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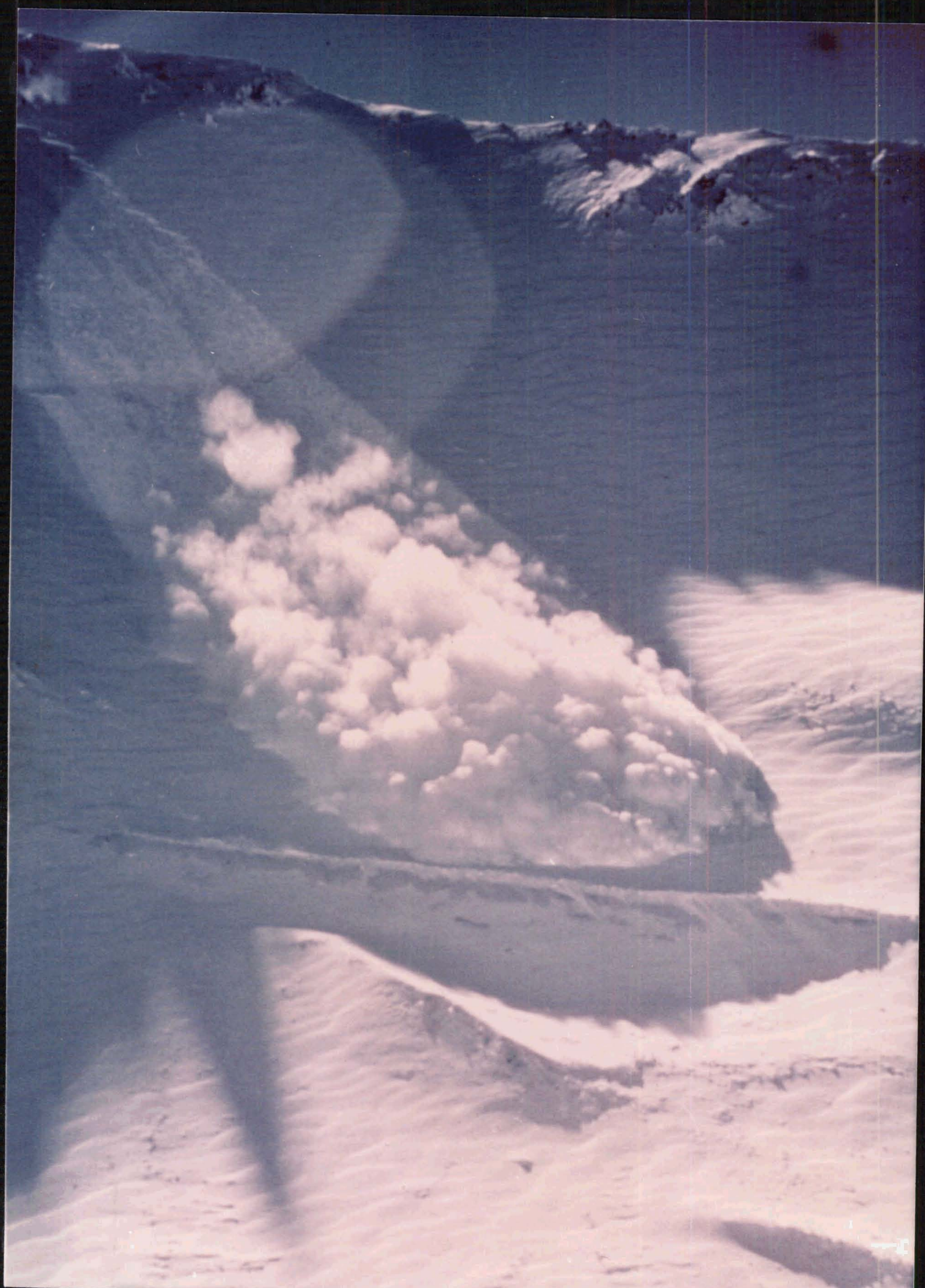
ARTIFICIALLY RELEASED MIXED MOTION AVALANCHE
(Photo: Dave McNulty)



SNOW AVALANCHE PHENOMENA
ON THE EASTERN SIDE OF THE
CRAIGIEBURN RANGE
NEW ZEALAND

A thesis submitted in partial fulfilment

of the requirements for the degree

of

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GLENN RUSSELL McGREGOR

University of Canterbury

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TABLE OF CONTENTS

Page

Page

Page

Page

LIST OF FIGURES

LIST OF TABLES

LIST OF APPENDICES

ABSTRACT

The role of terrain, meteorological and snowpack factors contributing to snow avalanche occurrence were investigated on the eastern side of the Craigieburn Range. Data on which this analysis was based were collected in the winters of 1979 to 1981.

The majority of slab avalanching occurred on lee slopes with large bowl shaped starting zones of 33 to 36 degrees inclination while frequency of avalanching was related to path size factors. Theoretical models adequately predicted avalanche runout distance on the majority of paths. However, these models proved only partially satisfactory for valley side paths with steep vegetated runout zones. An empirical model developed for predicting avalanche runout based on measured terrain characteristics performed well.

Threshold values for avalanching were established for precipitation, temperature and wind related variables. A discriminant analysis model for distinguishing dry, wet and non-avalanche days was established using maximum six hour precipitation intensity, variation in wind direction, total global radiation and two hour mean temperature. An independent test of this model produced acceptable results.

Synoptic weather types dominated by vorticity advection processes and characterised by cyclonic circulation were the most frequent storm types producing snowfall in the study period. However, systems typified by a mixture of warm advection and vorticity advection processes with

well developed troughs, associated fronts, high rates of thermal advection and combinations of stability and instability produced the greatest amounts of snow and snow avalanching.

A three phase snowpack development model was proposed and considered in relation to snow avalanching. The magnitude of positive and negative snowpack temperature gradients exceed those generally required for recrystallisation processes. Temperature gradient snow crystals were observed at the base and mid to upper parts of the snowpack. Fracture line profile, shovel test and shear frame analyses revealed that partially metamorphosed new snow undergoing equitemperature metamorphism, especially clusters of rimed needles, and snow in the early stages of temperature gradient metamorphism, were characteristically weak in shear. Three categories of potential sliding layers for avalanche release are presented. Consideration of snowpack and meteorological information indicates that the Craigieburn Range snow avalanche climate approximates most closely that of one transitional between a maritime and continental situation.

CHAPTER ONE

INTRODUCTION

1.1 SNOW AVALANCHES

Snow avalanches are a ubiquitous phenomenon in the alpine regions of the world and where they occur in conjunction with human activity they cause tremendous problems for public safety. In comparison with other countries, the annual death toll from snow avalanche remains low for New Zealand (Table 1.1), though a marked increase has occurred in the last two decades. This trend is accentuated if the 1980-83 data are extrapolated to the 1980 decade (Figure 1.1).

Table 1.1: Annual Avalanche Fatalities

Country	Average Fatalities	Period of Research
Austria	36	1949-1969
Japan	27	1918-1974
Switzerland	25	1940-1969
Norway	12	1836-1975
France	10	1949-1969
Italy	10	1949-1969
Canada	6	1970-1975
USA	6	1950-1974
Yugoslavia	5	1949-1969
Germany	3	1949-1969
Iceland	4	1800-1957
New Zealand	0.55	1900-1983
Scotland	0.16	1800-1979

Compiled from: Williams (1975a, 1975b), Dingwall (1977), Bjornsson (1980), Ward (1980).

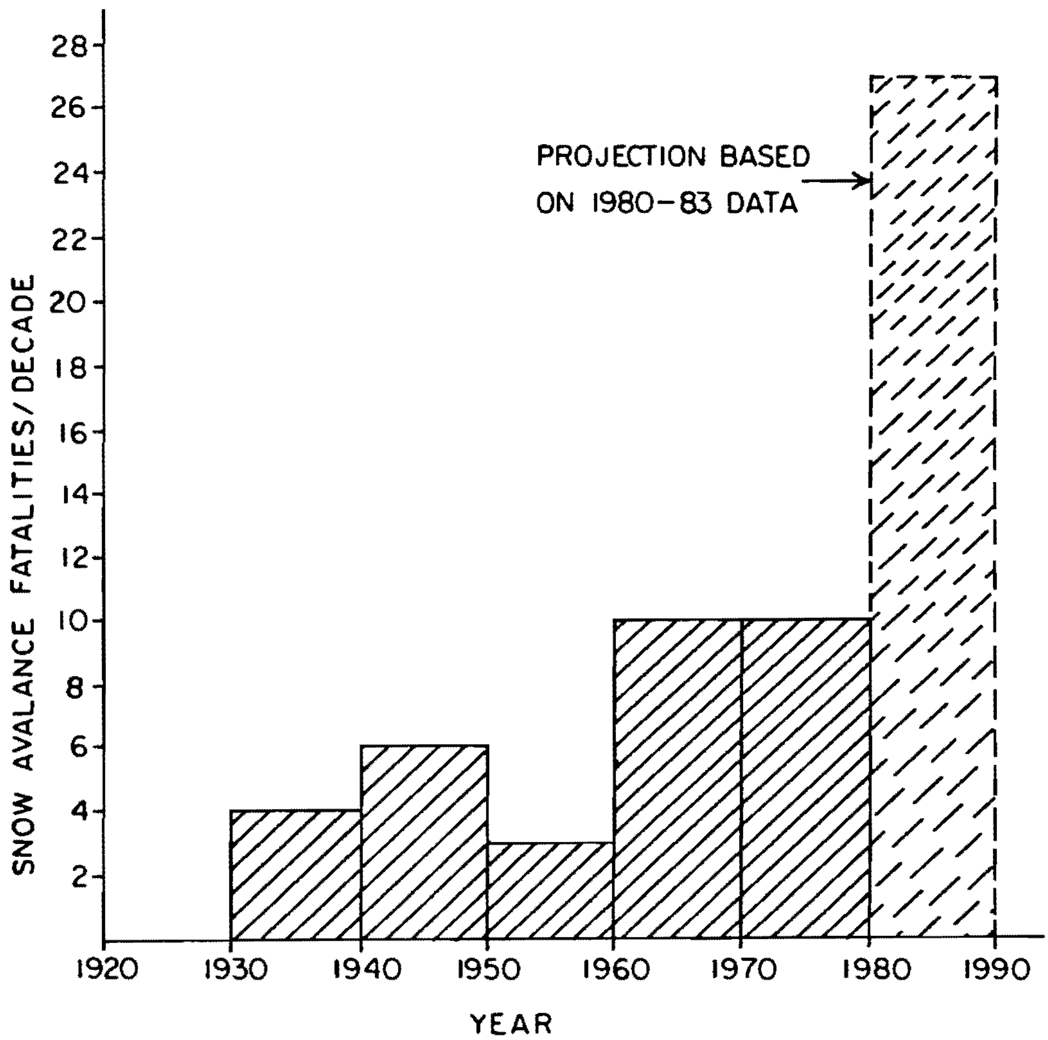


Figure 1.1: Number of avalanche fatalities by decade. Estimated fatality count for 1980-1989 is 27 given projected annual fatalities of 1980-1983.

Williams (1975a, 1975b) and Frutiger (1977) have noted similar trends in the United States and Switzerland respectively, and have suggested that this is a response to the increased recreational use of mountain environments in general. For New Zealand, Owens and O'Loughlin (1979) have noted an increased patronage of as much as 300 to 400 per-cent on some skifields over the last decade. This is of special significance, for while mountaineers have tradit-ionally recognised snow avalanche danger in the perennial snowpack (Simpson-Housley and Fitzharris, 1979) little is known of the behaviour of the seasonal snowpack and snow avalanching. With the extensive development of New Zealand skifields in the altitudinal zone of 1000-2000 m over the last decade, it has become increasingly apparent that avalanches are an active part of the seasonal snow cover. Accordingly, these must be dealt with on a day to day basis by skifield staff, climbers and mountain recreationists in general. While 75 percent of the avalanche fatalities recorded in New Zealand to date are climbing related, recent data suggest that skiers are more likely to be caught in an avalanche (Lefever, 1981, 1982, 1983). Presently the majority of skiing activity is restricted to skifield areas, but the problem of dealing with avalanches from both a commercial, management and forecasting point of view will become augmented as the sports of cross-country and heli-skiing gain popularity. This is further complicated by the fact that this rapid development of mountain recreation has paralleled a period of generally light snowfall winters (LaChapelle, 1979). Clearly then, there is a need for a greater understanding of the factors involved in avalanche occurrence in the New Zealand snow environment.

1.2 CURRENT STATE OF KNOWLEDGE

LaChapelle (1977) has noted that snow avalanches have long represented one of the most active areas of applied glaciology. Literature dealing with avalanches therefore appears in a wide range of publications. The first English language volume in which snow avalanches are given considerable coverage was that produced by Seligman (1936). However, reference to earlier works in other languages, on snow avalanches are made in this volume (Coaz, 1881). More recently, Mellor (1968) has presented an extensive review of avalanches while Perla and Martinelli (1976) have produced a handbook which deals with the practical aspects of slope stability evaluation, avalanche forecasting, control and rescue as well as many of the theoretical aspects of avalanche science. Apart from recent symposiums dealing with various aspects of snow avalanches (National Research Council of Canada, 1970, 1978, 1981; IAHS, 1966, 1975a, 1975b; United States Forest Service, 1973; International Glaciological Society, 1977, 1980, 1983; CRREL, 1981), the only other English volumes which the author is aware of that give over sections to snow avalanches are Shartran (1977), Colbeck (1980), and Gray and Male (1981). Research on avalanches has dealt with the following main themes.

(1) General area studies: Several studies have been conducted on the general characteristics of avalanching within distinct geographic areas. These usually assess the terrain, meteorological and snowpack factors but often emphasise the meteorological elements of avalanche formation. Early studies of this type have been carried out in

Europe and North America, but more recently studies from countries such as China, Ireland and Scotland are appearing in the literature. These general area studies are summarised in Table 1.2.

(2) Avalanche terrain mapping and zoning: General guidelines for the identification and mapping of avalanche terrain have been presented by Martinelli (1974) and Mears (1976). One of the main problems in avalanche terrain studies is the determination of the extent of avalanche runout. In the absence of good vegetation and geomorphological evidence (Bovis and Mears, 1976; Burrows and Burrows, 1976) historical records must be resorted to for the identification of the maximum runout distance of avalanches with return periods of 1 in 100 years or greater (Frutiger, 1975; Lied and Bakkehoi, 1980; Hestnes and Lied, 1980; Bjornnson, 1980). For the purpose of delimiting zones of likely avalanche damage for land use planning, avalanche mapping and zoning has been used widely (de Crecy, 1980; Frutiger, 1980; Ives and Plam, 1980; Freer and Schaerer, 1980). Such studies have drawn on a wide range of information for this purpose (Ives et al., 1976).

(3) Avalanche Dynamics: This is a rapidly expanding field of interest within the general area of avalanche studies. Most models for replicating or simulating avalanche flow have been based on a fluid flow analogy for avalanche flow (Voellmy, 1955; Lang et al., 1979; Cheng and Perla, 1979; Perla et al., 1980; Dent and Lang, 1980). These models are dependent on the setting of coefficients of friction representative of snow and terrain conditions for stopping avalanche flow. Mears (1980) debates the applicability of a fluid flow analogy to avalanche flow and

Table 1.2: General Area Studies Dealing With Snow Avalanching

Author	Area	Type of Study
Coaz (1881)	Swiss Alps	Consideration of general conditions associated with snow avalanching and implications.
Allix (1924)	French Alps	Review of avalanching in literature and causes and effects of avalanches to that date. Attempts at avalanche classification and prediction.
Akkouratov (1966)	Khibiny, Russia	General consideration of meteorological conditions involved in snow avalanche formation for this area with special attention to the role of snow drifting.
LaChapelle (1966)	Western United States	Partitioning of this area into 4 distinct areas of avalanche climate with associated contrasts in snowpack development and avalanche forecasting practices.
Wakabayshi and Yamamura (1968)	Hokkaido, Japan	General features of snow slab avalanches in Hokkaido.
Kosarov (1969)	Western Tien Shan, China	Evaluation of results on study of general meteorological conditions accompanying snow avalanche formation.
Yegimov and Kozik (1975)	Dukcant River Basin Western Tyan'Shan China	Relation of meteorological elements to avalanche formation.
Kotlyakov et al. (1977)	Khibins, Russia	General consideration of snow avalanche formation processes.
Bjornsson (1980)	Iceland	Evaluation of the areas general avalanche climate and terrain with identification of several areas with distinct terrain and weather characteristics as related to avalanche formation.
Ward (1980)	Cairngorm Mountains Scotland	General consideration of avalanche hazard and related avalanche formation processes.
Xiangsong and Yanlong (1980)	Tuen Shan China	Physical and mechanical properties of snowcover as related to snow avalanching plus consideration of general avalanche activity.

has presented a model based on the flow of fragments. Many of these models have been subsequently tested as to their applicability for predicting avalanche runout distance (Martinelli and Leaf, 1976; Buser and Frutiger, 1980; Martinelli et al., 1980; Bakkehoi et al., 1981, 1983; Schieldrop, 1983). Lied and Bakkehoi (1980) favoured a probabilistic approach for the calculation of runout distance and have presented an empirical model based on a range of terrain variables. Bovis and Mears (1976) likewise presented a similar empirical model relating runout distance to starting zone area. The most recent approach to predicting avalanche runout has been the combination of empirical and theoretically based techniques (Bakkehoi et al., 1983). Also related to the field of avalanche dynamics is the problem of estimating avalanche velocity for a variety of engineering and planning applications. Though theoretical models have gone some way in solving this problem (LaChapelle and Lang, 1980) several attempts have been made to measure actual avalanche velocity (Schaerer, 1973; Heimgartner, 1977; Schaerer and Salway, 1980; Shimizu et al., 1980).

(4) Avalanche forecasting: Meteorological and snowpack structure data comprise the two basic sources of information for forecasting avalanches and as pointed out by LaChapelle (1966) the emphasis placed on either of these factors may vary between areas of contrasting snow climates. Traditionally, emphasis has been placed on meteorological factors for avalanche forecasting owing to the difficulty of monitoring and assessing the significance of snowpack development processes during stormy periods. Two broad approaches to avalanche forecasting exist. These are the conventional

methods, based on cause and effect relationships between weather, snowpack and avalanche conditions, and the statistical approaches which depend on an objective analysis of accumulated snowpack, weather and avalanche occurrence data. The decision making processes involved in conventional methods of avalanche forecasting are complex and iterative in nature with some redundancy of data implicit in this approach (LaChapelle, 1980). Meteorological factors on which the conventional methods rely have been established for some time (Atwater, 1952; Judson, 1964, 1966; Perla, 1970). Perla (1970) and Obled (1970) presented the first statistically based approaches to assessing the relative importance of variables for forecasting avalanches. Perla (1970) analysed the effect of a range of meteorological variables on avalanche hazard probability while Obled (1970) used an elementary index method of uitilising meteorological parameters to identify avalanche days for a limited region. Bois and Obled (1972) subsequently applied principal components analysis to the problem of avalanche forecasting. Apart from Judson and Erickson (1973) who used a regression technique and Salway (1979) who applied time series analysis for the forecasting of avalanches, the majority of recent studies have been based on discriminant function analysis (Bois et al., 1975; Bovis, 1976, 1977; Fohn et al., 1977; Drozdovskaya, 1979; Obled and Good, 1980). Though statistical methods will not provide a 100 percent efficiency for avalanche forecasting they provide the avalanche forecaster with a framework from which decisions based on conventional methods may be modified. As noted by LaChapelle (1977) statistical methods are not often used, with preference given to numerically based quantitative precipitation prediction models as an aid in avalanche forecasting. Such models have been developed to improve mountain weather forecasts (Daly, 1978; Gigliotti and Parent, 1978). Associcated with this have been attempts to identify synoptic weather patterns associated with snowfall production and snow avalanching (Hutcheon and Lie, 1978; Grischenko and Tokmakov, 1978; Fitzharris and Schaerer, 1980; Nakajima, 1981). In addition to attempts to improve mountain forecasts and identify critical synoptic situations some process oriented models (Tesche and Yocke, 1976, 1978; Judson et al., 1980; Gubler, 1980) have been developed for simulating or predicting potential avalanche release.

(5) Snow structure and snow mechanics: A vast amount of literature dealing with the mechanics of snow exists (Mellor, 1977; Salm, 1982). Some of this is of direct relevance to snow avalanche research (Perla, 1969, 1971; Brown et al., 1972; Lang and Brown, 1973; Curtis and Smith, 1974; McClung, 1977, 1980; Sommerfeld, 1969, 1971, 1980; Sommerfeld and King, 1979; Armstrong, 1980; Singh and Smith, 1980; Yamada, 1980; Perla et al., 1982; Conway and Abrahamson, in press) but as outlined by Lang and Dent (1982) much of the literature is given over to consideration of snow trafficability and snow surface resistance, and studies of snow and ice properties. In the general area of snow metamorphism Colbeck (1982a) has suggested a modification of the snow metamorphism terminology originally developed by Sommerfeld and LaChapelle (1970) on the grounds that equitemperature metamorphism may also take place in the presence of a temperature gradient less than that generally required for temperature gradient recrystallisation in the snowpack. The importance of snow structure in snow avalanching research has been recognised for some time

(Seligman, 1936). The majority of studies to date, however, have been concentrated in areas of continental climates. These works (de Quervain, 1958; Martinelli, 1971; Trabant and Benson, 1972; LaChapelle and Armstrong, 1976; Armstrong, 1977, 1980; Bradley et al., 1977; Colbeck, 1982b) have highlighted the importance of temperature gradient forms in the snowpack. This has subsequently directed research to consider the exact processes and patterns involved in temperature gradient crystal growth (Akitaya, 1975; Voitkovsky et al., 1975; Marbouty, 1980; Adams and Brown, 1982, 1983). In addition to the importance of temperature gradient development and its implications for snowpack instability, some evidence has been presented of the role that melt freeze layers play in avalanche release in areas dominated by maritime climates (Moore, 1975; Ward, 1980). One of the basic problems facing avalanche research is the incorporation of snowpack structure data into an objectively based avalanche forecasting scheme. Attempts have been made by Good (1975) to quantify crystal types using factor analysis while LaChapelle and Ferguson (1980) have produced a stability index based on snow properties for inclusion in avalanche forecasts.

(6) Documentation of avalanche accidents: A common approach to assessing the factors involved in avalanche formation is the compilation of avalanche accident reports (Gallagher, 1967; Williams, 1975a, 1975b; Stethem and Schaerer, 1979, 1980). These document not only conditions leading up to avalanche release and probable cause of release but also make recommendations on how future hazardous situations may be avoided or suitable adjustments to the hazard made.

(7) Control of avalanche effects by structures and/or artificial release: Structures of various types have been used in Europe for at least 300 years to protect highways, railroads and inhabited areas from snow avalanches. European literature dealing with control of avalanche effects by means of structural engineering is therefore vast (Frutiger, 1977). Structural controls are also used widely in the United States (Schaerer, 1960, 1962, 1966; Frutiger and Martinelli, 1966; Mears, 1981). In the absence of structural control, artificial avalanche release, usually by explosive, is used for avalanche control (Gardner and Judson, 1970; Mellor, 1973). Some early work in the United States on the effects of explosives on snow has been presented by Fuchs (1957). Perla (1978a) has reviewed the use of explosives on snow for avalanche control by artificial release for North America, while Perla and Everts (1983) have evaluated the use of helicopter bombing and preplanted charges in avalanche control.

(8) Geomorphic significance of avalanches: Though literature on snow avalanche formation, release and mechanics implies that snow avalanches have little contact with the ground (Gardner, 1983) they have been shown to be important geomorphic agents for slope erosion (Matthes, 1938; Davis, 1962; Markgren, 1964; Gardner, 1970, 1983; Luckman, 1977) and deposition (Rapp, 1959; Caine, 1969; Gardner, 1970; Luckman, 1978). Associated with the geomorphic activity of avalanches is a distinct suite of landforms such as avalanche boulder tongues (Rapp, 1959; Gardner, 1970; Luckman, 1978), debris tails and avalanche mounds (Peev, 1966) and avalanche impact landforms (Fitzharris and Owens, 1980; Corner, 1981). Such diagnostic

features are often used as aids in delimiting the extent of avalanche runout (Ives et al., 1976).

1.3 AVALANCHE RESEARCH IN NEW ZEALAND

Though Smith (1947) noted snow avalanche related problems in the construction of the Homer Tunnel, Milford Road Fiordland, it has not been until recently that the general avalanche hazard has been formally recognised (Dingwall, 1976, 1977). Bovis (1977a) has also considered snow avalanches and the implications for public safety in New Zealand. New Zealanders' general perception of the snow avalanche hazard is considered by Simpson-Housley and Fitzharris (1979) to be far removed from reality, while LaChapelle (1979) and Owens and O'Loughlin (1979) have stressed the need for realisation of the snow avalanche hazard in New Zealand. The human and physical adjustments to the snow avalanche hazard in New Zealand have been discussed by Prowse et al. (1981).

While these papers are valuable in setting a suitable context for the awareness and perception of the hazard and the formulation of policies and procedures to deal with the hazard, there has been little written on the processes of avalanche formation in New Zealand. This contrasts markedly with areas overseas for which accounts of the contributory factors involved in avalanche formation are available.

Some consideration has been given to avalanche terrain. This however, has been largely from a hazard mapping point of view (Hatherley-Greene, 1978; Waters, 1980; Fitzharris and Owens, 1980) with little work on the specific effects of terrain on avalanche occurrence. McNulty (1983) used much of the criteria established by overseas workers to map avalanche terrain for a part of the northern Craigieburn Range. For the same area Conway (1977) and McIntosh (1979) used vegetational evidence to map avalanche paths with particular emphasis on vegetation damage in avalanche runout zones. Fitzharris (in press) tested the applicability of a range of theoretical and empirical models for prediction of avalanche runout distance for a small number of paths in the Mt Cook area. Though firmly established as a technique for the use in planning and siting of activities in the alpine zones of overseas areas, these methods have not yet been fully tested in New Zealand for a wide range of avalanche terrain.

Fitzharris (1976) presented the first detailed account of meteorological and snowpack variables involved in a specific avalanche event in the Mt Cook area for New Zealand. Prior to this the only studies specifically directed at snowpack and snow accumulation processes, but with no implications to avalanching being drawn, were those of Heine (1962) and Morris and O'Loughlin (1965). It has not been until recently, following the publication of the report by LaChapelle (1979), that studies of snowpack and meteorological processes with implications to snow avalanche research or related specifically to snow avalanche formation and occurrence have appeared in the New Zealand literature (McNulty and Fitzharris, 1980; Weir and Owens, 1980; Owens and Prowse, 1980; Prowse, 1981; Ho, 1982; Moore and Marcus, 1983; Owens et al., 1983). Conway and Abrahamson (in press) have departed from the general theme of these studies and taken a rigorous approach to the mechanics of slab avalanche release in the Mt Cook area.

From the studies conducted to date, it has become increasingly apparent that different snow climates exist throughout New Zealand with associated different snowpack development processes. The importance to be placed on different snowpack and weather factors for assessing snowpack stability and forecasting avalanches has therefore become quite evident (Prowse and Owens, 1984). Avalanche forecasting in New Zealand is presently based on conventional causal-intuitive methods with forecasters relying heavily on inputs of meteorological data (Fitzharris et al., 1983). To date no statistically based precipitation or avalanche forecasting techniques have been developed to aid avalanche forecasting.

1.4 REASONS FOR STUDY AND AIMS

While specific avalanche studies have partly fulfilled some of the proposed research objectives of LaChapelle (1979), no broadly based study which considers most aspects of avalanches for a given geographic area has yet been carried out in New Zealand. A study of this nature is essential if a sound information base for avalanche forecasting and the prediction of avalanche runout distance for planning purposes is to be established. The broad objective of this research is therefore to establish for the Craigieburn Range the factors involved in snow avalanche formation, release and runout. While this is a reconnaissance study, comparison with other snow avalanche environments will allow some fundamental features of snow avalanching in New Zealand to be elucidated. The specific aims of this thesis are:

(1) to identify and map avalanche paths paying particular attention to factors affecting avalanche formation and

frequency, and maximum extent of runout;

(2) to document avalanche events and investigate the role of meteorological processes in their occurrence;

(3) to outline the synoptic weather patterns and atmospheric processes associated with heavy snowfalls and avalanche release;

(4) to consider seasonal and short term developments in snowpack structure as they relate to avalanche formation and release.

1.5 ORGANISATION

The remainder of the thesis is organised into six chapters, four of which deal with the specific aims outlined above. The study area, methods of data collection and the statistical techniques used in this study are outlined in Chapter 2. The effect of avalanche terrain on avalanche formation and frequency, characteristics of the avalanche terrain, a consideration of avalanche dynamics and the evaluation of empirical and theoretical models for predicting avalanche runout distances are dealt with in Chapter 3. Chapter 4 assesses the influence of a range of independent meteorological variables on avalanche occurrence. An objective statistical technique for identifying those variables which in combination best discriminate between a set of dry, wet and non-avalanche days is assessed in the latter half of this chapter. The role of synoptic scale weather patterns and atmospheric processes in snowfall production and avalanche occurrence are evaluated in Chapter 5. Chapter 6 deals with snowpack structure, the factors controlling it, and its role in actual and potential avalanche release. A review of the main findings of this research and suggestions for future research appears in Chapter 7.

CHAPTER TWO

STUDY AREA AND METHODS

2.1 STUDY AREA LOCATION AND CHOICE

The Craigieburn Range lies east of the main divide of the Southern Alps of New Zealand by approximately 20 km and runs north-south for a distance of 26 km. The centre of the range is located at 43°10'S, 170°40'E (Figure 2.1). The eastern side of the Craigieburn Range was chosen as the study area for three main reasons. Firstly four skifields located here provide good access to the seasonal snow zone. Two of these fields were identified by LaChapelle (1979) as possessing significant avalanche problems. Secondly some basic information regarding snow avalanches already existed for parts of the Craigieburn Range, namely Porter Heights skifield and Broken River-Allan's Basin area. Associated with this latter area, the Forest Research Institute (FRI) has maintained a climate station at the Broken River skifield (BR) 1550 m a.s.l. since 1966 (Figure 3.4b). This is the only daily serviced year round climate station above the winter snow- line in New Zealand. Finally as there is a large number of personnel in the Craigieburn Range during winter, especially on skifields, this offered advantages from the point of setting up a widespread avalanche observation network.

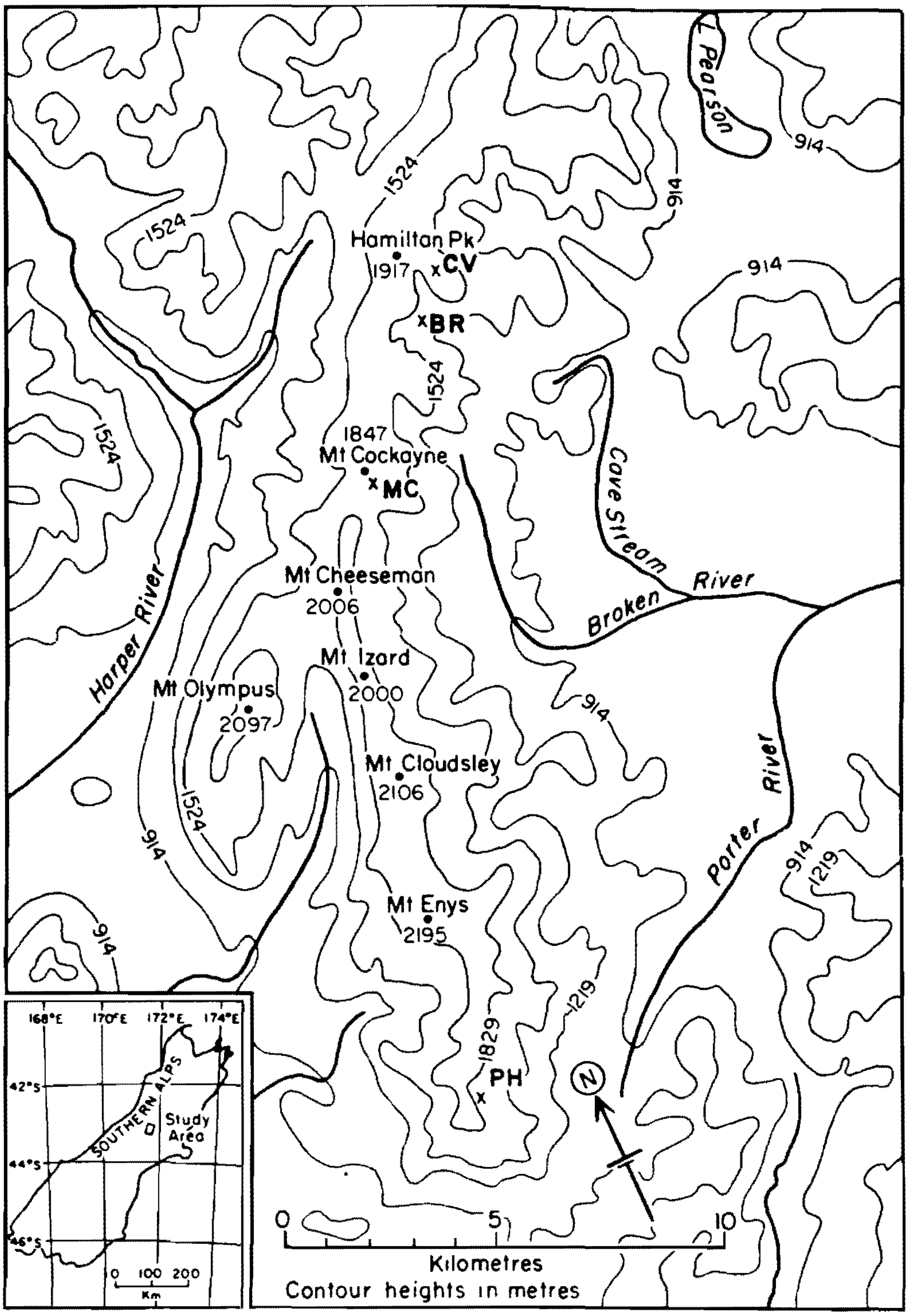


Figure 2.1: Location of Study Area

PH - Porter Heights Skifield

MC - Mt Cheeseman Skifield

BR - Broken River Skifield (location of meteorological site)

CV - Craigieburn Valley Skifield

2.2 PHYSICAL GEOGRAPHY

2.2.1 Geology-Geomorphology

The lithology of the Craigieburn Range is dominated by strongly indurated sandstone of the Torlese supergroup. These rocks are mainly of Triassic to Jurassic age. Gage (1980) has noted that beds are often tilted and faulted with numerous shatter belts. Such shatter belts are evident in the field as extensive tracts of bedrock showing complex jointing systems. Valley sides are typified by extensive talus slopes with upper slopes in the range of 30-40^{o}. Many of these slopes remain active with the highly jointed bed-rock outcrops supplying debris.

Evidence of former cirque glaciation abounds in many of the upper valleys of the Craigieburn Range (Chinn, 1975) with hummocky cirque moraines at the lower limit of upper valley bowl shaped basins. Though valley glaciation in the Late Pleistocene would have been active little evidence exists today owing to contemporary mass movement and fluvial erosional processes.

2.2.2 Soils

Soils in the upper valleys of the Craigieburn Range are often thin with only skeletal horizon development. The soils are classified as Spencer under grassland and Bealey under forest cover. Both have a skeletal, steepland yellow brown appearance. The upper slopes on which snow avalanches occur are generally devoid of soils.

2.2.3 Vegetation

Grasslands in the Craigieburn Range extend up to 1850 m and are dominated by Chionochloa pallens and C. macra . Much of the northern end of the Craigieburn Range is covered by beech forest (Nothofagus solandri var. cliffortioides) with the upper limit of timberline at approximately 1370 m a.s.l. (McCracken, 1980). In the southern end of the Range, beech forest is mainly restricted to isolated areas below 1000 m. As avalanches frequently descend below treeline extensive tracts of beech forest may sustain damage due to avalanche. This is especially so in the Craigieburn Valley area (Conway, 1977 and Burrows et al., 1979).

2.2.4 Climate

The general climate of the Craigieburn Range is influenced by the progression of anticyclones over the South Island separated by troughs of low pressure (de Lisle, 1969). The broader synoptic patterns influencing the avalanche climatology of the Craigieburn will be discussed in a later chapter.

As all avalanche starting zones occur above 1500 m in the Craigieburn Range this study is predominantly concerned with climate of the area above 1500 m. McCracken (1980) has summarised climate data for the Broken River Ski Basin (BR) at 1550 m for the years 1966 to 1979 (Table 2.1). The coldest month at the BR site is July followed by August and June. Winds reach a mean daily maximum in September with August and October displaying equal daily speeds. Mean daily windspeeds at BR during strong north-west flows of 10 ms^{-1} and gusts up to 67 ms^{-1} at 1800 m a.s.l. have been

Table 2.1: Average Mean Monthly Climate Data Broken River Skifield 1966-79 (After McCracken, 1980)

	J	F	M	A	M	J	J	A	S	O	N	D	Annual mean of total
Temperature (^oC)													
Absolute max.	24.0	25.0	22.4	18.2	12.9	11.1	9.1	13.5	14.6	14.5	20.2	20.2	
Mean max.	12.8	14.1	11.6	8.2	4.8	2.5	1.4	1.7	3.3	5.9	8.6	11.5	7.2
Mean	8.7	9.7	8.4	5.4	1.9	-0.3	-1.4	-1.1	0.1	2.5	4.9	7.2	3.8
Mean min.	4.7	5.5	4.4	1.8	-1.0	-3.4	-4.0	-3.9	-2.7	-0.9	1.1	3.3	0.4
Absolute min.	-3.3	-5.8	-7.0	-8.6	-13.1	-11.5	-12.4	-14.5	-12.9	-10.3	-9.0	-5.7	
Grass min.	2.9	2.0	1.3	-1.2	-4.2	-6.6	-7.1	-7.1	-4.7	-2.0	0.6	1.1	-2.2
No. yr in record	10	9	11	11	12	10	10	5	6	8	8	9	
Days ground frost	6	7	9	16	25	27	28	27	22	19	12	10	208
Days screen frost	3	3	4	9	18	26	29	27	24	19	11	6	179
Rainfall (mm)													
Mean total	124	100	100	167	131	93	102	144	193	153	128	151	1586
No. yr in record	14	14	14	14	11	9	6	6	6	8	10	12	
Windspeed													
Mean daily (m/sec)	2.4	2.3	2.5	2.5	2.5	2.7	2.9	3.0	3.4	3.0	2.8	2.5	2.7
Radiation													
1969-74													
Mean daily (MJ/m^2)	21.3	19.0	14.1	9.7	5.5	4.2	4.7	8.5	12.7	18.0	20.5	21.9	4870

reported by Rowe (1968) attesting to the windiness of this environment. Precipitation amounts received at 1500 m a.s.l. are greatest in the months of September, October and August respectively. Such a trend may be related to the increased frequency of north-westerly flows during this period (Rowe, 1968). Morris and O'Loughlin (1965) noted that snow is likely to fall at any time of the year with approximately 33 to 36 percent of the precipitation above 1500 m being received as snow. Snow begins to accumulate on the ground above treeline on the average in late May or early June and reaches a maximum in September or October (Figure 2.2). Snow may remain on the ground from 5 to 9 months.

Though the general climate of the Craigieburn Range is considered 'warm' by Morris and O'Loughlin (1965) compared to other alpine areas, the winter months of June, July, August and September for the Craigieburn Range tend to be dominated by below freezing temperatures, above average precipitation amounts and general windiness above 1500 m. Prowse (1981) has considered the representativeness of the climate of the Craigieburn Range for the period 1975 to 1980 in relation to the climate of the area over the last two decades. He concludes that this period is generally representative of the period 1950 to present characterised by light snowfall and warmer temperatures. This contrasts markedly with the late nineteenth and early twentieth century as evinced by Burrows (1976). Though the recent climatological data may be temporally representative of the last 10 to 20 years, the question arises as to the spatial representiveness of the BR site. During periods of field work, differences in the amounts of precipitation received

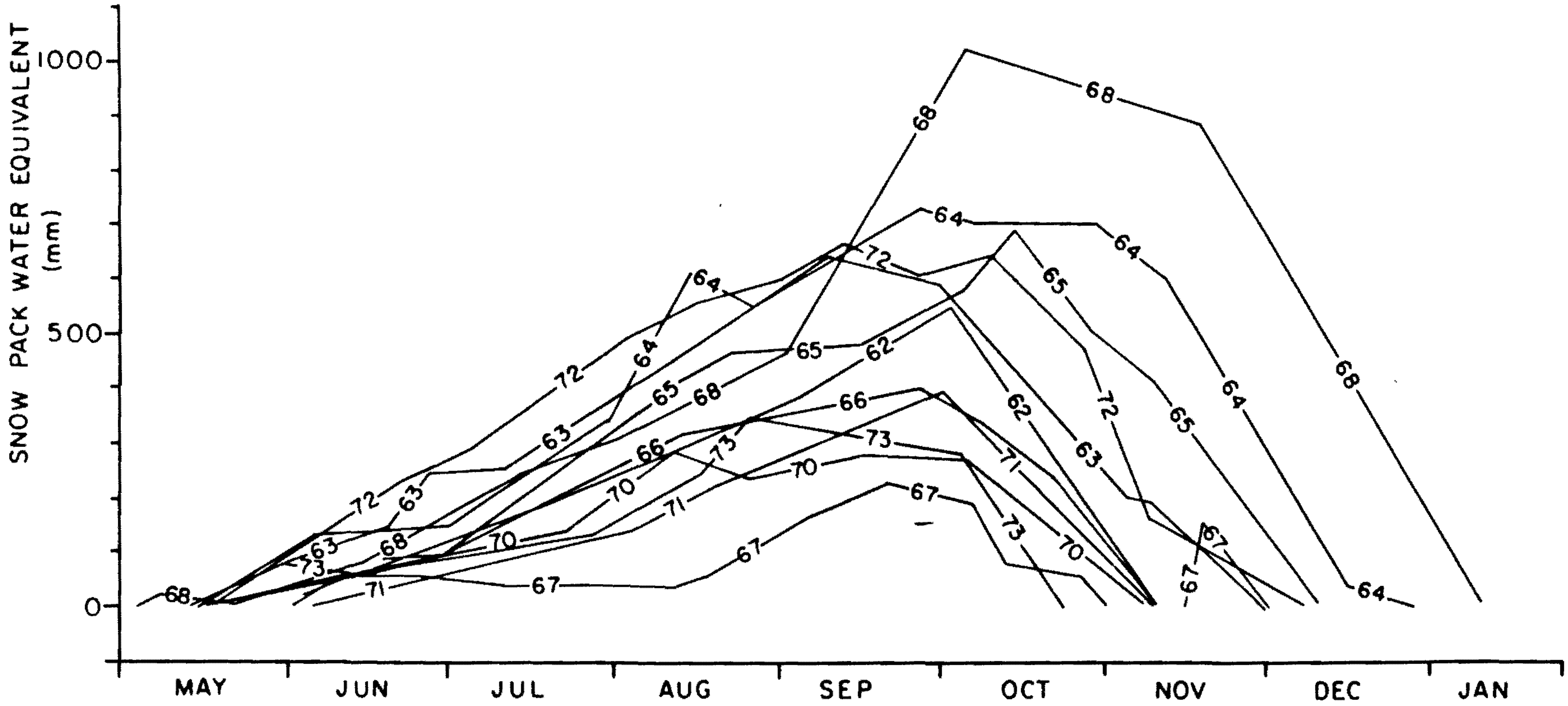


Figure 2.2: Total snowpack water equivalent for 1964-1973 Allan's Basin Craigieburn Range (After O'Loughlin, 1969; Prowse, 1981).

at either end of the range were noted. Areas in the northern part of the Craigieburn Range on some occasions received greater amounts of precipitation from the north-west, while areas in the southern parts of the range received greater amounts from the south to south-west direction. No comparative quantitative precipitation records exist to confirm or reject such a hypothesis. Until further evidence demonstrates anything to the contrary, the meteorological readings taken at the BR site will be considered as representative of the Craigieburn Range.

2.3 METHODS OF DATA COLLECTION

As the basic aim of this research is to assess terrain, meteorological and snowpack variables as factors in avalanche formation, data collection involved a variety of techniques. Briefly, data regarding terrain characteristics were collected during summer months by field survey, snowpack structure analysis was performed manually during the winter months of 1979, 1980 and 1981, while the majority of the meteorological data used in this study were automatically recorded at the BR site. Daily checks for instrument malfunction and for the purpose of instrument recalibration were made at this site by FRI staff. Where inconsistencies were notable in the daily meteorological records trends of the data were established from a lower elevation site at 914 m in the Craigieburn Forest Park (CF) (Figure 3.4b). Time checks were also used to adjust the readings accordingly. Initially an attempt was made to establish a meteorological site on Mt Cockayne at 1680 m a.s.l. (Figure 3.4b). Operational difficulties owing to the rigours of the environment and general inaccessibility during times of major storms when constant checks should have been made necessitated the abandonment of this site. Synoptic weather maps and upper air data were obtained from the New Zealand Meteorological Service. Specific methods and data used in the following analysis are detailed to a greater extent in the appropriate chapters. The following subsections however, outline the major instruments and methods used in data collection and some of the problems encountered.

2.3.1 Precipitation

Initially all precipitation data were to be collected using a Weather Measure propane heated snow gauge. This was located at 1680 m a.s.l. on Mt Cockayne. Problems associated with snow drifting and freezing of snow in the aperture necessitated the removal of this instrument. Heavy reliance was therefore placed on the precipitation record collected at BR. A Belfort weighing bucket dual traverse precipitation gauge was regularly maintained by FRI field staff. An anti-freeze salt solution was used in this gauge. The large 200 mm orifice of the Belfort gauge and the weighing bucket design are generally considered to minimize the effect of snow 'capping'. The periods for which data were required showed no evidence of 'capping' of the snow gauge. The Belfort gauge was equipped with an Alter shield subsequent to this study. Such shields have been shown to improve the catch efficiency of gauges (Goodison, 1979). Snow boards were installed for recording new snow depths and water equivalents. These however proved unsatisfactory as snow was easily eroded from these in windy periods during and following snow deposition. Additionally sheltered sites in which snow boards should be set up are generally not available at the elevation of the meteorological site. Total snowpack depth was measured using a total snow stake. Measurements of new snow depth were also taken from this. Even though measurements from these are also susceptible to inaccuracies owing to wind effects they are considered adequate in the absence of other reliable techniques. As this study relies on a single point measurement of precipitation it must be assumed that this site is representative of total precipitation increments in the Craigieburn Range. Ideally a network of measurement sites should be used in a study of this nature considering the spatial variabililty of snow accumulation in rugged terrain (Grant and Rhea, 1974; Hendrich et al., 1979). Such a network of recording sites has subsequently been established in the Craigieburn Range in conjunction with the New Zealand Mountain Safety Council's Pilot Avalanche Forecasting Scheme (Fitzharris et al., 1983).

2.3.2 Air Temperature and Humidity

A Casella hygrothermograph was located at the BR site in a Stevenson screen 1.5 m above ground level. A short incomplete record was obtained from a Weather Measure hygrothermograph located in a Stevenson screen at 1680 m on Mt Cockayne. Constant filling of the screen by snow during precipitation events associated with high winds, however, rendered this instrument and record completely unreliable. Reliance was therefore placed on the hygrothermograph located at 1550 m at the BR site. Periods for which temperature and humidity records at this site were incomplete were supplemented by records at 1550 m on Mt Cockayne from an identical instrument or the appropriate lapse rates of Prowse (1981) were applied to temperature records collected at 914 m. The normally low error of $1^{o}C$ for the bimetallic temperature sensors on the Casella hygrothermographs may increase markedly if the sensor becomes iced. Daily temperature checks were performed by FRI field staff. Any discrepancies between temperature traces and point temperature checks were subsequently adjusted. An Assman aspirated hygrometer was used to check humidity readings from the hygrothermograph. As the Casella hygrothermograph relies on a hygroscopic hair as the humidity sensor, it may be subject to errors due to icing hence affecting the mechanical response of the hygroscopic hairs. Presence of ice on the hairs may also result in slightly different rates of vapour transfer. With icing, vapour transfer will be over ice as opposed to water. Prowse (1981) however regards this problem as minimal as the difference between saturation vapour pressure over ice versus over water reaches a maximum of only 0.27 mb at $-10^{o}C$ with negligble pressure differences near the freezing point. If sensors are also not continually ventilated lags of up to an hour may result between the surrounding air and the instrument (Bergen, 1968). Periods in which the sensors may be susceptible to such errors were during storms when screens became blocked with snow. Though constant attempts to clear screens from blocking snow were made some large irregularities did appear in the humidity record. For these periods trends in humidities were established from lower elevation stations on Mt Cockayne.

2.3.3 <u>Radiation</u>

A Feuss bimetallic actinograph situated at 914 m was used to measure global radiation. Prowse (1981) has compared the ratio of global radiation received at 1550 m and 914 m and has produced ratios for the various winter months. These ratios were used to adjust the total global radiation recorded at 914 to that of 1550 m as the bimetallic recorder was removed from the BR site prior to this study. The corrected values of global radiation were used in the analysis.

2.3.4 <u>Wind Speed and Direction</u>

A Lambrecht anemograph located at 914 m was initially envisaged as supplying all wind direction and wind speed data. Analysis of wind direction and speed records from an identical anemograph located at 1500 m on Mt Cockayne however showed some gross discrepancies between the two sites. A chi-square test of the association of wind direction at 1500 m with that at 914 m showed no statistical agreement between wind directions recorded at each site. Reasonable correlations however were obtained between hourly windspeed readings for the two sites. The anemograph at 914 m was therefore abandoned as a source of wind flow information. As the nature of the study necessitates good reliable ridge top wind data it was considered that the record from the anemograph on Mt Cockayne at 1500 m would suffice. As this was not installed until August 1980 no record was available for 1979 and the early part of the 1980 winter. Data from the 1980-1981 record were therefore compared with upper wind data at the 800 mb level over Christchurch for wind speed and direction. Acceptable correlations between the 800 mb level winds and that at the Mt Cockayne site were obtained. Consequently, the 800 mb wind speed and directions above Christchurch were used as surrogates for the free flow of wind over the Craigieburn Range. As average ridge top elevation of the Craigieburn Range is approximately 1800 m these data was also considered to be a reasonable measure of the average wind speed and wind direction at ridge top. Symbols used in displaying windspeed data appear in Appendix 1.

2.3.5 Upper Air Data and Synoptic Weather Maps

As the majority of the weather systems affecting the Craigieburn Range come from the west, information regarding general atmospheric thermal and moisture structure should be derived from a west coast situation. However, the nearest New Zealand Meteorological Service station to the Craigieburn Range from which regular radiosonde soundings are made is Christchurch International Airport, positioned 80 km east of the Craigieburn Range. Information collected at this site may therefore be subject to lee effects. This is especially so under strong westerly conditions when upper level air may become decoupled from cool air ponded over the Canterbury Plains. Adiabatic warming may also occur as a result of the Fohn effect. Freezing level, atmospheric temperature and moisture structure, height of constant pressure surfaces, atmospheric layer thicknesses and wind flow data were subsequently considered with this fact in mind. Similar information from Invercargill, located 470 km south-west of Christchurch, was also considered and in some instances may be a better indicator of free atmospheric conditions, although some lee effects may also be experienced here.

Information used in the analysis of weather patterns was also obtained from the New Zealand Meteorological Service. Surface and 500 mb charts at a scale of 1 to 2 million were analysed for synoptic scale development of avalanche producing storms. Surface charts were available at 6 hourly intervals whereas the 500 mb charts were available only at 12 hourly intervals.

2.3.6 Snowpack Structure

Snowpack structure data were collected by the standard method of snowpit analysis. This involves the noting of snow depth, layer thickness, density, temperature, crystal type, crystal size and hardness. This information was subsequently plotted as a snow profile. The procedures involved in snowpit analysis are outlined in Appendix 2.

2.3.7 Avalanche Classification and Occurrence

Avalanches were classified using a similar approach to that of the United States Forest Service (Perla and Martinelli, 1976 and Ives and Armstrong, 1976). This classification system was chosen in preference to the system proposed by the International Commission on Snow and Ice (de Quervain et al., 1973) for two reasons. Firstly the avalanche reporting forms distributed by the Avalanche Committee of the New Zealand Mountain Safety Council follows closely the procedure of avalanche classification of the United States Forest Service and secondly this classification is easy to comprehend for untrained field observers. The avalanche reporting form and notes accompanying the form appear in Appendix 3.

The determination of the effect of causative factors of snow avalanche formation and release on avalanche occurrence inherently involves the collection of reliable avalanche occurrence data. Systematic documentation of type, size, location and time of occurrence is mandatory. To achieve this an avalanche observation network was established throughout the Craigieburn Range. The main observation points were from skifields and the Craigieburn Forest Park. Staff at these areas were circulated with the avalanche recording sheets appearing in Appendix 3. A regular discipline of observation and recording was emphasised to the staff involved. Regular visits to the various skifields were made to assess the efficiency and regularity of observations. Invariably only bare details concerning the nature of the release were noted such as time of release, morphology, size, and type of release in addition to the basic path data. This precluded any analysis on actual avalanche magnitudes such as that performed by Shimizu (1967). Forms were collected on a regular basis and subsequently processed. The initial discrimination between soft and hard slab avalanches was abandoned because the nature of the debris was often difficult to discern over long distances. Soft and hard slab avalanches were therefore classified together as slab avalanches, which comprised 40 percent of all avalanches recorded (Figure 2.3). Sixty six percent of all natural avalanches recorded occurred in the size range of magnitude 2 to 3 (Figure 2.4). Size two avalanches are considered large enough to cause threat to human life. (Appendix 3)

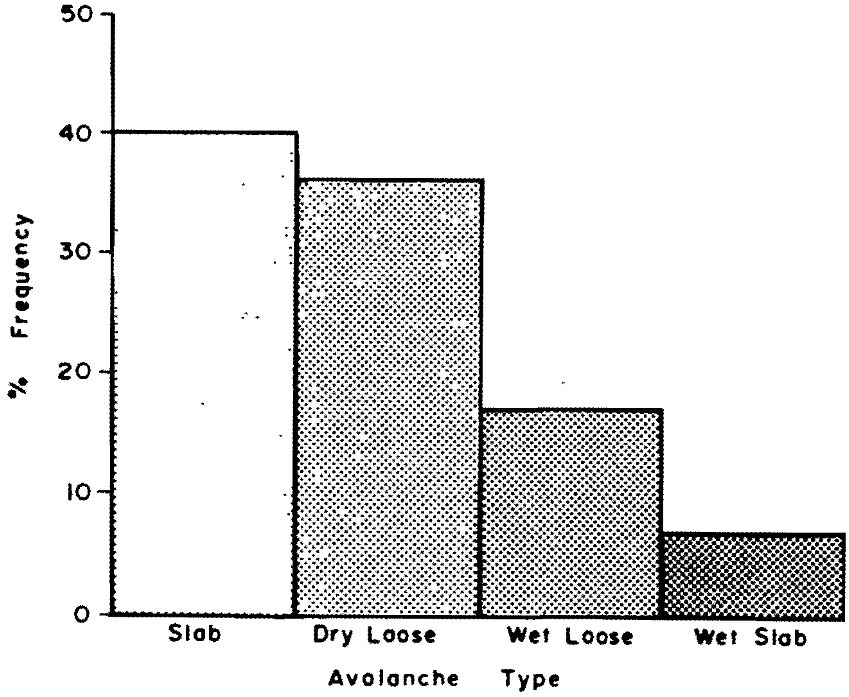


Figure 2.3: Percentage frequency of avalanche occurrence by avalanche type.

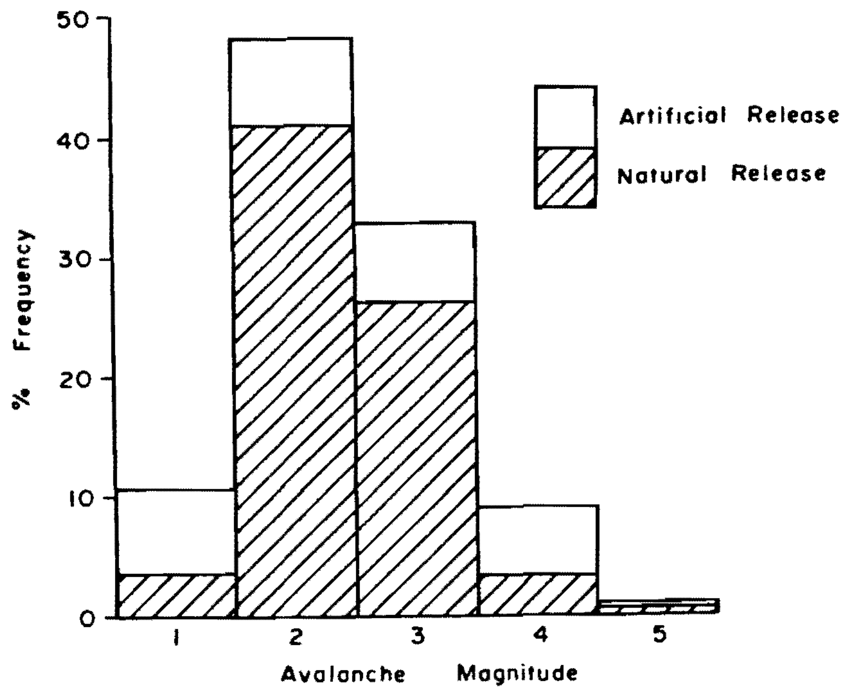


Figure 2.4: Percentage frequency of avalanche occurrence by avalanche magnitude (refer Appendix 3)

2.4 STATISTICAL ANALYSIS

Data were subjected to statistical analysis to establish any significant differences or relationships between variables. For testing differences between samples the Student 't' test and F test were used to compare group means and variances respectively. The procedure followed was that as outlined by Davis (1973).

Simple linear regression and multiple regression analyses were used to examine the relationship between a dependent variable and a set of independent or predictor variables. These procedures require that; data be measured at an interval or ratio scale, each array of the dependent variable Y for a given combination of X values follows the normal distribution, the regression of Y and X values are linear and all the Y arrays have the same variance. Homogeneity of variance is determined by examination of plots of residuals against the predicted Y for visible patterns of irregularity (Kim and Kohout, 1975 and Norcliffe, 1975). For multiple regression some problems may exist with multicollinearity. This may arise if one of the independent variables is a perfect linear function of one or more other independent variables in the equation. Coefficients may therefore not be able to be determined uniquely. Problems with inversions of the correlation matrix also arise with intercorrelations in the 0.8 to 1.0 range. To avoid the inherent problems of collinearity correlation matrices were examined and where appropriate variables displaying intercorrelations of greater than $r = \pm 0.8$ were removed from the subsequent analysis. All simple and multiple regression analyses performed and presented in this analysis comply with the above restrictions and assumptions.

Discriminant function analysis was used to distinguish between two or more groups of cases. For distinction between groups, a set of discriminating variables is used. The main objective of such a technique is to weight and linearly combine the discriminating variables so that the groups are forced to be as statistically distinct as possible. Discriminant analysis attempts to do this by forming one or more linear combinations of the discriminating variables. The discriminant function takes the form of

$$Di = di_1 Z_1 \; di_2 Z_2 + \ldots\ldots + di_n Z_n + C \qquad (2.1)$$

where Di is the discriminant score, di_1 is the discriminant coefficient for variable 1, Z_1 is the raw value of variable 1 etc, and C is a constant. This technique will be discussed in more detail in Chapter Four.

CHAPTER THREE

AVALANCHE TERRAIN ANALYSIS

3.1 INTRODUCTION

Perla and Martinelli (1976) have cited terrain as one of the three major contributory factors in avalanche formation. This chapter is concerned with the influence of avalanche terrain characteristics on avalanche formation, frequency and runout distance.

3.2 PREVIOUS WORK

The avalanche path, the basic unit of avalanche terrain is composed of three distinct zones (Martinelli, 1974) (Figure 3.1). Starting zones generally possess angles in the range 30 to 45 degrees (Table 3.1) while avalanche tracks which may be confined or unconfined (Figure 3.1) are commonly inclined at angles of 20 to 25 degrees (Martinelli, 1974; Miller et al., 1976), though there are some major exceptions in areas which have been oversteepened by previous glacial activity (Fitzharris and Owens, 1980). Schaerer (1972, 1975b) has also noted that for avalanches to maintain their momentum, avalanche tracks must be at least 25 to 28 degrees. Studies dealing with the starting zone have mainly dealt with establishing the role of a range of factors such as slope angle, aspect and general nature of starting zone morphology in avalanche formation. The avalanche track has traditionally received little attention in the literature, although Matthes (1938), Rapp (1959) and

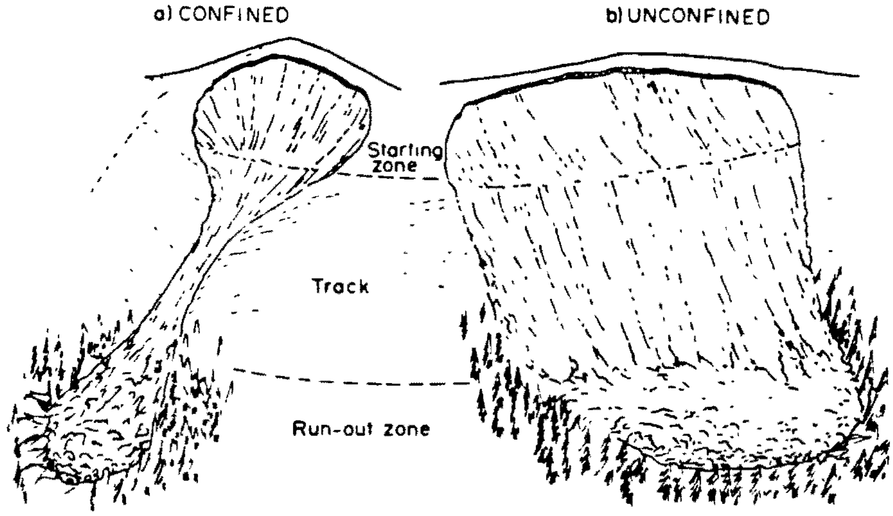


Figure 3.1: Definition of avalanche path terminology

Table 3.1: Summary of Starting Zone Angles

Author	Location	Starting Zone Angles
Martinelli (1974)	Rocky Mountains Colorado	20-45°
Perla and Martinelli (1976)	Various	30-45°
Miller et al. (1976)	San Juans Colorado	60% 20-40°
LaChapelle and Armstrong (1976)	San Juans Colorado	49% 34-41°, 72% 30-45°
Perla (1977)	Whistler British Columbia	Mean of 38.3°
Hatherley-Greene (1978)	Otira Valley New Zealand	24-45°
Weir (1979)	Mt Hutt New Zealand	30-40° > 40°
Waters (1980)	Arthurs Pass New Zealand	25-45°
Stethem and Perla (1980)	Whistler British Columbia	40.2°
Fitzharris and Owens (1980)	Milford Road New Zealand	Mean of 40° Range 29-50°

Markgren (1964) have noted the significance of this zone as an area of debris transportation and geomorphic work.

The runout zone, the area in which the avalanche comes to rest and generally characterised by angles of 5 to 15 degrees (Martinelli, 1974; Miller et al., 1976), is of particular interest in avalanche terrain studies. This is the area in which maximum damage to any human related activity by avalanche is likely to occur. Consequently, avalanche hazard maps have been used for defining the area and extent of the runout zone and likely areas of maximum risk (Freer and Schaerer, 1980; Hestnes and Lied, 1980; Ives and Plam, 1980). Where relevant vegetation damage, historical or geomorphic evidence is unavailable, various theoretical models (Voellmy, 1955; Salm, 1966, 1979; Lang et al., 1979; Perla et al., 1980), developed on the assumption that flowing snow acts as a fluid, may be used. These runout distance predictions are based on the response of avalanche flow to various friction coefficients, the values of which may vary between areas given contrasts in avalanche terrain and snow climates (Owens et al., 1984). Coefficient values therefore require recalibration for the use in different geographic areas (Buser and Frutiger, 1980; Martinelli et al., 1980). In contrast empirical models developed by Bovis and Mears (1976) and Lied and Bakkehoi (1980) use a combination of avalanche terrain characteristics to predict avalanche runout distance. The philosophy behind these empirical approaches is that runout distance is solely the response of the flowing snow to various terrain features. Schaerer (1972), has similarly shown that avalanche frequency may be influenced by certain terrain features.

3.3 AIMS

The aims of this analysis follow closely those of the studies cited above. These are:

(1) to describe avalanche terrain characterisitcs of the study area and consider the role of these in avalanche formation and frequency of release;

(2) to examine the applicability of the theoretical models developed by Voellmy (1955) and Perla, Cheng and McClung (1980) (hereafter referred to as the PCM model) for predicting avalanche runout distance in the Craigieburn Range;

(3) to present an empirical model for avalanche runout distance prediction based on terrain variables.

3.4 DESCRIPTION OF AVALANCHE PATH CHARACTERISTICS

3.4.1 Methods

Good evidence for identification and location of avalanche paths exists in the northern end of the Craigieburn Range. Here where large areas of beech forest still remain, avalanche runout zones can be easily identified by vegetation damage (Figure 3.2). Conway (1977) and McIntosh (1978) have mapped avalanche paths according to the vegetation evidence for this area. On the eastern side of the Craigieburn Range from Mt Cheeseman south to Porter Heights skifield (Figure 2.1), where vegetation is sparse or completely lacking, identification of avalanche paths is very difficult. Field investigation in the summer months revealed only meagre information as to the active avalanche paths. Debris found on scree slopes was difficult to

Figure 3.2: Avalanche damage in Nothofagus solandri var. cliffortioides Craigieburn Valley Skifield. Note vegetation stripped to below crown on the tree in the middle right foreground.

identify as either avalanche, debris slide or rockfall in origin, though in some cases (Figure 3.3) perched debris provided good evidence. Heavy reliance was therefore placed on winter observation of actual avalanche paths. These were subsequently mapped at a scale of 1:15840 (Figures 3.4a,b,c) This map represents the avalanche paths identified as active in the years 1979 to 1981. However, paths which may have a lower frequency higher magnitude occurrence of avalanches may not have been identified during the study period. This was proven correct in 1982 when widespread avalanching was reported from an intense storm depositing up to a metre of new snow. Those areas identified were not used in the subsequent analysis of characteristics of avalanche terrain or analysis of runout distance Analysis is therefore limited to those paths observed as being active in the years 1979-1981.

Schaerer (1972), noted that determination of terrain variables for avalanche paths may be achieved by map analysis using special topographical maps. He found that conventional maps of 1:50000 were of insufficient accuracy and used maps of 1:5000 and 1:2500 with contour intervals of 7.6 m and 5.0 m respectively. As the only maps available for the Craigieburn Range were 1:63360, 1:15840 and one map of a small area of the northern end of the range at 1:4000, slope profiling was used in terrain analysis of 30 paths. Comparison of statistics extracted from the profile analysis of the 1:15840 map of 30.5 m contour intervals (Table 3.2) reveal discrepancies between field survey and map analysis as exemplified by higher mean angles and greater variation in slope angles.

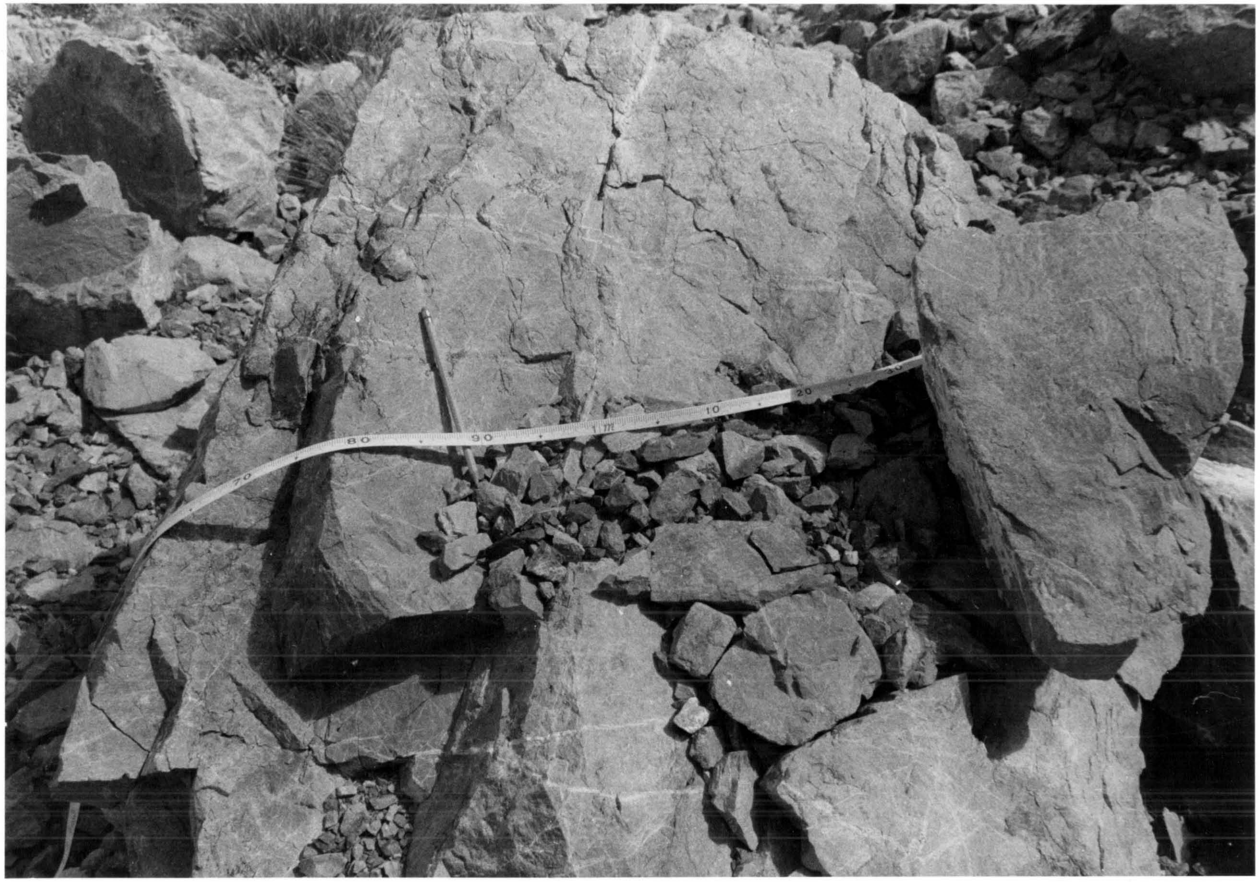


Figure 3.3: Perched avalanche debris, Middle Basin Craigieburn Valley Skifield. Note wide range in material size.

Table 3.2: Variation in Slope Angle Between Field Survey and Map Analysis at 1:15840 and 1:4000. (n = 30)

	Field Survey $\bar{\theta}$	s^2	1:15840 $\bar{\theta}$	s^2	1:4000 $\bar{\theta}$	s^2
Starting Zone	34.8	4.0	38.3	44.9	36.1	7.3
Track	29.5	6.2	32.1	20.2	31.6	8.4
Runout Zone	20.3	24.0	23.9	57.8	22.8	37.2

$\bar{\theta}$ = mean slope angle
s^2 = variance

3.4.2 Results

3.4.2.1 The Starting Zone

The upper limit of starting zones on the eastern side of the Craigieburn Range occur in an altitudinal band from 1616 to 2134 m. Mean starting zone for the 30 avalanche paths profiled was 34.8 degrees with a variance of 4.0 degrees (Table 3.2). This falls within the range of slope angles generally accepted as sufficient for avalanche release (Table 3.1). Four types of starting zone were identified. These were open lee slopes below ridges (Type 1) open bowl shaped lee slopes below ridges (Type 2), open slopes below a gully (Type 3) and gullies below ridges (Type 4) (Figure 3.5a, b, c). These starting zone types conform closely to the gullies and lee slopes identified by Schaerer (1972), the gullies and bowls noted by Martinelli (1974) and the depressions of Lied and Bakkehoi (1980) as preferential sites for avalanche release. The close association of avalanche starting zones with ridges and the fact that Type 1 starting zones have the highest frequency of release (Table 3.3) highlights the importance of open lee slopes as areas of snow redeposition and avalanche formation as noted by

Figure 3.5a: Type 1 starting zone. Large open lee slopes below ridges Hamilton Face, Craigieburn Valley Skifield.

Figure 3.5b: Type 2 starting zone. Open bowl shaped lee slopes below ridges Porter Heights Skifield.

Figure 3.5c: Type 3 and 4 starting zones. Open slopes below gullies left and gullies below ridge right. Tim's Basin, Craigieburn Forest Park.

Martinelli (1974) and LaChapelle (1978).

Table 3.3: Frequency of Avalanche Type with Starting Zone Type (Percentages)

	Avalanche Type				
Starting Zone Type	Dry Slab	Dry Loose	Wet Loose	Wet Slab	Total
Type 1	29.0	5.0	1.5	4.0	39.5
Type 2	7.0	2.0	1.5	3.0	13.5
Type 3	4.0	13.0	2.0	-	19.0
Type 4	-	16.0	12.0	-	28.0
Total	40.0	36.0	17.0	7.0	100.0

Starting zone aspects for the Craigieburn Range ranged from 30 to 240 degrees (Figure 3.6). Slope aspect may influence avalanche occurrence through its control on snow deposition and structural changes in the snowpack. Though easterly aspects generally predominate, total avalanche occurrence peaks on north-easterly and south-easterly aspects (Figure 3.7). As the majority of heavy snowfalls in the Craigieburn are associated with north-westerly wind flows, south-easterly aspects are favoured sites for snow deposition. Most avalanching on these aspects is made up of dry loose avalanches representing sloughing off of snow in the early parts of north-westerly storms. While slab avalanches are also important on these slopes, they have the greatest occurrence on north-easterly aspects which may represent wind loading during periods of south-to-south-westerly flow following deposition of snow from the north-west. Though south-easterly aspects are shady for the majority of winter and are preferential sites for the development of structural instability in basal layers, no preferential release of slabs on shady slopes, as noted by

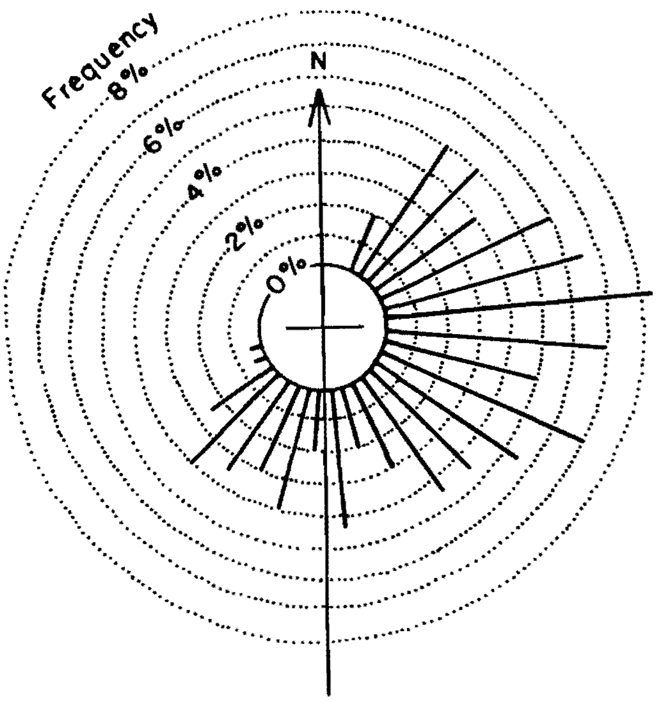


Figure 3.6: Range of starting zone slope aspects. Majority of starting zones occur with north-easterly to south-easterly aspects.

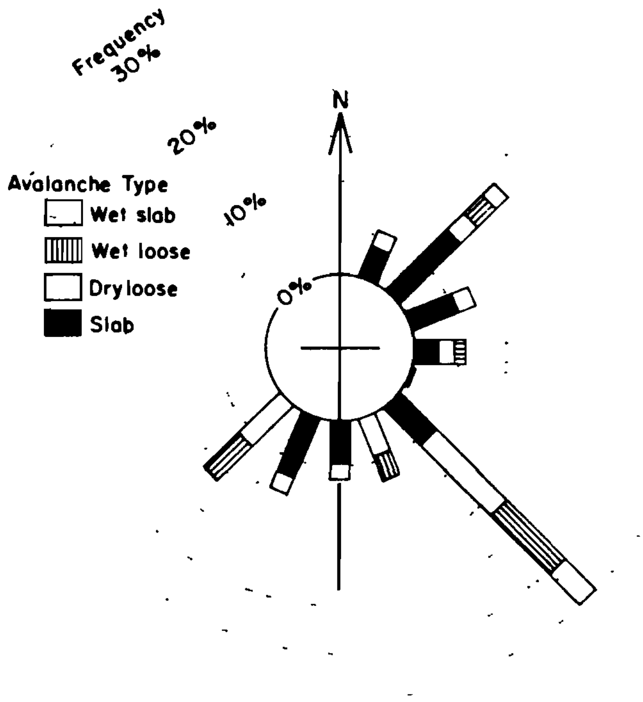


Figure 3.7: Percentage frequency distribution of avalanche type with slope aspect. Note majority of all avalanches occur on north-easterly and south-easterly aspects coincident with snow transportation from the south-west and north-west respectively.

Bradley (1970) for the Northern Rocky Mountains, Montana, was evident in this study. The majority of wet avalanches occurred on south-easterly to south-westerly aspects which is unusual as these are typically shady slopes. As 86 percent of all wet avalanche cycles were associated with rain, and wet avalanche starting zones are typified by gullies, precipitation would be preferentially channelled into these gully situations, thus increasing the likelihood of bond breakdown by free water in these areas over open bowl shaped areas. Additionally snow may remain on these shady aspects late into the season thus increasing the likelihood that any release from these aspects may be of the wet type as a result of generally warmer temperatures.

Redeposition of snow on the easterly slopes of the Craigieburn Range, which lie on the lee of strong westerly flows, will increase the likelihood of avalanche occurrence on these slopes. This agrees well with Martinelli (1974) who noted that lee slopes are more prone to avalanche release than windward slopes. Akkouratov (1966) and Kotlyakov and Plam (1966) have also noted lee slopes as prime sites for slab formation as snow drifting will be maximised in these areas under windy conditions. Periods of slab avalanching have been noted by Weir (1979) for the Mt Hutt Range coincident with snow redeposition on lee slopes. The aspect control on avalanche formation found in this study contrasts with that of Schaerer (1975b) who has found no variation in avalanching with slope aspect, and also Fitzharris and Owens (1980) who have reported a full range of aspects for avalanching in the Milford Road area, Fiordland. Contrasts in the role of aspect in avalanche formation for the

Craigieburn and Mt Hutt Ranges to that of the Milford Road area, is possibly a function of total storm precipitation amounts received. Large new snow increments are received in the Milford area with all slopes being heavily loaded while in contrast new snow increments for the Craigieburn and Mt Hutt Ranges are frequently less than 0.5m. Wind redistribution of these relatively small amounts of new snow is therefore required to load lee slopes for slab formation. Starting zone characteristics identified in this section were used to map areas of potential avalanche release (Figure 3.4a, b, c).

3.4.2.2 Avalanche Tracks

Average avalanche track slope in the Craigieburn Range for those paths sampled was 29.5 degrees with a variance of 6.2 degrees. The range of avalanche track slopes was wide, ranging from 27.0 degrees typical of the larger paths (Figure 3.8) to 34.0 degrees, typical of short steep valley side paths (Figure 3.9). The mean slope of the Craigieburn avalanche tracks is high compared to that noted by Martinelli (1974) for the Rockies, and higher than the slope suggested by Schaerer (1972) for avalanches to maintain momentum. Though avalanche tracks approach 34 degrees for the Craigieburn Range they are nowhere near as steep as those noted by Fitzharris and Owens (1980) for the Milford Road area. This is a consequence of the nature of the tracks in these two areas. Most tracks in the Craigieburn Range are developed on scree slopes while many of those in the Milford area possess steep plunging cliffs.

Three types of avalanche track were identified for the Craigieburn Range. Unconfined tracks make up 59 percent of the paths studied, while confined tracks and a combination

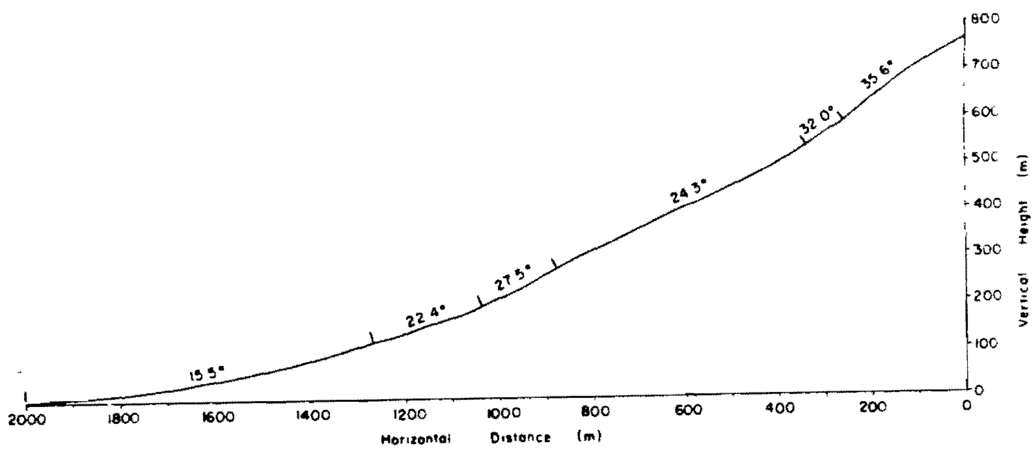


Figure 3.8: Longitudinal profile of Fourth Gut avalanche path, Middle Basin, Craigieburn Valley Skifield. This profile form is generally representative of those paths in excess of 400 m vertical relief.

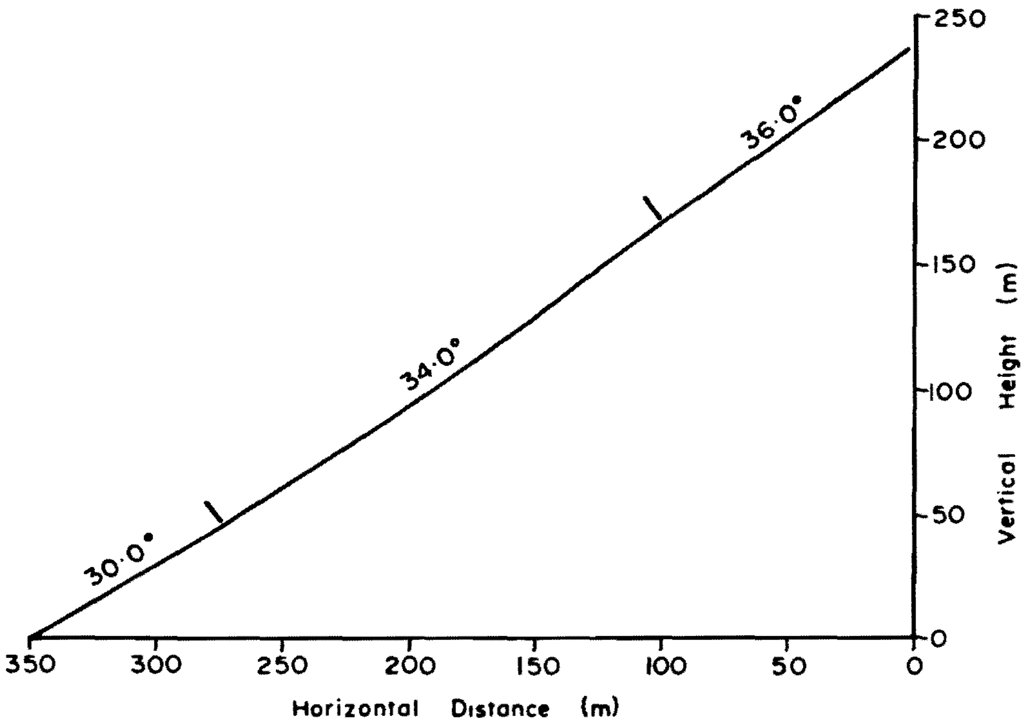


Figure 3.9: Typical steep valley side avalanche path. Note exceptionally steep runnout zone of 30 degrees. These rectilinear forms resemble closely rockfall scree slopes.

of unconfined and confined composed 17 percent and 24 percent respectively. The large proportion of unconfined tracks for the Craigieburn Range, in contrast to that found by Miller et al.(1976) for the San Juans and Schaerer (1972) for the Rogers Pass area, British Columbia, reflects the predominance of open scree slopes below both Type 1 and 2 and Type 3 starting zones (Figure 3.10). Confined tracks consist mostly of either valley side avalanche tracks bound by small interfluves (Figure 3.11) or main valley tracks where the main valley acts as the actual avalanche track. Avalanche tracks possessing both characteristics of confinement and unconfinement usually possess upper track chutes which are composed of a predominant bedding plane outcrop in the steeply dipping bedrock.

3.4.2.3 Runout Zones

Mean runout zone slope for the Craigieburn paths is high, (Table 3.2) as many of the paths have runouts on steep valley side slopes, some of which are densely vegetated by low beech forest. This contrasts with values of runout slopes found for other areas of 5 to 15 degrees (Martinelli, 1974; Miller et al., 1976; Fitzharris, in press; Huber, 1982) with runouts occurring on open valley floors or fans adjoining open valley floors. Some of the larger paths in the Craigieburn Range do however possess runout zone angles similar to that found elsewhere (Huber, 1982), though during the study period, avalanches did not run to the full extent of these paths as defined by vegetation damage. Mean runout zone angle for the Craigieburn paths approximates most closely that found for the Milford Road avalanche paths

Figure 3.10: Unconfined avalanche tracks on open scree slopes, Note type 4 starting zones in middle foreground. Tim's Basin, Craigieburn Forest Park.

Figure 3.11: Confined avalanche track. Two low interfluves produce the confinement. Distance from figure to end of runout zone in bushline approximately 750 m.

(Fitzharris and Owens, 1980). However, basic difference in the morphology of the runout zones exists for these two areas. Runout zones for the Milford Road area are frequently characterised by steep well developed debris fans below cliffs or steep avalanche tracks. These contrast markedly to the open scree slope or valley bottom runouts of the Craigieburn Range.

The high variance of runout zone angles suggests that two types of runout zone may exist. High angle runout zones ranged from 19 to 25 degrees with a mean of 23 degrees while lower angle runout zones possessed a mean of 13 degrees ranging from 6.5 to 16 degrees. The former runout zones all occurred on valley side avalanche paths, the lower angle runout zones however, occurred on paths that either occupied an entire valley bottom or possessed a valley side starting zone and upper track, with the lower track and runout zone in the valley bottom. A difference of means test was conducted for these high and low angle runouts with a significant difference in mean slope angle at the 0.05 level evident.

3.4.3 General Nature of Craigieburn Range Avalanche Terrain

While the Craigieburn Range avalanche paths generally resemble those found by other researchers, the short steep avalanche paths developed on valley side slopes, for which runout slopes may be up to 30 degrees, have not been previously discussed in the literature. In such cases, where runout zones are so steep, the avalanche paths are composed of active scree slopes, with very little variability in slope profile. They are essentially rectilinear with very little debris cone or fan development at their distal ends and resemble more closely rockfall slopes (Caine, 1969) than avalanche slopes. As these paths also possess small starting zones it is likely that the high angle runouts may be accounted for by the small mass of snow involved in avalanching. Thus on release the mass of snow does not gain enough momentum to travel onto lower slopes or down the main valley once it has reached the bottom of the scree slope. The mass of snow involved in avalanching on these small avalanche paths may also be insufficient to do much geomorphic work thus accounting for their similar shape to rockfall slopes.

While the large avalanche paths of the Craigieburn Range and Mt Cook areas show some similarity in starting zone angle, concave profile form and runout slopes generally less than 16 degrees, the avalanche paths of the Milford area contrast markedly with these. Many of the starting zones for these areas occur within relict cirque basins. For the Mt Cook and Craigieburn areas the original cirque basins and associated valleys have been modified by post-glacial fluvial and mass movement processes to give a more uniform profile. This contrasts with the avalanche paths of the Milford area, the upper parts of which occupy hanging valleys, with tracks often characterised by plunging cliffs. Such contrasts in form may arise from contrasts in the response of the very resistant granites and gneisses of the Milford area and the highly erodible greywackes typical of the Mt Cook and Craigieburn Range areas to the post-glacial erosional regime.

Some contrast exists between the avalanche paths of the northern and southern ends of the Craigieburn Range. This is not only with respect to vegetation damage but the more frequent occurrence of avalanche starting zones in cirque basins for the northern end of the Craigieburn Range. This may result from two factors. The majority of the northern cirque basins have a general south-easterly aspect. Presumably, cirque glaciers would have developed to a greater degree in these basins as a result of their aspect. Ice may also have remained longer after general climatic warming. If cirque basins did develop in the southern end of the range to a similar extent, very little contemporary evidence of their existence is present. This may be a function of the post-glacial erosional history of the area. Great variability in the degree of fracturing of the argillaceous greywackes exists between either end of the Craigieburn Range with the most fractured rocks occurring in the southern end. These fractured rocks may therefore have been more susceptible to post-glacial erosion with many of the cirque basins losing their original form due to erosion and scree accumulation. Though large amounts of scree are present in some of these basins many large outcrops of bedrock occur. In some cases steeply dipping bedding planes may form the interfluves for confinement in tracks. In other cases bedding planes cutting across open slopes form mid-track chutes and gullies. Gullies formed in bedrock outcrops below ridges and above open slopes are a response to preferential erosion along fractured joints in the bedrock.

A combination of processes has therefore produced the avalanche terrain of the Craigieburn Range. While many of the main valleys may possess the remnant shape of glaciated valleys dating from the Late Pleistocene (Chinn, 1975), it is the cirque basins developed in the Late Pleistocene and Holocene and subsequent post-glacial erosional history that has moulded the present avalanche terrain.

3.5 INFLUENCE OF AVALANCHE TERRAIN ON AVALANCHE FREQUENCY

Schaerer (1972) has shown that the nature of the terrain may play a role in the frequency of avalanching. In particular he showed that track slope and exposure to wind were of special significance, while slope angle at the fracture point, roughness of the starting zone area and variability of the avalanche track incline showed secondary effects. This section applies a similar approach to that of Schaerer (1972) for the investigation of terrain-avalanche frequency relationships.

3.5.1 Methods

Variables used in the analysis (Table 3.4) were measured either directly in the field or derived from avalanche path profiles. Mean, standard deviation, maximum and minimum values for these are shown in Table 3.5. Twenty paths were used in the analysis. The majority of these paths were located on Porter Heights, Mt Cheeseman and Craigieburn Valley skifields or were easily observable from the Craigieburn Forest Park Headquarters. The avalanche occurrence data are thus considered to be quite accurate. Observations included only natural avalanche releases of

Table 3.4: Terrain Variables Investigated

Symbol	Dimension	Variables	Description
F	(yr^{-1})	Avalanche Frequency	Average number of avalanches of size 2 or greater for 1979-1981
θ_F	$(^{\circ})$	Fracture Point Angle	Average of two angles measured 30 m along the slope above and below avalanche fracture point
$\overline{\theta}_T$	$(^{\circ})$	Mean Angle of Track	Average slope angle from bottom of starting zone to top of run-out zone
S_T	$(^{\circ})$	Standard Deviation of Track Angle	Standard deviation of 30 m segments of track slope
C	( )	Convexity/Concavity	Number of breaks of slope of 4° or greater
$\overline{\theta}_S$	$(^{\circ})$	Mean Starting Zone Angle	Average angle of starting zone
S_S	$(^{\circ})$	Standard Deviation of Starting Zone Angle	Standard deviation of 30 m (or shorter) segments in the starting zone
θ_{Smax}	$(^{\circ})$	Maximum Starting Zone Angle	Maximum angle of 30 m segments in the starting zone
R	(m)	Ruggedness	Difference in height between top and bottom of starting zone
f_S		Coefficient of Friction of Starting Zone	Ratio of starting zone relief and horizontal extent
f_T		Coefficient of Friction of Avalanche Track	Ratio of track relief and horizontal extent
GD	(m)	Starting Zone Ground Distance	Ground distance from top to bottom of starting zone
Y"	(m^{-1})	Path Curvature	Second derivative of the parabola $Y = ax^2 + b$ describing path form
α	$(^{\circ})$	Path Gradient	Defined by $\alpha = \tan^{-1}(H/L)$ where H is vertical height and L horizontal distance between top of starting zone and distal point of runout zone
W_{NW}		North West Wind Exposure	1 - Cos (315 - Slope aspect) Perfect lee slope = 2
W_{SW}		South West Wind Exposure	1 - Cos (225 - Slope aspect) Perfect lee slope = 2

Table 3.5: Terrain Variables - Mean, Standard Deviation, Maximum and Minimum

Variable	Mean	Standard Deviation	Maximum	Minimum
θ_F (°)	34.8	0.73	37.0	33.0
$\overline{\theta}_T$ (°)	29.5	0.95	32.4	27.0
S_T (°)	2.84	1.16	5.93	1.61
C ( )	2.00	2.22	7	0
$\overline{\theta}_S$ (°)	34.8	1.24	36.5	30.8
S_S (°)	1.01	0.92	3.7	0.7
θ_{Smax} (°)	35.7	1.03	37.0	34.0
R (m)	62.3	30.1	108	28
f_S	0.60	0.10	0.93	0.50
f_T	0.49	0.05	0.60	0.43
GD (m)	113.1	50.3	195	30
Y'' (m^{-1})	0.002	0.0011	0.0048	.00042
α (°)	28.3	3.31	33.8	23.7
W_{NW}	1.58	0.37	1.98	1.01
W_{SW}	1.41	0.42	1.99	0.1

size two or greater as these are most likely to cause loss of life or damage to buildings.

A multiple stepwise regression procedure was used to determine the relative contributions of the various terrain variables to the explanation of the variance of avalanche frequency. Some preliminary sifting of the variables was performed before the final running of the regression analysis. This included the removal of variables with intercorrelations of greater than 0.80. Such a procedure is recommended as outlined in Section 2.4.

3.5.2 Results

Average avalanche frequency ranged from 0.33 to 2.7 avalanches per track per year. Sixty nine percent of avalanches recorded for the tracks studied were of magnitude three or greater. The distribution of avalanche magnitudes for tracks used in this analysis is quite different for all avalanches (Figure 2.4). This reflects the influence that wet and dry loose avalanches, originating mainly from Type 4 starting zones, have on the lower end of the distribution of avalanche magnitude for all tracks.

Examination of independent correlations of F with the 12 terrain variables considered revealed eight with correlations at the 0.05 level or greater. Two of these variables (f_T and Y") show significant negative correlations with F. The analysis was subsequently rerun with these twelve variables. The resulting three term regression (Table 3.6) explains 72 percent of the variance of avalanche frequency and is significant at the 0.01 level.

Table 3.6: Variables Selected and Variable Coefficients For Avalanche Frequency - Terrain Regression

Variable	Regression Coefficient	r^2	r^2 change	Standard Error
GD	0.0125	0.43	0.43	0.56
F_T	-5.365	0.60	0.17	0.48
α	-0.0897	0.72	0.12	0.41
Constant	5.3727			

3.5.3 Physical Significance of Variables Selected

Avalanche starting zone ground distance was selected as the first variable. As all starting zones in this analysis are of a similar shape, ground distance of the starting zone is a measure of the volume of snow available for avalanching. The positive correlation therefore shows that larger the starting zone the more likely it is to release.

On theoretical grounds f_S and f_T would be expected to be inversely related to avalanche frequency. This holds true in this analysis but f_T shows the only significant correlation. Mean f_T was 0.49, equivalent to an avalanche track gradient of 26 degrees. The minimum value of 25 degrees suggested by Schaerer (1972, 1975a) for the maintenance of flow is equivalent to f_T = 0.46. Only two of the four paths which have frequencies equal to two releases per track per year have f_T less than or equal to 0.46. Three of these tracks, however, show f_T less than the mean of 0.49. From this it appears that for F greater than the mean of 1.2, f_T for the Craigieburn Ranges should be less than 0.49. This presumably reflects size of the avalanche path as the large paths (large relative relief and long runout distances) have smaller f_T values. This contrasts with the results of Schaerer (1972, 1975a) whose analysis implies that greater frequency is expected on steeper tracks with higher f_T values, but no evidence is given by Schaerer of any avalanche path size avalanche frequency relationships.

Avalanche paths with lesser gradients display higher avalanche frequency. Like f_T this presumably reflects the effect that overall avalanche path size has on avalanche frequency. This is so as the paths that have lesser gradients also display the greatest relief, runout distances and largest starting zones. Apart from these size variables no correlations exist between α and any other variables in the analysis, thus testifying to its true independent contribution to the explanation of avalanche frequency.

Consideration of a range of terrain variables and their effect on avalanche frequency has revealed the dependence of F on overall avalanche path size. This is reflected most in the selection of starting zone ground distance as the single most powerful predictor of avalanche frequency. Though Schaerer (1972) found an avalanche path wind exposure term to be of importance in determining avalanche frequency, no such relationship was evident for the Craigieburn Range. This results mainly from the fact that the paths used in the analysis showed very little variation in slope aspect (Table 3.5).

3.6 MODELS USED FOR AVALANCHE RUNOUT DISTANCE PREDICTION IN THE CRAIGIEBURN RANGE

3.6.1 The Voellmy Model

Until recently the most common approach to predicting avalanche runout was that developed by Voellmy (1955) as described by Leaf and Martinelli (1977). The Voellmy approach based on the general properties of open channel flow involves three underlying assumptions. The initial velocity is calculated using an expression similar to the Chezy equation for fluid flow (Chow, 1959):

$$V = C\sqrt{RS} \qquad (3.1)$$

where V = velocity (ms^{-1})

C = Chezy coefficient ($m^{\frac{1}{2}}s^{-1}$)

R = hydraulic radius (m)

S = slope

The Voellmy equation based on (3.1) above is

$$V = \sqrt{\varepsilon h' (\sin\psi - \mu \cos\psi)} \qquad (3.2)$$

where ε = coefficient of turbulent friction (ms^{-2})

μ = coefficient of kinetic friction (dimensionless coefficient)

h' = flow height (m)

ψ = slope angle

For this approach the avalanche path is broken up into segments based on either a major break in slope or change in slope morphology. Where such changes occur approximate equations can be used to estimate the change in velocity and flow height from one segment (n) to another (n-1):

$$\frac{V_n}{V_{n-1}} = \frac{h'n-1}{h'n} \simeq \left(\frac{\sin\psi n}{\sin\psi n-1}\right)^{\frac{1}{3}} \quad (3.3)$$

The final expression for calculating runout distance is:

$$S_R = \frac{V^2}{[2g(\mu \cos\psi_R - \tan\psi_R) + V^2 g/\varepsilon\, hm]} \quad (3.4)$$

where S_R = runout distance (m)

g = gravitational acceleration (9.8 ms^{-2})

ψ_R = slope angle in the runout zone (degrees)

hm = $h' + V^2/4g$ (when debris piled in a steep cone)

Voellmy (1955) considered that flow height (h') varies in accordance with flow regime. For flowing avalanches h' is approximated by the height of the avalanche fracture line, whereas for mixed motion avalanches

$$h' = 4\ hD \quad (3.5)$$

where hD is the average depth of the avalanche debris. Most avalanches observed in this study were of the flowing type, but some of the avalanches noted on the larger paths were definitely mixed motion with a considerable airborne component. For these, equation (3.5) was used, while all other calculations were performed using h' equal to fracture height. The kinetic friction coefficient (μ) used in equation (3.4) was that as estimated by Schaerer (1975a) and subsequently used by Leaf and Martinelli (1977) and approximating the kinetic fricton term of Martinelli et al.(1980):

$$\mu = 5/V \quad (3.6)$$

A computer program was written with input of slope angle for each track segment, flow height and actual runout distance. From this, velocity and kinetic friction coefficients were calculated with equation (3.2) and (3.4) being iterated until an appropriate turbulent friction coefficient (ε) was found to match the known stopping position. An error of ± 5 m was allowed for in the estimation of stopping position. This process was performed with a number of flow heights. The analysis was performed for eleven paths possessing good evidence of the maximum limit of runout distance, and consequently deals with larger magnitude events than considered in the frequency analysis of Section 3.5.

3.6.1.2 Results and Discussion

Results of the Voellmy analysis are presented in Table 3.7. Similar to the studies of Schaerer (1975a) and Leaf and Martinelli (1977) (Table 3.8) ε, values for the Craigieburn Range paths range widely (301 to 1382 ms^{-2}). The highest values of ε and those corresponding to a hydrodynamic surface characterised by smooth snow cover with no trees (Table 3.8) occur for flow heights of 0.5 m. For the large paths with vertical reliefs (H) greater than 450 m, mixed motion flows of up to 4.5 m and tracts of forest or avalanche protection works in their runout zone, ε values fall into the category of slopes with boulders, trees and forest (Table 3.8). These low ε values therefore represent high friction surfaces in the runout zone.

Associated with ε, the values of μ for individual paths always exceeded tan ψR. If μ was allowed to fall

Table 3.7: Flow Height, Kinetic and Turbulent Friction Coefficients and Runout Distance Using The Voellmy Model

Path	Flow Height (m)	Runout Distance (m)	Runout Slope (degrees)	ε (ms^{-2})	μ
Avalanche Corner	1.50	145-155	20.6	502-508	0.42
Grunt 2	2.00	175-185	18.0	370	0.35
Hamilton	2.00	320-330	12.3	675-695	0.25
Face	4.50	320-330	12.3	325-330	0.25
Fourth	2.00	410-420	15.5	510	0.29
Gut	4.00	410-420	15.5	272	0.28
Ridge	1.50	170-180	24.0	401	0.48
Route	2.00	170-180	24.0	301	0.48
Scorpio	0.50	110-110	22.0	1356	0.47
	2.00	100-110	22.0	340-343	0.45
Big Mama	1.00	390-400	6.4	1310-1320	0.23
	2.00	390-400	6.4	685-695	0.22
	4.50	390-400	6.4	335	0.21
RHSRHS	1.00	90-100	16.6	750-770	0.38
	2.00	90-100	16.6	392-396	0.38
Bluff	1.00	150-160	14.5	751-756	0.41
Face	1.50	150-160	14.5	507-511	0.41
	2.00	150-160	14.5	385-388	0.41
Libra	0.50	145-155	21.0	1251-1255	0.44
	2.00	145-155	21.0	321-323	0.44
Barricade I	0.50	115-120	18.0	1375-1382	0.40
	2.00	115-120	18.0	357-362	0.38

Table 3.8: Range of Turbulent Friction Coefficients Found For Other Studies

Study	ε (ms^{-2})	Slope conditions
Voellmy (1955)	400-600	Decreasing roughness
Salm (1972)	400-600	
Schaerer (1975b)	150-300 400-600 500-750 1200-1800	Slope with boulders, trees forest Average gully Average, open mountain slope Smooth snow cover, no trees
Shen and Roper (1970)	750	
Leaf and Martinelli (1977)	500-1800	Rough bouldery to smooth snow slopes
Buser and Frutiger (1980)	1360	
Martinelli et al. (1980)	1500 or greater	Snow covered slopes
Schaerer and Salway (1980)	1420-1810	Snow filled gully
Fitzharris (in press)	150-600	Bouldery, snow free runouts

below $\tan\psi R$ this resulted in simulated avalanches overshooting their respective runouts. This agrees well with Perla (1981) who has noted that μ should always be greater than $\tan \psi R$ because of this problem. Voellmy (1955) has suggested that low values of μ down to 0.1 represent powder avalanches with well developed dust clouds. Likewise Martinelli et al. (1980) have suggested low values of an internal friction term (ν), similar to μ used in this analysis, represent fresh dry snow, while values of μ of 0.40 are more realistic for low velocity flows. Taking the formulation of Schaerer (equation 3.6) used in this analysis, a velocity range of 10 to 50 ms^{-1} gives μ ranging from 0.5 to 0.1. Low values of μ for the Craigieburn paths are thus associated with faster moving avalanches typified by soft dry snow. This contrasts with Perla (1981) who has suggested that $\mu < 0.1$ represents flows with significant quantities of water. As wet avalanches may maintain momentum on slopes less than 10 degrees (Perla, 1981), which are also typical of large dry avalanche runout zones, the relationship between μ and the moisture status of the snow may be non-linear with low μ values appropriate for both wet and dry avalanches.

Though a range of μ, ε, and h' values reconcilable with field evidence give accurate predictions of avalanche runout distance, it is suggested that for flowing avalanches, which typically occur on paths with $H < 450$ m, h' should be set at 2.0 m. This approximates the average depth of snow found in the starting zones of the Craigieburn Range. If μ is set at $\mu > \tan\psi R$ and $h' = 2.0$ m for all paths with $H < 450$ m, ε is found to lie in the range of 300-695 ms^{-2} which agrees well with that originally recommended by Voellmy (1955) and subsequently suggested by Salm (1979). The range of ε also coincides with those values found by Schaerer (1975) to represent average mountain slopes. This seems reasonable, as on a subjective basis those avalanche paths with open scree slopes and relatively uniform long profiles could be considered as average mountain slopes.

On large paths where H > 450 m and values of h' = 2.0m seem inappropriate as a mixed motion regime is likely, h' should be set a h' > 4.0 m. For this case ε falls below 330 ms^{-2} for the Craigieburn paths. This lies within the range of ε values for mixed motion avalanches with flow heights up to 14.0 m found by Fitzharris (in press) for the Mt Cook area. Applying similar flow heights to the large Craigieburn paths, μ values down to 135 ms^{-2} are indicated with $\mu = 0.18$, comparable to those values found by Fitzharris (in press).

3.6.2 The PCM Model

Perla et al. (1980) considered the a priori subjective selection of a mid-path reference point, from which runout distance is calculated, an unsatisfactory feature of the Voellmy (1955) model. An alternative two parameter model is presented by these workers with the starting zone as the reference point. Distance thus enters into the acceleration and deceleration computations with velocity of the snow mass becoming a function of slope. This function is calculated from the differential equation of motion:

$$\tfrac{1}{2} \frac{dV^2}{dS} = g(\sin\psi - \mu \cos\psi) - D/M\, V^2 \qquad (3.7)$$

where S = position of the centre of mass measured along the path from the reference point (m)

D/M = mass drag ratio (m)

The solution of this function relies on the two adjustable factors of μ and D/M. The coefficient of kinetic friction (μ) is the same as that used by Voellmy (1955), while D/M is a mass to drag ratio where M equals avalanche mass and D is a drag parameter. The other essential difference between this approach and that of Voellmy (1955) is that in the latter an infinitely long flow moving past a point is assumed whereas in the PCM model the motion of snow is treated as a finite mass. McClung (1983) has subsequently derived the Voellmy (1955) equations from the PCM model showing that the PCM model is a more general model.

Numerical solution of equation (3.7), as discussed by Korner (1976) and Cheng and Perla (1979), is achieved by dividing the avalanche path into segments of relatively constant slope angle (ψ). Once segments have been defined, values of slope angle (ψ), segment length (L), coefficient of friction (μ) and mass-drag ratio (M/D) are assigned.

If the speed at the beginning of the ith segment is V_i^A, and the avalanche does not stop somewhere in the middle of the segment, then the speed V_i^B at the end of the ith segment is the solution of equation (3.7) (Perla et al., 1980).

$$V_i^B = [\alpha_i (M/D)_i (1-\exp\beta_i)+(V_i^A)^2 \exp\beta_i] \qquad (3.8)$$

where $\alpha_i = g(\sin\psi_i - \mu_i \cos\psi_i)$

$\beta_i = -2L_i/(M/D)i$

If the avalanche stops at a mid-segment position, the stopping distance S from the beginning of the ith segment is the equation (3.7) solution (Perla et al., 1980).

$$S = \frac{(M/D)i}{2} \quad \ln\left[1 - \frac{(V_i A)^2}{\alpha_i (M/D)i}\right] \tag{3.9}$$

The computation is iterated down to the runout zone until the stopping position is reached.

As no accurate velocity data were available for the Craigieburns it was assumed following Perla et al. (1980) that wet avalanches commonly have velocities of 10 ms^{-1} whereas dry avalanches reach up to 50 ms^{-1}. Given $0.1<\mu<0.5$ as suggested by Perla et al. (1980) and Bakkehoi et al. (1981) and substituting M/D for εh of Voellmy, the possible range of M/D may be established by

$$M/D = \frac{V^2}{g(\sin\psi - \mu\cos\psi)} \tag{3.10}$$

Using equation (3.10) M/D values range from 20 m to 2049 m for the tracks studied. To establish the correct μ, M/D combinations μ was fixed and M/D allowed to vary. Velocities generated by the two parameter model were then compared with known avalanche velocities to establish the feasibility of the μ, M/D combination.

3.6.2.1 Results and Discussion

For all paths a range of μ, M/D combinations were found to match the known avalanche stopping position. Values of μ ranged from 0.15 to 0.45 with μ always greater than tan ψR. As snow failure may be initiated on slopes as low as 25 degrees an upper limit of μ = 0.45 was adopted as suggested by Perla et al. (1980). The range of μ , M/D pairs is displayed in Table 3.9. Though a continuum of values exists for the μ, M/D pairs only the upper and lower pairs and those at μ = 0.05 intervals are shown along with the respective maximum velocities generated.

Mean M/D for μ = 0.35 was 483 m with a range from 115 m to 1590 m. Similarly for μ = 0.45 $\overline{M/D}$ was 651 m with a range of 350 m to 1450 m. For avalanche paths at Mt Cook, Fitzharris (in press) found that the highest μ reconcilable with known physical evidence of avalanche velocity and terrain conditions was 0.40 with an associated M/D of 500 m. For μ = 0.40, M/D was found to range from 170 to 3050 m with $\overline{M/D}$ of 950 m for the Craigieburn Range. These values do comply with those of Fitzharris (in press), but the overall lower values of μ for the Mt Cook paths attests to their much lower runout zone angles. A similar range of M/D values for the Craigieburn Range was also found by Perla et al. (1980) for 25 avalanche paths along the North Cascades Highway Washington, in the United States.

For several paths in the Craigieburn Range it was found that M/D values up to 10^3 were applicable. Velocities generated by these high μ, M/D pairs are comparable to velocities reported by Schaerer (1962, 1973), Kotlyakov et al. (1977), Heimgartner (1977), Schaerer and Salway (1980) and Shimizu et al. (1980) for high speed avalanches. However, radical changes in M/D with small changes in μ bring about negligible increases in velocity (Table 3.10). Perla et al. (1980) have discussed this as a problem arising from the competing influence of the μ and M/D parameters. Bakkehoi et al. (1981) also noted similar problems with this model.

Table 3.9: Friction Coefficients, Mass-Drag Ratios and V_{max}

Path	tan ψR	μ	M/D (m)	V_{max} (ms^{-2})
Avalanche Corner	0.37	0.40	190	15.39
		0.45	570	18.38
Grunt 2	0.40	0.41	115	14.45
		0.45	350	16.69
Hamilton Face	0.21	0.22	120	21.22
		0.25	360	31.55
		0.30	710	36.57
		0.35	1300	40.57
		0.36	1480	41.80
		0.40	3000	43.53
		0.43	9100	44.16
Fourth Gut	0.27	0.28	207	24.19
		0.30	420	31.26
		0.35	910	35.34
		0.40	2070	37.00
		0.43	3250	38.09
		0.44	9999	38.50
Ridge Route	0.34	0.35	100	16.33
		0.40	270	22.32
		0.45	520	25.42
Scorpio	0.34	0.35	98	15.68
		0.40	360	23.07
		0.45	1050	24.60
Big Mama	0.11	0.15	315	35.02
		0.20	500	40.85
		0.25	710	43.74
		0.30	1000	45.46
		0.35	1590	47.49
		0.40	3050	49.35
		0.44	9900	50.08
RHSRHS	0.25	0.35	150	18.15
		0.40	310	20.26
		0.45	660	20.79
Bluff Face	0.25	0.27	115	18.91
		0.30	215	24.26
		0.35	405	28.26
		0.40	700	30.54
		0.45	1450	31.71
Libra	0.38	0.39	125	17.33
		0.40	170	19.59
		0.45	390	23.91
Barricade I	0.32	0.35	115	16.90
		0.40	225	19.44
		0.45	430	20.15
LHS Huts	0.46	0.50	240	19.11
		0.55	620	21.54

Table 3.10: Insensitivity of V_{max} to Increases in M/D

Path	μ	M/D	V_{max} (ms^{-2})
Hamilton Face	0.36	1480	41.80
	0.38	1720	42.37
	0.40	3000	43.55
	0.43	9100	44.16
Fourth Gut	0.40	2070	37.00
	0.43	3250	38.09
	0.44	9999	38.50
Big Mama	0.36	1740	47.90
	0.38	2200	48.58
	0.40	3050	49.35
	0.44	9900	50.08

Given the negligible increase in V_{max} with related large increases of M/D a possible upper limit of M/D may exist. Perla et al. (1980) noted for dry avalanches an upper limit of $M/D = 10^4$ may be set. Bakkehoi et al. (1981) have proposed an upper limit of M/D in terms of the total vertical relief (H) of the avalanche path and suggested that large values of M/D will necessitate the setting of μ larger than the tangent of the starting zone. For the Craigieburn paths the maximum M/D in terms of H considering negligible increases in V_{max} is 3.2H. This lies well within the range of $0.1H < M/D < 10H$ proposed by Bakkehoi et al. (1981). The minimum value of 0.16H also falls within those limits prescribed by Bakkehoi et al. (1981), with the combined range of μ, M/D pairs being (0.15, 0.16H) to (0.45, 3.2H). This agrees well with the ten best μ, M/D pairs found for describing stop positions on the 25 Cascade paths by Perla et al. (1980) of (0.46, 3H) $\geq$, M/D $\leq$ (0.12, 0.1H). Bakkehoi et al. (1981) also noted that their results are fully consistent with this range.

The importance of H with respect to V_{max} is demonstrated when regressing V_{max} against H with a resultant r^2 of 0.76, significant of the 0.01 level. Dependence of V_{max} on M/D is also apparent (r^2 = 0.57, significant at the 0.05 level). High speed avalanches will therefore be produced on paths with great vertical relief which is to be expected given potential energy considerations. Such flow conditions have been observed on a number of the large Craigieburn Range paths. For a few paths possessing runout zones of 25 degrees and heavily vegetated with low beech forest, μ had to be set in excess of 0.45. For these cases μ was set at 0.50. Associated with these high μ values were low M/D values demonstrating that frictional drag due to energy dissipation by the vegetation is very important in stopping the avalanche flow.

3.6.3 Empirical Models

3.6.3.1 Methods

Though the two theoretical models tested gave satisfactory results, an empirical model relating terrain variables to avalanche runout distance was developed using the same paths and terrain variables as in the avalanche frequency analysis. Similarly the variables removed from the analysis of avalanche frequency were likewise not used in this analysis of runout distance. In contrast to the approach of Lied and Bakkehoi (1980), who used products of independent variables for a similar study, only simple independent variables were used here. Runout distance is that distance from the top of the avalanche starting zone to the end of the runout zone as defined in Section 3.4.2.3.

3.6.3.2 Results

Univariate correlation analysis revealed five variables significantly correlated with avalanche runout distance at the 0.01 level (Table 3.11). The variable R was removed from any subsequent analysis as it was highly correlated with GD and Y". Stepwise multiple regression produced the results shown in Table 3.12.

3.6.3.3 Physical Significance of Variables Selected

Individual correlation coefficients show that the less concave avalanche paths possess the greatest runout distances. Such paths maintain steeper slopes over longer distances allowing avalanche snow to gain greater momentum, thus allowing longer runout. Path shape is therefore influential in determining V_{max}. Individual path curvatures for the Craigieburn Range paths were regressed against the predicted V_{max} for the PCM model. The resultant regression, $V_{max} = 43.1 - 9628\ Y''$ (3.11) ($r^2 = 0.51$, significant at the 0.05 level) revealed dependence of V_{max} on this path shape variable. Bakkehoi et al. (1981) have proposed that V_{max} is dependent on a path shape variable:

$$V_{max} = \sqrt{2gH\left(1 - \frac{\tan\alpha}{\tan\theta}\right)} \qquad (3.12)$$

where V_{max} = maximum velocity

H = total vertical relief

α = angle of line connecting the end points of the largest event

Table 3.11: Variables Significantly Correlated with Avalanche Runout Distance at the 0.01 level

Variable	Correlation Coefficient (r)
θ_F	0.62
S_T	0.72
R	0.71
GD	0.71
Y"	-0.85

Table 3.12: Multiple Regression Analysis of Avalanche Runout Distance

Independent Variable	Regression Coefficient	r^2	r^2 Change	Standard Error
Y"	-187910.40	0.72	0.72	199
GD	2.47	0.81	0.09	169
S_T	93.29	0.85	0.04	153
Constant	209.11			

θ = angle of line connecting top point of path to point at top of runout zone

This is similar to equation (3.11), except H is included along with the shape factor $\left(1 - \frac{\tan\alpha}{\tan\theta}\right)$

The Bakkehoi et al. (1981) model was subsequently used with Craigieburn Range path data to predict V_{max} (Table 3.13). Correlation of V_{max} from equation (3.12) with that predicted by the PCM model gave $r^2 = 0.93$. However it must be remembered that both approaches assume velocity as a function of path geometry and that no actual mesurements of V_{max} are involved. Bakkehoi et al. (1981) suggested that variation of V_{max} from path to path may be more a function of path shape than H. However, for the Craigieburn paths H is marginally better at predicting V_{max} as determined by the PCM model ($V_{max} = 8.48 + 0.15H$, $r^2 = 0.69$, significant at 0.01 level) than Y" ($r^2 = 0.66$), which is similar to $\left(1-\frac{\tan\alpha}{\tan\theta}\right)$ used by Bakkehoi et al. (1981). As Y" is highly negatively correlated with H, the selection of Y" in the regression analysis for runout prediction also reflects the role of avalanche path size in determining avalanche runout distance.

In addition to path shape V_{max} relationships and the resultant effect on runout distance, Lied and Bakkehoi (1980) have shown that path shape is also a good predictor of the point at which the avalanche path gradient intersects the end of the actual avalanche path profile. Application of the regression of Lied and Bakkehoi (1980) of $\alpha = 1.07 \times 10^4 Y" + 21$ (3.13) gave extremely high values of α for short steep valley side avalanche paths (H < 450 m) but values approaching reality for the larger less concave paths (Table

Table 3.13: Comparison of Maximum Velocity Produced From Two Models

Path	H (m)	α (degrees)	θ (degrees)	V_{max}(PCM) (ms^{-2})	V_{max} (Bakkehoi et al) 1981 (ms^{-2})
Avalanche Corner	235	27.0	29.0	18.38	19.20
Grunt 2	225	27.5	29.4	16.69	19.25
Hamilton Face	750	22.0	26.0	44.16	45.70
Fourth Gut	725	23.50	27.0	38.50	45.60
Ridge Route	230	27.0	31.6	24.42	26.10
Scorpio	300	27.4	29.3	24.60	21.10
Big Mama	600	24.7	29.7	50.08	47.70
RHSRHS	280	26.9	28.7	20.79	20.07
Bluff Face	437	26.5	30.2	31.70	34.70
Libra	295	30.6	32.6	23.90	20.80
Barricade I	276	28.2	30.12	20.15	16.80

Table 3.14: Comparison of α Predicted From Lied and Bakkehoi (1980) and This Study's Model For Craigieburn Range Avalanche Paths

Path	α (degrees)	Y"	Predicted α (degrees) Lied and Bakkehoi (1980)	Predicted α (degrees) This Study
Avalanche Corner	27.0	.0020	42.4	27.3
Grunt 2	27.5	.0020	42.4	27.3
Hamilton Face Face	22.0	.00018	22.9	23.4
Fourth Gut	23.5	.00042	25.4	24.0
Ridge Route	27.0	.0030	53.1	29.4
Scorpio	27.4	.0014	35.9	26.0
Big Mama	24.7	.00104	32.1	25.3
RHSRHS	26.9	.0022	44.5	27.8
Bluff Face	26.5	.0012	33.8	25.6
Libra	30.6	.0020	42.4	27.3
Barricade I	28.2	.0020	42.4	27.3

3.14). An identical regression analysis to that of Lied and Bakkehoi (1980) was also performed for the Craigieburn Range paths with $\alpha = 2.124 \times 10^3 Y'' + 23$ (3.14). The coefficient of determination was 0.56 with predicted values matching closely those measured from avalanche path profiles (Table 3.14). The higher regression coefficient for the Craigieburn paths reflects the much steeper shorter paths for this area in comparison to those used by Lied and Bakkehoi (1980).

The second variable included in the regression (GD), as shown earlier, is a measure of initial snow volume. Avalanche paths with larger starting zones are therefore likely to run greater distances. However, avalanche runout distance was consistently underpredicted for those paths with multiple starting zones. On these large paths runout distance is more likely to be a product of multiple starting zone release. A more precise measure of starting zone size, such as that used by Bovis and Mears (1976), should therefore be used in this sort of analysis. GD also reflects the role of path size, which through its relationship to Y" has been shown to be important in determining avalanche runout distance.

High S_T values found for those paths with variable track slopes or large contrasts in upper and lower track slope angles act to increase avalanche runout distance. Snow may be accelerated in the steep upper parts of the track or pulverised as a result of variable track slopes. Both of these effects will induce a marked airborne component into the flow. Frictional resistance in such cases will be much less than that of a simple flowing avalanche with greater velocities and further travel of the snow resulting.

3.6.4 Applicability of Voellmy, PCM and Empirical Models to Craigieburn Range Runout Distance Predicition

While the two theoretical models tested give acceptable results reconcilable with overseas studies some features of these models need to be treated carefully. The establishment of a mid-path reference point for the Voellmy model from which all runout distances are calculated may cause some problems. As many of the avalanche paths in the Craigieburn Range are rectilinear there are often few distinguishing features between the track and the runout zone. This is especially so on those steep valley side paths where runout zone inclinations resemble more closely avalanche track slopes. Fixing a mid-path reference point thus becomes very difficult and may vary drastically between observers.

The Voellmy model includes in its calculations a flow height term (h'). As meagre evidence of actual flow heights exists for the Craigieburn Range, except for occasional observations, a range of flow heights could be used. However a set flow height, for those paths on which a flowing regime is more likely, is considered applicable. This arises from the great variability in assignment of kinetic (μ) and turbulent (ε) coefficients of friction with varying flow heights. The flow height recommended for these paths is taken as equivalent to the average depth of snow found in the starting zones of the Craigieburn Range. The coefficients found to match the known stopping position for this flow height therefore represent those applicable for a maximum event on these paths. For the Craigieburn Range if

μ is set below $\tan\psi R$ and ε is allowed to exceed 695 ms^{-2} simulated avalanches will overshoot their runouts with a flowing regime of 2.0 m depth. The general range of recommended for the Craigieburn Range paths for which flowing avalanches are most likely then is ε = 300-695 m^{-2}, h' = 2.0 m and $\mu > \tan R$. Generally the μ and ε values applicable to these paths are higher and lower respectively than those found for other studies and reflect the steeper occasionally snow free and often rougher runout zones of the Craigieburn Range.

For the large Craigieburn Range paths where H exceeds 450 m, a flowing regime of 2.0 m depth with ε = 300-695 m^{-2} resulted in simulated avalanches stopping part way down the avalanche path. These limits of h' and ε are therefore not applicable to the large paths in which mixed motion flow regimes are frequent. For these paths h' up to 12.0 m with ε of 135 to 150 m^{-2} and $\mu > \tan\psi R$ replicates the known stopping position of the maximum events on these paths. Flow heights of up to 12.0 m have been observed on these paths and in addition to the values of μ and ε are reconcilable with those found for large paths in the Mt Cook area.

Unlike the Voellmy model, computations in the PCM model do not involve a flow height term. Also the reference point from which all runout distances are calculated is the top of the starting zone which seems more appropriate for the paths of the Craigieburn Range given the problems encountered in establishing consistent mid-path reference points. In the absence of any accurate velocity data limits of velocity and therefore M/D were established. Applicable M/D values for the Craigieburn Range paths were found not to exceed 3.2H, as above this, further large increases in M/D produced insignificant increases in V_{max}. The PCM model with the related range of μ, M/D combinations of (0.15, 0.16H) $\leq$ μ, M/D $\leq$ (0.45, 3.2H) is considered suitable for the Craigieburn paths and agrees well with the range of values established for overseas studies.

A feature of both the Voellmy and PCM model was the inability to solve for a small number of paths the correct μ, ε, and M/D combinations. This was specifically for those valley side paths in which forest vegetation was present in the runout zone. The Voellmy model proved completely unsuitable for these paths with no solutions established. However, using the PCM model, the setting of μ at 0.5 with associated low M/D values, indicative of high surface friction runout surfaces, gave partially satisfactory results. For the same valley side paths an empirical model based on the terrain variables slope curvature, a surrogate measure of starting zone area and variability of avalanche track slope, accurately predicted avalanche runout distance. One inherent weakness is apparent in this empirical model, in that single starting zone release is presumed. Underprediction of avalanche runout distances using this model therefore occurs for those large paths with multiple starting zones.

3.7 SUMMARY

The Craigieburn avalanche terrain has been shown to resemble that of other areas, with regard to starting zone characteristics such as slope angle, size, morphology and aspect. In addition to the active avalanche paths identified during the study period, these critical terrain features were used to map areas of potential avalanche release. Areas identified are typically open bowl shaped lee slopes in excess of 30 degrees occurring above 1600 m elevation. Many of the potential starting zones identified have subsequently been noted to release.

The main contrasting feature of the avalanche paths in this study compared to those found elsewhere is the wide range of runout zone angles. This may be attributed to the wide range of avalanche path sizes considered in this analysis compared to the large paths typical of overseas studies. Both high and low angle runout zones are evident in the Craigieburn Range with the former typical of short steep valley side paths. These steep runouts are suggested to be a function of either the smaller masses of snow involved in release, as many of these paths possess small starting zones, or the presence of dense tracts of vegetation on these slopes. Such vegetation will stop flowing snow on slopes, that otherwise are typical of avalanche tracks. Low angle runouts are associated with large avalanche paths occupying main valleys or paths with starting zones and upper parts of their tracks in a valley side situation with the remainder of the path occupying the main valley. These avalanche paths are concave in form and represent most closely paths cited in the overseas literature.

Relict cirque basins feature predominantly as areas in which avalanche starting zones are found in the Craigieburn Range. In contrast to areas such as Fiordland, where much of the original glacial terrain remains and dominates avalanche path form, the Craigieburn paths reflect equally the post-glacial erosional history in an area of fractured sandstones and siltstones. Many valley side avalanche paths are therefore characterised by open active scree slopes often rectilinear in form and resembling rockfall slopes. This rectilinear form is believed to be a response to the small amount of geomorphic work done by the small snow masses involved in avalanching on these slopes.

The effect of a range of terrain variables on avalanche frequency was assessed by multiple regression. The frequency of avalanches greater than magnitude 2 per path is dependent on variables representing overall avalanche path size, especially that of starting zone size. Though some aspect control on determining avalanche type and frequency through snow loading has been shown for the whole study area, a wind exposure term was found to have little effect on individual path avalanche frequencies. This is most likely a result of the similar starting zone aspects used in the avalanche frequency analysis.

The application of theoretical models for prediction of avalanche runout distance in the Craigieburn Range produced acceptable results for the majority of paths studied. Unsatisfactory features noted were the insensitivity of the flow equations for velocity calculation to wide variations in ε for the Voellmy model and M/D for the PCM model, and the inability of the Voellmy model to solve the runout distance for steep vegetated valley side paths. Large M/D values represent those flows with large masses and little frictional drag and typify conditions found on large paths with mixed motion avalanches in which high velocities are attained. Overall μ and M/D values found for the Craigieburn paths are generally lower than elsewhere, reflecting the snow free terrain often associated with avalanche runouts in the Craigieburn Range.

The empirical approach established by Lied and Bakkehoi (1980) for predicting avalanche path gradient and hence maximum extent of the avalanche runout zone using a path shape factor, proved unsatisfactory for small avalanche paths with H < 450 m but gave acceptable results for large avalanche paths. However, an empirical model developed for predicting avalanche runout distance using a surrogate measure of starting zone area, a path shape factor and the variability of avalanche track angle proved satisfactory for all paths apart from some under prediction on large paths. This under prediction is most likely a response of the model's insensitivity to multiple starting zone release.

CHAPTER FOUR

THE ROLE OF METEOROLOGICAL FACTORS IN AVALANCHE FORMATION

4.1 INTRODUCTION

In the absence of a sound meteorological and avalanche occurrence record, avalanche forecasters often rely on those meteorological variables critical for avalanche formation established for other areas. To date this has been the case for the Craigieburn Range and for other areas in New Zealand. The aims of this chapter then are:

(1) to assess the significance, to avalanche release, of a range of meteorological variables conventionally used in avalanche forecasting;

(2) to apply a discriminant function analysis approach to distinguishing dry, wet and non-avalanche days;

(3) to develop an avalanche forecasting model.

4.2 PREVIOUS WORK

Two main approaches to avalanche forecasting using meteorological data can be identified in the literature. These are causal-intuitive methods often referred to as conventional methods of avalanche forecasting (LaChapelle, 1980) and the purely statistical methods. In using conventional methods forecasters rely on present and past knowledge of meteorological conditions to qualitatively infer conditions of instability. The avalanche forecasting process using this method is iterative in nature (Businger et al., 1981) as feedback from new data and observations continually refines the initial assessment of avalanche potential. With such a practice a large amount of data may become redundant (LaChapelle, 1980) as new data in combination may supplement existing inaccurate data. LaChapelle (1966) has also noted that it may be necessary to adopt a variety of approaches for forecasting avalanches given the variation of snow climate from region to region.

Though a variety of variables are used in conventional avalanche forecasting (Atwater, 1952, 1954; Judson, 1964, 1966; and Perla, 1970), Armstrong and Armstrong have (1977) noted that there is limited agreement on which factors are the most important in avalanche release, although new snow types, precipitation rates and amounts, wind direction/speed and temperature patterns are generally accepted. Similar factors are emphasised by the Swiss (Zingg, 1966; de Quervain, 1975) with the possible exception of crystal type. The key to successful avalanche forecasting using these conventional methods relies on the collection of accurate and timely data from the field and expedient decision-making by experienced forecasters (Williams, 1980). Inputs from weather forecasts may also supplement this data (Daly, 1978). To date these methods have been used with success in avalanche warning systems (Moore and Marriot, 1981; Businger et al., 1981; and Weir, 1983).

Statistical methods differ from the conventional methods in that quantitative analysis of weather, snow and avalanche data is used to establish forecasts. The main method used has been discriminant function analysis (Judson and Erickson, 1973; Bois et al., 1975; Bovis, 1976, 1977; Drozdovskaya, 1979 and Obled and Good, 1980). These studies have all identified comparable variables traditionally used in conventional avalanche forecasting with a similar predictive ability of approximately 80 percent. Salway and Moyse (1978) and Salway (1979) have considered time series analysis to be more appropriate as it accounts for time lag autocorrelations and a wide range of the dependent variable. Judson and Erickson (1973) have attempted to use regression analysis but, as pointed out by Bois et al. (1975) the distribution of the dependent variable with respect to the predictor variables is highly non-symmetric and is a serious drawback for common regression techniques. In contrast to discriminant analysis regression techniques do not take into account the type of avalanche. Salway (1979) and Bovis (1976) also believe that regression is not appropriate for avalanche forecasting.

4.3 METEOROLOGICAL VARIABLES AND AVALANCHE PROBABILITY

4.3.1 Methods

To assess the influence of a range of independent meteorological variables or the trend of these variables on avalanche probability in the Craigieburn Range a similar approach to that of Perla (1970) was adopted. This involved the construction of probability diagrams for a range of meteorological variables (Figure 4.1a-1). Variables used in the analysis are means derived from the 24 hour period prior to 1200 hours on the day of the event. Ten dry and seven wet avalanche storms were used in the analysis. The sum of the magnitude of all avalanches greater than size 2 per storm (MAG) was used as the measure of avalanche activity.

In this analysis the probability of an event with MAG >10 (P_{10}) occurring was used and calculated by

$$P_{10} = \frac{\text{Number of storms with MAG greater than 10}}{\text{Total number of storms within variable interval}} \quad (4.1)$$

4.3.2 Results

Two dominant trends in P_{10} were apparent in the probability plots. These were firstly a steady increase in P_{10} with a related increase or decrease of the contributory variable, and secondly an oscillation of P_{10} around a probability of 0.5 (Figure 4.1 a-l). Perla (1970) classified these as first order and second order effects respectively. For the Craigieburn Range those variables showing first order effects were new snow depth, total precipitation, maximum six hour precipitation intensity, mean storm temperature, minimum storm temperature, temperature range and maximum six hour windspeed increase. All of these except maximum six hour windspeed increase have also been identified by Perla (1970) as showing first order effects. Relative humidity, mean six hour windspeed, maximum storm temperature and a wind component variable all demonstrate second order effects, of which Perla (1970) has identified only relative humidity as demonstrating second order effects. These variables will be discussed with regard to their role in avalanche occurrence in the following sections.

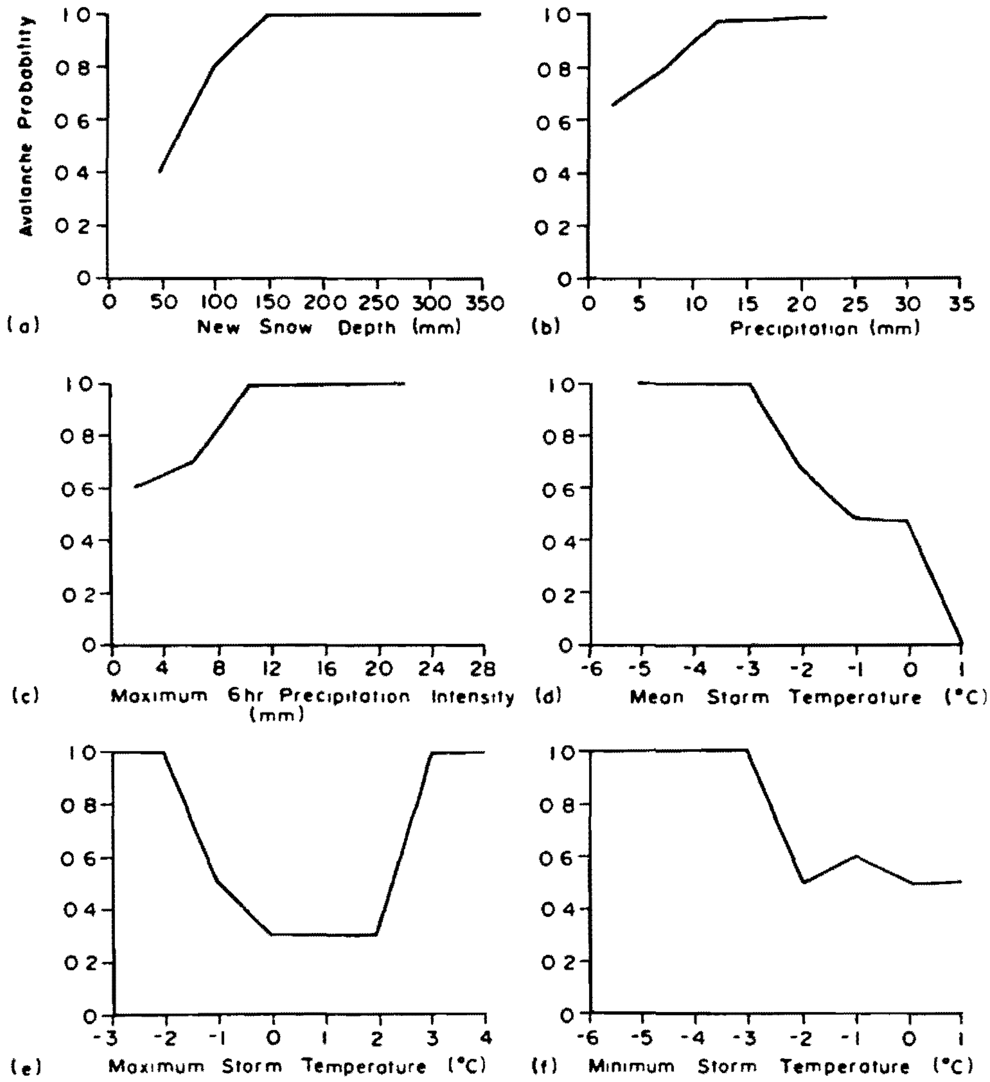


Figure 4.1: The effect of meteorological variables on avalanche probability.

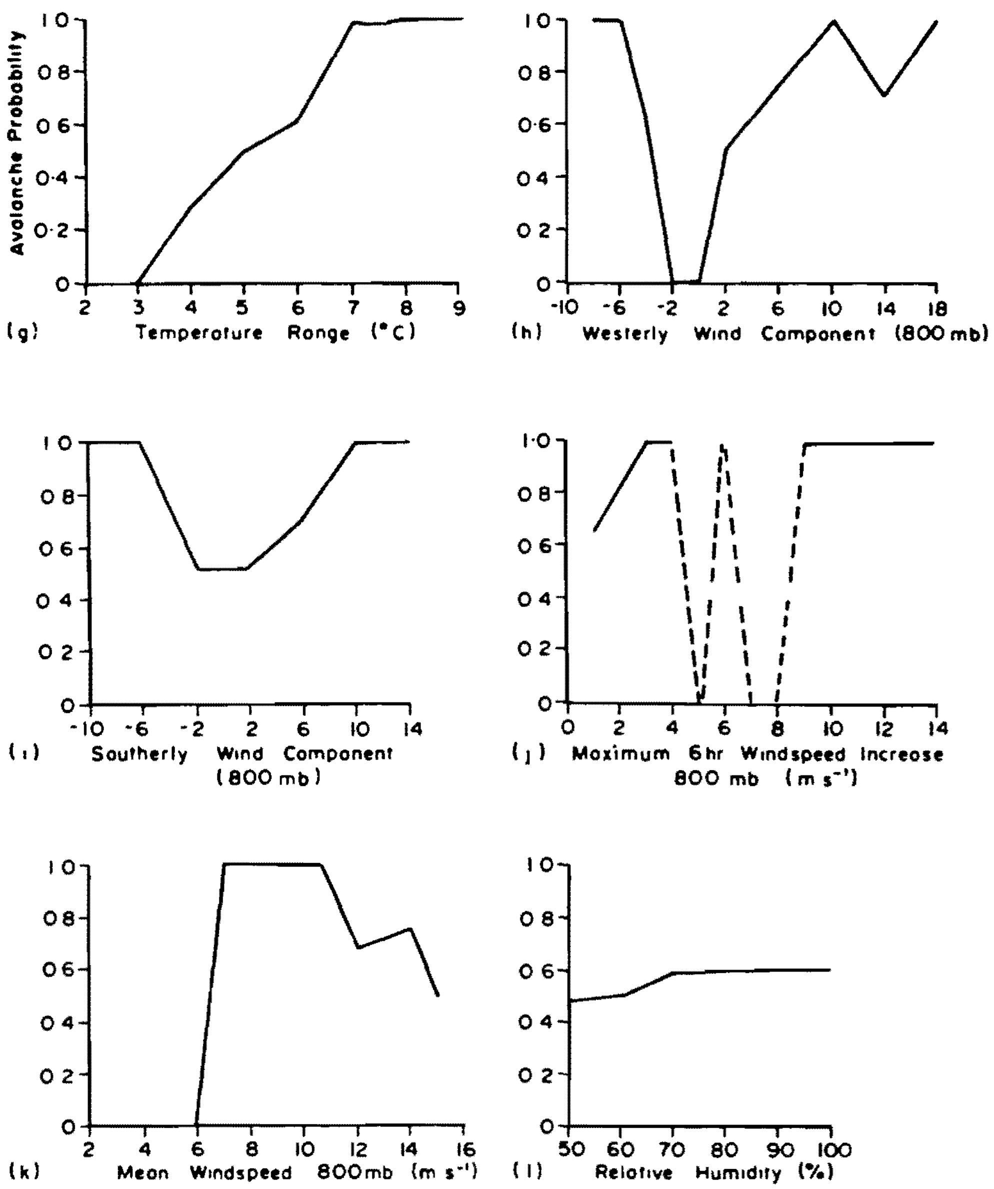


Figure 4.1: The effect of meteorological variables on avalanche probability.

4.3.3 Influence of Individual Meteorological Variables on Avalanche Occurrence

4.3.3.1 Precipitation Related Variables

New snow increments which produced cycles of avalanching in this study ranged from 20 mm to 320 mm with a strong relationship being exhibited between total new snow and MAG. (Figure 4.2) Ninety four percent of releases in the Craigieburn Range were of the direct action type, typical of a maritime location. (Appendix 4a-i). This contrasts markedly with areas where climax release predominates, such as the San Juan mountains of Colorado, where there is little correlation between avalanche frequency and new snow increments (Armstrong and Ives, 1976). Except for two cases all total avalanche storm precipitation amounts for the Craigieburn Range fell below the 250 mm and 300 mm suggested by Judson (1964, 1966) for the Western United States and Perla (1970) for Alta, Utah, respectively to generate instability. Perla (1970) has however noted that with strong winds only 100 mm new snow is necessary to generate instability for the Loveland-Berthoud Pass area. For the Craigieburn Range a similar situation may exist where strong winds in association with relatively small precipitaion increments may also be important in enhancing instabililty. The effect of wind will be discussed in Section 4.3.3.3.

Judson (1964) has suggested that snow water equivalent is a far better indicator of total load imposed on a slope than simply new snow depth. However, in this study the difference between correlations of MAG with new snow depth (Figure 4.2) and water equivalent were not significant.

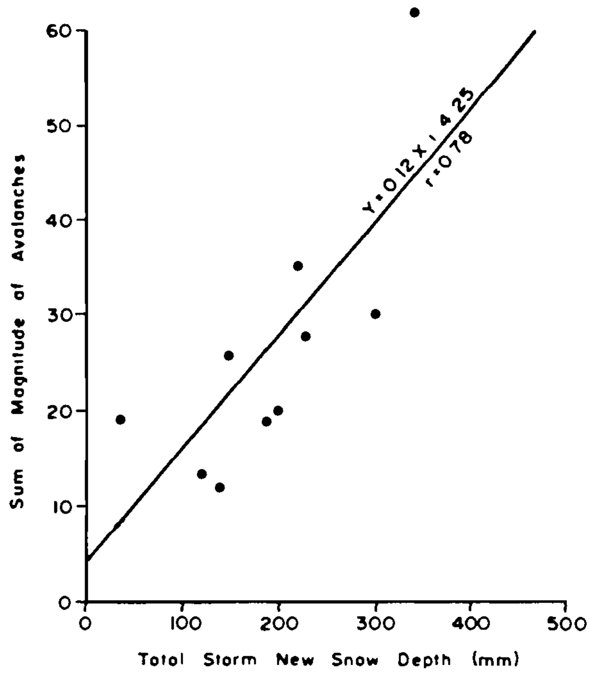


Figure 4.2: Regression of sum of all avalanches greater than magnitude two per storm against total storm new snow depth.

This may be accounted for by the small sample size or the reduced catch efficiency of the Belfort raingauge during windy conditions. In addition to total new snow depth and precipitation water equivalent, Atwater (1952) suggested that precipitation intensities of 2.5 mm hr^{-1} at wind velocities greater than 7 ms^{-1} produced a high degree of avalanche hazard. Judson (1964), Perla (1970) and Judson and Erickson (1973) likewise noted the importance of precipitation intensity. For the Craigieburn Range a weak correlation (r = 0.43, significant only at the 0.10 level) was found between maximum six hour precipitation intensity and MAG. Overall average precipitation intensity for the dry avalanche storms was 0.93 mm hr^{-1}; slightly higher than the 0.8 mm hr^{-1} recorded by Prowse (1981) for 80 snowstorms in the Craigieburn Range irrespective of avalanche occurrence over a five year period. The avalanche storms of this study are also much shorter with an average duration of 32 hours in contrast with 46 hours for the snowstorms studied by Prowse (1981). Dry avalanche storms also showed greater average precipitation amounts (30 mm) compared to the 21.8 mm recorded by Prowse (1981). Approximately 25 percent of the storms studied by Prowse (1981) produced precipitation intensities greater than 2.5 mm hr^{-1} though only for short durations. At no time during this study did mean storm precipitation intensity exceed 2.5 mm hr^{-1} but 40 percent of the dry avalanche storms did have intensities greater than 2.5 mm hr^{-1} for three hours or more. It therefore appears that much shorter duration maximum six hour precipitation intensities of 1.5 to 2.0 mm hr^{-1} are effective in initiating instability in the Craigieburn

Range. This critical figure has been derived from the mean maximum six hour precipitation intensity for the ten dry avalanche storms producing soft slabs.

4.3.3.2 Temperature Related Variables

The main temperature related phenomena causing snowpack instability in the Craigieburn Range appears to be low temperatures during and immediately after a storm through their effects on hindering settlement and sintering. The probability plots of mean and minimum storm temperatures (Figure 4.1 d and f) demonstrate these effects. As mean storm temperature increases above -3^{o}C avalanche probability decreases. Conversely, minimum storm temperatures of -3^{o}C or less result in increased probabilities. Temperatures below -3^{o}C then are effective in preventing settlement and sintering within the snowpack through a decrease in intergranular bond growth and compactive creep. In such conditions a snow deposit of low mechanical strength will be produced (Seligman, 1936, p.146; Ramseier and Sanders, 1966; Perla and Martinelli, 1976, p.76). Judson (1964), LaChapelle (1966) and Mellor (1968) also noted that the likelihood of avalanching is increased with low or falling temperatures, preventing overall slope stabilisation.

For this study minimum temperature gave a negative correlation of r = -0.36 with MAG while temperature range gave a positive correlation of r = 0.44 (Figure 4.1 g). These were significant at the 0.10 and 0.05 levels respectively. As most storms in the Craigieburn Range start off warm and cool down (Prowse, 1981), a large temperature range may imply lower temperatures towards the end of a storm. The effects of this will be similar to those above as low temperatures hinder compaction in the absence of strong winds. The effects of low temperatures maintaining instability over several days is demonstrated well by the two storms of 25 and 27 August, 1981 (Figure 4.3). The first cycle of avalanching on 25 August resulted from low temperatures preventing settling of the snowpack in conjunction with high rates of wind redistribution of the snow. The second cycle of avalanching on 27 August, 1981 occurred during a period of falling temperatures (Figure 4.3). Such an effect is further augmented when large amounts of snow redistribution take place as the redistributed snow will remain unsintered resulting from low air and snowpack temperatures.

In contrast to the above, Perla (1970) and Bovis (1977) have found that an increase in the probability of dry avalanche release is accompanied by an increase in temperature. Bovis (1977) believes that a rise in air temperature concurrent with slope loading may increase the rate of secondary creep in new and old snow, provided that rapid densification and stabilisation have not already taken place. No distinct examples of rising temperatures causing natural dry avalanche release were apparent during the study period. The only example of an 'inverse storm' of the type described by Perla and Martinelli (1976, p.36) resulted in no natural avalanching (Figure 4.4), but artificial control by explosives produced some limited activity. In this example, precipitation was initially light and temperatures remained below zero. A change in wind direction to the east brought increased precipitation from 0600 hours on 26 August 1979.

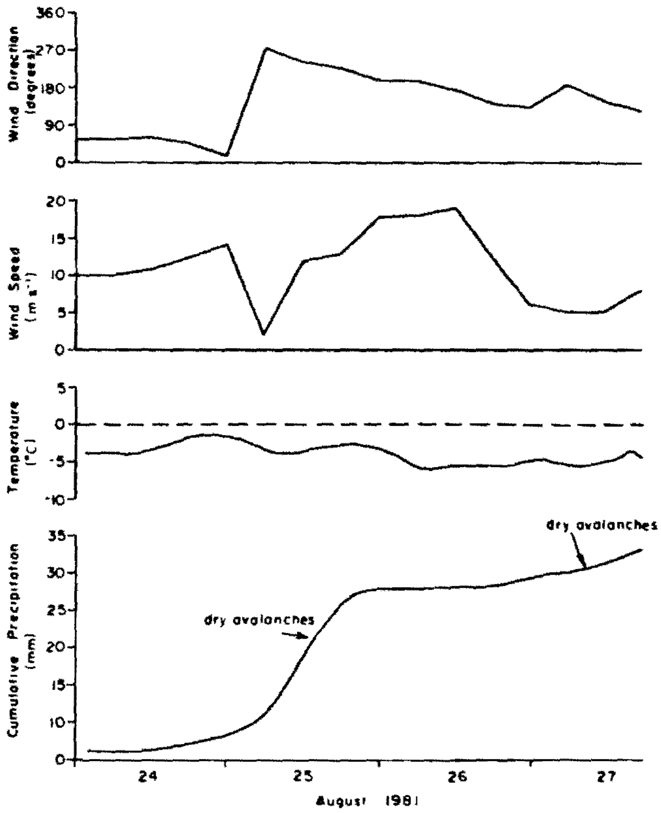


Figure 4.3: Storm plot of 24 August, 1981 to 27 August, 1981. Note avalanche occurrence following periods of high windspeeds and decreasing temperatures.

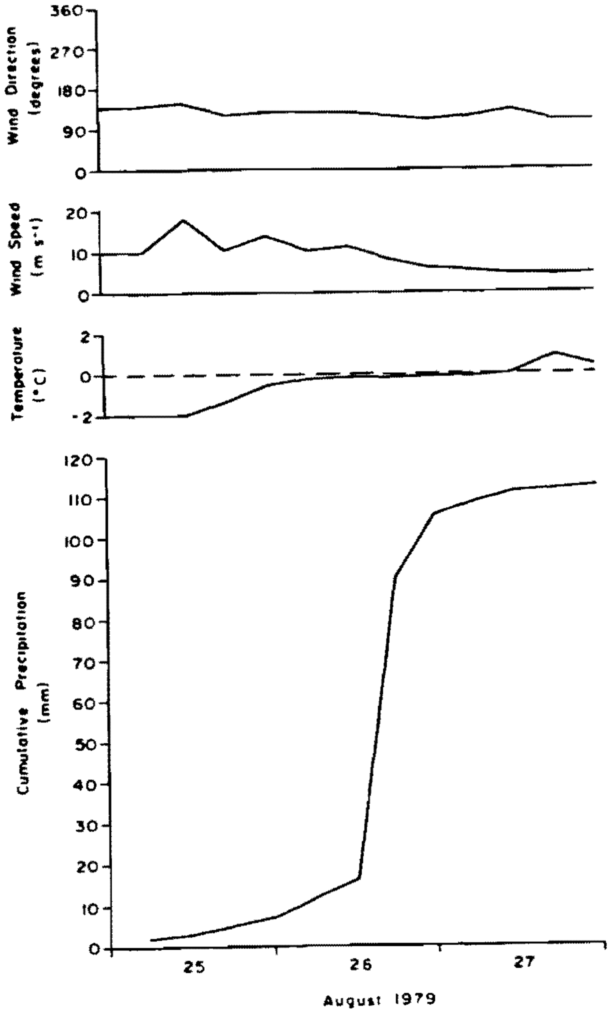


Figure 4.4: Storm plot of 25 August, 1979 to 27 August, 1979. Note majority of precipitation received during a period of rising temperatures. This situation is typical of an inverse storm.

For the period following this, in which the majority of the precipitation was received, temperatures rose consistently. The effect of such a trend in temperature on the structure of the new snow and implication for stability are discussed in Chapter 6. Unlike dry avalanche release in the Craigieburn Range, wet avalanche release is distinctly related to periods in which temperatures rise from near $0^{o}C$ to well above $0^{o}C$. Often associated with these periods of rising temperatures are strong north-westerly winds which may be accompanied by rain. Wet avalanche release in such cases results from bond breakdown due to excess free water in the snowpack.

4.3.3.3 Wind Related Variables

Judson (1964) found windspeed to be the single most important factor contributing to slab formation, while Bovis (1977) noted little difference in windspeed between avalanching and non-avalanching situations. Seligman(1936), Mellor (1968), Perla and Martinelli (1976) and Salway (1979) all agree that above a critical level wind becomes less important to avalanche activity. For the Craigieburn Range avalanche probability increases rapidly with mean windspeeds in the range of 6 to 11 ms^{-1} (Figure 4.1k) but above this latter figure the probability drops from 100 percent to 75 to 50 percent with mean windspeeds of 12 to 15 ms^{-1}. In such cases, snow redeposition may be reduced due to sublimation effects or increased turbulence on lee slopes. Although there is an absence of data points in the 5 to 8 ms^{-1} range (Figure 4.1j) increases in maximum windspeed of greater than 3 to 4 ms^{-1} are shown to be effective in increasing the avalanche probability. These increases may determine the relative degree of snow transport and redeposition during a storm. This agrees well with Akkouratov (1966) and Kotlyakov and Plam (1966) who noted that marked periods of snow drifting associated with windspeed increases produce definite periods of avalanching. Such wind redistribution process and the high mean dry avalanche storm windspeed of 11 ms^{-1} for the Craigieburn Range compared with that found by Perla (1970) and Salway and Moyse (1978) may account for the smaller amounts of total precipitation required for avalanche release in the Craigieburn Range.

Of the ten sequences of wind direction/speed and precipitation data (Appendix 5, a-o) for dry avalanche cycles, five show the majority of the precipitation from the north-west. While a generally westerly component has been previously identified by O'Loughlin (1969) for bringing precipitation to the Craigieburn Range, an important easterly component has been revealed in this study for precipitation production. Using wind vector components as suggested by Barry and Perry (1973, p.22), the importance of strong north-westerly and south-westerly flows in producing a high avalanche probability is evident (Figure 4.1h and i). Additionally, the significant easterly component related to strong southerly to easterly conditions (Figure 4.1i) is shown as important for increasing avalanche probability. All wet avalanche cycles except one were related to north-west events, the remaining one being related to rising temperatures associated with a warm anticyclonic south-westerly flow. The implications of the general air flows associated with cycles of avalanching in the Craigieburn Range will be discussed in Chapter 5.

4.3.3.4 Relative Humidity

For this study relative humidity displays little effect on avalanche probability verifying the findings of Perla (1970). With relative humidities of 70 percent or greater probability remains at 60 percent (Figure 4.1 2). Seligman (1936) proposed that snow sinters more readily into a slablike texture when humidity is high. Salway (1979) suggested that during high wind transport high humidity would supress sublimation, leaving more snow to be redistributed. As noted by LaChapelle (1977), the effect of relative humidity on avalanche occurrence appears to be unclear. This also holds true for this study. Though a higher mean relative humidity is apparent for dry avalanching compared to non-avalanching situations, (83 and 70 percent respectively) this most likely reflects the higher humidities naturally associated with new snow increments.

4.3.4 Critical Levels For Dry Avalanche Formation

Though it has been shown from the foregoing analysis that the wind and temperature variables for avalanche formation generally parallel those critical values and trends proposed by Perla (1970), the precipitation related variables for the Craigieburn Range show some contrasts. Average avalanche storm precipitation intensities for the Craigieburn Range fall far below those of the Alta, Utah area but approximate more closely those of 1.33 mm hr^{-1} for

Loveland-Berthoud Pass, Colorado. For this latter area precipitation intensities required to initiate instability are reduced to 0.88 mm hr^{-1} during periods in which windspeeds exceed 4.2 ms^{-1}. This is comparable to the average avalanche storm precipitation intensity for the Craigieburn Range which may be considered a windy environment with average windspeeds of 11 ms^{-1} during periods of avalanche formation. Similarities between new snow amounts required to initiate instability also exist for these two areas. Only 100 mm is required for the Loveland-Berthoud Pass area while a 100 percent avalanche probability is realised in the Craigieburn Range with total new snow of 150 mm. Both these values contrast markedly with Alta, Utah and attest to the importance of wind as a contributory variable to avalanche formation in the Craigieburn Range. Similar to Perla (1970), critical values for dry avalanche formation are proposed for the Craigieburn Range (Table 4.1). These have been derived from an analysis of individual storm plots and the points at which P_{10} attains a 100 percent probability of occurrence. Such values may be used as useful guidelines for qualitative assessment of avalanche potential in the Craigieburn Range.

Table 4.1: Critical Values For Dry Avalanche Formation (24 Hourly Averages)

Variable	Critical Value
New Snow (mm)	150
Total Storm Precipitation Water Equivalent (mm)	12.5
Maximum 6 Hour Precipitation Intensity (mm hr^{-1})	1.5
Mean Storm Temperature (^{o}C)	-3 or less
Maximum Storm Temperature (^{o}C)	-3 or less (dry avalanche 3 or above (wet avalanche)
Temperature Range (^{o}C)	7 or greater
Maximum 6 Hour Windspeed Increase (ms^{-1})	4 or greater when 24 hour precipitation is 12.5 mm, or mean windspeed in previous 24 hours has been at least 3 to 4 ms^{-1}
Mean Windspeed (ms^{-1})	6 to 7 over 24 hours during precipitation
Wind Direction	A strong north-westerly, south-westerly or south-easterly component with precipitation of 12.5 mm or greater in 24 hours

4.4 APPLICATION OF DISCRIMINANT FUNCTION ANALYSIS TO AVALANCHE FORECASTING

In this study discriminant function analysis was chosen because:

(1) more than one situation may be considered at a time, whether it be a dry, wet or non-avalanche situation;

(2) as the number of non-avalanche days are greater than avalanche days, these can be reduced;

(3) the effects of observational errors may also be controlled;

(4) where non-linearity of explanatory variables and the dependent variable in regression analysis presents a problem, no such problems are apparent in discriminant analysis;

(5) comparability with other studies is enhanced.

4.4.1 Methods

4.4.1.1 Nature of the Data

Avalanche occurrence and meteorological data were collected as outlined in Section 2.3.7. As discriminant analysis involves the grouping of data into 'a priori' groups the data set was split into three, comprising dry, wet and non-avalanche days. A dry and wet avalanche day were defined as a day on which an avalanche of magnitude 2 or greater occurred. The non-avalanche day set should be strictly bounded by the first and last recorded avalanche occurrences. This was not possible, as the 'avalanche season' proved to be very short so days immediately outside the time limit were selected (Table 4.2). Those days chosen

Table 4.2: Periods From Which Days Selected For Discriminant Analysis

1979	15 July to 20 September
1980	11 June to 2 September
1981	25 July to 19 September

lying outside the strict 'avalanche season' were those for which total snowpack depth was greater than 1 m and precipitation occurred as snow. An equal number of avalanche and non-avalanche days could not be selected from the respective years as some seasons had a higher frequency of avalanche cycles thus reducing the available days from which non-avalanche days could be selected. Fifteen non-avalanche days were selected by random sampling. This approach reduces serial correlation between days as might arise with the persistence of a weather pattern (Bois et al., 1975; Bovis, 1976, 1977). Sampling errors for parameters in each population of events are also equalised. There were 17 avalanche days in the sample of which seven were wet avalanche days. Once avalanche and non-avalanche days had been identified, reduction of meteorological data was performed to produce the variables displayed in Table 4.3.

4.4.1.2 Variable Selection

The discriminant analysis was initially performed using all variables recorded in the period 24 hours prior to

Table 4.3: Variables Initially Considered in Discriminant Analysis

Variable Number	Variable Symbol	Variable Description
1	TEMP	Mean of 2 hourly temperatures in period of 24 hours up to 1200 hours on sampled day (^oC)
2	MAX	Maximum 2 hourly temperature in same period as (1) above (^oC)
3	MIN	Minimum 2 hourly temperature in same period as (1) above (^oC)
4	$TEMP_{1,2,3}$	Mean 2 hourly temperautre over 1, 2 and 3 day period prior to sampled day (^oC)
5	$MAX_{1,2,3}$	Maximum 2 hourly temperature in same period as (4) above (^oC)
6	$MIN_{1,2,3}$	Minimum 2 hourly temperature in same period as (4) above (^oC)
7	RH	Mean of 2 hourly relative humidity in same period as (1) above (%)
8	$RH_{1,2,3}$	Mean relative humidity over 1, 2 and 3 day period prior to sampled day (%)
9	PPT	Total precipitation (water equivalent) in 24 hours prior to 1200 hours on sampled day (mm)
10	$PPT_{1,2,3}$	Cumulative precipitation over 1, 2 and 3 day period prior to sampled day (mm)
11	PI	Maximum 6 hourly precipitation intensity in same period as (9) above (mm)
12	RAD	Total global radiation in same period as (9) above ($MJm^{-2}d^{-1}$)
13	NS	Total new snow in 24 hours up to 0900 hours on sampled day (mm)
14	NS_2	Total new snow in 48 hours up to 0900 hours on sampled day (mm)
15	Z	Total snowpack depth at 0900 hours on sampled day (mm)
16	U_1	Mean 800 mb westerly wind component in 24 hours prior to 1200 hours on sampled day (ms^{-1})
17	V_1	Mean 800 mb southerly wind component in same period as (16) above (ms^{-1})
18	SD.WD	Standard deviation of 800 mb wind direction from a reference point of 60 degrees in same period as (16) above (degrees)
19	M.In	Maximum 6 hourly 800 mb wind speed increase in same period as (16) above (ms^{-1})
20	WD	Mean 800 mb wind speed in same period as (16) above

1200 hours on the sampled day. Fourteen variables were therefore included in the analysis. The criteria for variable selection relies on the maximisation of the F ratio and the coincident minimisation of the Wilks' Lambda. A stepwise discriminant procedure (Klecka, 1971, p.447) was used in variable selection. When the F ratio and associated Wilks' Lambda fell below a prescribed level (significance of 0.01) then computation and further variable selection ceased. Bois et al. (1975) used a Wilks' U test and the Mahalanobis distance as criteria for variable selection. Bovis (1976, 1977) and Drozdovskaya (1979) both used the Mahalanobis distance for assessing variable significance. Wilks' Lambda has been used by Fohn et al. (1977) and Obled and Good (1980) for discriminating between avalanche and non-avalanche days. Eisenbeis and Avery (1972) noted that Wilks' Lambda is the best selection criterion given the accuracy of the F and Chi-squared tests. However, if the assumption of equal dispersion matrices is violated both the F and Chi-squared tests and hence the Wilks' Lambda are biased.

Initially, simple discriminant analyses were performed between dry and non-avalanche days, dry and wet avalanche days, and wet and non-avalanche days. Variables selected per discrimination differed either in their order of selection or their specific effect on the likelihood of avalanching.

4.4.2 Results and Discussion

4.4.2.1 Dry Avalanche Day Non-Avalanche Day Discrimination

Six variables were included in the discriminant function giving a significant discrimination between dry and non- avalanche days:

$$D_i = 0.123\ PI + 0.033\ SD.WD - 0.239\ RAD + 0.001\ Z$$
$$-0.104\ MIN + 0.020\ PPT - 2.240$$
$$\text{with } D_o = 0.77 \qquad (4.2)$$

PI was selected as the best of the range of variables for the discrimination. This has been widely recognised as a critical factor in snow stability (Atwater, 1952; Judson, 1964, 1966; Perla, 1970; Tesche and Tocke, 1976; LaChapelle, 1978). The second most important variable selected is that of SD.WD. Low SD.WD values are related to consistent north-westerly conditions in which PI is high. High SD.WD values represent those storms in which there is a major wind direction change. Non-avalanche days display low SD.WD values but low PI in contrast to that of dry avalanche days. The remaining precipitation variables (PPT and Z) probably represent increased loading, while increased Z reduces roughness of potential sliding surfaces. The inclusion of RAD reflects the overcast conditions ($\overline{RAD}$ = 3.3 $MJm^{-2}d^{-1}$) and low cloud bases associated with precipitation conditions. This contrasts with the often cloudless and precipitation free non-avalanche days ($\overline{RAD}$ = 6.6 $MJm^{-2}d^{-1}$). Although MIN values are similar for avalanche and non-avalanche days ($\overline{MIN}$ = -3.8 and -3.2^{o}C respectively). Persistence of low temperatures during periods of snowfall are important in hindering new snow settlement and sintering as discussed in Section 4.3.3.2.

For the majority of previous discriminant analysis studies (Judson and Erickson, 1973; Bois et al., 1975; Bovis 1976, 1977; Fohn et al., 1977; Drozdovskaya, 1979) it has also been shown that precipitation, minimum temperature and wind variables are of significance in discriminating between dry avalanche and non-avalanche days. The role of global radiation though has only been previously identified in European work (Bois et al., 1975; Fohn et al., 1977; Drozdovskaya, 1979) but not in North America (Judson and Erickson, 1973; Bovis, 1976, 1977; Salway and Moyse, ,1978; Salway, 1979).

Two avalanche day events are misclassified and lie in the non-avalanche day side of the discriminating index (Figure 4.5a). Overall, 92 percent of grouped cases are classified correctly, with 20 percent of dry avalanche days misclassified, similar to that of Drozdovskaya (1979). Bovis (1977) has found that 32 percent of dry avalanche days and 9 percent of non-avalanche days were misclassified while Salway (1979) presents misclassification percentages of 17 percent for dry avalanche and 24 percent for non-avalanche days. For this study, the probability of misclassifying a given avalanche day to the non-avalanche day group (and vice versa) is 17 percent and compares favourably with those of Judson and Erickson (1973) ranging from 21 to 30 percent for eight avalanche paths. Bois et al. (1975) have obtained probabilities of misclassification as low as 5 and 8 percent for dry and wet avalanche days respectively. Confidence limits were fitted to the range of discriminant scores for

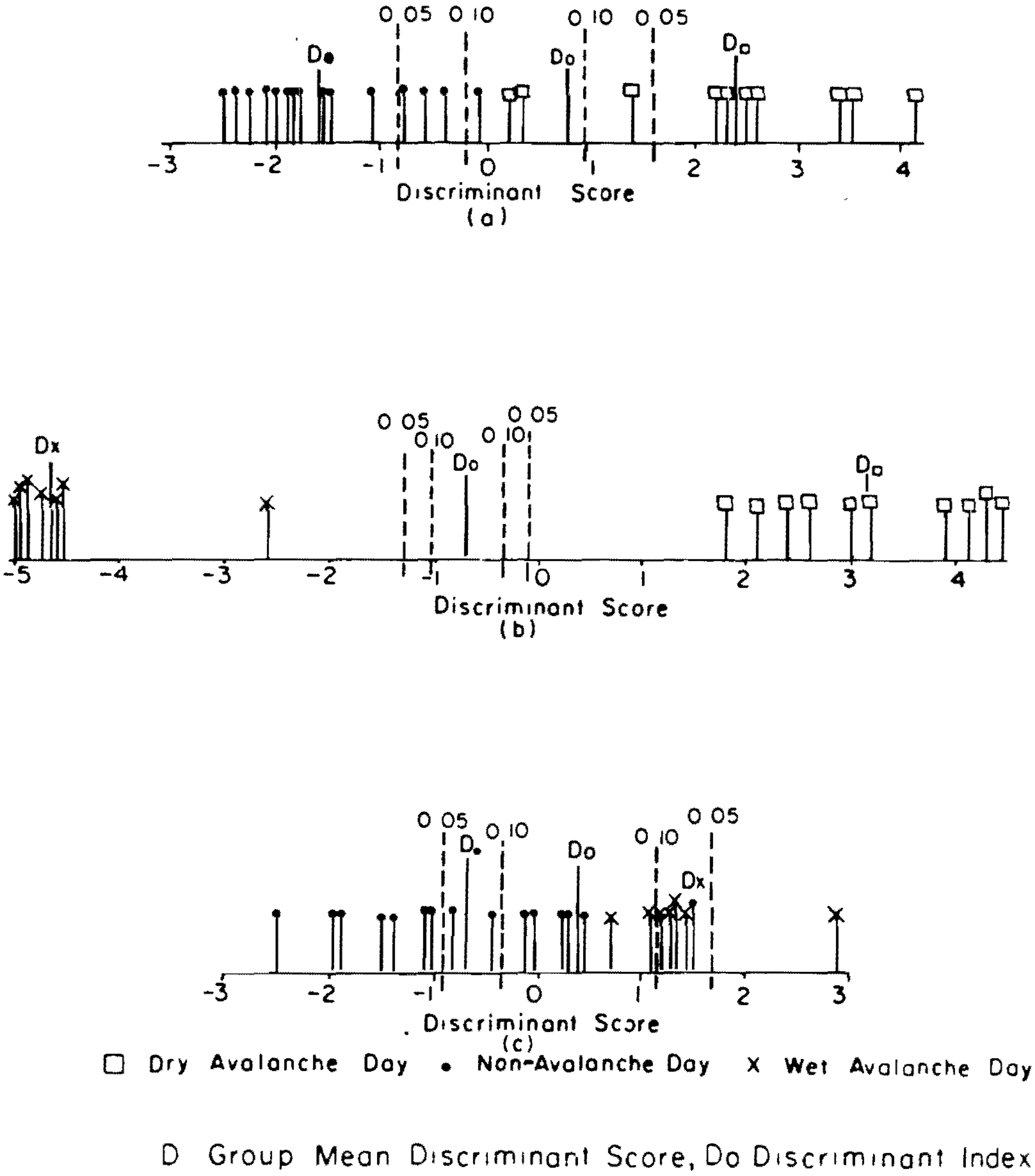


Figure 4.5: Distribution of discriminant scores.

(a) dry avalanche day non-avalanche day discrimination

(b) wet avalanche day dry avalanche day discrimination

(c) non-avalanche day wet avalanche day discrimination dry and non-avalanche days according to the method of Freeze (1964) (Figure 4.5a). Individual event days were also assigned a probability of being classified as dry or non-avalanche day according to the method of Bovis (1976). (Appendix 6). The individual percentage probabilities are shown in Table 4.4. These range for dry avalanche days from 2 to 96 percent. The two events showing probabilities of less than 3 percent, are the ones misclassified in the discriminant model and also lie in the region of uncertainty defined by the confidence limits. The three largest probabilities coincide with the three largest and most widespread avalanche cycles. A correlation of r = 0.61 (significant at 0.01 level) is revealed if avalanche probability (P) is used to predict MAG; MAG = 11.6 + 0.27 P. (4.3). Large probabilities are therefore associated with widespread occurrence.

4.4.2.2 Dry Avalanche Day Wet Avalanche Day Discrimination

Days classified as wet avalanche days included both wet loose and wet slab avalanches. In all cases, wet avalanche cycles included both morphological types. All dry and wet avalanche days were classified correctly with an overall probability of misclassification of 8 percent. The discriminant equation produced was:

$$D_i = -0.991\ \text{TEMP} + 0.04\ \text{SD.WD} + 0.151\ \text{PI} - 0.340\ \text{RAD} + 0.167\ \text{M.In} - 0.075\ U_1 - 2.955 \qquad (4.4)$$

with $D_o = -0.70$ (Figure 4.5b).

Table 4.4: Percentage Probability of Classification as Dry Avalanche or Non-Avalanche Day

Date	Percentage Probability of Classification Dry Avalanche Day	Non-Avalanche Day
15/7/79	48	<1
29/7/79	86	<1
05/8/79	64	<1
14/8/79	88	<1
11/6/80	53	<1
16/8/80	96	<1
16/8/81	<2	<3
25/8/81	57	<1
27/8/81	<3	<3
05/9/81	39	<1

The selection of TEMP emphasises the importance usually placed on air temperature for classifying days as potentially wet or dry avalanche days. Mean TEMP for wet avalance days (1.4^{o}C) is much greater than dry avalanche days (-2.6^{o}C). PI, SD.WD and RAD feature again in this discrimination and are related to dry avalanche release. Both RAD and TEMP have negative discriminant coefficients indicating that increases in these variables will increase the likelihood of wet avalanche activity. This agrees well with Bois et al. (1975) who noted the importance of temperature at 1300 hours on the day of an event and absorbed radiation flux. Bovis (1976) also noted the importance of mean air temperature for wet avalanche release through melt-water generation and bond breakdown. For dry and wet avalanches M.In and U_1 are very similar showing that wind-speed increases and winds with a strong westerly wind component are both important for these types of avalanching. Lower SD.WD values for wet avalanches days (SD.WD = 30.6) in contrast to dry avalanche days (SD.WD = 51.0) attest to the consistent north-west flows typical of wet avalanche cycles. When associated with low SD.WD values, high PI and TEMP values increase the probability of wet avalanching through the effects of rain. However, if TEMP decreases or remains low with an increase in PI the likelihood of dry avalanching is increased. A greater frequency of wet avalanche release in the spring was noted by Bovis (1977) for the San Juan mountains. No such trend was evident for this study with the main features distinguishing dry from wet avalanche days being above freezing temperatures, increased global radiation and strong north-west flows. Under such conditions wet snow avalanches may occur at any time during the winter in the Craigieburn Range.

4.4.2.3 Wet Avalanche Day Non-Avalanche Day Discimination

All wet avalanche days were classified correctly while 7 percent of non-avalanche days were incorrectly classified. Overall probability of misclassification was 24 percent. The discriminant equation produced consists of four variables:

$$D_i = 0.987\ \mathrm{TEMP} - 0.632\ \mathrm{MIN} + 0.037\ \mathrm{RH} + 0.098\ \mathrm{WS} - 3.99 \qquad (4.5)$$

with $D_o = 0.34$ (Figure 4.5c)

Temperature variables feature predominantly in this discrimination. The selection of TEMP and MIN and consideration of coefficient signs indicates that increases in both these values will increase the likelihood of wet avalanches. Of note is the absence of any precipitation variables in this discrimination. Though wet and non-avalanche days are characterised by low precipitation amounts, there are great differences in the means of TEMP and MIN (Table 4.5).

Table 4.5: Mean and Minimum Two Hourly Temperatures For Wet and Non-Avalanche Days - Variances in Brackets

	Avalanche Day	
	Wet	Non
TEMP (^{o}C)	1.4 (3.0)	-2.2 (8.0)
MIN (^{o}C)	-0.2 (5.5)	-3.2 (9.9)

Increases in RH and WS values also result in an increase of the likelihood of wet avalanches. The role of humidity in wet avalanche release is not as clear as for other variables in the analysis. Higher mean humidities and lower variances are displayed by wet avalanche days ($\overline{RH}$ = 77 percent s^2 = 49). This lower variance attests to the constant atmospheric moisture conditions during periods of north-westerly flow. Likewise, higher humidities are probably associated with cloud cover frequently present during strong north-westerly conditions, or precipitation related wet avalanche cycles. Under such conditions high humidities may be important for melt through latent heat production from condensation (Prowse, 1981; Moore, 1983). Associated with high WS values are strong westerly wind components. In addition, significant inverse relationships exist between TEMP and V_1 indicating that warmer air comes from the north. Wet avalanche days then are characterised by warm strong north-westerly conditions. Such conditions are conducive to high sensible heat production as suggested by Prowse (1981) and Moore (1983) who note that 57 percent and up to 78 percent respectively of total heat inputs for snowmelt may come from sensible heat production.

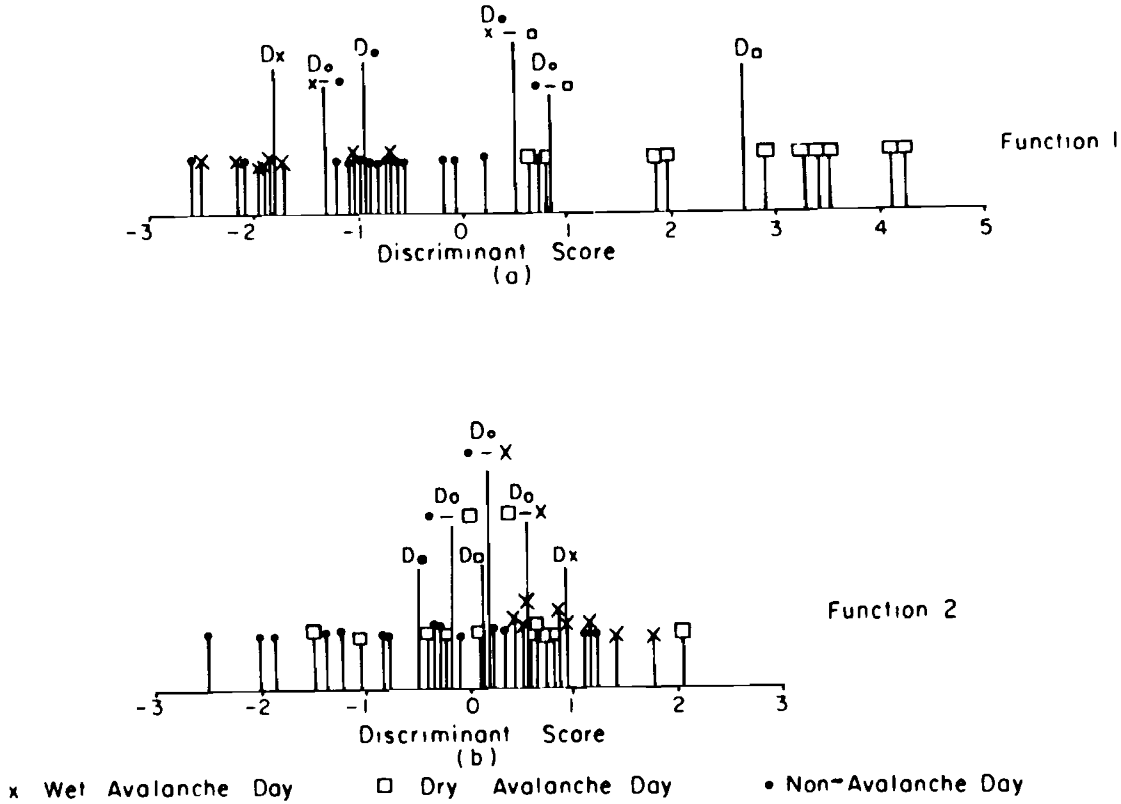
4.4.3 Two Discriminant Function Forecasting Model

Bovis (1977) noted that where there is a separation between dry and wet avalanche seasons, no problem exists in applying a single discriminant function. However, as wet avalanches are likely to occur at any time throughout the winter in the Craigieburn Range, and given the logistics of using different input variables for two simple discriminant functions, a discriminant model utilising the same variables but allowing classification into one of three possible situations appears more attractive. A two discriminant function analysis was therefore performed with dry, wet and non-avalanche day groups. The same variables and procedures outlined in Section 4.4.1 were also used in this analysis.

4.4.3.1 Results and Discussion

The two discriminant functions produced were significant at the 0.01 level. Initially each function comprised seven variables with 91 percent of all three grouped classes being classified correctly. The number of variables was subsequently reduced to four, as such a large number of variables may prove cumbersome in any real time application of the model. The four variables and related coefficients resulting in 78 percent of all grouped classes being classified correctly appear in Table 4.6. Individual class percentage classification results appear in Table 4.7.

Analysis of the standardised discriminant coefficients (Table 4.8) allows some inference to be drawn as to the variables these functions may represent. This is analogous to factor 'naming' in factor analysis (Klecka, 1975, p.443). The high loadings on PI and SD.WD and the relationship between these two variables discussed previously indicates that function one is primarily a 'precipitation function'. Function two in contrast, is dominated by TEMP. The similar function one scores (Figure 4.6a) for wet and non-avalanche days attests to their similarity in having very little or no precipitation. In contrast function two scores (Figure 4.6b) show that wet and non-avalanche days are most dissimilar.



D Group Mean Discriminant Score, Do Group Discriminant Index

Figure 4.6: Distribution of discriminant scores for three avalanche day discrimination.

(a) Function 1 scores

(b) Function 2 scores

Table 4.6: Variables and Discriminant Function Coefficients For Three Group Avalanche Day Discrimination

Variable	Coefficients Function 1	Function 2
PI	0.177	0.067
TEMP	-0.269	0.389
SD.WD	0.031	0.003
RAD	-0.151	0.013
Constant	-1.443	0.095

Table 4.7: Classification Results Using Four Variable Discriminant Model

		Predicted Group Membership (%)		
Group	No.of Cases	Dry Avalanche Days	Wet Avalanche Days	Non-Avalanche Days
Dry Avalanche Days	10	80.0	0.0	20.0
Wet Avalanche Days	7	0.0	72.0	28.0
Non-Avalanche Days	15	0.0	20.0	80.0

Table 4.8: Standardised Discriminant Function Coefficients

Variable	Coefficients Function 1	Function 2
PI	0.909	0.343
TEMP	-0.606	0.876
SD.WD	0.874	0.095
RAD	-0.525	0.046

This reflects the difference in the many clear cool non-avalanche days to that of the warmer overcast wet avalanche days. The power of function one, which explains 90 percent of the variance, is best realised when considering the probabilities of misclassification for the separate functions (Table 4.9).

Table 4.9: Percentage Probability of Misclassification for Function One and Function Two

		Group	
	Dry Avalanche Day	Wet Avalanche Day	Non-Avalanche Day
		Function 1	
Dry Avalanche Day		15	17
Wet Avalanche Day	33		34
Non-Avalanche Day	35	23	
		Function 2	

While function one is effective in discriminating between dry and wet and dry and non-avalanche days the high probability of misclassification for the wet non-avalanche discrimination points to the weakness of this 'precipitation function' for these two groups. The associated probability of misclassification for these two groups however is greatly improved when using function two or the 'temperature function'. The combined effects of these two functions are best demonstrated in a bifunctional plot of the discriminant scores on a territorial map with borders of regions of like classification (Figure 4.7). Greatest difference between wet and non-avalanche days is along the vertical temperature function axis. Likewise greatest similarity is shown between dry and non-avalanche days along this axis as their

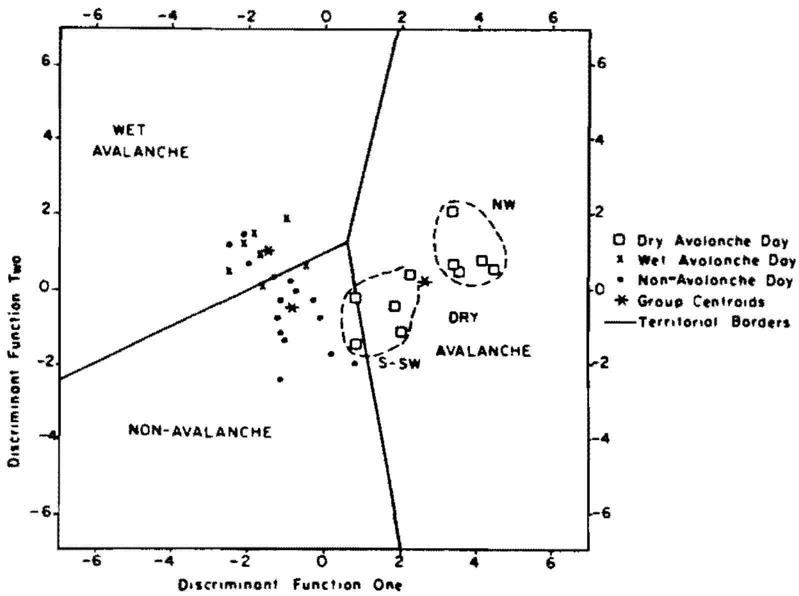


Figure 4.7: Bifunctional plot of discriminant scores for dry, wet and non-avalanche days. Note higher function one and two scores for NW storms.

respective mean temperatures are very similar. The 'precipitation function' is instrumental in separating dry and non-avalanche days on the grounds of the contrasting effects of PI and SD.WD for these two days as discussed previously. Compared to wet and non-avalanche days, dry avalanche days are dispersed along both the precipitation and temperature function axes. This is also reflected in the larger variance of function one scores of dry avalanche days (s^2 = 1.56), in comparison to that of the wet and non-avalanche days, (s^2 = 0.31 and 0.73 respectively). Some difference in the meteorological characteristics of dry avalanche days may therefore exist given such a large variance.

Fohn et al. (1977) and Obled and Good (1980) found that avalanche day groups may contain subgroups. They identified separate dry avalanche day groups related to the specific characteristics of the meteorological variables generating instability. Dry avalanche storms for the Craigieburn Range were split into north-west and south to south-west groups according to the relative strengths of the westerly and southerly components. The temperature and precipitation characteristics of individual storms were subsequently assessed. All north-west storms were found to have higher function one and function two scores indicating higher precipitation intensities and warmer temperatures. Likewise south to south-westerly storms with lower function one and two scores have colder temperatures and smaller amounts of precipitation compared to that of north-westerly storms (Table 4.10).

Table 4.10: Mean Meteorological Variables For North-West and South to South-West Dry Avalanche Storms

Variable	North-West	South to South-West
PI (mm)	17.6	7.1
TEMP (^{o}C)	-1.5	-3.6
PPT (mm)	29.2	9.2
Precipitation Intensity over 6 hours (mm hr^{-1})	2.93	1.18
over 24 hours (mm hr^{-1})	1.21	0.38

For this two discriminant function analysis the two days misclassified belong to the south to south-west group of storms. These two storms in the simple discriminant analysis possessed probabilities of classification as a dry avalanche day of <3 percent due to the low 24 hour precipitation intensities of 0.08 mm hr^{-1} and 0.26 mm hr^{-1}. Avalanching on these two days occurred mainly because of high windspeeds of 7.6 ms^{-1} and 9.6 ms^{-1} respectively resulting in high rates of wind loading.

4.4.4 Real Time Application of an Avalanche Forecasting Model

Most real time forecasting of avalanche occurrence is performed between 0600 and 0800 hours. As the variable coefficients for avalanche day discrimination (Table 4.6) are produced from variables in the time period 24 hours prior to 1200 hours on the sampled day this may appear to pose some problems given a 0600 to 0800 hour time of issue of any avalanche forecast. No problem exists with this as variable values should be integrated from 1200 hours on the day prior to the time the forecast is being issued. In addition projected values of PI and SD.WD obtained from analysis of synoptic and numercial weather forecasting models may also be substituted into the model with avalanche probability prediction being made up to 24 hours ahead. Once the forecast has been issued it can be updated with further addition of data. At the end of the current forecast day the discriminating variables would have been evaluated for the period up to 2400 hours thus forming part of the integrated input for the following day's forecast. Using this procedure a continual evaluation of the forecast may be made with the integration of meteorological data over the preceding 24 hours.

Though two levels of discriminant function models have been presented it is recommended that for quick estimates of the likelihood of a day being classified as a dry, wet or non-avalanche day, the two function model in conjunction with the territorial map (Figure 4.7) should be used. However, if a specific level of hazard is required and assuming a high percentage probability represents a high and widespread occurrence the procedure outlined in Figure 4.8, using the three simple discriminant functions should be followed.

4.4.5 Model Testing

To test the validity of the two function discriminant model an independent set of data for 1982 was used. Avalanche occurrence and meteorological data were collected as part of the Craigieburn Range Forecasting Scheme instigated by The New Zealand Mountain Safety Council (Fitzharris et al.,1983) and now run by The Forest Research Institute (Weir, 1983). The dry and wet avalanche days used in the

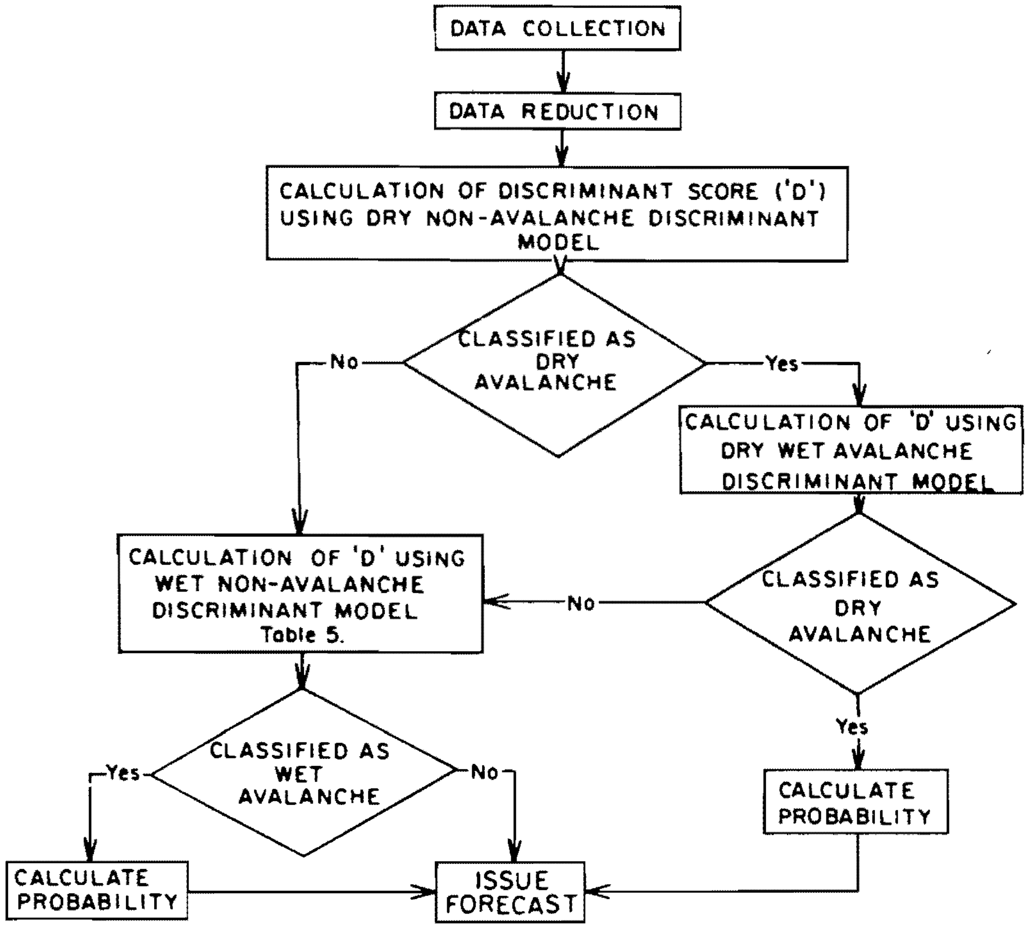


Figure 4.8: Suggested procedure for issuing avalanche forecast using simple two avalanche day discriminant model.

analysis and the related meteorological variables are shown in Table 4.11. Only 24 hourly as opposed to two hourly means were available for temperature in this analysis. Also adjusted radiation values allowing for elevational differences between the CF and BR sites were not available.

Table 4.11: Variable Values Used In Independent Test of Two Function Discriminant Model

		Variable		
Date	PI	TEMP	SD.WD	RAD
25/6/82	6	-1.05	49	1.4
14/7/82	7	-6.8	13	6.9
19/7/82	12	-1.1	18	2.4
20/7/82	6	-4.8	21	4.3
22/7/82	3	-7.5	28	7.0
30/7/82	6	0.5	119	5.1
03/8/82	0	-2.2	6	8.2
26/8/82	12	-2.6	70	11.3
01/9/82*	7	-1.3	29	10.6
11/9/82*	10	2.9	80	8.3
12/9/82*	6	-1.1	43	6.6
28/9/82*	0	-1.5	28	18.7
29/9/82*	0	-3.8	32	17.9

*Wet Avalanche Days

Three of the eight dry avalanche days are misclassified (Figure 4.9). This corresponds to a misclassification of 37 percent. One of the misclassified cases, however, lies very close to the border discriminating between avalanche and non-avalanche situations. This event (22 July 1982) followed close on a major avalanche cycle 24 to 36 hours prior to the meteorological readings for that day. In such a situation conditions should still be considered marginal with an appropriate hazard warning being posted. All wet and non-avalanche days except one for both groups respectively were classified correctly. Overall, 75 percent

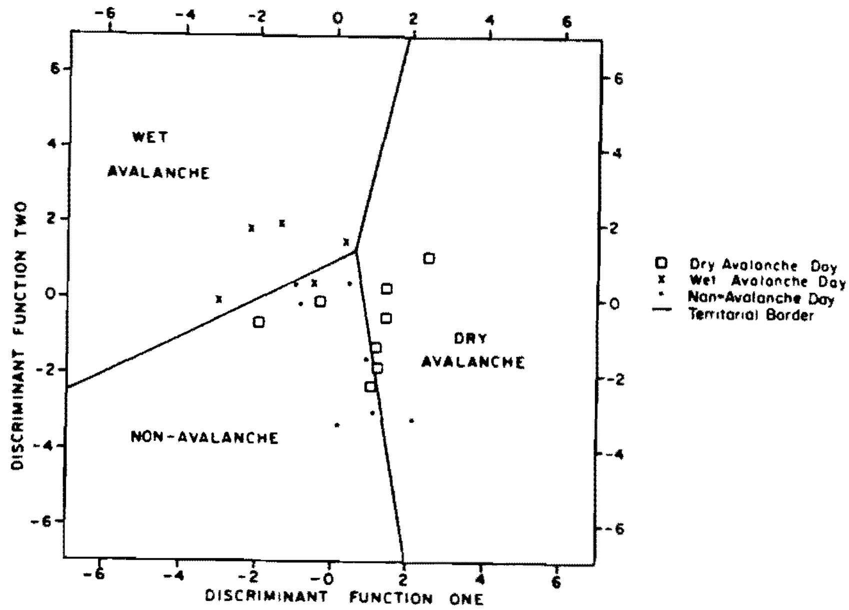


Figure 4.9: Bifunctional plot of independent set of dry, wet and non-avalanche day discriminant scores for 1982.

of all grouped classes were classified correctly, 2 percent less than with the original data. Testing with the 1982 independent data set has shown that the model is valuable in a forecasting situation. The slightly increased percentages of misclassification for dry avalanche days may be accounted for by the contrasts in the winters of 1979 to 1981 from which the model was developed, to that of 1982 characterised by an early season cold period with very little fresh snow. This resulted in a very weak snowpack susceptible to failure under small loadings of snow (Weir, 1983).

4.5 SUMMARY

Two broad approaches to assessing the importance of a variety of meteorological variables in avalanche formation and release have been discussed. Assessment of the role of various meteorological variables in avalanche formation through the consideration of avalanche probability and storm plots has indicated that precipitation and temperature related variables are the most important. Of prime significance are precipitation intensity and mean storm temperature. However, while precipitation intensity has been shown to be significant in avalanche formation, the threshold values for the Craigieburn Range fall into the lower end of the range identified by Perla (1970). This is thought to be a function of the windiness of the Craigieburn Range snow environment. The relative importance of variables has been shown to differ given the contrasts between wet and dry avalanches. This is highlighted most in the statistical analysis of avalanche producing conditions by the use of discriminant function analysis. Two group discrimination between wet and dry avalanche and non-avalanche days respectively revealed that TEMP was the most important discriminator. Actual temperature trends however, differed in their effects on dry and wet avalanching. A decrease in temperature coincident with precipitation increased the likelihood of dry avalanches whereas the probability of wet avalanche release is enlarged with a rise in temperature. Precipitation related variables (notably PI) proved to be the most important in discriminating dry avalanche from non-avalanche days. The two function discriminant function model presented for a three avalanche day discrimination uses the same variables as in the three two group discriminations and are similar to those found by other workers. In the order of importance these are:

(1) maximum 6 hour precipitation intensity in period 1200 hours on day prior to 1200 hours on sampled day;

(2) mean 2 hour temperature in same period as (1) above;

(3) standard deviation of wind direction in same period as (1) above;

(4) global radiation in same period as (1) above.

An independent test of the model produced acceptable results. The model's applicability as an aid in avalanche forecasting therefore appears justified.

Assessment by both qualitative and quantitative methods has facilitated the establishment of the relative importance of a range of meteorological variables in avalanche formation. The significance of variables identified by the objective discriminant function analysis is mainly understood through the results of the subjective analysis of meteorological factors and avalanche occurrence. This highlights the significance of using these two methods in conjunction as important non-linear meteorological effects on avalanche occurrence may be masked in a purely objective analysis.

CHAPTER FIVE

SYNOPTIC STORM TYPES, ATMOSPHERIC PROCESSES AND SNOW AVALANCHING

5.1 INTRODUCTION

The previous chapter showed that heavy precipitation was the most important meteorological factor for avalanche release. Identification of the synoptic types and atmospheric processes associated with heavy precipitation events, in conjunction with those meteorological factors already identified, should therefore enable greater predictability of snow avalanches. The basic aims of this chapter then are to:

(1) classify those storms producing cycles of avalanching in the Craigieburn Range;

(2) examine the physical factors and processes contributing to high intensity and large precipitation events;

(3) consider some methods and factors useful in predicting such events to aid avalanche forecasting.

5.2 PREVIOUS WORK

Some work on synoptic scale patterns associated with heavy snowfalls and in some cases dealing specifically with snow avalanche release exists for both overseas and New Zealand (Table 5.1). The majority of the synoptic situations described feature fronts associated with well developed troughs or low pressure systems. These storm

Table 5.1: Synoptic Scale Situations Associated With Heavy Snowfall and Avalanching

Author	Location	Synoptic Scale Situation
Goree and Younkin (1966)	Western United States	Warm advection or vorticity advection storm associated with enclosed lower level cyclonic circulation with upper level meridional trough.
O'Loughlin (1969)	Craigieburn Range New Zealand	Low pressure/cold front passage, deepening depression.
Fox (1933)*	Cascade Mountains Washington State U.S.A.	Intense warm/cold front with a warm sector. Slow moving warm front, occlusions and cold front passage.
Thompson and Kells (1973)	Mt Ruapehu New Zealand	Low pressure systems.
Fitzharris (1976)*	Mt Cook New Zealand	Cold front passage in low pressure trough.
Neale and Thompson (1977)	South Island New Zealand	Quasi-stationary fronts, slow moving low pressure systems.
Hutcheon and Lie (1978)*	Juneau, Alaska	Intense pressure gradient between high pressure cell over northern Canada and low pressure cell over Gulf of Alaska.
Grischenko and Tokmakov (1978)*	Carpathians, Ukraine	Movement of fronts in troughs of cyclones from cold north-east sector.
Fitzharris and Schaerer (1980)*	British Columbia Canada	Quasi-stationary frontal systems and intensifying low pressure systems.
Owens and Prowse (1980)*	Coronet Peak New Zealand	South-west airflow following cold front passage
Prowse (1981)*	Craigieburn Range New Zealand	Deepening depression moving east of country and well developed trough and cold front.
Nakajima (1981)	Japan	Cyclones/cold fronts, expanding continental anticyclones, strong cyclonic flows with steep pressure gradients.

*Studies dealing specifically with snow avalanching or having implications for snow avalanche formation.

types appear to be particularly significant in New Zealand for snowstorms reaching to low elevations, where they have been referred to as warm advection and vorticity advection storms by Neale and Thompson (1977). It has also been shown that the processes associated with ascending motion, atmospheric thermal structure and moisture content are important in snowfall production. Secondary to these are topography, snowflake production and downward penetration of snow (Neale and Thompson, 1977).

Several attempts have been made to predict heavy snowfall using a variety of parameters such as temperature, dewpoint depression, pressure and winds from the surface to approximately the 10 mb level, 500 mb surface heights and 1000-500 mb thickness. The main aim of these is to improve forecasts for avalanche forecasting (Daly, 1978; Gigliotti and Parent, 1978). Results from such forecasts supplement information collected at the local scale, thus forming a firmer information base from which an avalanche forecast may be established. Such an approach is used in New Zealand with the Meteorological Service providing valuable upper level data to assist avalanche forecasting at the local scale (Fitzharris et al. 1983).

5.3 NATURE OF DATA AND METHODS

Precipitation, avalanche occurrence, synoptic scale weather pattern and upper air data used in this analysis has been collected as outlined in Chapter 2. Problems associated with these data are also discussed. Some significant avalanche producing storms from 1978 and 1982 have also been included with those from 1979 to 1981 in this analysis to augment the record.

Several approaches to classifying storm types were considered. The methods of Lamb (1950) were considered too subjective while the objective statistical techniques of Lund (1963) and McCutchan and Schroeder (1973) are beyond the scope of this analysis because of the short nature of the record. Storms were classified according to the method of Neale and Thompson (1977) which is based largely on rates of thermal advection (TA) during storms. These were calculated as outlined by McIntosh and Thom (1969). Vertical velocities (ω) were also calculated for the storms in this analysis using the method suggested by Barry and Perry (1973, p.45-46). This technique is especially suited to data collected at constant levels if diabatic effects are assumed negligible. Errors may arise if soundings are not strictly vertical with adiabatic lapse rates being overestimated by up to $1^{o}C\ km^{-1}$ (Staley, 1966). Temperature lapse rates, atmospheric layer thickness and heights and atmospheric moisture contents were extracted from radiosonde soundings while tephigrams were plotted and checked for conditions of stability or instability. These were also used to aid classification of storms.

The problem of assessing methods for predicting heavy snowfall events and periods of likely avalanche occurrence was approached at two levels. Firstly, individual indices of circulation intensity, proximity and thickness, and their effects on precipitation amounts for the following six hours within each storm were analysed. Secondly, multiple stepwise regression was used to establish on a storm by storm basis which upper level factors affect total storm precipitation and storm avalanche magnitude.

5.4 SYNOPTIC SCALE STORM TYPES

Storms which showed a variation in TA above and below $1^{o}C\ hr^{-1}$ over the period of precipitation, were classified as mixed warm advection/vorticity advections (WA/VA) storms. These produced the largest amounts of precipitation and resulted in widespread avalanche release along the Craigieburn Range. The remaining storms with TA consistently less than $1^{o}C\ hr^{-1}$ fall into the category of vorticity advection (VA) storms. These make up the majority of dry avalanche storms in this study (Table 5.2).

VA storms are related to the development of centres of low pressure over or in close proximity to the South Island, while WA/VA storms possess pronounced frontal development in a low pressure trough situation. These two types resemble closely the storm types identified by O'Loughlin (1969) and Prowse (1981). The storms with a well developed front/low pressure trough and low pressure centre identified by O'Loughlin (1969) and Prowse (1981) resemble the mixed WA/VA storms in this analysis. The remaining storms identified by these workers were generally characterised by a deepening depression moving east away from the country and match the situation of the VA storms identified here. Combining the results of the present analysis with those of O'Loughlin (1969) and Prowse (1981) (Table 5.3), 53 percent of storms producing snow in the Craigieburn Range are typified by the passage of a trough of low pressure and front over the South Island. This is similar to the situation found at Mt Ruapehu where 50 percent of the days with snowfall were accompanied by low pressure systems (Thompson and Kells, 1973).

Table 5.2: Classification of Dry Avalanche Storms

Storm Number	Date of Avalanching	Total Storm Precipitation (mm)	800 mb windspeed (WS)/ wind direction (WD) at time of maximum precipitation WD	WS	Maximum Rate of Thermal Advection $^{o}C\ hr^{-1}$	800-500 Thickness gpm	Storm Type	Magnitude
1	15/7/79	10.3	172	14	0.65	3471	VA	26
2	29/7/79	28.1	163	06	0.95	3403	VA	35
3	05/8/79	1.8	240	03	0.31	2538	VA	8
4	14/8/79	36.8	294	21	1.04	3575	Mixed WA/VA	30
5	11/6/80	51.2	313	32	3.86	3585	Mixed WA/VA	22
6	16/8/80	68.5	320	32	1.47	3707	Mixed WA/VA	63
7	16/8/81	2.4	216	07	0.56	3553	VA	18
8	25/8/81	21.0	068	11	0.83	3568	VA	20
9	27/8/81	28.0	177	20	0.53	3536	VA	24
10	05/9/81	19.7	258	16	0.30	3499	VA	13

Table 5.3: Frequency of Synoptic Storm Types Producing Snowfall in the Craigieburn Range (Percentage in Brackets)

Author	Types: Low Pressure Trough/ Frontal Systems	Low Pressure Systems	Total
O'Loughlin (1969)	23 (76)	7 (24)	30
Prowse (1981)	3 (20)	12 (80)	15
This Study	3 (30)	7 (70)	10
Total	29 (53)	26 (47)	55

5.4.1 Warm Advection/Vorticity Advection Storms

Storms 4, 5 and 6 (Table 5.2) show the features typical of WA/VA storms. These storms produced the greatest amounts of precipitation in the Craigieburn Range during the study period and also widespread avalanching. Storm 6 not only produced the largest avalanche cycle in the Craigieburn Range, but also resulted in widespread avalanching throughout the South Island. The development of storms 4 and 6 are outlined in Appendix 7.

Typical of these situations is a well developed trough of low pressure and cold front. Associated with this is a depression which may lie north-west or south-east of the South Island which generally moves eastward (Figure 5.1a and b). Preceding the front in such cases are strong moist north-westerly winds which are often coincident with periods of maximum precipitation (Table 5.2). Precipitation intensities and amounts reach a maximum when rates of thermal advection are highest as a result of the strong north-westerly flow.

Several distinct phases are present in the development of these mixed WA/VA storms (Figure 5.2a-e).

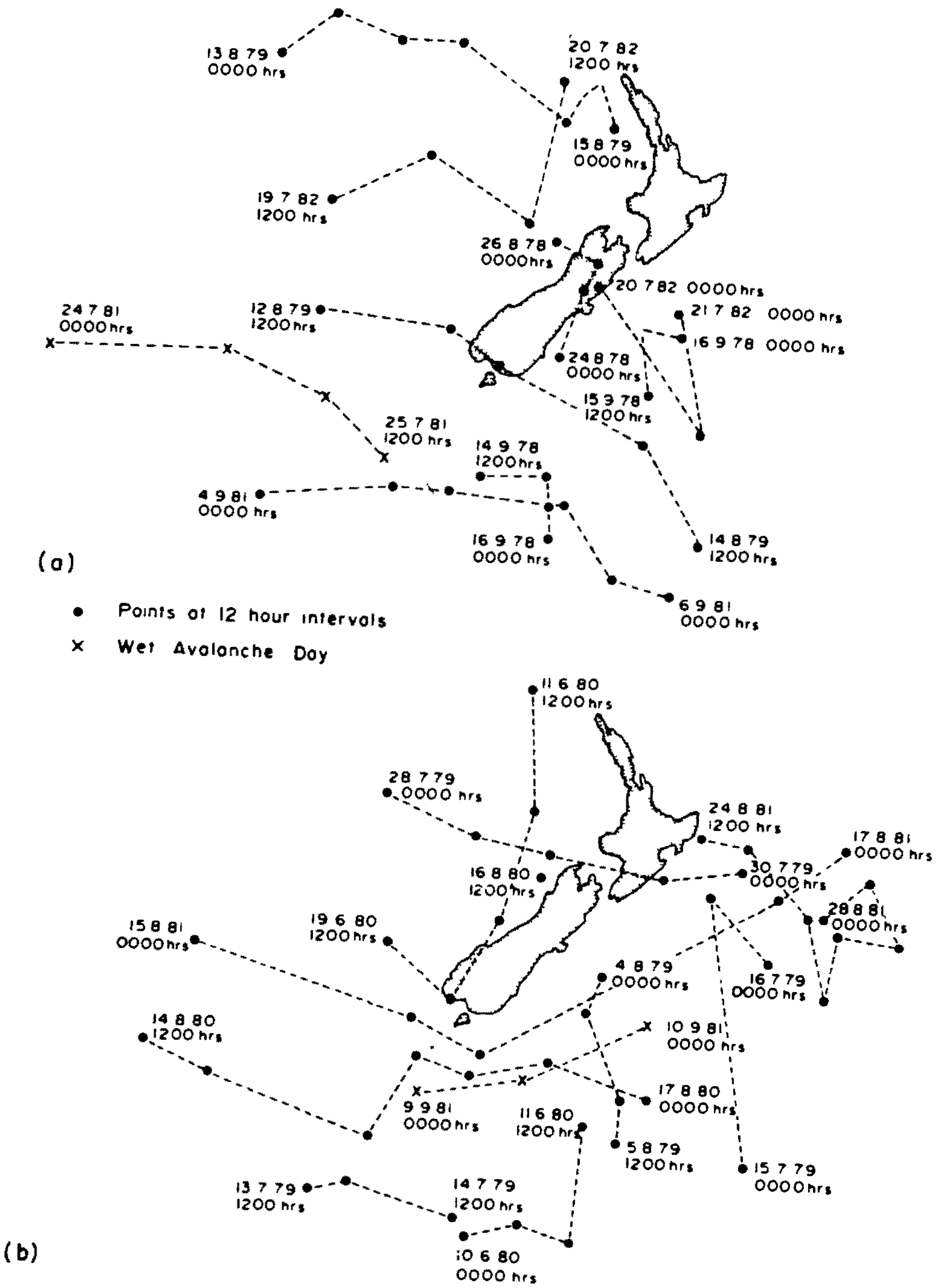


Figure 5.1a and b: Tracks of low pressure centres associated with avalanche storms. Note movement of dual low pressure centres for 12-15/8/79 (a) and 19-21/7/82 (b), also static nature of 24-25/8/81.

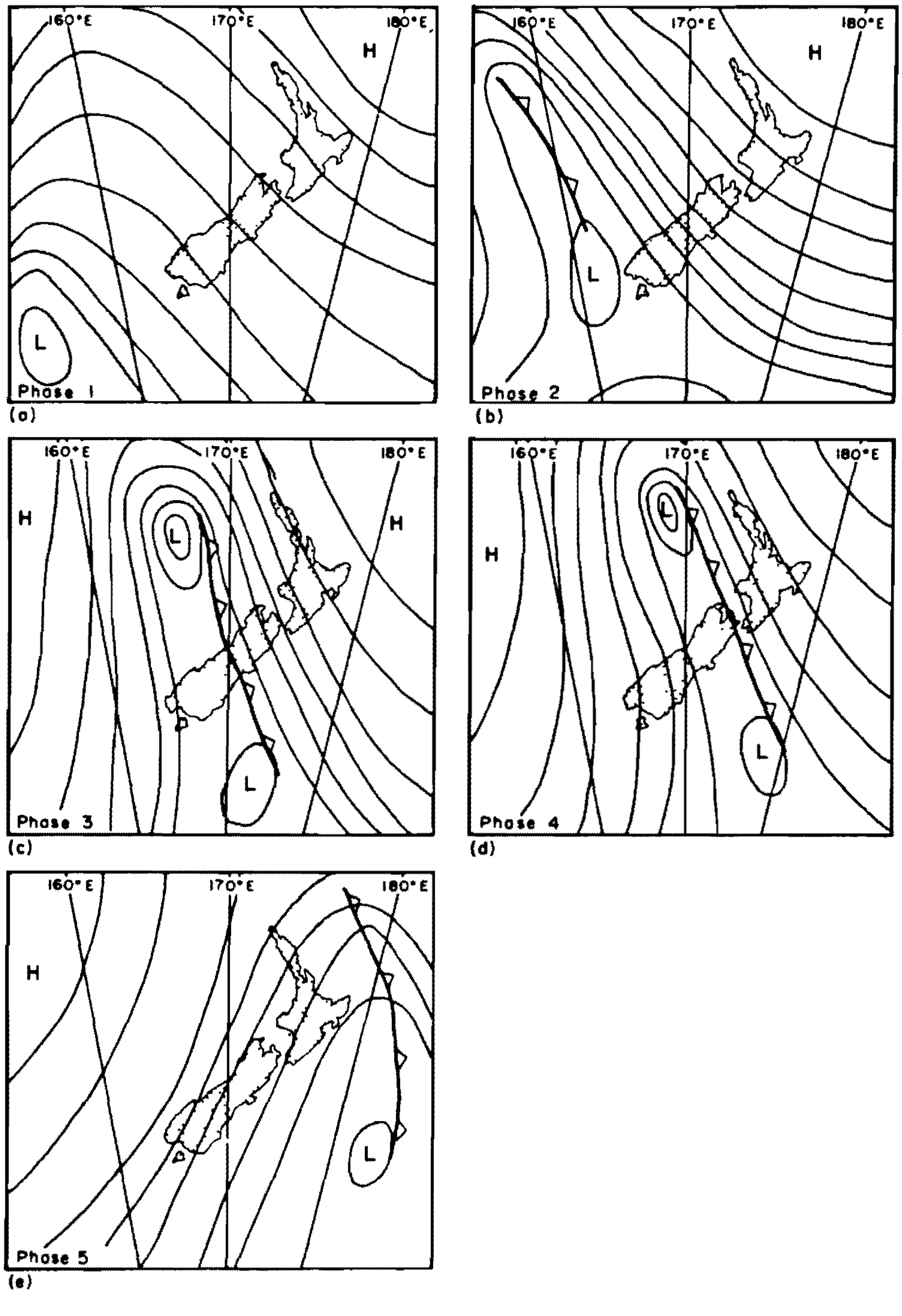


Figure 5.2: Five phase development of mixed warm advection vorticity advection storms. Note progressive movement of front and trough with associated north-westerly flow. Maximum avalanche potential realised in phases 3 and 4.

Phase I - A low pressure centre south-west of the country rapidly approaches with a general westerly flow over the South Island. North to north-east of New Zealand lies a ridge of high pressure extending from the Tasman Sea. This may remain as a ridge, or a distinct cell of high pressure may form north-east of New Zealand related to general subsidence of air in the area of 30-35^{o}S.

Phase II - The low pressure centre has moved onto the southern tip of the country with north-westerly wind velocities increasing. Associated with the depression is a distinct front separating warm north-westerly air from cooler southerly air. At this stage precipitation may commence in the Craigieburn mountains and general inland Canterbury High Country. The area of high pressure remains north to north-east of the country and is slow moving. Rates of thermal advection are increasing over Christchurch while at Invercargill they remain high.

Phase III - Two definite centres of low pressure have formed at this stage lying north-west and south-east of the country. A front lying over the central parts of the South Island joins the two low pressure centres. Strong north-westerly conditions prevail at the leading edge of the front while cool southerly air is replacing warm air at the rear. At this stage the front may be slow moving due to surface friction or the slow eastward movement of the area of high pressure north-east of the country. As a consequence, the front may become stationary or warm as the intense north-westerly flow holds back the colder southerly air. Rates of thermal advection and vertical motion reach a maximum in and around this phase as warm air is rapidly advected over cool air at the surface. Vertical ascent is also aided by the areas of convergence north-west and south-east of the country. Precipitation reaches a maximum in the Craigieburns at this stage.

Phase IV - The trough, low pressure frontal system shifts north. Precipitation in the Craigieburns still continues but at a reduced intensity from the south to south-west.

Phase V - Rapid replacement of warm air by cold air at the surface shifts the centre of low pressure off the east coast of the South Island. If a front still exists at this stage it will lie over the Pacific Ocean east of New Zealand. Precipitation gradually ceases with rates of thermal advection becoming negligible. A general south to south-westerly flow prevails over the South Island.

Phase III of this development is probably the most important phase as regards avalanche formation. Not only is it the stage of peak precipitation and therefore likely time of maximum precipitation intensities (correlations between TA and PI revealed a correlation of $r = 0.69$, significant at the 0.05 level),but strong north-westerly winds prevail. Redistribution of snow in this phase will be maximal. Immediately following Phase III and frontal progression the southerly wind change will result in lower temperatures. Such a change may hinder any sintering of newly deposited snow. Depending on the strength of the southerly wind change, redistribution of snow after frontal passage may also occur.

Fox (1973) has described a progression of a front of avalanche release along the Washington Cascades related to frontal precipitation. Though the Craigieburn Range does not cover a large enough longitudinal range to show a distinct relation of frontal passage to avalanche release, the evidence presented here demonstrates that avalanche release in the Craigieburn Range for those storms with well developed fronts is definitely related to frontal progression. Fitzharris (1976) has described an avalanche event in the Mt Cook area which is closely related to frontal passage.

During the study period storm systems possessing well developed fronts were noted to produce a progressive wave of avalanching from south to north throughout New Zealand. The timing of avalanche release for areas in which avalanche occurrence information is available can be correlated with frontal passage. Storms 4 and 10 are presented in Appendix 7 as examples of this.

5.4.2 Vorticity Advection Storms

The remaining storms producing dry avalanching in the Craigieburn Range have been classified as vorticity advection storms and are characterised by the development of low pressure systems. Though fronts may be present in these examples they are not associated with strong north-westerly flows in a meridional trough. The development of storms 2, 8 and 9 plus one other from outside the study period are described in Appendix 7.

Centres of low pressure appear to develop in close proximity to the east coast of the South Island in flows of southerly air on the downstream side of high pressure centres and eventually track eastward (Figure 5.1a and b). Development may also be in a weak trough situation. In strong contrast to the WA/VA storms, precipitation in the VA storms is received from the easterly through to south to south-westerly direction (Table 5.2) as the leading edge of the cyclonic flow moves onto or passes over the area. Easterly to southerly air is replaced at the leading edge by cooler south to south-westerly air, sometimes possessing greater velocities. Precipitation ceases at this stage. The movement of high velocity cool south to south-westerly air into the area immediately following precipitation results in high rates of snow redistribution.

The development of these low pressure systems and the relation to potential avalanche release can be considered in three distinct phases (Figure 5.3a-c).

Phase I - A low pressure centre develops in a southerly air stream either to the east or west of New Zealand. Precipitation begins from a northerly to easterly quarter.

Phase II - The centre of low pressure is positioned near or over the central New Zealand-South Island area. Precipitation intensities and amounts may reach a maximum in the Craigieburn Range as the leading edge of the cyclonic flow passes over the area. Moist east to south-east air is drawn into the area with the majority of the precipitation coming from this direction. Complete saturation of the atmosphere up to the 700 mb level may result. At this stage the centre may be slow moving and hence prolong the period of precipitation.

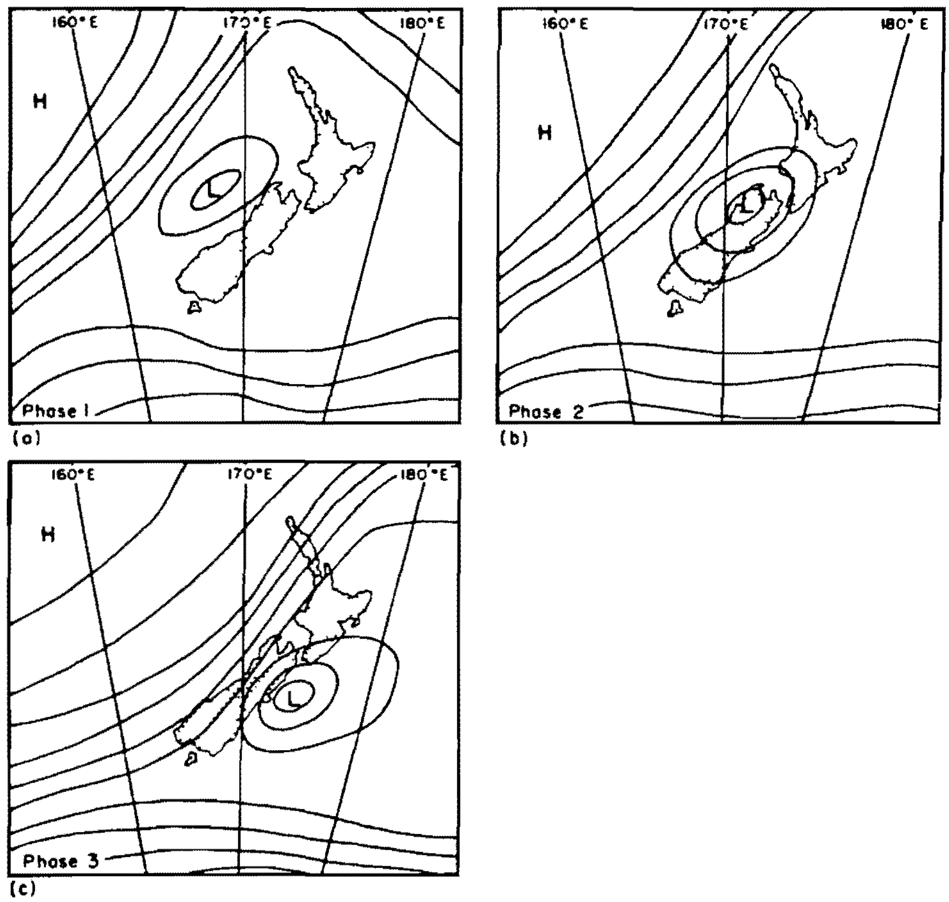


Figure 5.3: Three phase development of vorticity advection storm. Maximum avalanche potential realised in phase 2 as low pressure system moves over the South Island.

Phase III - Cool south to south-westerly air replaces air from the easterly to southerly quarter as the centre of low pressure moves off the area. The passage of the leading edge of cyclonic flow off the east coast brings precipitation in the Craigieburn Range to a halt.

Because maximum precipitation intensities occur in Phase II, this is the most likely time for avalanche release. The possibility for wind slab generation in Phase III also exists as cold wind blown snow may be packed into localised areas.

5.5 ATMOSPHERIC PROCESSES PRODUCING LARGE SNOWFALLS

Three main factors have been noted by Neale and Thompson (1977) as important for snowfall production. These are: (1) ascending motion

(2) thermal structure of the atmosphere

(3) atmospheric moisture.

The importance of these in producing snow and their relation to periods of maximum precipitation will be discussed in the following subsections.

5.5.1 Ascending Motion

Strong ascending motion is very important for generation of heavy snowfall as it provides a mechanism by which water vapour is drawn in at lower levels and accelerated vertically thus allowing condensation. Ascending motion has been attributed to the process of warm advection (WA) or vorticity advection (VA) by Goree and Younkin (1966). In the absence of significant WA most ascending motion, disregarding any orographic effects, may be attributable to the ascent of unstable cold air by VA. Rates of thermal advection (TA) and vertical velocities (ω) were calculated for all dry and wet avalanche storms and appear in Table 5.4

Several days on which snow fell in the Craigieburn Range in the 1979 to 1981 study period, TA exceeded $1^{o}C\ hr^{-1}$ and in all cases ω exceeded 50 mm s^{-1}. Storm 5 possessed the highest TA of $3.86^{o}C\ hr^{-1}$ with an associated ω of 210 mm s^{-1}. Such a high TA as this has been noted by Thompson and Neale (1977) as typical of WA storms. Also ω for this storm exceeds that noted by Neale and Thompson (1977) but approaches that noted by Sanders (1955) for intense surface fronts (Table 5.5). High TA and ω values for this study are typical of those storms (4, 5 and 6) dominated by strong north-westerly flows in a meridional trough with an associated low pressure cell. These storms also produced the highest precipitation intensities and amounts for this study. In such cases, a low pressure centre was located both north-west and south-east of the country, with ascending motion attributable to warm north-westerly air flowing over cooler southerly air at the frontal edge. Ascending motion may have been enhanced further, by cool unstable southerly air flowing into the areas of low pressure. Strong updrafts and high rates of ascending motion, developed in these situations, have been noted to increase the rate of crystal riming when cloud liquid contents are high (Jiusto and Weickmann, 1973). Heavily rimed needles have been observed in the Craigieburn Range in the early stages of WA/VA and VA storms when either or both TA or ω values are high. Occurrence of these crystal forms at the

Table 5.4: Rates of Thermal Advection and Vertical Velocities Above Christchurch

Date	Thermal Advection (C^o hr^{-1})	Vertical Velocity (mm s^{-1})
14/7/79	0.50	27
15/7/79	0.65	81
28/7/79	1.28	87
29/7/79	0.59	39
01/8/79	1.47	93
02/8/79*	0.32	25
04/8/79	0.19	16
05/8/79	0.31	28
13/8/79	1.04	106
14/8/79	1.07	70
04/9/79	0.0	7
05/9/79*	0.16	10
10/6/80	1.10	91
11/6/80	3.86	210
29/6/80	0.10	23
30/6/80*	1.40	128
11/8/80	0.46	44
12/8/80*	1.08	62
15/8/80	1.47	75
16/8/80	0.46	35
24/7/81	0.02	9
25/7/81*	0.55	55
15/8/81	0.44	39
16/8/81	0.56	51
24/8/81	0.57	48
25/8/81	0.83	48
26/8/81	1.19	91
27/8/81	0.53	11
04/9/81	0.05	10
05/9/81	0.03	0
09/9/81	0.67	54
10/9/81*	1.15	129
12/9/81	0.61	79
13/9/81*	0.51	28

*Wet avalanche days

Table 5.5: Rates of Ascending Motion Associated with Precipitation Production

Author	Type of Development	Vertical Velocity ($mm\ s^{-1}$)
Sanders (1955)	intense surface front	250
Petterssen (1956)	major cyclogenesis	100-200
Lyall (1972)	polar low	200
Barry and Perry (1973)	normal depressions	50
Jiusto and Weickmann (1973)	localised convectional storms	1000
Trenberth and Neale (1976)	major cyclogenesis	100
Neale and Thompson (1977)	surface fronts	70
Auer and White (1982)	winter cyclone	65

bottom of newly deposited slabs attests to their early storm origins. The role these crystal types play in avalanche release will be discussed in Chapter 6. The zone of convergence and hence accelerated updraft producing these rimed forms is replaced as the front and associated trough of low pressure move away from the Craigieburn Range.

Storms classified as VA gain the majority of their ascending motion through cyclonic vorticity. Vertical velocities, during periods of precipitation, for all VA storms generally exceeded 50 mm s^{-1} (Table 5.4). These velocities range between those considered typical for normal depressions (Barry and Perry, 1973) and approach those found for periods of major cyclogenesis (Petterssen, 1956; Trenberth and Neale, 1976) (Table 5.5). Incorporation of cold southerly air in the lower troposphere within a well developed cyclonic circulation typifies the situation producing these rates of ascent in the VA storms for this study. This resembles the processes characteristic of the Polar Low described by Lyall (1972). The two VA storms producing the largest amounts of precipitation (Storm 1 and 2) also possessed the greatest ω for this class of storm.

Though both the processes of WA and VA are important in producing the necessary ascending motion for snowfall in the Craigieburn Range, a combination of these two processes appears to be most conducive to production of high precipitation amounts and intensities as exemplified by the WA/VA storms 4, 5 and 6.

5.5.2 Thermal Structure

Discussions of the role of atmospheric thermal structure in snowfall production have largely involved considerations of stability/instability (Lumb, 1961, 1963; Power et al., 1964, Ferguson, 1970; Juisto and Weickmann, 1973). For this study, all VA storms were found to be conditionally unstable with lapse rates greater than 6.5°C km^{-1} but less than the dry adiabatic rate of 10° C km^{-1}. This was established by tephigram analysis. The WA/VA storms all showed overall stability between the 800 and 500 mb levels (Table 5.6). Analysis of individual layers however, revealed distinct layers of instability occurring below significant inversions (Table 5.7) for storms 4, 5 and 9.

The vertical sounding for storm 4 demonstrates this effect well (Figure 5.4). Instability occurred in the layers 810 to 710 mb with an inversion present at 680 to 640 mb. The unstable layer is marked by a flow of cool southwesterly air, the cooling effect of which would aid both downward penetration of snow and vertical motion of air moving in at a lower level. Both these processes would enhance any condensation in the lower parts of the advected north-west air. The soundings of storms 5 and 9 show similar patterns of inversions developed in a north-westerly flow capping unstable layers characterised by south-westerly flow. The magnitude of these inversions may, however, be exaggerated as a result of lee effects, as westerly air subsides and warms adiabatically after passing over the Southern Alps as discussed in Section 2.2.6. Storm 6 in contrast, displays complete stability with a marked inversion, features noted by Goree and Younkin (1966) and Neale

Table 5.6: 800-500 mb Lapse Rates Above Christchurch

Date	Lapse Rate $^{o}C\ Km^{-1}$
15/7/79	7.31
29/7/79	6.90
02/8/79*	5.69
05/8/79	7.34
14/8/79	5.87
05/9/79*	5.89
11/6/80	5.29
30/6/80*	6.51
12/8/80*	5.65
16/8/80	5.90
25/7/81*	6.53
16/8/81	7.31
25/8/81	8.40
27/8/81	6.80
05/9/81	8.57
10/9/81*	9.82
13/9/81*	6.58

*Wet avalanche day

Table 5.7: Layers Showing Instability and Inversions Above Christchurch on Day and Day Before Avalanche Occurrence

Date	Unstable (mb)	Inversion (mb)
13/8/79		690-665
14/8/79	810-710	680.640
04/9/79		
05/9/79	990-890	
10/6/80		720-695
11/6/80	996-820, 660-640	640-590
11/8/80		
12/8/80		770-740
15/8/80		780-710
16/8/80		690-660
15/8/81		
16/8/81		740-620
24/8/81		
25/8/81		930-740
26/8/81		
27/8/81	860-840	765-710
09/9/81		
10/9/81		880-580

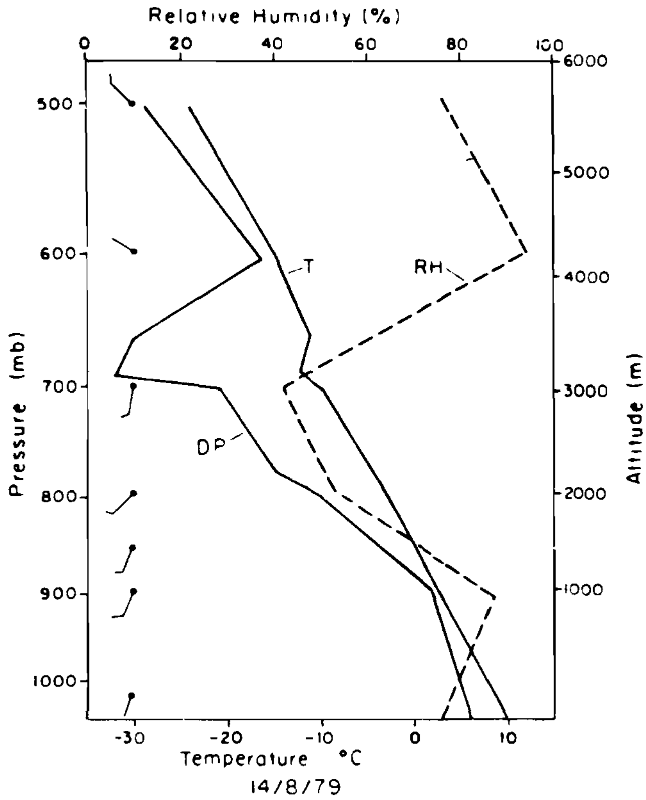


Figure 5.4: Radiosonde sounding 14 August, 1979 above Christchurch at 1200 hours. Note inversion at 680-640 mb with north-westerly and south-westerly flow above and below the inversion respectively.

and Thompson (1977) as typical of warm fronts and warm advection storms. The magnitude of the inversion reached a maximum on 15 August coincident with high TA and windspeeds from the north-west (Figure 5.5a) but had decreased markedly by 1200 hours on 16 August as windspeeds dropped to 7 ms^{-1} and the wind backed to a more westerly direction (Figure 5.5b). The majority of the precipitation received in the Craigieburn Range from this storm occurred within this time interval.

Conditions of instability occurring below inversions (storms 4, 5, 9; Table 5.7) have been noted by Power et al. (1964) to be highly conducive to snow crystal growth, as these areas are often characterised by supersaturation. Thirty-four percent of the storms producing snow studied by Power et al. (1964) possessed a similar structure. Further, Power et al. (1964) have noted that areas of supersaturation where stable layers lie below inversions, (storms 6, 7 and 8), also feature predominantly in snowfall production. The evidence presented for this study suggests that a combination of conditions of stability and instability is effective for producing significant amounts of snow in the Craigieburn Range. This contrasts with the conditions of overall stability suggested by Ferguson (1970) and Jiusto and Weickmann (1973), and instability by Lumb (1961, 1963) for snowfall production.

5.5.3 Atmospheric Moisture Content

Though large amounts of precipitable water may be available in a storm, this does not necessarily mean that large amounts of precipitation will be received at the

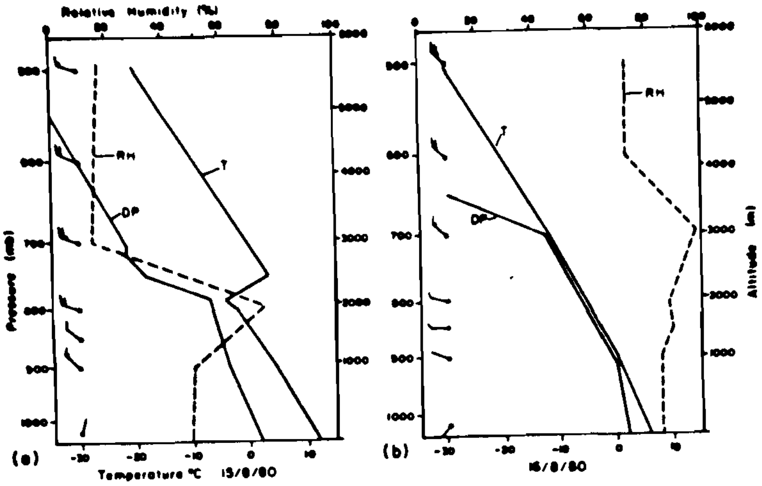


Figure 5.5: Radiosonde sounding 15 and 16 August 1980 Christchurch.

(a) Note large inversion and strong north-westerly winds.

(b) Note near saturation in layers dominated by westerly winds.

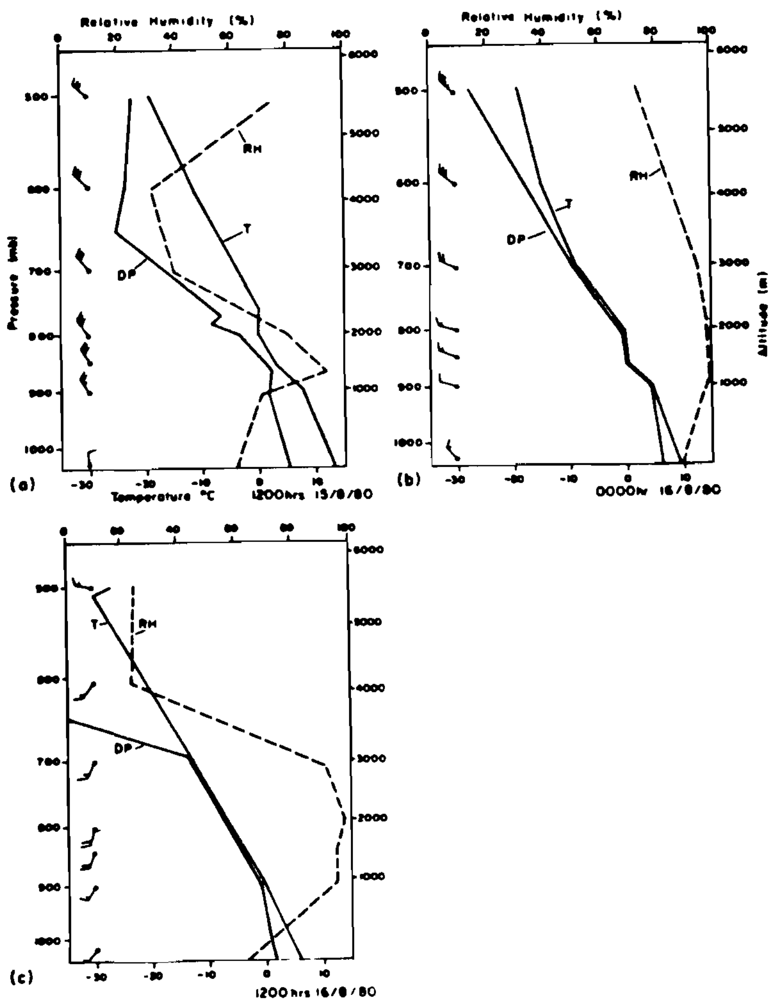


Figure 5.6: Radiosonde sounding 15 and 16 August 1980 Invercargill.

(a) Note inversion at 790-760 mb level coincident with that in Figure 5.5a.

(b) Inversion still present but at lower level. Saturation increasing as winds back to west.

(c) Note complete saturation in layers dominated by south-westerly wind.

earth's surface. This is highlighted in this study. While storms 1 and 2 possessed the greatest amounts of precipitable water, it was storms 4, 5 and 6 with marked inversions and quasi-stationary fronts that produced most precipitation. As discussed by Sevele (1969) and Neale and Thompson (1977) and noted by Fox (1973) and Fitzharris and Schaerer (1980), the relatively static nature of this type of system can lead to large amounts of precipitation being received at the surface. The thermal structure above Christchurch for storm 6 and its implications to snowfall production has been discussed in the previous section. A similar pattern of atmospheric moisture content for this storm is also shown by the Invercargill sounding (Figure 5.6 a-c). A small inversion was present on 15 August, 1980 at 1200 hours at the 790-760 mb level (Figure 5.6a) and corresponds with the marked inversion above Christchurch at the same level (Figure 5.5a). Moisture content below this layer was approaching saturation. By 0000 hours on 16 August, 1980 (Figure 5.6b) a very moist layer of air existed up to the 650 mb level with TA exceeding $3^{o}C\ hr^{-1}$. Complete saturation existed up to the 715 mb level by 1200 hours on 16 August, 1980 (Figure 5.6c), similar to that of Christchurch.

On occasions when TA was high above Invercargill and Christchurch with large amounts of precipitation produced in the Craigieburn Range moist north-west air was involved. In contrast, situations in which TA remained low (storms 1 and 2) and atmospheric moisture contents were high with near saturation up to the 700 mb level (Figure 5.7a and b) moist

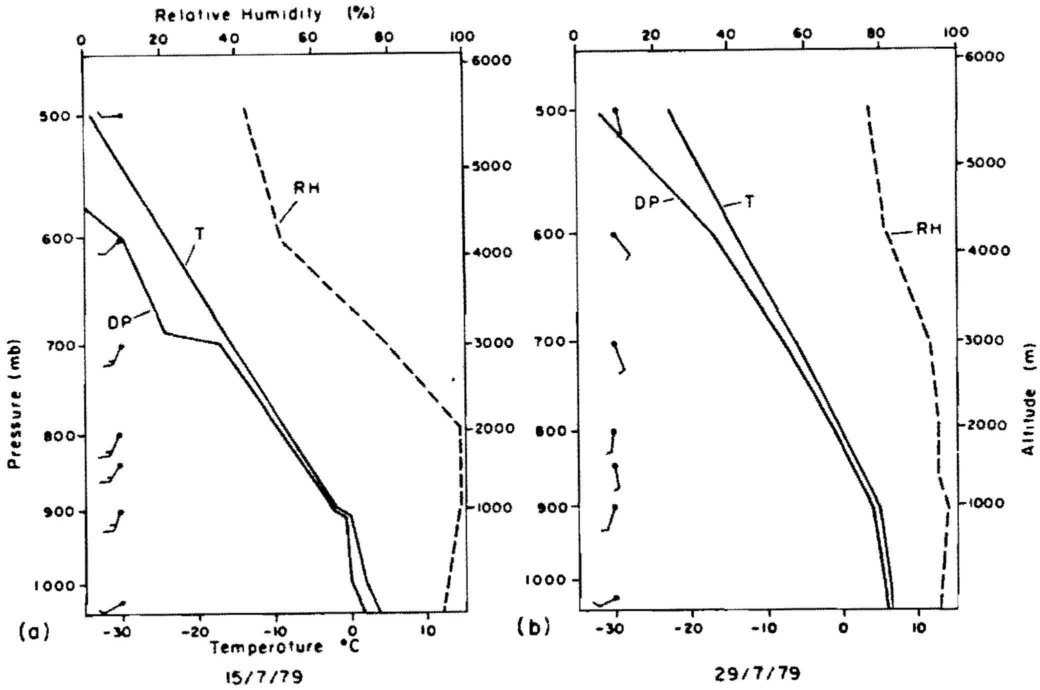


Figure 5.7:

(a) Radiosonde sounding 15 July, 1979 above Christchurch, 1200 hours.

(b) Radiosonde sounding 29 July, 1979 above Christchurch, 1200 hours.

Note almost complete saturation in south-westerly flow (a) and south-easterly flow (b) up to the 700 mb level.

south to south-westerly air was involved. This was typical of VA storms.

5.5.4 Wet Avalanche Storms

Though this chapter is mainly concerned with snowfall production and dry avalanche storms a brief mention of wet avalanche storms will be made here. The tracks of two storms which typify those resulting in wet avalanche release are shown in Figure 5.1a and b. Of the wet avalanche events six involved westerly to northerly flows with fronts developed in these flows. These systems brought brief periods of rain to the Craigieburn Range. The remaining wet avalanche cycle occurred in south-westerly conditions with a centre of high pressure lying east of New Zealand.

Rates of ascending motion varied markedly between wet avalanche days with ω varying from 7 mm s^{-1} with TA of 1.12^{o}C hr^{-1} to 129 mm s^{-1} with associated TA of 1.15^{o}C hr^{-1}. Though ω may be high, with the rapid cooling of air expected, upper air temperatures in conjunction with surface temperatures remained too high for production of snow. Various conditions of stability and additional instability were displayed by the wet avalanche storms. All wet avalanche storms showed a mean increase in the 1000-500 mb thickness of 55 gpm over the 24 hour period preceding wet avalanche release, indicative of the movement of warm air associated with westerly to northerly flows which were typically moist.

5.6 PREDICTION OF HEAVY SNOWFALL AMOUNTS AND ASSESSMENT OF AVALANCHE POTENTIAL

The purpose of this section is twofold. Firstly, some additional easily measurable synoptic scale factors will be evaluated with respect to precipitation production. Secondly, correlation and stepwise multiple regression will be utilised to assess the significance of variables for the prediction of heavy snowfall amounts and dry avalanche potential.

5.6.1 Circulation and Atmospheric Thickness Effects on Precipitation Amounts

Schallert (1962) has demonstrated relationships between an intensity index of cyclonic flows and precipitation for Colorado storms. The index derived by Schallert (1962) was used in this study to analyse circulation intensity-precipitation relationships. The index is calculated by

$$CI = \frac{P_1 + P_2 + P_3 + P_4 - 4P_0}{H^2} \qquad (5.1)$$

where P_1, P_2, P_3, P_4 = The pressure value at four vertices at 3° latitude from the centre of a diamond grid orientated north-south.

P_0 = The pressure at the central point (Craigieburn Range)

H = Latitude of the grid (3°)

Circulation indices for the 1000 mb (CI_{1000}) and 500 mb (CI_{500}) surfaces were calculated for the avalanche producing storms and plotted over time with precipitation. At times of precipitation CI_{1000} and CI_{500} appear to be out of phase. This is especially so in Figure 5.8a and d.

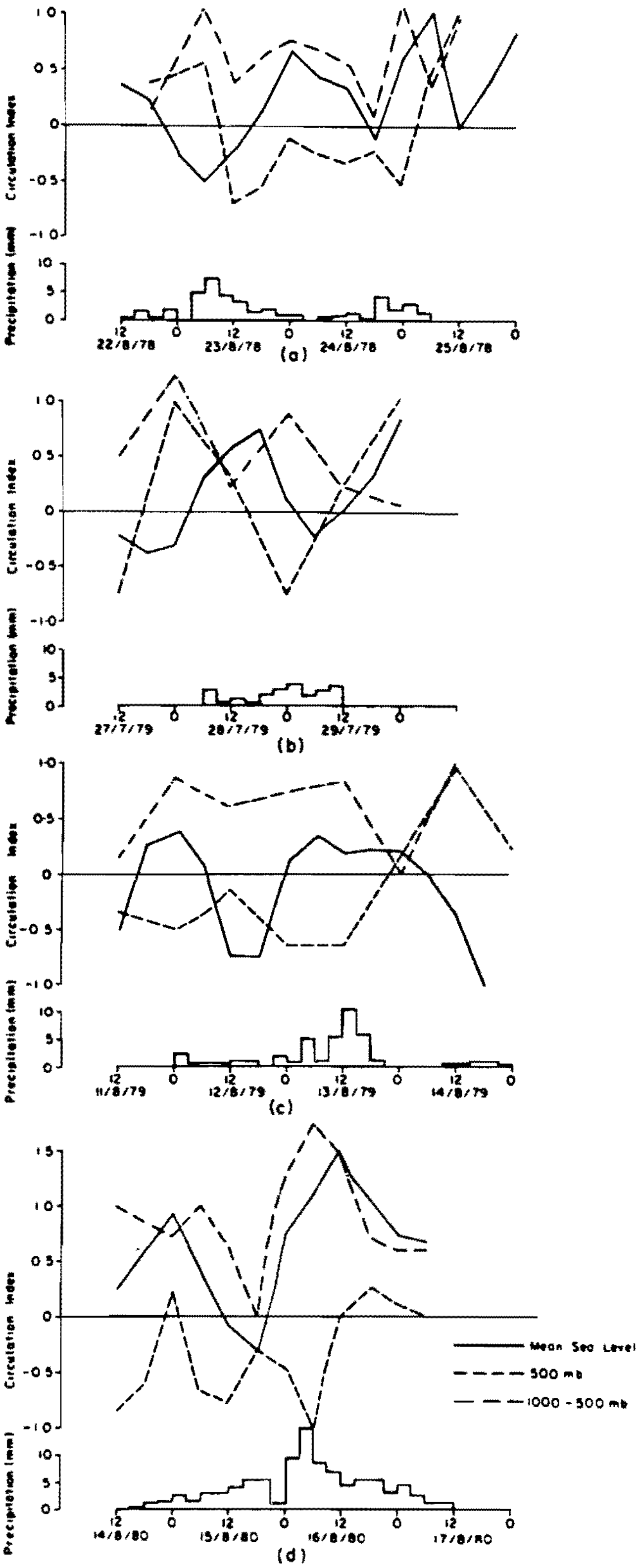


Figure 5.8: Circulation indices at various levels. Note precipitation coincident with those periods when sea level and 500 mb indices are out of phase.

Correlations between the difference in CI_{1000} and CI_{500} and precipitation in the next six hours were subsequently performed for the dry avalanche storms. Eight of the 13 storms show correlations significant at the 0.05 level or greater (Table 5.8). These correlations imply that precipitation is related to those periods when lower level and upper level low pressure or pressure gradient patterns are spatially out of phase. Klein et al. (1968) noted that with increasing intensity of a cyclonic system the frequency of precipitation increases markedly. They further noted that centres of the most intense low pressure centres are closest to the maximum precipitation frequency and become displaced towards the pole at higher levels. Displacement of higher level lows away from low pressure centres results from the closed circulation towards cooler air. Younkin (1968) likewise noted that areas of maximum precipitation are often up to 10^{o} latitude downstream from upper level trough or low pressure centres. As noted in Sections 5.4.1 and 5.4.2 precipitation often begins with the approach of the leading edge of an upper level trough or low and as these areas passed over the Craigieburn Range and moved east of the South Island precipitation ceased. Correlations were therefore performed between distance of the surface low pressure centre and precipitation in the next six hours. Seven storms show significant correlations at the 0.05 level or above (Table 5.9). Five of these storms possessed fronts developed either in a low pressure trough or low pressure system. The remaining two storms were dominated by cyclonic flow. These correlations indicate the importance of the approach of a centre of low pressure in precipitation

Table 5.8: Correlation Between Difference in 1000 - 500 mb Circulation Indices and Precipitation in the Next Six Hours

Storm Date	Correlation Coefficient (r)	Length of Period (hrs)
22/8/78-25/8/78	0.44*	66
15/9/78-16/9/78	0.94**	42
14/7/79-15/7/79	0.34	36
28/7/79-29/7/79	0.78*	42
04/8/79-05/8/79	0.36	36
12/8/79-14/8/79	0.71**	72
10/6/80-11/6/80	0.80*	36
14/8/80-16/8/80	0.55*	60
15/8/81-16/8/81	0.27	48
24/8/81-25/8/81	0.78*	42
26/8/81-27/8/81	0.10	30
04/9/81-05/9/81	0.60*	66
18/7/82-20/7/82	0.56*	66

* significant at 0.05 level
** significant at 0.01 level

Table 5.9: Correlation Beteween Distance of Low Pressure Centre and Precipitation in the Next Six Hours

Storm Date	Correlation Coefficient (r)	Length of Period (hr)
22/8/78-25/8/78	-0.30	66
15/9/78-16/9/78	-0.63**	42
14/7/79-15/7/79	-0.32	42
28/7/79-29/7/79	-0.60*	42
04/8/79-05/8/79	—	36
12/8/79-14/8/79	-0.50*	72
10/6/80-11/6/80	-0.90**	36
14/8/80-16/8/80	-0.70*	60
15/8/81-16/8/81	-0.18	48
24/8/81-25/8/81	-0.24	48
26/8/81-27/8/81	-0.24	30
04/9/81-05/9/81	-0.57*	66
18/7/82-20/7/82	-0.54*	60

* significant at 0.05 level
** significant at 0.01 level

production whether it has an associated front or not (Figure 5.9, 5.10). Maximum precipitation may therefore be expected as a front or leading edge of a cyclonic flow approaches the Craigieburn Range. For the former case precipitation will be frontal while in the latter precipitation will be largely related to vorticity advection.

Murray (1952), Boyden (1964) and Lowndes et al.(1974) have all stressed the significance of the critical thickness of various atmospheric layers for the production and occurrence of snow to low levels. Boyden (1964) has presented a variety of critical thicknesses of atmospheric layers as related to the probability of snowfall occurrence. Neale and Thompson (1977) have noted that low elevation snowstorms in New Zealand have occurred with 1000-500 mb thicknesses of 5150 to 5400 gpm. Average 1000-500 mb thicknesses during periods of snowfall for this study ranged from 5303 to 5453 gpm. Using the probabilities of Boyden (1964) these values correspond to snow probabilities of 30 and 10 percent respectively, similar to that found by Neale and Thompson (1977). The WA/VA storms of this study possessed greater thicknesses than the VA storms also verifying the results of Neale and Thompson (1977). A difference of means tests was conducted for mean 1000-500 mb and 800-500 mb thickness for dry and non-avalanching situations. The non-avalanche days were the same as the non-avalanche set used in Section 4.5. No difference in either thickness were found indicating that as a separate measure for predicting snowfall they do not appear very useful. However, if the trend in 1000-500 mb thickness is considered over the 48 hour period before avalanching, it is

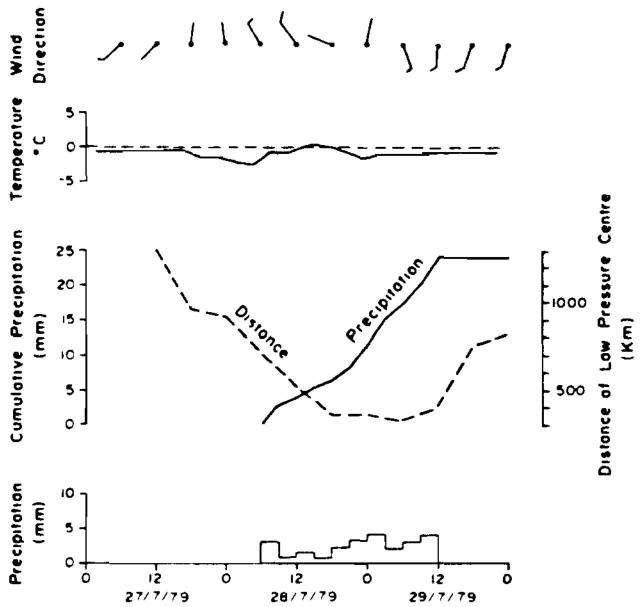


Figure 5.9: Storm plot 27 to 29 July, 1979.
Note majority of precipitation received as low pressure centre approaches area in this vorticity advection storm.

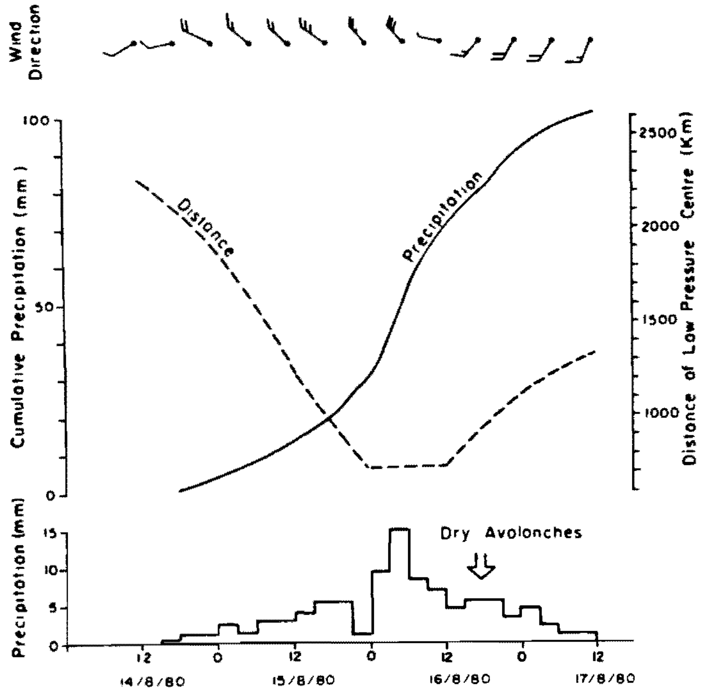


Figure 5.10: Storm plot 14 to 17 August, 1980.
Note rapid increase in cumulative precipitation with the approach of low pressure centre. Also majority of precipitation produced from a north-westerly direction in this warm/vorticity advection storm.

shown that for dry avalanche storms there is a mean decrease of 51.6 gpm in contrast to an increase of 33.0 gpm for non-avalanching situations. This presumably reflects the approach of a cold front, or centre of cyclonic vorticity and associated cooler south to south-westerly air, which has been shown above to play an important role in the timing of precipitation in the Craigieburn Range.

5.6.2 Correlation and Regression Analysis of Factors Affecting Water Equivalent and Avalanche Magnitude

Correlations between those variables introduced in the previous subsection and Section 5.5. (Table 5.10) with total storm precipitation water equivalent (WE) and MAG were performed (Table 5.11). Apart from WE the only variable showing significant correlation with MAG is CI. It has been shown above that periods of precipitation appear to be related to those times when the circulations at the 1000 mb and 500 mb levels are out of phase. Such a relationship indicates that the greater this phase difference the greater the likelihood of large precipitation amounts and potential avalanche activity. The significant correlation of MAG with WE necessitates consideration of those variables which may produce a large WE. From the correlation coefficients TA and ω both appear to be important substantiating their role in precipitation production discussed in Section 5.5.1. Of the independent variables, RES seems of secondary importance compared to CI, TA and ω. This however reinforces the conclusions reached earlier in this section that the slower moving the centre of low pressure, whether it is related to a front/trough or cyclonic system, the greater amounts of

Table 5.10: Variables Used In Stepwise Multiple Regression Analysis

Variable Symbol	Variable Description	Units
TA	Rate of thermal advection on day of maximum precipitation	$^oC\ hr^{-1}$
ω	Vertical velocity on day of maximum precipitation	$mm\ hr^{-1}$
PW	Precipitable water on day of maximum precipitation	mm
Z	800-500 mb thickness	gpm
RES	Residence time of low pressure centre within 750 km of BR	hrs
T	Mean 800-500 mb temperature	^oC
CI	Maximum Difference between 1000 and 500 mb circulation indices during period of precipitation	$mb(deg.lat)^{-3}$
LAP	800-500 mb temperature lapse rate on day of maximum precipitation	$^oC\ km^{-1}$
WE	Total storm preciptation water equivalent	mm
MAG	Sum of magnitude of all avalanches greater than magnitude 2 for individual storm	dimensionless
DIS	Minimum distance of low pressure centre from BR	km
P_{A-I}	Pressure difference between Auckland and Invercargill	mb

Table 5.11: Correlation Coefficients For Sum of Avalanche Magnitude and Total Storm Water Equivalent with Table 5.10 Variables.

	MAG	WE
MAG	1.00	
WE	0.61*	1.00
TA	0.25	0.72*
ω	0.28	0.60*
PW	0.35	0.11
Z	0.01	-0.42
RES	0.48	0.38
T	0.07	-0.35
CI	0.74*	0.54
DIS	-0.27	-0.50
LAP	-0.41	-0.33
P_{A-I}	0.23	-0.05

* significant at 0.05 level or greater

precipitation to be expected. This in turn will increase the likelihood of avalanche occurrence.

While the independent variables may be related to MAG and WE, a combination of these variables should improve the explanation of the variance of MAG and WE. Stepwise multiple regression analysis was used for this. Two regressions were produced (Table 5.12 and 5.13), both significant at the 0.025 level. CI and ω variables selected for the prediction of MAG have both been identified earlier as important for precipitation production. The inclusion of T is puzzling as it possesses no correlation with MAG at all but has a negative correlation with WE. Examination of standardised beta coefficients however shows that an increse in T will serve to increse MAG. In addition, partial correlation analysis (Blalock, 1972, p.437) shows that 27 percent of the variance of MAG is explained by T if CI is controlled for. This presumably reflects the influence of the warmer atmospheric temperatures associated with WA/VA storms which produced the large avalanche cycles.

CI appears again in the regression for the prediction of WE. In addition to this are TA, LAP and RES. High TA values have been shown to be important for producing heavy snowfall and consequently widespread avalanching, which typify those storms classified as WA/VA. The negative correlation between LAP and WE attests to the importance of lower lapse rates, again typical of WA/VA storms, in producing large precipitation amounts for avalanching. The inclusion of RES signifies the importance of the time spent by that part of the storm associated with precipitation production in close proximity to the Craigieburn Range for producing large storm precipitation amounts.

Table 5.12: Variables and Coefficients For Prediction of Sum of Avalanche Magnitude (MAG)

Variable	Coefficient	Standard Error	r^2	Change in r^2
CI	22.98	10.7	0.56	0.56
T	2.83	9.2	0.71	0.15
ω	0.18	7.8	0.82	0.11
Constant	24.52			

Table 5.13: Variables and Coefficients For Prediction of Total Storm Water Equivalent

Variable	Coefficient	Standard Error	r^2	Change in r^2
TA	16.52	13.0	0.53	0.53
CI	15.11	7.5	0.86	0.33
LAP	1.18	7.3	0.88	0.02
RES	-0.08	7.1	0.89	0.01
Constant	91.03			

The applicability of the variables identified in this analysis to an avalanche forecasting situation depends largely on their predictability. The variables identified here are all easily estimated from numerical weather prediction models such as that presented by Trenberth (1973). Variables such as ω, TA, T and LAP may be derived either directly or indirectly from estimates of atmospheric layer temperatures and windspeeds produced by such numerical models. Prognoses from the same numerical models would also be used to calculate the relevant circulation indices. If such variables were to be used in a discriminant model, as presented in Section 4.4, in conjunction with meteorological variables measured at the local scale, a data set comprising variables measured on non-avalanche and wet avalanche days would also be required. Inclusion of this type of data would require the discriminant model to be re-run with new discriminant functions being produced. Though beyond the scope of this analysis, the writer believes that a combination of variables measured at these two scales in such an avalanche forecasting model would greatly improve the statistical forecasting of avalanche occurrence.

5.7 SUMMARY

Two main synoptic patterns have been identified that are associated with snow avalanching in the Craigieburn Range. These are similar to those found by O'Loughlin (1969) and Prowse (1981) to be important for producing snowfall in the Craigieburn Range. Similar to Prowse (1981) but unlike O'Loughlin (1969), storms associated with a cyclonic flow system and having negligible frontal development, predominated. However, it was those systems possessing a well developed trough and front which produced the largest amounts of precipitation and the most widespread avalanche cycles in this study. These systems were found by O'Loughlin (1969) to be the most frequent for bringing snow to the Craigieburn Range. The general synoptic situations identified and related to dry avalanche occurrence in this study resemble most closely those systems described by Fox (1973), Grischenko and Tokmakov (1978) and Fitzharris and Schaerer (1980) for overseas studies.

For the production of snow three main processes or conditions have been considered. Ascending motion due to warm advection or vorticity advection plays an important role in lifting moist air to aid condensation as have been identified by Goree and Younkin (1966) and Neale and Thompson (1977). Rapid rates of ascent in combination with high rates of thermal advection typify these situations which have a well developed depression and associated front, the front being preceded by a flow of moist north-westerly air. In these cases vorticity advection at lower levels due to inflow of cool southerly air augments uplift of warm north-westerly air being advected over the South Island. Characteristic of these mixed warm-vorticity advection storms are inversions related to the advection of north-west air capping cool unstable south to south-westerly air. Similar thermal structures have been identified by Power et al. (1964) and Auer and White (1982) for production of heavy snowfalls.

Storms characterised purely by vorticity advection and hence possessing low rates of thermal advection are associated with low pressure centres sitting near or over the country and lacking any distinct trough-front development. In these, vertical velocities are sufficient given the conditional instability typical of these storms to produce large amounts of condensation for precipitation. The inflow and convergence of cool moist air at lower levels often produces near saturation if not complete saturation for considerable depths in the atmosphere.

Most precipitation is received in the Craigieburn Range when upper and lower level circulations are out of phase. This substantiates the findings of Goree and Younkin (1966) and Klein et al. (1968). Precipitation in the Craigieburn Range has also been shown to be related to the proximity of the centre of low pressure. The passage of this, whether associated with a trough/frontal system or enclosed cyclonic circulation, plays an important role in timing of precipitation and thus direct action avalanching.

Atmospheric layer thickness appears to be of little use as a predictor of snowfall probability in the Craigieburn Range. Daly (1978) has also noted the inadequate nature of atmospheric layer thicknesses for predicting snowfall. Layer thicknesses for this study correspond to snowfall probabilities of 10 to 30 percent compared to studies of Murray (1952), Boyden (1964) and Lowndes et al. (1974). Periods associated with upper and lower level circulations out of phase, high rates of thermal advection, 800-500 mb lapse rates showing stable conditions and slow moving low pressure trough/frontal systems, characteristics all typical of mixed warm-vorticity advection storms, have been shown in this study to be the most important determininants of heavy snowfall production and thus increased dry avalanche potential. Such variables may be included along with meteorological variables measured at the local scale in an avalanche forecasting model.

CHAPTER SIX

THE ROLE OF SNOWPACK STRUCTURE IN SNOW AVALANCHING

6.1 INTRODUCTION

Though the role of snowpack structure in snow avalanching has been recognised for some time (Seligman, 1936), avalanche forecasting has traditionally depended on meteorological data as a source of information for assessment of likely avalanche occurrence. This may be attributed to the difficulty of monitoring and assessing the significance of snowpack development processes. However, in inland and continental areas, characterised by extremely low temperatures and shallow snowpacks, and where climax avalanche release predominates, the role of snowpack structure in snow avalanching may be very important (LaChapelle, 1966). Studies conducted in such areas (Perla, 1971; Bradley et al., 1977; LaChapelle and Armstrong, 1976; Stethem and Perla, 1980) have highlighted the importance of temperature gradient processes in producing snowpack instability for avalanching. Fewer studies have been undertaken in maritime climates, (LaChapelle, 1966; Moore, 1975; Ward, 1980) which in contrast to inland and continental areas, are characterised by deep snowpacks and the predominance of direct action avalanching (LaChapelle, 1966). Moore (1975) and Ward (1980) have also shown that rain on snow events with slabs releasing over melt freeze crusts are typical of these environments. Further, Ward (1980) has noted the complete absence of temperature gradient forms in this type of snow environment.

As weather systems affecting New Zealand are of a maritime origin, and the majority of avalanches recorded for areas in New Zealand where a good avalanche occurrence record exists are direct action (Ho, 1982; this study), it may be expected that structural characteristics involved in snow avalanching in New Zealand may resemble those typical of a maritime climate, as described by LaChapelle (1966). The obvious implication of this is that snowpack structure may be only secondary in determining avalanche release. However, from the few studies conducted to date it is apparent that snow structure may play an important role in avalanche release in New Zealand. McNulty and Fitzharris (1980) have described a large climax avalanche event in the Craigieburn Range resulting from collapse of temperature gradient crystal layers developed early in the season. Ho (1982) has also presented evidence of snowpack failure due to temperature gradient crystal failure. Other workers (Weir, 1979; Weir and Owens, 1981; Prowse, 1981; Owens et al., 1983; Prowse and Owens, 1984) have noted the presence of temperature gradient forms in the snowpacks of inland Canterbury and have commented that these are more typical of an inland-continental climate. Conversly, Ho (1982), Moore and Marcus (1983) and McGregor et al. (in prep) have reported the widespread occurrence of melt freeze layers, a feature typical of maritime snow climates, and their role in avalanche release for several areas in New Zealand. Clearly a mixture of snowpack characteristics representative of both maritime and inland-continental snow climates exist for the snowpacks of New Zealand. This is certainly true of the Craigieburn Range snowpack which has been classified by

Prowse and Owens (1984) as one transitional between a coastal and inland situation with the implication that a mixture of meteorological and snowpack information should be used for avalanche forecasting in such an environment.

Though previous work in New Zealand has revealed a variety of snowpack structural features as important in snow avalanching, it has dealt mainly with a limited number of case studies. Also, apart from the work of Conway and Abrahamson (in press) these studies have given only a peripheral consideration to snow structure's role in snow strength and avalanche occurrence. The basic aim of this chapter then is to assess the role of snowpack structure in snow avalanching for the Craigieburn Range through:

(1) a description of the development of the Craigieburn snowpack;

(2) examination of processes affecting the Craigieburn snowpack;

(3) identification of specific snowpack structural characteristics diagnostic of weak layers.

6.2 SEASONAL DEVELOPMENT OF SNOWPACK STRUCTURE

6.2.1 Nature of Data and Terminology

During the three winter field seasons of 1979 to 1981 standard snowpit observations were conducted at weekly intervals. The majority of snowpits dug were at an elevation of 1800 m on Mt Cockayne (Figure 2.1). Additional pits were also dug at Porter Heights and Craigieburn Valley skifields to make up a total of 73 pits on which this analysis is based. Though analysis of the seasonal development of the snowpack is based largely on information collected at one site, snowpits dug elsewhere in the Craigieburn Range coincident with these observations show that this site is representative of general conditions and trends. Snowpits were analysed as outlined in Section 2.3.6. Symbols used in displaying snowpack information appear in Appendix 2b and 2c. A symbol for early equitemperature gradient snow has been included with internationally understood and accepted symbols recommended for use in displaying snowpack data (UNESCO/IASH/WMO 1970). Frequent observations of snow resembling equitemperature snow but not possessing the partially metamorphosed form of new snow or the fully rounded form of advanced equitemperature snow were made during the field season. The inclusion of this symbol therefore appears justified.

Snow metaphorphism terminology used in this analysis follows that proposed by Sommerfeld and LaChapelle (1970). Reviews of the actual processes will not be presented here as they appear in detail elsewhere (LaChapelle, 1969; Akitaya, 1975; Voitkovsky et al., 1975; Perla and Martinelli, 1976; Male, 1980; Prowse, 1981; Colbeck, 1982a). Equitemperature (ET) metamorphism involves the change of crystal structure within a snowpack with a uniform temperature or temperature gradients less than those generally required to induce temperature gradient metamorphism. Colbeck (1982a) has alternatively referred to this process as progression towards an equilibrium form. The eventual product of ET metamorphism is an overall reduction in grain size with formation of bonds between grains at points of contact.

Temperature gradient (TG) metamorphism involves the change of crystal form under an active temperature gradient. The main effects of TG metamorphism is initial faceting of the crystal followed by development of stepped surfaces and hollow crystals. Colbeck (1982b) has subsequently termed these kinetic growth forms. Melt freeze (MF) metamorphism is the process by which grains are formed by refreezing of free water in the snowpack. In contrast to ET and TG grains which form bonds by necks, MF grains are welded together with little intergranular space and appear in the snowpack as distinct layers of aggregated grains.

6.2.2 Results and Discussion

Evaluation of the trend of various snowpack parameters for the three winter field seasons over time (Figure 6.1a-e) show that July has the highest observed mean temperature gradients ($0.013^{o}C\ mm^{-1}$) and mean percentage of total snowpack depth with TG snow (38 percent). This is much lower than the 57 percent found at a comparable stage of the winter for the San Juans by LaChapelle and Armstrong (1976). Likewise July has the lowest mean snowpack density ($250\ kg.\ m^{-3}$), mean ramsonde number (120 N) and percentage of ET snow (37 percent). In contrast, higher snow densities, snow strengths and greater proportions of ET snow are shown by the early season snowpack. Though there is some variation within and between years, the data presented in Figure 6.1 a-e implies that the Craigieburn Range snowpack undergoes distinct phases of development from a snowpack dominated by ET to one with a high proportion of TG to one dominated by an isothermal temperature regime.

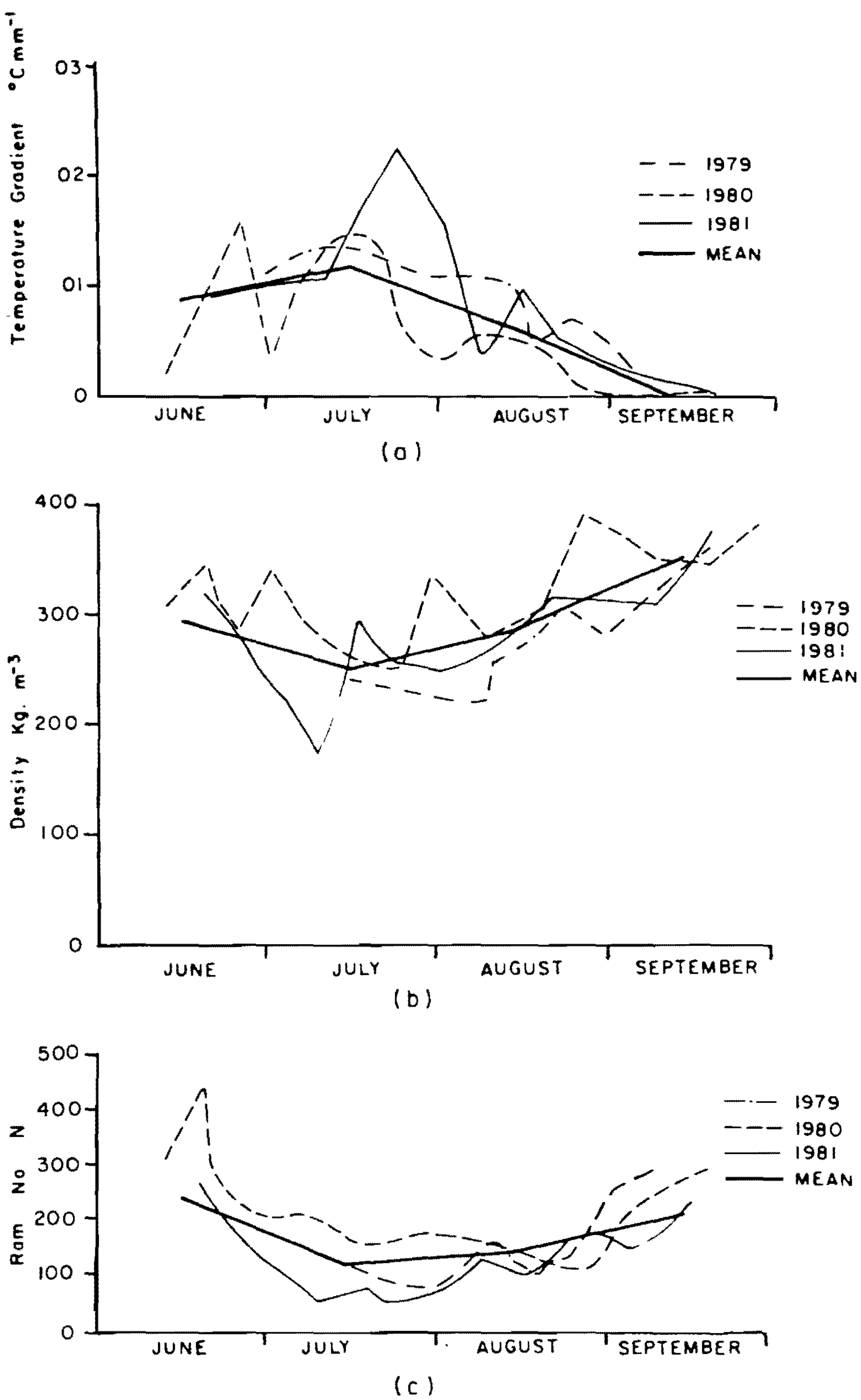
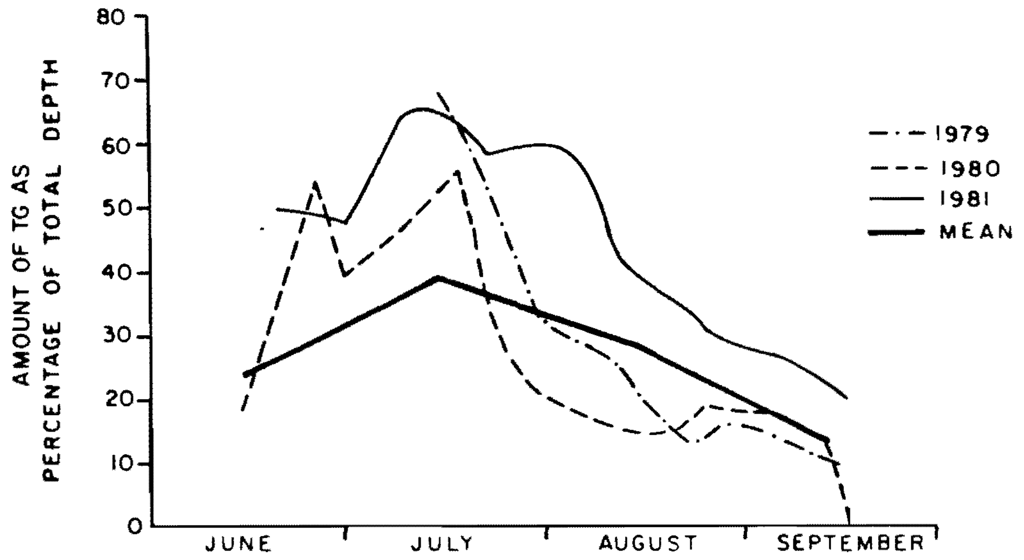


Figure 6.1a-c: Trend of snowpack parameters June to September, 1979-1981.

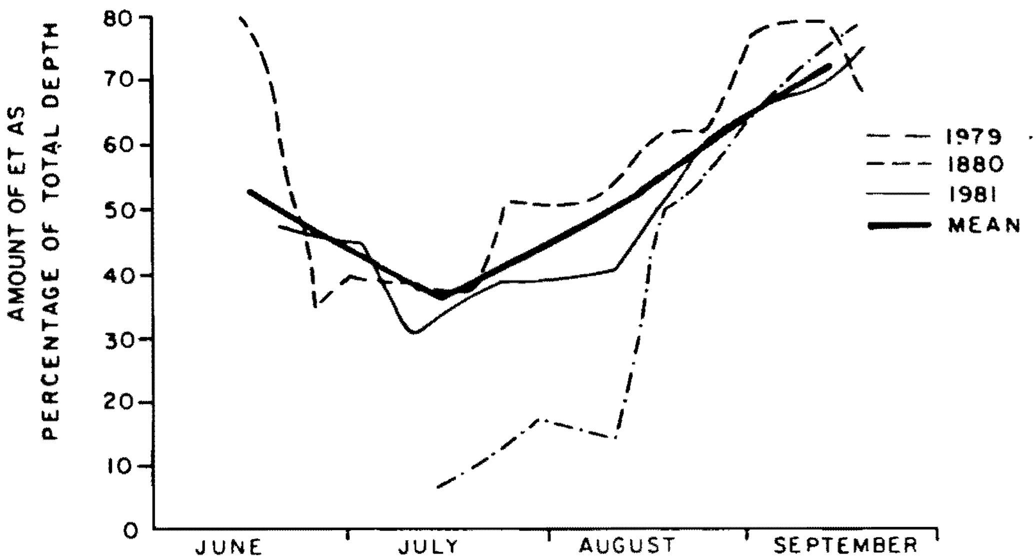
(a) Total snowpack temperature gradient

(b) Mean snowpack density

(c) Ramsonde peretrometer number



(d)



(e)

Figure 6.1d-e: Trend of snowpack parameters June to September, 1979-1981.

(d) Proportion of total snowpack depth possessing TG forms

(e) Proportion of total snowpack depth possessing ET forms

A three phase model of snowpack development is presented in Figure 6.2. The time of onset of the individual phases are not rigid and will vary from year to year according to the time of arrival of the first permanent winter snow in phase 1 and the onset of general isothermal conditions in phase 3. The time partitions are not necessarily strict but are suggested ones given field evidence from 1979 to 1981. This model therefore represents the development of mean snowpack conditions for the Craigieburn Range. Though phases 1 and 2 are typified by general anisothermal conditions, isothermal conditions may be realised at any time within these phases.

July has the highest mean proportion and lowest variation of total snowpack less than $-1^{\circ}C$ (Table 6.1). This indicates that snowpits at this time of the year, coincident with Phase 2, are consistently anisothermal apart from small variations most likely the result of surface radiation affects. The thermal regime of the snowpack at this time of the year is therefore more conducive to the development of TG forms as exemplified in Figure 6.1d. Greatest variation occurs in June when skeletal snowpacks may become completely isothermal in a matter of hours as a result of surface radiation. Variability is also caused at this time of the year by inputs of snow thus reducing surface temperatures. September has the lowest mean proportion of the total snowpack depth less than $-1^{\circ}C$, with the majority of the snowpack dominated by isothermal conditions. This is also evinced by the fact that September is the only month in which temperatures exceed $-1^{\circ}C$ for the total snowpack depth (Table 6.1). ET and MF metamorphism are therefore likely to predominate

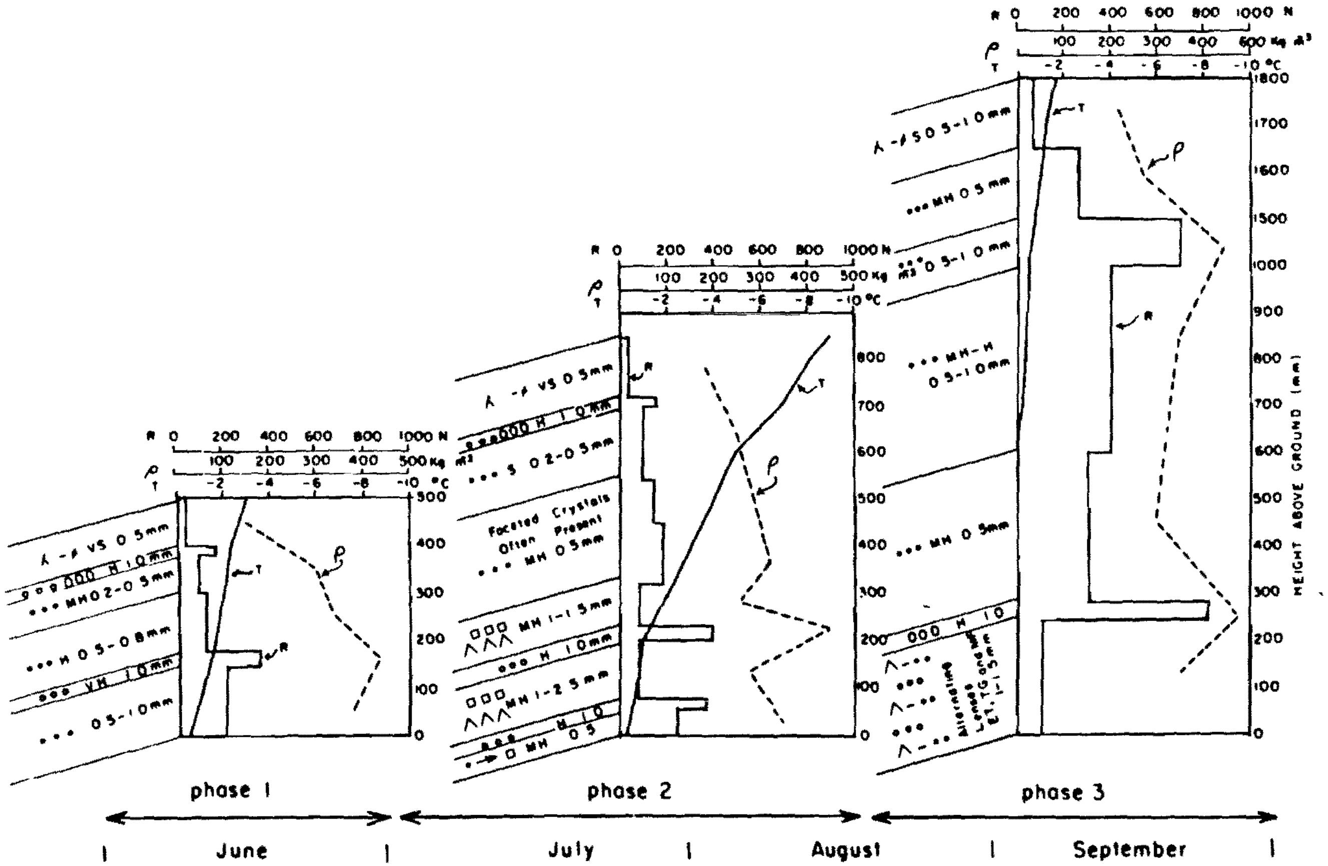


Figure 6.2: Three phase development of Craigieburn Range seasonal snowpack. Note large temperature gradients attained in Phase 2 with the occurrence of some mid-pack TG forms. Also note weak layers of basal TG from Phase 2 preserved in Phase 3.

in Phase 3 as this is a period of increasing air temperatures and greater occurrence of warm north-westerly winds.

Table 6.1: Proportion of Total Snowpack Depth Below -1°C By Month

	Month			
	June	July	August	September
Mean (Percentage)	45.8	88	53	18
Standard Deviation	23.4	6.7	21.5	21.4
Number of Observations	12	11	33	16
Number of Pits >-1°C	0	0	0	5

The three phase model presented shows that TG snow formed in the latter stages of Phase 1 and early stages of Phase 2 may be maintained in the pack until late August. Lower parts of the pack at this time of the year (Phase 3) are characterised by interdigitating layers of TG changing to ET and MF. While retaining a relatively high density of around 350 $kg.m^{-3}$ these layers remain weak compared to the rest of the snowpack and may be susceptible to collapse either under heavy new snow increments, more common at this time of the year, or as a result of melt and bond breakage. By late August-early September (Phase 3) the snowpack is dominated by large strong slabs of ET snow, some reaching densities as high as 450 $kg.m^{-3}$. This contrasts with the upper parts of the snowpack in Phase 2 which, while dominated by ET snow, has lower densities and faceted crystals which presumably result from persistence of strong temperature gradients.

Owens et al. (1983) have presented a similar model for the development of snowpacks for the eastern inland region of the South Island. While the general progression of development in the two models is similar, the major difference arises for the snowpacks in July to September. Evidence presented here suggests that the snowpack for July to mid August (Phase 2) is quite different to that of mid August to late September (Phase 3), with this latter phase resembling more the characteristics suggested by Owens et al. (1983) for the snowpacks of July to September inclusive.

The three phase model of snowpack development is further exemplified by a consideration of the densification of individual ET and TG layers for 1980 and 1981 (Figure 6.3). ET layers show continuous densification with a rapid density increase in the latter stages of the season coincident with Phase 3. In contrast TG layers do not densify rapidly but remain relatively uniform (Phase 2) until a rapid density increase paralleling that of the ET. As an increase in density generally parallels an increase in strength, the TG layers do not gain much strength as the overlying ET layers are gaining strength and bulk density. Basal TG layers therefore gain strength through densification slowly compared to overlying TG layers. As discussed by Prowse and Owens (1984) TG snow remains brittle and does not deform, while ET snow deforms and densifies gradually. This is probably a result of the high compressive strength of TG snow (Akitaya, 1975; LaChapelle and Armstrong, 1976) which negates any densification effects due to overburden pressure. These basal layers, however, remain weak in shear and are susceptible to failure in the latter parts of the season when shear stresses may rapidly increase as a result of mechanical load. The wet slab event described by McNulty and

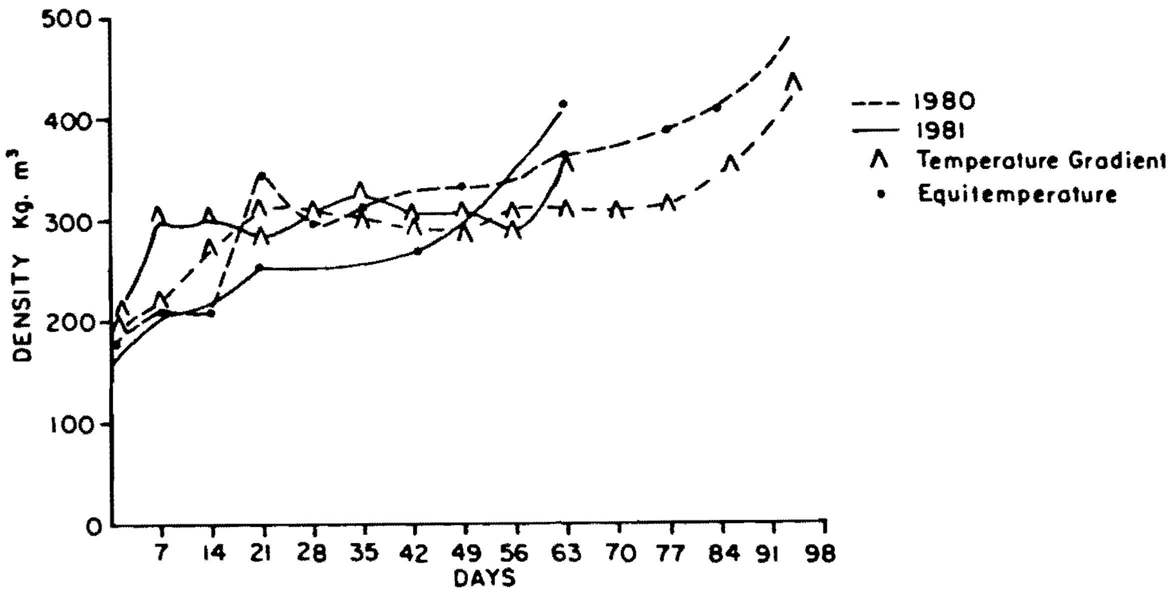


Figure 6.3: Densification of ET and TG layers over time for 1980 and 1981. Note minimal densification of TG layers after initial rapid densification.

Fitzharris (1980) is a good example of initial failure in the upper parts of the snowpack with subsequent loading and shearing of a basal TG layer. In such cases the volume of snow involved in avalanching may be increased markedly.

6.3 SNOWPACK TEMPERATURE GRADIENTS

The previous section showed that TG snow forms an important part of the Craigieburn Range snowpack. This verifies the findings of McNulty and Fitzharris (1980), Prowse (1981), Owens et al. (1983) and Weir (1983) who have similarly reported the occurrence of TG forms in this area. Prowse (1981) has calculated mean monthly snowpack temperature gradients for the Craigieburn Range according to the method of Akitaya (1974). However, all gradients noted are less than the proposed value of $0.01^{o}C\ mm^{-1}$ (LaChapelle, 1969) for active recrystallisation in the snowpack. Using the formulation suggested by LaChapelle and Armstrong (1977) the time required (t) for TG formation may be approximated by

$$t = K/\frac{dpv}{dz} \qquad (6.1)$$

where K = the coefficient of recrystallisation ($mb\ d\ mm^{-1}$)

$\frac{dpv}{dz}$ = the vapour pressure gradient ($mb\ mm^{-1}$)

Given $K = 0.04$, as suggested by LaChapelle and Armstrong (1977) for the beginning of TG development and $\frac{dpv}{dz} = 0.005\ mb\ mm^{-1}$, the length of time required for TG development would be 8 days. If $\frac{dpv}{dz} = 0.0035\ mb\ mm^{-1}$ (equivalent to a temperature gradient of $0.01^{o}C\ mm^{-1}$, with temperature at the ground-snow interface assumed to be $0^{o}C$) TG development should take place within 11.5 days. Though the time for TG development appears to lie between 8 and 11.5 days, LaChapelle and Armstrong (1977) have also noted that the effects of recrystallisation can appear in new snow within 24 hours, while mature depth hoar may form within 72 hours if the vapour pressure gradient is large enough. As these forms are particularly weak in shear (Colbeck, 1982b), the identification of periods in which large temperature gradients appear, with the related development of TG forms bears significant implications for potential avalanche release. Clearly an analysis of short term temperature gradients is warranted to establish the magnitude of these within the Craigieburn snowpack with possible implications for TG development.

6.3.1 Methods

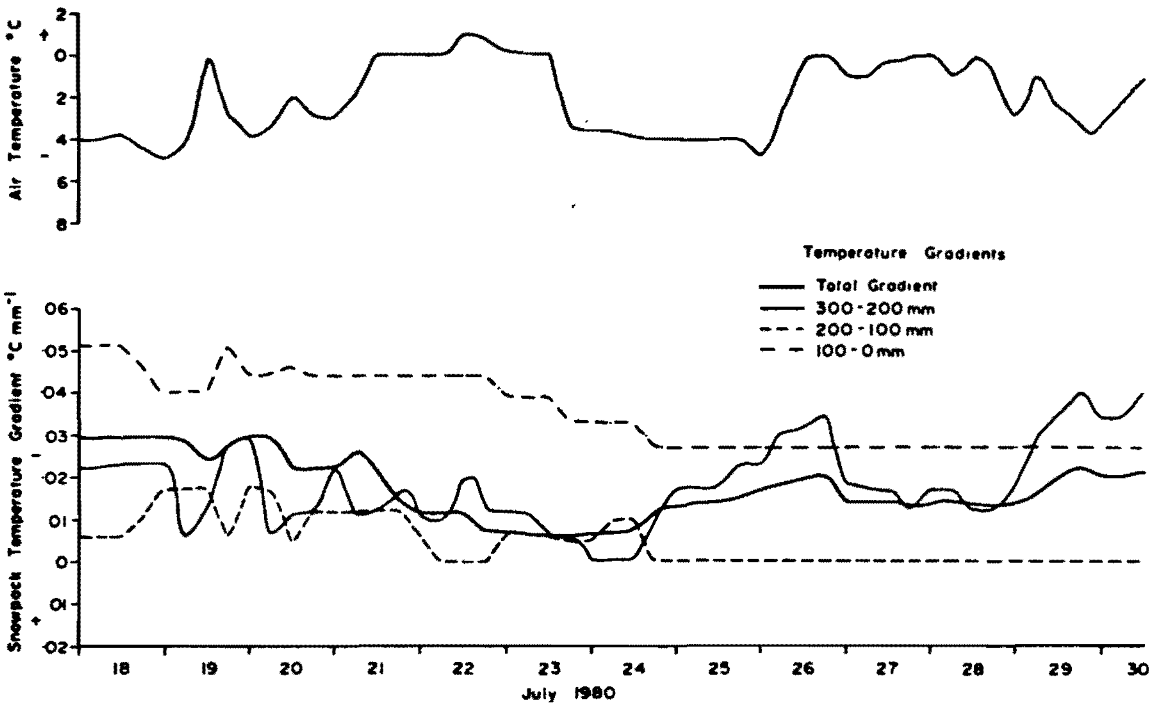
A limited record of the snowpack temperature gradient was obtained for short periods in 1980 and 1981 from a three point distance thermograph situated at 1680 m on Mt Cockayne on an unvegetated scree slope of 28 degrees with an easterly aspect. The majority of the record is for June and July. Probes were located initially at ground level, 100 mm and 200 mm above the ground. These were later adjusted to 100 mm, 200 mm and 300 mm above the ground as snowpack depth increased.

Negligible new snow increments were received at the study site for June to July 1981, the period for which the majority of the record exists. Total snowpack depth at this site rarely exceeded 0.40 m. Frequent checks on the positioning of the top probe relative to the surface of the snowpack revealed that it was consistently near the surface of the pack. Total snowpack temperature gradient for this specific study therefore applies to that gradient between ground level and 300 mm. The top probe was occasionally exposed resulting in large diurnal fluctuations of the temperature trace. Such periods in the record were excluded from the analysis.

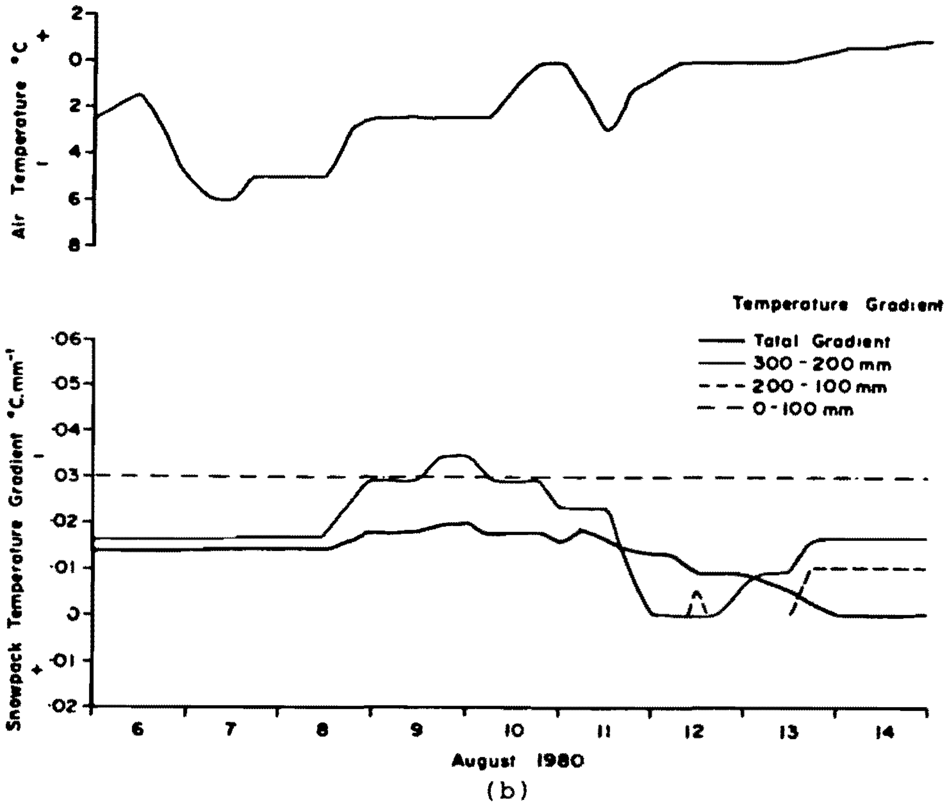
6.3.2 Results

During the six week period of record for 1981, the only place in the snowpack to show a relatively constant temperature was at ground level. Here temperatures were maintained within the range of $-0.6^{\circ}C$ to $-1.2^{\circ}C$. In only one instance did ground level temperatures fall below these. In this case, snowcover had been stripped to 100 mm, with ground level temperatures of $-3.5^{\circ}C$ to $-4.1^{\circ}C$. Ground below the surface remained frozen during the early parts of winter but thawed as lower parts of the snowpack became isothermal from mid-August.

Plots of six hourly temperature gradient values demonstrate well the effect of air temperature trends (Figure 6.4 a-e). This is especially so for the period 20 to 28 June, 1981 (Figure 6.4c and d). From 1200 hours on 20 June general cooling took place with a drop in air temperature. This was accompanied by a rise in total snowpack temperature gradient to $0.033^{\circ}C\ mm^{-1}$ by 0600 hours on 23 June. Air temperatures reached their lowest at 0400 - 0600 hours on 24 June with an associated temperature gradient of $0.026^{\circ}C\ mm^{-1}$. Following this a rapid warming phase ensued with destruction of the total snowpack temperature gradient as near surface snow



(a)

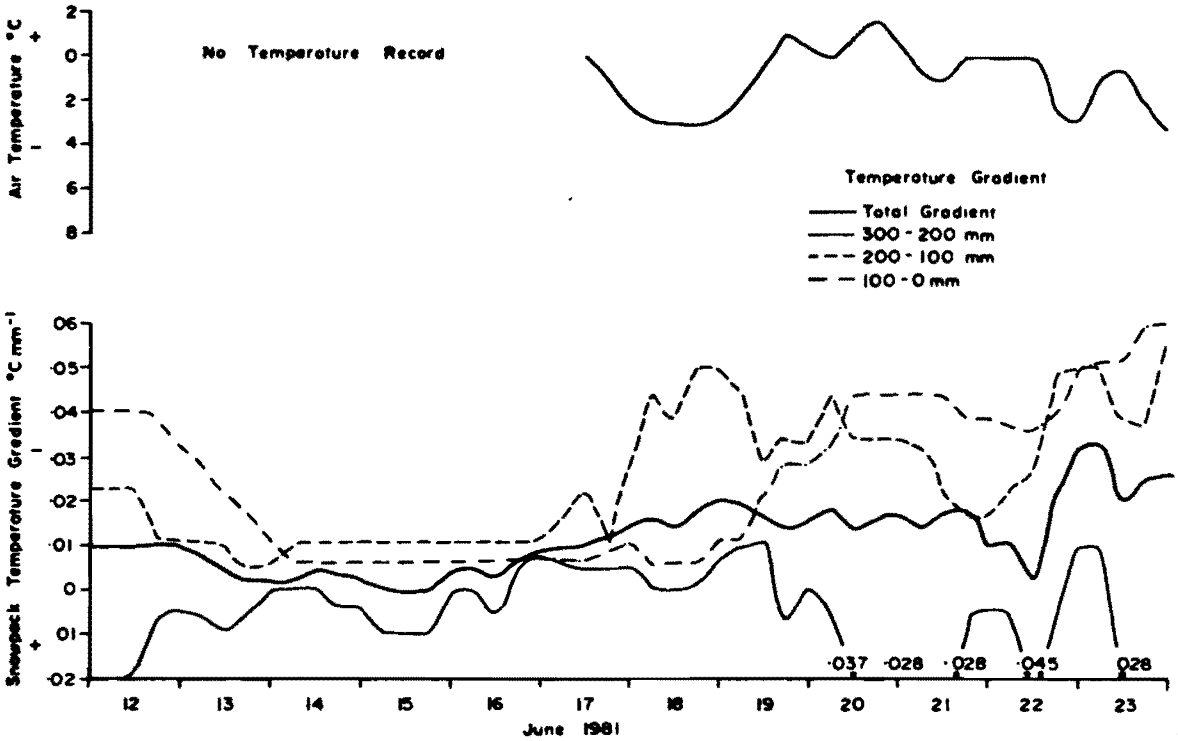


(b)

Figure 6.4: Plot of air temperature and six hourly snowpack temperature gradients.

(a) 18 July, 1980 to 30 July, 1980

(b) 6 August, 1980 to 14 August, 1980



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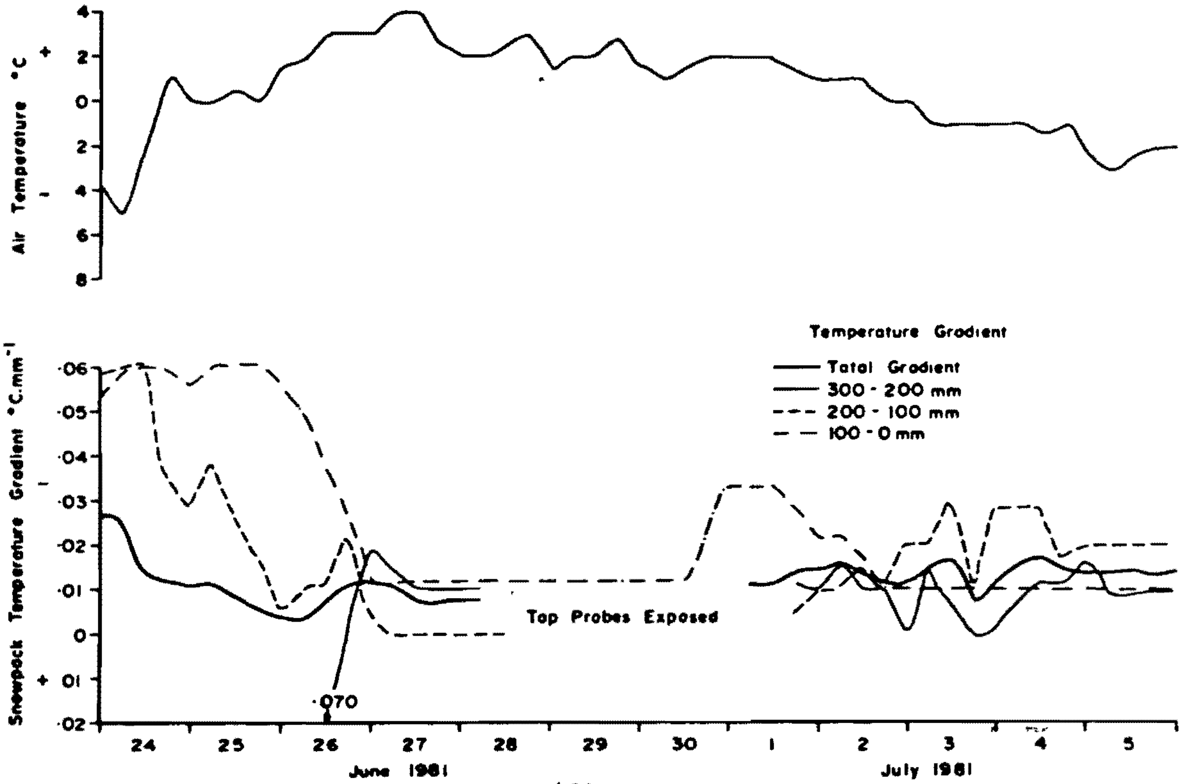
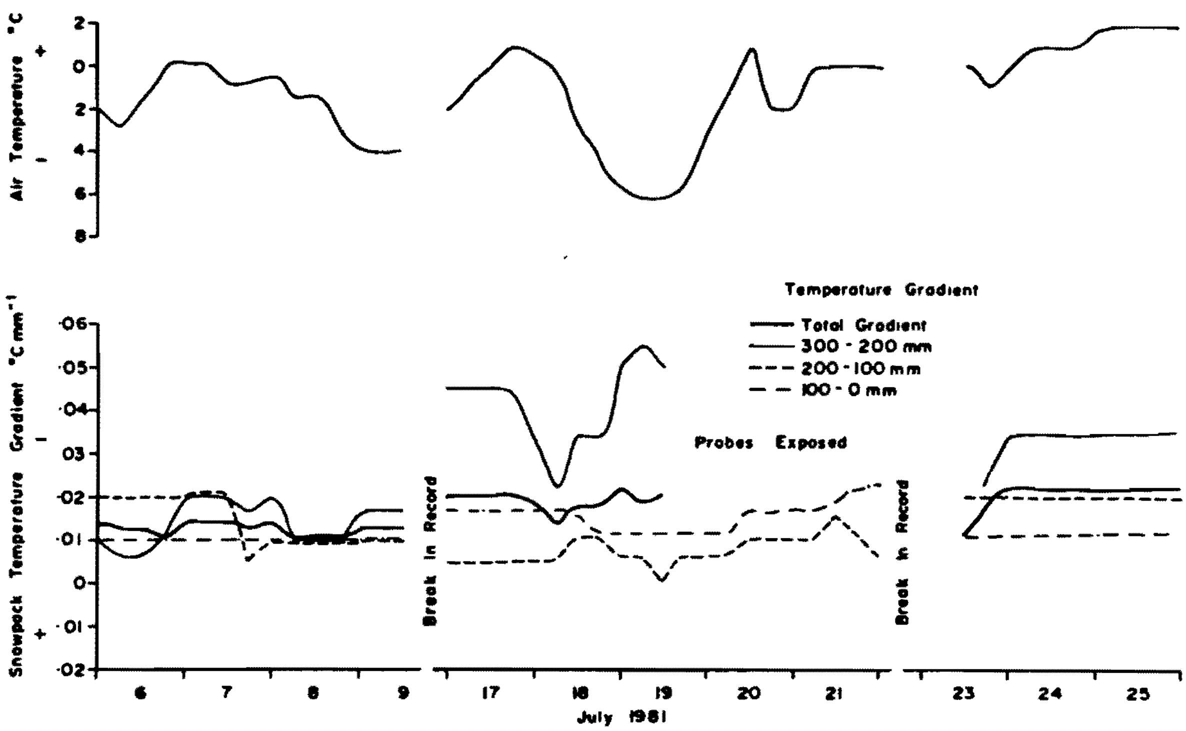


(d)

Figure 6.4: Plot of air temperature and six hourly snowpack temperature gradients.

(c) 12 June, 1981 to 23 June, 1981. Note strong positive gradients

(d) 24 June, 1981 to 5 July, 1981.



(e)

Figure 6.4: Plot of air temperatures and six hourly snowpack temperature gradients.

(e) 6 July, 1981 to 25 July, 1981

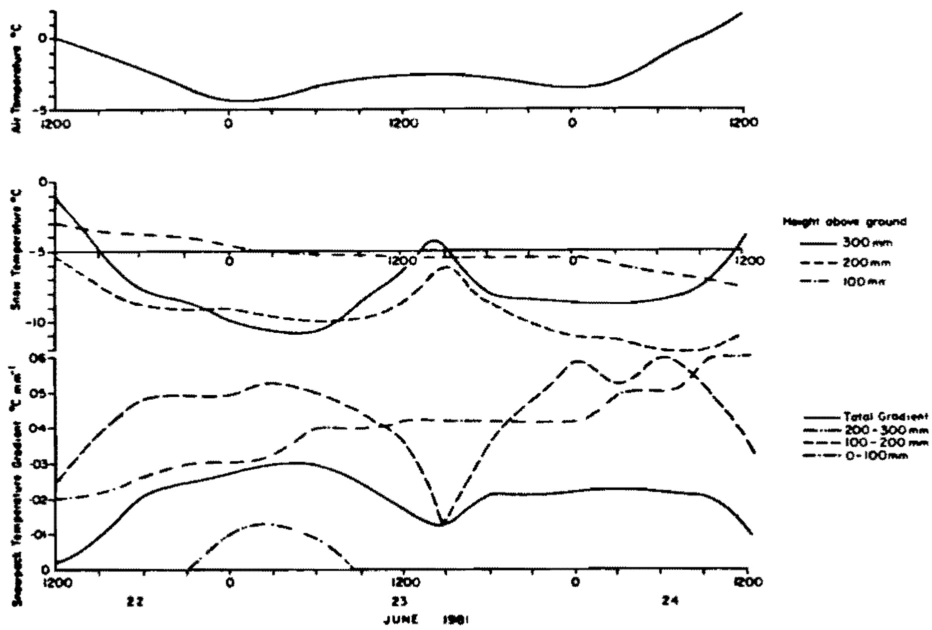


Figure 6.5: Continuous record of snowpack temperatures and temperature gradients, 1200 hours 22 June, 1981 to 1200 hours 24 June, 1981.

temperatures rose, with air temperatures reaching 3-4^{o}C above zero. Accordingly, individual temperature gradients within the snowpack decreased rapidly. Probably the most dramatic of these was the gradient in the lowest 100 mm of the snowpack. With the rapid warming, temperature gradients in this layer dropped from 0.06^{o}C mm^{-1} to 0.011^{o}C mm^{-1} within 24 hours.

A 48 hour period was monitored to establish the effects of diurnal air temperature fluctuations on the snowpack temperature gradient and snowpack temperatures (Figure 6.5). During this period, temperatures in the snowpack immediately above the ground were consistently -0.5^{o}C. Weather during this period was clear, with negligible wind and air temperatures consistently below 0^{o}C. Total snowpack depth during the 48 hours was 350 mm with the snowpack dominated by TG snow (Figure 6.6). Total snowpack temperature gradient shows diurnal variation as does air temperature (Figure 6.5). If individual layer temperature gradients are considered, the entire snowpack plus layers at 0-100 mm and 100-200 mm show negative gradients for the entire 48 hours, with total snowpack temperature gradient reaching a maximum of 0.03^{o}C mm^{-1}. The layer 200-300 mm, however, shows negative gradients for only a very short time. This may have implications for net direction of vapour transfer. Positive temperature gradients occurred in this layer for the majority of the time over 48 hours reaching a maximum after 1200 hours, when air temperatures and near snow surface temperatures were highest. In addition to the period of 20 to 26 June 1981, positive temperature gradients also occurred from 12 to 16 June, 1981 (Figure 6.4c). The implications of

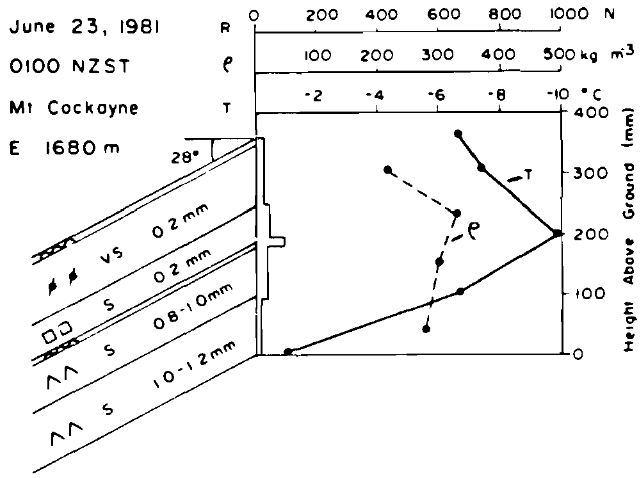


Figure 6.6: Snowpack structure adjacent to three point distance thermograph.

persistent near surface positive temperature gradients in the snowpack will be discussed in the next section.

In contrast to near surface positive temperature gradients, several periods also exist in the record of near surface negative temperature gradients (Table 6.2).

Table 6.2: Periods When Mean Surface Gradient Exceeded Mean Total Snowpack Temperature Gradient.

Period	Mean Temperature Gradient $^{o}C\ mm^{-1}$ Total	Near Surface
24-30 July 1980	.015	.025
06-11 August 1980	.016	.024
26-28 June 1981	.010	.028
17-19 July 1981	.019	.040
23-25 July 1981	.022	.034

Calculations of daily snowpack temperature gradients were also made (Akitaya, 1974) for the 1979, 1980 and 1981 study periods using snowpack and temperature data collected at the BR site (Figure 6.7a-c). The only periods which showed temperature gradients greater than $0.01^{o}C\ mm^{-1}$ were the early months of May and June 1979, June 1980, and early July 1981. In all cases, total snowpack depth was well below 0.5 m. The longest period in which gradients greater than $0.01^{o}C\ mm^{-1}$ were sustained was from 7 June to 20 June, 1980. Temperature gradients for this period reached $0.07^{o}C\ mm^{-1}$ and represent gradients typically found in skeletal snowcover of less than 100 mm.

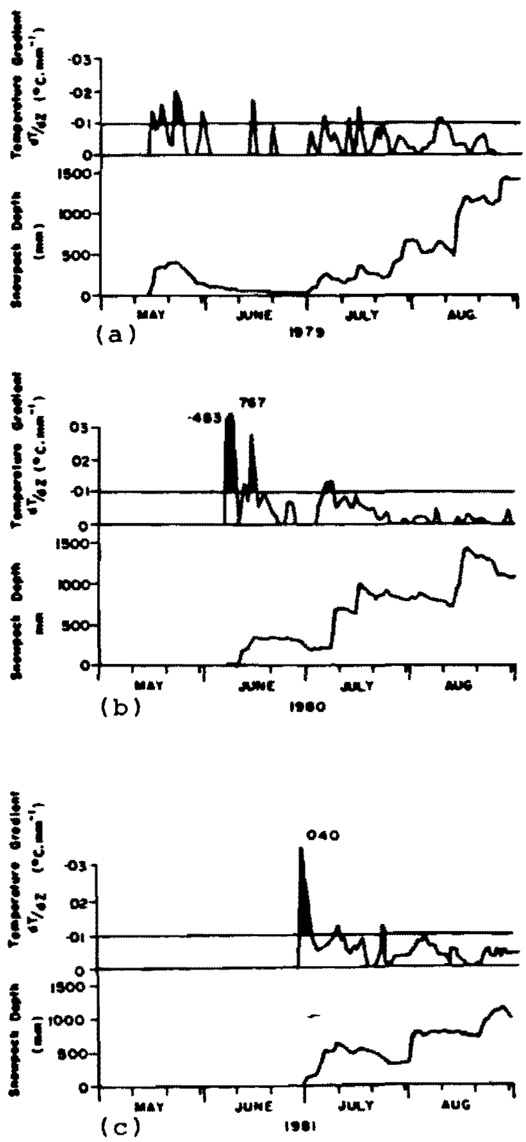


Figure 6.7: Daily snowpack temperature gradients May to August 1979 to 1981. Note the majority of temperature gradients exceeding 0.01^{o}C mm^{-1} occur when snowpack depth less than 500 mm.

6.3.3 Discussion

While vapour transfer and TG recrystallisation may take place under the strong negative near surface and total snowpack temperature gradients identified in this study, it is proposed that vapour transfer may also take place under positive near surface temperature gradients. Such a process may enhance TG recrystallisation in the upper parts of the snowpack. Sustained periods of positive vapour pressure gradients of up to 0.02 mb mm^{-1} would be sufficient to impart the effects of recrystallisation in the upper parts of the snowpack within 2 days by sublimation of vapour from the upper pack onto colder grains in the middle of the pack. Six hourly positive vapour pressure gradients for this study ranged from 0.012 mb mm^{-1} to 0.02 mb mm^{-1} (equivalent to $0.005^{o}C$ mm^{-1} to $0.07^{o}C$ mm^{-1}) which are sufficient to impart the effects of recrystallisation in the snowpack within 3.2 to 2 days respectively. Although not as high as the positive vapour pressure gradients noted by Armstrong (1981), gradients in excess of 0.012 mb mm^{-1} are evident for three days between 1200 hours on 23 June 1981 to 1200 hours on 26 June 1981. Such prolonged periods of positive vapour pressure gradients contrast with the highly variable ones of short duration noted by Granberg (1979) and Armstrong (1981). The presence of a strong negative vapour pressure gradient in association with a positive gradient may also lead to a concentration of vapour in the middle of the snow-pack, thus increasing the likelihood of TG recrystallisation in this area (Figure 6.8).

In addition to this mid-pack sublimation process, the formation of faceted crystals may also be enhanced by the

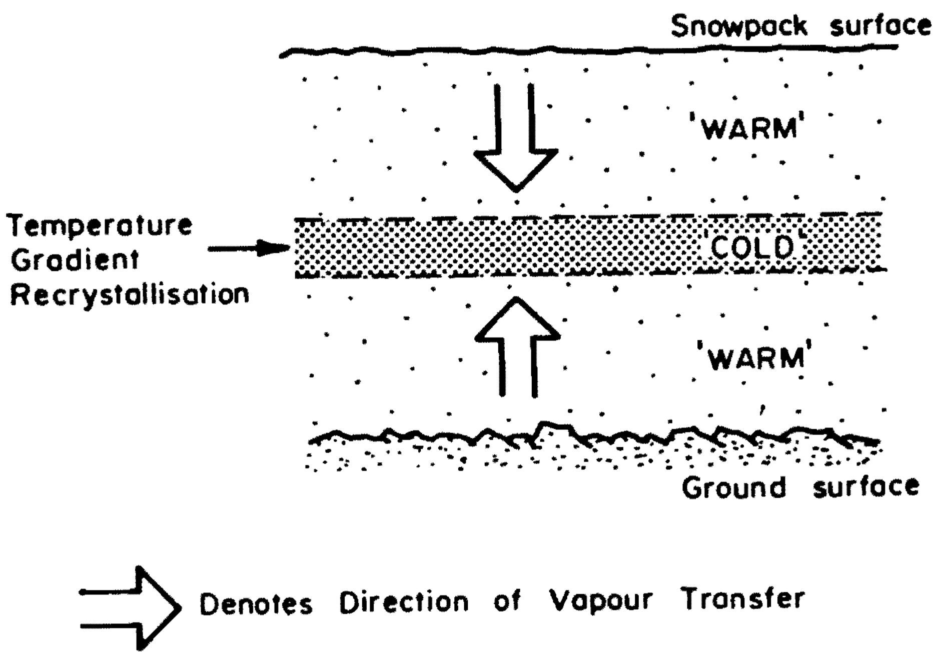


Figure 6.8: Development of upper pack positive and lower pack negative temperature gradients with concentration of vapour in the middle of the pack and mid-pack temperature gradient metamorphism.

presence of a MF layer. Vapour moving in response to a positive or negative temperature gradient would be inhibited by this impermeable barrier. Recrystallisation in such a case would take place on the top or underside of the MF layer, depending on the direction of the vapour transfer. Such a process of flow restriction is suggested by de Quervain (1958) and Prowse (1981). Colbeck (1982a) has also noted that TG snow may form above MF layers as a result of the high saturation vapour pressures associated with these features. During the period of field work TG forms were frequently found associated with MF layers. Also as noted by Bradley et al (1977), Prowse (1981), Adams and Brown (1983) and Prowse and Owens (1984) temperature gradient anomalies were found associated with these areas in the snowpack and are the result of the metamorphic state of the snow (Bradley et al., 1977).

Armstrong (1981) suggested that strong upper pack temperature gradients may arise from different combinations of meteorological events. An adequate temperature gradient may be set up between a layer of isothermal snow which will naturally have high vapour pressures and a layer of cold newly deposited snow, if the weather remains clear and cool. Conditions such as these would be transient (Armstrong, 1981). It is also suggested by Armstrong (1981) that on a smaller scale, a similar situation may develop if a melt-freeze crust is buried by a cold snowfall, with the melt-freeze layer retaining its heat. Vapour pressure differences will therefore exist between the newly deposited snow and the 'warmer' melt-freeze crust. This latter process is probably unimportant in the Craigieburn Range as most storms start off warm and cool down (Prowse, 1981). Snow temperatures are therefore not expected to be low enough for a significant temperature gradient to be set up.

Though a range of processes governing mid-pack TG development exist, the importance of these is envinced by the occurrence of significant layers of this form, 10 to 50 mm in thickness, in 11 of the 73 snowpits analysed in this study. Such forms are believed to resemble those found by LaChapelle and Armstrong (1977) and Armstrong (1981) for the San Juan Mountains, Colorado.

6.4 SNOW STRENGTH STUDIES

This section attempts to identify specific snowpack structural characteristics diagnostic of weak layers. Identification of these characteristics in the snowpack prior to snowfall will allow evaluations to be made of likely snowpack behaviour during periods of heavy precipitation and slope loading. Fracture line profile analysis, shovel and shear frame tests were used to assess the different types of weak layers and the relative strengths of these.

6.4.1 Fracture Line Profile Analysis

The most effective method for establishing snow stratigraphic relationships involved in avalanche release is by fracture line profile analysis. This involves the excavation of a snowpit at the point where the avalanche has broken away. Only nine fracture line profiles were recorded during the study period and appear in Appendix 8. This is not as many as was initially planned due to fracture line inaccessibility and lack of suitable personnel for recording when the writer was not in the field. The crystal types involved in avalanche release, snowpack and weather conditions prior to release, and probable release mechanisms are summarised in Table 6.3. The high proportion of climax releases in this analysis is a result of the inclusion of two artificial releases and the fact that sampling of direct action release fracture lines was not always possible because of weather constraints. The 'below, at, above' terminology used in this analysis refers to the position of layers relative to the weak layer. The 'at' layer, is therefore the weak layer in which failure has occurred. Weak layers for this study are composed predominantly of partially metamorphosed new snow undergoing ET with clusters of rimed needles or faceted crystal types. The atmospheric processes responsible for formation and reasons for the occurrence of clusters of rimed needles towards the base of freshly deposited slabs have been discussed in the previous chapter. Stethem and Perla (1980) have also noted the common occurrence of partially metamorphosed new snow, which may be rimed, in the weak layers of slab avalanches at Whistler Mountain, British Columbia. Faceted crystals have also been shown by Stethem and Perla (1980) to be important as weak layers. For the Craigieburn Range, the faceted crystals in profile 5 are most likely remnant surface hoar as they consitute a distinct thin layer, and occur above a former surface MF layer. Profile 2 with weak layers composed of early TG snow in the upper pack, resembles most closely those release types noted by LaChapelle and Armstrong (1976) for the San Juan mountains, where failure of TG snow, often associated with fragile freeze-thaw crusts, predominates. As discussed in Section 6.3.3 such TG-MF associations are of common occurrence in the Craigieburn Range snowpack. Bed surfaces for

Table 6.3: Summary of Crystal Types and Conditions Leading to Slab Avalanche Release

Fracture Profile No	Crystal Types in Relation To Weak Layer			Snowpack and Weather Conditions Leading to Avalanche Release	Type of Release	Probable Release Mechanism
	Below	At	Above			
1	MF	Combination of early to advanced TG	ET	Temperature gradient of $0.01^{\circ}C$ mm^{-1} for pack 0-500 mm AGL. Development of TG at base above MF later	Artificial Release (climax)	Artificial control performed to release slab at 530 and 730 mm AGL. Failure occurred with collapse of TG layer at 60-130 mm layer.
2	ET to early TG	ET with faceted grains	ET to early TG	Severe temperature gradient developed above crust at 310 mm AGL. Deposition of slab at 335 to 550 mm AGL followed by slab at 500 to 670 mm AGL	Climax	Shearing in layer subject to TG and dominated by faceted grains, possibly some surface hoar, as a result of additional loading by slab at 500 to 670 mm AGL
3	MF	Partially metamorphosed new snow undergoing ET	Same as 'At' layer	New snow deposited above 215 mm AGL. MF layer at 305 to 350 mm AGL formed from oscillating freezing level	Direct Action	Compactive creep and general overburden causing collapse of low density snow layer above MF layer
4	MF	Partially metamorphosed new snow undergoing ET	Same as 'At' layer	Deposition of slab 630-1140 mm AGL on hard MF layer	Direct Action	Poor cohesion between cold new slab and old MF surface
5	MF and some faceted crystals	Surface hoar	New snow to early ET	Surface hoar development on pre-slab surface at 215 mm AGL Deposition of slab 225-700 mm	Direct Action	Shear of slab over remnant surface hoar developed over old MF crust
6	ET	Clusters of rimed needles	Partially metamorphosed new snow to early TG	Deposition of high density snow over lower density snow at 450 to 1200 mm AGL on compact ET surface	Artificial Release (climax)	Initial shearing in rimed needles with further collapse from load inducement of TG and ET layer at 0-200 mm AGL
7	ET	Clusters of rimed needles	Partially metamorphosed new snow undergoing ET	Deposition of slab 560-780 mm AGL on very compact ET slab	Direct Action	Poor cohesion between slab ET surface plus shearing in layer of rimed needles
8	ET	Partially metamorphosed new snow and clusters of rimed needles	ET	Deposition of high density wind compact slab over unconsolidated layer	Direct Action	Collapse of unconsolidated layer of rimed needles from overburden pressure. Similar to failure mechanics of hard slab.
9	ET	Clusters of rimed needles	ET	Deposition of small increments of snow over 84 hours with intervening cold periods	Artificial Release (Direct Climax)	Shearing in layer composed exclusively of well defined clusters of rimed needles

AGL = Above Ground Level

the nine profiles were composed predominantly of ET snow or MF layers, similar to that noted by Ho (1982) for the Mt Cook area.

A summary of slab measurements for the Craigieburn Range shows some contrasts with those presented by Perla (1977) and Stethem and Perla (1980) for a range of predominantly continental type snow climates (Table 6.4).

Table 6.4: Comparison of Slab Measurements

	This Study			Perla (1977)			Stethem and Perla (1980)		
Parameter	n	$\bar{x}$	s	n	$\bar{x}$	s	n	$\bar{x}$	s
Slope Angle (^{o})	9	34.6	1.05	194	38.3	4.8	25	40.2	5.22
Slab Thickness (m)	9	0.51	0.20	193	0.67	0.43	30	0.86	0.61
Mean Slab Density ($kg.m^{-3}$)	9	239	88	121	206	77	25	218	33.6
Density at Bed Surface ($kg.m^{-3}$)	9	365	93	72	231	76	8	226	44.1
Temperature at Bed Surface (^{o}C)	9	-3.1	1.7	111	-4.5	3.0	21	-4.6	2.2

n = number of cases
$\bar{x}$ = mean
s = standard deviation

Mean slope angle recorded by Perla (1977) and Stethem and Perla (1980) is much higher than that noted for this study. As a result of the small sample size and the tendency to sample fracture lines on gentler slopes, the actual mean for the Craigieburn Range may be higher. Greater mean slab thicknesses recorded by Perla (1977) and Stethem and Perla (1980) may reflect the greater depths of snow involved in climax release types in a continental climate as opposed to the shallower, direct action slabs involving only new snow increments for the Craigieburn Range. This is especially so for Stethem and Perla (1980). Contrasts in mean slab density for the Craigieburn Range may represent the greater role of wind and possibly the effects of riming associated with warmer ambient temperatures during periods of slab deposition in the Craigieburn Range. Differences in mean bed surface densities attest to the contrasting nature of bed surfaces. For this study, ET and MF predominate as bed surfaces, while for areas with a continental snow climate (LaChapelle and Armstrong, 1976; Perla, 1977; Stethem and Perla, 1980), TG is more prevalent. Mean bed surface temperature for the Craigieburn Range matches well the range recorded by Perla (1977) and Stethem and Perla (1980). Though only based on a small sample size, the basic difference between the slab parameters noted for this study and those recorded by Perla (1977) and Stethem and Perla (1980) are believed to reflect the contrasts between the coastal-inland snow climate of the Craigieburn Range (Prowse, 1981; Owens et al., 1983) compared to the predominantly continental type snow climates for the study of Perla (1977) and Stethem and Perla (1980).

Assessment of fracture line profiles has shown that faceted crystals associated with crusts, cold slabs overlying well developed MF layers and occurrence of thin layers of clusters of rimed needles within new snow, appear to be the predominant physical features associated with weak layers causing avalanche release in the Craigieburn Range.

6.4.2 Shovel Tests

An alternative quick and effective method for identifying weak layers in the snowpack is the shovel test. This involves the excavation of a column of snow on the uphill side of a snowpit and the observation of layers that shear off when a shovel is wedged in behind the column of snow (Figure 6.9). Weak layers once identified were subject to standard snowpit analysis techniques. Results for this analysis are presented in the form of histograms (Figure 6.10 to 6.12). Strong layers are those in which no shear occurred.

The most predominant crystal types susceptible to shear by the shovel test are partially metamorphosed new snow undergoing ET, snow in the early stage of TG and advanced TG snow changing to ET (Figure 6.10a). These crystal types only occur infrequently in strong snow layers (Figure 6.10b). All weak layers characterised by ET in the early stages of TG metamorphism were usually thin (10-50 mm) and often occurred above a MF crust in the mid to upper parts of the snowpack. As discussed in Section 6.3.3 upper pack temperature gradients may be responsible for the formation of these forms above MF layers and crusts. The virtual absence of all faceted forms from the strong snow layers supports the findings of Martinelli (1971), LaChapelle and Armstrong (1976) and Bradley et al. (1977) who have noted that these crystal types are often the most unstable. Twenty three percent of all weak layers possessed faceted forms in the early stages of TG metamorphism. These are equivalent to the subhedral forms found by Bradley et al. (1977), Stethem and Perla (1980) and Adams and Brown (1982, 1983) to be the weakest of the TG forms.

Fracture line profile analysis revealed the frequent occurrence of partially metamorphosed new snow undergoing ET

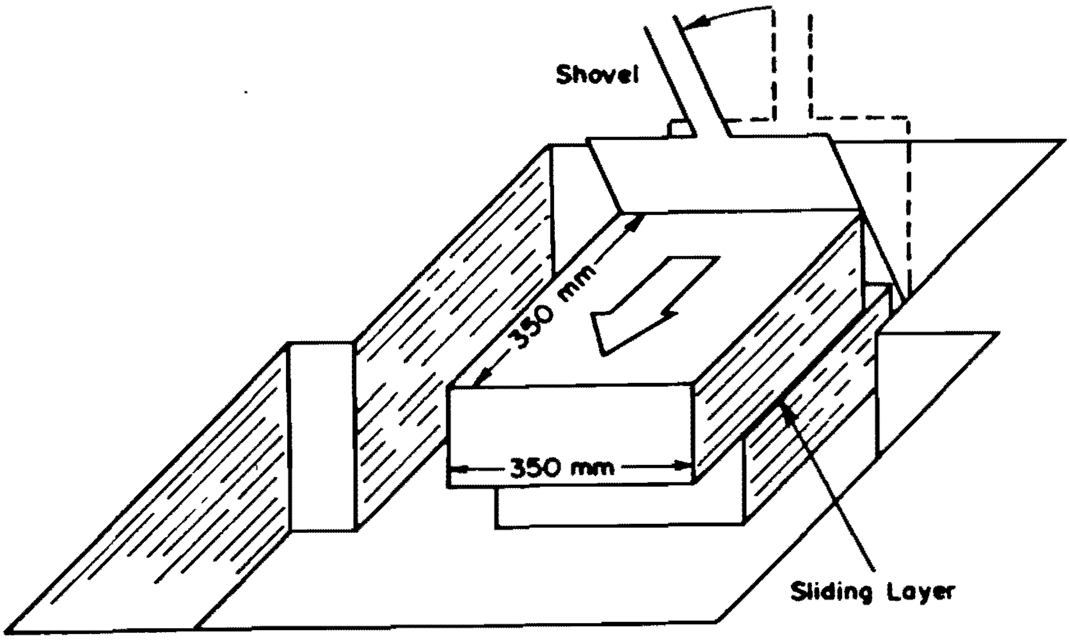
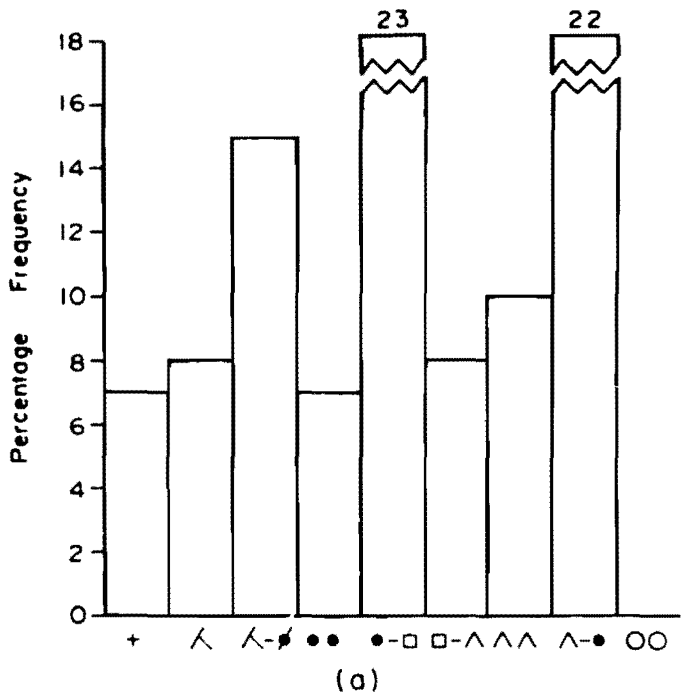


Figure 6.9: A shovel test was employed to help identify potential sliding layers. The technique involves excavating a vertical three sided column of snow in the face of a snow pit. The column dimensions are approximately 350 mm wide by 350 mm deep. A shovel is gradually inserted at the rear of the column and levered forward until the column fails. The crystal structure of the upper and lower surfaces of the sliding layer is then examined. (After Prowse, 1981)



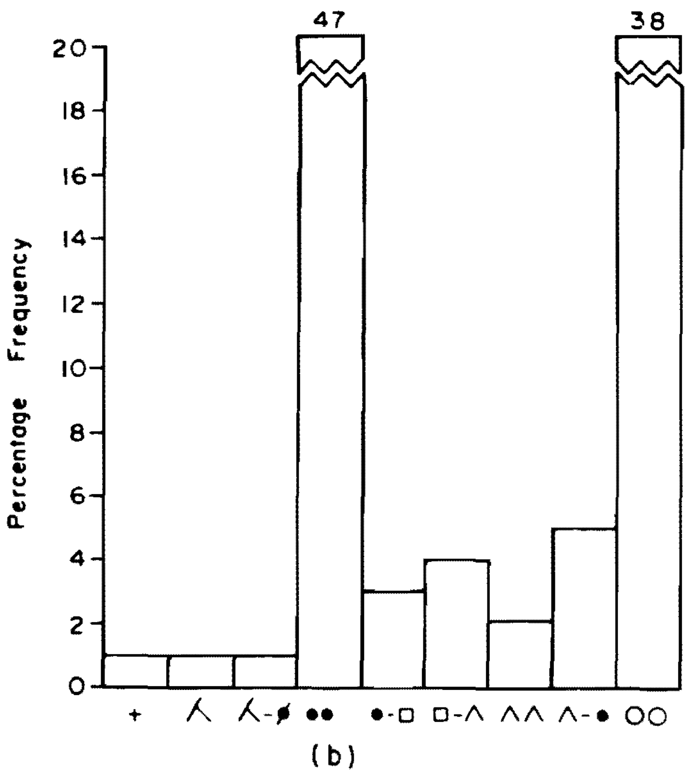


Figure 6.10: Frequency of crystal types.

(a) Weak layers. Note high frequency of early TG compared with advanced TG.

(b) Strong layers with ET and MF predominant.

in weak layers. Shovel tests confirmed this, with many layers consisting of this crystal type shearing very easily. Predominant in these layers were clusters of rimed needles constituting thin, but distinct layers. Their susceptibility to shear may result from the large levers between crystals and lack of sintering. As the degree of riming increases, the chance for multiple sintering of bonds decreases with the result that heavily rimed crystals tend to remain as singular grains. This is similar to the affect of graupel. Rimed crystals were also found in the slabs overlying the weak layers, and as noted by LaChapelle (1967) are of importance for building up thick slabs. From the evidence presented for this study it appears that the effect of crystal riming on snowpack stability may be non-linear. Up to a point riming will enhance bonding as found in the overlying slabs for this study and LaChapelle (1967), but beyond this the chance of multiple sintering will be decreased as crystals or collections of rimed crystals remain as singular grains and are thus susceptible to shear.

Advanced TG changing to ET was found as a predominant type in the weak layers. Theoretically there should be a gain of strength with this transition, but as noted by LaChapelle (1969), even though this type of snow may attain considerable strength, it always remains weaker mechanically than snow that has not been subject to a strong temperature gradient. Weakness of these crystal types may result from the loss of compressive strength owing to breakdown of necks between crystals and the preservation of a quasi-cup form with little sintering between grains typical of advanced TG. The strong layers are dominated by ET and MF snow (Figure

6.10b). This confirms the results of the fracture line analysis. The nature of the respective bonding processes for ET and MF snow make these inherently stable.

Contrasts in ram resistance between weak and strong layers also exist (Figure 6.11a and b), with weak layers showing a marked difference in the distribution of strength to that shown by the strong layers. A distinct difference between the density of strong and weak layers was also found (Figure 6.12a and b) as would be expected from the frequently referred to relationship between snow strength and density (Bull, 1956; Keeler and Weeks, 1967, 1968; Martinelli, 1971; Weir and Owens, 1981; Prowse, 1981). Mean density for the weak layers was 268 $kg.m^{-3}$ in contrast to 346 $kg.m^{-3}$ for strong layers. Fifty two percent of weak layer densities are in the range 200-300 $kg.m^{-3}$ (Figure 6.12a). Stethem and Perla (1980) reported a predominance of densities in the range of 200-250 $kg.m^{-3}$ for unstable layers. The high proportion of strong layers with densities exceeding 400 $kg.m^{-3}$ (Figure 6.12b) reflects the occurrence of ET and MF crystal types. Though contrasts in density exist, no difference in regression coefficients was apparent when ram number and density were regressed against each other for the weak and strong layers. The overall regression and correlation coefficients for this relationship (Figure 6.13) appear in Table 6.5. The correlation coefficient agrees well with those of other workers. The regression intercept and coefficient for this study resembles more closely those for the "interior snow" deposits compared to the higher density and strong "wind surface" deposits of Prowse (1981) and the snow, excluding depth hoar, less than four months old of Martinelli (1971).

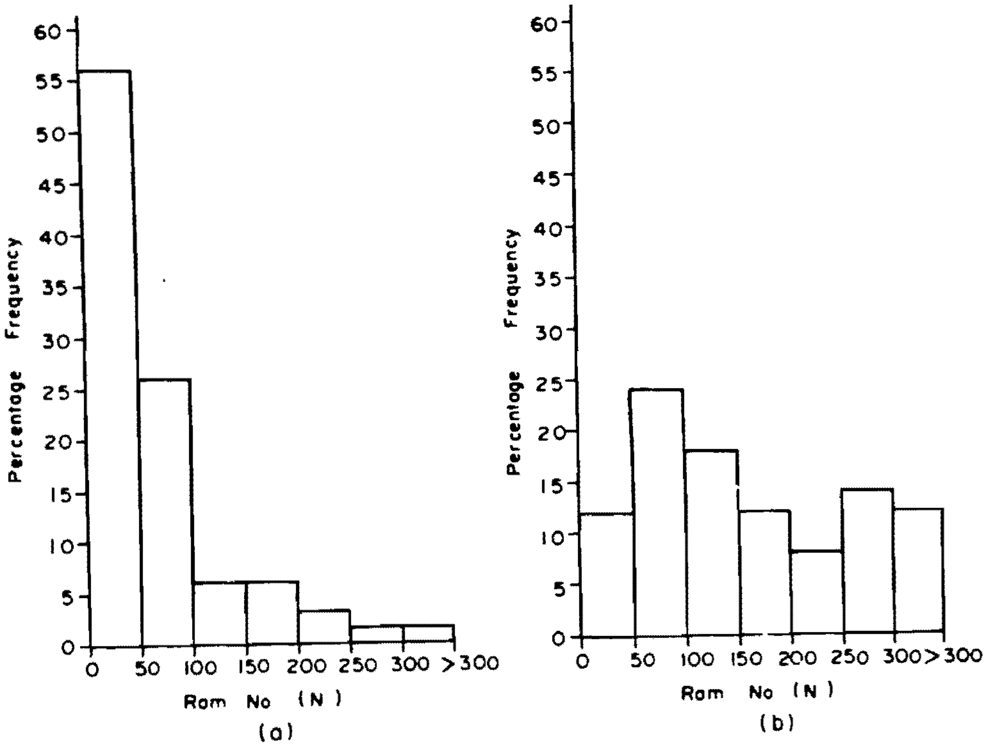


Figure 6.11: Distribution of Ramsonde Penetrometer Number.

(a) Weak layers. Note high frequency of ram numbers below 50.

(b) Strong layers. Note 46 percent of strong layers have ram numbers greater than 150 N compared to 19 percent for weak layers.

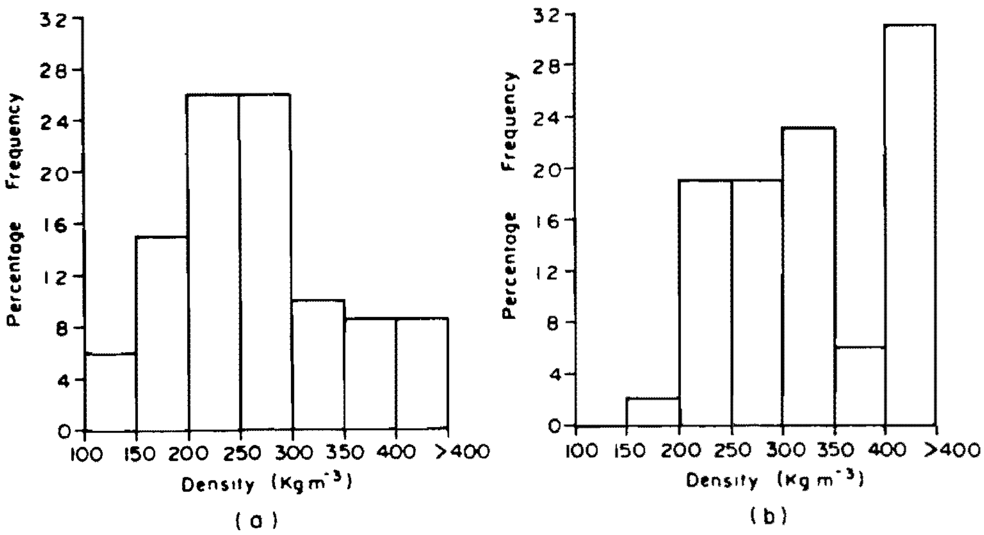


Figure 6.12: Distribution of Density.

(a) Weak layers. Majority of layers in 200-300 kg m^{-3} class which correspond to densities often found for early TG.

(b) Strong layers. 62 percent occur with densities greater than 300 kg m^{-3} demonstrating predominance of ET and MF in strong layers.

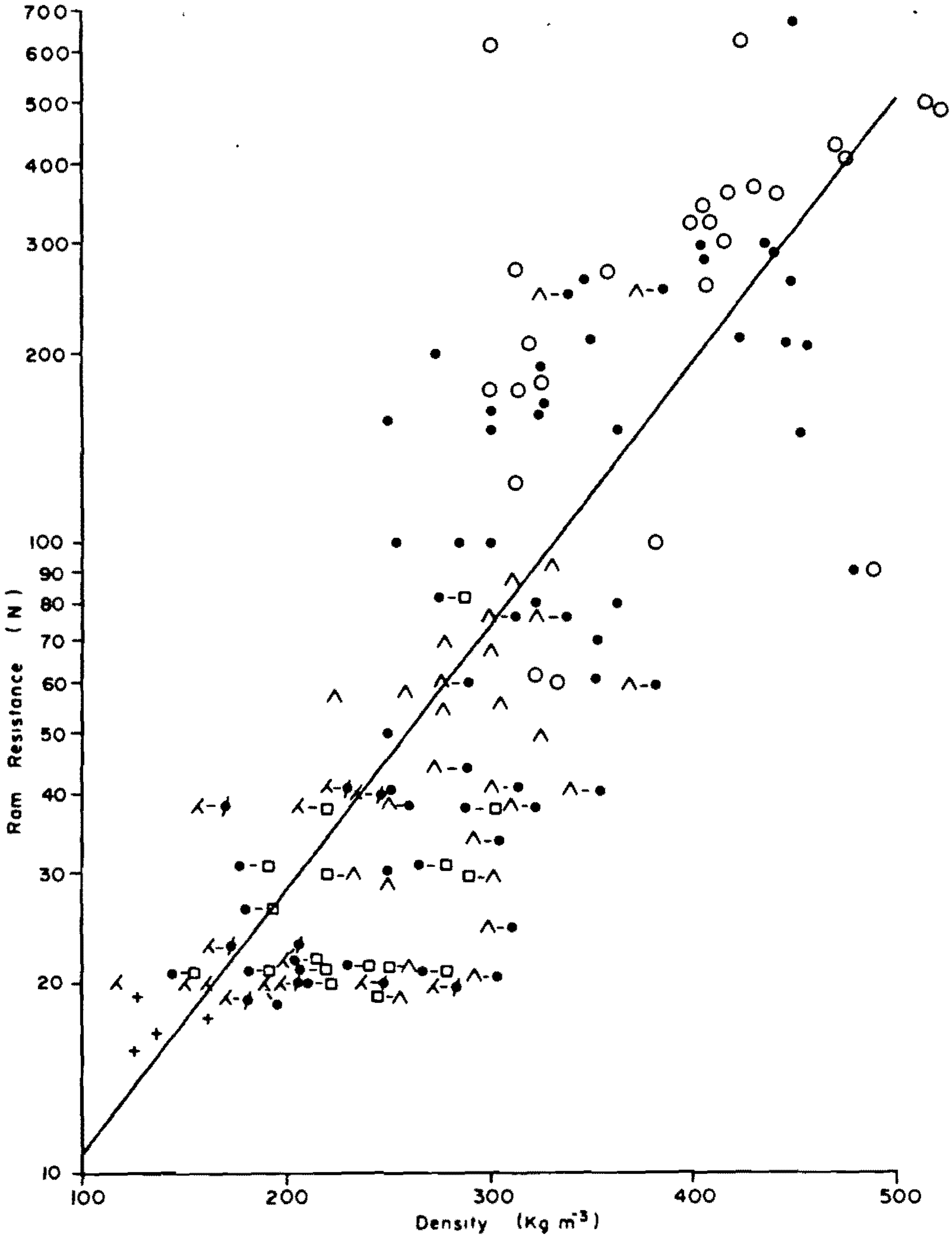


Figure 6.13: Regression of Ram Resistance and Density. Note distribution of crystal types showing a crystal type strength relationship.

Table 6.5: Regression and Correlation Coefficients For Log Ram Number-Snow Density Relationships

Source	a	b	r	Location
Bull (1956)	-0.6107	0.00531	0.80	Greenland
Keeler and Weeks (1967)	-0.8446	0.00640	0.94	Montana
Keeler (1968)	-0.7482	0.00599	0.89	Montana
Martinelli (1971)	-0.463	0.00421	0.68	Colorado
Weir and Owens (1981)	-0.477	0.0050	0.85	Mt Hutt
Prowse (1981) All samples	-0.408	0.0046	0.79	Craigieburn Range
Prowse (1981) Interior Snow	-0.392	0.0045	0.83	Craigieburn Range
Prowse (1981) Surface Wind Deposits	-0.485	0.0051	0.78	Craigieburn Range
This Study	-0.354	0.0041	0.77	Craigieburn Range

a = regression intercept
b = regression coefficient
r = correlation coefficient

Although no difference between the regression coefficients for the weak and strong layer exists in contrast to differences found by Martinelli (1971) between young (less than 14 days) and older (less than one month) snow, an interesting spread of the data occurs which may represent possible crystal type-strength relationships (Figure 6.13). Partially metamorphosed new snow undergoing ET and ET snow in the early stages of TG occur in association in the plot. For the partially metamorphosed new snow undergoing ET sintering has not progressed to the stage where well developed bonds are evident, whereas for ET undergoing TG, bonds are being reduced with the formation of facets. In both cases the nature of the bonding is weak, susceptible to shear and not very resistant to vertical compression. Both these therefore appear as the weakest of the metamorphosed crystal types. Positioning of TG crystals in the early stages of metamorphism relative to advanced TG, demonstrates the much weaker nature of early TG snow. This again reinforces the contention that subhedral forms are typically weaker than euhedral or advanced TG forms.

Snow in the early stages of TG metamorphism occurring above MF layers was found to shear easily over MF layers. This is considered to be potentially more unstable than a situation in which TG occurs below a MF layer, as stress imposed on a MF layer may not be transferred to layers below it. This is analogous to the capacity of hard slab to bear considerable stress as described by LaChapelle (1978). However, as with the hard slab, collapse of a rigid MF layer onto a layer of TG grains will result in failure. Shovel tests which showed shear of a MF layer over a TG layers were those in which a layer of TG above the MF layer was absent, with the slab being firmly attached.

In terms of relative strength, it has been shown that advanced TG snow has the greatest strength of the TG forms. However, as suggested in Section 6.2 the early season development of TG may have important ramifications for either late season snowpack stability or stability during periods of thaw. Shovel tests revealed that TG undergoing ET was susceptible to easy shearing. Such metamorphic changes occurred under an isothermal regime and in many cases metamorphism had been accelerated by the effects of free water. Rapid warming phases may destroy well developed TG layers by bond breakdown as a result of free water. An example of this is presented in Appendix 9.

No contrasts in grain size occurred between the weak and strong layers as many of the well developed TG and MF grains possessed similar grain sizes. Similarly, grain size for snow undergoing ET, closely resembled snow in the early stages of TG.

6.4.3 Shear Frame Tests

The main aim of the shear frame tests was to quantify the relative strengths of the weak crystal types identified in the shovel tests. The shear frame constructed for this purpose (Figure 6.14) is identical to the Alta shear frames discussed by Perla (1969). The shear frame was placed in the weak layer with force being applied by a rapid smooth pull on a spring scale attached to the shear frame. Force at the time of failure was noted. The rapid application of stress used in this method, as noted by Kinosita (1967), presumes brittle type failure. Shear strength is given by

$$\tau = a.F \qquad (6.2)$$

where τ = shear strength (Nm^{-2})

a = area of the frame (0.05 m^2)

F = maximum force (N)

N = Newtons

Determination of the strength of all weak layers was not possible as many of the sliding layers were too thin to be sampled by the shear frame. Problems were also encountered with the insertion of the frame into weak low density layers. In such cases these layers collapsed. A knife was therefore used to cut out the basic shape of the shear frame in the snow sample with the frame being subsequently placed

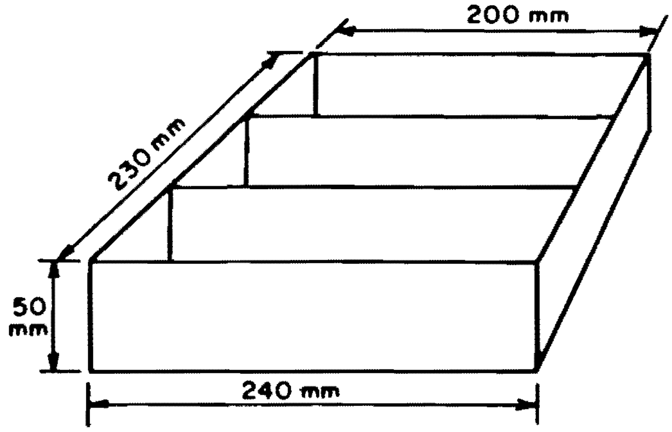


Figure 6.14: Alta shear frame used in shear strength tests.

in the sample by way of the knife cuts. For high density snow, attempts to shear often resulted in the frame plucking out a sample of snow. Repetitions were performed until a planar shear was obtained. Though some problems do exist with this method, the technique is considered suitable for strength determination and has been used widely by other workers (Perla, 1977; McClung, 1977; Montmollin, 1982; Perla et al., 1982).

The mean strengths obtained for a range of crystal types are displayed in Table 6.6.

Table 6.6: Mean and Range of Shear Strength for Different Crystal Types (Nm^{-2})

Crystal Type	Maximum	Minimum	Mean	Standard Deviation	Number of Observations
⅄ − ⌀	784	176	434	197	6
• − □, □ − Λ	1176	333	637	245	9
Λ − •	1254	646	1043	244	5
Λ	1509	784	1119	265	7
•	1430	980	1288	179	5

⅄ − ⌀	Partially metamorphosed new snow undergoing early ET
• − □, □ − Λ	Early stages of TG
Λ − •	Advanced TG changing to ET
Λ	Advanced TG
•	ET

Partially metamorphosed new snow undergoing the early stages of ET was the weakest of all crystal types followed by snow in the early stages of TG metamorphism. The range of shear strengths for partially metamorphosed new snow undergoing ET for this study falls within the range of 138 to 1020 Nm^{-2}

noted by Perla (1977) for newly deposited snow. Plotting of shear strength against density (ρ) (Figure 6.15) reveals similar crystal type-strength relationships to that of Figure 6.13. The linear relationship of LogN = 2.32 + 0.0018 (6.3) gave a correlation coefficient of r = 0.67 significant at the 0.01 level. Similar relationships for shear strength and density have been demonstrated by Perla and Martinelli (1976), Mellor (1975) and Perla et al. (1982). Perla and Martinelli (1976) have however commented that ET snow may be ten times the strength of TG snow of a comparable density. They also suggested, all other things being equal, the larger the grain size the weaker the snow. This is partly true for the crystal types in this study. However, size strength anomalies occur between the finer grained weak new snow undergoing ET and the larger, stronger higher density advanced TG grains. For this case the strength-density relationship appears to overshadow any crystal size-strength relations. Keeler and Weeks (1968) have also noted a similar pattern with dry finer grained low density snow exhibiting lower strengths than dry larger depth hoar. The ease of shearing of the lower density new snow compared to the higher density, larger depth hoar may be a function of porosity. Bond contact in a more porous media will be much less than in a comparatively less porous media, and thus will be more susceptible to the leverage effects of any applied shear stress. As applied force is redistributed as high stress concentrations at neck boundaries, the more porous media with a greater grain size to neck cross section ratio will be more susceptible to shear. The same principles apply to the contrast in the ease of

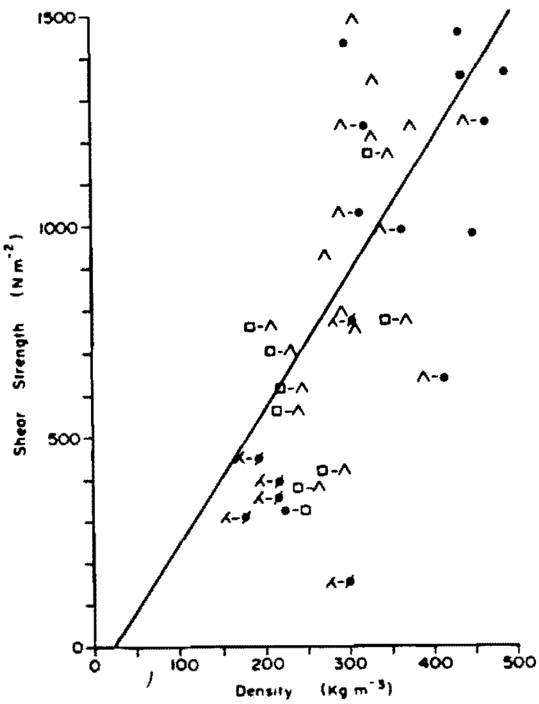


Figure 6.15: Regression of shear strength and density. Note similar distribution of crystal types as in Figure 6.13.

shear between advanced TG and ET grains as suggested by Perla and Martinelli (1976). No shear strength ram resistance relationships were found for this study as also noted by Keeler and Weeks (1968), Keeler (1969) and Mellor (1975).

In summary the shear frame tests bear out the qualitative conclusions of the shovel tests. Crystal types most susceptible to shear are partially metamorphosed new snow in the early stages of ET metamorphism, snow in the early stages of TG metamorphism and advanced TG changing to ET. In addition, the data further reinforce the contention that subhedral TG grains are much weaker in shear than advanced TG grains.

6.5 EVALUATION OF POTENTIAL SLIDING LAYERS

The foregoing analysis has mainly addressed the problem of assessing the factors influencing the occurrence of strong or weak layers in the snowpack. However, the information presented bears significant implications for assessing the behaviour of avalanche slopes possessing weak layers during periods of snowfall and hence stress application. Three categories that represent potential sliding layers in the snowpack are suggested. These categories are quite independent and therefore evaluation of snowpack behaviour in the field with reference to any one of these will involve assessment of a specific set of variables.

6.5.1 Weaknesses In New Snow Stratigraphy

Weaknesses in new snow are a consequence of sharp contrasts in the mechanical properties of individual layers deposited in the same storm. Conditions conducive to this are, deposition of high density snow on low density snow in an inverse storm resulting from air temperature increases with storm passage (Profile 6), or deposition of distinct layers of clusters of rimed needles in the early parts of storms (Profile 8). In an inverse storm, snowpack failure is likely to occur as a result of overburden pressure. The high degree of riming displayed by clusters of rimed needles results in minimal sintering. These forms therefore have little cohesion and are easily susceptible to shear. During storms the continual monitoring of air temperature trends and crystal type is important for identifying these potentially weak layers. Atmospheric conditions with high vapour pressure and temperatures approaching zero are especially conducive to riming. Shovel tests on easily accessible, but safe slopes should be used to identify early storm layers of rimed needles if information pertaining to crystal type and temperature is absent or lacking.

6.5.2 Weakness Of New Snow At And Between Old Snow Surface

This category is related to the character of the pre-slab deposition surface or nature of the new snow overlying this surface. Deposition of new snow on slopes which possess extensive areas of surface hoar represent potentially unstable situations (Profile 5). The thin neck like bonds between surface hoar and the surface they are attached to are very susceptible to any leverage effects imposed by shear stress. As surface hoar forms as a response to sublimation of water vapour onto cold surfaces, periods of cool clear weather preceding snow deposition should be noted. Extensive areas of surface hoar will be preserved on slopes that remain in shade for the majority of the day. These represent areas where the development of potential sliding surfaces may be the greatest. The shovel test is an appropriate method for establishing the degree of stability during periods of slope loading. MF crusts may form poor bonding surfaces for newly deposited snow, especially if the temperature of the MF crust is well below zero. Deposition of cold snow on such a surface will result in little cohesion between the newly deposited slab and MF crust through lack of sintering at the crust (Profile 4). Weak new snow crystal types may be deposited directly on compact cold surfaces of ET or MF and preserved for several days while overlying slabs undergo metamorphic changes. These represent potential sliding surfaces as subsequent loading of these layers may initiate failure leading to climax avalanche release (Profile 9).

6.5.3 Temperature Gradient Forms

Occurrence of these as distinct layers are considered to represent a potential sliding surface. TG grains occurring at the bottom of the snowpack and remaining there for the majority of the season are especially susceptible to shear during periods of increased air temperatures and snowpack free water content (Appendix 9). Free water may invade these layers destroying bonds. Also failure in the upper parts of the pack may induce collapse of the lower TG layers as a result of overburden pressure from over-riding snow during avalanche flow. Extensive areas may therefore breakout to the ground markedly increasing the volume of snow involved in avalanching. TG grains may also be found in the mid to upper parts of the snowpack as distinct layers (Profile 1) or associated with MF layers. TG grains in the early stages of metamorphism, lying above MF crusts, have been found to be susceptible to easy shear. Identification of distinct layers of TG greater than 50 mm thickness may be made by quick soundings with the ramsonde penetrometer. By this method, these weak layers may be identified rapidly over a wide area. Identification of thin laminates of TG grains above MF crusts can be made by the shovel test. Early season monitoring of air temperatures and average snow depths on all aspects should be undertaken in order to establish the magnitude of the temperature gradient and hence development of early season TG grains. Largest temperature gradients will be attained on shady slopes. Once the snowpack has become established, regular checking for the development of mid-pack TG grains should be made. In periods of rapid stress application, these TG forms will shear easily.

6.6 SUMMARY

A synthesis of all snowpack data has facilitated the production of a three phase model for the seasonal development of the Craigieburn Range snowpack. The early season snowpack is often characterised by mechanically and structurally strong layers of snow. Imposition of large temperature gradients however, soon reduces the mechanical strength of the snowpack with maximum snowpack temperature gradients and minimum snowpack strengths appearing in June and July. However, because heavy snowfalls are uncommon in June and July, climax releases are more likely from mid August to September.

Continuous temperature gradient records have shown that in contrast to the usually negative temperature gradients developed in the snowpack in response to radiative cooling, and responsible for TG formation, significant positive temperature gradients may persist for up to 72 hours. Vapour transfer may therefore become reversed in the upper parts of the pack with sublimation of vapour on to cooler grains in the middle of the pack aiding TG formation. Snowpack temperature records have also shown that anisothermal conditions are predominant in July. The thermal regime at this time of the year will therefore favour development of TG forms in contrast to September when the majority of the snowpack is isothermal and ET and MF metamorphism predominates.

Snow strength studies were performed to assess the strength properties of different types of snow layers. Clusters of rimed needles deposited in the early parts of snow storms and occurring at the bottom of newly deposited slabs frequently occur in the lubricating layers of the limited number of fracture line profiles analysed. Shovel test data also revealed thin deposits of rimed needles as being susceptible to easy shear under applied stress. Other crystal types shown to be weak were grains in the early stages of TG metamorphism, and correspond to the subhedral type shown by other workers (Bradley et al., 1977; LaChapelle and Ferguson, 1980; Stethem and Perla, 1980; Adams and Brown, 1982) to be naturally weak in shear. Grains changing to ET from TG in an isothermal snowpack were also revealed as being weak.

Much of the shearing in the shovel tests was associated with MF layers and crusts. Frequently a thin layer of TG crystals in the early stages of metamorphism acted as the lubricating layer. Regular snowpit analysis also revealed the frequent occurrence of TG crystals above and below ice crusts. Shear strength tests showed partially metamorphosed new snow undergoing ET metamorphism to be the weakest of all metamorphosed crystal types in shear. This emphasises the importance of these as potential sliding layers in newly deposited slabs. Snow in the early stages of TG metamorphism was shown to be more susceptible to shear than advanced TG, thus reinforcing the contention that subhedral TG grains are typically weaker than advanced TG grains.

Analysis of snowpack structural properties has shown that the structural properties of fresh snow and that of the existing snowpack play an important role for potential and actual avalanche release in the Craigieburn Range. Three independent snow stratigraphic categories are presented, representing potential sliding surfaces. The identification of these in the snowpack prior to snowfall or in the initial stages of snowfall will allow possible snowpack behaviour to be gauged in periods of slope loading.

Evidence presented in the foregoing analysis and discussion shows that the Craigieburn Range snowpack possesses characteristics, such as the frequent occurrence of MF layers, typical of maritime snow climates. In addition to this though, the extensive development of TG snow in the lower and upper levels of the snowpacks was observed and is more typical of continental snow climates. The occurrence of snowpack failure and development of weaknesses associated with both these snowpack characteristics, typical of quite different snow environments, supports the classification of the Craigieburn Range snowpack as one transitional between a coastal and inland situation as suggested by Owens et al. (1983) and Prowse and Owens (1984).

CHAPTER SEVEN

CONCLUSION

Conclusions discussed at the end of Chapters 3 to 6 are summarised here in point form with discussion devoted to their implications. Suggestions for future research are also discussed.

7.1 SUMMARY OF MAIN FINDINGS AND IMPLICATIONS

7.1.1 Terrain Analysis

(1) Preferential sites for avalanche formation are typified by open bowl shaped slopes in excess of 30 degrees possessing aspects in the range of 35 to 65 and 130 to 150 degrees, positioned downwind or in the lee of a gully or ridge, and occurring above 1600 m. These diagnostic features were used to map areas of potential avalanche release.

(2) Runout zones in the Craigieburn Range displayed a wide range of angles. High angle runouts were associated with short steep valley side paths, often vegetated, while low angle runouts were typical of avalanche paths occupying main valleys or possessing lower track and runout zones in the main valley.

(3) The Craigieburn Range avalanche terrain generally reflects the effect of post glacial fluvial and mass movement processes on highly fractured greywacke in an area of former cirque and limited valley glaciation.

(4) Annual frequencies of avalanches greater than size 2 were shown to be dependent on path size factors especially that of starting zone area.

(5) Theoretical models of Voellmy (1955) and PCM (1980) were found to be generally applicable for estimating avalanche runout distance.

(6) Coefficients for these models are generally lower than those found by other workers reflecting the often rougher and snow free runout zones of the Craigieburn Range.

(7) An empirical model based on the terrain variables starting zone size, curvature of the avalanche path and variability of avalanche track incline was shown to satisfactorily predict avalanche runout distance. Some underprediction occurred for paths with multiple starting zones.

As further development and the extension of existing skifield facilities is planned for the Craigieburn Range, attention should be paid to those areas proposed as likely sites. Slopes identified and mapped as areas of potential and actual avalanche release occur within these areas. LaChapelle (1979) has also noted some existing skifields in the Craigieburn Range as being unsuitably located. If development is to proceed, careful planning will be necessary for the siting of routes of access and skifield facilities, with possible construction of avalanche protection works. With regard to this the New Zealand Mountain Safety Council, Avalanche Committee has compiled guidelines for skifield safety and development (Williams, 1984). This is considered essential if the avalanche hazard is to be minimised from a physical point. Many of the upper slopes of existing skifields possess terrain typical of avalanche slopes. These will therefore represent areas of hazard to skiers following heavy snowfalls or periods of slope loading by wind redistribution of snow if artificial control has not been performed. The failure of the Voellmy model and the setting of exceptionally high coefficients for the PCM model for stopping avalanches on steep heavily vegetated slopes attests to vegetation's role as a dissipator of kinetic energy. Runout zones on these slopes occur on slopes more akin to avalanche track slopes. This bears implications for dwellings presently located immediately downslope from the present runouts as defined by vegetation damage. While some protection is afforded by the vegetation these dwellings are considered to be at risk to damage by avalanche. Removal of any vegetation upslope from these dwellings, or the receipt of larger amounts of snow in the starting zones of these paths will represent conditions under which greater extent of avalanche flow and hence potential damage to buildings will be realised.

7.1.2 The Role of Meteorological Factors in Avalanche Formation

(1) Precipitation and temperature related variables were shown to be important determinants of avalanche occurrence. Threshold precipitation values for avalanche formation in the Craigieburn Range fall in the lower end of the range found for overseas areas. This is a function of the windiness of the Craigieburn avalanche climate.

(2) Periods in which storm precipitation intensities exceeding 0.93 mm hr^{-1} or mean 6 hour precipitation

intensities greater than 1.5 mm hr^{-1} coincident with low or decreasing temperatures and associated with windspeed increases are considered critical for avalanche formation in the Craigieburn Range.

(3) Discriminant function analysis, using the variables, maximum 6 hour precipitation intensity, mean 2 hour temperature, standard deviation of wind direction and global radiation, was found to be a suitable technique for distinguishing between dry, non and wet avalanche days for the study period.

(4) Two types of storm were identified. These were firstly, storms producing the greatest precipitation amounts and intensities with the majority of precipitation from the north-west, accompanied by strong winds, mean storm temperatures just below freezing and large temperature ranges. Secondly those with air flow predominantly from the south-east to south-west direction, with smaller precipitation amounts and low temperatures.

Identification by objective statistical analysis of those meteorological variables, often used in conventionally based avalanche forecasts for distinguishing between dry, non and wet avalanche days confirms the use of these in conventional avalanche forecasting. This may imply that statistical models of avalanche forecasting are not necessary given the accuracy achieved by competent and experienced avalanche forecasters. However, given the difficulty of forecasting mountain weather in New Zealand, and the fact that these forecasts comprise a large part of the information base on which assessments of likely avalanche occurrence are made, it is proposed that a statistically based model will aid avalanche forecasting in the Craigieburn Range. In such a case an objective model would be used to verify or modify assessments based on conventional methods with a specific level of probability being assigned to an event day.

7.1.3 Synoptic Weather Types, Atmospheric Processes and Snow Avalanching

(1) Mixed warm advection/vorticity advection storms characterised by strong frontal development, high rates of ascending motion and thermal advection, combinations of stability and instability associated with inversions in north-westerly air flows, produced the largest amount of precipitation in the Craigieburn Range.

(2) Vorticity advection storms typified by cyclonic circulations developed in flows of cool southerly air on the downstream side of anticyclones lying to the west of New Zealand, were the most frequent storms bringing snow to the Craigieburn Range but were of secondary importance with respect to total storm avalanche frequency.

(3) Phase differences between upper and lower level circulation patterns, high rates of thermal advection, stable atmospheric conditions typical of WA/VA storms and slow moving centres of low pressure have been shown by statistical analysis to be important for heavy snowfall production and hence potential avalanche release in the Craigieburn Range.

Analysis of the movement of frontal and low pressure systems bears implications for time of precipitation arrival and maximum potential for direct action avalanche release in the Craigieburn Range. For WA/VA storms maximum avalanche potential as a result of slope loading exists when the front, typical of these systems, is approaching and passing over the area. Precipitation intensities and amounts reach their highest at this point. Strong cold southerly winds following frontal passage may augment potential release as sintering is hindered owing to snow redistribution at low temperatures. Maximum avalanche potential in VA storms is realised as the leading edge of the cyclonic flow passes over the Craigieburn Range from a westerly direction bring large amounts of precipitation from an easterly direction, or if a cyclonic system intensifies off the east coast of the South Island bringing moist south-easterly to south-westerly air into the region. Strong cold southerly winds often associated with this type of development will be conducive to the development of hard or wind slab. If either of these systems become stalled and movement to the east is restricted by a blocking anticyclone the period of precipitation will be prolonged with large amounts of precipitation produced. In such cases potential for avalanche is likely to be high. Incorporation of those variables identified as important in heavy snowfall production in an avalanche forecasting model such as that presented in this analysis, is considered would markedly increase the accuracy of statistically based avalanche forecasting. This would effectively tie together the use of meteorological variables collected at the local scale, and the information derived from weather forecasts, commonly used in conventional methods of avalanche forecasting, to give a broader based and more effective avalanche forecasting model.

7.1.4 <u>The Role of Snowpack Structure in Snow Avalanching</u>

(1) The seasonal snowpack of the Craigieburn Range has been shown to possess three phases of development, the timing of which will depend on the arrival of the first snow and the onset of general isothermal conditions.

(2) Anisothermal snowpack conditions in July produce temperature gradients large enough for recrystallisation processes in the Craigieburn Range snowpack. Strong positive temperature gradients for up to 72 hours will have implications for direction of net vapour transfer.

(3) Development of TG forms as distinct layers in the mid to upper parts of the snowpack and often occurring in association with MF layers is an important feature of the Craigieburn Range snowpack.

(4) Partially metamorphosed new snow undergoing ET, especially clusters of rimed needles, featured frequently as weak layers in avalanche fracture lines. In addition to this crystal type, TG snow undergoing ET and snow in the early stages of TG were particularly weak in shear. Early TG was shown to be naturally weaker than advanced TG.

(5) Snowpack structural characteristics involved in actual avalanche release, and identification of structural situations typical of both maritime and continental snow climates implies that the Craigieburn avalanche climate resembles one transitional between a coastal and inland continental type.

Development of anisothermal conditions in the snowpack early in the season is likely to bear implications for late season snowpack stability and potential climax avalanche release. Weak basal layers composed of TG and developed in July and early August may collapse under large new snow loads or become susceptible to bond destruction by free water in late August and September when the greatest inputs of precipitation and frequencies of warm north-westerly winds occur. The magnitude of individual avalanches may also be increased markedly if release in the upper parts of the snowpack causes further collapse of weak layers at the base owing to the overburden pressure of avalanching snow. In such cases avalanches may break through to the ground significantly increasing their damage potential.

Development of strong positive temperature gradients in conjunction with strong negative temperature gradients will aid concentration of vapour in the middle parts of the snowpack thus increasing the chances of recrystallisation in this area as a result of mid pack sublimation. Developments of this nature represent potentially unstable situations in the snowpack. Observation of the snowpack during prolonged periods in which there is no signficant new snow increments and clear nights with sub-freezing temperatures prevail, is essential for the detection of mid pack TG forms. Subsequent loading of slopes on which this form has developed may increase the frequency and magnitude of avalanching as a result of both climax and direct action releases. The susceptibility of clusters of rimed needles, produced in the early parts of storms with high vapour pressures and rapid rates of ascent, to shearing bears implications for direct action avalanche forecasting. As the atmospheric conditions conducive to the development of these forms are often maximised in storms possessing well developed fronts or strong moist cyclonic circulations, special attention should be given to crystal types produced in the early stages of these storms.

The confirmation that snow in the early stages of TG is much weaker than advanced TG and the common occurrence of these in the snowpack as distinct layers or in association with MF layers suggests that more attention should be directed to the identification of this crystal type for the assessment of potential avalanche release. This highlights the implication to avalanche forecasting of classifying the Craigieburn snowpack as one transitional between a coastal and inland-continental situation. The joint consideration of meteorological and snowpack data is therefore recommended for efficient avalanche forecasting in the Craigieburn Range. Though few climax avalanches were noted during the study period, it is believed a greater length of avalanche occurrence record will firmly establish climax avalanche release, as a result of latent snowpack instability, as an important feature of the Craigieburn Range avalanche climate.

7.2 SUGGESTIONS FOR FUTURE RESEARCH

Given the limited time period for sampling avalanche events in this study, many areas are apparent where improvements could be made or future research conducted. These will be discussed in the following subsections.

7.2.1 Terrain

For the establishment of terrain avalanche frequency relationships a greater length of record is required. An avalanche observation programme is to be continued in the Craigieburn Range with particular attention being paid to avalanche frequency on specific paths; given this, the extension of the frequency analysis presented in this research is possible. As most field surveying of avalanche paths was confined to skifields, further sampling of other major paths identified during this study and in some cases coincident with areas proposed for future skifield development, is necessary. This would allow additional testing of the theoretical models described in this study and allow other models such as Lang et al. (1979) to be investigated. Artifical avalanche control by avalauncher at Porter Heights skifield presents an opportunity for good empirical data to be collected on avalanche velocity, impact pressures, flow heights and density of flowing snow. This could be used in conjunction with detailed terrain and survey data for establishing accurate kinetic and turbulent coefficients of friction. A useful comparison of actual and predicted avalanche velocities would also be facilitated with an assessment of the influences of the denser snow and rougher avalanche runout of the Craigieburn Range on general avalanche dynamics. In addition more information on flow types is necessary to reinforce or dispute the contention that a flowing regime for the Voellmy analysis be assumed for the majority of paths. Estimates of flow heights on paths with mixed motion avalanches is also needed. No attempts to monitor or calculate impact pressures were made for this study. However this is clearly needed if results are to be used in land use planning. Possibilities of monitoring actual impact pressures exist in the Craigieburn Range as some of the large paths are suitable for construction of structures on which impact pressure plates could be located. This would partially fulfill some of the basic engineering research requirements not yet undertaken for avalanche run-out zones in New Zealand.

7.2.2 Avalanche Meteorology

The avalanche forecasting model presented in this study is based on both a limited avalanche occurrence and meteorological record. This model may therefore be only representative of the years for which it was developed. This was partly vertified by independent testing of the model. A greater length of record is therefore needed to account for year to year fluctuations in weather patterns. In addition to the temporal representativeness of this model is the problem of spatial variability of snowfall and its effect on snow avalanche release. Until recently the number of high elevation meteorological stations in the Craigieburn Range has been limited to one. However, in conjunction with the New Zealand Mountain Safety Council's Pilot Avalanche Forecasting Programme (Fitzharris et al., 1983) the network has expanded to include many skifield locations. This will allow an objective comparison of precipitation amounts to be made for these sites. From this the effect of storm source and related variability of precipitation on avalanche hazard could be established. Though meteorological data collected at the BR site is considered accurate within the range expected of the instruments and harsh conditions, several improvements are possible. As precipitation amounts and intensities have been shown to be of prime importance in avalanche formation the accurate measurement of these is essential. Any subsequent precipitation gauges installed above snowline in the Craigieburn Range should be equipped with Alter snow shields to improve gauge catch efficiencies during periods of strong wind. These in addition to other standard meteorological instruments such as Stevenson screens should be mounted on adjustable height platforms so they can remain free of the snow surface. During the study period the accurate measurement of solid precipitation using snow boards was not possible as sites at which these were located resulted in the snow board being blown clear of snow. For the measurement of solid precipitation and determination of water equivalents well sheltered sites will have to be found for snow board measurements. As there is a distinct lack of such sites at and above 1500 m, artificial enclosures should be built or snow boards located in isolated forest clearings on lower slopes. Wind is also an important factor in the Craigieburn Range snow environment. Lower elevation sites for wind recordings are completely inadequate from an avalanche forecasting point of view as gross discrepancies in wind direction and speed exist between these sites and those at higher elevations. While 800 mb wind directions and speeds have proved satisfactory for determining the effects of wind on avalanche formation and occurrence in this study, wind speeds and directions should be recorded at a ridge top site. However owing to the rigours naturally associated with a mountain snow environment this may not be possible given the lack of a wind recording instrument to weather the effects of riming and exceptionally high windspeeds. For the day to day assessment of avalanche occurrence interrogation of remote stations by telemetry would offer tremendous advantages. This is especially so with regards ridge top windspeeds and precipitation amounts.

In view of the importance of synoptic weather types and atmospheric processes for producing heavy snowfall for avalanche release, evaluation of computer based predictions of vertical motion and thermal structure, currently being used by the New Zealand Meteorological Service for general weather forecasting, would be useful. Such numerical models could easily be applied to a snow forecasting situation. Analysis of individual storms has highlighted the importance of atmospheric structure and processes for snowfall production. As the majority of weather systems affecting the Craigieburn Range come from the west, university based radiosonde studies on the west coast of individual storms with possible Meteorological Service assistance should be encouraged. From this a useful comparison would also be facilitated between soundings on a west and east coast situation with the affect of weather modification by the Southern Alps being established.

7.2.3 Snowpack Studies

The role of snowpack structure in avalanching is an important one with many potential areas of study possible. As much artificial avalanche control is practised in the Craigieburn Range, ample opportunity exists for trained skifield personnel to collect a large amount of fracture line information. As found elsewhere, new snow types were the weakest of all crystal types and appeared frequently in the fracture profiles recorded. However, only qualitative observations of new snow crystal types were made in this study. This also applies to other studies dealing with snow in New Zealand. Though it is expected that new snow crystal types will not be very variable in the New Zealand snow environment given the narrow range of snow storm temperatures, the degree of riming is expected to be quite variable as a result of stability contrasts between and within storms. A co-ordinated study of detailed snow crystal indentification by microphotography or crystal replication with upper atmospheric thermal and moisture information in conjunction with detailed storm plots would yield useful information on processes involved in crystal riming. This would bear special significance to avalanche studies as it has been shown that rimed crystal forms are important in avalanche formation and release.

In association with detailed storm plots and fracture line profile analyses shear frame tests could be conducted on weak layers to establish any significant crystal type-strength relationships, with strength-stress ratios also being established. This would allow some comparison with snow stability indices produced for the Mt Cook area (Conway and Abrahamson, in press). The analysis of snowpack temperatures is based on a limited record for a shallow snowpack. Ideally a complete seasonal study should be undertaken of snow metamorphism as it relates to the thermal regime of the snowpack with progression of the season. From this the three phase model proposed for snowpack development between June and late September may be verified or modified.

The frequent occurrence and shearing of early TG forms associated with MF layers in the Craigieburn Range snowpack requires further investigation. Specifically this would involve an analysis of the effect of MF layers on vapour flow. Though more suited to a laboratory study extensive field checking of situations in which TG occurs in association with relatively impermeable MF layers would help direct any laboratory based investigation of these processes. The effects of positive temperature gradients on vapour flow could also be established in such a study.

The classification of the Craigieburn Range avalanche climate as one transitional between a coastal and inland-continental one presents possibilities for a much wider snow climate classification study. For the Canterbury-West Coast regions the spatial extent of snow structure characteristics found for the Craigieburn Range needs to be delimited. This would involve a consideration of the east to west trend in snowpack depth, structure and thermal regime. At some point east of the main divide a limit to the eastern-inland snow region as proposed by Owens et al. (1983) would be defined. Characteristics of avalanching west of this are expected to differ markedly to that of the Craigieburn Range. If such a supposition is true factors used in avalanche forecasting will vary accordingly, similar to that identified by LaChapelle (1966) for the United States.

7.2.4 Wet Avalanches

Attention in this analysis has been mainly directed at dry avalanches. A more detailed study of snowpack and weather conditions leading to wet avalanche release would not only yield interesting results about this infrequently studied phenomenon, but would also have implications to snow hydrology as wet avalanche release is often coincident with periods of snowmelt. In addition the role of melt freeze layers in inhibiting vertical flow of water in the snowpack and causing bond breakdown needs to be assessed in conjunction with a general consideration of the effects of rain on snow metamorphism, especially that of temperature gradient metamorphism. Though there was not a pronounced seasonal trend in wet avalanche activity for the Craigieburn Range during the study period, wet avalanches occurring in the spring have the potential to do a large amount of geomorphic work on slopes on which they occur. The significance of these as active geomorphic agents also needs to be assessed further.

7.2.5 Future Avalanche Research in New Zealand

At present in New Zealand no central government department is responsible for snow research, of which snow avalanches comprises only a small part. Currently the only hope of furthering our understanding of the general New Zealand snow environment and avalanche formation in general lies mainly with the universities, though recently the Forest Research Institute has shown some interest. These institutions in conjunction with the Avalanche Committee of the New Zealand Mountain Safety Council appear to be the only bodies willing to establish and direct research on snow avalanches and snow in general in New Zealand and have been instrumental in setting up a pilot forecasting scheme for the study area. This in conjunction with other forecasting schemes set up in the Fiordland, Mt Cook and Tongariro National Park areas under the auspices of the New Zealand

Mountain Safety Council, Avalanche Committee represent a vast improvement of the avalanche data base for New Zealand. A seminar is to be convened by this Committee to evaluate these forecasting schemes and identify any modifications required.

The main objective of this study was to consider the role of terrain, meteorological and snowpack factors in contributing to avalanche occurrence in the Craigieburn Range. In addition to the work by Morris and O'Loughlin (1965), McCracken (1980), Prowse (1981), Owens et al. (1983), Fitzharris et al. (1983), Weir (1983), Owens and Prowse (1984), this research contributes to the general understanding of the snow environment processes of the Craigieburn Range. Though mainly of a reconnaissance nature this study has identified areas of research to which future attention should be directed. As the Craigieburn Range only represents one small area of New Zealand's alpine regions, similar studies to this one are recommended for other mountainous areas. Such studies in addition to the specific research needs identified here will yield profitable results for the general understanding of snow avalanches and will set a suitable context for more detailed investigation of snow avalanche phenomena in New Zealand.

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APPENDICES

APPENDIX 1

WIND VELOCITY SYMBOLS (ms^{-1})


0 - 5


6 - 10


11 - 15


16 - 20


21 - 25


26 - 30


31 - 35


36 - 40

APPENDIX 2

SNOWPIT ANALYSIS

2a	Procedure
2b	Snow Crystal Type Symbols
2c	Hardness

APPENDIX 2a : PROCEDURE

Following pit excavation the wall of the snowpit was brushed as an aid for identifying contrasts between any snow layers. Height above ground level (AGL) of these were recorded. Individual layers were analysed as follows:

Density - a 0.1 m^3 sampler was used to sample density. Once samples had been extracted they were immediately weighted on a Yamato 0-0.1 kg letter scale. Readings were obtained down to an accuracy of 1 g. Sampling of consolidated snow proved no major problem but that of highly crystalline temperature gradient layers often proved difficult due to total disaggregation of the layer. Density values are presented in $kg.m^{-3}$.

Temperature - the normal procedure for determining snow temperature at depth within the snowpack is by placing bimetallic stem thermometers at 100 mm intervals down the snow column. These have an approximate measurement accuracy of $\pm 0.25^oC$. Initially used in snowpit analysis these were subsequently replaced with a Digitron digital probe thermometer. The digital display gave temperatures down to a scale of 0.1^oC with an accuracy of ±0.2% of 0.5^oC. This instrument was particularly useful due to the rapidity with which temperatures could be taken. It also has a great advantage for determining small scale temperature differentials, especially in and around ice crusts.

Crystal Size and Type - a Pentax monocular (8 x 30 magnification) was used in determining crystal size and type. This monocular had a built-in 0.1 mm scale grid which made it particularly useful for determining accurate crystal sizes; a problem often confronted when using a hand lense and coarse standard 1.0 mm grid. Crystal types were identified and classified according to the procedure of the International Snow Classification (UNESCO/IASH/WMO 1970) as appears in Perla and Martinelli (1976). The literary translation of these symbols appears in Appendix 2b.

Hardness - this was evaluated by two methods. Resistance to horizontal penetration and hence a measure of horizontal compressive strength was made by the standard hand test as outlined by Perla and Martinelli (1976) (Appendix 2c). Resistance to vertical penetration and hence a measure of the vertical distribution of compressive strength through the snow column was estimated using the ramsonde penetrometer. A detailed description of the instrument is given by Bader et al. (1939). Though the precise physical interpretation of the meaning of the ram number is unclear, clear relationships have been obtained between ram number and unconfined compressive strength (Abele, 1974) and density (Weir and Owens, 1981). The main problem associated with the ramsonde penetrometer is its insensitivity to low density weak layers. This is especially true of new snow. Keeler and Weeks (1969) also noted the insensitivity of the ram in the range of low ram numbers less than 3.

APPENDIX 2b : SNOW CRYSTAL TYPE SYMBOLS

+

Freshly deposited snow, initial forms can be easily recognised.

Partially metamorphosed new snow retaining some of its initial form but definited reduction in convexities on crystals. Structure often felt like.

Early stages of ET metamorphism. Crystals have lost the majority of their initial form, but may show signs of elongation. Rounding process definitely in progress with overall decrease in grain size. These grains are commonly less than 0.5 mm.

●

Advanced stages of ET metamorphism. Grains definitely rounded with all convexities gone with elongation absent. Definite bonds developed between grains. Grains increasing in size attaining a size no greater than 2 mm.

○

Grains formed by MF metamorphism with definite welded bond structure. Grains usually greater than 1 mm.

□	Early stages of TG metamorphism. Grains show initial faceting with grains being angular with flat sides or faces. Grains less than 1 mm.
∧	Angular grains with abundance of stepped faces, initial ET, MF or new snow form completely unrecognisable. Presence of hollow cups. Structure very sugar like. Grain size greater than 1 mm and increasing.
∿∿	Sun or rain crust, less than 10 mm in thickness.
▬	Ice layer greater than 10 mm in thickness and composed of MF grains. Usually dense and very solid.
→	In the process of actively changing forms from one type to another.
—	A mixture of two grain types.

APPENDIX 2c : HARDNESS

SYMBOL RANGE	HAND TEST	RAM NUMBER (kg)
VSS	Fist	0 - 2
VS		
MSS	Four fingers	3 - 15
MS		
M	One finger	26 - 50
MH		
MHH	Pencil	51 - 100
H		
HH	Knife	Over 100
VH		
VHH	Thick Ice	-

In the hand test the specified object is pushed into the snow with a force of about 5 kg. A gloved hand is used in the first three levels. The symbol range refers to the ease by which the object enters the snow.

APPENDIX 3

AVALANCHE RECORD SHEET

(Adapted from Armstrong and Ives 1976)

Department of Geography

University of Canterbury

Avalanche Record Sheet
(Adapted from Armstrong & Ives 1976)

Skifield:

Observer:

Date of Avalanche: (Actual or Estimate) Time:

Path Data

Name of Avalanche Path: ______________________

Aspect of Slope: ______________________

Slope Angle: ______________________

Wind Direction: Cloud Type:

% Cloud Cover:

Weather Conditions Before Release (Rain, Snow, Wind, Temperature etc.)

Circle Where Appropriate

Avalanche Morphology: (See additional notes)	Hard Slab Soft Slab Wet Slab Loose Wet Loose
Type of Release: (See additional notes)	Direct Action Climax
Trigger:	Natural Artificial Skier Explosive Avalauncher

Size of Release: Magnitude in brackets (See additional notes)	Slough (1) Small (2) Medium (3) Large (4) Major (5)
Running Surface:	On old surface in starting zone Ran to ground in starting zone
Motion: (See addition notes)	Sliding Flowing or tumbling Mixed motion
Layers:	Only New Snow Includes older snow layers
Height of Fracture Line	____________ m
Percent of Path Effected	____________ %
Starting Zone:	Area When Viewed From Below Top Middle Bottom Left Centre Right

---

Vertical Fall	____________ m
Where Debris Stopped:	Fracture or Starting Zone Transition or Bench Partway Down Track Bottom of Track or Runout Zone
Debris Depth at Centreline	____________ m

Length of Centre Line (Slope distance) __________ m

Additional Comments (observers opinion on cause of avalanche etc.)

Additional Notes

Avalanche Morphology

Hard Slab:- Slab avalanche starts at a line. Broken snow layer is medium hard to very hard and of high density. A hard slab preserves chunks or blocks over longer avalanche paths.

Soft Slab:- Slab avalanche start at a line. Broken snow layer is very soft and of low density. Slab disintegrates into loose material immediately after start.

Wet Slab:- Avalanche debris is blocky and usually containing liquid water at the time of deposition.

Loose:- The loose avalanche starts at a point. Triggering at the point may be caused by snowfall from a cornice or rock overhang, rock or stonefall, or any other falling object, or by a skier.

Wet Loose:- Same as loose form above but avalanche debris containing liquid water at time of deposition.

Types of Release

Direct Action:- This type falls during or within 24 hours after a storm, and involves only the snow of that storm at the release point (LaChapelle, 1966).

Climax:- This type falls as the result of internal structural weaknesses within the snowcover which may develop over long periods of time. It may be triggered by a new snowfall but involves snow layers at the release point deposited by more than one storm.(LaChapelle, 1966).

Size of Release

Classification of size is based on the volume of snow avalanche relative to path size and potential damage. This is a combination of the methods used by the United States Forest Service and recommended by the National Research Council of Canada.

Slough (1) - This is any slide running less than 150 feet or 50 m slope distance regardless of avalanche path dimensions and generally harmless to humans and structures.

Small (2) - Harmful to humans but not structures.

Medium (3) - Harmful to humans and could damage structures.

Large (4) - Probably damage large structures.

Major or Maximum (5) - Would devastate buildings and general structures.

Motion

If avalanche is actually observed, the motion will be one of the following types:

Sliding:- occurs when snow breaks loose and moves downslope without rolling or tumbling.

Flowing or Tumbling:- snow whether granular or in blocks, moves along the snow or ground surface in a rolling turbulent motion.

Mixed Motion:- Mixed airborne and ground motion.

APPENDIX 4

MONTHLY AVALANCHE AND WEATHER OBSERVATION SHEETS

(a) July, 1979

(b) August, 1979

(c) September, 1979

(d) June, 1980

(e) July, 1980

(f) August, 1980

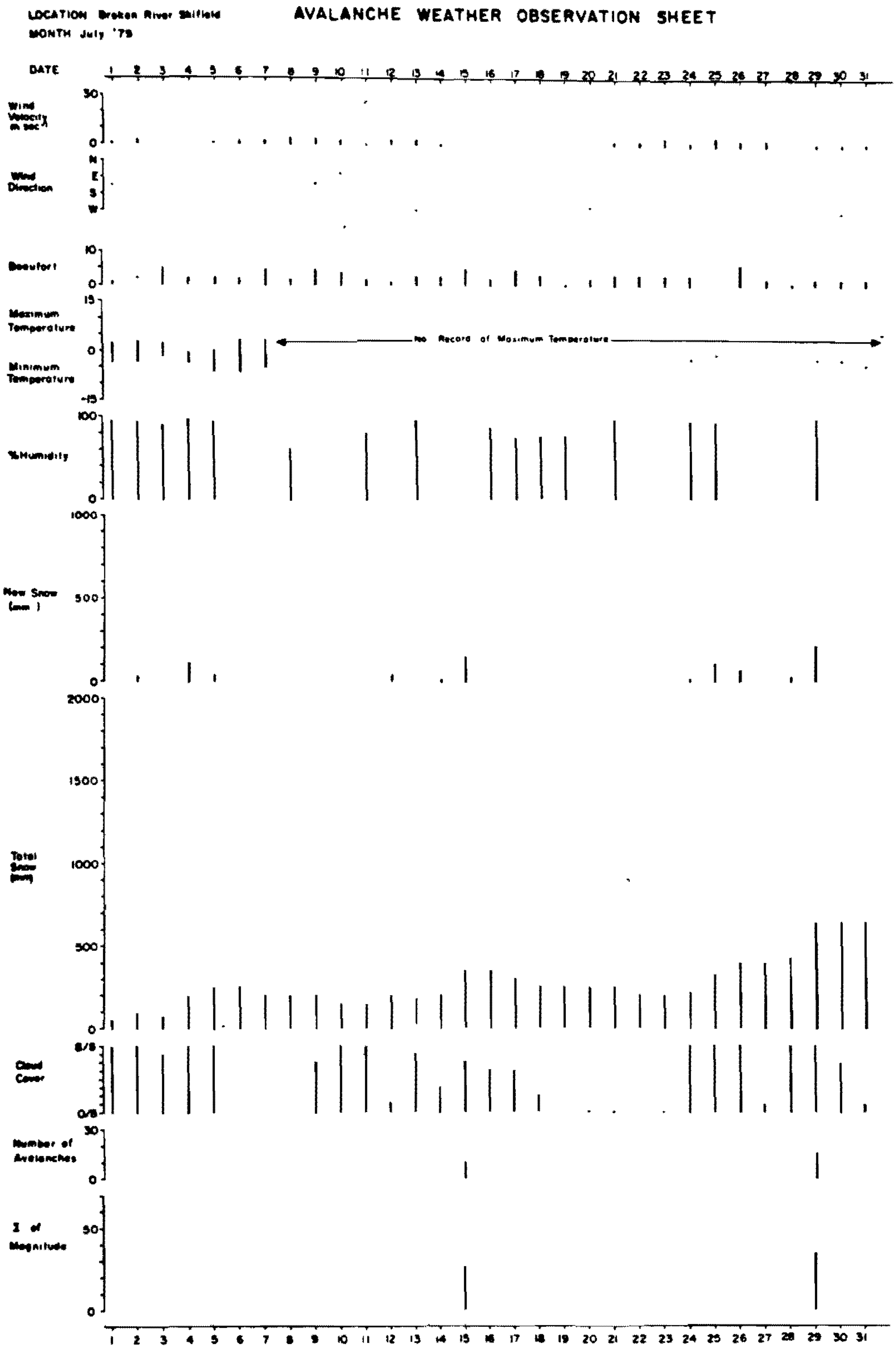
(g) July, 1981

(h) August, 1981

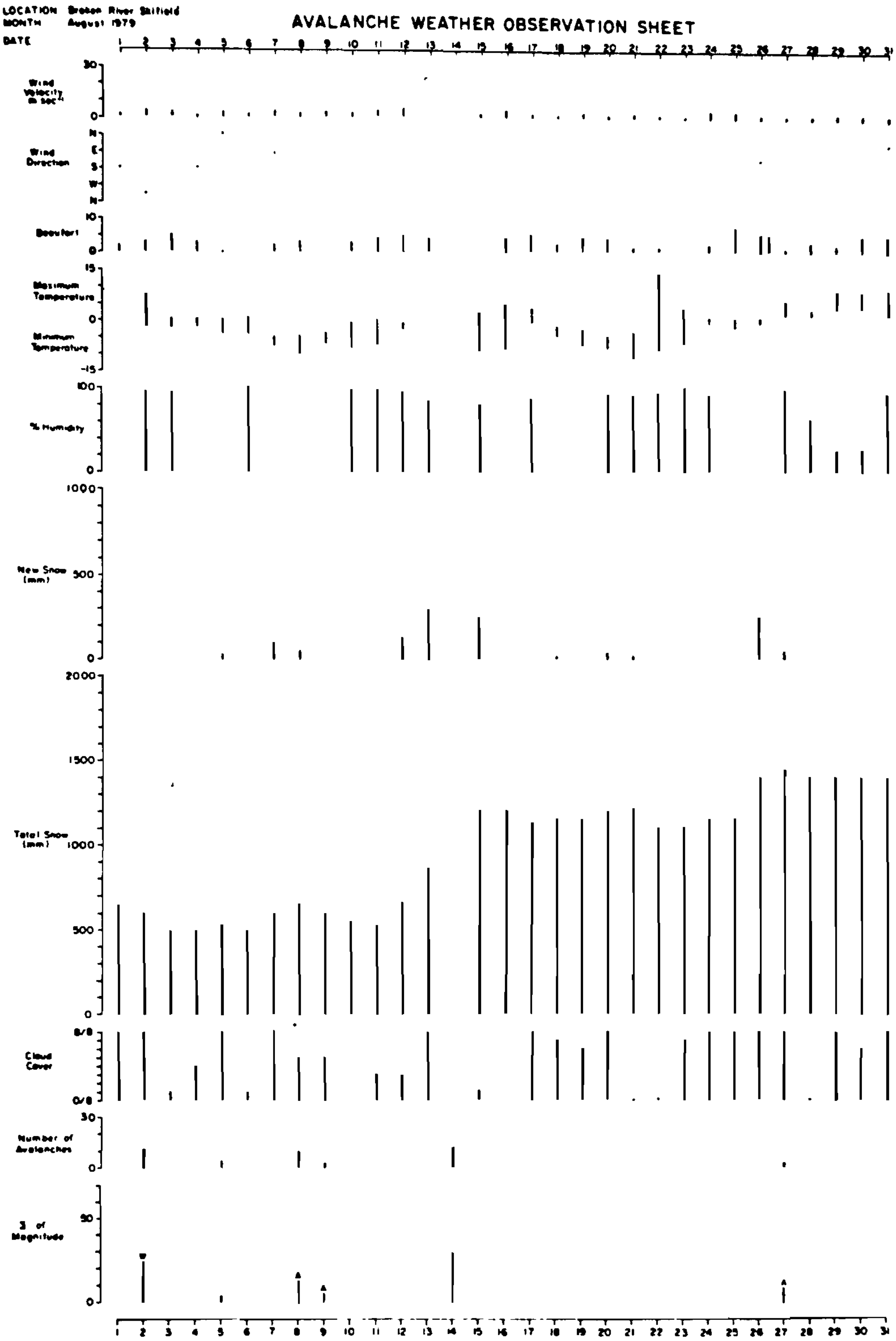
(i) September, 1981

Gaps in humidity record represent days on which no recording was available.

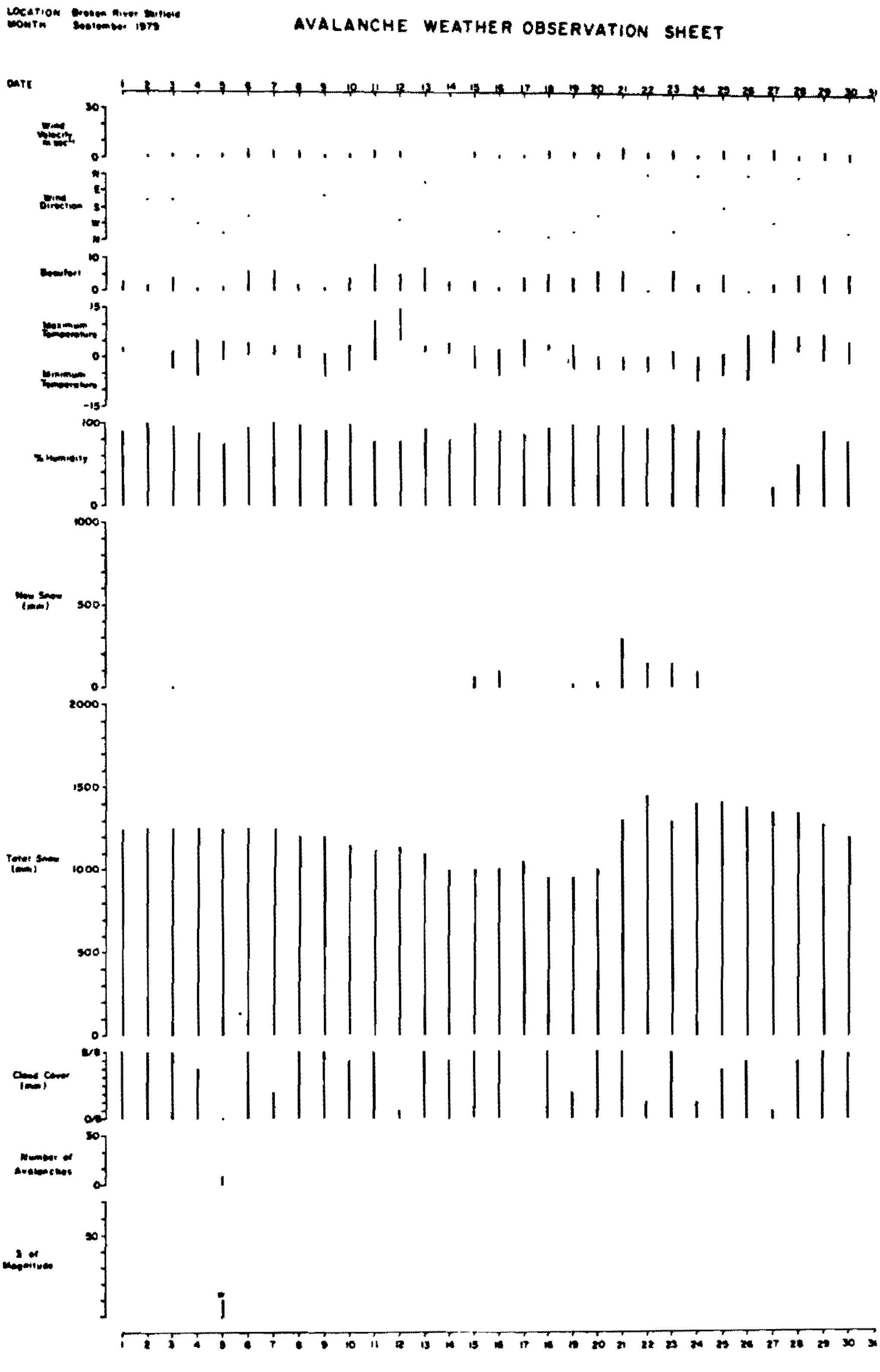
a


LOCATION Broken River Skifield
MONTH July '79
AVALANCHE WEATHER OBSERVATION SHEET
DATE
Wind Velocity m sec⁻¹
Wind Direction
Beaufort
Maximum Temperature
Minimum Temperature
No Record of Maximum Temperature
%Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
Number of Avalanches
Σ of Magnitude

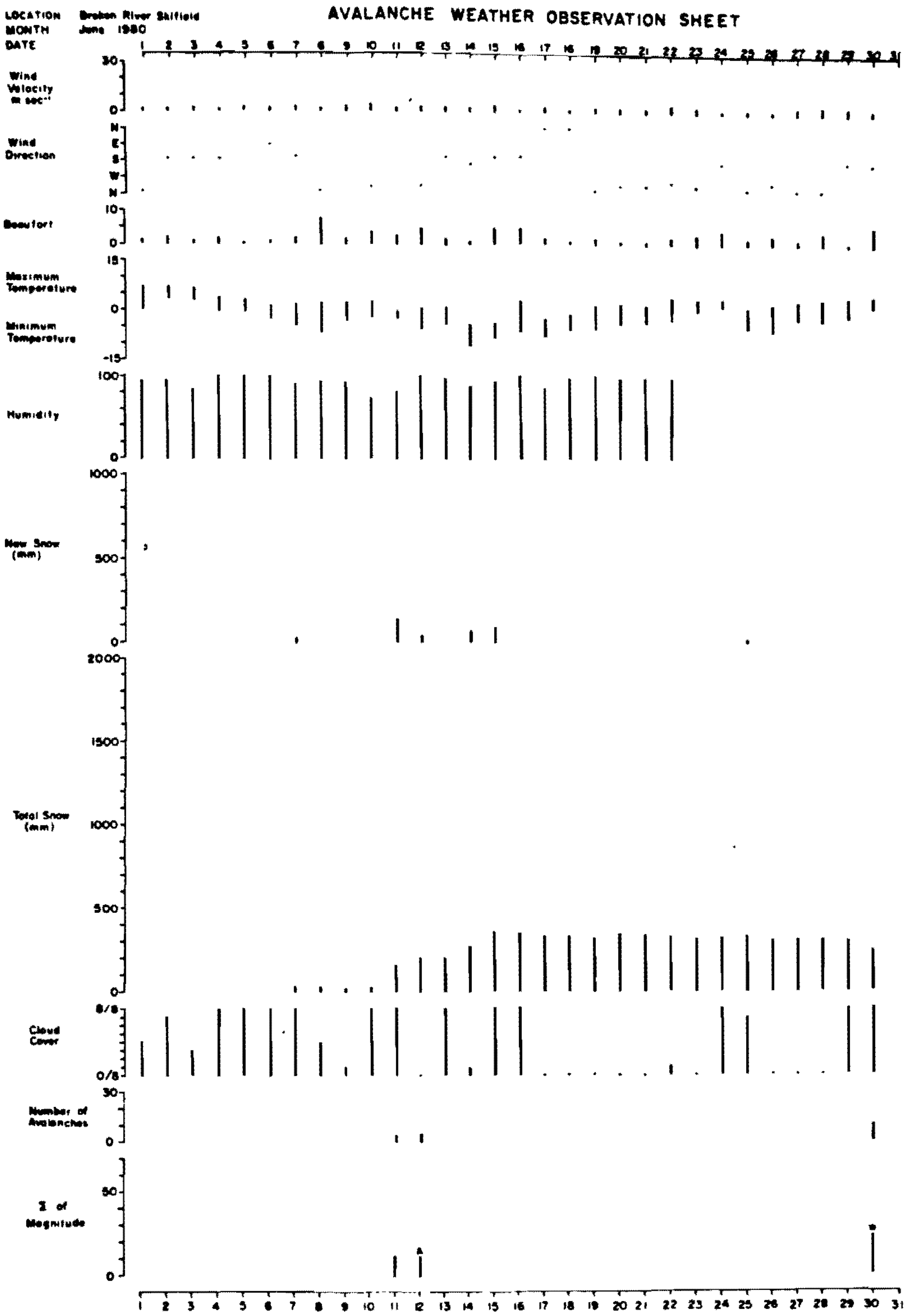
b


LOCATION Broken River Skifield
MONTH August 1979
AVALANCHE WEATHER OBSERVATION SHEET
DATE
Wind Velocity
Wind Direction
Beaufort
Maximum Temperature
Minimum Temperature
% Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
Number of Avalanches
Magnitude

C

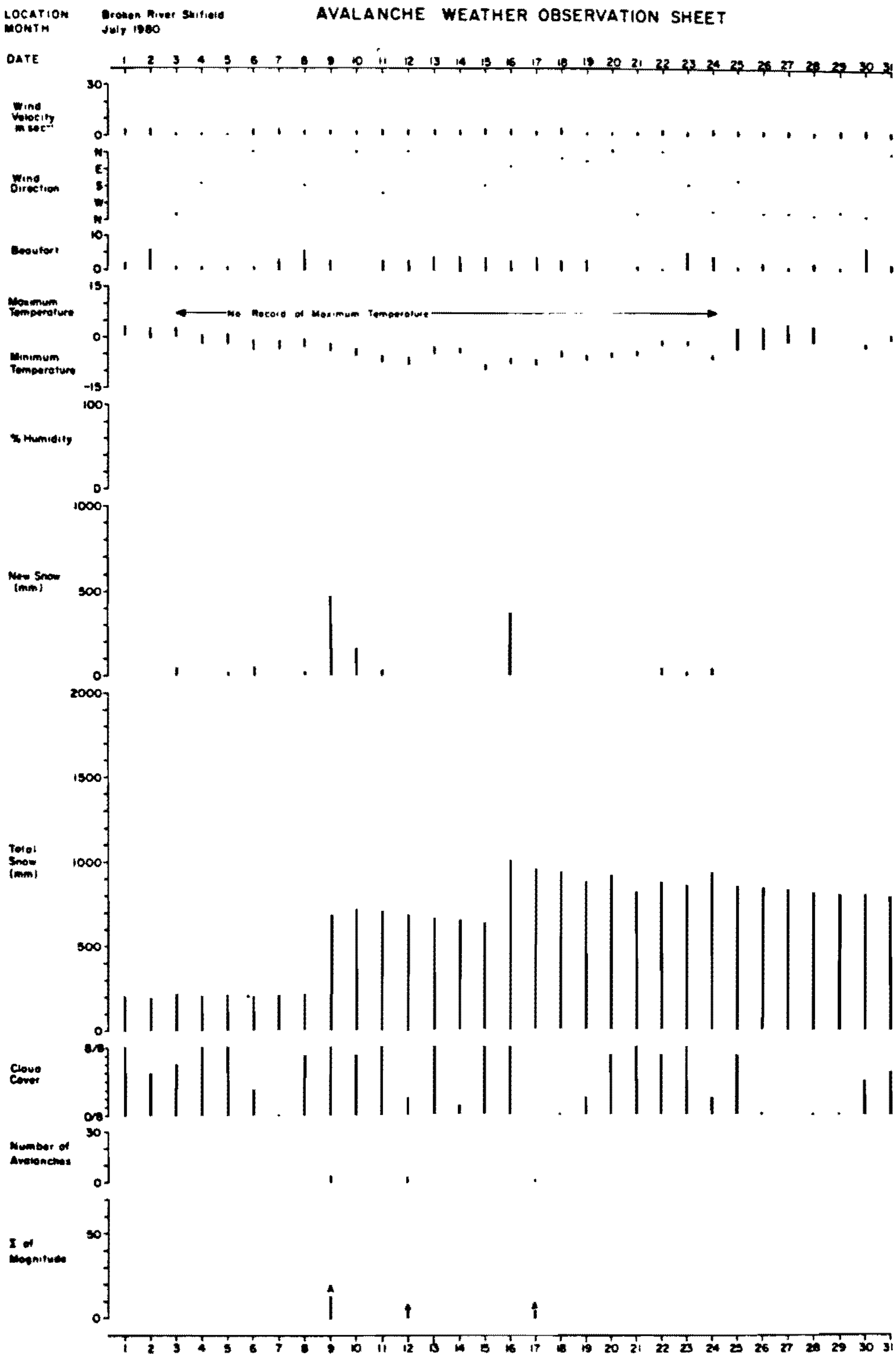

LOCATION
MONTH
AVALANCHE WEATHER OBSERVATION SHEET
DATE
Beaufort
% Humidity
0
500
1000
1500
2000
50

d

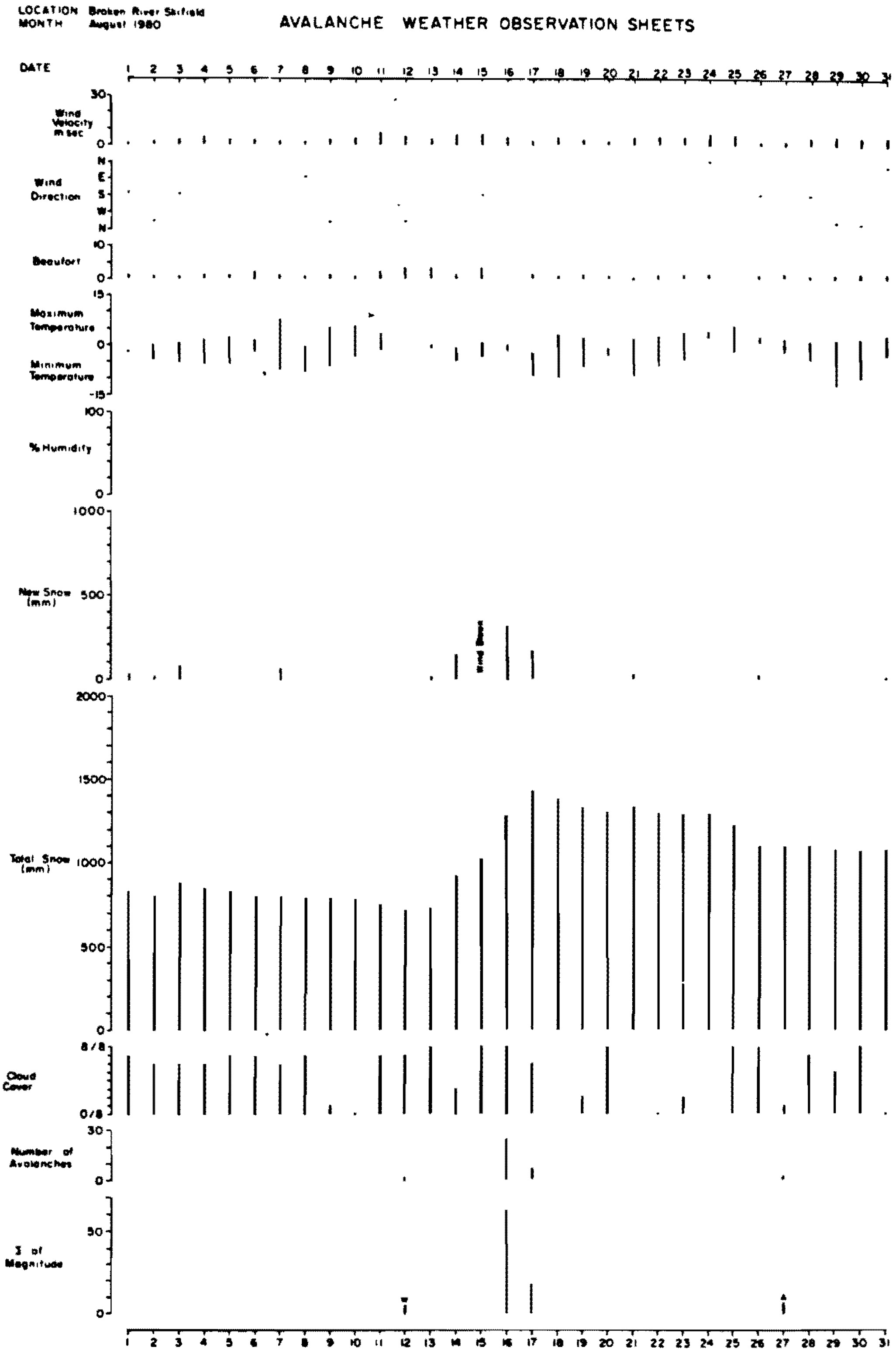


AVALANCHE WEATHER OBSERVATION SHEET
LOCATION Broken River Skifield
MONTH June 1980
DATE
Wind Velocity m sec⁻¹
Wind Direction
Beaufort
Maximum Temperature
Minimum Temperature
Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
Number of Avalanches
Σ of Magnitude

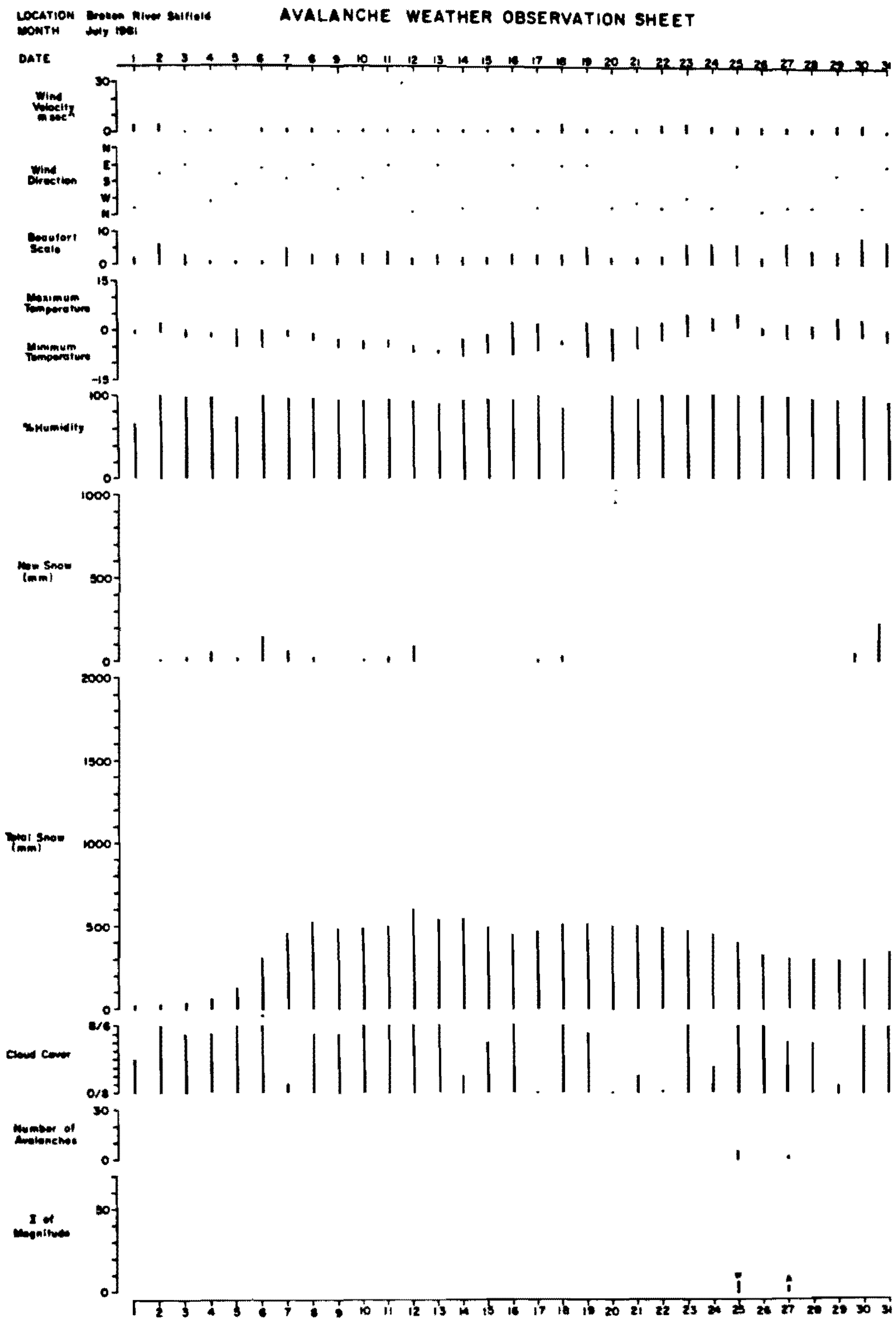
e


LOCATION Broken River Skifield
MONTH July 1980
AVALANCHE WEATHER OBSERVATION SHEET
DATE
Wind Velocity m sec⁻¹
Wind Direction
Beaufort
Maximum Temperature
No Record of Maximum Temperature
Minimum Temperature
% Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
Number of Avalanches
Σ of Magnitude

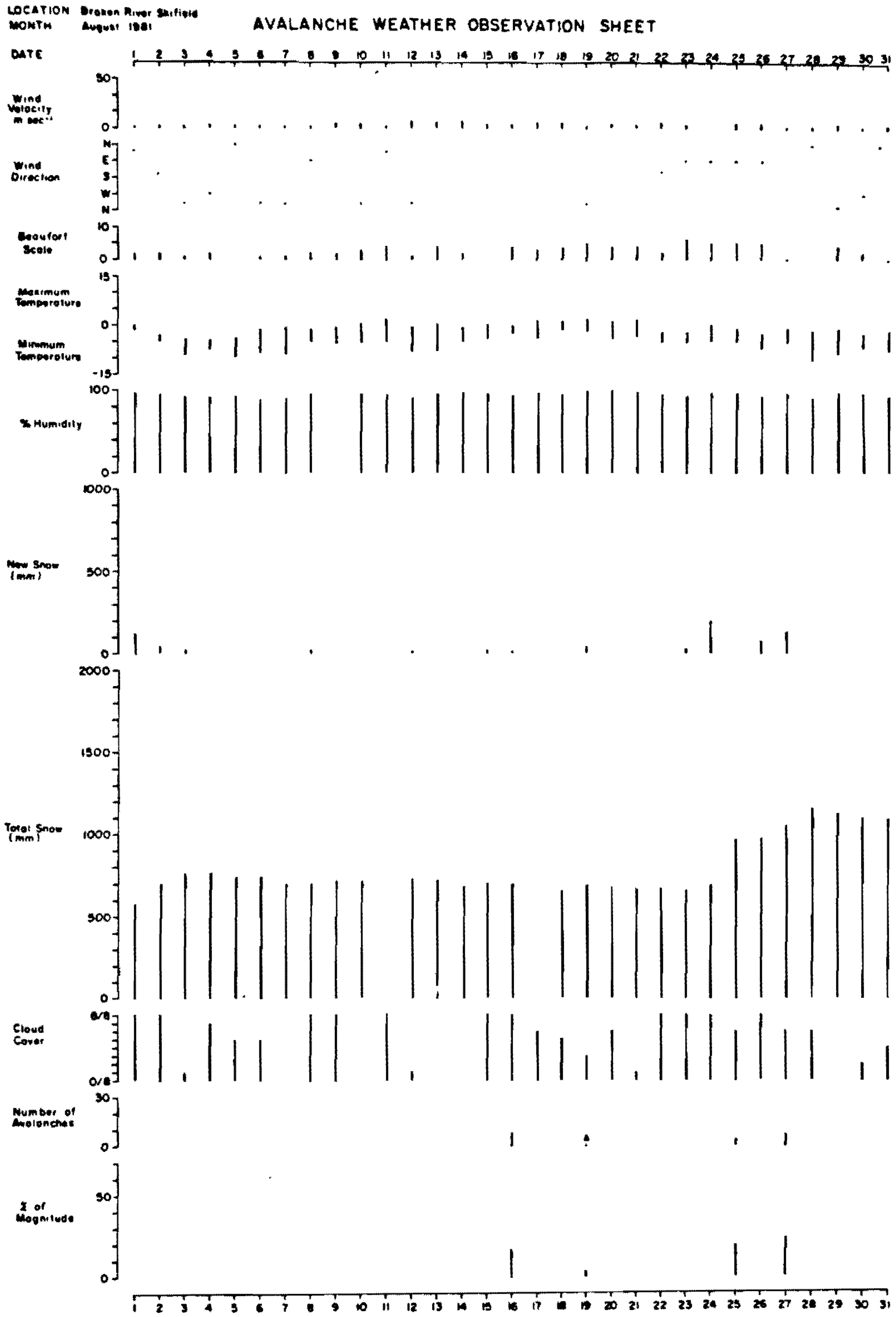
f


LOCATION Broken River Skifield
MONTH August 1980
AVALANCHE WEATHER OBSERVATION SHEETS
DATE
Wind Velocity m sec
Wind Direction
N
E
S
W
N
Beaufort
Maximum Temperature
Minimum Temperature
% Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
8/8
0/8
Number of Avalanches
Σ of Magnitude

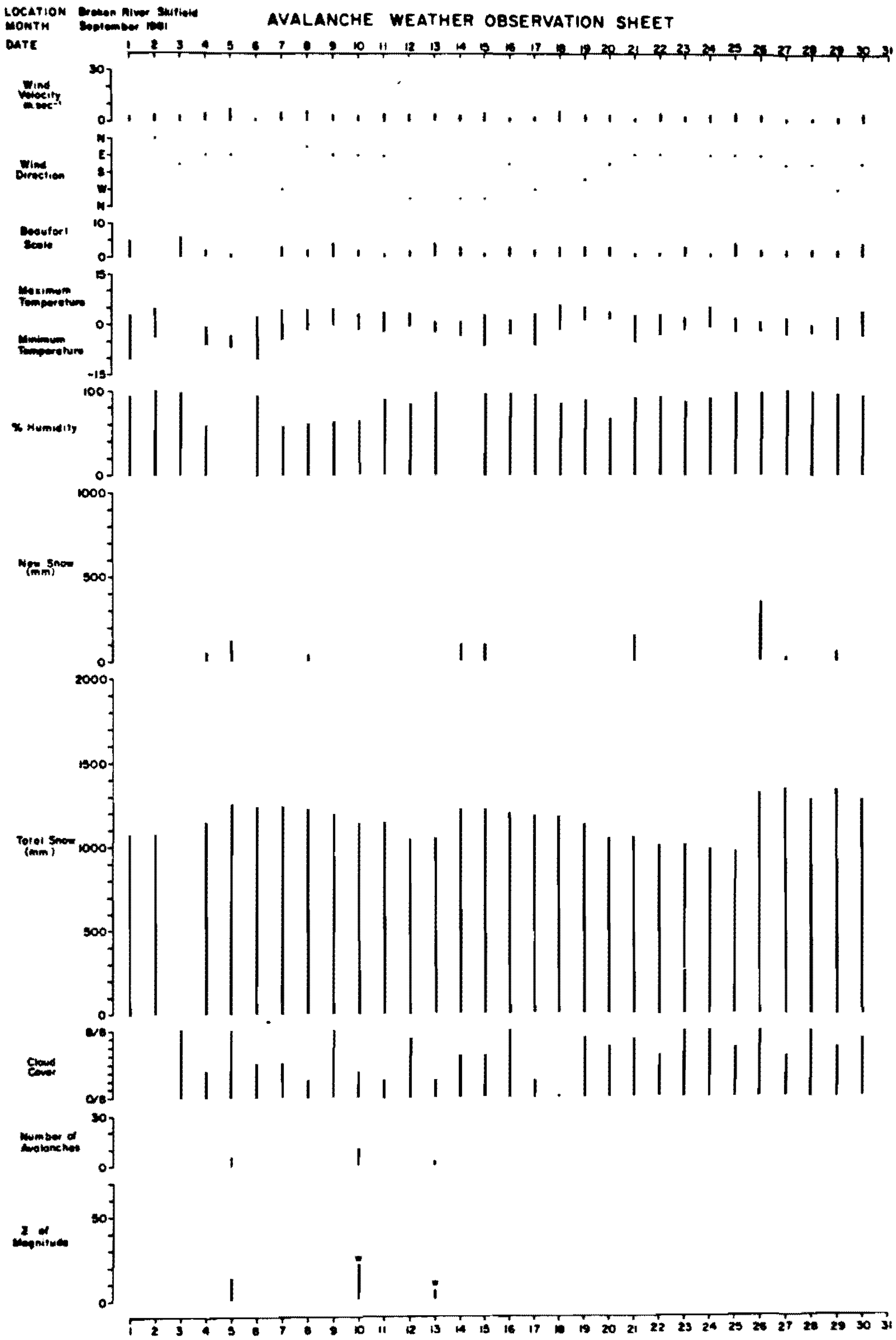
g


LOCATION Broken River Skifield
MONTH July 1981
AVALANCHE WEATHER OBSERVATION SHEET
DATE
Wind Velocity m sec⁻¹
Wind Direction
Beaufort Scale
Maximum Temperature
Minimum Temperature
%Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
Number of Avalanches
Σ of Magnitude

h


LOCATION Broken River Skifield
MONTH August 1981
AVALANCHE WEATHER OBSERVATION SHEET
DATE
Wind Velocity m sec⁻¹
Wind Direction
Beaufort Scale
Maximum Temperature
Minimum Temperature
% Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
Number of Avalanches
Σ of Magnitude

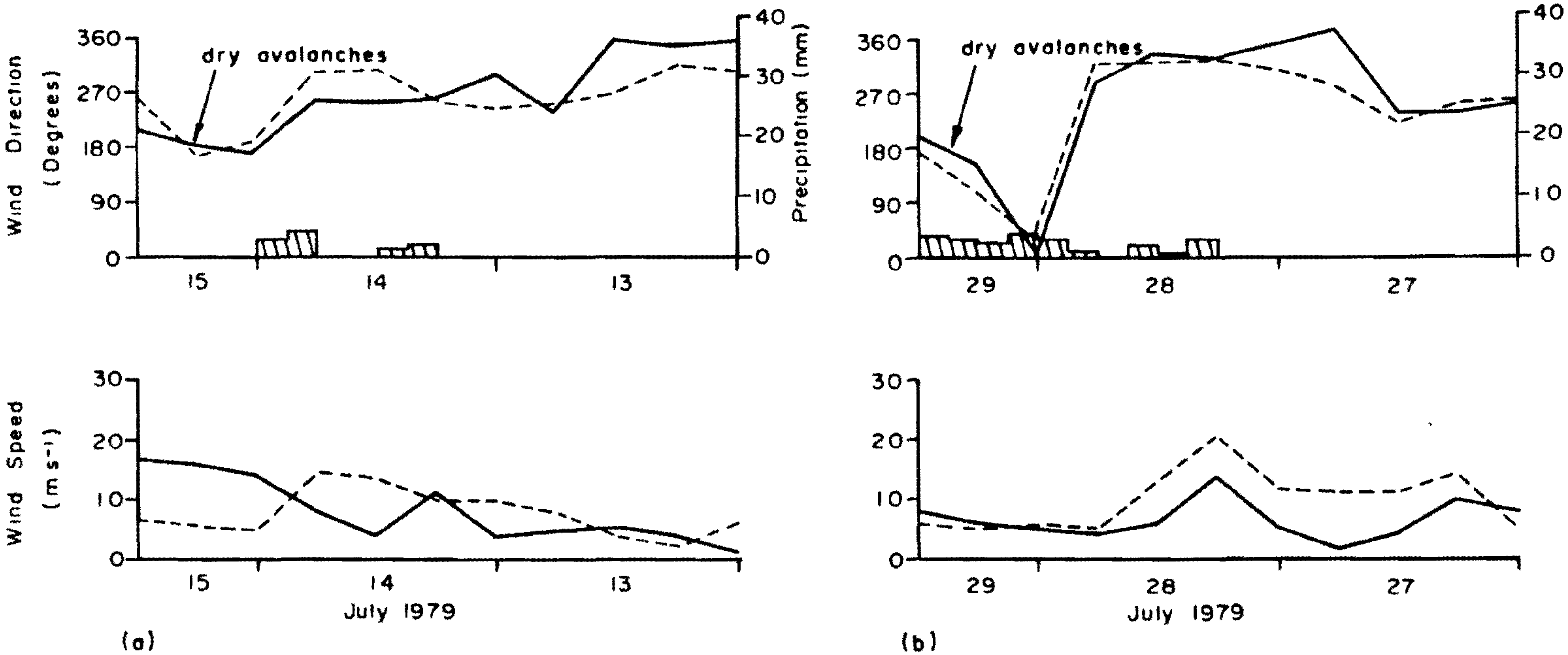
i


LOCATION Broken River Skifield
MONTH September 1981
AVALANCHE WEATHER OBSERVATION SHEET
DATE
Wind Velocity m.sec⁻¹
Wind Direction
Beaufort Scale
Maximum Temperature
Minimum Temperature
% Humidity
New Snow (mm)
Total Snow (mm)
Cloud Cover
Number of Avalanches
Σ of Magnitude

APPENDIX 5

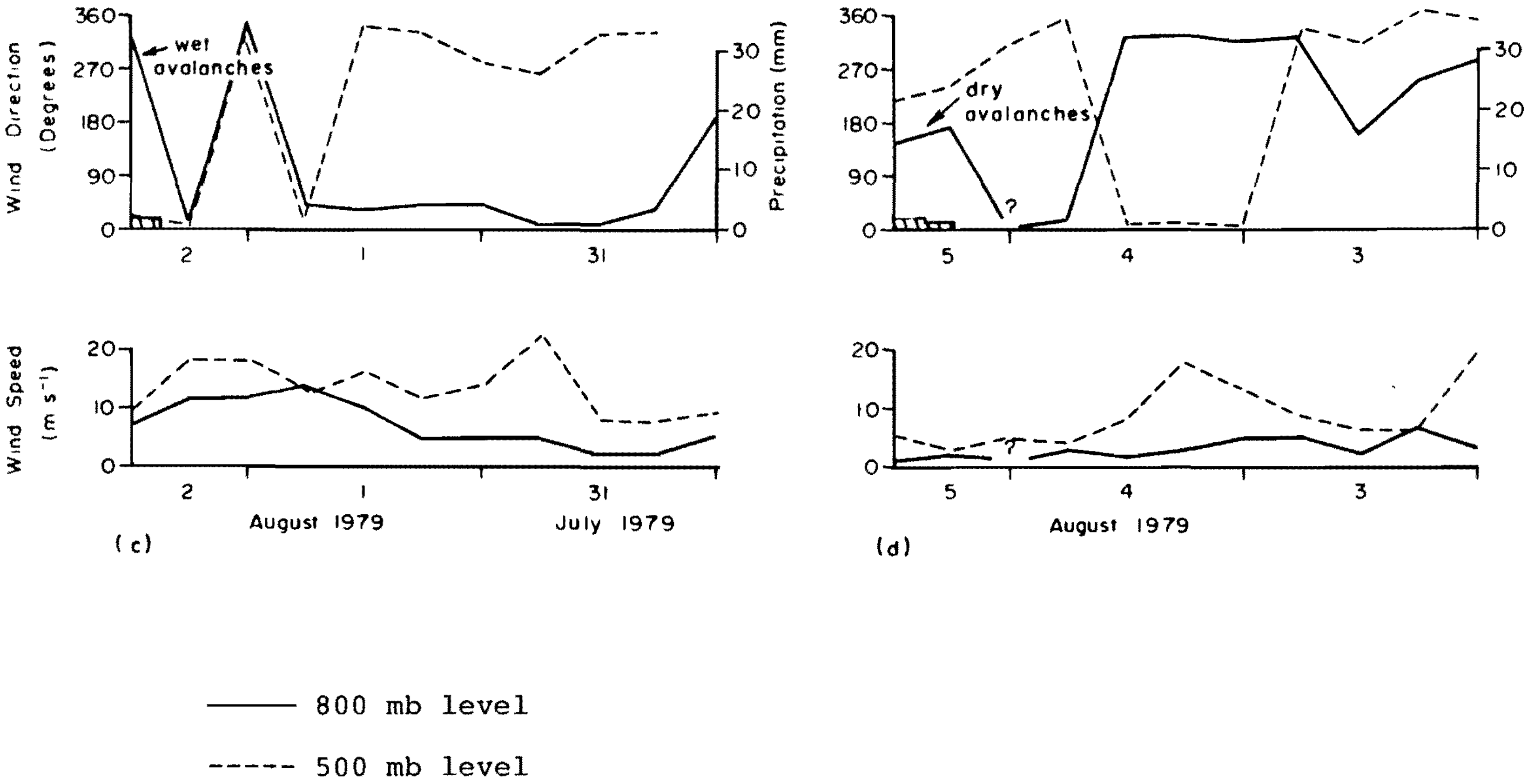
WIND DIRECTION/SPEED AND PRECIPITATION PLOTS AND TIME OF AVALANCHE RELEASE

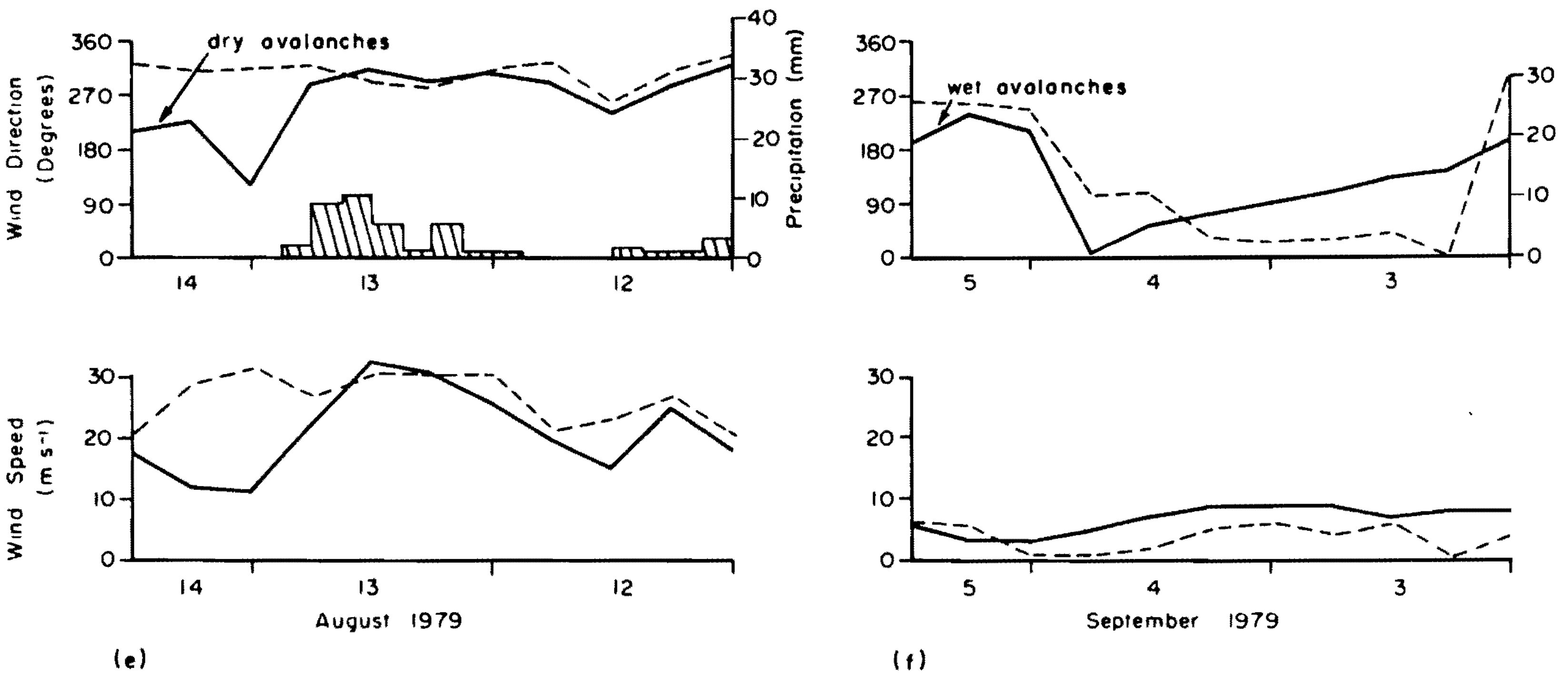
(a) 13-15 July, 1979	(dry avalanches)
(b) 27-29 July, 1979	(dry avalanches)
(c) 31 July, 1979 - 2 August, 1979	(wet avalanches)
(d) 3-5 August, 1979	(dry avalanches)
(e) 12-14 August, 1979	(dry avalanches)
(f) 3-5 September, 1979	(wet avalanches)
(g) 9-11 June, 1980	(dry avalanches)
(h) 28-30 June, 1980	(wet avalanches)
(i) 10-12 August, 1980	(wet avalanches)
(j) 14-16 August, 1980	(dry avalanches)
(k) 23-25 July, 1981	(wet avalanches)
(l) 14-16 August, 1981	(dry avalanches)
(m) 23-27 August, 1981	(dry avalanches)
(n) 3-5 September, 1981	(dry avalanches)
(o) 9-13 September, 1981	(wet avalanches)



——— 800 mb level

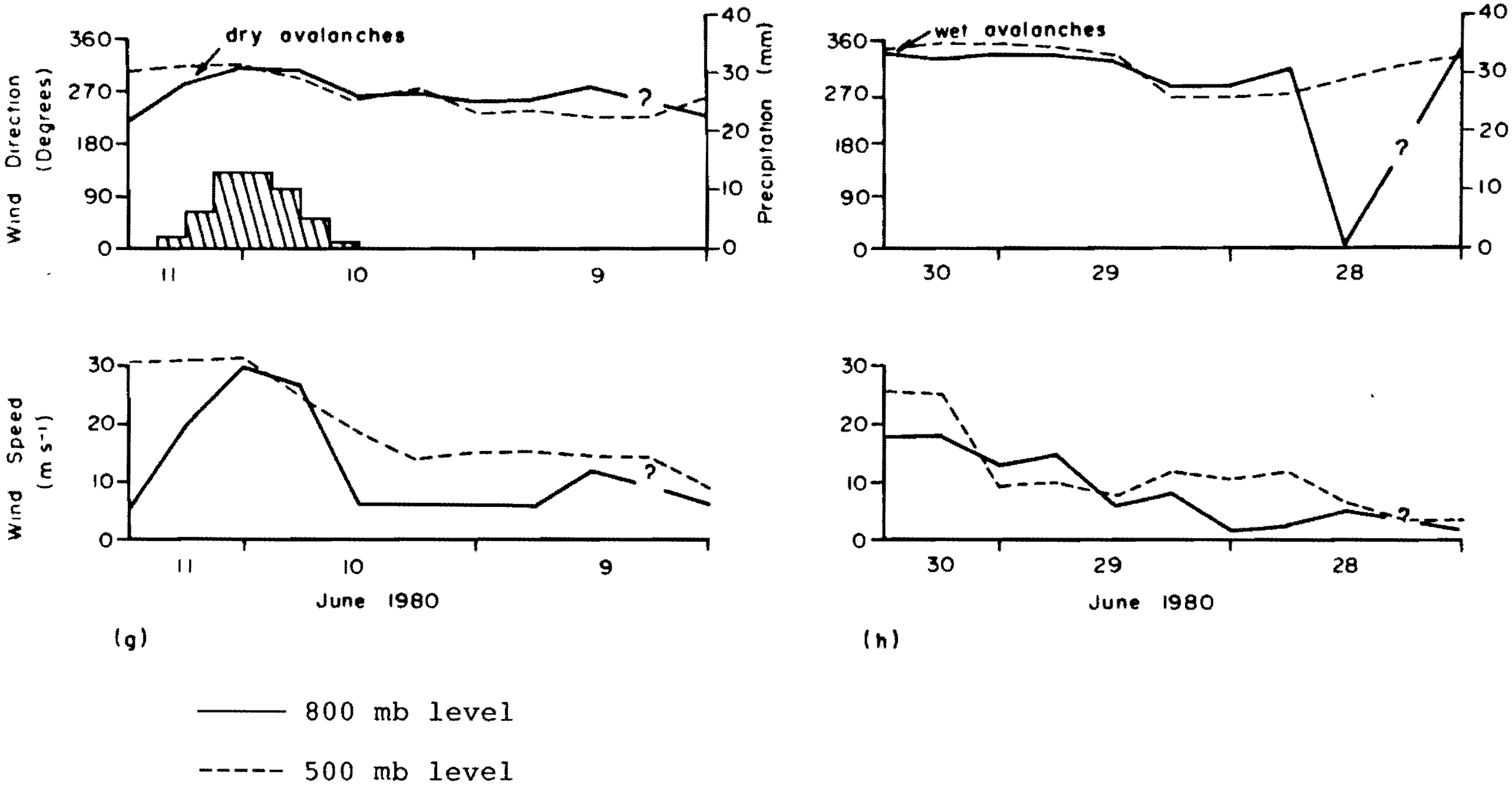
----- 500 mb level

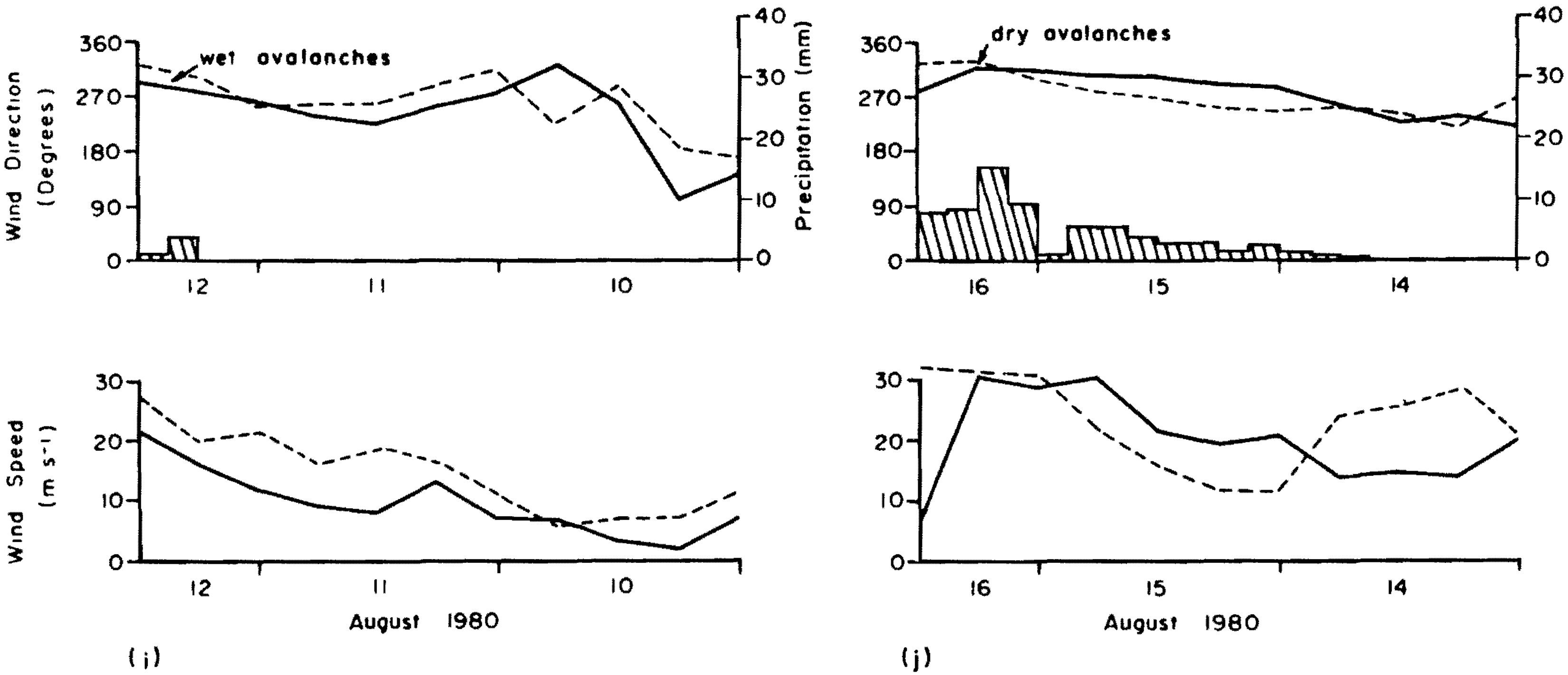

Wind Direction (Degrees)
Precipitation (mm)
Wind Speed (m s⁻¹)
wet avalanches
dry avalanches
?
360
270
180
90
0
30
20
10
2
1
31
5
4
3
August 1979
July 1979
(c)
(d)
800 mb level
500 mb level



——— 800 mb level

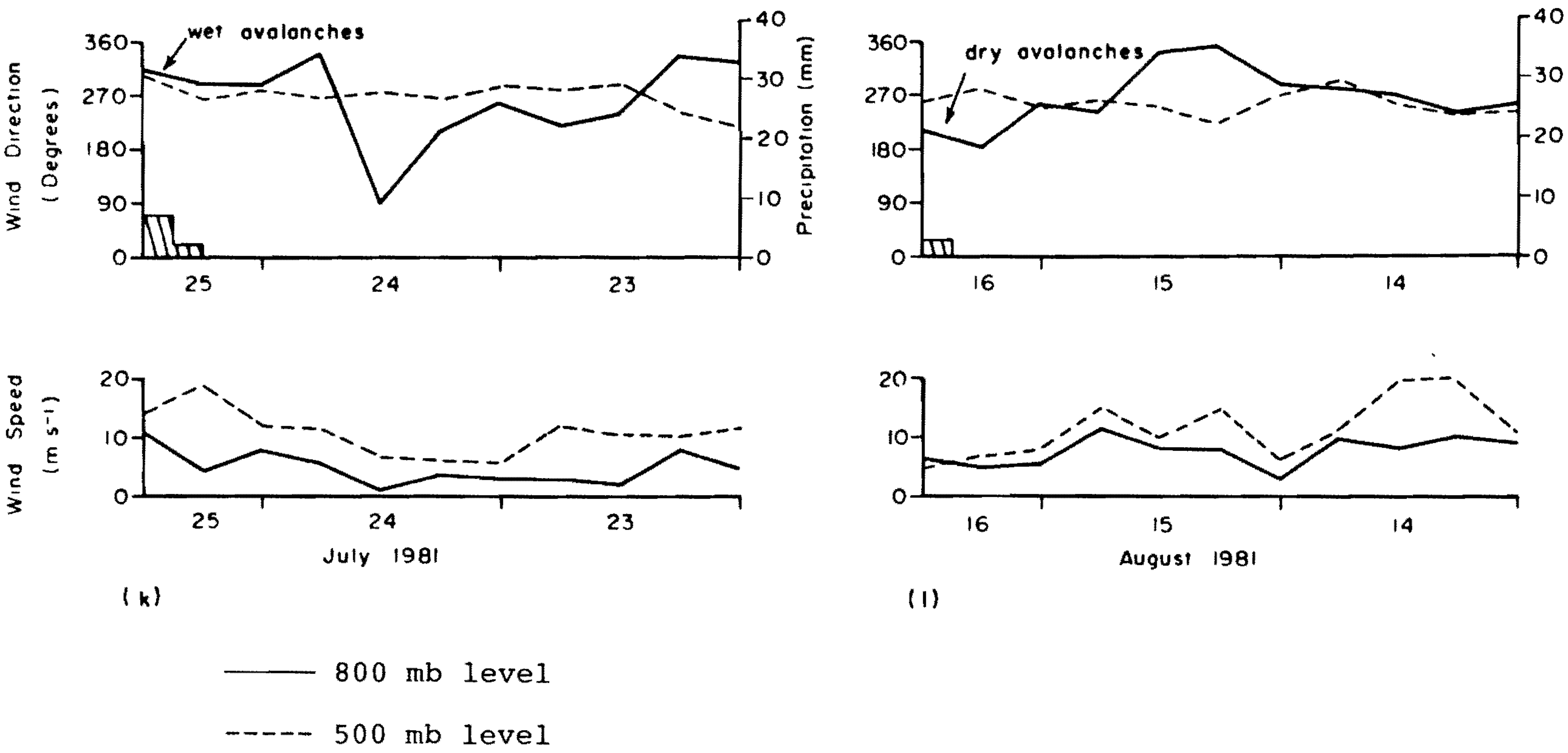
- - - - 500 mb level

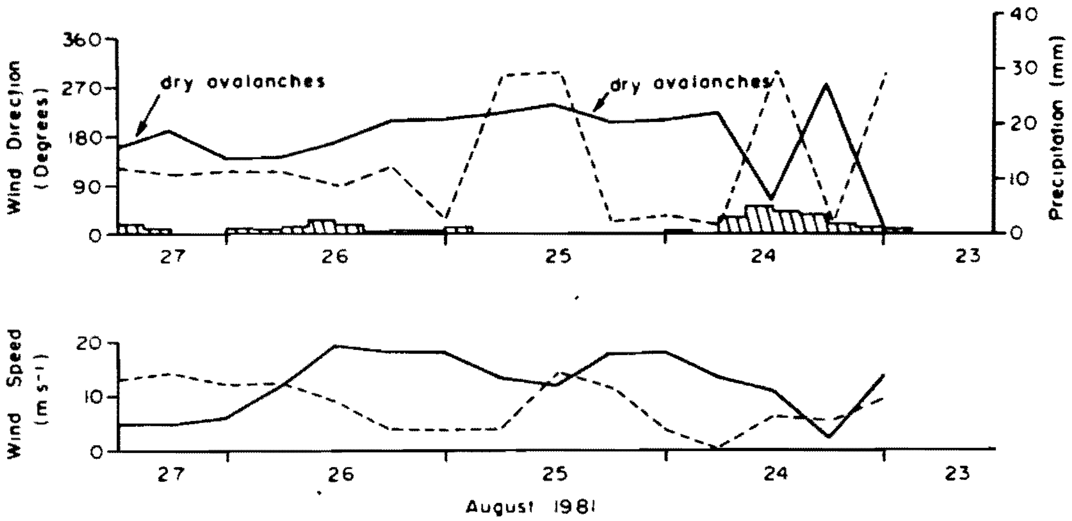

dry avalanches
wet avalanches
Wind Direction (Degrees)
Precipitation (mm)
Wind Speed (m s-1)
360
270
180
90
0
40
30
20
10
11
10
9
30
29
28
June 1980
?
(g)
(h)
800 mb level
500 mb level



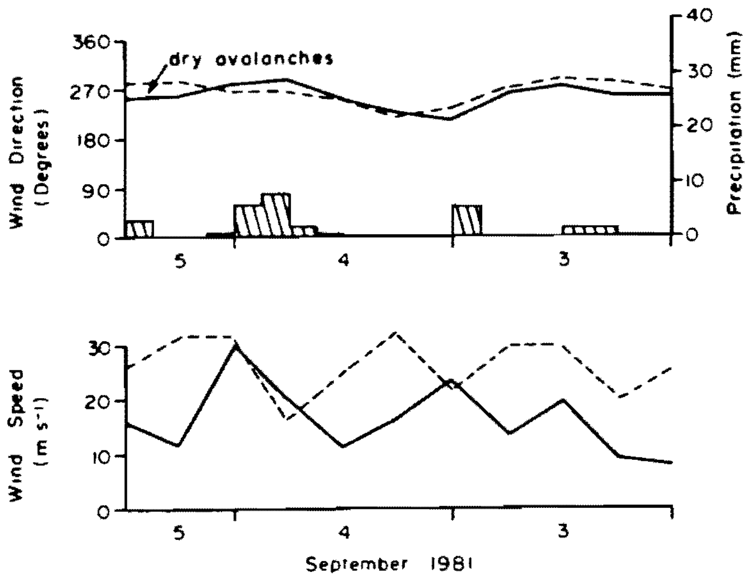
——— 800 mb level

- - - - - 500 mb level

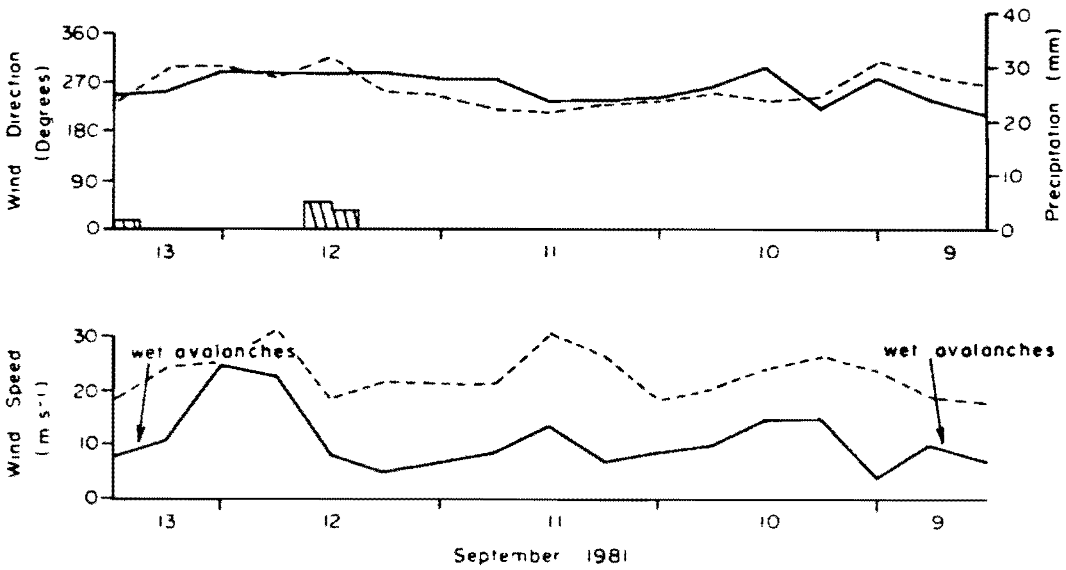

wet avalanches
dry avalanches
Wind Direction (Degrees)
Precipitation (mm)
Wind Speed (m s⁻¹)
July 1981
August 1981
(k)
(l)
800 mb level
500 mb level



(m)



(n)



(o)

——— 800 mb level

----- 500 mb level

APPENDIX 6

PROCEDURE FOR CALCULATING AVALANCHE NON-AVALANCHE DAY PROBABILITIES (BOVIS, 1976) USING NORMAL DEVIATES OF FUNCTION ONE SCORES OF TWO FUNCTION DISCRIMINANT MODEL

If u_1 and u_2 are the mean scores of the non-avalanche and avalanche groups respectively (Figure 6.a), then the avalanche normal deviate for p_2 is

$$\Delta = (p_2 - u_2) = (1 - a_3) \tag{1}$$

which is the probability of avalanche group membership for p_2. The difference $(p_1 - u_1)$ yields a negative number, since the avalanche probability for p_1 is less than 50

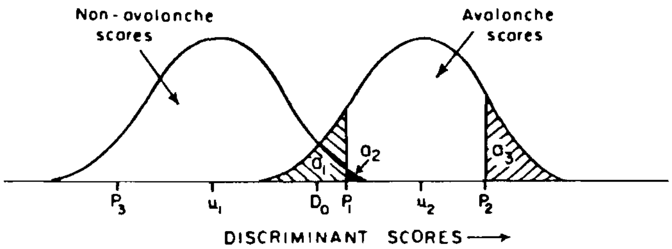


Figure 6.a

percent. The method for computing non-avalanche day probabilities differs slightly from the above since non-avalanche probabilities increase to the left of D_0. Therefore non-avalanche normal deviates are found from

$$\Delta^1 = -(p_1 - u_1) = a_2 \tag{2}$$

This ensures that points falling to the left of u_1 are positive deviates. From Figure 6.a, point p_2 would have a large negative Δ^1 score, indicating a very small probability of belonging to the non-avalanche group, whereas point p_3 has a high probability for this group. In practice, the user need not be concerned with areas under each curve since probabilities can be read from a table of the standard normal distribution once the appropriate normal deviates have been computed.

Once the data for the disciminant model has been collected the following steps should be carried out.

(1) Compute the discriminant score for function one by producing the sum of the products of the coefficients presented in Table 4.6 for function one and the real time variables.

(2) Find the difference between the discriminant score and the appropriate avalanche day group mean score (Table 6.a) with the appropriate avalanche and non-avalanche deviates being calculated. Bovis (1976) notes that particular attention should be paid to the size of the difference as this will establish whether probability is less than or greater than 50 percent.

Table 6.a: Mean Function One Discriminant Scores For The Three Avalanche Day Groups

Group	Mean
Dry Avalanche Day	2.679
Wet Avalanche Day	-1.711
Non-Avalanche Day	0.987

(3) Using the deviates calculated above the probabilities pertaining to the deviates can be read from a table of the standard normal distribution.

(4) Once the above has been established a forecast may be issued.

To illustrate the above, a worked example is presented with the raw meteorological variables and related discriminant score presented in Table 6.b.

Table 6.b: Raw Variable Values and Discriminant Score

	Variable				Function One
	PI	TEMP	SD.WD	RAD	Discriminant Score
Value	7.0	-4.8	35.3	1.30	1.987

Firstly the dry avalanche non-avalanche day deviates are estimated from (1) and (2) above such that:

$$\begin{aligned} \Delta &= 1.987 - 2.679 \\ &= -0.692 \end{aligned}$$

Given the negative sign, the probability is less than 50 percent. This day has a 24 percent probability of being classified as an avalanche day.

For the non-avalanche deviate:

$$\begin{aligned} \Delta^1 &= -(1.987 - 0.987) \\ &= -(1.00) \end{aligned}$$

This is equivalent to a 15 percent probability of being a non-avalanche day.

On the basis of percentage probabilities this case would be classified as an avalanche day. Whether it is dry or wet avalanche day depends on the calculation of the wet avalanche deviate:

$$\begin{aligned} \Delta^1 &= -(1.987 - -1.711) \\ &= -(3.698) \end{aligned}$$

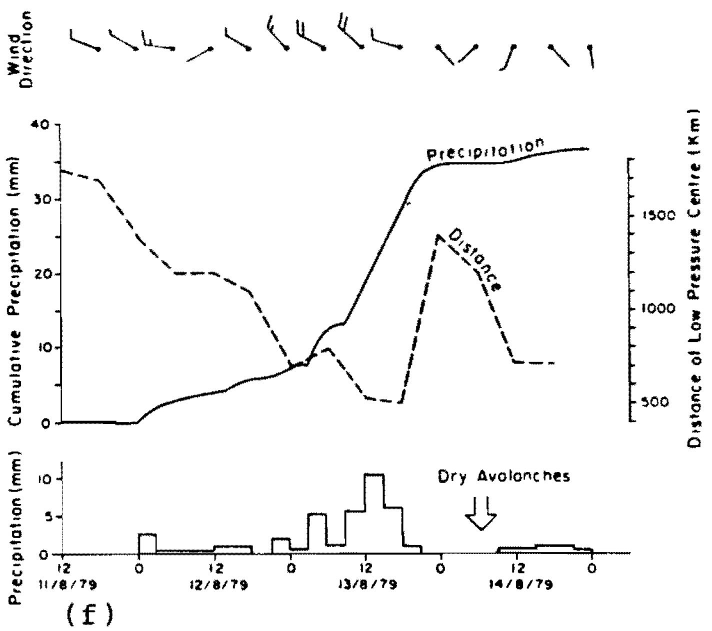
This value gives a probability of less than 1 percent of being a wet avalanche day. The day is therefore classified as a dry avalanche day.

If the procedure outlined in Figure 4.8 is to be used, the following avalanche day mean discriminant scores and disciminant functions are to be used. The same procedure is to be followed as outlined above with the appropriate discriminant function being used per day type discrimination. (Table 6.c)

Table 6.c: Day Mean Discriminant Scores and Discriminant Equations

Day Type Discrimination	Day Mean Discriminant Score		Discriminant Function Equation
Dry - Non	2.34	-1.57	4.2
Wet - Dry	-1.73	3.21	4.3
Non - Wet	1.68	1.52	4.4

APPENDIX 7.1 : DEVELOPMENT OF STORM 13 TO 14 AUGUST, 1979



(f)

(a) A centre of low pressure was located immediately off the south-west coast of the South Island. Associated with this was an occluded portion as cool southerly air in addition to modified cool northerly air replaced warmer northerly air at the centre. Precipitation began to fall steadily in the Craigieburn Range at this point. (f)

(b) An upper level depression at the 500 mb level had developed in association with the lower level low (a above).

(c) Replacement of warm north-westerly air being advected over the South Island by cool southerly air lead to cold front formation. A low pressure centre also developed south-east of the South Island with the inflow of cool southerly air.

(d) Precipitation had begun again (f) as the cold front passed over the South Island and the depression to the north-west intensified.

(e) Warm north-west air continued to be advected over the South Island at upper levels.

APPENDIX 7.2 : DEVELOPMENT OF STORM 15 TO 16 AUGUST, 1980

(a) A low pressure approached the South Island rapidly from a south-west direction (Figure 5.10). A strong north-westerly flow persisted over the South Island. A cell of high pressure remained to the north-east of New Zealand.

(b) A well developed centre of low pressure lay immediately to the south of the South Island. Associated with this was a well developed cold front and trough. Strong north-westerly conditions prevailed.

(c) The trough of low pressure still remained over the South Island with low pressure centres north-west and south-east of the country.

(d) Associated with the development of the lower level trough (a) was that of an upper level trough, the axis of which lay to the west of the lower level trough.

(e) Precipitation in the Craigieburn Range decreased markedly (Figure 5.10) as the downstream side of the upper level trough passed over the South Island.

APPENDIX 7.3 : SINGLE FRONTAL PASSAGE AND PROGRESSIVE AVALANCHE RELEASE THE STORM OF 13 TO 14 AUGUST, 1979

Treble Cone Skifield - Wanaka (W)

13/8/79 (1200 hrs) - a cold front approached the Treble Cone skifield area at Wanaka. Precipitation at Wanaka was reported from the north-west at this point.

(1800 hrs) - the cold front had passed over the Wanaka area. South to south-westerly winds predominanted. Avalanching was first reported at 1500 hrs on 13/8/79 at about the time the front passed over the area. Avalanching at Treble Cone continued with the flow of cool southerly air into the area.

Craigieburn Range Skifields (C)

13/8/79 (1200-1500 hrs) - precipitation reached a maximum as a warm front lay north of the Craigieburn Range. A cold front also lay south of Wanaka at this stage. Strong north-west flow predominated between these two fronts.

14/8/79 (0000 hrs) - the warm front had decayed with a cold front lying over the Craigieburn Range. Winds backed to the south to south-west.

(0600 hrs) - the cold front had passed over the Craigieburn Range. Cool southerly air flowed into the area. Avalanching was first noted around 0800 hrs as the cold front passed over the Craigieburn Range.

Tongariro National Park (T)

15/8/79 (0000 hrs) - a dual front lay north-east and south-east of an area of low pressure immediately off the west coast of the North Island.

(1200 hrs) - the dual front had become stationary with strong south to south-westerly winds being advected over the Tongariro National Park area. Avalanche release began soon after this.

Summary

Timing of avalanching for the Wanaka, Craigieburn and Tongariro areas appears to be related to a few hours after frontal passage and maximum precipitation. Presumably this is related to wind redistribution of the new snow by southerly winds.

APPENDIX 7.4 : MULTIPLE FRONTAL PASSAGE AND AVALANCHE RELEASE THE STORM OF 3 TO 5 SEPTEMBER, 1981

(a) Front 1 had passed over the Craigieburn Range by 0000 hrs on 4/9/81. Precipitation from this front had been received in the Craigieburn Ranges at 1800-2100 hrs on 3/9/81 (e).

(b) Front 1 had passed off the country with a general south-westerly flow over the Craigieburn Range area. A second front was rapidly approaching the Fiordland area at this stage. Strong north-westerly wind preceded the front bringing precipitation to Fiordland. Avalanching in the Fiordland area was noted soon after 1200 hrs on 4/9/81 with the passage of front 2.

(c) Precipitation in the Craigieburn from the north-west reached a maximum between 1800 hrs, 4/9/81 and 0000 hrs on 5/9/81 (e) as front 2 passed over the area. Avalanching was noted in the Craigieburn Range soon after the passage of front 2. A third front had developed in the south-westerly air following front 2. Front 3 passed over the Fiordland area bringing further precipitation and avalanche release sometime between 0000 and 1200 hrs on 5/9/81.

(d) Front 3 had passed over the Craigieburn Range by 1200 hrs 5/9/81 producing further precipitation between 0900 and 1200 hrs on 5/9/81. Further avalanching in addition to that noted with the passage of front 2 occurred in the Craigieburn Range as south-westerly air moved into the area following the passage of front 2. Front 2 at this stage lay over the North Island. Avalanching was recorded in the Mt Egmont and Tongariro National Parks as Front 2 passed over and Front 3 rapidly approached these areas.

APPENDIX 7.5 : DEVELOPMENT OF STORM 24 TO 26 AUGUST, 1978

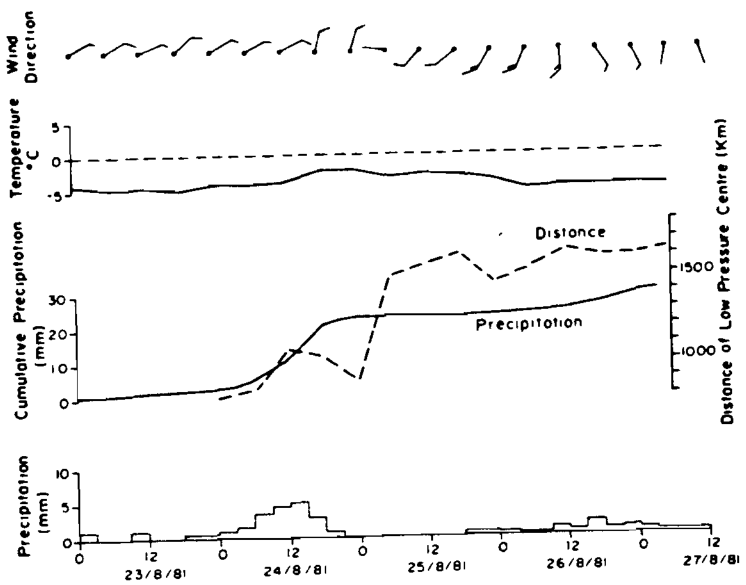
(a) A cyclonic flow dominated the situation over New Zealand with a centre of low pressure lying over central New Zealand. Precipitation commenced in the Craigieburn Range at 0300 hrs from a north-easterly direction. (e)

(b) The low pressure system intensified with movement of the low pressure centre just north-west of the Craigieburn Range. Precipitation in the Craigieburn Range reached a maximum around this time.

(c) An upper level low had developed at the 500 mb level in association with the lower level low. The centre of the 500 mb low was displaced slightly north-west of the surface low.

(d) Movement of the centre of low pressure off the South Island brought the end of precipitation in the Craigieburn Range (e). Increasingly strong southerly winds invaded the area at this stage until avalanche release as a result of overloading by wind redistribution was reported at 1000-1200 hrs on 26/8/78 (f).

APPENDIX 7.6 : DEVELOPMENT OF STORM 24 TO 27 SEPTEMBER, 1981



(f)

(a) A centre of low pressure lay over the North Island while a weak cell of high pressure lay south-east of the South Island. Precipitation reached a maximum on 24/8/81 in the Craigieburn Range at this point. (f)

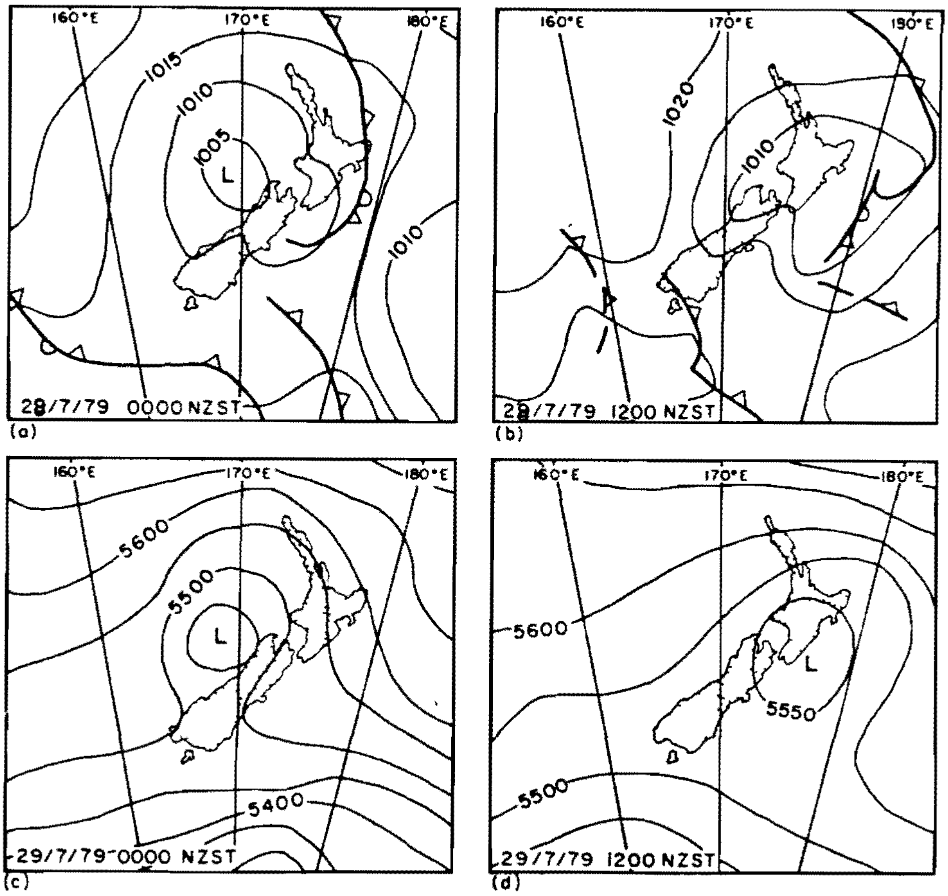
(b) Cyclonic and anticyclonic flows from the areas of low and high pressure produced convergence of air from a north-easterly and south-easterly direction east of the South Island. South-westerly air moving into this zone of convergence produced a cold front lying the length of the South Island.

(c) The zone of convergence and associated front shifted east away from the coast of the South Island. Precipitation stopped in the Craigieburn Range at this point with the movement of cool southerly air into the area. Avalanching was noted soon after 1200 hrs on 25/8/81 with increased rates of wind redistribution.

(d) A centre of low pressure developed in the flow of cool southerly air off the east coast of the South Island. The centre of low pressure remained relatively static with precipitation coming from the south to south-east.

(e) A small cycle of avalanching was noted in the Craigieburn Range as south-easterly air flowed into the area from the leading edge of an approaching high pressure system. Air temperatures continued to fall with the inflow of this southerly air (f).

APPENDIX 7.7 : DEVELOPMENT OF STORM 29 JULY, 1979



(a) A centre of low pressure moved onto the South Island. Precipitation in the Craigieburn Range at this point was from a northerly direction (Figure 5.9).

(b) The centre of low pressure moved over the area bringing further precipitation from the south (Figure 5.9). Avalanching was noted in the Craigieburn Range at this point coincident with the southerly flow and peak precipitation.

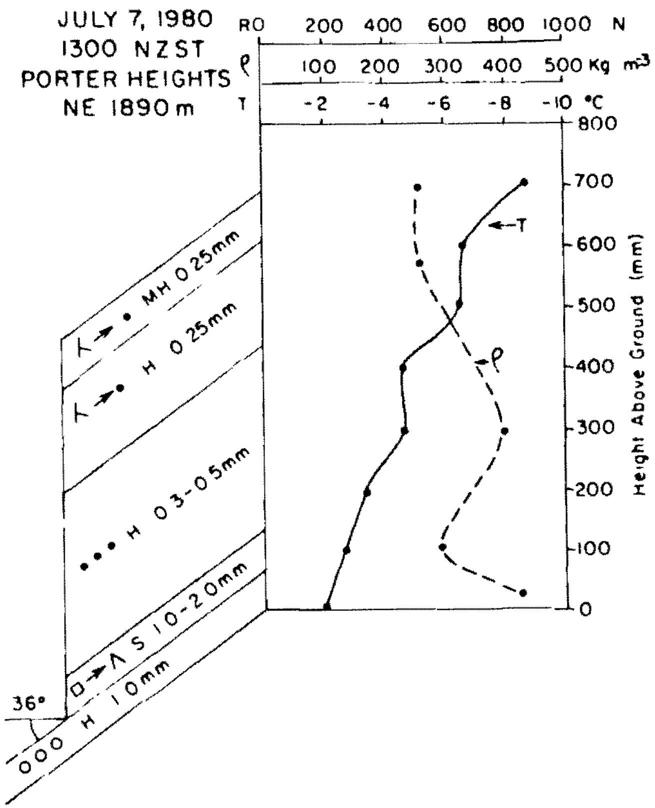
(c) An upper level low had developed in conjunction with the surface low (a).

(d) Precipitation had ceased in the Craigieburn Range by the time the upper level low had shifted off the South Island and had become less intense.

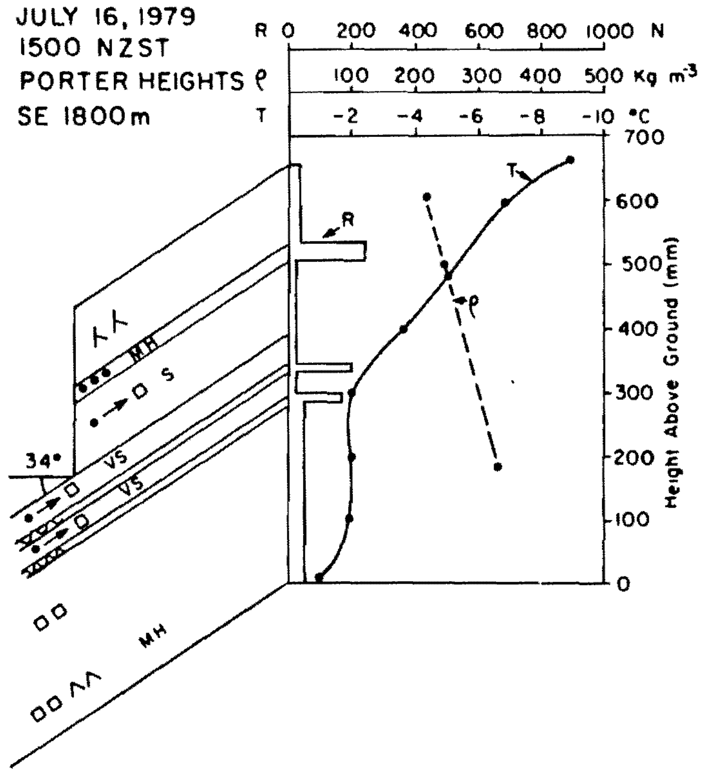
APPENDIX 8

FRACTURE LINE PROFILES

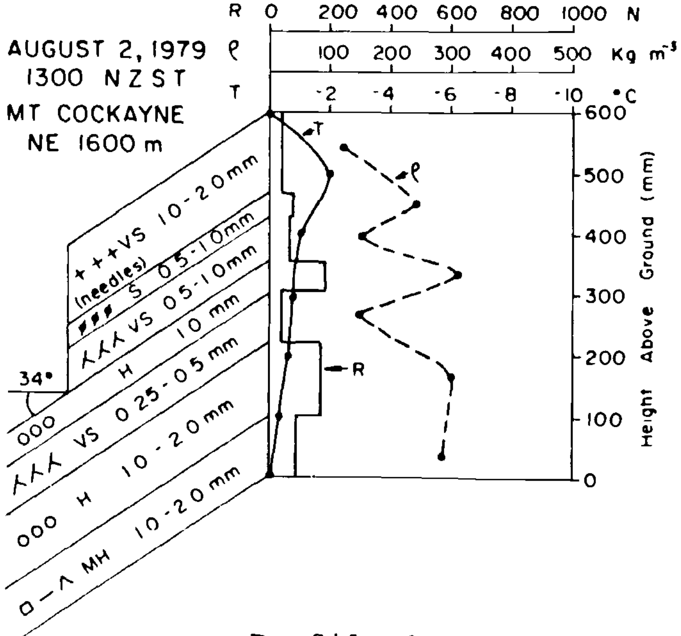
NUMBERS 1 - 9



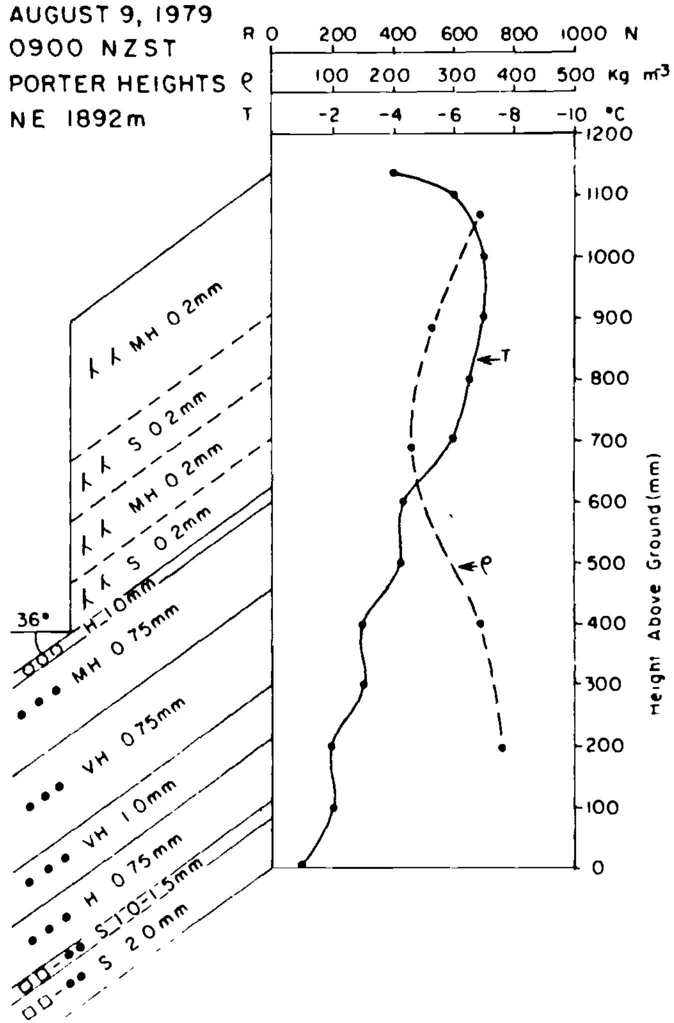
Profile 1



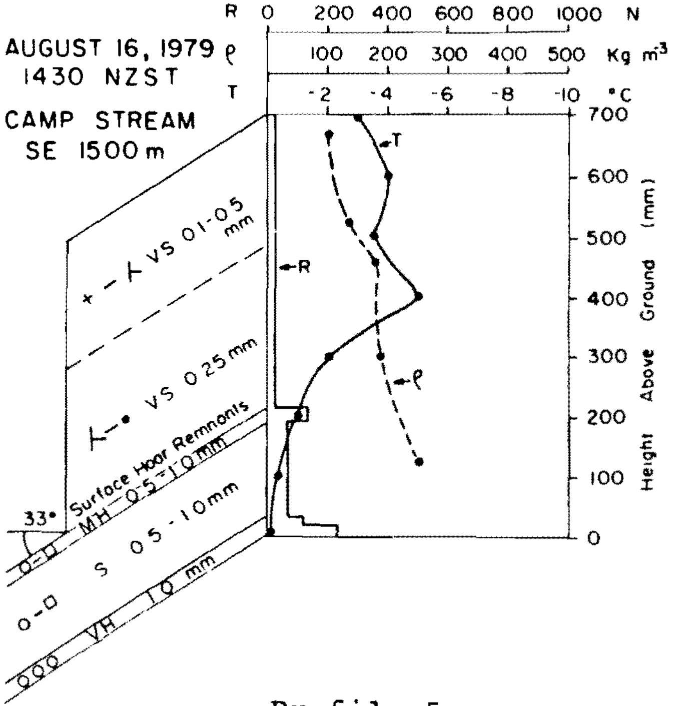
Profile 2



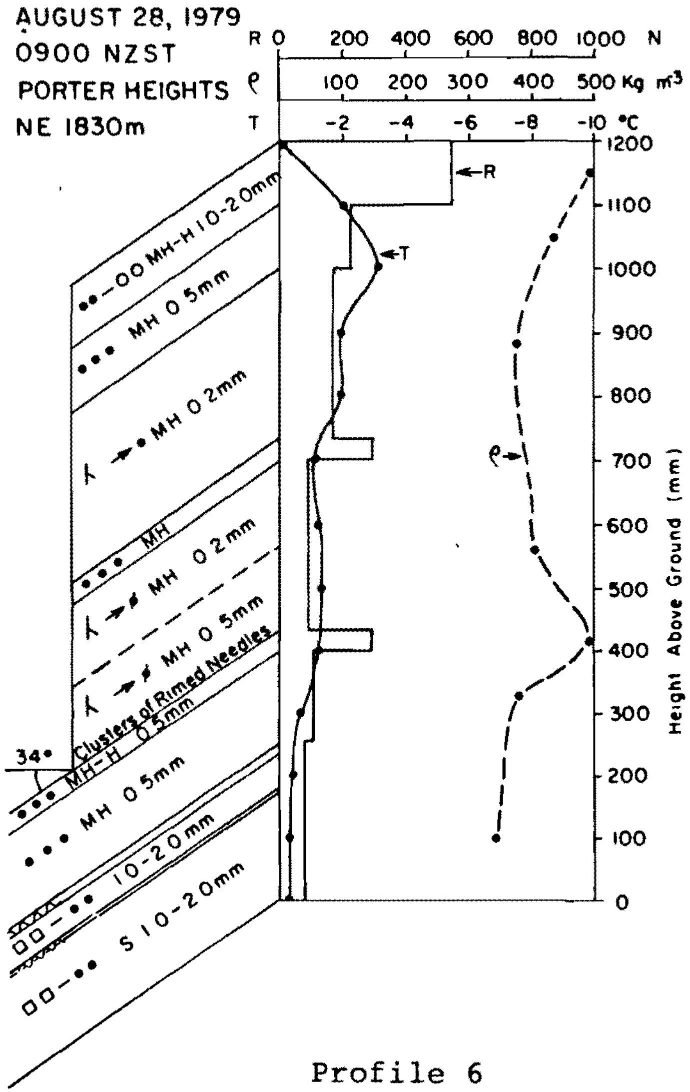
Profile 3



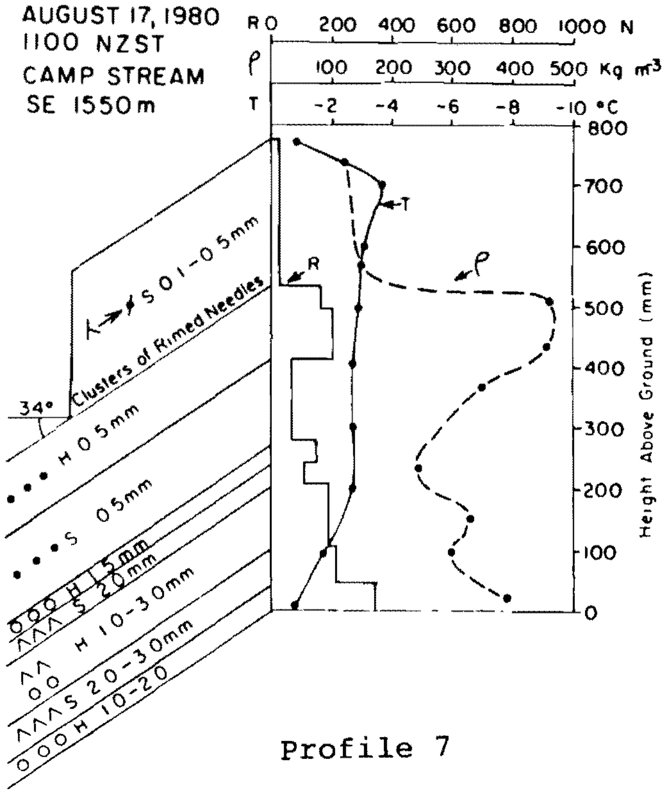
Profile 4



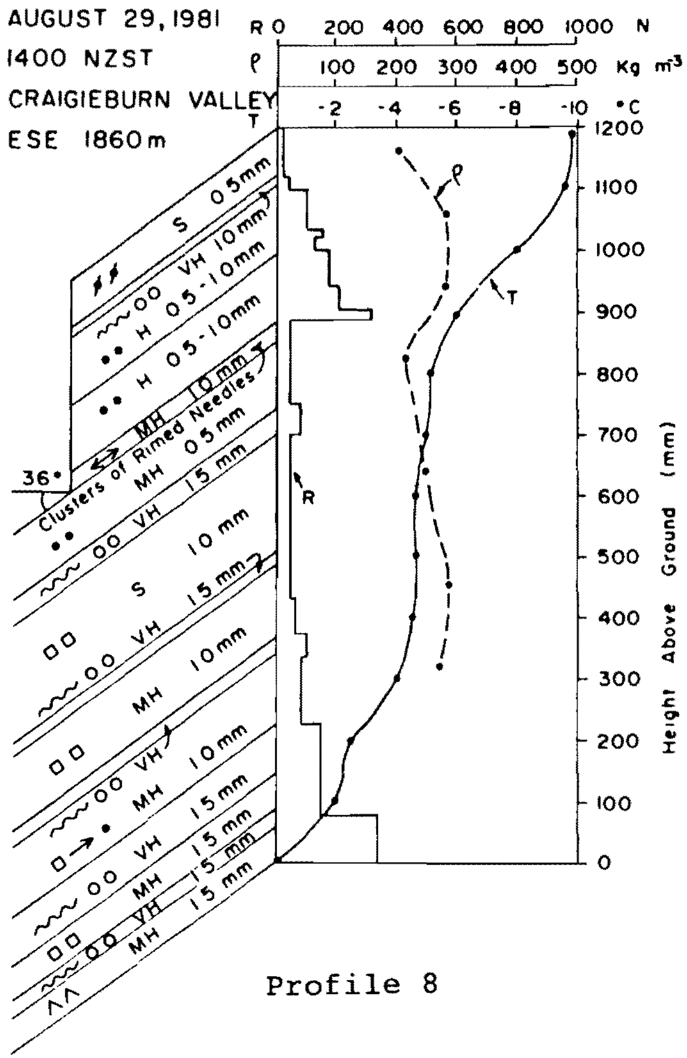
Profile 5



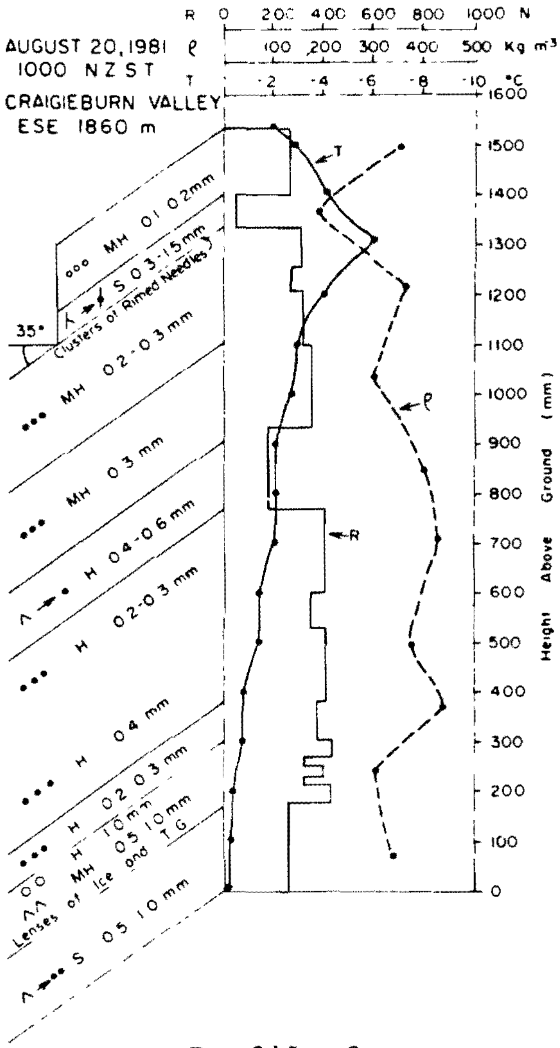
Profile 6



Profile 7



Profile 8


AUGUST 20, 1981
1000 NZST
CRAIGIEBURN VALLEY
ESE 1860 m
R 0 200 400 600 800 1000 N
ρ 100 200 300 400 500 Kg m³
T -2 -4 -6 -8 -10 °C
Height Above Ground (mm)
MH 01 02mm
S 03-15mm
Clusters of Rimed Needles
35°
MH 02-03mm
MH 03mm
H 04-06mm
H 02-03mm
H 04mm
H 02 03mm
H 10mm
MH 05 10mm
Lenses of Ice and TG
S 05 10mm

Profile 9

APPENDIX 9

EXAMPLE OF TEMPERATURE GRADIENT CRYSTAL DEVELOPMENT, MF LAYER BREAKDOWN, AND COLLAPSE OF SNOWPACK AS A RESULT OF BOND DESTRUCTION BY FREE WATER

Temperature gradients over the period of the three snowpits (a,b,c) changed from 0.007^{o}C mm^{-1} to 0.016^{o}C mm^{-1} to 0.005^{o}C mm^{-1} respectively. Decrease in overall snowpack depth is largely a result of snowpack settlement as evinced by the increase in upper snowpack densities from snowpit 'a' to 'c'. No new snow increments were noted over this period.

A period of sub-freezing temperatures and clear nights between snowpits 'a' and 'b' resulted in the destruction of the MF layers in snowpit 'a' at 200-240 mm above ground level (a.g.l.) by the large temperature gradient noted in snowpit 'b'. The effect of this massive vapour flux is manifest in the reduction in ram resistance of the MF layer in snowpit 'a' from 2010 to 80N in snowpit 'b'. Also, the well rounded and welded bond structure of the MF layer in snowpit 'a' had been reduced to a collection of MF grains and large faceted crystals in snowpit 'b'. These large faceted crystals are probably the original MF grains but show the effects of acting as foci of sublimation during vapour transfer. The severity of the temperature gradient recorded in snowpit 'b' had reduced the upper parts of the well compacted ET layer at 0-200 mm a.g.l. in snowpit 'a' from a density of approximately 400 kg.m^{-3} to 280 kg.m^{-3} in snowpit 'b' with associated development of mid to advanced TG snow. The majority of the snowpack in snowpit 'b' possessed faceted crystals as a result of this extreme temperature gradient.

From snowpit 'c' there is no evidence of the large temperature gradient of snowpit 'b'. Snow densities had increased in the upper parts of the snowpack and a major proportion of the pack was at melting point. Such a rapid contrast to snowpit 'b' resulted from a rain on snow event, one day prior to snowpit 'c'. As a result of warm air temperatures and free water draining through the top layers of the snowpack at 320 to 340 mm a.g.l. in snowpit 'c' all faceted grains had been destroyed with rapid ET metamorphism predominating. Ponded water was also evident at 180 to 310 mm a.g.l. in snowpit 'c' possibly indicating remnants of the former MF layer. This is further supported by the density increase immediately above the former position of the MF layer at 200 mm a.g.l. and the lower density TG snow below at 100 to 190 mm a.g.l. showing only a minimal increase in density.

The overall effect of free water in the pack at 200 to 300 mm a.g.l. was the destruction of bonding in this layer. Failure initially in this layer with further failure in the advanced TG layer at 100 to 190 mm a.g.l. Extensive areas of wet slab activity were noted as a consequence of this set of developments.

ERRATUM

For Figure 3.4a-c Avalanche Tracks should read Avalanche Paths.