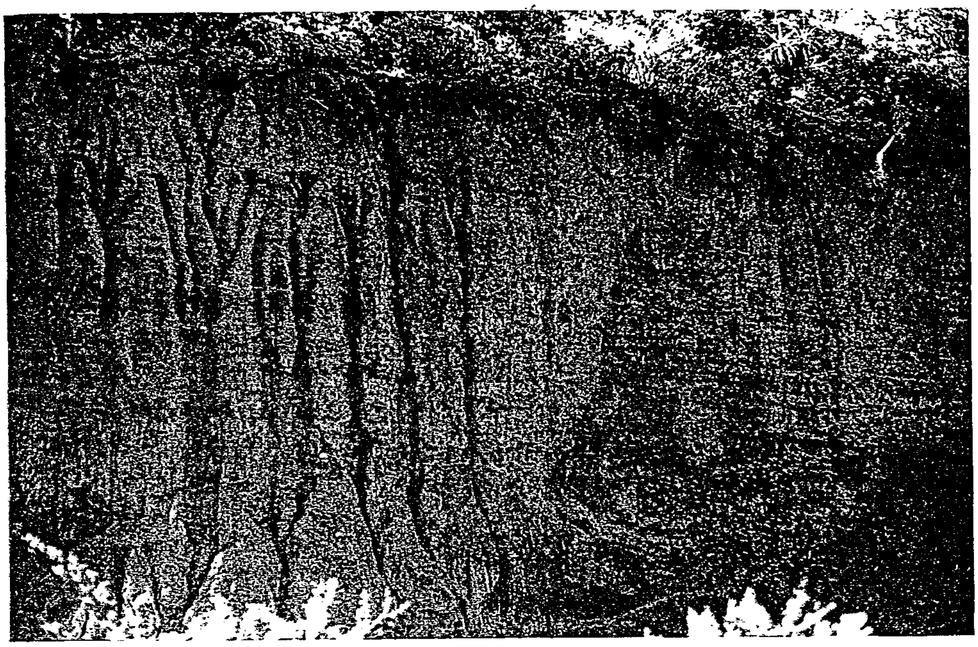

Figure 6.17. The intercalated fine grained colluvium and highly lenticular gravel/diamict deposits at Spring Grove.

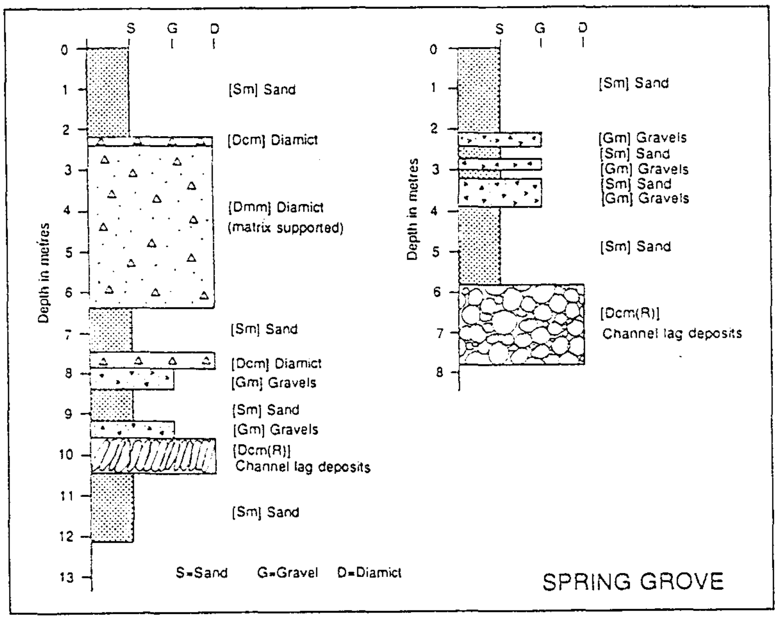

Figure 6.18. Two sedimentary logs illustrating the typical character of the sediments at Spring Grove.

Figure 6.19. The basal section of fine grained colluvium at Spring Grove. Note the abundance of isolated clasts and the large boulder with its flat lower surface parallel to the hillslope.

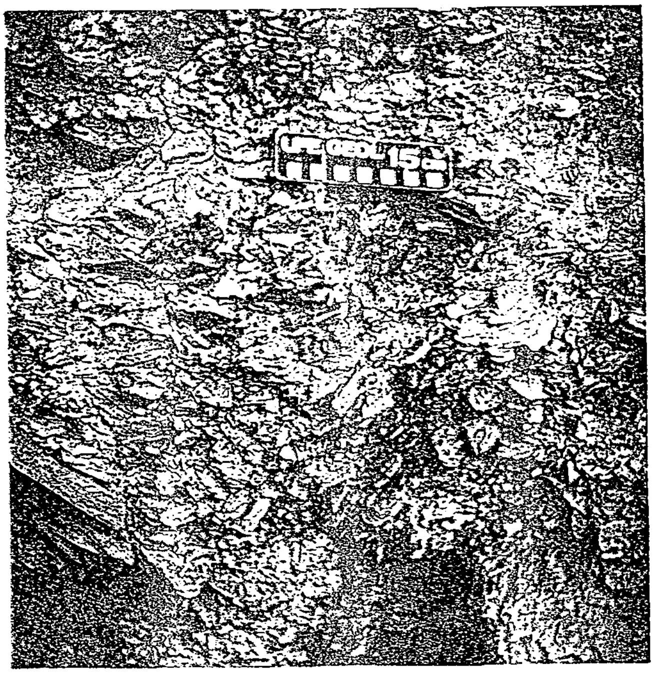

Figure 6.20. An example of unoriented clasts within a gravel bed at Spring Grove.

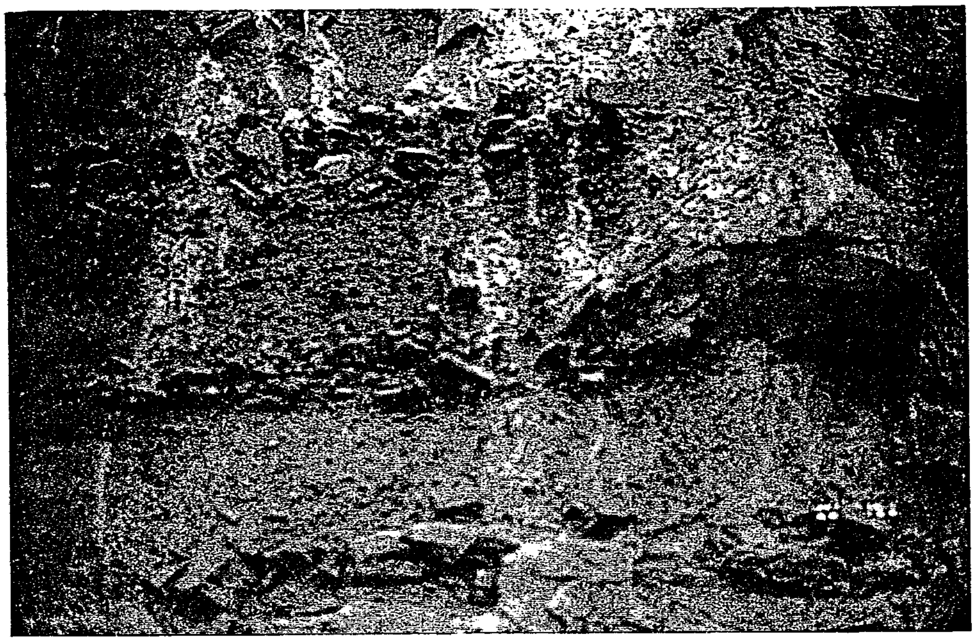

Figure 6.21. Highly lenticular inversely graded gravel bed interpreted as a debris flow. Note the stoneline below the gravel bed.

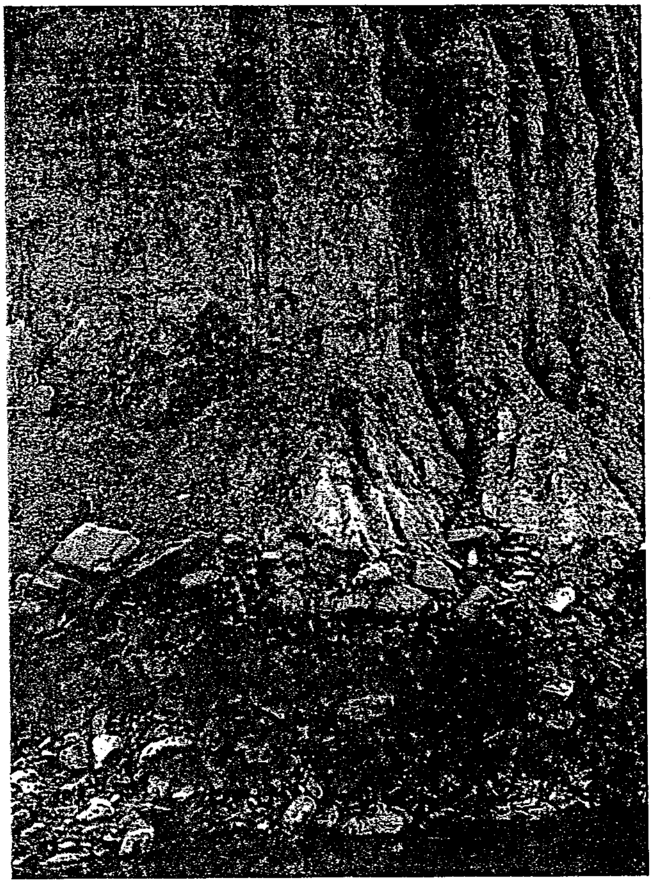

Figure 6.22. The basal diamict interpreted as a channel lag deposit at Spring Grove underlying the rhythmically bedded colluvial sediments.

Figure 6.23. The basal river lag deposit showing the rounded nature of the clasts.

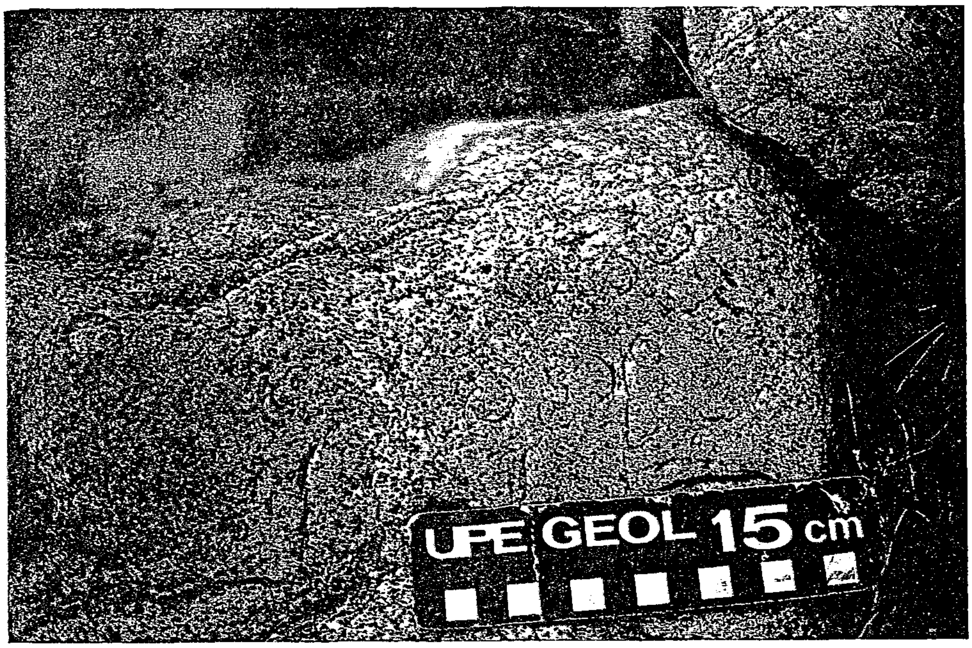

Figure 6.24. A quartzite cobble from the basal diamict which displays classic percussion marks.

6.4.2. Grain Size Distributions

6.4.2.1. Burchleigh

The grain size distributions of samples from the lower slopes are statistically unimodal (e.g. 0m & 40m, Fig.6.25.), while that of the mid and upper slopes is bimodal (e.g. 80m, 129m, 160m & 200m, Fig.6.25.). Thus bimodality decreases downslope, with a concomitant decrease in the significance of the coarser mode relative to the finer mode as a percentage of the total sediment weight (Fig.6.25.). Sediments thus become finer downslope. The finer of the two modes has a peak at 0.125mm, except that from 200m which has a peak at 0.25mm. There is a weakly developed secondary coarse mode at 4mm for the 40m sample. All other samples have coarse modes at 16mm or coarser (Fig.6.25.).

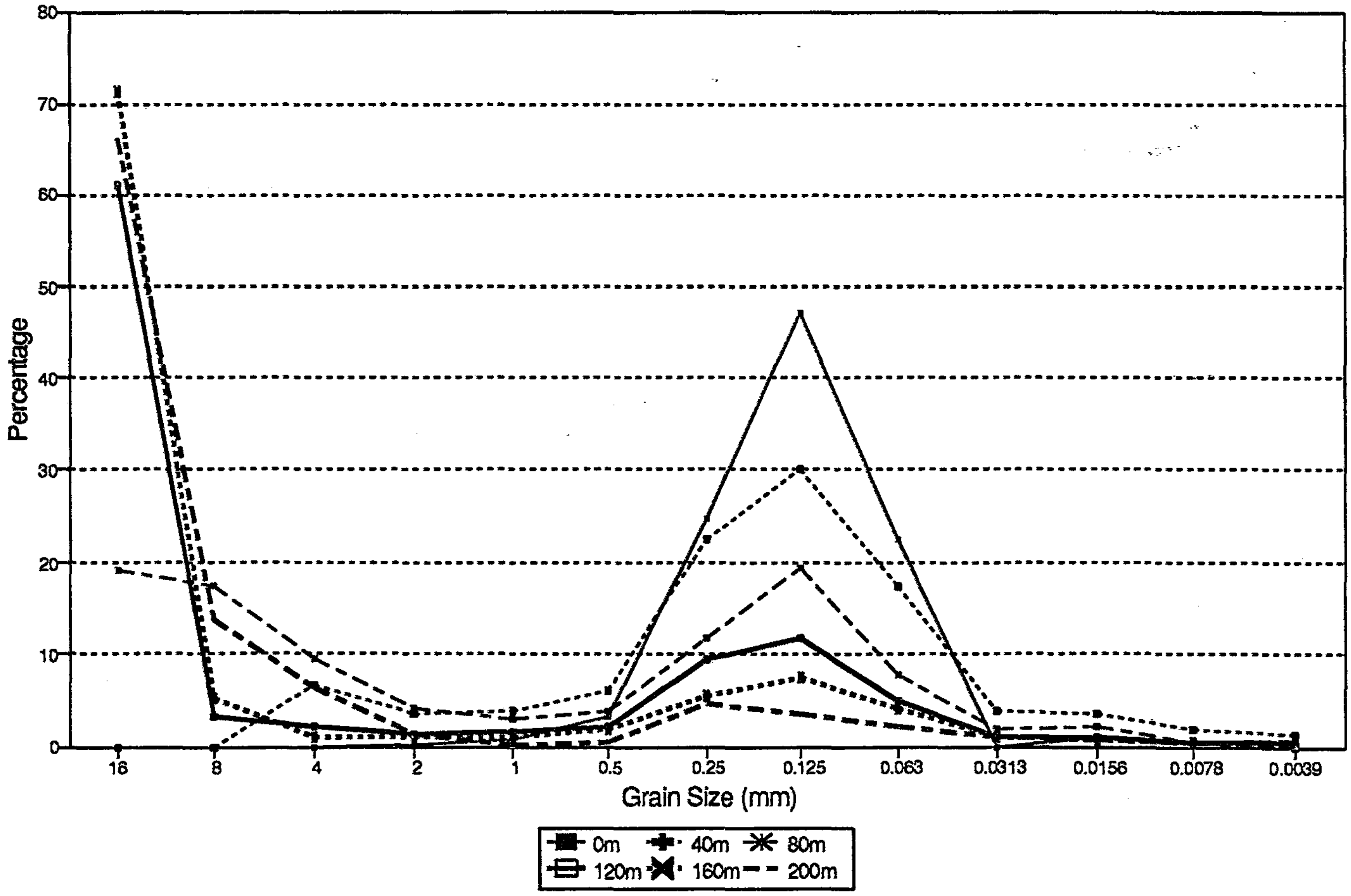

Figure 6.25. The grain size distribution for sediments at 0m, 40m, 80m, 120m, 160m and 200m (top of slope) from the base of the slope at Burchleigh.

The standard deviation of the lumped data (two modes together) shows a decrease down slope from a high of 18.00 at 200m to a low of 7.12 at 80m. From 80 metres the standard deviation increases again to a value of 14.71. The sorting of the hillslope sediments improves downslope to the lower reaches of the slope where upon it deteriorates again. Gale and Hoare (1991) suggested that the greater the standard deviation the less sorted a sediment.

The river sediments at the base of the section have a unimodal grain size distribution (Fig.6.26.). The mode is slightly finer than the "coarse mode" in the slope sediments, but much coarser than the "fines mode". The standard deviation of these sediments is 8.7. The river sediments are thus better sorted than the hillslope sediments, apart from the sample taken at 80m.

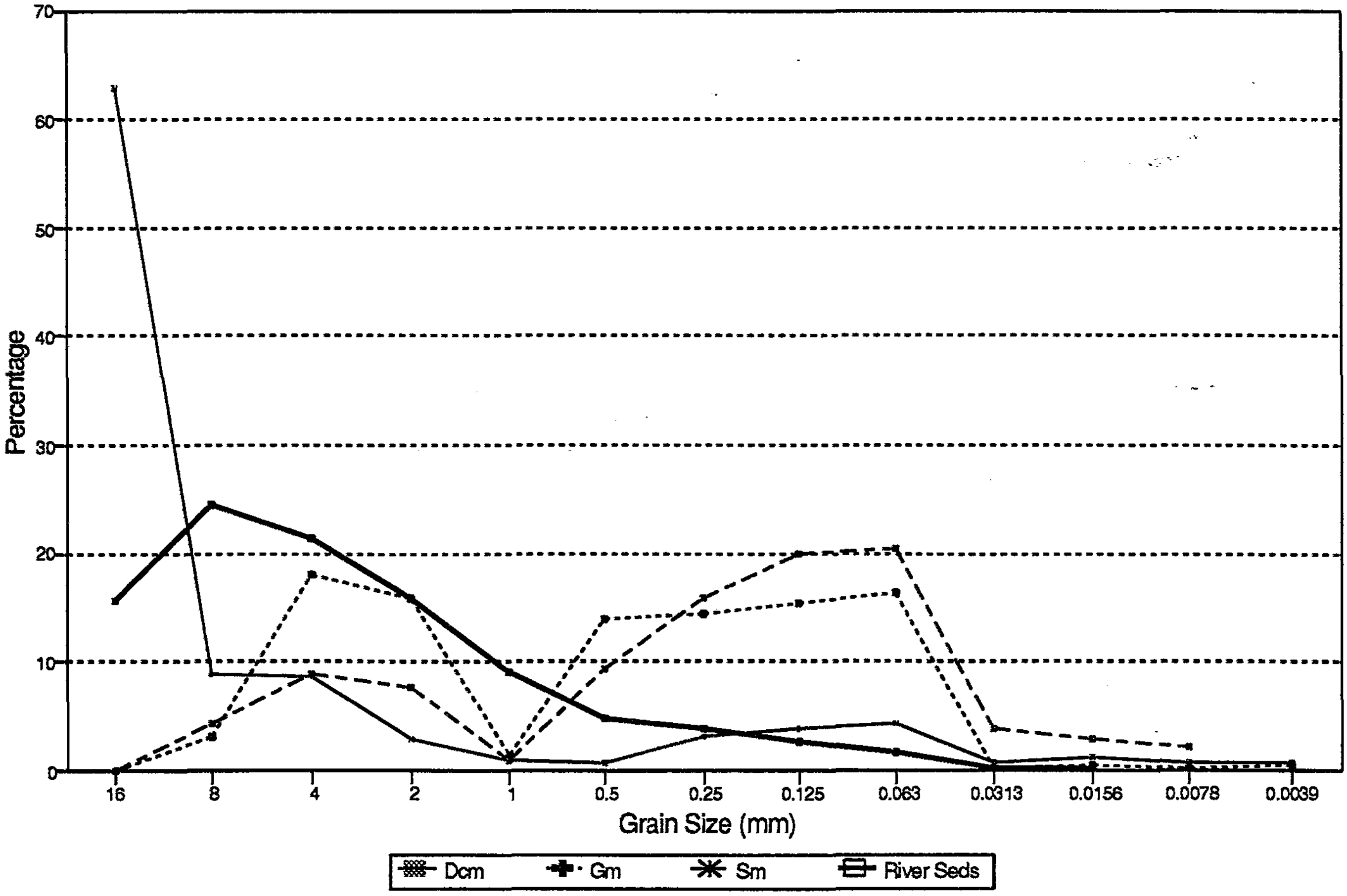

Figure 6.26. The mean grain size distribution for diamicts (Dcm), gravels (Gm), colluvial sands (Sm) and river sediments (River Seds) at Burchleigh.

Within the section, the diamict (Dcm) beds show a very weak bimodal distribution (Fig.6.26.). This bimodality is well developed in the gravels (Gm) and sands (Sm) (Fig.6.26.). The fine mode in the diamict is finer than that found on the hillslope and river sediments (i.e. 0.063mm as opposed to 0.125-0.25mm and 8mm respectively). This fine mode represents the matrix. The coarser fraction is however very similar to that of the mid and upper hillslope sediments in terms of total weight percentage per sample.

The gravel sediments have a coarse modal peak at 4mm and a broad fine modal peak, which has a range from 0.5mm to 0.063mm (Fig.6.26). The two modes are almost equally well developed, with the finer mode containing slightly more sediment (Fig.6.26.). The finer of the two modes which represent the sands is better developed than the coarse mode. The coarse mode peaks at 4mm and the fine mode over a range from 0.125mm to 0.063mm. The standard deviation of the sediments within the section ranges from a high of 16.82 for the diamict to a low of 7.01 for the sands. The gravels have an intermediate standard deviation close to that of the sands, namely 7.74. The fine mode for the sands occurs at 0.063mm.

Table 6.3. The percentages of gravel (>2mm), sand (<2mm>0.063mm), and silt+clay (<0.063mm) at Burchleigh for the sediment types discussed in the text.

Sample	Gravel%	Sand%	Silt+Clay%
0m	0.14	98.26	1.60
40m	10.06	79.95	9.99
80m	50.06	45.49	4.45
120m	67.78	29.40	2.82
160m	78.39	19.53	2.08
200m	87.55	10.60	1.85
Dcm	83.45	13.15	3.40
Gm	36.92	61.45	1.63
Sm	21.09	66.64	12.27
R.Seds.	77.23	21.90	0.87

If the sediments are represented in terms of gravel (>2mm), sand (<2mm>0.063mm), and silt+clay (<0.063mm) the relationship between distance upslope and grain size becomes clearer (Table.6.3.). The percentage of gravel within the sediments increases upslope, while the percentage of sand and silt+clay decreases. The value for silt+clay at the base of the slope (0m) does not follow the trend. The percentage of gravel decreases from the diamicts (Dcm) to the sands (Sm). The diamicts (Dcm) have the lowest percentage of sands, while the sands (Sm) have the highest (Table.6.3.). The gravels (Gm) have the least silt+clay, while the sands (Sm) have the most. The river sediments at the base of the slope have sediment characteristics between those of the diamicts (Dcm) and the gravels (Gm), with an intermediate amount of gravel and a lower silt+clay content than either.

6.4.2.2. Spring Grove

The sediments are bimodal on the upper and lower slopes and unimodal on the mid slopes (Fig.6.27.). At the top of the slope (90m) the sediments are very weakly bimodal with the fines mode poorly developed. The fines mode has a peak at 0.063mm (Fig.6.27.). The unimodal peak at 50m is located at 0.5mm. This is also the value of the coarser mode at the base of the slope (Fig.6.27.). The finer mode at the base of the slope is at 0.063mm, which is the same as that for the sediments at the top of the slope (Fig.6.27.). The weight percentage of the 0.063mm modal fraction for a sample thus increases downslope as the sediments become finer (Fig.6.27.). The coarse mode for sediments at the top of the slope is greater than 16mm. The standard deviation decreases downslope as the sediments become better sorted from a value of 21.19 at the top of the slope (90m) to 6.72 at the base (0m).

Four sediment types were identified in the sedimentary sequence. These were diamict (Dcm), gravels (Gm), sands (Sm) and palaeo-river sediments (PRS). The diamict and gravel are both weakly bimodal, whereas the sands and PRS are unimodal (Fig.6.28. & 6.29.). The diamict has a strongly developed coarse mode greater than 16mm and a subsidiary mode at 0.125mm. The primary mode for the gravels is located within the fines at 0.125mm. The lesser, coarser mode has a peak at 8mm. The lesser peak however is quite broad and starts at the 2mm interval (Fig.6.28.). The standard deviation of the diamict is greater than that for the gravels, 9.18 as opposed to 7.03. The gravels are thus better sorted.

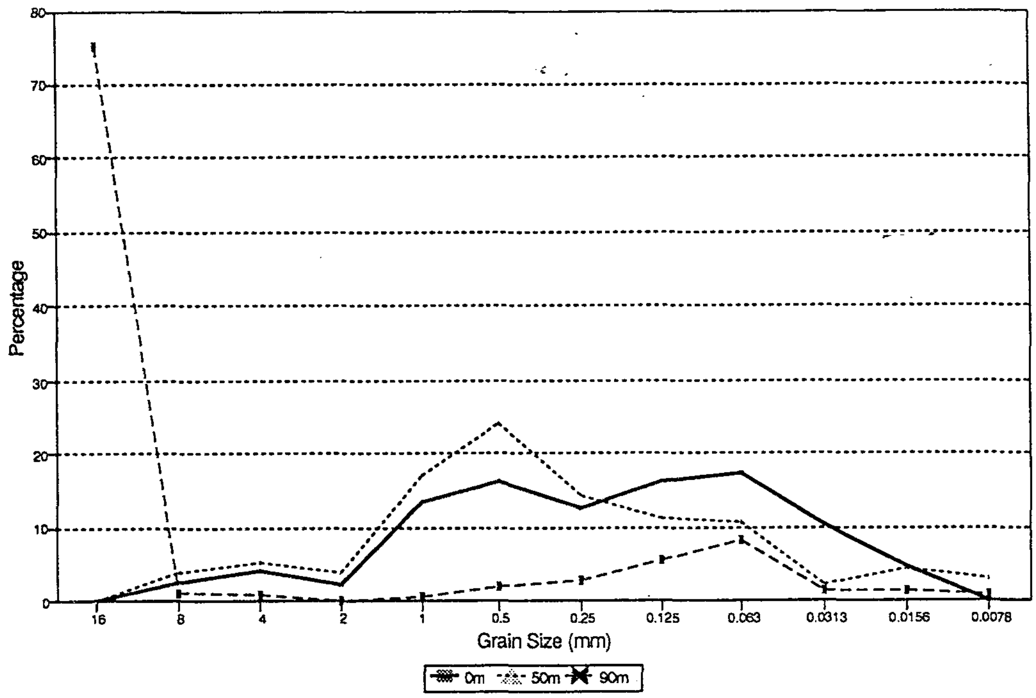

Figure 6.27. The grain size distribution for samples collected at the base of the slope (0m), the midslope (50m) and top of the slope (90m) at Spring Grove.

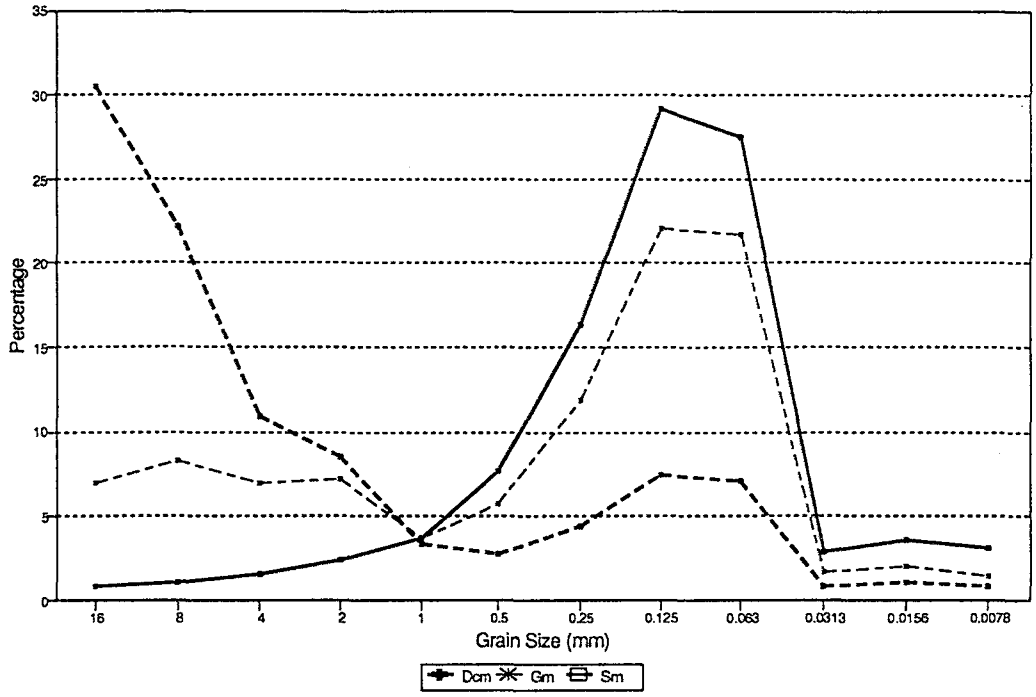

Figure 6.28. The mean grain size distribution for diamictic (Dcm), gravel (Gm) and colluvial sand (Sm) sediments at Spring Grove.

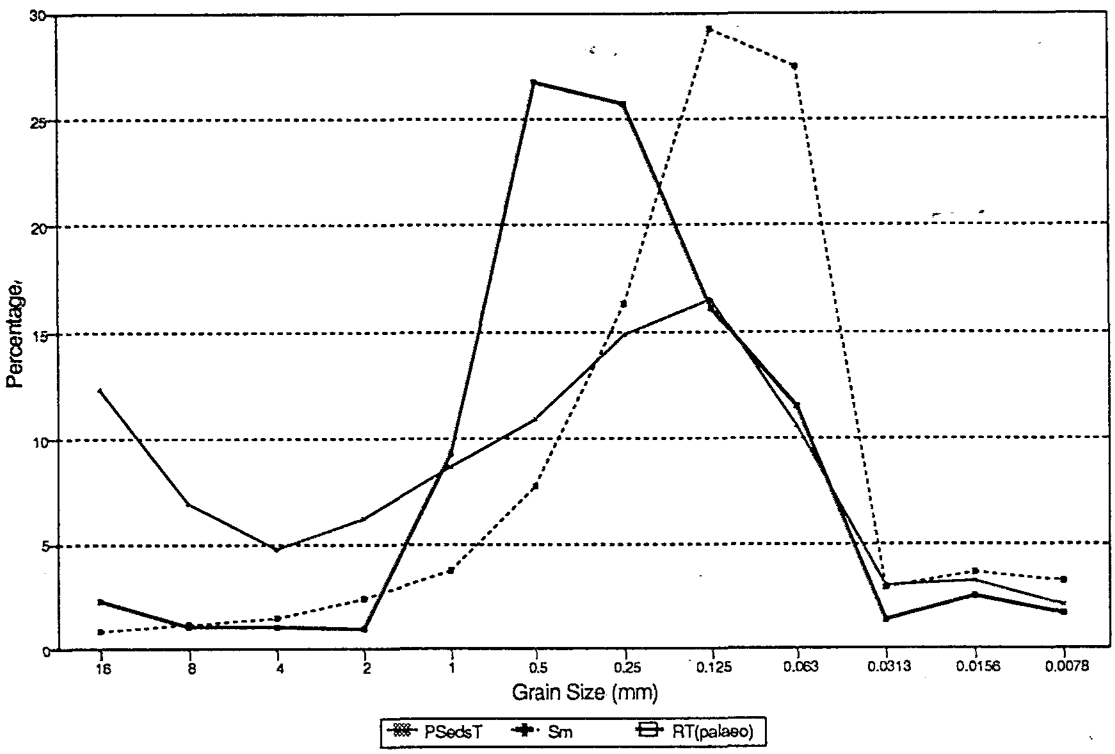

Figure 6.29. The mean grain size distribution for pool sediments (PSedsT), colluvial sands (Sm) and palaeo-river sediments [RT(palaeo)] at Spring Grove.

The unimodal sands and palaeo-river sediments have modal peaks at 0.125mm and 0.5mm respectively (Fig.6.29.). The palaeo-river sediments are better sorted than the sands. The former have a standard deviation of 9.7 while the latter has one of 10.25. Pool sediments collected from the river below the sedimentary sequence are better sorted than the above, with a standard deviation of 4.8. They are bimodal with a strongly developed peak in the fines at 0.125mm (Fig.6.29.). There is a subsidiary peak in the coarser sediments at 16mm. Slopewash sediment which supplies material to the stream had a unimodal distribution with a peak at 0.063mm. The slopewash sediments have a standard deviation of 7.73. They are therefore better sorted than the sands from which they are derived, but less well sorted than the pool sediments in which they end up.

If the sediments are classified in terms of gravel, sand, and silt+clay they have the following characteristics. There is an increase in the percentage of gravel upslope and a decrease in the percentage of silt+clay (Table 6.4.). The sands reach a peak at midslope

Table 6.4. The percentages of gravel (>2mm), sand (<2mm>0.063mm) and silt+clay (<0.063mm) for sediments at Spring Grove discussed in the text.

Sample	Gravel%	Sand%	Silt+Clay%
0m	8.78	75.96	15.26
50m	13.15	77.36	9.49
90m	77.28	19.20	3.52
Dcm	72.10	24.99	2.91
Gm	29.37	65.35	5.28
Sm	5.96	84.48	9.56
PRT	5.29	89.27	5.44
Pool Seds.	30.07	61.75	8.18
Slope Wash	0.0	82.02	17.98

and then decrease rapidly upslope (Table 6.4.). There is a decrease in gravel content from the diamicts (Gcm) in the sedimentary sequence to the sands (Sm). The sand and silt+clay contents increase upslope. The palaeo-river sediments have a lower gravel content than the hillslope sands (Table 6.4.), but a higher percentage of sand than any other sediment type within the sequence. The percentage of silt+clay for the palaeo-river sediments (5.44%) is in between that for the sands (Sm) (9.56%) and gravels (Gm) (5.28%). The pool sediments from the river at the base of the slope have a very similar grain size composition to that of the gravels (Gm) within the sedimentary sequence. The gravel and silt+clay percentages (gravel=30.07%, silt+clay=8.18%) are slightly greater than those for the gravels (gravel=29.37%, silt+clay=5.28%) within the sequence, while the percentage of sand is slightly less (pool sediments=61.75%, gravel [Gm]=65.35%).

6.4.3. Atteberg Limits

6.4.3.1. Burchleigh

The liquid limits range from a low of 13.5 at the base of the slope (0m) to a high of 39.0 at 150m. There is a general decrease in liquid limits from the top to the bottom of the slope (R^2=0.4)(Fig.6.30.). Although the plastic limits of the slope show a high correlation with the liquid limits (R^2=0.89) they have to be discarded as they are above the liquid limits, except those values corresponding to samples at 0m, 20m and 40m. By definition plastic limits must have a lower water content than the liquid limits. This matter is dealt with in more detail in the discussion.

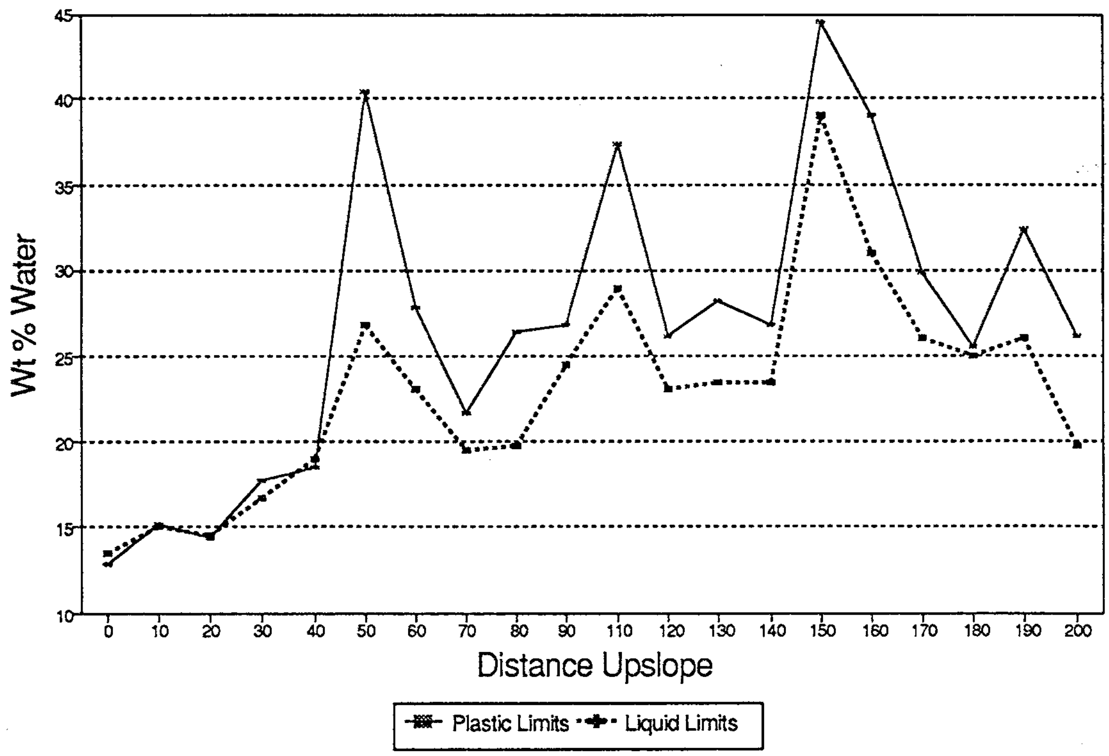

Figure 6.30. The Atteberg limits at 10m intervals upslope for Burchleigh.

6.4.3.2. Spring Grove

The liquid limits range from a low of 22.5 at 0m to a high of 38.5 at 60m. There is a very low correlation between the liquid (R^2=0.05) and plastic limits (R^2=0.06) and distance downslope. The liquid and plastic limits are highest on the mid slope between 20m and 70m upslope (Fig.6.31.). There is a very strong correlation between the plastic and liquid limits (R^2=0.93). Two anomalies occur at 20m and 60m where the plastic limits have a higher water content than the liquid limits.

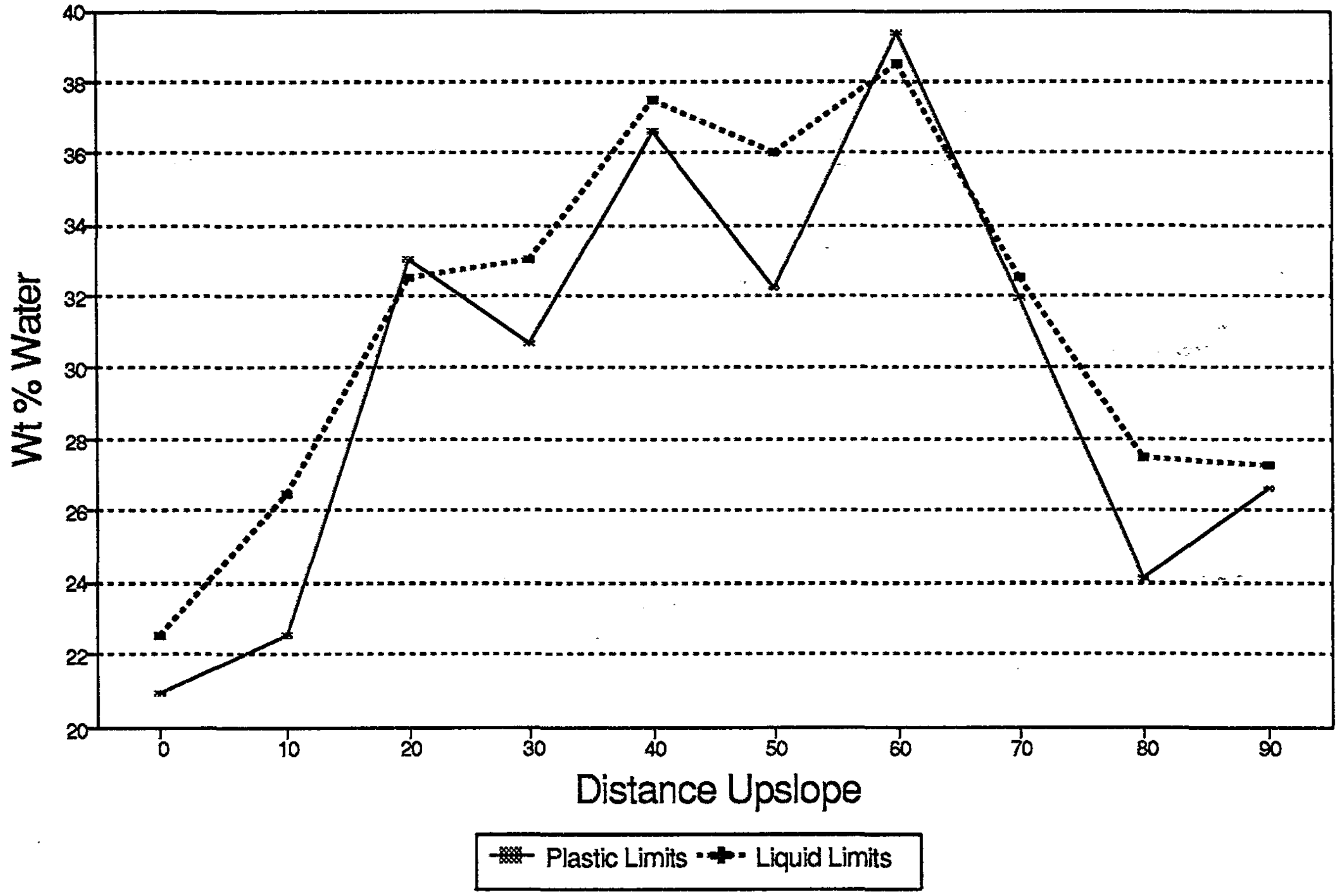

Figure 6.31. The Atteberg limits at 10m intervals upslope for Spring Grove.

CHAPTER 7

DISCUSSION

7.1. CLIMATE

7.1.1. Temperature

The Alexandria district is presently located within a humid mesothermal climatic zone which is affected by both tropical and polar air masses (Strahler, 1975). Mean annual air temperatures range from 16.4°C at Grahamstown to 19.7°C at Southey's Hoek. The temperature regime at the coast is more equitable than that inland. The mean monthly temperatures for most stations are greatest in February and lowest in July. The mean monthly temperatures range from a high of 24.15°C at Southey's Hoek in February to a low of 12.0°C at Grahamstown in June. At Rockhurst mean monthly minimum temperatures fall below 5°C in July. The highest daily maximum temperature in the region was recorded at Vaalkranz (46°C) and the lowest daily minimum temperature at Grahamstown (-8°C). There is no clear trend in terms of an increase or decrease in temperatures for the length of the records at each station.

If a drop of 5.5°C (Tyson, 1986) or 14.0°C (Lewis and Hanvey, 1993) did take place within the Last Glacial Maximum (LGM) (25 000-15 000 BP, Butzer 1984) as noted in the literature review, MAAT could have ranged from 10.9°C or 2.4°C at Grahamstown to 12.2°C or 5.7°C at Southey's Hoek. Estimated mean monthly temperatures in terms of a 14°C drop (most extreme scenario - Lewis, 1988) in temperature were also calculated to determine those months which potentially may have had mean monthly temperatures below 0°C. Three stations had at least one month with an estimated mean below freezing, namely Grahamstown-vet (3 months - June[-2.0°C], July [-1.95°C], August[-0.6°C]), Rockhurst (2 months - June[-1.2°C], July[-0.8°C]) and Midhurst (1 month - July[-0.35°C]).

Under the most extreme scenario, Grahamstown would have fallen into French's (1976) marginal periglacial domain (i.e. areas MAAT >-2°C but <+3°C). Within this domain frost action processes are present but do not dominate, although seasonally and diurnally frozen ground would occur. Karte (1983) suggested that seasonally and diurnally frozen ground may be characterised by such features as seasonal frost crack polygons, micro frost crack polygons, earth hummocks, thufur, "thermokarst", sorted circles and stripes and microforms of gelisolifluction. However, none of these features are likely to have formed on slopes as steep as those examined in this study (Washburn, 1973). Therefore the absence of identifiable periglacial features at Burchleigh and Spring Grove does not negate the possibility that the study sites may have occupied a marginal periglacial regime during the LGM or similar cold periods in the past. In particular when one considers that present temperatures have been known to drop as low as -8°C (Grahamstown, 1 November 1970) and that snowfalls have been recorded in Grahamstown on 6 occasions from 1883 to 1994. Frost action must have taken place during the LGM when temperatures were colder than at present. Diurnally frozen ground would at least have occurred during the winter months of the LGM, assuming a drop of at least 5.5°C in MAAT, although the depth of freezing is impossible to estimate due to lack of evidence.

7.1.2. Precipitation and Evaporation

Mean annual rainfall is greatest at Alexandria (913.3mm) and least at Langholm (518.5mm). Most rainfall occurs during the equinoxes, especially October. The driest months are January, June and July. The greatest rainfall for a 24 hour period occurred on 25 September 1952 at Alexandria when 290.6mm fell. High rainfall events are associated with cold fronts and principally cut-off lows (Tyson, 1986). Evaporation is lowest in winter and highest in summer. Rainfall exceeds evaporation by the most during October. As a result soil moisture conditions can be expected to be greatest during this month and hence it is the month in which slope failures are most likely to occur.

Using Caine's (1980) conservative estimate (slope failure occurs when rainfall >103mm in 24 hours), rainfall of sufficient intensity to cause slope failure can be expected on average once every ten years. Within the last 1000 years alone (during which climate has been approximately the same as at present according to Tyson, 1986) this amounts to over 100 potential slope failures in the region of the study sites.

Snow has fallen in Grahamstown and its immediate environs 6 times since the first record on the 9 August 1883. The heaviest fall occurred on 20 June 1976, which blanketed the whole Grahamstown region with a few centimetres of snow.

Evaporation is greatest during the summer months and least during the winter months. The driest month, when evaporation exceeds rainfall by the greatest amount is most often January, while the wettest month, when rainfall exceeds evaporation by the greatest amount, is generally October. Soil moisture is therefore expected to be greatest in October and least in January. Hillslopes are therefore most likely to fail in October. Intense or persistent rainfall is expected to proceed any slope failure, although in exceptional cases this may have been induced by snow melt. The latter is only likely to have occurred during the LGM.

If one considers that during the LGM the winter rainfall region would have expanded to include the area of the study sites (Tyson, 1986), that evaporation rates over the oceans were 20-30% less than today (Martin, 1981) and that a drop in temperature within the winter rainfall regions is accompanied by an increase in rainfall (Tyson, 1986), it could safely be concluded that soil moisture conditions were much greater during the LGM than at present. As a result one would expect slope runoff, slope failures and debris flows to have been more common than at present during the LGM. One would thus expect evidence for slope failure and debris flows to be a common feature of hillslopes in the Alexandria district where there is suitable hillslope sediment available. As will be seen later in section 7.3, which discusses the hillslope sedimentology at the study sites, this was found to be the case.

7.2. SEISMIC ACTIVITY

Seismic events within the time of historical records (1850-1994) occur at a rate of 10.5/100 years in the Alexandria district. The minimum Modified Mercalli Scale Intensity for slope failure (IV) has been exceeded 7 times (Table.6.2.). The largest earthquake recorded in the Alexandrian hinterland occurred on 21 May 1850. It had a Modified Mercalli Scale Intensity of V-VI, which was large enough to trigger lateral spreads and flows (Keefer, 1984). Although the seismic record is too short (approximately 150 years) to calculate recurrence intervals for given earthquake intensities in the Alexandria district a rough estimate can be obtained by using the diagrams in Fernandez and Guzman (1979b). The diagrams indicate that there is a 10% probability that an intensity of V will be exceeded within the next 100 years or a 10% probability that an earthquake of VI will be exceeded within the next 500 years.

The slope gradients at each study site are well in excess of those required for soil lateral spreads and flows, namely 0.3-10° (Keefer, 1984). If seismic events have occurred with the same frequency and intensity over the last 1000 years as those recorded in the last 150 years, which is a reasonable assumption given the tectonic characteristics of the region, then one can conclude that slope failure has been triggered by seismic events within the late Quaternary at Alexandria. Slope failures triggered by earthquakes have been recorded at Esikheleni in the Transkei within the last 10 years on gradients <20° (Beckedahl *et al.*, 1988). Esikeheleni is similar to the Alexandria region in terms of seismicity and geomorphology.

Given the estimated recurrence intervals for rainfall and seismic events great enough to initiate slope failure the probability of the two occurring together would be approximately 5 times every 1000 years near Alexandria. As this is a very conservative estimate, coarse clastic debris flow deposits resulting from combined rainfall/earthquake events, should be a prominent feature of many hillslope sedimentary sequences in the study region. This assumes that the hillslopes have coarse clastic sediment available for transport. Coarse clastic material is a common feature of the debris slope material mantling the hillslopes at both the study sites. This would appear to indicate that seismic activity (possibly in combination with appreciable rainfall) has contributed to sediment movement and deposition in the Alexandria district.

7.3. SEDIMENTOLOGY

7.3.1. Slope Gradient

The lower slopes at Burchleigh are inclined at 26°, the steepest midslope section is 53°, while the gradient at the crest is <10°. The gradient at the base of the slope at Spring Grove is 22°, the steepest midslope section is inclined at 30°, while the slope just below the crest slopes downwards at an angle of 15°. Sand has an angle of repose between 32° and 34°, while angular gravel on talus and similar slopes has an angle of repose up to 45° (Pye, 1994). Given the sandy nature of both hillslopes, they are both below the angle of repose, apart from the steep upper midslope regions, especially that at Burchleigh (53°). Gravitational processes would therefore appear to play a subsidiary role to fluvial transport processes for most of the length of both slopes.

The slopes at Burchleigh and Spring Grove both have an abundance of sediment and are therefore transport limited. Abrahams *et al.* (1985) have suggested that the presence of fines within hillslope sediments is indicative of a transport limited slope. Fines are abundant on the hillslopes at both study sites. If one takes all the above into account one would expect the hillslope sedimentology to be adjusted to the hydraulic processes active on the hillslope.

7.3.2. Hillslope Surface Sediment

7.3.2.1. Burchleigh

The sediments become finer downslope and, to midslope, better sorted. The degree of bimodality decreases downslope as the significance of the coarser mode decreases in relation to the finer mode. However, from the midslope downwards the sorting becomes progressively poorer, contrary to expectations (Abrahams *et al.*, 1988). This could be the result of the increasing frequency with which isolated larger clasts start to appear which have rolled downslope under the influence of gravity. The larger the clast, the further it will tend to roll downslope and hence the greater the influence on the calculation of standard deviation of a sediment distribution. Standard deviation was the measure used in this study to determine the degree of sorting within a sediment sample. The presence of rock outcrops on the midslope is not thought to have had a significant impact on the standard deviation trend. Any sediment derived from them would lead to an initial increase in the standard deviation whereafter it would be expected to decrease again downslope. If the sediment grain sizes are lumped together into the classes: gravel, sand, and silt+clay, the downward fining and increased sorting trends are much clearer. The percentage of gravel decreases while the percentage of sand and silt+clay increases downslope. This corroborates the view above that the increase in the standard deviation on the lower slopes is due to the presence of large clasts which have rolled downslope. The increased standard deviation on the lower hillslope is thus an artefact of the marriage between gravitational and fluvial hillslope processes.

The equation developed by Abrahams *et al.* (1988) and the Hjulstrom diagram were used in an attempt to quantify the nature of the overland flow regime which deposited the hillslope sediments. Abrahams et al (1988) used the diameter of the heaviest particle available for transport. In this study D_{84} (grain size for which 84% of grains are smaller) was used in order to exclude those large clasts which may have rolled downslope under the influence of gravity and that have not been transported by fluvial processes. However, only shear stresses for the sediment at the base of the slope (<40m) were calculated as all the other sediment samples higher upslope (>40m) had a D_{84}>16mm. This prevents one from determining downslope trends of any sort, because the 16mm sieve was the largest used in the grain size analyses and is thus the largest grain size recorded in this study.

The samples at 0m and 40m were exposed to shear stresses of $0.012 N.m^{-2}$ and $0.068 N.m^{-2}$ respectively, while those samples above 40m upslope would have been moved by shear stresses $>35.49 N.m^{-2}$. An estimate of the average flow velocity on the slope was obtained using the D_{50} and Hjulstrom's diagram (Reineck and Singh, 1980), which suggests that flow velocities ranged from $25 cm.s^{-1}$ at the base of the slope to $55 cm.s^{-1}$ at midslope (80m). Above this point (80m upslope) grain sizes (D_{50}) are all above 16mm in diameter, which indicates that overland flow velocities must have been $>150 cm.s^{-1}$. If one excludes the coarser clasts from the flow velocity estimate within each sediment sample one obtains a constant value of $25 cm.s^{-1}$ for the sediments of the finer mode in each grain size distribution. This is probably a better estimate of the overland flow velocity than the use of D_{50} as it targets those sediments almost certainly transported by overland flow. The clasts within the coarser fraction may have moved downslope through a process of soil creep.

7.3.2.2. Spring Grove

The hillslope sediments become finer and better sorted downslope. The standard deviation decreases downslope as the sediments become better sorted. The percentage of gravel decreases while the percentage of silt+clay increases downslope. The percentage of sand reaches a maximum at midslope.

The sediments are bimodal at the top and base of the slope and unimodal on the midslope (Fig.6.26.). The unimodal peak for the midlsope sediments is located at 0.5mm. The fines mode for the sediments at the top and bottom of the slope is located at 0.063mm. The percentage weight of the 0.063mm fraction thus increases downslope. This trend differs from that at Burchleigh where the sediments at the base of the slope had a unimodal grain size distribution, while those on the mid- and upper slopes were bimodal. The size of the finer mode is also larger at Burchleigh, 0.125mm as opposed to 0.063mm at Spring Grove. However, unlike the bimodal distributions at Burchleigh which have the coarser mode >16mm, the coarse mode at Spring Grove at the base of the slope is located at the same place as the unimodal peak at midslope, namely 0.5mm. This is interpreted as a decrease downslope in the significance of the transport mechanism responsible for the deposition of the coarse mode in relation to that of the fine mode at Spring Grove.

The smaller grain size of the fine mode at Spring Grove in relation to that at Burchleigh and the lack of a significant coarse mode >16mm appears to be a reflection of the weathering characteristics and fine grained nature of the sandy shale bedrock. The shales weather easier than the quartzites at Burchleigh. The clasts on the hillslope at Burchleigh are blocky whereas those at Spring Grove are generally flat and platy. This would make it more difficult for them to roll downslope under the influence of gravity than similar size clasts at Burchleigh. As a result there is a paucity of large (>16mm) clasts on the slopes at the base of the Spring Grove hillslope. This situation is amplified by the fact that the quartzites are more resistant to weathering than their shale counterparts and would therefore tend to remain intact on the surface of the hillslope for longer periods than the shales.

An attempt was made to quantify the overland flow regime at Spring Grove using the same methods as applied at Burchleigh. Estimated minimum shear stresses range from $0.14N.m^{-2}$ at the base of the slope (0m) to $0.31N.m^{-2}$ at midslope (50m) and $>35.49N.m^{-2}$ at the top of the slope (90m). Average flow velocities range from $25cm.s^{-1}$ at the base of the slope (0m) and $30cm.s^{-1}$ at midslope (50m) to $>150cm.s^{-1}$ at the top of the slope (90m). The fine mode (0.063mm) has an estimated flow velocity of $25cm.s^{-1}$ and the coarse mode (0.5mm) a flow velocity of $30cm.s^{-1}$.

Summary of Hillslope Surface Sediment Trends:

At Burchleigh the hillslope sediments become finer and better sorted in the downslope direction of transport. The progressive weakening of this relationship with distance below the midslope was ascribed to the tendency of the larger clasts rolling downhill to travel greater distances than smaller clasts on to the lower slopes. However the degree of bimodality decreases downslope as expected. These apparently contradictory features of the hillslope appear to be due to the resistant nature of the quartzite clasts. A coarse mode although it decreases in prominence remains an important feature of the grainsize distribution downslope, while the degree of sorting within the fine mode increases, but at a lesser rate than the decrease in prominence of the coarse mode. This increase in disparity between the coarse and fine modes may explain the increase in standard deviation below the midslope.

The sediments at Spring Grove become finer and better sorted downslope. The sediments at the top and bottom of the slope are bimodal, while those of the midslope are unimodal. However, the distribution at the base of the slope is very weakly bimodal. Unlike Burchleigh the coarse mode at the base of the slope is considerably smaller than that at the top of the slope, namely 0.5mm at the base of the slope and >16mm at the top. Furthermore this value (0.5mm) is the same as that for the unimodal peak. This enigmatic feature of the hillslope sediments could possibly represent reworking of the sediment deposited by an earlier flow.

Burchleigh, located on weathering resistant quartzites, has a coarser fine mode than the sediments at Spring Grove, located on more easily weathered sandy shales. Thus the grain size distribution of the hillslope sediments appear to reflect the bedrock lithology. At Burchleigh there is no correlation between distance downslope and slope gradient (R^2=0.01) and only a weak relationship with the standard deviation (sorting) of the sediments (R^2=0.36). There is also no relationship between the slope gradient and standard deviation (sorting) (R^2=0.01). However the weak relationship between the standard deviation and distance downslope is deceptive as there is a very high correlation between the percentage of sand within the sediments and distance downslope (R^2=0.95). The hillslope at Burchleigh therefore appears well adapted to the downslope fluvial transport processes.

At Spring Grove there is a weak correlation between distance downslope and slope gradient (R^2=0.4), but a strong correlation with standard deviation (R^2=0.72). This adjustment of the sediments to the fluvial processes on the hillslope is corroborated by the fairly high correlation between the distance downslope and the percentage of sand within the sediments (R^2=0.67). However the strongest correlation is found between the slope gradient and the standard deviation (R^2=0.9), although the correlation is even higher for the percentage of sand within the sediments (R^2=0.93). Slope gradient at Spring Grove is therefore of greater importance than distance downslope in terms of the fluvial processes active on a hillslope and their ability to entrain and deposit sediment. Bed slope is known to have a significant impact on grain entrainment (Pye, 1994). The greater the slope the easier a grain is entrained all else equal.

Abrahams *et al.* (1985) found that "...slopes developed in widely jointed and mechanically strong rocks are poorly adjusted to contemporary processes, weathering-limited, and characterised by little or no relationship between S [slope gradient] and D [grain size]." On the other hand they suggested that "...debris slopes underlain by closely jointed and/or mechanically weak rocks...are well adjusted to extant processes, transport-limited, and exhibit a strong relationship between S [slope gradient] and D [grain size]." The very weak relationship between the percentage of sand within the sediments at Burchleigh and the slope gradient (R^2=0.0007) and the strong correlation value for the same relationship at Spring Grove (R^2=0.93) therefore appears to indicate that the sediments at Spring Grove are much better adjusted to the hillslope transport processes than those at Burchleigh. This explains the higher correlation between the degree of sorting at Spring Grove with distance downslope (R^2=0.72) than that at Burchleigh (R^2=0.36).

The liquid limits for the slopes at Burchleigh and Spring Grove are lowest at the base of the slope and highest on the upper midlsope. Less rainfall should therefore be required for the sediments at the base of the slopes to deform than higher up. However, the increased slope failure potential at the base of the slope is mitigated by the lower slope gradient in relation to that further upslope. The significance of the Atteberg limits in terms of slope failure potential at Burchleigh and Spring Grove is thus unclear.

In conclusion, therefore, sedimentological analyses indicate that sediments at Burchleigh and Spring Grove were deposited as a result of both gravitational (rolling) and overland flow processes, and possibly soil creep. All these processes could occur under present climatic conditions.

7.3.3. Hillslope Stratigraphy

7.3.3.1. Burchleigh

The fine grained colluvium within the sedimentary sequence is here interpreted as material transported downslope by overland flow and rainsplash. Some downward translocation of sediment has also probably taken place. This together with weathering would have skewed the lower horizons towards the fines. However these two processes (i.e. weathering and downward translocation) are not thought to have been significant as no palaeosol horizons were evident within the sedimentary sequence. This applies equally well to the site at Spring Grove. The large clasts which sometimes form stone lines within the sedimentary sequence are interpreted as lag deposits, marking a hiatus in hillslope deposition. The clasts appear to have rolled downslope after breaking off one of the rock outcrops higher up, coming to rest with one of their flatter sides parallel to the hillslope surface.

The highly lenticular gravels and diamicts are interpreted as the remnants of mass flow deposits, in this case debris flows, on the basis of their coarse clastic nature, their tendency to form convex upper surfaces, their laterally extensive nature and the presence of non-erosive bases which form sharp contacts with the underlying sediments. Gloppen and Steel (1981) have shown that the relationship between bed thickness (BTh) and maximum particle size (MPS) may indicate whether or not sediments were deposited as debris flows. A regression line plotted for this relationship, using data from Burchleigh, has an R^2 value of 0.61. This indicates that there is a moderate positive correlation between BTh and MPS and provides support for the debris flow interpretation above. Gloppen and Steel (1981) also found that debris flows have a BTh/MPS ratio less than or equal to 3. The 5 beds interpreted as debris flows fall within this range apart from 2, each of which may be a composite of more than one debris flow. This is particularly true of the thickest "debris flow" bed measured (i.e. BTh=2.140m, MPS=0.189m) where indistinct stratification is evident (Fig.7.1.).

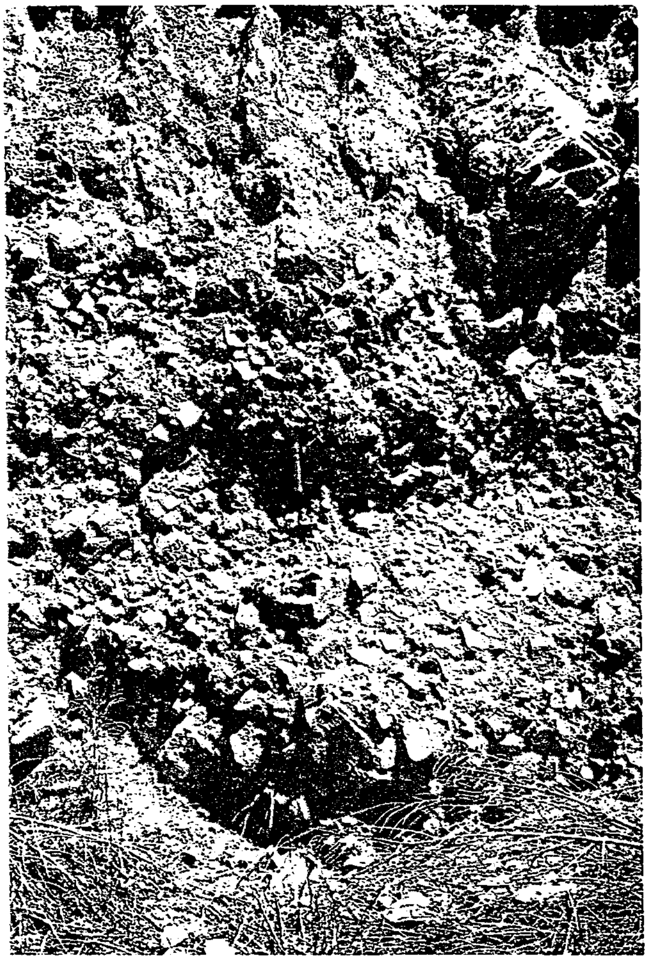

Figure 7.1. A diamict at Burchleigh which appears to be a composite of a number of three debris flows.

Only one deposit interpreted as a debris flow appears to occupy a channel. This may have been a former erosion gully as the bed has a convex upper surface where it has "overflowed" the "channel" (Fig.6.15.). This bed lies just below the current hillslope surface. Gully erosion is a relatively common feature of the Alexandria region, at present.

According to Shultz's (1984) classification scheme the debris flows represented in the sedimentary sequence are primarily of Type II and III. Type II, which are clast-rich debris flows, have high clast concentrations with inverse grading sometimes well-developed. Some of the debris flows within the sequence are both clast supported and inversely graded. Type III, pseudoplastic debris flows, cannot support large clasts as well as Type II flows although large clasts may sometimes be suspended during high velocity flow stages. These clasts are deposited or moved as bedload on flow deceleration. The "channelised debris flow" (Fig.6.14.) described above may be an example of a Type III flow as it has large clasts at the top of the flow and as bedload with finer sediment in between.

Scott (1971) found that mudflows had a bimodal grain size distribution and debris flows a weakly bimodal or unimodal distribution. Pierson (1980) also found debris flows to be unimodal. The diamicts within the sedimentary sequence thus agree well with this data if the debris flow interpretation is correct. The highly lenticular gravel deposits which have also been identified as debris flows, which have bimodal characteristics, may therefore be mudflows, which are fine grained variants of debris flows. The percentage of gravel (83.45%), sand (13.15%), and silt+clay (3.40%) for the diamicts (Dcm) which are interpreted as debris flows is in broad agreement with that recorded by Pierson (1980) in an active debris flow during surges (i.e. gravel=70%, sand=20% and silt+clay=10%).

An attempt has not been made in this study to determine the competence of these debris flows using the largest clast rafted by the flow as an indicator, as the error involved in making the requisite assumptions would be too great. The calculations assume that the largest clast matches flow competence. This is only true if there is a range of clast sizes equal to and greater than the clast size matching flow competence available. There is no guarantee that this was the case at the study site.

Two diamictic beds at the base of the sequence appear to be fluvial and not debris flow deposits. The lowermost diamict occurs in close association with an imbricated gravel and fine grained sands. This bed exhibits features characteristic of a point-bar sequence: a coarse channel lag, with imbricated gravels and sands on the inside of the meander bend. The diamict just above this bed, but slightly "downstream" is interpreted as a channel lag deposit. These sediments therefore appear to be remnants of the former position of the present river channel. As the stream cut into the bank opposite the hillslope dissected by the road cutting (located on the inside of the stream bend) and eroded downwards, the channel came to "occupy" successively lower positions within the sedimentary sequence.

The river sediments which occupy the present day stream channel have a unimodal grain size distribution, unlike the well developed bimodality in the gravels and sands of the sedimentary sequence. This unimodal nature also differs from that of the sediments mantling the hillslopes, apart from the sediments near its base which are also unimodal. The river sediments are 77.23% gravel, 21.90% sand and 0.87% silt+clay. The coarse clastic nature of the sediments and the very low silt+clay content appear to indicate that the sediments were deposited by a high flow velocity. The low silt+clay content could also be an indication of the bedrock material as these sediments at Burchleigh form a much smaller fraction (0.87%) in the river gravels than in the pool sediments (8.18%) and palaeo-river gravels (5.44%) at Spring Grove. The silt+clay content on the hillslopes is also lower on average at Burchleigh (3.8%) than at Spring Grove (9.42%). The bedrock at Burchleigh is quartzite, whereas the bedrocks at Spring Grove are sandy shales.

An attempt was made to obtain an estimate of the palaeovelocity of the palaeo-river and the present river channel sediments by using the Thompson-Campbell equation and Hjulstrom's diagram. In each case the average grain size (D_{50}) was taken as an indication of "average flow velocities". According to the former the palaeo-river gravels were associated with a flow velocity of $10.31 m.s^{-1}$ and the present stream sediments with a flow velocity of $8.44 m.s^{-1}$. These figures appear very high as the stream only flows episodically. If the above flow velocities are correct to within an order of magnitude it would still appear to indicate that the sediments are only moved during flash floods. The Hjuslstroms diagram gives much lower and more realistic flow velocities of $95 cm.s^{-1}$ for the stream sediments and $30 cm.s^{-1}$ for the palaeo-river gravels.

The fine grained sands (Sm) which are interpreted as sediment deposited by overland flow would have been exposed to a minimum shear stress of $1.25 N.m^{-2}$ and an average flow velocity of
$25 cm.s^{-1}$.

7.3.3.2. Spring Grove

The sedimentary sequence is characterised by alternating layers of fine grained sands (Sm) and coarser clastic deposits (Fig.6.16). In places these beds are of equal thickness, but on most occasions the sands form the thicker beds. The fine grained sands are interpreted as colluvial sediments deposited by overland flow and rainsplash (Mucher and de Ploey, 1977; Mucher *et al.*, 1981). The stone lines within the sequence therefore appear to mark a hiatus in deposition when erosion was dominant on a slope, with the clasts remaining as a lag deposit.

At the base of the sedimentary sequence is a clast-supported diamict comprising large well-rounded quartzite clasts (Fig.6.22.). These clasts are larger, rounder and composed of a different lithology to the diamicts and gravels higher up (Fig.show comparative difference). The lithology of the clasts which compose this basal diamict are composed almost entirely of quartzite as opposed to the quartzite and sandy shale mixture of the other diamicts. This basal diamict is interpreted as a palaeo-river channel lag deposit. This view is corroborated by the presence of percussion marks on some of the clasts (Fig.6.23.). Percussion marks have been attributed to river deposits (Gale and Hoare, 1991).

The coarse clastic diamicts (Dcm) and gravels (Gm) within the sequence have been identified as debris flow deposits. They display typical debris flow features such as highly lenticular bed geometry (Nemec and Steel, 1984), poor sorting (Crozier, 1986) and sharp contacts with the underlying sediments. All 4 of the debris flow beds apart from one have a BTh/MPS ratio <3, which is within the realm of debris flow deposits as suggested by Gloppen and Steel (1981). The reason for the anomalous bed's large BTh/MPS ratio (8.05) is problematic, but the bed may be a composite of more than one debris flow. Alternatively it may have been deposited by a different process, such as a sheetflood after heavy rains or a slushflow after snow melts (Elder and Kattelmann, 1993). The beds have a very low R^2 correlation value for BTh/MPS, namely 0.29. This is probably due to the inclusion of the enigmatic bed discussed above in the very small data set.

The diamicts and gravels are both weakly bimodal, which is typical of debris flows (Scott, 1971). The diamicts have a strongly developed primary mode >16mm and a subsidiary mode at 0.125mm. The primary mode for the gravels is located at the same grain size (0.125mm) as the subsidiary mode for the diamicts. The coarse, subsidiary mode for these gravels is 8mm. The higher standard deviation of the diamicts (standard deviation=9.18) indicates that they are more poorly sorted than the gravels (standard deviation=7.03). The diamicts and gravels at Spring Grove all appear to have been clast-rich or pseudoplastic debris flows (Shultz, 1984).

In contrast to the weak bimodality of the diamicts and gravels, the sands (Sm) and palaeo-river sediments are unimodal with peaks at 0.125mm and 0.5mm respectively (Fig.6.28.). The palaeo-river sediments (standard deviation=9.7) are better sorted than the sands (standard deviation=10.25), but less well sorted than the pool sediments (standard deviation=4.8) from the present river channel. The pool sediments are bimodal with a strongly developed primary mode at 0.125mm and a subsidiary mode >16mm. The sediment washed from the hillslope and cliff into the present stream channel has a unimodal grain size distribution with a primary mode of 0.063mm. It is better sorted (standard deviation=7.73) than the hillslope sands (Sm) from which it originates but less well sorted than the river sediments in which it is deposited.

The diamicts have the highest gravel content and the sands the lowest of the hillslope sedimentary sequence (Table.6.4.). The diamicts also have the lowest sand and silt+clay contents, while the sands have the highest (Table.6.4). The palaeo-river sediments at the base of the exposure have a higher sand content than any of the bed types above, a lower gravel content and an intermediate silt+clay content. The pool sediments within the present river channel have a gravel, sand and silt+clay content very similar to that of the hillslope gravels (Gm), although they are deposited by fluvial action and are not a mass flow. It is unlikely the gravels themselves are river sediments as they do not occupy channels, are highly lenticular, do not display imbrication of any sort and are comprised of angular clasts.

The palaeoflow conditions associated with the deposition of the fluvially deposited sediments were estimated using the techniques outlined in the section on Burchleigh. The average palaeo-flow velocity of the palaeo-river sediments amounted to 27.5cm.s^{-1}, while the sediments within the present channel were deposited by a flow velocity of $30cm.s^{-1}$. The sands (Sm) within the sedimentary sequence were deposited by an estimated flow velocity of $25cm.s.^{-1}$ and exposed to shear stress of $0.035N.m^{-2}$ This is a lower shear stress than the estimates for the current basal hillslope sediments, although the palaeo-flow velocity estimate of $25cm.s^{-2}$ is within the range of that estimated for the sediments at the base of the slope.

<u>Summary of Hillslope Architecture at the Two Study Sites:</u>

The stratigraphy at both study sites is dominated by fine grained colluvium, with subordinate highly lenticular diamicts. The colluvium is thought to represent sediment transport and deposition by rainsplash and overland flow. Stonelines occupy a number of horizons within the sedimentary sequences, with a boulder or two present in exceptional cases. The diamicts have been interpreted as debris flow deposits. The diamictic beds represent tangential sections of the terminal lobes of the debris flows. At the base of each section are fluvial deposits, which have been interpreted as channel lag deposits and river sediments. All these processes can occur under present climatic conditions.

Although temperatures during the LGM were low enough, no unequivocal evidence was found for a geomorphic process restricted to frost or ice action.

7.4. ATTEBERG LIMITS

7.4.1. Burchleigh

Only the liquid limits are considered for Burchleigh as the sandy nature of the sediments prevented an accurate assessment of the plastic limits. The Liquid limits ranged from a low of 13.5% at the base of the slope to 39% at 150m, which is the transition between the midslope and upper slope. The general decrease in the liquid limits from the top to the bottom of the slope could indicate that the lower slopes are more susceptible to slope failure than the upper slopes. Less rainfall would be required for the sediments at the base of the slope to reach their liquid limits than for the sediments near the top of the slope.

Most of the sediments are thus cohesionless (liquid limit<20%) or of low plasticity (Harris, 1987). The liquid limits for this sediment therefore fall within the range of the "poorly sorted diamicts resulting from viscous flow (gelifluction)" that, in fossil form, Lewis and Hanvey (1991) have studied in the Drakensberg. Harris (1987) has interpreted such liquid limits as diagnostic of gelifluction deposits and has used them to indicate fossil gelifluction deposits in Britain. Gallop (1991, reported in Ballantyne and Harris, 1994) has shown experimentally that gelifluction is associated with similar liquid limits to those at Burchleigh. Liquid limits thus suggest that at least some of the deposits at Burchleigh may be of gelifluctional origin, although their is no other sedimentary evidence to support this.

7.4.2. Spring Grove

The plasticity limits at Spring Grove are excluded here for the same reason the plasticity limits were excluded from the discussion on the Atteberg limits at Burchleigh. Again the plastic limits seem inordinately high. The high correlation between the trends for the plastic and liquid limits at both research sites is taken as an indication of consistent bias in the methods adopted by the researcher. The subjectivity in determining the water content for "rollability" of the sediment must have been incorrectly applied.

The liquid limits at Spring Grove range from a low of 22.5% at the base of the slope to a high of 38.5% at 60m on the midslope. The sediments at the base and top of the slope therefore require the least rainfall to reach the liquid limit and begin movement downslope. The liquid limits therefore indicate sediments of low to medium plasticity according to Harris (1987). These sediments fall within the range of sediments recorded by Lewis and Hanvey (1991) which were associated with gelifluction and periglacial shearing in clay soils.

There is no correlation between the liquid limits and gravel (R^2=0.01), sand (R^2=0.04) or the silt+clay (R^2=0.11) fractions at Spring Grove or for the silt+clay fraction at Burchleigh (R^2=0.02). However, there is a weak correlation at Burchleigh between the gravel (R^2=0.48) and sand (R^2=0.50) fractions with the liquid limits of sediment samples. These relationships are problematic as it was expected that a decrease in grain size and hence pore space would have lead to a decrease in the liquid limits. The silt and clay fractions were lumped together for the calculations. This may have masked a strong correlation between the clay content and the liquid limits as clays are known to have important geotechnical properties (Crozier, 1986). As with the sediments at Burchleigh, the liquid limits indicate that some of the sediments at Spring Grove may have been emplaced by gelifluction, although other evidence is lacking.

CHAPTER 8

CONCLUSION

The study sites at Burchleigh and Spring Grove have been influenced by a similar suite of geomorphic processes. These processes were deduced from the sedimentary deposits present within each hillslope section. The following conclusions can be made in this regard:

* The fine grained colluvial sediments represent deposition by overland flow/sheetwash.

* The highly lenticular diamicts composed of angular clasts probably represent the terminal lobes of debris flows.

* The diamicts and gravels at the base of both sedimentary sequences are channel lag deposits.

* Most slope failures which precede debris flows were probably triggered by intense or extended periods of rainfall associated with the incursion of sub-polar air masses, cold fronts. Melting snow also has the potential to trigger slope failure in the Alexandria region, but it is an extremely unlikely scenario.

* Seismic events may have triggered slope failure either with or without the hillslopes sediments being saturated.

* The sediments are probably of Holocene age. The paucity of carbonaceous material and absence of stone artefacts suitable for dating purposes prevented the age of the hillslope sediments from being determined accurately. However, the sediments occupy the flanks of valleys which have been cut into the Nanaga Formation palaeodunes, which are of Pliocene to early Pleistocene age (Lewis, 1995). The Nanaga Formation palaeodunes were deposited on top of a series of marine erosion surfaces cut between the early Cretaceous and the end of early Miocene age (Partridge and Maud, 1987).
  The sedimentological, climatic and seismological evidence collected for this thesis indicates that all the processes identified within the sedimentary sequence could have been active under present conditions. Given the lack of any evidence for sedimentary

processes active during a colder thermal regime, the sediments presumably post date the LGM.

The results of this study appear to be in agreement with existing research on debris slopes in southern Africa. Botha *et al.* (1990) studied slope sediments in northern Natal, which is a subtropical, summer rainfall region, while Butzer studied hillslope sediments within the winter rainfall regions of the western Cape which have a Mediterranean climate. Botha *et al.* (1990) suggested that the bulk of the sediments that they investigated in northern Natal were deposited by sheetwash with mass movement processes playing a subsidiary role. Butzer (1979) identified grezes litees on 15° footslopes associated with several mid-Pleistocene cold intervals at Elandsbaai, western Cape. Butzer (1979) attributed the periodic mobilization of the gravels to frost or water-assisted gravity movements or colluvial deposition.

The sediments investigated in this study therefore appear to provide a link in a broad continuum from the sheetwash dominated deposits within the present summer rainfall regions of Natal to those in the winter rainfall regions of the western Cape where mass movement processes are important. The sediments at Alexandria indicate an almost equal importance in the processes of overland flow/sheetwash and debris flow.

The objective of this study was therefore achieved successfully, namely to identify the surface processes which resulted in the deposition of two unconsolidated late Quaternary deposits near Alexandria, eastern Cape. Potential trigger mechanisms in terms of seismic events and extreme meterological conditions were identified. However the inability to obtain accurate dates for the hillslope sediments prevented one from relating the hillslope sediments to known past climate change with any degree of accuracy.

The work completed for this thesis has filled a gap in the knowledge of the hillslopes within the eastern Cape coast and its immediate hinterland. No comprehensive study of this nature had been done before in the region. Most work of this nature has been carried out in northern Natal and the western Cape, with a few studies in the Orange Free State and the Transvaal. The eastern Cape falls within a climatic transition zone between the subtropical summer rainfall regime of the eastern seaboard and the Mediterranean climate of the western Cape. As such it can be expected to provide a sensitive record of climatic changes. Although this was not immediately apparent within this study, further research will no doubt reveal such changes in climate.

The approach used in this thesis was unusual in a southern African context. It represents to the authors knowledge the first time an attempt has been made to identify *and* quantify the trigger mechanisms and palaeoslope processes active on the hillslopes. Seismic events have received little attention as a geomorphic process in southern Africa. This study suggests they should be investigated more carefully. It is also the first time the snow records for the region have been collated and recorded.

This study makes a small, but important contribution to the knowledge of hillslope processes and the internal architecture of the hillslopes. Overland flow/sheetwash and debris flows have therefore been the dominant processes on the hillslopes within the Alexandria region during the Holocene.

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