UNIVERSITY OF CALGARY

Observations of Faceted Crystals in Alpine Snowpacks

by

Gregory T. Johnson

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES IN PARTIAL

FULFILLMENT OF THE REQUIREMENTS FOR

THE DEGREE OF MASTER OF SCIENCE

DEPARTMENT OF CIVIL ENGINEERING

CALGARY, ALBERTA

NOVEMBER, 2000

0-612-55284-5

Abstract

During the winters of 1993 to 2000 in the Columbia Mountains and the Rocky Mountains of southwestern Canada, over 100 strength measurements from 16 time series of faceted crystals were made with the shear frame. Snowpack variables were also measured at the same times and locations as the strength measurements. Rank correlations and physical arguments show load is the primary snowpack variable that affects the shear strength of layers of faceted crystals.

During the winter 1999-2000 bond diameters between faceted crystals were measured. One of these layers was observed in the Columbia Mountains and one in the Rocky Mountains. Bond diameters were qualitatively compared with snowpack variables. The shear strength, load, and bond diameters are inter-related. Physically, more load pushes faceted crystals closer together causing and increase in layer strength because of increased number of bonds, and / or increased bond diameters.

Acknowledgements

I would like to thank Dr. Bruce Jamieson who provided supervision, expertise, patience, and funding for this project.

For their encouragements, proof reading, and help with field studies I would like to thank Jeff Bonderchuck, Thomas Chalmers, Eric Dafoe, Michelle Gagnon, Torsten Geldsetzer, Jeff Goodrich, Rowan Harper, Phil Hein, Jill Hughes, Crane Johnson, Alan Jones, John Kelly, Kalle Kronholm, Paul Langevin, Bruce McMahon, Tom Riley, Dave Skjönsberg, John Schliess, Steve Thomas, and Adrian Wilson.

For their advice on measuring bonds I would like to thank Dr. Charles Fierz and Dr. Ron Perla. I would also like to thank Dr. Karl Birkeland for many discussions on the mysterious ways faceted crystals develop.

For their assistance with field studies, I thank the avalanche control section of Glacier National Park, Mike Wiegele Helicopter Skiing, and Dr. Dave McClung of the University of British Columbia for use of the cold laboratory and equipment at Rogers Pass.

Finally, I would like to thank my family and Andrea Kortello for endless patience and encouragement.

Table of Contents

List of Tables

List of Figures

Notation

A	Empirical constant analogous to the strength of ice
Age	Number of days since the layer of faceted crystals formed
B	Empirical constant
Emax	Maximum length of faceted crystals
Emin	Minimum length of faceted crystals
Δt	Time interval between measurements ranging from 1 to 8 days
Δt_{series}	Time interval over the time series
H	Thickness of the slab, the depth of snow over the layer of faceted crystals
H_{avg}	Thickness of the slab averaged over the time interval Δt
ΔH_{series}	Thickness of the slab averaged over the time interval Δt_{series}
$\Delta H/\Delta t$	Change of thickness of the slab over the time interval Δt
$\Delta H/\Delta t_{series}$	Change of thickness of the slab over the time interval Δt_{series}
HS	Depth of snow
HS_{series}	Depth of snow averaged over the time interval Δt_{series}
HS_{avg}	Depth of snow averaged over the time interval Δt
$\Delta HS/\Delta t$	Change of the depth of snow over the time interval Δt
$\Delta HS/\Delta t_{series}$	Change of the depth of snow over the time interval Δt_{series}
ΔHS_{series}	Difference in depth of snow over the time interval Δt_{series}
L	Thickness of the layers of faceted crystals
Load	Weight of overlying snow per unit horizontal area
$Load_{avg}$	Load averaged over the time interval Δt
$Load_{series}$	Load averaged over the time interval Δt_{series}
$\Delta Load/\Delta t$	Change in load over the time interval Δt
$\Delta Load/\Delta t_{series}$	Change in load over the time interval Δt_{series}
$\Delta Load_{series}$	Difference in Load over the time interval Δt_{series}
ρ	Density
ρ_i	Density of ice
ρ_{slab}	Density of slab overlying a weak layer

$\rho slab_{series}$	Density of slab overlying a layer of faceted crystals averaged over the time interval Δt_{series}
R	Crystal (grain) diameter
R_{wl}	Hardness of layers of faceted crystals measured with the hand hardness test
R_{wlavg}	Hand hardness averaged over the time interval Δt
σ	Overburden stress (load) on layers of faceted crystals
Σ	Shear strength of snow
Σ_{inter}	Shear strength of faceted crystals in intermountain snowpack climates
Σ_{cont}	Shear strength of faceted crystals in continental snowpack climates
Σ_i	Adjustable constant representing the shear strength of ice
Σ_1	Shear strength of layers of faceted crystals on day 1
TA	Air temperature
TA_{avg}	Air temperature averaged over the time interval Δt
TA_{series}	Air temperature averaged over the time interval Δt_{series}
$\Delta TA/\Delta t$	Change in air temperature over the time interval Δt
TA / HS	Temperature gradient averaged over the entire snowpack
TA / HS_{series}	Temperature gradient averaged over the entire snowpack and the time interval Δt_{series}
TG	Temperature gradient
TG_{series}	Temperature gradient averaged over the entire time series
TG_{avg}	Temperature gradient averaged over the time interval Δt
$\Delta TG/\Delta t$	Change of the temperature gradient over the time interval Δt
TG_{avg}/Twl_{avg}	The temperature gradient divided by the temperature of the faceted layer averaged over the time interval Δt
Twl	Temperature of the layer of faceted crystals
Twl_{avg}	Temperature of the layer of faceted crystals averaged over the time interval Δt
Twl_{series}	Temperature of the layer of faceted crystals averaged over the time interval Δt_{series}

$\Delta Twl/\Delta t$	Change of the temperature of the layer of faceted crystals over the time interval Δt
X	Bond diameter

1 Introduction

1.1 Avalanche Accidents and Costs

In North America between 1985 and 1998 there was an average of 28 avalanche fatalities per year (Atkins. In press). In southwest Montana, 59 percent of investigated avalanches between 1990 and 1996 (Birkeland et al., 1998) and 26 percent of the fatal avalanche accidents in Canada between 1972 and 1992 (Jamieson and Johnston, 1992) were a result of failures of layers of faceted crystals. Statistics for Canada show the number of avalanche fatalities is greatest in the continental snowpack of the Rocky Mountains where weak layers persist and least in the coastal snowpack climate of the Coast Mountains (Jamieson and Geldsetzer, 1996).

In Canada, the estimated economic cost associated with avalanches is $23.5 million per year (Jamieson and Stethem, 2000). Property damage caused by avalanches accounts for approximately 2.1 percent of the annual cost (McClung and Schaerer, 1993). In addition to property damage avalanches can destroy timber in forests. Estimated costs of destroyed trees in Canada due to a single avalanche can amount to $500,000 (Jamieson, In press).

Avalanche mitigation and control are expensive. Morrall and Abdelwahab (1992) estimated closing the Trans-Canada Highway at Rogers Pass, British Columbia, in western Canada for two hours costs between $50,000 and $90,000. Rogers Pass is typically closed to traffic 100 hours per winter (Jamieson and Stethem, 2000). In addition, approximately 70 highway areas require avalanche control in different areas of British Columbia (McClung and Schaerer, 1993). The total cost of vehicle closure to highways in British Columbia amounts to several million dollars per winter (Jamieson and Stethem, 2000). Other indirect costs include litigation, reduction in real estate values in areas threatened by avalanches, railway closures, ski area closures, and liability for the helicopter and snowcat ski industry (Voight, 1990; McClung and Schaerer, 1993).

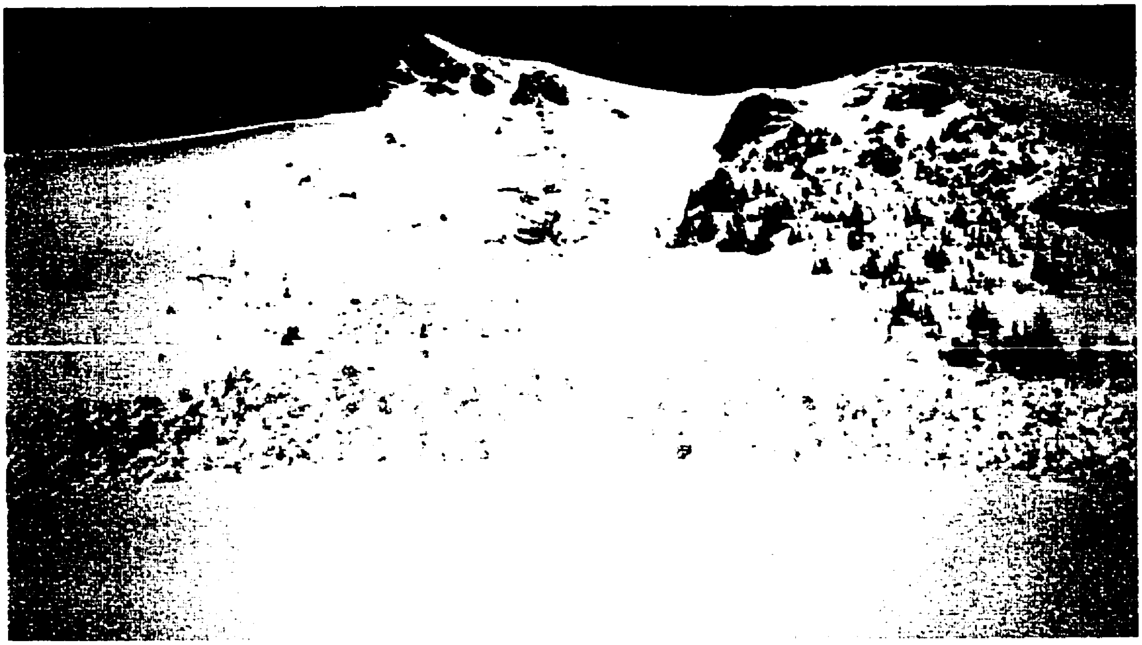

Figure 1.1 Loose snow avalanche.
(A. Jones photo)

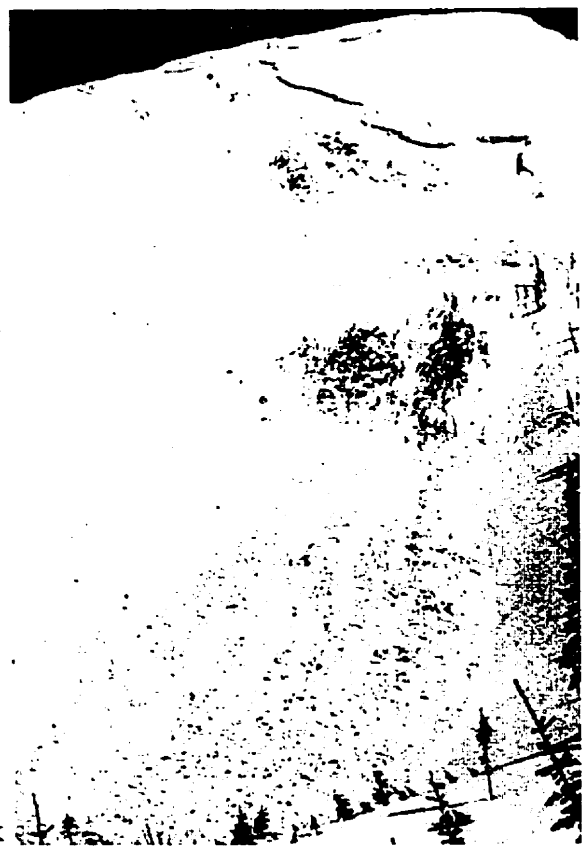

Figure 1.2 A slab avalanche.
(B. Jamieson photo)

1.2 Types of Avalanches

Two common types of avalanches are loose snow and slab avalanches. Loose snow avalanches (Figure 1.1) start at a point when a small amount of snow near the surface slips. After initial failure, the moving snow entrains more snow and spreads out as it travels downslope. Loose snow avalanches usually occur during or soon after storms when the surface snow is unstable.

Slab avalanches occur when a weak layer of snow fails beneath cohesive layers. The result is a relatively large mass of snow releasing (Figure 1.2) as a slab. Slab avalanches can occur in either wet or dry snow. The failure may be initiated by natural or artificial causes. Artificial causes include skiers, snowmobiles, explosives, and helicopters. Excluding explosives, skiers are the most common cause of artificial triggering in Canada (Jamieson, 1995). Skiers trigger slab avalanches by inducing critical stresses on weak layers that initiate failure (Föhn, 1987). Failure is thought to initiate in shear (McClung, 1979) and rapidly spread as brittle fracture.

1.3 Snow Metamorphism

The snowpack structure is dependent on the history of meteorological conditions, load due to new snowfall, and heat from the ground. After new snow particles are deposited they change in shape and size under the influence of three metamorphic processes: rounding (equilibrium), faceting (kinetic growth), and melt-freeze. The dry snow metamorphic process depends on the magnitude of the temperature gradient. Rounding metamorphism occurs when the magnitude of the temperature gradient, measured vertically in the snowpack, is below 10°C/m. Faceting metamorphism occurs when the magnitude of the temperature gradient is above 10°C/m (Akitaya, 1974). The extent of faceting or rounding metamorphism is dependent on snowpack properties such as temperature, temperature gradient, density, surface area-to-volume ratios of the crystals, load, and time. Under melt-freeze metamorphism snow melts and refreezes into clusters, poly-crystals, or crusts. In this study, only dry layers of faceted crystals were studied.

Figure 1.3 New precipitation particles have large surface-to-volume ratios that cause rapid decomposition.

1.3.1 Rounding Metamorphism

Initially, new snow breaks down because large surface-to-volume ratios (Figure 1.3) cause thermodynamic instabilities. These instabilities cause

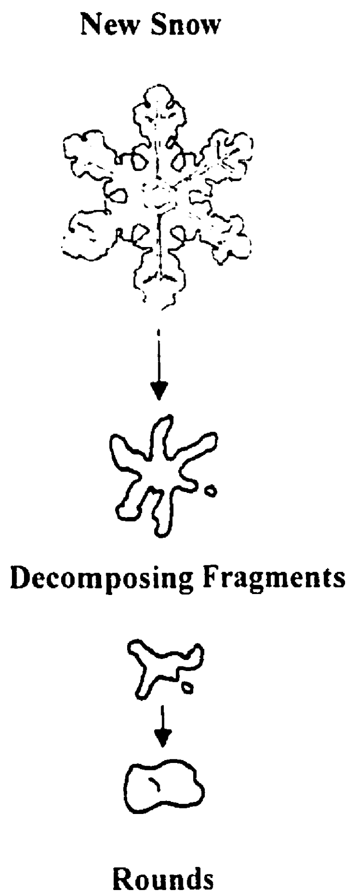

Figure 1.4 The evolution of a snow crystal during rounding metamorphism.

rapid decomposition of the new snow and rounding within the first few days (Colbeck, 1980). Decomposed and fragmented precipitation particles (class 2) are the intermediate stage between precipitation particles and rounded snow (Figure 1.4). After initial reduction in surface-to-volume ratios, decomposing fragments slowly decay over the winter to form rounded crystals if the magnitude of the temperature gradient remains below 10°C/m (Figure 1.4).

The rate at which snow crystals round increases with temperature (de Quervain, 1958). Generally, layers of rounded grains buried deep in the snowpack are warm and the grains round quickly. During rounding large rounded grains form at the expense of smaller crystals. This process causes an increase in strength by transferring water vapour from convex to concave areas, thereby promoting bond formation (de Quervain, 1963; Perla and Sommerfeld, 1987; Kurowia, 1975; Colbeck, 1997).

1.3.2 Faceting Metamorphism

Temperature gradients within the snowpack drive vapour pressure gradients which control water vapour movement. However, vapour pressure is difficult to measure within the snowpack. Instead, the temperature gradient is used as the index for water vapour transport.

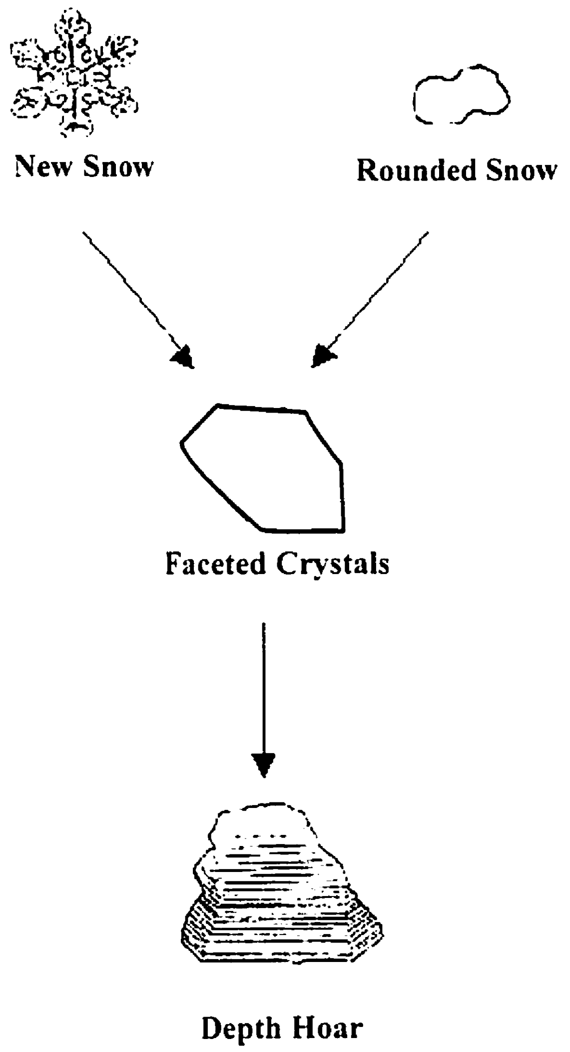

Figure 1.5 If a temperature gradient persists in new snow or rounded snow faceted crystals form and then depth hoar.

Temperature gradients are primarily a result of differences between ground temperatures and temperatures near the snow surface. The temperature profile of alpine snowpacks in temperate climates is usually warm ($0^{\circ}C$) at their base and cool towards their surface (Armstrong, 1981). When the magnitude of the temperature gradient is $10^{\circ}C/m$ or greater vapour transport is sufficient to cause the development of faceted snow crystals (Akitaya, 1974). As a result, water vapour sublimates upwards from the top of one crystal and deposits on the bottom of another (Armstrong, 1985). Faceted snow crystals are angular and have flat faces (Colbeck et al., 1990) (Figure 1.5). After faceted crystals form they may be persistent weak layers within the snowpack. These layers are responsible for some difficult to forecast slab avalanches.

The rate at which faceted crystals grow and the extent of their formation are primarily a function of snowpack temperature, the magnitude of the temperature gradient, and snow density (Akitaya, 1974; Marbouty, 1980; Fukuwaza and Akitaya, 1993).

The relationship between temperature gradient and crystal growth rate has only been shown for temperature gradients between 100°C/m and 300°C/m. Within this temperature gradient range the crystal growth rate increases as the given temperature gradient increases (Fukuwaza and Akitaya, 1993). The effects of the magnitude of the temperature gradient below this range and below 10°C/m are unknown.

The temperature of the snow also effects crystal growth rates (Akitaya, 1974; Lachapelle and Armstrong, 1977; Marboughty, 1980). Air within the pore spaces of snow is near saturation. The amount of water (mass) that air can hold is dependent on temperature. Cold air cannot hold as much water vapour as warm air. As a result when faceting occurs in cold snow (<-10°C) there is not as much water vapour available to deposit on crystals and hence growth is slower (LaChapelle and Armstrong, 1977; Armstrong, 1985; Stull, 1995).

The density of the snowpack affects crystal growth because the size and the network of pore spaces between the crystals affect the transport of heat and water vapour through the snowpack. As snow density increases, the size and network of pore spaces decreases, which limits water vapour migration. In addition, thermal conductivity increases with increased snow density allowing heat to flow more easily through the snow crystals and reducing the temperature gradient (Moore, 1982).

The combined effect of temperature gradient, temperature, and density is complex. In general, high crystal growth rates occur when the snow density is low, there is a large temperature gradient, and the snowpack temperatures are relatively warm.

1.3.3 Melt-freeze Metamorphism

Melt-freeze metamorphism is a result of melting and refreezing of liquid water between snow crystals. In the presence of water from rain, fog, or warm air temperatures, strengths of snowpack layers decrease because the bonds warm and start to melt. When the snow layers refreeze they form ice layers (crusts) or poly-crystals with large bonds and gain strength. Crusts are of concern, because they may promote faceting in adjacent layers (Moore, 1982; Colbeck 1991; Colbeck and Jamieson, In press) and may provide sliding layers for avalanches.

1.3.4 Snow Crystal Descriptions

The sequence of crystal shapes in the faceting process is shown in Figure 1.5. In the presence of a temperature gradient, rounded crystals and / or new snow form faceted crystals and / or depth hoar. Common sizes of faceted crystals and depth hoar reported by field workers range from 0.25 – 4 mm (Birkeland et al., 1998; Jamieson, 1995). The International Classification for Seasonal Snow on the Ground (Colbeck et al., 1990) defines four types of faceted crystals. Solid faceted crystals usually take the form of hexagonal prisms, Class 4fa. Near-surface faceted crystals are faceted crystals that form during near-surface faceting processes. These faceted crystals are defined as class 4sf and have a maximum size of 0.5 mm (Colbeck et al., 1990). However, Birkeland (1998) reports that these crystals often grow bigger. When rounded crystals are subject to an increasing temperature gradient, and faceted crystals are forming, they are classified as mixed forms, Class 3mx. These crystals change from rounded edges to sharp angular edges. Faceted crystals that are rounding are classified as mixed forms, Class 4mx. These crystals change from sharp angular edges to rounded edges.

Depth hoar occurs in dry snow when the temperature gradient is sustained long enough to cause striations or hollow forms (Figure 1.5). The shapes of the crystals include hollow cupped prisms, sector plates, and long prisms (Akitaya, 1974; Colbeck et al., 1990) and are usually 2-10 mm in length. The cup crystal (class 5a) is a hollow striated crystal that grows under a large temperature gradient and is sometimes found in weak layers. If cup crystals are further subjected to a temperature gradient they grow into columns losing their lateral bonds and become very weak (class 5b). The final stage of depth hoar growth occasionally occurs in continental snowpacks, but commonly in polar snowpacks (Class 5c)(Sturm and Benson, 1997). These depth hoar crystals are columnar crystals 10-20 mm in length, that are stronger than the other classes due to a network of bonds.

1.4 Bonds Between Snow Crystals

The formation and growth of ice bonds between snow crystals is complex and not well understood. Past studies have shown the number of bonds per crystal, bond diameters, bond shape, and bond orientation are probably the primary factors that determine the strength of snowpack layers (Yosida, 1963; Keeler 1969; Brown, 1980; Gubler, 1982; Edens et al. 1991; Colbeck, 1997; Colbeck, 1998). These factors are affected by load and temperature (Keeler, 1969; Edens et al., 1991; Colbeck, 1997). In the absence of temperature gradients greater than 10°C/m, rounding of crystals slowly occurs. During rounding metamorphism bonds grow and contribute to the strength of the layer (Colbeck, 1997). However during faceting, bonds both disappear and grow.

1.5 Snowpack Climates of the Columbia and Rocky Mountains

In western Canada there are three mountain climate zones: coastal, intermountain, and continental (Fitzharris, 1981; Claus et al., 1984; Mock, 1995). Each of these climate zones has different weather and snowpack conditions. The Coast Mountains have a coastal climate characterized by relatively warm temperatures and heavy snowfall. The Rocky Mountains have a continental climate associated with cold temperatures and shallow snowpacks. The intermountain climate of the Columbia Mountains is due to a mixture of coastal and continental weather systems. As a result, the intermountain climate has less snowfall than the Coast Mountains, but more than the Rocky Mountains. Typical mid-winter snowpack depths at tree-line are 2-4 m in the Columbia Mountains and 1-1.5 m in the Rocky Mountains. The exception is east of the Columbia Mountain divide where in certain locations (e.g. Bobby Burns) the continental snowpack climate exists.

The continental snowpack of the Canadian Rockies is known for its layers of faceted crystals and depth hoar. Typically the temperature gradient averaged over the snowpack is above the faceting threshold of 10°C/m from the first snowfall until the end of February (Appendix A). By March, facets and depth hoar crystals deep in the snowpack are warm, under a low temperature gradient, and start to round. However, faceting may still occur due to localized temperature gradients caused by diurnal fluctuations of air temperature and radiation.

In general, the overall snowpack temperature gradient never reaches the threshold value for faceting in the thick snowpack areas of the Columbia Mountains. Most of the faceted crystals form quickly due to one of the three near-surface processes described by Birkeland (1998). After layers of faceted crystals are buried, low snowpack temperature gradients promote rounding and strengthening. The snowpack climates of the Rocky Mountains and Columbia Mountains are summarized in Appendix A.

1.6 Field Observations of the Snowpack

The objective of a snow profile is to provide detailed information about each snowpack layer (Figure 1.6). Layer information includes grain type, grain size, density, thickness, hand hardness, and liquid water content (Colbeck et al., 1990). Snowpack temperatures are taken at specified depths and not at each layer. The location of a profile must be carefully chosen to best represent the surrounding snowpack. In Canada, snow profiles are performed according to the *CAA Observation Guidelines and Recording Standards for Weather, Snowpack and Avalanches* (CAA, 1995).

Figure 1.6 A field worker observing snow crystals while performing a snow profile.

Strength tests are used to measure the strength of snowpack layers. The shear frame test was used in this study to measure the shear strength of layers of faceted crystals. The test uses a shear frame which "grips" the snow just above the weak layer and a force gauge to record the force required to produce fracture.

1.7 Topic of Study

To prevent avalanche accidents avalanche forecasters rely on snowpack data, weather forecasts, and previous avalanche activity. Because weather, snowpack conditions, and time restrict data collection, there is an increasing interest in snowpack evolution models. However, the shear strength of layers of faceted crystals is often poorly predicted by such models (Fierz, 1998). In addition, the failure of layers of faceted crystals are difficult to predict because their failure strength is typically unknown and they may persist in the snowpack for months.

The mechanical properties of snowpack layers are directly related to the properties of bonds between snow crystals (Ramseier and Sander, 1966; Keeler, 1969; Kry, 1975; Gubler, 1978; de Montmollin, 1982). Failure of weak snowpack layers occurs when a sufficient number of bonds fail between individual grains. The number of bonds, their size, shape, and orientation determine the strength of snowpack layers (Colbeck, 1997). This study also focuses on the evolution of bonds of layers of faceted crystals over time and their relationships with shear strength and snowpack variables.

The objectives of this study are:

- to monitor the evolution of bonds over time between faceted crystals.
- to relate the changes in bonds to snowpack variables and shear strength.
- to determine snowpack variables that correlate with the shear strength of layers of faceted crystals.

1.8 Overview of Chapters

In Chapter 1, avalanche impacts, types of avalanches, metamorphism of crystals bonds between crystals, and the objectives of this study are presented. Chapter 2 reviews the theories of bond growth, studies of bonds over time, textural evolution of layers of faceted crystals, near-surface faceted crystal formation, and rate-of-strength-change forecasting models. The methods required to collect data are presented in Chapter 3. In chapter 4, the results of bond diameter measurements, the geometrical evolution of bonds and their relationship with snowpack properties for two layers of faceted snow crystals are discussed. In chapter 5, changes in shear strength of layers of faceted snow crystals and their relationships with easily measured snowpack properties are presented and strength is regressed as a function of time. The conclusions and suggested research are presented in Chapter 6.

2 Literature Review

2.1 Introduction

This review focuses on the evolution of bonds between faceted crystals, the evolution of layers of faceted crystals, the theory of bond development, faceted crystal formation processes, shear strength over time models, and correlations between shear strength and snowpack variables. Section 2.2 describes studies on bond evolution. The theory and hypotheses for bond growth are presented in Section 2.3. The evolution of layers of faceted crystals is reviewed in Section 2.4 and the formation processes and case studies of near-surface facet crystal formation in Section 2.5. In Section 2.6 models, for change in shear strength over time are reviewed. Previous work on rate of strength change models is described in Section 2.7. The summary of the literature review is presented in Section 2.8.

2.2 Studies of Bonds Over Time

Bonds between snow crystals are important because their properties are the primary factors that affect the mechanical properties of layers of snow crystals (Ramseier and Sander, 1966; Keeler, 1969; Kry, 1975; Gubler, 1978; de Montmollin, 1982; Colbeck, 1997; Pielmeier, 2000). A bond is usually defined to be the minimum cross-sectional area of the necks between snow grains. In previous research, the evolution of bonds was typically studied with plane or thin sections. Two-dimensional stereological parameters were used to quantify bonding properties such as bond diameter, bonds per unit area, and joint order.

Seligman (1936, p. 151) photographed the process of firnification from precipitation particles. Although the photographs are not from a time series, they represent the time dependent densification process. Bonds are described in terms of crystals being "frozen" or "cemented" together. Keeler (1969) gives credit to Eugster (1952) for first recording the number and size of bonds and correlating them with density. In Akitaya's (1974, p. 7) extensive study of faceted crystal and depth hoar development, bonds were observed to be thin and disappearing during the faceting process.

Yosida (1963) studied the evolution of joint order in rounded snow with an initial density of 120 kg/m^3 over a period of 31 days. Joint order uses the ice network of snow, to show the amount of bonding per crystal. The method requires lines be drawn on photographs of thin sections bisecting each arm of snow crystals. The points created where the lines meet represent bonds between snow crystals. The number of these lines intersecting per crystal is the joint order. The study was limited to two-dimensional analysis, but Yosida concluded three-dimensional results would be similar due to the isotropic structure of rounded snow. It was found for the rounding metamorphism of snow the mean joint order remained approximately 3.4. The constant joint order indicates the structure of the snow did not significantly change over the 31-day period (Keeler, 1969).

Keeler (1969) was the first to systematically monitor bonds in a natural snowpack throughout an entire winter (Colbeck, 1997). Data were collected in the field from December to March at Alta, Utah and supplemented with laboratory tests. Collected samples of snow were studied by thin section analysis in a nearby cold laboratory. From the thin sections joint order, bond diameter, number of bonds per square millimetre, and number of bonds per crystal were measured. Similar to Yosida (1963), Keeler found the joint order did not significantly change for rounded snow. The number of bonds per square millimetre was used to calculate bond area per unit area. This calculation assumed all bonds had an equal diameter and a circular cross sectional area. Even though Keeler reported the "crudeness" of measuring bonds per unit area, he reported the importance of the calculated bond area per unit area increasing with shear strength. In addition, both bond diameter and the number of bonds per crystal were found to increase with time. Increases were expected because rounding was the predominant metamorphic process.

In Keeler's (1969) study on bond evolution only two measurements were taken within the first 59 days. The snowpack layer he monitored was initially new snow that became decomposed and rounded particles. Despite only two data points in the first 59 days, the bond diameters were shown to reach a maximum diameter for their

Figure 2.1 Traditionally researchers have assumed bonds were concave necks between grains. In this case, a grain boundary groove is present at the bond.

"environment". He concluded the maximum bond diameter indicated most of the structural arrangement of rounded crystals occurred within two months.

2.3 Bond Growth Theory for Dry Snow

Sintering is the process by which ice and snow particles bond together below the melting point (Ramseier and Keeler, 1967; Colbeck, 1997). There are two theories describing the physical processes of bond growth. The first is taken from powder metallurgy and describes bond growth in terms of temperature, time, and crystal radius (Kuczynski, 1949). It does not consider overburden pressure (load), temperature gradient, crystal shape, or bond geometry. The driving force for the transfer of ice to the bonds is a gradient in chemical potential between concave and convex surfaces. Alpine snow is unique in the sense that it is usually encountered near its melting temperature (>-10°C). Its relatively high temperatures promote quick metamorphic changes. The relatively high temperature and chemical gradient allow water molecules to migrate to bonds by volume diffusion, surface diffusion, viscous flow, and evaporation – condensation. The equation for bond growth is expressed in general form as

$$X = R(f(T)\, t / R^{m})^{1/n} \tag{2.1}$$

where $f(T)$ is a temperature term, t is time, R is the crystal radius, X is neck radius, and m and n (Table 2.1) are functions of individual growth processes. The equation above assumes geometry of the ice crystals are spherical and the bonds have geometries of circular concave necks (Figure 2.1).

Kingery (1960) proposed that of the four mechanisms in which ice can be transported, surface diffusion was responsible for bond growth between two ice spheres. In contrast, Hobbs and Mason (1964) found the primary growth mechanism is sublimation-condensation. Ramseier and Keeler (1966) confirmed this by testing snow in air and liquid silicone. The crystals submerged in silicone grew very slowly compared to those exposed to air because they prevented ice sublimating from one snow crystal to the next.

Keeler (1969) found that bond growth in low density snow was faster than the sintering theory predicted. Keeler (1969b) suggests the accelerated growth rate was due to overburden pressure and Colbeck (1997, 1998) proposed growth rates were due to an increased vapour flow from temperature gradients.

Colbeck (1997, 1998) presents the second theory of bond growth. He argues the classical theory of sintering describes snow as a non-crystalline material without grain-boundary grooves. In order to describe snow as a crystalline material, boundary grooves must be present in the neck between two crystals. Colbeck introduces grain-boundary diffusion as another and the primary ice transport mechanism contributing to bond growth.

Grain-boundary diffusion is controlled by the imbalance in stresses in the boundary groove (Figure 2.2). The imbalance in stresses causes ice molecules to diffuse out of the grain boundary. The diffusion of ice molecules causes the crystals to move

Table 2.1 Bond growth mechanisms and their descriptions (Kuoiwa, 1975; Hobbs and

Bond Growth Mechanism	**Description**	**m**	**n**
Surface Diffusion	Difference in absorbed ice molecules between neck and crystals (set up by surface tension forces)	4	7
Volume Diffusion	Lattice vacancies	3	5
Viscous Flow	Surface tension forces	1	2
Evaporation Condensation	Convex surface to concave necks	2	3

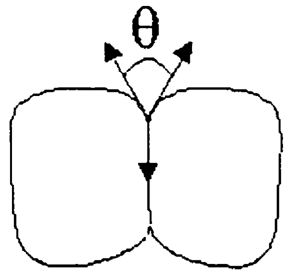

Figure 2.2 The grain boundary groove is located where the two crystals meet. The forces at the grain boundary are shown in 2-dimensions with arrows and the dihedral angle with θ. The three forces are balanced when the dihedral angle is 145°. If the dihedral angle is not 145° then the two upwards forces pull water molecules out of the grain boundary.

closer together. When the crystals move closer together the bonds grow and the dihedral angle between the crystals increases. When new bonds form this angle is small and the stress imbalance is large. This causes rapid bond growth. As the dihedral angle enlarges the stresses are closer to equilibrium and bond growth slows. The diffusion of ice molecules continues to occur until the dihedral angle reaches 145°. When the equilibrium dihedral angle is reached the bonds have grown to their maximum size. However, bonds may continue to grow if the crystals they connect grow.

Colbeck also postulates the rate at which bonds grow is dependent on vapour flow that is driven by temperature gradients not pressure sintering. He argues a greater rate of vapour flow should cause faster bond growth, just as it does for crystals. This may be why the rate of bond growth in Keeler's study was greater than the classical theory predicted.

2.4 Evolution of Layers of Faceted Crystals

Textural properties of snow crystals include shape, size, and intergranular relationship and arrangement (Arons and Colbeck, 1995). Throughout the history of snow studies textural properties have been analyzed. Avalanche forecasters monitor the evolution of texture to help predict snowpack stability. While layers of faceted crystals are relatively weak, field workers usually associate faceted crystals that are rounding with strengthening of layers because bonds are developing. Studying faceted crystal texture is important because conclusions may be drawn about bonds and strength.

Recently, attention has been given to the evolution of layers of faceted crystals (Stock et al., 1998: Jamieson and Johnston, 1997). In a recent study, Stock et al. (1998) followed numerous layers at Red Mountain Pass, Colorado from December to March. He measured stuffblock scores (Birkeland and Johnson, 1996), and hand hardnesses on valley bottoms and south, and north-facing slopes. The study found faceted crystals were larger on north-facing slopes and slower to gain strength.

Jamieson and Johnston (1997) recorded faceted layer shear strength, temperatures, temperature gradient, load, crystal type, and thickness of the slab for the winter of 1996 in the Columbia Mountains and Coast Mountains of southwestern Canada. The layer formed in November, but the temperature gradient was less than 10°C/m in magnitude by mid-December and continued to decrease throughout the winter. Some of the crystals rounded to mixed forms in December. Dry slab avalanches on this layer diminished in March, but but there were observations of wet slab avalanches on this layer in May. As expected from observations of decreasing temperature gradients, increasing loads, increasing temperatures, and increasing thickness of slabs the shear strength increased over time throughout the winter.

2.5 Near-Surface Faceted Crystal Formation and Case Studies

Near-surface faceted crystals were mentioned in literature by Seligman (1936 p.66, Fig. 52, p. 70) and Bader et al. (1939 p. 207). The specific metamorphism processes of near-surface faceted crystals are important because they determine initial crystal textures and their bond characteristics. Birkeland (1998) describes faceted crystal formation processes for radiation recrystallization, melt-layer recrystallization, and diurnal recrystallization. All three processes include temperature gradients in magnitude up to 300°C/m, which results in rapid crystal growth (Sommerfeld and LaChapelle, 1970; Armstrong, 1985; Fukuzawa and Akitaya, 1993; Jamieson and Johnston, 1997; Birkeland et al., 1998). Conversely, faceted crystals that form at the basal layers of the snowpack are generally subject to temperature gradients less than 100°C/m, densities greater than 185 kg/m^3, and may take a few weeks to develop (Akitaya, 1974; Fukuzawa and Akitaya, 1993).

Temperature gradients that drive near-surface faceting are a result of shortwave radiation, radiative cooling, and or the heat added by new snow or rain. Radiation recrystallization is typically found at high elevations, south aspects, low latitudes, and in the top few centimeters of snow (Sommerfeld and LaChapelle, 1970). Shortwave radiation is absorbed in the snow a few centimeters under the surface, often melting it. At the same time, out going longwave radiation cools the snow surface. Extreme temperature gradients occur and faceted crystals may form in hours (Sommerfeld and LaChapelle, 1970; Armstrong, 1985; Birkeland, 1998). Radiation recrystallization often causes crusts to form beneath the faceted snow (Birkeland et al., 1998) providing sliding layers for avalanches.

Diurnal recrystallization occurs from large swings in temperature gradients near the snow surface. Usually the process occurs during periods of cold clear nights and clear but subfreezing days. The temperature gradient switches direction from negative to positive forcing a change in vapour flow, which causes bi-directional faceting.

Melt-layer recrystallization occurs when the surface of the snow is melted and cooler snow falls on top. The result is a temperature gradient between the wet snow and the new snow sustained by the latent heat of the wet layer (Fukuzawa and Akitaya, 1993; Jamieson and Johnston, 1997; Colbeck and Jamieson, Submitted). Eventually the wet snow freezes and forms a crust. The process is enhanced when the new snow is followed by cold clear nights (Birkeland, 1998).

Faceted crystal layers often form under or above crusts (Seligman, 1936; Akitaya, 1974; Stratton, 1977; Moore, 1982; Colbeck, 1991; Birkeland et al., 1996; Birkeland, 1998; Birkeland et. al., 1998). Moore (1982) gives a theory on why crusts may promote faceted crystal growth. Crusts are dense layers that have high thermal conductivities relative to the surrounding snow. As a result, heat is easily transferred. Strong temperature gradients are created from heat being transferred away from the bottom of the crust and heat being added to the top. Colbeck (1991) adds the heat at the top of the crust promotes evaporation and vapour transport. Colbeck and Jamieson (Submitted) argue faceted crystal growth above wet or moist crusts is promoted by an abundant heat supply. After the faceting process has started it may be self perpetuating through a decrease in layer density.

Fukuzawa and Akitaya (1993) measured growth rates of near-surface faceting in the field and laboratory. In the field, quick growth was observed during a cold clear night when a thin layer of low density snow was overlying older dense snow. A strong temperature gradient was caused by the absorption of the shortwave radiation during the day that melted a layer a few centimetres below the surface. The melted snow remained at 0°C until well after sunset while the surface cooled rapidly. The laboratory experiment involved a snow sample subject to temperature gradients ranging from 100°C/m to 300°C/m and sustained for 50 hours. Rapid growth of faceted crystals occurred and depth hoar formed within hours.

Jamieson and Johnston (1997) gave a case study of a layer of faceted crystals that formed above a crust in the Columbia Mountains and Coast Mountains in November 1996. Rain fell on the 13th of November, wetting the snow surface. In the following 10 days, arctic air moved into the region and up to 87 mm of water equivalent precipitation fell in the South Columbia Mountains, 40 mm in the North Columbia Mountains, and 90 cm of snow in the Coast Mountains. Faceting was predominant in the North Columbia Mountains where the new snowfall was least. Faceting occurred because the cold new snow created a temperature gradient with the warm wet snow. The result was a near-surface layer of faceted crystals that created avalanches with human fatalities until January and natural avalanches until May.

During the winter of 1995-1996, Birkeland and Johnson (1996) investigated temperature gradients associated with diurnal recrystallization. Conditions yielded temperature gradients exceeding 200°C/m during the night and 100°C/m during the day. Within 36 hours of sustained but fluctuating temperature gradients faceted crystals, approximately 1 mm in size, formed in the top 25 cm of snow. Near-surface faceting occurred on all aspects and elevations. Avalanche activity occurred within a few days after the layer was buried with 50 cm of snow.

2.6 Strength Change Models

In past studies the shear strengths of snowpack layers have been modelled as a function of density. Perla et al. (1982) fitted strength Σ and density ρ data for individual crystal types to the equation:

$$\Sigma = \Sigma_i(\rho/\rho_i)^k \quad (2.2)$$

where Σ_i is the ultimate shear strength of ice, ρ_i is the density of ice and k is a coefficient. They found distinct regression coefficients that show crystal types gain strength at different rates.

Recent work by Jamieson and Johnston (1999) ranked factors associated with the rate of shear strength change of surface hoar. The study correlated snowpack variables with strength changes of layers of surface hoar. Correlations found the best predictor variables are height of the snowpack, the maximum crystal size, and thickness of the slab. Jamieson and Johnston also discuss physical processes of predictor variables that affect the rate of strength change. In general, bigger surface hoar crystals took longer to gain strength than smaller ones. This is probably due to few bonds per unit area relative to surrounding layers. Surface hoar layers were faster to gain strength the deeper they were buried. As more snow accumulates the surface hoar layer experiences an increase in load. The load is probably related to an increase in the number and size of bonds (Jamieson and Schweizer, 2000).

2.7 Summary

The relationships between the evolution of faceted crystal texture, bonds, and strength are not well known. This is probably due to the difficulty of qualitatively and quantitatively studying bonds.

The theory of bond growth between snow crystals has been derived from tests on uniform ice spheres under controlled conditions in a cold laboratory. Evaporation-condensation was found to be the primary crystal growth process. In low density natural snow. Keeler (1969) found sintering occurred faster than theory predicted. Colbeck (1997) disagrees with Keeler's (1969) suggestion that load causes larger bonds to form. Instead he argues temperature gradient is responsible for the growth rate. Recently, Colbeck (1997, 1998) proposed the classical bond growth theory presented by Hobbs and Mason (1964) may not be valid because it does not consider grain boundary groves.

Keeler (1969) was the first and only person to systematically study the evolution of bonds in rounded alpine snow. The studies were conducted at Alta, Utah in the field and in a cold laboratory during the winter of 1966-1967. From thin section analysis, bond diameter, number of bonds per unit area, and number of bonds per crystal were measured. It was shown that bonds reached a maximum diameter within the first two months.

Three types of formation processes for near surface faceting are radiation, melt-layer, and diurnal recrystallization. All three processes form faceted crystals and depth hoar quickly because they usually involve temperature gradients well above the critical value for faceting and low density snow. Conversely, faceted crystals and depth hoar that form in basal layers may take a few weeks to develop. After faceted crystals form, the evolution of their texture is important because of its effects on bonding and change in strength.

In past studies, the shear strengths of snowpack layers have been modelled as a function of density. Perla et al. (1982) found distinct regression coefficients for different types of snowpack layers indicating the variation in their shear strengths.

Only one study has focused on predicting rate of strength change of weak layers. Jamieson and Johnston (1999) found predictor variables that correlated with strength change for layers of surface hoar, but not for layers of faceted crystals.

3 Methods

3.1 Laboratory and Field Work

Snow profile observations and shear strength measurements on layers of faceted crystals were conducted at several study plots during the winters of 1993 through 2000. In addition, during the winter of 1999-2000 faceted crystals and their bonds from one layer at Bow Summit, Alberta and one layer at Mt. Fidelity, British Columbia were photographed in the Rogers Pass cold laboratory.

Study plots listed in Table 3.1 are located in the Columbia Mountains of British Columbia and the Rocky Mountains of Alberta. The Columbia Mountains are known for their intermountain snowpack climate. Average air temperatures are slightly cooler than the Coast Mountains, but warmer than the Rocky Mountains. The Rocky Mountains are associated with a continental snowpack climate. Typical mid-winter snowpack depths at tree-line are 2-4 m in the Columbia Mountains and 1-1.5 m in the Rocky Mountains. The exception is east of the Columbia Mountain divide where in certain locations (e.g. Bobby Burns) the continental snowpack climate exists.

At the Mt. St. Anne and Vermont (Table 3.1) study plots "air boxes" were used. Air boxes were constructed with plywood and were 14.5 m long, 2.5 m wide. One was 1 m high and placed at the Vermont study plot were the snowpack typically reaches 1.5 m. The other was 2 m high and placed at the Mt. St. Anne study plot where the snowpack typically reaches 3 m. During early winter the snow was shoveled off the top of each box until the snowpack reached the height of the box. Subsequently, the snow was allowed to accumulate on top of the box. One the layers of faceted crystals formed in the snowpack on top of the box, they were tested in the same manner as in the adjacent full-depth snowpack, except that the distance between snow pits was reduced to 0.5 m and the snow pits were carefully back-filled after the measurements were completed (Jamieson and Johnston, 1999).

Alpine snowpacks vary over time and space. Important requirements of study plots are (Figure 3.1) relatively constant slope angles, open areas without trees, minimal ground vegetation, and little snow drifting. These properties are necessary to minimize spatial variability to facilitate monitoring changes over time.

Table 3.1 Study Plots

Organization	Location	Study Plots	Snowpack	Elevation (m)	Aspect	Slope (°)
Glacier National Park, Park Canada 1999-2000	Rogers Pass, Selkirk Mountains,	Mt. Fidelity	Intermountain	1890	N	6
Mike Wiegle Helicopter Skiing 1996-2000 *	Cariboo Mountains, BC	Mt. St. Anne (MSA)	Intermountain	1900	E	0
		Cut-Block	Intermountain	1900	E	0
				1600	E	23-32
				1600	E	23-32
Canadian Mountain Holidays Bobby Burns 1993-1998 *	Purcell Mountains, BC	Moose	Intermountain	1900	NW	20-34
		Vermont	Continental	1600	N	0-6
		Elk	Continental	2300	NE	10-24
		Lodge Plot	Continental	1370	-	0
Banff National Park 1999-2000	Rocky Mountains	Bow Summit North	Continental	2020	N	0

* Observations of layers of faceted crystals prior to 1998 were performed by others.

Figure 3.1 Study plots should have snowpacks that represent the surrounding terrain.

3.2 Equipment

3.2.1 Microscope

In the cold laboratory at Rogers Pass, a Wild Heerbrugg polarizing stereomicroscope and Nikon SLR camera were used to take photographs. Two adjustable fiber optic lights provided transmitted light with little heat. Polarizers could be attached to the microscope mount and the objective. A Wild Heerbrugg HU phototube was used as a camera mount. The phototube allowed the camera to use the microscope as its lenses. Photographs could be taken at 8.10x, 16.20x, 31.25x, or 62.50x magnification.

3.2.2 Iso-Octane

Iso-octane was used to transport specimens of faceted crystals from study plots to the Rogers Pass cold laboratory (Brun et al., 1991) during the winter of 1999-2000. This chemical is a liquid with a freezing point of −60°C. Submersing snow crystals in iso-octane below 0°C prevents metamorphism of snow crystals.

3.2.3 Shear Frames and Force Gauges

The shear frames had an area of 250 cm^2 and were trapezoidal (Figure 3.2) to reduce friction between the sides and the snow. Each of the three cross members consisted of 0.6 mm thick stainless steel soldered to the frame. The lower edges of the shear frame were sharpened to ease placement. Chatillon and Imada force gauges (Figure 3.2) were used for the shear frame test. Table 3.2 lists the type of force gauge and the manufacturer rated accuracies. A load dial allows the gauge to be set to zero before each test. During each test, the load dial needle recorded the maximum force required to fracture the layers of faceted crystals with the shear frame.

3.2.3 Temperature Recording Equipment

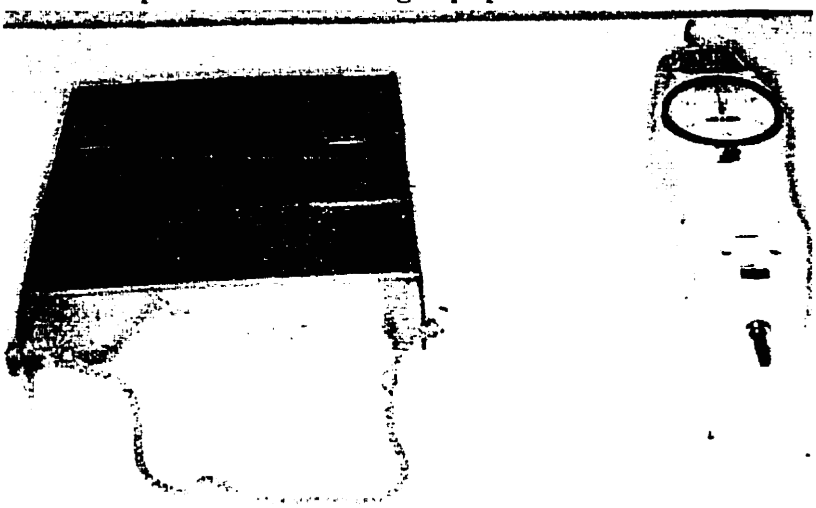

Figure 3.2 A 250 cm² shear frame and force gauge

Snowpack temperatures were measured using hand held digital thermometers and thermister pairs connected to a datalogger (Figure 3.3). Digital thermometers were used to incrementally record snowpack temperatures through the depth of the profile. Before the winter field season each thermometer was calibrated in an ice bath. Although the manufacturer rated accuracy was ± 1.0°C, a thermometer was not used unless it was within ± 0.2°C of 0°C during calibration.

Two thermisters, rigidly spaced 10 cm apart by a wooden block, were used to continually measure the temperature gradient across layers of faceted crystals. The thermisters were located at the end of plastic sheaths approximately 20 cm from the wooden blocks. Three sets of thermisters recorded hourly temperatures and stored the data on a Campbell Scientific CR10 datalogger. The accuracy of the thermisters was ±0.1°C.

Table 3.2 Pull Gauge Type and Accuracy

Pull Gauge Type	Maximum Load (kgf)	Manufacturer accuracy
Chatillon	2.5	1% max load
Imada	3	0.1% of reading
Chatillon	10	1% of max load.
Imada	10	0.3% of reading
Chatillon	25	1% max load
Imada	30	0.1% of reading

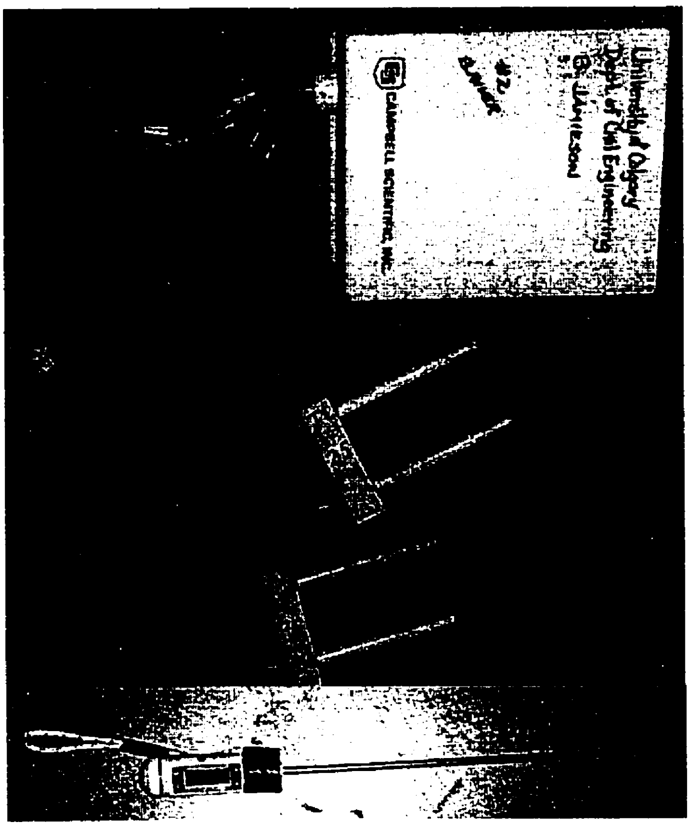

Figure 3.3 Pairs of thermisters attached to a Campbell Scientific CR-10 datalogger. Below is a digital thermometer.

3.3 Procedures

Snow profiles were observed and the shear strength measured depending on the depth of the layer of faceted crystals and the amount of time buried. If the layer of faceted crystals was deeper than 75 cm and had been buried for two weeks or more layer and snowpack tests were performed once per week. Measurements were made twice per week in cases where layers were shallower than 75 cm or had been buried for two weeks or less. Each observation of a snow profile and measurements on a particular layer were done at an undisturbed location within the study plot each time, typically

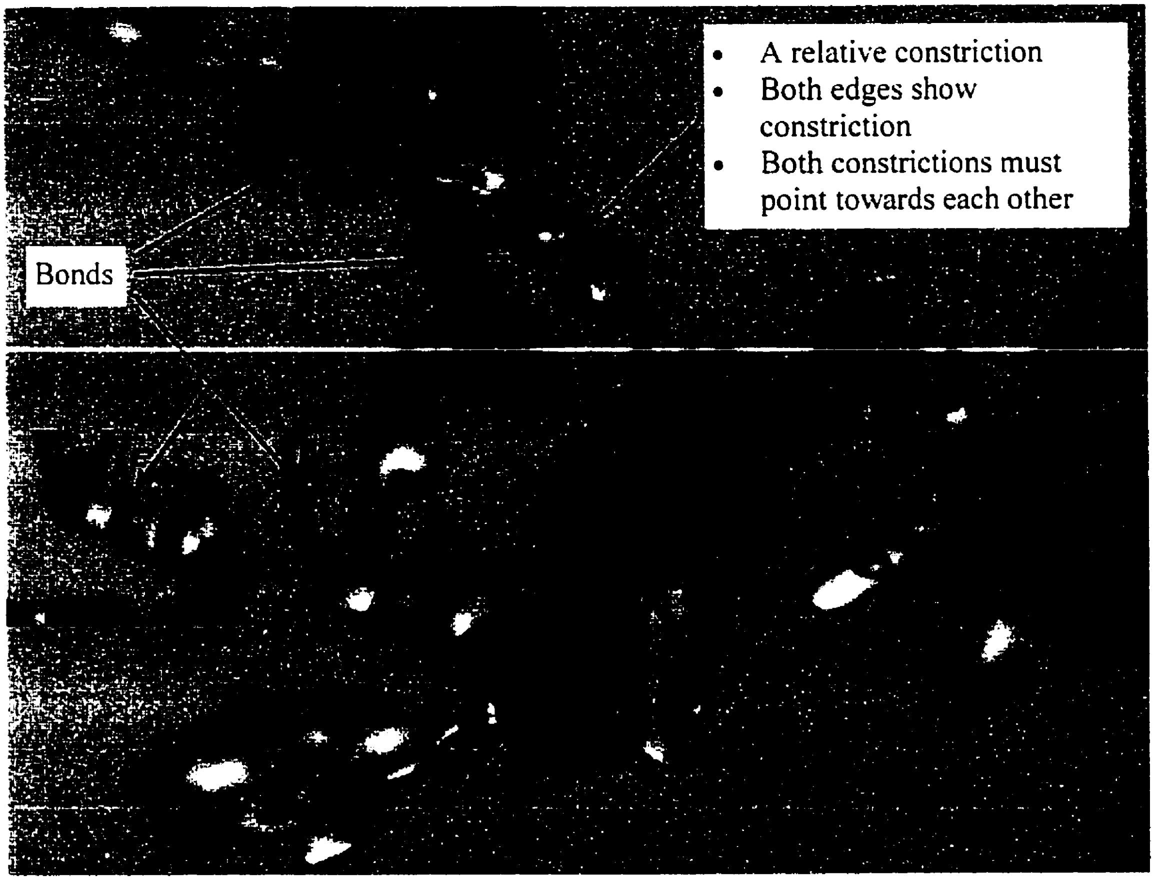

Figure 3.4 Guidelines for locations of bonds between snow crystals.

1.5 m - 2 m from the previous place.

3.3.1 Collecting and Photographing Faceted Crystal Aggregates

Faceted crystal aggregates are small clusters of faceted crystals bonded together. After the layer of faceted crystals was identified in a snow profile, a crystal plate was used to carefully remove the aggregates. They were then placed in chilled iso-octane, maintained below freezing, and transported to the cold laboratory for subsequent photography.

In the cold laboratory specimens were partially dissagregated manually until aggregates with 2-15 bonds were obtained. Each aggregate was then placed on a glass slide and photographed. The magnification on the microscope was set between 8.10x and 62.50x to capture crystals and their bonds.

Photographs were taken using Fuji Sensia II ISO 100 and Velvia ISO 50 slide film. Exposure times varied between 1/8 and 60 seconds depending on the magnification and amount of polarization. Each slide was then scanned into digital format and Image Pro software (Media Cybernetics, Silver Springs, MD) was used to measure bond diameters and crystal sizes.

Identifying bonds was somewhat subjective. As a guideline Kry's (1975) definition for 2-dimensional stereological analysis was used (Figure 3.4). Kry identified bonds in section plane photographs with the following criterion:

1) A relative constriction must exist; 30% constriction is required on the plane.
2) Both edges of the crystals must show a constriction
3) The constrictions on each edge must approximately point towards each other.

3.3.2 Slab Load

The slab weight per unit area on a weak layer is often referred to as the load. Each time shear frame tests were performed slab loads were measured by two methods. The first method requires all snowpack layers above layers of faceted crystals be identified with a snow profile. A cylindrical 100 cm^3 sample of snow from each layer is taken. If layers are thicker than 10 cm samples are taken vertically. When the layers are thick enough to take multiple vertical samples the results are averaged. In the case where the layers are less than 10 cm thick, but greater than 4 cm samples are taken horizontally. If the layers are less than 4 cm thick samples could not be taken. After removal from the snowpit wall, each sample is weighed using digital scale with an accuracy of 0.1 g. The load of the overlying slab is calculated by taking the product of gravity and the sum of layer thickness and densities.

$$\text{Load} = g\,(\rho_1 h_1 + \rho_2 h_2 + \ldots..) \qquad (3.1)$$

where g is gravity, ρ is density, and h is the layer thickness.

The second method is a "core load". It measures the bulk density of the overlying slab by using a cylinder with a cross sectional area of 0.0028 m^2. Vertical core samples

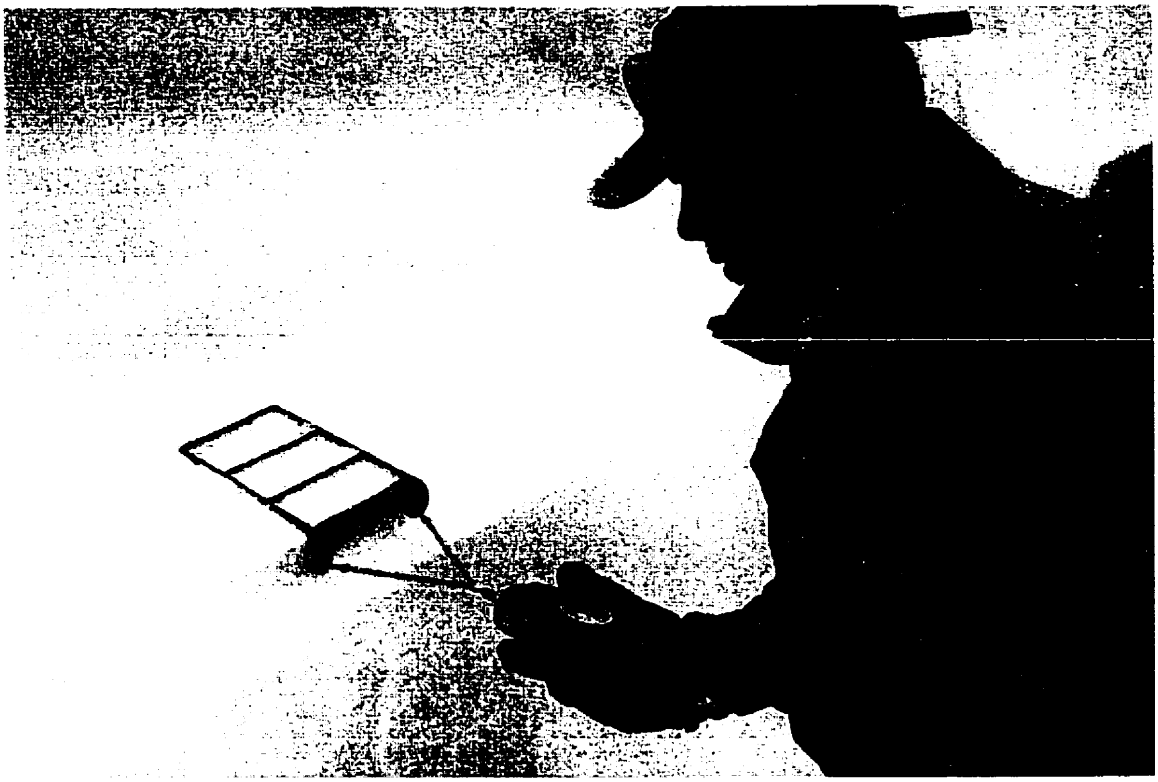

Figure 3.5 The shear frame test should be performed quickly to ensure brittle failure.

of snow overlying the desired layer are extracted from the snowpack and weighed. Load is calculated using the cross sectional area of the cylinder.

3.3.3 Shear Frame Test

The shear frame test was used to measure the shear strength of the layers of faceted crystals. Measuring the shear strength of layers of faceted crystals typically involves 12 individual shear frames tests in one or two rows across snowpit walls (Jamieson and Johnston, In press). The overlying snow was removed, leaving approximately 4-4.5 cm of undisturbed snow overlying the weak layer. A 250 cm^2 stainless steel shear frame with sharpened edges was inserted in the overlying snow so the bottom of the frame was 2-5 mm above the top of the weak layer. A thin blade was passed around the sides of the frame to remove snow that was in contact with the frame. A force gauge was attached and pulled parallel to layer quickly (< 1 sec) to produce brittle failure (Figure 3.5). If the fracture had a divot under the rear compartment of 1 cm or deeper, the test was rejected (Jamieson and Johnston, In press). The shear strength of the layer of faceted crystals was calculated by dividing the recorded force by the area of the frame. The shear strength is then adjusted for size effects of the shear frame (Sommerfield, 1980; Föhn ,1987).

The variation in shear strength among the 12 shear frame tests is from the shape of the fracture surface, the distance between the bottom of the shear frame and the weak layer (Jamieson and Johnston, In press), and spatial variability in snowpack layering. Jamieson and Johnston (In press) report for 342 sets of shear frame tests at level study plots and various types of snow crystals, the average coefficient of variation is 0.144.

According to Jamieson and Johnston (In press), fracture surfaces with a back divot 10 mm deep or greater under the rear compartment have strength measurements significantly different from planar fractures. The most common fracture surface of the November 18th, 1999 and February 9th, 2000 layers of faceted crystals was planar, indicating that the fracture surface was not a source of variation.

Strength measurements are sensitive to where the shear frame is placed with respect to the weak layer (Jamieson and Johnston, In press). The shear frame should be placed 2-5 mm above the weak layer. When the shear frame is placed within the weak layer, shear strength measurements may decrease by 20 %. If the shear frame is placed 10 mm above the weak layer, a 41 % increase in shear strength has been observed when compared with a shear frame placed within the weak layer.

3.3.4 Snowpack Temperatures

After, the layer(s) of faceted crystals was identified with a snow profile, thermisters pairs were then carefully inserted 5 cm above and below the center of the layer of faceted crystals. Care was taken to ensure the thermister pairs were placed with the correct slope angle and distance from the layer of faceted crystals. Two to three pairs of thermisters were usually used to reduce placement error. After they were placed snow was shoveled back into the snow pit to insulate the thermisters. Dense snow was used as fill at the bottom of the pit and progressively less dense snow was used towards the top. Upon removal every 7 to 14 days, distances between the thermisters and the center of the layers of faceted crystals were measured.

The temperature profile of the snowpack was measured with digital thermometers in 10 cm intervals. Temperatures were also measured 5 cm above and below as well as in the center of the layer of faceted crystals. Surface temperatures and snowpack temperatures measured down to a depth of 30 cm were shaded with a snow shovel on the surface to block incoming shortwave radiation.

3.3.5 Snowpack Parameters for Strength Change Models

Snowpack variables were measured at study plots the same day shear frame tests were performed. Snowpack parameters that were not described in the previous sections include air temperature. layer thicknesses. layer hardnesses. snowpack depth, crystal type. crystal size. and water content. Measurement procedures are found in CAA (1995). These variables are used in Chapter 4 to study bonds and in Chapter 5 to model the shear strength of buried layers of faceted crystals.

4 Faceted Crystal Bond Evolution and Textural Properties

4.1 Introduction

The shear strength of layers of faceted crystals is dependent on bond area, number of bonds per crystal (coordination number), bond shape, and bond orientation relative to other bonds and snow crystals (Keeler, 1969; Colbeck, 1998; Fierz, In press). In this study, bond diameters from two layers of faceted crystals were measured during the winter of 1999-2000. One of these layers was observed in the Columbia Mountains and one in the Rocky Mountains of western Canada. Various snowpack properties were measured at the same time and location as the specimens for bond measurements were collected. In this chapter, these snowpack properties are qualitatively analyzed for relationships with bond diameter and bond growth. In addition, the evolution of bond shapes from cold laboratory photographs is discussed.

4.2 Formation of the November 18th, 1999 and February 9th, 2000 Layers of Faceted Crystals

4.2.1 November 18th, 1999 Layer of Faceted Crystals

On the evening of the November 11th, 1999, at Mt. Fidelity in the Columbia Mountains, air temperatures warmed above freezing and rain began to fall. From November 11th to the 17th there were only a few morning hours of temperatures below freezing and associated snowfall. On November 15th, a snow profile revealed the snow pack had a depth of 130 cm and was entirely isothermal (Figure 4.1). The snowpack structure consisted predominantly of moist rounded crystals, moist melt-freeze crusts, moist poly-crystals, and moist rounded crystals with some water percolation channels.

Sub-freezing air temperatures on November 17th and 18th froze the snow surface creating an upper melt-freeze crust before any new snowfall. As a result, layers of faceted poly-crystals, faceted crystals, and crusts formed below, but not above, the upper crust. A possible explanation for faceting below the upper crust is that it provided a

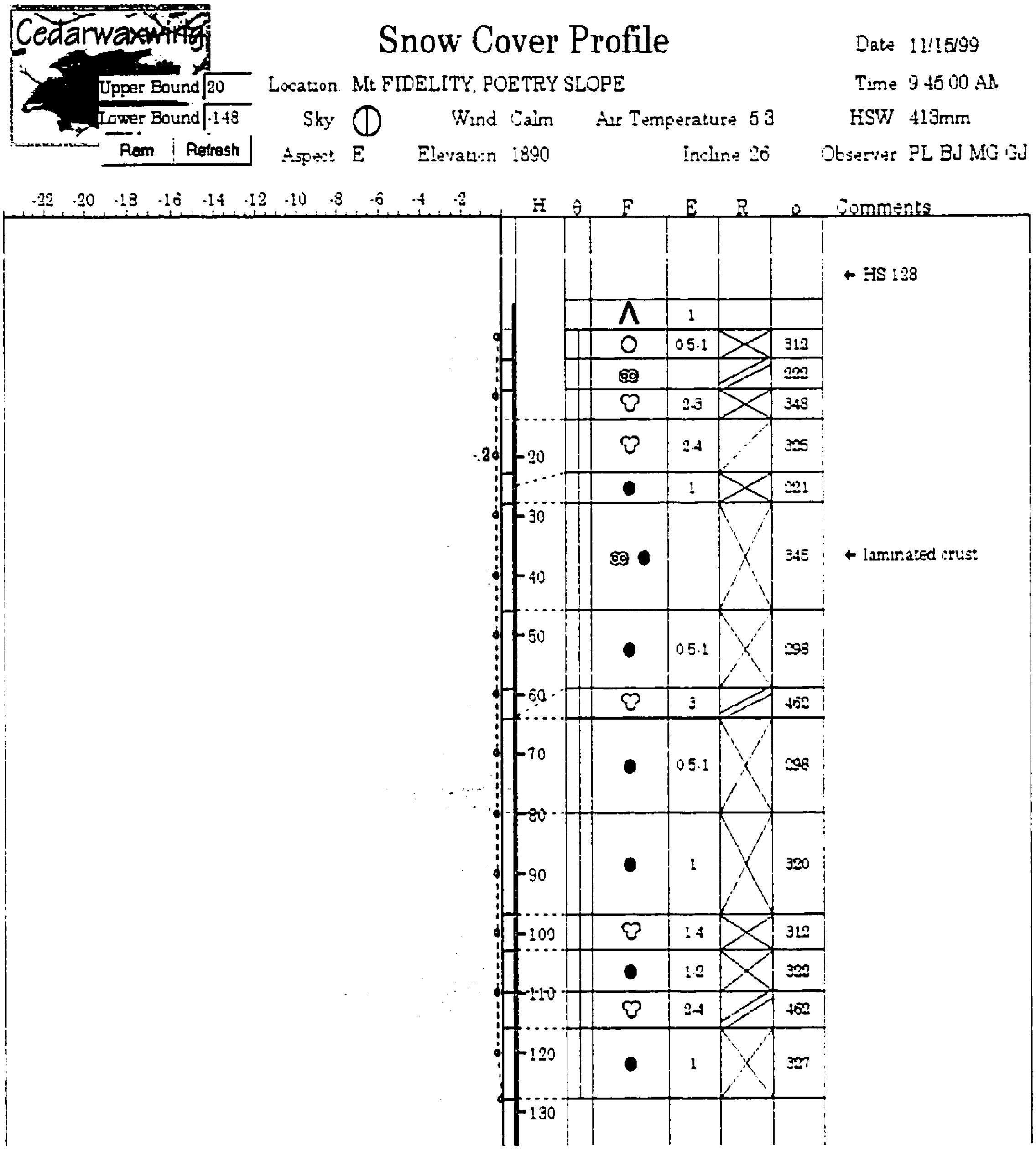

Figure 4.1 Snow profile of an isothermal snowpack at Mt. Fidelity on November 15, 1999. The grain types in column F are described in Colbeck et al. (1990).

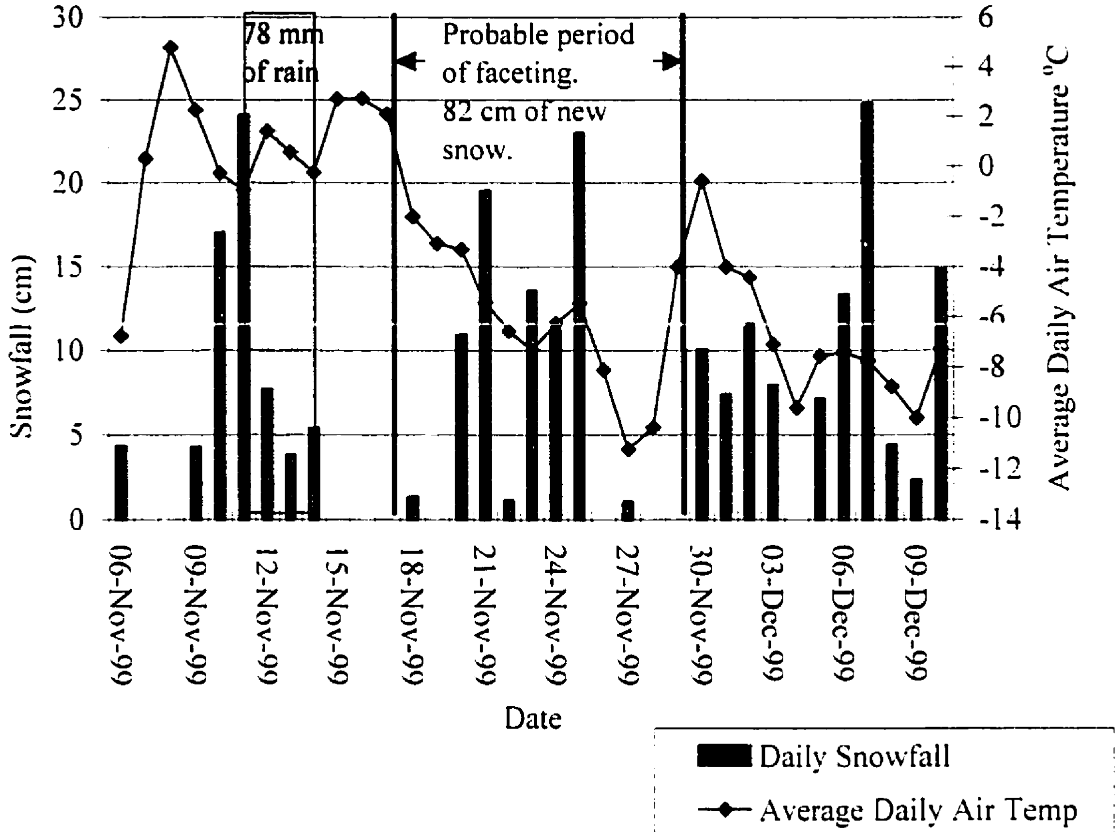

Figure 4.2 Daily snowfall and average daily air temperature at Mt. Fidelity from November 6th to December 10th, 1999.

barrier to the transport of vapour from the moist snow below creating an area of vapour deposition (Seligman, 1936; Colbeck, 1991). In addition, the magnitude of the temperature gradient was prolonged above 10°C/m because of the heat stored in the 130 cm isothermal snowpack. Faceting above the crust was probably limited because the lack of a prolonged temperature gradient of sufficient magnitude. From November 18th to November 30th, relatively warm air temperatures and 82 cm of new snowfall prevented a high temperature gradient from forming (Figure 4.2). After the layer of faceted crystals formed below the crust, subsequent warm snowpack temperatures caused the faceted crystals to round by mid-December.

4.2.2 February 9th, 2000 Layer of Faceted Crystals

On February 7th, 2000 at Bow Summit in the Rocky Mountains, air temperatures were above freezing and 2 mm of surface hoar was reported on the surface. By February 9th snow began to fall and the temperatures dropped to approximately –30°C. During the period from February 9th to February 14th, snow equivalent to 26 mm of water fell. The cool air temperatures caused a temperature gradient with a magnitude greater than 10°C/m that persisted for 12 days, causing the surface hoar to evolve into faceted crystals. These faceted crystals were not reported as rounding facets (class 4c) until March 6th, 2000.

4.3 Sources of Error

4.3.1 Spatial Variability

Measurements of bond diameters may vary spatially within study plots. Wind, rain, or ground cover often cause uneven snow surfaces and layering that result in metamorphism of snow crystals at different rates. To quantitatively express the spatial variability in layers of faceted crystals, the coefficient of variation of shear strength measurements from the shear frame test is used (Jamieson and Johnston, In press).

The February 9th layer of faceted crystals was approximately 1 cm thick and had no visual structural variations. The coefficient of variation for this layer ranged from 0.02 to 0.18, with an average of 0.09. This is below the average value of 0.144 that Jamieson and Johnston (In press) report. Considering the relative uniformity of the February 9th layer and the high degree of experience of shear frame operators, it is assumed the shear frame was placed consistently 2–5 mm above the weak layer and frame placement was not a source of variability. The coefficient of variation of 0.09 mostly represents the spatial variability with a set of shear frame tests because all other sources of variation are negligible. The coefficient of variation of 0.09 is substantially lower than 0.144 and shows the February 9th layer was relatively uniform in the study plot. Therefore bonds collected from this layer, within the study plot, are considered to be relatively uniform over space.

The November 18th layer of faceted crystals had structural variations observed across snowpit walls. Snowpit observations at Mt. Fidelity throughout the winter of

1999-2000, showed the layer of faceted crystals was not level across the snowpit wall due an uneven ground cover and water percolation channels. The average coefficient of variation for the November 18th layer ranged from 0.20 to 0.38, with an average of 0.28. These observed structural variations and the large coefficient of variation indicate the faceted crystals and presumably their bonds were not uniform over space.

4.3.2 Preservation in Iso-Octane

After the faceted snow crystals were collected and placed in iso-octane, they were maintained below freezing in a cooler and transported by skis and vehicle to the Rogers Pass cold laboratory for storage. At the end of January, the crystals were transported to the University of Calgary cold laboratory. In early February the specimens were transported back to Rogers Pass and stored until March 29th, 2000. The iso-octane preserved the crystals and their bonds. Figure 4.3 shows a snowflake that was accidentally preserved with the faceted crystals on Jan 3rd, 2000. It shows no signs of damage.

4.3.3 Measurement Errors

Bond diameter measurements were biased because large bonds were often difficult to measure and non-cylindrical bonds were not measured. Non-cylindrical shaped bonds are not linear on microphotographs, meaning they have sharp corners, or have irregular shapes (Figure 4.4). Visual inspection showed these bonds were often large and connected more than two crystals. Bonds connecting more than two crystals were not considered.

The bond measurements were manually performed on the photographs using Image Pro imaging software (Media Cybernetics, Silver Springs, MD). Typical magnifications were 100x to 1000x. At these magnifications, bonds were clearly visible and measurement errors are negligible.

4.4 Bond Diameter Measurements

Bond diameters and associated statistics are listed for each measurement in Table 4.1 for the February 9th, 2000 time series of faceted crystals at Bow Summit and the November 18th, 1999 time series at Mt. Fidelity. The median bond diameters, the upper and lower quartiles, and maximum and minimum for both layers are shown in Figure 4.4.

Figure 4.3 New snowflake preserved in iso-octane for 3 months.

The median bond diameters of the February 9th series increase slowly over the period of 47 days from 0.070 mm to 0.109 mm. The minimum bond diameters remained nearly constant at 0.022 mm. The maximum bond diameters increased from 0.20 mm to 0.29 mm. The median bond diameters of the November 18th series remained constant at approximately 0.130 mm from December 15th to March 3rd. The minimum bond diameters were approximately constant over time at 0.025 mm. However, the maximum bond diameters increased from 0.26 mm to 0.43 mm.

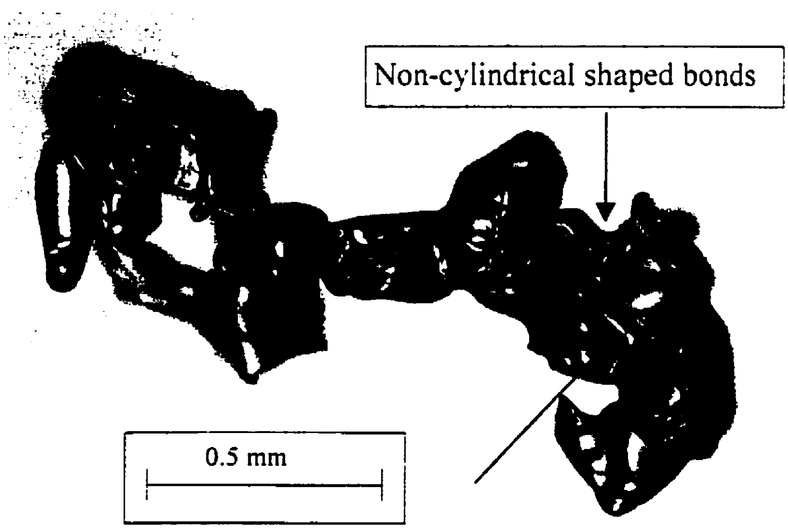

Figure 4.4 Non-cylindrical shaped bonds.

4.5 Snowpack Variables Associated with Bond Growth

4.5.1 Height of Snowpack

The height of snowpack (HS) at a tree-line site, with little wind effect, is associated with the regional snowpack climate (Appendix A).

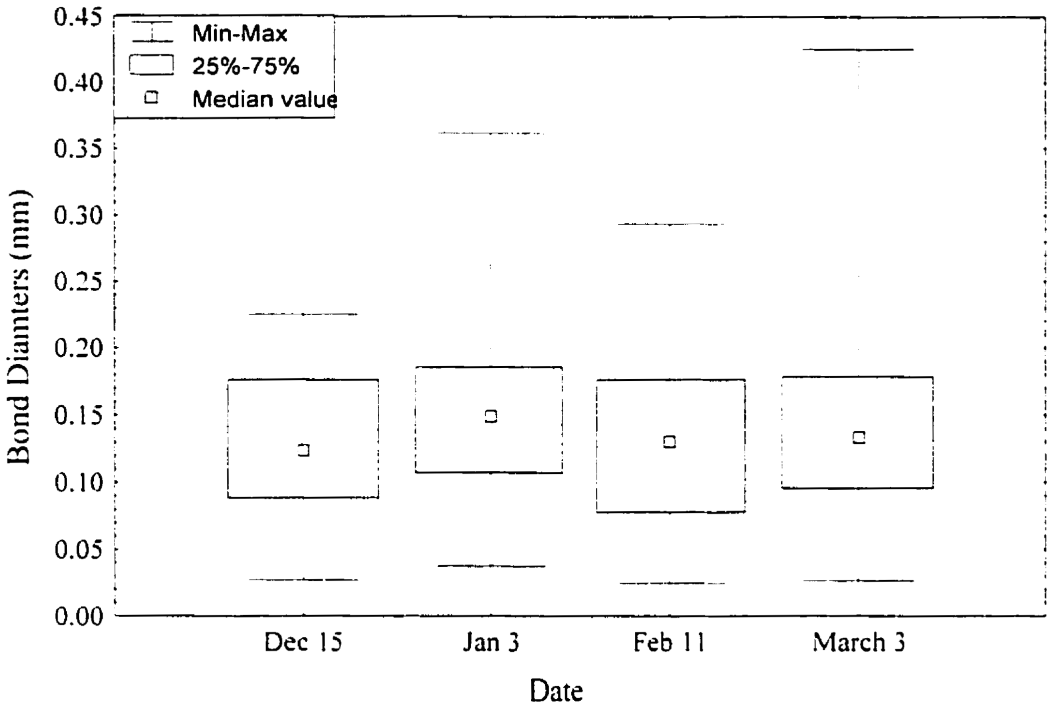

Figure 4.5a November 18, 1999 layer of faceted crystals

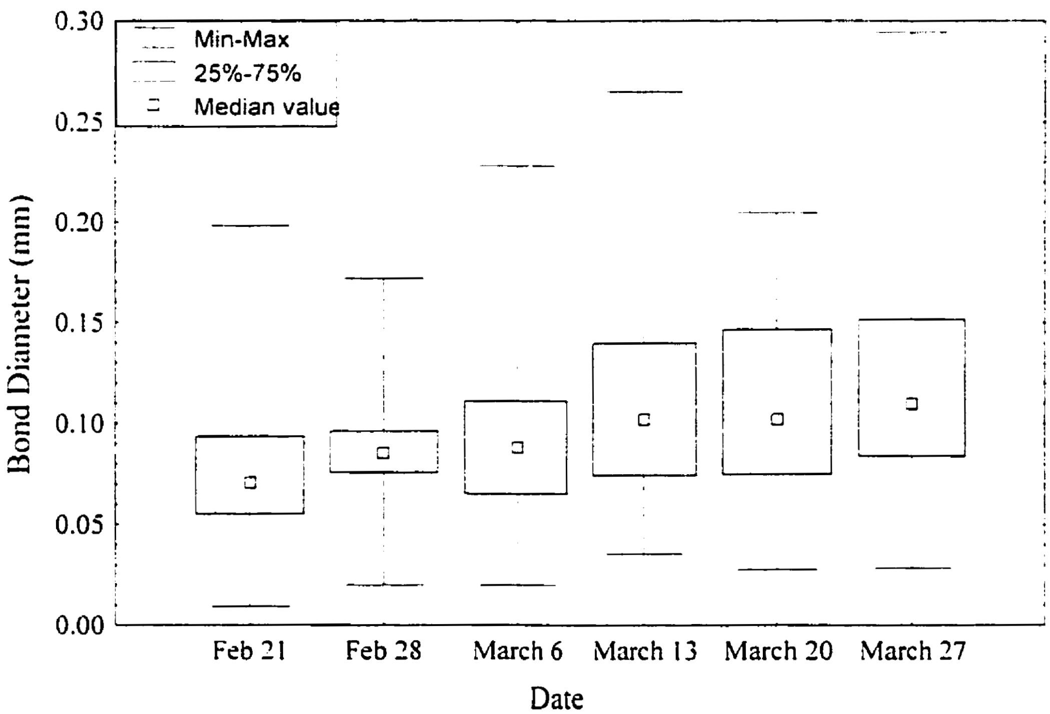

Figure 4.5b February 9, 2000 layer of faceted crystals

Figure 4.5 Time series of bond measurements for the November 18th, 1999 and February 9th, 2000 layers of faceted crystals

Table 4.1 Bond diameter measurements and associated statistics.

February 9, 2000 at Bow Summit, Banff National Park							
Test Date	Layer Age (days)	N	Min (mm)	Max (mm)	Median (mm)	Lower Quartile (mm)	Upper Quartile (mm)
21-Feb	12	37	0.009	0.198	0.071	0.055	0.093
28-Feb	19	38	0.020	0.172	0.085	0.075	0.096
6-Mar	26	110	0.020	0.228	0.088	0.065	0.111
13-Mar	33	59	0.036	0.265	0.102	0.074	0.139
20-Mar	40	52	0.028	0.204	0.102	0.075	0.146
27-Mar	47	92	0.022	0.294	0.109	0.083	0.151
November 18, 2000 at Mt. Fidelity, Rogers Pass, Glacier National Park							
Test Date	Layer Age	N	Min	Max	Median	Lower Quartile	Upper Quartile
	(days)		(mm)	(mm)	(mm)	(mm)	(mm)
15-Dec	27	47	0.027	0.225	0.123	0.088	0.176
3-Jan	46	62	0.037	0.361	0.149	0.107	0.186
11-Feb	85	47	0.024	0.293	0.130	0.078	0.176
3-Mar	106	60	0.026	0.425	0.120	0.095	0.178

Figure 4.6 shows that larger bonds were observed at Mt. Fidelity, where the snowpack is deeper than Bow Summit.

4.5.2 Temperature Gradient and Temperature of the Weak Layer

Faceted crystals subject to warm temperatures and temperature gradients with magnitudes below 10°C/m are expected to round, and the bonds between crystals grow (Colbeck, 1997). The magnitude of the temperature gradient on the February 9th series persisted above the critical value of 10°C/m until it was 19 days old (February 28th), but snowpack temperatures were approximately −13°C, slowing the faceting process. During this period, the median bond diameters increased from 0.076 mm to 0.088 mm. After the temperature gradient weakened and the snowpack temperatures warmed to approximately −4°C, the bonds continued to increase in diameter.

The magnitude of the temperature gradient was not measured for the November 18th series prior to December 15th. However, the formation conditions indicate the high temperature gradient was short lived and the subsequent warm temperatures caused the faceted crystals to undergo rounding metamorphism. When measurements on this layer began on December 15th, the crystals had rounded. Over the next 105 days the crystals continued to round, but the median bond diameters did not increase (Figure 4.4a).

4.5.3 Load

At Mt. Fidelity, the load on the layer of faceted crystals increased at a greater rate than at Bow Summit. The load on the February 9th series at Bow Summit slowly increased throughout the winter from 0.34 kPa to 1.04 kPa (Figure 4.7), while median bond diameters showed an increasing trend. At Mt. Fidelity the loads dramatically increased from 2.87 kPa to 8.37 kPa, but the median bond diameters remained constant.

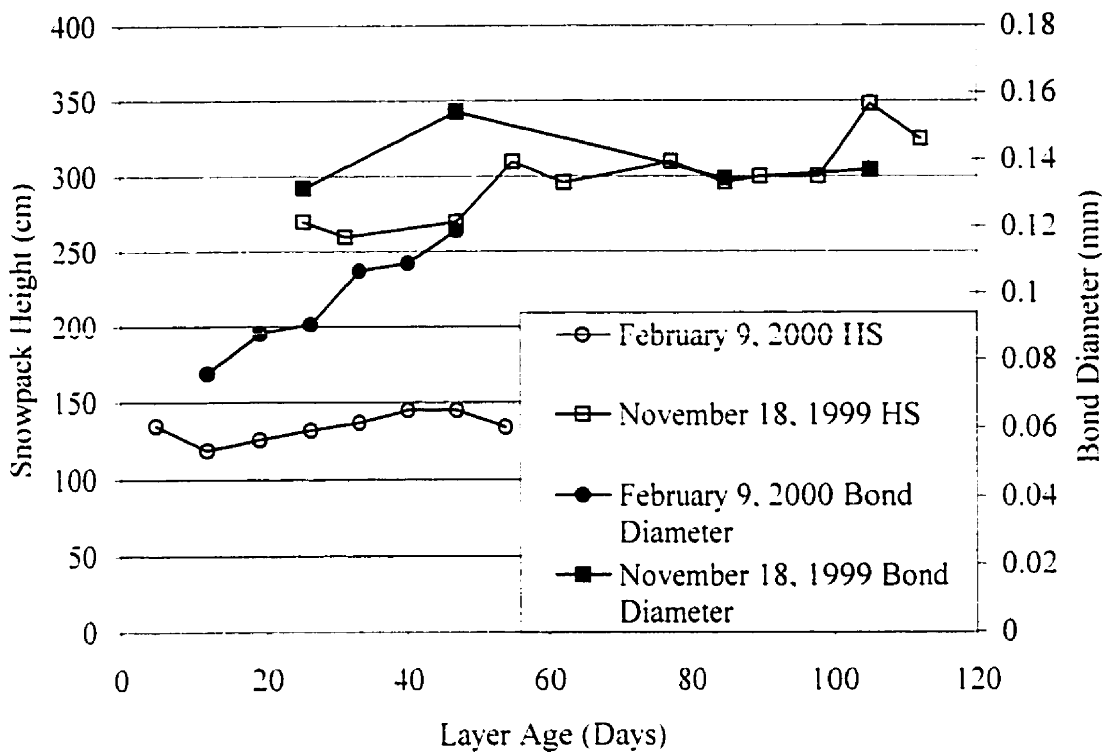

Figure 4.6 Depth of snowpack and bond diameters over time.

4.5.4 Shear Strength

When faceted crystals increase in strength their bonds are expected to grow and / or increase in numbers. At Bow Summit, the February 9th series increased in shear strength from 0.11 kPa to 2.97 kPa over 47 days (Figure 4.8). During this time, the median bond diameter increased as well (Figure 4.5). However, the November 18th, 1999 series did not show an increase in shear strength and increased in bond diameter over time. The median bond diameters of the faceted crystals on December 15th had already reached their maximum diameter of 0.13 mm and the shear strength of the layer was 1.7 kPa. Over the next 105 days the shear strength of the layer increased to 6 kPa.

4.6 Textural Evolution of Faceted Crystal Bonds

A time series analysis of the textural evolution of bonds from the February 9th layer at Bow Summit was conducted from February 14th to March 27th, 2000. On February 21st, the snow crystals were faceted and the bonds were small and mostly

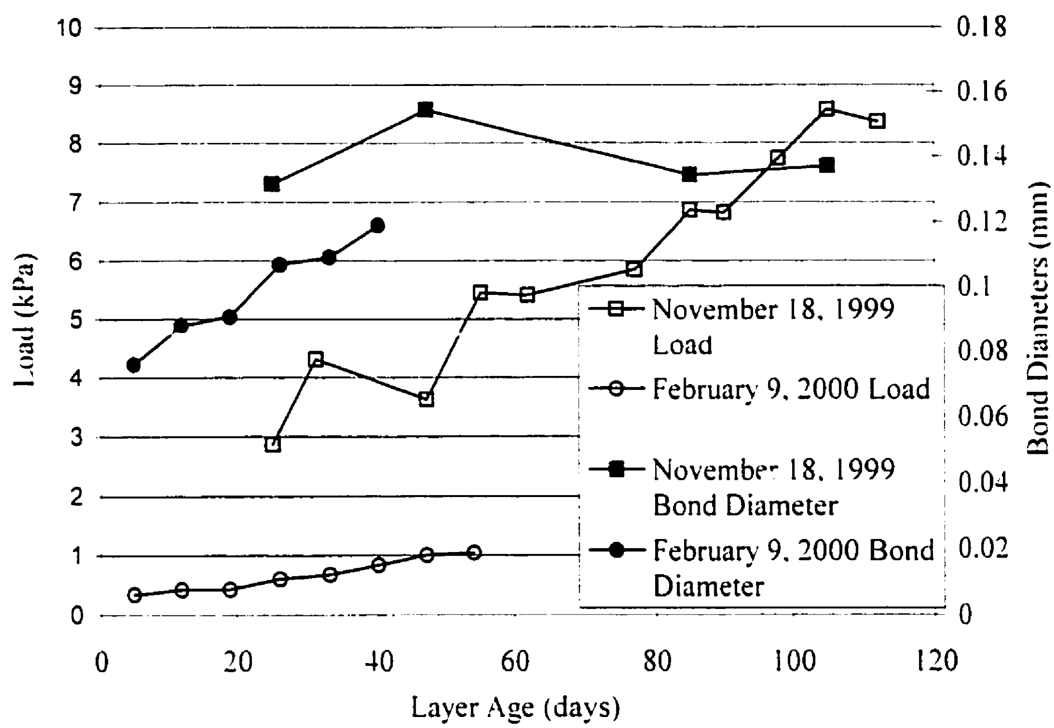

Figure 4.7 Load and median bond diameters over time.

cylindrical in shape (Figure 4.9). The temperature gradient varied between $-31^{\circ}C/m$ to $+8^{\circ}C/m$ from February 9th (layer burial) to February 28th. After February 28th the magnitude of the temperature gradient remained below $10^{\circ}C/m$ and field workers reported the faceted grains rounding by March 6th (Figure 4.10). Still, the bonds of the rounding faceted crystals appeared to be cylindrical in shape. By March 20th, bonds between the faceted crystals had started to change their shape. Instead of the cylindrical shaped bonds, irregular shapes started to appear (Figure 4.11). The last observation of bonds was made on March 27th (Figure 4.11), one week before the layer became isothermal. The faceted crystals in most cases were rounded, and in many cases, the bonds had irregular shapes.

4.7 Measurements of Faceted Crystals

Measurements of crystal size of faceted crystals were based on field observations under low magnification and on photographs taken in the cold laboratory. Both types of crystal size measurements were recorded from the February 9th and November 18th series

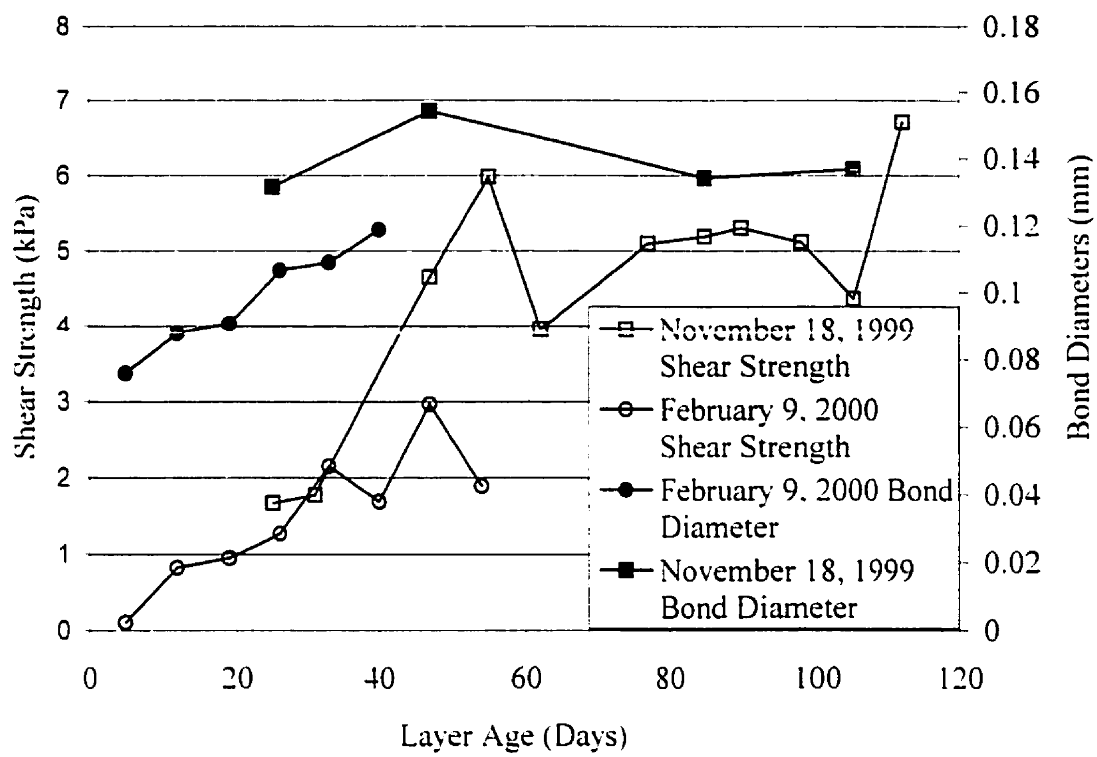

Figure 4.8 Before bonds reach their maximum diameters they increase with shear strength. After they reach their maximum diameter the shear strength continues to increase.

of faceted crystals. Measurements from cold laboratory are considerably smaller than field observations for the February 9th, 2000 layer (Figure 4.13). The magnifications used during image processing were approximately 100 to 1000x, while magnification in the field was 8x or 30x. Under high magnification faceted crystals smaller than 0.25 mm are clearly visible, but under 8x they are not. In some cases under 8x magnification, instead of small individual crystals being measured in the field, groups of crystals are probably being measured as one crystal. In both the cold laboratory and the field, the only observation day that showed reasonable agreement in crystal size was on February 21st. On this date, the crystals were observed under 30x magnification in the field. Although not common practice in this study, it suggests that with higher magnifications smaller crystal sizes are reported.

Despite the difference in cold laboratory and field crystal size measurements the

Figure 4.9 Faceted crystals of the February 9^{th} layer at Bow Summit preserved on February 21^{st}, 2000.

Figure 4.10 Faceted crystals from the February 9^{th} layer at Bow Summit preserved on March 6^{th}, 2000.

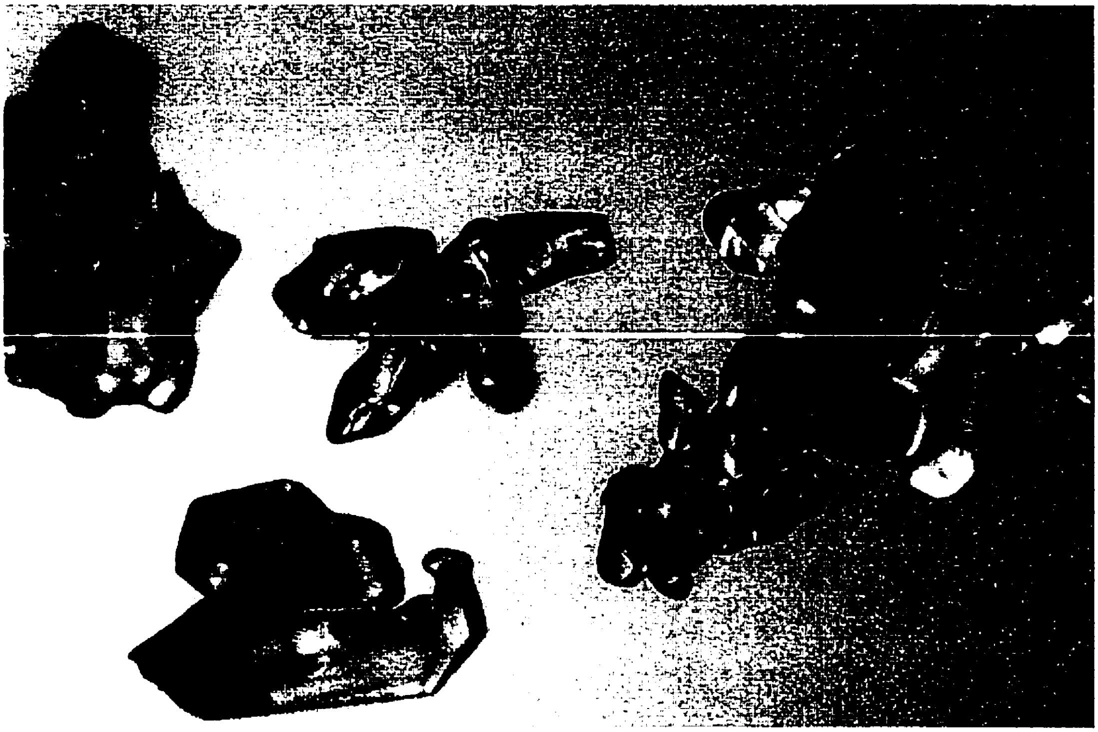

Figure 4.11 Faceted crystals from the February 9^{th} layer at Bow Summit preserved on March 20^{th}, 2000.

Figure 4.12 Faceted crystals from the February 9^{th} layer at Bow Summit preserved on March 27^{th}, 2000.

results from the cold laboratory show the rate at which rounding faceted crystals grow (class 4c) is similar to the growth of rounded crystals (class 3) with a temperature gradient in magnitude less than 10°C/m. The average rate of crystal growth for the February 9th series was 0.0023 mm per day, the November 18th series was 0.0013 mm per day, and Keeler's (1969) series was 0.0025 mm per day.

4.8 Bond Diameter-to-Crystal Size Ratio

The ratio of bond diameter-to-crystal size over time provides a relationship between faceted crystal growth and bond growth. The size of crystals was measured along the long axis using photographs taken in the cold laboratory. The bond diameter-to-crystal size ratios for the February 9th, November 18th, and Keeler's series are shown in Figure 4.14. For the February 9th series, the ratio remained fairly constant at 0.4, which indicates the bonds were growing at the same rates as crystal sizes. For the

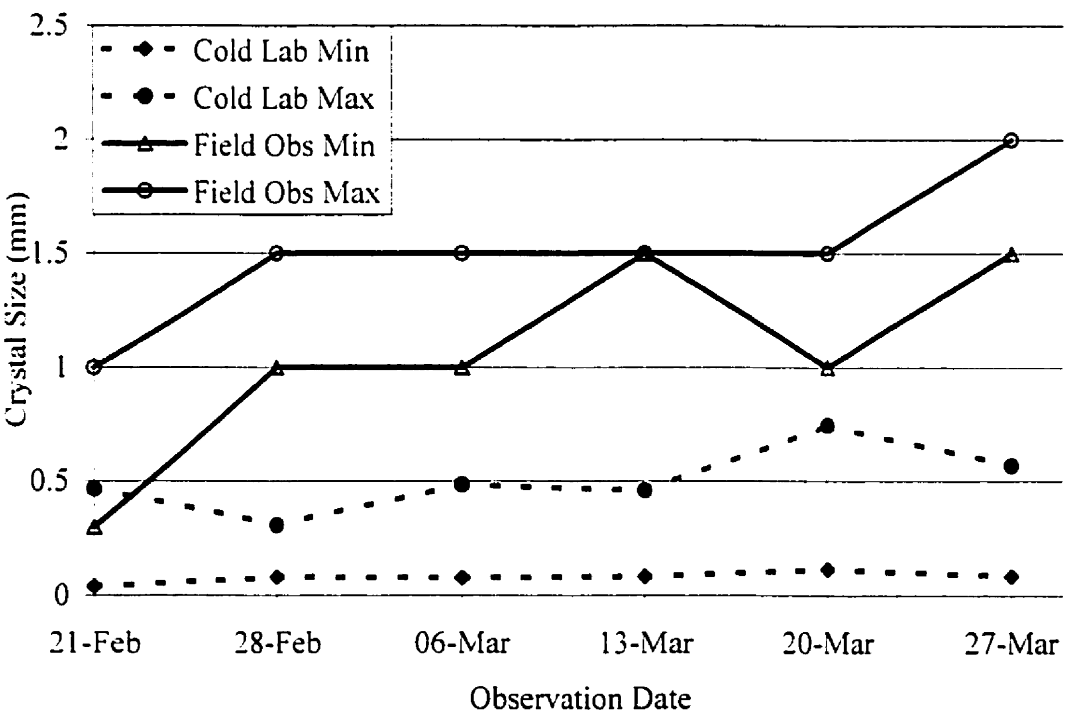

Figure 4.13 The differences between faceted crystal measurements in the field and the cold laboratory are substantially different for the February 9th layer of faceted crystals.

November 18th series, the ratio decreased from just over 0.4 to approximately 0.3. This trend is expected because bond sizes were constant, but the crystal sizes slightly increased. Keeler's bond diameter-to-crystal ratios represents new snow decomposing and rounding metamorphism of rounded grains. As expected, the first two data points show the bonds growing at much faster rates than the crystals. This is because new snow initially decreases in size from either the rounding or faceting processes. However, Keeler's data are hard-to-explain because crystal sizes slightly increase in size from Day 2 to Day 12. After Day 59 Keeler's data showed both the crystal sizes and bond diameters did not substantially increase in size and the bond to crystal ratio remained constant.

4.9 Discussion

In the Columbia Mountains, the November 18th series of faceted crystals reached a maximum bond diameter of approximately 0.13 mm when the layer was 25 days old (December 3rd, 1999). During this period (November 18th to December 3rd) the load increased dramatically, the magnitude of the temperature gradient dropped below 10°C/m, and the snowpack temperatures were warm.

Bond diameter measurements from the February 9th series in the Rocky Mountains showed the bonds continued to increase in diameter (Figure 4.5b). This layer of faceted crystals was subjected to snowpack temperatures below -10°C until February 28th, and temperature above –10°C, until March 27th, 2000. In addition, this layer was subject to small loads (February 9th to March 27th), and a prolonged temperature gradient (February 9th to February 28th). This series warmed to 0°C one week after the last measurement was collected.

In past studies, it has been shown theoretically (Colbeck, 1997) and experimentally (Keeler, 1969) that bond growth is initially rapid and then slows with time. Bonds were not measured for about 1 month after the November 18th, 1999 series formed at Mt. Fidelity and 12 days after the February 9th layer formed at Bow Summit. The bonds of the November 18th series did not grow substantially after the first measurement on December 15th. At Bow Summit, the bonds continued to grow slowly over time until the last measurement on March 27th. The differences in the rate of bond

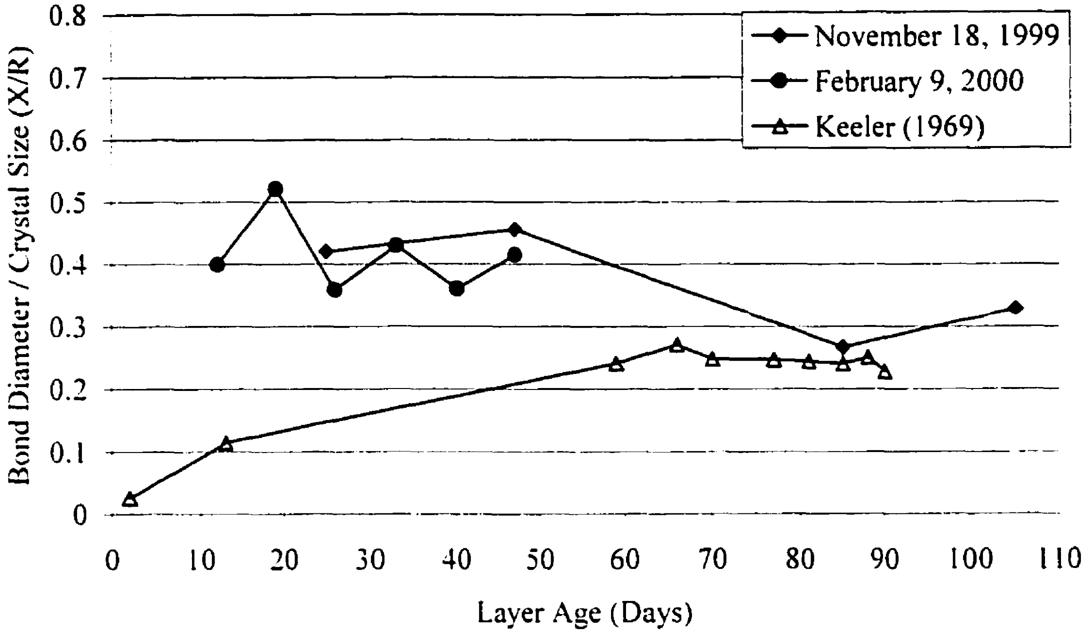

Figure 4.14 Ratio of median bond diameter-to-crystal size.

growth and bond diameters between the November 18th and February 9th series may be associated with differences in load and temperature gradient.

The effects of load and temperature gradient on bond growth between faceted crystals are difficult to separate based on field measurements. The persistence of the high temperature gradient was considerably longer in the Rocky Mountains than in the Columbia Mountains. As a result, the faceted crystals in the February 9th series in the Rocky Mountains were more angular than the crystals in the November 18th series in the Columbia Mountains. Over the period of a week, from February 21st to February 28th, when the magnitude of the temperature gradient exceeded 10°C/m, bond diameters from the February 9th grew and the layer gained strength. Bond measurements (Table 4.1, Figure 4.5b) do not indicate a difference in growth rate during faceting (February 21st to February 28th) and rounding (February 28th to March 27th) metamorphism. This indicates that load has a greater affect on bond diameters than temperature gradient.

The rate at which faceted crystals gain strength is non-linearly dependent on bond diameter (Keeler, 1969). Figure 4.15 shows that the shear strength to bond area ratio increases at approximately the same rate for both the February 9th and November 18th

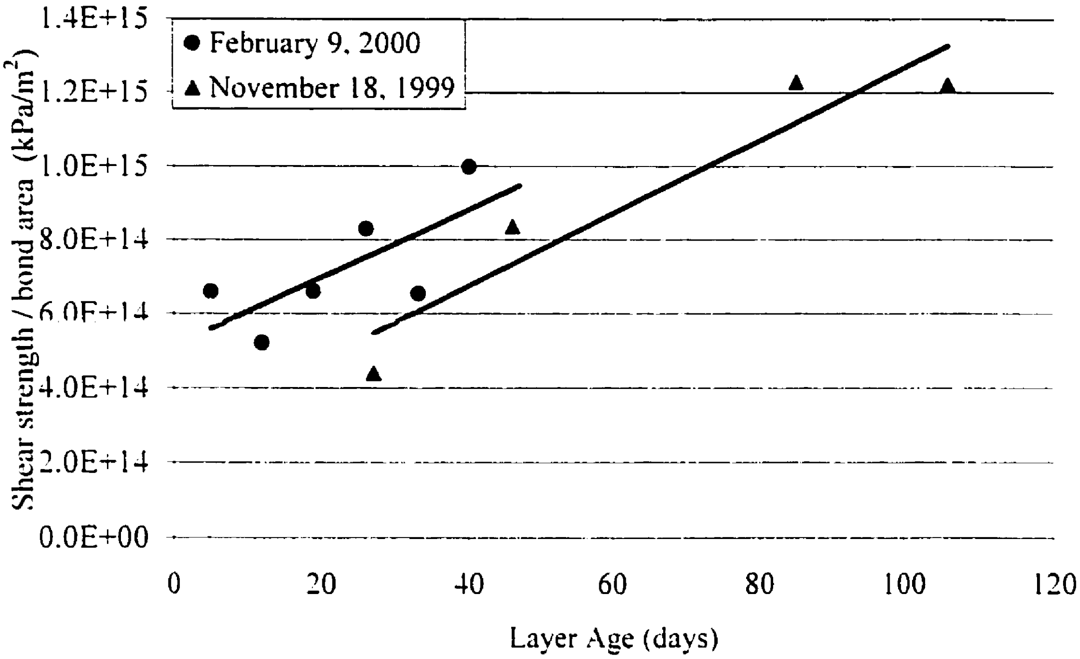

Figure 4.15 Shear strength-to-bond area ratio.

series. Figure 4.8 shows that the shear strength for the November 18th series increases over time and Table 4.1 and Figure 4.5 show that median bond diameter for this series does not increase over time. This relationship, and the fact that minimum bond diameters for both layers remain constant and have similar values throughout the season, suggests that new bonds are forming. Bonds are expected to grow over time and not remain constant in size. The average minimum bond diameters for the February 9th and November 18th series are 0.024 mm and 0.028 mm, respectively. The formation of new bonds implies the number of bonds per crystal is increasing with time. The number of bonds per unit volume could not be measured to verify the number of bonds per crystal. Keeler did measure bonds per grain for rounded snow and found a substantial increase initially, but then the number of bonds per grain reached a steady value of 2.61. For dense rounded snow this makes sense because of the relatively small pore spaces and spherical shapes. In faceted snow, pore spaces are relatively large and the crystals are relatively large and angular. New bonds may continue forming due to densification associated with load and an increase in bonds per crystal.

The non-cylindrical shaped bonds observed on March 20th and March 27th may have resulted from two bonds growing together (Figure 4.16). Stresses at the bonds are as much as two orders of magnitude higher than the bulk stresses in the snowpack (Brown and Edens, 1991). At level study plots, load is the primary applied stress because creep does not exist. Consequentially, much of the snowpack deformation occurs at the bonds (Brown and Edens, 1991) and is due to load.

Any type of dry snow crystal has the potential to become faceted given a sufficient temperature gradient and time. The crystal type that is undergoing faceting and its size may be important parameters that determine the size of bond diameters and their shapes. In recent studies, attention has focused on near-surface faceting of new snow crystals. Not all layers of faceted crystals form from new snow. For example, consider poly-crystals and surface hoar under similar temperature gradients, temperatures, and depths. Surface hoar crystals characteristically have small bonds at their base, large bonds near their top, and few bonds within the layer connecting the crystals due to their shape (Jamieson and Schwiezer, 2000). Frozen poly-crystals have strong intergranular bonds due to the melt-freeze process. When faceting starts, the poly-crystals already have large strong bonds. Consequently, it will take more time for the poly-crystals to become as weak as the surface hoar under faceting metamorphism.

Figure 4.16 Two bonds possibly growing together.

4.10 Conclusions

The shear frame test was used to estimate the spatial variability of two layers of faceted crystals and their bonds. The November 18th layer was shown to have spatial variability. Despite the spatial variability, its bond diameters were constant from the first measurement date when it was 25 days old until the last measurement when it was 105 days old. The February 9th layer did not have substantial spatial variability and its bonds continually increased in diameter, but were smaller than the bonds in the November 18th layer.

The continental and intermountain snowpack climates have temperature gradients and loads of different magnitudes. These differences are used to study the effect of temperature gradient and load on bond growth. Two time series measurements of layers of faceted crystals were observed. There was no substantial difference in the bond growth rate during faceting and during rounding metamorphism, indicating load was the dominant variable. At Mt. Fidelity, which has an intermountain snowpack climate, the load and bond diameters increased faster than at Bow Summit, which has a continental snowpack climate.

The shear strength of both time series of faceted crystals increased at a faster rate than bond area increased. The fact that minimum bond diameters for both layers remained constant over time, suggests that new bonds are forming. Load may be responsible for these new bonds. Keeler (1969) observed a constant number of bonds per crystal during rounding metamorphism, indicating that the structure of the snow remained unchanged. In contrast, faceted crystals have relatively large pore spaces that may allow for further re-arrangement and crystal contact over similar periods.

The shapes and evolution of bonds between faceted crystals may influence the rate at which faceted crystals gain strength. A change in bond shape from the February 9th, 2000 layer at Bow Summit was observed. Initially, the bonds appeared cylindrical but as the layer aged, large bonds with irregular shapes, and bonds between more than two crystals started to appear.

5 Snowpack Properties that Influence the Shear Strength of Layers of Faceted Crystals in the Intermountain and Continental Snowpack Climates

5.1 Introduction

During the winters of 1993-2000 in the Columbia Mountains and the Rocky Mountains of southwestern Canada, over 100 strength measurements from 16 time series of faceted crystals (Table 5.1) were made. Each time series consists of measurements at 3 to 8 day intervals of a particular layer at a specific study plot. In this chapter, rank correlations are used to establish relationships between the shear strength of layers of faceted crystals and snowpack variables.

Of the 16 time series, the eight with the most complete data sets are used for regression analysis to determine relationships between shear strength and time. Rank correlations were also performed on the regression coefficient Σ_1, which represents the series initial strength. Table 5.1 lists the locations, number of data points, and the identification dates for layers of faceted crystals in each climate region.

5.2 Rank Correlations

Figure 5.1 shows the time profiles of shear strengths for individual layers of faceted crystals in the intermountain and continental snowpack climates. Two different strength trends are apparent. Layers of faceted crystals in continental snowpacks are considerably weaker than in intermountain snowpacks.

Some snowpack properties are referred to as predictor variables because they might be associated with shear strength (Jamieson and Johnston, 1999) of layers of faceted crystals. Predictor variables are directly measured snowpack properties (Table 5.2) or calculated combinations of these measurements, (such as TA / HS). Some predictor variables, (such as Δload / Δt), represent the average rate of change over the interval Δt which ranges from 3 to 8 days. In addition, predictor variables, (such as Load_{avg}), represent the average of measurements, at the start and end of an interval.

To assess which snowpack properties are associated with changes in shear strength, the predictor variables were rank correlated with three response variable

Table 5.1 Time series of layers of faceted crystals in the continental and intermountain climates.

Intermountain snowpack climate			Continental snowpack climate		
Location	*Identification Date*	*N*	*Location*	*Identification Date*	*N*
Mt. St. Anne *	November 20, 1997	9	Vermont Air Box *	January 17, 1997	7
Mt. St. Anne	December 24, 1998	5	Vermont Study Plot *	January 18, 1997	7
Mt. St. Anne *	November 22, 1996	13	Bow Summit	November 18, 1999	10
Mt. Fidelity	November 18, 1999	9	Bow Summit	February 9, 2000	8
			BB-Lodge *	December 5, 1993	11
			BB-Lodge*	December 15, 1993	5

* Observations of faceted crystals prior to 1998 were performed by others.

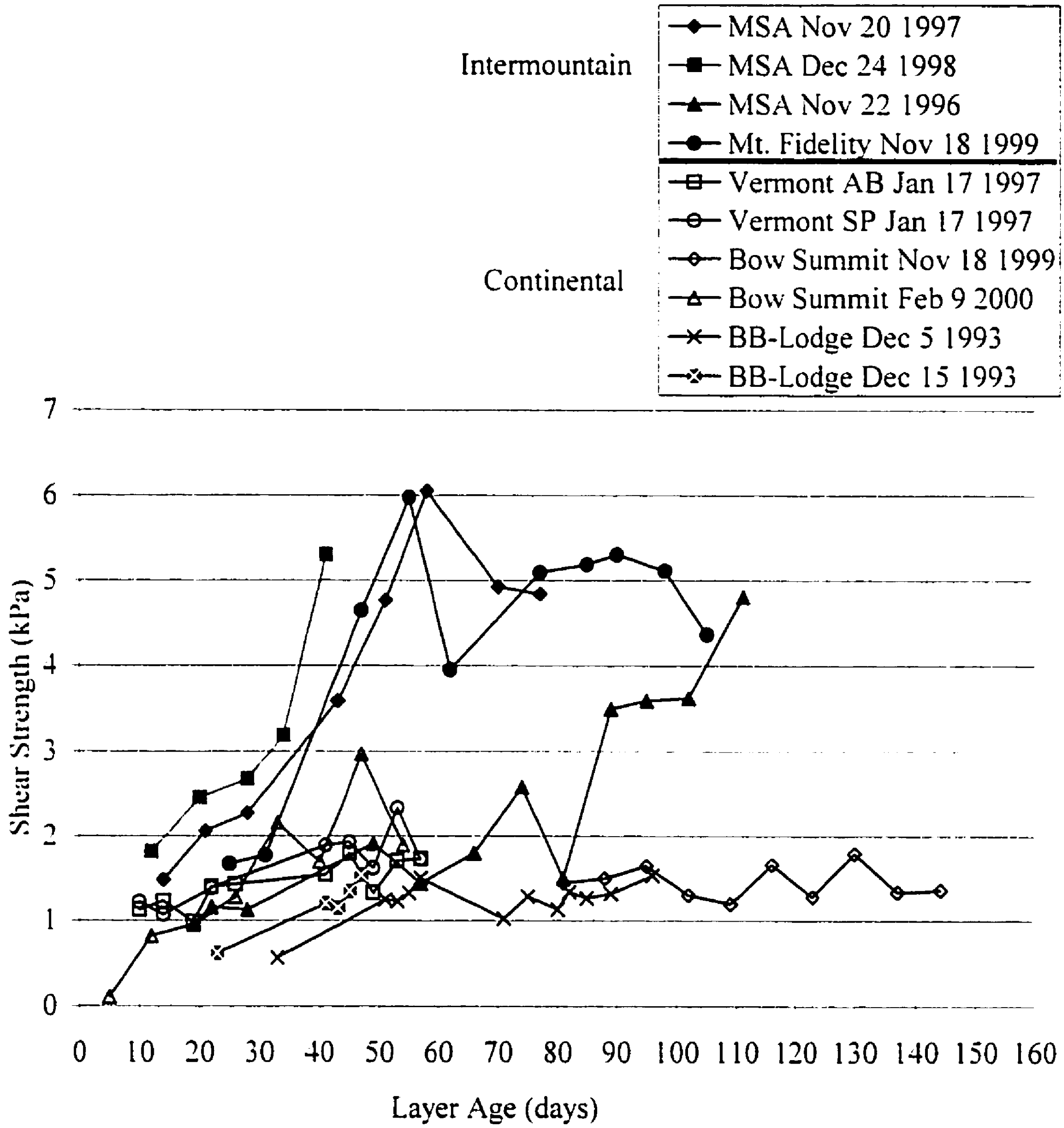

Figure 5.1 Shear strength of layers of faceted crystals over time.

categories:

1) continental snowpack climate
2) intermountain snowpack climate
3) both snowpack climates combined.

The response variable Σ_{cont} represents shear strength in continental snowpack

Table 5.2. List of predictor factors and response variables.

Predictor Variables	
Load	Weight of the overlying snow per unit horizontal area (kPa)
ρ_{slab}	Averaged slab density of snow above faceted layer (kg/m^3)
H	Thickness of the slab. the depth of snow over the faceted layer (cm)
R_{wl}	Hardness of the faceted layers measured with the hand hardness test (CAA. 1995)
HS	Depth of snowpack (cm)
Age	Number of days since the faceted layer formed
TG	Temperature gradient measured across faceted layers. Measurements are taken 5 cm above and 5 cm below the faceted layer (°C/m)
Twl	Temperature of the faceted layer (°C)
L	Thickness of the faceted layer (cm)
TA/HS	Snowpack temperature gradient (°C/m)
Emax	Maximum length of faceted crystals (mm)
Emin	Minimum length of faceted crystals (mm)
TA	Air temperature (°C)
Response Variables	
Σ	Shear strength of layers of faceted crystals measured with the shear frame (kPa)
Σ_{inter}	Shear strength of layers of faceted crystals measured with the shear frame in inter-continental snowpack climates (kPa)
Σ_{cont}	Shear strength of layers of faceted crystals measured with the shear frame in continental snowpack climates (kPa)

climates. Σ_{inter} for intermountain snowpacks, and Σ for shear strength in both snowpack climates combined.

The frequency distribution of all shear strength measurements is shown in Figure 5.2. These strength measurements failed the Kologorov-Smornov test of

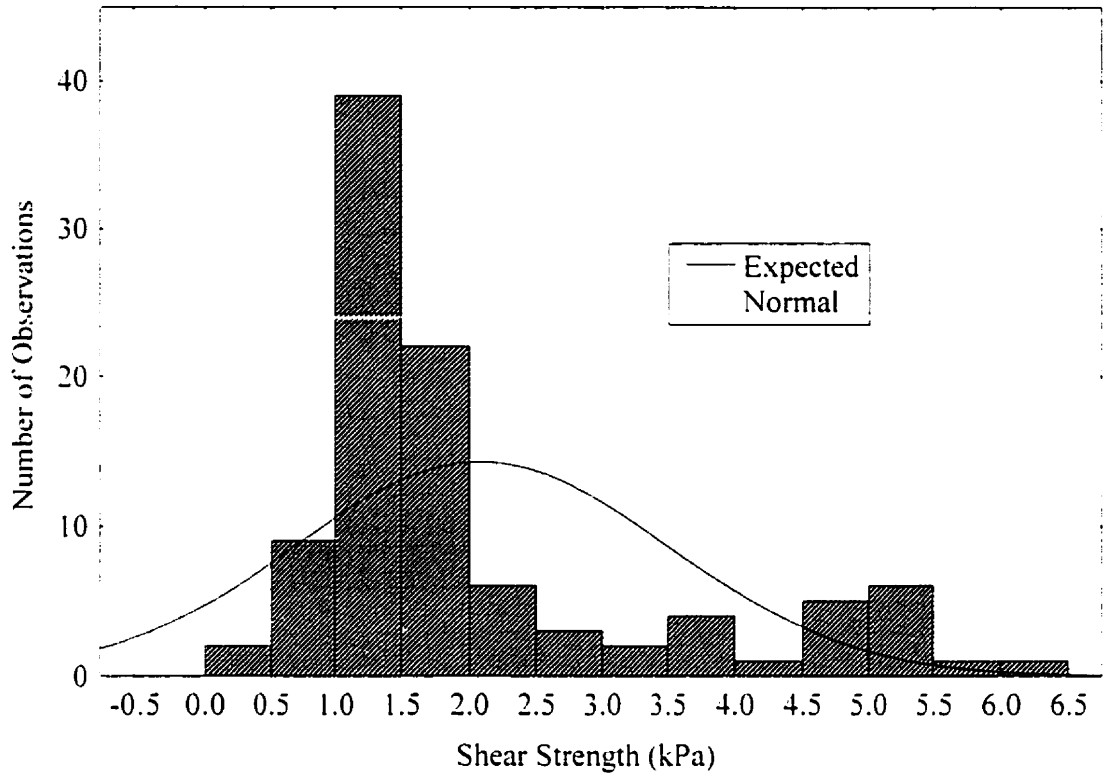

Figure 5.2 Distribution of measured shear strength values of layers of faceted crystals.

normality ($d = 0.258$, $p < 0.01$). Since Pearsons correlations require the response and predictor variables to be normally distributed, Spearman rank correlations are used to assess the relationship between shear strength and predictor variables. This correlation technique only requires the data to be at least on an ordinal scale. The Spearman statistic R ranges from −1 to 1 with −1 and 1 being a perfect correlation and 0 indicating no correlation. The significance level *p* represents the reliability of a correlation. A *p*-level of 0.05 indicates there is a 5% probability of getting correlations between predictor variables and response variables in uncorrelated data (Statistica, 1999).

A serial correlation is a measure of the relationship between past observations and present observations of a variable (Chatfield, 1980, pg 23-32). Serial correlations are present in the data and cause p-values to be underestimated. In this study, statistical analysis of the serial correlations is difficult because of the limited number of data in individual time series and irregular time intervals. Physical arguments between strength and significant predictor variables are used to explain the correlations.

Table 5.3 Spearman rank correlations between strength and snowpack variables. Correlations are significant if $p < 5.0E\text{-}2$ (Shown in bold)

Correlations with shear strength (Σ) for faceted layers within both continental and intermountain snowpacks.				Correlations with shear strength (Σ_{inter}) for faceted layers within continental snowpacks.				Correlations with shear strength (Σ_{cont}) for faceted layers within continental snowpacks.			
	N	R	*p-level*		N	R	*p-level*		N	R	*p-level*
Load	**101**	**0.71**	**7.5E-17**	**H**	**48**	**0.82**	**1.0E-12**	**ρ_{slab}**	**53**	**0.52**	**5.7E-05**
Load$_{avg}$	**86**	**0.73**	**1.1E-15**	**Load**	**48**	**0.79**	**2.1E-11**	**R$_{wlavg}$**	**44**	**0.48**	**9.1E-04**
ρ_{slab}	**101**	**0.68**	**3.7E-15**	**H$_{avg}$**	**40**	**0.83**	**3.8E-11**	**Load$_{avg}$**	**46**	**0.45**	**1.9E-03**
H	**101**	**0.68**	**6.5E-15**	**Age**	**48**	**0.76**	**2.8E-10**	**Load**	**53**	**0.42**	**1.9E-03**
H$_{avg}$	**86**	**0.72**	**7.7E-15**	**Load$_{avg}$**	**40**	**0.79**	**1.1E-09**	**R$_{wl}$**	**52**	**0.40**	**3.2E-03**
R$_{wl}$	**97**	**0.61**	**4.8E-11**	**ρ_{slab}**	**48**	**0.71**	**2.2E-08**	**H$_{avg}$**	**46**	**0.41**	**4.4E-03**
R$_{wlavg}$	**79**	**0.64**	**1.8E-10**	**Twl**	**47**	**0.68**	**1.3E-07**	**H**	**53**	**0.34**	**1.3E-02**
HS$_{avg}$	**85**	**0.61**	**4.4E-10**	**Twl$_{avg}$**	**40**	**0.70**	**4.9E-07**	**Age**	**53**	**0.32**	**2.1E-02**
HS	**101**	**0.57**	**4.7E-10**	**HS**	**48**	**0.64**	**1.2E-06**	**Emax**	**53**	**0.29**	**3.5E-02**
Age	**101**	**0.44**	**5.0E-06**	**R$_{wl}$**	**45**	**0.64**	**1.8E-06**	**HS$_{avg}$**	**45**	**0.30**	**4.8E-02**
TG	**101**	**0.39**	**5.8E-05**	**HS$_{avg}$**	**40**	**0.66**	**3.3E-06**	TA	53	0.25	7.2E-02
TG$_{avg}$	**85**	**0.39**	**2.2E-04**	**R$_{wlavg}$**	**35**	**0.69**	**5.0E-06**	ΔTwl/Δt	45	-0.26	8.7E-02
TG$_{avg}$/Twl$_{avg}$	**85**	**-0.31**	**4.4E-03**	**TG**	**48**	**0.33**	**2.2E-02**	HS	53	0.23	9.8E-02
Twl	**99**	**0.28**	**4.9E-03**	**Emax**	**46**	**0.32**	**3.2E-02**	TA/HS	45	0.24	1.1E-01
L	**92**	**-0.23**	**3.0E-02**	**ΔTG/Δt**	**40**	**-0.33**	**3.6E-02**	TA$_{avg}$	46	0.24	1.1E-01
Twl$_{avg}$	**85**	**0.23**	**3.6E-02**	ΔTwl/Δt	40	-0.30	6.1E-02	ΔHS/Δt	45	0.18	2.3E-01
TA/HS	**80**	**0.23**	**3.7E-02**	Emin	46	0.27	6.6E-02	L	49	-0.16	2.6E-01
ΔH/Δt	85	0.21	5.2E-02	L	43	-0.27	7.6E-02	TG	53	0.16	2.6E-01
ΔTwl/Δt	85	-0.20	7.3E-02	TG$_{avg}$	40	0.28	7.7E-02	ΔH/Δt	45	0.16	2.9E-01
Δload/Δt	85	0.19	8.3E-02	TA$_{avg}$	35	-0.19	2.7E-01	Twl	52	0.12	3.8E-01
Emax	99	0.17	9.4E-02	TG$_{avg}$/Twl$_{avg}$	40	0.18	2.7E-01	TG$_{avg}$	45	0.13	4.1E-01
TAavg	81	-0.17	1.3E-01	ΔHS/Δt	40	-0.17	2.8E-01	TG$_{avg}$/Twl$_{avg}$	45	-0.12	4.2E-01
ΔTA/Δt	80	-0.16	1.5E-01	ΔTA/Δt	35	-0.11	5.1E-01	Twl$_{avg}$	45	0.12	4.5E-01
ΔTG/Δt	85	-0.07	5.3E-01	TA	45	-0.10	5.1E-01	ΔTA/Δt	45	-0.10	5.3E-01
ΔHS/Δt	85	0.06	5.7E-01	TA/HS	35	0.03	8.5E-01	Δload/Δt	45	0.08	6.1E-01
Emin	99	0.05	5.9E-01	ΔH/Δt	40	0.00	9.8E-01	Emin	53	0.02	8.7E-01
TA	98	-0.05	6.1E-01	Δload/Δt	40	0.00	1.0E+00	ΔTG/Δt	45	0.02	8.9E-01

5.3 Results and Discussion of Spearman Rank Correlations

Snowpack variables that correlated with shear strength of layers of faceted crystals are listed in order of their statistical significance in Table 5.3. In continental snowpacks shear strength Σ_{cont} significantly ($p < 0.05$) correlated with 10 snowpack variables. In intermountain climates, shear strength Σ_{inter} significantly correlated with 15 snowpack variables. When the data were not partitioned into a climatic region, Σ correlated with 17 snowpack variables. Physical arguments for the correlations with shear strength are supported by cross-correlations between the predictor variables shown in Table 5.4 and discussed in Sections 5.3.1 to 5.3.9.

5.3.1 Snowpack Depth (HS_{avg})

Snowpack depth strongly correlates with shear strength Σ_{inter} in the intermountain climate (p = *1.2E-6*, Table 5.3). This is not surprising since depth of snow (HS) positively correlates with load (p = *8.9E-12*, Table 5.4), TG (p = *8.2E-4*, Table 5.4), and Twl (p = *4.8E-15*, Table 5.4). Deep snowpacks have large loads, temperature gradients of low magnitude, and warm temperature of the weak layer. The depth of snow (HS) weakly correlates with shear strength Σ_{cont} in the continental snowpack climate (p = *0.04*,

Table 5.4 p-values for cross-correlations between snowpack factors. Correlations are considered significant if p < 5.0E-2.

	Age	HS	Load	H	R_{wl}	TG	Twl	Emax
HS	3.1E-01	-	-	-	-	-	-	-
Load	**8.9E-12**	**3.1E-14**	-	-	-		-	-
H	**2.7E-07**	**6.1E-20**	**1.0E-27**	-	-	-	-	-
R_{wl}	**1.9E-02**	**2.0E-06**	**6.2E-08**	**3.9E-07**	-	-	-	-
TG	**8.2E-04**	**1.3E-05**	**2.1E-08**	**5.0E-09**	**9.0E-03**	-	-	-
Twl	**4.8E-15**	5.9E-01	**3.5E-06**	**1.2E-04**	2.9E-01	**4.7E-05**	-	-
Emax	**8.0E-14**	7.4E-01	**9.7E-04**	**9.6E-03**	3.1E-01	4.5E-01	**2.1E-04**	-
ρ_{slab}	**9.8E-19**	**2.0E-04**	**1.5E-22**	**1.6E-12**	**3.9E-06**	**1.8E-03**	**1.7E-07**	**1.1E-04**

Table 5.3).

5.3.2 Load

Load has a greater positive correlation with the shear strength of layers of faceted crystals in the intermountain climate ($p = 2.1E\text{-}11$, Table 5.3) than the continental climate ($p = 1.9E\text{-}3$, Table 5.3). Load also correlates more significantly with shear strength Σ (both climates) ($p = 7.5E\text{-}17$, Table 5.3) than any other snowpack property. These three

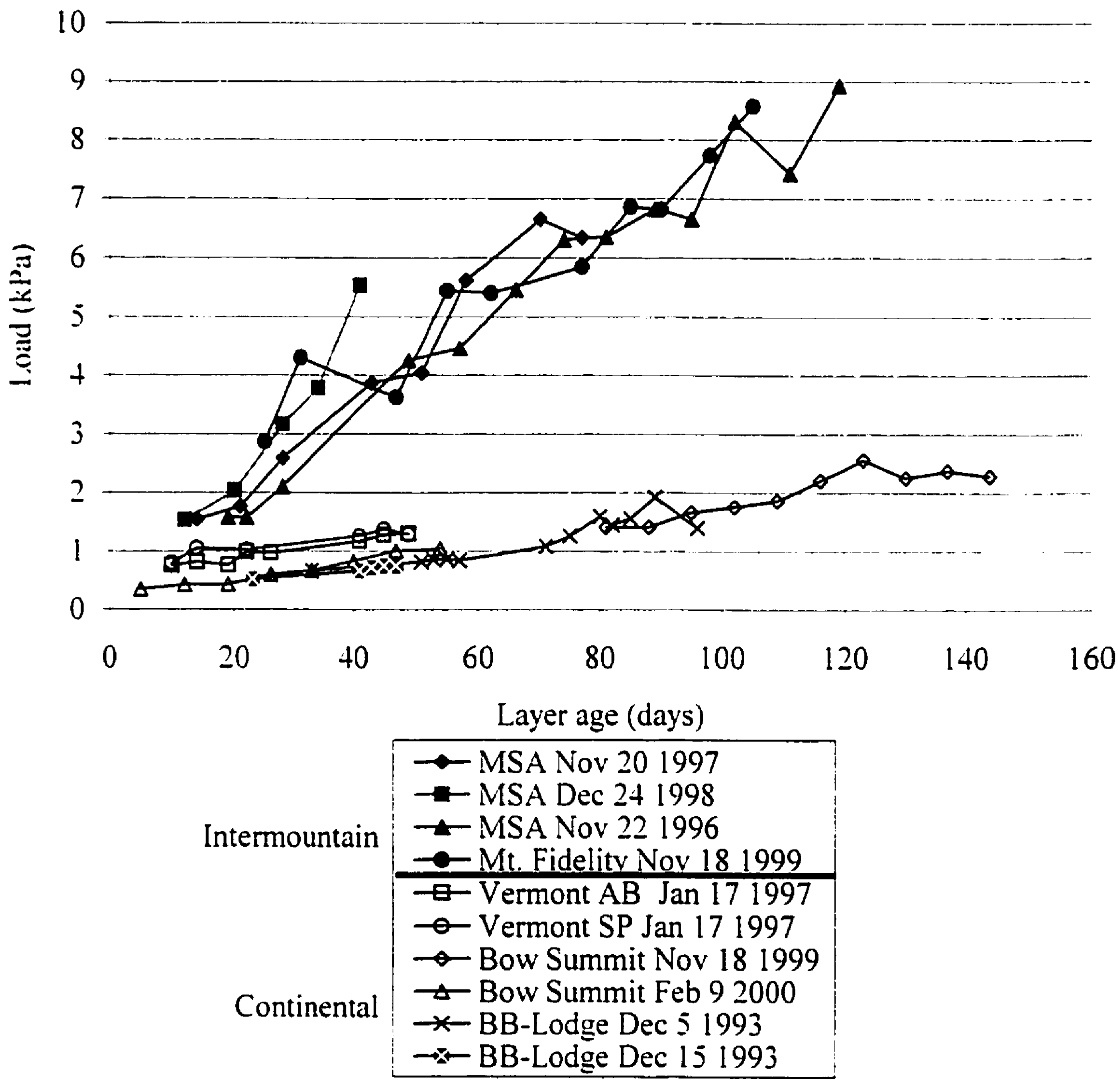

Figure 5.3 Load over time is clearly different in the continental and intermountain climates.

correlations show the strong dependence of shear strength on load, or at least until maximum shear strengths are reached.

In each climate, maximum shear strengths are reached after different periods of time. Maximum shear strength was reached after approximately 50 days in the continental climate and approximately 70 days in the intermountain climate (Figure 5.1). After 70 days the load in the intermountain climate averaged 6.24 kPa. After 50 days the load in the continental climate averaged 0.94 kPa (Figure 5.3). After maximum shear strengths are reached, loads continued to increase, but other factors (such as layer temperature, previous temperature gradient, and crystal size) may prevent further increases in strength.

5.3.3 Slab Thickness (H)

Slab thickness strongly correlates with shear strength Σ_{inter} (p = 1.0E-4, Table 5.3) for intermountain snowpacks and Σ_{cont} (p = 4.3E-3, Table 5.3) for continental snowpacks. Slab thickness is expected to affect strength because thicker slabs are associated with larger loads (p = 1.0E-27, Table 5.4), warmer temperature of the weak layer (p = 1.2E-4, Table 5.4), and temperature gradients with low magnitude (p = 5.0E-9, Table 5.4).

5.3.4 Slab Density (ρ_{slab})

Shear strength positively correlates with slab density ρ_{slab} (Table 5.3) for all three climatic categories (Σ_{inter}, Σ_{cont}, Σ). This is not surprising since slab density is a function of slab thickness (p = 1.7E-12, Table 5.4) and load (p = 1.5E-22, Table 5.4).

5.3.5 Hand Hardness (R_{wl})

Hand hardness roughly measures the resistance to penetration. It is an index of strength. Shear strength positively correlated with hand hardness in all three climatic categories (Σ, Σ_{inter}, Σ_{cont}).

In this study, layers of faceted crystals in continental snowpacks reached maximum hand hardnesses of 1F+, but rounded facets in the intermountain snowpack reached P+. The difference in hand hardnesses for each snowpack climate is primarily due to load ($p = 6.2E\text{-}8$, Table 5.4) because load contributes to bond formation (Section 4.7).

5.3.6 Temperature of the Weak Layer (Twl)

The temperature of the weak layer (Twl) positively correlates with shear strength Σ_{inter} in intermountain snowpacks ($p = 1.3E\text{-}7$, Table 5.3), but not in continental snowpacks. In the deep snowpacks of the intermountain snowpack climates, temperatures gradually increase (Figure 5.4) over the winter, promoting rounding of the faceted crystals and bond growth. The temperature of the weak layer (Twl) correlated with slab thickness (H) ($p = 1.2E\text{-}4$, Table 5.4) and Age ($p = 4.8E\text{-}15$, Table 5.4). This suggests that as layers age, they are buried deep and are thereby well insulated from cold air.

In the continental snowpack climate of the Rocky Mountains, the temperatures of weak layers slowly warmed throughout the winter. However, thin overlying slabs reduce insulation from cold air. As a result, the temperatures of weak layers undergo fluctuations that may reduce correlations with shear strength.

5.3.7 Temperature Gradient (TG)

The temperature gradient weakly but positively correlated ($p = 0.021$, Table 5.3) with the shear strength Σ_{inter} of layers of faceted crystals in intermountain snowpacks. However, an effect of temperature gradient on shear strength is not expected since most of the temperature gradient data has magnitudes less than 10°C/m.

The temperature gradient did not correlate with shear strength in continental snowpacks Σ_{cont}. Temperature gradients in continental climates fluctuate and may remain greater than 10°C/m for months. As a result, weak layers of well developed faceted crystals form. After temperature gradients dissipate in February (Appendix A) weak layers are slow to gain strength because the crystals are large and they are subject to relatively light loads.

5.3.8 Crystal Size (E_{max})

The maximum crystal size weakly and positively correlated ($p = 0.03$, Table 5.4) with strength in both snowpack climates, but field workers report that larger faceted crystals are slower to gain strength. The positive correlation indicates that larger crystals have higher strengths. Field observations and cold laboratory observations show increases in crystal sizes when faceted crystals are rounding (Figure 4.13). This suggests growth of rounding faceted crystals is associated with strengthening.

5.3.9 Crystal Type (faceting or rounding faceted crystals)

Faceted crystals are classified as solid angular particles (4a) or angular particles that are rounding (4c) (Colbeck et al., 1990). Field workers usually assume increases in strength when faceted crystals show signs of rounding. In each of the layers of faceted crystals from the Columbia Mountains, rounding was observed from mid-December through the winter. Layers from continental snowpack areas were not reported as rounding until mid-to-late March.

5.4 Shear Strength – Densification Model

An attempt was made to predict shear strength changes from layer densification. Kojima (1967) showed strain rate is a function of layer densification if snow is considered a Newtonian material. The relationship is interpreted as:

$$\rho_{i+1}/\rho_i = (\eta/\sigma\Delta t)+1 \qquad (5.1)$$

where ρ_i is the initial density for a time interval, ρ_{i+1} is the final density for a time interval, η is the compactive viscosity, and σ is the change in load during a time interval.

Numerous studies have found shear strength correlates well with density for dry snow (Keeler, 1969; Perla, 1982; Jamieson and Johnston, In press). Perla (1982) and

Jamieson and Johnston (In press) expressed shear strength as function of density in general form as

$$\Sigma = A\rho^B \tag{5.2}$$

where A and B are constants.

Substituting Equation 5.1 into Equation 5.2 gives a relationship between strength and densification.

$$\Sigma_{i+1}/\Sigma_i = ((\eta/\sigma\Delta t)+1)^B \tag{5.3}$$

Compactive viscosity is usually expressed as a function of density (Shapiro et al., 1997). Most layers of faceted crystals were less than 4 cm thick, and thus too thin for density

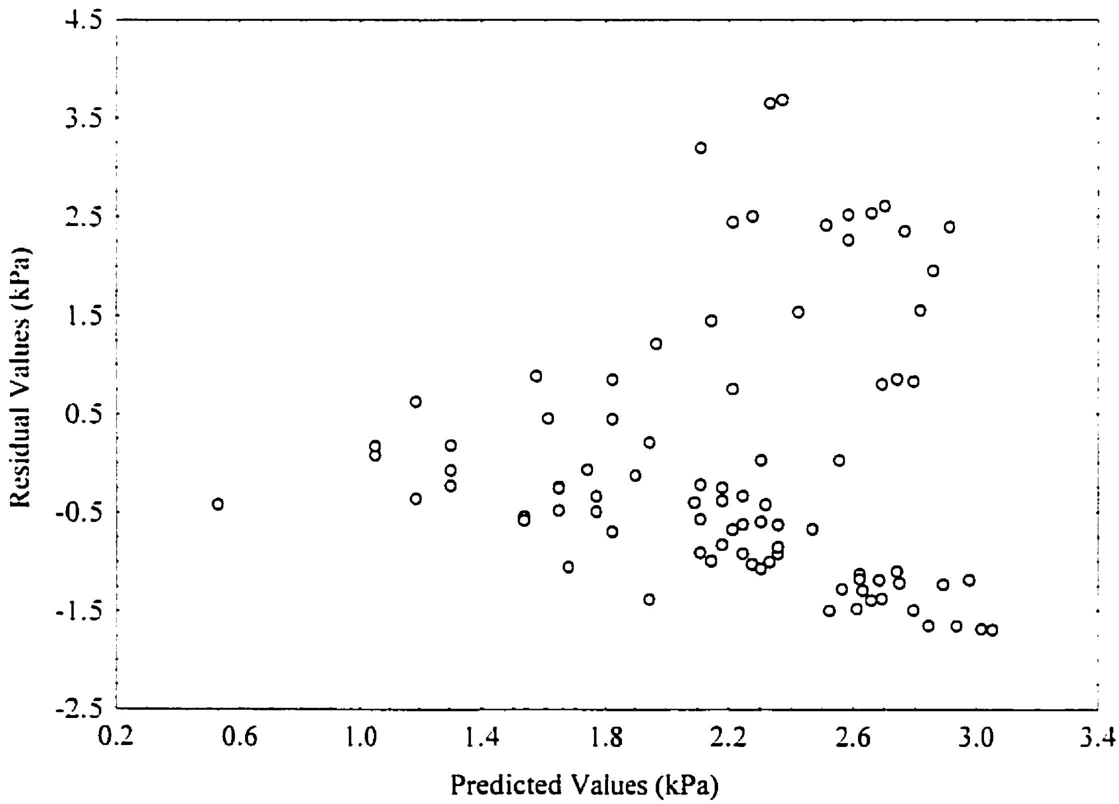

Figure 5.4 Residual values increase with predicted shear strength values.

measurements, eliminating the possibility of estimating viscosity. Viscosity was treated as a constant in the regression analysis. For each series (Table 5.1), Equation 5.3 was fitted using the Quasi-Newton least squares method to estimate η and B. Neither of the regression coefficients were significant for any of the faceted layers. There is no further discussion of this regression model in this study.

5.5 Shear Strength – Time Model

Initially, shear strength increases quickly, and then slows in rate of strength gain over time (Figure 5.1). For this reason a power law of the following form appears promising to predict the shear strength of layers of faceted crystals over time:

$$\Sigma/A = t^{B} \tag{5.4}$$

where t is time and A and B are constants. A relationship between shear strength and

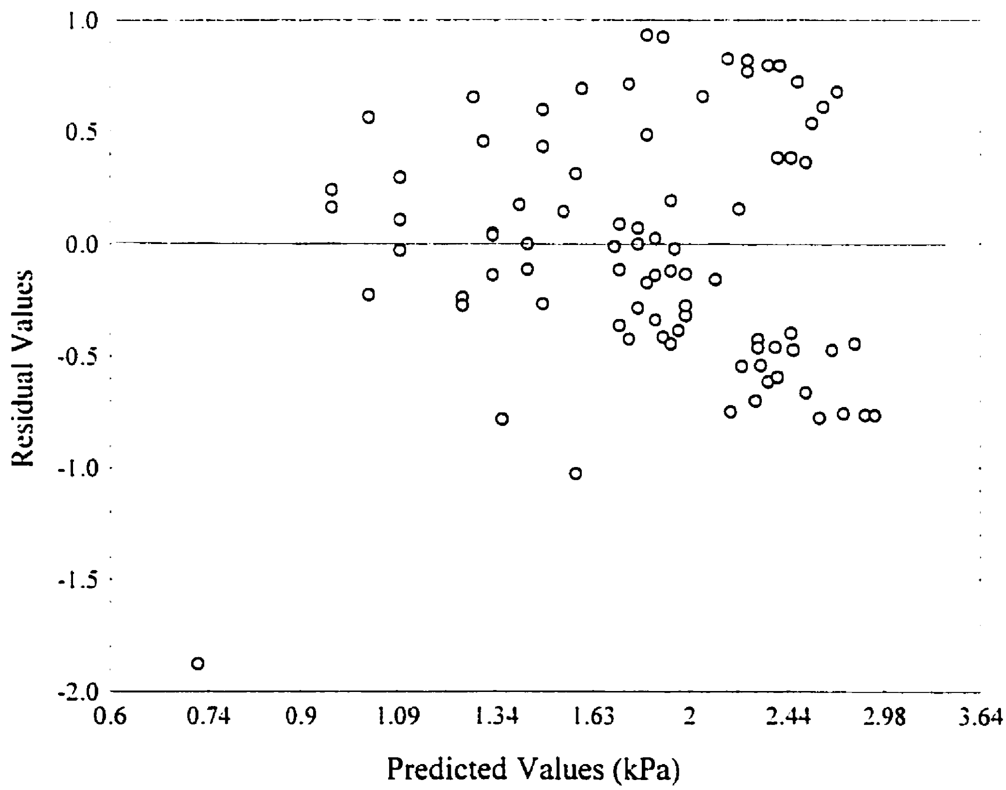

Figure 5.5 After the logarithmic transformation the residuals are randomly distributed with a mean of approximately zero.

time may be expressed in dimensionless parameterization as:

$$\Sigma/\Sigma_1 = (t/t_1)^B \tag{5.5}$$

where Σ is shear strength, t is time, Σ_1 is a constant that represents the strength on day 1 (or the initial strength), t_1 is day one, and B is an empirical constant.

When shear strength data is fitted to Equation 5.5 by the method of least squares the variance increases with predicted strength (Figure 5.4). Constant variance of residuals is one of the standard assumptions of least squares (Chatterjee and Price, 1977, pg 9-10). Constant variance is indicated by the residuals lying randomly about zero in residual plots. In order to alleviate the effects of non-constant variance, a logarithmic transformation of the variables is used (Chatfield, 1980, pg 14-15). Equation 5.4 is transformed and is expressed as:

$$\ln \Sigma = \ln \Sigma_1 + B \ln (t/t_1) \tag{5.6}$$

and the plotted residuals (Figure 5.5) show less dependence on the dependent variable $\ln \Sigma$. Logarithmic values of shear strength and the regression coefficient Σ_1 are transformed into actual values by taking the exponential of both sides of Equation 5.6.

After this correction for variance, another analysis of residuals is useful for detecting outliers (Chatterjee and Price, 1977). In this case, studentized residuals are used. The studentized residuals should lie randomly about zero and should fall between −2 and +2, otherwise they are outliers (Figure 5.5). Outliers affect regression results and must be examined individually before they are removed. The data point for which shear strength is 0.72 kPa (Figure 5.5) is almost an outlier and may influence the fit.

5.6 Model Results

Strength data were regressed using Equation 5.6 for each of four layers of faceted crystals from intermountain snowpack climates and six from continental climates. The

Table 5.5 Regression coefficients and the associated average values.

Location	ID Date	Σ_1 (kPa)	Std. Error	p-level	B	Std. Error	p-level
Intermountain							
Mt. St. Anne	**20-Nov-96**	**0.304**	**0.36**	**1.0E-05**	**0.80**	**0.10**	**1.7E-03**
Mt. Fidelity	**18-Nov-99**	**0.169**	**0.77**	**1.6E-04**	**0.77**	**0.19**	**3.4E-03**
Mt. St. Anne	**24-Dec-98**	**0.265**	**0.63**	**3.0E-03**	**0.74**	**0.19**	**3.1E-02**
Mt. St. Anne	**22-Nov-97**	**0.072**	**0.55**	**8.0E-06**	**0.83**	**0.13**	**5.9E-05**
	Averages	0.203		*Averages*	0.78		
Continental							
Vermont Air Box	17-Jan-97	0.426	0.44	1.0E-05	0.34	0.12	2.9E-02
Vermont Study Plot	17-Jan-97	0.487	0.71	9.4E-04	0.35	0.19	1.4E-01
Bow Summit	18-Nov-99	1.947	1.13	1.5E-04	-0.06	0.24	8.0E-01
Bow Summit	**9-Feb-00**	**0.024**	**0.57**	**1.4E-03**	**1.21**	**0.18**	**4.6E-03**
BB-Lodge	**15-Dec-93**	**0.016**	**0.47**	**9.9E-03**	**1.17**	**0.13**	**2.9E-03**
BB-Lodge	**5-Dec-93**	**0.109**	**0.83**	**2.0E-04**	**0.57**	**0.20**	**1.6E-02**
	Averages	0.050		*Averages*	0.98		

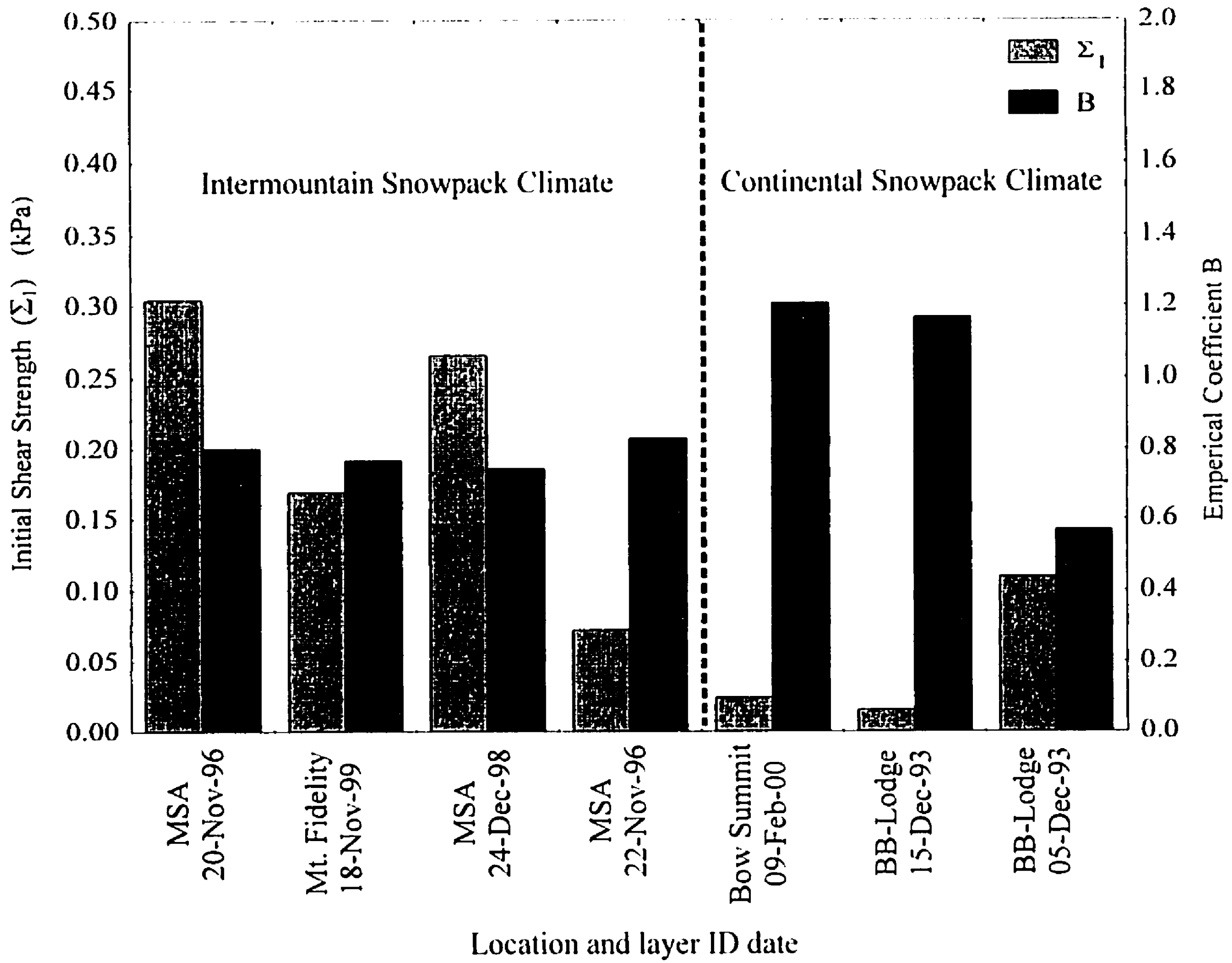

Figure 5.6 Regression coefficients for the continental and intermountain snowpack regions.

regression coefficients Σ_1 (strength on day 1) and B for each layer are listed in Table 5.5. These coefficients are only accepted at the 1% significance level ($p<0.01$). If either coefficient was not significant then the layer was not considered further analysis. Figure 5.6 shows the significant coefficients for intermountain and continental snowpack climates: the coefficients for intermountain snowpack regions are strikingly similar. The average initial shear strength is 0.203 kPa and average B is 0.78. The consistency of the coefficients is probably due to the similar climate of the study sites at Mt. St. Anne and Mt. Fidelity. Both of these study sites are located in the Columbia Mountains at approximately 1900 m.

The average initial strength for continental snowpack regions is approximately 0.05 kPa, which is considerably less than the value for intermountain snowpacks. The average value of the coefficient B is 0.98.

5.7 Results and Discussion of Correlations Between Snowpack Variables Averaged Over the Series and Initial Shear Strength

Once the significant regression coefficients Σ_1 and B were determined, they were rank correlated with snowpack properties (Table 5.6). The regression coefficient B did not significantly correlate with any of the snowpack variables. Spearman rank correlations were used because shear strength and the measured snowpack properties were not normally distributed. Significant snowpack properties ($p < 0.05$) are marked in bold type in Table 5.6. Despite there being only seven data points, significant correlations with Σ_1 still emerge. The initial strength

Correlations with the initial strength (Σ_1) for faceted layers in continental and intermountain snowpack climates.

	N	R	p-level
$\Delta Load/\Delta t_{series}$	**7**	**0.89**	**6.81E-03**
$\Delta H/\Delta t_{series}$	**7**	**0.89**	**6.81E-03**
ΔHS_{series}	**7**	**0.89**	**6.80E-03**
$\Delta HS/\Delta t_{series}$	**7**	**0.89**	**6.80E-03**
TG_{series}	**7**	**0.75**	**5.21E-02**
HS_{series}	**7**	**0.75**	**5.20E-02**
ΔH_{series}	7	0.64	1.19E-01
HS_{series}	7	0.61	1.48E-01
$Load_{series}$	7	0.54	2.15E-01
$\Delta Load_{series}$	7	0.50	2.53E-01
TA/HS_{series}	7	0.36	4.32E-01
Twl_{series}	7	0.25	5.89E-01
TA_{series}	7	-0.43	3.37E-01
$pslab_{series}$	7	-0.50	2.39E-01

Σ_1 positively correlated with HS_{series}, ΔHS_{series}, $\Delta HS/\Delta t_{series}$, $\Delta Load/\Delta t_{series}$, $\Delta H/\Delta t_{series}$ and TG_{series}. These represent differences and averages over the time period between the first measurement and the last measurement, Δt_{series}.

The series averaged snowpack variables are representative of snowpack climate. Values for these variables, with the exception of the temperature gradient, are greater in intermountain climates. The value of temperature gradient is greater in continental climates. The positive rank correlations between various series average snowpack variables and initial strength Σ_1 indicate that layers of faceted crystals in intermountain climates have greater initial strengths than layers in continental climates.

5.8 Discussion of model results

The regression coefficient Σ_1 represents the shear strength on Day 1 and B is a empirical coefficient. The regression curves in Figure 5.7 use the average values of the regression coefficients Σ_1 and B for each snowpack climate. Each curve accurately describes strength until the layers reach an age of approximately 50 days for the continental snowpack climate and 70 days for the intermountain snowpack climate. After these ages the predicted strength values are substantially higher than observed values.

The average value of Σ_1 is greater in intermountain climate sites than at continental climate sites. When a given layer is covered by new snow, greater amounts typically fall in intermountain regions than in continental regions. The thick slabs are associated with large loads and moderate temperature gradients. Load correlates with strengthening and moderate temperature gradients with slower faceting and smaller grains. Continental snowpack climates are associated with less cloud cover and cooler temperatures than intermountain climates. The result is more near-surface faceting. These properties are significant factors in the initial strength of layers of faceted crystals.

5.9 Conclusions

Relationships between the shear strengths of weak layers of faceted crystals and various snowpack variables were assessed by Spearman rank correlations. To assess associations with strength, the data were divided into (a) intermountain and (b) continental, and (c) both climate regions combined. Snowpack variables that correlated best with shear strength of layers of faceted crystals are load and slab thickness.

The slab thickness is associated with strengthening of layers of faceted crystals through its association with load and temperature gradient. Thick slabs insulate weak layers of faceted crystals from cold air temperatures, causing warmer weak layer temperatures and lower temperature gradient. Thick slabs also apply more load to underlying layers of faceted crystals.

Load had a greater correlation with shear strength in the intermountain climate than in the continental climate. This difference in correlation shows that large loads cause layers of faceted crystals to gain strength more quickly. Physically, more load pushes faceted crystals closer together resulting in increased density, number of bonds, and / or bond diameter. However, load showed less effect on shear strength after 50 days in continental climates and 70 days in intermountain climates. Beyond these points, loads continued to increase, but other factors (such as layer temperature, previous temperature gradient and crystal size) may prevent further increases in strength.

Shear strength for each layer of faceted crystals was fitted to a power function of time. The regression coefficient Σ_I is the initial shear strength of the layer and the coefficient B is an empirical coefficient. The initial strengths Σ_I of the layers of faceted crystals in the intermountain climate are larger than in the continental climate. The average initial shear strengths Σ_I are 0.203 kPa and 0.05 kPa for intermountain and continental climates, respectively. The average climatic constant B was 0.78 and 0.98 for intermountain and continental climates, respectively. The model fit the shear strength for 50 days in the continental climate and 70 days in the intermountain climate. After these periods of time, predicted shear strength values were substantially higher than average measured values.

The initial strength Σ_I was rank correlated with snowpack variables. The snowpack variables are indicators of climate regions because they represent differences and averages over the period Δt_{series}. These variables, except temperature gradient, are greater in the intermountain climate. Temperature gradient is greater in the continental climate. The positive rank correlations between various series average snowpack

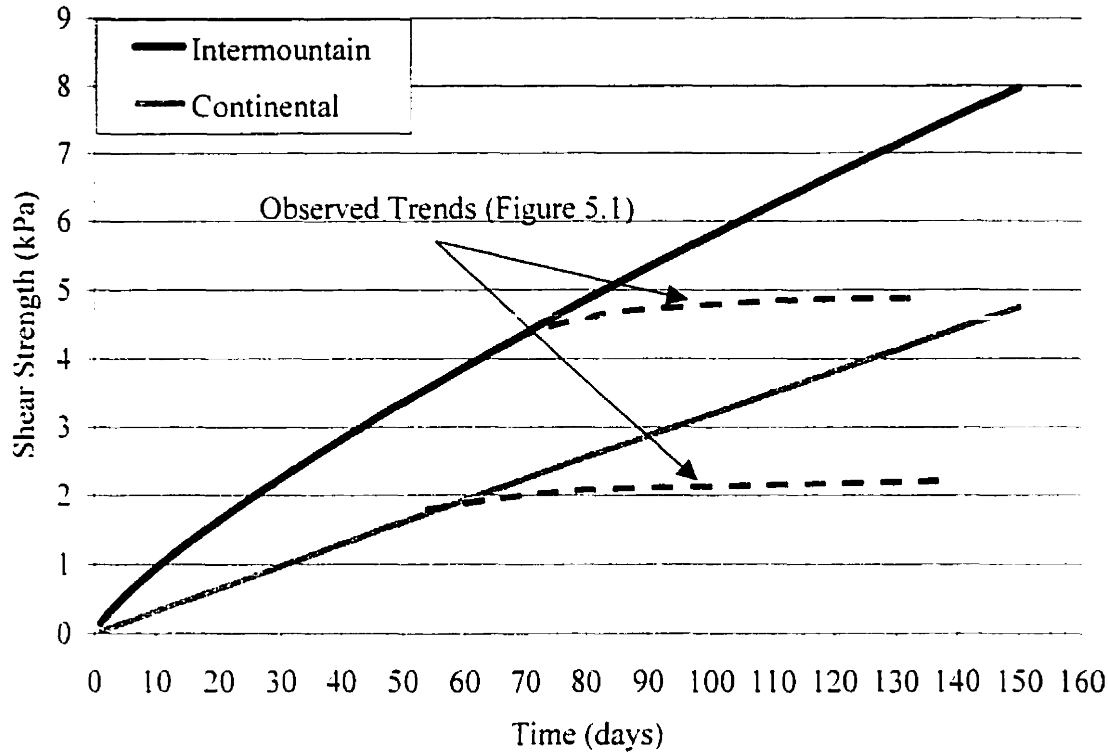

Figure 5.7 Predicted strength trends for the intermountain and continental snowpack climates. The predicted trends are shown with solid lines.

variables and initial strength Σ_1 indicate that layers of faceted crystals in intermountain climates have greater initial strengths.

6 Conclusions

6.1 The Studies of Bonds Between Faceted Snow Crystals

At Mt. Fidelity, which has an intermountain snowpack climate, the median bond diameters were larger at any age than at Bow Summit, which has a continental snowpack. The difference in median bond diameters is because continental and intermountain snowpack climates have snowpack variables, notably load, of different magnitudes.

There was no substantial difference in the bond growth rate during faceting and during rounding metamorphism, indicating temperature gradient has little affect on bond growth under observed conditions. Since this field study showed bond diameter and shear strength increasing with load during the initial 50 days in the Columbia Mountains and 70 days in the Rocky Mountains and because Gubler (1982) observed the strength of bonds under load increasing over times of several minutes in the cold laboratory, load is an important factor in bond growth.

The shear strength of both layers of faceted crystals increased at a faster rate than bond area increased. In addition, minimum bond diameters for both layers remained constant, with similar values, over time. These two arguments suggest that new bonds between grains are forming. This is in contrast to Keeler (1969) who observed a constant number of bonds per crystal during rounding metamorphism and densification. However, faceted crystals have relatively large pore spaces that may allow for further grain re-arrangement during densification and new grain bonds.

A change in bond shape from the February 9th, 2000 layer at Bow Summit was observed from February 21st to March 27th. Initially, the bonds appeared cylindrical, but as the layer aged, large bonds with non-cylindrical shapes, and bonds between more than two crystals (non-cylindrical) started to appear.

6.2 Snowpack Variables Associated With the Shear Strength of Layers of Faceted Crystals

To physically assess Spearman rank correlations with shear strength, the data were divided into (a) intermountain and (b) continental, and (c) both climate regions combined. Snowpack variables that correlated best with shear strength of layers of faceted crystals are load and thickness of the slab.

The thickness of the slab is associated with strengthening of layers of faceted crystals because thicker slabs apply more load to weak layers. Thick slabs insulate weak layers of faceted crystals from cold air temperatures, causing warmer weak layer temperatures and temperature gradients of low magnitudes. However, temperature gradient and temperature of the weak layer did not correlate as significantly with shear strength as load did. Based on these two weak rank correlations, and physical arguments relating load to bond formation and bond growth, load is the dominant snowpack variable associated with strengthening of the layers of faceted crystals.

Load had a greater correlation with shear strength in the intermountain climate than in the continental climate. This difference in correlation shows that faster loading caused layers of faceted crystals to gain strength more quickly. However, load had less effect on shear strength after 50 days in continental climates and 70 days in intermountain climates. Beyond these points, loads continued to increase, but other factors (such as bond size, layer temperature, previous temperature gradient, and crystal size) may limit further increases in shear strength.

6.3 Modeling the Shear Strength of Layers of Faceted Crystals over Time

The shear strength for each layer of faceted crystals was fitted to a power function of time with two regression coefficients. The regression coefficient Σ_I is the initial shear strength of the layer and the coefficient B is an empirical coefficient. The initial strength Σ_I of the layers of faceted crystals in the intermountain climate are larger than in the continental climate where the layers of faceted crystals consisted of larger crystals.

The initial strength Σ_I was rank correlated with climatic snowpack variables for significantly correlated time series. These variables, including load, depth of snow, and thickness of the slab, but excluding temperature gradient, are greater in intermountain climates. Temperature gradient is greater in continental climates. The positive rank correlations between these snowpack variables and initial strength Σ_I, indicate that the layers of faceted crystals in intermountain climates have greater initial strength than in a continental climate.

6.4 Effect of Load on Bonds and Shear Strength

In this study, load was shown to be the predominant snowpack variable that affects the shear strength and the bonds of layers of faceted crystals. When new snow is deposited more load is applied to layers of faceted crystals. Under rounding metamorphism, this causes the grains to slowly re-arrange themselves and new bonds form. At the same time, the existing bonds grow.

Load and thickness of the slab were the most significant snowpack variables that correlated with shear strength. The thickness of the slab generally indicates how much load is overlying a layer of faceted cyrstals. Physically, more load pushes faceted crystals closer together resulting in increased density, increased number of bonds, and / or increased bond diameters. As a result, the shear strength of layers of faceted crystals increases.

6.5 Suggested Research

The properties of bonds between snow crystals are not well known. This is probably because bonds are difficult to identify. As a result, bonds have not been extensively studied with cold laboratory experiments or field observations. Previous studies and this study only considered cylindrical bonds. As bonds age, their geometry evolves and many become non-cylindrical. Three-dimensional analysis of bonds would be useful for predicting their evolution.

Observations of bonds between different types of snow crystals under controlled conditions in the laboratory are needed to isolate the effects of snowpack variables (such as temperature, temperature gradient, and load). Field observations of bonds between different types of snow crystals are needed to verify the cold laboratory results. In addition, bond growth and bond geometries need to be studied at the interfaces between snowpack layers that usually fail as the first stage of slab avalanche release.

Slab avalanches often occur when layers of faceted crystals are overlying or underlying crusts. Often the growth of layers of faceted crystals is extensive near crusts. Field measurements of temperature gradients, temperature, crystal growth, the thickness of crusts, and shear strength would provide insight into the physical effects of crusts on adjacent layers.

Skier-triggered avalanches that release on weak layers of faceted crystals generally occur within the first two weeks of burial. However, the majority of shear strength and snowpack variable measurements used in this study are more than two weeks old. The amount of loading on the weak layer during this time period is important. Daily or 12 hour measurements of the shear strength and snowpack variables, over a period of two weeks after burial, would provide insight into the effect of loading rate. If the measurements are sufficiently long and continuous, the time lag between the changes of snowpack variables and changes in shear strength may be determined.

Snowpack climates determine characteristic values of snowpack variables. However, there has been no extensive study of alpine snowpack climates in southwestern Canada. Snowpack climates in southwestern Canada are thought to be extensions of the snowpack climates in the western United States. However, these two sets of climates may be different. An extensive study of snowpack climates would be useful for determining relationships between climate zone, snowpack structure, avalanche activity, avalanche run-out models, and comparisons with the snowpack climates of the United States.

References

Avalanche Control Section (ACS). 1998. Mt. Fidelity weather records. Parks Canada. Glacier National Park. British Columbia.

Akitaya. E. 1974. Studies on depth hoar. Contributions from the Institute of Low Temperature Science. Series A. 26. 1-67.

Armstrong, R.L. 1981. Some observations on snowcover temperature patterns. Proceedings of the Avalanche Workshop in Vancouver. 3-5 November 1980. National Research Council of Canada Technical Memorandum 133. 66-75.

Armstrong. R.L. 1985. Metamorphism in a subfreezing. seasonal snowcover: the role of thermal and vapor pressure conditions. Ph.D. Thesis. University of Colorado. Boulder. 175 pp.

Armstrong. R.L and Armstrong. B.R. 1987. Snow and avalanche climates of the western United States. Avalanche Formation. Movement. and Effects. IAHS Pub. 162 Wallingford. Oxen: International Association of Hydrological Sciences. 281-294.

Arons. E.M. and S.C. Colbeck. 1995. Geometry of heat and mass transfer in dry snow: a review of theory and experiment. Reviews of Geophysics 4(33). 463-493.

Atkins. D. In press. Human factors in avalanche accidents. International Snow Science Workshop 2000. Big Sky. Mt. American Avalanche Association, PO Box 1032. Bozeman. Mt. 59771. USA.

Baldwin. J. 1994. Exploring the Coast Mountains on Skis. John Baldwin. Vancouver. BC.

Bader. H., R. Haefeli. E. Bucher. J. Neher. O. Eckel and C. Thams. 1954. Snow and its Metamorphism. Snow. Ice and Permafrost Research Establishment Translation 14. Cold Regions Research and Engineering Laboratory. Hanover, New Hampshire. 313 pp.

Ballard, G.E.H. and E.D. Feldt. 1965. A Theoretical consideration of the strength of snow. Journal of Glaciology 6(43). 159-170.

Barry, R. 1992. Mountain Weather and Climate. Routeledge, New York, New York.

Birkeland, K. and R. Johnson. 1996. The stuffblock snow stability test. USDA Forest Service report 9623-2836-MTDC. 15 pp.

Birkeland, K. 1998. Terminology and predominant processes associated with the formation of weak layers of near-surface faceted crystals in the mountain snowpack. Arctic and Alpine Research. 2(30). 193-199.

Birkeland, K. R. Johnson and D.S. Schmidt. 1998. Near-surface faceted crystals formed by diurnal recrystallization: A case study of weak layer formation in the mountain snowpack and its contribution to snow avalanches. Arctic and Alpine Research. 2(30) 200-204.

Brown, R.L. 1980. A volumetric constitutive law for snow based on a neck growth model. Journal of Applied Physics. 51(1). 161-165.

Brown, R.L. and M.E. Edens. 1991. On the relationship between neck length and bond radius during compression of snow. Journal of Glaciology 37(126). 203-208.

Brun, E. and E. Pahaut. 1991. An efficient method for delayed and accurate characterization of snow grains from natural snowpacks. Journal of Glaciology 37 (127). 420-422.

Canadian Avalanche Association (CAA). 1995. Observation Guidelines and Recording Standards for Weather, Snowpack and Avalanches. Canadian Avalanche Association. P.O. Box 2759, Revelstoke, BC, Canada. 98 pp.

Claus, B.R, S.O. Russel, and P. Schaerer. 1984. Variation of ground snow loads with elevation in southern Bristish Columbia. Canadian Journal of Civil Engineering. 3 (11). 480-493.

Chatfield, C. 1980. The analysis of time series: an introduction. Chapman and Hall, New York, New York.

Chatterjee, S. and B. Price. 1977. Regression analysis by example. John Wiley and Sons, New York, New York.

Colbeck, S.C. 1979. Grain clusters in wet snow. Journal of Colloid and Interface Science. 72(3) pg. 371-384.

Colbeck, S.C. 1980 Thermodynamics of snow metamorphism due to variations in curvature. Journal of Glaciology (26)94. 291-301.81-96.

Colbeck, S. C. Akitaya, R. Armstrong, H. Gubler, J. Lafeuille, K. Lied, D. McClung and E. Morris. 1990. International Classification for Seasonal Snow on the Ground. International Commission for Snow and Ice (IAHS), World Data Center A for Glaciology, U. of Colorado, Boulder, CO, USA.

Colbeck, S.C. 1991. The layered character of snow covers. Reviews of Geophysics 29(1).

Colbeck, S.C. 1997. A review of sintering in seasonal snow. CRREL report 97-10. 11 pp.

Colbeck, S.C. 1998. Sintering in a dry snow cover. Journal of Applied Physics. 84(8). 4584-4589.

Colbeck, S.C. and J.B. Jamieson. Submitted. The formation of faceted layers above crusts. Cold Regions Science and Technology.

Daffern, T. 1992. Avalanche Safety for Skiers and Climbers. Rocky Mountain Books, Calgary, AB.

de Quervain, M. 1951. Strength properties of a snow cover and its measurement. U.S. Army Snow Ice and Permafrost Research Establishment, Translation 9. 9 pp.

de Quervain, M. 1958. On metamporphism and hardening of snow under constant pressure and temperature gradient. International Association of Scientific Hydrology, Proceedings of the General Assembly in Toronto, 1957, Vol. 4. 225-239.

de Quervain, M. 1963. On the metamorphism of snow. In: Ice and Snow: Properties, Processes and A pplications, M.I.T. Press, Cambridge, Mass. 377-390.

de Montmollin. V. 1982. Shear tests on snow explained by fast metamorphism. Journal of Glaciology 28(98). 187-198.

Edens. M.Q. and R.L. Brown. 1991. Changes in microstructure of snow under large deformations. Journal of Glaciology 37(126). 193-202.

Fierz. C. 1998. Field observation and modeling of weak-layer evolution. Annals of Glaciology 26. 7-13.

Fitzharris B.B. 1981. Frequency and Climatology of Major Avalanches at Rogers Pass. 1909-1977. Ottawa National Research Council. Canada Association Committee on Geotechnical Research. DBR Paper No. 956.

Föhn. P.M.B. 1987. The stability index and various triggering mechanisms. In: B. Salm and H. Gubler. eds.. Avalanche Formation. Movement and Effects. International Association of Hydrological Sciences. Publication No. 162. 195-211.

Fukuzawa. T. and E. Akitaya. 1993. Depth-hoar crystal growth in the surface layer under high temperature gradient. Annals of Glaciology 18. 39-51.

Gadd. Ben. 1986. Handbook of the Canadian Rockies. Corax Press. Jasper. AB

Gubler, H. 1978. Determination of the mean number of bonds per snow grain and of the dependence of the tensile strength of snow in stereological parameters. Journal of Glaciology 20(83). 329-341.

Gubler. H. 1982. Strength of bonds between ice grains after short contact times. Journal of Glaciology 28(100). 457-473.

Hay. John. and Oke. Tim. 1976. The Climate of Vancouver. Tantulus Research Limited. Vancouver. BC.

Hobbs. P.V. and B. Mason. 1964. The sintering and adhesion of ice. Philosophical Magazine. 9(98) 181-197.

Jamieson. J.B. and C.D. Johnston. 1992. Snowpack characteristics associated with avalanche accidents. Canadian Geotechnical Journal 29. 862-866.

Jamieson, J.B. 1995. Avalanche prediction for persistent snow slabs. PhD Thesis, Dept. of Civil Engineering, University of Calgary, 275 pp.

Jamieson, J.B. and T. Geldsetzer. 1996 Avalanche Accidents in Canada Volume 4 1984-1996. Canadian Avalanche Association, Revelstoke, B.C Canada.

Jamieson, Bruce and Colin Johnston. 1997. The facet layer of November 1996. Avalanche News 52, 10-15.

Jamieson, J.B. and C.D. Johnston. 1999. Snowpack factors associated with strength changes of buried surface hoar layers. Cold Regions Science and Technology, 30, 19-34.

Jamieson, J.B. and C.D. Johnston. In press Evaluation of the shear frame test for weak snowpack layers. Annals of Glaciology 33.

Jamieson, J.B. and C.S. Stethem. 2000. Snow Avalanche Hazards in Canada: Challenges and Progress. Submitted to Journal of Natural Hazards.

Jamieson, J.B. and J. Schweizer. 2000. Texture and strength changes of buried surface hoar layers with implications for dry snow-slab avalanche release. Journal of Glaciology, 46(152), 151-160.

Keeler, C.M. 1969. Some physical properties of alpine snow, U.S. Army Cold Regions Research and Engineering Laboratory, Technical Note, 67 pp.

Keeler, C.M. 1969. The growth of bonds and the increase of mechanical strength in a dry seasonal snow-pack. Journal of Glaciology, 8(54), 441-450.

Kingery, W.D. 1960. Regelation, surface diffusion, and ice sintering. Journal of Applied Physics, 5(31).

Kojima, K. 1967. Densification of seasonal snow cover. Physics of Snow and Ice, International Conference on Low Temperature Science. Hokkaido Universtiy. pp. 929 – 952.

Kry, P.R., 1975. Quantitative stereological analysis of grain bonds in snow, Journal of Glaciology, 14(72), 467-477.

Kuczynski, G. 1949. Self-diffusion in sintering of metallic particles. Journal of Metals, pp. 169-178.

Kuroiwa, D. 1975. Metamorphism of snow and ice sintering observed by time lapse cine-photomicrography. Proceedings of the Grindewald Symposium, International Association of Hydrological Sciences, Washington, D.C. Publication No. 114, 42-48.

LaChapelle, E. and R.L. Armstrong. 1977. Temperature patterns in an alpine snow cover and their influence on snow metamorphism. Institute of Arctic and Alpine Research, University of Colorado, pp. 33.

Marbouty, D. 1980. An experimental study of temperature-gradient metamorphism. Journal of Glaciology, 94(26), 303-312.

McClung, D.M. 1979. Shear fracture precipitated by strain softening as a mechanism of dry slab avalanche release. Journal of Geophysical Research 84(B7), 3519-3526.

McClung, D.M. and P.A. Schaerer. 1993. The Avalanche Handbook. The Mountaineers, Seattle, p. 271.

Mock, C.J., 1995. Avalanche climatology of the continental zone of the southern Rocky Mountains USA. Physical Geography 16(3), 165-187.

Mock, C. and K. Birkeland. 1999. Avalanche climate regions of the western United States: An Update. Avalanche Review 17(3), 3-5.

Mock, C. and K. Birkeland. 2000. Snow avalanche climatology of the western United States mountain ranges. Bulletin of the American Meteorological Society, 81(10), 2367-2392.

Moore, Mark. 1982. Temperature gradient weakening of snowpacks near rain crusts or melt-freeze layers. Presented at the 1982 International Snow Science Workshop in Bozeman, Montana.

Morrall, J.F., and W.M. Abdelwahab. 1992. Estimating traffic delays and the economic cost of recurrent road closures on rural highways. Logistics and Transportation Review 29(2), 159-177.

Perla, R.I., T.M.H. Beck and T. Cheng. 1982. The shear strength index of alpine snow. Cold Regions Science and Technology 6. 11-20.

Perla, R.I., and R.A. Sommerfeld. 1987. On the metamorphism, morphology and microstructure of snow. In: Proceedings of International Snow Science Workshop at Lake Tahoe, ISSW Workshop Committee, Homewood, California, 98-102.

Pielmeirer, C. M. Schneebeli T. Stucki. In press. Snow Texture: a comparison of empirical versus simulated texture index for alpine snow. Annals of Glaciology 33.

Ramsier, R. and C. Keeler. 1966. The sintering process in snow. Journal of Glacology 6 (45). 421-424.

Ramsier, R.O and G. Sander. 1966. Temperature dependence and mechanism of sintering. US Army Cold Regions Research and Engineering Laboratory, Research Report 189 pp. 16.

Shapiro, L.H., J.B. Johnson, M. Sturm, and G.L. Blaisdell. 1997. Snow Mechanics: Review of the State of Knowledge and Application. US Army CRREL Report 97-3.

Statistica. 1999. Electronic Manual. Statsoft, Tulsa, OK.

Sommerfeld, R.A. and E.R. LaChapelle. 1970. The classification of snow metamorphism. Journal of Glaciology 9(55). 3-17.

Sommerfeld, R.A. 1980. Statistical models of snow strength. Journal of Glaciology 26 (94). 217-223.

Seligman, G. 1936. Snow Structure and Ski Fields. International Glaciological Society, Cambridge. 555 pp.

Stock, J., K. Elder and K. Birkeland. 1998. The temporal variations of near-surface faceted crystals, Red Mountain Pass Corridor, Colorado. Proceedings of the International Snow Science Workshop, Stevens Pass Ski Area PO Box 98 Skykomish WA. 98288. USA. 265-269.

Stratton, J. 1977. "Development of upper level temperature gradient crystals" and "Upper level termperature gradient avalanche cycle-February 1-2 1977." Short, unpublished papers that were circulated among Utah avalanche workers in the late-1970s and 1980s by John Stratton, Snowbird Snow Safety.

Stull, R. B. 1995. Meteorology Today For Scientists and Engineers. West Publishing Company, St. Paul, MN.

Sturm, M. and C. Benson. 1997. Vapor transport, grain growth and depth-hoar development in the subarctic snow. Journal of Glaciology 143(43), 42-59.

Voight, B., Armstrong, B. R., Armstrong, R. L., Bachman, D., Bowles, D., Brown, R.L., Faisant, R. D., Ferguson, S. A., Fredston, J. A., Kennedy, J., Kiusalaas, J., LaChapelle, E. R., McFarlane, R. C., Newcomb, R., Penniman, R., and Perla, R., 1990. Snow-avalanche Hazards and Mitigaition in the United States. Washington, D.C.: National Academy Press.

Yosida, Z., 1963. Physical Properties of Snow. In Ice and Snow: Properties, Processes and Applications. Edited by W.D. Kingery. M.I.T. Press, Mass. 485-527.

Appendix A: Contrasting the Snowpack Climates in the Columbia and Rocky Mountains of Southwestern Canada

A.1 Introduction

In the western contiguous United States there are three snowpack climate zones: coastal, intermountain, and continental (Armstrong and Armstrong, 1987; Mock and Birkeland, 2000). Each of these climate zones has different weather and snowpack conditions. Similar snowpack climate and avalanche studies (Fitzharris, 1981; Claus et al., 1984; Mock, 1995) have suggested the same snowpack climates exist in southwestern Canada. Despite these assumptions there has not been similar study verifying the snowpack climate zones in southwestern Canada.

In the western United States the three snowpack climate zones are described in terms of snowpack and weather characteristics. The Coast Mountains are a coastal climate due to relatively warm temperatures and heavy snowfall. The Rocky Mountains are a continental climate because they are associated with cold air temperatures and a shallow snowpack. The intermountain climate of the Columbia Mountains is due to an overlap between the coastal and continental weather systems (Armstrong and Armstrong, 1987). As a result, the intermountain climate has less snowfall and cooler temperatures than the Coast Mountains, but more snowfall and warmer temperatures than the Rocky Mountains.

In this appendix snowpack depth, new snowfall, and air temperatures are summarized from two sites in the Rocky Mountains and three sites from the Columbia Mountains. The contrasts of these snowpack and air temperatures are used to distinguish between the intermountain climate of the Columbia Mountains and continental climate of the Rocky Mountains in southwestern Canada.

A.2 Geographic and Topographic Characteristics of the Columbia Mountains and Rocky Mountains

The weather and snowpack climates of the Columbia and Rocky Mountains are partially due to geographic location, orientation, and topographic characteristics. Both mountain ranges are oriented in a general southeast to northwestern direction. The Columbia Mountains are located in eastern British Columbia and are approximately 550 km from the Pacific Ocean. Parallel to the Columbia Mountains but further east are the Rocky Mountains. The Rocky Mountains straddle the British Columbia - Alberta border and are about 800 km from the Pacific Ocean. Both the Columbia and Rocky Mountains have maximum elevations of about 3400 to 3700 meters.

A.3 Mountain Weather Patterns

In western Canada three major pressure systems determine the character of the weather (Fitzharris, 1981) (Figure A.1). The Aleutian low often dominates southwestern Canada during the winter (Fitzharris, 1981). Aleutian low pressure systems develop in the Gulf of Alaska, with typical diameters of 100-3000 km (Daffern, 1992), and travel from the southwest intercepting the west coast.

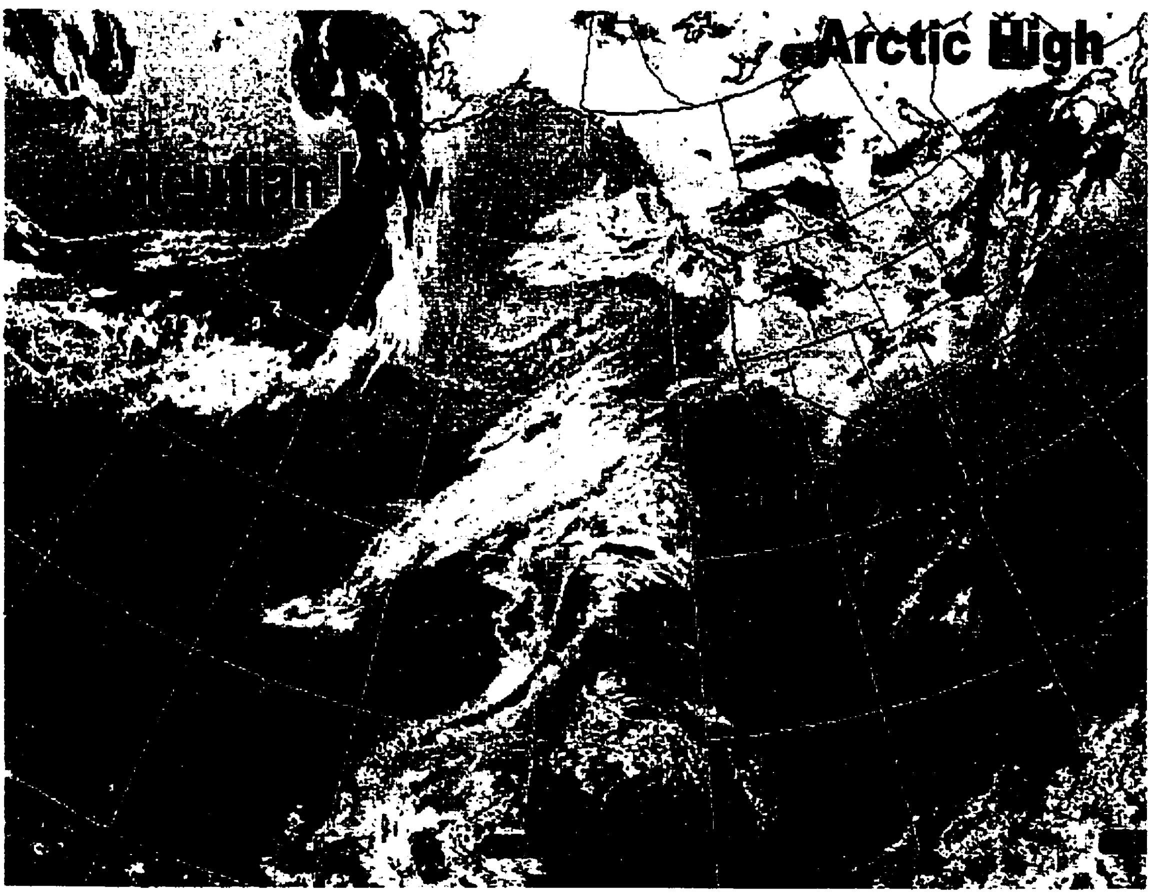

Figure A.1 Dominant pressure systems in western Canada.

Air from high pressure systems travels outwards from the center and sinks (Barry, 1992). There are two high pressure systems that affect western Canada. The Pacific High pressure system is responsible for moderate temperatures and dry periods (Fitzharris, 1981) and when present is usually located off the coast of Washington or British Columbia. Pacific High pressure systems typically do not travel extensive distances inland (Hay, 1976). The Arctic High pressure system effects both the Columbia and Rocky Mountains. During the winter months it is the dominant pressure system over the arctic and often moves to the southwest and causes clear skies and cold temperatures.

A.4 Climate Characteristics

Each of the three climate zones in western Canada has different weather regimes and snowpack characteristics. The Coast Mountains are the subject to the brunt of Aleutian lows and infrequent high pressure systems. Vancouver, BC experiences 72% of its days with low pressure disturbances and 22% with high pressure (Hay, 1976). The frequency of low pressure systems results in a deep snowpack. The annual precipitation in the Coast Mountains varies from 0.5 m to 3.5 m water equivalent (Baldwin, 1993).

The Rocky Mountains are considered to have a continental climate. The snowpack is thin and the temperatures are cold. The annual precipitation at the valley floors varies between 0.4 m and 0.6 m (Gadd, 1986). Locations in the western Rocky Mountains receive an additional 0.1 m of precipitation compared with eastern regions (Gadd, 1986). These conditions are often due to the predominant Arctic high pressure system. When weather patterns from the Pacific Ocean reach the Rocky Mountains their moisture content is reduced which leads to less precipitation (Barry, 1992).

The position of the cold Arctic highs and warm Aleutian lows determine the character of the weather in the Columbia Mountains (Fitzharris, 1981). The annual precipitation in Revelstoke, BC is 1 m water equivalent (Gadd, 1986). The Columbia Mountains receive less precipitation than the Coast Mountains, but more than the Rocky Mountains. At Mt. Fidelity, in the Columbia Mountains, precipitation occurs an average of 23 days per month in the winter (Avalanche Control Section, 1998).

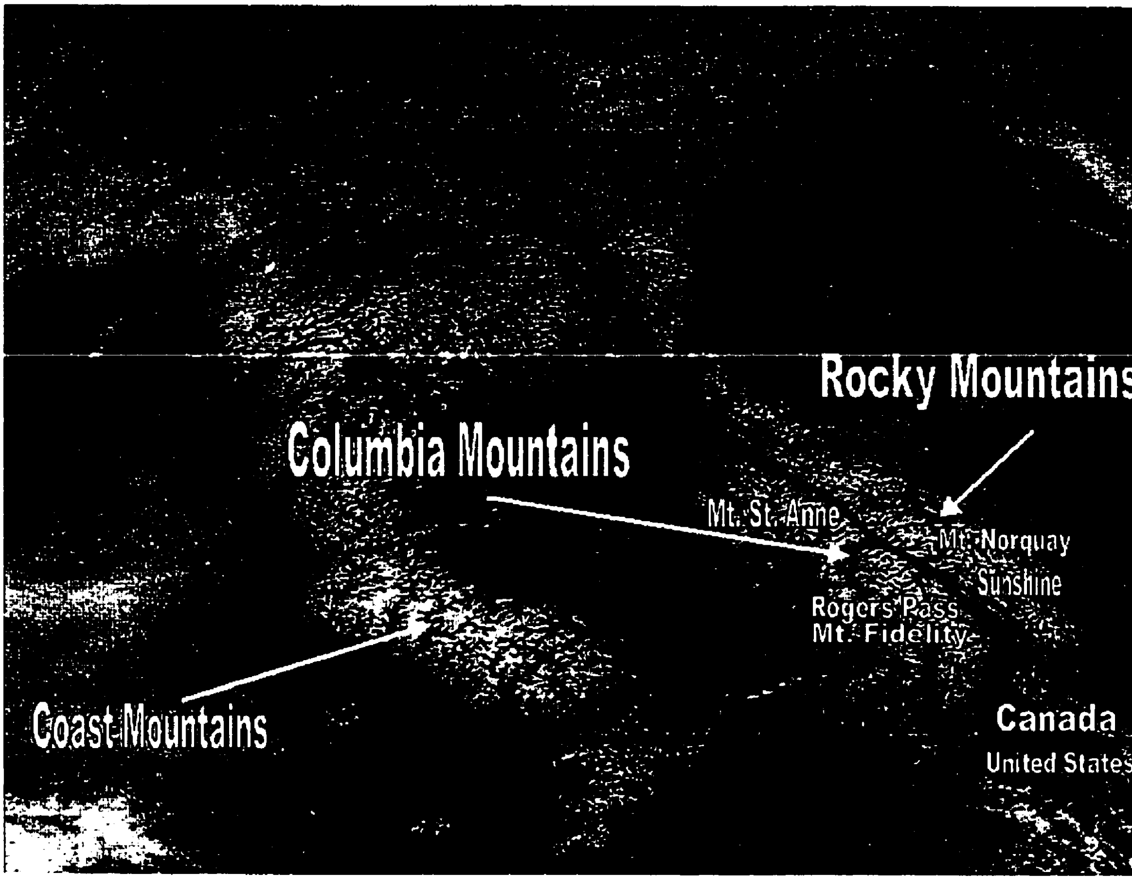

Figure A.2 Locations of mountain ranges and study sites

A.5 Data Collection

To contrast the snowpack climates of the Columbia Mountains with the Rocky Mountains data were collected from ski areas, Parks Canada, and the Snow and Avalanche Research Project at the University of Calgary. Sunshine Village Ski Area and Mt. Norquay Ski Area provided data for the Rocky Mountains. The Avalanche Control Section (ACS), a branch of Parks Canada at Rogers Pass, and the University of Calgary (Mt. St. Anne) provided data from the Columbia Mountains. Figure A.2 shows the locations of each study plot and Table A.1 gives site elevations and aspects for each study plot.

Sites were selected that had sufficient data over many years and that were considered representative of the ranges. The data collected from each site includes daily maximum temperatures, daily minimum temperatures, daily morning temperatures,

Table A.1 Site elevations, locations, and observation times.

	Aspect	Elevation (m)	Time of observations			Data from	
			Min. Temp.	Max. Temp.	Morning Temp.	Months	Years
Intermountain							
Rogers Pass	VB*	1315	Hourly	Hourly	7:00 AM	November - March	1966- 1997
Mt. St. Anne	E	1900	5:30 AM	5:30 AM	5:30 AM	December - March	1991-1997
Mt. Fidelity	E	1905	Hourly	Hourly	7:00 AM	November - March	1966- 1997
Continental							
Sunshine	VB*	2225	-	-	-	November - March	1970-1998
Mt. Norquay	VB*	1729	8:00 AM	4:30 PM	8:00 AM	November - March	1981-1996

*Valley bottom.

monthly snowfall, snow depth at the end of the month.

At Rogers Pass (1315 m) and Mt. Fidelity (1905 m) data were collected from November through March, 1966-1997. Data from Mt. St. Anne (1900 m) was collected from January through March, 1990-1991 and December through March 1991-1997. In most cases data from Mt. St. Anne were not complete for each month. If data were collected more than 15 days per month they were used for averaging. The morning temperatures from Mt. St. Anne were taken at 5:30 am.

Data from Mt. Norquay were collected from November through March, 1981-1996. The minimum and morning temperatures were recorded at 8:00 am. Maximum daily temperatures were observed at 4:30 pm. The Mt. Norquay snow and weather records contain very few days with missing data. At Sunshine Village Ski Area, Parks Canada recorded total daily snowfall and mean monthly temperatures from November through March 1970-91. Data were recorded from Sunshine Village Ski Area for daily maximum temperature, minimum temperature, monring temperature, snowfall, and end of the month snow depth from November through March 1989-1990, and 1994-1998.

A.6 Results and Discussion

To contrast the snow climates of the Columbia and Rocky Mountains in Canada, monthly values of two air temperatures variables and two snow variables are listed in Table A.2 and Table A.3 and shown in Figures A.4, A.5, A.6, A.7.

A.6.1 Minimum and Present Air Temperatures

Average monthly minimum temperatures in Figure A.4 indicate consistently cooler temperatures for each month in the Rocky Mountains. Cooler temperatures are expected because of the often clear skies and relatively low humidity of a continental climate. The average monthly present temperature also follows a similar trend. In each month the temperatures are colder in the Rocky Mountains than the Columbia Mountains (Table A.1). Lower present temperatures are also expected because of cool nights.

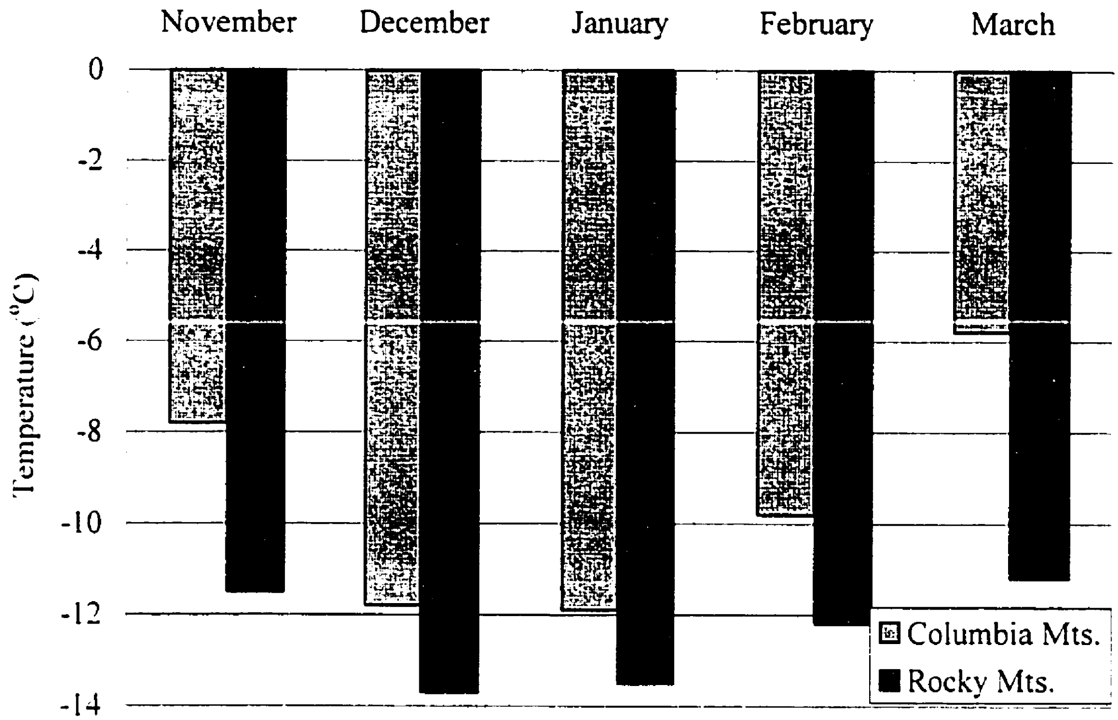

Figure A.3 Monthly average of daily minimum air temperatures.

A.6.2 Monthly Snowfall and Snowpack Depth

The average monthly snowfall and the average end of the month snow depth for the Rocky Mountains are approximately one-half of the values for the Columbia Mountains. Figure A.4 shows the average maximum monthly snowfall is 87 cm for December in theRocky Mountains and 228 cm for January in the Columbia Mountains. Figure A.5 shows the end-of-month snow depth is 132 cm in the Rocky Mountains and 291 cm in the Columbia Mountains at the end of March. However, both maximum snow depths occur in March. These data may not represent the maximum snow depths because April and May are not considered in this study. The higher average monthly snowfall and end-of-month snow depths in the Columbia Mountains are expected because moist air masses typically travel from the west or southwest and encounter the Columbia Mountains before the Rocky Mountains.

Table A.2 Intermountain climate data.

Mt. St. Anne Study Plot (1900 m)

Parameter	November	December	January	February	March	Season
Average Maximum Temperature (^oC)	-	-7.3	-6.6	-4.4	-1.9	
Average Minimum Temperature (^oC)	-	-11.6	-11.0	-10.0	-9.0	**-10.4**
Average Present Temperature (^oC)	-	-9.8	-9.4	-8.4	-7.2	**-8.7**
Precipitation: Snow (cm)	-	202.8	201.8	134.9	141.3	**680.8**
Precipitation: Rain (mm)	-	-	-	-	-	
Snow Depth (cm)	-	202.5	260.5	281.6	290.8	**258.9**

Mt. Fidelity Study Plot (1905 m)

Parameter	November	December	January	February	March	Season
Average Maximum Temperature (^oC)	-4.0	-7.2	-7.1	-4.4	-1.6	**-5.1**
Average Minimum Temperature (^oC)	-8.6	-11.9	-12.1	-9.7	-7.7	**-10.0**
Average Present Temperature (^oC)	-6.3	-9.6	-9.6	-7.0	-4.7	**-7.4**
Precipitation: Snow (cm)	250.2	268.7	262.3	205.8	171.7	**1158.7**
Precipitation: Rain (mm)	10.3	2.2	2.4	1.1	3.2	**19.2**
Snow Depth (cm)	124.0	-	-	-	-	**124.0**

Table A.2 Intermountain climate data.

Rogers Pass Study Plot (1315 m)						
Parameter	November	December	January	February	March	**Season**
Average Maximum Temperature (^{o}C)	-3.0	-7.3	-7.7	-3.6	1.7	**-4.2**
Average Minimum Temperature (^{o}C)	-7.0	-11.9	-12.7	-9.7	-6.8	**-9.6**
Average Present Temperature (^{o}C)	-5.0	-9.6	-10.2	-6.6	-2.6	**-6.8**
Precipitation: Snow (cm)	168.9	208.3	219.2	154.6	97.7	**848.7**
Precipitation: Rain (mm)	28.4	6.6	8.0	5.7	16.4	**65.1**
Snow Depth (cm)	76.0	-	-	-	-	**76.0**

Columbia Mountains Averages						
Parameter	November		January	February	March	**Season**
Average Maximum Temperature (^{o}C)	-3.5	-7.3	-7.1	-4.1	-0.6	**-4.8**
Average Minimum Temperature (^{o}C)	-7.8	-11.8	-11.9	-9.8	-5.8	**-9.4**
Average Present Temperature (^{o}C)	-5.7	-9.6	-9.7	-7.3	-4.8	**-7.4**
Average Temperature Gradient (^{o}C/m)	7.6	5.8	4.6	3.5	2.0	
Precipitation: Snow (cm)	209.5	226.6	227.8	164.9	136.9	**965.7**
Precipitation: Rain (mm)	19.4	4.4	5.2	3.4	9.8	**42.2**
Snow Depth (cm)	103	202.5	260.5	281.6	290.8	**258.9**

Table A.3 Continental climate data.

Norquay Ski Area Study Plot (1729 m)						
Parameter	November	December	January	February	March	**Season**
Average Maximum Temperature (^{o}C)	-2.9	-6.6	-6.1	-3.6	0.4	**-3.8**
Average Minimum Temperature (^{o}C)	-10.5	13.5	-13.2	-11.4	-9.7	**-6.3**
Average Present Temperature (^{o}C)	-6.1	-10.2	-10.1	-7.8	-4.7	**-7.8**
Precipitation: Snow (cm)	50.0	45.4	48.8	38.8	48.9	**231.9**
Precipitation: Rain (mm)	-	-	-	-	-	-
Snow Depth (cm)	29.7	43.9	59.1	68.4	72.4	**54.7**

Sunshine Ski Area Study Plots (2250 m, 2225 m)						
Parameter	November	December	January	February	March	**Season**
Average Maximum Temperature (^{o}C)	-3.0	-3.7	-7.7	-6.0	-3.7	**-4.8**
Average Minimum Temperature (^{o}C)	-12.3	-13.9	-13.9	-12.9	-12.6	**-13.1**
Average Present Temperature (^{o}C)	-9.3	-12.2	-12.9	-9.8	-7.0	**-10.2**
Precipitation: Snow (cm)	114.0	128.0	115.0	92.0	86.0	**535.0**
Precipitation: Rain (mm)	-	-	-	-	-	-
Snow Depth (cm)	82.7	124.0	156.3	182.4	192.0	**147.5**

Table A.3 Continental climate data.

Rocky Mountain Averages						
Parameter	November	December	January	February	March	**Season**
Average Maximum Temperature (°C)	-3.0	-5.2	-6.9	-6.1	-1.7	**-5.0**
Average Minimum Temperature (°C)	-11.5	-13.7	-13.5	-12.2	-11.2	**-12.4**
Average Present Temperature (°C)	-7.7	-11.2	-11.5	-7.4	-5.9	**-8.7**
Average Temperature Gradient (°C/m)	20.5	16.3	12.5	9.7	8.5	
Precipitation: Snow (cm)	56.2	86.7	81.9	65.4	67.5	**357.7**
Precipitation: Rain (mm)	0.0	0.0	0.0	0.0	0.0	**0.0**
Snow Depth (cm)	56.2	84.0	107.7	125.4	132.2	**101.1**

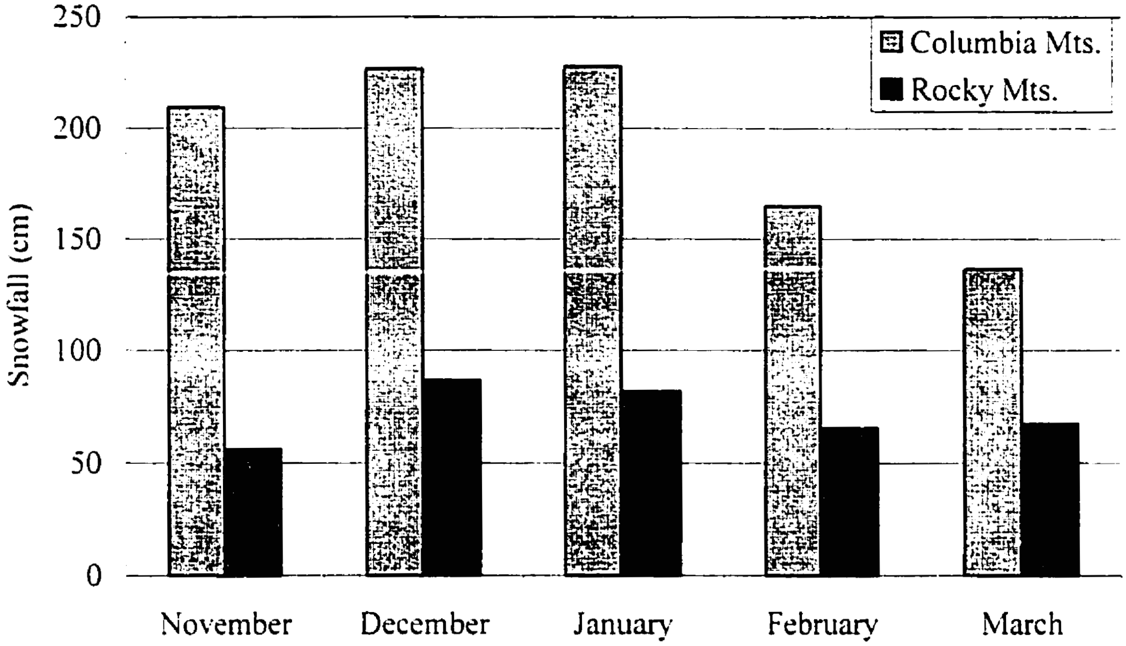

Figure A.4 Total monthly snowfall

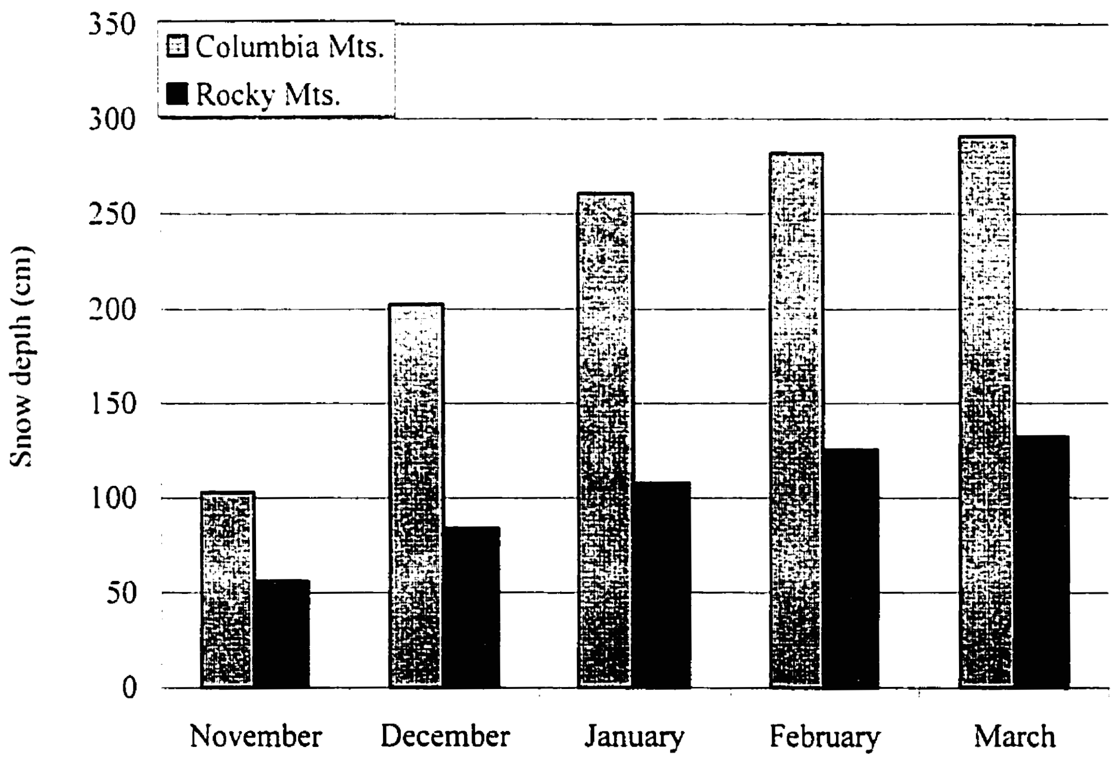

Figure A.5 End-of-month snow depth..

A.6.3 Snowpack temperature gradient

The average monthly temperature gradient averaged over the snowpack was calculated by dividing average monthly minimum temperature by the end of the month snowpack. This calculation estimates the maximum snowpack gradient (Akitaya, 1974; Armstrong and Armstrong, 1987):

$$\text{Temperature Gradient} = HS_{month}/TA_{min} \tag{A.1}$$

where HS_{month} is the end of month snowpack and TA_{min} is the average monthly minimum temperature (Figure A.6).

The snowpack of the Rocky Mountains of Canada is known for its layers of faceted crystals and depth hoar. Typically the temperature gradient averaged over the snowpack is greater than the faceting threshold of 10°C/m (Akitaya, 1974) from the first snowfall

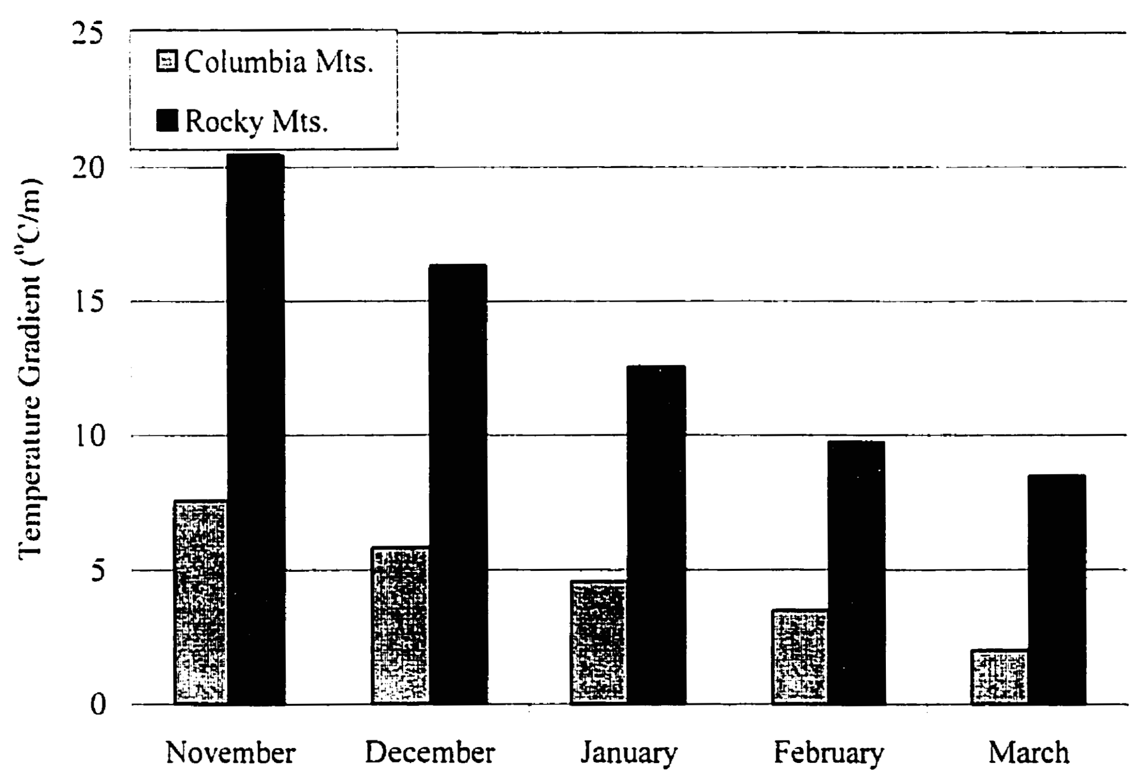

Figure A.6 The monthly snowpack temperature gradient. The two snowpack climates are clearly distinguished.

until the end of February (Figure A.6). The result is many layers of depth hoar and faceted crystals. By March, faceted crystals and depth hoar deep in the snowpack are warm, under a low temperature gradient and start to round. However, near-surfacing faceting still occurs due to localized temperature gradients caused by diurnal fluctuations of air temperature and radiation.

In general, the snowpack temperature gradient never reaches the threshold value of 10°C/m for faceted crystal formation in the Columbia Mountains (Figure A.6). Most of the faceting is due to near-surface temperature gradients (Birkeland, 1998). After layers are buried, a thick snowpack promotes low temperature gradients that are associated with

Figure A.7 Snowpack climates of the western United States and southwestern Canada.

rounding of crystals and strengthening of layers.

A.7 Conclusions

This study has summarized long-term weather and snow data to characterize the intermountain and continental climates of southwestern Canada. The results are the basic step in finding a relationship between climate zone, air temperature, snow depth, snow fall, and temperature gradient.

The data presented should only be used for an indication of snowpack climates in the Columbia and Rocky Mountains of western Canada. Climate data was collected from only two sites in the Rocky Mountains and three sites in the Columbia Mountains.

Despite the limitations associated with this study, the snowpack climates of both mountain ranges are clearly distinct. These identified climates are assumed to be an extension of the intermountain and continental snowpack climates of the United States (Figure A.7). The trends from the average monthly maximum temperatures, minimum temperatures, present temperatures, calculated temperature gradient, snowfall, and the end of month snow depth indicate that the Columbia Mountains are generally warmer and receive heavier snowfall than the Rocky Mountains in winter. The exception is east of the Columbia Mountain divide where in certain locations (e.g. Bobby Burns) the continental snowpack climate exists.

Additional data collected would produce more representative results from each climate zone. In the future a study summarizing continuous data from at least 10 sites in each mountain range (Coast Mountains, Columbia Mountains, Rocky Mountains) would be useful for quantitatively verifying the three climate zones. The results would be useful for determining relationships between climate zone, snowpack structure, avalanche activity, avalanche run-out distances, and comparisons with the snowpack climates of the United States.