PERIGLACIAL FEATURES IN THE VICINITY OF TIFFINDELL SKI RESORT, NORTH EAST CAPE DRAKENSBERG, SOUTH AFRICA, AND THEIR IMPLICATIONS FOR THE DEVELOPMENT OF THE RESORT.

THESIS

Submitted in fulfilment of the

requirements for the Degree of

MASTER OF SCIENCE

of Rhodes University

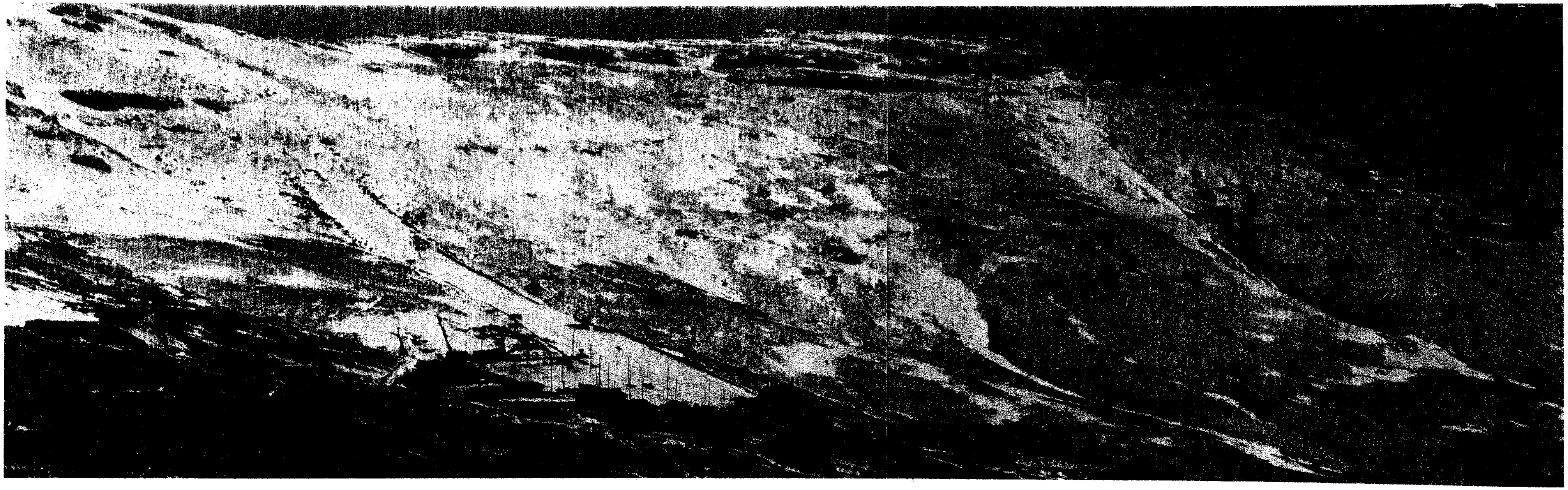
by

KAREN M. KÜCK

Department of Geography,

Rhodes University, GRAHAMSTOWN, South Africa

December 1996



Tiffindell Ski in the East Cape Drakensberg (July 1996).

ABSTRACT

This thesis provides a description of the periglacial environment and features in the vicinity of Tiffindell Ski resort, on the slopes of Ben MacDhui (3001.2m), the highest point of the East Cape Drakensberg, South Africa. Active and inactive periglacial features were located, mapped and described. Of particular interest were periglacial slope deposits including gelifluction turf-banked lobes and stone lobes, and cryoturbation features including polygons and thufur. Local environmental factors, such as aspect, moisture, topography, soil texture and depth of freezing, appear to act as important controls on the spatial distribution of the periglacial features.

Identification and quantification of periglacial processes in the regolith was investigated using temperature and soil moisture sensors coupled to dataloggers. Research was undertaken over a 16 month period from June 1995 to September 1996 so that comparisons between the winter conditions of 1995 and 1996 could be drawn. The Tiffindell area was observed to be characterised in the winter months by diurnal 'freeze-thaw days', as well as by 'ice days', 1996 experiencing colder temperatures than 1995. With more than 78% of the days from May to September 1996 being 'ice days', and simultaneously experiencing high soil moisture contents, freezing penetration to a depth of greater than 0.2m was observed to occur in the Tiffindell area, causing frost heave and gelifluction.

The summer thaw of ice lenses that developed in the cold winter months caused surface movement downslope of gelifluction lobes of up to 39mm over an 18 month period, although movement declined rapidly with depth and was essentially restricted to the uppermost 130mm of the regolith. Other features such as sorted and non-sorted polygons and thufur were identified and found to be active under the present climatic conditions and depth of frost penetration at Tiffindell.

Stone lobes were identified on the south and southeast-facing slopes at Tiffindell, but are apparently inactive under present climatic conditions. Their existence suggests the presence of severe seasonal frost in the past.

The implications of the air and ground surface temperatures, and of seasonal frost penetration for the development of Tiffindell Ski resort were considered, and suggestions regarding their economic significance are presented.

ACKNOWLEDGEMENTS

Thanks are due to those who willingly spent many days with me on the slopes at Tiffindell. Thank you Warren Scholtemeyer for all your continued support, although I demanded so much from you. To Pete Illgner for his input, both on site and in the lab.

Thank you to all those from Tiffindell Ski for their co-operation and enabling this thesis to be possible. Special thanks go to Ivan van Eck for encouraging this study, as well as always ensuring everything I needed was always available. To Martin Olckers for "babysitting" my loggers while I was not there. To 'Kaptein' and the others who helped with arduous field work snow, rain or shine.

Thanks to all those who helped with supplying visual evidence and allowed their photographs to be included in this thesis. To Dan Lieberman for the macrophotography, Gwyn Jones, Professor Colin Lewis and Warren Scholtemeyer. Thanks go to Rycherde Walters for all his help in developing the black and white photographs.

I would also like to acknowledge Ivan Hansen for allowing me to use the contour map of Tiffindell. Also Stuart Piketh for use of the WITS/CSIR meteorological data. And Lenny Olyott for his help with plant identifications.

Special thanks go to Robyn Cretchley for helping me with the statistics and the graphs. Your help was invaluable. Thanks also for helping me through the 'final stages' of production, and helping me to stay sane!

I am deeply indebted to Professor Colin Lewis for his advise and guidance. Also to my parents and Warren: thank you for your endless encouragement and support. I could not have done this without you all.

Thank you to the Rhodes University Geography and Hydrology departments for the loan of equipment and which was vital to the project. Also to Rhodes University for financial support.

INTRODUCTION CHAPTER 1

1.1. RATIONALE

The mountains of the East Cape and Natal Drakensberg and of the adjacent areas of Lesotho rise to an elevation of over 3 000m, and form the "largest and highest mountain mass in Southern Africa" (Deacon and Lancaster, 1988). This thesis examines and describes the distinctive periglacial features existing at and adjacent to Tiffindell Ski Resort in the vicinity of Ben MacDhui, East Cape Drakensberg, especially gelifluction lobes, thufur and patterned ground phenomena. These landforms have already been shown to be well developed in various high altitude areas of the Drakensberg range of South Africa and in surrounding mountainous areas by a number of workers (Sparrow, 1967a; Harper, 1969; Marker and Whittington, 1971; Hastenrath and Wilkinson, 1972; Sänger, 1988; Hall, 1991a; Boelhouwers, 1991a, 1991b, 1994; Grab, 1994; Lewis, 1988a&b, 1990, 1994a, 1996), their regional and altitudinal distributions being comprehensively discussed in Chapter 2.

The distribution of periglacial features reflects the interplay of microclimate, soil depth and type, vegetation cover, and aspect in an area where today the July mean temperature is around 3°C and frost occurs on 90-150 days of the year (Harper, 1969). The slopes of Ben MacDhui display a variety of periglacial features (Lewis, 1996), the majority of them appearing to be currently active, and reliant on the combination of lithology, climate and other aspects mentioned above. A comparison of the nature and of the controls to which these features at Tiffindell are subjected with others occurring in southern Africa and abroad, may assist in the accurate identification of the features, as well as their relative activity, and (in the case of gelifluction lobes) their mass wasting potential.

It is widely recognised in areas outside Africa that an understanding of periglacial processes and conditions is essential when planning upland development in cold climate areas (Carson and Kirkby, 1972; Williams, 1982; Sidle *et al.*, 1986; Burn, 1991; King *et al.*, 1992; Haeberli, 1992). Experience elsewhere in the world has shown that roads, buildings, pipelines and other man-made structures may be adversely affected by periglacial conditions and the effects of freezing temperatures in the ground (Williams, 1982), and therefore need to be designed for the periglacial conditions of the area in which they are established. This is currently of great geotechnical interest to engineers and development planners, and the need for this awareness by South Africans is continually growing. During the 1990's major economic developments were initiated in periglacial areas in Lesotho (particularly the Lesotho Highlands Water Project) and in South Africa (particularly Tiffindell Ski resort). Therefore, there is a growing need for southern African geomorphologists to study periglacial features and the processes involved in their formation.

The entrepreneurs of Tiffindell Ski are aware of the importance of the region's climate, as well as the regolith and the processes operating within it, for the development and success of the resort. They therefore invited this study of the climate and environmental conditions at both ground level and within the ground (the regolith) and the processes operating on the slopes of the resort and those adjacent.

Although over the past 30 year period, the location of periglacial features has been studied in southern Africa, and suggestions have been made for the origins of these features, empirical data is scarce. This study provides a new set of data regarding several periglacial features and processes occurring in one particular region in southern Africa, and thus contributes to the knowledge of the periglacial environment in this area.

1.2. THE STUDY SITE

The Ben MacDhui area is situated in the East Cape Drakensberg, South Africa, and falls within the middle latitudes: 30°39'S; 27°55'30"E. Tiffindell farm is located on the slopes of Ben MacDhui, which peaks at 3 001.2m, and is the highest mountain in the East Cape Province. Tiffindell Ski resort, the lodge of which is at an altitude of 2720m, is in an extremely remote part of South Africa, being approximately 22km from the nearest village, Rhodes. Rhodes is located in the Bell River valley at approximately 1800m above sea level, at the southern edge of the Drakensberg, and is approximately 796km from Johannesburg, 716km from Durban and 1105km from Cape Town (Figure 1.1).

Tiffindell farm covers an area of approximately 1 300 hectares (Science Applications[cc], 1993), the former-Transkei forming its Northern boundary. The ski resort is located on the south facing slopes of the Ben MacDhui - Breslin's Kop ridge. The ridge forms a watershed between drainage to the north, forming the Telle River system, and to the south forming the Bell River system.

The study area for this research commences at the level of the resort itself, that is 2720m, and incorporates the rest of the valley up to Ben MacDhui at an altitude of 3001.2m a.s.l., as well as beyond the border down into the former-Transkei (altitudes across the border ranging between approximately 2910 and 2930m). The area of the site is approximately 119 hectares (Appendix I).

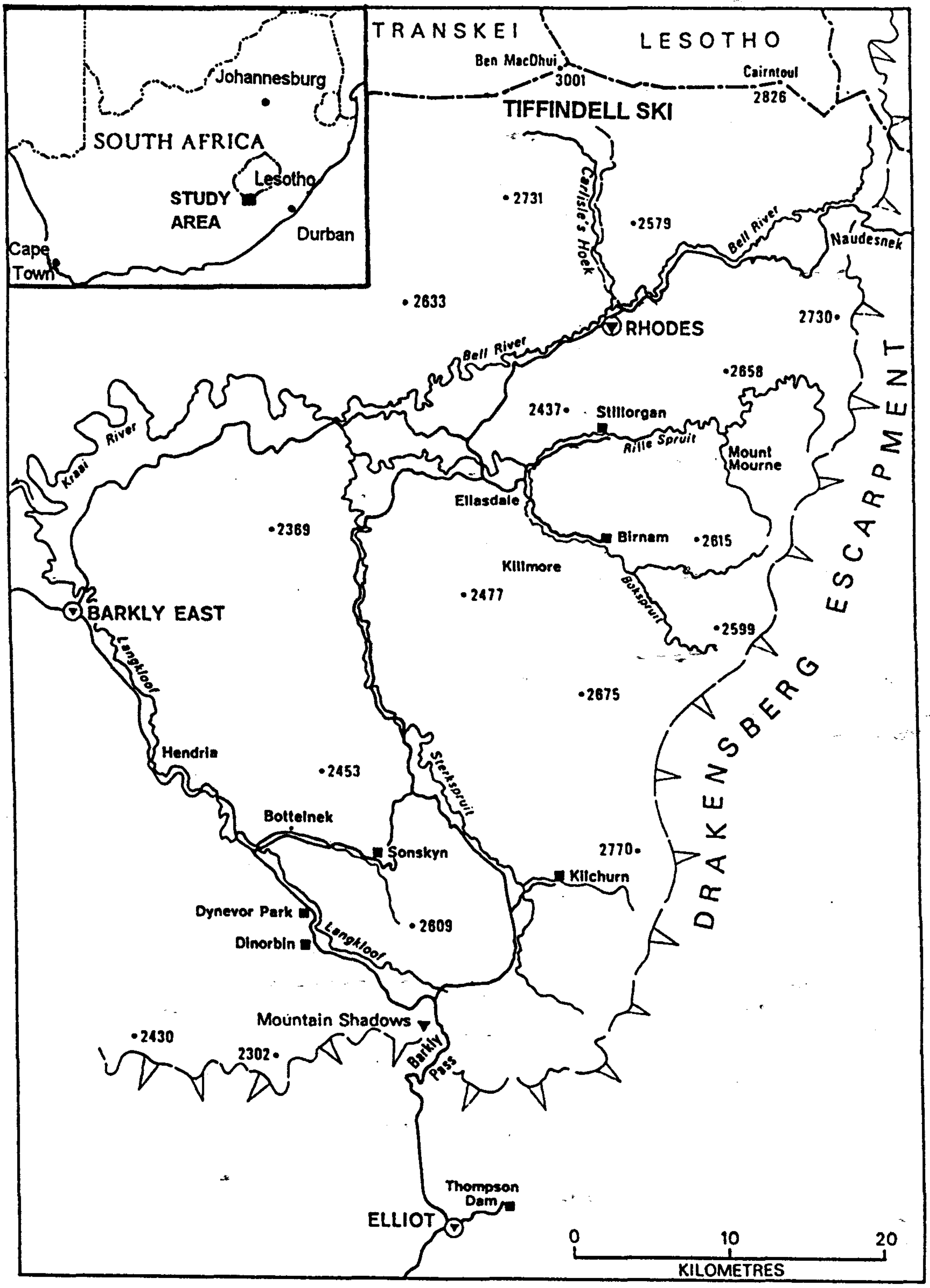


Figure 1.1: ***Location map illustrating the position of Tiffindell Ski within South Africa and the East Cape Drakensberg.***

(from Lewis, 1994a)

1.3. OBJECTIVES

The primary objective of the study was to determine whether periglacial features and processes occur in the Tiffindell area of the East Cape Drakensberg. Secondary objectives were:

1. To locate, map and describe the various periglacial features present on the south and southeast-facing slopes of Ben MacDhui, on the Ben MacDhui - Breslin's Kop ridge and in adjacent areas of the former Transkei.

2. To assess the present state of activity of the various features; and if found to be active, to assess the conditions and/or physical processes that may play an important role in determining the activity.

3. To determine the controls on the spatial distribution of specific features located in the area.

4. To examine and assess the implications of the various active periglacial processes on the development of the resort; as well as attempting to infer implications on other developments in the vicinity.

5. To assess possible anthropogenic effects due the development of the resort on the environment.

1.4. SPECIFIC AIMS

In order to accomplish the objectives, specific aims were outlined:

1. To gather data on the present climatic conditions of the area, as well as site-specific microclimatic data; to investigate the day-to-day conditions and/or

physical conditions that may play an important role in determining periglacial activity.

2. To research the relevant literature, which would form a basis on which the Tiffindell features could be assessed.
3. To produce geomorphic maps of the study site, which display the locations of the features under research.
4. To examine the processes occurring at the various sites of periglacial activity.
5. To provide visual evidence of periglacial processes in the area, which will clearly illustrate what the text describes.
6. To conduct sediment analyses on samples taken from representative periglacial features identified within the study site.
7. To relate the climatic data collected in the area with the periglacial features in the Tiffindell region.

PERIGLACIAL ENVIRONMENTS AND THE PERIGLACIATION OF SOUTHERN AFRICA

CHAPTER 2

This chapter provides a broad outline of periglacial and mountain geomorphology in order to set a base upon which the subsequent descriptive and process work within the Tiffindell area can be compared and assessed. A broad summary of the periglacial features found elsewhere in southern Africa allows for comparisons between these high altitude areas.

2.1. INTRODUCTION

Geocryogenic processes are currently active at high altitudes in Southern Africa and there is evidence for their more widespread activity during the Pleistocene (Corté and Hall, 1991). Southern Africa has a range of environments, processes, landforms and sediments, both current and fossil, that are associated with cold climates - otherwise referred to as periglacial environments.

2.1.1. Introduction to periglacial geomorphology.

Geomorphology can be defined as the scientific explanation of natural landforms, or assemblages of landforms, collectively known as the landscape. Periglacial geomorphology is then, according to Thorn (1992:2), used to "...identify a definable subset of processes, landforms, or landscapes ... which exhibit distinctive and substantive attributes".

The term *periglacial* was first proposed by Loziński (1910) to describe features that originate under the predominant influence of frost action, and has since been modified to refer to a wider range of non-glacial processes and features of cold climates *regardless of age or proximity to glaciers* (Washburn, 1973). In fact, the present definition is

> *"The conditions, processes and landforms associated with cold, nonglacial environments"*
> (Harris *et al*, 1988:63).

A true 'periglacial' area is one which has *not* been glaciated but has been subject to cold-based processes *without* the pre-conditioning or provision of debris by glaciers (Hall, 1991a). Formerly glaciated areas may, however, have been subjected to periglaciation, as in Britain (Ballantyne and Harris, 1994) and Ireland (Lewis, 1985). According to Washburn (1973), the periglacial environment is typically associated with a climate sufficiently cold as to leave physical evidence of its influence, and is therefore said to be dominated by frost action processes, that is, the alternate freezing and thawing of ground water. The frequency and amplitude of these freeze-thaw cycles may vary and give rise to different types of periglacial processes or different intensities of operation, but the occurrence of such cycles is imperative (Thorn, 1992; Ballantyne and Harris, 1994). Frost processes occur at intensities sufficient to generate detectable forms (results) within the majority of the highland areas of South Africa (Lewis, 1988a,b).

Within periglacial areas, the significance of the effects of frost action (that is, frost-heave and freeze-thaw processes) varies with climatic conditions and material types (Goudie, 1993). The most hazardous zone is that in which winter temperatures are low enough for freezing to occur to a great depth, and summer temperatures are high enough for thaw to take place, that is, for the development of an *active layer*. The active layer refers to the layer of ground which is subject to annual freezing and thawing, that is the ground that is frozen in winter and thaws in summer, and so does not remain permanently frozen (Hall, 1991a). In many areas of the world, the active layer is underlain by *permafrost*, which is subsurface material in which the temperature remains below 0°C for at least two consecutive years (Harris *et al.*, 1988; Brown and Péwé, 1973 cited in King *et al*, 1992:73). Yet, there is no perfect correlation between intensive frost action and permafrost, and locations which experience freeze-thaw oscillations for all or part of the year frequently exist without the presence of continuous permafrost (Washburn, 1973). As a result, the periglacial environment can be divided into two contrasting regions: the continuous permafrost of the polar periglacial areas; and the alpine areas of lower latitudes subject to freeze-thaw cycles, *seasonally frozen ground* (frozen ground that melts in summer) and also possible sporadic permafrost.

Determining a concise definition of what constitutes a periglacial environment, and the associated landforms, is the subject of much debate. Thorn (1992:1) provides a comprehensive definition of a periglacial environment, and thus periglacial geomorphology:

> *"Periglacial geomorphology is that part of geomorphology which has, as its primary object, physically based explanations of the past, present, and future impacts of diurnal, seasonal, and perennial ice on landform and landscape initiation and development."*

This definition provides a framework for discussion by unifying the geomorphic role of ground ice, as well as the interaction between ground temperature and ground moisture. It also addresses the question of the relative significance of ground ice processes in the development of cold region landscapes.

According to Washburn (1973), in addition to frost action, certain aspects of mass wasting, nivation, fluvial action, marine action and wind may also contribute to the production of periglacial features. Therefore the respective contributions of periglacial and non-periglacial processes to landform development need to be assessed. However, the distinction between landform initiation processes and landform growth (or maintenance) is often complex. It is important to remember that any landform labelled periglacial may be periglacial only in origin, growth, or maintenance; or may be periglacial throughout its development (Thorn, 1992).

2.1.2. Environmental factors for periglacial environment development

Several important factors govern periglacial environments and processes, as well as the development and distribution of frozen ground. The most influential of these include the topography and regional climate, their degree of control being dependant on scale. Three major locational factors determine the characteristics of global climates in general, and of periglacial climates in particular. These are latitude, altitude and continentality (Ballantyne and Harris, 1994). Individually, and in combination, these factors account for much of the range in present-day periglacial climates.

As atmospheric temperatures fall with height (known as the lapse rate, generally around 0.6°C per 100m rise), distinct contrasts in mean annual temperatures exist between lowland and upland regions at similar latitudes, and cold climates are known to occur in mountainous regions, even at low latitudes. Temperature differences are also noted between areas subjected to maritime influences, and the continental interiors. Maritime influences tend to modify the temperature extremes and the result is generally a subdued seasonal climate (Tyson, 1986).

Local factors will also modify the regional climate and strongly influence periglacial processes. Tricart (1969) recognised various periglacial climates, and provided one of the most useful classification systems:

A. *Dry with severe winters*: associated with permafrost, little or negligible running water, and strong wind action.

B. *Humid with severe winters*
   1. *Arctic variety:* associated with less permafrost, many more freeze-thaw cycles than A, and much more snow, which provides protection from wind in winter and abundant snowmelt in summer.
   2. *Mountain variety:* associated with frequent frost action, more running water, and little wind effect because of higher precipitation and more irregular topography.

C. *Cold with little seasonal temperature change*

1. *High-latitude island variety:*
2. *Low-latitude mountain variety:* both (1) and (2) are associated with abundant short-period frost cycles penetrating only to shallow depths and with reduced wind effectiveness because of high humidity. Absence of permafrost.

Therefore the following contrasts can be deduced from the above classification (Davies, 1969). Permafrost is characteristic of A-type climates; infrequent occurrence of permafrost is associated with B; and permafrost is absent in C. Climate A experiences seasonal, high amplitude frost cycles, penetrating to great depths, while C experiences diurnal, low amplitude frost cycles affecting only shallow depths. Freeze-thaw cycles are much fewer in A than C. B is typically intermediate. Running water is much less important than wind for climate A; yet for climates B and C, the reverse tends to be the case.

The overall influence of climate is dependant on the scale of operation, that is zonal, local or microclimates. Zonal climates reflect latitudinal and altitudinal criteria, local climates are a function of the local topography (aspect), and the microclimate incorporates the influences of ground surface characteristics such as vegetation and moisture.

Topography influences climate, as outlined above, and also exerts a direct effect on periglacial processes (Washburn, 1973). For example, the configuration (manner of arrangement; shape; outline) and gradient of a hillside can determine the dominant mass-wasting process (*e.g.* landsliding, solifluction or frost creep). A very low angle favours retention of moisture and development of certain forms of patterned ground.

Bedrock is essentially an independent factor in relation to periglacial processes, since it is usually a pre-existing, older feature.

Other environmental factors such as snow cover, soil moisture and vegetation can be critical controls of periglacial processes, especially of frost action. The amount and distribution of snow cover are functions of climate and topography, the effect of the latter being important in the distribution of snowdrifts (Washburn, 1973). The insulating effects of snow cover moderate the ground temperature and moisture contents in the soil.

The amount of liquid moisture and its seasonal distribution are functions of climate, topography and the parent rock material. Climate exhibits large scale control, topography tending to modify the climatic influences, the rock material determining the amount of moisture entering and remaining in the ground.

Vegetation, like soil moisture, is also a function of climate, topography and rock material. Zonal climate sets the broad vegetation patterns, but a strong relationship exists between vegetation and the local or microclimate. Vegetation influences the amount and effects of periglacial processes through its insulating and binding of the soil. The height of the treeline (altitudinal limit of trees regardless of species) is considered as a lower boundary for the alpine climatic zone (Ballantyne and Harris, 1994), and typically acts as a critical boundary for certain periglacial processes (Washburn, 1973).

2.1.3. Features associated with periglacial environments

As periglacial areas are characterised by geomorphic features and geological structures that form under the varying influence of frost action, the features associated with cold climates are typically highly varied. The formation of periglacial phenomena are dependent on climatic conditions (Washburn, 1973; Thorn, 1992; Ballantyne and Harris, 1994), as all geomorphic processes are effectively controlled by frost action and the presence of frozen ground, but are often independent of the presence of permafrost. Some of these features form landforms and others, although reflected on the surface, are well developed as subsurface structures (for example, ice wedge casts; Lewis, 1977). Periglacial features may be classified according to their mode of formation and present or fossil morphology. In South Africa, they can be broadly divided into periglacial slope processes and deposits (including mass movement phenomena such as gelifluction lobes and terraces, blockstreams and stone lobes (garlands); Lewis, 1988b; Marker, 1995a) and cryoturbation phenomena (such as features formed by needle ice activity and patterned ground, including stone polygons and thufur; Lewis, 1988b; Marker, 1995a), nival features (such as protalus ramparts; Lewis 1994b, 1996), and possibly permafrost creep features (notably rock glaciers; Lewis and Hanvey, 1993).

Cryoturbation phenomena in particular include a number of different features, and are useful in showing that frost action has, at some time or other, occurred beneath the ground surface. However, each individual type of periglacial phenomenon is dependent on specific climatic and edaphic conditions (Karte, 1983). Table 2.1 summarises Karte's conclusions. Therefore, because of this dependence on specific environmental conditions, periglacial phenomena may also be considered as diagnostic indicators for such conditions (Karte, 1983; Thorn, 1988).

PERIGLACIAL PHENOMENA	INDICATION OF MEAN ANNUAL AIR TEMP.	EDAPHIC CONTROL AND INDICATION
Thermal contraction features:		
Ice wedge polygons	<-4° to <-8°C	Fine-grained sediments with high moisture contents; poorly drained sites
Seasonal frost crack polygons and micro non-sorted polygons	<0° to -4°C	Silty to loamy, frost susceptible sediment; moderately well drained
Frost mounds:		
Tundra hummocks	<-10°C	Fine grained silty, frost susceptible; seasonally saturated sediment
Earth hummocks, thufur	<+3°C	
Seasonal frost mounds	<+1° to <-3°C	Fine grained; water permeable
Rock glaciers	<+2° to 0°C	Angular bedrock debris with fines
Sorted patterned ground (small forms)	<+1°C	Frost susceptible fine sediments; moderately moist
Microforms of geli-solifluction	<-2°C	Silty to loamy sediments and debris

Table 2.1: Summary of climatic and edaphic conditions required for periglacial phenomena formation (from Karte, 1983)

2.2. THE PERIGLACIATION OF SOUTHERN AFRICA

Active periglacial phenomena exist at altitudes above 2 550m in the KwaZulu-Natal (usually referred to as 'Natal') and East Cape Drakensberg mountains and adjoining areas, and at 1800m and above on the Matroosberg in the western Cape, and have been described on numerous occasions by a number of workers. Active periglacial slope phenomena have been described (*e.g.* Marker and Whittington, 1971; Lewis, 1988a,b, 1996; Marker, 1989, 1990a; Boelhouwers 1991a&b, 1994, 1995), as well as active cryoturbation features (*e.g.* Harper, 1969; Hastenrath and Wilkinson, 1972; Dardis and Granger, 1986; Grab, 1994, 1996). Several of the above mentioned workers have described both periglacial slope and patterned ground phenomena that are currently active, as well as those that are considered inactive under present-day climatic conditions (*e.g.* Hastenrath and Wilkinson, 1972; Nicol, 1973; Lewis and Dardis, 1985; Lewis, 1988a,b; Lewis and Hanvey, 1993; Lewis, 1996).

Due to the dependence of periglacial processes on climatic conditions, numerous attempts have been made to define periglacial processes using climatic parameters. Peltier (1950), employing the variables of mean annual temperature (MAT) and mean annual precipitation (MAP), suggested a general classification scheme to indicate where periglacial activity occurs, that is, where the MAT is below 0°C and the MAP in the range 0-1500mm. Lewis (1988a) extrapolated data from Mokhotlong and Thaba Tseka (Lesotho) and Cathedral Peak (KwaZulu-Natal Drakensberg) to indicate the geomorphic processes, based on Peltier's scheme, that probably occur at present-day at high altitudes within South Africa and Lesotho (Figure 2.1).

The annual average temperatures of all three stations (which are at or above 2000m) were extrapolated by Lewis (1988a) using a lapse rate of 0.6°C (ignoring the local site effects), and the stations' original recorded precipitation values were used without any changes. From this climatic data, Lewis suggested that frost action, mass movement, wind action and chemical weathering are all at play at various levels in southern Africa. Without the consideration of local conditions occurring in high mountain areas, however, a true reflection of the processes operating there can not be gained.

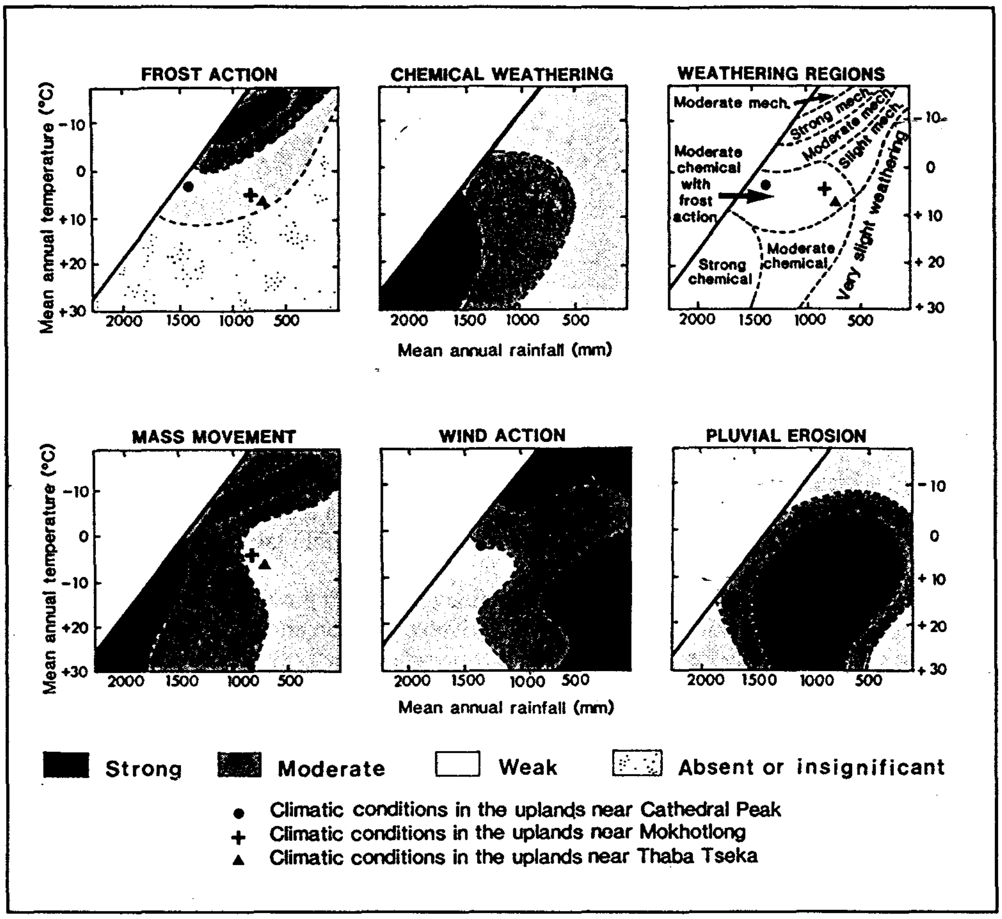


Figure 2.1: Climatic parameters for geomorphic processes and regions (after Peltier, 1950). Conditions on the uplands of Lesotho and the KwaZulu-Natal Drakensberg are indicated for the Mokhotlong, Thaba Tseka and Cathedral Peak areas.

(from Lewis, 1988a)

Within southern Africa, great diurnal ranges of temperature exist in mountain areas, especially in the winter months. At Rhodes (1800m), it is not unusual for temperature ranges of up to 30°C to occur between midday and night-time temperatures in the winter (Lewis, 1988a). The shaded south-facing slopes, especially at altitudes above 2500m, may remain frozen for two months or more over the winter months, as witnessed in the

Drakensberg range (Granger, 1976, cited in Lewis, 1988a; Lewis, 1996). It is therefore not surprising that a range of periglacial phenomena (although perhaps somewhat limited) are active at high altitudes in the area at present.

Lewis (1988b), in reviewing the nature of periglacial features in southern Africa, has suggested that subperiglacial conditions exist at present within high altitude areas. The presently active phenomena tend to be small-scale, typically occurring in upland areas that are only marginally suited to periglaciation under present conditions. Several fossil periglacial features are, however, found to exist in high altitude areas and within the sediments of lower altitudes. These inactive features generally indicate that cold climatic conditions prevailed during the Pleistocene, and allowed for the occurrence of deep penetrating frost action and nival processes. These include features such as blockfields (Boelhouwers, 1994), rock glaciers (Lewis and Hanvey, 1993) and protalus ramparts (Lewis, 1994b; 1996).

At least fifteen different features that have formed under periglacial conditions, identifiable as landforms, sediments, or features within sediments (Lewis, 1988b), have been recorded in southern Africa over the past 30 years. They are categorised as active or inactive (fossil), depending on their present state, and form an important part of the periglacial history of South Africa. A summary of those features reported before 1988 was presented by Lewis (1988b, Tables 1 and 2).

Table 2.2 below provides a summary list of these known active and fossil periglacial features up until 1996, as well as their general locations within South Africa and Lesotho.

ACTIVE PERIGLACIAL FEATURES		**FOSSIL PERIGLACIAL FEATURES**	
Feature	**Location**	**Feature**	**Location**
Solifluction or gelifluction terraces/lobes	East Cape Drakensberg Natal Drakensberg Lesotho	Stone lobes	East Cape Drakensberg
Turf-banked lobes	East Cape Drakensberg Natal Drakensberg Lesotho Western Cape	Stone-banked terraces/lobes	Natal Drakensberg Lesotho Western Cape
Thufur	Natal Drakensberg Lesotho	Rock glaciers	East Cape Drakensberg
Polygons/Stripes	East Cape Drakensberg Natal Drakensberg Lesotho Western Cape	Blockstream/ blockfields	East Cape Drakensberg Natal Drakensberg Lesotho Western Cape
		Head/ solifluction	East Cape Drakensberg Natal Drakensberg Lesotho Western Cape
		Debris cone	East Cape Drakensberg
		Ice-wedge cast	East Cape Drakensberg
		Compacted soil horizon indicating permafrost	East Cape Drakensberg Natal Drakensberg
		PROTALUS RAMPARTS	EAST CAPE DRAKENSBERG

Table 2.2: Occurrence and general location of active and fossil periglacial features identified in South Africa and Lesotho.

Based on Lewis (1988a&b)

Lewis (1988b), in using this information, indicated that freeze-thaw activity must have been sufficiently effective to form periglacial features. Boelhouwers (1991a) has described present-day periglacial activity in the Natal Drakensberg and has drawn climatic inferences from his data. He has also shown (1991b; 1995) that periglacial activity occurs at high altitudes in the Western Cape mountains. Hanvey and Marker (1992) have also described thufur and other periglacial microforms at altitudes above 3 000m in Lesotho. These descriptions, together with all other records, support the contention that subperiglacial conditions exist at high altitudes in the Drakensberg of Natal and the Eastern Cape, as well as in some other upland areas further west (Lewis, 1988a).

Lewis (1988b) also presented the distribution of the reported periglacial phenomena in southern Africa as of 1988 on maps (Lewis, 1988b, Figures 1, 4). A summary of the distribution of several reported active and inactive periglacial features in southern Africa is shown on Figure 2.2. Lewis (1988b) also produced graphs to indicate the altitudinal distributions of the features in relation to latitude and longitude (Lewis, 1988b, Figures 2,3,5,6). His analyses reveal that periglacial features in southern Africa are found primarily within the high altitude areas (1 800m to 3 500m), and with the exception of those in the Hexriver Mountains (Western Cape) are restricted to areas between 30°39'E and 27°56'E, coincident with Lesotho and the adjoining upland areas.

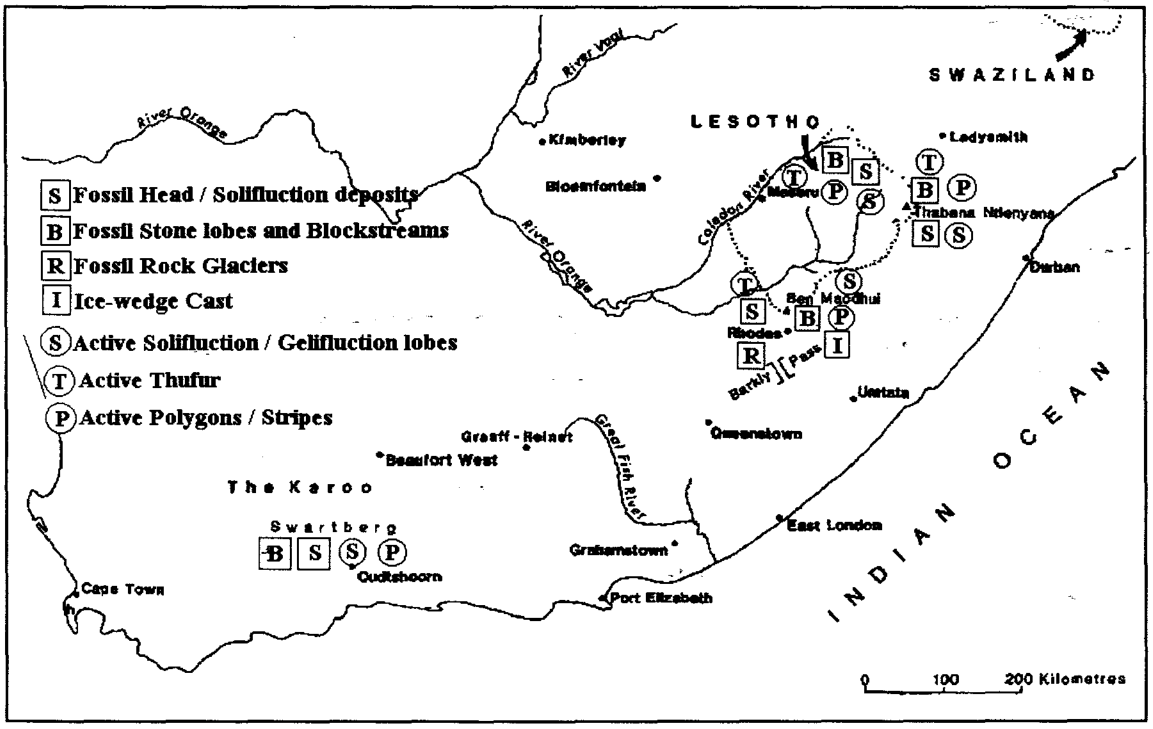


Figure 2.2 Location of active and inactive periglacial features in southern Africa. Symbols indicate the features occurring in the general vicinity of the East Cape, Natal and Lesotho Drakensberg, and the Western Cape mountains.

A summary of the relevant literature on fossil and active periglacial features occurring in southern Africa is presented in Table 2.3. This table includes only those features examined within this thesis and only those papers or personal observations that provide sufficient information to enable the locations and altitudinal distributions to be recorded.

2.2.1. Regional and altitudinal distribution of active and inactive periglacial phenomena.

Gelifluction (turf-banked) lobes, polygons (which have been noted merging into stripes) and thufur are known to be active at present in southern Africa (Table 2.2, Table 2.3). They are located in the mountains of Lesotho and in the adjacent uplands of the Drakensberg of Natal and the Eastern Cape, as well as in the Hexriver mountains in the Western Cape uplands, although at somewhat lower altitudes (Table 2.3).

The active periglacial features in southern Africa range from 1 800m to little less than 3 500m, their distribution being influenced by elevation and topography. The lobes formed by gelifluction (or in some cases solifluction) exhibit the greatest range, from 1800m to 3300m, although the lower limit for current activity is considered to be either 2000m or, more probably 2550m. Evidence in the published literature relating to the lower level is unclear and requires confirmation.

Within the Drakensberg active sorted and non-sorted polygons range from 2550m to above 3353m, although those recorded at the lower altitude differ from many others in that they are reported as occupying the floors of perennial ponds, and are non-sorted. The elevation of 2800m appears to be the lowest limit in the Drakensberg for polygons that do not occupy such a preferential position. Fossil forms are found at lower elevations in the uplands of the Western Cape (Sänger, 1988). Thufur actively occur at altitudes between 2900 and 3100m, and none are reported to occur within the lower areas of the Western Cape Mountains. Stone-banked lobe features are found to be active, yet predominantly within the Natal Drakensberg. As they may be of similar origin to the fossil stone lobes reported by Lewis (1996) at Tiffindell, they are considered within Table 2.3.

FEATURE	F/A	LOCATION	ALTITUDE	SOURCE
Turf-banked lobes/ gelifluction lobes	F	Lesotho	1680-2290m	Sparrow (1967; 1971)
	A	Sani Pass, Lesotho	2990-3109m	Marker & Whittington (1971)
	A	Lesotho	2000-3300m	Hastenrath & Wilkinson (1972)
	F	Lesotho	2700m+	Hastenrath & Wilkinson (1972)
	A	Lesotho	*c.* 3050m	Fitzpatrick (1978)
	A	Western Cape Mnts	1900m	Borchert & Sänger (1981)
	F	Barkly Pass, E Cape	1980m	Lewis & Dardis (1985)
	A	Natal Drakensberg	3000-3120m	Dardis & Granger (1986)
	F	Carlisle's Hoek, E.Cape	1800m	Hanvey, Lewis & Lewis (1986)
	F	Rhodes, East Cape D	1800m	Lewis & Hanvey (1988)
	F	Golden Gate, OFS	2000m	Marker (1989)
	A	Natal Drakensberg	3000m	Boelhouwers (1991a)
	A	Western Cape Mnts	1800m+	Boelhouwers (1991b)
	A	Hexriver, W Cape	1800-1850m	Boelhouwers (1994)
	A	Tiffindell, E Cape D	2800-3000m	Lewis (1996)
Sorted and non-sorted polygons/ micro-patterned ground	A	Lesotho	3100m	Van Zinderen Bakker (1965)
	A	Lesotho	3050m+	Harper (1969)
	F	Lesotho	3354m+	Harper (1969)
	A	Sani Pass, Lesotho	3353m+	Marker & Whittington (1971)
	A	Lesotho	3100-3300m	Hastenrath & Wilkinson (1972)
	A	Western Cape Mnts	1900m	Borchert & Sänger (1981)
	A	East Cape D	2800-3000m	Lewis (1985, pers obs)
	A	Natal Drakensberg	*c.* 3000m	Dardis & Granger (1986)
	F	Matroosberg, W Cape	1800m	Sänger (1988)
	A	Sani Pass, Lesotho	3000m+	Boelhouwers & Hall (1990)
	A	Natal Drakensberg	3000-3200m	Boelhouwers (1991a)
	A	Lesotho	3000m+	Hanvey & Marker (1992)
	A	Hexriver, W Cape	1850-1948m	Boelhouwers (1995)
	A	Natal Drakensberg	3380-3410m	Grab (1996)
	A	Tiffindell, E Cape D	2900m	Lewis (1996)
Thufur	A	Lesotho	3100m	Van Zinderen Bakker (1965)
	A	Sani Pass, Lesotho	2960-2990m	Marker & Whittington (1971)
	A	Lesotho	2990-3100m	Hastenrath & Wilkinson (1972)
	A	Sani Pass, Lesotho	3100m	Boelhouwers & Hall (1990)
	A	Natal Drakensberg	3100m+	Boelhouwers (1991a)
	A	Lesotho	3000m+	Hanvey & Marker (1992)
	A	Mohlesi valley, Lesotho	2900-3270m	Grab (1994)
Stone lobes/ stone-banked lobes	A	Natal Drakensberg	3000-3120m	Dardis & Granger (1986)
	F	Matroosberg, W Cape	1800m+	Sänger (1988)
	A	Natal Drakensberg	3150-3240m	Boelhouwers (1988, pers obs.)
	A	Western Cape Mnts	1800m+	Boelhouwers (1991b)
	A	Natal Drakensberg	3200m	Boelhouwers (1994)
	A	Hexriver, W Cape	1830-1860m	Boelhouwers (1995)
	F	Tiffindell, E Cape	2700m+	Lewis (1996)

A = Active feature F = fossil feature D = Drakensberg OFS = Orange Free State

<u>Table 2.3:</u> Active and fossil periglacial features existant at or adjacent to Tiffindell, and that have also been recorded elsewhere in southern Africa.

Other periglacial features that are no longer active have also been reported from many areas in the uplands of eastern and central Lesotho and from adjacent areas of Natal and the Eastern Cape. Those features attributed to previous periglacial conditions include gelifluction deposits (head) and patterned ground phenomena and are reported from both high and low altitudes (Table 2.3).

Due to the spatial extent of the sub-continent, as well as to the altitudinal differences from east to west and the climatic division between eastern and summer rainfall and western winter rainfall areas, periglacial features in eastern and western areas of the subcontinent are assumed to have developed under different climatic regimes.

2.2.2. Geomorphic evidence for Quaternary periglacial activity in the East Cape Drakensberg

A series of papers on fossil periglacial features in the Drakensberg of the Eastern Cape (Lewis and Dardis, 1985; Hanvey, Lewis and Lewis, 1986; Lewis and Hanvey, 1988; Hanvey and Lewis, 1990; Lewis and Hanvey, 1991; Hanvey and Lewis, 1991; Lewis and Hanvey, 1993; Lewis, 1994b) has shown that head, gelifluction, solifluction flows, rock glaciers and protalus rampart deposits exist in the area above 1800m.

Lewis and Hanvey (1991) have suggested that periglacial conditions existed at and above 1900m in the East Cape Drakensberg in the time between 24 300 and 12 200 B.P., as evidenced by solifluction fans at Glen Orchy, near Rhodes. They have further shown (1993) that rock glaciers were active, down to altitudes of about 1 800m, at around 21 000 B.P. in Bottelnek (Figure 1.1), and thus at least a discontinuous permafrost zone extended to at least that altitude. They have presented evidence of interstadial conditions in the vicinity of Bottelnek at 35 000 B.P., that were replaced by stadial conditions (in which rock glaciers were active) subsequent to 27 000 B.P.. These stadial conditions, which Lewis and Hanvey (1993) have named the *Bottelnek Stadial*, were replaced by warmer, interstadial conditions before 13 000 B.P.. Therefore, they infer cold stadial conditions from 27 000 B.P. to 13 000 B.P. (Lewis and Hanvey, 1993).

Lewis and Dardis (1985) reported a section of 'head' at Dynevor Park, near Barkly East (1980m) (Figure 1.1) that evidenced more than one phase of former Quaternary periglacial activity. During this stage, therefore, the zone of systematic freeze-thaw extended down to at least 1980m (at the section), and was sufficiently effective to form periglacial features in southern Africa. Therefore, it can be concluded that mean annual air temperatures at 1980m were +1°C or less, yet the existence of what appeared to be ice wedge casts within the lower part of the section indicates that mean annual air temperatures must have initially been between -4°C and -8°C, and ameliorated to reach +1°C during the formation of the heads. Marker (1994a) also reported that debris of periglacial origin overlies palaeosols dating to 33 000 to 29 200 B.P. in the Golden Gate Highlands National Park in the Free State portion of the Drakensberg, which supports findings in the Eastern Cape.

Titkov (1988, cited in Lewis and Hanvey, 1993) maintains that rock glaciers are active when between 10% and 20% of the surrounding area is covered by glacier ice. Consequently, it is likely that at least small glaciers existed in the East Cape Drakensberg in the Bottelnek Stadial.

Lewis (1994b) has identified protalus ramparts at 2000m at Killmore (Figure 1.1) in a valley that is tributary to the Bokspruit. He presents evidence suggesting that the ramparts accumulated during the Bottelnek Stadial.

The discovery of fossil rock glaciers and of protalus ramparts, believed to date to the Bottelnek Stadial (of between 27 000 and 13 000 B.P.), indicates that in the East Cape Drakensberg within the Stadial there were times in which:

i) at least discontinuous permafrost existed at and above 1 800m.
ii) mean annual air temperatures (MAAT) rose no higher than -2° (the highest MAAT at which rock glaciers can form, according to Haeberli (1985, cited in Lewis,1994b)).
iii) conditions were probably semi-arid.
iv) perennial snowbeds occurred down to at least 2 000m.

2.2.3. Climatic information derived from the periglacial phenomena.

Karte (1983) has classified the climatic conditions under which different periglacial phenomena develop (Table 2.1). It is thus possible to determine at least some climatic conditions by reference to periglacial features.

Figure 2.3 (as presented by Lewis, 1988b) indicates present climatic conditions in the Natal, Lesotho and East Cape Drakensberg at present, as evidenced by active periglacial features in the area. The zone of systematic freeze-thaw that is sufficiently effective to form periglacial features extends down to at least 2 550m (Lewis, 1994a), and possibly to 2 000m, the mean annual air temperatures thus ranging from a little below freezing to +1°C at this elevation. Recorded air temperatures are often far below freezing in these high altitude areas, for example, the lowest nocturnal temperature recorded in the past decade at Rhodes (*c.*1 800m) was -22°C, so it is not surprising that active periglacial features exist in the vicinity.

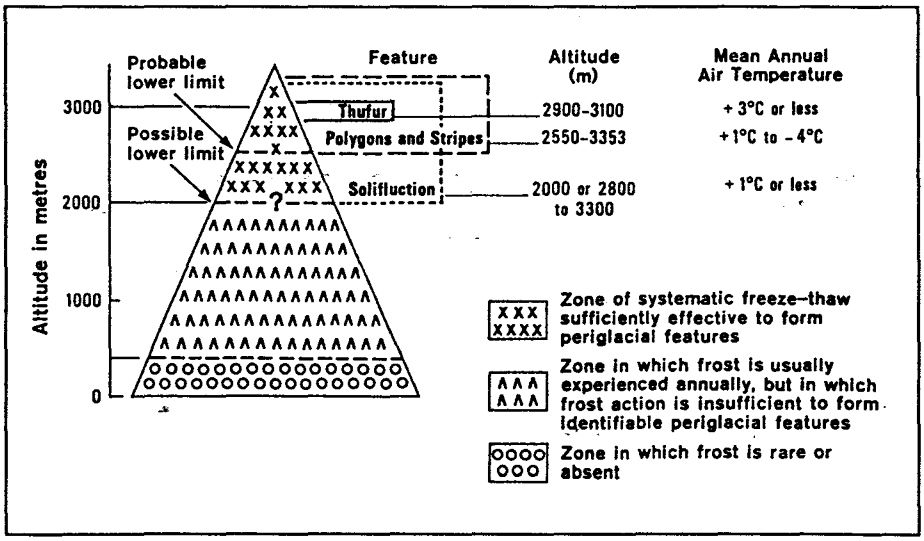


Figure 2.3: *Modern climatic conditions in the KwaZulu-Natal, Lesotho and East Cape Drakensberg as evidenced by active periglacial features.*

(From Lewis, 1994a)

2.2.4. Evidence of past glaciation

Although there is no evidence to indicate that the summit of the East Cape Drakensberg has ever been glaciated, Lewis (1994a) has discovered glacial features of Quaternary age at lower elevations in the East Cape Drakensberg. Striations on a rock surface at Eliasdale (Figure 1.1) indicate ice flow, and thus the former existence of a glacier, in that valley (Lewis, 1996). The glacier ice presumably fed off an ice cap on the high plateau at the head of Bokspruit and Rifelspruit, which rises above 2500m, and terminated downvalley of Eliasdale (1760m), and is thought Lewis (1996) to pre-date 35 000 B.P.

In addition to these striae and to striae near Rhodes in the Bell River valley (Lewis, 1996), other geomorphological evidences of glacial activity exist in the East Cape Drakensberg. These include trough-like valleys in which the spurs on the lower slopes have been truncated (Lewis, 1996), glacial erosional features (such as p-forms) and unconsolidated depositional sediments, (such as kame terrace deposits, a kame moraine and dropstones suggestive of a former pro-glacial lake in the Bell River valley near Rhodes), which indicate that valley glaciers formerly extended into the Bell River and Riflespruit valleys from ice caps or more extensive ice masses on the Ben MacDhui-Cairntoul and Cairngorm Plateaux (Lewis, 1994a, 1996).

Lewis and Hanvey (1988) have also located debris deposits (subsequently suggested by Lewis (1996) to be the remains of lobate rock glaciers) downvalley of Rhodes. Lewis (1996) has shown that these deposits are younger than 34 1000 B.P. and probably accumulated during the Bottelnek Stadial (that is, between 24 300 and 13 000 B.P.; Lewis and Hanvey, 1993). The palynological evidence presented by Lewis, Hill and Papaloizou (submitted) shows that interstadial conditions (the Birnam Interstadial) preceded the Bottelnek Stadial. Lewis (1996) has suggested that the valley glaciers existed during the Eliasdale Stadial, which preceded the Birnam Interstadial, and which Lewis, Hill and Papaloizou (submitted) show predates 38 000 B.P.

2.2.5. Summary and conclusion

Active and fossil periglacial features exist at altitudes above 2550m in the Drakensberg of South Africa at present. The modern climatic conditions are considered to be 'subperiglacial', with small-scale active features typically occurring only at high altitudes. Although fossil periglacial features are reported within the high altitude areas, evidence for their previous existence has been located in sediments at lower altitudes. This provides geomorphic evidence that colder climatic conditions prevailed during the Pleistocene, which allowed for the occurrence of deep penetrating frost action at lower altitudes than at present.

The occurrence of both fossil and active periglacial features have been recorded in the East Cape Drakensberg. Evidence for periglacial activity above 1800m during the Bottelnek Stadial (27 000 B.P. to 14 000 B.P.) was presented by Lewis and Hanvey (1993). Evidence has also indicated that at least discontinuous permafrost existed at altitudes above 2000m during that stadial and that perennial snowbeds existed at that altitude (Lewis, 1996). Prior to that stadial, before 34 100 B.P., valley glaciers existed in the vicinity of Rhodes (Lewis, 1994b) although no evidence of periglacial activity contemporary with or predating that valley glaciation has yet been identified (Lewis, 1996).

3.1. LOCATION

Tiffindell farm which is the focus of this study, is situated in the mountains of the East Cape Drakensberg. The 1 300 hectare farm extends from Breslin's Kop (2863m) on the western side, to the junction of the border fences between Lesotho, the Republic of South Africa, and the former-Transkei - which lies just beyond Ben MacDhui (3001.2m) (Figure 1.1; Appendix I).

3.2. GEOLOGY

The geology of the area is dominated by deposits of the Karoo Supergroup, which comprises the following units:

Sequence	Group	Formation	Composition
Karoo Super-Group	Drakensberg* Lebombo*	Drakensberg	basalt lavas
	Stormberg	Clarens Elliot Molteno	sandstone sandstone; shale sandstone; shale
	Beaufort	Burgersdorp Katberg Balfour Middleton Koonap	shale; mudstone sandstone sandstones
	Ecca	Fort Brown Ripon Collingham Whitehill Prince Albert	shales (turbidites)
	Dwyka		glacial tillites

* = *These are subgroups, which make up the Volcanics group.*

Table 3.1: Formations comprising the Karoo Sequence. (complied from SACS, 1980)

Karoo sedimentation ended with the eruption of volcanics (Drakensberg and Lebombo Subgroups) in the early Jurassic, forming the capping for the Karoo Supergroup. The main volcanic episode extended between 172-155 million years ago in the vicinity of Lesotho and the surrounding areas (Schmitz and Rooyani, 1987). These basaltic outpourings produced one of the largest continental volcanic fields in the world, of which the present 140 000km^2 Drakensberg plateau is an erosional remnant (Tankard *et al*, 1982). The most extensive outcrops of these rocks are confined to Lesotho, to the immediately neighbouring portions of South Africa, and to the Lebombo Mountains on the border between Swaziland and KwaZulu-Natal. The preserved thickness of the Drakensberg basalts was recorded as 1 400m at Mont aux Sources in the northeast of Lesotho, and even here the top of the succession is not seen (Lock *et al*, 1974). At present, the lavas outcrop over 70% of the surface of Lesotho, forming the Lesotho mountain area and the higher foothills, as well as the bordering mountains within South Africa (Schmitz and Rooyani, 1987).

The study area lies within the East Cape Drakensberg, and is dominated by a portion of the Drakensberg Subgroup. Basalt of the Drakensberg Formation (and to a lesser extent, dolerite and gabbro) forms the dominant bedrock at Tiffindell. The Drakensberg Formation basalt is typically dark and fine textured, which influences the soils in the area. The dominant mineralogy of the basalt is plagioclase (a sodium/calcium aluminosilicate), with pyroxene (an iron/magnesium silicate) and olivine (a magnesium/iron silicate) next in abundance (Binne and Partners, 1971, cited in Schmitz and Rooyani, 1987). Zeolite, agate and quartz are also present in the basalts, and formed as secondary minerals through precipitation, filling the pores of the original lava.

Below an altitude of approximately 1 900m in the lowland areas, bedrock is predominantly of sedimentary origin. The Clarens, Elliot and Molteno sedimentary Formations are comprised mainly of shale and sandstone (Schmitz and Rooyani, 1987).

3.3. PHYSIOGRAPHY

The main part of Tiffindell farm occupies a south facing valley (Appendix I). The southeast-facing slope rises to 2930m, while the south-facing slope rises to 3001.2m at Ben MacDhui. The valley is asymmetric, with the southeast-facing slope being the steeper of the two (Figure 3.1). Average gradients were measured at approximately 27° on the west slope, decreasing to an average of less than 25° on the south-facing slope. The far east part of the valley, directly beneath the Ben MacDhui summit (Appendix I), is one of the steepest sections of the Tiffindell valley, and is characterised by slope gradients of 30° or more. The general valley slope is concave. Below 2700m, the valley flattens considerably, and it is in this area that Tiffindell Ski resort is located.

Around the head of the valley, above 2900m, bedrock outcrops, forming vertical or near-vertical cliffs up to 16m high. These 'headwalls' often completely freeze over with thick sheets of ice during the winter months (Lewis, 1996; Figure 3.2). The summit of the basin is characterised by a relatively flat ridge which runs from east to west, and is longitudinally divided by the former-Transkei/South Africa boundary.

The valley lies within the upper-Bell catchment, and contributes to the headwaters of this river. Five or more sporadic rivulets are evident within the Tiffindell valley, and at present these feed the man-made dams situated around the resort. There is little evidence of previous large-scale disturbance or erosion (for example, by landslides etc). Yet, recent development of the resort has disrupted the natural surroundings due to excavations and landfills, and the development of ski-runs, water storage facilities and so on.

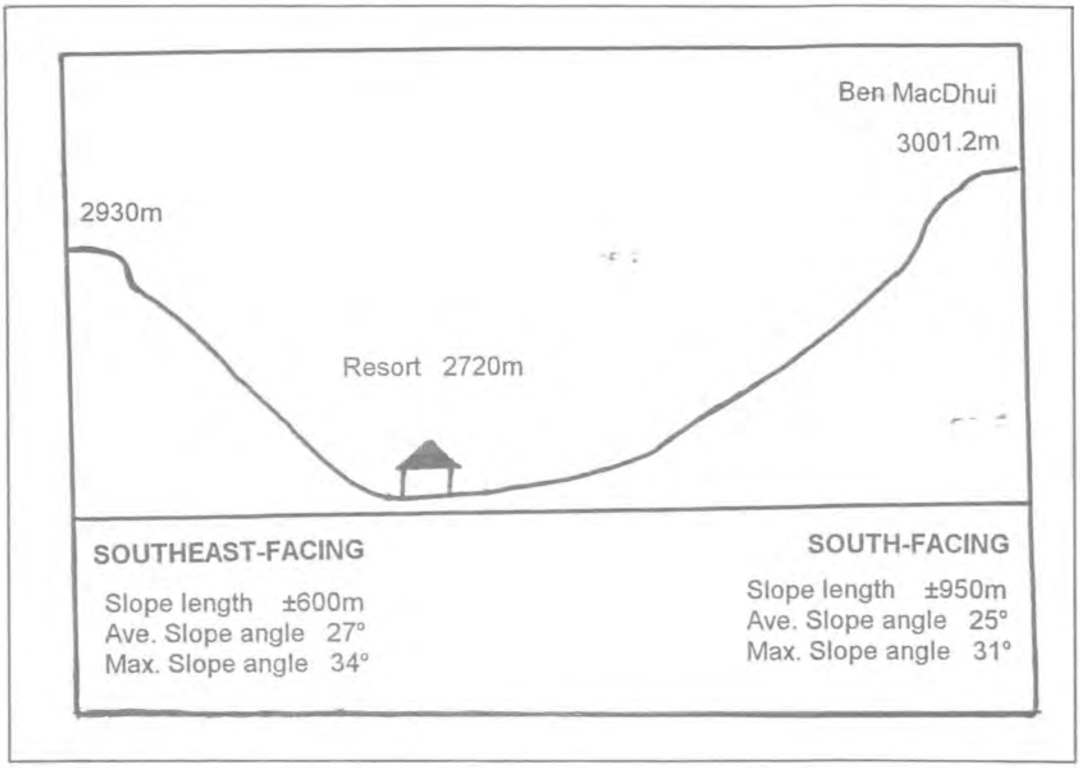


Figure 3.1: Schematic representation of valley asymmetry at Tiffindell

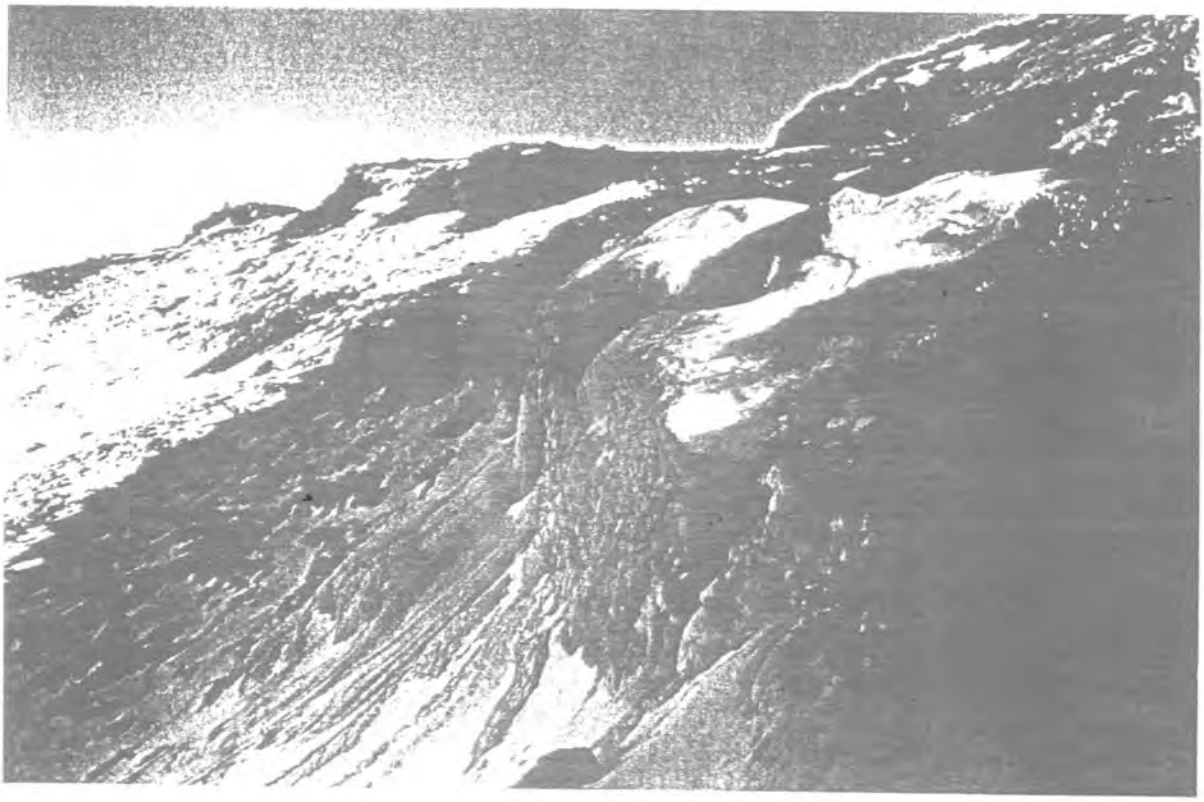


Figure 3.2: Frozen cliff faces at Tiffindell in winter.

3.4. CLIMATE

3.4.1. Climate classification

The climate of the region can be summarised by the classification scheme proposed by Köppen, which is the most widely used system for geographical purposes (Strahler, 1975). The climate is defined according to fixed values of precipitation and temperature, based on yearly or monthly averages. Strahler (1975) describes this scheme in more detail. The East Cape Drakensberg and adjacent Lesotho area may, according to Köppen, be classified as a *Cwb* type, that is a warm temperate (mesothermal) climate with a dry winter. The mean temperature of the warmest month being approximately 22°C, the coldest month experiencing an average temperature under 18°C, but above -3°C (Schulze, 1947; Strahler, 1975). The term *mesothermal* is widely substituted for *temperate* to imply intermediate temperatures as compared with the extreme heat of dry deserts, and the extreme cold of polar and arctic climates.

3.4.2. Present climate in southern Africa

The climate of southern Africa is strongly influenced by the latitudinal position of the subcontinent, which tapers southwards approximately from the Tropic of Capricorn to 35° south. The subcontinent's position in relation to the major circulation features of the southern hemisphere results in those areas south of 20°S latitude experiencing the effects of temperate disturbances originating in the circumpolar westerlies, and with tropical disturbances influencing the lower latitudes (Deacon and Lancaster, 1988). Southern Africa therefore lies in a zone of interaction between these atmospheric circulations which influences the climate in all seasons, and is a major factor affecting rainfall.

Isohyets assume an almost north-south alignment (except in the southern Cape), 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). Topography influences precipitation in most areas. For example, the orographic effects of the Drakensberg escarpment and the Cape Fold Belt are marked (Deacon and Lancaster, 1988). Rainfall is seasonal (Schulze, 1984), the northern and eastern areas experiencing more than 80 percent of the annual rainfall during the summer months, and the southwestern Cape experiencing predominantly winter rainfall. The southern Cape coast lies within an all-seasons rainfall belt. The winter rainfall is most commonly associated with the passage of cold fronts (that is, dominant westerly disturbances). The summer rains are largely a product of convective activity. Over the eastern and south-eastern Escarpment rain may occur up to 130 days per year (Tyson, 1986).

Air temperatures in southern Africa show great spatial diversity, and are largely dependent on altitude, configuration of the land, and the proximity to oceanic influences. The subcontinent spans about 20° of latitude, and varies from sea-level to mountains exceeding 3000m in height. Variations in the recorded mean annual temperatures (MAT) indicate these ranges. For example, MAT's of 23.3°C at Goodhouse in the western Orange River valley compared to 11.5°C at Mokhotlong in the mountains of Lesotho. The frequency of days with maximum temperatures exceeding 30°C shows a strong west to east gradient with often less than ten occurrences over the Escarpment region (Tyson, 1986). These upland areas are known to experience more than 90 days a year with temperatures below 0°C (Deacon and Lancaster, 1988).

Mean temperatures do not, in general, give a true reflection of the extremes experienced at a particular location. Especially in areas of diverse relief, katabatic flow and local winds can cause rapid and considerable temperature variations over short distances. These variations are especially noticeable in the minimum temperatures (Tyson, 1986). Frequencies of below-freezing minimum temperatures are highest over the Drakensberg and Escarpment areas, these areas also being prone to cold snaps (often up to 40 such occurrences) and snowfall in the winter months. Temperature falls exceeding 10°C may be expected more than five times a year over a large part of southern Africa. If these frequencies were to rise with an increasing frequency of westerly perturbations, then overall temperature conditions would be expected to change, and a more severe climate similar to that which prevailed about 16 000 years ago in southern Africa would predominate (Deacon, 1983; Tyson, 1986).

With regards to the range of periglacial climates summarised by Tricart (1969; Chapter 2, section 2.1.2.), the alpine periglacial climate of South Africa's high altitude areas corresponds to the mountain subtype of Tricart's cold with little seasonal temperature change zone (C-type climate). This is characterised by higher winter temperatures than those of arctic or polar climates, as well as seasonal and diurnal temperature fluctuations which result in abundant, typically diurnal, frost cycles often penetrating only to shallow depths. Orographic effects result in generally high precipitation totals, with a large majority of the winter precipitation in the form of snow. Permafrost is absent.

3.4.3. Present climatic conditions at Tiffindell Ski

The Tiffindell area forms a part of the subperiglacial zone of southern Africa, the alpine climate of the region being characterised by warm, wet summers, and cold drier winters with the majority of the winter precipitation falling as snow. The amount and frequency of snowfall in the area is typically erratic, and with often more than eight snowfalls a year (Tyson, 1986). Snowfalls of up to 0.5m have been recorded in the spring and summer months (Tiffindell precipitation log book, 1996).

There is a lack of meteorological data for the area, so for long-term records it is necessary to extrapolate from a weather station located at Rhodes (1800m, approximately 22km from Tiffindell). From data collected for this research project from May 1995, the mean annual temperature of the Tiffindell area can be calculated as approximately 7.5°C. Average seasonal temperatures show a mean temperature of 17.5°C for summer and -2.9°C for the winter months (Chapter 5), but both seasonal and diurnal variations are considerable. Temperature extremes vary from -14°C to 32°C (Barrie Low and Rebelo, 1996; pers. obs.).

The area is predominantly (approximately 70% of the time) exposed to winds from a north westerly direction under the influence of a continental high. When a ridging high occurs, the wind direction shifts to south easterly and when a cold front passes the wind is from the south west for approximately half a day (S.Piketh, pers.comm.). At all times high wind speeds may be experienced at the field site. Wind speeds at an altitude of 2758m (measured at the WITS weather station at Tiffindell) averaged 8.48ms^{-1} during a month-long period in winter 1996 (CSIR Environtek, 1996). On the Breslin's Kop - Ben MacDhui ridge, wind speeds have been observed to increase markedly (pers. obs.).

Precipitation is chiefly rainfall, which predominantly falls in the summer months from October to March, often in the form of thunderstorms which may be intense but also local. Median annual rainfall has been recorded as 800mm (Dent *et al.,* 1989) for the area, yet a large percentage is lost in immediate run-off. The high winds also have a severe drying influence. The wettest months are January, February and March (Tiffindell precipitation log book, 1996), yet the drier winter months of June and July are typically associated with snowfall, and according to Tyson (1986), only 10% of the total precipitation can be expected within these winter months.

Cold snaps and snowfall are common with the passing of a cold front, which usually last from 2 to 4 days, and are followed by characteristic clear conditions often resulting in heavy frosts (Tyson, 1986). Usually the falls of temperature associated with cold snaps are of the order of 5°C, but have been recorded as severe as 19.7°C by Schulze (1972, cited in Tyson, 1986). The snow is seldom long-lasting and only covers the mountains for a period of days to a few weeks. It has been suggested that only a slight temperature decrease would be necessary to produce a snow-cover lasting for more than seven months (Sänger, 1988).

3.5. VEGETATION

The Ben MacDhui area forms part of the Alti Mountain grassland biome (Barrie Low and Rebelo, 1996) which occurs between altitudes of 2500 and 3480m in the upper mountain region of Lesotho and the adjacent Drakensberg. This biome consists mainly of tussock grasses, dwarf shrubs and creeping or mat-forming plants, with trees being completely absent.

The alpine grassland is dominated by several different xeromorphic grasses, which include *Merxmuellera disticha* (Mountain Wiregrass), *Pentzia cooperi, Thermeda triandra, Pentaschistis curvifolia, Harpechloa falx, Koeleria capensis* (Junegrass) and *Merxmuellera drakensbergensis*. These perennial grasses comprise more than half of the total community cover, and tend to be the most widespread species. They are typically tufted, between 150-450mm tall, giving about 30% basal cover. They were deemed unsuitable as a ski slope base because of their restricted ground cover (Science Applications, 1993), and were replaced in 1993 with Kentucky Blue Grass on the ski-run slopes. Although the indigenous grasses do not offer complete basal cover, they tend to bind the soil together effectively with their shallow fibrous root systems, and thus assist in lessening the acute danger of erosion of the fine-grained soil during the growing season due to high rainfall. However, modified areas of the resort have had additional vegetation introduced in order to instill stability in these areas. For example, Midmaar Rye grass has been planted on the banks of the resort's man-made dams (I.van Eck, pers.comm.).

Cyperaceae, Ericaceae and Asteraceae families are common in the area, the *Helichrysum* species being particularly abundant (Barrie Low and Rebelo, 1996). These plants, as well as many others, exhibit several morphological adaptations in response to climate (Chelchinskey, 1996). There is a general trend towards dwarf-shrubs with reduced leaves, many of which have become hairy, or woody stemmed in order to prevent excess water loss. For example, *Helichrysum trilineatum* are noted to have relatively hairless stems at lower altitudes, while those established above 2700m are extremely 'woolly' in appearance. The plants' evolution in response to the climatic conditions has resulted in the growth of hairs, which may contribute to the reduction of transpiration, retention of heat, as well as to the protection of the plant from increased levels of insolation, and from damaging frosts and snow (Hilliard *et al.*, 1987). Other plants have reduced flowers, and many of the *Senecios* species have formed a pappus for seed dispersal, using the windy conditions to their advantage (Chelchinskey, 1996). *Euphorbia clavaroides* are able to store water as they are succulents, and can therefore cope well with water stress.

The general stature of the plants also reflects the conditions in the area. The plants tend to become almost compact at higher altitudes, such as *Helichrysum sessilioides*, which is well branched and forms dense compact cushions which are tightly pressed against the rock on which it grows, the roots inhabiting the crevices. This plant has adapted to the harsh conditions, and is found on exposed rock faces up to 3001m. Others, such as *Lotonis galpinii* var. *prostrata*, grow extremely close to the ground in order to minimise damage by wind (Hilliard *et al.*, 1987).

There are several wetland, or marshy, areas, and these high altitude vlei communities are dominated by *Kniphofia thodei* (Red Hot Poker), *Carex aethiopica*, *C. globerabilis* and *Senecio erubescens*. These marshes tend to occur at altitudes of about 2650m or even lower, and are only rarely associated with periglacial features. Yet, they hold important information on the vegetation history, and therefore palaeoclimatology of the region during the time of sediment accumulation. Discovery of a peaty soil sequence near the ski lodge (at an altitude of ±2750m; Appendix I) allowed for radiocarbon dating and palynological analyses (Rosen and Lewis, in preparation) to be conducted for the area.

Radiocarbon dating revealed that the analysed peaty sediments encompassed the time period 460±30 B.P. to just over 2730±45 B.P. Palynological analysis on these sediments revealed few marked vegetation changes from the present vegetation, which is not surprising given the recent time period of the soil sequence.

The relative abundances of the identified taxa within the peaty sequence give an indication of the varying climatic conditions at specific times. Cyperaceae pollen dominates throughout the profile, thus indicating that moist conditions prevailed throughout the period of sedimentation. At ±2000 B.P., an increase in the pollen of Ericaceae and Asteraceae types may indicate that more humid and cooler conditions prevailed, at least locally (*cf.* Van Zinderen Bakker, 1989; Scott *et al.*, 1995). With the major climatic variations in southern Africa occurring prior to the period encompassed by the organic soil sequence (Partridge *et al.*, 1990), no dramatic variations in temperature and wetness are expected to have occurred locally in the past 5000 years, nor are they indicated by the fossil pollen record from Tiffindell.

The typical vegetation of the Tiffindell area resembles that of other high altitude alpine areas in southern Africa (Hilliard *et al.*, 1987; Hill, 1992; Grāb, 1994), nevertheless, the overall vegetation at Ben MacDhui tends to be characteristic of, and able to cope with, a somewhat drier climatic regime than that of, for example, the Natal Drakensberg.

3.6. SOILS / UNCONSOLIDATED SEDIMENTS

The basalt of the Drakensberg Formation forms the dominant bedrock for the area, and being typically non-argillaceous, but dark and relatively fine textured, influences the properties of the unconsolidated sediments of the area. The grain size of the regional regolith is predominantly fine-grained, and largely consists of silty-sand, (approximately 60% to 75% by mass). The surficial sediments therefore, in general, have high frictional strength but little cohesion and low saturation water content.

The unconsolidated sediments are shallow at Tiffindell, ranging in thickness from 0.30m at the summit to greater than 1.0m at the base of the ski slope (2720m) and consist of dark brown to blackish brown (10YR 3/3 to 10YR 2/2) silty-sand permeated by grass roots and often containing large rocks. The organic content, as indicated by ignition loss, varies from 4% in unvegetated areas, to approximately 20% on the vegetated slopes and flatter sections, to higher contents varying from 20% to 50% within patterned ground areas.

The Atterberg limits of the surficial sediments, which give an indication of consistency under different moisture conditions, are determined by the soil[1] characteristics. The Liquid Limit (LL) denotes the moisture content at which soil passes from the plastic to the liquid state, as determined by the liquid limit test (BS 1377, 1975), and is characteristically low in these soils (20%-35%), which generally perform as cohesionless soils (Harris, 1987). The Plastic Limit (PL) is the moisture content at which a soil becomes too dry to be in a plastic condition, as determined by the plastic limit test (BS 1377, 1975) and is also low. Therefore, the Plasticity Index (PI) (ie. the numerical difference between the LL and the PL of a soil) is characteristically less than 20%, generally in the range 0-15%. No soils were deemed as non-plastic, that is, a soil with a plasticity index of zero, or PL>LL (BS 1377, 1975).

The low LL and PI of the debris derived from non-argillaceous bedrock makes it particularly sensitive to changes in water content, and susceptible to loss of strength and flowage when water contents are high. Effective soil drainage, especially in winter, is often temporarily impeded by the presence of ice lenses within the soil profile or by bedrock which may lie relatively close to the surface. The surficial sediments at Tiffindell are often saturated, especially in those seasons of thaw of the ice lenses (Chapter 5) and of substantial rainfall. As a consequence, soil leaching is common. This reduces the levels of natural fertility of the soils as well as the development of characteristically acidic podsols (Barrie Low and Rebelo, 1996). The small amounts of plant nutrients available are confined almost exclusively to the top soil which may be relatively rich in organic matter as translocation of organic matter is typically slow.

---

1 *soil* is used here in the same sense as in Soil Mechanics, *i.e.* it is unconsolidated surficial sediment.

Pedogenic processes such as biological and chemical weathering tend to operate more slowly in periglacial sediments than in those of warmer environments (Ballantyne and Harris, 1994), yet due to the prevailing cold, moist conditions, enhanced mechanical weathering of the bedrock and soil constituents due to frost shatter (Chapter 5) is common (Buckman *et al.*, 1969). Of the minerals originally inherent to the parent material, quartz constitutes the largest percentage present in the unconsolidated sediments in the upland areas at present (Schmitz and Rooyani, 1987).

The soil profile at Tiffindell tends to lack distinct soil horizons. This may be a result of churning of the soil due to seasonal freezing and thawing associated with ice segregation (Chapter 5), frost sorting (Chapter 5) and the formation of patterned ground (Chapter 7 and 8), which effectively breaks-up and deforms the soil horizons (Reiger, 1974, cited in Ballantyne and Harris, 1994).

4.1. SELECTION OF THE STUDY SITE

The study site was selected with reference to the following criteria that were necessary for the research to be undertaken.

i) The site had to support recognizable periglacial features.
ii) The site should not have previously been studied in detail.
iii) The site had to be accessible.
iv) Accommodation at or near the site was essential.
v) Adequate labour had to be available for installation and maintenance of equipment.
vi) The site had to be secure, i.e. free from theft or major risk of physical attack.

In addition it was preferable that the study site should be in the process of economic development, so that the impact of periglacial processes on development, and visa versa, could be assessed.

The study site was at and adjacent to Tiffindell Ski resort, which is located at an altitude of 2720m. The slopes of Ben MacDhui between altitudes of 2600m and 3001m contain a variety of periglacial features, as does the summit area between Ben MacDhui and Breslin's Kop (Appendix I). This site hosts many of the periglacial features and processes known to occur within the East Cape Drakensberg (Lewis, 1996). The slopes are also predominantly south-facing, therefore tending to support more cold climate features than north-facing slopes which are exposed to greater periods of insolation. Consequently, these slopes were selected to research the occurrence of active and fossil features in a high altitude area in the East Cape Drakensberg.

Less extensive research has been conducted at high altitudes within this highland area than in the Natal Drakensberg range, only Lewis (1988a,b, 1996) reporting the occurrence of active periglacial features, at altitudes in excess of 2500m. Fossil periglacial features have been studied in detail at lesser altitudes (Hanvey *et al.*, 1986; Hanvey and Lewis, 1991; Lewis and Hanvey, 1991; Lewis and Hanvey, 1993; Lewis, 1994; Lewis, 1996). The present research provides detailed data on active, as well as on some fossil features in one of the highest areas of South Africa and therefore contributes to the examination of the Drakensberg Range as a whole.

Access to Tiffindell Ski was by 4WD vehicles along either the Carlisle's Hoek road from the village of Rhodes (1800m), or from Bidstone farm (2000m). From the resort, the field research area (approximately 119 Ha) was accessed on foot, even in snowbound conditions. Accommodation at the research site was provided at the resort. Other conveniences of the resort were the availability of people to observe the monitoring equipment on a regular basis, as well as the freedom and security to conduct field research anywhere within the boundaries of the farm. The borders between South Africa and the neighbouring countries were one of the few limiting factors (Figure 1.1) and resulted in less detailed research being conducted in those territories.

In addition to the contribution to the knowledge of periglacial features in southern Africa, the research also provides information which is of economic value to the resort developers.

4.2. CLIMATE

Climatic conditions in the Tiffindell area were established from several sources. An approximate 16 month data set (May 1995 - September 1996) was recorded for the study period using two MC Systems 120 electronic data loggers with temperature sensors 1.2m above ground level (the details of which will be discussed further in section 4.5). Further data was obtained from hygrometer and snowfall records taken by employees of the resort during the skiing season, as well as data for June 1996 obtained from the meteorological station set up at Tiffindell by the University of the Witwatersrand (WITS) and CSIR

Environtek. A more generalised view of the regional climatic conditions was obtained from various existing weather stations within similar latitudes, and/or altitudes through the Computing Centre for Water Research (CCWR), Pietermaritzburg. From these stations, daily and monthly temperature, and rainfall data was obtained from the CCWR via FTP (File Transfer Protocol) on the internet.

4.3. TOPOGRAPHY

The general topography of the research area on Tiffindell farm (119 Ha) was surveyed in 1994 by A. King and I. Hansen, who produced a 1:2000 topographic map of the Ski Resort area. This map, in conjunction with surveys by the author, was used to produce topographic cross-sections of the area. The location of periglacial features identified by the author was plotted on the map and on the cross-sections (Appendix I; Chapter 10).

Surveys to measure changes in hillslope profile with altitude at Tiffindell were conducted on both the south and southeast-facing slopes using an altimeter, theodolite, measuring tape and ranging rod (profile lines A and B; Appendix I). A 20m section was surveyed, followed by 75m being paced out and described, but not accurately surveyed with the theodolite. This was standardised for both transect lines in order to build up adequate and comparable data sets (Young *et al.*, 1974). The sites for survey were selected for apparent visual representation of the area as a whole. Practical difficulties were also taken into consideration (Cox, 1981), and areas characteristic of large, steep rocky outcrops and headwalls were avoided.

Extensive measurement and mapping of the natural features was undertaken to produce an accurate and comprehensive map displaying the topography. This also required the use of an altimeter, as well as a brunton compass for bearings and gradient measurements. The Barigo electronic altimeter which was used has a range of -500m to 8000m above sea level, with a 1m resolution. It functions on the principle of a barometer by sensing atmospheric pressure, and consequently calculates the relevant local altitude. The altimeter initially requires the input of a known altitude, and as it is sensitive to changes in atmospheric pressure, variations need to be corrected in the event of pressure changes, for example the advancement of a frontal system.

4.4. SEDIMENT ANALYSIS

Sediment samples were collected from the following sites, details of which are given in the relevant chapters:

i) At 95m intervals on a transect line upslope on both the south and south-east facing slopes;
ii) Gelifluction lobes: from treads, risers and at varying depths within lobes;
iii) Thufur site: from within individual thufa, and below thufa at 5cm intervals to a depth of 45cm. Samples were also collected from inter-hummock depressions;
iv) Sites of 'frost mound' development;
v) Sites of frost heave activity and polygon formation;
vi) Stone lobes: from the 'fines' of the lobe;

A 0.5-1kg sediment sample was collected at each sampling site with the use of a geology-pick and a trowel, placed in labelled plastic sample bags and sealed tightly.

4.4.1. Grain size determination

The sediment samples were dried slowly in an oven of temperature approximately 50°C. The individual particles were then separated for the sediment analysis using a pestle and mortar. Care was taken in order to preserve the samples' constituents by not over-grinding the samples and thereby affecting the results by causing a bias towards the fines.

Particle-size analyses of the sediment samples was based on the techniques suggested by Gale and Hoare (1991). Sieves ranging from the *gravel (16mm)* to the *silt (63μm)* particle-size category were selected for particle-size analysis by sieving. Those size classes <63μm were separated using a standard pipette analysis technique (Gale and Hoare, 1991).

4.4.2. Organic content determination

Organic content was determined using the method of Davies (1974), which is described by Gale and Hoare (1991). Each sample was placed in a crucible in a furnace of temperature 430±°C for 24 hours. After allowing the samples to cool to room temperature in a desiccator, they were weighed, and the percentage by mass lost on ignition was calculated for each sample.

4.4.3. Moisture content determination

Moisture content is the ratio of the weight of water to the weight of solid in a given volume, and is expressed in percentage terms. Moisture content in a given sample was determined by the loss of weight when a 5-10g sample was oven dried at 105-110°C for a minimum of 16 hours (McGreal, 1981). This property is valuable in geomorphological analysis, as soil behaviour and strength vary with moisture content, and the analysis forms the basis of Atterberg limits which represent changes in soil state.

4.4.4. Determination of Atterberg limits

At certain water content values, soils exhibit characteristic behaviour and strength conditions depending on their particular particle size ranges. The values delimiting the soil properties are called Atterberg limits, the determination of which gives an indication of consistency under different moisture conditions, that is solid, plastic or liquid. The limits determined are known as the *liquid limit* and the *plastic limit.*

For the Tiffindell sediment, the Atterberg limits were measured according to the methods proposed by British Standards (BSI 1377, 1975). The Casagrande method was used for the determination of liquid limits (W_L). The soil liquid limit is the "water content 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" (Kézdi, 1980).

The plastic limit refers to the "minimum water content (W_P) of the soil at which threads of 3-4mm thickness can be rolled out without crumbling" (Kézdi, 1980:80). Therefore, Kézdi's method of rolling the soil into threads was used, a strict uniformity of the procedure being adopted as no controlled apparatus is used (Vickers, 1978).

The difference between the liquid and the plastic limits can be used to calculate a *plasticity index* (I_P):

$$I_P = W_L - W_P$$

These Atterberg limits suggest the conditions under which the studied soil may be expected to display movement. If the moisture content exceeds the Liquid limit, the sediment can be expected to lose strength and be susceptible to 'flow'.

Therefore, these, together with other analyses mentioned above, were used as an indication of the soil (*i.e.* unconsolidated sediment) properties within the research site. The significance of the environmental conditions measured within the study area can be assessed with reference to these properties (Harris, 1973).

4.5. PROCESS STUDIES

For process studies, several sites were selected at random, so as to obtain a true reflection of the degree of activity of the various features and processes occurring in the area. The following process studies were chosen as they appeared to be representative of the dominant processes occurring within the study site, and would also determine whether the features being studied were fossil, or active features. Figure 4.1 and Appendix I display the location of several of the process study sites.

TIFFINDELL SKI

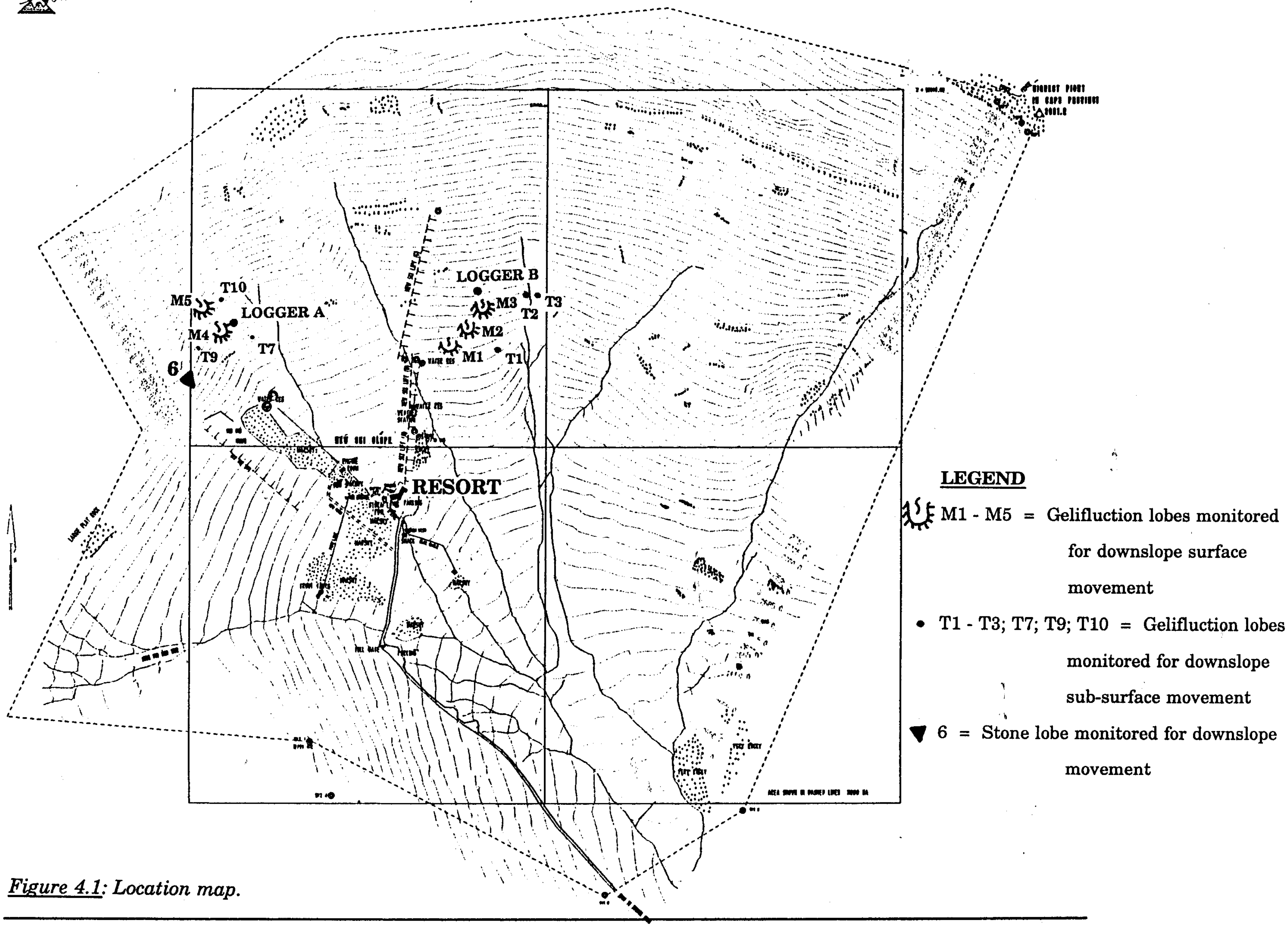


Figure 4.1: Location map.

4.5.1. Frost heave and needle ice activity

Two experimental sites (sites A and B; Figure 4.1), were set up to measure and monitor frost action and needle ice activity for a 10 day period during mid-June 1996. The experiments were initiated on June 8 1996, and monitored until 19 June 1996. Continuous *diurnal frost heave* experiments were undertaken at each location.

The markers were wooden dowels 6cm long and 0.25cm in diameter. They were ringed in black at every centimetre mark to facilitate uniform and accurate insertion into the ground and to facilitate gathering of results. Four rows of 8 markers each were arranged in a compact grid pattern as shown on Figure 4.2. Individual markers were spaced 2cm apart within the rows, and the rows were 3cm apart. Row 1 was inserted to a depth of 1cm; row 2 to a depth of 2cm, row 3 to a depth of 3cm, and row 4 to a depth of 4cm.

The markers were not disturbed for the duration of the experiment and so the results are cumulative. That is, the results from each day would have been affected by the previous day's activity, and will in turn, have had an effect on the following day's results.

The main difference between the two locations selected was the difference in altitude. Experimental site A was at 2721m, site B at 2794m (Figure 4.1). Both sites were established on essentially flat ground (gradient measured to be 3°). Site A was established on a footpath, and site B on the tread of a terrace. As both these areas were in danger of being disturbed, they were well marked out with chevron tape in order to minimise disturbance by the resort's visitors.

Before the experimental sites were set up, the selected areas were smoothed over and all existing ice crystals were destroyed. As the experimental sites were set up between 3 and 4pm, exposure to the sun for the entire day had melted the majority of the needle ice at the sites, as they were established in areas unprotected from the sun and wind.

4.5.2. Frost action and gelifluction processes within the Tiffindell sediment

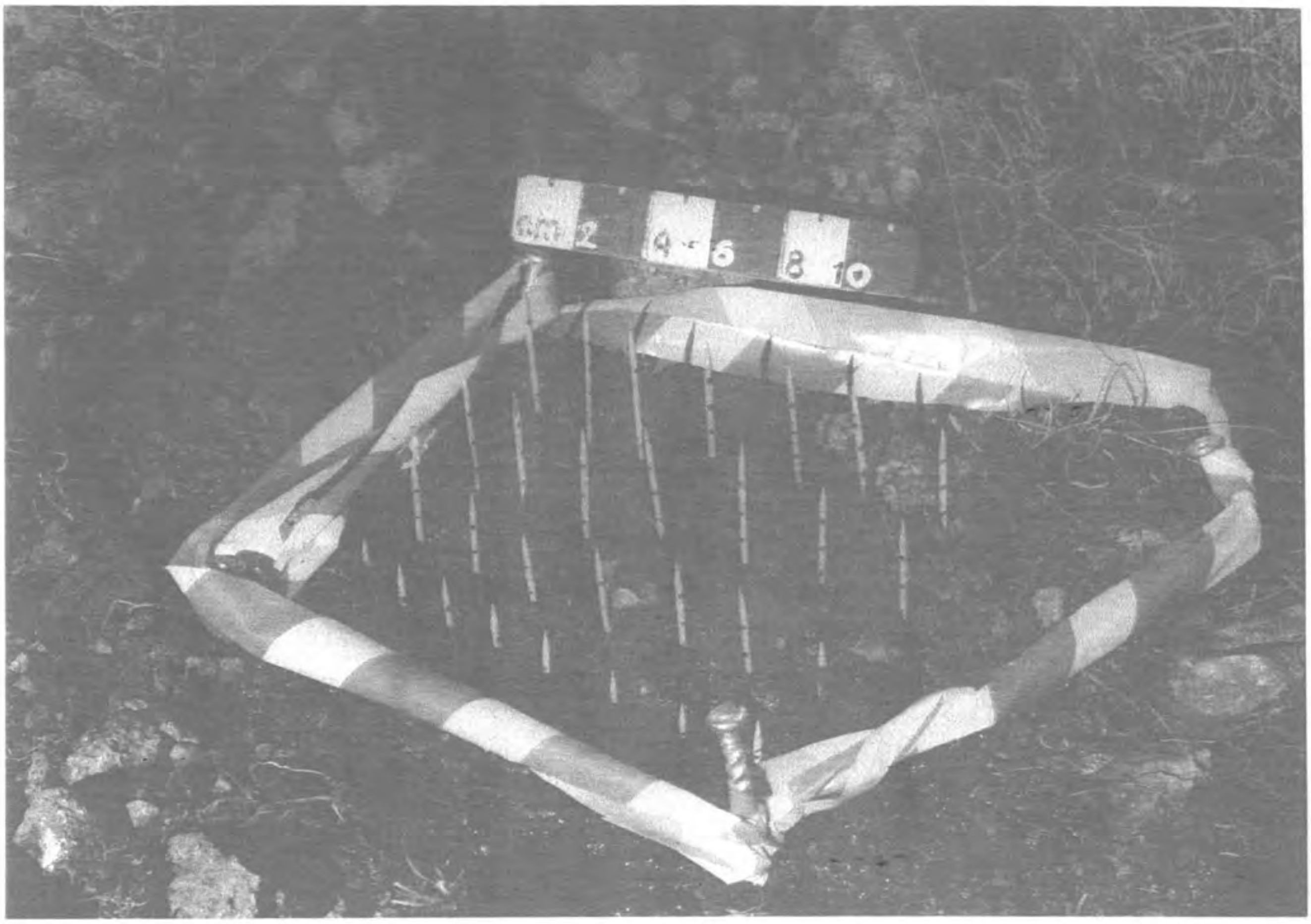
Frost action processes and downslope sediment movement within the Tiffindell area was recorded using various techniques as described below. In order to obtain an idea of the relative number of freeze-thaw cycles and depth of freezing, and soil moisture content occurring at various depths throughout the year, three relatively large turf-banked lobes were selected (Figure 4.1; Figure 4.3) and the climatic conditions at each monitored for more than one year (that is, May 1995 to September 1996).

4.5.2.1. Recording microclimatic data

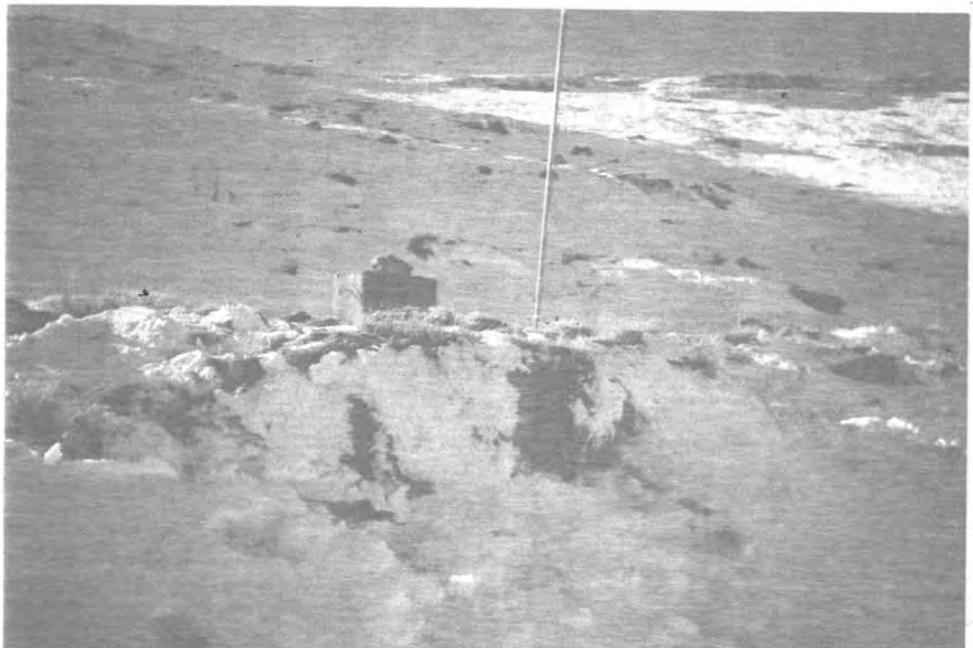
Microclimatic data was measured using several temperature probes and soil moisture sensors, which were connected to two MCS-120 electronic data loggers. Station A was monitored by logger A, and stations B and C by logger B (Figure 4.1).
At each station, the following was recorded:

- air temperature 1.2m above the ground;
- soil temperature at ground level;
- soil temperature at depths 0.05m, 0.20m below ground level;
- soil moisture at depths 0.05m, 0.20m below ground level.

Each stations' set of sensors thus consisted of four temperature sensor probes, and two soil moisture sensors. The sensors connected to logger B were arranged into two sets, one for station B, and the other for station C, and allowed for the simultaneous recording of the conditions at both stations B and C. Figures 4.3 and 4.4 present a visual impression of the situation at station B.



<u>Figure 4.2:</u> Experimental site A (2721m) displaying the arrangement of wooden dowels for the diurnal frost heave experiment. Those inserted to a depth of 4cm are in the foreground, and those to 1cm in the background (8 June 1996).



<u>Figure 4.3:</u> Station B at 2788m, showing the positioning of Logger B. The red pole supports the air sensor.

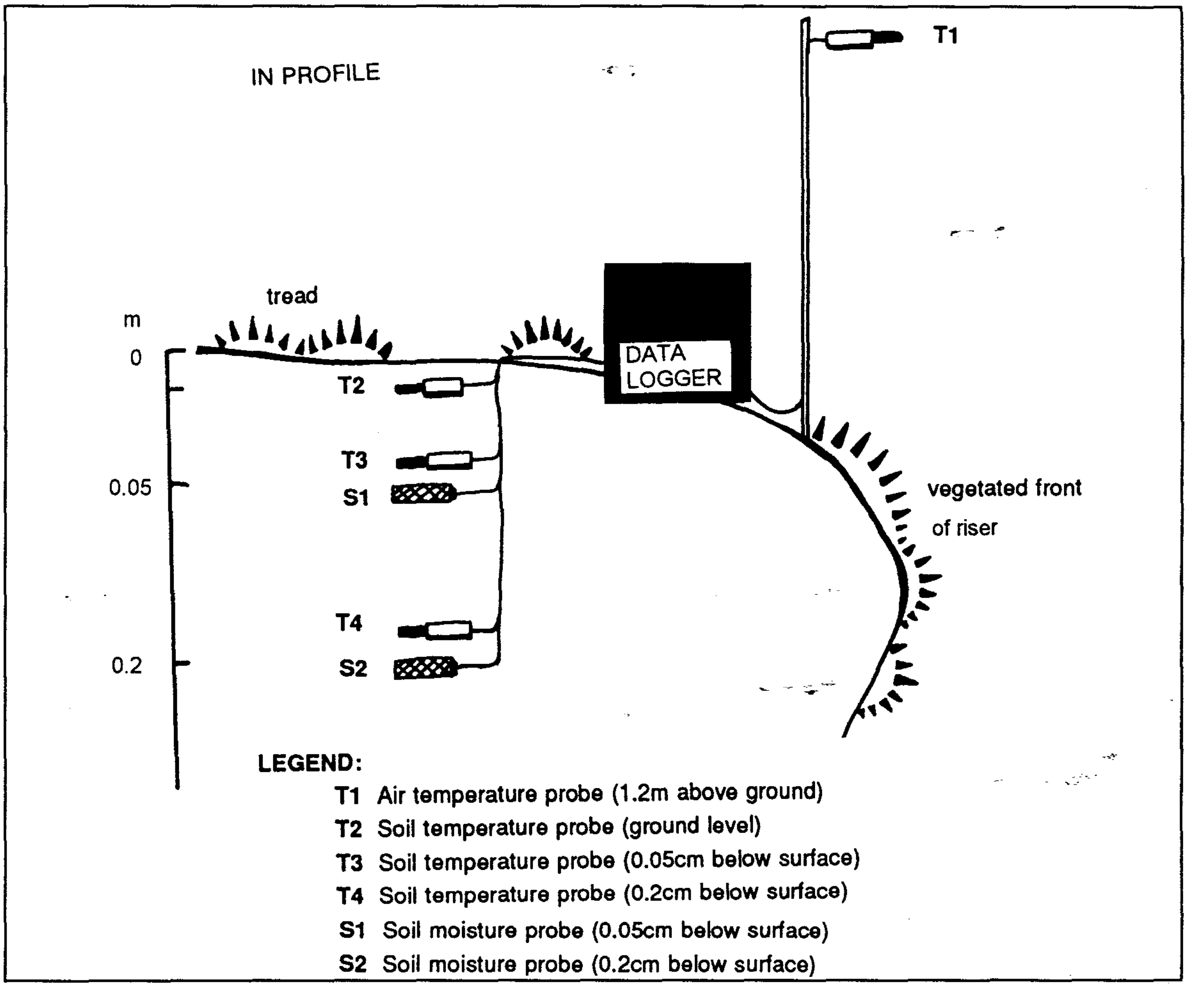


Figure 4.4: Location of datalogger, temperature, and soil moisture probes within a turf-banked lobe at Station B (2788m elevation).

All the temperature probes were of the MCS 151 series, which have "a thermal inertia approximately equal to a large mercury thermometer" (MC Systems, 1990a). The probe (or 'sensor'), is mounted in a 8cm long thick-walled aluminium tube, and is intended for measuring average temperature changes. The operating temperatures are in the range -20°C to +70°C with an accuracy of ±0.2°C at 25°C and a resolution of ±0.1°C (MC Systems, 1990a).

The 8mm (diameter) x 75mm (length) temperature probes were inserted horizontally, at their respective depths, into undisturbed soil that had been temporarily exposed by the excavation of a small pit. Good contact between the soil and the sensor was confirmed, and the pit was immediately refilled to ensure minimal disturbance. The probes measuring the soil surface temperatures were inserted just below the surface of the soil, that is, only a thin layer (<5mm) of soil covered the sensor. The air temperature probe was attached to a pole at a height of 1.2m above ground level.

Soil moisture was measured using the MCS 159 Nylon Soil Moisture sensor (NSM), which is considered to be highly suitable for use in moist soils (MC Systems, 1990c). This sensor is ideal for continuous measurement of soil moisture content, and is relatively sensitive to small changes in concentration of the soil solution. The 70x18x4mm NSM sensor consists of an inner mesh electrode wrapped in nylon material, and an outer mesh electrode enclosing the sensor. The moisture content of the nylon material equilibrates with the soil moisture, and the resistance between the two electrode meshes is recorded. Therefore, the drier the nylon material, the greater the resistance between the two electrode meshes. The moisture content is recorded as a percentage, and can fall anywhere in the range between 0-100% range, 100% representing saturated conditions. The optimum operating temperature range of the NSM sensors is -10°C to +50°C, and the accuracy of each sensor is completely soil dependent (MC Systems, 1990b). The sensors were inserted into the soil at the same time as the temperature probes, and care was taken to ensure that the 25cm^2 active area of each sensor was in full contact with the soil.

All the sensors were calibrated and checked in the laboratory before being taken into the field, and were allowed to equilibrate for a 24 hour period once put in place. The sensors remained in place throughout the recording period.

The loggers were monitored, and the memory modules and power packs were regularly replaced by the researcher or by Tiffindell staff. Temperature and soil moisture conditions were recorded onto memory modules at two hour intervals over the period from 27 May 1995 to 15 September 1996. Electronic problems with logger A resulted in an incomplete data set from that station for the second half of the study period. The data was read from memory modules into a spreadsheet package, QuattroPro 5, for manipulation of the results.

4.5.2.2. Surface gelifluction lobe movement

Five lobes were selected (M1 to M5), two on the southeast-facing slope, and three on the south-facing slope (Figure 4.1) in order to monitor surface downslope movement of the gelifluction lobes.

A series of triangular metal markers (Figure 4.5) were placed on the selected lobe treads in a straight line, just behind the lobe front, in order to act as markers for movement. They were 3mm thick, 150mm long on each side, and had a 100mm spike on each corner in order to prevent the marker from being dislodged by snowfall, excessive rain or wind. A 300mm long central rod (painted red) was used as an indicator when orientating the markers in a straight line. The central rods were later cut shorter before insertion in an attempt to minimise excessive disturbance of the markers by inquisitive people.

The markers were inserted with the flat edge flush with the soil surface, in a straight line approximately 15-20cm from the front of the lobe edge (Figure 4.6a). They were placed approximately 100mm apart from one another (Figure 4.6b). Permanent reference points of heavy metal stakes were driven into the ground on either side of the selected lobe to a depth of at least 1m. A length of nylon cord was attached to these reference points across the lobe front and behind the central rods of the markers, to avoid any disturbance in their downward movement. Therefore, the taut line between the stakes acted as a guide in aligning the markers, and was used periodically in measuring the displacement of the metal markers. Final measurements of movement were made in November 1996.

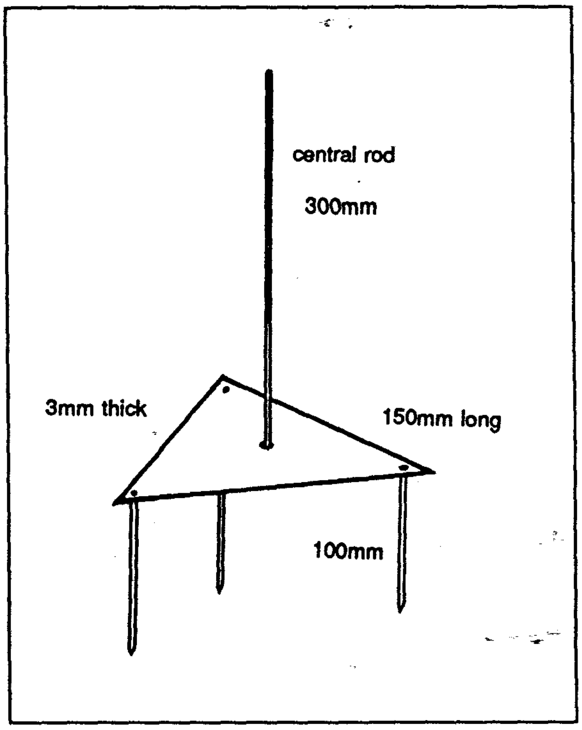
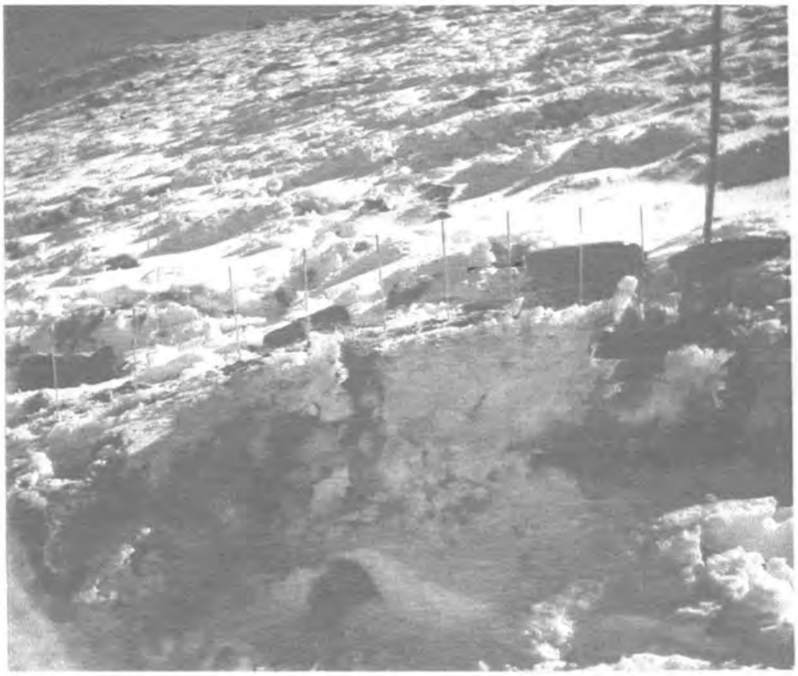


Figure 4.5: Schematic representation of the metal markers used in soil surface movement determination.

4.5.2.3. Subsurface gelifluction lobe movement

Subsurface rates of lobe movement were also recorded with the use of eleven flexible polyethylene tubes, which were inserted in June 1995 vertically into the lobes. The thick-walled hollow tubes (10mm outside diameter) were 1m in length, and heat sealed at their bottom ends (Smith, 1992; Price, 1991). They were implanted to a depth just short of 1m into a slightly larger hole (made by a metal auger being hammered into the ground; Figure 4.7), and then kept straight by a thin steel rod that was removed once the soil had been repacked around the tube and it was secure. The top end was then also heat sealed in the field, and the site was marked with a red pole.

<u>Figure 4.6a</u>: Metal markers emplaced on the lobe front (M1) in a straight line.



<u>Figure 4.6b</u>: Front view of lobe M4 showing the positions of the metal markers.

<u>Figure 4.7</u>: Preparation of the ground for polyethylene tube insertion (May 1995).

<u>Figure 4.8:</u> Vertical and horizontal lines painted across lobe 6 to test for movement of the boulders

After two winter and two spring seasons, six tubes were excavated in November 1996 to provide a measure of soil displacement. Each excavation extended below the bottom of the tube and left it exposed along a vertical face. A vertical plumb-line was dropped from the surface to the bottom of the tube, which was assumed not to have moved since the time of installation. The horizontal distance from this line to the leading edge of the tube was then measured to the nearest millimetre to give an indication of the displacement from the vertical. Annual movement rates were established by dividing these horizontal displacements by the elapsed time since installation. Five tubes were left *in situ* so that long-term movements can be readily monitored in the future.

4.5.3. Stone lobes

One stone lobe was chosen at random in order to investigate the relative movement rates (if any) of the lobe itself, or alternatively of the boulders that constitute it. Several lines were painted vertically and horizontally on the stone lobe in June 1995 in order to test whether or not there is surface movement of the stone lobe (Figure 4.8). After two winter seasons in November 1996, the lobe was re-investigated for visible movement of the boulders.

4.6. DATA ANALYSIS

The recorded temperature data was summarised by taking daily maximum and minimum of the twelve temperature recordings per day. The data was then tested using the Statgraphics (version 7) computer package, and a variety of statistical tests.

An Analysis of Variance (ANOVA) was used, as this tests for the effect of a single factor (e.g. depth, aspect or season) on a variable (e.g. temperature). A Multifactor Analysis of Variance is a simultaneous analysis of the effect of more than one factor on the means. Although an ANOVA tests the null hypothesis (that there is no difference between temperatures), its rejection does not imply that all means are different. Therefore, in cases where a significant difference was found, a Multiple Range Test was conducted to determine where the differences occurred.

5.1. INTRODUCTION

The most consistent and typical feature of periglacial regions is the presence of a surface layer of ground which is subject to seasonal or daily freeze-thaw, and is thus characterised by ground ice (Washburn, 1973; Davies, 1969; Ballantyne and Harris, 1994). This is known as the "active" layer, and is described by Hall (1991a:136) as being typically frozen in winter, and thaws in summer (ie. does not remain frozen). Although distinctive of permafrost areas, ground ice (and thus an active layer) is not limited by permafrost, and is frequently referred to as the 'seasonally-frozen layer' in nonpermafrost areas (Ballantyne and Harris, 1994). Seasonally frozen ground is defined as "ground frozen by low seasonal temperatures and remaining frozen only through the winter" (Muller, 1947:221 in Washburn, 1973). This is the situation within high altitude areas in southern Africa.

The occurrence of ground ice within the active layer may be present from one year to the next (in the form of permafrost); it may appear and disappear seasonally, often in the form of ice lenses; or it may only be of nightly occurrence and produce little more than a crust at or near the surface of the soil (Davies, 1969).

Frost action is the action of frost in repeated freezing and thawing cycles (Washburn, 1973), and is commonly involved in both feature-forming and weathering processes, although the latter often play a secondary role. When water freezes, its volume increases by approximately 9% (Goudie, 1993; Thorn, 1992), so that ground with a large moisture content will display a proportionately greater increase in volume than ground with a small moisture content. During freezing, the growth of ice crystals attract the migration of unfrozen soil moisture to the freezing front, causing gradual segregation of groundwater into favoured areas (Taber, 1930), thereby contributing to the growth of segregated ice lenses (Thorn, 1992). In unconsolidated sediments, this results in local expansion and contraction so that there is upward heaving of some sections to relieve others, large scale frost action activity frequently resulting in the formation of distinctive landforms and sedimentary structures (Ballantyne and Harris, 1994). In this way, ground ice may be considered to act partly as a structure-forming phenomenon (Davies, 1969).

Descriptions already exist of frost action landforms and of other associated features that are currently active and developing under relatively shallow, seasonal ground freezing conditions in the higher mountain ranges of South Africa and Lesotho (e.g. Boelhouwers, 1994; 1995; Grab, 1994; Lewis, 1996). These include patterned ground and cryoturbation phenomena, as well as periglacial slope deposits, such as are discussed in chapters 6 to 9. The nature and significance of these features needs to be considered in terms of the degree of frost action operational in these upland areas.

5.1.1. Active layer processes

The depth of the active layer is often not much more than 0.5m (Ballantyne and Harris, 1994) and responds rapidly to changes in temperature (Kane *et al*, 1991 cited in Koster, 1993), and therefore to the duration and intensity of winter freezing and summer thawing. The depth of the active layer varies from year to year depending on the yearly climatic conditions, and the ice segregation layers may disappear during the warmer seasons (Solomatin and Xu, 1994). The depth of freezing and thawing within the active layer is controlled by several factors, and is therefore subject to considerable local variation (Washburn, 1973). These factors include topography, slope aspect, nature of the substrate, its moisture availability, and vegetation cover, as well as the local climate, and are extensively discussed by Washburn (1973), Harris (1972) and Ballantyne and Harris (1994).

The depth of freeze depends largely on the cumulative number of degree-days below 0°C, whereas thaw results from temperatures rising above the 0°C isotherm. Thawing of the active layer takes place from the surface downwards in response to above-zero air temperatures. Refreezing during winter in seasonally-frozen areas occurs primarily from the surface downwards, as heat is conducted out of the ground by subzero air temperatures (Ballantyne and Harris, 1994).

The geomorphological significance of the active layer undergoing freeze-thaw processes depends largely on the nature of the substrate. In areas of exposed bedrock material, mechanical weathering is likely to be enhanced. In those areas of unconsolidated sediments forming the substrate, the soil granulometry plays a significant role. In fine-grained soils (fine sands, silts and clays), it has been demonstrated that porewater does not freeze at 0°C, but at some temperature slightly below zero (Williams and Smith, 1989; Hall, 1992) and this freezing does not take place instantaneously at a specific temperature, but occurs over a temperature range. Therefore the soil water progressively freezes as temperatures fall. Some soil water may therefore remain unfrozen even at temperatures well below zero (Williams and Smith, 1989). A direct consequence of this is the occurrence of *ice segregation* during freezing of fine-grained soils.

5.1.2. The process of ice segregation

The formation of ground ice is primarily dependent on temperature. Gradual cooling of the ground surface leads to the initiation of soil freezing as the 0°C isotherm moves slowly downwards into the soil and behind it soil water undergoes progressive freezing. Unfrozen water within the soil is drawn to the freezing front (Thorn, 1992), leading to uneven freezing in the soil, and the development of lenses of clear ice within the frozen soil (Ballantyne and Harris, 1994). This is known as *ice segregation*. Thick ice lenses can form as a result of continued contribution of moisture to the freezing front due to pore water migration.

With the occurrence of very rapid freezing of the soil column, the ability of the pore water to migrate is reduced, and the water tends to freeze *in situ* (Washburn, 1973) as interstitial ice.

The amount of ice segregation in the seasonally-frozen layer depends largely on soil granulometry, ground water conditions and moisture supply, and the rate of freezing (Williams and Smith, 1989). Pore size, and therefore grain size, strongly affects the growth and form of ice in the soil (Beskow, 1935) by influencing the amount of water available for freezing, the freezing temperature of the soil water and the movement of additional water to the freezing front (Washburn, 1973; Hall, 1992). There is an increase in contact area between the soil particles and the pore water surrounding them as grain size is reduced, and an increased tendency for surface effects to keep pore water from readily crystallising (Washburn, 1973). Fine-grained soils are characteristic of small pores between the granules, resulting in these soils experiencing a considerable depression of the freezing point, allowing the pore water to remain mobile for longer. Silty soils in particular display an increase in potential for drawing water to the freezing front (Williams and Smith, 1989). Coarse grained sediments lack this suction potential and tend to experience the development of interstitial ice rather than ice segregation. In an open system, water is continuously available for movement to the freezing front, facilitating the development of large ice lenses which are responsible for frost heave,which is at its maximum within fine grained silty soils (Beskow, 1935). These soils are therefore termed *frost susceptible* (Beskow, 1935; Harris, 1981a; Ballantyne and Harris, 1994).

Beskow (1935) presented an empirically derived grading such that soils with finer grain size were defined as frost susceptible, and soils with coarser grain size as non frost susceptible (Figure 5.1). Although clay soils may be susceptible to frost, low permeability may restrict water migration and hence overall lens growth.

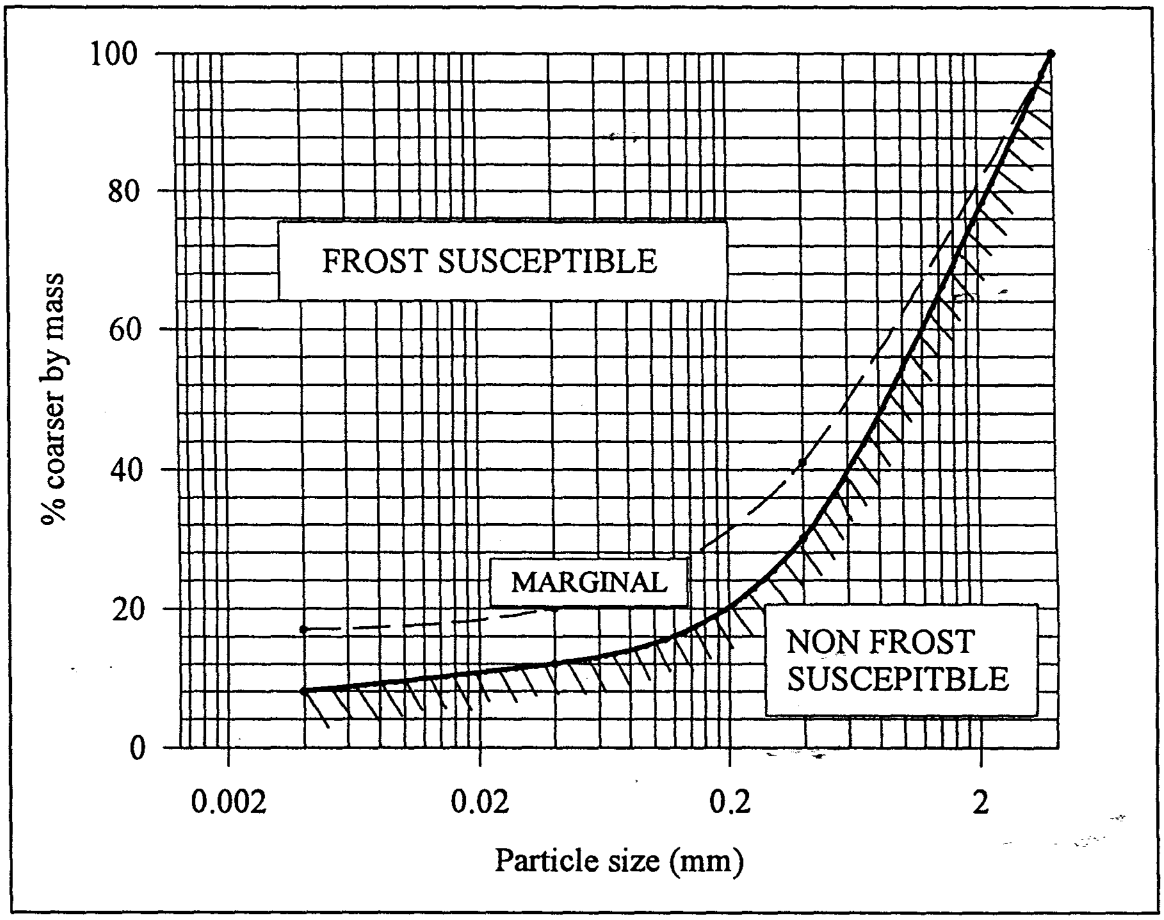


Figure 5.1: Beskow's frost susceptibility limit curve in relation to grain size (Beskow, 1935).

Soil freezing is a complicated thermodynamic process, and is largely influenced by the climate, soil moisture and rate of freezing. Firm explanations for the processes acting are not completely established, and detailed theoretical explanations of the mechanisms operating to produce segregation ice are complex. Much of the research on the topic has been summarised by Washburn (1973), Harris (1981a), Williams and Smith (1989), Ballantyne and Harris (1994).

The majority of ice lenses associated with ice segregation appear as horizontal layers of ice within the sediment. These lenses exhibit great structural variations, and, with adequate water supply, ice lenses ranging in thickness from less than a millimetre

(Washburn, 1973) to many centimetres are commonly produced in the active layer. Such lenses can be extensive, for example, a continuous 13m lens was discovered in a trench in permafrost in Thompson, Manitoba (Egginton and Dyke, 1990). With other conditions being equal, ice lenses normally decrease in size with depth (Washburn, 1973).

Ice segregation within seasonally-frozen soil causes a marked increase in volume, resulting in upward expansion of the ground surface in order to relieve the subsurface stress. Such *frost heave* progressively raises the level of the ground surface by an amount equal to the total thickness of ice lenses, plus additional for any interstitial freezing that may have occurred. If groundwater seepage in the lower part of the seasonally-frozen layer maintains a water supply during the entire period of freezing, segregation ice may develop throughout this layer, and frost heaving of the surface will be considerable. Where there is no such external water supply available, ice lenses are generally concentrated near the ground surface. In its frozen state therefore, the seasonally-frozen layer (or parts of it) may contain a considerable volume of clear ice lenses (Ballantyne and Harris, 1994). This volume of frozen water is in excess of the natural saturated water content of the soil in its unfrozen state, due to excessive migration of additional pore water towards the freezing fronts during freeze-up. The presence of such ice lenses has important consequences during thaw, as an amount of water that cannot be accommodated in the available pore space is released into the soil profile. Discharge of this excess water is necessary before the soil can readjustment to a stable balance, and is often liberated by gradual seepage away (Ballantyne and Harris, 1994). In cases where discharge is inhibited and the soil remains saturated, movement processes such as gelifluction may occur (Chapter 6).

Ice segregation during winter freezing and the release of excess water during summer thaw consolidation are responsible for the formation of a variety of periglacial phenomena, as well as for cryogenic deformation of near-surface sediments (*cryoturbation*), in both permafrost and nonpermafrost areas.

5.1.3. Frequency of freeze-thaw cycles

The frequency of freeze-thaw cycles is a very important control in the effectiveness of frost action processes. The frequency of freeze-thaw cycles is taken as the number of times the air temperature falls below the freezing point (0°C) and then subsequently rises above it again (Nyberg, 1993), but the differences between air and ground temperatures must be regarded when applying the results to frost action in the ground (Washburn 1969, 1973; Harris, 1972; Hall, 1980; Boelhouwers, 1991a; Thorn, 1992; Nyberg, 1993).

Large discrepancies, in the range of 30°C, between air and ground temperatures as a result of insolation, are well illustrated in the Mesters Vig district of Greenland by Washburn (1969). It is possible for ground temperature to fall below 0°C at air temperatures above freezing, and vice versa. The frequency of cycles beneath the surface of the ground may vary widely from the surface frequency because of insulating effects and the nature of the substrate, and tend to be reduced even at slight depths (Washburn, 1969). As air temperatures are regarded as an inadequate measure of ground conditions (Thorn, 1992), soil freeze-thaw activity requires multiple *in situ* measurements (Nyberg, 1993). For ice nucleation to occur at the freezing plane, Lawler (1988) has suggested that the surface temperatures must oscillate between 2°C and -2°C.

5.2. FROST HEAVE AND NEEDLE ICE ACTIVITY

The significance of an environmental process is often evaluated in terms of its magnitude (or intensity) and frequency of operation (Wolman and Miller, 1960)

5.2.1. Frost heave

Hopkins and Sigafoos (1954) described frost heaving as being the predominantly upward movement of mineral soil during freezing. Taber (1952) showed that pressure generated by the growth of ice crystals is perpendicular to the freezing isotherm and that freezing extends downwards from the ground surface, displacing mineral soil upwards.

The occurrence of seasonal ground ice is extremely important with respect to the way in which the differential expansion and contraction (that is, frost heave) induces sorting of particles and their movement in lateral, upward and downward directions (Corté, 1966). Superficial ice segregation formed diurnally also result in lateral movement and sorting, but may often be formed in ground too shallow for the vertical movements they cause to be of major significance.

5.2.2. Needle Ice activity

The most typical form of superficial ice segregation is that of needle ice, a localised and small-scale heave phenomenon produced by ice segregation at or just beneath the surface. Their growth relies on a balance between temperature, moisture and soil conditions (French, 1976). Needle ice, or 'pipkrake' (Taber, 1918, cited in Washburn, 1973), has been defined by Washburn (1973:81) as "an accumulation of slender, columnar, bristle-like ice crystals (needles) practically at, or immediately beneath, the surface of the ground".

As it is a direct result of the cooling of the ground surface which leads to their formation, the ice crystals grow upwards in the direction of heat loss, and are therefore generally oriented normal to the cooling surface. They are capable of lifting earth particles and stones several centimetres above the ground surface. The needles can range in length from a few millimetres to several centimetres. Commonly reported needle lengths vary between 10 and 50mm, which is consistent with a single night of freezing (Lawler, 1988). According to Beskow (1935) needle ice has been observed to grow up to about 20cm in height. The needles produced are commonly near vertical structures, except for long needles that tend to curve (Washburn, 1969).

Needle ice development is widespread throughout the world, the global distribution of which has been examined in detail by Lawler (1988). However, as was noted by French (1976), needle ice is particularly common in alpine locations where the frequency of freeze-thaw cycles is at its greatest.

Needle ice development is restricted by thermal controls, which provides a basis for defining possible needle ice areas. There are several requirements for needle ice formation. Outcalt (1971:395) suggests that

1. an equilibrium surface temperature at least as low as -2°C is necessary for ice nucleation (although subfreezing temperatures maintain needle growth);
2. low soil water tensions (high soil moisture contents) are required for ice segregation to take place; and
3. unfrozen moisture must migrate to the freezing front at a sufficiently fast rate to match the liberation of latent heat at the plane of segregation.

This last condition can only be fulfilled if the pore size of the growth medium is appropriate, as has been discussed above (section 5.1.2.). Meentemeyer and Zippin (1981) found optimum needle ice development in soils with 12% to 19% silt-clay content (ie. grain size <0.063mm).

The occurrence of needle ice is largely restricted to shadow-rich sites with ample moisture supply and sparse vegetation cover. Boelhouwers (1991a) has reported widespread needle ice activity particularly on the south-facing slopes and the poorly drained summit plateaux, and at footpaths and at risers of terracettes in the Natal Drakensberg, South Africa. Although the effects of needle-ice activity are commonly seen in a number of places, needle ice itself is seldom observed in its full extent due to early morning thaw (Washburn, 1969). It is also possible for needle ice activity to persist beyond the winter months and to occur at almost any time of the year if the correct balance between temperature and soil moisture exists (Figure 4 in Lawler, 1988).

Little information exists on needle ice growth rates, and the need for comprehensive research on absolute growth rates, as well as their controls, such as cooling rate, moisture supply, and hydraulic conductivity of the host medium is stressed by Lawler (1988). Local site factors will play a significant role in determining the nature of needle ice activity. For instance, as shown by Lawler (1986a) river banks seem highly conducive locations because of relatively high soil-moisture contents and the high availability of additional abundant water to the ice segregation regions. In addition, they may exhibit a general lack of vegetation, and be subject to lower wind speeds due to their position within the landscape. These controls are shown by Outcalt (1973, cited in Lawler, 1988) to increase needle ice length.

In the field, needle ice growth is usually associated with diurnal freezing and thawing (Outcalt, 1969), and is an important factor in periglacial creep. Needle ice activity also assists the sorting process through small-scale differential heaving on the surface of miniature polygons, and may give rise to the development of microhummocks.

5.2.3. Field studies of frost heaving

A need to monitor the changes in ground surface elevation has been expressed by Lawler (1988). The majority of frost heave studies in the field have, however, been made in areas of patterned ground rather than on slopes (Harris, 1981a). The measurement of these frost-action and mass-wasting effects have used several techniques, including employment of the heavometer (James, 1971), frost heave frames (Jahn, 1961, cited in Harris, 1972) and target cones (Washburn, 1958; 1969). Detailed descriptions of these forms of apparatus are reviewed by James (1971).

Resultant data from these experiments shows that the moisture content of the mineral soil, the vegetation, and the depth of the target insertion are critical variables controlling target heave, and are thus reflected in the results. Greatest amounts of recorded heave were confined to characteristically wet areas, with thin snow cover and sparse vegetation.

5.3. FROST ACTION AND FROST SHATTERING

Frost action is not only evidenced in sediments, but also in the provision of debris by mechanical breakdown. Shattered rock is a typical feature of cold climate environments, and in periglacial regions is considered as a major process in landscape development (Tricart, 1969; French 1976). Frost shatter is considered one of the basic geomorphic processes operating to produce such debris (Lautridou and Seppälä, 1986). However, the efficiency of the freeze-thaw weathering processes in the role of bedrock disintegration requires careful consideration with regards to its place of operation, its relationship to other weathering processes, and its role in landform development (French, 1981; Hall, 1991b).

Weathering of the bedrock is rarely by freeze-thaw alone, and several weathering processes such as freeze-thaw, wetting and drying, salt weathering and natural biological activities are usually found to act in combination with one another (Hall and Otta, 1990; Hall, 1991c). An understanding of the interaction between freeze-thaw processes and other processes will result in a more accurate view of the processes aiding bedrock disintegration and the formation of typically angular clasts.

The nature and efficiency of freeze-thaw weathering is affected by rock temperature, rock moisture content, physical properties (including strength) of the rock, and the rate of operation - the quantification of which has recently been attempted in the field (*e.g.* Fahey and Lefebure, 1988; Matsuoka, 1990). The rate of frost shatter is dependent on the frequency and magnitude of the frequency of freeze-thaw cycles (Davies, 1969; Watson, 1969), the abundance of freeze-thaw cycles being noted in laboratory experiments as more important than their duration in causing frost shatter (Thorn, 1988; Goudie, 1993).

An assessment of the effects of frost shattering processes with regards to rock breakdown in an area requires an evaluation of the degree of temperature depression, the amount of water within rock, and the occurrence of any other operative processes.

With respect to the potential for freeze-thaw weathering being operational at present, actual rock temperature data, rock moisture data and clear evidence of interstitial rock water freezing are required for evaluation to be possible. It has been shown by many workers that minimum temperatures of -5°C for approximately 10 hours or more are needed for rock damage to result because of frost weathering (Hall, 1991a).

5.4. FROST ACTION IN SOUTH AFRICA, AT TIFFINDELL

Although widespread frost heave and needle ice activity is often recorded within periglacial environments, there is a distinct need for an improved knowledge of the actual periglacial conditions occurring at specific sites (French, 1976; Thorn, 1988; Nyberg, 1993). In order to fulfil this need, an assessment of the controls on soil frost processes must be undertaken, and two groups of factors must be considered. These are climatic factors and material properties (Krantz, 1990; Boelhouwers, 1995). Relative air and regolith temperatures as well as the properties of the regolith, including the amount of soil moisture, need to be recorded. This data can then be used to establish the intensity, frequency and duration of freeze-thaw cycles (that is annual, seasonal or diurnal) occurring within the area, the soil moisture availability during these cycles, and thus the effectiveness of formation of ground frost features (Krantz, 1990).

Apart from isolated studies of temperature conditions in the field (*e.g.* Mark and Bliss, 1970; Harris, 1972; Hall, 1980; Matsuoka, 1990; Nyberg, 1993; Orwin, 1993; Mark, 1994), temperature data are still rarely available from true high mountain terrain. In the event of meterological screen data from a mountain locality existing, studies of frost action within the ground are still limited, as an often large discrepancy exists between the air and ground temperatures (Washburn 1969; Harris, 1972; Hall, 1980; Boelhouwers, 1991a; Thorn, 1992).

Within the high altitude areas of the East Cape Drakensberg, no measurements of air and regolith temperatures, or of soil moisture were recorded prior to the present study. The closest established recording station to the uplands is at Rhodes Police Station, at a comparatively low altitude of 1800m. Therefore, in order to report on freeze-thaw activity, and discuss the geomorphic consequences of frost action in the Tiffindell area, air and soil temperatures, soil moisture (Chapter 4) and rates of gelifluction (Chapter 6) were measured in 1995 and 1996. Although the records are of a short-term nature, they still provide some grounds for a discussion of freeze-thaw activity and its geomorphic effects at high altitudes in the area.

5.4.1. Temperature variability

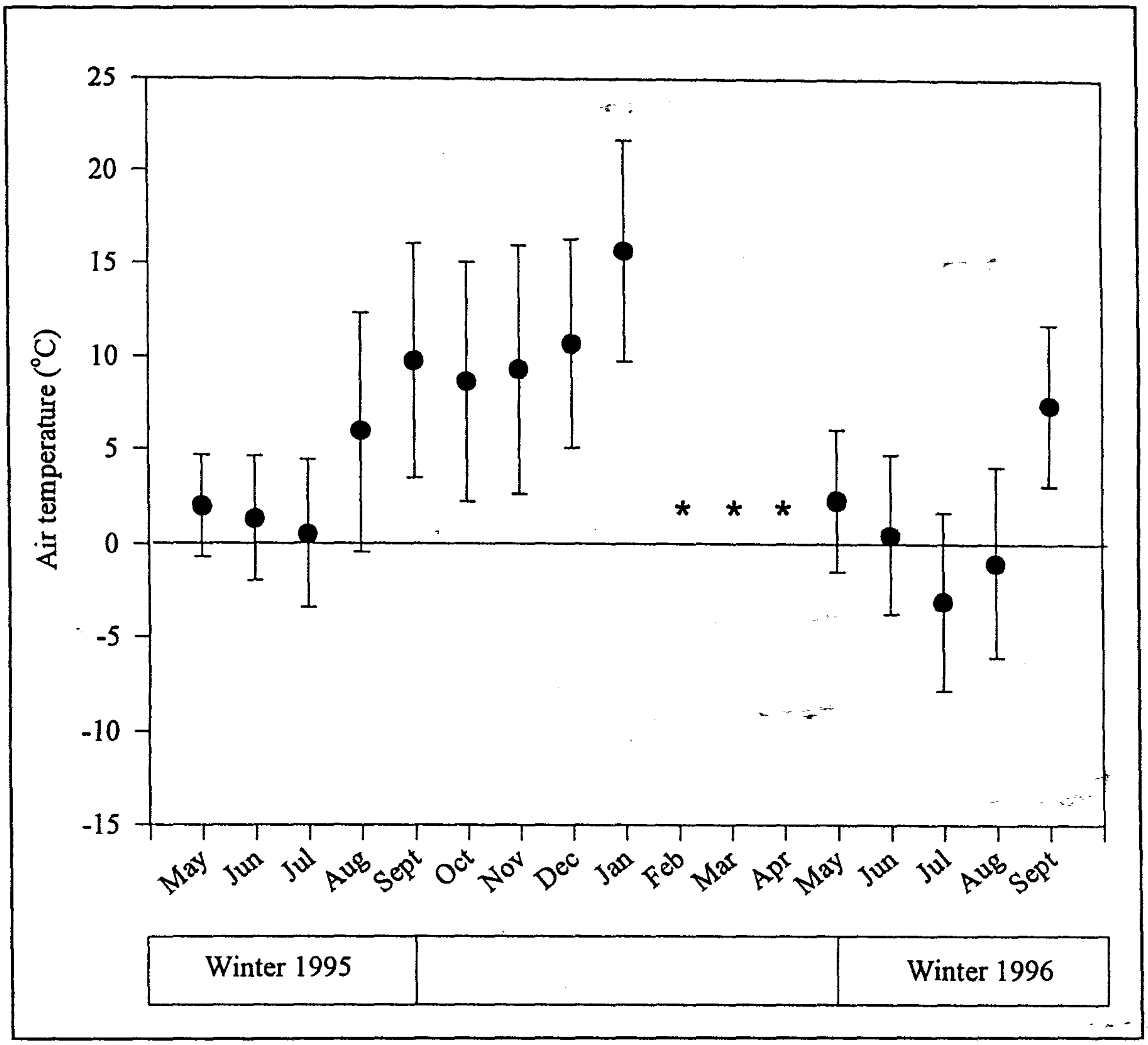
A nearly continuous record, extending from 27 May 1995 to 15 September 1996, of air (1.2m) and soil (ground level, 0.05m and 0.2m) temperatures was obtained with a long-term recorder at Tiffindell. Both the winter and spring conditions for each year were monitored for the air and the microclimatic environment within the soils. These temperature conditions were recorded at three different locations within the Tiffindell area. One station (Station A) was located at 2822m on the southeast-facing slope, the other two on the south-facing slopes at 2788 and 2791m (Stations B and C respectively; Figure 4.1; Appendix I). The data for these locations was not completely continuous due to equipment failure, and station B had the best record for the time period (Appendix II). Data from station A is the most incomplete, and is therefore used infrequently in the following discussion. Additional records from air temperature records taken at 2735m at Tiffindell resort (usually only within the skiing season), the station at Rhodes, and other high altitude stations obtained via the Computing Centre for Water Research (CCWR) provided additional data for the region, and also served to confirm the values recorded by the data loggers.

5.4.1.1. General air temperature conditions at Tiffindell

Air temperature extremes during the 16 month period at station B ranged from 26.22°C (14 January 1996) to -12.8°C (16 July 1996). Mean daily maxima for monthly periods were measured from 21.3°C to -0.12°C; mean daily minima from 9.75°C to -5.9°C. The recorded mean daily values fell from 16.3°C in January 1996 to -3°C in July 1996. Therefore, the range in mean monthly air temperatures was approximately 19°C.

Variations in soil temperature were also evident. Monthly extremes were relatively wide during mid-summer, and ranged over more than 10°C (*e.g.* January maximum and minimum means (1996) ranged from 21.9°C to 11.14°C respectively), but tended to narrow during winter (± 5°C range; July maximum and minimum means (1996) from -1.8°C to -6.57°C).

Figure 5.2 shows the monthly range in air temperature at station B from May 1995 to September 1996 at Tiffindell. A clear seasonal range in temperatures is evident from the recorded values, with the winter seasons displaying near zero temperatures. Summer temperatures predominantly remained above zero at this altitude of almost 2800m, with few occasions of air temperatures fluctuating below zero being recorded over the 16 month period. However, the daily range could be more than 10°C between maximum and minimum temperatures. The summer air temperatures showed a considerable decrease with the onset of autumn and winter.



* indicates no data

Figure 5.2: Mean monthly air temperatures (with standard deviation bars) over the 16 month study period showing seasonal variation. Data recorded at station B (altitude 2788m) at Tiffindell.

Temperature data recorded over 16 months at the three stations (A, B, and C) at Tiffindell were tested using an Analysis of Variance (ANOVA) to determine whether the maximum and minimum temperatures recorded at 1.2m in the air and at depths within the soil profile (Appendix II) varied significantly between the three stations. The results indicated that the temperatures varied significantly with depth and between the three stations (Table 5.1). This evidence indicates that temperature conditions are different on the south and southeast-facing slopes.

Maximum air temperatures were significantly lower at station A than at station B, while minimum air temperatures were significantly lower than those at stations B and C (which were similar). Therefore, in general, air temperatures were found to be lower on the southeast-facing than on the south-facing slope at Tiffindell. Maximum temperatures at ground level were similar at all three stations, however minimum ground level temperatures were significantly higher at station B. Maximum and minimum temperatures at 0.05m and 0.2m all showed similar trends, with temperatures at station A being significantly lower than those at station B, and station B temperatures being significantly lower than at station C (that is, A < B < C; Table 5.1). Within the regolith temperatures were lower on the southeast-facing slope than on the south-facing slope.

In summary, temperatures were found to be lowest on the southeast-facing slope at Tiffindell, while there was little difference between the slightly higher temperatures recorded at the two stations on the south-facing slope. Therefore, by predominantly using the data collected from station B, which is situated on the south-facing slope, an idea of the conditions occurring on this slope can be ascertained, and it can be anticipated that conditions on the southeast-facing slope may be more severe. Thus, recorded intensities and duration of freeze-thaw cycles and frost activity at station B can be expected to be higher for station A. Station B would be likely to experience less freeze-thaw activity than stations A or C.

Temperature data extending from May to December 1995 from station B was tested using a Multifactor Analysis of Variance (MANOVA) to determine whether the maximum and minimum temperatures varied significantly between the air and at different depths within the ground (over all seasons), and between different seasons at all measured levels.

Depth	Maximum temperatures				Minimum temperatures			
	Station	**Mean**	**Groups**	**Sig. level**	**Station**	**Mean**	**Groups**	**Sig. level**
Air	A	11.94	X		A	1.32	X	
	C	12.90	X X	p=0.031	B	3.09	X	p<0.0001
	B	13.78	X		C	3.94	X	
Ground level	A	10.22	X		A	0.79	X	
	B	10.45	X	p=7.63	C	0.99	X	p<0.0001
	C	11.07	X		B	3.2	X	
0.05m depth	A	5.36	X		A	2.31	X	
	B	9.81	X	p<0.0001	B	3.49	X	p<0.0001
	C	13.87	X		C	4.94	X	
0.2m depth	A	4.32	X		A	3.2	X	
	B	7.59	X	p<0.0001	B	4.4	X	p<0.0001
	C	8.07	X		C	5.61	X	

<u>Table 5.1:</u> Summary of ANOVA and Multiple Range Analysis results to determine whether there is variation between the temperature conditions in the air, at ground level, at 0.05m and at 0.2m depths between the stations A, B, and C (thus the south and southeast-facing slopes). The mean winter temperatures at 4 levels are significantly different at the 3 stations studied over winter 1995.

Note: the 'X' illustrate the homogeneous groups.

Daily maximum and minimum temperatures at station B vary significantly with depth (MANOVA, $p<0.0001$ and $p<0.014$ respectively; Table 5.2). The mean maximum daily air temperature (8.83°C) was significantly ($p<0.001$, MANOVA) higher than the mean ground temperature (0.53; Table 5.2). No significant difference was found between daily mean maximum temperatures at ground level (0.53°C), 0.05m (0.63°C) and 0.2m (0.87°C) depth. Mean daily minimum temperatures tended to increase slightly with depth. There was no significant difference between the daily minimum mean air temperatures (-0.31°C) and those at ground level (-0.19°C) and 0.05m (0.14°C). The mean temperature of the ground at 0.2m (0.7°C) was, however, significantly higher ($p<0.02$) than air and surface temperatures (Table 5.2).

Depth	Mean maximum	Homogeneous groups		Mean minimum	Homogeneous groups	
Air	8.83		X	-0.31	X	
Ground level	0.53	X		-0.19	X	
0.05m depth	0.63	X		0.14	X	X
0.2m depth	0.87	X		0.69		X
Sig. level	$p<0.0001$			$p<0.015$		

Table 5.2: Summary of results of a MANOVA and Multiple Range Analysis to determine differences in mean daily maximum and minimum temperatures between the air, ground level, 0.05m and 0.2m depth for 1995.

During 1995, differences between the mean daily maximum and minimum temperatures for winter, spring and summer were highly significant (MANOVA $p< 0.0001$ and $p<0.0001$ respectively). Mean winter daily maximum temperatures (2.93°C) were significantly lower than mean spring temperatures (11.82°C) which were, in turn, significantly lower than mean summer temperatures (17.9°C) (Table 5.3). Mean daily minimum temperatures followed the same trend with mean winter minima (0.25°) being significantly lower the in spring (3.64°C) which were significantly lower than mean summer temperature (8.11°C) (Table 5.3.)

Season	Mean maximum	Homogeneous groups		Mean minimum	Homogeneous groups	
Winter	2.93	X		0.25	X	
Spring	11.82		X	3.64		X
Summer	17.9		X	8.11		X
Sig. level	p<0.0001			p<0.0001		

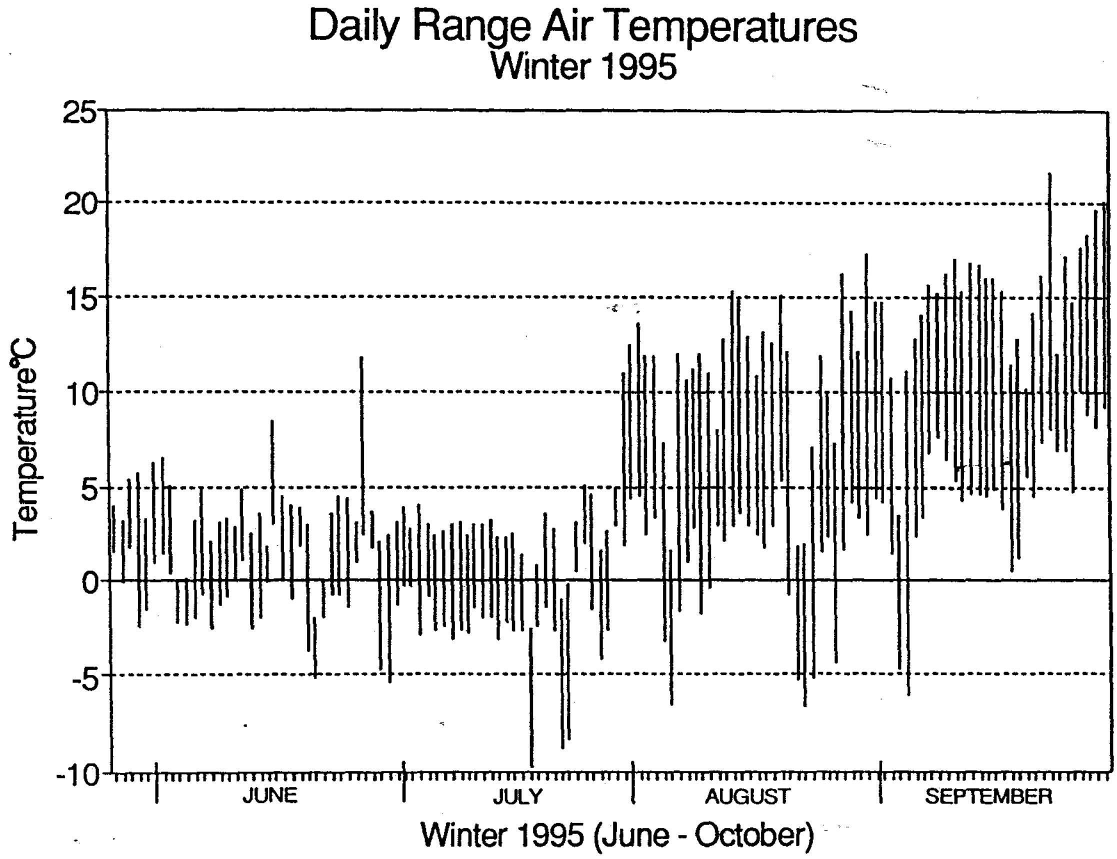
Table 5.3: Summary of the results of a MANOVA and Multiple Range Analysis to determine differences in the mean daily maximum and minimum temperatures over winter, spring and summer for 1995.

5.4.1.2. Winter temperature variations

Figures 5.3a to 5.6b graphically display the daily ranges in air and soil temperatures (at ground level, 0.05m and 0.2 respectively) for winter 1995 and 1996 for the Tiffindell area (note the differences in scale of the graph axes). Table 5.4 provides a summary of the temperatures and freeze-thaw activity occurring at station B for the winters (May to September) of 1995 and 1996 respectively.

	1995				1996			
	Air	Ground	0.05m	0.2m	Air	Ground	0.05m	0.2m
Mean max. T°	11.96	1.78	1.92	0.53	3.66	-0.215	0.289	-0.46
Std. deviation	4.05	2.13	2.16	0.36	4.49	4.66	4.65	4.79
Mean min. T°	1.71	0.45	0.37	0.27	-2.573	-3.712	-3.462	-3.84
Std. deviation	3.64	0.65	0.54	0.38	4.456	4.348	4.407	4.358
Absolute max. T°	17.22	7.89	7.51	1.71	15.7	12.94	13.34	12.87
Absolute min. T°	-6.62	-2.04	-1.84	-1.19	-12.97	-14	-13.81	-14.2
# freeze-thaw days	62	34	45	43	65	34	34	32
# ice days	9	24	10	18	34	74	78	82

Table 5.4: Summary of the temperature and freeze-thaw activity occurring at station B for the winter months (May - September) of 1995 and 1996.



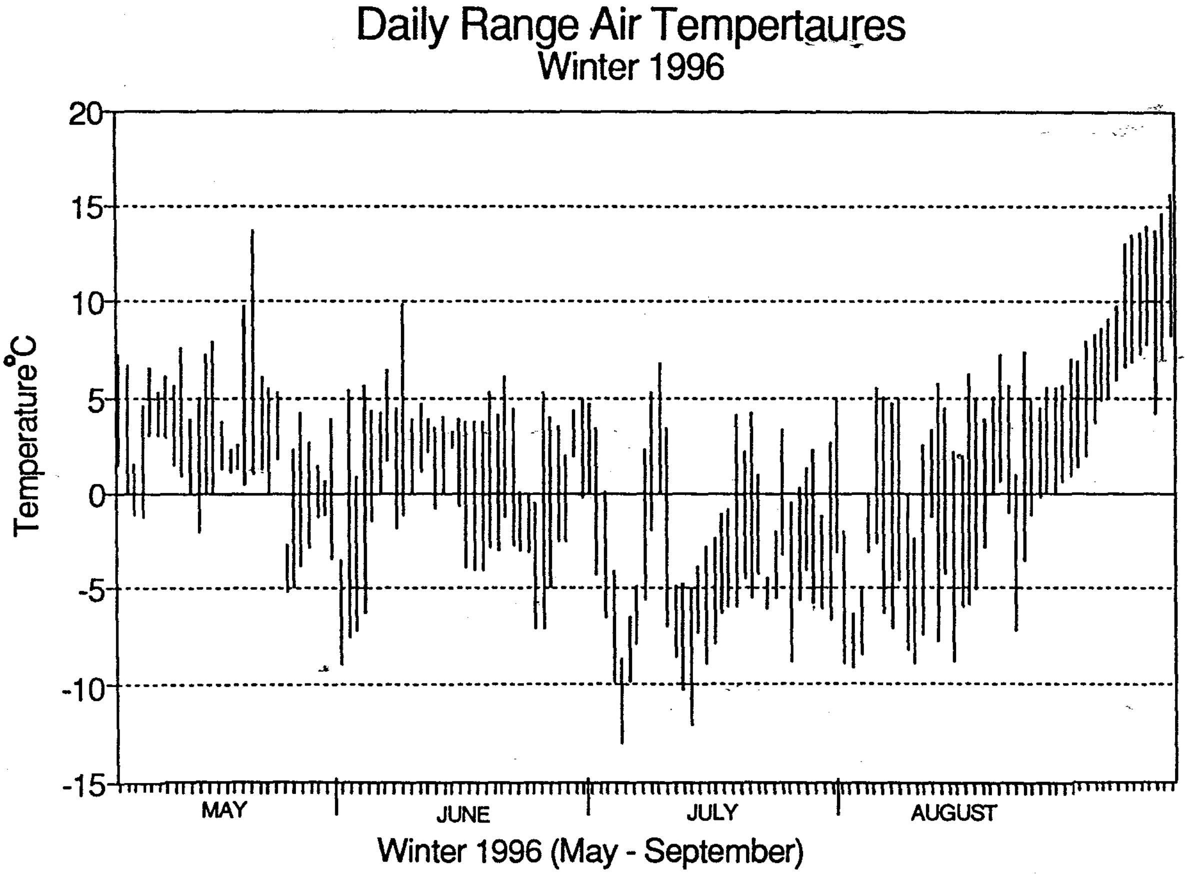
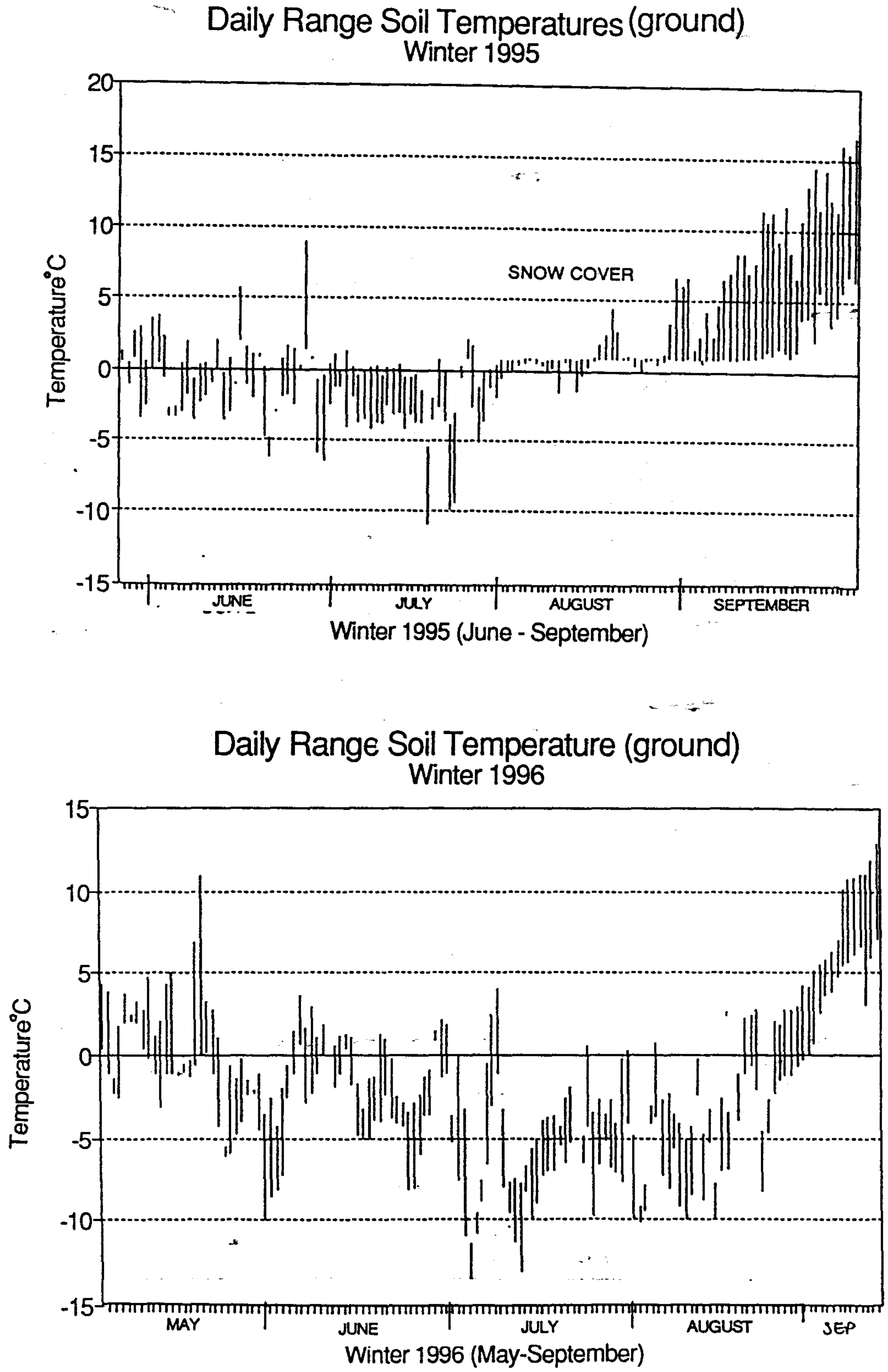
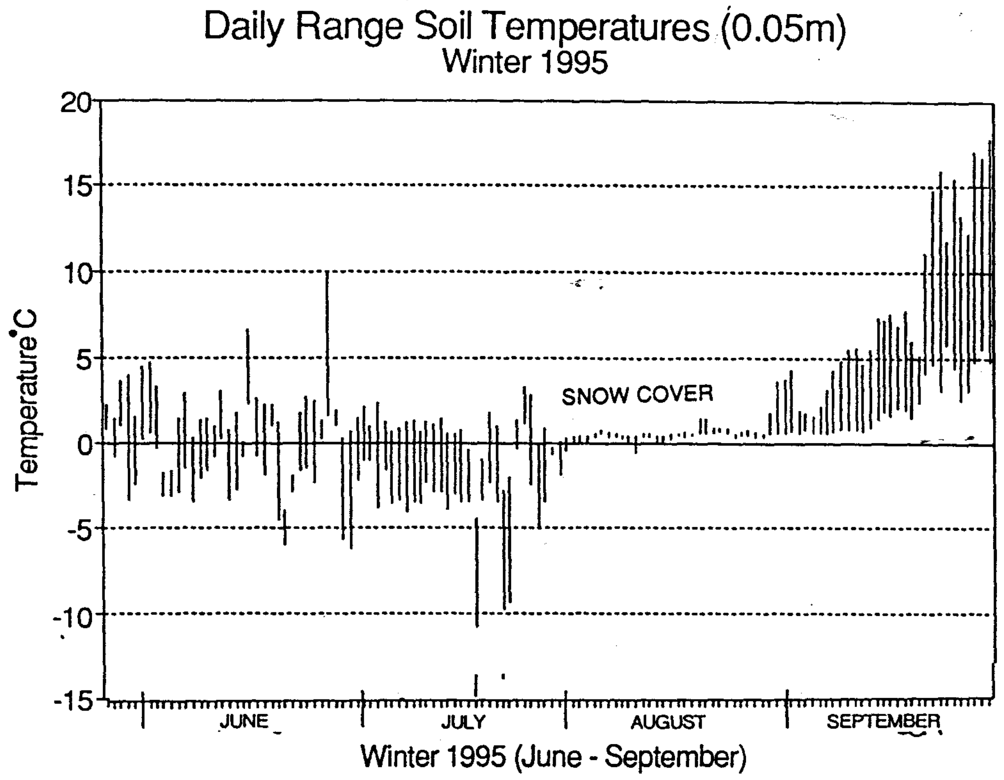
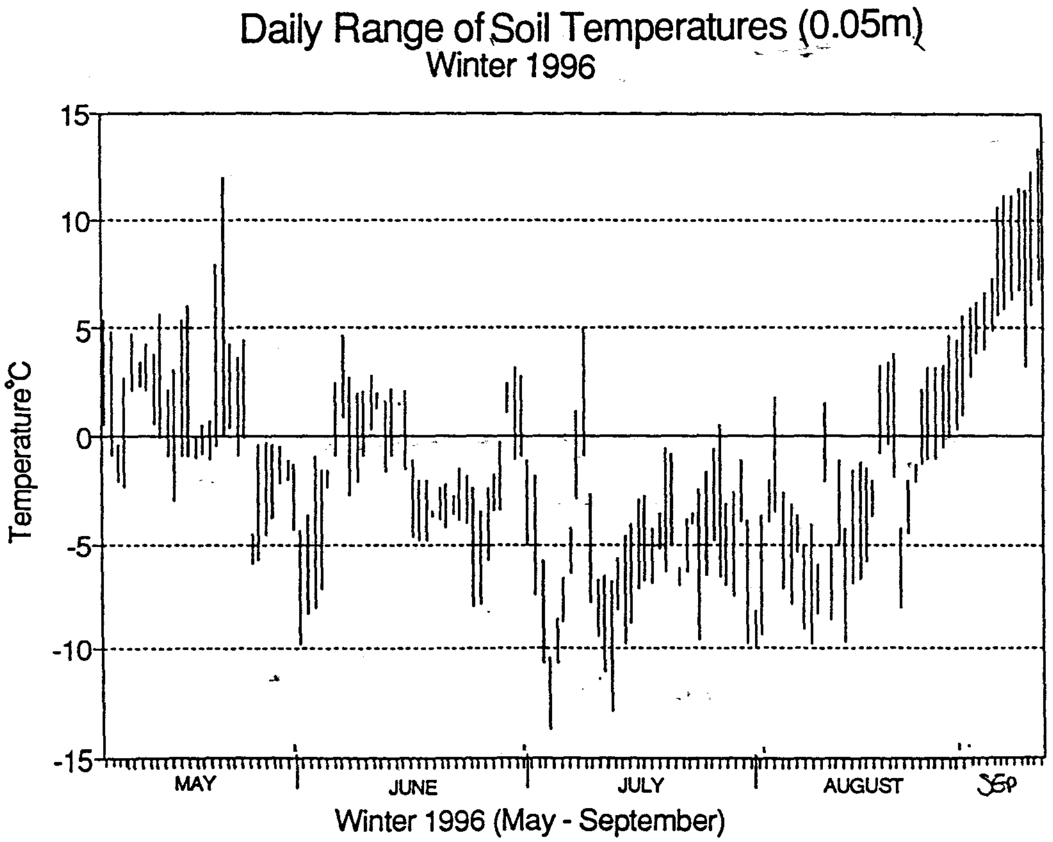


__Figure 5.3 a&b:__ Daily range of air temperatures for winter 1995 and 1996 respectively (at Station B)

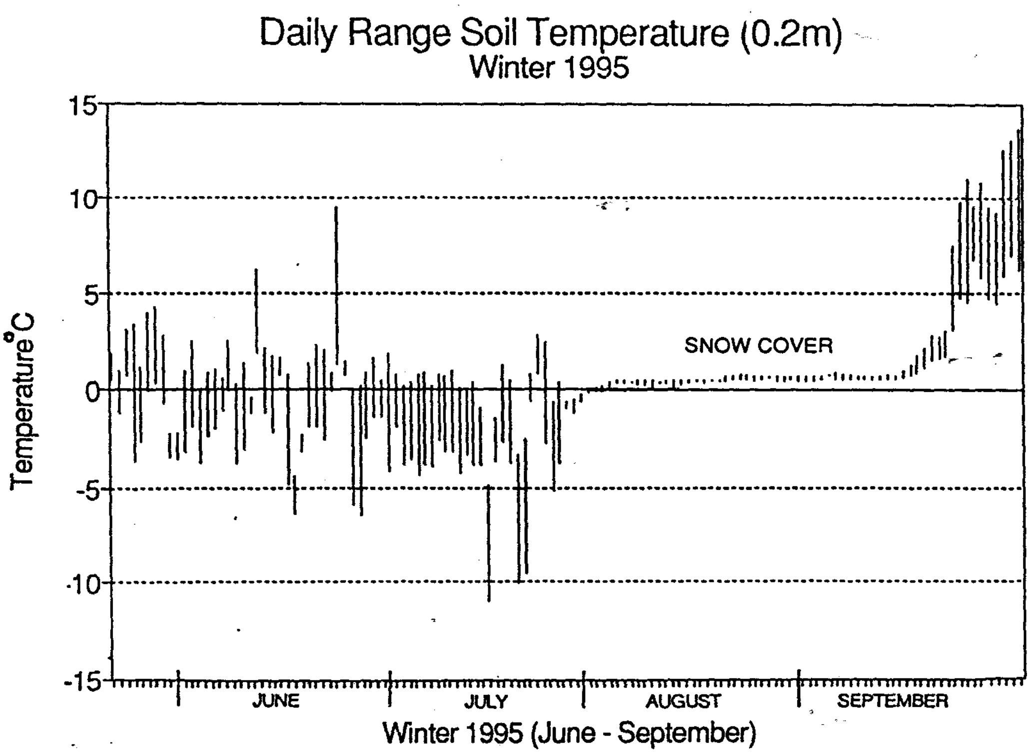


<u>Figure 5.4 a&b:</u> Daily range of soil temperatures at ground level for winter 1995 and 1996 respectively (at Station B)





<u>Figure 5.5 a&b:</u> Daily range of soil temperatures at 0.05m depth for winter 1995 and 1996 respectively (at Station B)



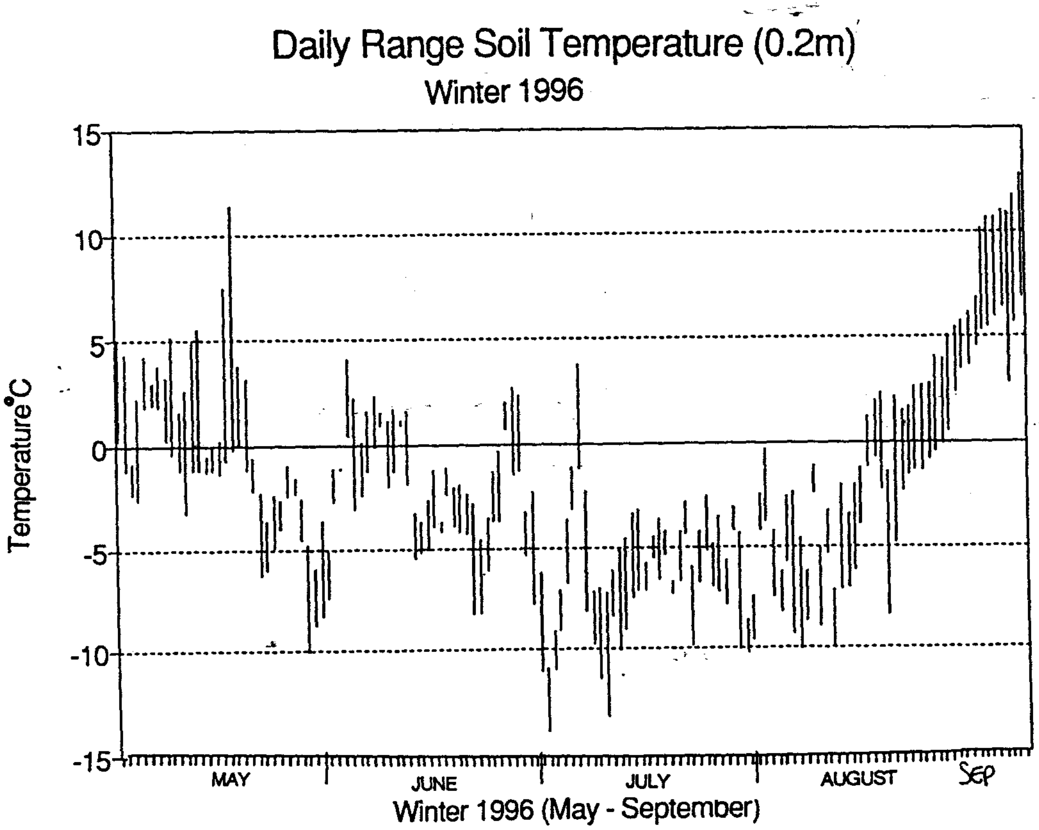


Figure 5.6. a&b: Daily range of soil temperatures at 0.2m depth for winter 1995 and 1996 respectively (at Station B)

Figure 5.3(a & b) display the daily ranges for air temperatures for the winter months of 1995 and 1996. The air temperature data was tested using an Analysis of Variance (ANOVA) to determine whether the maximum and minimum winter air temperatures varied significantly between the two years. Mean maximum and minimum winter air temperatures were found to vary significantly (ANOVA $p<0.0001$ and $p<0.0001$ respectively) between 1995 and 1996 (Table 5.5). Mean maximum winter temperature (2.7°C) and mean minimum winter temperatures (-1.93°C) in 1996 were found to be significantly lower than those for 1995 (10.8°C and 2.28°C respectively; Table 5.5)

	Maximum temperatures			**Minimum temperatures**		
	Mean	**Groups**	**Sig. level**	**Mean**	**Groups**	**Sig. level**
1996	2.7	X		-1.93	X	
1995	10.8	X	p<0.0001	2.28	X	p<0.0001

Table 5.5: Summary of ANOVA and Multiple Range Analysis results for station B to determine whether winter maximum and minimum air temperatures vary between 1995 and 1996.

Figures 5.3a and 5.3b show similar variation in temperature between May and July, July being the coldest month. Extreme differences between the two winters are, however, seen for August and September. 1995 showed a warming trend after the cold month of July, whereas the temperatures for 1996 remained cold, and only began increasing at the beginning of September. Both years showed a relatively rapid rise of temperatures with the onset of spring.

Figures 5.4 to 5.6 show the temperature conditions in the soil over the two winter seasons. These reflect that air and soil temperatures follow similar patterns with regards to possible temperature fluctuations, but the magnitude of variation is less, and appears to decrease with depth. Figure 5.4 displays the daily range of soil temperature at ground level from May to September. As reflected in the air temperatures, May and June conditions were similar for the two years. The recorded July temperatures were also coldest for both years, with 1996 experiencing a daily minima of approximately 3°C less than for 1995 (-6.57°C and -3.42°C respectively).

The conditions recorded in July and August 1996 reflect a cold temperature regime, 47% of the total days being below 0°C (that is, ice days). The conditions for 1995 showed an increase in temperature at the end of July, and August temperatures were recorded above 0°C. This pattern is reflected by the air temperatures and continues with depth.

Figures 5.5a and 5.6a display the soil conditions at 0.05m and 0.2m respectively for winter 1995, the amelioration in August and September temperatures being reflected, as well as a slight delay in temperature increase with depth. This minimal temperature fluctuation with depth may be attributed to a blanket of snow covering the mountains during this period, which insulated the ground from air temperature fluctuations. The result was a reduced variation in soil temperature which rose above freezing level. A few relatively short periods of snow cover were recorded in June and July during 1996, and allowed for greater temperature variations with depth in the soil within that year, as are displayed in Figures 5.5b and 5.6b. This variation between the years' temperatures are marked in the August and September months.

Soil temperature in the winter months of 1996 rarely rose above 0°C, resulting in a considerable number of ice days. Only 13% of the daily temperature ranges recorded for the winter of 1995 remained below 0°C, yet in 1996 the soil experienced largely frozen conditions, with at least 57% of the daily ranges not rising above 0°C. Winter soil temperatures for 1996 showed less daily variation with depth than the air temperature, as was noted for 1995, as well as fewer cycles rising above and falling below 0°C (freeze-thaw cycle). Mean daily soil temperatures were recorded as being below freezing for approximately 3 months of the year (June-August) for 1996.

A period extending from 6 to 9 July 1996 experienced minimal (± 10cm) snowfall, yet experienced continuous below freezing temperatures and frost. The severity of the air temperatures experienced was indicated by the freezing and subsequent bursting of the resort's water-pipes (I.van Eck, pers.comm.). The soil temperatures during this period were recorded as frozen, the maximum being -0.9°C at ground level, and the minimum -14.13°C at 0.2m depth. Soil temperatures showed little rise over the ensuing period, even as air temperatures rose, the minimum recorded soil temperatures all being below -3°C. The fluctuations in soil temperature essentially reflected conditions at the surface, that is the air temperatures. Air temperatures during this period remained cold and continued to drop below freezing at least at night until the beginning of September (Figure 5.3b).

Although a thin cover of snow existed from the 6 to 13 July 1996, temperatures persisted below freezing and continued to fluctuate, and so it can be assumed that the effect of air temperatures must have been greater than the insulating effect of the snow cover, unlike the conditions observed for 1995. As less than 15cm total snowfall was recorded during this period, the relative thickness of the snow cover was minimal by the standards of other alpine areas, and thus its effects could be expected to be overridden by the effects of great air temperature ranges. In August 1995, when the ground temperatures were moderated by the effect of a thicker snow cover, the air temperatures were also generally high, with maxima reaching more than 15°C and few minima falling below 0°C (Figure 5.3a). Therefore, with a less harsh air temperature regime, the soil conditions stabilised somewhat, and reflected only minor temperature changes occurring in the air. As the ground was still being influenced by the snow cover, the temperatures did not increase as much as they would have under snow-free conditions. Therefore, it appears that the amount of soil freezing at Tiffindell is more greatly influenced by air temperature variability than by snow cover. In both years, rapid warming, and thus thawing, of the soil was noticed at the onset of spring and the marked increase in air temperatures.

Large variations were noted between the winter data collected for the air and at various depths within the soil from 1995 and 1996, the latter year experiencing colder temperatures for longer periods of time than 1995. This is illustrated for air temperatures in Table 5.3. The lowest recorded air temperature was -12.97°C in 1996, yet only -9.89°C in 1995. These temperatures obviously had a direct effect on the temperatures experienced within the regolith. These colder air temperatures in 1996 may reflect the cold conditions experienced throughout South Africa in 1996, as evidenced in Figure 5.7.

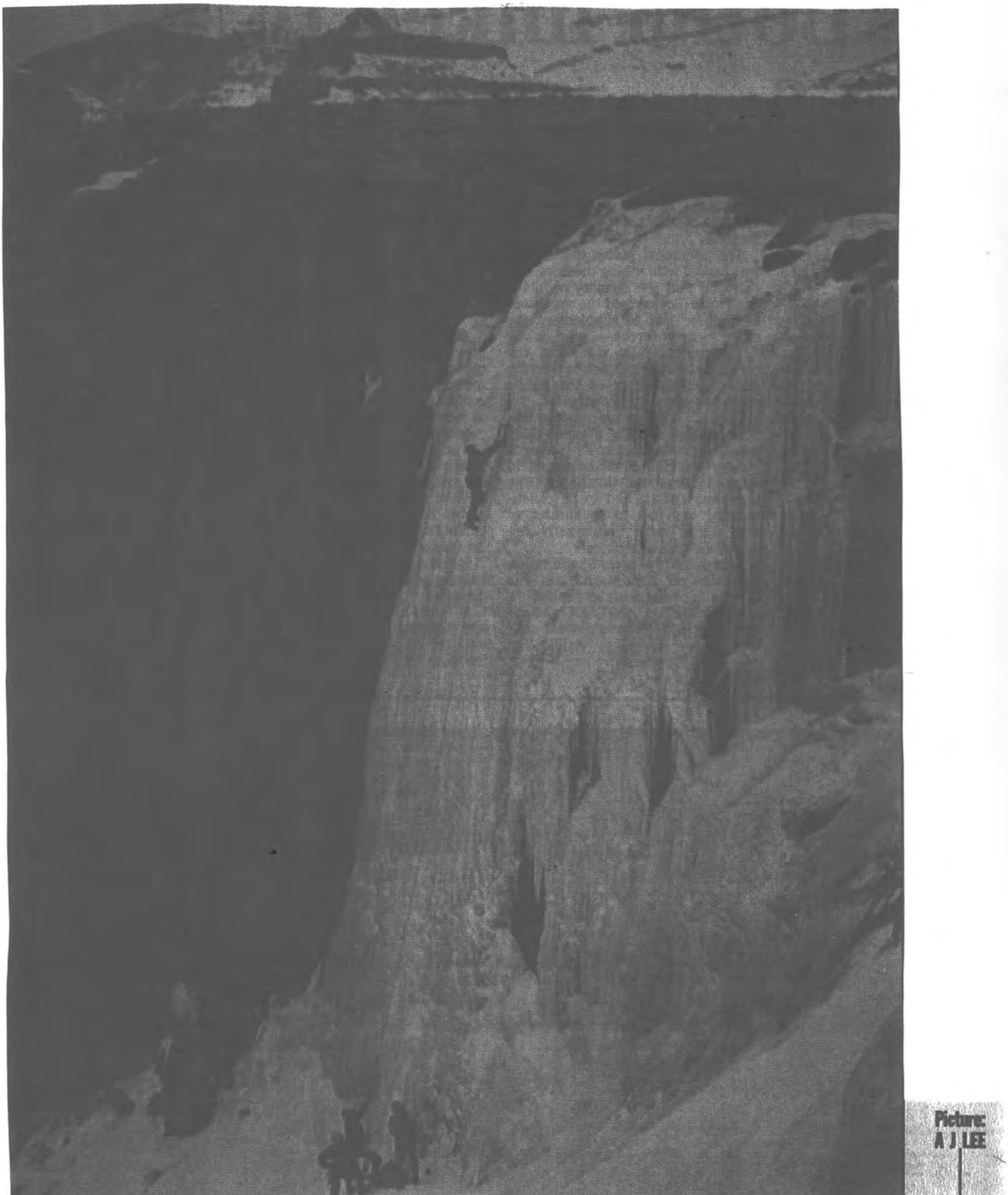


Figure 5.7: Conditions at Natal Drakensberg on the 11 August 1996 (Sunday Times newspaper).

5.4.1.3. Freeze-thaw cycles

A freeze-thaw cycle in this study is defined as a fall in temperature across 0°C and an ensuing rise above. The continuous temperature records for air and soil at the Tiffindell stations indicate the existence of seasonal, diurnal frost heave cycles running from May to September, and sometimes beyond, and were used to determine the number of days when temperatures remained either below freezing point (ice days), or above (frost-free days), or fluctuated above and below (freeze-thaw days). The number of freeze-thaw cycles recorded by air and soil temperatures at various depths were calculated separately, as those freeze-thaw cycles monitored in the air were often not mirrored by cycles at the ground surface. The relative number of cycles occurring from May to October for stations B and C for both 1995 and 1996 are reflected in Table 5.6.

This seasonal period enables effective comparison between the two years, with regards to the periods in which continuous data was collected. It also reflects the most intensive period with regards to the frequency of freeze-thaw occurrence. Few freeze-thaw cycles were recorded during the warmer months after this period, with only 18 additional days experiencing temperatures below 0°C after November 1995 until mid-January 1996 (when the loggers malfunctioned), and recording only resumed again in May 1996. The results from station A are not complete enough to accurately assess the number of cycles, and have therefore been excluded.

The air was frost-free on an average of 52.5% of the days between May and the end of September in 1995, but only on 27% of the days in 1996. These days were distributed over 5 months of the year, but were most common at the start of winter during the beginning of May and then in September with the onset of spring. Approximately 44.5% (± 4.69 s.d.) of the days fluctuated across the freezing point for both 1995 and 1996. The remainder of the days (10.5% and 24.5% for 1995 and 1996 respectively) were frost (ice) days when air temperatures were below freezing. As already noted, 1996 experienced colder temperature conditions than the winter of 1995, and there were more days with temperatures below freezing. During the coldest month, July 1996, 19 (61%) of the days remained below freezing, and 11 (35.5%) experienced freeze-thaw cycles. The longest frost-day period over both winters lasted for 9 days. Data from a meterological station at Ox-Bow Lake (altitude 2591m) revealed that this station only experienced freeze-thaw cycles less than 24 hours in duration, although 90% of the days in July experienced these cycles (P.Illgner, pers.comm.).

YEAR	STATION		FROST-FREE DAYS		ICE DAYS		FREEZE-THAW DAYS	
1995	B	air	82	53.3%	9	6%	62	40.5%
		ground	95	61.5%	24	16%	34	22.5%
		0.05m	98	64%	10	6.5%	45	29.5%
		0.2m	92	60%	18	12%	43	28%
	C	air	79	51.5%	12	8%	62	40.5%
		ground	85	55.5%	22	14.5%	46	30%
		0.05m	96	63%	9	6%	48	31%
		0.2m	91	59.5%	18	12%	44	28.5%
			n = 153 days					
1996	B	air	39	28%	34	24.5%	65	47.5%
		ground	30	22%	74	53.5%	34	24.5%
		0.05m	26	19%	78	56.5%	34	24.5%
		0.2m	24	17%	82	59.5%	32	23.5%
	C	air	36	26%	34	24.5%	68	49.5%
		ground	26	19%	75	54%	37	27%
		0.05m	25	18%	78	56.5%	35	25.5%
		0.2m	24	17%	82	59.5%	32	23.5%
			n = 138 days					

Table 5.6: Freeze-thaw cycle activity in winter (May to October) for Stations B and C at Tiffindell during 1995 and 1996.

The sensor at ground level at station B was markedly colder than the sensor at 1.2m above the ground, and recorded more than twice as many ice days in both 1995 and 1996 (24 against 9, and 74 against 34 respectively), yet less than half the number of freeze-thaw cycles (34 against 62, and 34 against 65, for 1995 and 1996 respectively). With depth within the soil, the incidence of freeze-thaw cycles is reduced (Table 5.6). Approximately 25% of the days between May and September experienced freeze-thaw cycles at depth, compared to 40.5% days in the air. The incidence of ice days, however, shows variation between the years due to the differences in overall temperature. In 1995, only ±11% of the days remained below freezing point at depth, whereas in 1996 an average of 56.5% days were recorded. For both 1995 and 1996 though, more days on average existed below the freezing point with depth in the soil than in the air. Fewer frost-free days therefore existed for the recorded period in 1996 than in 1995 (Table 5.6).

A Multiple Range Analysis test (Table 5.5) indicated that temperatures at stations B and C were statistically similar. Temperatures are, however, likely to vary between stations that are on different slopes or are at different altitudes (Table 5.5). The records from one station may be extrapolated for use in the immediate area surrounding the station, but it is highly probable that they will be inaccurate with any substantial distance from the recording site. This is due to a variety of factors (besides temperature) such as aspect, relative soil composition and moisture contents.

At stations B and C, the majority of the recorded air freeze-thaw events for both years were shorter than 24 hours in duration (Table 5.6). However, at depth, freeze-thaw cycles spanning several days were common (ice days). In 1996, these ranged from 2 days to 9 days, as shown on Figures 5.4b, 5.5b and 5.6b. A notable result from station B was that most freeze phases within the soil were of low intensity, seldom reaching below -5°C in 1995, but decreased temperatures in 1996 resulted in a large majority fluctuating between -5°C and -10°C. The high frequency of freeze-thaw cycles >24 hours in 1996 may be attributed to the lack of a substantial snow cover and to low temperatures.

Figure 5.8 displays a model of the duration and intensity of freezing within the air and soil at various depths for station B in the winter of 1996. This indicates the intensity of freezing temperatures experienced within a cold winter season. Temperatures lower than -5°C were not common within the months approaching winter, and after winter (*i.e.* May and September). As summer progressed, the air temperatures rose quickly. Soil temperatures also showed this increase, although at a delayed rate with depth.

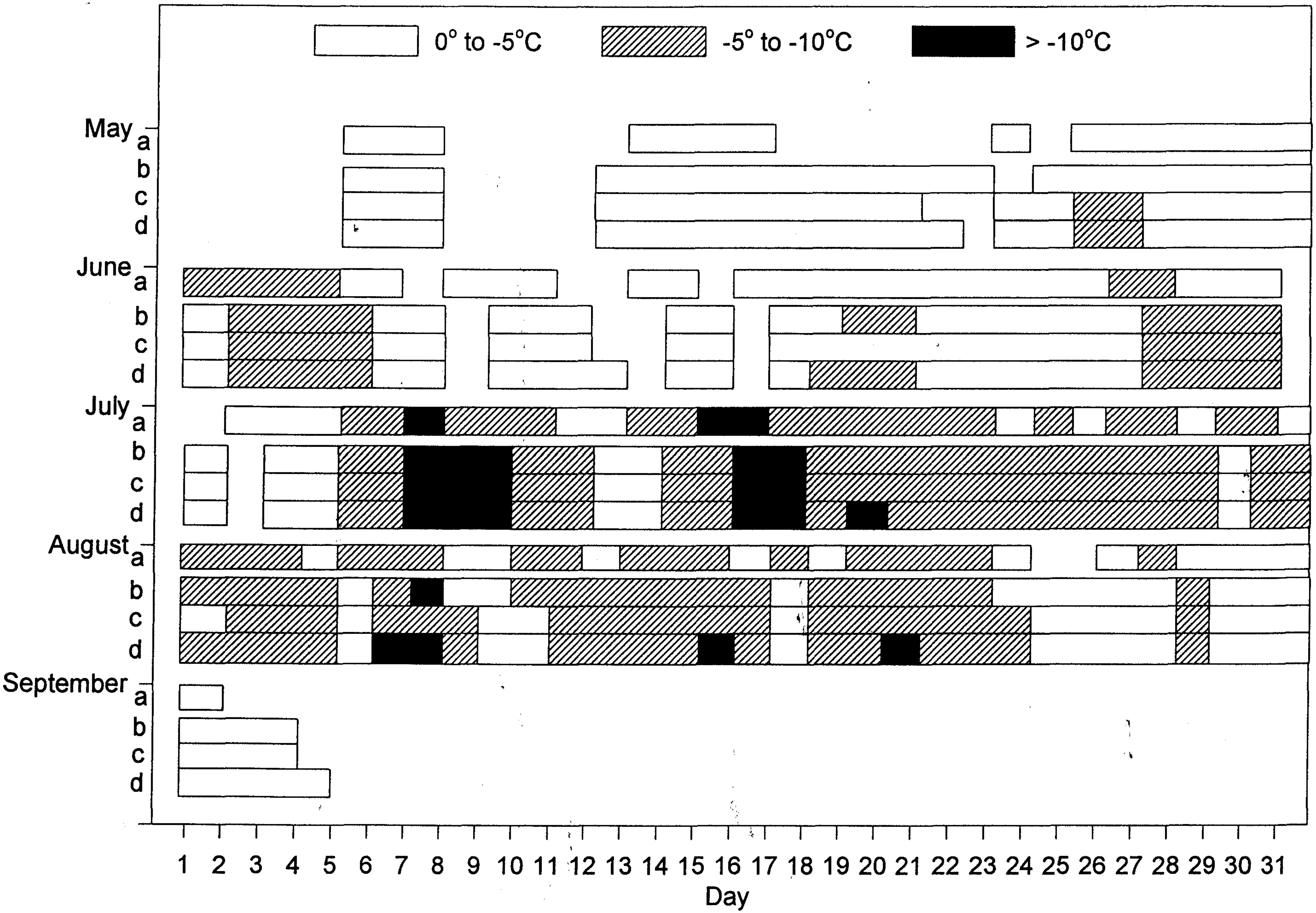


Figure 5.8: Duration and intensity of freezing a) in air, b) at ground level, c) at 0.05m and d) at 0.2m at station B for winter 1996.

This slight lag within the soil temperatures was evident in both a drop or rise of temperatures (Figure 5.8). Therefore, air temperatures can not simply be used to determine those occurring at depth within the soil. The greatest freezing temperatures (>-10°C) were reached in July, which was the coldest month at Tiffindell in 1995 and 1996, and Figure 5.8 shows that the soil maintains these colder temperatures for 2 days after the air temperature has risen again.

5.4.1.4. Conclusions

Three principle aspects of temperature conditions and freeze-thaw activity emerge from the data collected at Tiffindell:

i) monitoring of air temperatures gives a poor indication of temperature fluctuations occurring at or within the ground,

ii) within the periglacial environment at Tiffindell, differences in freeze-thaw regime may occur from site to site, and slope to slope.

iii) frost action processes occur within the soil over a large majority of the days in winter, and although pore water tends to freeze at temperatures somewhat below zero (Hall, 1980), temperatures, especially during 1996, were cold enough to allow for substantial pore water freezing.

5.4.2. Material properties

The rate at which soil moisture migrates through a sediment and towards a freezing plane, resulting in the consequent growth of segregation ice, is known as the *frost susceptibility* of the sediment (French, 1976). Beskow's (1935) established limit for frost heaving in unconsolidated sediments suggests that fine silts and sands are most frost susceptible. Meentemeyer and Zippin (1981) point out that natural frost heaving often occurs in sediments slightly coarser than this. They summarise the conclusions of various other workers by stating: "Soil texture must be neither too fine to retard flow of water to the freezing plane nor too coarse to impede capillarity. The actual movement of water to the freezing plane is controlled by the interaction of moisture availability and texture" (Meentemeyer and Zippin, 1981:114). From their analysis the authors conclude that a minimum of 8% fines (ie. the fraction <63μm) is required for needle ice growth in water saturated soils, although the optimum content is between approximately 12% and 19%.

The combined percentage for silt and clay, which constitutes the fine fraction, has been measured as greater than 8% (*cf.* Meentemeyer and Zippin, 1981) for the unconsolidated surficial sediments at Tiffindell, and thus satisfies the grain size prerequisite for needle ice formation. Sediments sampled at eight sites at Tiffindell consisted largely of silty sand, containing varying proportions of gravel, sandy and silty sized material, the exact proportions varying depending on the area from which the samples were taken (Figures 5.9 and 5.10). Those samples containing a higher percentage of fines suggest that much of the Tiffindell area may be highly frost susceptible (*cf.* Beskow, 1935) and consequently liable to frost heave (Figures 5.1 and 5.9).

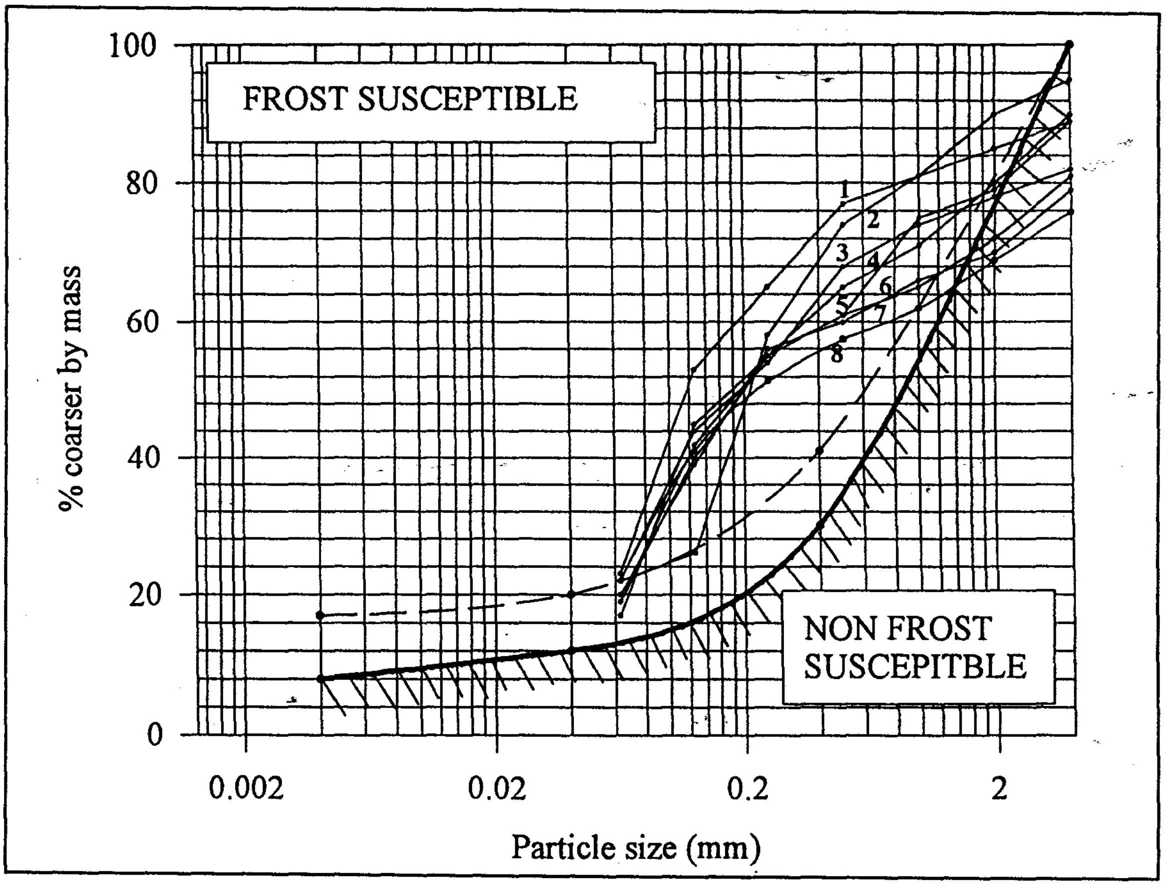


Figure 5.9: Plot of sediment composition of 8 sample points at Tiffindell in relation to Beskow's frost susceptibility limit (cf. Figure 5.1)

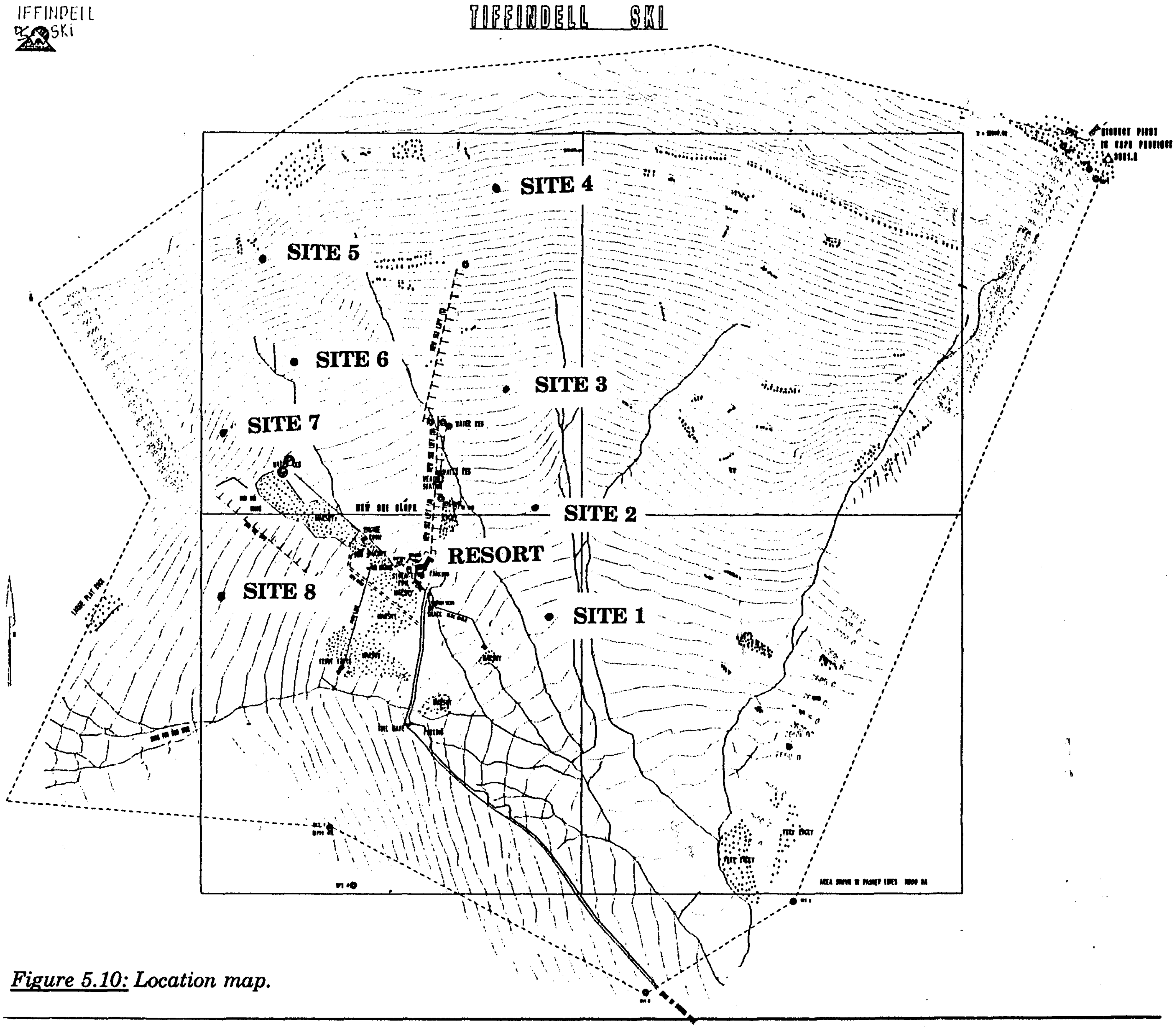


Figure 5.10: Location map.

Liquid limits for the 8 samples were measured below 25%, and plasticity less than 3.5%. The unconsolidated sediments are therefore non-plastic and generally have very low liquid limits. The index properties were found to be fairly constant at several sites, and are considered in Chapter 6, as are the implications of the liquid limits and plasticity investigations.

5.4.3. Soil moisture

Much of the winter precipitation at Tiffindell falls as snow, which normally melts within a few days, resulting in continuous saturation of surficial sediments, producing high soil moisture percentages. The soil moisture levels for station B from June - September in both 1995 and 1996 are presented in Figure 5.11. Figure 5.11 (upper) shows that during the months of June and July of 1995, in which approximately 50 freeze-thaw cycles occurred, the ground soil moisture contents were approximately 50% for both 0.05m and 0.2m depth. As the soil moisture probes record the percent moisture, it could be expected that a frozen probe would measure a value below 100% moisture. From the beginning of August, when soil temperatures rose above 0°C (Figure 5.4 to 5.6), the soil moisture content increased rapidly at 0.05m depth, the soil moisture at 0.2m rising less quickly. This higher percentage of moisture in the upper layers of the soil may be due to snow cover and to moisture associated with its melting. Thawing of the regolith may also be implied due to the apparent 100% saturation of the soil at the upper levels. Trapping of this soil water within the top 0.1m is also indicated by the upper section being typically 20% more saturated than at 0.2m depth. This may indicate that the ground at 0.2m remains frozen for longer than the upper 0.05m (Figures 5.5a and 5.6a), thus forming an impermeable layer, resulting in the water released during thaw of the top soil being trapped, causing 100% saturated conditions.

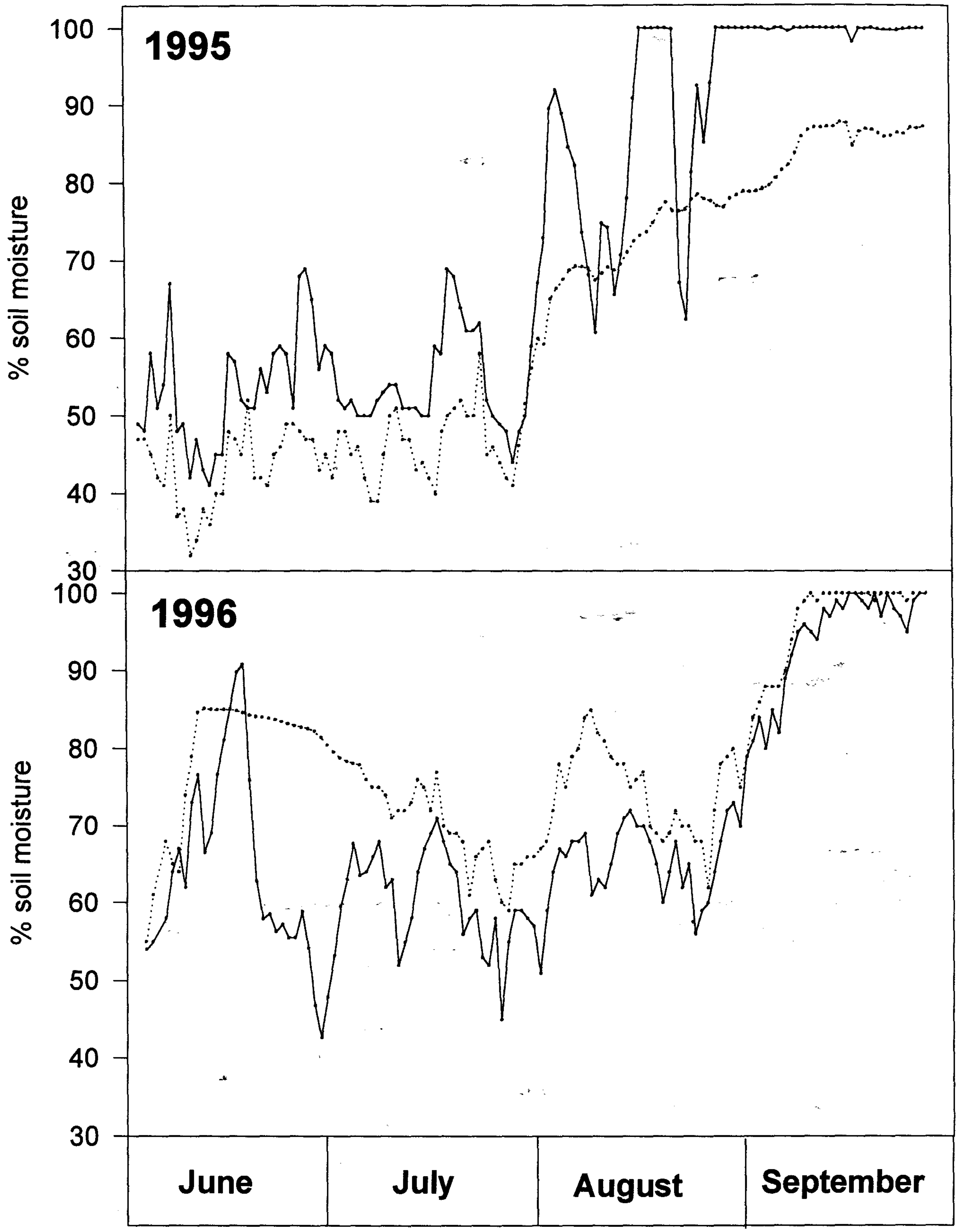


Figure 5. 11: Daily percentages of soil moisture at 0.05m (solid line) and at 0.2m (dashed line) at station B during the winters of 1995 and 1996.

The measured soil moisture contents ranged between 70% and 100% for the remainder of the recorded period (*i.e.* until 15 January 1996). Figure 5.11 (lower) indicates that the winter of 1996 experienced a longer period of freezing before thaw began, as the soil moisture contents only show a marked rise in early September, which correlates with an increase in the air and regolith temperatures respectively (Figures 5.3b, 5.4b, 5.5b, 5.6b). With the onset of spring and warmer temperatures, the soil moisture contents suggest that thawing preceded from above, the soil at 0.05m thawing first, resulting in greater soil moisture percentages in the top 0.05m than in the lower 0.2m. Unfortunately, no data was recorded for autumn, so the relationship between lowering soil temperatures and decreasing soil moisture cannot yet be established.

A difference between relative soil moisture contents between the depths of 0.05m and 0.2m can be seen for 1995 and 1996 (Figure 5.11). This variation can be related to differences in the rates of thaw occurring in the two years, thaw taking place earlier and less suddenly in 1995 than in 1996. The blanket of snow in 1995 contributed to the high soil moisture content, whereas the soil moisture content of 1996, being less affected by snowfall, was a direct indication of the amount of moisture that was frozen within the ground. In both years, impeded drainage due to ground ice at or below 0,2m may have contributed to the high soil moisture values due to meltwater saturating the unfrozen upper soil layers.

In 1996 thaw reached the 0.05m and 0.2m depths almost simultaneously, as evidenced by the rapid rise of ground temperatures (Figures 5.5b and 5.6b). The higher soil moisture content at 0.2m depth may have been due to the lack of an insulating snow cover, so that the upper layers of soil (0.05m) were subjected to greater insolation and wind, and thus greater evaporation than soil at greater depth, resulting in a greater loss of soil moisture.

Saturated conditions existed in both 1995 and 1996 as a result of the onset of warmer temperatures and thaw. Under these conditions, the upper portion of the soil would be susceptible to soil flowage as discussed in Chapter 6.

The ground remained near water saturation levels throughout the warmer seasons of 1995 and 1996. The high saturation levels indicate that moisture availability was not a restricting factor on soil frost action, neither was the soil granulometery. The climatic controls at the logger site also appeared distinctly favourable for the occurrence of effective diurnal freeze-thaw action at the soil surface, as well as within the soil profile, in winter. Surficial sediments were therefore highly susceptible to needle ice activity and to frost heave.

5.4.4. Needle ice activity

The growth of needle ice and evidence of its effects are widespread at Tiffindell throughout the winter seasons. During the winters of 1995 and 1996 needle ice occurred in abundance at altitudes between 3 001m and 2 300m. Needles tended to form nightly rather than on a seasonal basis, and were generally between 0.5cm and 7cm long and were observed to grow within one freeze-thaw cycle (Figure 5.12).

Those of greater dimensions were observed in areas lacking vegetation, yet possessing abundant soil moisture (for example, downslope from watertanks, on the banks of the dams, on the poorly drained summit plateaux, at footpaths and at risers of terracettes or lobes where snow commonly gathers). These areas were often characterised by a series of columnar needles separated by a fine layer of organic soil (Figure 5.13). Outcalt (1969, 1971) has suggested that when material becomes incorporated within the needle ice crystal, this may indicate a period of growth stress, possibly related to a decrease in the amount of available soil moisture, or to an increase in soil temperature.

Where needle ice was observed within vegetated areas, growth commonly resulted in localized vegetation disruption. In bare areas, during growth the needles lift soil particles and small stones several centimetres, and on thaw, the ground surface has a broken and wrinkled appearance (Figure 5.14), forming "nubbins" (Washburn, 1969).

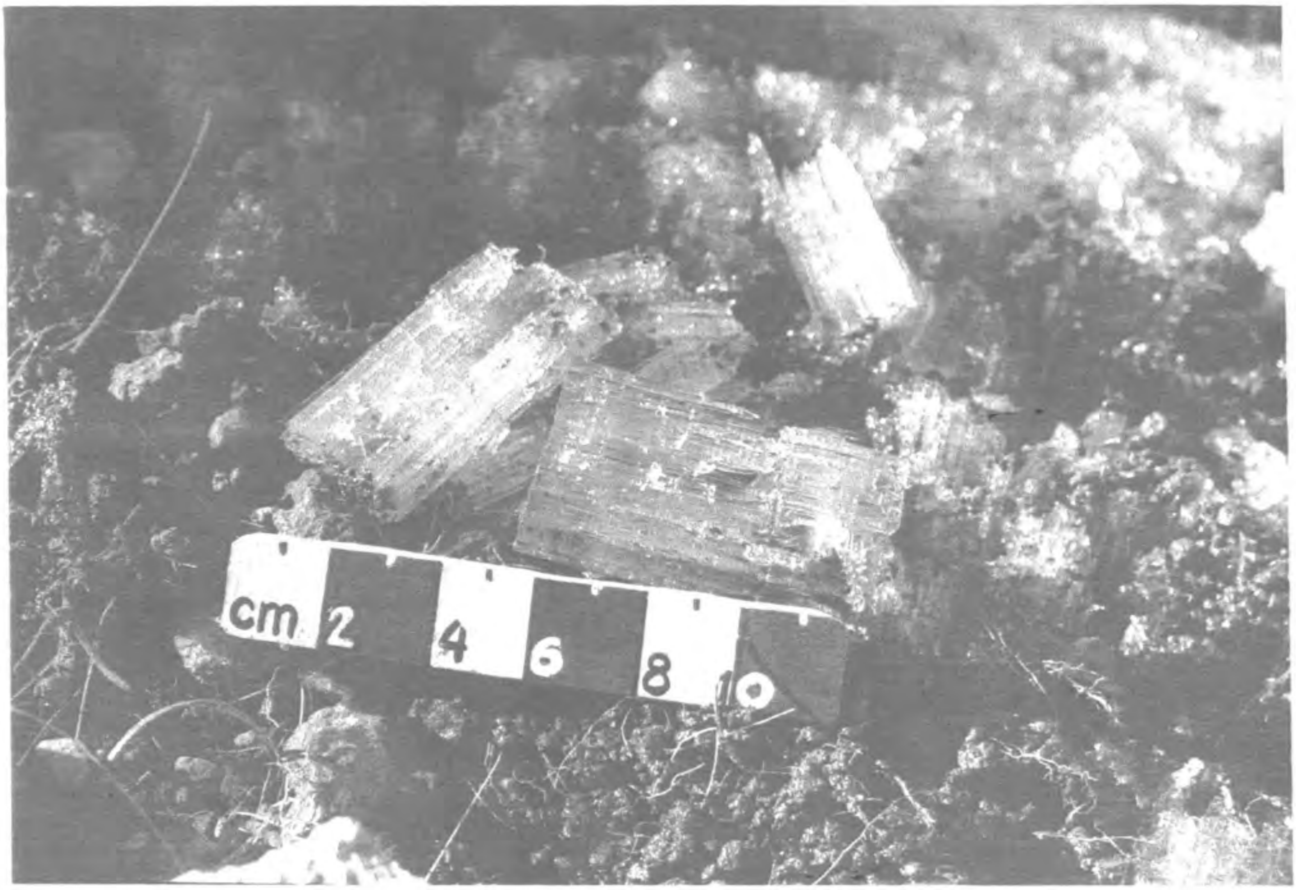


Figure 5.12: Diurnal needle ice development at 2600m in an area of abundant moisture. Needles are up to 7cm in length (July 1996).

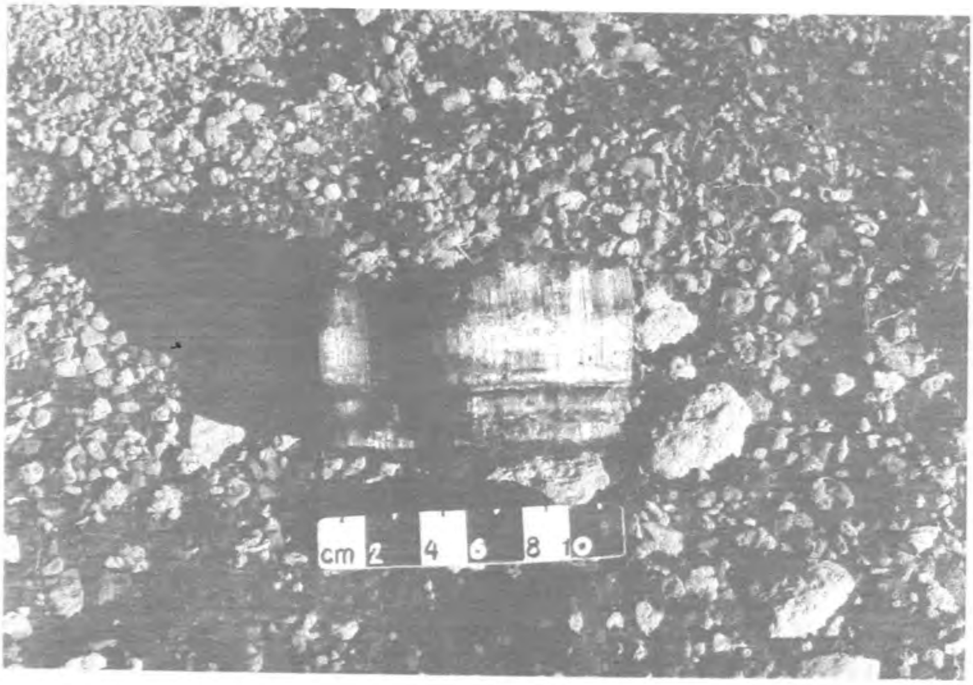


Figure 5.13: Columnar needle ice separated by fine layers of organic soil. The distinct layers of development can be distinguished 5cm, 0.5cm and 0.7cm respectively (2760m July 1996).

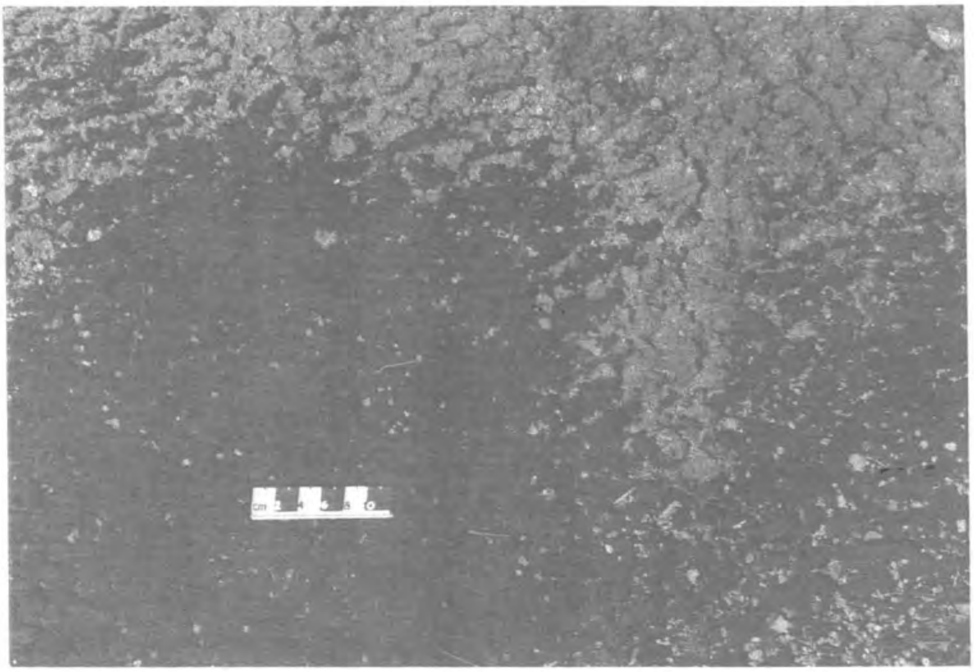
Although needle ice appears to be most common in the winter months, it has also been noted in the warmer seasons, yet with shorter ice crystals than in winter, and only in areas of abundant soil moisture. Boelhouwers (1991a; 1994) has reported widespread needle ice activity at high altitudes on south-facing slopes in the Natal Drakensberg, Marker and Whittington (1971), Hastenrath and Wilkinson (1973) and Boelhouwers and Hall (1990) reporting the occurrence of needle ice in the Lesotho highlands in winter.

5.4.5. Heave

The amount of moisture available for freezing, as well as the relative rates of soil freezing due to temperature conditions, have been found to vary between sites at Tiffindell, as well as from year to year. Thus, the amount of soil heave is not consistent from year to year, or even from day to day.

The relative amount of soil freezing may vary from site to site due to the winter soil temperatures and/or the insulating effects of snow. Mean monthly soil temperatures in the winter of 1995 were considerably higher than those of the winter of 1996, with significant depressions of soil temperature only occurring in mid-July (Figures 5.4a, 5.5a, 5.6a). The rate and intensity of soil freezing was therefore slower in the winter of 1995 than in that of 1996. Low temperatures (<-5°C) were reached within the same soil profile at the end of May 1996 (Figures 5.4b, 5.5.b, 5.6b), encouraging the development of ice lenses and producing an active layer.

High soil moisture values at the beginning of the winter freeze-up in 1996 (Figure 5.11) due to abundant summer and autumn precipitation in the area (approximately 400mm from March to May 1996; Tiffindell precipitation logbook, 1996) resulted, together with the cold temperatures, in a substantial amount of ice lens formation in the area producing considerable frost heave of the soil surface, especially in areas of constant water supply. Ice lenses in excess of 10cm thick developed adjacent to the lodge and other resort buildings, where water runoff was impeded by the buildings (Figure 5.15). The ground was heaved so considerably by these ice bodies that doors were unable to be opened.



<u>Figure 5.14</u>: *Frost heave disruption in the form of nubbins ("cryptogamic soil buds", Perez, 1996) develops a wrinkled appearance.*

<u>Figure 5.15</u>: *Ice lenses up to 10cm thick adjacent to the buildings at an altitude of 2720m.*

In 1996, a section through a gelifluction lobe on the southeast-facing slope at Tiffindell, at an altitude of 2820m revealed the existence of a series of ice lenses (Figure 5.16).

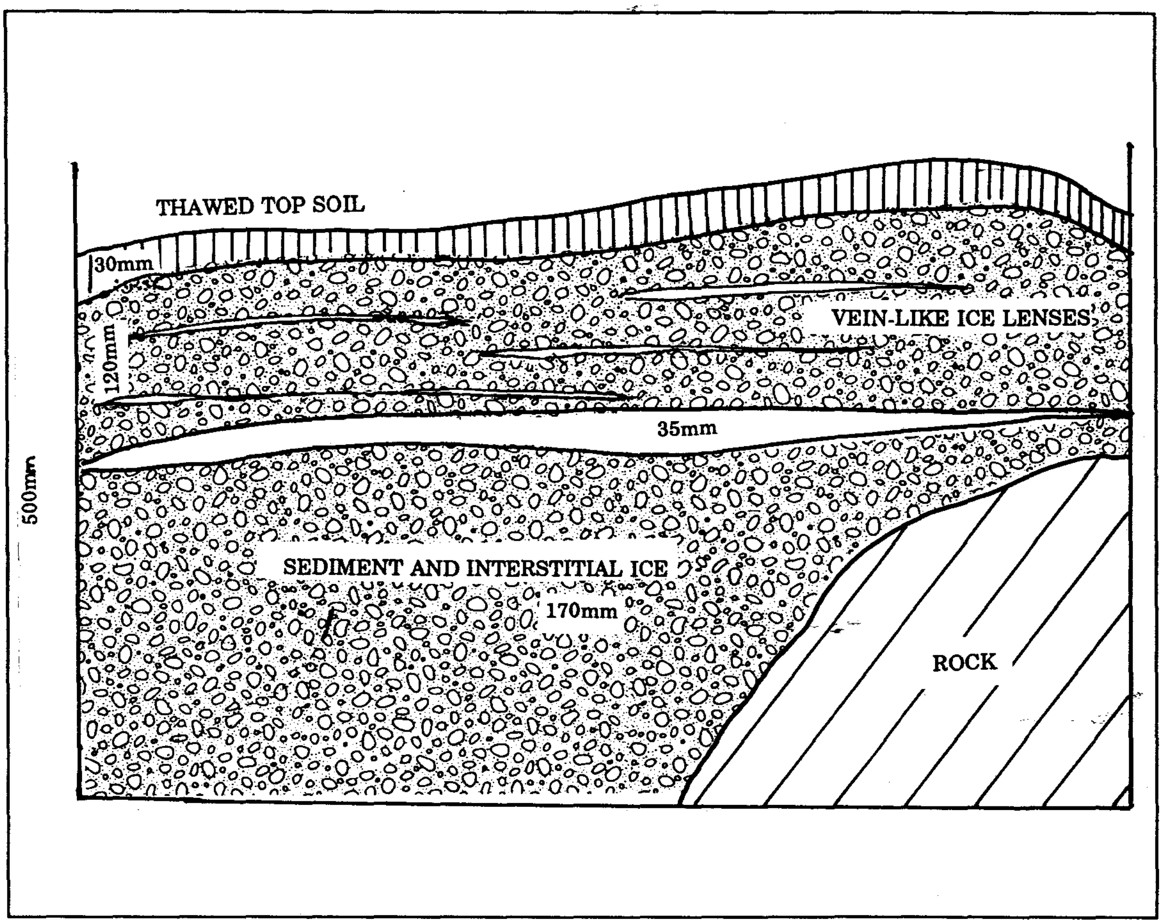


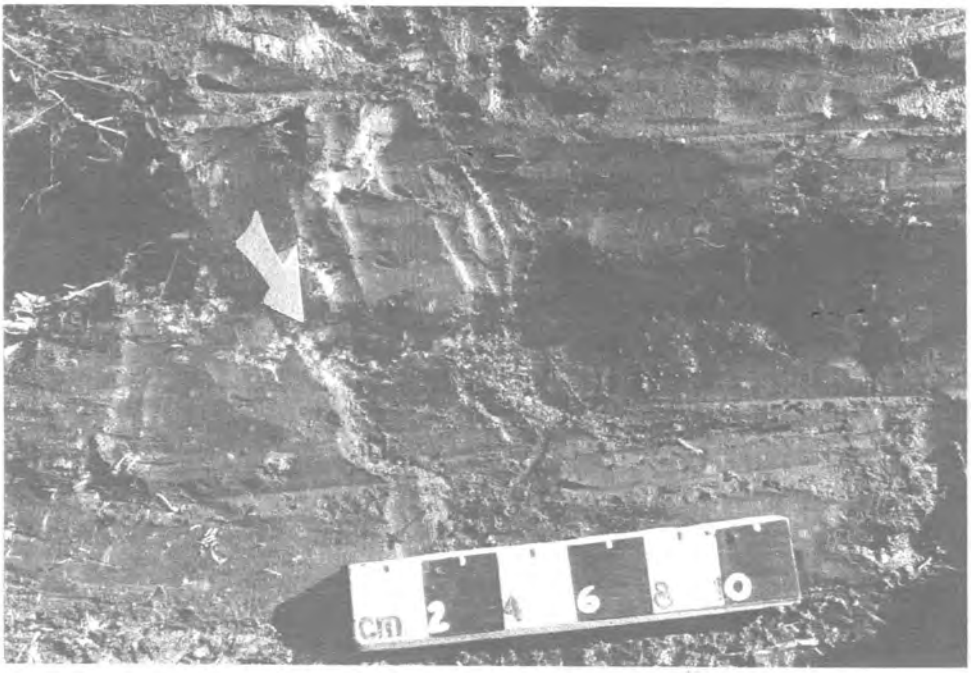
Figure 5.16: Cross-section of a gelifluction lobe (2820m) at Tiffindell displaying the presence of ground ice lenses. The lobe was excavated on 7 June 1996.

The top 30mm of the section was composed of roots and saturated sediment. Below this existed four fine ice segregation vein-like lenses within 120mm of sediment and interstitial ice, the largest lens only measuring 6mm in thickness (Figure 5.17). A larger 35mm thick ice lens was recorded at 150mm depth (Figures 5.18), and below this at least 170mm of interstitial ice and sediment. Below 355mm, no additional substantial ice development was seen within the profile. Therefore, ice segregation was observed to a depth of 185mm, and further interstitial ice to a depth of 355mm. As this section was excavated in early-June 1996 at the onset of winter, it is likely that further ice segregation would have occurred later in that season, due to decreases in temperatures after that date. Therefore, the depth of freezing at Tiffindell in 1996 can be estimated to have reached to at least 200mm below the surface of the ground.

Soil temperatures recorded during the winters of 1995 and 1996 respectively support the estimated depth of freezing given in the previous paragraph (Figures 5.4 to 5.6). As Table 5.6 indicates, the majority of the winter season of 1996 was subjected to freezing temperatures within the soil. Ice days alone were recorded for more than 78% of the days during May to September 1996, and as depth of freeze is largely dependant on the cumulative number of days below 0°C (Ballantyne and Harris, 1994; Solomatin and Zu, 1994), it can be suggested that considerable freezing took place down to at least a depth of 0.2m in the Tiffindell area, from the beginning of June to the end of August.

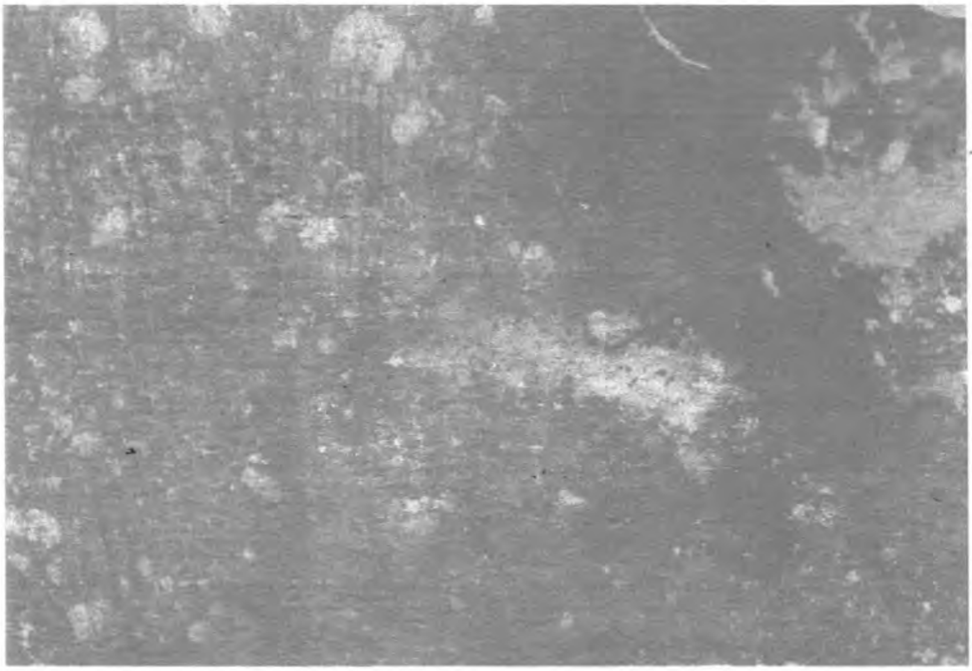
The effects of snow cover appear to be of limited importance at Tiffindell due to the short period in which the ground is normally snow covered. No complete record is available of the number of days with snow cover although, for 1996, a record of the days on which snow fell was kept by the resort's snowmaker. Assuming (as a result of personal observations) that snow lasts for 3-4 days, then the number of days with snow cover can be deduced for 1996. Between May and September 1996, 9 days were recorded on which snowfall occurred. Therefore, the Tiffindell area may have been snow-covered approximately 27 to 36 days out of a possible 153 days (18%-24%). This suggests that the blanketing affect that would be produced by prolonged snow cover is not normally experienced at Tiffindell, and that air temperature fluctuations are of greatest importance in influencing surface and sub-surface temperatures.

In conclusion, the Tiffindell area is characterised by seasonal temperature changes, as well as by short-term freeze-thaw cycles, both of which promote frost creep in moisture-rich soils due to needle ice and frost heave activity.



Dan Liebberman

Figure 5.17: Vein-like ice lens (18mm thick) within the soil profile of a gelifluction lobe. This lens occurred within the top 150mm of the lobe (June 1996; 2820m).



Dan Liebberman

Figure 5.18: Ice lens development within a gelifluction lobe. The lens was 35mm thick and occurred at 170mm depth (June 1996). Note surrounding interstitial ice development.

5.4.6. Geomorphic implications

The temperature and soil moisture data from Tiffindell implies a current freeze-thaw activity of only approximately 3 months per year. In order to establish whether frost heave and needle ice is important at Tiffindell, two short-term experimental sites were established in 1996 at 2721m and 2794m respectively to monitor diurnal needle ice activity and heave at different altitudes (Chapter 4, section 4.5.1.). Both sites were located (Figure 4.1) in areas of low gradient (approximately 3°), bare of vegetation, and characterised by fine-grained soils. They were located on a footpath and on a lobe tread respectively. Air and surface freeze-thaw cycles were recorded at Station B, and at both experimental sites the amount of heave of wooden dowels was measured after each recorded freeze-thaw cycle. The results at the upper site showed that dowels inserted to a depth of less than 3cm exhibited greatest heave with the occurrence of diurnal freeze-thaw cycles (Figure 5.19a), and those inserted more than 3cm showed greatest heave with freeze-thaw cycles lasting more than 48 hours. As those at deeper depths were heaved more out the ground, they also became susceptible to additional heave, and eventually all the dowels were heaved out of the ground completely (Figure 5.19b).

The greatest heave was recorded at the lower site which was the moister of the two sites, being on a footpath downslope from watertanks and adjacent to the ski slope. Although no natural precipitation fell during the experimental period, 'artificial snowfall' was received when the snow-guns were used on the ski slope. Greatest heave was also measured during the periods when artificial snow was produced, but since 'man-made snow' is generated during the colder periods (air temperature <-2°C, I.van Eck, pers.comm.), the recorded frost heaving may have been a result of the combination of prolonged freeze-thaw cycles as well as of the provision of man-made moisture. Whatever the case may have been, the effects of freeze-thaw cycles and of assorted frost heave and needle ice activity are of geomorphic importance at Tiffindell.

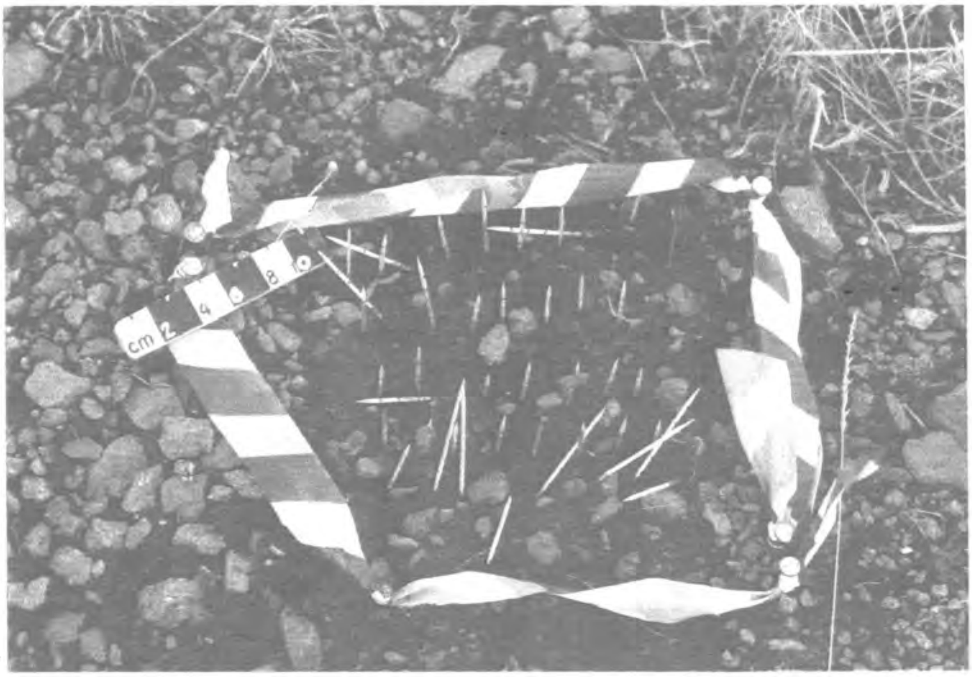


Figure 5.19a: *Diurnal frost heave experiment A after freeze-thaw cycle. Dowels inserted 1-2cm into the soil showed the greatest heave (foreground).*

Figure 5.19b: *After 10 days and several freeze-thaw cycles lasting more than 48 hours, all the dowels have been heaved, and the majority have fallen over. Therefore heave to a depth of greater than 4cm is indicated.*

6.1. MOVEMENT PROCESSES

In mountainous areas in periglacial environments, mass wasting due to solifluction is usually of greater importance than other forms of downslope movement (mass wasting) of superficial soil or rock debris. Solifluction is a distinctive form of periglacial mass wasting in both permafrost and nonpermafrost areas (Harris, 1981a; 1987b; Kirkby, 1995), and is responsible for the slow, downslope flow-like soil movement of saturated unconsolidated sediments, resulting from the combined effects of frost creep and gelifluction mechanisms (French, 1976; Harris, 1981a; Lewkowicz, 1992; Smith, 1992). Solifluction occurs within many alpine areas (Washburn, 1973), including the Tiffindell area, and is partly caused by frost creep.

6.1.1. Frost creep

Frost creep is defined as

> *"...the net downslope displacement that occurs when the soil, during a freeze-thaw cycle, expands normal to its surface and settles in a more nearly vertical direction"*
> (Benedict, 1970:170).

The process differs from other forms of soil creep as it is dependent upon diurnal freeze-thaw processes and upon ice segregation as motive forces. The resultant heaving is proportional to the total thickness of segregated ice layers, or of needle ice, that form in the freezing soil, and is generally directed at right angles to the ground surface. On melting, the heaved particles fall back perpendicularly under the influence of gravity and tend to follow an almost zig-zag shaped path downslope during the freeze-thaw cycle.

Most movement takes place in the upper layers of regolith (Gerrard, 1981), the actual amount of frost creep clearly decreasing with depth. Heaving is favoured by saturated conditions and by slow, deep freezing, and is important only in soils that contain sufficient fine-textured material to permit water to move upward to the base of the frozen layer (that is, in soils that are *frost susceptible*; Benedict, 1970; French, 1976). Movement rates of up to 0.3 $cmyr^{-1}$ have been recorded on moderate (± 10-15°) slopes (Goudie, 1993).

Frost creep is typically divided into two components, which act in opposite directions at different times during the frost heave cycle:

(1) *potential frost creep* is the downslope component of movement caused by expansion of unconsolidated sediments during freezing. It varies with the angle of slope and the magnitude of heaving, and is influenced by all of the factors that govern ice segregation in the freezing overburden;

(2) *retrograde movement* is the upslope component of movement caused by non-vertical settling of the overburden during thaw.

Both components of frost creep are typically active on slopes, although downslope movement is generally more prominent. Indications of this are revealed in the measurement of sediment movement, yet greater upslope (retrograde) movement has been recorded at several sites (for example, Verster and van Rooyen 1988).

6.1.2. Gelifluction

Solifluction is the "slow downslope flow of saturated unfrozen earth materials" (Harris *et al.*, 1988), and following the original definition by Andersson (1906), need not require either a frozen substrate or even freezing and thawing. Gelifluction is "solifluction associated with frozen ground" (Washburn, 1967 cited in Benedict, 1976). The movement of unconsolidated sediments is due to an upward expansion associated with freezing of a surface, or near surface layer; a flow of this material in a downslope direction with the spring thaw; and a final contraction associated with drainage of the thawed ice layers (Kirkby, 1995). Andersson's (1906) original definition of solifluction ('soil flow') can be applied to soil movement in warm as well as cold climates. Subsequently, a form of solifluction that occurs under cold climatic conditions, was named *gelifluction* and defined as:

> *"the slow flowing from higher to lower ground of water-saturated waste above a seasonally or perennially frozen substrate".*
> (Lewkowicz, 1988).

The source of the moisture required during the autumn freeze up for ice lens development is often derived from a number of sources including snowbanks, ground ice lenses or precipitation. The importance of ice lensing, and therefore the amount of moisture available for freezing, is illustrated by the often restricted occurrence of gelifluction, for example within the Niwot Ridge, Colorado (Benedict, 1970). Benedict (1970) has shown that although much of the alpine region is saturated during the spring thaw, significant gelifluction occurs only in areas where the water table remains high enough during the autumn freeze to permit thick ice-lens development. At the onset of warmer temperatures, the thawing ice lenses not only supply water, but create discontinuities that reduce the cohesive strength of the soil (Williams, 1959): these discontinuities are probably required in order for significant gelifluction to occur (Benedict, 1970).

The prominence of gelifluction in cold climates can be partly attributed to:

(1) the role of an impermeable substratum, such as a layer of frozen ground, which limits the downward movement of water through the soil and thereby promotes saturation of the soil; and

(2) the role of the thawing segregated ice lenses in providing excess moisture which reduces internal friction and cohesion in the soil.

Under these conditions gelifluction is possible on gradients as low as 1° (Harris, 1981a). Significant gelifluction probably only occurs at moisture values approximating or exceeding the Atterberg liquid limit, that is, at values at which unconsolidated sediments have little if any shear strength, as will be discussed later. The importance of moisture is also demonstrated by the fact that, within limits, its influence is more significant than the binding effect of vegetation or the effect of gradient. Gelifluction in high and medium latitudes is an annual process, associated with seasonal climatic conditions. In low latitudes it may be a diurnal process at high altitudes, as on Mount Kenya (Mahaney, 1988). Gelifluction varies from year to year in its speed, and therefore in the amount of sediment movement. Recorded rates of movement are 10-100 times faster than those for soil creep (Selby, 1982), and are often in the order of 0.5-10 $cmyr^{-1}$ or more (French, 1988).

6.1.3. The interrelationship of the processes

As both frost heave and gelifluction processes are influenced by ice segregation in periglacial environments, they commonly occur simultaneously and it is difficult to separate the two processes in field studies in order to evaluate the relative importance of each as a mass wasting agent (Benedict, 1970; Harris, 1972, 1981a; Smith, 1992). The relative importance of gelifluction and of frost creep varies from site to site and from year to year, but it appears that gelifluction is more sensitive to soil moisture conditions during thaw, so that frost creep tends to be more important in drier sites. Rates of solifluction show considerable within-site variation, soil moisture conditions during thaw being the most important controlling factor (Harris, 1981a). Measurements from a turf-banked lobe in the Colorado Rocky Mountains (Benedict, 1970, 1976) illustrate that gelifluction accounted for more than 80% of net movement along the saturated axis of the lobe; and that in drier localities frost creep was the major component of movement, although often being equalled or exceeded by retrograde displacement.

Many attempts (Table 17 in Harris, 1981a) have been made to gauge the rate and character of present-day solifluction movements in periglacial environments by direct field measurement. Unfortunately most attempts have been conducted over short periods (Benedict, 1970; Harris, 1973; Price, 1991; Smith, 1992) and consequently the assessment of solifluction as a mass wasting agent is rarely possible.

Measurements of downslope movement are generally based upon annual resurveys of surface or subsurface markers, and do not differentiate between frost creep, creep due to wetting and drying, and gelifluction. They indicate, instead, the combined effect of these processes. Surface movement has been monitored by marked stones and pegs placed in a straight line on the surface or protruding through the surface (e.g. Pissart, 1964; Washburn, 1967 cited in French, 1976; Egginton and French, 1985 cited in Lewkowicz, 1992). Variations of movement with depth have been studied by the insertion of flexible plastic tubing which subsequently deforms (*e.g.* Harris, 1972; Price, 1991; Smith, 1992), by probes attached to strain gauges (Williams, 1962) and by utilising pillars (Smith, 1987, cited in Lewkowicz, 1992). Detailed surveying of marker positions relative to a stable benchmark during subsequent excavation allows a velocity profile of average movements to be generated. Consideration needs to be given to:

i) the basic installation, as it may prove to be difficult not to disturb the study area too much;

ii) the destructive removal of the markers, which results in no further monitoring being possible at the sites; and

iii) the lack of accuracy of surveys of short duration (Lewkowicz, 1992).

The values recorded for downslope movement are not always comparable at all sites, due to the use of different measuring techniques, site moisture availability and relative slope angle (Benedict, 1976). Despite the measurement differences, Benedict (1976) reported maximum rates and depths of movement from 102 experimental sites (summarised in Figure 2, Benedict, 1976) which revealed that the median maximum velocity for solifluction lobes and terraces is 3.0cm per year, and largely tends to involve movement of only the upper 50 cm of the soil, a decrease in velocity occurring with depth. Viewed in section, (Figure 6.1) a typical velocity profile shows greatest movement rates near the surface, diminishing with depth, producing a concave downslope profile.

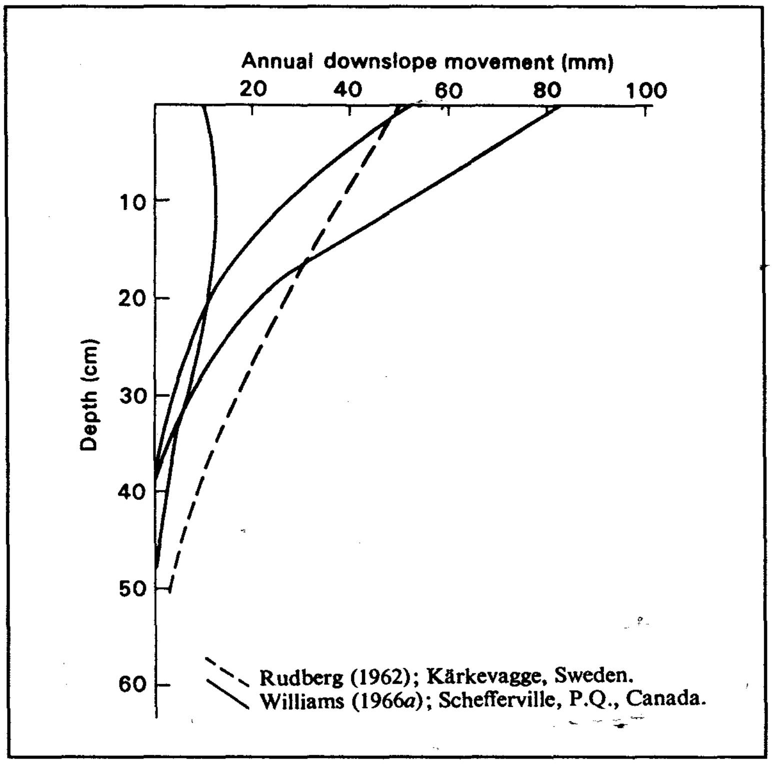


Figure 6.1: Velocity profiles for soil movement during solifluction.

(from Carson and Kirkby, 1972)

Research into rates of mass movement in alpine areas has concentrated on slopes where mass-wasting is clearly active, such as those exhibiting well-developed solifluction lobes and terraces. Considering the inherent variability of rates of mass movement both spatially and temporarily, much more data is required from areas where such clearly visible topographic features may not be present, but where mass-wasting is undoubtedly an important agent of sediment movement on slopes (Harris, 1981a).

Solifluction is best displayed on slopes of approximately 10° to 20° where distinct tongues, lobes and lobate terraces may form (Price, 1991). On slopes in excess of around 25° solifluction may be replaced by more rapid mass movement, such as mudflows and slides. The location of solifluction lobes are not controlled so much by gradient as by concentrations of soil moisture, as already discussed. This was well illustrated by Washburn (1967, cited in Washburn, 1973) at Mesters Vig, Greenland, who recorded average rates of solifluction movement on a consistent 10°-14° gradient which ranged from a minimum of 0.6cmyr^{-1} in relatively dry sections to a maximum of 6.0 cmyr^{-1} in wet areas.

Solifluction is also strongly affected by grain size. The high porosity and permeability of gravel and coarse sand promote good drainage and do not favour saturated flow. Fines tend to remain wet longer than coarse grain sizes, and silt is particularly subject to flow because it lacks the cohesion of clays and slakes readily. The Atterberg liquid limit is also lower in silt than in clay, so that less moisture is required for flow. Silty soils tend to predominate over clays in cold climate areas as mechanical weathering tends to be more important than chemical weathering. This may be one of the reasons for the prevalence of solifluction in such climates.

Vegetation has been regarded as having an impeding effect on rates of solifluction movement although, by restricting surface movement, it may cause greater subsurface movement (French, 1976).

Solifluction deposits tend to be diamictons but some deposits show crude stratification parallel to the slope (Washburn, 1973; Harris, 1981a). Diamictons are considered as poorly sorted clast-sand-mud sediments regardless of the depositional environment (Eyles *et al.*, 1983). Solifluction material is commonly angular and the deposits may show a sorting from fine grained at the bottom of the section, to coarse grained at the top of the section. The clasts within the deposit tend to lie with their long axis oriented in the direction of movement (Washburn, 1973), thus offering the least resistance to movement. The long axis therefore indicates the direction from which the material has come (Lewis, 1966).

6.1.4. The influence of permafrost

Permafrost prevents free drainage of water through unconsolidated sediments and so maintains saturation within the profile until freezing begins. The occurrence of permafrost therefore encourages sediment movement downslope. In areas of highly permeable debris or when there is an inadequate source of water, permafrost may be required in order for significant frost creep and gelifluction to occur. Where moisture is not a limiting factor, permafrost is not required by either movement process, and may even inhibit movement as it truncates the depth of influence of the frost heave cycle. Benedict (1976, figure 4) shows that heaving is greatest where freezing is deep (±2m), and where it continues throughout the winter. In environments characterised by shallow permafrost, frost creep is discouraged due to the reduction of the unconsolidated sediment layer available for ice segregation.

6.2. CLASSIFICATION AND DESCRIPTION OF ASSOCIATED LANDFORMS

Frost creep and gelifluction produce a variety of lobate and terrace-like landforms which can most conveniently be classified on the basis of their sorting and topographic expression (Benedict, 1970), and are shown in Table 6.1.

	No surface expression	Lobate	Terrace-like
Non-sorted	Non-sorted sheet	Turf-banked lobe	Turf-banked terrace
Sorted	Blockfield	Stone lobe or stone-banked lobe	Stone-banked terrace

Table 6.1. Classification of landforms produced by downslope soil movement.

Moving regolith lacking topographic expression is the least studied, but possibly the most widespread form of mass movement, particularly in the high Arctic where the absence of vegetation enables solifluction to act uniformly (French, 1976). In sub-Arctic and alpine zones vegetation cover and soil drainage are much more variable than in the high Arctic, favouring the development of more localised lobes and terraces. The lobate forms are generally smaller than terraces, terrace-like features often being formed by the coalescing of several lobes (French, 1976). This study pays particular attention to lobate features, which are typical of the Tiffindell area.

6.2.1. Turf-banked lobes

These widely distributed forms can be defined as *"lobate accumulations of moving soil that lack conspicuous sorting"* (Benedict, 1970:170). The term 'turf-banked lobes' as used by Galloway (1961) encompasses other terms such as 'soil lobes', 'solifluction/gelifluction lobes' (Washburn, 1973), and 'non-sorted lobes' (Ballantyne and Harris, 1994). The lobes are best developed in relatively snow-free areas, where moisture, instead of being uniformly distributed across the slope, is confined to linear drainageways (Benedict, 1970). Thus the lobes typically form one below the other wherever moisture is channelled along definite drainage routes.

Turf-banked lobes are composed of a fairly homogenous mass of debris which is predominantly fine grained and therefore frost susceptible. The material shows little or no evidence of lateral or vertical sorting. Such features can support a complete vegetation cover, and thus can be called vegetation-covered lobes (Ballantyne and Harris, 1994), although the vegetation is generally thickest on the fronts (risers). The treads (central areas) of the lobe surfaces are less well vegetated and may even be bare of vegetation. At high altitudes large stones and boulders may be concentrated at the lobe front (Harris, 1981a), typically orientated with their long axes parallel to the direction of flow, and slightly imbricate to the ground surface.

Lobes develop on slopes of between 4° and 25°, yet the tread gradients are generally only 1.5°-7° (Ballantyne and Harris, 1994). Several authors describe lobes occurring on slopes below late lying snow patches and other ground moisture sources.

Within small active forms, greatest downslope movement generally occurs in the centre of the lobe, but there may be considerable variation (Washburn, 1969; Benedict, 1970). Movement may be more rapid in one area because of the varying texture or water content of the regolith itself, or the material may be affected by external factors such as gradient or the damming effect of rocks (Price, 1974). The level of activity within lobes is usually dependant on the degree of soil moisture. Less active areas are usually quite rocky and/or dry. The fronts of a few particularly active lobes tend to be oversteepened, bulge outward over the lower sediments and curl over (Price, 1973). Thus they commonly overlie buried organic material which provides considerable potential for both process studies and palaeoenvironmental reconstruction, particularly by the ^{14}C dating of such organic material (Matthews *et al.*, 1986).

Where the downslope rates of soil movement decrease, turf-banked terraces are common. These large, bench-like accumulations, are essentially the consolidation of several lobes and tend to form arcuate fronts trending parallel to the contours (Benedict, 1976).

6.2.2. Stone lobes or stone-banked lobes

Stone lobes, or stone-banked lobes, occur on slopes of similar gradient to those on which turf-banked lobes develop. The major contrast between the two is the relative abundance of coarser material and the lack of vegetation in the stone lobes compared with the turf-banked forms. In alpine areas stone-banked lobes and terraces tend to occupy a higher altitudinal zone than their turf-banked counterparts (Harris, 1981a). Stone lobes will be discussed comprehensively in Chapter 9.

6.3. DOWNSLOPE SEDIMENT MOVEMENT AND GELIFLUCTION AT TIFFINDELL

6.3.1. Terracettes

Large portions of both the south and the south-east facing slopes of the Tiffindell valley are characterised by scattered terracette-like features which form steps in the unconsolidated debris that overlies the basalt bedrock. These terracette fronts are often crescentic in shape and are aligned essentially across the slope. Terracettes exist between 2750m and 2870m on the *c.*23° south-facing slope, and between 2790m and 2880m on the southeast-facing slope at a slightly steeper gradient of *c.*26°. The height of the risers varies between 25cm and 55cm, while the average tread width is 70cm (Appendix III). The tread angle is between 4° and 7° and they commonly support a grassy cover which ends at the riser of the next terracette on the upslope side. The dense root network of the vegetation binds the soil to form a 'scarp' along the length of the riser. The Tiffindell terracettes are generally short in length (mean length 1.6m), compared to those of the Natal Drakensberg and Sani Pass areas (mean length 4m, but extend up to 10m; Killick, 1963; Verster *et al.*, 1985; Verster and van Rooyen, 1988; Watson, 1988; Boelhouwers and Hall, 1990), and do not exceed 2.5m (Appendix III; Figure 6.2).

The terracette sediments at Tiffindell appear to be presently subject to erosion by frost action processes. Clear evidence of needle ice activity was observed during the winters of 1995 and 1996 at the unvegetated sections of the riser and therefore may be responsible for the backward erosion of the riser front, as well as causing creep of the bare soil occurring on the treads. The retreat of the risers due to frost action processes has also been noted in the Natal Drakensberg (Killick, 1963; Harper, 1969; Watson, 1988; Boelhouwers and Hall; 1990; Boelhouwers, 1991a). Surface wash of these bare sections of tread was observed, particularly during periods of high rainfall (pers.obs.).

Watson (1988) has reported the existence of terracettes in the Natal Drakensberg that were analogous in appearance to those of Tiffindell. She suggests that identification of a specific process of terracette origination is difficult, but that the Natal features apparently originated as a result of soil creep processes, and are not considered to be due to frost creep or gelifluction. No evidence has been discovered at Tiffindell to account for the origins of the terracettes, although needle ice appears to be important for their existence at Tiffindell.

Animal trampling may have contributed to terracette formation and maintenance in the past, but this area has not been used for grazing since at least 1991 (I.van Eck, pers.comm.), so trampling cannot be of importance to the maintenance of the present terracettes.

6.3.2. Turf-banked lobes

The individual slope environments of the Tiffindell valley are quite different due to differences in exposure to solar radiation and meltwater from snow. Therefore, it is on the more moist south and southeast-facing slopes, within the same area as terracettes occur, that there is the distinct development of larger, isolated lobe-like features, which can be classified as turf-banked lobes. These, however, generally occur on the more moderate slopes (*c.*20°), and instead of being uniformly distributed across the slope such as terracettes, tend to be confined to linear drainageways. The lobes are therefore often formed one below the other wherever moisture is channelled along definite drainage routes.

The lobes have been identified in a 'mid-zone', on the south-facing slope predominantly between 15° and 26° at an altitude of approximately 2765m to 2855m, and on the southeast-facing slope between 2800m and 2880m with slope angles between 19° and 28° (Figure 6.3). None were observed to occur on the north-facing slope.

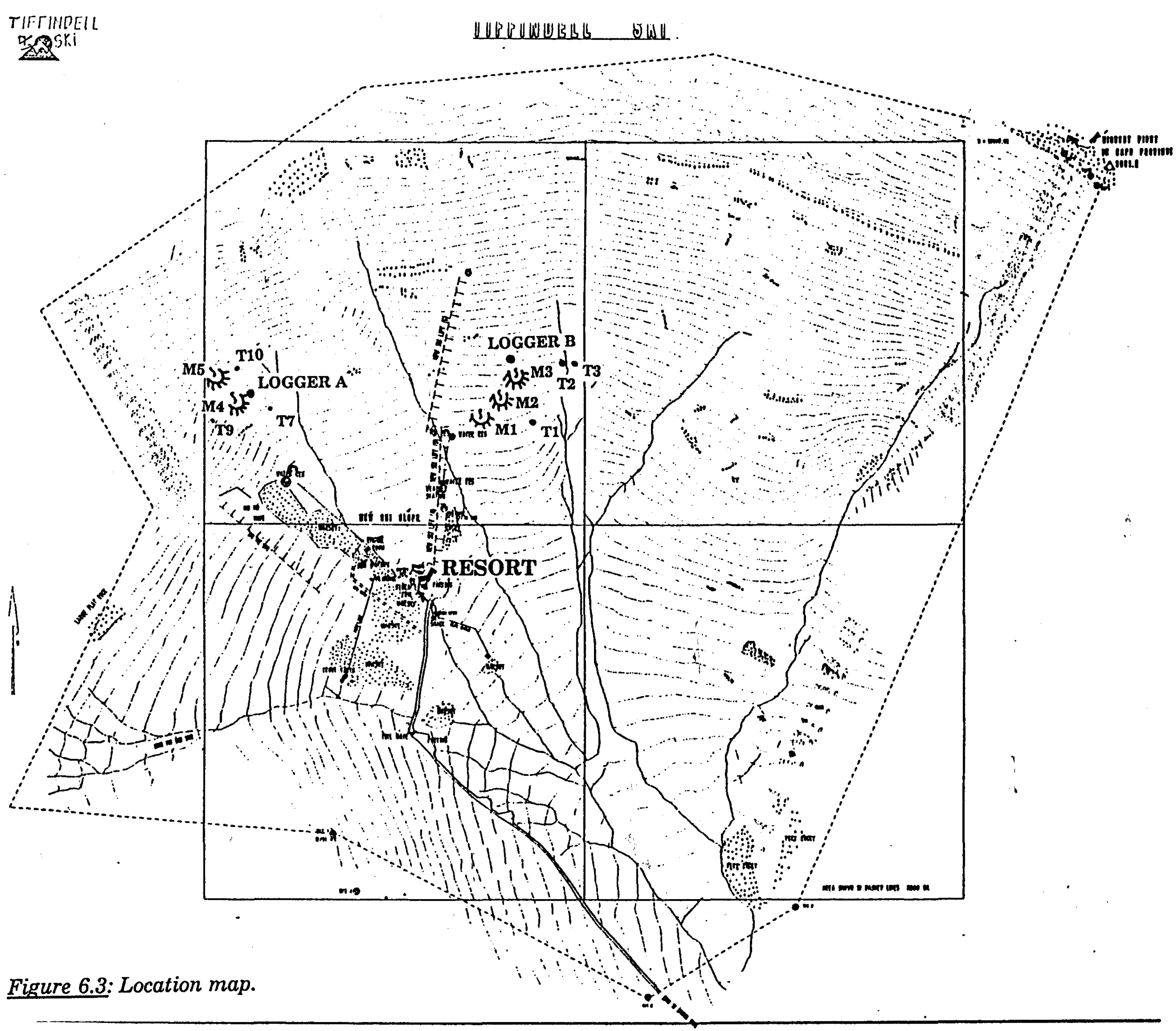


Figure 6.3: Location map.

They differ in form from the terracettes as their fronts are narrow and crescentic in outline rather than broad and straight. They are generally small-scale, and are synonymous with those features called 'turf-banked steps' reported elsewhere in southern Africa (Marker and Whittington, 1971; Hastenrath and Wilkinson, 1973; Dardis and Granger, 1986; Boelhouwers and Hall, 1990; Boelhouwers 1991a; Boelhouwers, 1995; Lewis, 1996). In contrast to the terracettes, they are regarded as positive evidence for gelifluial activity (Lewis, 1988a,b), and were therefore selected for a study of downslope sediment movement rates at Tiffindell due to gelifluction processes (Figure 6.4).

The gelifluction lobes which are similar to those described from many other areas of the world (*e.g.* Lewis and Lass, 1965; Quinn, 1975; Mark, 1990; Ballantyne and Harris, 1994) are typically formed in fine sandy soil of thickness ranging from 0.3m at the summit to 1m or more at the base of the ski slope (2720m) and often incorporate large rocks. The treads vary between 0.9m and 4.2m in length (n=30) and 0.7m and 3.4m in width (Appendix III). In profile, the lobes are step-like, with a spacing between the risers varying between 1m and 5m. The treads are gently inclined downslope at angles of 2°-14°.

The near-vertical lobe fronts vary in height from 0.41m to 1.03m (average 0.75m; ±0.194 s.d.) and are characterised by a vegetated frontal riser. Individual plants grow on both the riser and tread, and include the *Helichrysum* species, as well as abundant grasses such as *Pentaschistis curvifolia*, *Themeda triandra* and *Harpechloa falx*. Vegetation is, however, predominantly concentrated on the riser (average 80% cover) which, in providing a sharp micro-environmental gradient, typically support those species requiring protection or abundant moisture. The treads, by comparison, may be almost devoid of vegetation (Figure 6.4), often poorly drained, and thus tend to promote frost heave activity (Chapter 5) and as a result the surfaces often show signs of amorphous sorting. The heaving of the bare ground due to frost action makes colonisation of the treads by plants difficult, as well as indicates the development of ground ice, thus supporting the hypothesis that these may be of gelifluction origin.

Figure 6.2: Terracettes at altitudes around 2700m at Tiffindell Ski Resort, the backpack indicates the height of the terracette fronts.

Figure 6.4: The front of a turf-banked lobe at 2825m. Note the well vegetated front, and the bare tread of the lobe in the foreground. The lip is curved over, and indicates that the sediment is actually moving downslope.

Needle ice growth and shallow frost heave (i.e. freeze-thaw activity), and thus surficial frost creep is active in at least the upper 0.05m of the regolith (Chapter 5), and thus contributes to downslope sediment movement. The vegetation at the lobe fronts makes the surface more cohesive, leading to a certain amount of stability in the terraces and as a result they are generally stable underfoot.

In the immediate lee of the riser of each lobe lies the zone with maximum protection. The wind is greatly reduced and snow tends to accumulate and persist in these micro-environments for weeks after snowfall. The slow melt of the snow in these environments effectively acts as a supply of soil moisture.

The occurrence of turf-banked gelifluction lobes suggests that sediment movement is active on the Tiffindell slopes. Below 2765m and 2800m on the south and southeast-facing slopes respectively, only terracettes occur, and no features appearing to be of gelifluction origin were observed.

6.3.3. Rates and processes of downslope movement

Studies of downslope sediment movement outside of South Africa by those such as Benedict (1970), Harris (1972) and Smith (1992) have attempted to take environmental conditions into account and to assess their affects on the measured rates of movement. The writer undertook similar research at Tiffindell, South Africa, in 1995 and 1996.

Two different sets of experimental equipment were installed at Tiffindell on both slopes to provide information on the surface and subsurface soil movement occurring within the study area during the period of temperature recordings. Surface movement of five lobes (M1 to M5; Table 6.2; Figure 6.3) was recorded by measuring the relative displacement of metal markers aligned across the treads of lobes, just behind the riser (Chapter 4; section 4.5.2.2.). Subsurface movement rates were assessed using flexible tubes inserted vertically to a depth of 1m within the soil (Chapter 4; section 4.5.2.3.). Eleven stations were originally set-up in June 1995, yet only six were excavated in November 1996 (Stations T1, T2, T3, T7, T9, T10; Table 6.2; Figure 6.3). One station was disturbed and the tube removed, another two were 'lost' due to their surface markers being removed, and the other two have been left *in situ*. These, together with the surface markers (which have also being left *in situ*), potentially form part of a long-term study within the Tiffindell area. Table 6.2 provides a summary of the physical characteristics of the eleven lobes investigated for surface and sub-surface movement at Tiffindell.

6.3.3.1 Surface soil movement

Measurement of the displacement of the surface markers on lobes M1 to M5 in November 1996 revealed obvious downslope movement over the 18 month period that they had been in place (Figure 6.5). This movement gave the net movement during the study period. Of the original 45 markers, only 4 showed no visible surface movement and 6 were disturbed by burrowing animals or resort visitors. The displacements of the remaining 78% of the markers are presented in Figures 6.6 to 6.10.

The mean values for the movement of the markers at the five sites are calculated in Table 6.3.

Lobe #	Altitude	Mean surface movement (mm)	Standard Deviation
M1	2770m	44	108
M2	2777m	131	47
M3	2786m	256	158
M4	2825m	19	33
M5	2835m	7	22
OVERALL		217	142

Table 6.3: Summary of the measured rates of surface movement at five localities.

Lobe #	Altitude (m)	Gradient (degrees)	Tread angle (°)	Riser ht (m)	Tread length (m)	Tread width (m)	Lobe characteristics
Surface movement experiemnt sites							
M1	2770	20	2	0.94	3.60	3.4	Bedrock boulders on either side of the lobe front, and may 'dam-up' soil water in the area; riser vegetated; tread generally bare; several markers dislodged due to rodents' burrows.
M2	2777	23	4	0.62	1.55	1.20	Well vegetated riser and tread; marker's central rod cut to 5cm in length.
M3	2786	22	4	0.68	2.20	1.70	Vegetated riser; tread bare and stony.
M4	2825	25	7	0.72	3.40	3.10	At logger A, which measures climatic and soil conditions; vegetated riser; generally bare tread.
M5	2835	21	5	0.56	1.87	1.05	Well vegetated riser; little visible sign of frost heave.
Subsurface movement experiment sites							
T1	2779	23	12	0.74	2.90	2.10	Well vegetated riser; tread generally bare.
T2	2789	22	6	0.84	2.30	1.50	Vegetated riser and partly vegetated tread.
T3	2790	22	8	0.89	2.35	1.60	Vegetated riser; tread generally bare; visible bedrock at upper end of tread.
T7	2813	24	9	0.68	1.30	1.20	Vegetated riser; and partly vegetated tread; boulder observed within the profile during excavation; may dam water and influence the regional soil moisture.
T9	2826	25	7	0.51	0.92	0.79	Small lobe with well vegetated riser and tread.
T10	2834	21	14	0.62	1.25	0.96	Vegetated riser with partly vegetated tread.

Table *6.2: Summary of the physical characteristics of the eleven lobes investigated for surface and subsurface movement at Tiffindell from 1995 to 1996*

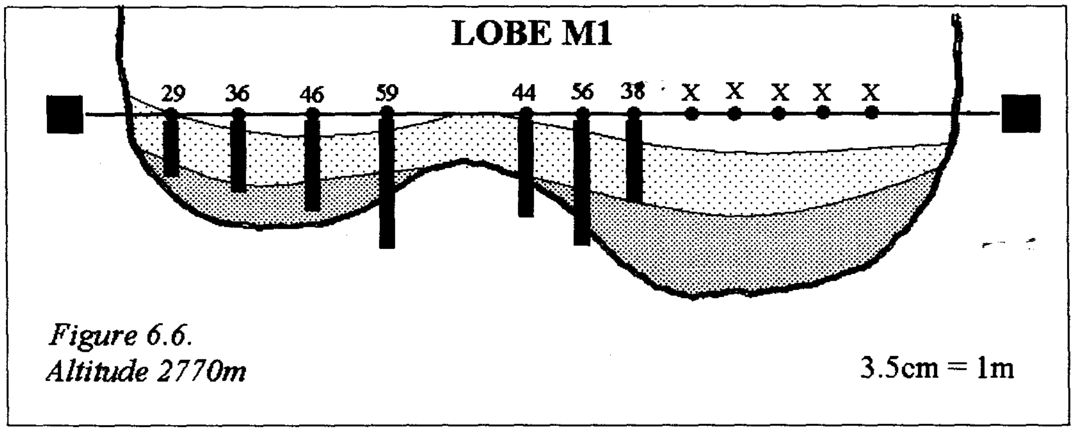


Figure 6.6.
Altitude 2770m

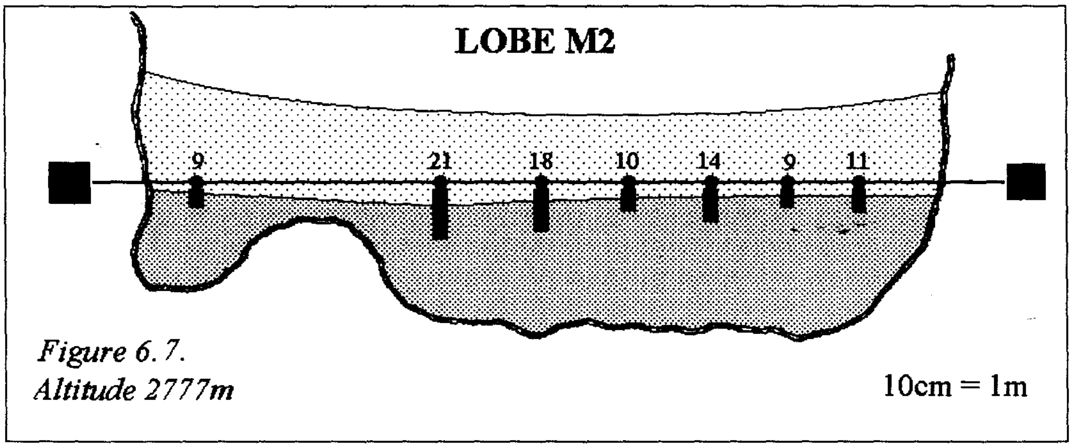


Figure 6.7.
Altitude 2777m

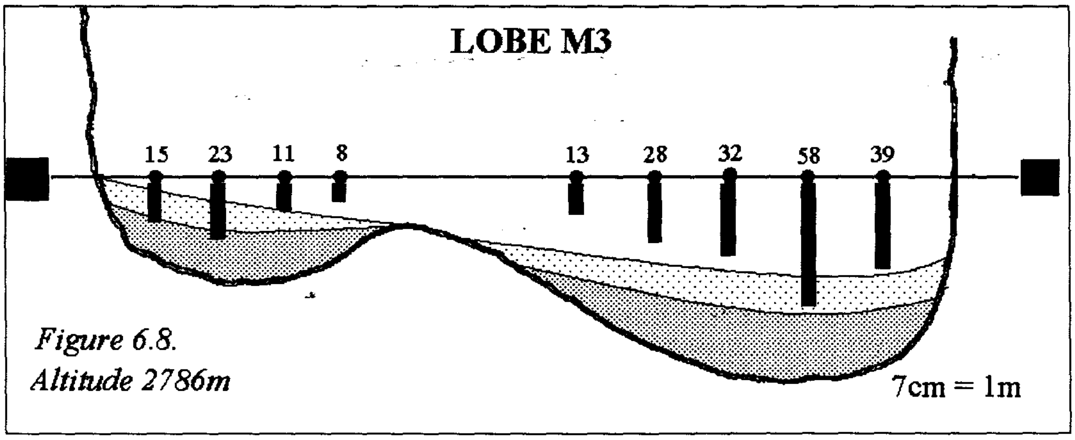


Figure 6.8.
Altitude 2786m

Permanent reference point **Dense vegetation** **Sparse vegetation** **No vegetation**

Figures 6.4, 6.5, 6.6: Rates of surficial movement at gelifluction lobes M1, M2 and M3 during the period May 1995 to November 1996. Displacement of markers (illustrated by bars) is measured in mm and does not indicate the direction of movement.
Bars not to scale of the lobes. X represents disturbed markers.

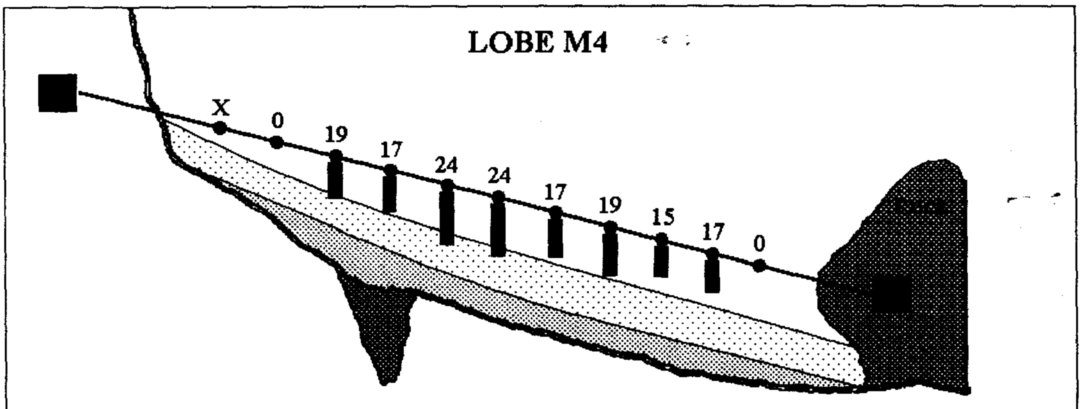


Figure 6.9.
Altitude 2825m

4cm = 1m

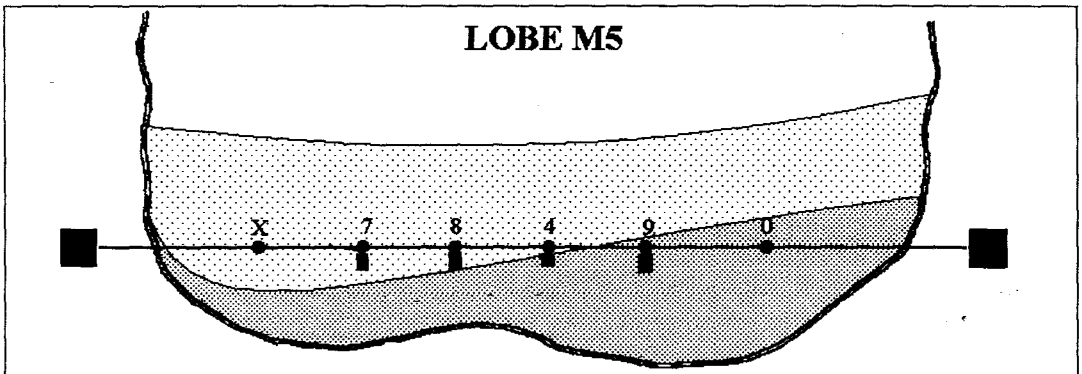


Figure 6. 10.
Altitude 2835m

10cm = 1m

Permanent reference points **Dense vegetation** **Sparse vegetation** **No vegetation**

Figures6.9,6.10: Rates of surficial movement at gelifluction lobes M4 and M5 during the period May 1995 to November 1996. Displacement of markers (illustrated by bars) is measured in mm and does not indicate direction of movement.
Bars not to scale of the lobes. X represents disturbed markers.

Those lobes with similar characteristics (Table 6.2) have generally shown similar amounts of movement, for example, lobes M2 and M5 both display well vegetated risers and treads, as well as the lowest amounts of recorded movement. The others with generally bare treads show a larger amount of movement. The lobe at the lowest altitude (M1) and least slope angle has the greatest mean recorded surface movement (Figure 6.6). The overall mean value (217mm ±142 s.d.) may be used to calculate an annual mean surface movement of a lobe tread over the 18 month period. Within 200mm of the lobe front, a displacement of approximately 109mm yr^{-1} can be anticipated.

The calculated standard deviations for the individual sites, as well as overall, indicate relatively large variations. This can be explained by the relative distance of the markers to the lobe axis and edges. Those in the vicinity of the lobe axis characteristically show greater variability in movement than those on the edges. This was also reported by Mark (1994).

At Tiffindell, those markers in axial positions on the lobe displayed significant movements of up to 59mm over the study period (M1, Figure 6.6). Those at the lobe edges generally showed less movement, and often recorded no movement (M4, Figure 6.9; M5, Figure 6.10).

The displacement of the markers was measured in November 1995, and again in November 1996 (Appendix IV). In 1996, it was observed that several of those markers inserted in areas bare of vegetation were no longer flush with the ground surface and had been heaved to some extent, although not greater than 19mm. This implies that the colder conditions experienced at Tiffindell in winter 1996 compared to 1995 (Chapter 5), would have resulted in greater amounts of frost activity in the regolith, greater frost heave and the development of thicker ice lenses, and thus induced vertical displacement of the plates. A comparison of the relative rates of horizontal movement for each year also indicates that the net movement in 1996 was higher than in 1995. This may be a result of the higher moisture content in the ensuing spring thaw of 1996 due to thicker ice lens development during winter, as well as to higher precipitation (Tiffindell precipitation log book, 1996). Although seasonality of movement was not recorded on the Tiffindell lobes, it is likely to be confined to the period of maximum thaw in spring when soils are supersaturated and readily puddled.

As the markers were initially inserted to 10cm depth in the soil, and had not been notably lifted by frost heave in 1995, and several experienced only minimal heave in 1996, yet had moved along the surface, it follows that the soil layer at 10cm depth was mobile and not just the uppermost few centimetres. This observation is significant for the discussion of the mechanism of mass movement.

The measurements of surface movement (Figure 6.5) clearly indicate that the turf-banked lobes were active features over the studied periods, their relative movement resulting from the combined effects of frost creep and gelifluction. Such downslope movement is confirmed by the concave profile of many lobe fronts, as well as by the existence of overhung lips on some of the smaller lobe fronts (Figure 6.4). These often experience local collapse, as are also described from New Zealand by Mark and Bliss (1970) and Mark (1994). Despite these obvious signs of recent movement in some of the smaller lobes, it is assumed that these lobes are relatively slow moving, but with the uppermost soil experiencing more rapid movement than the lower layers. Therefore, an attempt was made to measure the sub-surface rate and pattern of movement within the Tiffindell turf-banked lobes.

6.3.3.2. Subsurface soil movement

Soil movements recorded at each of the 6 excavated sites at Tiffindell are summarized in Table 6.4 overleaf.

Tube #	Depth of tube insertion (mm)	Max.depth of movement (mm)	Maximum recorded movement (mm)*	Approx.annual movement (mm yr^{-1})
T1	950	70	11	± 55
T2	930	no visible movement recorded at this site		
T3	940	40	4	± 2
T7	960	130	4	± 2
T9	970	40	3	± 15
T10	970	no visible movement recorded at this site		

* = Recorded movement at the ground surface

Table 6.4: Summary of short-term sub-surface sediment movement at Tiffindell from May 1995 to November 1996.

Subsurface movement rates show a significant range over the 18 month period. For example, 11mm total displacement was observed within the study period (1995-1996) at site T1, and no movement was recorded during the same interval at sites T2 and T10. In the majority of the cases (sites T3, T7; T9) the excavations revealed surface movements closer to an average of 25mm yr^{-1} (±0.4mm yr^{-1} s.d.) for the interval between May 1995 and November 1996.

Figure 6.11 shows the extent of subsurface movements at the 4 sites where movement was recorded. The velocity profiles show soil movement declining with depth. All the displacements were shallow and were mainly restricted to the upper 40mm to 70mm of the sediment (Table 6.4). Deeper movements were visible at site T7 (see lobe T7, Figure 6.11; 6.12a,b), but die out at between 70mm and 130mm below the surface. Excavation of the tubes revealed that those that had moved displayed a concave downslope shape (see lobe T1, Figure 6.11; Figure 6.12). This type of velocity profile is associated with "either creep or compressive shearing in soliflucting sediments" (Benedict, 1970; Smith, 1992). In general, movements are significantly smaller and shallower than those recorded over a short period of time in other parts of the world (*e.g.* Harris, 1972).

Figure 6.5: Downslope displacement of metal markers of lobe M2 in November 1996. The markers have moved out of line, and are also no longer flush with the ground surface due to frost heave.

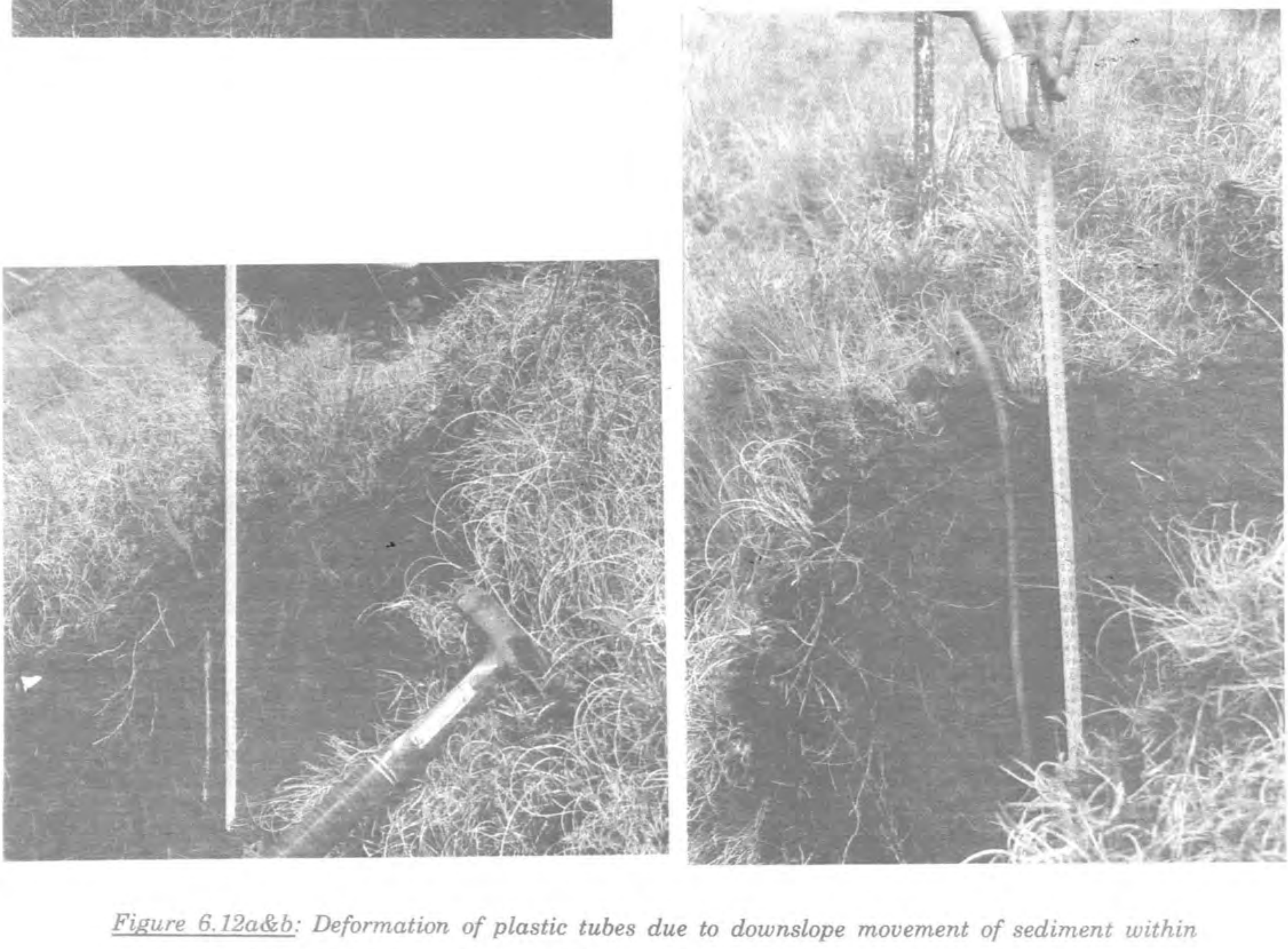


Figure 6.12a&b: Deformation of plastic tubes due to downslope movement of sediment within gelifluction lobes after 18 months (lobes T1 & T7; November 1996).

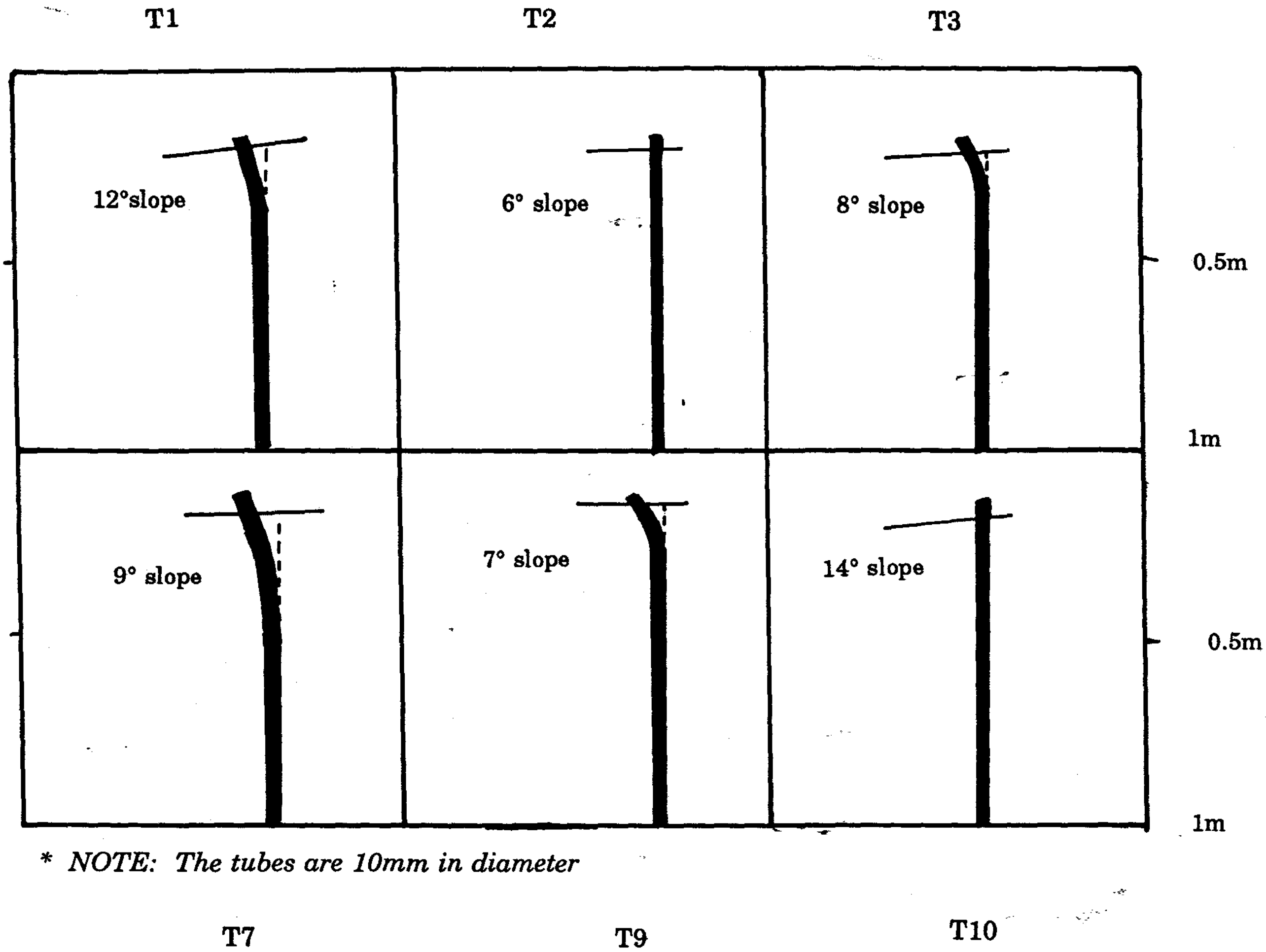


Figure 6.11: *Soil displacements recorded at depth from May 1995 to November 1996*

6.3.3.3. Field observations

In an attempt to analyse the process of sediment movement within a southern African context, soil properties and climatic conditions within the study area were measured in addition to the rates and depth characteristics of soil movement. The physical properties, including grain size, organic content, and Atterberg index properties were measured for samples of the moving sediment from each of the 11 experimental sites. Table 6.5 provides a descriptive summary of these 11 sites.

It was found that the soil samples from the 11 sites displayed a similar particle size distribution. The sediment essentially consisted of silty sand, containing varying proportions of gravel, sandy and silty sized material (Table 6.5). The 11 samples were therefore all found to be frost susceptible according to Beskow's limits (Beskow, 1935; Figure 6.13), and therefore, winter frost heaving is likely.

The measured Atterberg limits indicate non-plastic soil and sediments, with plasticity indices no higher than 14% (Table 6.5). Low Liquid limits were measured between 24% and 28%, the higher values generally corresponding to higher organic contents. Low plasticity reflects the lack of clay and indicates cohesionless soils, particularly sensitive to changes in water content, and susceptible to loss of strength and flowage when water contents are high and exceed the Liquid limit. Therefore, the lobes are susceptible to gelifluction when wet conditions prevail during thaw (Harris, 1981a).

6.3.4. Comparison of short-term and long-term movement

It is acknowledged that short-term measurements of downslope sediment movement lack accuracy (Lewkowicz, 1992), and that long-term records do clearly provide a more adequate appraisal of average geomorphic activity (Smith, 1992). However, due to time constraints, this study could only be conducted over a two-winter and two-spring period. Consequently, the assessment of mass wasting in the Tiffindell area for this period may be highly inaccurate.

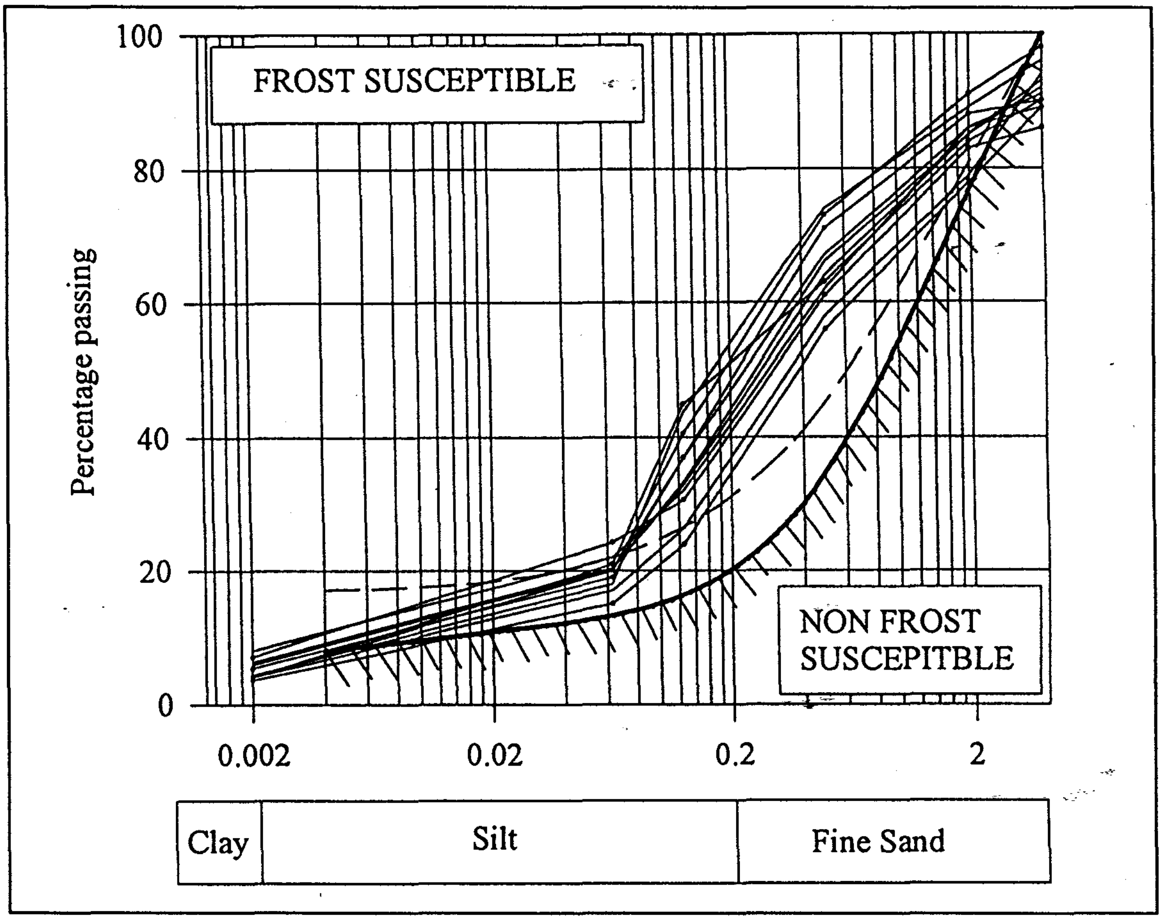


Figure 6.13: Particle size distribution of samples taken (at 0.2m depth) from the lobes at Tiffindell used for surface and subsurface movement measurement. The curves are plotted against Beskow's frost susceptibility limit (cf. Figure 5.1).

It has been suggested by Smith (1992) that a 3 to 4 year period is needed before the short-term rate approximates the long-term rate, and that short-term measurements tend to overestimate displacements (often by an average of 125%). Caine (1981) illustrated a similar bias associated with surface marker stones and attributed it to the time needed for tracer material to be incorporated into the soil. This suggests that the surface markers used at Tiffindell would also require a similar stabilization period.

Site number	Locational Characteristics			Physical Properties						
	Altitude (m)	Aspect	Gradient (degrees)	LOI %	% Clay ≤0.002mm	% Silt >0.002 - ≥0.063mm	% Sand & Gravel ≥0.063mm	Liquid Limit (%)	Plastic Limit (%)	Plasticity Index (%)
Surface Movements										
M1	2770	S-facing	20	4.91	5.9	31	63.1	25	11	14
M2	2777	S-facing	23	4.3	6.3	27	66.7	24	15	9
M3	2786	S-facing	22	6.2	4.4	39.4	56.2	26	19	7
M4	2825	SE-facing	25	8.42	4.1	22.3	73.6	28	19	9
M5	2835	SE-facing	21	6.11	8	24	68	26	15	11
Subsurface Movements										
T1	2779	S-facing	23	4.11	6	26.8	67.2	24	16	8
T2	2789	S-facing	22	6.53	4.2	40.8	55	27	20	7
T3	2790	S-facing	22	5.96	4.1	36.5	59.4	26	15	11
T7	2813	SE-facing	24	4.16	5.4	31.6	63	24	16	8
T9	2826	SE-facing	25	7.27	3.6	21.9	74.5	27	18	9
T10	2834	SE-facing	21	4.86	7.1	23.5	69.4	25	15	10

LIO = loss on ignition

Table 6.5: Geomorphological characteristics of the sites of periglacial mass movement measurement

The techniques of measurement of sub-surface movement also display several problems:

i) installation of the tubes results in substantial disturbance to the surrounding sediment, and several years may pass before the markers or tubes will be fully incorporated into the soil. The tubes may, therefore, have a stabilization period that extends beyond the duration of most mass wasting studies in which they have been employed (Saunders and Young, 1983). In view of the distinct structures identified in solifluction soil (Harris, 1981b), the impacts of installation almost certainly continue to influence solifluction for an indefinite period.

ii) there has been little consideration of the effectiveness of plastic tubing for measuring solifluction, for example, with regards to the relative densities of the plastic tube and the soil material. Great differences may result in the tubes resisting deformation and thereby generating false information about soil displacement (Williams, 1959). Similarly, the tendency of tubing material to curl can obviously lead to indeterminate measurement errors. It is therefore suggested that an alternative method be utilised at Tiffindell in the future, for example test pillars, as used by Harris (1972).

For these reasons, some of the existing equipment has been left *in situ* at Tiffindell, and it is suggested that additional experimental equipment be installed, thus initiating a long-term study. In the event of long-term measurement being undertaken, the short and long-term data can be used together to examine the impact of time on the absolute annual measure of both surface and subsurface solifluction rates. The possibility, however, that any variation in tube measurements may display the natural variability in solifluction behaviour still remains.

6.3.5. Geomorphic implications

Due to the obvious bias associated with the results obtained over the study period, it is difficult to attempt a reasonable assessment of the average work done by solifluction per year within each of the study sites, or even within the area as a whole.

The values of surficial soil movement, recorded in the year 1995-1996 by the remeasurement of the metal markers inserted into the regolith displayed considerable spatial and temporal variation. Recorded amounts of movement for 1995 were less than those recorded in 1996 (Appendix IV), and this can be related to differences in temperature and soil moisture conditions, which were recorded for the study period (Chapter 5), although the majority of the measurement were recorded at a station which was not measured for movement.

Freeze-thaw activity did not occur below 0.1m depth for extended periods in 1995 within the lobe at station B. The majority of the ground freezing that took place was surficial (less than 0.1m deep) and produced small frost heave effects. This is evidenced by the lack of heave of the surface markers observed in November 1995. Nevertheless, fairly rapid soil movements in the downslope direction of up to 20mm were recorded in the uppermost 0.1m during this period (Appendix IV). This seems to favour the significance of gelifluction of water saturated thawing sediment, instead of frost creep related to frost cycles, as the most important mechanism for the movement of the lobes in 1995.

In 1996, a greater depth of freezing recorded over the winter suggested that a greater amount of frost heave occurred within this year, as is evidenced by the heaving of the markers. The ice lens development suggested to occur below 0.1m depth would have enhanced this heave. More rapid rates of surface movements of up to 39mm were recorded in 1996 for the upper 0.1m of the regolith. These measurements may reflect a combination of vertical frost heave, and horizontal soil flowage, resulting in a greater amount of sediment movement for 1996, relative to 1995.

The temporal variation of the recorded movement of the lobes can be explained by reference to the temperature conditions recorded at various depths within the soil at station B. It can be suggested that soil temperatures with depth in 1995 showed a slow, gradual warming trend (Figures 5.4a, 5.5a, 5.6a) from August to the beginning of spring thaw, and while insulated by a snow cover during this time, experienced slow melting of the relatively few ice lenses that had developed within the profile over the winter months. Although the top 0.05cm of the regolith was moisture laden (Figure 5.11), from soil thaw and from snow-melt, the water was able to seep away unrestricted by lower ice lenses, as the soil was not subjected to additional freeze-thaw cycles once the thaw began (Chapter 5).

In contrast, the soil profile underwent major freeze-thaw activity in winter 1996, and greater ice lenses presumably developed with depth than in 1995. The regolith was also subjected to a rapid rise in air temperatures at the onset of spring (Figure 5.3b), and induced thaw within the ground. As the soil thaws from the surface down, it is evident that frozen ground at depth within the profile would exist longer, inhibiting soil water seepage, and causing saturated conditions to persist from the surface down to approximately 0.2m depth. The moisture within the upper 0.05m was subjected to other factors such as insolation and wind, which may have resulted in evaporation of some of the surface soil moisture , as indicated in Figure 5.10. This, therefore, may have increased the potential for gelifluction activity within the turf-banked lobes during 1996, as evidenced by the higher rates of surface movement (Appendix IV).

Spatial diversity of the movement of the lobes is similar for both 1995 and 1996 (Appendix IV). This indicates that factors other than climatic conditions influence the rate of movement of the Tiffindell turf-banked lobes. These factors include aspect, vegetation and available moisture, as discussed previously. Lobes M1 and M3 on the south-facing slope exhibit the greatest mean surface movement over the two years. Both lobes M2 and M5 are bound with dense vegetation, and both display the least recorded movement, lobe M5 on the southeast-facing slope displaying the least movement overall. Soil moisture conditions are site specific, and require measurement at each site in order to assess the spatial varaition.

Factors such as the vegetation cover, seasonal snow cover characteristics, the depth of ground freezing in autumn and the thawing conditions in spring, effectively determine the rate of movement that can be anticipated at a specific site. Therefore, as movement appears to be highly site specific at Tiffindell (Table 6.3) soil movement needs to be recorded at the same locations in future years to be able to compare respective annual rates of movement. Another implication of the results is that assessments of the geomorphic effects of freeze-thaw activity should be based on multiple *in situ* measurements and not on data from a nearby meterological station.

6.3.6. Conclusions

Turf-banked lobes at altitudes of *c.*2800m at Tiffindell move downslope due to gelifluction. Surface rates of movement of up to 39mm per year were recorded, although movement declines rapidly with depth and is essentially restricted to the upper 130mm of the lobes. The rapid downhill movement of gelifluction material is of economic importance at Tiffindell, and will be discussed in Chapter 10.

"A very low-angle slope favours retention of moisture and development of certain forms of patterned ground"
(Washburn, 1973:11)

7.1. INTRODUCTION

The presence of patterned ground phenomena within a landscape provides a visually impressive range of landforms which are frequently regarded as the most obvious characteristic of the periglacial environment (Williams and Smith, 1989). A wide variety of patterned ground features of a symmetrical nature are known. They are usually differentiated according to five criteria: their plan-form geometry (ie circles, polygons, networks); the presence or absence of sorting between coarse and fine debris; the degree of vegetation cover (ie sorted and non-sorted forms); their current status (ie relict or active); and their dominant formative mechanism(s) (Wilson, 1995). The formation of these features is predominantly attributed to frost-related processes (Wilson and Sellier, 1995) at or near the surface of mineral soils and rock material, which is influenced by the properties and behaviour of these substrates.

Patterned ground is subject to a number of problems regarding nomenclature, with considerable debate and lack of consensus among authors (Washburn, 1956; Thorarinsson, 1951; Thorn, 1992). For the purpose of this study, the purely descriptive definition of patterned ground proposed by Washburn (1956) will be used:

"...a group term for the more or less symmetrical forms, such as circles, polygons, nets and stripes, that are characteristic of, but not necessarily confined to, mantle subject to intensive frost action."

The use of a purely descriptive classification avoids consideration of the various origins and dominant formative mechanism of the features (Ballantyne and Harris, 1994). Patterned ground is generally evident in locations where temperatures frequently oscillate above and below freezing. Frost action and the formation of ground ice may both be considered to be fundamental factors in the origin of patterned ground phenomena, always subject to lithology and to topographic controls. Other mechanisms resulting in patterned ground, such as contraction and cracking due to drying and low temperatures, local differential heaving, the operation of cryostatic pressures, ejection of stones towards the freezing surfaces and the eluviation of fines, are comprehensively discussed by Washburn (1956). Much patterned ground is polygenetic in origin (Wilson, 1992), and similar forms can be created from different processes operating together or individually (French, 1976).

Cracking by desiccation, dilation and seasonal frost (that is, thermal contraction) significantly contributes to the development of sorted, as well as non-sorted patterned ground phenomena worldwide. Sorted forms initiated by cracking obviously involve the addition of a sorting process. Desiccation cracking is essentially fissuring due to contraction by drying, and has been cited as an important control on the development of small-scale sorted polygons (Ballantyne and Matthews, 1983). Desiccation cracks develop in response to internal tensile stress applied during the drying phase of the soil due to drainage from evaporation or the withdrawal of moisture to loci of ice formation or cooling of the soil (Warburton, 1990). Dilation cracking is fissuring due to stretching of surface materials, and is generally related to frost heaving of the ground, yet according to French (1976) may be of little importance in relation to the formation of patterned ground. Seasonal frost cracking is the fissuring confined to a seasonally frozen layer, the thermal contraction of frozen ground dividing the soil surface up into polygonal nets. This process is widespread, especially as it is not confined to permafrost areas. The conditions for frost cracking varies for different soils with different moisture contents, and is commonly affected by the rate of temperature change and by the absolute temperature drop (French, 1976). Desiccation, however, does not occur in gravels and coarse sands lacking silt and clay (Ballantyne, 1987) because they are free-draining sediments with large pore spaces (Beskow 1935). The properties of the soil thus influence the frequency of thermal contraction cracking, the spacing of cracks and the size of the patterned ground features which result (Washburn, 1973).

Vertical cracks develop due to horizontal (lateral) contraction of the surface soil layer. Sorting often occurs by movement of coarse material to the soil surface, and subsequently towards and into the surface cracks, as has been shown by experiments by Pissart (1974, cited in Williams and Smith, 1989).

7.2. CLASSIFICATION OF PATTERNED GROUND

The essential elements of the classification of patterned ground have been suggested by Washburn (1956) as being pattern (i.e., dominantly circular, polygonal, intermediate, stepped or striped), and the presence or absence of sorting between stones and fines. These characteristics of patterned ground can be easily ascertained in the field, or from photographic evidence. Patterned ground is affected by gradient (Williams and Smith, 1989), most circles, nets and polygons occurring on essentially horizontal ground, and steps and stripes on slopes. However, some of the former features have also been noted to occur on appreciable slopes (Washburn, 1956).

Circles, polygons, and nets are characteristically composed of several "mesh, or ... cell" units (Washburn, 1973), which tend to become elongated over a transition gradient of 2°-7°. Depending on the conditions, they may grade into stripes with an increase in the slope gradient. Stripe forms are confined to slopes, and can occur without merging upslope or downslope into "mesh" forms. All patterns exhibit both sorted and non-sorted varieties (Washburn, 1956, 1973). Sorted patterns are those in which the mesh of the pattern consists of a border of stones surrounding or demarcating the limits of finer debris. Non-sorted patterns are those in which there is no visible sorting of coarse and fine material (French, 1976).

South African periglacial areas are limited in the patterned ground features that they display. Both active and fossil patterned ground features have, however, been reported from a variety of locations in southern Africa, as reviewed by Lewis (1988b). They occur predominantly above 2900m altitude in regions that include the Lesotho Highlands and Sani Pass (Marker and Whittington, 1971; Boelhouwers and Hall, 1990; Grab, 1996), the Natal Drakensberg (Boelhouwers, 1991a; Boelhouwers, 1994) and the East Cape Drakensberg (Lewis, 1996). The active features rely on present day frost sorting activity for their existence and predominantly include non-sorted and sorted polygons, as well as thufur (Chapter 8). Only those active features which are found within the study site will be considered in more detail. None of the fossil features so far identified in southern Africa are indicative of the former existence of permafrost.

7.3. SORTED AND NON-SORTED POLYGON DEVELOPMENT

- ***Sorted polygons***

Sorted patterns develop on sparsely vegetated soils that contain abundant clasts embedded in fine sediment (Ballantyne and Harris, 1994). Sorted polygons consist of a dominantly polygonal mesh and have a sorted appearance commonly due to a border of stones surrounding finer material (Washburn, 1956). The border stones may range from pebbles to boulders, and tend to influence the widths of the polygonal features (Goldthwaite, 1976). Small polygons (with a diameter of about 10cm) develop in seasonally frozen ground where permafrost is absent, but larger forms (>2m in diameter) appear to be reliant on deep annual freezing, and often (but not always) on permafrost (Ballantyne and Harris, 1994). Therefore, differences in origin between small and large polygon forms can be attributed to the amount of freeze-thaw activity (that is, the intensity of regelation processes and active layer depth) in the area (Nicholson, 1976; Ray *et al.*, 1983), as well as to the overall climate, to moisture availability and to the granulometry of the host sediments (Quinn, 1987).

Research has established that the formation of sorted polygons is caused by differential frost heaving and contraction of the surface layer due to drying and to very low temperatures. Over time, these processes lead to a sorting of the debris in the surface layer into a mesh-like pattern. The coarser debris forms the edge of each cell, while the finer sediment, with or without stones, occupies the interior of each cell (Lewis, 1996) and is occasionally domed up (Buckle, 1990). The origins of these patterned ground phenomena are not fully understood, and their development is often attributed to a 'sorting' circulatory model involving upward and outward movement of coarser sediment in cells to adjacent borders, and compensatory movement of finer material towards the cell centres at depth (Ballantyne and Harris, 1994). The causes of such a circulatory movement are the subject of much debate among researchers.

Ballantyne and Matthews (1983) proposed that the polygonal cracks due to the contraction of the ground act as freezing centres and as the sites of the concentration of clasts, thus forming polygons. Goldthwaite (1976) suggested that randomly spaced concentrations of fines are prone to greater frost heaving than are coarser sediments, as fine sediments are more susceptible to frost action (Beskow, 1935). With frost sorting occurring within the profile, clasts are heaved to the surface away from the fine, often domed, material, and thus surface clasts experience lateral gravity-induced sorting. The above ideas have been further developed by van Vliet-Lanoë (1988) for the Spitsbergen area. She suggests that a combination of differential frost susceptibility and ground cracking, associated with textural contrasts within the sediment, are causes of mass displacement during freezing.

Sorted polygons never appear to develop singly, but occur as a 'network' (Ballantyne and Matthews, 1983). They range in size from large forms extending up to approximately 10m across to 'micropolygons' having a diameter of often less than 10cm. Polygon size tends to increase with increasing severity of climate and with the availability of water (French, 1976).

- ***Non-sorted polygons***

Non-sorted polygons are patterned ground whose mesh is dominantly polygonal, yet lack the visible sorting of coarse and fine debris, and thus the border of stones. Their borders are commonly delineated by wedge-shaped fissures or cracks which narrow downwards, and the polygon centres may exhibit slight updoming (Williams and Smith, 1989) resulting in a hummocky micro-relief (French, 1976). They may often take on the appearance of 'desiccation cracks', small non-sorted desiccation polygons being commonly found worldwide, the drainage of the site affecting the polygon size (van Vliet-Lanoë, 1988). Non-sorted polygons include ice-wedge polygons, frost crack polygons and vegetation polygons (Washburn, 1956). Frost crack and desiccation polygons (Romanovskij, 1973) differ from the others due to the absence of any wedge fillings (Washburn, 1969).

Like the sorted variety, non-sorted polygons are group forms and never develop singly (Ballantyne and Matthews, 1983). Vegetation may be concentrated along their borders, emphasising the pattern, but is usual only in larger forms. These polygons may develop as large forms greater than 100m in diameter (*e.g.* ice wedge polygons, Washburn (1956)), or as small forms of as little as 5cm across. Commonly the smaller forms have a pentagonal or hexagonal mesh and may occur in the central areas of larger polygons of both varieties. They are commonly found to form in most soil types, including well-sorted fines, sand, gravel or diamicton which are associated with an abundant water supply (van Vliet-Lanoë, 1988).

All polygon forms occur most frequently on nearly horizontal surfaces and are generally not found in the polygon form on slopes steeper than about 4° (Lewis, 1996; Davies, 1969). Small sorted forms occur in environments ranging from polar regions to temperate highlands, and commonly develop in permafrost free areas in which there is appreciable frost action. For polygon formation to occur, it is essential that suitable sediments as well as suitable climatic conditions should exist at sites of low slope angle. Large sorted polygons are best developed in permafrost environments, therefore fossil forms may be regarded as reasonable evidence of a former permafrost area (Washburn, 1973; Karlstrom, 1990).

Some polygons may be ephemeral (short-lived) features, forming during the winter and spring when frost action is common at high altitudes, and being erased by other processes in the summer, when frost is rare or absent (Lewis, 1996). According to Karte (1983) small polygons, such as those common in high altitude areas of South Africa (Hastenrath and Wilkinson, 1972; Boelhouwers, 1991a, 1994; Lewis, 1988b, 1996), form when mean annual air temperatures do not exceed 3°C.

7.4. POLYGON DEVELOPMENT AT TIFFINDELL SKI

Both sorted and non-sorted polygons were observed within the study site, although few locations characterised by suitable sediments, climatic and moisture conditions, and low slope angle exist. Their occurrence at Tiffindell is limited to locations of abundant moisture where the angle of slope, with the exception of one site, is less than 4°. The fresh appearance and surface micro-relief of the sorted patterns is indicative of frost activity under present climatic conditions within the Tiffindell area. The sites of polygon activity were described with respect to their relative altitudes, slope angles and mesh sizes. Sediment samples were taken for selected sites, a field assessment being made for the rest. Table 7.1 provides a summary of sites displaying active patterned ground (Appendix I), with notes on their relative distribution and size.

TYPE	S#	ALT.	GRADIENT	MESH WIDTHS	GRAIN SIZE	OBSERVATIONS
SORTED POLY-GONS	1	2922m	2-3°	± 7cm	pebble border; fine grained centre	On the ridge (waterdivide); in a bare area with scarce protective vegetation.
	2	2899m	1-2°	≤ 6cm	pebble border; fine grained centre	45x40cm; vegetation-free areas surrounded by low protective vegetation.
NON - SORTED POLY-GONS	3	3001m	2°	± 10cm	sandy silt	In slight depressions in between bedrock outcrops (shallow pools); vegetation scarce.
	4	2942m	1-2°	± 13cm	mainly sandy; with pebbles & fines	On the ridge; in a bare area with a moderate amount of protective vegetation.
	5	2935m	1-2°	± 12cm	mainly sandy; with pebbles & fines	On the ridge; little vegetation; in slight depressions in between bedrock outcrops.
	6	2916m 2913m	1-2°	large variations ± 10cm	silt - fine grained	Several sites; in areas surrounding thufa site; under shallow water in summer; exposed in winter months.
	7	2783m	2-3°	± 5-8cm	sandy silt with pebbles	At 'Ice Pub'; South-facing slope; exposed area due to excavations and ski run.
	8	2759m	± 11°	range from 4x9cm to 10x17cm	medium grained sand	Disturbed ground adjacent to the ski run; no vegetation; extremely moist due to snowmelt and water from snowguns.
	9	2759m	1-2°	large variation from ±3cm to ≥10cm to ≥25cm	sandy silt	North-facing slope; little vegetation; prominent gaps around stones. Large polygons with cracks ±5cm deep.
	10	2755m	3°	± 8cm	silty sand	South-facing slope; downslope from water tanks; abundant moisture available; moderate vegetation.
	11	2755m	± 7°	average ± 7cm	silty sand	On the banks of the Big Dam; highly disturbed area; but very moist.

KEY: #S: Site number; see Figure 7.1 **ALT:** Altitude

Table 7.1: Polygon sites identified in 1995 and 1996 at and adjacent to Tiffindell.

7.4.1. Small non-sorted polygons

This polygon type was found in greater abundance than were sorted polygons (Table 7.1; Appendix I). They were most commonly observed in extensive vegetation-free areas, characterised by sands and silts with few pebbles, but were not confined to such areas. The dominant soils were generally fine to medium sands, the sand content of which ranged from 30% - 68% at the sampled sites. They were largely found in locations characterised by abundant ground moisture, such as downslope from water storage tanks (site 10), on the banks of a dam (site 11), adjacent to the ski run (site 8) and in areas of poor drainage (site 6).

Sharp-edged fresh cracks outlined non-sorted polygons with mesh sizes ranging from 3x4cm to 20x25cm throughout the study site. Measurement of the widths of the dividing cracks at the surface ranged from 0.1 to 0.7cm, at sites 8 and 9 respectively. Most of the cracks had a surface width of about 0.3cm. The depths of the cracks of the polygons also varied, and some could be probed with a wire to depths greater than 5cm. The average depth of the polygon cracks was recorded as 2.5cm. Ballantyne and Matthews (1983) suggest that fresh cracks are no more than a few centimetres deep. Excavation of site 4 showed that several fine cracks extended to a depth greater than 16cm.

The Tiffindell area is subject to diurnal freeze-thaw cycles on a seasonal basis (Chapter 5). The winter season experiences the most frequent and intense freeze-thaw cycles. Given the relatively high precipitation and amount of ground moisture, as well as the cool winter temperature regime of the Tiffindell area, desiccation cracking is less likely for pattern initiation than is seasonal frost activity in the cold months. However, it is noted that the development of non-sorted polygons is not limited to the winter season, therefore, due to the lack of freeze-thaw activity in the warmer seasons, desiccation cracking must be an important process in the development of these polygons.

Frost cracking and desiccation polygons (Figure 7.1), such as those of the Tiffindell area, are also of widespread distribution in other areas of similar climatic conditions in southern Africa such as the Lesotho Highlands (Hanvey and Marker, 1992) and the Natal Drakensberg (Boelhouwers, 1991a). Within the winter seasons of 1995 and 1996, small non-sorted polygons were observed to form at Tiffindell under cold and typically clear sky conditions that allowed freezing and thawing of the ground surface to take place, although often only on a diurnal basis. These polygons occurred mainly on bare silty mineral soil, often in the close vicinity of other patterned ground and of frost heave phenomena (such as sorted polygons, thufur, frost heaved and needle ice lifted debris, and gaps around stones) and are considered to form primarily as a result of seasonal frost cracking processes. Desiccation polygons, between 5cm and 10cm in diameter, were observed on the floors of shallow depressions where water periodically gathers and stands (sites 3 and 6). The occurrence of polygons under water has been noted by other workers elsewhere (Washburn, 1956, 1969; Lewis, 1988b, Thorn, 1992) and their formation has been ascribed to the withdrawal of soil moisture towards ice nuclei as a result of frost action (Ballantyne and Matthews, 1983) subsequent to draining or evaporation of the surface water that flooded the depressions.

At site 8 polygon development was noted in an elongated area displaying frost heave, adjacent to the ski run at an altitude of 2759m. This area has been exposed due to the construction of the ski run, and is completely free of vegetation. It is constantly moist from groundwater seepage from melting snow on the ski run, as well as from water that is expelled from the snowguns during the snow making process. The surface of the elongated heaved area, measuring 1.2mx0.9m, was cracked into numerous polygons, despite the steep gradient of 11°. The polygons, which were elongated parallel to the length of the heaved area, were developed in damp sand, and had sizes ranging from 4x9cm to 10x17cm, a few being equidimensional. The dividing cracks were up to 0.6cm wide, and appeared fresh. The fact that polygons were absent in the adjacent unheaved surfaces, which were underlain by similar materials, and that the polygonal pattern was so largely dependent on the shape of the heaved area, warrants the conclusion that these polygons were consequent on dilation of the surface during heaving (Washburn, 1973), although this is an uncommon formative process. A similar scenario on a much larger scale, was reported by Washburn (1969) for the Mesters Vig District, Greenland.

In some areas, a fresh pattern was seen to have developed over an underlying (and hence older) pattern (*e.g.* at site 6). The age of the older pattern is difficult to ascertain, as it may have developed within the same season, or may be a remnant from a previous cold period. The fresh polygons were noted to be mainly flat, but some of the older polygons were slightly domed. This may be a product of erosion of the border cracks, or may indicate the persistent presence of localised frost heaving in the area.

The non-sorted form of polygons at Tiffindell frequently exhibit fast (often overnight) development, especially those related to diurnal freeze-thaw cycles. Coupled with this quick development is often a short-lived existence, especially when subjected to thaw, as they are typically soft and vulnerable when the surface substrate is not frozen. They appear to be readily eroded by rainfall, snowfall, runoff, as well as people. Some non-sorted polygons in undisturbed areas, especially those developed in fine sediment, were noted to become strongly desiccated and hard during the summer, and thus resistant to erosion (for example, those at site 6).

No evidence was found to indicate the presence of ice-wedge polygons or of ice-wedge casts in the Tiffindell area. This suggests either that permafrost (which is necessary for ice-wedge development) did not exist within the study area in the past or, if it did, that the evidence has been destroyed by more recent cryogenic activities.

7.4.2. Small sorted polygons

Well developed sorted polygons were observed along the Ben MacDhui - Breslin's Kop ridge in the Tiffindell area at altitudes of around 2900m (Table 7.1; Appendix I, sites 1 and 2). Sorted polygons have been reported at similar altitudes in the Natal Drakensberg (Boelhouwers, 1991a) and in the Lesotho Highlands (Hanvey and Marker, 1992).

The polygonal (straight borders between angular bends) shape of the polygons at sites 1 and 2 is delineated by a border of small pebbles and edgewise gravel. The mesh size of the micropatterns range from ≥3cm to ≤7cm in diameter (Figure 7.2). The mineral soil of the central areas is characteristically silty, with a low gravel content. The grain size distribution for an active sorted polygon from site 1 is shown in Figure 7.3.

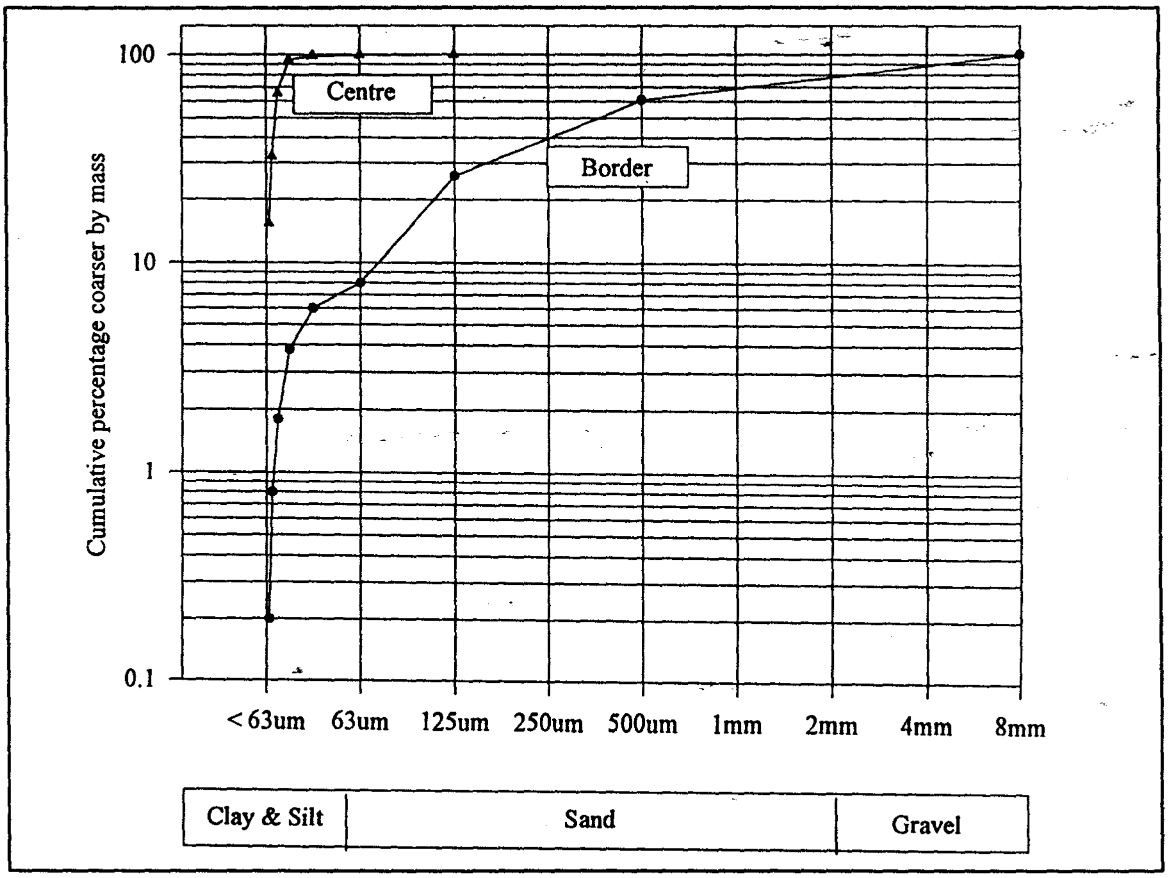


Figure 7.3: Particle size distribution of centre and border sediments of a sorted polygon from site 1 (altitude 2922m) at Tiffindell.

The border and central area are displayed separately, and indicate that the feature is moderately well sorted. The centre fines have a mean grain size of 0.1875mm (medium to fine sand), and tend to be poorly sorted. The majority of the clasts (73%) forming the polygon border are between 9 and 16mm a-axis diameter (pebble size) and well sorted. This maximum size of sorted clasts is generally consistent within the sorted polygon forms. Excavation at site 1 indicates sorting to a depth of between 5cm and 8cm, and suggests that the shallow ground freezing limits the maximum depth of sorting, which in turn dictates the dimensions of the resultant patterns and the maximum size of sorted clasts (Ballantyne and Matthews, 1983). Only regolith containing clasts that are predominantly smaller than approximately 16mm would appear to experience sorting into patterns under present climatic conditions.

Pentagonal forms predominate within these small features, although square forms are also present. They are common on slopes with gradients of approximately 2°, which are essentially unvegetated. This lack of vegetation may indicate that intense frost action occurs in the ground in those areas of polygon formation. The development sites are often surrounded by low-lying dense vegetation which appears to protect the sorted polygon areas from the often harsh winds and direct sunlight which enhance groundwater evaporation.

Active sorting in the soils of the Lesotho highlands and the Natal Drakensberg has resulted in the formation of micro-polygons with mesh widths varying between 5cm and 20cm (Boelhouwers, 1991a, 1994). These polygons are of similar size to those presently active in the Tiffindell area, as well as to those discovered in 1994 by Lewis (1996). Lewis recorded polygons of mesh widths varying from 6cm to 12cm along the Ben MacDhui - Breslin's Kop ridge in the vicinity of site 1, whereas polygons of 6cm to 7cm mesh were identified in 1995/1996 (Figure 7.2).

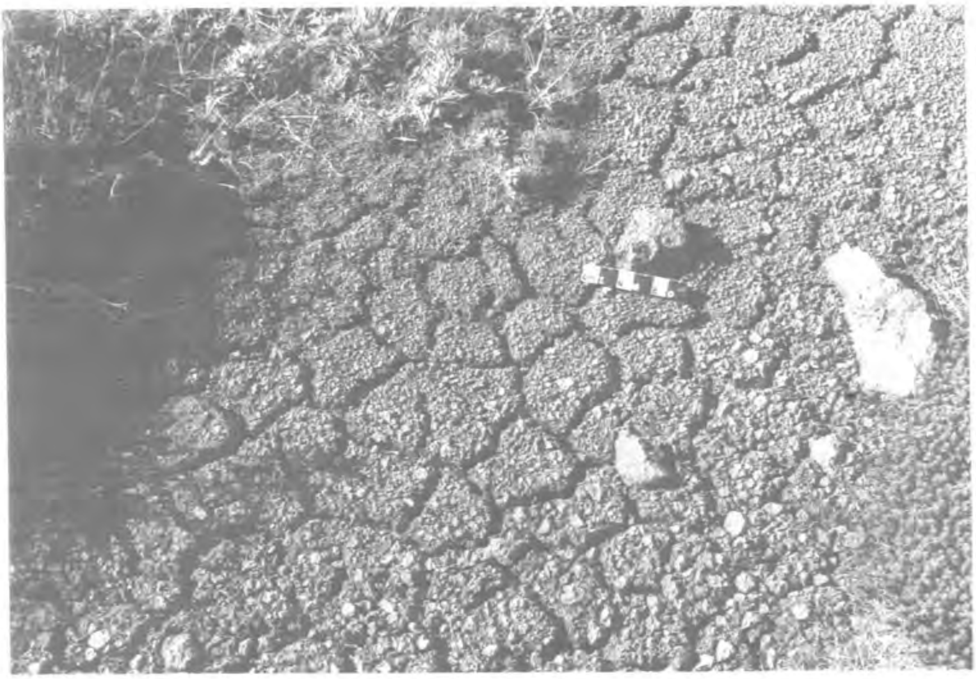


Figure 7.1: Non-sorted polygon development at 2942m in July 1992 at Tiffindell. Mesh widths approximately 10cm in diameter.

Warren Scholtemeyer

Figure 7.2: Sorted polygon with mesh widths of up to 7cm on the summit of the Ben MacDhui - Breslin's Kop ridge at an altitude of 2922m. Note elevated gravel borders and moist, finer soil in the centres.

Development of sorted polygons similar to those of the Ben MacDhui - Breslin's Kop ridge has been reported to occur in the St Mary's Range, Otago, New Zealand, within days (Orwin, 1993), usually following the passage of a cold front. Orwin (1993) experimented with a trial plot, and discovered that sorted polygons may develop within two or three days if suitable conditions occur. His results indicated that most freeze-thaw activity responsible for patterned ground formation in the St Mary's Range occurs only on a diurnal basis when abundant moisture and below 0°C temperature conditions occur together, regardless of season. Therefore, as long as the specific conditions are fulfilled, polygon development at Tiffindell can be anticipated for the winter months, as well as occasionally in the warmer months.

Orwin's (1993) experiment suggests that the sorted polygons observed at sites 1 and 2 on the Ben MacDhui - Breslin's Kop ridge in November 1995 may have been of recent origin, considering the passage of a frontal system at the time of discovery. The location and observation of polygons within the research area is often difficult, as removal of the features by wind, rain and snowfall may be as rapid as their formations. Another major problem, especially during the Resort's skiing season, is their destruction by the large number of visitors to the resort who display little respect for patterned ground, and other sensitive periglacial phenomena!

The evidence of sorted polygon formation indicates present day frost activity within the Tiffindell area. The depth of frost penetration at 2800m has been recorded to a depth of at least 0.2m (in 1996, Chapter 5), and therefore, can be expected to be of similar depths along the high-lying areas. This correlates well with values recorded by Boelhouwers (1994) in the Natal Drakensberg. By using the air temperature data recorded at 2800m with the 0.6°C/100m lapse rate, winter air temperatures for the ridge (1900m) can be extrapolated (without considering any additional factors). Mean minimum air temperature for May to September (1996) was recorded as -2.573°C (Chapter 5, Table 5.4), and therefore probable mean minimum winter air temperatures on the ridge for the same period can be suggested as being below -3°C. Thus, seasonal frost penetration can be expected within the Tiffindell area, those level areas on the ridge being highly susceptible in the winter seasons due to low mean temperatures (Washburn, 1973)

7.4.3. Summary: sorted and non-sorted polygons

Both sorted and non-sorted polygons are of similar size in the Tiffindell area, although the variation in sorted polygon sizes is greater. Non-sorted polygons form in a wide range of soil types and they are commonly found in areas of high soil moisture content. For example, site 6 was located in an area of poor drainage, site 10 was downslope from water tanks, site 11 was on the banks of a dam, and site 8 was adjacent to the ski slope, where moisture from snow melt and snowmaking was plentiful. Sorted polygons appear to form only where silty sediments suitable for the retention of soil moisture exist. Polygons exist only above an altitude of 2755m in frost susceptible soils.

THUFUR CHAPTER 8

> Thufur have been classified as a special non-sorted net form, whose mesh is "*characterised by a three-dimensional knoblike shape...*" (Washburn, 1956).

8.1. INTRODUCTION

Dome-shaped hummocks (thufur), averaging 50cm in height and 100cm in diameter, are a widespread micro-relief patterned ground phenomenon, and occur in "arctic, alpine and subarctic environments where climate allows ... seasonal ground frost patterns" (Schunke and Zoltai, 1988:231). Thorarinsson (1951) designated those forms which were supposedly produced under permafrost conditions as *earth hummocks*, and those occurring in seasonally-frozen soils as *thufur* (singular *thufa*). Tarnocai and Zoltai (1978) and Washburn (1956, 1973) regarded earth hummocks and thufur as being different names for the same landform. Harris (1988), however, has shown that earth hummocks form under different environmental conditions to thufur. The intense heave associated with permafrost environments commonly results in earth hummock development, thufur being produced by the "local displacement of surface soil material due to seasonal frost penetration" (Grab, 1994:112), permafrost not being a prerequisite (Thorarinsson, 1951; Harris, 1988, Pollard, 1988). The term 'thufur' is therefore used to describe the features observed in southern Africa, since permafrost does not exist in the sub-continent at present.

No *one* unequivocal mechanism for hummock, or thufur development has been identified with certainty, although much research has been undertaken on these features, especially in polar and subpolar regions (*vide* Schunke and Zoltai, 1988). Within southern Africa, thufur have been observed in the Lesotho Mountains (Marker and Whittington, 1971; Hastenrath and Wilkinson, 1973; Boelhouwers, 1991a; Hanvey and Marker, 1992) and in the adjacent Drakensberg of South Africa (Grab, 1994; Lewis, 1996). Field observations and measurements by Grab (1994) provide empirical information on the external and internal morphology of thufur at Mohlesi Valley, in the Natal Drakensberg, South Africa, which serves as a basis for comparison for the data obtained for Ben MacDhui.

8.2. THUFUR CHARACTERISTICS

8.2.1. Distribution

As thufur are not dependent on a permafrost layer for development, they are consequently relatively widespread throughout the world's alpine areas (Tarnocai and Zoltai, 1978). Thufur have been reported from many locations, including Iceland (Gerrard, 1992), Greenland (Raup, 1965 cited in Washburn, 1973), Canada (Sharpe, 1942 cited in Washburn, 1956), Sweden (Lundqvist, 1969), the Snowy Mountains of Australia (Costin and Wimbush, 1973), the St Mary's Range of New Zealand (Orwin, 1993), the mountains of Kazakhstan in central Asia (C.A.Lewis, pers.comm.), as well as in southern African alpine areas (Marker and Whittington, 1971; Hastenrath and Wilkinson, 1973; Boelhouwers, 1991a; Hanvey and Marker, 1992; Grab, 1994; Lewis, 1996).

The primary process for thufur development is frost heaving, therefore thufur distribution is dependant on elements such as heave-sensitive soil, abundant soil moisture and seasonal frost penetration. Should any of these conditions be lacking, thufur formation does not take place. Therefore, both local and global thufur distribution is dependent on these criteria.

The distribution of thufur is also reliant on factors such as altitude, local climatic conditions, topography, vegetation types and the availability of deep soils to sustain their development. Thufur development typically occurs on flat or gently sloping areas of low gradient (approximately 2° or less) which are subject to mean daily winter temperatures of below freezing (section 8.3.1.). They are also typical of fine-grained soils which are poorly drained or have a continuous supply of groundwater (Pollard and van Everdingen, 1992). Thufur usually occur on meadow grassland, which is of limited occurrence at the high altitudes at which periglacial conditions prevail in alpine areas, thereby contributing to the relative rarity of occurrence of these features (Lorenzo, 1969).

Thufur usually occur in large closely spaced groups in distinct fields (Lundqvist, 1969), although they may be found on occasion as scattered individuals. When closely spaced they are separated by troughs that are referred to as 'inter-hummock' depressions.

In an attempt to establish the upper and lower altitudinal boundaries for thufur development, those factors that influence thufur distribution should be considered. Upper limits have been suggested by Schunke (1977, cited in Grab, 1994) to exist where the amount of fine material covered by meadow grasses becomes so restricted that development of thufur is no longer possible. Grab (1994) suggests that, for the Mohlesi Valley, the availability of deep soils and abundant near-surface moisture determines the upper limit, rather than the occurrence of meadow grassland, which occurs above and below the altitudinal limits at which thufur exist at Mohlesi. One of the major considerations for the determination of lower limits for thufur distribution is the regional climate, which displays great control over the relative depth of frost penetration.

8.2.2. External morphology

The term 'thufur' describes a type of patterned ground that is essentially a form of unsorted net, that is, patterned ground whose mesh is neither dominantly circular nor polygonal (Washburn, 1956; 1973). In plan view, individual thufa appear generally circular or oval, yet on gently sloping sites, thufa sometimes elongate downslope. Where the gradient exceeds about 6°, such elongate thufa may grade into non-sorted stripes that form a ridge-and-furrow pattern following the line of maximum slope (Ballantyne and Harris, 1994).

In southern Africa, the average heights of located thufur have been recorded as 0.13m at Tina Head (±41km from Tiffindell), near the border of South Africa and Lesotho (Lewis, 1996) and 0.17m in the Mohlesi valley (Grab, 1994), the basal diameters ranging between 0.2m and 0.98m (average 0.48m) for individual mounds (Grab, 1994; Lewis, 1996). Considerable variations in size ranges between the alpine, and the Arctic and sub-Arctic areas (Schunke and Zoltai, 1988) reveal that thufur in alpine areas are generally smaller than those in high latitudes (Costin and Wimbush, 1973). This may be explained by the relatively brief freezing period and the shallow freeze penetration characteristic of most alpine areas (Grab, 1994).

Individual thufa mounds are usually covered with vegetation. In Alpine areas they are cloaked by meadow grass, which includes *Koeleria capensis* and *Abelia galpini* in the high alpine areas of southern Africa (Costin and Wimbush, 1973; Grab, 1994; pers.obs.). Lundqvist (1969) has suggested that the vegetation cover may help to keep the silty to sandy material collected as mounds. Often, different vegetation may develop on the mound itself and in the lower damp inter-hummock depression due to different drainage conditions caused by microrelief (Lötschert, 1974 cited in Schunke and Zoltai, 1988; Lorenzo, 1969). Such vegetation differences may cause major contrasts in the thermal regime of various parts of the thufa (Schunke and Zoltai, 1988).

8.2.3. Internal morphology

The internal morphology of thufur is characterised by "disrupted and displaced horizons and strata" (Schunke and Zoltai, 1988:237), the whole structure indicating that cryoturbation plays an active role in generating and preserving the shape of thufur. This is often visible in variation of soil colours within mounds. The individual layers, however, show less complex thrusting of the soil than in earth hummocks characteristic of permafrost areas (Harris, 1988). The intruded material often consists of parts of soil horizons or organic layers, and scattered organic carbon is common within the unfrozen mound material, the concentration often reaching 2-5% (Zoltai and Tarnocai, 1974 cited in Schunke and Zoltai, 1988).

The interiors of thufur typically display a core of fine-grained mineral soil that is virtually stone-free, and a rather thick humus mantle. A few have been observed to have a stone core (Raup, 1965 cited in Washburn, 1973; Boelhouwers and Hall, 1990). Silt and clay sized particles represent the greatest proportion of the soil (58-99% of the soil in Canada; 60% in Iceland; Schunke and Zoltai, 1988), their remainder consisting of fine sand. The soil is usually homogenous within a thufa, but there may be a slight increase in the coarser particles on the sides and the tops of a mound. Except for the work by Grab (1994), and to a lesser extent by Hastenrath and Wilkinson (1973), the internal characteristics of thufur in southern Africa have not been studied in detail.

8.3. THUFUR FORMATION

The conditions favouring formation of thufur include:

1. Perennial discharge of groundwater providing suitable hydrologic conditions,
2. The presence of a low permeability layer close to the ground surface which acts as an aquitard, and
3. A cold winter with daily mean temperatures below freezing (Pollard and van Everdingen, 1992).

8.3.1. Climatic controls for formation

A large majority of hummocks develop in areas underlain by permafrost while, alpine thufur, which are dependant on seasonally-frozen ground, are of less frequent occurrence (Scotter and Zoltai, 1982, cited in Schunke and Zoltai, 1988). Since permafrost does not exist in southern Africa at present, only the climatic conditions necessary for alpine thufur development will be considered.

According to Karte (1983) thufur form when mean annual air temperatures do not exceed +3°C. Lundqvist (1969), however, has suggested that thufur may form at mean annual air temperatures at least as high as +5°C. Local climatic conditions, especially the distribution of snow, is important for thufur formation. The insulating effect of deep and late-lying snow, which may gather in depressions around individual hummocks that afford shelter from wind, tends to decrease frost penetration. The absence of thufur at such sites was noted by Thorarinsson (1951). Frost action is essential for thufur formation and minimum temperatures must therefore fall below freezing at least seasonally.

In Iceland, thufur are known to be capable of forming within a few years or decades under present conditions (Schunke and Zoltai, 1988), although it is not yet known whether the same is true of southern Africa.

8.3.2. Thufur development

The mechanism of hummock formation is not known with certainty. Research has revealed that the mounds are produced by the permanent displacement of local surface soil material in frost-sensitive soil in the presence of plentiful moisture under climatic conditions that generate seasonal frost penetration (Schunke and Zoltai, 1988). That is, their formation is due to the upwelling of soil (pore) water in a seasonally active mineral soil by cryoturbation (frost-generated soil movements) and cryostatic pressure (Pollard, 1988), and to some extent by the product of ice segregation processes (Nelson *et al.*, 1992). This produces an initial miniature mound. Schunke (1981, cited in Schunke and Zoltai, 1988) suggested that initiation of the original mound could be connected with the movement of moisture towards the freezing front, which transfers fine-grained mineral soil particles in suspension (Lundqvist, 1969). These may then congregate to form a relatively homogenous fine-textured soil, and thus a cellular centre could develop by water movement associated with freezing. Another mechanism for the initiation of hummocks may be the random development of frost-heaved spots, as suggested by Mackay (1980, cited in Schunke and Zoltai, 1988).

The small newly-developed mounds are slightly better drained than the intervening areas, and as a result, an insulating, more mesic vegetation develops on them than in the moister troughs. Thus it is suggested that the uneven surface, as well as the differences in vegetation, may create thermal variations in the soil, the drier mound tending to lose heat more slowly than the moist inter-hummock depressions. In the more moist sections, the freezing front will depress faster and deeper into the soil profile than under the warmer mound, thus establishing lateral pressures towards the centre of the mound, displacing more materials and eventually forming the thufur (Schunke and Zoltai, 1988). Once formed, differences in moisture content, insulative vegetation cover and soil texture act to create cryoturbation processes that will maintain the hummocks.

Due to the more restricted depth of cryoturbation in non-permafrost areas, maximum deformation of thufur in alpine areas takes place within the upper section of the mound, the layers at a depth greater than 60cm being left undisturbed (Schunke, 1977b cited in Schunke and Zoltai, 1988). Therefore, Schunke (1977b) argues, the depth and rate of penetration of seasonal frost is responsible for the maintenance of the thufur. Therefore, as the freezing depth is dependent on the availability of moisture and the occurrence of cold temperatures, the occurrence of dry years or mild winters will result in limited or no cryoturbation (Zoltai and Pettapiece, 1974).

The material underlying the deposit in which the thufur develop also plays an important role, as an impervious layer will impede drainage, increasing the moisture conditions, and will resist inter-granular pore pressures, directing the resulting forces upward.

8.3.3. Thufur breakup

The breakup of a thufa involves the disruption of one or more of the individual's sides. With the removal of vegetation and soil, a disrupted surface forms and subsequent disintegration of the thufa is the result. Several factors may influence thufur breakup.

Where the wetlands are retreating, those individuals on the periphery are particularly susceptible, and will experience breakup first (Grab, 1994). Frost cracking or internal pressures (Crampton, 1977, cited in Grab, 1994) may result in many thufa showing slight cracking and splitting during the winter months, disturbing the protective covering of vegetation. The prolonged presence of surface water and/or ice may also cause disruption, particularly in thufa individuals occurring on level (1°) ground (Grab, 1994). This has been noted to contribute towards breakup on the upslope and cross-slope sides of thufur, as ice commonly becomes attached on these sides. Those thufur broken on their upslope sides often experience removal of the internal soil, and the replacement thereof with water or, more commonly, ice (Grab, 1994).

Needle-ice activity often detaches turf from the underlying soil on both the mound, as well as the inter-hummock depression. Consequently, the soils are exposed and no longer held together by turf, making them more susceptible to erosion. Van Zinderen Bakker and Werger (1974) suggested that needle ice was important in the development of 'crater-like thufur' associated with high-altitude bogs in Lesotho.

Van Zinderen Bakker and Werger (1974) also suggested that wind action is an important erosional agent, as thufur disintegration in Lesotho is common on the westward sides (that is, the windward sides) of the hummocks. Animal interference may also contribute to the disruption of thufur, especially trampling by the grazing animals which frequent these wetland areas due to the favourable grasses and high moisture availability.

8.4. THUFUR DEVELOPEMENT IN THE TIFFINDELL AREA

Thufur exist at altitudes above 2900m in the East Cape Drakensberg near Ben MacDhui (Appendix I; Figure 8.1), where the general absence of snow during the winter months and the occurrence of subzero temperatures during most nights from May to August creates favourable conditions for the localised development of frozen ground.

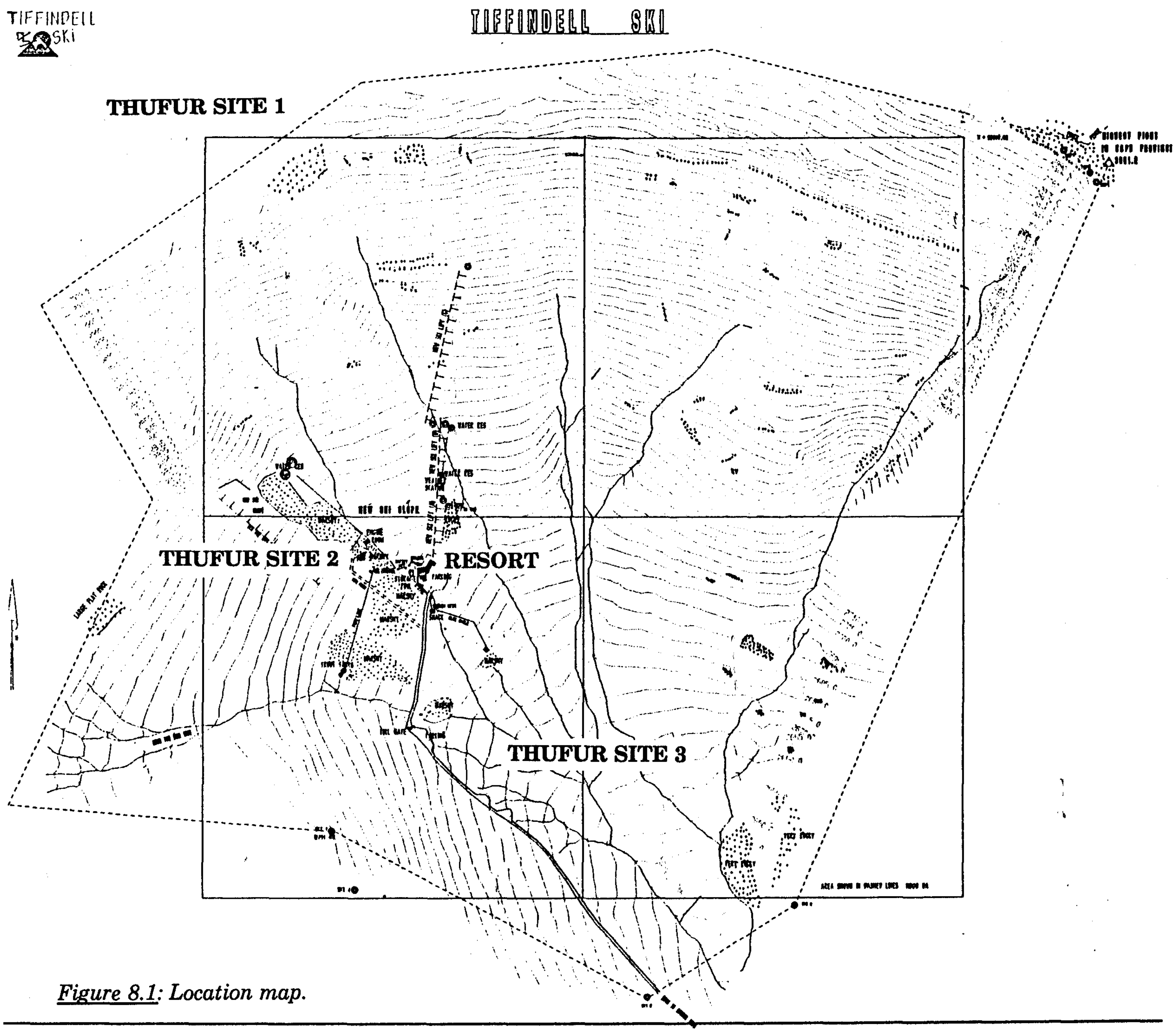


Figure 8.1: Location map.

8.4.1. Distribution and characteristics

Thufur were investigated at a site located at 2916m, across the Tiffindell border in the former-Transkei, as well as at two lower sites (altitudes of 2779m and 2650m; Figure 8.1). Observations of these features have been restricted to only a few sites in southern Africa, Lewis (1996) reporting the location of "frost mounds" at Tina Head; others being reported for the Natal Drakensberg (Boelhouwers, 1991a; Grab, 1994) and Lesotho mountains (Hastenrath and Wilkinson, 1972; van Zinderen Bakker and Werger, 1974). An estimation of the upper and lower limits for occurrence of the Tiffindell thufur can be based on those limits suggested by Schunke (1977, cited in Grab, 1994) and Grab (1994). That is, by the availability of deep fine soils suitably covered with grassy vegetation, as well as by abundant near-surface moisture. The Tiffindell area is not predominantly covered by meadow grass vegetation (Chapter 3, section 3.5) as is the Mohlesi valley (Grab, 1994), and has many exposed areas lacking grassy vegetation. The paucity of suitable plant cover therefore acts as a major restriction for Tiffindell thufur development. The availability of ground moisture also affects the spatial distribution of wetlands and consequently areas of thufur development.

8.4.2. Site 1 - 2916m

A population of approximately 30 individual thufa have formed within a slightly depressed, yet level (gradient between 1° and 3°), wetland area (approximately $10m^2$) that experiences poor drainage conditions throughout the year. This wetland appears to serve as a watering hole for the local fauna. The area is well vegetated, and is commonly characterised by hydrophillic species, such as *Kniphofia thodei* (Red Hot Poker). The thufur, occurring in the centre of the wetland, are well vegetated with meadow grasses and cushion plants (cover >80%) (Figure 8.2), the dominant species including *Koeleria capensis* and *Abelia galpini*. An as yet unidentified grass species at the site appears to be more hydrophillic than the other identified species. This hydrophillic grass species most commonly occurs in the inter-hummock depressions, and may therefore indicate that predominantly wetter conditions prevail there. These areas between thufa range from less than 8cm to often more than 30cm. The central grassy area upon which the thufur have developed is surrounded by a narrow unvegetated area. This bare area, during the wet season is normally saturated with water and contains pools of water on its surface.

In the cold winter months of 1995 and 1996, when the site was examined, the saturated bare areas contained ice lenses and displayed frost heave and needle ice activity, as well as non-sorted polygon development which persisted even through the warmer seasons.

8.4.2.1. External morphology

The external dimensions of 20 individual thufa were measured at Site 1. The height of each thufa was measured from the soil surface at its base to the corresponding surface at its apex, as were both cross-slope and downslope basal diameters (Table 8.1).

The Tiffindell thufur at Site 1 have an average height of 15.05cm (range 7 to 27cm; ±5.104 s.d.) and an average diameter of 37.675cm (17 to 90cm; ±15.687 s.d.; Table 8.1). Other smaller forms, which were essentially little more than bumps on the surface of the ground, were also noted in the area, but not recorded. These may be what Schunke (1977, cited in Schunke and Zoltai, 1988) called 'embryonic hummocks'. They probably indicate that the processes of thufa formation are active.

Thufur in alpine areas are reported as being generally smaller that those of arctic regions due to the relatively brief freezing period and the shallow freeze penetration experienced in alpine areas (Costin and Wimbush, 1973). At Tiffindell thufur site 1, ground freezing to a depth of 17-24cm was recorded over the June-July (1996) period, but ground freezing recorded in the valley adjacent to the Tiffindell ski slope (2800m) during the winters of 1995 and 1996 occurred to depths of at least 20cm and probably deeper. This freezing extended from May until early September.

THUFA No.	GRADIENT (degrees)[a]	THUFA HT (cm)	THUFA DIAMETER D_d (cm)	D_c (cm)	RATIO D_d to D_c
1	3	27	72	54	1.3
2	1	24	56	50	1.12
3	3	20	50	31	1.61
4	2	19	49	39	1.26
5	3	18	68	40	1.7
6	2	18	48	38	1.26
7	1	17	41	37	1.11
8	2	17	35	29	1.21
9	1	16	32	30	1.06
10	2	15	43	35	1.23
11	1	14	31	29	1.06
12	3	13	54	32	1.69
13	3	13	90	45	2
14	1	12	30	29	1.03
15*	1	12	34	30	1.13
16*	2	10	40	23	1.74
17	1	10	25	24	1.04
18	1	10	20	20	1
19	1	9	20	19	1.05
20	2	7	18	17	1.06

* = individuals removed for analysis.
a = gradient of surface on which thufa exist
D_d = measured downslope diameter
D_c = measured cross-slope diameter

n = 20

Height : mean 15.05cm range 7-27cm sd = 5.104
Diameter : mean 37.675cm range 17-90cm sd = 15.867

Table 8.1: Individual characteristics for 20 thufur at Site 1 (2916m)

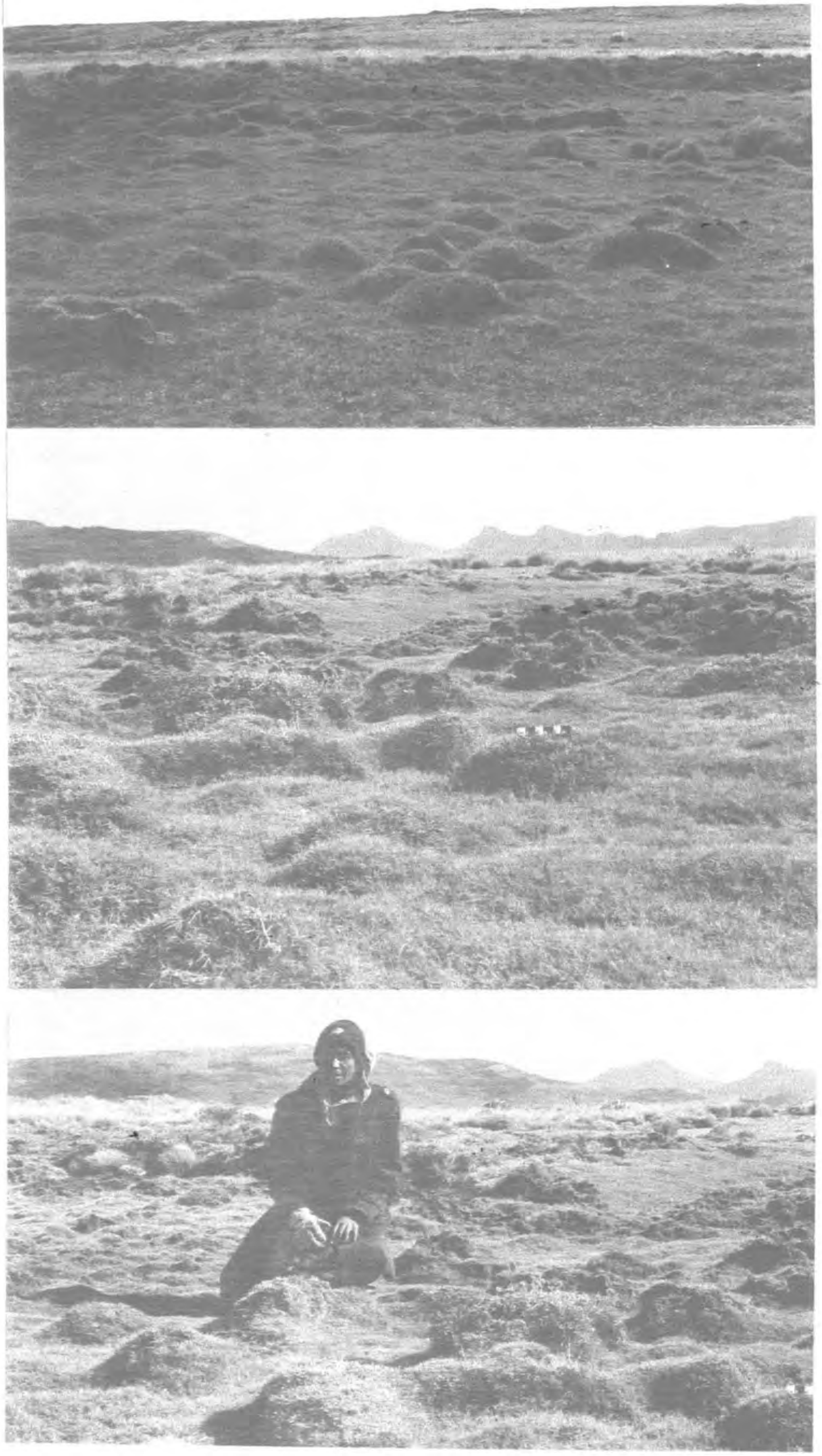


Figure 8.2a,b&C: *Thufur development at an altitude of about 2900m near Ben MacDhui.*

8.4.2.2. Internal characteristics

The internal characteristics of thufur in southern Africa have rarely been subject to intense investigation, the most detailed and valuable data being produced from a study by Grab (1994) of the Mohlesi valley thufur. Therefore, it is this data set against which the Tiffindell thufur will be compared.

On initial discovery of Tiffindell site 1, during the summer of 1995, an individual mound was removed for examination and analytical determination of its internal characteristics (sample A). In winter, 1996, a partly thawed thufa was sectioned to a depth of 30cm below the ground surface, samples being removed at 10cm intervals (sample B). The soil samples were analysed, the moisture, organics and soil contents being determined as a percentage of the mass of the wet sample (Table 8.2). Although only two features have been intensively investigated at this site (and only one to a depth of 30cm), correlations between the findings of these features, as well as correlations with the Mohlesi valley thufur, provide information on the internal characteristics of thufur in this region, which may be applicable to the entire southern African region.

Measurements from samples A and B indicate that most rooting extends down to 2.2cm on the mound apex, and 3.9cm or more on the sides of the mounds. This corresponds closely to the depth of typically unfrozen soil covering the thufa, and thus allows for root penetration. During a warmer period in mid-winter however, this depth of thaw was seen to double at site 1. Grab (1994), working on thufur in the Natal Drakensberg, reported values of 1.8cm on a mound apex and 3cm on the lower sides, but as the depth of thaw is seen to be variable, and largely dependent on the climatic conditions, little information can be acquired from this correlation.

DEPTH (cm)	Wet sample (% mass)			Dry sample (% mass)		% Fines	Soil description
	Moisture	**Organics**	**Soil**	**Organics**	**Soil**		
Thufa top	55.5	8.17	36.33	22.06	77.4	61.74	Leached white in colour (2.5YR 8/2)
Thufa centre	57.4	11.85	30.75	31.99	68.01	62.96	Grey (10YR 7/2), with distorted layers
Ground level	59.7	13.90	26.4	37.53	62.47	56.56	Darker in colour than mound (10YR 5/2)
10cm below	59.4	14.36	26.24	38.77	61.23	47.65	Darker soil, high root content (7.5YR 4/4)
20cm below	60.3	14.02	25.68	37.85	62.15	41.68	Dark brown soil (7.5YR 3/4)
30cm below	61.3	6.12	32.58	16.52	83.48	46.22	Dark soil meeting reddish soil (2.5YR 3/4)
Inter-Hummock depression	64.1	24	11.9	52	48	42.41	Dark brown soil with a high organic content (10YR 2/2)
MEAN	59.67	13.2	27.13	33.82	66.11	51.32	
Std. Deviation	2.75	5.72	7.78	11.73	11.65	8.98	

Note: % ***fines*** includes grain sizes ≤ 0.125mm, that is, fine sands and smaller.

__Table 8.2:__ Internal characteristics for Sample B at Site 1 (2916m), sampled in July 1996.

(adapted from Grab, 1994)

A relatively homogenous fine-grained grey soil (Munsell Soil Colour chart, 10YR 7/2) constitutes the inside of the mound, and similar soil has also been described by Hastenrath and Wilkinson (1973), Hanvey and Marker (1992) and Grab (1994). The colouring indicates that this is a typically gleyed soil, which is characteristic of poorly drained environments under a cold climate (Strahler, 1975). The soil forming the apex and sides of the thufa is almost white in colour (2.5YR 8/2) compared to the mottled grey interior, and as it is readily influenced by climatic conditions, as well as by the vegetation covering the mound, it may be highly leached and lacking organic matter. A cross-section of the mound displays additional distinct colour variations, which may be attributed to the displaced soil horizons due to cryoturbation processes. The soil darkens in colour from light grey (10YR 7/2) within the mound to a brown colour (7.5YR 4/4) as a result of an increasing organic content (Table 8.2.), and at approximately 30cm depth contacts a lower horizon of reddish-brown soil (2.5YR 3/4), which may contain small quantities of iron compounds.

The content of fine-grained soil within the actual thufa is over 60%, which is comparable with that of the Mohlesi valley thufur (Grab, 1994), as well as with results from Arctic and sub-Arctic regions (Tarnocai and Zoltai, 1978). The high percentage of fines diminishes with depth within the soil profile (Table 8.2.). This phenomenon was also noted by Grab (1994). The 5% increase in fines at 30cm depth may be attributed to the fine-textured red soil which forms the border of the following soil horizon. The ground level depressions surrounding the thufur are commonly composed of medium to fine grained sands and tend to be matted together with an abundance of shallow roots. The distribution of moisture showed limited variation (±2.75 s.d.) within the profile of sample B (Table 8.2.), although the average moisture content for the soil of the mound at and above groundlevel is 56.45%, compared to 64.1% for the interhummock depression. As the entire soil profile was saturated at the time of sampling, the differences in percent moisture content could be attributed to the grain size distribution within the profile. Although silt and clay fines typically retain more water, during saturated conditions the soils with greater pore spaces may effectively contain more water, and thus display a greater soil moisture content.

The amount of organic matter was measured at six levels within sample B, as well as at 5cm below the surface of the inter-hummock depression (Table 8.2). Relatively high organic contents were measured throughout the profile. The soil of the depression contained the highest percentage (24% mass of the wet sample). This could be due to the collection of organic matter around the mounds. The percentage of organic matter at ground level, 10cm depth and 20 cm depth were all measured close to 14%, the percentage decreasing to 6.12% at 30cm depth. The sharp decline in organic content displayed at 30cm may be because this sample was taken at the intersection with the differing lower soil horizon. Within the mound itself, organic matter content was measured as 8.17% at the apex and near the sides, the centre of the hummock itself containing an average of 11.85% (Table 8.2).

These values are almost the inverse of the organic contents measured by Grab (1994), his highest values (14%) being measured at the apex, with progressively decreasing values down to 5.7% at 30cm depth; and the inter-hummock depression only measuring an organic content of 9.1% (Table 8.3). This difference could be attributed to the fact that individual thufa contain different amounts of organic material, although analysis of sample A from Tiffindell revealed organic contents of 9.2% and 13.12% for the apex and the centre respectively.

Within the individual hummock that was removed for analysis, a layer of organic material which resembled a palaeosol was identified: it may be possible for this layer to be dated using radiocarbon dating techniques, although that has not yet been done due to financial and temporal constraints.

DEPTH	TIFFINDELL		MOHLESI VALLEY
	Sample A	Sample B	
Thufa apex	9.2%	8.17%	14%
Thufa centre	13.2%	11.85%	
Ground level	*	13.90%	8.7%
10cm below	*	14.36%	
20cm below	*	14.02%	6.4%
30cm below	*	6.12%	5.7%
Inter-hummock depression	*	24%	9.1%

* = No Data

Table 8.3: Table of comparison of % organic content within a thufa at site 1 at Tiffindell (samples A and B) and at Mohlesi Valley (Grab, 1994).

8.4.2.3. Other characteristics of thufur at Site 1

The majority of thufur at Site 1 are completely insulated by a grass covering (Figure 8.2), only about 20% showing evidence of slight splitting, possibly due to frost cracking or internal pressures. Of the 20 individuals inspected, six showed some signs of breakup. Four of the thufur displayed disruption on the windward sides (ie. the north and north-west sides), which were undergoing further breakup due to needle ice activity on the exposed sections. The other two disrupted thufur appeared to have been trampled by animals. Hoofprints up to 8cm deep were observed on the thufur tops and sides, producing disrupted surfaces that were exposed to needle ice activity. Thus wind action, animal interference and needle ice activity are considered to be important erosional agents affecting thufur at Tiffindell. Grab (1994) reported large-scale disruption within the Mohlesi valley thufur, attributing the majority of thufur breakup to frost cracking and upslope disruption processes (section 8.3.3).

The inter-hummock areas of the Tiffindell thufur at Site 1 have also been subjected to disturbance. The initial disturbance was probably a result of trampling by grazing animals, as well as by needle ice activity at these exposed sites. Yet during the summer months these sites tend to recover well as the vegetation is fast-growing and can cover disturbed areas before too much erosion takes place. This 'recovery system' was noted in winter 1996 when, on returning to the site, the area from which the thufa individual had been removed in the summer of 1995 was completely covered with vegetation, and minimal erosion was observed. Several irregular areas at Site 1 may well have been excavated by humans in order to provide clay for building purposes and may therefore not be of natural origin.

8.4.3. Site 2 - 2779m

Several 'thufur-like features' were observed on a section of the old ski slope (Figure 8.1), which is characterised by fine, short grasses and a gradient of 6-9°. These features do not exceed 11cm in height. They are all elongated in the downslope direction. Detailed measurement by Grab (1994) show that hummocks on gentler slopes (1-5°) have greater vertical development than those on steeper slopes (8-9°), indicating that steeper slopes in the alpine regions may impede vertical hummock growth. In fact, Tarnocai and Zoltai (1978) have suggested that vertical development of thufur rarely takes place where the slope gradient is over 7°, but they rather become increasingly elongated downslope on these steeper gradients.

Few other signs of frost activity were visible at Site 2, except for small scale frost heave and needle ice activity in the vegetation-free areas. There is little evidence of abundant soil moisture, and given the relatively low altitude (2779m), as well as the slope gradient, it is presumed that these features are currently inactive, although activity may be resumed under slightly colder and wetter conditions.

8.4.4. Site 3 - 2650m

This site is located near the man-made dams below the resort complex (Figure 8.1), and is characterised by abundant moisture and a gradient of less than 4°. The area supports several hydrophillic plant species such as *Kniphofia thodei* and *Senecio erubescens* and can be classified as a wetland (Olyott, 1996). The 'thufa-like' mounds are covered by *Carex glomerabilis* and do not measure more than 10cm in height. They are associated with needle ice activity, and also exhibit internal ice lens development during cold phases, and are therefore considered to be active on a seasonal basis, depending on the climatic conditions.

8.4.5. Discussion

Evidence of ground instability and frost action include the occurrence of non-sorted polygons (Chapter 7), gaps around stones, frost heave and needle ice activity (Chapter 5). Thufur at Sites 1 and 3 appear to be currently active, although those at Site 2 may be essentially dormant or fossil features.

Zoltai and Pettapiece (1974), working in Canada, also found that in dry years active thufur may show limited or no cryoturbation. However, in years where more moisture is available during the frost periods, they recorded an increase in thufur development. Similarly, thufur development in the Drakensberg and the Lesotho Highlands may be restricted during mild winters, but become active again during exceptionally cold periods (Grab, 1994). The upper limits of thufur activity may be primarily determined by the occurrence of suitably vegetated wetlands. At Tiffindell, as in Mohlesi valley, such wetlands are of limited extent, due to the relatively steep south-facing slopes which support predominantly moister conditions.

Karte (1983) has suggested that thufur activity occurs at MAAT of 3°C or lower. Nevertheless, Koaze *et al.* (1974) have located active thufur in Japan where the MAAT is about 6°C. Although the MAAT at altitudes around 2900m at Tiffindell has not been recorded, from measurements at 2800m (Chapter 5), it is suggested that the MAAT is also approximately 6°C (calculated using the monthly means as described by Miller, 1931). The air temperature data is characterised by both large seasonal variation and winter diurnal frost cycles, and intense frost action is predominantly restricted to the winter months, which have a mean temperature of approximately 0.5°C (Chapter 5). The general absence of cloud cover, and of long-lasting snow cover due to rapid melting rates and strong winds, and the occurrence of subzero temperatures during most nights from May to September (producing a mean minimum air temperature of approximately -3°C during June and July 1996), allows for ground freezing to depths of 0.2m or more which may persist for two or three winter months. This would imply that thufur activity on the higher summits is possible under present climatic conditions, but that it is restricted to the four coldest months, namely May to September.

Although the age of the thufur has not yet been determined, it is desirable that they be carefully monitored on a yearly basis so as to gain a better understanding of local and regional environmental conditions and the way in which these conditions change through time.

8.4.6. Other features

Within the Tiffindell area several unvegetated frost mounds of tabular or domed shape were located where water was plentiful. They were, however, only recorded during periods characterised by low temperatures and/or diurnal freeze-thaw cycles.

A tabular frost mound at an altitude of 2786m had developed upon a terrace tread which was well watered by ice and snowmelt runoff from the surrounding bedrock areas. An exposed section 8.5cm high and 23cm wide revealed a series of crystals up to 6cm in length which had developed to form the mound. As the site was located in the late afternoon, several exposed crystals has already partially melted, but the structure was still robust and able to support a person of 60kg without caving in.

A group of seven unvegetated mounds were located at 2756m in an area fed by a groundwater channel. The mounds ranged between 19x17cm and 25x20cm, the average height being approximately 10cm. The cores were frozen to a depth of approximately 10cm, however, the top few centimetres forming the mound were already melted. One individual removed for grain size determination revealed that an approximate 40% by mass consisted of pebble-sized particles, the majority of the remainder consisting of medium to fine-grained sands and silts.

A single ephemeral frost mound was located at an altitude of 2775m in the vicinity of the 'top dam'. The feature measured 20x15x12cm and was exposed on one side to reveal the interior lens of clean ice. Lewis (1996) has suggested that frost mounds may be the basic groundwork for the formation of more pronounced thufur, or earth hummocks containing ice-bodies. Pollard and van Everdingen (1992) state that frost mounds develop during a single winter season in response to water being forced between soil layers of relatively weak cohesion (otherwise referred to as a water injection). Subsequent freezing results in the formation of a mound on the surface with a core of polycrystalline ice, which as a rule, is very clean and contains little particulate matter (Nelson *et al.*, 1992). The Tiffindell features appear to be similar, yet of a smaller scale, to those described by Pollard and van Everdingen (1992) and Nelson *et al.* (1992).

8.4.7. Conclusion

Thufur have been identified at altitudes above 2900m on slope gradients of *c.*2° in the Tiffindell area. The thufur average 15.05cm in height and 37.675cm in diameter. They are active under present climatic conditions.

STONE LOBES — CHAPTER 9

9.1. INTRODUCTION

Stone lobes are composed of coarse detritus and are amongst the most impressive of all upland periglacial landforms. They commonly occur on debris-mantled slopes where the regolith contains a high concentration of large boulders, and can be described as lobate accumulations of stones and boulders overlying a relatively stone-free subsoil. These features are also referred to as 'stone garlands' (Antevs, 1932, cited in Benedict, 1970), 'boulder lobes' (Ballantyne and Harris, 1994) and 'stone-banked lobes' (Galloway, 1961; Benedict, 1970) in the literature, although the latter are usually associated with the presence of fines (Quinn, 1987).

9.2. STONE LOBE CHARACTERISTICS

9.2.1. Distribution

Stone lobes typically occupy moist snow-accumulation slopes with a slope gradient ranging between 12° and 30° (Benedict, 1970; Ballantyne and Harris, 1994). There is a partial overlap between the slope and exposure requirements of turf-banked lobes and stone lobes, the latter typically occupying the higher and steeper sections of the slopes (Harris, 1981a; Quinn, 1987).

9.2.2. Stone lobe morphology

The fronts of stone lobes are lobate, rocky, and often steep in profile, sloping at angles of 20-50° (Benedict, 1970). The riser heights typically reach 1m or more (Harris, 1981a), and rarely exceed 3m. Variations appear to be related to the size of the component boulders (Galloway, 1961), which show diversity due to the lithology of the parent rock material. Typical mean boulder diameters were recorded as 0.2-0.5m for lobes in the Cairngorms, and greater than 1m on Lochnagar, which resulted in risers up to 6m high occurring on these granite mountains (Ballantyne and Harris, 1994).

Some lobes may support vegetation between the constituent boulders, but more extensive vegetation cover is restricted to those lobes with finer-grained treads, that is, to 'stone-banked lobes'. Surface clasts may also exhibit lichen or moss growth. A lack of signs of disturbance of the lichen cover, or the general occurrence of vegetation within stone lobes may be an indication of inactivity of the features.

Individual lobes display great variation in width and tread length. Widths may vary from as little as 4m to over 20m (Table 11.1 in Ballantyne and Harris, 1994), and treads of up to 60m in length have been reported (Harris, 1981a). The size of lobes may be influenced by the size of the component boulders (Galloway, 1961). The treads of stone lobes are predominantly composed of stones and boulders without visible fine material (Benedict, 1970), this often being due to a downward fining sequence within the lobes (Boelhouwers, 1995). The surface boulders of stone lobes often display a preferred downslope orientation parallel to the gradient, as well as an increase in clast size in the downslope direction (Boelhouwers, 1995).

9.3. STONE LOBE FORMATION

Two main schools of thought exist with regards to the formation of stone lobes:

i) that they originate as stone-fronted gelifluction lobes or terraces, with gelifluction occurring within the fine material behind the stone fronts. These fronts are considered to be due to sorting processes as larger material characteristically moves downslope faster than finer material, and frost action processes are believed to lift the frontal stones into semi-vertical positions (Quinn, 1987; Boelhouwers, 1995).

ii) that they originate as a result of intense frost action and creep, which requires autumn saturation and a temperature regime that encourages deep freezing and thawing (Benedict, 1970).

The different formative processes outlined above may result in features with different attributes. Stone-banked lobes, which are characterised by fines and often have vegetated treads, may result from gelifluction (Galloway, 1961; Boelhouwers, 1994, 1995; Hall, 1981). Under different climatic conditions the fine material within these lobes may be removed by water action, leaving only the lobe of boulders as a remnant (Quinn, 1987; Lewis, 1996). Debris lobes which lack the presence of vegetation and fines, and are commonly known as stone or boulder lobes (Ballantyne and Harris, 1994), may develop as a result of frost action and frost creep. The formation of these lobes does not require a vegetation cover, and is probably favoured by its absence (Benedict, 1970).

Both formative processes rely on the availability of source material and on the presence of sufficient moisture for development and subsequent movement. Movement of these features would be greatest along the axis of the lobe, with maximum values occurring just below the lobe midpoint, as these areas typically display the highest available groundwater levels. Directions of movement within the lobes tend to preserve the surface profile, as flow is generally directed towards the lobe axis where movement is accelerating, and away from the edge where movement is retarded (Benedict, 1970). Therefore, the accumulations of boulders tend to remain in a lobate shape.

The lobate accumulations of boulders found within stream channels are related in origin to other stone lobes, although they probably formed under conditions of total saturation.

9.4. STONE LOBES AT TIFFINDELL

From their topographical position and material composition, as well as the visible longitudinal sorting of the surface material, the features occurring on the Tiffindell slopes are interpreted as stone lobes.

9.4.1. Distribution and morphological characteristics

A series of stone lobes occurs on the south and southeast-facing flanks of the Ben MacDhui - Breslin's Kop ridge at an altitude of between 2730m and 2895m, and form a major part of the Tiffindell landscape (Appendix I). The uppermost lobes are located approximately 50m to 70m below the summit of the ridge. There are at least 30 stone lobes, all the features being characteristically lobate in shape. They are arranged in seven groupings across the slopes, and are distributed below rocky outcrops and headwalls, and often in stream channels or drainageways (Appendix I). Due to their location, composition and lobate shape, the Tiffindell stone lobes are not considered to be a product of *in situ* weathering, as there are no visible large core rocks within the features, nor are there any tor-like features on the slopes within the vicinity of the lobes. Rather, their constituent boulders are considered predominantly cryoclastic in origin, derived from frost shattering upslope of individual lobes.

Twenty-two stone lobes were investigated (Appendix I), their relative slope position, orientation of the lobe long axis, altitude and gradient being recorded, as well as their physical dimensions (tread length and width, and riser height). This data is displayed in Table 9.1.

The tread lengths of the measured lobes range between 3.30m and 45m, the mean tread length 14.94m (±10.84 s.d.). The tread widths show less variation, ranging from 2.05m to 16.90m (mean 7.53m; ±4.17 s.d.). The mean riser height was recorded as 0.68m (±0.32 s.d.) for the 22 lobes, with a range between 0.39m and 1.34m being observed. The long axes of the lobes are orientated between 162° and 240°, parallel to the maximum slope. That is, south to southeast-facing orientations are favoured on gradients of between 15° and 27.5° (Figure 9.1).

Although an equal number of stone lobes was measured for both the south and southeast-facing slopes, more occur on the former slope. Variations in the lobe characteristics of those occurring on the two slopes are summarised in Table 9.2.

LOBE #	ASPECT	ORIENT-ATION[a]	ALTITUDE[b]	GRADIENT[c]	LENGTH (m)	WIDTH (m)	RISER HT (m)	IN CHANNEL?
1	SE-facing	162°	2890m	20°	11.40	9.10	0.44	No
2	SE-facing	174°	2884m	17°	9.80	7.30	0.39	No
3	SE-facing	180°	2874m	15°	12.70	8	0.49	No
4	SE-facing	170°	2831m	26°	9	4.60	0.47	No
5	SE-facing	163°	2827m	25°	7.40	5.60	0.57	No
6*	SE-facing	170°	2813m	27.5°	10.40	6	0.59	No
7	SE-facing	185°	2807m	24°	7.90	7.60	0.48	No
8	SE-facing	160°	2797m	22°	19.40	13.10	1.3	No
9	SE-facing	192°	2843m	20°	19.50	13.60	0.64	Yes
10	SE-facing	215°	2798-2782m	20°	± 34	± 11.50	1.34	Yes
11	SE-facing	212°	2774m	19°	9.80	5.40	0.49	Yes
12	S-facing	165°	2817m	24°	3.50	2.10	0.44	Yes
13	S-facing	160°	2810m	27°	4.60	2.45	0.39	Yes
14	S-facing	225°	2800m	27°	3.30	2.05	0.42	Yes
15	S-facing	220°	2790m	25°	4.60	3.55	0.47	Yes
16	S-facing	185°	2780m	24°	On ski slope - Two stone lobes removed			No
17	S-facing	225°	2858-2844m	22°	45	5.5	1.22	No
18	S-facing	235°	2855m	20°	27.70	7.80	0.79	No
19	S-facing	232°	2850m	19.5°	15.30	10.60	0.59	No

LOBE #	ASPECT	ORIENT-ATION$_a$	ALTITUDE^b	GRADIENT$_c$	LENGTH (m)	WIDTH (m)	RISER HT (m)	IN CHANNEL?
20	S-facing	233°	2840m	21°	21.60	12.10	0.84	No
21	S-facing	216°	2785m	17°	25.50	16.90	1.29	No
22	S-facing	240°	2735m	22°	11.40	4.30	0.69	Yes

a = Magnetic North bearing of lobe long axis
b = Altitude of the middle of the lobe (except lobes 10 and 17 which show the full altitude variation)
c = Gradient of the actual slope, not of the surface of the lobe.
* = Lobe painted with cross-lines

Table 9.1: Characteristics of some of the stone lobes (1-22) at Tiffindell.

LOCATION	ALTITUDE RANGE (m)	GRADIENT RANGE (°)	RISER HEIGHT (m)				TREAD LENGTH (m)				LOBE WIDTH (m)			
			min	mean	max	sd	min	mean	max	sd	min	mean	max	sd
S-facing slope	2875-2730m	17° - 27°	0.39	0.71	1.29	0.33	3.30	16.3	45.0	13.7	2.05	6.74	16.9	5.02
SE facing slope	2895-2775m	15° - 27.5°	0.39	0.65	1.34	0.34	7.40	13.7	34.0	7.88	4.60	8.35	13.6	3.13

Table 9.2: Summary of the characteristics of the Tiffindell stone lobes.

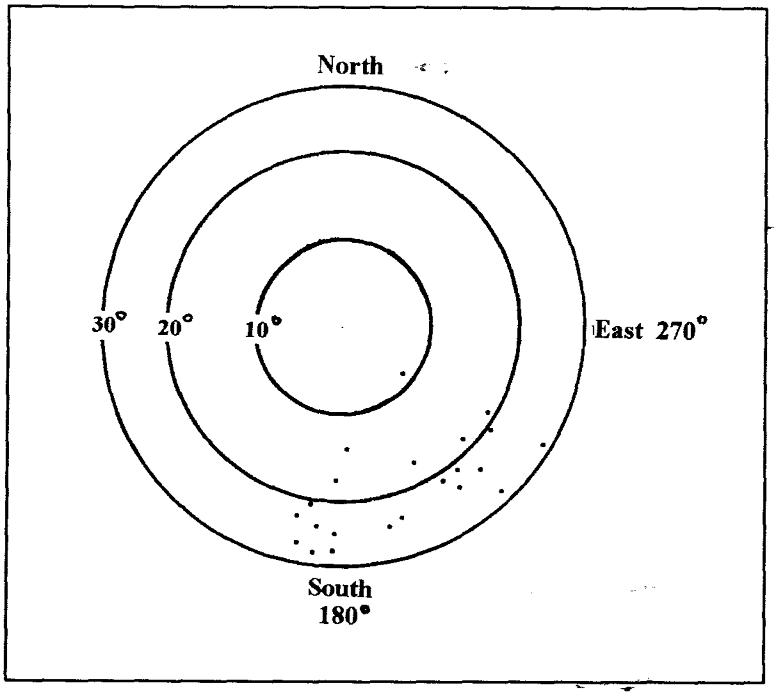


Figure 9.1: Scatter diagram showing the slope angle and the downslope orientation of the long-axes of 22 stone lobes at Tiffindell. South to southeast-facing orientations are clearly favoured.

The lobes on the south-facing slope occur over 145m elevation, extending down to 2730m, whereas those on the southeast-facing slope only occur over 120m elevation and do not extend beyond 2775m in altitude. Despite this difference in altitude, there is no great variation between the gradient range of the slopes (mean gradients 23° and 21.4° respectively), and both slopes exhibit small and large forms (Table 9.1). The south-facing slope, in general, shows a larger scope of lobe sizes, which range from a minimum tread length of 3.30m to a maximum of 45m, and a lobe width from 2.05m to a maximum of 16.9m. The mean riser height is also greater in comparison to the southeast-facing slope (0.71m compared to 0.65m), although the treads on that slope appear to be generally wider.

Profiles taken parallel to the maximum slope showed that the lobes have convex longitudinal profiles (Figure 9.3). They are composed of loose scree which is considered to be derived from the basaltic cliffs at the head of the valley, the boulders being typically angular, not often exceeding 1.5m in long axis length, and appear to have undergone frost shattering processes, albeit in the past. They display little interstitial fine material at the surface, except that from recent mechanical weathering.

The longitudinal profile data reveal the existence of differing riser heights, and tread lengths and widths for stone lobes at different altitudes, and angle of slope (Table 9.3).

		LENGTH (m) mean	LENGTH (m) s.d.	WIDTH (m) mean	WIDTH (m) s.d.	RISER HT (m) mean	RISER HT (m) s.d.
Altitude	2900m-2800m	14.7	11	6.96	3.48	0.60	0.23
	2800m-2700m	17.45	10.98	9.125	5.48	0.93	0.42
Gradient	≤ 20°	18.4	8.81	10.02	3.56	0.72	0.36
	> 20°	12.34	11.84	5.74	3.62	0.66	0.31

Table 9.3: Stone lobe profile data from Tiffindell with regards to variation in altitude and gradient

As evidenced in Table 9.3, those lobes developed at lower altitudes (2800m - 2700m) display a mean tread length almost 3m greater than those developed at higher elevation, as well as a 2m greater tread width. The corresponding lobe risers are also on average 0.33m higher. Stone lobes occurring on gradients less than or equal to 20° can be significantly longer in tread length than those on steeper gradients (mean length 6.06m greater than lobes on slopes >20°), whereas the riser heights show less variation between slope gradients. Those lobes on less steep gradients are on average almost double the width of those features occurring on steeper slopes.

Several of the stone lobes with relatively longer treads are found to occur on less steep gradients within stream channels (Table 9.1), and display evidence of two or more lobes 'running-in' to one another (for example lobes 10 and 22). This may be a result of a reduction in movement of the originally lower lobe due to a decrease in gradient, thus allowing lobes from above to coalesce, essentially forming one lobe of significant length.

9.4.2. Composition of the stone lobes

From investigation of 7 stone lobes, being 3 disturbed stone lobes from the ski slope, as well as 4 intact lobes elsewhere, their basic composition was determined. Figure 9.2 shows the stratigraphy of lobe 19 (Appendix V). The 3 other lobes examined were composed of similar sediments. The tread of the Tiffindell stone lobes is composed of a layer of cobbles and boulders (unit I), without interstitial fines, to a depth of 60cm to 90cm. Beneath this surface layer is a thinner and finer textured unit (unit II) with abundant small stones (67% particles ≥2mm) and an organic content of 16%. This unit extends approximately 18cm, and grades into a unit of dark brown (10 YR 3/3) soil with a higher percentage of fines (52% ≤ 125µm), yet only 8.75% organic content (unit III). This unit contains more than 22% silt, and is consequently frost susceptible. The scarcity of coarse material in unit III is, therefore, a reflection of vertical sorting processes, which carry stones upwards through this unit and add them to the base of the rubble layer.

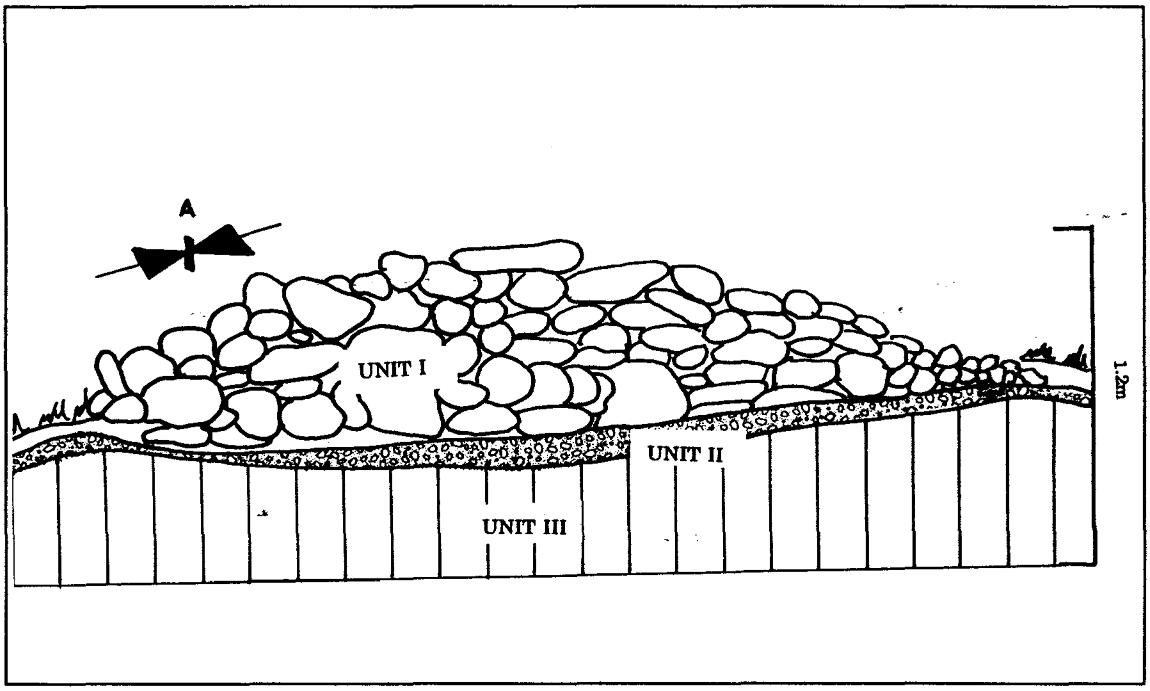


Figure 9.2: Cross-section through a stone lobe at Tiffindell (2850m), including profile characteristics and stone orientation data (A)

The lobe fronts are often composed of slightly larger blocks than the remainder of the lobe (Figure 9.3; Figure 9.4), and thus display a coarsening of the surface material towards the riser. Boulders were sampled at the upper end, in the centre and close to the riser of lobe 8 to establish whether sorting had actually occurred downslope, and boulder sample histograms were drawn to indicate the percentage of the boulder lengths at each position respectively (Figure 9.5). The material at the rear (upper end) of the tread is a mixture of coarse soil and rock, and appears to be 'trailing' behind the rest of the lobe in two tails extending from the outer edges of the lobe itself (Figure 9.4). This lateral sorting is also well developed on the sides of several of the lobe individuals.

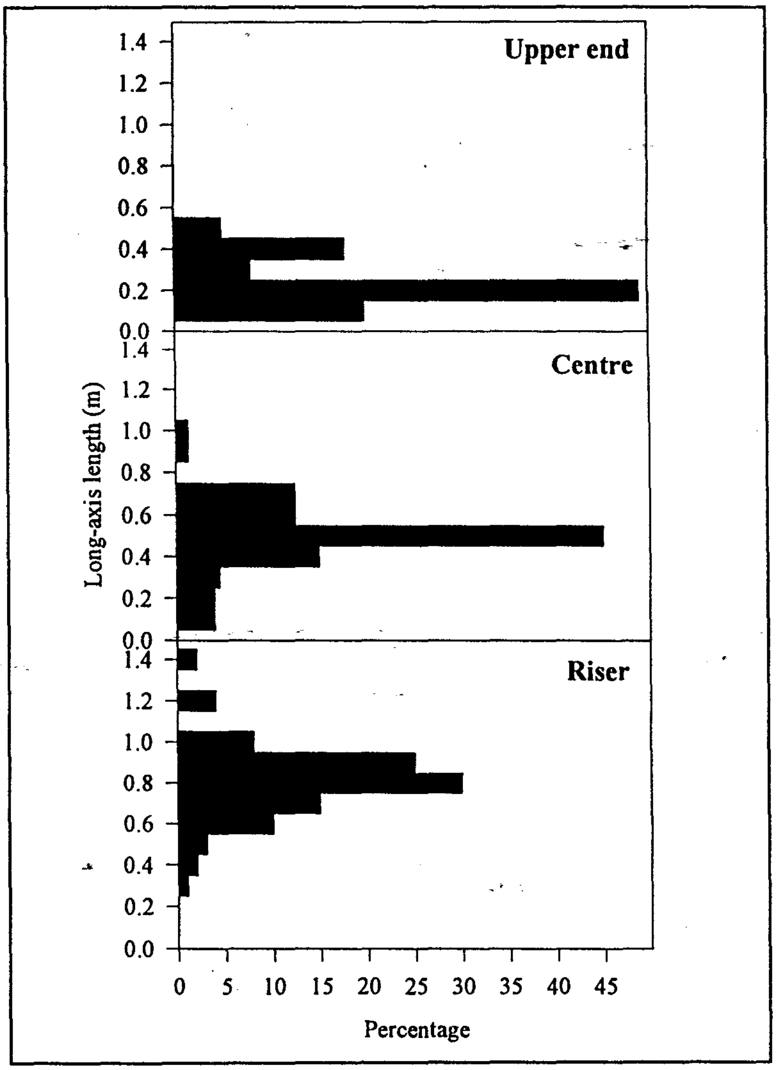
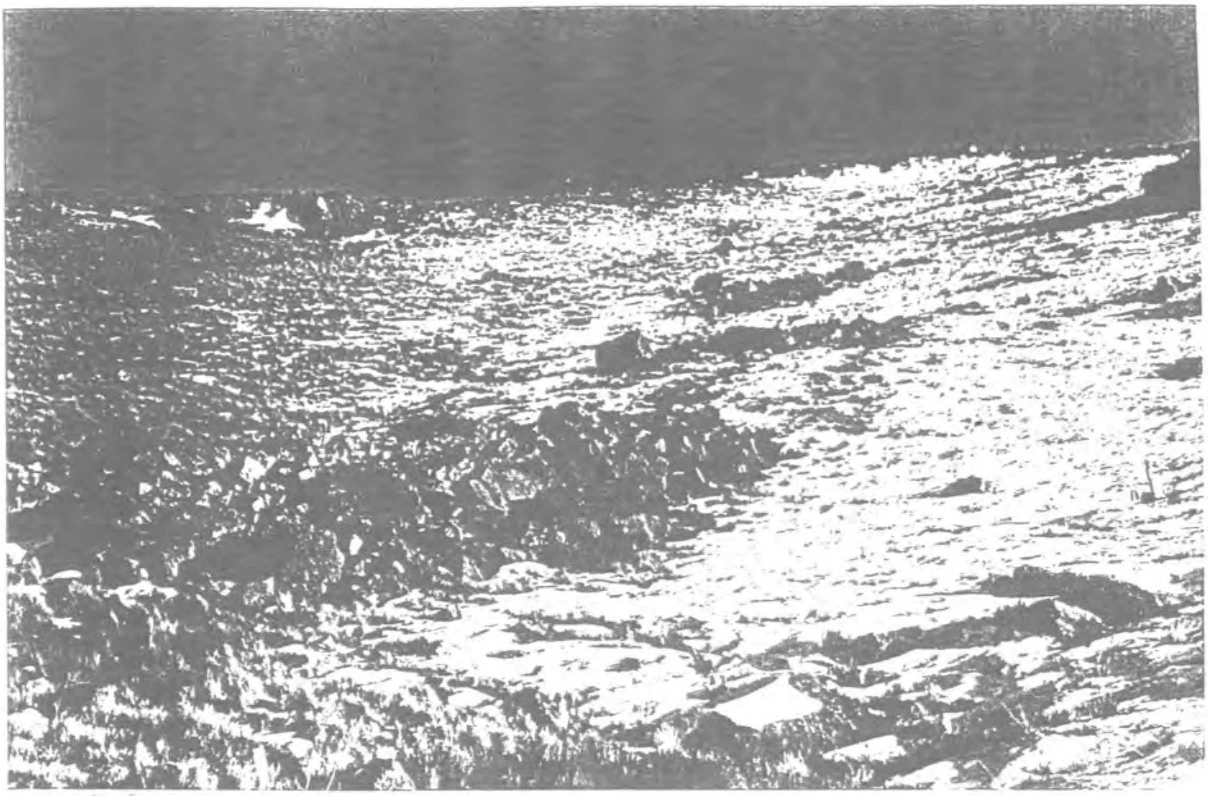


Figure 9.5: Boulder sample histograms for lobe 8 showing the percentage boulder long-axis length at the upper end, centre and riser, respectively.



C.A. Lewis

Figure 9.3: Stone lobes at altitudes above 2800m on the south-facing slope at Tiffindell Ski Resort. The lobe in the foreground is 12m wide, with a front that varies from 1.6 to 2.1m in height.

Figure 9.4: A large stone lobe (lobe 20) 21.6m in length at an altitude of 2840m on the south-facing slope at Tiffindell. Note the trailing stones at the upper end of the lobe (indicated by arrows).

The component boulders display a downslope orientation of the their long axes, and their flat faces tend to lie parallel to the lobe surface. The fabric data (*n*=25; Appendix V) for the boulders of one of the larger stone lobes (lobe 19) indicates this preferred downslope orientation parallel to the gradient of the slope (see A in Figure 9.2). At the riser of several individual lobes, the long axes of the boulders were rotated into orientations paralleling the fronts (that is, dipping into the lobe) and therefore a slight imbrication was observed within the boulders of the risers. This imbrication may indicate retarded clast movement. Similar observations have also been made by Boelhouwers for lobes in the Natal Drakensberg (1994) as well as for those in the Hexriver Mountains, Western Cape (1995).

Lobes with distinctly large boulders forming their fronts and occurring within their treads showed a tendency to be of greater overall size. Therefore, a relationship between the size of the lobes and the size of the component boulders may exist, as suggested by Galloway (1961).

9.4.3. Conclusion: Formation of the stone lobes

After detailed investigations of the Tiffindell stone lobes, the following conclusions with respect to their original formation could be drawn:

a) source of the component boulders was a rock outcrop or scattered boulders on the slope above the lobe;
b) debris within each lobe has been sorted;
c) a mechanism for downslope movement must exist (or must have existed).

The manner of transport and deposition of the stone lobes at Tiffindell is debateable. Initially, lobe formation may have been the result of the gathering of boulders within a depression, such as a drainage line, forming lobate features. As the flanks of the Tiffindell valley are not uniform, slope irregularities result in divergent "flowpaths" of boulders moving downslope, and the collection of boulders in channels and drainageways incised into the slope. This is evidenced by their forms and positions today. This phenomenon was also observed by Marker (1986) for what she describes as scree tongues on Elandsberg in the uplands of the Amatola mountains.

No evidence for existing downslope movement of the boulder accumulations was observed at Tiffindell in 1995 or 1996. These features are considered to be fossil under present climatic conditions. However, under more severe conditions (for example, during the Last Glacial Maximum) these sites would have been subject to greater periglacial conditions than at present, and with the presence of interstitial fines or ice, a downslope 'flowing' movement of these accumulations as lobes could be expected. A former flow mechanism is presently evidenced by:

a) the lobate form of the features;
b) the variable tread lengths and riser heights, which may be attributed to the successive overriding of lower lobes in the sequence by faster flowing lobes from above; and
c) the presence of imbricated blocks which appear to 'emerge' from the lobe fronts, which are suggestive of a flow mechanism.

An increase in precipitation levels may have resulted in the eluviation of fines from the Tiffindell stone lobes, and possibly in the retardation of the movement of these features.

The Tiffindell stone lobe features presumably developed under colder climatic conditions than those of the present. Movement of the block debris may have been due to frost creep, and the movement of the adjacent fine-textured material by solifluction or gelifluction. Stone lobes in Colorado investigated by Benedict (1970) are characteristic of a dominant frost creep environment, with solifluction occurring within the fines.

As frost creep is most effective where vegetation is absent and moisture levels high (Chapter 6), the preferred southern aspect of the features may indicate an origin related to frost and frost cycles. At present, autumn saturation is mainly limited to those areas on shaded slopes that are irrigated with water from runoff drainageways. Stone lobe formation outside these saturated areas must have occurred at a time in which isolated snowbanks were present, which supplied the upper slopes with moisture, causing saturation and hence conditions conducive for the development of lobes.

Movement of the stone lobes was probably terminated by a change to warmer, and possibly drier, climatic conditions (Deacon and Lancaster, 1988; Partridge *et al.*, 1990). Localised retardation resulted from a decrease in gradient causing thickening and lateral spreading of the lobe forms.

Despite present-day differences in appearance, and perhaps past differences in formation, in terms of size and form the stone lobes at Tiffindell can be compared to 'stone-banked lobes' recorded by Galloway (1961), Boelhouwers (1994, 1995), Dardis and Granger (1986) and Hall (1981) (although these features are largely associated with the presence of fines), and stone lobes investigated by Quinn (1987; Table 9.4). This comparison is illustrated in Table 9.4. The Tiffindell stone lobes occur at lower altitudes than those of the Natal Drakensberg described by Boelhouwers (1994) and Dardis and Granger (1986), yet are developed on a steeper gradient and are generally of smaller dimensions. The lobes described from Scotland by Galloway (1961) are of similar size, but also occur on slopes of lesser angle. The dimensions of the Tiffindell stone lobes are smaller than those of the south west of Ireland described by Quinn (1987), which may be anticipated as the Irish features supposedly formed from several adjacent, individual lobes coalescing, and are thus of a larger terrace-like nature. The lobes described from the Hexriver Mountains (Boelhouwers, 1995) and Marion Island (Hall, 1981) are found at much lower altitudes than those occurring at Tiffindell, and are largely composed of fines and display significantly smaller dimensions.

LOCATION	ALTITUDE	MEAN LENGTH (m)	MEAN WIDTH (m)	MEAN RISER HEIGHT (m)	MEAN GRADIENT (°)
Tiffindell - E. Cape Drakensberg; South Africa	2895m - 2730m	14.94 (range 3.30-45)	7.53 (range 2.05-16.9)	0.68 (range 0.39-1.34)	22.22° (range 15-27.5°)
Giant's Castle - Natal Drakensberg; South Africa (Boelhouwers, 1994)	3200m	± 19 (range 8-30)	± 4.5 (range 2-7)	± 1.65 (range 0.3-3)	± 16° (range 14-18°)
Natal Drakensberg - South Africa (Dardis and Granger, 1986)	3000m - 3120m	(range 10-60)	± 4 (range 2-6)	± 0.35 (range 0.2-0.5)	no value given
Hexriver Mountains - Western Cape; South Africa (Boelhouwers, 1995)	1800m - 1850m	1.49	1.93	0.20	16°
Marion Island (Hall, 1981)	70m - 450m	0.93	no value given	0.32	13°
Knocknadobar - south west Ireland (Quinn, 1987)	603m - 633m	22.31	no value given	5.63	8.24°
Scotland (Galloway, 1961)	1200-4000ft	± 16.8 (range 3.6-30)	± 6	± 1.8 (range 0.66-3)	± 14° (range 8-20°)

Table 9.4: Comparison of the dimensions of the Tiffindell stone lobes with other similar reported features.

9.4.4. Movement within the Tiffindell stone lobes

The growth of vegetation around and within the stone lobes, as well as the occurrence of lichen of considerable sizes, suggest that the features are inactive (fossil) under the present climatic conditions. Lines were painted across lobe 6 and were used to estimate the relative disturbance or movement of the component boulders. The majority of the painted rocks remained stationary for the two winter period (1995 - 1996), although a few showed slight downslope motion, possibly under the influence of gravity. Only one was displaced significantly, and its movement was possibly the result of disturbance by a rhebok or by a person walking across the lobe, and was not considered as a modern minor remobilisation of the surface blocks of the lobes.

Mean daily air temperatures in the vicinity of the stone lobes range from -3°C in July to 16.3°C in January (in 1996; Chapter 5). Such a regime is not consistent with the formation or maintenance of such periglacial forms (Washburn, 1973; Williams, 1961; French, 1976; Hall, 1981). Therefore it is concluded that the formation of the stone lobes dates to a former colder climatic stage extant within the Pleistocene. The stone lobes are essentially inactive at present.

9.4.5. Conclusion

Stone lobes of periglacial origin up to 16.90m wide, 1.34m high and 45m long, exist at altitudes above 2750m at Tiffindell. The lobes are concentrated along drainage lines and are essentially formed of debris derived from rock outcrops upslope of the lobes. The lobes are inactive at present and are therefore fossil elements of the landscape.

PERIGLACIAL CONDITIONS AND FEATURES OF THE TIFFINDELL AREA

CHAPTER 10

SUMMARY AND IMPLICATIONS FOR ECONOMIC DEVELOPMENT

10.1. ZONATION OF PROCESSES AND FORMS

Three main zones can be recognised at Tiffindell on the basis of altitude, topography and periglacial features, as shown on Figure 10.1: the lower region at and below the level of the resort (zone I), a 'midzone' between the resort level and the summit area (zone II), and the summit itself (zone III; Table 10.1).

Zone I extends from the plateau at 2500m, downslope of the ski resort, up to approximately 2750m. This area is characterised by well vegetated gentler slopes than those at and above the altitude of the ski runs. This is where the drainageway channels coalesce and have been dammed to provide water for use by the resort for snowmaking and for residential consumption. The zone supports a few terracettes and non-sorted polygons, and is subject to diurnal needle ice activity.

Zone II occurs between altitudes of *c.*2750m and *c.*2900m. This zone is characterised by a series of turf-banked gelifluction lobes and terracettes, as well as by stone lobes. The zone supports a vegetation cover of 60% (or more), although the treads of the gelifluction lobes are characteristically unvegetated. Frost heave and non-sorted polygon development occurs within these bare areas, as well as needle ice activity. Air and soil temperatures were recorded within this zone, and support the physical evidence that frost action processes are active, especially during the winter months.

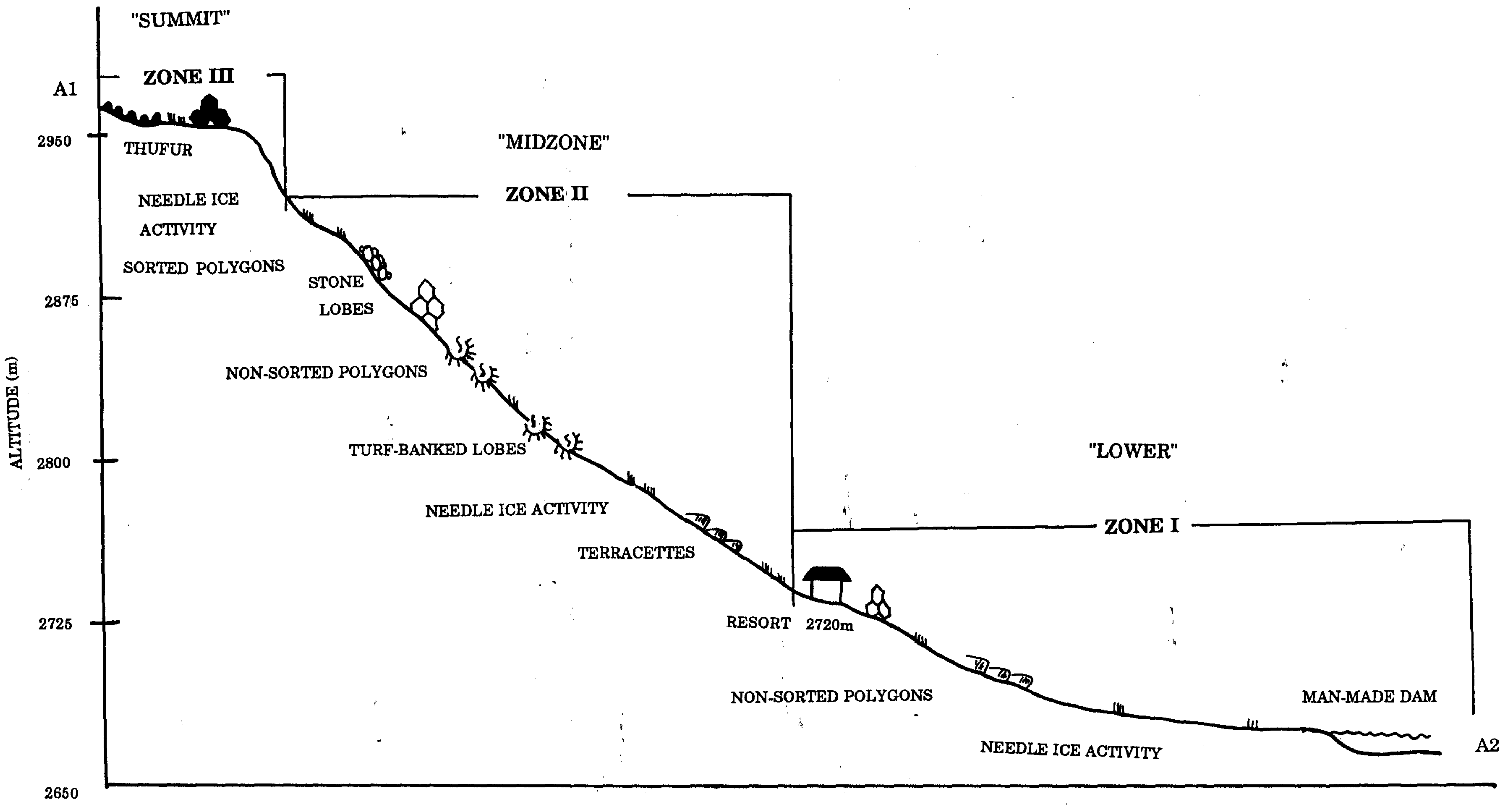


Figure 10.1: *Schematic representation of the three zones identified on the Tiffindell slope A1-A2 according to topography, altitude and processes.*

The upper zone (zone III; 2900m to 3001.2m) consists of a broad and gently sloping summit plateau, which acts as a waterdivide. The summit area is separated from the lower zone by steep headwall outcrops at approximately 2900m. The area is characterised by sparse vegetation cover, some of the ground being completely bare. The bare areas are characterised by frost action, frost heave, needle ice activity and the formation of (micro) patterned ground phenomena. Thufur occur where the ground is grass covered and particularly moist.

ZONE	ALTITUDE	VEGETATION	SLOPE	DOMINANT FEATURES
I	<2500m - 2750m	Well vegetated	Moderate (10-15°)	Terracettes; non-sorted polygons; needle ice activity
II	2750m - 2900m	Moderate vegetation (60%)	Steep (15-28°)	Terracettes; turf-banked gelifluction lobes; stone lobes; non-sorted polygons; needle ice activity
III	2900m - 3001.2m	Sparse vegetation	Gentle slope (<10°)	Sorted and non-sorted polygons; thufur; needle ice activity

Table 10.1: Zonation of the Tiffindell area by elevation, topography and dominant periglacial features.

10.2. ECONOMIC SIGNIFICANCE AND IMPLICATIONS FOR TIFFINDELL SKI RESORT

The climatic data (Figures 5.3a to 5.6b) clearly show that low temperatures may be expected at Tiffindell for at least three consecutive months per annum. Minimum air temperatures in 1995 and 1996 were below freezing on 45% and 72% of the days from the beginning of June to the end of August respectively, while ground temperatures (which are critical for the survival of snow cover) were at or below freezing on approximately 40% and 81% of the days respectively during the same period. On the basis of the climatic data recorded in 1995 and 1996, it is likely that (provided sufficient artificial snow is produced) the ski run and nursery slopes at Tiffindell Ski resort could be snow covered for at least three months per annum (*i.e.* early June to late August). This study did not investigate whether sufficient water is available for adequate snow making at the resort.

The rapid downhill movement of gelifluction material measured over 1995 and 1996 at Tiffindell should be considered in any future development of the ski resort, or any other developments above the plateau lying below the resort. Attempts in 1994 to erect a small building adjacent to the ski-run (at an altitude of 2783m), to be used as a pub (the 'Ice Pub'; Appendix I), were nullified when moving gelifluction deposits caused the upslope wall of the building to collapse in 1995 (Figure 10.2).

Resort buildings, which were erected in 1993 on stilts driven deep into the regolith, have shown no downslope displacement to the end of 1996. Balconies of the sleeping chalets, however, which were erected on foundations cut into the regolith, had warped before the end of 1995. Displacements of the balconies away from the chalets was measured as being 3cm or more (M.Olckers, pers.comm.). Frost heave and the development of thick ice lenses also distorts the ground surface, and has affected prefabricated buildings resting directly on the ground, making it difficult (and sometimes impossible) to open or shut doors. Future construction plans should take cognizance of the special problems created by the periglacial environment at Tiffindell and the advice of periglacial geomorphologists should be sought before further development takes place (*cf.* Harris, 1987b). Care should be taken to ensure that water is not discharged where it may exacerbate the development of ice lenses and/or stimulate increased rates of gelifluction.

Run-off from the ski slope, which is artificially provided with snow as often as the climatic and water source conditions will allow (Figure 10.3), should be led away from the resort buildings otherwise gelifluction may be facilitated on the slope on which the buildings exist. In 1996, Tiffindell experienced at least 1000 possible snow-making hours (I.van Eck, pers.comm.) and especially with warmer conditions prevailing during the days, man-made as well as natural snow is a major potential source of run-off. In order to minimise the deleterious effects of such gelifluction as is inevitable at Tiffindell, it is recommended that buildings of significant value be erected on stilts which are anchored on bedrock, or driven deeply into the regolith.

With large amounts of water being added to the ski slope in the form of man-made snow, the possibility of loss of cohesion of the regolith which may cause rapid soil flow, slumping, or even landslides, should be considered. These features occur "primarily as a result of shear failure at the boundaries of the moving mass" (Brunsden, 1971, in Gardiner and Dackombe, 1983) and include both sliding and flowing motions. This form of slope failure is normally related to extremely wet conditions (Crozier, 1986, cited in Illgner, 1995). The importance of creating and maintaining a vegetation cover at the resort, in order to stabilise the surficial sediments as much as possible, is stressed. The use of Kentucky Bluegrass on the ski slope in order to bind the soil is supported, although it is preferable that indigenous grass of equal or better soil-binding qualities should be utilised (if such grass exists).

10.3. Conclusion

This research project provides new information regarding periglacial features in southern Africa, and contributes towards improving knowledge of the periglacial environment in the East Cape Drakensberg. This study provides an assessment of the processes occurring in the vicinity of Tiffindell Ski, as well as locates, identifies and describes the periglacial features associated with those processes. The implications of the study for the development of Tiffindell Ski resort have been presented. Further geomorphological research, especially relating to the access roads to the resort, is desirable, since little is known in southern Africa of the effects of periglacial environments on roads and road construction.

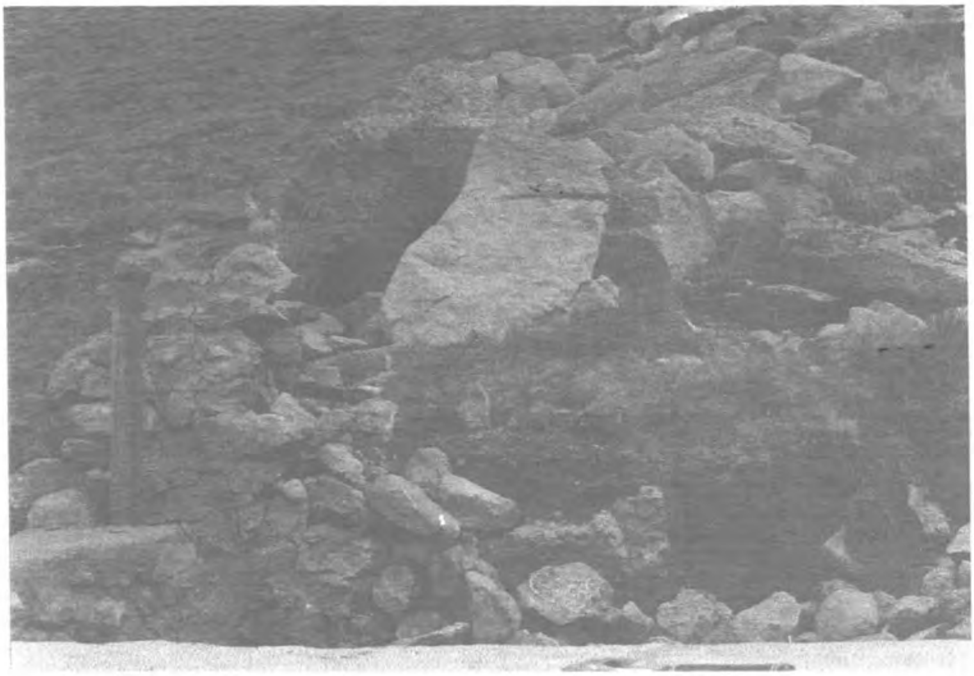


Figure 10.2: *Collapse of the upslope wall of the "Ice Pub" (2783m altitude) due to moving gelifluction deposits.*

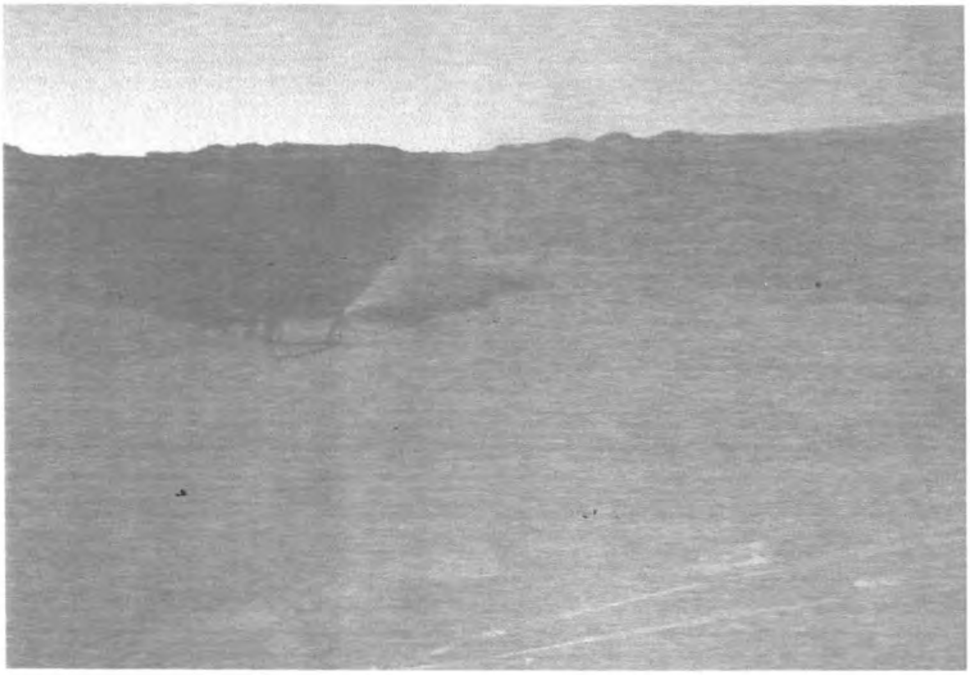


Figure 10.3: *Artificial snow production using a "snow gun" at Tiffindell Ski on the 31 May 1995. Snow is made by spraying water droplets, seeded with ice nuclei, into the air during cold periods (air temperatures <-2°C).*

LITERATURE CITED

ANDERSSON, J.G. 1906 Solifluction, a component of subaerial denudation. *J.Geol.* 14:91-112

BALLANTYNE, C.K. 1987 The present-day periglaciation of upland Britain. *In* Boardman, J (ed) *Periglacial processes and landforms in Britain and Ireland.* University Press:Cambridge, 113-125

BALLANTYNE, C.K. and HARRIS, C. 1994 *The periglaciation of Great Britain.* University Press:Cambridge.

BALLANTYNE, C.K. and MATTHEWS, J.A. 1983 Desiccation cracking and sorted polygon development, Jotunheimen, Norway. *Arctic and Alpine Research,* 15:339-349

BARRIE LOW, A.B. and REBELO, A.G. (eds.) 1996 *Vegetation of South Africa, Lesotho and Swaziland.* Department of Environmental Affairs and Tourism, Pretoria

BENEDICT, J.B. 1970 Downslope soil movement in a Colorado alpine region: rates, processes, and climatic significance. *Arctic and Alpine Research,* 2(3):165-226

BENEDICT, J.B. 1976 Frost creep and gelifluction features: a review. *Quaternary Research,* 6:55-76

BESKOW, G. 1935 Tjälbildningen och Tjällyftningen. *Sveriges Geol.,* Under 26. No.375 Series C, Stockholm (20 page English summary)

BOELHOUWERS, J. 1991a Present day periglacial activity in the Natal Drakensberg, Southern Africa: a short review. *Permafrost and Periglacial Processes,* 2(1):5-12

BOELHOUWERS, J. 1991b Periglacial evidence from the Western Cape Mountains, South Africa: a progress report. *Permafrost and Periglacial Processes,* 2(1):13-20

BOELHOUWERS, J. 1994 Periglacial landforms at Giants Castle, Natal Drakensberg, South Africa. *Permafrost and Periglacial Processes,* 5:129-136

BOELHOUWERS, J.C. 1995 Present day soil frost activity at the Hexriver Mountains, Western Cape, South Africa. *Z.Geomorph.N.F.,* 39:237-248

BOELHOUWERS, J. and HALL, K. 1990 Soil frost phenomena at Sani Pass. *In* Hanvey, P.M. (ed) *Field guide to geocryological features in the Drakensberg*. Grahamstown, Organising Committee, UNESCO/IGCP Project Number 297 pp 62-64

BORCHERT, G. and SÄNGER, H. 1981 Research findings of a Pleistocene glaciation of the Cape mountain-ridge in South Africa. *Zeitschrift für Geomorphologie*, 25:222-224

BRITISH STANDARDS. 1975 *Methods of test for soils for civil engineering purposes.* BS 1377, British Standards Institute: London

BUCKLE, C. 1990 *Landforms in Africa: an introduction to geomorphology*. Longman:Hong Kong

BUCKMAN, H.O. and BRADY, N.C. 1969 *The nature and properties of soils*. 7th ed. MacMillian: USA

BURN, C.R. 1991 Permafrost and ground ice conditions reported during recent geotechnical investigations in the Mayo District, Yukon Territory. *Permafrost and Periglacial Processes*, 2(3):259-268

CAINE, N. 1981 A source of bias in rates of surface soil movement as estimated from marked perticles. *Earth surface processes and landforms*, 6:69-75

CARSON, M.A. and KIRKBY, M.J. 1972 *Hillslope form and process*. Cambridge University Press: Cambridge, 475pp.

CHELCHINSKEY, C. 1996 *The taxonomy and morphological adaptations of Alti Mountain Grassland plants*, unpublished project, Rhodes University, Grahamstown

CORTÉ, A.E. 1966 Particle sorting by repeated freezing and thawing. *Biul.Perygl.*,15:175-240

CORTÉ, A.E. and HALL, K. 1991 Introduction: geocryology of the Americas - IGCP Project No 297. *Permafrost and Periglacial Processes*, 2:3

COSTIN, A.B. and WIMBUSH, D.J. 1973 Frost cracks and earth hummocks at Kosciusko, Snowy Mountains, Australia. *Arctic and Alpine Research*, 5:111-120

COX, N.J. 1981 Hillslope profile determination and analysis. *In* Goudie, A. (ed) *Geomorphological Techniques*. George Allen and Unwin, 62-66

CSIR ENVIRONTEK 1996 Meteorological data recorded at Tiffindell Ski.

DARDIS, G.F. and GRANGER, J.E. 1986 Contemporary periglacial phenomena in the Natal Drakensberg, South Africa. *Palaeocol.Afr.*, 17:89-93

DAVIES, J.L. 1969 *Landforms of cold climates*, Cambridge, Mass., MIT Press, 200pp

DEACON, H.J. 1983 Another look at the Pleistocene climates of South Africa. *S.Afr.J.Sci.*, 79:325-328

DEACON, J. and LANCASTER, N. 1988 *Late Quaternary Palaeoenvironments of Southern Africa*. Claredon Press:Oxford, 225pp

DENT, M.C., LYNCH, S.D. and SCHULZE, R.E. 1989 *Mapping mean annual rainfall statistics in southern Africa*. Department of Agricultural Engineering, University of Natal. ACRU Report No.27

EGGINTON, P.A. and DYKE, L.D. 1990 Apparent hydraulic conductivities associated with thawing, frost-susceptible soils. *Permafrost and Periglacial Processes*, 1:69-78

EYLES, N., EYLES, C.H. and MIALL, A.D. 1983 Lithofacies types and veryical profile models; an alternative approach to the description and environmental interpretaion of glcial diamict and diamictite sequences. *Sedimentology*, 30:393-410

FAHEY, B.D. and LEFEBURE, T.H. 1988 The freeze-thaw weathering regime at a section of the Niagara Escarpment on the Bruce Peninsula, southern Ontario, Canada. *Earth surface processes and landforms*, 13:293-304

FITZPATRICK, R.W. 1978 Periglacial soils with fossil permafrost horizons in southern Africa. *Ann.Natl.Mus.*, 23:475-484

FLÖHN, H. 1984 Climatic evolution in the Southern Hemisphere and the equatorial region during the Late Cenozoic. *In* Vogel, J.C. (ed.). Late Cainozoic Palaeoclimates of the Southern Hemisphere. A.A. Balkema, Rotterdam, 5-20

FRENCH, H.M. 1976 *The Periglacial Environment.* Longman:London

FRENCH, H.M. 1981 Periglacial geomorphology and permafrost. Progress in Physical Geography, 5:267-273

FRENCH, H.M. 1988 Active layer processes. *In* Clark, M.J. (ed) Advances in Periglacial Geomorphology. John Wiley and sons, 151-178

GALE, S.J. and HOARE, P.G. 1991 Quaternary sediments. Belhaven Press, New York, 323pp.

GALLOWAY, R.W. 1961 Solifluction in Scotland. Scottish Geographical Magazine, 77:75-87

GARDINER, V. and DACKOMBE, R. 1983 Geomorphological Field Manual, George Allen and Unwin: London, 254pp.

GERRARD, A.J. 1981 Soils and landforms: an integration of geomorphology and pedology. George Allen and Unwin, 49-60

GERRARD, J. 1992 The nature and geomorphological relationships of earth hummocks (Thufa) in Iceland. *Zeitschrift für Geomorphologie*, suppl Bd 86:173-182

GOLDTHWAITE, R.P. 1076 Frost sorted patterned ground: a review. Quat.Res., 6:27-35

GOUDIE, A.S. 1993 *The Nature of the Environment* (3rd Edition) Blackwell: Britain, 27-67, 86-114, 276-278

GRAB, S.W. 1994 Thufur in the Mohlesi Valley, Lesotho, Southern Africa. Permafrost and Periglacial Processes, 5:111-118

GRAB, S.W. 1996 A note on the morphology of miniature sorted stripes at Mafadi Summit, High Drakensberg. S.Afr.Geogr.J., 78(1):59-63

HAEBERLI, W. 1992 Construction, environmental problems and natural hazards in periglacial mountain belts. Permafrost and Periglacial Processes, 3(2):111-124

HALL, K. 1980 Freeze-thaw activity at a nivation site in northern Norway. Arctic and Alpine research, 12(2):183-194

HALL, K. 1981 Observations on the stone-banked lobes of Marion Island, S.A.J.Sci., 77:129-131

HALL, K. 1991a The significance of periglacial geomorphology in southern Africa: a discussion. South African Geographer, 18(1/2):134-137

HALL, K. 1991b The allocation of the freeze-thaw weathering mechanism in géocryological studies: a critical comment. S.Afr.Geog.J., 73:10-13

HALL, K. 1991c Rock moisture data from the Juneau Icefield (Alaska) and its significance for mechanical weathering studies. Permafrost and Periglacial Processes, 2:321-330

HALL, K. 1992 A discussion of the need for greater rigour in southern African cryogenic studies. S.Afr.Geog.J., 74(2):69-71

HALL, K. and OTTA, W. 1990 Observations regarding biological weathering on nunataks of the Juneau Icefield, Alaska. *Permafrost and Periglacial Processes*, 1:189-196

HANVEY, P.M., LEWIS, C.A. and LEWIS, G.E. 1986 Periglacial slope deposits in Carlisle's Hoek, near Rhodes, Eastern Cape Province. *S.Afr.Geo.Journal*, 68:164-174

HANVEY, P.M. and LEWIS, C.A. 1990 A preliminary report on the age and significance of Quaternary lacustrine deposits at Birnam, north-east Cape Province. *S.Afr.J.Sci.*, 86:271-273

HANVEY, P.M. and LEWIS, C.A. 1991 Sedimentology and genesis of slope deposits at Sonskyn, East Cape Drakensberg, South Africa. *Permafrost and Periglacial Processes*, 2:31-38

HANVEY, P.M. and MARKER, M.E. 1992 Present-day periglacial microforms in the Lesotho Highlands: implications for present and past climatic conditions. *Permafrost and Periglacial Processes*, 3:353-361

HARPER, G. 1969 Periglacial evidence in southern Africa during the Pleistocene epoch. *Palaeoecology of Africa*, 4:71-101

HARRIS, C. 1972 Processes of soil movement in turf-banked solifluction lobes, Okstindan, northern Norway. *In* Prive, R.J. and Sugden, D.E. (compilers) *Polar Geomorphology*, Institute of British Geographers Special Publication 4, London: Institute of British geographers, 155-174

HARRIS, C. 1973 Some factors affecting the rates and processes of periglacial mass movements. *Geogr.Annlr*, 55A(1):24-28

HARRIS, C. 1981a *Periglacial mass-wasting: a review of research.* British Geomorphological Research Group, Research Monograph. GeoAbstracts:Norwich, 4:204pp

HARRIS, C. 1981b Microstructures in solifluction sediments from south Wales and north Norway. *Biuletyn Peryglacjalny*, 28:221-226

HARRIS, C. 1987a Solifluction and related periglacial deposits in England and Wales. *In* Boardman, J (ed) *Periglacial processes and landforms in Britain and Ireland.* University Press:Cambridge, 209-223

HARRIS, C. 1987b Mechanisms of mass movement in periglacial environments. *In* Anderson, M.G. and Richards, K.S. (eds), *Slope Stability* John Wiley and sons, 531-559

HARRIS, C. 1988 The alpine periglacial zone. *In* Clarke, M.J. (ed) *Advances in periglacial geomorphology*. Wiley: Chichester, 369-413

HARRIS, S.A., FRENCH, H.M., HEGINBOTTOM, J.A., JOHNSTON, G.H. LADANYI, B., SEGO, D.C. and VAN EVERDINGEN, R.O. 1988 *Glossary of permafrost and related ground-ice terms.* National Research Council of Canada, Technical Memorandum 142, Ottawa

HASTENRATH, S and WILKINSON, J. 1972 A contribution to the periglacial morphology of Lesotho, Southern Africa. *Biuletyn Peryglacjalny*, 22:157-168

HILL, T.R. 1992 *Contemporay pollen spectra from the Natal Drakensberg and their relation to associated vegetation communities*, unpublished Ph.D. thesis, Rhodes University, Grahamstown, 288pp

HILLIARD, O.M. AND BURTT, B.L. 1987 *The botony of the southern Natal Draken.berg.* Vol 15. National Botanical Gardens. CTP Book Printers: Cape, 253pp

HOPKINS, D.M. and SIGAFOOS, R.S. 1954 Role of frost thrusting in the formation of tussocks. *Am.J.Sci*, 252:55-59

ILLGNER, P.M. 1995 *The morphology and sedimentology of two unconsolidated Quaternary debris slope deposits in the Alexandria District, Cape Province*, unpublished M.Sc. thesis, Rhodes University, Grahamstown

JAMES, P.A. 1971 The measurement of soil frost-heave in the field. *British Geomorphological Research Group Technical Bulletin*, 8, 43pp.

KARLSTROM, E.T. 1990 Relict periglacial features east of Waterton-Glacier Parks, Alberta and Montana, and their palaeoclimatic significance. *Permafrost and Periglacial Processes*, 1(3):221-234

KARTE, J. 1983 Periglacial phenomena and their significance as climatic and edaphic factors. *GeoJournal*, 7:329-340

KÉZDI, A. 1980 *Soil Testing: Handbook of soil mechanics.* Volume 2. Elsevier Scientific Publishing Company:Amsterdam

KILLICK, D.J.B. 1963 An account of the plant ecology of the Cathedral Peak area of the Natal Drakensberg. *Botanical Survey of South Africa*, Memoir no. 34, 156pp

KING, L. 1990 Soil and rock temperatures in discontinuous permafrost: Gornergrat and Unterrothorn, Wallis, Swiss Alps. *Permafrost and Periglacial Processes*, 1:177-188

KING, L., GORGUNOV, A.P. and EVIN, M. 1992 Prospecting and mapping of mountain permafrost and associated phenomena. *Permafrost and Periglacial Processes*, 3:73-81

KIRKBY, M.J. 1967 Measurement and theory of soil creep. *J.Geol.*, 75:359-378

KIRKBY, M.J. 1995 A model for variations in gelifluction rates with temperature and topography: implications for global change. *Geogr.Annlr.*, 77A(4):269-278

KOAZE, T., NOGAMI, M. and IWATA, S. 1974 Palaeoclimatic significance of fossil periglacial phenomena in Hokkaido, Northern Japan. *Quat.Res.*, 12:177-191

KOSTER, EA. 1993 Introduction - present global change and permafrost, within the framework of the International Geosphere-Biosphere Programme. *Permafrost and Periglacial Processes*, 4:95-98

KRANTZ, W.B. 1990 Self-organisation manifest as patterned ground in recurrently frozen soils. *Earth Sci.Rev.*, 29:117-130

LAUTRIDOU, J-P. and SEPPÄLÄ, M. 1986 Experimental frost shattering of some precambrian rocks, Finland. *Geogr.Annlr.*, 68A (1-2): 89-100

LAWLER, D.M. 1986 Bank erosion and frost action: an example from South Wales. *In* Gardiner, V. (ed.), *International Geomorphology 1986.* Part I, Chichester: Wiley, 575-590

LAWLER, D.M. 1988 Environmental limits of needle ice: a global survey. *Arctic and Alpine Research*, 20(2):137-159

KOSTER, EA. 1993 Introduction - present global change and permafrost, within the framework of the International Geosphere-Biosphere Programme. *Permafrost and Periglacial Processes*, 4:95-98

KRANTZ, W.B. 1990 Self-organisation manifest as patterned ground in recurrently frozen soils. *Earth Sci.Rev.*, 29:117-130

LAUTRIDOU, J-P. and SEPPÄLÄ, M. 1986 Experimental frost shattering of some precambrian rocks, Finland. *Geogr.Annlr.*, 68A (1-2): 89-100

LAWLER, D.M. 1986 Bank erosion and frost action: an example from South Wales. *In* Gardiner, V. (ed.), *International Geomorphology 1986*. Part I, Chichester: Wiley, 575-590

LAWLER, D.M. 1988 Environmental limits of needle ice: a global survey. *Arctic and Alpine Research*, 20(2):137-159

LEWIS, C.A. 1966 *The periglacial landforms of the Brecon Beacons*. unpubl. Ph.D. thesis, University College, Dublin:Ireland

LEWIS, C.A. 1977 Ice-wedge casts in North East County Wicklow, *Scientific proceedings of the Royal Dublin Society*,6(3):17-35

LEWIS, C.A. 1985 Periglacial features. *In The Quaternary history of Ireland*. Edwards, K.J. and Warren, W.P. (eds.) Academic Press: London, p95-113

LEWIS, C.A. 1988a Periglacial landforms. *In* Moon, B.P. and Dardis, G.F. (eds), *The geomorphology of Southern Africa*. Southern Book Publishers:Johannesburg, 103-119

LEWIS, C.A. 1988b Periglacial features in southern Africa: a review. *Palaeoecology of Africa*, 19:357-370

LEWIS, C.A. 1990 Introduction. *In* Hanvey, P.M. (ed) *Field guide to geocryological features in the Drakensberg*. Grahamstown, Organising Committee, UNESCO/IGCP Project number 297 pp1-7

LEWIS, C.A. 1994a *Field guide to the Quaternary glacial, periglacial and colluvial features of the East Cape Drakensberg*, South African Society for Quaternary Research, Rhodes University, Grahamstown.

LEWIS, C.A. 1994b Protalus ramparts and the altitude of the local equilibrium line during the last glacial stage in Bokspruit, East Cape Drakensberg, South Africa. *Geogr.Annlr.*, 76A (1-2):37-48

LEWIS, C.A. 1996 Periglacial features. *In* Lewis, C.A. (ed) *The geomorphology of the Eastern Cape, South Africa*. Grocott and Sherry: Grahamstown, 120-134

LEWIS, C.A. and DARDIS, G.F. 1985 Periglacial ice-wedge cast and head deposits at Dynevor Park, Barkly Pass area, north-eastern Cape Province. *S.Afr.J.Sci.*, 81:673-677

LEWIS, C.A. and HANVEY, P.M. 1988 Sedimentology of debris slope accumulations at Rhodes, Eastern Cape Drakensberg, South Africa. *In* Dardis, G.F.and Moon, B.P. (eds) *Geomorphological Studies in Southern Africa*. Balkema:Rotterdam, 365-381

LEWIS, C.A and HANVEY, P.M. 1991 Quaternary fan and river terrace deposits, Glen Orchy, East Cape Drakensberg, South Africa. *Permafrost and Periglacial Processes*, 2:39-48

LEWIS, C.A and HANVEY, P.M. 1993 The remains of rock glaciers in Bottelnek, East Cape Drakensberg, South Africa. *Trans.Roy.Soc.S.Afr.*, 48(2):265-289

LEWIS, C.A., HILL, T.R. and PAPALOIZOU, G. 1996 The palynology of interstadial deposits of the Rhodian (Last Glacial) stage, East Cape Drakensberg, South Africa. In prepartion.

LEWIS, C.A. and LASS, G.M. 1965 The drift terraces of Slaettaratindur, The Faeroes. *Geogr.J.*, 131(2):247-253

LEWKOWICZ. A.G. 1988 Slope processes. *In* Clark, M.J. (ed) *Advances in periglacial geomorphology*. John Wiley and sons p325-360

LEWKOWICZ, A.G. 1992 A solifluction meter for permafrost sites. *Permafrost and Periglacial Processes*, 3:11-18

LOCK, B.E., PAVERD, A.L. and BRODERICK, T.J. 1974 Stratigraphy of the Karoo volcanic rocks of the Barkly East District. *Geol.Soc.S.Afr.Trans.*, 77:117-129

LORENZO, J.L. 1969 Minor periglacial phenomena among the high volcanoes of Mexico. *In* Péwé, T.L. (ed.), *The Periglacial Environment*. McGill-Queen's University Press: Montreal, 161-175

LOZIŃSKI, W. 1910 Die periglaziale Fazies der mechanischen Verwitterung *11th Internat.Geol.Congress, Stockholm, compte rendu*, 1039-1053

LUNDQVIST, J. 1969 Earth and ice mounds: A terminological discussion. *In* Péwé, T.L. (ed), *The Periglacial Environment*. McGill-Queen's University Press: Montreal, 203-215

MAHANEY, W.C. 1988 Holocene glaciations and paleoclimate of Mount Kenya and other East African mountains. *Quaternary Science Reviews*, 7:211-225

MARK, A.F. 1990 Ecological and nature conservation values: the case for a conservation park. *In* Kearsley, G. and Fitzharris, B. (eds) *Southern Landscapes*. University of Otago

MARK, A.F. 1994 Patterned ground activity in a Southern New Zealand High-Alpine Cushionfield. *Arctic and Alpine Research*, 26(3):270-280

MARK, A.F. and BLISS, L.C. 1970 The high-alpine vegetation of central Otago, New Zealand. *N.Z.J.Bot.*, 8:381-451

MARKER, M.E. 1986 Pleistocene evidence for the Eastern Cape, South Africa: the Amatola scree tongues. *In* Gardiner, V. (ed), *International Geomorphology*. Part II, John Wiley and Sons: New York, 901-913

MARKER, M.E. 1989 Periglacial geomorphology at Golden Gate Highlands National Park: a note on its fieldwork potential. *S.A.Geographer*, 16(1/2):147-153

MARKER, M.E. 1990a Discussion of periglacial geomorphology at Golden Gate Highlands National Park, Orange Free State, South Africa. *S.A.Geographer*, 17:130-131

MARKER, M.E. 1990b Amatola scree tongues. *In* Hanvey, P.M. (ed) *Field guide to geocryological features in the Drakensberg*. Grahamstown, Organising Committee, UNESCO/IGCP Project Number 297 pp 11

MARKER, M.E. 1994 Dating of valley fills at Golden Gate Highlands National Park. *Suid-Afrikaanse Tydskrif vir Wetenskap*, 90:361-363

MARKER, M.E. 1995 Futher data for a Pleistocene periglacial gradient in southern Africa. *Trans.Roy.Soc.S.Afr.*, 50(1):49-58

MARKER, M.E. and WHITTINGTON, G. 1971 Observations on some valley forms and deposits in the Sani Pass area, Lesotho. *S.Afr.Geog.J.*, 53:96-99

MATSUOKA, N. 1990 The rate of bedrock weathering by frost action: field measurements and a predictive model. *Earth surface processes and landforms*, 15:73-90

MATTHEWS, J.A., HARRIS. C. and BALLANTYNE. C.K. 1986 Studies on a gelifluction lobe, Jotunheimen, Norway: ^{14}C chronology, stratigraphy, sedimentology and palaeoenvironment. *Geogr.Annlr.*, 68A(4):345-360

McGREAL, W.S. 1981 Bulk fabric and structure. *In* Goudie, A.S. (ed) *Geomorphological Techniques* Allen and Unwin:London, p99-100

MC SYSTEMS. 1990a *MCS 151/2/3 temperature sensor - Users Manual*. MC Systems, Cape Town, 14pp

MC SYSTEMS. 1990b *MCS 159 soil moisture sensor - Users Manual*. MC Systems, Cape Town, 15pp

MC SYSTEMS. 1990c *MCS 120-02 data logger - Users Manual*. MC Systems, Cape Town

MILLER, A.A. 1931 *Climatology.* Jarrold and sons, p8-9

MEENTEMEYER, V. and ZIPPIN, J. 1981 Soil moisture and texture controls of selected parameters of needle ice growth. *Earth Surface Processes and Landforms*, 6:113-125

NELSON, F.E., HINKEL, K.M. and OUTCALT, S.I. 1992 Palsa-scale frost mounds. *In* Dixon, J.C. and Abrahams, A.D. (eds.), *Periglacial Geomorphology*. John Wiley and sons, 305-325

NICHOLSON, F.H. 1976 Patterned ground formation and description as suggested by low arctic and subarctic examples. *Arctic and Alpine Research*, 8(4):329-342

NICOL, I.G. 1973 Land forms in the Little Caledon Valley, Orange Free State. *South African Geographical Journal*, 55:56-68

NYBERG, R. 1993 Freeze-thaw activity and some of its geomorphic implications in the Abisko Mountains, Swedish Lappland. *Permafrost and Periglacial Processes*, 4:37-47

OLYOTT, L.J.H. 1996 *Aquatic plants of the Eastern Cape*. unpublished project, Rhodes University, Grahamstown

ORWIN, J.F. 1993 *The periglacial geomorphology of the St Mary's Range, North Otago*. Unpubl. MSc thesis, University of Canterbury, New Zealand.

OUTCALT, S.I. 1969 Weather and diurnal frozen soil structure at Charlottesville, Virginia. *Water Resources Research*, 5:1377-1382

OUTCALT, S.I. 1971 An algorithm for needle ice growth. *Water Resources Research*, 7:394-400

PARTRIGDE, T.C., AVERY, D.M., BOTHA, G.A., BRINK, J.S., DEACON, J., HEBERT, R.S., MAUD, R.R., SCHOLTZ, A., SCOTT, L., TALMA, A.S., VOGEL, J.C. 1990 Late Pleistocene and Holocene climatic change in southern Africa. *S.Afr.J.Sci.*, 86:302-306

PELTIER, L.C. 1950 The geographic cycle in periglacial regions as it is related to climatic geomorphology. *Annals of the Association of American Geographers*, 40:214-236

PÉREZ, F.L. 1996 Cryptogamic soil buds in the Equatorial Andes of Venezuela. *Permafrost and Periglacial Processes*, 7(3):229-255

PISSART, A. 1964 Vitesse des mouvements du sol au Chambeyron (Basses Alpes). *Biuletyn Peryglacjalny*, 14:303-309

POLLARD, W.H. 1988 Seasonal frost mounds. *In* Clarke, M.J. (ed.) *Advances in periglacial geomorphology*. John Wiley and sons: Great Britain, 201-229

POLLARD, W.H. and VAN EVERDINGEN, R.O. 1992 Formation of seasonal ice bodies. *In* Dixon, J.C. and Abrahams, A.D. (eds.), *Periglacial Geomorphology*. John Wiley and sons, 281-304

PRICE, L.W. 1974 The developmental cycle of solifluction lobes. *Annals of the Association of American Geographers*, 64(3):430-438

PRICE, L.W. 1991 Subsurface movement on solifluction slopes in the Ruby Range, Yukon Territory, Canada - A 20 year study. *Arctic and Alpine Research*, 23:200-205

QUINN, I. 1975 *Glacial and periglacial features: NW Iveragh, County Kerry*. unpublished MA thesis, U.C.Dublin.

QUINN, I.M. 1987 The significance of periglacial features on Knocknadobar, South West Ireland. *In* Boardman, J (ed), *Periglacial processes and landforms in Britain and Ireland*. University Press:Cambridge, 287-294

RAY, R.J., KRANTZ, W.B., CAINE, T.N. and GUNN, R.D. 1983 A model for sorted patterned ground regularity. *Journal of Glaciology*, 29(102):317-337

ROMANOVSKIJ, N.N. 1973 Regularities in formation of frost-fissures and development of frost-fissure polygons. *Biul. Peryglac.*, 23:237-277

ROSEN, D. and LEWIS, C.A. 1996 An analysis of Holocene peat deposits at Tiffindell, East Cape Drakensberg, South Africa. In preperation.

RYDÉN, B.E. 1986 Winter soil moisture regime monitored by the time domain reflectometry technique (TDR) *Geogr. Annlr.*, 68A(3):175-184

SÄNGER, H. 1988 Recent periglacial morphodynamics and Pleistocene glaciation of the Western Cape folded belt, South Africa. *In* Dardis, G.F and Moon, B.P. (eds) *Geomorphological Studies in Southern Africa*. Balkema:Rotterdam, p383-388

SAUNDERS, I. and YOUNG, A. 1983 Rates of surface processes on slopes, slope retreat and denudation. *Earth surface processes and landforms*, 8:473-501

SCHMITZ, G. and ROOYANI, F. 1987 *Lesotho: Geology, Geomorphology, Soils*. National University of Lesotho, 191pp.

SCHULZE, B.R. 1947 Climates of South Africa according to the classification of Köppen and Thornthwaite. *S.Afr. Geogr. J.*, 29:32-42

SCHULZE, B.R. 1984 Climate of South Africa: Part 8: General Survey, *WB28, South African Weather Bureau*, 5th ed.

SCHUNKE, E and ZOLTAI, S.C. 1988 Earth Hummocks (Thufur) *In* Clark, M.J. (ed) *Advances in periglacial geomorphology*. John Wiley and sons: Great Britain, 231-244

SCIENCE APPLICATIONS[cc] 1993 *Viability study for a snow ski resort development in the Eastern Cape*. Unpubl. Report

SCOTT, L., STEENKAMP, M. and BEAUMONT, P.B. 1995 Palaeoenvironmental conditions in South Africa at the Pleistocene-Holocene Transition. *Quaternary Science Reviews*, 14:937-947

SELBY, M.J. 1982 *Hillslope materials and processes.* Oxford University Press

SIDLE, R.C., PEARCE, A.J. and O'LOUGHLIN, C.L. 1986 *Hillslope stability and land use*, Water resources monograph series 11, American geophysical union:Washington D.C. 122pp.

SMITH, D.J. 1992 Long-term rates of contemporary solifluction in the Canadian Rocky Mountains. *In* Dixon, J.C. and Abrahams, A.D. (eds) *Periglacial Geomorphology.* John Wiley and sons, 203-221

SOLOMATIN, V.I. and XU, X. 1994 Water migration and ice segregation in the transition zone between thawed and frozen soil. *Permafrost and Periglacial Processes,* 5:185-190

SOUTH AFRICAN COMMITTEE FOR STRATIGRAPHY (SACS) 1980 Stratigraphy of South Africa. Part 1 (Comp. L.E. Kent) *Lithostratigraphy of the Republic of South Africa, South West Africa/Namibia, and the Republics of Bophuthatswana, Transkei and Venda.* Handb. geol. Surv. S. Afr., 8

SPARROW, G.W.A. 1967 Pleistocene periglacial topography in Southern Africa. *J.Glaciol.*, 6:551-559

SPARROW, G.W.A. 1971 Some Pleistocene studies in Southern Africa. *S.A.Geog.*, 3:809-814

STRAHLER, A.N. 1975 *Physical Geography (4th Ed).* John Wiley and sons: New York, 643pp

STRöMQVIST, L. 1983 Gelifluction and surface wash, their importance and interaction on a periglacial slope. *Geogr.Annlr.*, 65(A):245-254

TABER, S. 1930 The mechanics of frost heaving. *J.Geol.*, 38:303-317

TABER, S. 1952 Geology, soil mechanics and botony. *Science*, 115:713-714

TANKARD, A.J., JACKSON, M.P.A., ERIKSON, K.A., HOBDAY, B.K., HUNTER, D.R. and MINTER, W.E.L. 1982 *Crustal evolution of Southern Africa: 3.8 billion years of earth history.* Springer-Verlag p400

TARNOCAI, C. AND ZOLTAI, S.C. 1978 Earth hummocks of the Canadian Arctic and sub-Arctic. *Arctic and Alpine Research*, 10:581-594

THORARINSSON, S. 1951 Notes on patterned ground in Iceland, with particular reference to the Icelandic 'flás'. *Geogr.Annlr.*, 33:144-156

THORN, C.E. 1988 Nivation: a geomorphic chimera. *In* Clarke, M.J. (ed) *Advances in periglacial geomorphology.* John Wiley: Chichester, pp3-31.

THORN, C.E 1992 Periglacial geomorphology: what, where, when? *In* Dixon, J.C. and Abrahams, A.D. (eds) *Periglacial Geomorphology.* John Wiley and sons, 1-30

TRICART, J. 1969 *Geomorphology of cold environments.* (Translated by E. Watson), Macmillian: London, 320pp

TYSON, P.D. 1990 Modelling climatic change in southern Africa: a review of available methods. *S.Afr.J.Sci.*, 86:318-330

VAN VLIET-LANOë, B. 1988 The significance of cryoturbation phenomena in environmental reconstruction. *Journal of Quaternary Science*, 3(1):85-96

VAN ZINDEREN BAKKER, E.M. 1965 Über Moorvegetation und den Aufbau der Moore in Süd-und Ostafrika. *Bot.Jahrb.* 84:215-231

VAN ZINDEREN BAKKER, E.M. 1989 Middle Stone Age palaeoenvironments at Florisbad (South Africa). *Palaeoecol.Afr.*, 20:133-154

VAN ZINDEREN BAKKER, E.M. and WERGER, M.J. 1974 Environment, vegetation and phytogeography of the high-altitude bogs of Lesotho. *Vegetatio*, 29:37-49

VERSTER, E., VAN ROOYEN, T.H. and LIEBENBERG, E.C. 1985 Relations of terracette morphology with soil properties and slope angle. *S.A.Geographer*, 13(2):113-120

VERSTER, E. and VAN ROOYEN, T.H. 1988 Measurement of soil movement on two hillslopes displaying terracettes in humid South Africa. *In* Dardis, G.F. and Moon, B.F. (eds) *Geomorphological studies in Southern Africa*. Balkema: Rotterdam, p311-320

VICKERS, B. 1978 *Soil Mechanics: Laboratory work in civil engineering*. Granada Publishing:Britain, 1-34

WARBURTON, J. 1990 Secondary sorting of patterned ground. *Permafrost and Periglacial Processes*, 1:313-318

WASHBURN, A.L. 1956 Classification of patterned ground and review of suggested origins. *Bulletin Geological Society of America*, 67:823-856

WASHBURN, A.L. 1958 Instrumentation for mass-wasting and patterned ground studies in Northeast Greenland. *Biuletyn Peryglacjalny*, 5:59-64

WASHBURN, A.L. 1969 Weathering, frost action, and patterned ground in the Mesters Vig District, Northeast Greenland. *Meddelelser om Gronland*, 176. 303pp.

WASHBURN, A.L. 1973 *Periglacial processes and environments*. Edward Arnold: London

WATSON, E. 1969 The slope deposits in the Nant Iago valley near Cader Idris, Wales. *Biuletyn Peryglacalny*, 18:95-113

WATSON, H.K. 1988 Terracettes in the Natal Drakensberg, South Africa. *In* Dardis, G.F. and Moon, B.F. (eds) *Geomorphological studies in Southern Africa*. Balkema: Rotterdam, p299-310

WILLIAMS, P.J. 1959 An investigation into processes occuring in solifluction. *Am.J.Sci.*, 257:481-590

WILLIAMS, P.J. 1961 Climatic factors controlling the distribution of certain frozen ground phenomena. *Geogr.Annlr.*, 43:339-348

WILLIAMS, P.J. 1962 Quantitative investigations of soil movement in frozen ground phenomena. *Biuletyn Peryglacjalny*, 11:353-360

WILLIAMS, J. 1978 A brief comparison of model simulations of glacial period maximum atmospheric circulation. *Palaeogeography, Palaeoclimatology and Palaeoecology*, 25:191-198

WILLIAMS, P.J. 1982 *The surface of the Earth: an introduction to geotechnical science*. Longman:London

WILLIAMS, P.J. and SMITH, M.W. 1989 *The Frozen Earth: Fundamentals of geocryology*. University press:Cambridge

WILSON, P. 1992 Small-scale patterned ground, Comeragh Mountains, Southeast Ireland. *Permafrost and Periglacial Processes*, 3:63-70

WILSON, P. 1995 Forms of unusual patterned ground : examples from the Falkland Islands, South Atlantic. *Geogr. Annlr.* 77A (3):159-165

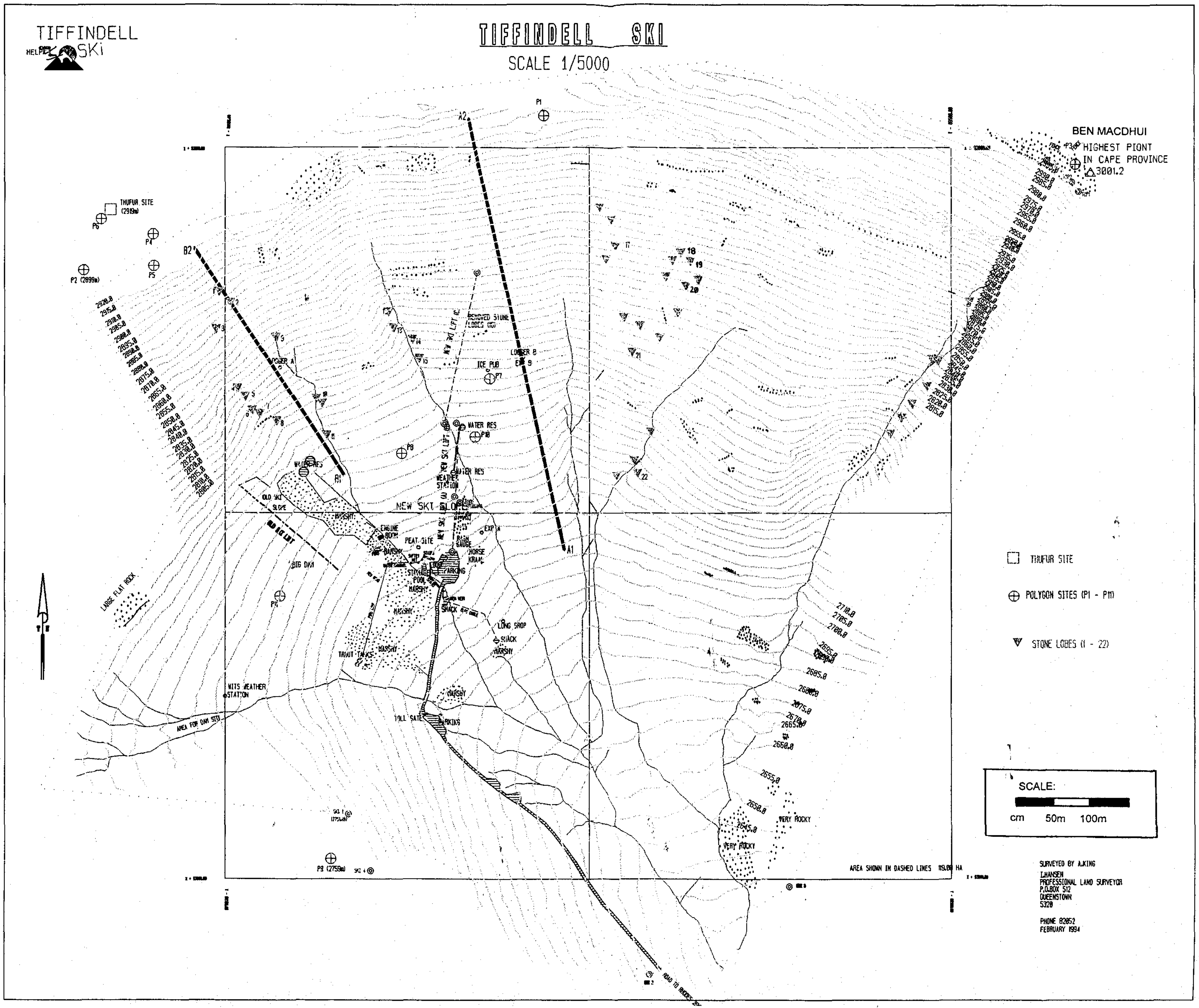
WILSON, P. and SELLIER, D. 1995 Active patterned ground and cryoturbation on Muckish Mountain, Co. Donegal, Ireland. *Permafrost and Periglacial Processes*, 6:15-25

WOLMAN, M.G. AND MILLER, J.P. 1960 Magnitude and frequency of forces in geomorphic processes. *J.Geol.*, 68:54-74

YOUNG, A., BRUNSDEN, D. and THORNES, J.B. 1974 Slope profile survey. *Brit. Geomorph. Res. Group Tech. Bull.*, 11. Norwich: GeoAbstracts, University of East Anglia.

ZOLTAI, S.C. and PETTAPIECE, W.W. 1974 Tree distribution on perennially frozen earth hummocks. *Arctic and Alpine Research*, 6:403-411

APPENDIX I



TIFFINDELL
SKi
TIFFINDELL SKI
SCALE 1/5000
BEN MACDHUI
HIGHEST PIONT
IN CAPE PROVINCE
3001.2
THUFUR SITE
(2919m)
P2 (2899m)
NEW SKI LIFT
REMOVED STONE LOBES
ICE PUB
WATER RES
WEATHER STATION
NEW SKI SLOPE
OLD SKI SLOPE
OLD SKI LIFT
ENGINE ROOM
PEAT SITE
MARSHY
RAIN GAUGE
HORSE KRAAL
PARKING
BIG DAM
STOKER POOL
SHACK
LONG DROP
TROUT TANKS
LARGE FLAT ROCK
WITS WEATHER STATION
AREA FOR DAM SITE
TOLL GATE
VERY ROCKY
ROAD TO RHODES
P9 (2759m)
AREA SHOWN IN DASHED LINES 119.00 HA
THUFUR SITE
POLYGON SITES (P1 - P11)
STONE LOBES (1 - 22)
SCALE:
cm
50m
100m
SURVEYED BY A.KING
I.HANSEN
PROFESSIONAL LAND SURVEYOR
P.O.BOX 512
QUEENSTOWN
5320
PHONE 82052
FEBRUARY 1994

APPENDIX II

Station B

J-Day	Air Temp. Max	Air Temp. Min	Ground temp Max	Ground temp Min	0.05m soil temp. Max	0.05m soil temp. Min	0.2m soil temp. Max	0.2m soil temp. Min
211	4.955	2.883	-0.168	-0.962	-0.282	-0.755	-0.649	-1.048
212	11.04	1.9	0.132	-2.035	0.077	-1.84	-0.474	-1.193
213	12.46	4.385	0.399	-0.525	0.232	-0.539	-0.195	-0.583
214	13.53	4.567	0.522	0.05	0.357	-0.033	0.014	-0.298
215	11.91	2.392	0.534	0.069	0.44	0.002	0.123	-0.133
216	11.97	3.334	0.499	0.289	0.374	0.082	0.139	-0.115
217	7.276	-3.175	0.659	0.392	0.513	0.314	0.366	0.094
218	1.63	-6.545	0.711	0.566	0.699	0.481	0.486	0.294
219	12.03	-1.639	0.681	0.366	0.633	0.32	0.441	0.299
220	10.64	0.922	0.577	0.418	0.551	0.286	0.413	0.245
221	11.18	2.811	0.506	0.246	0.469	0.195	0.451	0.182
222	12.03	-1.774	0.568	0.212	0.465	0.138	0.434	0.155
223	10.97	-0.364	0.459	-0.946	0.35	-0.574	0.486	0.08
224	8.05	2.842	0.611	0.393	0.505	0.28	0.413	0.272
225	12.81	2.102	0.636	0.196	0.518	0.221	0.434	0.26
226	15.3	2.929	0.462	-0.391	0.4	-0.019	0.489	0.12
227	14.84	3.571	0.533	0.062	0.447	0.103	0.451	0.163
228	12.92	2.955	0.614	0.34	0.538	0.2	0.506	0.2
229	10.84	2.407	0.675	0.468	0.58	0.415	0.498	0.368
230	13.15	1.762	0.74	0.48	0.647	0.301	0.483	0.3
231	12.58	2.897	0.718	0.53	0.593	0.392	0.521	0.343
232	15.09	5.311	0.683	0.496	1.45	0.595	0.48	0.328
233	12.12	-0.69	0.94	0.621	1.461	0.551	0.593	0.384
234	1.899	-5.27	0.861	0.743	0.81	0.595	0.657	0.517
235	1.981	-6.622	0.89	0.678	0.816	0.608	0.706	0.504
236	7.098	-5.125	0.867	0.202	0.839	0.515	0.768	0.52
237	11.92	1.593	0.512	-0.075	0.496	0.252	0.642	0.4
238	9.96	2.302	0.647	0.558	0.617	0.437	0.62	0.44
239	7.374	-4.355	0.803	0.579	0.757	0.439	0.67	0.536
240	16.27	1.622	0.649	0.237	0.66	0.289	0.654	0.377
241	14.33	4.211	0.67	0.432	0.592	0.315	0.592	0.325
242	12.12	3.323	1.005	0.603	1.757	0.472	0.594	0.42
243	17.22	2.475	2.074	0.726	3.624	0.596	0.654	0.391
244	14.76	4.323	2.754	0.671	3.665	0.57	0.609	0.402
245	14.71	4.212	3.471	0.873	4.283	0.623	0.64	0.411
246	10.76	1.398	1.587	0.704	1.896	0.507	0.64	0.466
247	3.508	-4.748	1.387	0.792	1.786	0.697	0.751	0.653
248	11.12	-6.047	0.848	0.654	1.573	0.508	0.809	0.453
249	12.79	2.28	1.08	0.678	2.085	0.511	0.712	0.452
250	14.1	3.355	2.038	0.72	3.114	0.511	0.691	0.463
251	15.66	6.71	3.876	0.8	4.261	0.617	0.666	0.434
252	15.21	7.601	4.753	0.92	4.856	0.714	0.604	0.435
253	16.26	6.492	4.996	0.863	5.56	0.68	0.635	0.433
254	16.97	5.226	5.132	0.908	5.601	0.693	0.667	0.438
255	15.31	4.297	4.583	0.84	4.632	0.655	0.711	0.43
256	16.82	4.642	5.81	1.149	5.528	0.846	0.685	0.488
257	16.72	4.667	7.872	1.57	7.403	1.267	0.936	0.558
258	15.98	4.447	7.891	1.978	7.28	1.714	1.288	0.673
259	15.96	4.847	7.623	1.684	7.512	1.503	1.706	0.777
260	15.33	3.742	7.36	2.313	6.957	2.007	2.125	1.02
261	11.46	0.505	6.954	1.959	7.848	1.867	2.743	1.398
262	12.86	1.215	5.673	1.486	6.04	1.442	2.703	1.569
263	10.2	5.566	5.365	2.585	5.169	2.388	3.037	1.645
264	14.2	4.455	13.8	4.355	11.09	4.042	7.537	2.983
265	16.08	7.354	16.08	4.321	14.79	4.652	9.83	4.684
266	21.71	8.07	17.2	2.614	15.9	2.985	11.05	4.497
267	11.98	6.855	12.57	5.316	11.86	5.784	9.69	6.706
268	17.2	6.899	16.04	4.08	15.42	4.393	10.94	5.742
269	14.79	4.72	13.65	1.756	13.24	2.421	9.49	4.698
270	17.53	10.01	12.86	2.679	12.26	3.018	9.22	4.314
271	18.25	8.87	17.35	4.527	17.02	4.721	12.65	5.834
272	19.67	8.13	17.47	5.441	16.75	5.543	13.08	6.856
273	20.13	9.18	18.78	4.498	17.75	4.67	13.71	6.103
274	19.09	5.947	18.77	4.691	17.45	4.806	13.46	6.161
275	13.51	5.451	16.26	4.793	16.57	5.284	13.18	6.956
276	16.17	4.838	18.59	3.469	18.36	3.419	13.62	4.643
277	17.42	5.79	18.22	2.502	17.27	2.427	13.06	3.812

Station B

J-Day	Air Temp. Max	Min	Ground temp Max	Min	0.05m soil temp. Max	Min	0.2m soil temp. Max	Min
278	17.07	8.18	15.98	1.993	14.77	1.869	11.33	3.208
279	14.21	4.89	15.12	1.574	13.84	1.337	10.57	2.876
280	15.85	5.57	16.8	1.201	15.31	1.047	10.83	1.976
281	15.74	4.173	13.31	0.764	12.12	1.104	8.81	2.265
282	16.35	5.311	11.59	4.757	11.38	5.293	9.03	6.385
283	16.38	4.65	12.6	3.206	12.14	3.391	8.13	5.893
284	17.49	6.266	12.51	3.485	12.18	3.976	7.906	5.898
285	17.7	3.974	11.93	5.106	11.66	5.54	8.15	6.658
286	13.36	-0.823	13.41	4.278	12.97	4.571	8.34	6.413
287	13.9	-2.313	14.19	3.513	13.58	3.916	8.46	6.148
288	12.84	3.434	13.33	4.614	12.76	4.935	8.52	6.61
291	13.29	4.261	16.5	6.853	14.28	7.318	11.19	8.73
292	7.659	2.4	8.99	5.585	8.02	5.835	8.65	6.643
293	10.17	0.941	15.09	3.269	13.18	3.948	10.78	4.761
294	10.81	2.563	14.34	3.201	12.83	3.871	10.65	4.932
295	13.74	4	15.71	2.069	14.05	2.777	11.28	3.711
296	9.55	-1.326	10.11	2.643	8.73	3.277	8.14	5.276
297	9.11	-0.81	14.53	1.325	12.88	1.615	10.17	2.376
298	8.02	-1.837	13.95	1.67	11.94	2.142	9.54	3.048
299	8.44	-5.739	15.52	0.91	14.23	1.097	10.21	1.811
300	12.1	1.816	15.78	0.888	14.92	1.018	11.07	1.767
301	15.2	3.46	17.21	1.255	16.35	1.447	12.8	2.272
302	18.64	4.21	19.16	2.551	17.85	2.844	14.23	3.935
303	18.04	6.874	18.56	4.453	16.46	4.639	13.72	5.596
304	16.78	7.008	18.05	4.842	16.62	4.892	13.7	5.718
305	19.11	6.864	17.07	5.182	15.61	5.431	12.99	6.539
306	14.52	6.619	19.37	5.596	18.64	6.564	15.7	8.77
307	17.79	7.214	18.99	3.366	17.6	3.905	13.96	5.017
308	17.71	7.865	20.48	4.852	19.09	5.035	15.58	5.831
309	17.5	6.11	20.76	2.511	18.91	3.099	13.47	4.359
310	18.49	7.955	16.03	2.879	15.41	3.372	11.01	6.321
311	16.59	8.71	19.14	5.194	18.65	5.994	15.2	8.95
312	15.42		3.772		4.342	2.27	9.66	5.35
328	11.26	1.353	15.83		15.68	6.089	12.65	4.229
329	11.86	-1.811	17.22	4.924	16.62	5.433	12.12	9.51
330	12.61	-1.349	17.97	4.349	17.17	4.947	11.96	8.87
331	15.31	5.172	18.77	5.482	17.91	6.178	12.25	9.28
332	15.45	7.63	12.75	7.993	12.1	8.41	11.59	10.19
333	12.06	4.304	9.23	5.662	9.04	6.128	10.07	8.95
334	9.94	-3.135	7.233	3.247	6.75	3.776	8.74	6.984
335	11.9	-4.687	20.85	1.822	19.09	2.048	12.91	4.449
336	15.38	2.988	20.86	3.273	18.5	4.125	12.7	6.729
337	15.01	4.061	12.94	7.311	12.02	7.905	11.85	9.83
338	18.05	4.448	18.5	6.198	16.72	6.674	11.84	8.81
339	16.15	5.449	17.76	8.3	16.64	8.45	12.42	10.05
340	16.11	5.039	18.96	6.943	16.67	7.63	13.99	9.13
341	16.81	4.084	18.21	5.122	15.75	6.516	12.94	8.61
342	15.43	3.257	20.56	5.406	17.88	6.446	14.86	8.24
343	11.91	4.251	13.95	7.109	12.07	7.596	12.3	9.14
344	15.3	5.17	22.08	4.625	19.68	5.822	15.8	7.336
345	16.79	5.127	20.07	5.128	17.31	6.201	15.23	8.33
346	17.11	6.443	23.28	5.13	21.55	5.764	17.09	7.573
347	16.88	7.136	21.6	7.586	21.11	7.812	17.69	9.28
348	13.92	7.018	16.37	7.378	14.27	7.7	12.34	8.59
349	17.85	6.584	21.8	6.391	20.38	6.845	17.74	7.712
350	15.46	4.598	16.08	7.085	14.39	7.517	12.43	8.75
351	7.388	4.51	9.41	6.506	8.78	6.793	8.64	7.273
352	12.6	6.292	14.69	7.403	13.09	7.519	11.98	7.577
353	15.53	7.159	19.04	7.354	16.77	7.623	15.09	8.19
354	16.73	7.397	20.31	6.137	18.25	6.903	16.29	8.02
355	16.91	6.148	20.76	6.229	17.7	7.026	15.72	8.16
356	14.62	3.372	15.4	6.442	13.19	6.939	12.08	7.932
357	18.82	4.585	17.35	5.341	14.41	6.11	12.59	6.588
358	16.09	6.106	15.15	6.313	13.22	6.863	12.02	7.522
359	15.93	6.68	16.91	6.386	14.5	6.959	13.72	7.801
360	14.55	4.866	16.33	6.375	14.46	7.003	13.76	7.885
361	14.47	3.722	18.64	5.826	16.5	6.811	15.02	7.818
362	17.27	6.273	20.62	7.025	17.84	7.571	16.09	8.5

Station B

J-Day	Air Temp. Max	Min	Ground temp Max	Min	0.05m soil temp. Max	Min	0.2m soil temp. Max	Min
363	17.33	6.241	23.95	6.049	22.04	6.963	19.16	8.25
364	19.09	6.702	24.24	6.186	22.46	7.188	19.54	8.68
365	21.6	9.46	25.13		23.55	7.101	20.28	8.89
1	22.51	10.26	25.61	6.603	25.02	7.408	20.9	9.45
2	22.3	11.35	20.1	7.477	19.28	8.58	14.74	11.36
3	21.64	10.03	22.79	10.89	21.72	11.15	16.8	13.68
4	22.79	9.28	24.33	11.51	23.19	11.96	17.3	13.91
5	22.11	9.4	22.7	12.96	22.27	13.41	17.37	15.06
6	22.49	10.12	24.97	12.49	24.02	12.95	18.39	15.2
7	19.57	10.34	22.19	13.28	21.88	13.7	17.49	15.73
8	19.7	9.69	22.18	10.34	21.77	11.14	16.9	14.48
9	19.9	9.72	22.4	11.97	21.92	12.6	17.06	14.72
10	20.82	8.58	22.52	11.11	22.03	11.69	17.11	14.56
11	22.93	9.97	23.73	10.45	23.12	11.25	17.42	14.29
12	21.53	9.86	24.02	11.99	23.43	12.59	17.66	14.89
13	20.79	9.77	22.55	12.26	21.88	12.9	17.17	15.1
14	21.13	10.72	24.59	11.21	23.76	11.82	16.84	14.32
161	5.06	0.562	0.622	0.43	0.684	0.501	0.96	0.806
162	8.37	0.426	0.582	0.228	0.71	0.518	0.975	0.775
163	9.84	1.593	0.55	0.275	0.703	0.493	0.959	0.78
164	6.586	3.121	0.633	0.459	0.693	0.534	0.954	0.805
165	8.84	2.706	0.58	0.44	0.723	0.58	0.936	0.816
166	8.83	4.211	0.615	0.434	0.718	0.581	0.956	0.82
167	8.84	1.288	0.646	0.49	0.696	0.571	0.988	0.82
168	8.1	-0.859	0.63	0.4	0.732	0.598	1.037	0.815
169	8.59	-2.645	0.598	0.203	0.785	0.6	1.057	0.78
170	10.41	-2.053	0.556	-0.125	0.702	0.525	1.008	0.793
171	10.98	-0.353	0.462	-0.538	0.62	0.136	1	0.799
172	8.75	-0.796	0.48	-0.43	0.66	0.1	0.916	0.668
173	11.75	1.44	0.471	-0.757	0.555	-0.21	0.908	0.703
174	15.05	-0.874	0.459	-0.611	0.601	-0.15	0.836	0.61
175	4.133	-1.32	0.445	-0.421	0.583	-0.037	0.85	0.739
176	3.546	-0.974	0.574	-0.456	0.58	-0.105	0.827	0.758
177	1.81	-3.704	0.493	0.154	0.616	0.36	0.82	0.72
178	11.26	-4.939	0.463	-0.173	0.567	0.096	0.834	0.654
179	9.24	-1.557	0.357	-1.021	0.356	-0.55	0.878	0.595
180	12.01	-1.522	0.253	-2.419	0.454	-1.638	0.834	0.587
181	11.25	-1.179	0.397	-1.542	0.522	-1.049	0.754	0.537
182	8.67	0.611	0.393	-0.64	0.52	-0.304	0.698	0.513
183	8.38	2.027	0.559	0.05	0.618	0.22	0.634	0.574
184	9.46	1.333	0.6	0.141	0.6	0.326	0.691	0.531
185	6.7	-0.685	0.645	0.482	0.7	0.5	0.66	0.5
186	4.773	-3.916	0.619	0.056	0.75	0.422	0.715	0.607

Station C

J-Day	Air temp. Max	Min	Ground temp. Max	Min	0.05m soil temp. Max	Min	0.2m soil temp. Max	Min
211	5.402	3.011	0.305	-1.346	0.143	-1.17	-0.015	-0.251
212	9.17	2.155	2.666	-2.231	0.369	-1.88	0.041	-0.417
213	11.54	4.497	5.071	-0.744	0.702	-0.54	0.155	-0.243
214	12.53	3.872	5.674	-0.153	0.739	0.122	0.231	-0.131
215	9.46	3.066	6.393	0.325	0.709	0.127	0.258	-0.024
216	10.64	3.746	6.905	0.029	0.71	0.424	0.254	0.052
217	7.58	-2.868	1.089	0.228	0.805	0.576	0.443	0.207
218	-1.691	-6.345	0.745	0.631	0.894	0.772	0.637	0.466
219	11.73	-1.353	0.719	0.353	0.817	0.52	0.573	0.318
220	10.77	0.922	0.388	0.097	0.608	0.42	0.498	0.2
221	11.49	2.611	0.367	-0.22	0.689	0.146	0.46	0.195
222	9.01	-1.707	0.49	-0.822	0.785	0.198	0.54	0.246
223	8.85	-0.45	0.563	-2.463	0.738	-1.548	0.587	0.201
224	8.48	3.055	3.577	0.476	0.805	0.636	0.463	0.22
225	10.64	2.49	2.562	-0.498	0.899	-0.058	0.483	0.249
226	15.33	2.653	5.069	-1.662	0.78	-1.366	0.492	0.092
227	13.61	2.957	6.73	-0.615	0.825	-0.26	0.442	0.1
228	11.7	2.746	6.6	0.016	0.861	0.339	0.486	0.164
229	9.05	2.293	8.32	0.588	1.005	0.781	0.493	0.201
230	10.9	1.321	8.67	0.584	1.918	0.821	0.538	0.258
231	11.86	2.692	9.08	0.542	2.6	0.848	0.52	0.298
232	15.55	5.338	11.46	0.66	4.392	0.83	0.467	0.26
233	10.6	-0.404	6.124	0.652	2.727	0.84	0.574	0.353
234	0.394	-5.053	0.777	0.546	1.031	0.907	0.738	0.54
235	2.125	-6.344	0.812	0.542	1.067	0.864	0.802	0.536
236	7.631	-5.102	0.762	0.139	0.953	0.424	0.786	0.484
237	13.07	1.474	3.914	-0.246	0.86	-0.08	0.642	0.341
238	9.17	2.622	4.009	0.575	0.936	0.784	0.583	0.38
239	8.78	-3.963	6.059	-0.293	0.994	0.817	0.704	0.428
240	13.3	1.075	7.398	-0.193	0.989	0.552	0.622	0.264
241	14.39	4.309	11.64	0.401	1.201	0.753	0.552	0.285
242	11.59	3.116	8.62	0.58	3.322	0.944	0.536	0.389
243	15.88	1.888	15.12	0.447	6.532	0.893	0.62	0.252
244	14.07	4.682	12.39	0.664	6.025	0.868	0.574	0.319
245	14.82	4.208	13.41	0.573	6.59	0.928	0.556	0.298
246	9.47	1.349	8.23	-0.042	1.612	0.897	0.592	0.418
247	2.635	-4.598	7.392	-0.018	2.454	0.919	0.711	0.571
248	13.44	-6.474	13.29	-0.506	4.094	0.892	0.62	0.393
249	12.54	2.397	11.2	-0.261	2.445	0.886	0.57	0.32
250	14.47	3.188	12.81	0.005	4.598	0.898	0.58	0.33
251	16.43	6.872	13.73	0.594	6.474	0.943	0.521	0.265
252	15.01	7.874	13.79	0.815	6.925	0.92	0.535	0.28
253	16.71	6.07	16.53	0.855	8.29	0.883	0.541	0.248
254	16.54	4.362	15.97	0.614	8.26	0.939	0.602	0.307
255	15.94	3.759	14	0.146	6.89	0.948	0.636	0.289
256	15.08	4.566	13.17	0.876	7.527	0.99	0.606	0.359
257	15.85	4.165	17.51	0.789	11.38	1.121	0.639	0.324
258	14.31	4.772	17.84	1.185	10.52	1.376	0.62	0.333
259	14.96	5.248	17.2	0.889	11.2	1.23	0.649	0.308
260	13.18	3.937	13.96	1.594	9.21	1.665	0.626	0.376
261	11.73	0.357	18.84	0.964	11.66	1.4	0.68	0.49
262	13.27	1.038	16.25	0.075	8.37	1.084	0.639	0.37
263	11.13	5.858	10.09	1.286	6.576	1.494	0.577	0.127
264	12.42	4.723	14.62	4.034	10.61	3.734	0.859	0.235
265	17.44	7.084	18.23	2.67	13.03	3.879	1.587	0.582
266	20.31	7.337	20.69	1.279	14.44	2.227	3.847	0.907
267	12.19	7.094	13.66	4.556	11.47	5.658	6.004	4.051
268	18.33	7.162	19.37	3.745	14.24	4.86	6.283	5.075
269	16.91	4.812	16.59	1.282	12.2	3.27	6.187	5.131
270	17.73	10.19	14.98	2.906	11.35	3.933	5.823	4.859
271	18.98	9.02	21.57	4.567	16.03	5.643	7.18	5.445
272	19.5	8.41	19.99	5.603	15.45	6.826	7.694	6.573
273	20.8	9.46	21.9	4.808	16.62	6.4	8.11	6.815
274	18.51	6.159	21.86	5.403	16.06	6.79	8.7	7.308
275	14.05	5.206	19.14	5.145	15.87	6.756	8.99	7.435
276	17.87	4.877	22.31	3.438	17.1	5.327	8.48	6.88
277	17.66	5.518	21.69	2.414	16.19	4.585	8.15	6.549

Station C

J-Day	Air temp. Max	Min	Ground temp. Max	Min	0.05m soil temp. Max	Min	0.2m soil temp. Max	Min
278	18.18	8.46	18.35	2.08	13.73	4.218	7.938	6.498
279	15.21	5.159	17.2	1.628	12.54	3.594	7.405	6.318
280	17.04	5.658	19.06	1.552	13.87	2.976	6.855	5.708
281	17.37	4.399	21.86	1.665	16.03	2.993	7.16	5.636
282	15.54	4.629	17.75	3.147	13.49	4.714	7.413	6.343
283	17.55	3.882		2.535	14.07	3.173	7.168	5.966
284	19.15	6.277			13.97	3.655	7.263	6.052
285	16.67	4.224			13.02	5.264	7.486	6.448
286	13.5	-0.659			15.35	4.105	7.73	6.372
287	13.18	-2.248			16.23	3.454	7.625	6.283
288	14.92	3.652				4.57	7.779	6.576
291	13.1	2.458			13.85	5.895	8.62	7.459
292	5.701	0.957			9.59	5.268	7.376	6.254
293	8.52	-0.227			16.02	2.782	8.86	4.8
294	7.597				15	3.746	8.7	6.188
295					16.21	2.839	8.55	5.496
296	13.13				11.05	4.046	8.12	6.707
297	10.71	-1.086			13.49	2.735	7.713	4.229
298	7.785	-1.677			12.06	2.417	7.226	4.679
299	12.52	-5.448			13.6	1.467	6.93	3.638
300	15.03	2.015			13.24	2.018	7.085	4.149
301	16.79	3.368			14.36	2.61	7.862	4.712
302	18.16	3.709			15.76	4.4	8.71	6.051
303	15.96	6.519			15.19	6.633	9.52	7.485
304	16.73	7.175			16.29	7.366	9.86	8.04
305	19.04	7.567			35.65	7.949	21.89	8.52
306	17.38	6.083			42.46	13.54	28.42	2.405
307	17.28	7.246			44.2	20.52	30	9.31
308	18.35	8.11			42.46	9.23	27.78	0.987
309	18.29	5.623			39.24	5.238	22.74	8.23
310	18.32	8.33			41.07	13.63	25.13	4.571
311	19.18	8.6			41.83	11.55	26.82	3.31
312	15.38	6.876			42	28.41	26.14	16.34
328	13.22	1.828			15.84	6.48	11.86	3.935
329	12.44	-1.212			18.82	4.819	11.75	9.94
330	14.54	-0.871			20.06	4.226	11.36	9.42
331	17.96	5.541			20.49	5.408	11.47	9.56
332	15.76	7.834			12.14	8.01	11.38	10.3
333	12.63	4.597			9.25	5.693	10.25	9.35
334	11.14	-3.146			6.981	3.921	9.25	7.82
335	15.26	-5.651			21.35	3.222	9.92	6.905
336	17.74	2.767			20.15	4.858	10.58	8.38
337	15.38	4.148			12.46	7.356	10.51	9.4
338	16.88	4.244			18.91	6.24	10.76	8.87
339	16.17	5.623			18.47	8.29	11.68	9.83
340	14.97	4.786			19.03	7.301	11.86	9.83
341	17.58	3.271			18.12	6.52	11.89	10.22
342	17.92	2.967			20.96	7.501	12.42	10.54
343	12.02	4.429			13.61	8.82	12.34	11.03
344	17.05	4.74			24.1	6.173	12.64	9.64
345	18.71	4.719			20.35	8.24	13.21	11.33
346	17.18	6.624			20.6	8.43	13.17	11.55
347	17.42	7.266			20.38	10.59	13.82	12.22
348	15.68	7.08			16.99	10.02	13.56	11.67
349	18.74	6.644			21.24	8.49	13.85	10.7
350	16.9	4.963			16.67	8.51	12.9	10.81
351	7.605	4.722			11.35	7.367	10.27	8.4
352	12.83	5.528			16.51	7.97	13.71	8.27
353	14.79	6.142			21.56	8.07	15.93	9.02
354	17.65				23.63	7.515	15.35	9.58
1	20.08				21.19	8	15.09	10.09
2	17.88				17.17	6.932	13.44	9.62
3	10.02				18.73	5.387	13.58	7.598
4	26.8				15.85	6.851	13.49	8.53
5	26.79				18.82	7.506	13.69	9.23
6	25.9				18.24	8.13	13.68	9.88
7	26.8				19.71	7.808	13.83	10.03
8	25.72				20.1	9.36	13.97	10.85

Station A

	Air temp.		Ground temp.		0.05m soil temp		0.2m soil temp	
J Day	Max	Min	Max		Max	Min	Max	Min
148								
149			2.433	0.533	6.027	0.581	6.196	1.24
150			0.743	0.555	1.114	0.674	1.621	1.208
151			3.218	0.593	1.081	0.804	1.721	1.307
152			2.631	0.685	1.35	0.79	1.84	1.416
153			0.769	-1.78	1.402	0.937	1.784	1.441
154			1.083	-1.938	1.035	0.805	1.826	1.598
155			1.856	-1.676	0.883	0.658	1.63	1.467
156			1.627	-1.62	0.952	0.69	1.7	1.407
157			0.735	0.4	0.918	0.584	1.673	1.473
158			0.411	-0.171	0.945	0.663	1.692	1.353
159			0.442	-0.205	0.88	0.728	1.696	1.527
160			0.525	0.019	0.878	0.723	1.665	1.468
161			0.465	-0.301	0.851	0.467	1.717	1.3
162			0.363	-0.485	0.802	0.503	1.626	1.326
163			0.575	-1.036	0.683	0.434	1.406	1.09
164			0.553	-2.174	0.655	0.401	1.353	1.228
165			1.081	-2.092	0.696	0.449	1.538	1.186
166			2.023	-1.973	0.608	0.403	1.449	1.237
167			0.427	-0.994	0.585	0.18	1.346	1.08
168			0.524	0.178	0.615	0.41	1.328	1.154
169			0.687	0.471	0.627	0.362	1.389	1.27
170			0.667	0.317	0.661	0.333	1.628	1.061
171			0.34	-0.054	0.913	0.337	1.578	1.105
172			0.373	-0.126	0.69	0.259	1.43	1.004
173			0.372	-1.544	0.698	0.253	1.415	1.029
174			0.395	-2.121	0.537	0.26	1.296	1.057
175			0.197	-2.773	0.422	-0.076	1.208	0.993
176			0.433	-3.07	0.246	-0.232	1.247	1.023
177			0.632	-2.677	0.259	-0.545	1.203	0.986
178			1.558	-1.697	0.46	-0.342	1.163	1.017
			3.881	-2.325	0.483	0.016	1.177	1.011
211	3.188	0.706	0.537	-1.621	0.365	-0.122	1.158	0.964
212	9.87	0.181	3.94	-2.766	-0.008	-0.154	0.477	0.009
213	11.94	2.466	6.868	-0.705	0.3	-0.874	0.648	0.248
214	12.24	2.586	6.45	-0.19	0.431	0.105	0.814	0.487
215	10.55	2.789	6.923	0.467	0.528	0.325	0.753	0.632
216	10.72	1.893	6.079	0.206	0.575	0.448	0.84	0.697
217	6.377	-4.768	0.654	0.415	0.664	0.447	0.925	0.813
218	0.052	-7.999	0.663	0.528	0.626	0.453	0.931	0.717
219	9.85	-5.47	0.712	0.617	0.637	0.497	1.034	0.8
220	10.21	-0.924	0.669	-0.196	0.661	0.478	0.951	0.698
221	10.2	1.346	2.566	-1.196	0.687	0.45	0.925	0.63
222	11.06	-3.443	5.387	-0.827	0.563	0.407	0.982	0.729
223	9.41	-2.802	6.217	-2.759	0.605	0.477	0.985	0.773
224	6.899	0.571	6.668	0.364	0.737	0.293	1.014	0.854
225	11.86	-0.541	5.822	-1.967	0.727	0.542	1.099	0.758
226	13.63	1.248	9.65	-2.274	0.661	0.304	1.034	0.633
227	12.83	1.267	10.15	-0.749	0.544	0.365	0.905	0.685
228	11.57	1.424	9.24	-0.159	0.619	0.421	0.961	0.715
229	9.82	0.557	9.06	0.614	0.617	0.464	0.926	0.686
230	11.26	-0.213	10.38	0.602	0.674	0.487	0.972	0.79
231	11.02	1.03	9.27	0.598	0.779	0.476	0.948	0.793
232	13.7	3.666	11.15	0.678	0.657	0.532	1.006	0.811
233	10.13	-1.927	5.124	0.643	0.763	0.593	1.106	0.82
234	1.157	-7.018	0.757	0.632	1.113	0.664	1.051	0.892
235	0.684	-8.31	0.727	-0.83	0.826	0.63	1.122	0.846
236	6.39	-6.728	0.577	-2.825	0.871	0.667	1.144	0.949
237	10.75	-0.144	2.982	-2.642	0.801	0.389	1.153	0.79
238	7.815	0.447	5.53	0.762	0.608	0.039	1.074	0.751
239	6.497	-5.995	5.033	-0.941	0.735	0.509	1.145	0.866
240	13.93	-0.997	8.34	-1.316	0.797	0.441	1.129	0.745
241	13.38	2.145	11.2	-0.294	0.656	0.468	1.019	0.736
242	10.02	1.603	9.22	0.585	0.995	0.512	0.964	0.725
243	14.81	0.287	14.98	0.557	1.95	0.602	0.871	0.721
244	13	2.437	12.1	0.549	3.735	0.575	0.972	0.728
245	13.75	2.864	12.24	0.657	3.934	0.606	1.057	0.779

Station A

	Air temp.		Ground temp.		0.05m soil temp		0.2m soil temp	
J Day	Max	Min	Max		Max	Min	Max	Min
247	2.524	-6.079	7.745	0.444	2.608	0.641	0.991	0.854
248	11.52	-8.12	8.19	-1.467	2.958	0.671	1.12	0.907
249	11.91	0.583	8.56	-0.208	2.176	0.692	1.122	0.907
250	13.49	2.059	10.94	-0.163	3.118	0.696	1.068	0.895
251	14.97	5.018	11.75	0.674	3.965	0.679	1.057	0.905
252	14.9	5.854	12.64	0.766	4.4	0.722	1.091	0.863
253	15.96	5.463	15.59	0.715	4.818	0.865	1.084	0.871
254	15.05	3.076	14.47	0.575	5.933	0.834	1.294	0.951
255	14.8	2.042	13.46	0.352	5.765	0.83	1.599	1.014
256	14.38	2.958	13.42	0.825	5.367	0.822	1.71	1.041
257	14.5	3.572	17.04	1.143	5.868	1.01	2.002	1.139
258	13.79	2.928	17.22	1.948	7.992	1.487	2.811	1.352
259	13.79	3.398	15.85	0.884	8.03	1.932	3.376	1.708
260	14.25	2.068	15.83	1.902	7.788	1.791	4.172	2.158
261	10.25	-0.873	18.28	1.455	7.756	2.31	4.651	2.757
262	11.84	-0.943	15.11	0.484	8.23	2.185	5.083	3.179
263	8.44	1.889	9.63	1.172	6.277	1.7	4.584	3.157
264	11.58	2.994	15.33	4.596	5.391	1.4	4.376	1.01
265	14.75	5.502	17.17	3.353	8.46	4.234	6.1	4.162
266	19.27	5.371	19.43	2.067	10.17	4.836	7.168	5.129
267	9.37	4.855	12.93	4.629	11.33	4.249	7.776	5.325
268	14.91	4.795	17.35	3.488	9.29	6.365	7.847	6.716
269	13.32	2.35	15.45	1.362	10.76	5.392	8.14	6.465
270	15.11	8.04	13.9	2.393	9.64	4.439	7.761	6.306
271	16.69	6.771	19.17	4.493	8.99	4.524	7.648	5.987
272	17.34	6.124	18.65	5.74	11.59	5.867	8.88	6.61
273	18.06	7.589	21.18	4.879	12.27	7.151	9.83	7.837
274	17.35	4.32	20.51	5.434	13.25	7.009	10.49	8.05
275	12.52	3.504	18.29	5.024	13.4	7.558	10.45	8.57
276	14.97	2.781	20.74	3.498	12.37	7.256	10.17	8.57
277	15.46	4.139	20.08	1.842	12.88	6.275	10.15	8.05
278	15.86	6.312	16.61	1.958	12.6	5.506	9.86	7.89
279	13.07	3.219	15.67	2.126	10.72	5.242	9.2	7.447
280	14.44	3.672	17.31	1.758	10.21	5.067	8.73	7.147
281	14.5	2.319	18.85	1.74	10.54	4.185	8.3	6.443
282	14.58	3.635	16.06	2.562	11.41	4.212	8.67	6.373
283	15.01	2.653	16.98	1.814	10.44	5.112	8.49	6.897
284	16.41	4.695	17.15	1.935	10.86	4.33	8.45	6.626
285	15.61	2.281	16.66	3.79	11.07	4.708	8.67	6.701
286	11.75	-2.448	19.99	2.941	11.09	5.966	8.94	7.308
287	12.83	-3.667	21.61	1.991	11.97	5.164	9.27	7.161
288	13.22	1.968	19.14	2.619	12.83	4.531	9.46	7.073
289	10.78	5.061	13.47	4.418	12.14	5.494	9.47	7.483
290	10.79	5.035	14.29	6.35	9.81	6.749	9.03	7.908
291	12.47	2.332	16.8	5.987	10.19	7.045	8.96	7.972
292	5.496	0.046	9.01	4.683	11.11	6.669	9.55	8.21
293	9.11	-1.165	14.09	3.877	8.05	5.548	7.995	6.863
294	8.64	-0.558	15.94	2.089	12.34	3.735	11.92	5.224
295	11.42	1.313	18.56	0.861	11.48	4.016	10.45	5.53
296	7.676	-3.185	11.15	1.274	11.46	2.604	8.78	4.905
297	7.46	-2.998	17.12	0.817	8.73	3.493	8.31	6.454
298	6.192	-5.049	15.5	0.825	9.89	2.291	7.469	4.493
299	8.24	-8.08	18.16	0.074	9.47	2.667	6.729	4.89
300	11.76	-0.874	16.73	-0.018	8.42	2.286	6.242	4.204
301	13.52	0.989	18.64	0.217	8.5	2.58	6.353	4.183
302	14.87	0.832	21.45	0.703	10.18	2.519	7.375	4.137
303	13.99	4.47	20.69	3.91	12.19	3.895	8.91	5.439
304	13.2	5.062	19.75	5.251	12.23	6.473	10.03	7.574
305	15.46	5.314	21.97	5.107	12.76	7.39	10.73	8.43
306	12.82	4.351	22.57	3.712	14.31	7.845	11.5	8.99
307	14.39	4.46	23.05	1.882	14.53	8.78	11.89	9.93
308	15.19	5.312	22.23	3.372	13.95	6.547	11.2	8.85
309	15.08	2.446	22.44	0.385	14.2	7.346	11.09	8.98
310	15.17	5.45	23.48	2.047	14.6	5.296	11.11	7.817
311	15.66	6.186	21.21	3.209	14.47	5.877	10.95	7.928
312	14.28	5.049	16.55		14.26	8.7	11.71	9.54
					8.39	6.042	10.7	8.71

APPENDIX III

TERRACETTE MORPHOLOGY

TERR. #	TREAD LENGTH	TREAD WIDTH	TREAD ANGLE	RISER HEIGHT
1	1.3	0.7	4	0.25
2	1.2	0.6	4.5	0.55
3	1.6	0.65	7	0.50
4	0.7	0.45	6	0.35
5	0.9	0.8	4.5	0.54
6	0.51	0.3	4.5	0.26
7	0.75	0.65	4.5	0.32
8	0.67	0.52	5	0.54
9	1.7	1.2	5.5	0.45
10	2.3	1.87	5.5	0.53
11	2.35	1.90	6	0.35
12	1.76	1.02	6	0.46
13	1.94	1.68	7	0.48
14	1.6	1.59	4.5	0.42
15	0.95	0.45	4.5	0.34
16	1.56	1.23	6	0.52
17	1.47	1.02	6.5	0.51
18	1.83	1.37	5.5	0.41
19	2.03	1.02	6.5	0.58
20	2.5	1.19	6	0.3
21	0.67	0.32	6	0.38
22	0.42	0.1	7	0.54
23	0.69	0.6	4.5	0.42
24	0.54	0.3	5	0.46
25	1.9	0.78	6	0.55

TURF-BANKED LOBE MORPHOLOGY

LOBE #	TREAD LENGTH	TREAD ANGLE	RISER HEIGHT
1	0.7	2	0.48
2	0.9	5	0.41
3	1.5	7	0.73
4	1.6	9	0.78
5	1.54	3	0.86
6	1.39	5	0.82
7	1.9	6	0.79
8	1.03	9	0.79
9	2.67	8	0.56
10	2.02	10	0.7
11	1.98	10	0.4
12	2.69	11	0.41
13	2.76	14	0.6
14	2.86	2	0.87
15	3.14	7	1.02
16	3.01	5	1.03
17	2.9	3	0.89
18	3.7	9	0.69
19	2.01	10	0.57
20	3.4	10	0.73
21	2.7	7	0.69
22	2.51	8	0.74
23	3.01	8	0.73
24	1.78	5	0.63
25	0.92	4	0.59
26	0.8	6	0.67
27	0.7	3	0.54
28	1.24	2	0.46
29	2.56	8	1.01
30	2.4	5	0.92

TURF-BANKED LOBE MORPHOLOGY

LOBE #	TREAD LENGTH	TREAD ANGLE	RISER HEIGHT
1	0.7	2	0.48
2	0.9	5	0.41
3	1.5	7	0.73
4	1.6	9	0.78
5	1.54	3	0.86
6	1.39	5	0.82
7	1.9	6	0.79
8	1.03	9	0.79
9	2.67	8	0.56
10	2.02	10	0.7
11	1.98	10	0.4
12	2.69	11	0.41
13	2.76	14	0.6
14	2.86	2	0.87
15	3.14	7	1.02
16	3.01	5	1.03
17	2.9	3	0.89
18	3.7	9	0.69
19	2.01	10	0.57
20	3.4	10	0.73
21	2.7	7	0.69
22	2.51	8	0.74
23	3.01	8	0.73
24	1.78	5	0.63
25	0.92	4	0.59
26	0.8	6	0.67
27	0.7	3	0.54
28	1.24	2	0.46
29	2.56	8	1.01
30	2.4	5	0.92

APPENDIX IV

CLAST NUMBER	LONG-AXIS LENGTH (mm)	DIP	ORIENTATION (magnetic north)
1	490	6°	252°
2	240	15°	286°
3	680	54°	306°
4	250	23°	184°
5	440	24°	199°
6	390	7°	260°
7	920	6°	152°
8	680	4°	175°
9	540	21°	182°
10	380	28°	193°
11	280	6°	169°
12	780	20.5°	257°
13	330	31°	253°
14	230	36°	181°
15	380	2°	252°
16	690	19°	197°
17	650	8°	264°
18	210	21°	290°
19	730	33°	279°
20	460	2°	235°
21	590	12°	244°
22	920	19°	262°
23	540	33°	137°
24	1300	20°	219°
25	710	10°	190°

Fabric data for lobe 19.