CHAPTER I: INTRODUCTION

1.1 THEORIES OF HILLSLOPE DEVELOPMENT INVOLVING A FORM-PROCESS RELATIONSHIP

Studies of hillslopes[1] and their evolution have long occupied a central position in geomorphology. This must largely be a consequence of the fact that much of the landsurface of the earth is composed of slopes (King, 1966), so that as Scheidegger has pointed out, "If the development of individual slopes is understood, the development of a landscape can be synthesised" (Scheidegger, 1961, p. 1). The amount of effort concentrated on the study of slopes is reflected by the volume of its literature; recent monographs on the subject include those of Brunsden (Ed.,1971), Macar (Ed., 1964, Ed., 1970), Birot et al. (1964), Everard et al. (1964), Selby (1970), Carson & Kirkby (1972) and Young (1971).

However, the development of hillslopes (especially in terms of the geomorphic processes which produce a particular slope

1. Hillslopes may be defined, following Schumm (1966, p.98) as the surfaces between a drainage divide and a valley floor.

form) is still not understood, and a great deal of agrument has been generated among geomorphologists regarding the methodology of slope studies (Chorley, 1964).

In general, two broad approaches to the study of slope development have been recognised (Young, 1963; Scheidegger, 1964). The first may be termed the "deductive" approach. In this type of study, assumptions regarding the mode of operation of surface processing are made, and predictions of resultant slope forms and development formulated on the basis of these assumptions. The validity of the deductions is then tested by comparing predicted slope forms with actual slopes. The second approach, which may be termed "inductive", proceeds in the opposite direction: field observations and measurements are first made, and these are subsequently reviewed for any significant relationships or suggestions as to the mode of slope development which may become apparent. It may appear that purely inductive studies of slope development would be difficult to carry out, since some hypothesis must be present to guide the collection of field data; however, observations may be made merely with the hope that useful results will emerge (Young, 1971, p.20), and induction is therefore a viable approach in slope studies.

The present failure to understand slope development appears to the writer to be partly a consequence of the dominance of the deductive approach in slope studies. As Leopold, Wolman, & Miller (1964, pp. 500-503), Ahnert (1970), Wilson (1968), Chorley (1964), and others, have noted, similar slope forms could conceivably be produced in a vast number of different situations, so that agreement between deductions based on "reasonable assumptions" and actual slope forms in no way validates the deductions or the assumptions on which they are based. As Chorley (1964, p. 71) has noted, this fact has allowed a large number of hypotheses to be advanced to explain identical aspects of slope development, such as the formation of convexities or concavities. Each hypothesis may be consistent with actual slope forms, and because of this fact one is left uncertain as to which (if any) is the correct explanation, and further, almost all of the arguments presented are circular or tautologous. Ahnert (1970) has suggested that further confirmation of such deductive slope models could be obtained by incorporating in them predictions of regolith thickness as well as of slope inclination and form. Then, he suggests, a comparison of the actual regolith thickness on slopes in the field with that on the model slopes could provide a valuable additional criterion for assessing the model's reliability, and he has proceeded to develop such models. However, this approach may be as specious as the simple comparison of predicted and actual slope forms, since soil properties are in some cases correlated with slope inclinations (Acton, 1965; Swan, 1970), although in other cases they may be independent (Furley, 1968). Thus, if the model predicts slope form correctly, then it may also predict soil properties correctly, whether it is strictly valid or not. This is essentially a problem of multi-collinearity (Poole & O'Farrell, 1971). For additional confirmation of slope models, the parameters predicted by the model must clearly be entirely independent of each other, and also of slope inclination. Such parameters would be extremely hard to find.

Examples of difficulties created by the use of deductive slope models are common. For example, King (1967) has explained the formation of the upper slope convexity as follows -

"With increasing distance from the divide, the quantity of material requiring to be transported increases in direct proportion if the bedrock weathers at a uniform rate. Hence for disposal of the waste by creep the slope must steepen progressively, ie., the slope must become convex."
(King, 1967, p. 145).

In addition to being tautologous, this argument assumes -

a) that material cannot accumulate on the slope;
b) that creep is the process which moves material on the slope;
c) that volumetric creep rate increases with slope angle; and

d) that the bedrock weathers at a constant rate over the slope.

King presents no evidence to support any of these assumptions; nor does he present any actual measurements of either slope form or process, and the suggested explanation of the formation of the convexity remains unconvincing.

Fenneman (1908) offered an alternative explanation of the same feature. He proposed that convexities were produced by "unconcentrated wash", whose erosive power increased downslope as the volume of runoff increased. "The effect of increasing power in this case is increasing slope". (Fenneman, 1908, p. 752). This argument is unconvincing for several reasons: Fenneman presents no evidence to show that "unconcentrated wash" is the dominant process on slope convexities, or that it gives way to any other process where the convex portion of the slope ends; nor does he explain why the convexity does not extend continuously down to the stream at the foot of the slope. His argument therefore remains doubtful; it is entirely lacking in supporting evidence.

Gilbert (1909) proposed yet another explanation of the formation of convexities. He suggests that "On the upper slopes, where water currents are weak, soil creep dominates and the profiles are convex. On lower slopes water flow dominates and profiles are concave." (Gilbert, 1909, pp. 346-347). Gilbert's argument will not be presented in detail here; however, it is subject to the same criticisms as those of Fenneman (1908) and King (1967): it is entirely speculative, and no measurements either of actual slope forms or of the processes operating on them are presented; it is circular; assumptions such as equal rates of weathering over the entire slope, a uniform rate of ground lowering over the whole slope, and absence of change in regolith particle size downslope are made without explanation or justification. Again, therefore, one cannot rely on Gilbert's conclusions.[1] According to Young (1963), five different explanations have been given to account for the formation of slope convexities, and clearly only one can apply in any particular case.

Similarly varied explanations of the development of other slope forms, such as concave and rectilinear sections, have been presented; these have been summarised by Schumm (1966, pp.102-103).

A number of reasons for the dominance of purely theoretical or deductive hypotheses of slope evolution, such as those

1. His ideas are still quoted however, apparently with approval, in many texts (for example, Carson & Kirkby, 1972,306-7).

discussed above (noted, for example by Carson, 1969, and Young, 1963), may be suggested.

Ahnert (1970) has pointed out that quantitative discussions of slope development would require measurements of the rates of weathering and of the downslope transportation of material as functions of other slope properties, principally slope inclination; despite the increasing number of such studies, and the improvements of measuring techniques, direct measurements of slope processes still contain large error terms, and this limits their usefulness as a basis for the theoretical modelling of slope evolution. According toAhnert (1970), an important component of this error term is a consequence of the fact that the essentially continuous slope processes of weathering and downslope transport are too slow to be accurately measured in a limited period of observation, while the rapid processes (such as landslides) occur too rarely or discontinuously to permit accurate determination of their rates over the long periods of time involved in slope development.

A number of methods have been used in attempts to overcome the problem of the time involved in process measurements, including -

a) making measurements on slopes which are evolving very rapidly, such as badlands (for example, see Schumm,

1956) and attempting to infer conclusions about processes operating on morphologically similar but larger landforms;

b) Assuming that within a single landscape are present slopes of differing relative age which can be arranged in a time sequence, and hence their manner of evolution determined (for example, see Savigear, 1952, and Carter & Chorley, 1961);

c) the use of mathematical models (for example, see Bakker & Le Heux 1946, 1947, 1950); and

d) the theoretical assumption of time sequences, based on limited considerations of process, as developed by W.M. Davis and W. Penck.

In addition to the problem of the time required for process studies, two further difficulties in studies of the relationship of surface form and geomorphic process have been noted by Arnett (1971). Firstly, the polygenetic nature of landforms implies that slope form is not necessarily the result of contemporary processes, hence making causal relationships between the two difficult to establish by measuring the operation of current slope processes; second, relations between slope form and slope process are difficult to establish because of the complexity of the relationship - for example, on any given slope, several processes may be operating, and each may be affected differently by differences in vegetation, exposure, moisture conditions, and so on, along the slope.

An additional complication in defining the relationships between form and process is to establish the degree and nature of feedback between the two. Chorley (1964) explains this as follows -

> "Intuition prompts the geomorphologist to believe that slope forms should be capable of understanding on the basis of the processes producing them through time, yet any detailed study of an individual slope problem shows that the magnitude and often the character of the processes affecting a given slope seem to be conditioned by the existing slope geometry."
> (Chorley, 1964, p. 71).

The actual nature of such feedback relationships would be extremely difficult to establish in the periods of time during which slope studies must be conducted; however, Arnett (1971) has concluded that in the Rocksberg Basin, Queensland, slope process is determined by slope form, while Schumm (1956) concluded that slope form was governed by process. Presumably the actual situation is that form and process are intimately inter-related.

In view of the above difficulties in establishing process-form relationships by actual observation, it is perhaps not surprising that very few such studies have been made by geomorphologists, and that the purely theoretical studies referred to above have come to be dominant.

One of the few studies which attempts to identify the process acting to produce a given landform, and to infer the manner of slope evolution, is that of Schumm (1956), in which he studies erosion processes on rapidly-evolving badland slopes. Schumm found that badlands in South Dakota and Arizona displayed two types of topography - steep, sharp-crested slopes (mean maximum slope angle 44 degrees) on the Brule formation, and broadly rounded interfluves (mean maximum slope angle 33 degrees) on the Chadron formation. By observing rates of erosion on various parts of the slope on each of these topographic types, he infers that rainwash is the dominant process acting to produce the steep slopes on the Brule formation, and that creep produces the rounded forms on the Chadron formation. On the basis of the same measurements he concludes that rainwash produces parallel retreat and that creep causes slope decline. However, Schumm made no actual measurements of the slope processes operative; his conclusions are based only on the total amount of surface lowering at each point along a slope during a period of time, and his assessment of the process likely to produce such a pattern of lowering.

No evidence is presented which shows, for example, that creep is indeed dominant on the slopes developed in the Chadron formation, and as has been noted above, this would be difficult to achieve.

It must therefore be concluded that geomorphology has produced essentially no reliable information on the slope forms produced by the operation of particular processes, and on the mode of slope development under the action of these processes. Carson & Petley (1970) conclude that this is not surprising, since the neglect of "process" after the advent of the Davisian cycle has made such a task " ... too unattractive or too difficult for geomorphologists ... ". (Carson & Petley, 1970, p. 72).

However, engineers have made several advances in this field, and these are the subject of the next section.

1.2 STUDIES OF THE FORM-PROCESS RELATIONSHIP MADE BY ENGINEERS

Skempton (1953) made a study of the stability of slopes in boulder clay valleys near Shotton (Durham), which were subject to landsliding. Only those slopes which were not subject to stream undercutting (for example, those isolated behind extensive floodplains) were found to be free of landsliding. By field surveying Skempton found that all such stable slopes stood at inclinations between 20° and 25°; the average inclination was about 22°.

Two different types of landslides were observed. The most common were shallow surface slips; surveying revealed that these caused only a small decrease (of a few degrees) in the angle of the slope on which they occurred. The second type of mass movement was deep rotational slipping; these movements produced a much larger change in surface form.

Conventional stability analysis based on measured values of the shear strength of the clay showed that these slips should only occur when the height of the slope exceeds about 130 feet; this finding agreed with the observed field evidence.

Skempton therefore proposed the following sequence of landsliding in this area -

Stage 1: Downcutting and undercutting lead to surface slips or slumps. The valley deepens but the slopes remain at about 30°, controlled by the processes of surface slipping.

Stage 2: The valley depth increases to the point (at about 130 feet) at which slope height is sufficient for deep slips to occur.

Stage 3: When downcutting ceases, or declines, meandering widens the valley floor and some slopes are isolated by the floodplain. These slopes then flatten by further slumping to a relatively stable inclination, which Skempton called the "mature angle of repose". This angle was about 22°, as shown by the observation of stable slopes referred to above; it corresponds to the angle of internal friction of the clay material, which Skempton determined in the laboratory.[1]

Stage 4: If left isolated, the slopes will then be subject to soil creep, which causes extremely slow decline in the slope inclination.

Skempton (1953) suggested that the relationship of "mature angle of repose" to strength properties of the clay should be investigated further.

In a similar study, Skempton & De Lory (1957) found that in the London clay all unstable slopes (which were subject to shallow landsliding) stood at inclinations of greater

1. This finding agrees with the purely theoretical suggestion of Souchez (1965).

than 10^{o}, and that all slopes of less than 10^{o}were stable. Now a previous study (Henkel & Skempton, 1955) had shown that the stability of such clay slopes depended solely upon the angle of shearing resistance ϕ' (see Chapter 4) and that the clay behaved as if its cohesion intercept, C', was zero.

Skempton & De Lory measured values of ϕ' for the London clay, and found that it was approximately 20 degrees. They showed, by simple statics, that in such material if the water table rises to the ground surface, then the maximum stable slope angle is equal to about $\frac{1}{2}\phi'$. Assuming that C'= zero, following Henkel & Skempton (1955) this value thus explained the division of London clay slopes into unstable (above 10^{o}) and stable (below 10^{o}) as a function of the strength properties of this clay.

Hutchinson (1967) studied both inland slopes and protected coastal cliffs and found that both went through several phases of degradation by landsliding until they reached an angle (of about 8 degrees) at which they became stable against such movements. He suggests that since all material on the slopes will have been moved by landsliding when this stage is reached, that its strength will have fallen to its residual value (see Skempton, 1964, and Chapter 4), and therefore that "The limiting slope angle will thus be controlled by the residual angle of shearing resistance of the slope-forming material and by the maximum pore-water pressures that have occurred in the ultimate slope." (Hutchinson, 1967, p. 118).

It has been suggested in all of the studies referred to above that the inclination at which slopes in the London Clay finally become stable against any form of landsliding (known as the "angle of ultimate stability") is governed by the properties of the clay and the ground-water conditions, and that this angle may be predicted with reasonable accuracy from a knowledge of the shear strength of the clay. As Skempton, Hutchinson & Neville (1969) observe,

> "It would be of great interest to see whether the angle of ultimate stability in other clays can be correlated with the residual strength, as seems clearly to be the case with London Clay" (p.334).

This was essentially the object of the present field project. The procedures used in such studies are regarded as valuable for geomorphic research because of their essentially inductive nature: field evidence on slope form and processes is first collected, and hypotheses regarding the formation of the slopes developed on the basis of this evidence. These hypotheses may then be tested by laboratory determination of soil strength, and either accepted or rejected.

Independent confirmation of the stability analyses based on the laboratory measurement of soil shear strength may be obtained by observing the minimum inclination of slope on which landsliding will occur, and also the inclination of parts of the landscape which have already reached the ultimate angle of stability. If agreement between these variables is found, then it is possible to conclude that landsliding is the dominant process which has formed the slopes being studied, and to predict the angle to which the slopes will ultimately be reduced.
By study of present slope forms, it should also be possible to infer the mechanism of this decline - whether hinge decline, decline by replacement, and so on.

By using this method it should therefore be possible to predict with reasonable accuracy, the long term pattern of slope development in an area without the need to make process measurements, which are subject to the difficulties referred to above, and to establish a reliable relationship between geomorphic process and the resultant landforms by inductive methods.

The aim of the present study was, therefore, to attempt to relate the form and evolution of hillslopes in an area subject to landsliding to the strength properties of the slope mantle, and to assess the significance of the results from the point of view of geomorphology, rather than of engineering.

1.3 PREVIOUS STUDIES

The only previous studies in which this has been attempted are those of Carson & Petley (1970) and Carson (1971a).

Carson & Petley (1970) attempted to relate the angle of straight hillslopes in the southern Pennines and Exmoor to the strength of the soil mantle. They surveyed 24 slopes in each area, noting the presence of stability or instability on each slope. Straight segments of these slopes were then plotted as a frequency diagram against angle, and several peaks, or more frequently occurring slope angles, were identified. Three soil samples were then taken from these areas, and strength determinations made using the direct shear technique to be described in Chapter 4. Carson & Petley concluded from the results of these tests that the slope angles which were relatively frequent in the two areas studied were a reflection of the strength properties of different types of slope mantle, which reached their ultimate angle of stability against shallow landsliding at these angles.

However, the procedures used in this study are of questionable validity. A large number of specific procedural criticism could be made, but only those of a general nature are discussed here.

Perhaps the most fundamental criticism is that their survey of stable and unstable slopes shows that unstable slopes occur on the entire range of slope angles encountered. Carson & Petley do not seem to have regarded this as a problem - however, it clearly suggests that their predicted angles at which the slope mantle should become stable against landsliding are not reflected in the actual distribution of landslide occurrence. This must surely cast doubt upon the significance of the results; indeed, the absence of a cutoff point beneath which slopes are stable indicates that either no such relationship exists or that the stable slope angle is lower than that of any slopes measured in the field, so that all these slopes must be regarded as unstable, and any attempt to relate their "stability" to properties of the soil mantle is specious.

A second criticism in that having only surveyed 24 slopes in each area, the frequency histograms used are based on extremely low frequencies, the total number of angles plotted being less than 30; thus a frequency of 4 slopes in a particular one degree class is regarded as a "peak" in the distribution whereas the surrounding classes, having a frequency of only 2 slopes, are regarded as being insignificant. Such minor differences could easily be a function merely of the sample size, and thus are not a sound basis for any discussion of the slopes in this area. In the present study at least 1,500 - 2,000 readings were found to be necessary to reliably define the modal angle classes in a similar frequency distribution (see Chapter 3). Further, Carson & Petley give no indication of the slope forms encountered, so that the percentage of the slope profile represented by the rectilinear segments plotted in their frequency distributions cannot be assessed. They make no comment on this matter.

As may be seen from the two criticisms just outlined, Carson & Petley (1970) obtained no satisfactory evidence that slopes in the areas studied tended to stand at a particular angle or angles, or that the slopes became stable against mass-movement at any similar angle. These two pieces of information are pre-requisite to any attempt to relate slope processes to slope angle, for without them there is no basis against which to check predictions of slope stability, based on measured soil strength parameters. For this reason, the findings of Carson & Petley (1970) must be accepted with caution.

The selection and testing of soil samples in this study was also far from ideal: only three samples were tested, one from a slope representing each of the "peaks" recognised in the slope angle frequency histogram. The samples were taken from that part of the soil profile which was considered to have least shear strength; however, no check was made either on this or on the variability of strength properties along the slope profile.

Further, all large stones were removed from the samples prior to testing, hence altering the nature of the material whose strength was being investigated. In addition, most tests (for example, Figure 9, p. 80 of Carson & Petley, 1970) show no tendency for a constant residual strength (see Chapter 4) to be reached; Carson & Petley do not explain how they derived their values from these graphs. Indeed, having determined (by whatever method) the value of ϕ' for each sample, Carson & Petley found that in only one of the three cases did the predicted stable slope correspond with that actually found in the field (see p. 85 of Carson & Petley, 1970). They therefore suggest that the fitted regression lines from which the values of are obtained " ... may not be correct", and therefore adjust these lines (without justification) so that they pass in each case through only two of the data points; these lines then define values of ϕ' which correspond more closely to the actual slope angles measured in the field. Such a procedure amounts to adjusting the results until they support the conclusion which it is desired to make, and cannot be justified.

The stability equations applied by Carson (1971) apply only to shallow planar landslides (see Appendix A). Since such landslides played no apparent role in the development of slopes in the area studied by Carson, the use of these equations is not valid.

It is therefore felt that no reliable geomorphic study has been made using the ideas of Skempton, De Lory, Henkel, Hutchinson, and others, and for this reason a further investigation was undertaken in the present project.

CHAPTER II:

THE ENVIRONMENT OF THE STUDY AREA

2.1 LOCATION

For the purpose of this investigation, a small stream catchment[1] lying about 3 miles northwest of Picton, and about 45 miles southwest of Sydney, was selected for detailed study. It lies on the southern side of the Razorback Range, which is a major topographic feature running approximately northwest-southeast and reaching elevations of about 1100 feet.

The stream which drains the study area is an unnamed tributary to Racecourse Creek, which flows ultimately to join the Nepean River south of Picton. The northern boundary of the catchment is khows as Evelyns Range, which reaches an elevation of about 1000 feet; maximum local relief is approximately 500 feet (Figure 21).

1. Map area approximately 2.5 sq. miles.

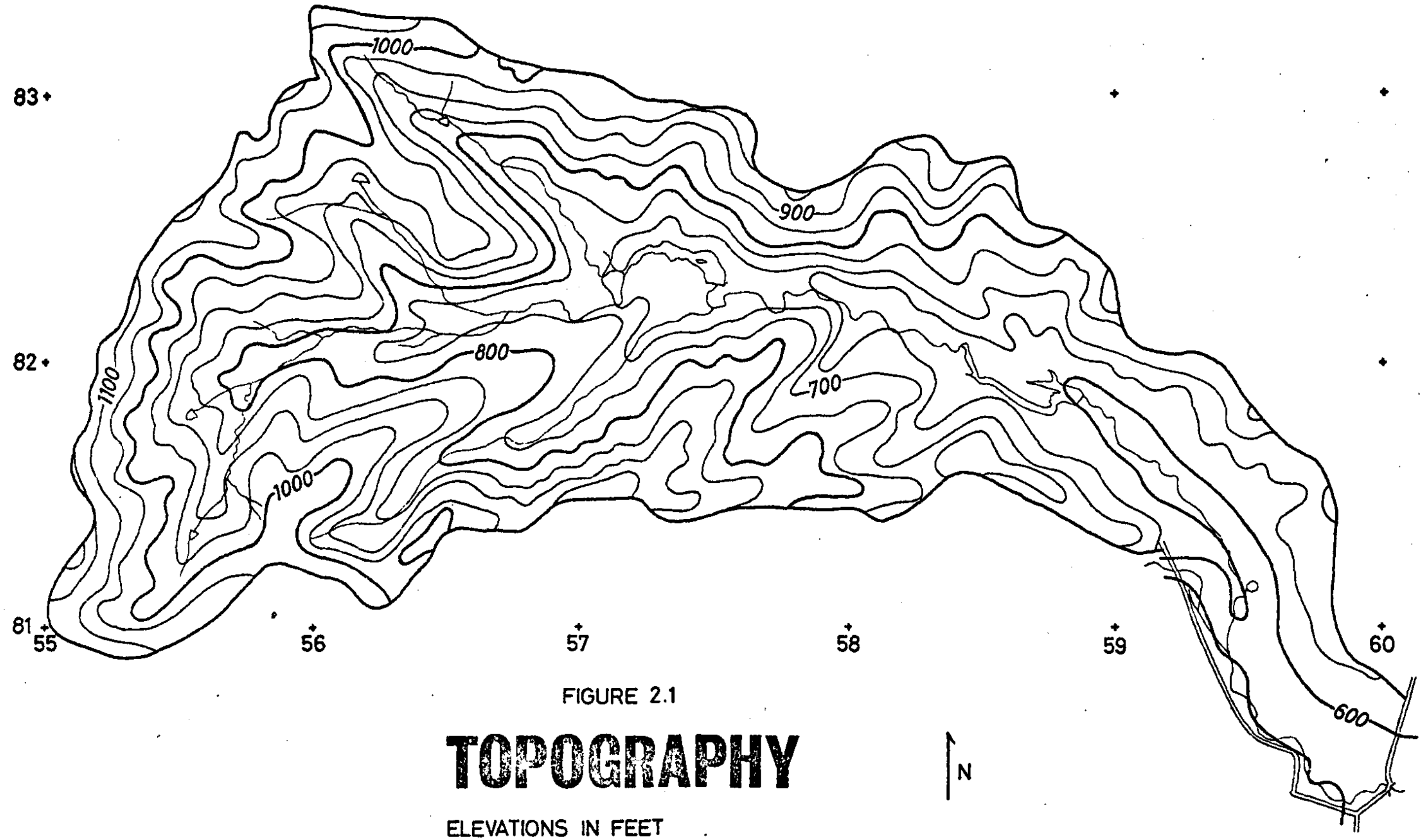


FIGURE 2.1

TOPOGRAPHY

ELEVATIONS IN FEET
CONTOUR INTERVAL 50 FT.

Enlarged from Military sheet No. 428, Zone 8 (1:63,360). Grid spacing = 3000 feet.

D. Dunkerley 1972

2.2 GEOLOGICAL SETTING OF THE STUDY AREA

The study area lies within the southern section of the Sydney Basin, a large structural basin which forms a north-westerly trending trough between the New England Fold Belt and the Lachlan Fold Belt, in eastern N.S.W. (Conolly, 1969b).

The Sydney Basin covers approximately 12,000 square miles on land, and may include a further 5,000 square miles offshore (Edwards, 1969; Stuntz, 1969). It represents part of a large depositional basin which formed in eastern N.S.W. during Permo-Triassic times, and was essentially an area of Permian and Triassic sedimentation (Conolly, 1969a). The basin consists of a central outcrop of Triassic rocks surrounded by Permian rocks on the north, west, and south (Figure 2.2). Little is known of the rocks seaward from the coast, but it is likely that they extend to the edge of the present Continental Shelf, which off the Sydney Basin, is about 15-30 miles from the coastline (Conolly and Ferm, 1971).

2.2.1 Stratigraphy of the Sydney Basin

The Permo-Triassic rocks of the Sydney Basin are generally divided into two groups - an upper non-coal-bearing sequence (Triassic), and a lower coal-bearing sequence (Permian). The Triassic sequence can be divided into three major rock units,

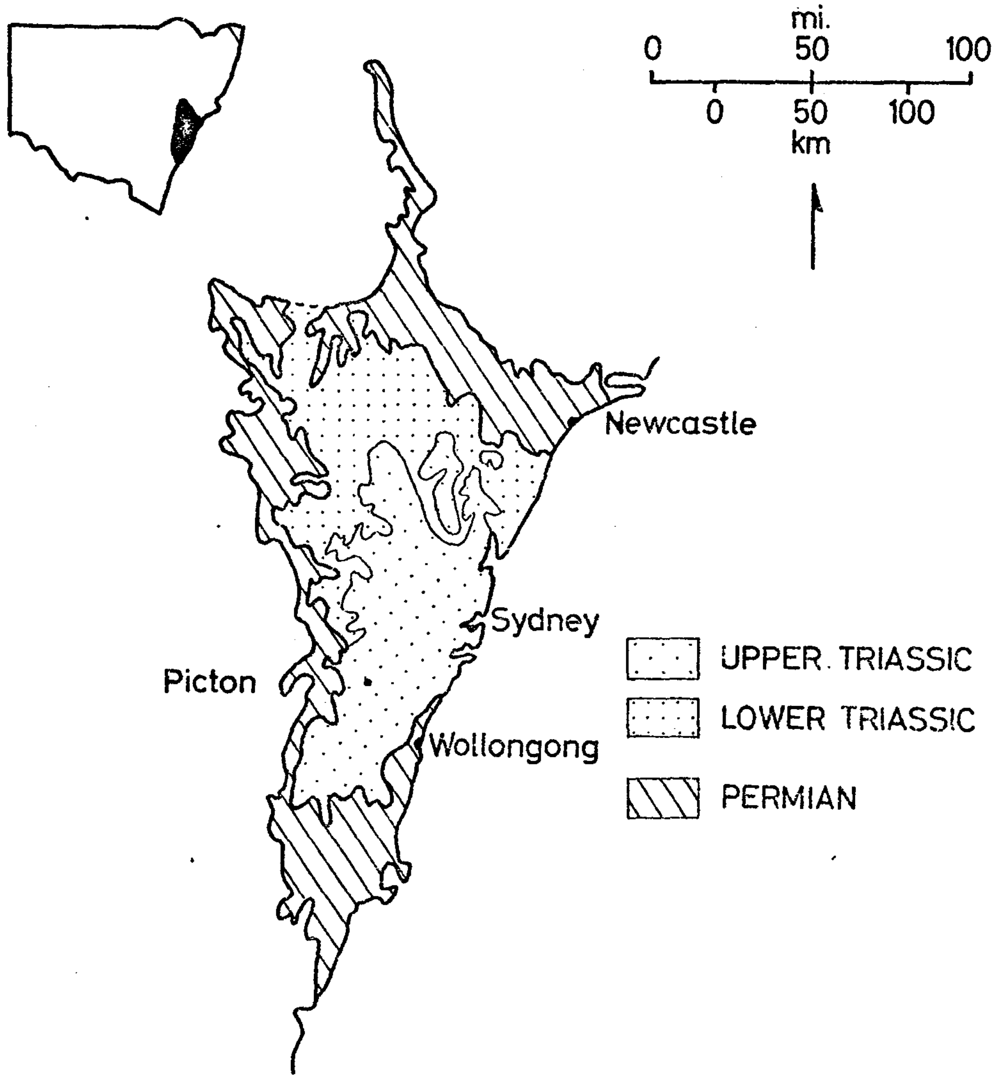


FIGURE 2.2

GEOLOGY OF THE SYDNEY BASIN
(partly after Packham, 1969, Figure 5.1)

the Wianamatta, Hawkesbury, and Narrabeen Groups, and the Permian sequence into an upper and a lower coal sequence, both overlying marine sediments (Conolly and Ferm, 1971). These relationships are shown in Table 2.1.

TABLE 2.1

Generalised Permian-Triassic Sequence, Sydney District

Wianamatta Group	)	
Hawkesbury sandstone	)	Triassic
Narrabeen Group	)	
Illawarra Coal Measures	)	Permian
? Shoalhaven Group	)	
? Carboniferous		

(After Table 5.1, Jnl. Geol. Soc. Aust., 16 (1), 311.)

The complete Permo-Triassic sequence is thickest in the vicinity of Sydney (at least 12,000 feet); however, the greatest thickness of Triassic rocks is in the north-western part of the basin. The present day basin shape is due largely to deformation during the Cainozoic Era, but is in part a reflection of the structure which existed during the Permian and Triassic (Packhan, 1969).

The study area is located entirely within the uppermost of the Triassic rock units referred to above, the Wianamatta Group.

2.2.2 Stratigraphy of the Wianamatta Group

The lithology and stratigraphy of the Wianamatta Group were first described comprehensively by Lovering (1954), and subsequent studies have been made by Dyer (1966), Herbert (1970), and Matson (1970).

It can be concluded from these studies that the stratigraphy is complex, the rocks displaying numerous rapid lateral facies changes. In addition, interpretation is made difficult by the lack of outcrop. As a result of these two factors, no agreement as to the detailed stratigraphy of the Wianamatta Group has been reached among the studies referred to above. However, the broader features of the stratigraphy have been fairly well established.

The Wianamatta Group consists of a conformable sequence of shallowly north to north-westerly dipping interbedded grey and black shales and lithic sandstones (Matson, 1970; Lovering and McElroy, 1969), and has a total thickness of approximately 900 feet (Herbert, 1970). Fossils are rare in all but the lowest 90 feet of the Group, and Lovering

(1954) was therefore uncertain as to the age of these rocks; however, recent palynological investigations have indicated a Mid to Late-Triassic age (Herbert, 1970; Lovering and McElroy, 1969).

Lovering (1954) divided the Group into two Sub-Groups, the Liverpool Sub-Group (the lower of the two, and of predominantly shale lithology), and the Camden Sub-Group (upper, dominantly of sandstone but with associated shales).

The Liverpool Sub-Group, which has a fairly constant total thickness of 400 feet (Lovering, 1954; Lovering and McElroy, 1969), lies immediately above the Hawkesbury sandstone (Table 2.1). Shale is the main rock type; Lovering (1954), Lovering & McElroy (1969), and Dyer (1966) have all recognised a single sandstone formation approximately 20 feet thick (the Minchinbury sandstone) separating two shale formations (the Ashfield shale and the Bringelly shale) as the formations comprising this Sub-Group; however, Chestnut and Pogson (1968) recognised three sandstone units, and Herbert (1970) and Matson (1970), four. (Herbert (1970) has therefore suggested that on the basis of recent drilling, Lovering's (1954) map shows the Minchinbury sandstone incorrectly plotted, possibly connecting several sandstone units at different stratigraphic levels). These schemes are illustrated in Figure 2.3.

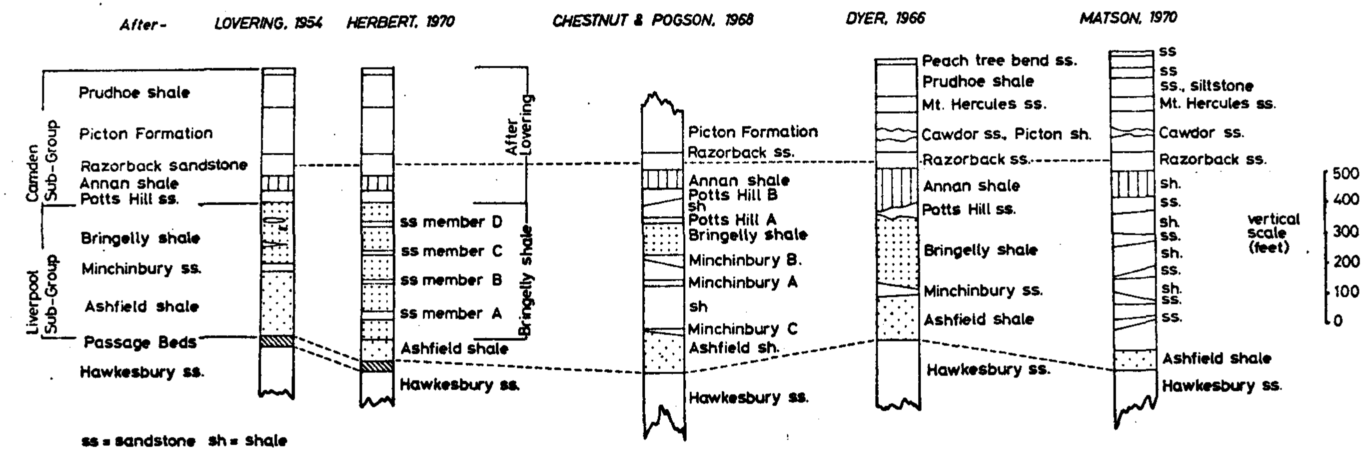


FIGURE 2.3

(Partly after Matson, 1970)

WIANAMATTA STRATIGRAPHY

The Camden Sub-Group was separated by Lovering (1954) at the base of the Potts Hill sandstone (Figure 2.3). This Sub-Group is composed of a series of alternating sandstones and shales; its thickness is rather variable, but reaches a maximum of about 350 feet in the Razorback Range (Lovering and McElroy, 1969). Again the detailed stratigraphy has not been agreed upon.

2.3 Geology of the Study Area

Field mapping, supplemented by diamond drill core logs, was undertaken in an attempt to identify the formations present in the study area. It was found that the geologic map of the Razorback produced by Lovering (1954, Map 8, p. 179) was incorrect, at least as far as the study area was concerned. Lovering had mapped the area as entirely located within the Liverpool Sub-Group, whereas the basal units of the Camden Sub-Group (the Potts Hill sandstone and Annan shale) are also present. It was also found that three (rather than one) sandstone members were present between the top of the Ashfield shale and the Potts Hill sandstone. The revised geological map is presented in Figure 2.4.

Difficulty was found in mapping the structure of the area, because outcrops were rare and often too weathered

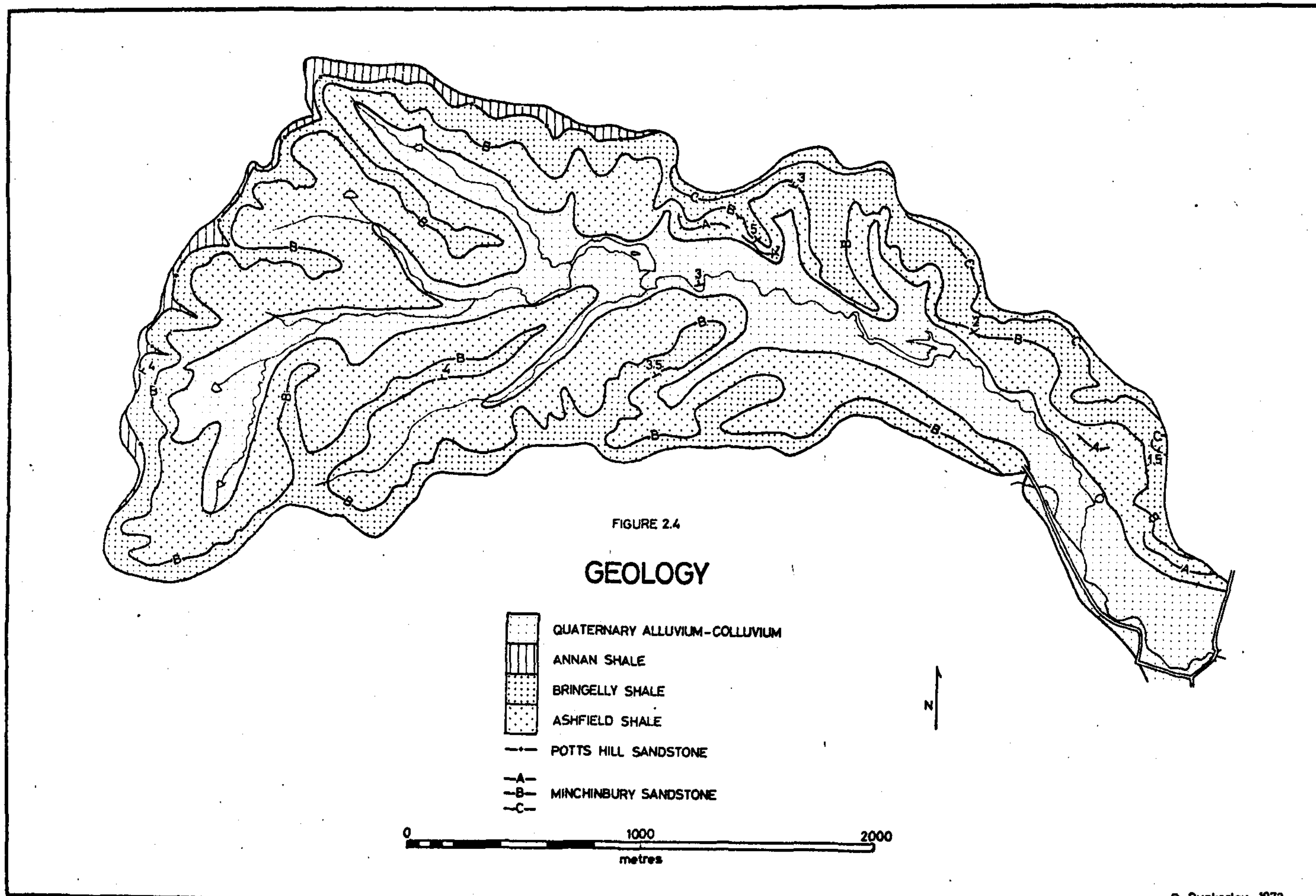


FIGURE 2.4

GEOLOGY

to give reliable readings. However, dip and strike measurements were made on all available outcrops, and these are plotted on the geological map. It was found that all units present dipped shallowly to the north or northwest. No folding was encountered. However, the presence of a major fault (presumably connected with the Nepean fault system, which runs past the eastern end of the study area) was inferred on the basis of the field mapping. The passage beds (or Mittagong formation) which mark the transition between the Hawkesbury sandstone and the Ashfield shale may be seen in the Hume Highway rail-underpass at Picton; at the same elevation in the study area, one mile to the northwest, the top of the Ashfield shale is present, and hence a fault of large vertical displacement must run northward past the eastern end of the study area. This agrees with the finding of Dyer (1966) who found that in the southern part of the Razorback, the Hawkesbury sandstone overlies the Ashfield shale; Dyer suggested that the faulting necessary to explain this was contemporaneous with the uplift of the Blue Mountains and the formation of the Lapstone Monocline. Branagan (1969) notes that although the present structure of this feature undoubtedly post-dates sedimentation, there is a suggestion of sedimentary control of the Mittagong formation by the monocline near Picton.

The rock units present in the study area were thus found to be the Ashfield shale, Minchinbury sandstone, Bringelly shale, and Potts Hill sandstone.

a) The Ashfield shale: This unit is composed almost entirely of well-bedded black shales, with some mudstones and siltstones locally abundant (Lovering, 1954). At several places within the study area, layers of hard sideritic mudstone were observed within the shale; this is a characteristic association of the Ashfield shale (Lovering, 1954). The black colour of these shales is due partly to organic matter, and in part to a high iron content, together with some pyrite (Lovering, 1954); in the study area the shale was found to be very micaceous.

According to Lovering (1954), the average thickness of the Ashfield shale is 200 feet; at 631770 Dyer (1966) found the thickness to be 160 feet.

b) Minchinbury sandstone: This is a grey, calcareous lithic sandstone; it is normally massive but some shale lenses are present (Lovering, 1954). Ironstone and sideritic nodules were found to be quite common in this sandstone. Its thickness varies from 10 to 20 feet (Lovering, 1954); however, three separate sandstone members were defined in the study area, in contrast to the single unit identified by Lovering (1954). The continuity of the upper and lower units (mapped as members A and C) could not be checked; the middle unit (B), which is more prominent, has been taken as the Minchinbury sandstone for purposes of defining formation boundaries.

c) Bringelly shale: This unit consists predominantly of black shale, but also has a subsidiary sandstone lithology, with many sandstone lenses from one inch to five feet in thickness (Dyer, 1966; Lovering, 1954). The unit has a fairly constant thickness of 200 feet (Lovering and McElroy, 1969).

d) Potts Hill sandstone: This unit is a massive and cross-bedded lithic sandstone, consisting of angular quartz grains and lithic fragments in a clay cement (Dyer, 1966). Since it is a calcareous sandstone, joints filled with secondary calcite are common (Lovering, 1954); these may be observed in the small quarry at 552819 on the western margin of the study area. The unit has an average thickness of 40 feet (Lovering, 1954).

2.4 CLIMATE OF THE STUDY AREA

2.4.1 Rainfall

The average total annual rainfall at Picton is only 31.5

inches, since Picton lies within a rainshadow area to the east of the Blue Mountains plateau (Corbett, 1972; Walker, 1960). The rainfall is characterised by a marked seasonal distribution (Figure 2.5), much of the annual total coming in summer and autumn. Winter and spring are drier seasons, and the typical short showers of these seasons are accompanied by strong westerly winds which lead to high evaporation rates. Summer rainfall comes in sharp thunderstorms and is of high intensity (Beirne, 1952; Huston, 1953). The rainfall is variable both in annual total and in seasonal distribution. Lawrence (1937) identified 8 unusually dry years in the period 1900-1935. Data for the rainfall at Picton since 1880 (obtained from the Commonwealth Bureau of Meteorology) show a marked decline in autumn and winter rainfall from about 1895 onward. There is a marked tendency for these seasons to be extremely dry on occasions, as in 1888, 1895, 1902, 1940, 1954, and 1968. Kraus (1954) noted that winter and spring rains along the south-east coast of Australia had shown a very marked decline at about 1893-1894, and suggested that a significant change in rainfall regimes occurred at this time, perhaps related to a change in the upper westerlies. He also noted that summer rainfall has increased gradually since 1893. Hence it may be misleading to speak of the average rainfall in this area, particularly as the tendency for autumn and winter to be extremely dry (noted above) appears to be reflected in the

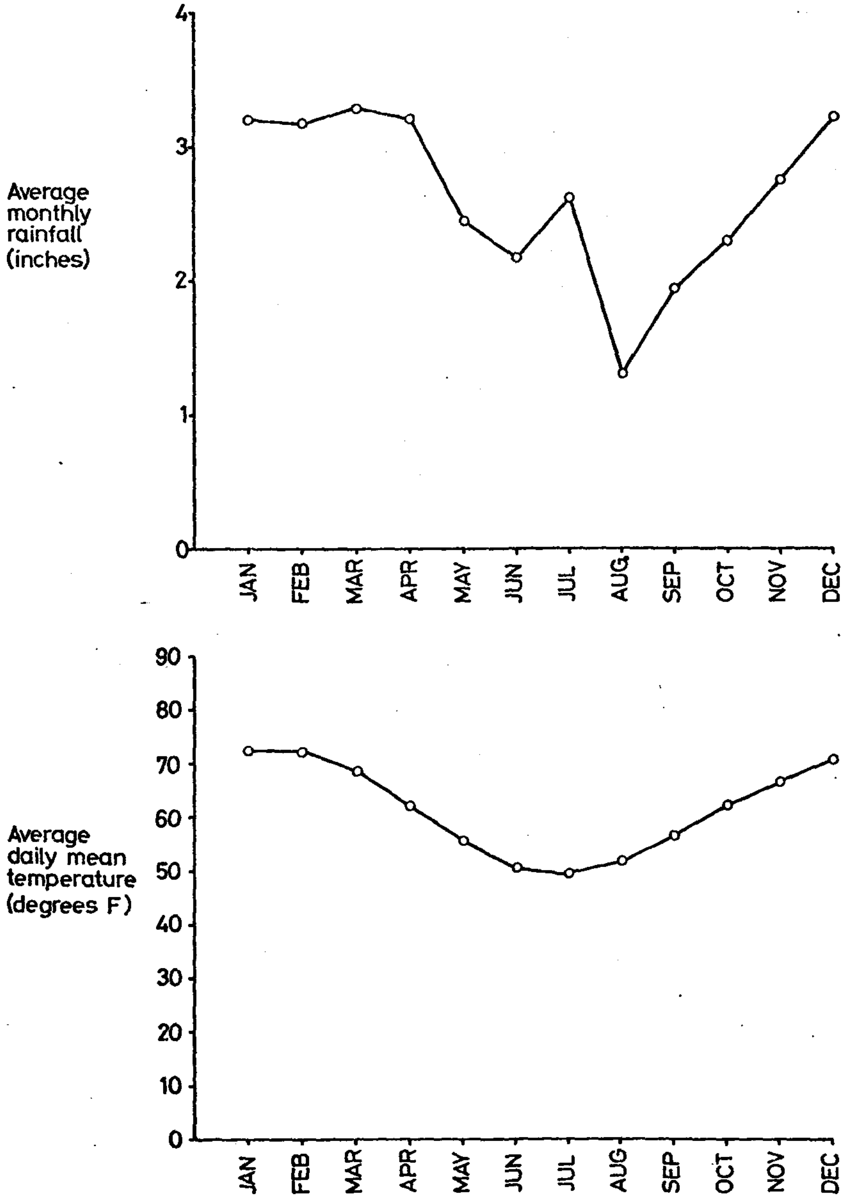


FIGURE 2.5
CLIMATIC DATA FOR PICTON

occurrence of landslides in the years following (Table 2.2).

Table 2

Relationship of dry seasons to landslide occurrence, Razorback Range.

Period of markedly dry Autumn-Winter seasons	Period of landsliding
1885, 1888	1880's (Cambage, 1924)
1895	1896 (" ")
1902,1908,1909,1910	1905-1915 (" ")
1954	1956 (Matson, 1970)
1968	1969 (" ")

2.4.2 Temperature

Temperatures at Picton range from a daily average of about 70^{o}F in summer to about 50^{o}F in winter. Daily maxima in summer average about 85^{o}F, and daily minima in winter about 40^{o}F. Complete temperature data are given in Figure 2.5.

2.5 VEGETATION OF THE STUDY AREA

The vegetation of the study area (and of most of the Razorback) has been altered by clearing and the introduction of new species since settlement. Cambage (1924) observed that the shale areas of the Razorback produced fairly open forest, and that the species included *Eucalyptus hemiphloia* (Grey Box), *E. tereticornis* (Forest Red Gum), *E. baueriana* (Round Leaf Box), and *Melaleuca styphelioides* (Paper-barked Tea Tree) in damp places. Cambage also noted that *Casuanina torulosa* (Forest Oak), after which The Oaks was named, was very common on the hills around Picton before settlement, since descriptions in 1802 and 1804 refer to the hills as "abounding with she-oaks" and "making good pasture". In addition to these species, Pidgeon (1937) recorded *E. siderophloia, E. sideroxylon, E. crebra, Angophora intermedia,* and *A. subrelutina,* as having belonged to the native vegetation.

Of these species, some examples of the genus *Melaleuca* (paperbarks) are still present in the study area today, particularly on the lower parts of slopes, which is taken to indicate that these are moist areas, since such conditions are favoured by this genus (Child, 1968). Also still present are a few grey box *(E. hemiphloia),* and ironbarks *(E. crebra)*.

The study area is largely cleared, and this area is now occupied by Wallaby grass *(Danthonia spp.),* coastal blue grass *(Dicantheum spp),* and paspalum *(Paspalum dilatatum).*

2.6 SOILS OF THE STUDY AREA

Following the definition of Hamel and Flint (1972, p. 167) that "Colluvium is geological material which has moved downslope under the influence of gravity, ie. landslide or creep debris", it would appear that a large percentage of the soils of the study area are colluvial. However, whilst some of the area has been mapped as "alluvium-colluvium" (Figure 2.4), much of it is occupied by soils which would generally be regarded as having developed *in situ*.

These soils have been classified as red podsolics, following the scheme of Stace and others (1968); they are members of the Cumberland Family mapped by Walker (1960). The weathered profiles are generally deep (Table 2.3), commonly more than six feet, even on the tops of hills. Hand augering has shown that the soil profile typically consists of a variable depth (four inches to two feet) of friable, dark brown A-horizon (commonly clay loam) which is subject to hardsetting (Northcote, 1971, p. 12), overlying a predominantly red (5YR 5/6) B horizon, composed of stiff clay. The B horizon is typically about three feet thick, becoming mottled with yellow and/or grey towards the base. The B horizon overlies up to ten feet of weathered shale fragments in a matrix of clay. A-horizons are often stony, containing both shale and sandstone fragments, but the B-horizons are generally free of such material.

Soil reaction is neutral to acid throughout the profile, typically being about 7 near the surface and decreasing to 5.5-6 at depth.

Table 2.3

Depth of soil profiles (including C-horizons) in the study area.

Borehole No.	Depth (feet)
RB 95	18'
RB 97	9'
RB102	8'6"
RB103	17'6"
RB104	6'6"
RB105	5'6"
RB106	12'9"
RB107	6'
RB108	11'9"
RB109	12'
RB110	9'6"
RB113	7'
RB114	14'
RB124	10'

(Source: Extracted from borehold data supplied by the Clutha Corp.)

2.7 SEISMICITY

In the years 1959-1967 inclusive, 181 tremors have been located in the vicinity of the Sydney Basin. Most epicentres lie near the present western boundary of the basin; some are associated with the Lapstone Monocline (and related structures) which is present as a fault in the Picton area (Doyle and others, 1968). Details of shocks of magnitude 4 (Richter) or greater so far recorded are given in Table 2.4.

The epicentre of the largest earthquake (the Robertson-Bowral earthquake) recorded so far was only 20 miles from Picton. Since the period of record is very short, it is reasonable to assume that shocks of greater magnitude occur from time to time.

2.8 CURRENT SLOPE PROCESSES

Five active slope processes have been recognised within the study area. These may be classified as follows -

a) Mass-movements
   - i) creep
   - ii) landslides

b) Particle-movements
   - iii) slopewash
   - iv) gullying
   - v) piping

Table 2.4

Seismicity of the Sydney Basin

Year	Source	S	E	Mag.	Approx. distance from Picton, miles
1919[1]	Burke-Gaffney	33.5	150.7	4.0	45
1925	"	33	152	4.5	110
1930	"	34.5	149	5.5	100
1934	"	34.5	149.5	5.5	70
1934	"	34.5	149.2	4.5	90
1947	"	35	149.5	4.0	85
1947	"	34	148.6	4.0	120
1949	"	34.8	149.3	4.5	85
1961[2]	Doyle and others	34.33	150.30	5.5	20

1. The Kurrajong earthquake.
2. The Robertson-Bowral earthquake.

Source: From Burke-Gaffney (1951) and Doyle et al. (1968)

In addition, some material is doubtless removed from the slopes in solution.

A) Mass movements.

i) Creep: soil creep was recognised on the basis of terracettes of various sizes on parts of many profiles. However, such visible evidence of creep was only observed where the slope was steeper than 22°; commonly such slopes were in the range 25-30°. It is assumed that creep occurs on gentler slopes without visible effect.

ii) Landslides: 33 prominent slides were observed within the study area. These will be discussed fully in a subsequent section.

B) Particle movements.

iii) Slopewash: Evidence for slopewash was only observed where the vegetation cover had been destroyed by stock, as along some ridge crests. Elsewhere the thick grass cover obscured the ground surface, and it is considered that under such a dense vegetation mat slopewash would be negligible.

iv) Gullying: 11 major discontinuous gully systems occur within the study area. Their location is shown in Figure 3.3. The headwalls of several of these were pegged and surveyed at the beginning of fieldwork, and were found to have advanced upslope by collapse by an amount of four to eight inches over a period of several months. However, these gullies generally only occur in the bodies of alluvium-colluvium at the base of slopes, and hence may be disregarded when considering the evolution of the upper slopes. Some details of the gully systems elsewhere in the Razorback Range have been given by Quick (1968).

Gullying is often closely associated with areas affected by landslides. This may be a consequence of the fact that groundwater is able to flow out over the slope after a landslide (So, 1971; Cambage, 1924), initiating gullying according to the process outlined by Kam (1965), even though the grass cover may not have been broken.

v) Piping: Several examples of pipes were noted. All were associated with earthflows. In cross-section the pipes were circular, having diameters of approximately two feet. Jackson (1966) has suggested that piping may contribute to the occurrence of landslides, but in the study area all those observed were located in material which had moved by landsliding, and hence they must have developed after the movement. It is considered that piping also is negligible as a slope-modifying process in this area.

2.9 LANDSLIDES IN THE STUDY AREA

Thirty-three major landslides were observed within the study area. A simple calculation, as given by Young (1971, p.86) is sufficient to show that these forms of mass movement must be the volumetrically dominant process on slopes on which they occur: a single landslide involving the movement of $10m^3$ of soil 10m downslope is volumetrically equivalent to a continuous process (such as creep or slopewash) acting at an average rate of 10 $cm^3/cm/yr$ for 10^6 years. Hence, if landslides occur on a given slope even at intervals of many thousands of years, they will be the dominant process on that slope.

Some actual data on the volume of material moved by landslides in the Razorback Range was given by the Department of Main Roads (1952), who removed at least 80,000 cubic yards of soil material from the old road over the Razorback during the early part of 1950. Huston (1953) has noted that during 1950, some entire hillslides moved, thus involving very large volumes of material.

2.9.1 Earthflow distribution and morphology

The landslides observed within the study area were all classified as belonging to the earthflow and slump-earthflow classes of Varnes (1958). In such movements the displaced mass of soil adopts a form which suggests that it moved as a viscous fluid. Examples are shown in Plates 2.1 and 2.2; both of these earthflows occurred on south-facing slopes within the study area. According to Varnes (1958), slip surfaces within such flows are generally poorly developed and short-lived, and the boundary between moving and stationary material may be sharp or it may be a zone of plastic flow. Almost all of the flows within the study area appeared to have moved without the formation of a major slip surface; however, slump-earthflows were also

Plate 2.1

Plate 2.2

Examples of earthflows in the study area recognised on the basis of backward-tilted blocks of soil, characteristic of slumping (Sharpe, 1938), encountered near the headwalls of some flows. It is probably that discrete shear surfaces are associated with these features. Crandell (1952) has suggested that the term earthflow should not include this upper part of the movement, since it does not involve true flowage.

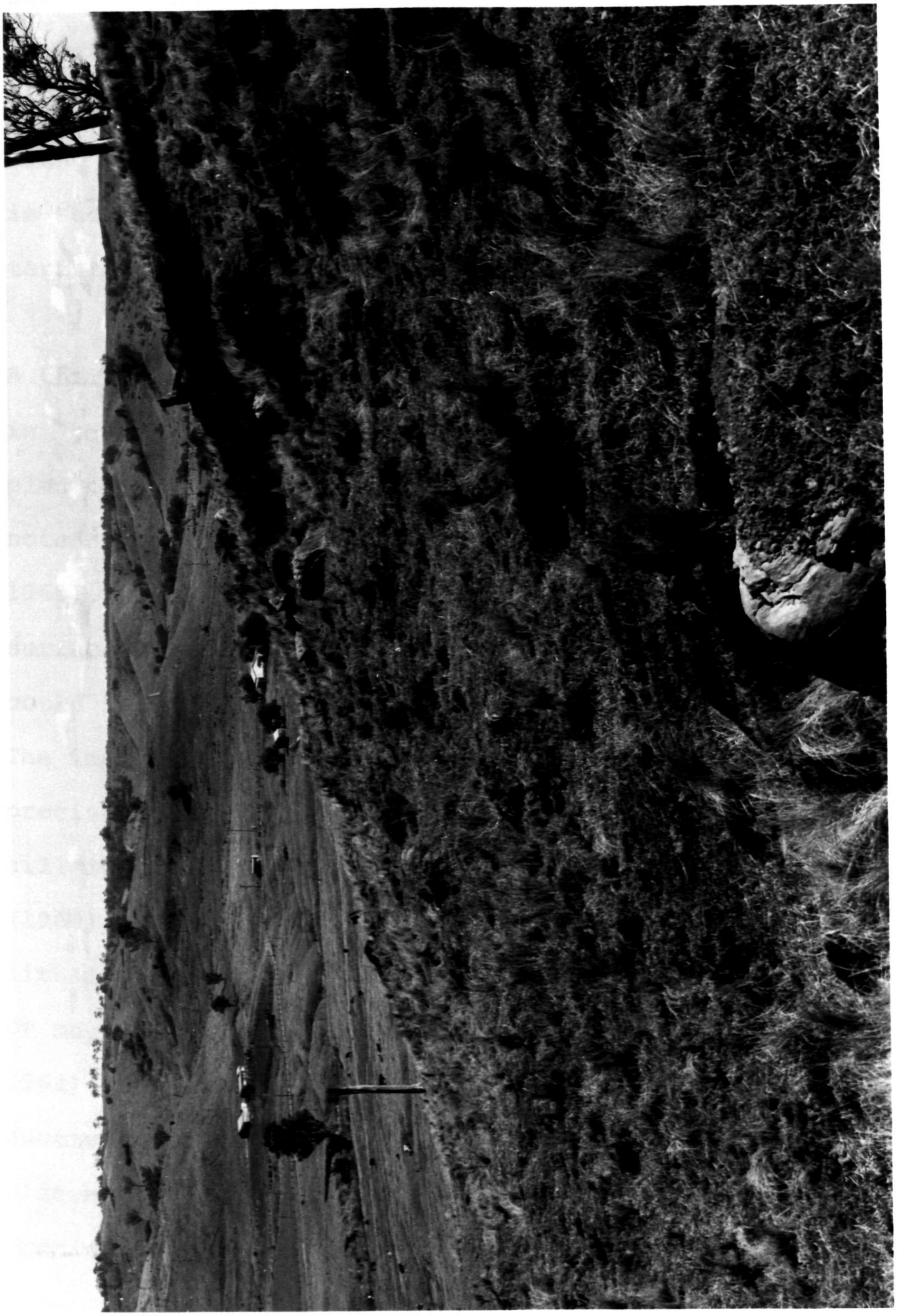
No morphological classification of the earthflows within the study area has been attempted; however, certain common characteristics are noteworthy. It is an almost universal characteristic (the exceptions being very small flows only a few metres in length) that the headwall area of a slide is occupied by an outcrop of a sandstone bed; often the sandstone outcrop completely surrounds the amphitheatre-like source area of the flow. In some cases only large displaced blocks of sandstone in the headwall area indicate the presence of a sandstone bed, as shown in Plate 2.3. The significance of this characteristic will be discussed in section 5.3.

A second common characteristic of the earthflows in the study area is that the erosional zone, or apparent source of the displaced material, occupies a relatively small percentage of the area affected by the flow; the bulk of this area is occupied by the depositional zone, that is, by the material deposited by the movement. This can be

Plate 2.3

Headwall area of a landslide in the study area showing displaced blocks of sandstone.

The prominent bench on the ridge in the middle distance is formed by the Minchinbury sandstone (member B); the skyline reflects the presence of the Potts Hill sandstone.



seen in Plates 2.1 and 2.2. In some cases, no such discrete zones of the movement are discernable, and entire hillsides appear to have failed simultaneously along numerous small shear surfaces. Such a situation is shown in Plate 2.4. The significance of these characteristics is considered in Section 5.3.

A third characteristic is that most of the earthflows occur in "coves" between spurs, rather than on slopes with marked plan convexity, such as spur-ends. This tendency has been noted in other studies (Waltz, 1970; Galpin, 1940; Rapp, 1960), where it is generally attributed to the control of surface runoff by the plan form of the slope; however, it could be equally significant in terms of sub-surface flow. The importance of the movement of sub-surface water in the precise location of an earthflow, for example on a particular hillside, has been emphasised by Crozier (1969) and Wright (1966). The movement of groundwater may be controlled by lithological junctions (Prior and Douglas, 1971, p.299), or may relate to the distribution of "percolines" (Bunting, 1964). Bailey and Rice (1969) found that hydraulic conductivity of soils at landslide sites was lower than that elsewhere; this also may be a significant factor in the precise location of earthflows in the study area.

Plate 2.4

In an attempt to define the part of the soil profile which generally failed in such flows, and hence governed the stability of the slopes, cross-sections of three of the smaller earthflows were prepared by hand-augering a series of five holes along a line running down the middle of the flow from the undisturbed area above to the lower limit of the visible erosional zone. Holes were placed three to five feet apart along this line. In two cases the failure appeared to have affected all of the profile above the lower part of the B-horizon, since only this lower zone and its contact with the C-horizon remained parallel to the ground-surface when plotted in section, whilst the remainder of the profile was disturbed and in general could not be correlated from one auger hole to the next. In these cases failure in the highly plastic B-horizon is indicated, the A-horizons having been disturbed merely by virtue of their location immediately above; however, in the third case examined, (a very small flow), only the A-horizon appeared to have been affected, since even the top of the B-horizon remained parallel to the topographic slope. The cross-section of this flow is given in Figure 2.6. In the largest flow in the study area, exposures in gully sections suggested that even the C-horizon may have been involved in the movement, since the deposited material contained abundant small fragments of weathered shale, typical of C-horizon material.

A significant characteristic possessed by the three flows examined was that the up-slope limit of the erosional zone occurred at a point at which the soil cover was unusually shallow. This is shown in Figures 2.6 and 2.7. The same relationship is evident in the section given by Sharpe & Dosch (1942, Figure 7, p. 318) who do not comment on this feature; its possible significance will be considered in section 5.3.

A striking feature of the distribution of earthflows, apart from their close relation to sandstone beds mentioned previously, is their general absence from northerly-facing slopes, and concentration on southerly-facing slopes. To obtain a more accurate impression of their distribution, the aspects of the 33 flows within the study area were measured from aerial photographs; it was found that 18 (or 54.4%) occurred on south-facing slopes, and a further 5 (15.1%) on south-easterly facing slopes (Table 2.5). To check whether this concentration might be due in part to a predominance of south-facing slopes in the study area, the orientation of the ridge crest above each flow was also recorded. It was found (Table 2.5) that no directional concentration was present in the ridge data, and hence, assuming that slopes are present on both sides of a ridge, that there was no predominance of south-facing slopes. The two distributions are shown in Figures 2.8 and 2.9.

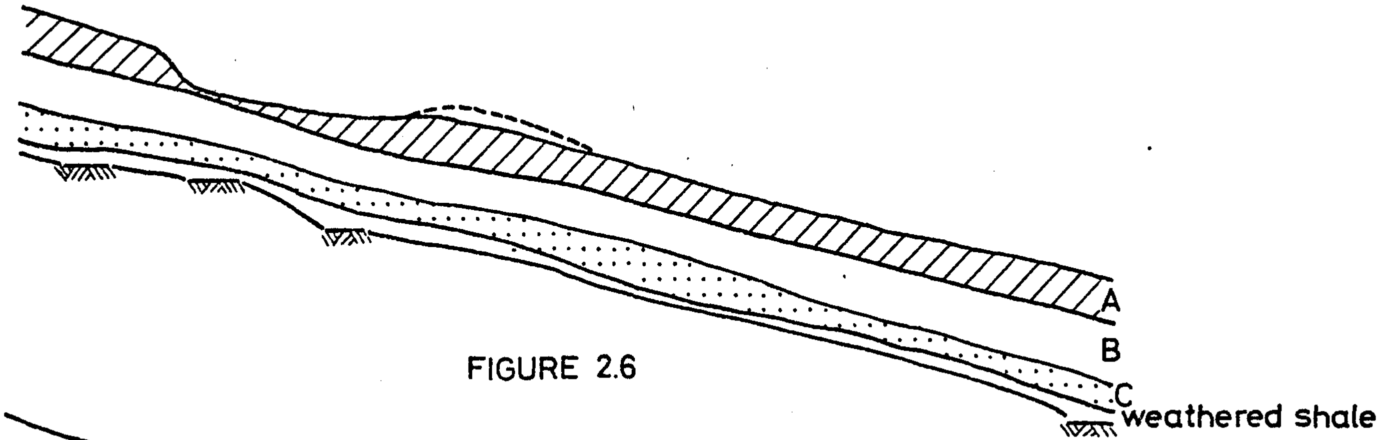


FIGURE 2.6

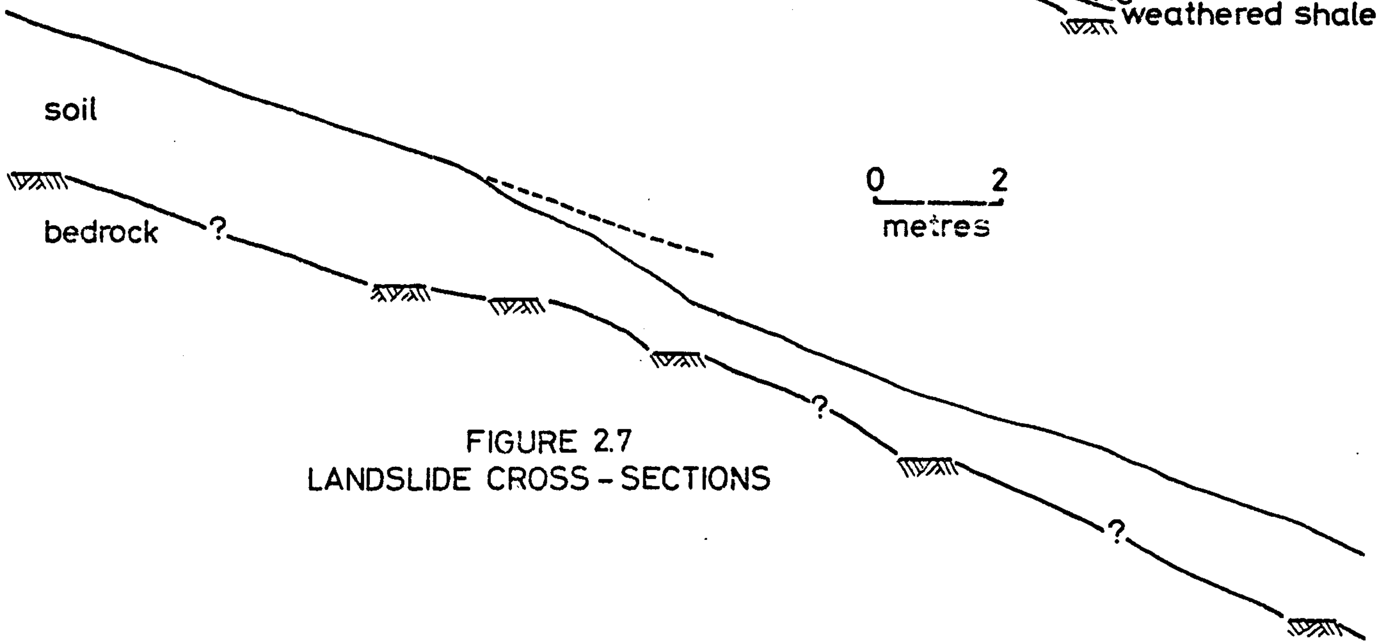


FIGURE 2.7
LANDSLIDE CROSS - SECTIONS

Table 2.5

Orientation of landslides and ridge crests in the study area.

Quadrant	Number of slides	Number of ridge crests
N	4 (12.1%)	4.5 (13.6%)
NE	1 (3.3)	3 (9.1)
E	1 (3.3)	5 (15.2)
SE	5 (15.1)	4 (12.1)
S	18 (54.5)	4.5 (13.6)
SW	1 (3.3)	3 (9.1)
W	3 (9.1)	5 (15.2)
NW	0 (0.0)	4 (12.1)
	N=33	N=33

Since this was a very small sample, a further 304 readings were made on flows (and corresponding ridge crests) throughout the Razorback Range. As can be seen from Table 2.6, the same pattern emerged: most flows occurred on slopes facing south and southeast, but this could not be explained by a predominance of slopes facing in this direction.

There are two possible explanations of this distribution - that is results from structural control by the north-northwesterly dipping strata, or that it reflects the control of slope processes by local microclimate.

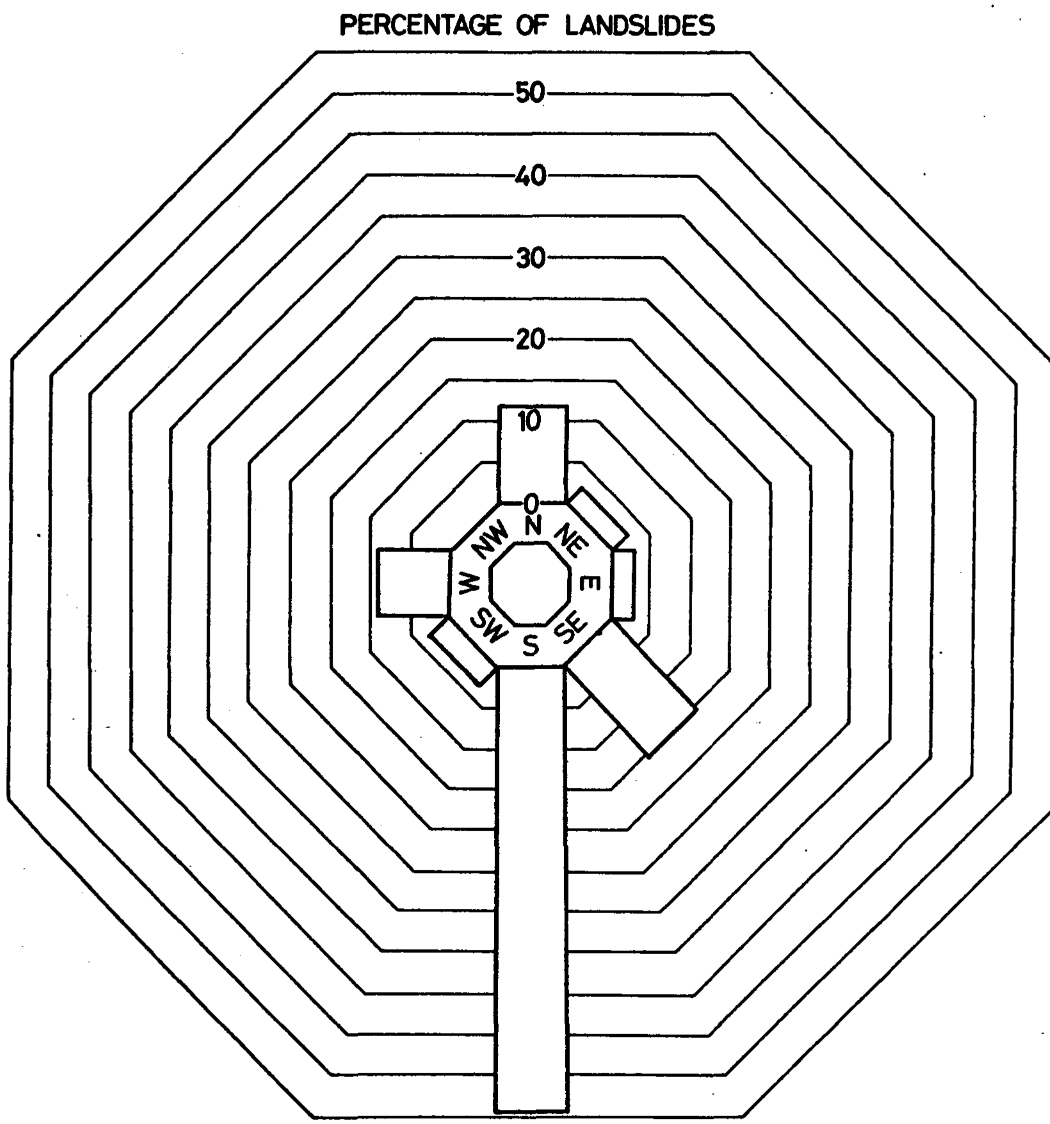


FIGURE 2.8

ORIENTATION OF LANDSLIDES SOUTH OF EVELYNS RANGE

PERCENTAGE OF RIDGE CRESTS

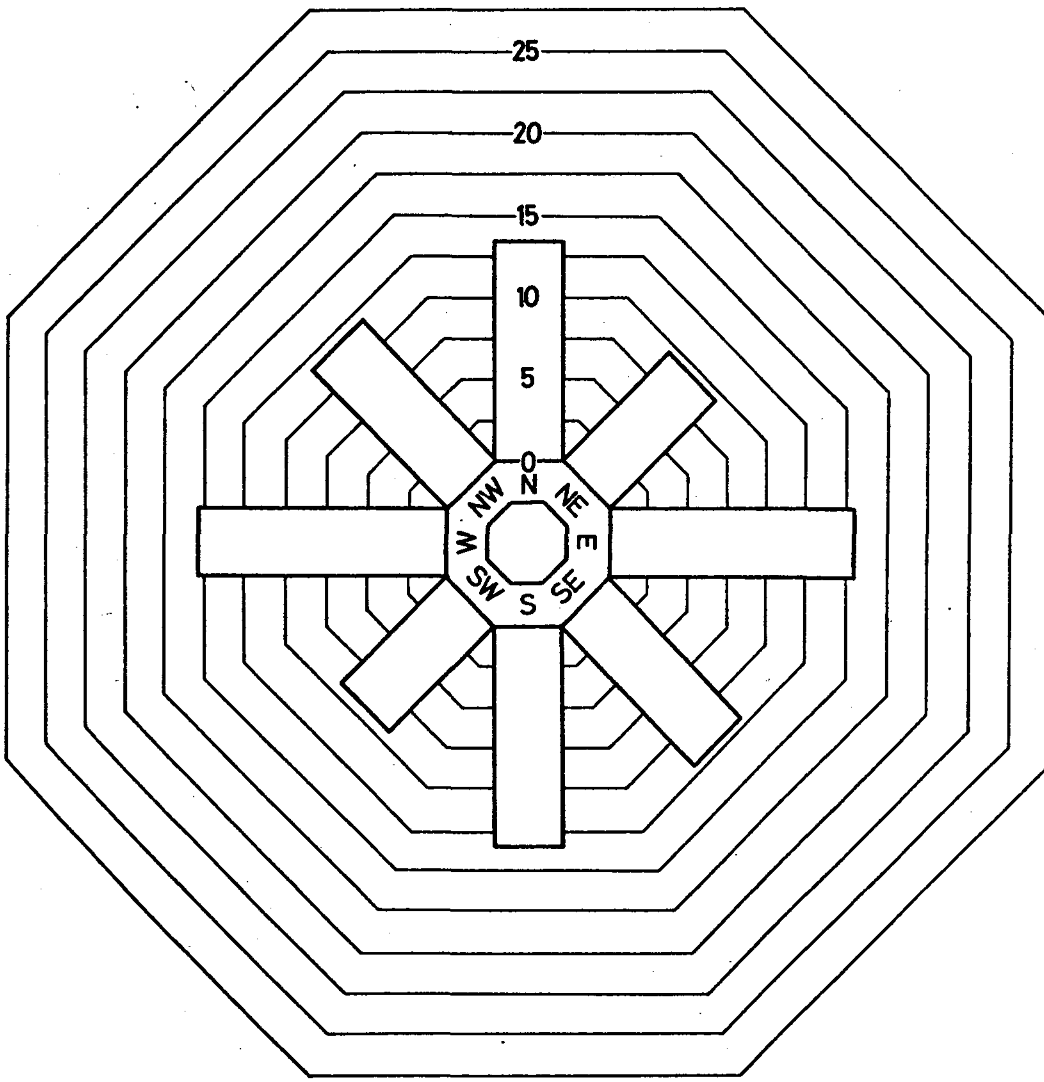


FIGURE 2.9

ORIENTATION OF RIDGE CRESTS SOUTH OF EVELYNS RANGE

Table 2.6

Orientation of landslides and ridge crests in the Razorback Range area

Quadrant	Number of Slides		Number of Ridge crests	
N	25	(8.2%)	40.5	(13.3%)
NE	30	(9.9)	47	(15.5)
E	37	(12.2)	31	(10.2)
SE	55	(18.1)	33.5	(11.0)
S	62	(20.4)	40.5	(13.3)
SW	44	(14.5	47	(15.5)
W	38	(12.5)	31	(10.2)
NW	13	(4.3)	33.5	(11.0)
	N=304		N=304	

The possibility of structural control is consistent with the stability analysis of layered systems developed by Henkel (1967), according to which slopes which receive water through permeable layers dipping away from the outlet (or slope) will become stable at a lower angle than those in which the permeable layer dips toward the outlet (or slope). Hence in the study area if water flows through the sandstone beds into the soil mantle, more frequent flows might be expected on south-facing slopes, since the sandstone beds dip away from these slopes, which are therefore less stable than the north-facing slopes. In this way the concentration of flows on south and southeast-facing slopes is explained by the opposite north-and northwesterly dips of the sandstone beds.

Microclimatic control (as suggested for landslides elsewhere by Beaty, 1965 and 1972) is also consistent with the distribution of flows, although more data would be required to check this relationship. The effect of microlimate would be to alter the soil moisture regime, tending to make south-facing slopes permanently moister, so that they could be saturated more frequently than other slopes. This could not be due to wind effects, since the principal drying winds in this area are the westerlies (Huston, 1953); however, it has been noted that in the Sydney region south and southeast-facing slopes are in shadow for part of the day, and hence remain moister (Pidgeon, 1937). Here also the correlation with flow distribution is very good, although actual climatic data for the study area are not available.

The pattern of earthflow distribution might also relate to the dominant storm-track directions, as suggested by Fisher and others (1968); however, no data on this were available for the study area.

Insufficient data were available to decide between these possibilities; the distribution of earthflows is considered to be important in terms of slope development, although no definite reason for the observed pattern can be advanced. In view of the very gentle geologic dip, the hypothesis of microclimatic control appears to be more likely.

2.9.2 Causes of earthflows in the Study Area

Almost all writers on landslides, and in particular earthflows, stress the correlation which may be observed between the occurrence of periods of intense rainfall and of major landslide activity. Examples are found in the work of Zaruba and Mencl (1969), Hanlon (1952), Walker (1963), Bailey and Rice (1969), Beaty (1956), Berry and Ruxton (1961), Crozier (1969), Engelen (1967), Fisher and others (1968), So (1971), Rice and Foggin (1971), Bailey (1967), Prior and Ho (1972), Prior and Douglas (1971), Vargas and Pichler (1957), Wright (1966), and others. Such a relationship of earthflow occurrence to periods of heavy rainfall is evident for the study area also; an example, during the heavy rains of 1950, has been documented by the Department of Main Roads (1952).

Terzaghi (1950, p.91) has shown that the production of landslides by heavy rainfall connot be related to any "lubricating" effect, since only a very thin film of water is required to reduce the inter-grain friction to its minimum valve, and sufficient moisture is generally present in soils to achieve this.

Sharpe (1938, p. 85), Terzaghi (1950, p. 91), Crozier (1969), Hanlon (1958), and Bailey and Rice (1969) have referred to "overloading" (ie, the increase in weight of the soil due to the additional water) as a possible factor contributing to landslide occurrence. Hacker (1940, p.275) has rejected this suggestion on the grounds that "... the friction increases at the same rate as the pressure." However, Hacker appears not to have understood the meaning of the Coulomb-Terzaghi equation (Section 3.2).

Considering a mass of soil on a slope inclined at angle β to the horizontal, it is clear that the component of weight normal to the slope is proportional to the cosine of β, while the component parallel to the slope is proportional to the sine of β. Now whilst the shearing stress is directly proportional to the downslope component of the weight of the soil, the strength of the soil mass (C remaining constant) is proportional to $\sigma \tan \phi$, as given by the Coulomb-Terzaghi equation, and

ϕ (the normal stress) is proportional to the cosine of the slope angle, β, as shown above. Hence we have -

shearing stress: increases as $\sin \beta$

shearing resistance: increases as $\cos \beta \tan \phi$.

Evaluating these relationships for various value of ϕ and β (Figure 2.10) shows that stress-strength behaviour as suggested by Hacker (1940) only occurs when β is of approximately the same value as ϕ; at ground slopes less than ϕ, the increase in strength due to additional weight more than counterbalances the increase in shear stress; however, when β exceeds ϕ, the increase in shear stress is greater than the increase in strength, and is thus potential cause of landsliding. It is therefore considered that the increase in weight of soil as a result of heavy rain is a possible cause of earthflows in the study area.

Heavy rainfall may tend to produce earthflows in a number of other ways. In unsaturated soils, the surface-tension of the water film between grains imparts considerable cohesion, which is lost upon saturation (Terzaghi, 1950; Terzaghi & Peck, 1948), hence reducing soil strength. Flow of water downslope through the soil exerts an additional downslope stress on the soil particles; and finally, the water entering the soil causes a rise of the piezometric surface, which leads to an increase in the pore-pressure, u, and a decrease in soil strength, according to the Coulomb-Terzaghi equation. This arises largely because part of the weight of the soil is borne by the pore fluid (Varnes, 1958, p. 38). A complete analysis of these last two effects will be referred to in a later chapter; the mathematical relationships are given in A Appendix A.

It has also been pointed out (Ward, 1945; Richter, 1958, pp. 123-124) that landslides may be triggered by earthquakes. As has been shown (Section 2.7), the study area is far from inactive seismically, and it is possible that seismic activity may have triggered some earthflow movements in association with periods of intense rainfall.

2.10 ENVIRONMENTAL CHANGE WITHIN THE STUDY AREA

2.10.1 Climatic Change

It has already been noted (Section 2.4.1) that a significant climatic change may have occurred in the study area in historical times; no information as to previous changes is available; however, Walker (1962)

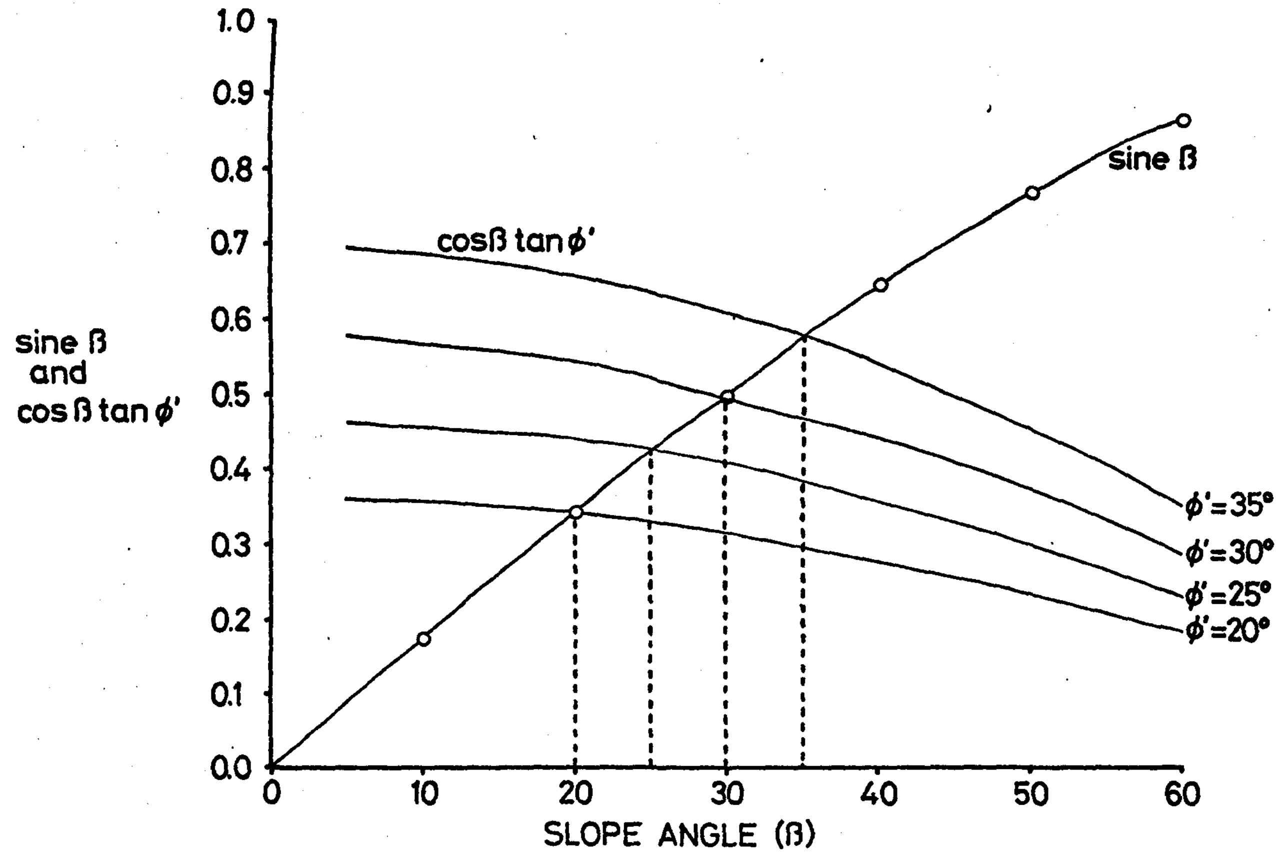


FIGURE 2.10
RELATIONSHIP OF NORMAL AND TANGENTIAL
STRESSES FOR VARIOUS VALUES OF ß AND ϕ'

has postulated the occurrence of a period of dry climate, associated with the deposition of alluvium in stream channels (such as that mapped in the study area); however, the precise age and nature of the change are not known. Evidence to be presented later shows that if the study area has been affected by major climatic changes during the period in which the hillslopes have been evolving, then this either had little effect on hillslope form, or else the effects were rapidly hidden on establishment of the present climate.

2.10.2 Vegetation Change

It was noted in Section 2.5 that the vegetation of the study area has been altered by clearing; however, no indication was given of the magnitude of the change. Pidgeon (1937) has studied the Eucalypt solerophyll forests of the Sydney region, and notes that except on sandstone, where it is interrupted by rock outcrops, the undergrowth of herbs and grasses forms a continuous ground cover. On the Wianamatta shales the ground cover is principally herbaceous. Pidgeon (1937, 1941) has identified an association, characterised by *Eucalyptus hemiphoia* and *E. tereticornis*, on the Wianamatta shales, which occupy a low-rainfall area of the Sydney basin. Referring to this association (which originally must have covered the study area), she notes from remnanta of the original vegetation that ".. it seems fairly certain that the area was originally thinly forested" (1937, p. 335), and that in the drier areas (as around Picton), it would have approached a woodland tree-spacing. Pidgeon (1937) emphasised that the *Eucalyptus hemiploia - E. tereticornis* association was not a true forest; in 1941 she described it as having a "parkland" appearance.

This description of the vegetation agrees with that of Cambage (1924, p. 211), and it is therefore reasonable to conclude that under the present climate, the study area has always possessed a continuous grassy ground cover, and has never been thickly wooded.

2.10.3 Surface Process Changes

In view of the above conclusions it may be deduced that the study area has been affected by surface processes of essentially the same intensity and character for a considerable period of time. Soil moisture conditions may be considerably different underdense forest and under grassland (Ovington, 1965, pp. 126-127), the soil generally being less moist under forest; and it has been found that forest clearing can promote slope failure by mass-movement, partly through loss of the mechanical re-inforcement of the forest root system (Gray, 1970). However, the presence of even a dense forest cover does not prevent the occurrence of landslides (Sharpe & Dosche, 1942; Waltz, 1970;

Galpin, 1940; Jackson, 1966). Certainly, slopewash under the natural vegetation would have been similar in intensity to that occurring today.

It therefore seems that the suggestion that the earthflows at present active around the Razorback Range and the study area have always been a natural process in the formation of the landscape (made by Hanlon, 1958; Cambage, 1924, and Huston, 1953) is supported by the evidence presented above that the natural vegetation was not markedly different in type from that which exists today, although it may be that earthflows occur more frequently today than prior to settlement.