WET LOOSE SNOW AVALANCHING IN SOUTHWESTERN MONTANA

by

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ABSTRACT

Wet loose snow avalanches are a significant hazard within many ski areas. Wet snow stability changes dramatically over short time periods which typically coincide with operating hours, and few quantitative tools exist for avalanche workers attempting to predict the onset of wet snow avalanching. Field work was conducted at two study sites in southwestern Montana during the springs of 2003, 2004, 2005, and 2006. The study is composed of three separate experiments. The first documents stratigraphic boundary conditions present during periods of wet loose instability. Results show that melt-water accumulation within the upper 15cm of the snowpack increases the likelihood of wet loose avalanche occurrence. The second focuses on the mean daily and minimum daily air temperatures, and how well each variable indicates wet loose avalanche activity. Results are consistent with prior research and clearly show that temperature alone is not a good indicator. The third relates wet loose snow avalanching to surficial shear strength. A 250cm^2 shear frame was used to make as many as 210 surficial shear strength measurements of melt-freeze snow per day. Changes occurred rapidly within the melt-freeze cycle as shown by highly significant changes in shear strength within half hour intervals. Most importantly, the data shows an apparent association between surficial shear strength and avalanche activity. When shear strength measurements dropped below 250 Pa wet loose avalanches were observed, and triggered, in the immediate vicinity of study slopes. Conversely, surficial stability on the study slope improved when shear strength values exceeded 250 Pa. This research provides insights into wet loose snow avalanching and the development of possible tools for better predicting wet loose snow avalanche occurrence.

CHAPTER 1

INTRODUCTION

Little research exists on wet snow avalanches. There are several reasons for this, but foremost is that historically, dry slab avalanches kill more people and have therefore generated more interest. As a result of the paucity of wet snow research, there is a limited scientific understanding of the spatial and temporal relationship between free water infiltration, its effect on snowpack structure, and the concomitant changes in mechanical properties that result in avalanche activity.

Wet snow avalanches occur in two distinct morphologies: 1) wet slab avalanches and 2) wet loose avalanches. While all wet avalanche activity appears to be a response to an input of free water (McClung and Schaerer, 1993), wet loose and wet slab avalanches have distinct boundary conditions that result in different mechanisms of release. This thesis will focus specifically upon wet loose snow avalanches.

Wet loose snow avalanches occur when the water content of near-surface snow increases to a point where surficial layers lose enough strength that the slope angle suddenly exceeds the static friction angle (McClung and Schaerer, 1993). In this condition, slopes can avalanche in response to a trigger such as skier traffic, or snow falling off rocks or trees. In southwestern Montana, wet loose avalanches generally occur in response to high water contents caused by elevated spring temperatures and solar radiation, but can occur at any time during the winter if elevated temperatures, radiation and/or rain provide a sufficient free water input (Carse, 2003; Romig, 2004; Romig et al., 2004). Wet loose snow avalanches are easily recognizable as point releases that form a triangular pattern on the descent. They can be differentiated from dry loose snow avalanches by the presence of liquid water in the avalanching snow, the presence of snowballs, and well defined 'scour' marks and striations along the bed surface of the avalanche (McClung and Schaerer, 1993). Wet loose snow typically exhibits a relatively high density (in comparison with dry snow) resulting in slower, albeit more destructive avalanches. For example: dry, flowing avalanches have been shown to have a mean impact force of 5 – 30 (ton/m^2) while wet avalanches have an estimated impact force of 30 - 40 (ton/m^2) (McClung and Schaerer, 1993).

Although wet loose snow avalanches are a significant hazard in many operational settings, research relating wet snow strength to avalanche activity is limited and few quantitative tools are available to forecasters. The following review will illustrate what is known about wet snow, demonstrate how that knowledge relates to wet loose avalanches, and develop research questions that form the basis of this thesis.

CHAPTER 2

WET SNOW

Metamorphism of Wet Snow

The introduction of liquid water into snow through warming, radiation, and/or rain directly affects the hydraulic properties and strength of snowpacks (Conway, 1988). Change in particle size, bond growth, and densification not only depend on the presence of liquid water, but also on how much liquid water is available. In this regard, snow scientists use terms developed in soil science to formally describe the free water content of snow (Colbeck, 1973). The pendular moisture regime has low free water content (less than 7% by volume). The funicular moisture regime exists when the free water content exceeds 7% (Denoth, 1980; Colbeck, 1982). In the pendular regime air is continuous throughout the pore spaces and water is found only in isolated cells. The amount of liquid water present is greater than the capillary requirement (irreducible water content), but less than the amount needed to have the capillary rings around neighboring grains coalesce and/or connect (Fig. 2.1). In the funicular regime, water is continuous and air occurs only in isolated cells (Fig. 2.1). The distinction between high and low water content is important because snow exhibiting funicular properties has a lower overall strength than snow at lower water contents (Colbeck, 1982). In addition, infiltration rates are faster when water is continuous throughout the pore spaces, causing a marked difference in flow rates and heat transfer between the two regimes (Salm, 1982).

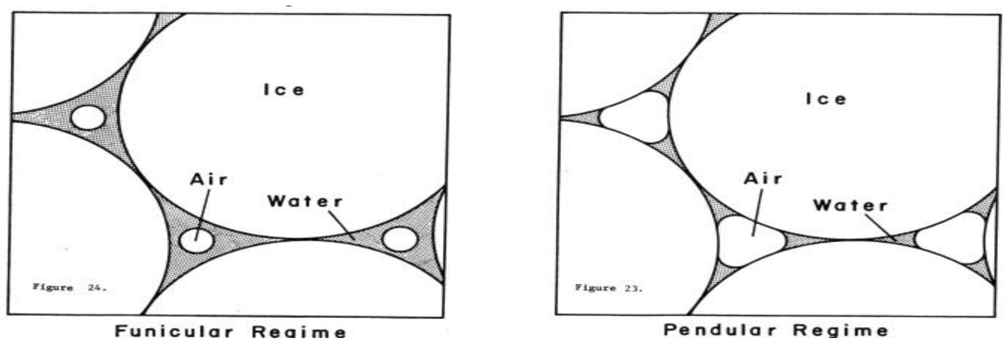


Figure 2.1. Idealized view of wet snow in the pendular and funicular regimes (from Armstrong, 1976, p. 73).

Liquid water serves as a highly efficient medium for heat transfer in wet snow (Salm, 1982). The overall result is the decay of smaller grains and growth of larger grains. In the pendular regime, although water is present, metamorphism is slower than in the funicular regime, but still much faster than that encountered when there is no free water present. Grain-to-grain strength in the pendular regime is highly dependent upon tensional forces between grains (Colbeck, 1982). Energy transfer is limited because the cross sectional area available for heat transport (through water) is small and is shared by areas of gas / solid interface. Minute temperature differences between regions of water tension (capillary forces) and grain surfaces drive the direction of heat flow and subsequent creation or destruction of bonds (Colbeck, 1973). If heat flow is directed towards areas affected by surface tension, melting occurs and bonds are broken. If heat flow moves away from inter-granular regions of surface tension, freezing and subsequent strengthening occur (Colbeck, 1973).

In contrast, the funicular regime is characterized by more rapid metamorphism (grain growth, loss of strength, densification) in response to inter-granular temperature gradients and efficient heat flow between different sized particles (Colbeck, 1973). At the transition between pendular and funicular, air bubbles decrease in size and distribution allowing heat transfer between adjacent grains to be governed by liquid water. Capillary forces cease to play a role in bond formation resulting in a general loss of cohesion between grains.

Movement of Water through Snow

By definition, snow must be at 0^{o} Celsius for free water to be present. For water to infiltrate previously dry snow, a continuous film of water called the 'irreducible' water content must be present between grain boundaries. Explanations of the irreducible water content of snow have varied, but recent research suggests that it depends only on the existing porosity or density of snow prior to wetting (Coleou and Lesaffre, 1998). Water penetrates previously dry snow through channels that occupy a small part of the snow volume (Kattelmann, 1984)(Figure 2.2). In groundwater studies these channels are called unstable saturated zone flow, or 'fingering' (Silio and Tellam, 2000). Preferential flow of this type occurs even in well sorted, homogenous sands, and has been shown to be highly affected by stratigraphic heterogeneity (Silio and Tellam, 2000). Snow is highly stratified, and a discussion of how water flows through other mediums is informative. Several relationships have been suggested for heterogeneous systems (Sililo and Tellam, 2000):

1) Stratification will tend to enhance rather than dissipate fingering.
2) In discontinuously layered systems, funneling will influence the location of fingers.

3) Lateral flow on top of fine grained layers promotes greater flux, and in turn more fingers in the down-dip direction.
4) In systems where a fine grained conductive layer has variable thickness, the amount of fingering and in turn flow will be greatest where the layer is thinnest.
5) Surface depressions in a conductive layer will concentrate flow, and fingers will form below these areas.

Saturated, or unsaturated flow in snow is more complex than that found in sands because porosity, grain size, and/or the resulting transmissivity is dynamic in snow, while these factors are static in sands. In snow, prolonged exposure to higher water contents increases grain size and permeability, allowing established fingers to become preferential flow channels.

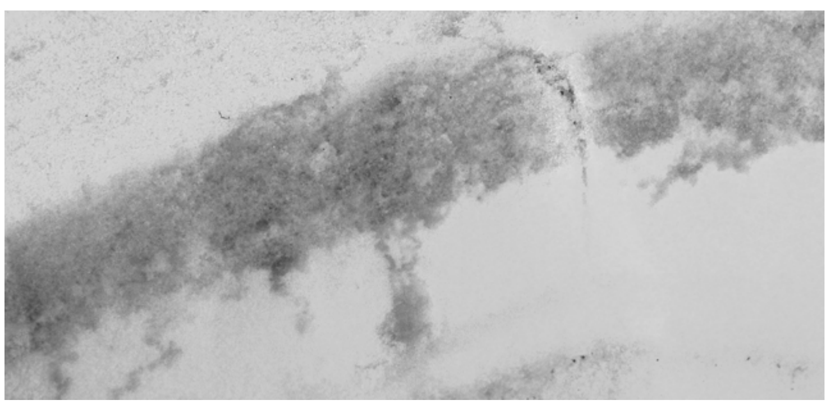


Figure 2.2. Melt-water infiltration of new snow. Powdered dye was placed on the snow surface immediately following snowfall. Fingering and the resulting wetting front can be seen.

Water flow through snow can be modeled using variations of Colbecks' gravity flow theory (Colbeck, 1972; Kattelmann, 1985; Bhutiyni, 1994). In short, the theory explains that water flow through snow is highly dependent upon gravity, and can be seen as an application of Darcy's Law (Colbeck, 1972, 1974, 1978):

$$V = -K\,(d\psi / dz + 1)$$

where V is the free water flux, K is the unsaturated hydraulic conductivity, $(d\psi / dz)$ is the capillary pressure gradient, and 1 is the gravitational pressure gradient. The capillary pressure gradient is negligible in conductive and homogenous snow, but becomes a factor at certain stratigraphic boundaries. This approach is highly simplified when compared to naturally occurring snowpacks and modeling becomes difficult when it is applied to snow (Kattelmann, 1984). By definition, the unsaturated hydraulic conductivity is a variable that changes with grain size and water content. When grain sizes change in a metamorphosing snowpack, the value of K also changes.

Although it has been shown that almost any stratigraphic change can interrupt water infiltration to one degree or another; ice layers, and layers exhibiting changes in grain size and density, can disrupt and/or control water flow in snow (Marsh and Woo, 1985; Conway and Benedict, 1994; Kattelmann and Dozier, 1999). Ice layers can form at the surface of the snow, or within the snowpack through refreezing of water produced by melt and/or rain. Water will move downward until it encounters a textural discontinuity where it spreads out and subsequently freezes (Kattelmann and Dozier, 1999).
Ice layers can effectively stop infiltration and route free water down slope for undetermined distances and times (Kattelmann, 1985). The spatial variability of ice layer characteristics account for distribution of water that encounters that layer. Fine grained snow overlying course grained snow is very effective at delaying and/or routing

infiltration because of the differences in capillary forces between layers (Colbeck, 1974). On a grain to grain scale, water is held more tightly in smaller pore spaces (through capillary attraction) than in larger pore spaces. Water will continue to accumulate in the upper, fine grained layer until the pressure between the two layers is equal (Kattelmann, 1985). For example, if there was only void space below the retaining layer, the pressure at the interface would have to reach atmospheric pressure (saturation) before water can drip into the void.

Saturated layers are examples of funicular regimes in which wet snow metamorphism is rapid. Grain size and conductivity increase and result in the establishment of a positive feedback loop in the system (Wakahama, 1968). Once a pattern of flow is established in a wet snowpack, water will continue to follow this pattern and preferential flow is established and becomes more efficient as time progresses. This process can be halted by freezing and precipitation, but research shows that in time flow will return to previously established patterns (Conway, 1994; Kattelmann and Dozier, 1999).

In highly evolved spring snowpacks, dendritic patterns form on the snow surface. These patterns can be recognized as surficial depressions parallel to the maximum inclination of the snow surface (Higuchi and Tanaka, 1982). Investigations of grain morphology and water content below these depressions revealed a coarser grain size and higher water content than adjacent snow samples. In three independent studies, spatial analysis of the distribution of melt-water flowpaths has shown correlation lengths of 5-7 meters in the mountains of Wyoming and Montana (Williams et al., 1999). These results suggest that the flow pattern in wet snowpacks develops into organized systems.

The Mechanical Properties of Wet Snow

The mechanical strength of wet snow is directly related to the amount of free water contained in a given unit of snow, and its interaction with the grain to grain structure. Techniques developed in order to measure the water content of snow are explained in Appendix A. In contrast with dry snow, where deformation is a result of the creep of ice (deformation and movement of individual snow grains), the major mechanism of deformation in wet snow is regelation (pressure melting and refreezing) at stress free grain boundaries (Salm, 1982). This distinction is important because regelation in wet snow has been shown to be a fast process that is dependent upon the stress per unit of snow (Colbeck, 1979) instead of the relatively slow process of the migration of individual particles of ice (Salm, 1982).

Several studies focus on the shear strength of wet snow (Perla et al., 1982; Brun and Rey, 1987; Bhutiyani, 1994). Although strong correlations exist between dry snow density and shear strength, poor correlation exists between wet snow shear strength and measures of density (Perla et al., 1982; Brun and Rey, 1987). Perla et al (1982) measured the shear strength index of alpine snow. The index was stratified according to snow crystal morphology, then correlated with measurements of snow density, temperature, and crystal size. The correlations of the shear index and density were significant for all crystal types ($r > 0.8$ for rounded granular and faceted granular snow) except those of melt-freeze morphologies ($r = 0.276$). This phenomenon was attributed to the observation that with little discernable change in crystal morphology, the strength of melt-freeze snow can change dramatically with the addition of liquid water. Brun and

Rey (1987) concluded that an estimation of shear strength based upon the physical description (primarily density) of a snow sample is only valid for dry, fine grained or fresh snow. In addition, they discussed the influence of water content on shear strength, and found little change in the shear strength of wet snow at or below 6% water content by volume (they did not sample anything with a water content higher than 6%). Bhutiyani (1994) reached similar conclusions, finding that density as a sole predictor of wet snow shear strength is inadequate. In addition, he showed that a basic assessment of grain size was a significant factor when correlating density and strength. Samples with crystal size smaller than 1mm provided a significantly better correlation than grain sizes larger than 1mm ($r = 0.87$ vs. $r = 0.53$ respectively), an observation that is plausible if the assumption is made that crystals smaller than 1mm have seen little melt-freeze metamorphism, and have more contact points between crystals per unit of snow. Bhitiyani (1994) improved the correlation for grain sizes larger than 1mm somewhat by developing a correction factor that included both density and water content, but the predictive capability was still fairly low ($r = 0.65$). Study of the shear strength in relation to water content showed no significant changes up to 6%, but dropped by a factor of 2 once the water content exceeded ~7% by volume (consistent with Colbeck, 1982). There is some question as to why no significant change is noted in shear strength between 0 and 6% water content. Although Colbeck (1973) theorized that in the pendular regime, capillary strength is not sufficient to compensate for the disappearance of bonds by melting, the fact that research has shown no appreciable change in shear strength with changes in water content (pendular snow) suggests that even though capillary pressure is not high enough to counteract the disappearance of bonds, it is strong enough to limit the amount of water available for bond degradation (Bhitiyani, 1994). In funicular snow, water exists continuously between contact points and bonds can be completely degraded.

Research has repeatedly shown that wet snow loses its strength at the transition between the pendular and funicular regimes, or when its water content reaches about 7% by volume (~14% pore volume)(Ambach and Howorka, 1965; Colbeck, 1982; Bhutiyani, 1994). Wakahama (1968) recorded a drop in hardness of 2 orders of magnitude in a natural surface layer from the morning before melt and at noon when the water content of that layer reached 20 - 25% pore volume. This rapid change in wet snow strength as it transitions from the pendular to the funicular regime helps explain why the onset of wet snow avalanches is often rapid and dramatic.

Wet Snow Stability

Previous research has highlighted several relationships pertaining to wet snow stability. Armstrong (1976, p. 74) states:

> 'when the bulk of the snowcover becomes isothermal, the immediate potential for wet avalanche release is greatly increased…Once the deteriorating strength of the snowcover (due to melt-water infiltration) reaches the point when it can no longer resist gravitational stresses, it will release as either a loose or wet slab avalanche, depending on boundary conditions. These boundaries may be caused by stratigraphic irregularities within the snow cover…While the slab type is often of greater magnitude, due to its release over a broader area, wet loose avalanches can also incorporate a large amount of snow depending upon how deep into the snow cover the percolation of melt-water has advanced prior to release, and how much additional snow may be released by the moving avalanche.'

Armstrong (1976) also noted that wet avalanche cycles at Red Mountain Pass, Colorado coincided with mean daily air temperatures above $0^{o}C$, though temperatures were taken at a study site and may or may not have reflected actual starting zones temperatures. Kattelman (1984) showed that stratigraphy provides a control on melt water movement throughout the snowpack that can ultimately result in the formation of instabilities associated with high accumulations of free water along stratigraphic boundaries. These boundaries may be ice layers or changes in hardness, changes in grain size (small grains overlying large grains), or in some cases simple textural discontinuities such as minute wind crusts (Wakahama, 1968; Colbeck, 1973; Kattelmann, 1984; Conway, 1994).

When free water accumulation reaches the funicular regime (above 7% by volume), there is a noticeable loss of strength and the potential for a resulting instability increases (Ambach and Howorka, 1965; Colbeck 1982). Unfortunately, although water content in the funicular regime seem to be the root of most wet loose avalanche events; direct measurements of water content, especially the water content of specific strata in the field, are difficult (Appendix A). Kattelmann (1984) suggests that in terms of forecasting wet snow avalanches, an assessment of where water might concentrate within the snowpack might be of much greater practical value than attempting to define the specific water content of any given strata. Kattelmann (1984) also noted increased stability as vertical flow paths were established within the snowpack. Theoretically, once flow between layers is established, the snow pack drains efficiently enough to prevent water accumulations from reaching the funicular regime and subsequent instabilities from occurring.

Wet Loose Snow Avalanching in Southwestern Montana

In southwestern Montana, wet loose snow avalanches primarily occur when surficial snow loses strength in response to intensive periods of melt water production in March and April (Johnson, 2002; Romig, 2004; Romig et al., 2004). Prediction is challenging. Water does not always accumulate along surficial stratigraphic interfaces, nor does avalanche activity or instability always occur on warm, sunny days. There are no carefully documented stratigraphic analyses (snow pit profiles) that coincide with periods of wet loose avalanche activity in southwest Montana.

Romig (2004; Romig et al., 2004) conducted a statistically based study using archived data collected by Bridger Bowl Ski Area for March from 1968 – 2001. Several predictor variables are linked to wet avalanche occurrence. These predictors include the minimum daily temperature, change in total snow depth (an estimate of settlement), and the three-day cumulative new snow water equivalent. Even though these predictor variables could be linked to wet snow avalanches, the relationships were too weak to provide an operational forecasting tool.

Wet loose avalanches occur when snow loses strength, but how much strength must be lost for avalanches to begin? Documented research is not available that directly relates the shear strength of wet snow to avalanche activity. Wet loose snow avalanches occur when near surface layers lose enough cohesion (inter-granular strength) that the slope angle exceeds the static friction angle. During a typical melt-freeze cycle surficial strength is high when layers are frozen, decreases during the day in response to elevated temperatures and radiation, and increases again as temperatures and radiation inputs decrease (Carse, 2003). Preliminary studies by this author show that use of a 250cm^2 shear frame can detect changes in the strength of surficial snow over the course of a day. If changes in the shear strength of surficial snow can be documented throughout targeted melt freeze cycles, and subsequently correlated to wet loose avalanche activity, it is possible that quantified methods of forecasting and prediction can be developed.

The literature suggests there are several field observations and measurements that might be used to predict periods of increased wet-snow-avalanche hazard. Bridger Bowl is particularly interested in this problem because mitigation and control techniques used in dry snow are not as effective in wet snow. This thesis reports on several experiments conducted to improve predictive capabilities for the wet loose snow avalanche hazard. The following three chapters are presented as independent modules that contain the research questions, methods, results, and discussion of that particular chapter. The first will address observations and stratigraphic documentation made during and between observed wet loose avalanche cycles in southwest Montana. The second will explore the relationship between mean daily air temperature, minimum daily air temperature, and observed wet loose avalanche activity. The third will relate surficial shear strength to wet loose snow avalanching.

CHAPTER 3

RELATING SNOWPACK STRATIGRAPHY TO WET LOOSE SNOW AVALANCHING

Wet loose snow avalanche forecasting and management can be improved by documented descriptions of wet loose avalanche activity and the characteristics of the snowpack revealed in snowpits. For example, snow in the funicular regime is weaker than snow in the pendular regime (Denoth, 1980; Colbeck, 1982), but we don't know if 'very wet' layers always result in instability. Bridger Bowl uses measurements of the snow temperature 15 cm below the surface to determine whether the surficial snow is isothermal, and to therefore help assess the risk of wet snow avalanches (Johnson, 2002), but the efficacy of this procedure has not been documented. Snowpit profiles are typically not available for days with wet loose avalanche activity, so no associations can be made between stratigraphy and avalanche occurrence. In summary, documentation is needed to show the timing of wet loose avalanche events in relation to patterns in stratigraphy that precede wet loose avalanche activity. The following questions were developed in order to assess the relationship between stratigraphy, melt-water accumulation, and wet loose avalanche release during March and April of 2003 and 2004 in the Bridger Range of southwestern Montana:

1. Is the depth of isothermal conditions (0^{o}C) an indicator of wet avalanche activity?
2. Do layers of very wet snow always result in wet loose avalanche activity?
3. Are there certain stratigraphic sequences that result in wet loose avalanche hazards?

The available literature states that surficial accumulations of free water lead to wet loose avalanche release through a decrease in the cohesion of surficial snow (McClung and Schaerer, 1993). The literature does not, however, talk specifically about snowpack temperature, or stratigraphic sequences that may influence the development of weak layers. I hypothesized that wet loose avalanche activity would occur in conjunction with the seasonal change from a 'cold' to 'warm' snowpack. In addition, the experiment was constructed to document snowpack stratigraphy during periods of stability, and periods of instability, in order to test whether the presence of a certain stratigraphy can be considered to be dangerous.

Methods

Snowpack stratigraphy was documented at least once a week, with additional documentation during periods of warming or as field time allowed. Snowpack profiles were documented at the North Boundary and South Bradley study sites (Fig. 3.1). The North Boundary site is east facing with a steepness of ~32 degrees and an elevation of 2347 meters. The South Bradley site is south east facing with a steepness of 30 degrees and an elevation of 2286 meters. Snow profiles were compiled through consecutive right to left (looking uphill) excavation starting at the lower end of each study area, and were conducted in accordance with the *International Classification of Seasonal Snow on the Ground* (ICSSG) (Colbeck et al., 1990) to include layer depth, temperature, hardness, density, crystal morphology, and water content. Density was calculated using a 200cm^3 triangular density kit. The presence of free water was determined by visualization with a hand lens and the relative water content assessed by the hand squeeze test as outlined in

ICSSG (Colbeck et al., 1990). In some cases it is difficult to classify wet and very wet snow using the hand test due to the large range of water contents covered by this method of classification (3-8% for wet snow and 8-15% for very wet snow). For example, common; to have a very wet layer underlain by 'wetter' snow along a stratigraphic interface. When this occurs, water can be seen as a darker (more gray) layer with the naked eye. Cases such as this were assigned a water content of 'slush', but the crystal types were categorized as seen. 'Snow Pilot' graphical software was used to document snow profiles. The term melt-water 'horizon' is used as a descriptor for this situation. Profiles from days showing significant instability were compared to profiles from days showing relative stability in order to determine if there were significant stratigraphic differences.

The presence of wet loose avalanche activity and timing during the study period was documented by onsite monitoring of the eastern Bridger Range from the North side of Saddle Peak to Wolverine Basin (Figure 3.1). In March and April of 2003 and 2004, observations were conducted as field time allowed; days with elevated temperatures were targeted, but a continuous stratigraphic record and documentation of avalanche activity was not achieved. The goal was to be in the field during various melt cycles in order to document stratigraphy in periods of stability and periods of instability. When the locations of witnessed avalanches were accessible, the depth to the bed surface and the bed surface wetness were assessed. These observations were then compared to the day's stratigraphy (snow profile for the day) and the unstable layers highlighted. Avalanche destructive potential was estimated using guidelines set by the Canadian Avalanche Association (McClung and Schaerer, 1993), and subsequently adopted as the "destructive" scale in the U.S. classification (Greene et al., 2004). This method of classification was chosen because it allows visualization and comparison of avalanches from a variety of avalanche paths.

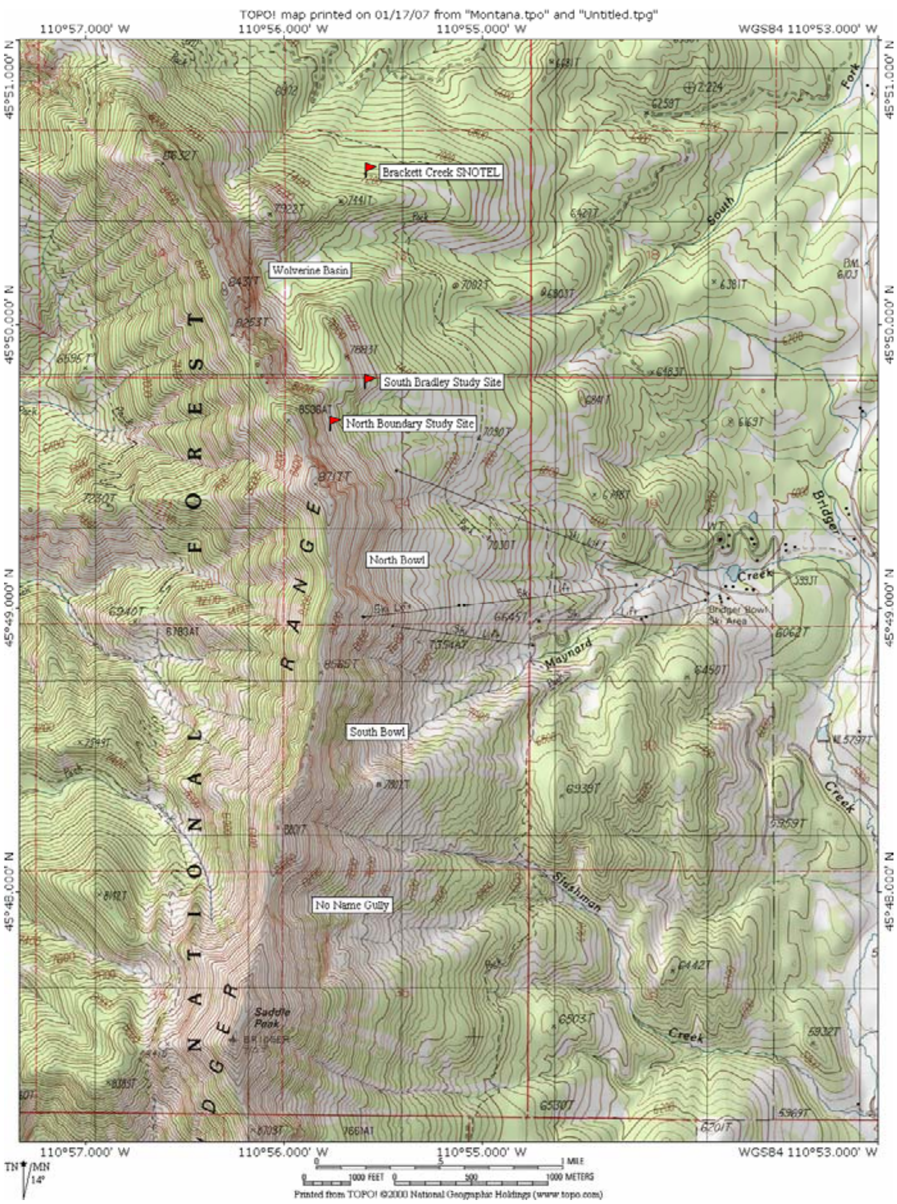


Figure 3.1. Study area map. Observations were made throughout April of 2003 and 2004 along the eastern range front between Saddle Peak in the south and Wolverine Basin to the north.

Results

Field observations of avalanche activity and snowpack stratigraphy were conducted in the spring of 2003 and 2004. Avalanche days versus non-avalanche days are defined by the presence of one or more observed wet loose avalanches over the course of that day. A total of 42 wet loose snow avalanches ranging in estimated size from D1 to D3 were documented. Figure 3.2 portrays observed wet loose avalanche occurrence during spring of 2003 and 2004. Twenty-seven class D1 avalanches, thirteen class D2 avalanches, and two class D3 avalanches were observed. The majority of these (with the exception of 4 avalanches in March of 2004) occurred in April. Table 3.1 lists the location and estimated destructive size of witnessed avalanches.

Table 3.1. Date, size, and occurrence of wet loose avalanches observed in the Bridger Range during April of 2003 and March and April of 2004. Forty-two wet loose avalanches were observed during this time period.

Date	# Avalanches	Size (D1, D2, D3)	Location
4/8/2003	2	1	No Name
04/11/03	5	1,2	North Boundary - 3 class 1; 2 class 2
04/12/03	4	1	South Bowl
04/13/03	1	2	North Boundary
04/21/03	6	1,2	North Bowl - 2 class 2, 2 class 1; Bridger gully - 1 class 2; Apron - 1 class 1
04/22/03	2	2,3	Bridger Gully - 1 class 2; Apron 1 - class 3
03/24/04	1	2	Hidden Fan
03/29/04	3	1	North Boundary
04/10/04	3	1,2	No Name - 2 Class 2; North Boundary - 1 class 1
04/11/04	2	2	South Bowl
04/12/04	3	1	North Boundary
04/14/04	1	1	Hidden Fan
04/21/04	3	1,2	South Bowl - 1 class 2; North Bowl - 2 class 1
04/22/04	1	2	Apron
04/23/04	2	1	North Bowl
04/26/04	1	3	Hidden Fan
04/27/04	2	1	North Boundary

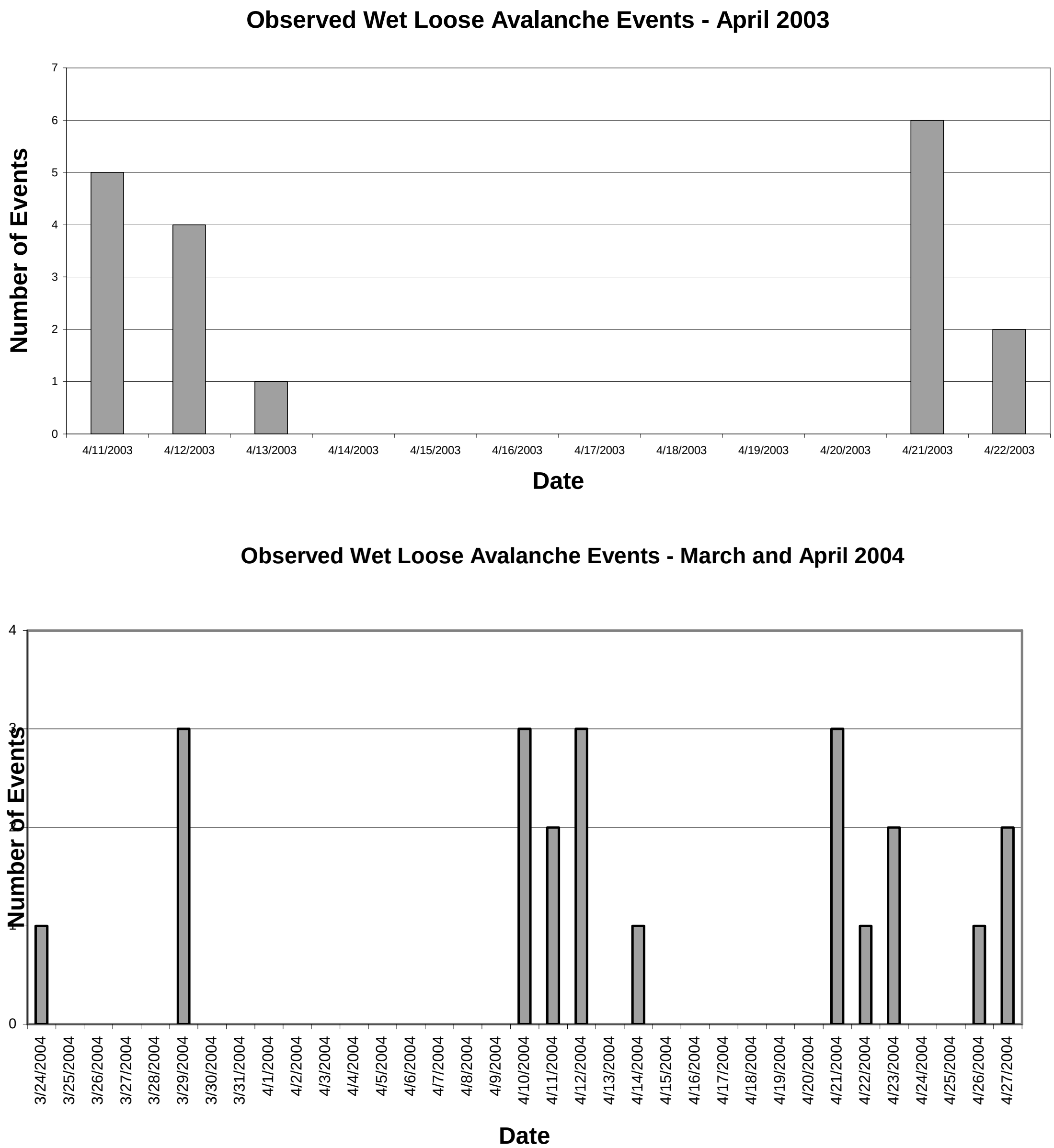


Figure 3.2. Observed wet loose avalanche occurrence in the Bridger Range, 2003-2004.

Snowpack stratigraphy documented during the study is located in Appendix B. Appendix B1 identifies the symbology used to develop snow profiles. Twenty profiles from March and April of 2003 and 2004 are included in this paper and can be seen in Appendix B2-B21. The maximum time between stratigraphic documentation was 6 days, but the average was 3-4 days during periods of stability, and 1-2 days during periods of instability. Ten of the profiles are from avalanche days and ten profiles are from non-avalanche days. Avalanche versus non-avalanche days are defined by the presence of one or more observed wet loose avalanches over the period of that day. Figure 3.3 shows the relationship between bed surface depth and wet loose avalanche size for eleven observed avalanche events for which data on the bed surface depth are available.

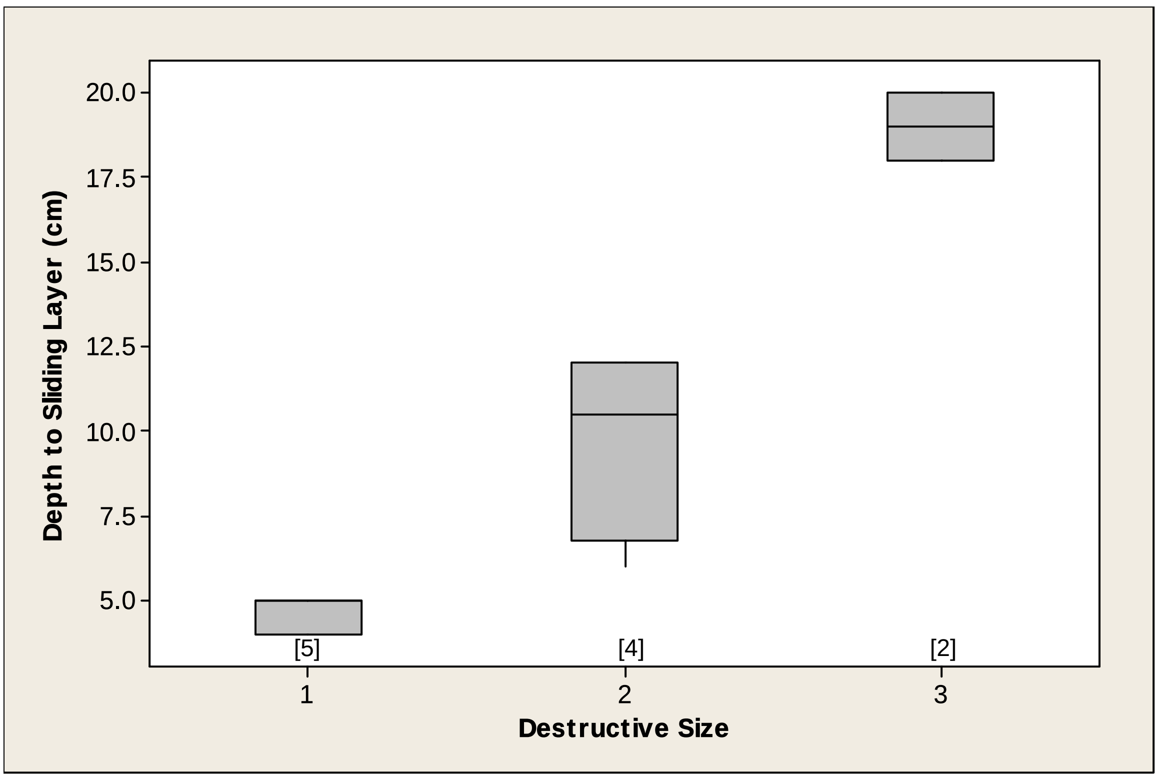


Figure 3.3. March 2003, March and April 2004 observed wet loose snow avalanche size in relation to bed surface depth. The line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. Bracketed numbers below each group are the number of individual shear frames in the sample.

Discussion

The observation set is biased because the numbers used in this study are based on field observations that were made as field time and logistics allowed. Avalanches might have occurred in the Bridger Range when no observers were present, or other avalanches may not have observed. However, the observations presented do generally reflect the conditions present during various melt cycles in the Bridger Range during April of 2003 and March and April of 2004.

Depth of Isothermal Conditions as a Factor in Wet Loose Avalanche Activity

Experience at Bridger Bowl has shown that 15 cm of isothermal surficial snow available for transport can result in wet loose avalanche hazards (Carse, 2003). In the early stages of the project, it was hypothesized that the depth of wet avalanche instability would correspond to the depth of isothermal conditions within the snowpack, or that as the snowpack changes from a ‘dry’ or ‘cold’ snowpack to an isothermal snowpack (‘wet’, warm), instabilities associated with wet avalanche activity occur at levels consistent with isothermal temperatures (and the resulting depth of melt water infiltration). This hypothesis was not tested due to the fact that both field seasons were initiated after the transformation from a ‘cold’ to a ‘warm’ snowpack (Appendix B, note that temperature profiles for all of the documented snowpits were within 0.5 degrees of $0^{o}C$). However, the observations presented show that wet loose avalanche activity is common in isothermal snowpacks, or snowpacks already subjected to prolonged melt-freeze metamorphism.

In a three phase system at zero degrees Celsius water can exist as ice, air, or water. Analysis of avalanche days in relation to documented snowpack stratigraphy suggests that, in an isothermal snowpack, surficial instability is not as dependent upon differential snowpack temperatures as it is upon the water content or accumulation in near surface snow. There is a semantic issue that is sometimes encountered in wet loose avalanche mitigation. While some practitioners consider danger levels to increase when 'surficial snow becomes isothermal', many documented non-avalanche days show wet snow at 0^{o}C (isothermal throughout the snowpack) that is not unstable. A more accurate way to describe the problem would be to say 'wet loose snow avalanche danger increases when a near surface layer or layers become very wet' (Appendix B), this relationship will be discussed further in the next section. In the Bridger Range, class D1 wet loose avalanches occurred when the bed surface was up to 5cm below the snow surface, class D2 avalanches occurred at depths between 6 and 12cm (the majority had depths above 9cm), and class D3 slides occurred at depths between 18 and 20 cm (Fig. 3.3).

Melt-water Accumulation and
Wet Loose Snow Avalanche Release

Prior research has shown that melt-water accumulates at a variety of stratigraphic boundaries (Wakahama, 1968; Kattelmann, 1986; Conway, 1994). The results of this research are consistent with previous documentation and show water accumulating along ice lenses, changes in crystal size (small to large was most prevalent, but examples of water stopping at large to small boundaries was also noted), changes in crystal type, and changes in hardness(Appendix B2-B21). With the exception of ice lenses, for every example of water accumulating at a specific type of boundary, there is an example of water moving through a similar boundary. Factors such as the presence of flow paths, the measure of existing conductivity versus the amount of free water available for transport, and slope steepness contribute to this phenomenon. Due to the dynamic nature of melt-water flow in snow, and the experience gained during this study, an assessment of where water is accumulating on a certain day is more important and more practical than where water may accumulate in the future in terms of forecasting avalanches.

Kattelmann (1985) hypothesized that snowpack stability is high when melt-water is passing through the snowpack unhindered, or when the input (of melt water) equals the output. In other words, when the amount of melt-water created is consistent with the flow of water through the snow pack, high levels of accumulation do not occur and weak layers are not present. All observed wet loose avalanche days with corresponding snow profiles exhibited an apparent 'horizon' or layer of visible water (appears darker than surrounding snow, and can be seen with the naked eye soon after excavation of the pit) that served as a weak layer (Appendix B-2, 3,7,9,13,14,16,19,20,21). Figures 3.4 and 3.5 depict how powdered dye was used to illustrate the occurrence of a weak surficial layer.

Figure 3.4. Melt-water accumulation, or ‘horizon’ along a surficial stratigraphic interface. Powdered dye was applied to the surface prior to the daily melt cycle.

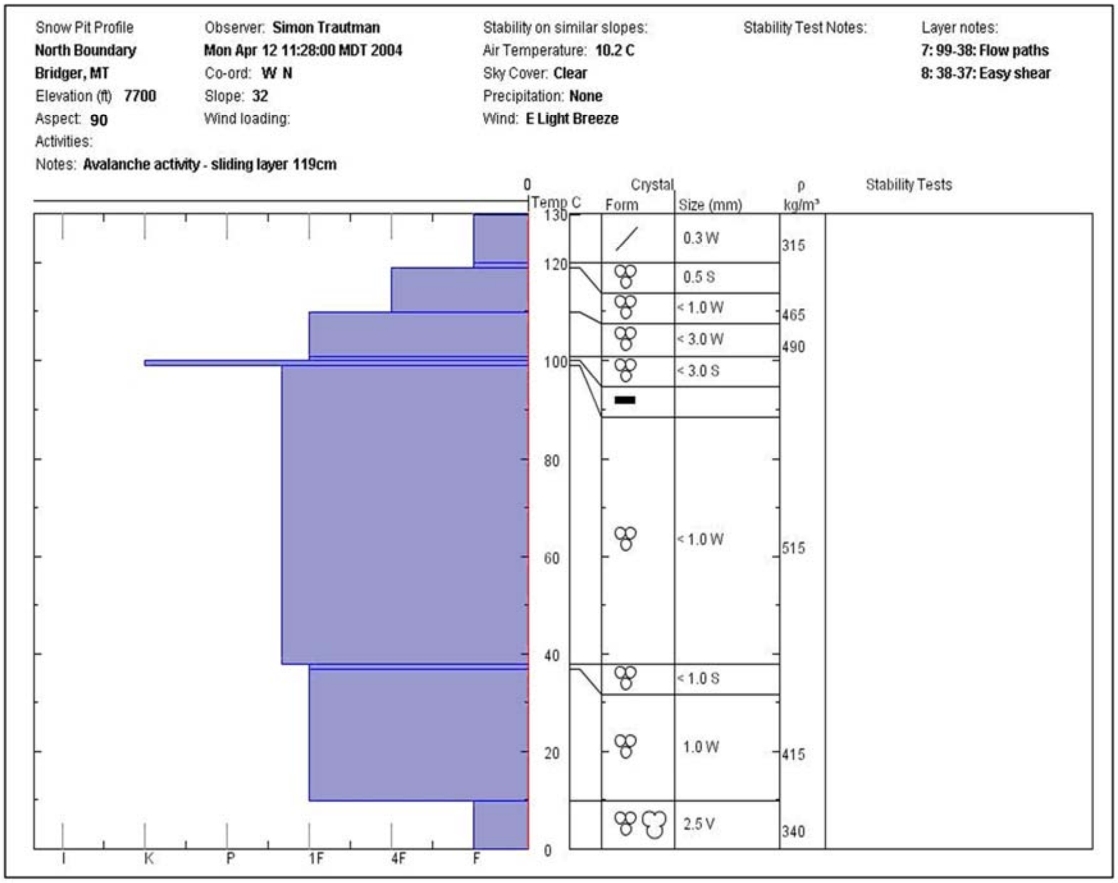


Figure 3.5. Melt-water accumulation at 119 cm resulted in wet loose avalanche activity. All days classified as 'avalanche days' can be identified by a slush layer (S), or melt-water 'horizon' in the upper 15cm of the snowpack. Water content is designated under the size category.

Surficial instabilities such as the one depicted in Figure 3.4 and 3.5 typically persist until the stratigraphic boundary has been compromised, at which point water can move freely along established flow paths (Figure 3.6). Documented non-avalanche days are easily identifiable by the lack of visible melt water accumulation along stratigraphic interfaces in the upper layers of the snowpack. (Fig. 3.7; Appendix B - 3,5,6,8,10,11,12,15,17,18). It is important to note that the upper layers of the snowpack can be classified as 'wet' and still be stable. Instabilities occurred when a layer with high water content or ‘slush’ layer was present in the upper 15cm. Although these layers can be seen with the naked eye soon after excavation of the snowpit, the determination between wet snow, very wet snow, and slush is somewhat ambiguous because the hand squeeze test does not offer the resolution needed for quantifiable results. In order to quantitatively determine the true wetness of individual layers, new technology must be developed for use in a field setting.

Figure 3.6. Melt-water penetration to the depth of the snowpack. In this picture water initially accumulated along the well defined near surface layer (see Figure 3.4), until that layer was compromised (in this case a dog walked over the study plot). After the surface layer was compromised, dye immediately penetrated to depth.

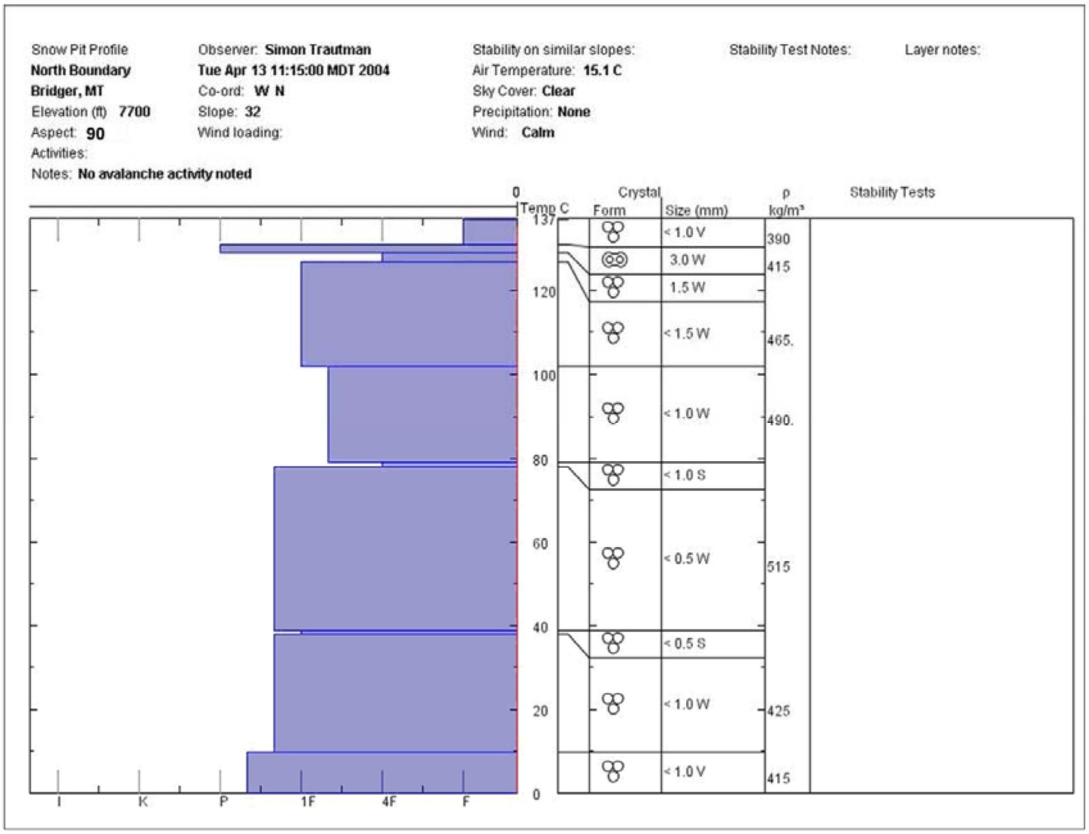


Figure 3.7. No visible accumulation of melt-water in surficial snow. Non-avalanche days are recognizable by the lack of high water content (in this case classified as slush (S)), or melt-water 'horizon' in the upper 15cm of the snowpack. Water contents are designated under the size category.

The snow profiles indicate that melt-water 'horizons' within 15cm of the snow surface coincide with the occurrence of wet loose avalanche activity. However, there are many examples of melt-water 'horizons' at deeper levels within the snowpack that did not act as a weak layer (Figure 3.8; Appendix B-15, B-16, B-17, and B-19). These layers reacted to shovel shear tests and displayed the typical lack of strength associated with high water contents in snow, but they did not produce avalanches. In all cases the layers in question were found between strong, dense layers composed of moist to wet, well rounded and clustered melt freeze crystals.

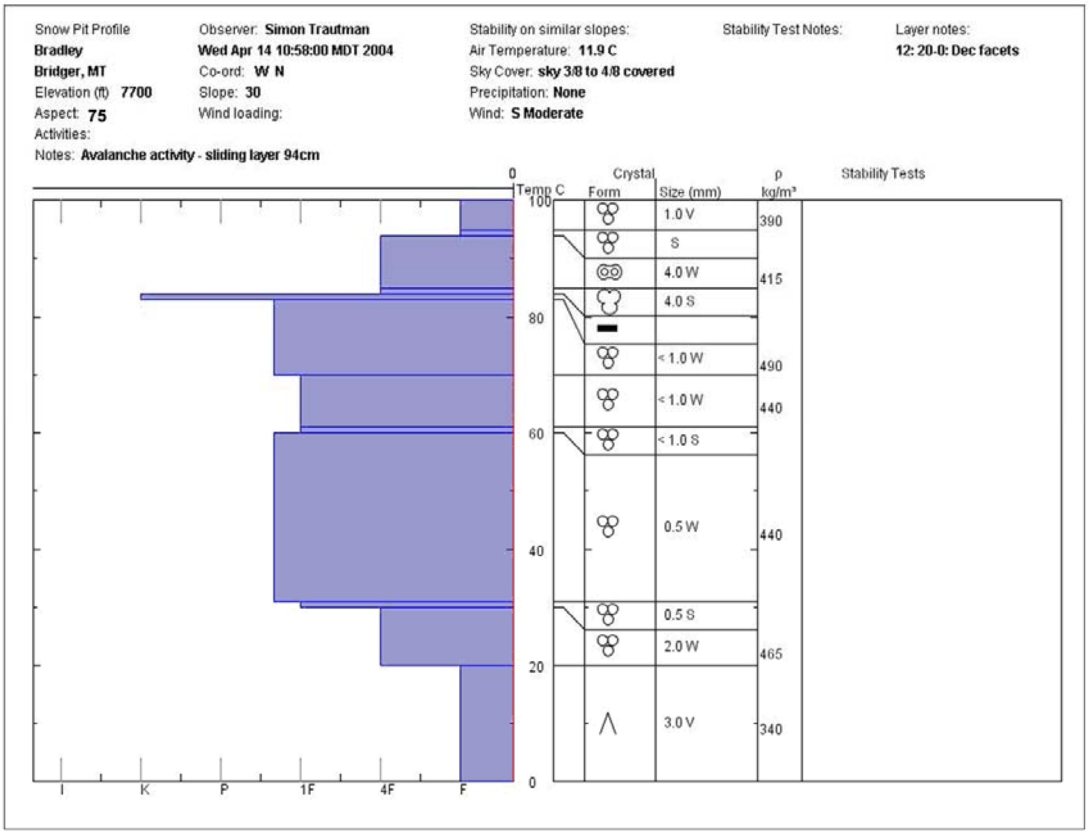


Figure 3.8. Example of melt-water accumulation at lower levels in the snowpack that was not associated with avalanche activity (60cm and 30cm layers). Water contents are designated under the size category.

Stratigraphy Present in Conjunction with Observed <u>Wet Loose Snow Avalanche Hazards</u>

Wet loose snow avalanches are generally not associated with the critical danger and inherent risk found in dry slab avalanches. There are several reasons for this, but the most important is that in most conditions they are small, relatively slow, and when triggered the resulting avalanche moves down and away from the trigger (instead of breaking above the trigger as is the case in many dry slab avalanches). That being said, wet loose avalanche hazards do exist, and based upon the density and water content of the snow involved, demand respect. For the purpose of this discussion, an avalanche hazard is defined as the ability of an avalanche, independent of terrain factors, to seriously injure, or kill, a person caught in the avalanche.

Avalanche size is dependent upon the amount of snow available for transport by that avalanche (McClung and Schaerer, 1993). The results presented in Figure 3.3 are consistent with this statement and suggest that 9 cm or more of wet surficial snow available for transport can result in class D2 and larger avalanches which, by definition, are large enough to bury, injure, or kill a person. Of the observed avalanches listed in the results, two (out of forty-two) were classified as D3 avalanches and represented serious avalanche hazards. The first of these occurred on 22 April 2003 (Appendix B-4) around 11:00. This avalanche entrained the upper 20 cm of the snowpack, moved at an estimated 22 m/s and had a debris pile 2.5 meters deep (Fig. 3.9). Several of the snowballs in the debris were over 2 meters in diameter. The second occurred on 26 April 2004 (Appendix A-20) at 14:30, entrained the upper 18cm, moved approximately 15 m/s, and had a debris pile of 1.5 meters deep (Fig. 3.10). In both cases the sliding surface was not the point at which water had been accumulating throughout the early part of the day, but was a lower interface between large (+3mm), very poorly bonded (fist hardness) poly-crystals and an underlying layer of well-bonded melt freeze rounded grains. The layer of poly-crystals had very little strength between individual crystals; a hand full could be sifted between the fingers. Melt-water accumulation occurred immediately above the poly-crystalline layer, within a well-defined melt freeze crust that seemed to loose much of its integrity immediately prior to both avalanches. For example, on 22 April 2003, boot penetration at 08:00 was 5cm and at 10:45 it had increased to 20cm, which corresponds to the depth of the bed surface (the avalanche occurred at 11:00). On

26 April 2004, use of a ram penetrometer revealed a substantial crust at 87cm (depth at which initial ram placement was stopping), at 14:00, ram placement dropped to the depth of the bed surface at 77cm. The avalanche occurred at approximately 14:30.

On both of these days, the snowpack had a surface crust and skiing was extremely difficult. Ski tips tended to dive intermittently with very little chance of coming back to the surface once the crust had been broken through. Ski pole tests revealed a ~10cm surficial crust underlain by another ~10cm of very poorly bonded poly-crystals. Mean daily air temperatures were warm, but unremarkable; 6 and 4^{o}C respectively. Snowpack stratigraphy used in this analysis was not collected from avalanche paths, but from representative slopes in the near vicinity. These slopes were chosen for similar aspect, steepness, and elevation.

Figure 3.9. Wet loose snow avalanche debris from the Apron slide path, Bridger Bowl, April 22, 2003.

Figure 3.10. Wet loose snow avalanche debris from the Northwest Passage avalanche path, Bridger Bowl, 26 April 2004.

A moderately hard melt freeze crust at the surface is a typical characteristic of the snow pack in the Bridger Range in April. The layer of large, poorly bonded poly-crystals immediately below the crust is atypical (Fig. 3.11 and 3.12).

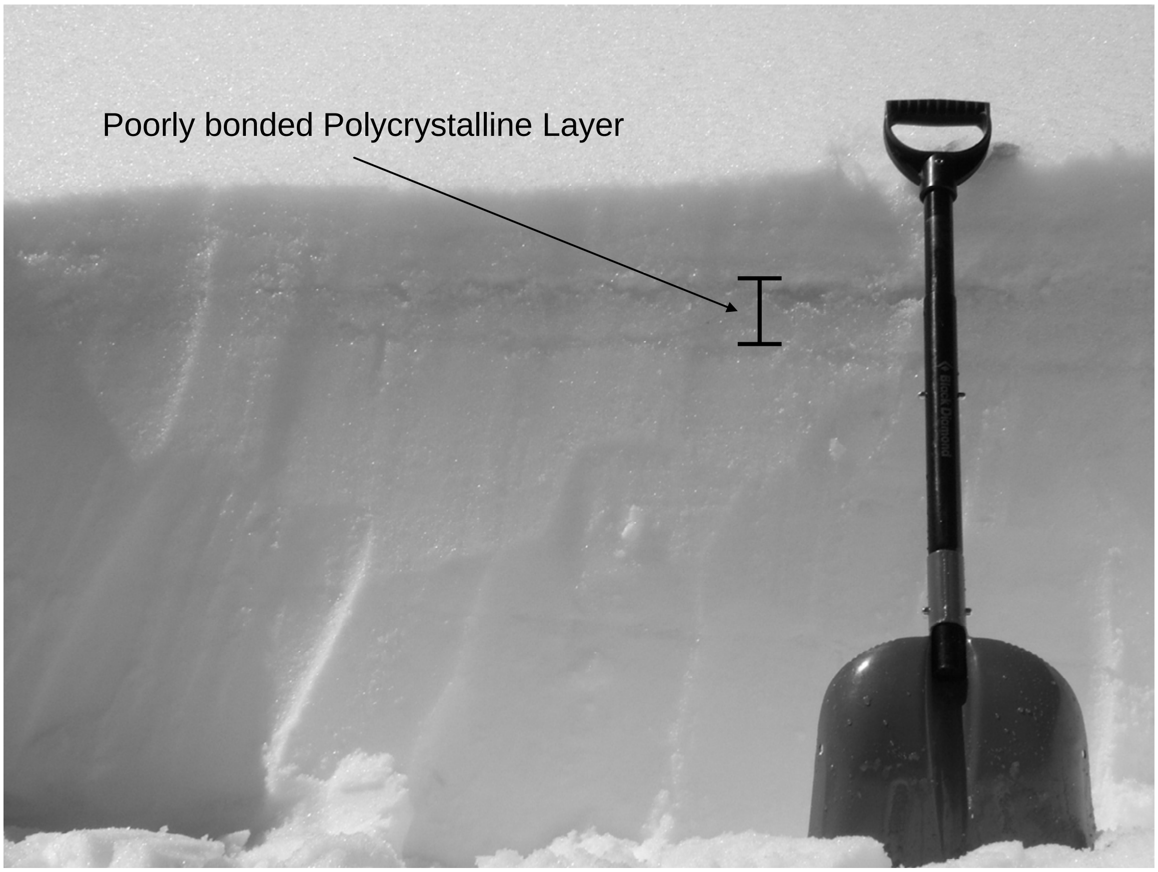


Figure 3.11. Atypical, weak polycrystalline layer involved in the 26 April 2004, wet loose snow avalanche (please refer to Figure 3.12 for stratigraphy related to this picture).

It is not known if the two avalanches described above resulted from a rapid loss of strength in the poly-crystalline layer due to a pulse of melt-water being released from the overlying crust when integrity was lost, or from simple entrainment of the weak poly-crystalline layer following avalanching in the surficial layers once the overlying crust lost strength. Either way, the presence of this poorly bonded layer allowed a much greater amount of snow to be incorporated resulting in much larger avalanches.

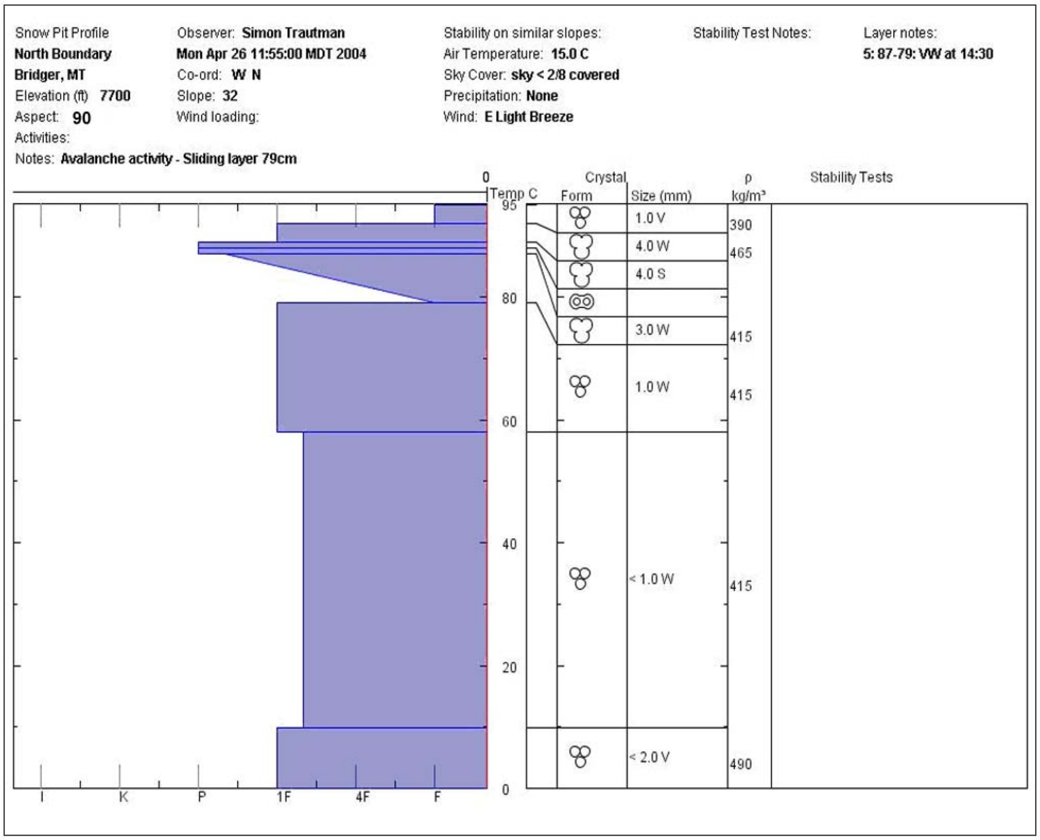


Figure 3.12. The 80cm – 90cm layer is an example of the atypical polycrystalline layer involved in the 26 April 2004 event.

Conclusions

Stratigraphic analysis can be very useful in operational settings when conducted on a given day. Snow pits must be placed in a location representative of the avalanche path in question. In the area studied, 8cm or more of snow available for transport can produce class D2 and larger avalanches. This observation may be highly dependent upon terrain factors. The availability of additional snow for transport (such as the polycrystalline layer shown in Figure 3.8) increases the existing hazard.

Wet loose avalanche instability is dependent on the amount of melt-water being retained in the upper layers of the snowpack and the existing stratigraphy is the dominating factor in melt-water accumulation. Once flow paths are established and the snowpack drains efficiently, stability improves. Springtime snowfall can result renewed instabilities until drainage to the lower snowpack is established. Although melt-water 'horizons' in the upper 15cm of the snowpack are associated with wet loose avalanche activity, prediction is not simple. In many cases it is difficult to tell wet from very wet snow, and water content can change dramatically in short time periods. In order to use water content as a predictive tool, technology must be developed that can determine the water content of thin (~1cm) layers in a field setting. The development of quantified tests dependent on water content, but possibly easier to assess, may simplify the process and provide tools that can be used to develop solid wet snow avalanche forecasting strategies.

CHAPTER 4

RELATING AIR TEMPERATURE TO WET LOOSE SNOW AVALANCHE ACTIVITY

The literature has highlighted simple relationships between air temperature and wet avalanche activity. For example, Armstrong (1976) noted high correlation between mean daily temperatures above $0^{o}C$ and wet avalanche activity at Red Mountain Pass, Colorado; and Romig (2004; Romig et al., 2004) found the minimum daily air temperature to be a good predictor value at Bridger Bowl, Montana. Operational experience (Carse, 2003) and research (Romig, 2004; Romig et al., 2004) have shown that temperature alone is not an adequate indicator of wet loose avalanche activity. I posed the following research question to test how the relationships described by Armstrong and Romig related to observed wet loose avalanche activity in the Bridger Range during 2003 and 2004, and to determine if those relationships are useful in a practical setting:

1. How well do the mean and minimum daily air temperatures indicate observed wet loose avalanche activity?

Air temperature as a stand alone forecasting tool is expected to be inadequate, and decisions in operational settings should not be based solely upon air temperature.

Methods

Field data and observations were collected during the spring months (March and April) of 2003 and 2004. Historically, wet avalanches have occurred in Bridger Bowl Ski Area from late February to April (Johnson, 2002; Romig, 2004; Romig et al., 2004).

Seasonal maximum, minimum, average, and three-hour air temperature data were obtained from the Brackett Creek SNOTEL site throughout the study period (Fig. 3.1). The mean daily air temperature and minimum daily air temperature were compared to avalanche activity (or lack thereof) and precipitation. The presence of wet loose avalanche activity during the study period was documented by onsite monitoring of the eastern Bridger Range-front from the North side of Saddle Peak to Wolverine Basin (Fig. 3.1). Avalanche days were determined either through observed avalanche events, or through the presence of wet loose avalanche debris that was deposited during observation days. Non-avalanche days were determined by the lack of observed avalanches during field days, or by the lack of debris from up to two days prior to the field day.

Results

Two wet loose snow avalanche cycles were witnessed in both April, 2003 and April, 2004. A total of 42 wet loose snow avalanches ranging in estimated sizes from D1 to D3 were documented (Fig.3.2). Avalanche occurrence was graphed in relation to the mean and minimum daily air temperature during the study.

Mean Daily Air Temperature in Relation to Wet Loose Snow Avalanching

Armstrong (1976) noticed a high correlation between mean daily air temperature and wet avalanche activity on Red Mountain Pass in Colorado. The following graphics were produced to test the applicability of this relationship to wet loose avalanche activity in Montana's Bridger Range. Figures 4.3, 4.4, 4.5 and 4.6 show the mean daily air temperature, precipitation, and avalanche activity on field days in April of 2003, March of 2004, and April of 2004 respectively. Solid lines represent days classified as avalanche days. Non-avalanche days are represented by dotted lines (days when no avalanches were observed) and by dashed lines (days when the lack of avalanche activity was established by the lack of debris). Observers were not present on all days during the month. A total of 10 days were observed in April of 2003, 15 days in March of 2004, and 18 days in April of 2004.

Figure 4.1 depicts two new snow events, one rain event, six field days above 0^{o}C (mean daily temperature) with wet loose avalanches, two field days at or above 0^{o}C without wet loose avalanches, one field day below 0^{o}C with wet loose avalanches, and one field day below 0^{o}C without wet loose avalanches. Six of seven observed avalanche days had mean daily temperatures of 6-10^{o}C. The remaining day had a mean temperature of -1^{o}C.

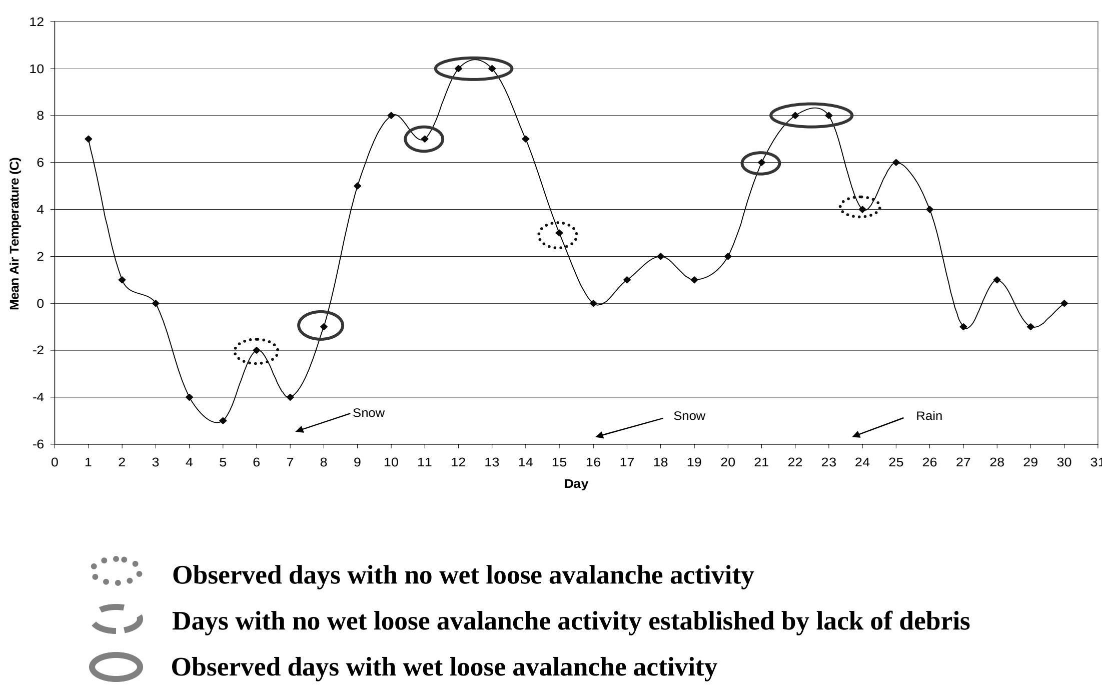


Figure 4.1. April 2003 average air temperature in relation to observed avalanche days, non – avalanche days, and precipitation events.

Figure 4.2 depicts one new snow event, one field day with a mean air temperature above $0^{o}C$ with wet loose avalanche events, 10 field days above $0^{o}C$ without wet loose avalanche events, one field day below $0^{o}C$ with wet loose avalanche events, and two field days below $0^{o}C$ without wet loose avalanche events. The mean air temperature varied dramatically between the two observed avalanche days this month, with one day having a mean air temperature of $-1^{o}C$ and the other having a mean air temperature of $10^{o}C$.

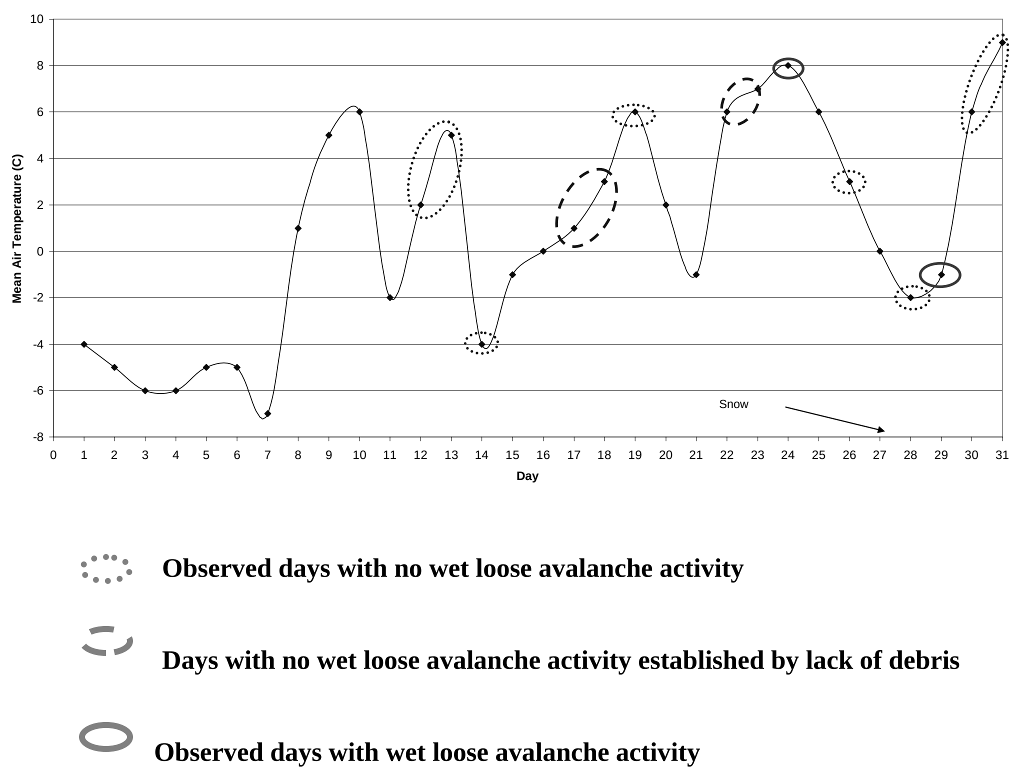


Figure 4.2. March 2004 average air temperature in relation to observed avalanche days, non – avalanche days, and precipitation events.

Figure 4.3 depicts two new snow events, eight field days at or above $0^{o}C$ with wet loose avalanches, ten field days at or above $0^{o}C$ without wet loose avalanches, and one field day below $0^{o}C$ with wet loose avalanche events. Four observed avalanche days occurred at mean temperatures between –1 and 1°C, and five observed avalanche days occurred between 3 and 10°C.

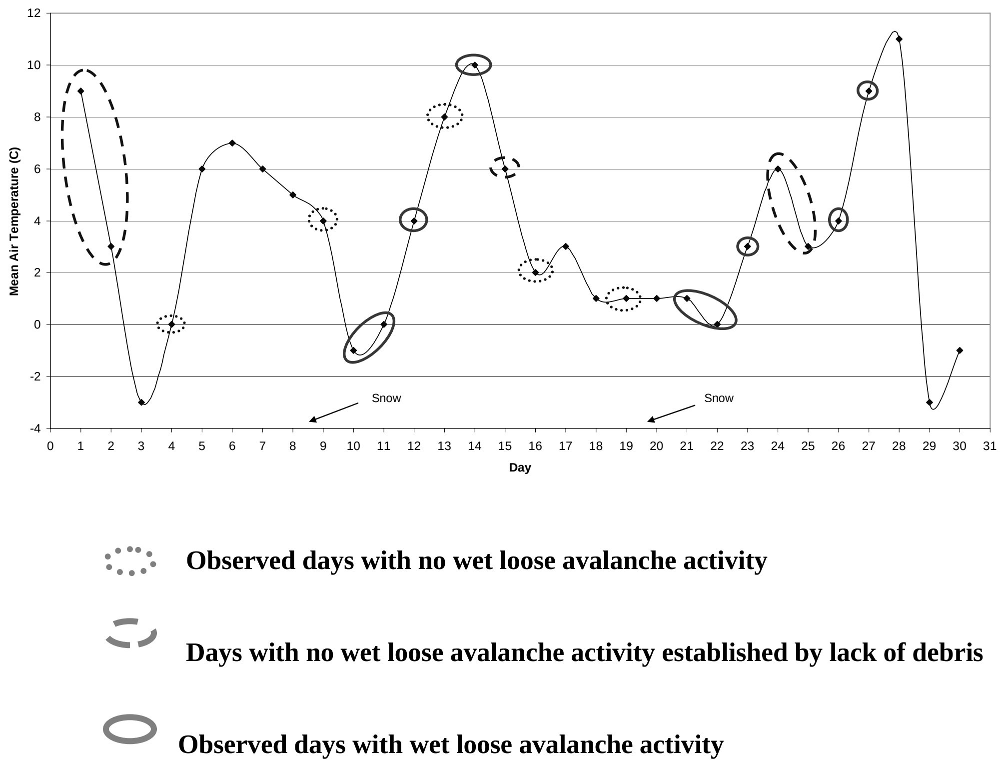


Figure 4.3. April 2004 average air temperature in relation to observed avalanche days, non – avalanche days, and precipitation events.

Figure 4.4 depicts the mean daily air temperature on observed avalanche days and non-avalanche days. Avalanche days have a median value of 5°C, while non-avalanche days have a median value of 3°C.

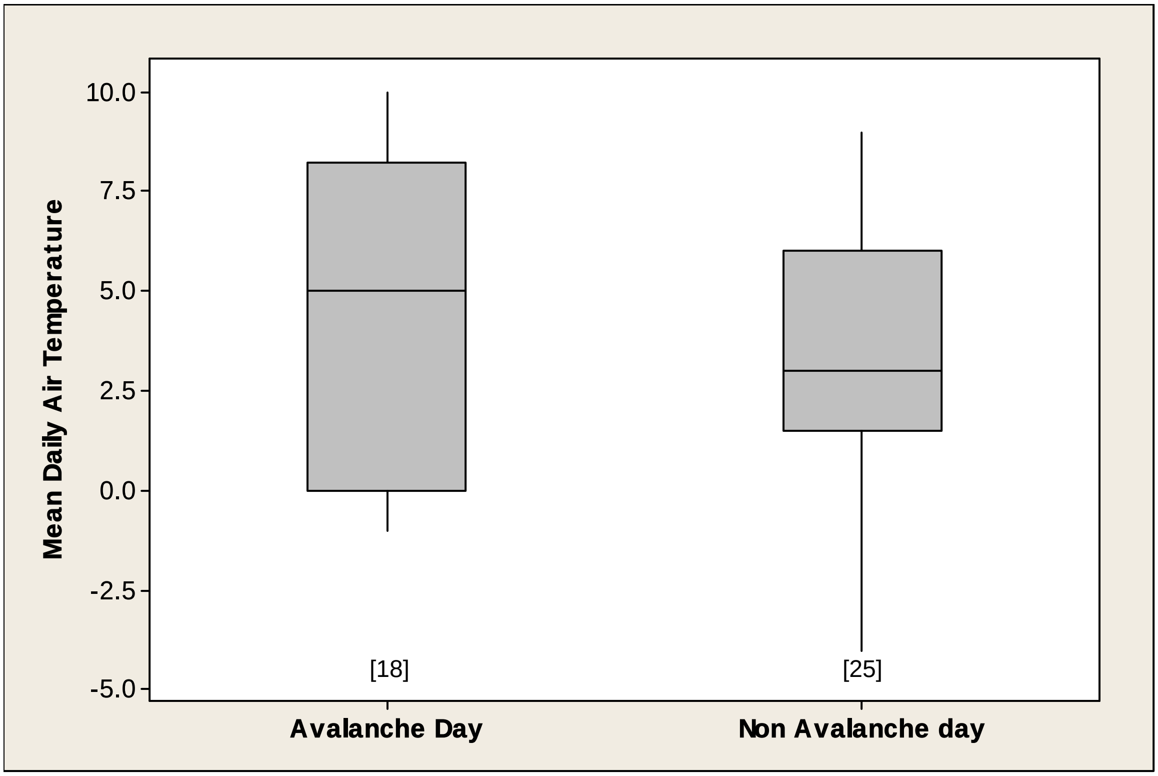


Figure 4.4 Mean daily temperature (°C) in relation to wet loose avalanche days and non-avalanche days in April 2003, March 2004, and April of 2004. The line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. Bracketed numbers below each group are the number of individual shear frames in the sample.

Minimum Daily Air Temperature in Relation to Wet Loose Snow Avalanching

Romig (2004; Romig et al., 2004) found the minimum daily air temperature to be a significant predictive variable of wet avalanche activity. In addition, practitioners have found that a 'soft freeze' (surficial snow incompletely frozen) is a good indicator of instability the following day (Carse, 2003). Figures 4.5, 4.6, 4.7, and 4.8 depict the minimum daily air temperature, precipitation, and the presence (or lack of) avalanche activity as assessed by fieldwork in April of 2003, March of 2004, and April of 2004 respectively. Solid lines represent days classified as avalanche days. Non-avalanche days are represented by dotted lines (days when no avalanches were observed) and by dashed lines (days when the lack of avalanche activity was established by the lack of debris). There were a total of 10 field observation days in April of 2003, 15 days in March of 2004, and 18 days in April of 2004.

Figure 4.5 depicts two days with snow, and one with rain. There were five field days above a $0^{o}C$ minimum air temperature with wet loose avalanche events, two field days above $0^{o}C$ without wet loose avalanche events, two field days below a minimum $0^{o}C$ with avalanche events, and six field days below a minimum air temperature of $0^{o}C$ without avalanche events. Six of seven observed avalanche days occurred with minimum daily air temperatures between –2 and $5^{o}C$.

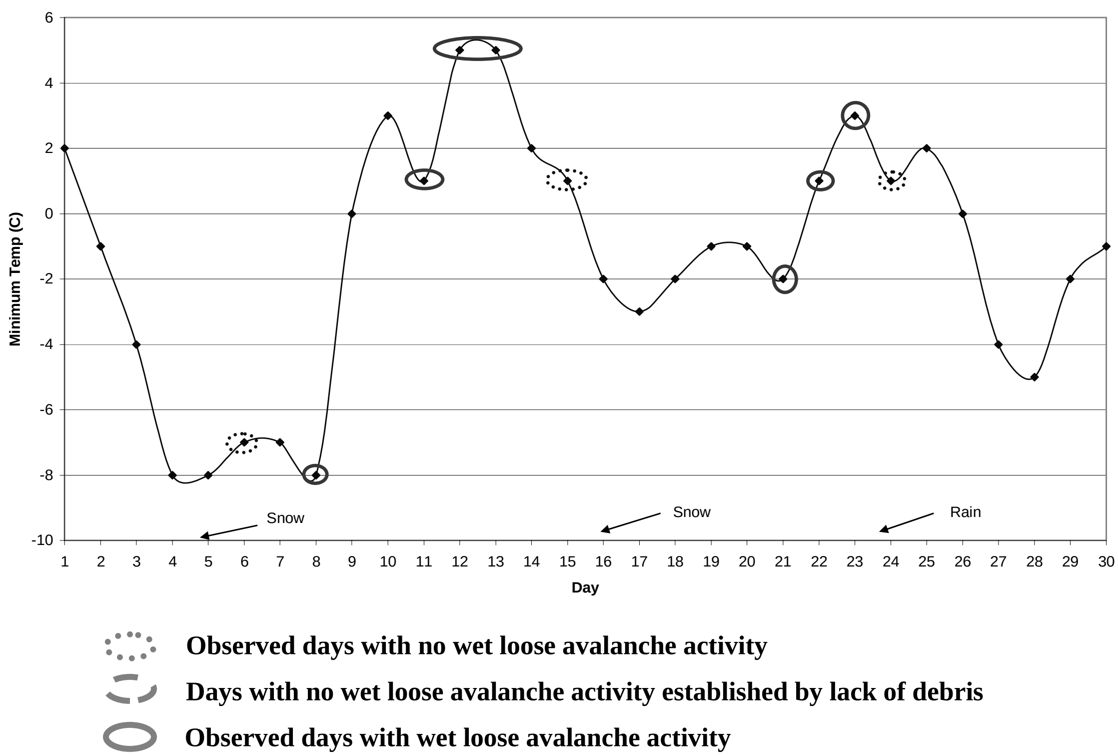
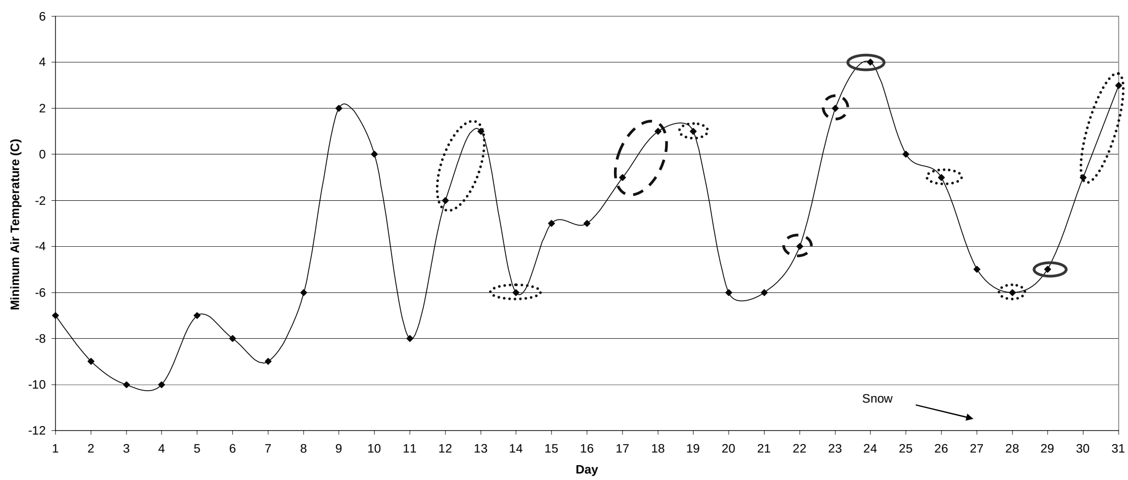


Figure 4.5. April 2003 minimum air temperature in relation to observed avalanche days, non – avalanche days, and precipitation events.

Figure 4.6 depicts one observed new snow event, one field day above 0°C minimum air temperature with wet loose avalanche events, five field days above 0°C minimum without wet loose avalanche events, one field day below 0°C minimum air temperature with wet loose avalanche events, and seven days below 0°C minimum air temperature without avalanche events in March 2004. Of the two observed wet loose avalanche days this month, one day had a minimum air temperature of –5°C and the other a minimum air temperature of 4°C.



Observed days with no wet loose avalanche activity

Days with no wet loose avalanche activity established by lack of debris

Observed days with wet loose avalanche activity

Figure 4.6. March 2004 minimum air temperature in relation to observed avalanche days, non – avalanche days, and precipitation events.

Figure 4.7 depicts two observed new snow events, two field days above a 0°C minimum air temperature with avalanche events, one field day above 0°C minimum without avalanche events, two field days at 0°C minimum air temperature without avalanche events, seven field days below 0°C minimum air temperature with avalanche events, and six field days below 0°C minimum air temperature without avalanche events in April of 2004. Seven of nine avalanche days occurred on days with a minimum air temperature between –1 and –5°C. Two of nine avalanche days had a minimum temperature between 3 and 6°C.

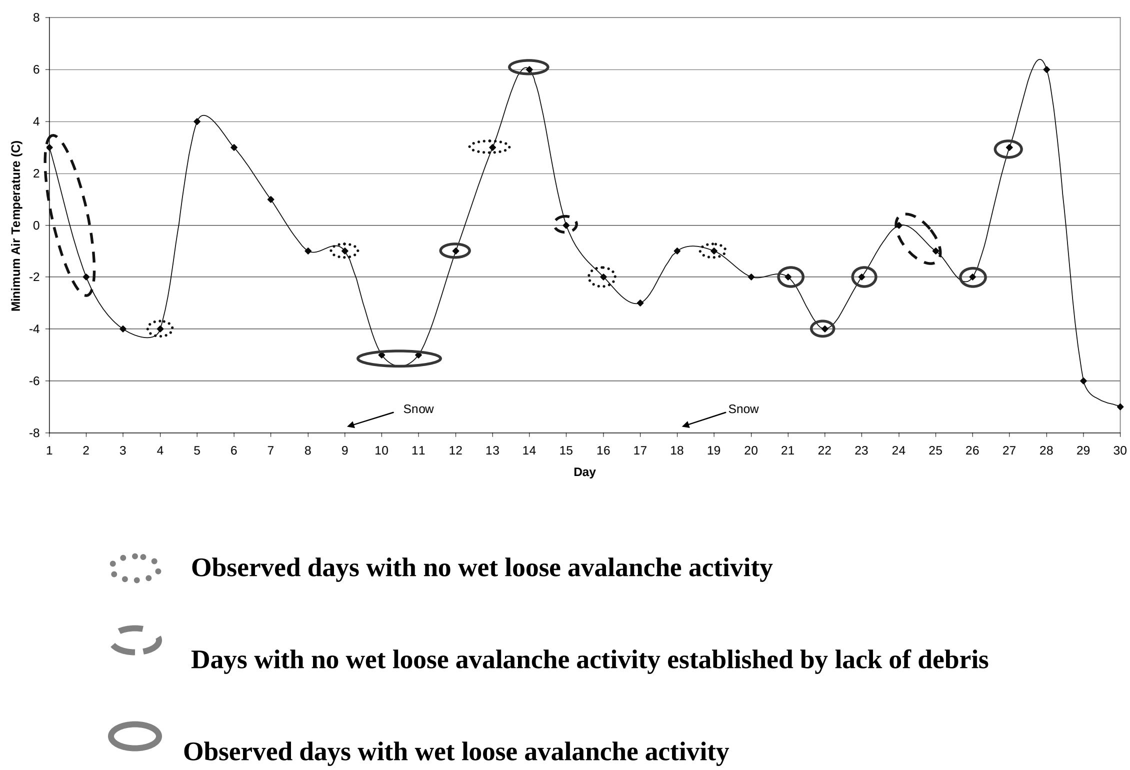


Figure 4.7. April 2004 minimum air temperature in relation to observed avalanche days, non – avalanche days, and precipitation events.

Figure 4.8 shows the relationship between all observed wet loose avalanche days and the recorded minimum daily temperature. Avalanche days have a median value of -1.5°C, and non-avalanche days have a median value of -1°C.

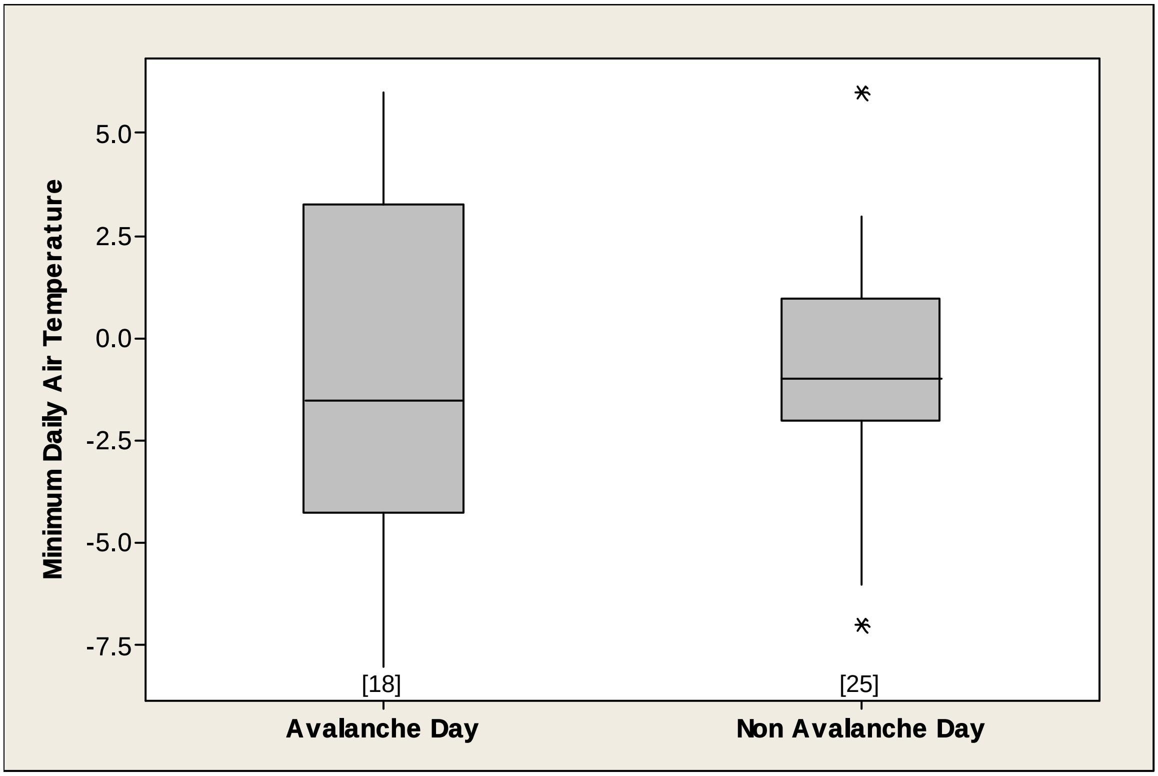


Figure 4.8. Minimum daily temperature in relation to wet loose avalanche days and non-avalanche days in April 2003, March 2004, and April of 2004. The line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. Bracketed numbers below each group are the number of individual shear frames in the sample.

Discussion

The observation set is biased because the numbers used in this study are based on field observations that were made as field time and logistics allowed. Avalanches might have occurred in the Bridger Range when no observers were present, or that were not encountered by an observer. However, the observations presented do generally reflect the conditions present during various melt cycles in the Bridger Range during April of 2003 and March and April of 2004.

Mean Daily Air Temperature

Armstrong (1976) suggested that wet avalanche cycles at Red Mountain Pass coincided with mean daily air temperatures above $0^{o}C$. In the Bridger Range, 83% of observed avalanche days occurred when the mean daily air temperature was above zero centigrade (17% occurred at a mean temperature below $0^{o}C$). This percentage does not adequately reflect the fact that not all days above zero centigrade were avalanche days. Figure 4.4 compares the mean daily air temperature distribution of observed avalanche days and non-avalanche days. Avalanche days have a higher median value than non-avalanche days, but there is obvious overlap between the spread of each data set. Although the range of each plot is similar, avalanche days are skewed toward warmer temperatures while non-avalanche days are skewed towards colder temperatures. Clearly, factors in addition to mean daily temperature are required to produce wet snow avalanches.

Minimum Daily Air Temperature

Romig (2004; Romig et al., 2004) identified the minimum temperature as a predictive variable and nightly temperatures above freezing are often used operationally as an indicator for possible wet avalanche activity the following day. The question becomes, “How cold does it have to get in order to sufficiently strengthen the snowpack?’ Based on the observations presented, use of the freezing threshold is not sufficient; 56% (10 of 18) of the recorded avalanche days occurred after minimum nightly temperatures of $0^{o}C$. Figure 4.8 compares the minimum daily air temperature distribution of avalanche days and non-avalanche days. The range of the data presented for non-avalanche days falls completely within that presented for avalanche days. It is clear that in the data presented, the use of the minimum daily air temperature as a predictive tool for wet loose avalanches is not effective.

Energy Balance at the Snow Surface

Once snow temperature is raised to $0^{o}C$, further energy input results in the conversion of ice to liquid water. To fully understand melt, the energy balance at the snow surface must be considered:

$$\text{Energy Balance} = R + H + LE + G + F$$

Where R is the net radiation input (short-wave and long-wave), H is the sensible heat exchange (conduction in response to temperature gradients, can be increased by wind), LE is the latent heat flux (energy transfer through condensation or sublimation), G is the ground heat flux (energy transfer from temperature gradients at the snow / soil interface), and F is the advective heat flux (mass transfer of energy into snow, i.e. rain on snow) (Williams, 1998). Although all of these factors need to be considered, radiation is the dominant driver. Air temperature is, and has been, used by practitioners because it is readily accessible. Unfortunately, it only partially represents both the sensible heat exchange and the radiation input. A more complete picture could be provided with radiation instrumentation, but these instruments are relatively expensive and, thus far, are not widely used by avalanche forecasting operations. However, in order to attempt to understand how meteorological variables affect wet loose avalanching, the radiation balance must be taken into account.

Conclusions

Experience and prior research have shown that monitoring air temperature is only partially useful when forecasting wet loose snow avalanches. The data presented in this paper show that mean and minimum daily air temperature values show a common range on avalanche days and non-avalanche days. This is not surprising since the snow surface temperature is more strongly affected by the radiation balance than by the air temperature. Daily radiation history is also important. Thus, in order to make a useful association between meteorological variables and avalanche days vs. non-avalanche days, data are needed on the net radiation balance and sensible heat exchange over time.

CHAPTER 5

RELATING SURFICIAL SHEAR STRENGTH TO WET LOOSE SNOW AVALANCHE ACTIVITY

The literature reports several studies which focus on the shear strength of wet snow (Perla et al., 1982; Brun and Rey, 1987; Bhutiyani, 1994). Although strong correlations exist between dry snow density and shear strength, only weak correlation exists between wet snow shear strength and measures of density (Perla et al., 1982; Brun and Rey, 1987). No one has explored shear strength during melting cycles, and research is not available that directly relates the shear strength of wet snow to avalanche activity. Shear strength is expected to change in response to water regime (Bhutiyani, 1994) and hypothetically will change during the day as melting occurs and near surface water contents increase. This research addresses the following research questions:

1. Is a 250cm2 shear frame effective at quantifying changes in wet snow shear strength ?

2. Based upon the methods applied, how fast do significant changes in the shear strength of wet snow occur?

3. Are targeted measurements of surficial shear strength a valid measure of slope scale strength, or do slope scale spatial patterns outweigh temporal changes?

4. Are changes in surficial shear strength correlated with wet loose snow avalanche activity? Is there a threshold strength at which wet loose avalanches can occur?

I hypothesize that there is a threshold strength at which wet surficial snow can avalanche on slopes of appropriate steepness. The questions listed above will give insight into whether or not shear frames can be used as a viable tool for operational forecasting. A portion of the following research has been published (Trautman et al., 2006).

Methods

Data Collection

Changes in shear strength spanning targeted melt-freeze cycles were documented during four separate April days in 2005 and 2006. Days were chosen based upon the presence of a well developed surficial melt-freeze crust, forecasted sunny weather and above freezing temperatures. Each study site was in a position where neighboring slopes could be monitored for avalanche activity. Testing began when frozen surficial snow had softened enough to allow shear frame placement. A 250 cm^2 shear frame was used in conjunction with a 5 kg Wagner force dial. In most cases, the frame could be inserted when the shear strength was greater than the maximum strength of the gauge. Under these circumstances, the pull was given the maximum value (5kg) of the gauge. In order to be as consistent as possible, a specific weak layer was not targeted; instead changes in strength were measured ~4cm below the surface (depth of a 250cm^2 frame). Shear frames were inserted to the depth of the frame, and the adjacent snow (outside the frame) was removed with a putty knife (Fig. 5.1). Pull time was 1 – 1.5 seconds. Shear force was converted to shear strength using the following equation:

$$\tau_e = f/a$$

Where τ_e is the effective slope parallel shear strength, f is the shear force in Newtons (N), a is the area of the shear frame in (m^2), and τ is the shear strength in Pascals (Pa). All shear frame results were adjusted for shear frame size using the following equation (Fohn, 1987; Greene et al., 2004):

$$\tau_{\infty} = 0.65\tau_{250}$$

Where 0.65 is the correction factor identified by Fohn, 1987, τ_{250} is the shear strength identified using a 250 cm^2 frame and τ_{∞} is the Daniels strength, or shear strength adjusted for shear frame size effects.

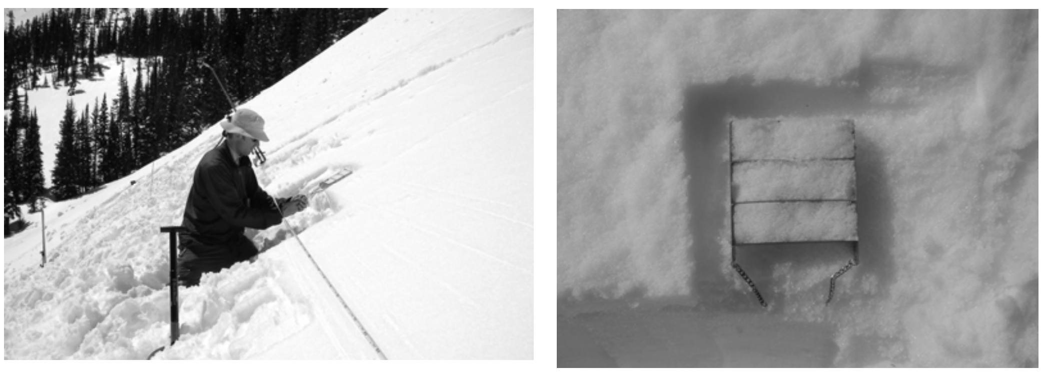


Figure 5.1. 20 April 2006 study site. The photographs depict the methodology used to document changes in the surficial shear strength of wet snow.

<u>Experimental Design Used to Answer Questions 1 and 4</u>

Field Day 1 and 2, 24 and 25 April 2005

Data was collected along the north boundary of the Bridger Bowl Ski Area, 24 km (15 miles) north of Bozeman, Montana (Fig. 3.1). The study site is east facing with an elevation of 2438 m (8000 feet) and a slope of 32 degrees. Tests were conducted in hourly transects consisting of 12 shear frame pulls. Each transect was completed in approximately fifteen minutes.

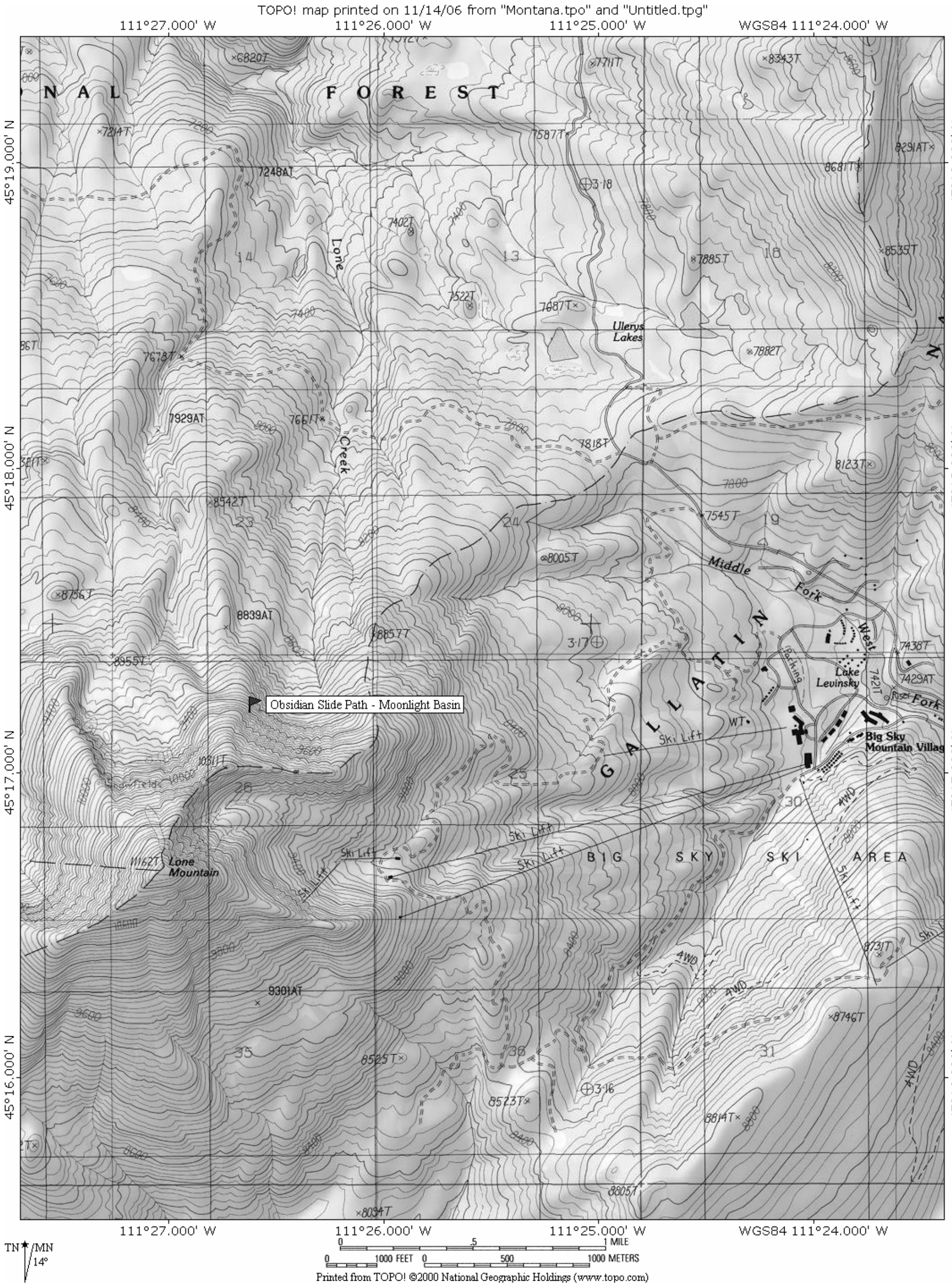


Figure 5.2. Location of the study site on Field Day 4 – 22 April 2006.

Field Day 4 - 22 April 2006

Data was collected in the Obsidian slide path at Moonlight Basin Ski Area. Moonlight Basin is 56 km (35 miles) south of Bozeman, Montana (Fig. 5.2). The study site is east facing, has an elevation of ~2743m (9000 feet), and a slope of 40 degrees. Shear strength measurements were initiated at 09:45 and then conducted hourly until 17:45. Tests were conducted in transects consisting of 10 shear frame pulls and were completed in approximately fifteen minutes (Fig. 5.3).

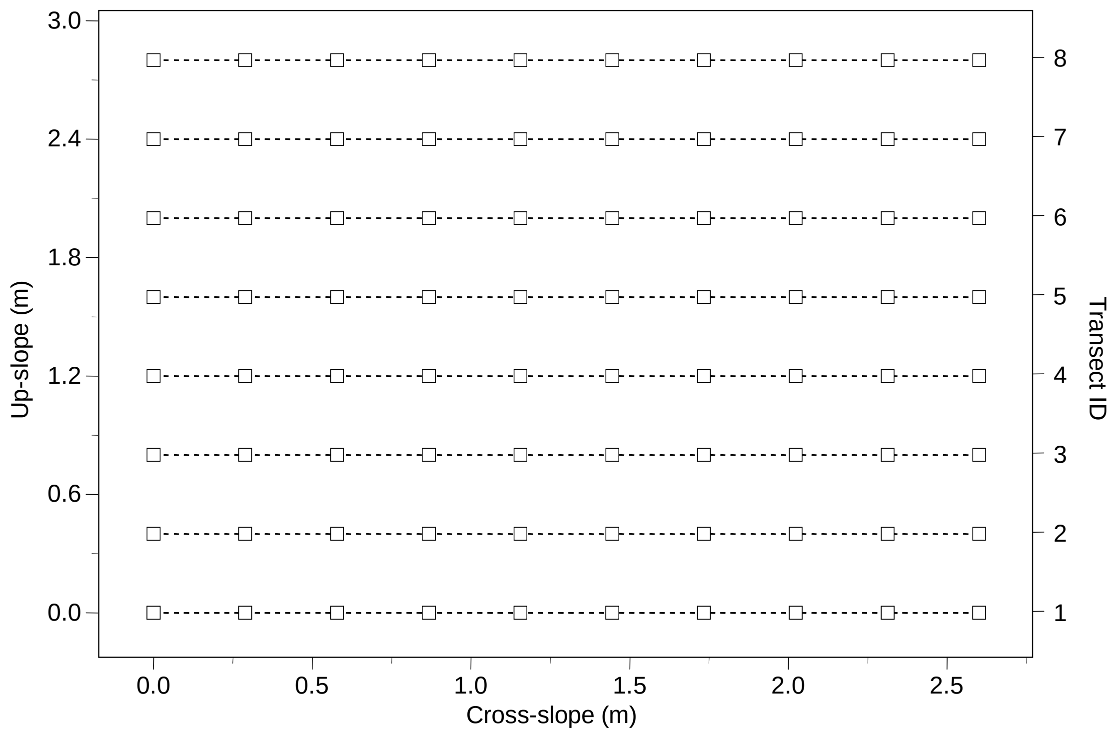


Figure 5.3. Conceptual sampling pattern used on Field Days 1, 2, and 4. Individual shear frames were placed approximately 30 cm apart. Transects were conducted hourly over the course of each melt-freeze cycle. The number of transects, and the number of tests per transect varied by day.

Experimental Design Used to Answer Questions 2 and 3

Field Day 3 - 20 April 2006

Data was collected north of Bridger Bowl Ski Area in the vicinity of Bradley meadows. Bradley Meadows is 24 km (15 miles) north east of Bozeman, Montana (Fig. 3.1). The study site is southeast facing, has an elevation of 2316 m (7600 feet), and a variable slope between 27 and 37 degrees. The sampling pattern was chosen in order to test both temporal and spatial controls on strength. The data collected was taken from five full 30 meter transects that were further divided into 3 sub-transects per transect. Each sub-transect consisted of 5 groups of 3 shear frames spaced 3 meters from the previous group. This pattern allowed almost continuous strength testing throughout the cycle (Fig. 5.4).

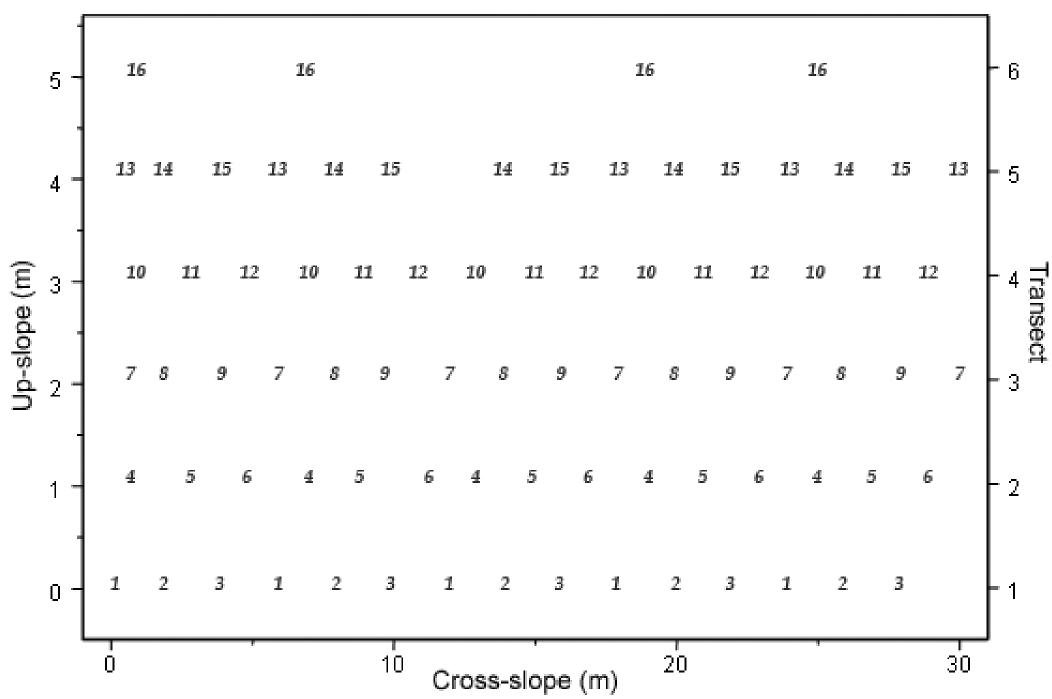


Figure 5.4. Sampling pattern used on Field Day 3, 20 April 2006. Five full transects composed of 45 individual shear frames were conducted. Each transect consisted of five sub-transects. The designation of 1, 2, 3, 4…16 are consecutive sub-transect sampling sites. Each sampling site was composed of 3 individual shear frames; in effect, each sub-transect is composed of 15 individual shear frames.

Data Analysis

Measurements were tested for significant temporal change in shear strength by transect or sub-transect using the non-parametric Mann-Whitney (Wilcoxon Rank-Sum) test. Statistics examined significant change in strength between consecutive groups over time. Box-plots were used to graphically identify outliers and differences in central tendency. Time between individual transects and sub-transects was noted. The presence or absence of avalanche activity, and the associated timing during each field day was compared to corresponding surface shear strength measurements.

Question 3 required a detailed representation of slope scale steepness. Nineteen slope angle measurements were taken at random locations across the slope. An ordinary Kriging procedure was used to interpolate slope angle in areas of the slope that were un-sampled. Kriging is a geostatistical method that provides estimates for un-sampled locations by computing weighted averages of sample values from nearby locations. The weights are determined on the basis of the semi-variogram, a statistical model of the relationship between spatial autocorrelation and the distance between pairs of sampled values.

Surficial Shear Frame Results

Shear Strength – Field Day 1 - 24 April 2005

April 24, 2005 was a warm, partly cloudy day with a minimum air temperature of -3^{o}C and a maximum air temperature of 14^{o}C. Testing began at 08:00 and continued on the hour until 18:00 (Fig. 5.5). Eleven transects, or 132 individual shear frame tests were conducted. Between 12:00 and 13:00 there was a brisk, cooling wind. The sun went down at ~16:20. No avalanche activity was noted.

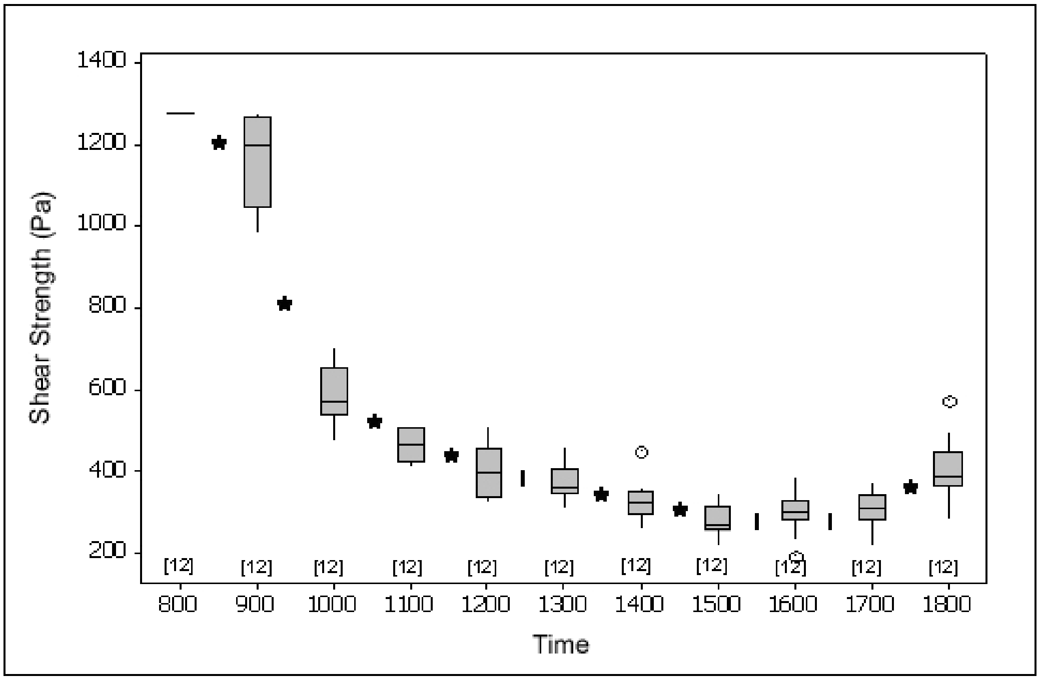


Figure 5.5. Hourly shear strength on 24 April 2005 at the North Boundary Study Site ate Bridger Bowl. Statistically significant changes between consecutive sets of measurements (Mann Whitney $p < 0.05$) are denoted by an "*", changes that are not significant are denoted by a "|". The horizontal line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. White circles denote outliers. Bracketed numbers below each group are the number of individual shear frames in the sample.

Shear Strength – Field Day 2 - 25 April 2005

April 25, 2005 was a warm, partly cloudy day with a minimum air temperature of -1°C and a maximum air temperature of 13°C. Testing began at 08:00 and continued on the hour until 17:00 (Fig. 5.6). Ten transects, or 120 individual shear frame tests, were conducted. Between 13:00 and 14:00 there were intermittent clouds and cool winds. At

15:30 a razor crust was noted. The sun went down at 16:17. No avalanche activity was noted.

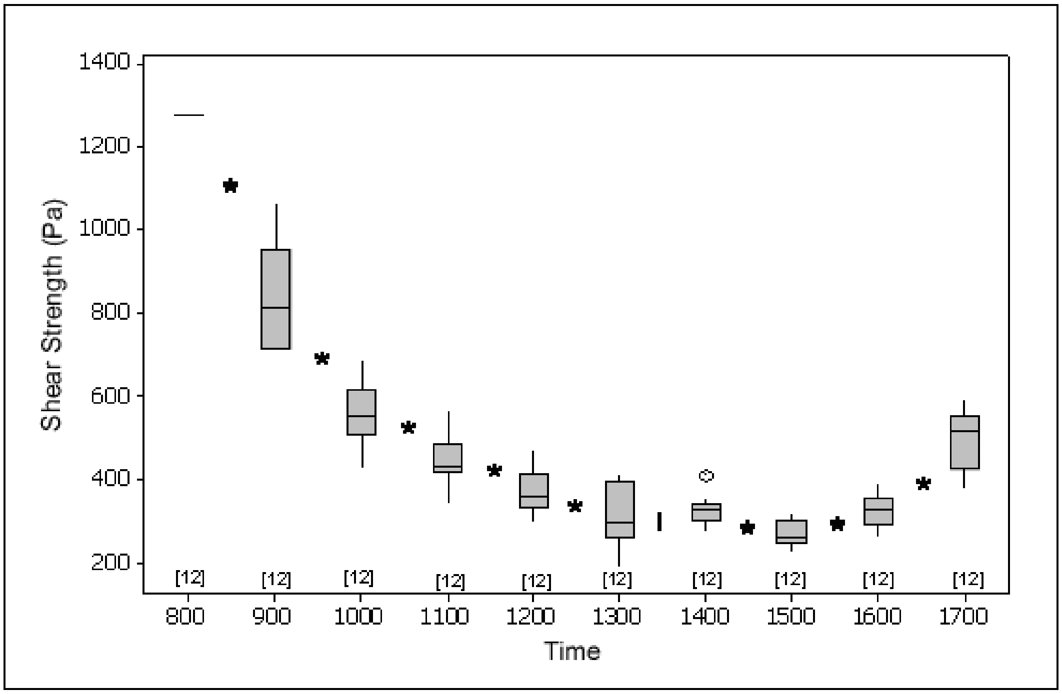


Figure 5.6. Hourly shear strength on 25 April 2005 at the North Boundary Study Site at Bridger Bowl. Statistically significant changes between consecutive sets of measurements (Mann Whitney $p < 0.05$) are denoted by an "*", changes that are not significant are denoted by a "|". The horizontal line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. White circles denote outliers. Bracketed numbers below each group are the number of individual shear frames in the sample.

Shear Strength – Field Day 3 - 20 April 2006

April 20, 2006 was a warm, mostly sunny day with a minimum air temperature of -1^{o}C and a maximum air temperature of 14.1^{o}C. Testing began at 09:45 and continued throughout the day until 18:11 (Fig. 5.7). A total of 210 individual shear frame tests were conducted. At 16:42 a razor crust was noted. The sun went down at 17:10. 'Roller ball' activity was common and two minor sluffs were noted in the vicinity of the study site at 13:40.

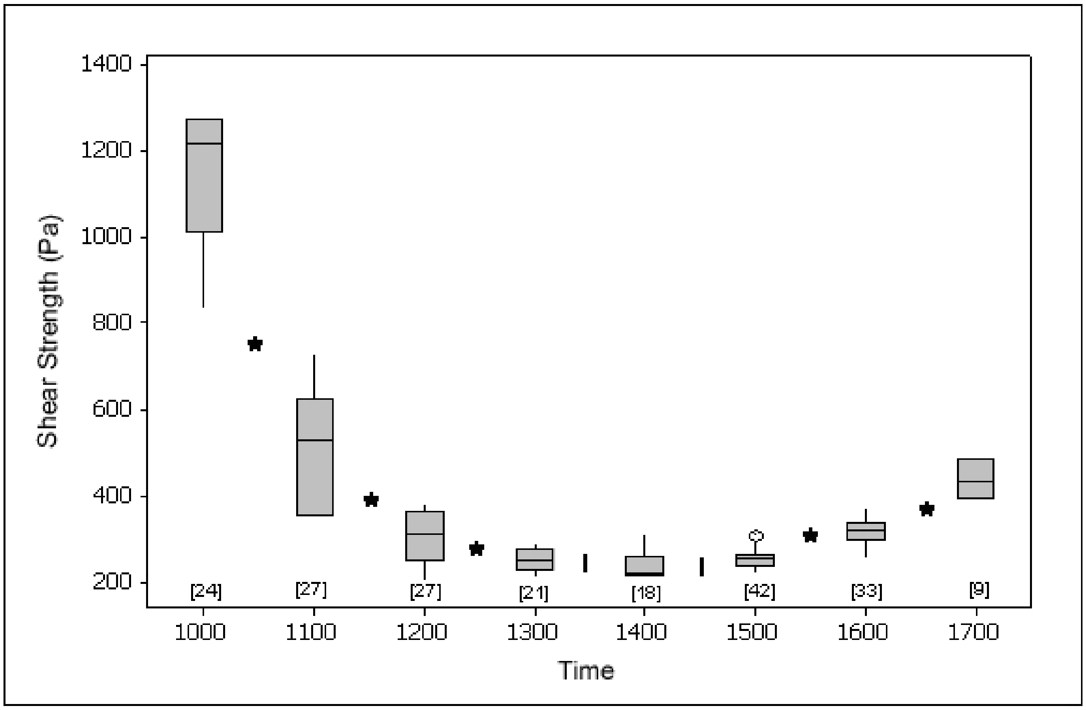


Figure 5.7. Hourly shear strength on 20 April 2006 at the South Bradley study Site at Bridger Bowl. Statistically significant changes between consecutive sets of measurements (Mann Whitney $p < 0.05$) are denoted by an "*", changes that are not significant are denoted by a "|". The horizontal line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. White circles denote outliers. Bracketed numbers below each group are the number of individual shear frames in the sample.

In order to address question 2, shear strength measurements were plotted at the sub-transect level (Fig. 5.8). The minimum time needed to complete a sub-transect was 20 minutes, the maximum time was 54 minutes, and the average time was 31 minutes. The greatest change in strength noted (~ 50%) occurred in 20 minutes between sub-transect 2 and 3.

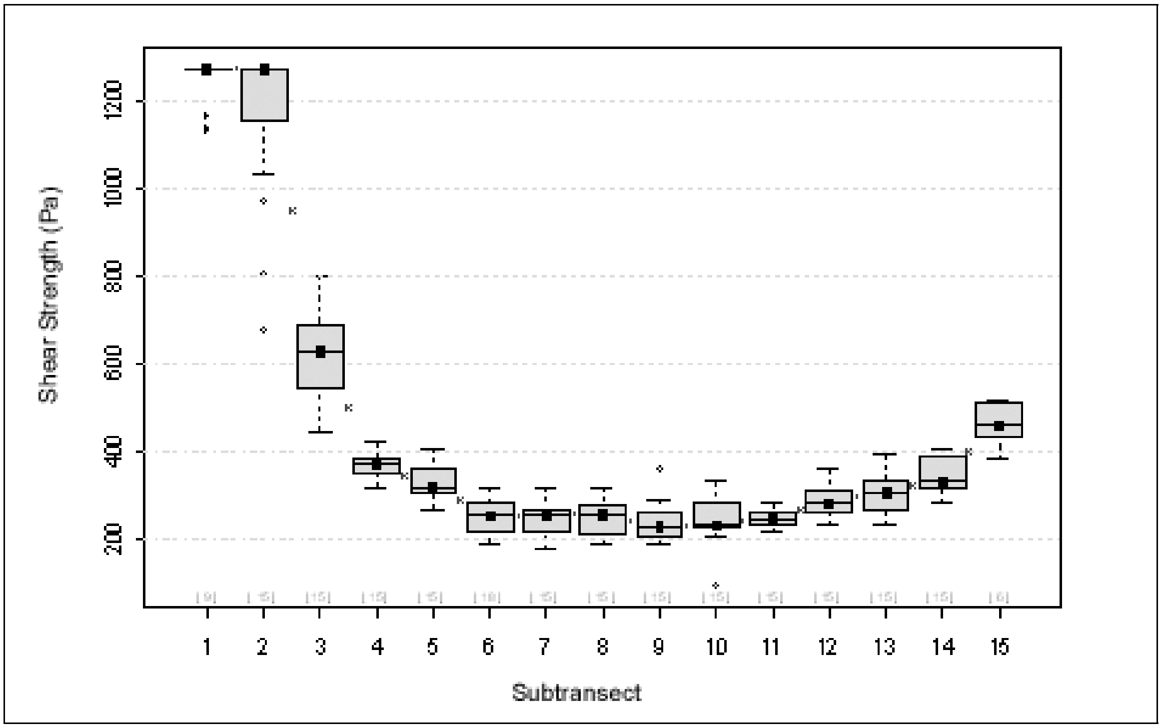


Figure 5.8. Shear frame results at the sub-transect level on 20 April 2006. Statistically significant changes between consecutive sets of measurements (Mann Whitney $p < 0.05$) are denoted by an "*", changes that are not significant are denoted by a "|". The horizontal line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. White circles denote outliers. Bracketed numbers below each group are the number of individual shear frames in the sample.

Shear Strength – Field Day 4 - 22 April 2006

April 22, 2006 was a warm, mostly sunny day with high cirrus clouds and a light breeze from the south east. The minimum air temperature was -0.4°C and the maximum air temperature was 15.2°C. Shear strength measurements were initiated at 09:45 and then conducted hourly until 17:45 (Fig. 5.9). At 15:54 there was a down-slope, cooling wind in conjunction with the topographic sunset. At 16:04 a razor crust was noted. Avalanche activity was noted between 11:40 and 15:45.

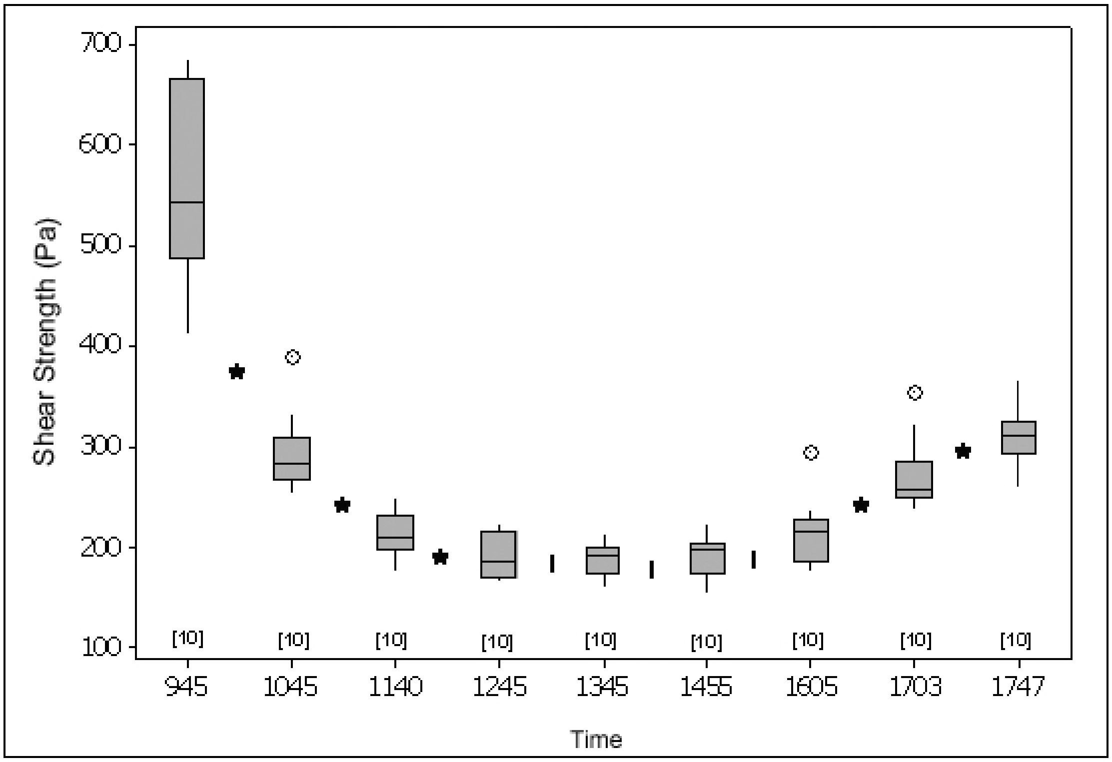


Figure 5.9. Hourly shear strength on 22 April 2006 at the Obsidian Study Site at Moonlight Basin. Statistically significant changes between sets of measurements (Mann Whitney $p < 0.05$) are denoted by an "*", changes that are not significant are denoted by a "|". The horizontal line represents the median, the box encompasses the 25th to 75th percentile of measurements and the whiskers are 1.5 times the interquartile range. White circles denote outliers. Bracketed numbers below each group are the number of individual shear frames in the sample.

Shear Strength and Avalanche Activity on Field Day 4 – 22 April 2006

On the morning of 22 April 2006 there was evidence of previous 'roller-ball' activity, but no avalanche debris was noted. Mean study slope shear strength dropped from 316 Pa at 10:45 to 229 Pa at 11:45. 'Roller-ball' activity was widespread at 10:45 and several avalanches (WL-N-D1/D2) were noted in and around the study site between 11:40 and 15:45. A ski cut within the study slope produced a small avalanche at 13:25 (WL-AS-D1). There was no significant change in strength until early evening when the mean strength increased from 232 Pa at 1605 to 293 Pa at 17:03. A ski cut on the study slope at 17:03 produced no results (Fig. 5.10).

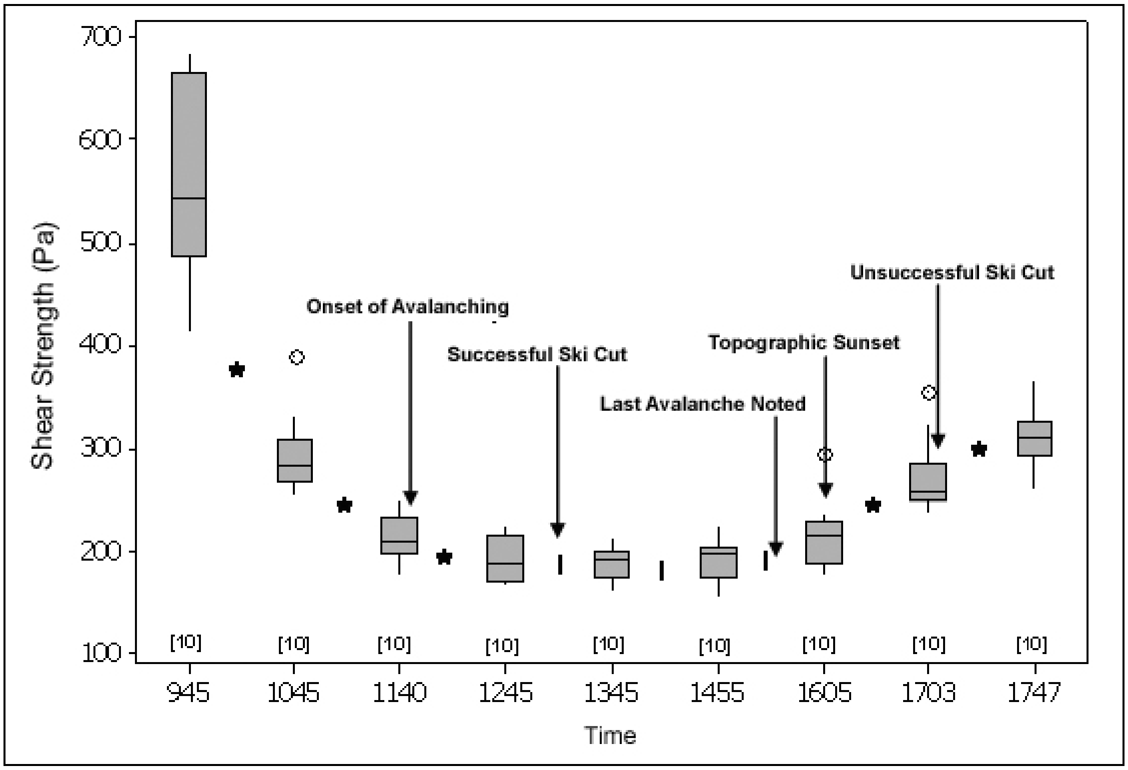


Figure 5.10 Avalanche activity and shear strength on 22 April 2006. Statistically significant changes between sets of measurements (Mann Whitney $p < 0.05$) are denoted by an "*", changes that are not significant are denoted by a "|". The horizontal line represents the median, the box encompasses the 25^{th} to 75^{th} percentile of measurements and the whiskers are 1.5 times the interquartile range. White circles denote outliers.

Discussion

Quantifying the Shear Strength of Wet Snow Using Shear Frames

Shear frames proved to be an effective way to document changes in the surficial shear strength of melt-freeze snow. The rate of weakening and strengthening varied by day, though similarities clearly exist (Fig. 5.5, 5.6, 5.7, 5.9). In particular, the data shows that the surficial layers of the snowpack can significantly weaken quickly (20 minutes). The methods used did not allow a full documentation of the strengthening trend of the cycle. In the morning frames could be inserted without compression of the underlying stratigraphy, but in the evening pressure on surficial ice layers compressed the weaker underlying snow. However, the initial strengthening trend could be tracked on all four field days.

Documenting the Time Between Significant Changes in Shear Strength

Figures 5.5, 5.6, 5.7, and 5.9 show that snow during melt freeze cycles commonly undergoes significant changes in shear strength over the course of one hour. In order to improve forecasting and mitigation techniques, finer temporal resolution is needed. At the South Bradley Study site near Bridger Bowl 20 April 2006, shear strength data was collected almost continuously throughout the melt-freeze cycle. A 50% loss in strength was noted in a 20 minute period, and a significant gain in strength was noted after a 25 minute interval (Fig. 5.8). This finding matches the experience of avalanche workers, who commonly observe extremely rapid changes from stable snow to unstable snow when dealing with wet snow avalanches; sometimes this change can occur between two short chairlift rides (Carse, 2003).

Targeted Measurements as a Valid Measure of Slope Scale Strength

Spatial trends in melt are noticeable in field settings and experience has shown that in many cases aspect can control which slopes receive sufficient energy to loose strength. Although rocks and other areas of concentrated radiant heat can result in small regions of instability, snow from these areas often releases and slides onto slopes below without producing an avalanche. In most operational settings, hazardous conditions occur when larger and more open slopes loose enough strength to avalanche when triggered from above. The key question is whether or not targeted measurements of shear strength on open slopes can be used to estimate the strength of specific slopes given small (slope scale) variations in steepness. If slope scale spatial patterns control slope-wide strength, data is expected to show a different pattern in the changes in strength between data collected in a small area and that collected from a larger area. Data collected on 20 April 2006 at the South Bradley study site (Fig. 3.1 and Fig. 5.4) were used to address this problem. Measurements on this day consisted of almost continuous strength tests back and forth across a somewhat variable slope throughout the day. The slope was 30m long by 5m wide (a total area of 150m^2 compared with only 9m^2 on days 1, 2, and 4) with a variable steepness between 26 and 37 degrees (Fig. 5.11).

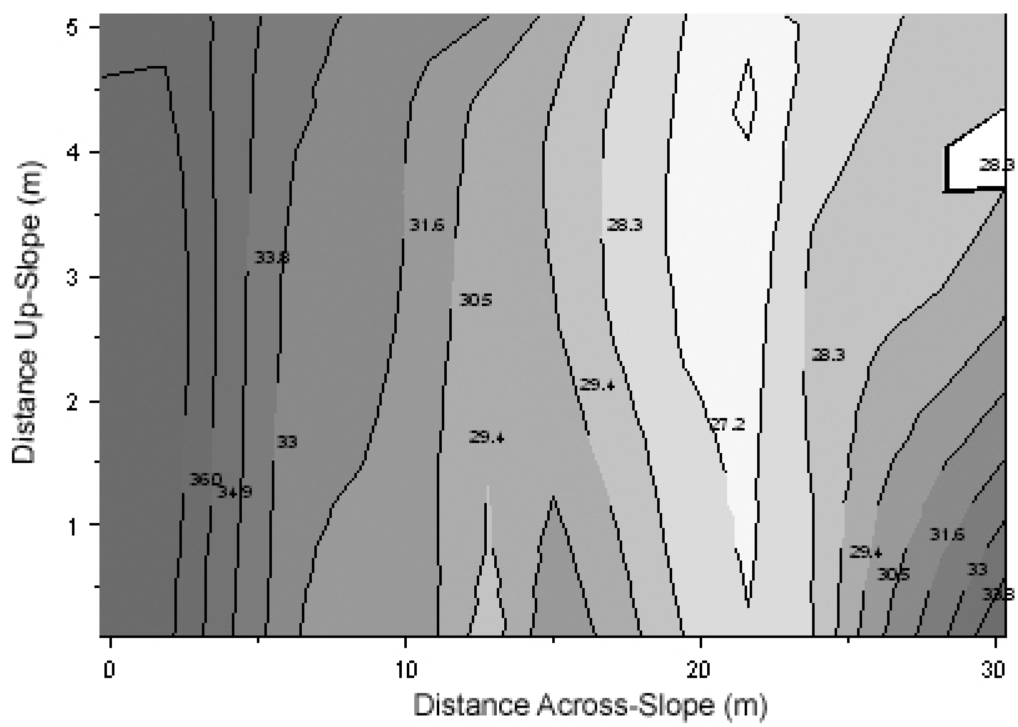


Figure 5.11. Slope angle map for the 20 April 2006 study slope.

Changes in strength over time across this slope are very similar to those documented with smaller sampling patterns at both the transect and sub-transect scale (Fig. 5.7, 5.8), and no trend in the data is noted when plotting slope angle in relation to shear strength (all data over the course of the day) (Fig. 5.12). These findings show that, for these data, temporal factors are the dominant driver of strength across a variable slope over the course of one melt-freeze cycle.

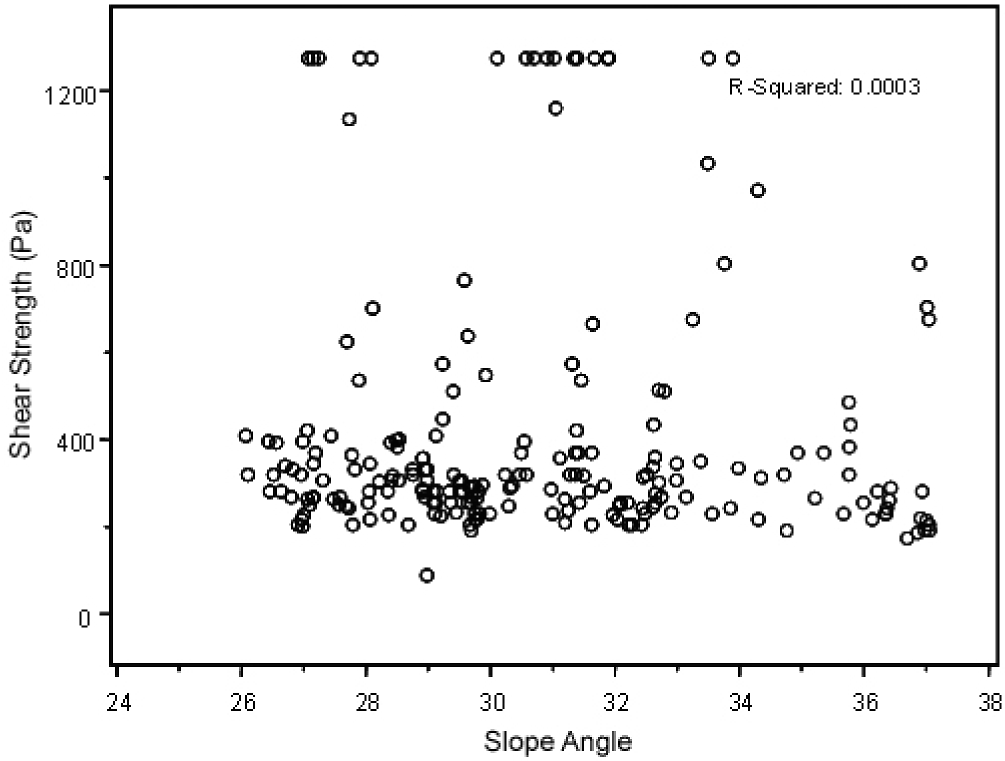


Figure 5.12. Field Day 3 - 20 April 2006. The scatterplot depicts shear strength values in relation to slope angle. Simple linear regression showed no apparent trend in the data.

Research has shown that steepness affects the shear strength measured on slopes (Perla, 1983). The shear stress (σ) equation is written as follows:

$$\sigma = \rho \, g \, z \, \mathrm{Sin} \, (\theta)$$

Where ρ is the density, g is the acceleration due to gravity, z is the height of snow above the shear plane, and θ is the slope angle. Based on this equation, it is apparent that slope angle is integral when calculating shear stress and suggests that during the shear strength experiment, spatial controls (such as slope angle) 20 April 2006, may be masked by primary changes in strength due to heating and cooling.

In order to determine the influence slope angle played on strength, an analysis of shear strength in relation to the Sine of the slope angle was conducted. The sine of the slope angle was used because it represents a close estimate of the affect of slope steepness on shear stress (compared with the actual slope angle). The data set was chosen based upon the six sub-transects in Figure 5.8 that exhibit no statistical change in strength between consecutive tests over time. The idea was to choose a period during the day when the strength was relatively static, and to document the affect slope angle has on measured shear strength during this time (Figure 5.13).

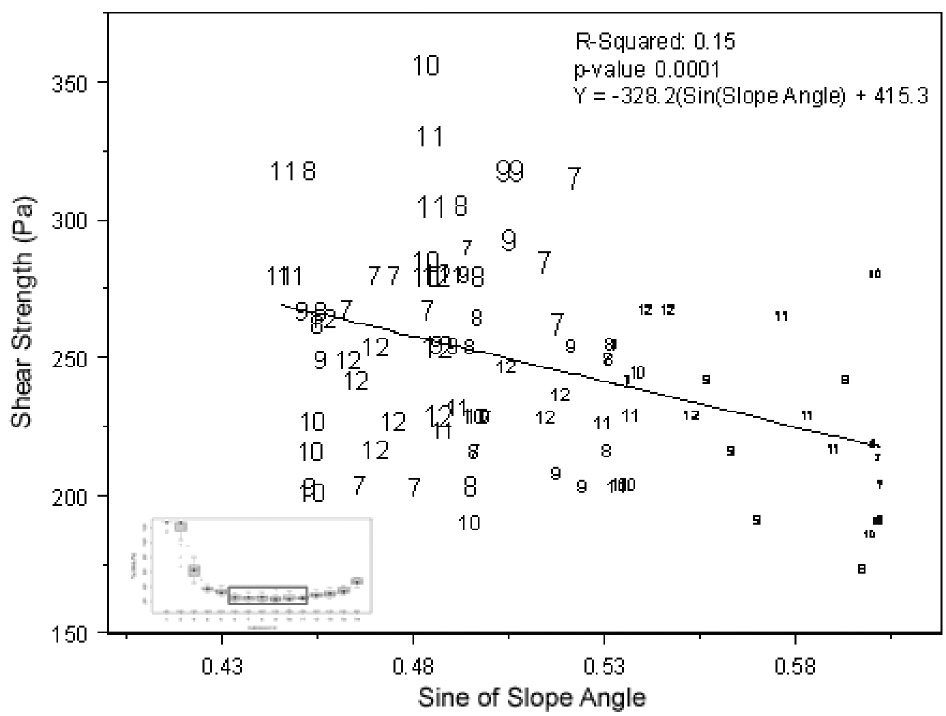


Figure 5.13. Shear strength in relation to the Sine of the slope angle from sub-transects 6 – 11 on Field Day 3 (Fig. 5.8). The numbers plotted relate to the sub-transect from which the data was taken (see box in inset, compare inset to Figure 5.8). Number size relates to the position in the sub-transect from which the individual test was taken. The smallest numbers represent strengths from the left side of the study slope and the largest numbers represent strengths from the right side of the study slope.

The linear representation is able to capture a minor slope scale spatial trend. However, the trend explains only 15% of the variance in shear strength during the time of day when shear strength had reached it's lowest and was relatively static.

The analysis in this section shows that on the slope measured (between 27 and 36 degrees), snow strength is less dependent on small changes in steepness than on temporal factors such as heating and cooling. Spatial variability in the slope scale strength of dry snow has been shown to be very high (Landry et al., 2004). The research presented in this thesis shows that in wet surficial snow, time is the driving factor in strength and far outweighs spatial changes. This finding is important because it suggests that site selection for measurements (of slopes representative of avalanche paths) is much easier than that encountered in dry snow.

Surficial Shear Strength in Relation to Avalanche Activity

Comparing the changes in shear strength between the four field days suggests an association between avalanche activity and surficial shear strength (Fig. 5.10 and 5.14). The data is limited, and comparisons are between data taken from different slopes. Slope elevation, steepness, and aspect vary. The data was collected in melt-freeze snow (grain forms of melt-freeze morphology).

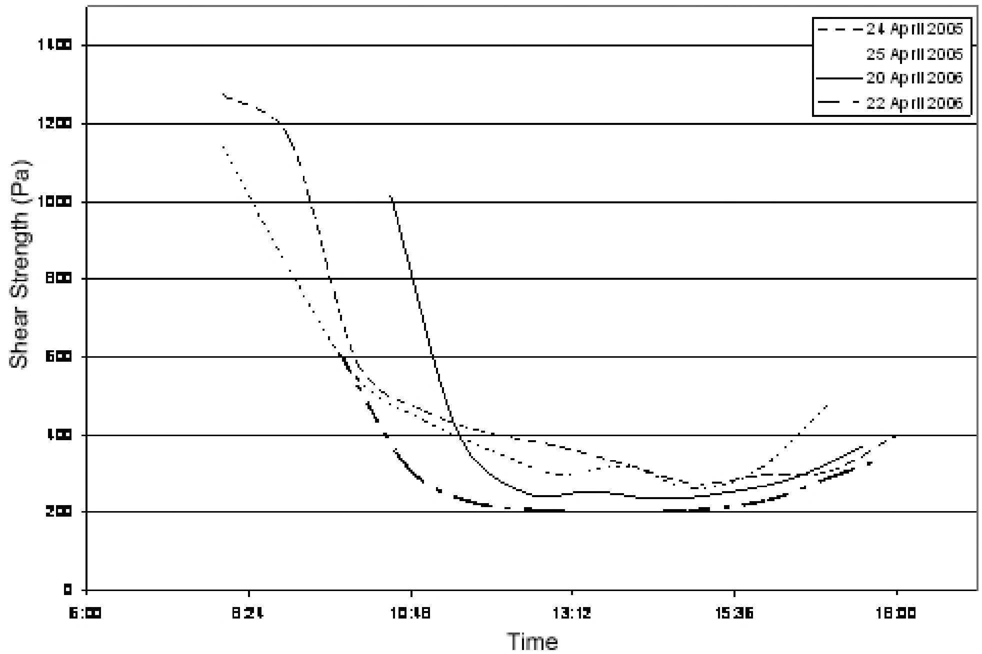


Figure 5.14**.** Mean hourly shear strength over time on each of the four field days (the lines presented span a scatterplot of the data). Minor sluffing was noted on 20 April 2006 and widespread wet loose avalanche activity was noted on 22 April 2006.

Although there is a dramatic decrease in shear strength on all four days, no avalanche activity was noted in the vicinity of the study site on 24 and 25 April 2005 at the North Boundary study site. On 20 April 2006 at the South Bradley study site, 'roller-ball' activity was noted and two minor sluffs occurred around 13:40. At this time the mean shear strength of the study slope was between 254 Pa and 237 Pa (Fig. 5.7, 5.15). Widespread surficial wet loose snow avalanching occurred on 22 April 2006 after the mean shear strength dropped from 316 Pa to 229 Pa and stability increased in conjunction with an increase in shear strength to 293 Pa (Fig. 5.10). These limited data suggest that avalanching may begin when strengths drop below 250 Pa for the slopes and dates studied. Clearly, more work needs to be done to see whether this threshold holds for more general situations. However, the estimate provided by this research presents a possible target value for future studies which might identify an operational threshold that could be used for forecasting wet loose snow avalanches in ski areas.

Conclusions

Changes in the shear strength of surficial wet snow can be documented using shear frames. The data presented shows that the shear strength of wet snow can change dramatically in as little as 20 minutes, suggesting that operationally, avalanche workers may need to monitor slopes as often as fifteen minute intervals. At the slope scale, snow strength is less dependent on small changes in steepness (27 -36 degrees), and spatial location, than on temporal factors such as heating and cooling. This relationship is important because it suggests that the selection of a representative site may be much easier when assessing wet, surficial snow, than when assessing dry snow instabilities. Wet loose avalanche activity coincided with mean shear strengths below 250 Pa, and no evidence of instability was noted when shear strengths exceeded that value. These results are encouraging and future studies would benefit from quantifying the energy balance, water content, and layer morphology to related shear strength measurements.

CHAPTER 6

SUMMARY AND CONCLUSION

Snowpack Stratigraphy

Wet loose snow avalanches are a common phenomenon in isothermal ($0^{o}C$) snowpacks during the late spring in southwestern Montana. Stratigraphy is a dominant factor in melt-water flow, and periods of instability are more dependent upon the amount of free water retained in surficial layers of the snowpack than on the time or the temperature of a given day (Appendix B).

Avalanche workers should conduct regular determinations of where melt-water is accumulating and how much snow is available for transport. Snow profiles show that melt-water 'horizons' within 15cm of the snow surface signal a high probability of wet loose avalanche occurrence (Appendix B-2, 3,7,9,13,14,16,19,20,21). Further, on the slopes studied ~8cm (or more) of weak surficial snow can result in avalanches that are hazardous to skiers (Fig. 3.3), though terrain factors should always be considered. In addition, the development and/or presence of large, poorly bonded, polycrystalline layers below surficial melt-freeze crusts are indicative of the potential for larger avalanches (Fig. 3.11 and 3.12).

Stratigraphy is as important in spring snowpacks as it is in winter snowpacks and should be treated as such. The understanding of wet avalanche prediction and mitigation can be improved by careful stratigraphic documentation. More specifically, analysis of how spring snow changes temporally both on a day scale, and a day to day scale would be beneficial.

Air Temperature

Comparisons of the mean and minimum daily air temperature range for avalanche days and non-avalanche days show an extensive overlap. For example: analysis of the mean daily air temperature shows that all values between the first and third inter-quartile range associated with avalanche days fall within the range of mean daily air temperatures associated with non-avalanche days. Air temperature readings do not adequately capture the energy exchange and subsequent melt, or freezing at the snow surface. Further research needs to apply a quantified snow surface energy balance to wet loose snow avalanche occurrence. In many cases the energy balance equation discussed in section 4.3.3 can be simplified to contain just the net radiation input (estimate of short wave and long wave) and an analysis of the sensible heat exchange (can be achieved through air temperature and wind).

Surficial Shear Strength

Changes in the shear strength of surficial wet snow can be documented using shear frames. The shear strength of wet snow can change dramatically in as little as 20 minutes, suggesting that operationally, avalanche workers may need to monitor slopes in fifteen minute intervals. Targeted strength measurements can be used on slopes with minor variations in steepness. Slopes between 30 and 40 degrees became unstable when the mean shear strength dropped below 250 Pa and showed significant increases in stability when the mean shear strength rose above that value. The data is limited, but these results are encouraging and future study in this area should be pursued. More research is needed to determine if threshold strengths can be applied to avalanche prone slopes. In addition, an understanding of how temporal changes in the energy balance at the snow surface relate to temporal changes in surficial shear strength, and/or hardness, should be should be conducted.

There is an immense amount of potential for research in how the mechanical properties of wet snow relate to energy flux, water flux, and avalanche activity. This thesis is an illustration of how even relatively simple experiments may be able to provide progress in solving relatively complex problems. Additional research can provide a better understanding of wet snow processes, and further the development of operational tools that will assist in the prediction of wet snow avalanches.

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APPENDICES

APPENDIX A

MEASUREMENT OF THE FREE WATER CONTENT OF SNOW

Layer density and the history of liquid occupation largely control properties such as reflectivity, rheology, and the flux of liquid water (Colbeck 1978). Colbeck separates the free water content (F) into two components:

1. % Liquid water saturation, Sw
2. Porosity, Φ

Their relationship can be seen as $F = Sw\Phi$.

Many difficulties are inherent in measurement of the free water content of snow. Snow is a heterogeneous material that may undergo rapid and extensive transformations in crystal properties, strength, permeability, and transmissivity. Because the presence of liquid water directly controls these transformations, measurements of liquid water content take place in a rapidly changing system. In addition, applied studies may have to be conducted in remote locations that require transportable equipment.

Traditional techniques for the measurement of liquid water in snow include centrifugal separation, melting calorimetry, and freezing calorimetry. These techniques, while applicable in certain field situations, are destructive in nature and make comprehensive sampling of water content impossible or difficult at best. Disadvantages stem from flow field disturbance due to active removal of the sample. This can create new surfaces within the flow field and subsequently renders study of spatial variation suspect. In addition, measurements using centrifugal separation and melting calorimetry are inherently inaccurate (Colbeck, 1978). Freezing calorimetry indicates the liquid water content of a snow sample by measuring the latent heat of fusion. In an insulated bottle, a freezing agent (silicon oil) is added to the snow sample. Measurement of the amount of heat gained is a direct measure of the amount of liquid water frozen (Stein et al., 1996). Although there is some argument that freezing calorimetry is inherently open to error in field applications (Denoth, 1996), This method is accurate to 1-2% by weight and is accepted for use in laboratory calibration of non-destructive methods (Stein et al., 1996).

The first relatively non-destructive methods used in alpine applications were centered around dielectric technique. Methodology used for the transition to snow had been around since the 1930's when capacitance meters were used to determine the liquid content of wheat and later other materials such as soil. Gerdel 1954 used a capacitance probe operating at frequency of 1.5 MHz and was probably the first to apply dielectric devices to snow (Cobeck 1978).

Snow can be considered a heterogeneous dielectric material consisting of ice, air, in some instances water, and is used as a capacitor in conjunction with a probe of known resistance. The density of snow and its liquid water content are correlated to calibrated dielectric properties (Coleou and Lesaffre, 1998). At 1MHz, water has a dielectric constant of 88 while that of ice is 3 (Stein et al., 1996). The snow dielectric constant is independent of the snow temperature around 0^{o}C, effectively allowing a variety of moisture contents at zero degrees during the melting process (Stein and Kane, 1983). Although the dielectric constant of snow is sensitive to small changes in the amount of liquid water present, it is difficult to interpret the constant of the solid-liquid-gas mixture because of 'the importance of shape factors on the contribution of each phase to the dielectric constant of the media' (Colbeck, 1978).

The accuracy of these devices is dependant on our understanding of the relationship between the volumetric liquid water content, the dielectric constant, and the density of snow. Denoth 1989, describes that relationship in the following way:

$K_{red} = 0.187\Theta + 0.0045\Theta^2$

$K_{red} = K - 0.00192\rho_s - 0.44 \times 10^{-6}\rho_s$

Solved for Θ:

$\Theta = -20.77 + (20.77^2 + 222.222 * K_{red})^{1/2}$

Where:

K = the dielectric constant

K_{red} = the dielectric constant reduced for variations in snow density

Θ = the volumetric water content

ρ_s = density of snow (kg/m^3) (Lundberg, 1996)

Denoth's relationship is used with capacitive methods to apply dielectric measurements to the determination of free water content and has been shown to be valid by Denoth in 1989 for contents below 6% and Perla in 1991 for contents beyond 6% (Lundberg, 1996).

A popular dielectric technique somewhat different than capacitive methods is Time Domain Reflectometry (TDR). TDR was initially developed for use in soils and has been quite successful in determining moisture content in that medium. TDR produces a fast voltage signal containing frequencies from 10MHz up to one GHz that is channeled through the snow by parallel metal rods (Stein et al., 1996). The TDR measures the propagation time of the pulse between the rods; the length of the rods is inversely proportional to the speed of the pulse (Stein et al., 1996). When the length of the rod is known, the permittivity of the insulator (in this case snow) around the probe can be determined. In 1983 Stein and Kane used the TDR technique to measure the relative dielectric constant of snow. Initially, they thought that just as in soil, the relationship between the water content and the dielectric constant would be independent of density. After experimentation, it was apparent that the snow dielectric constants vary significantly due to density alone (Stein and Kane, 1983). This observation is reasonable given the differences in the dielectric constant of air and ice. An advantage compared to capacitive methods is small sensitivity to variable conductivity. TDR sensors can be specifically configured to the medium of interest. The main disadvantage to this technique is interpretation of the complex signal received. Relationships of the ice, air, and water mixtures to the travel time of the magnetic pulse, are described by Schneeblei et al., 1998.

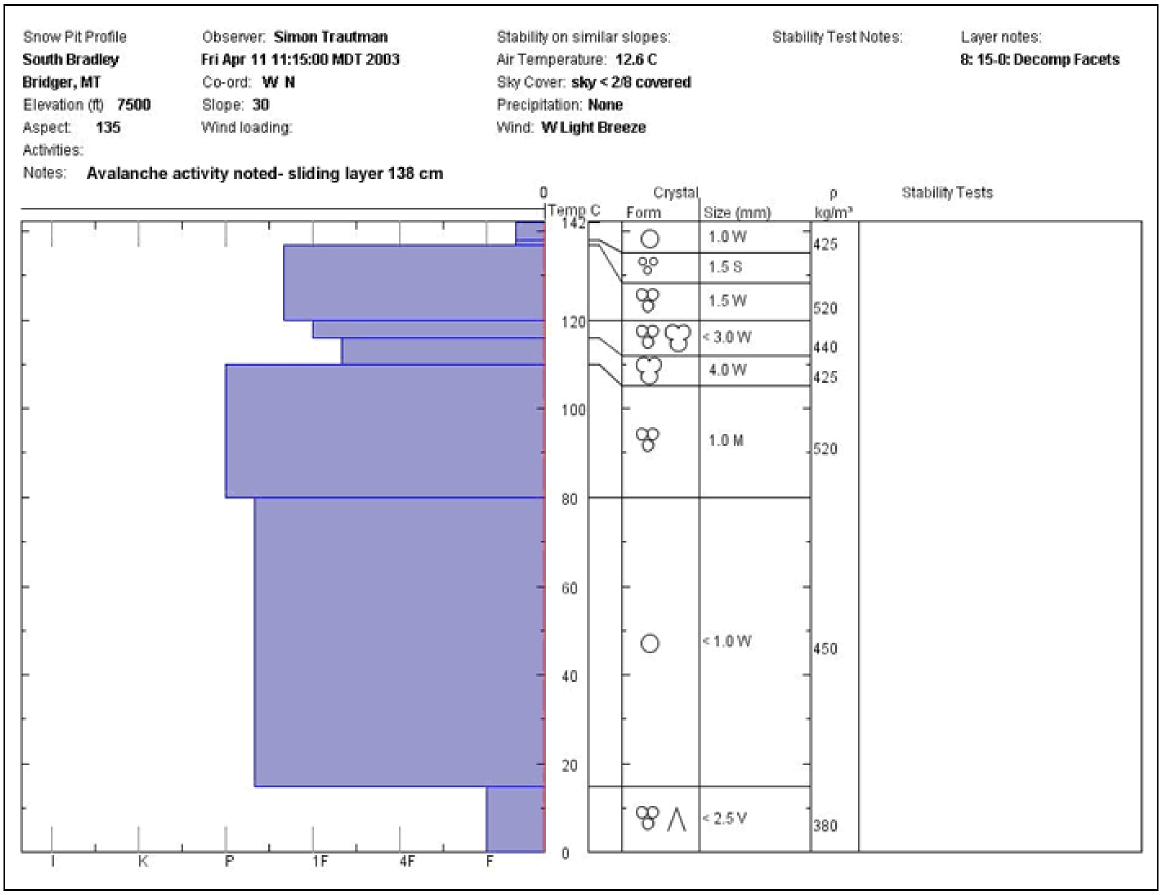
With the development of remote sensing, and technological advances in microprocessor assisted monitoring devices, there has been an increase in the demand for non-destructive measurement systems that can determine and record the liquid water content in situ with high accuracy and simplicity in operation (Denoth, 1996). Methods based on the determination of dielectric function at frequencies exceeding 10 MHz allow for relatively precise, rapid and non-destructive measurements in the field (Denoth, 1996). The design of dielectric sensors depends on their operating frequencies: flat plate sensors are used in for radio frequencies (20-50 MHz), monopole antennas and snow forks operate in the 100MHz – 3GHz range, and microwave X and K bands (8-16 GHz) have been used (with little success). For sensor operations in excess of 4 GHz, the sensitivity to changes in liquid water content decreases significantly (Denoth, 1996).

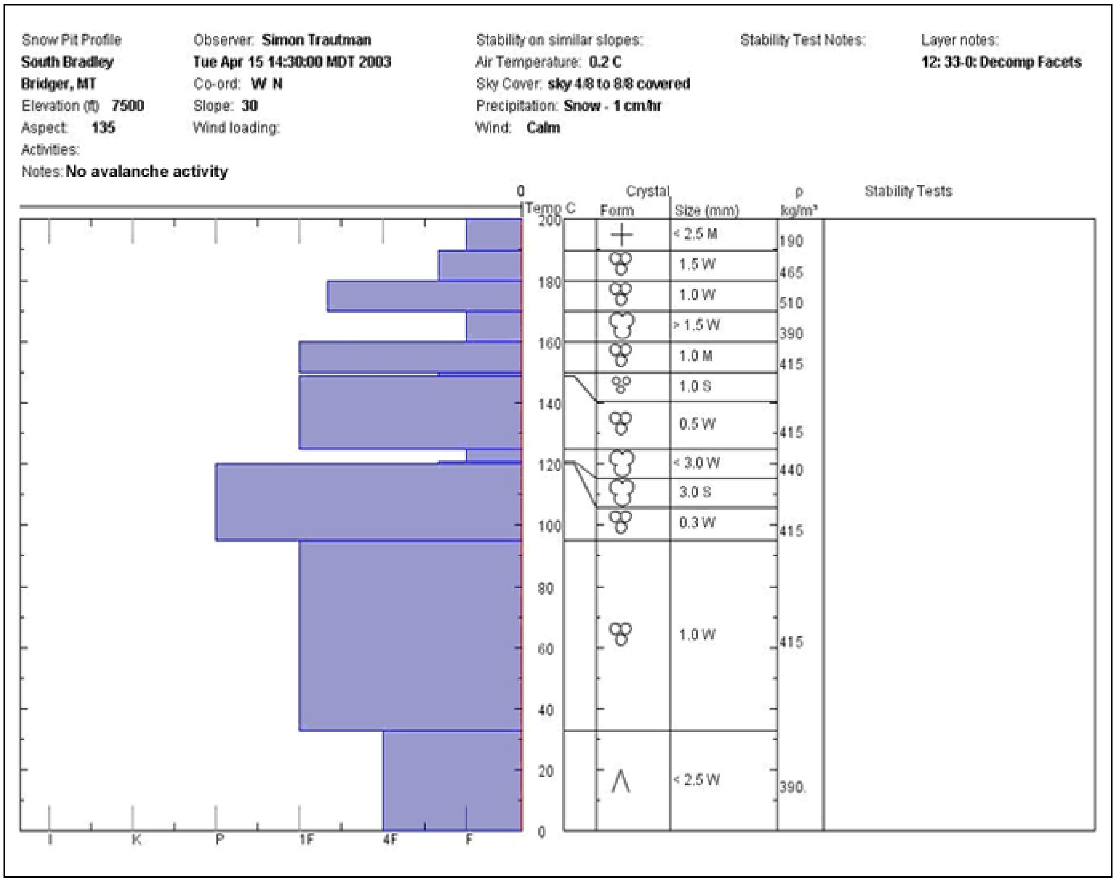
Currently, it appears that flat plate sensors such as the Denoth meter (a capacitance probe which measures an area of 13x9 cm2 and operates at 27 MHz) offer the highest resolution for the measurement of liquid water in the snowpack. These devices have had a reasonable degree of success and are used by organizations such as AAR at the University of Colorado. The problem with this device is that it requires separate density measurements in order to calculate water content. The sensors are of higher specific density than seasonal snow and can absorb a high amount of solar radiation when located near the surface (Schneeblei et al., 1998), making it difficult to set up a full-pack wetness measurement system and/or long-term measurements.

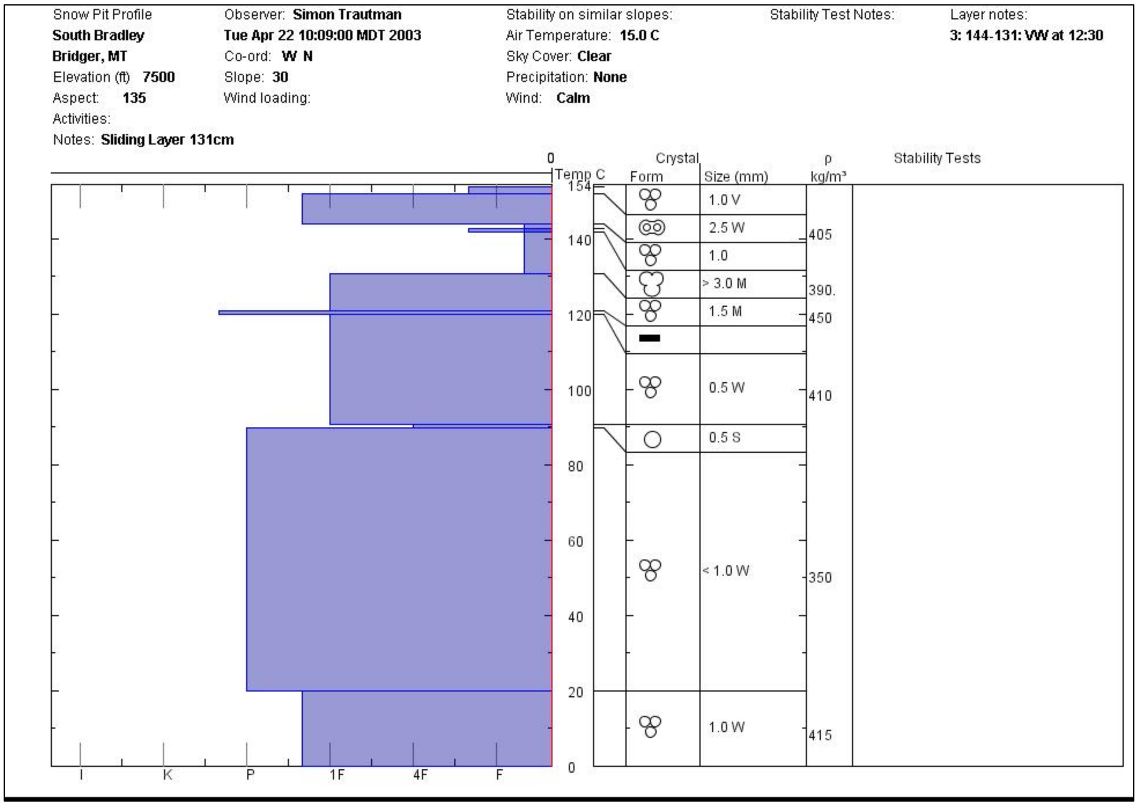
Two pronged snow probes such as the *snow fork* and *snow probe* are currently in use with TDR technology to monitor snow density and water content. They are advantageous due to multiplexing capabilities and flexibility in construction, and have proved effective in tracking infiltration patterns in snow (Schneeblei et al., 1998). Difficulties in manufacturing laboratory samples of wet snow for calibration have led to problems estimating free water content (Stein, 1996; Lundberg 1996; Schneeblei et al., 1998). Results have been reasonable, but differ consistently with those found using the Denoth meter and freezing calorimetry.

APPENDIX B

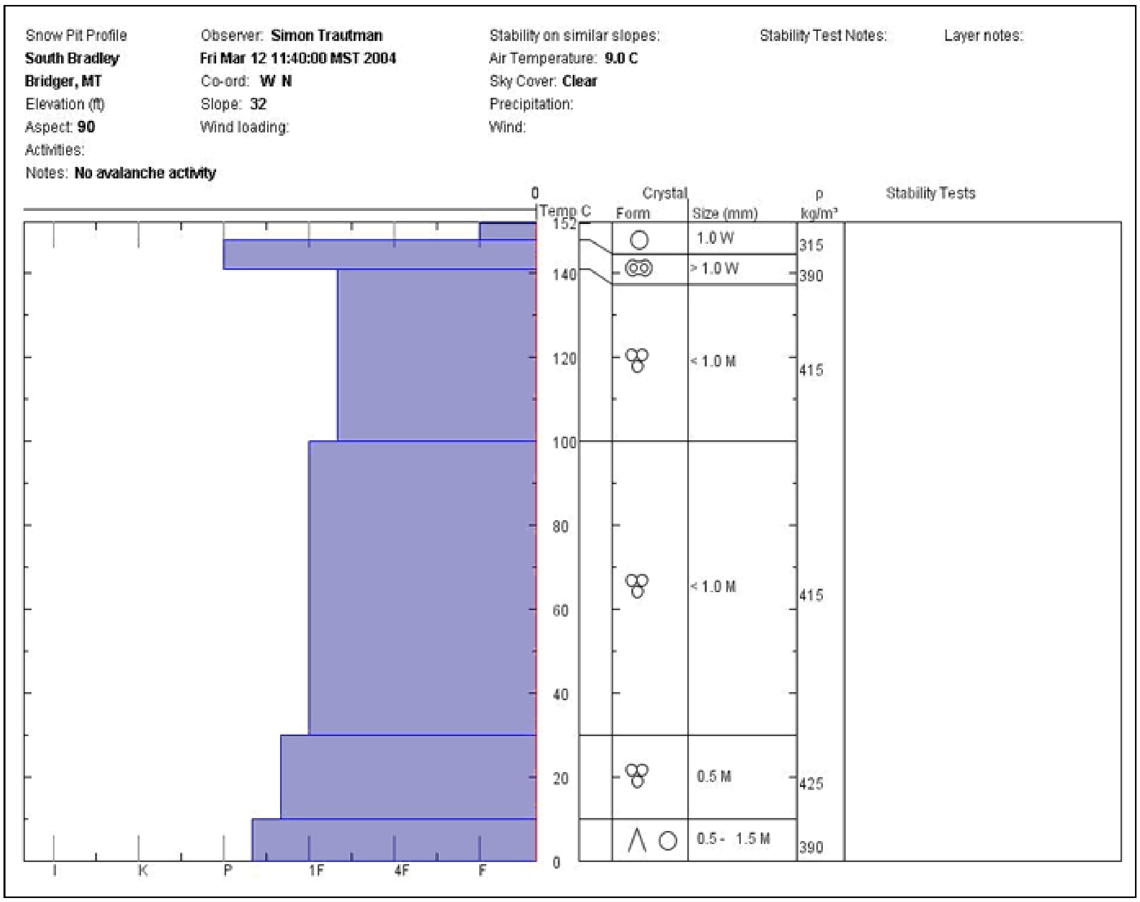
SNOW PIT PROFILES 2003 AND 2004


Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect: 135
Activities:
Notes: Avalanche activity noted- sliding layer 138 cm
Observer: Simon Trautman
Fri Apr 11 11:15:00 MDT 2003
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 12.6 C
Sky Cover: sky < 2/8 covered
Precipitation: None
Wind: W Light Breeze
Stability Test Notes:
Layer notes:
8: 15-0: Decomp Facets
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
1.0 W
1.5 S
1.5 W
< 3.0 W
4.0 W
1.0 M
< 1.0 W
< 2.5 V
425
520
440
425
520
450
380
142
120
100
80
60
40
20
0
I
K
P
1F
4F
F

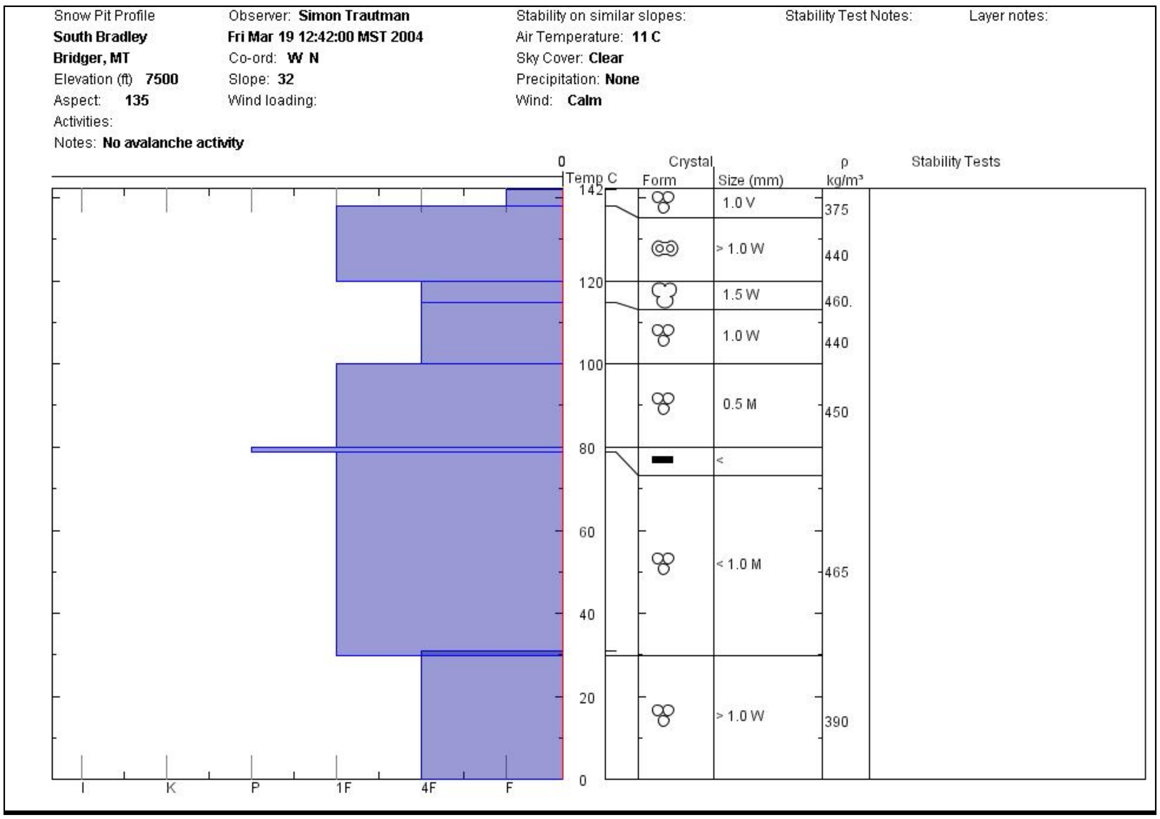

Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect 135
Activities:
Notes: No avalanche activity
Observer: Simon Trautman
Tue Apr 15 14:30:00 MDT 2003
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 0.2 C
Sky Cover: sky 4/8 to 8/8 covered
Precipitation: Snow - 1 cm/hr
Wind: Calm
Stability Test Notes:
Layer notes:
12: 33-0: Decomp Facets
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
< 2.5 M
1.5 W
1.0 W
> 1.5 W
1.0 M
1.0 S
0.5 W
< 3.0 W
3.0 S
0.3 W
1.0 W
< 2.5 W
190
465
510
390
415
415
440
415
415
390.
I
K
P
1F
4F
F



Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect: 135
Activities:
Notes: Sliding Layer 131cm
Observer: Simon Trautman
Tue Apr 22 10:09:00 MDT 2003
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 15.0 C
Sky Cover: Clear
Precipitation: None
Wind: Calm
Stability Test Notes:
Layer notes:
3: 144-131: VW at 12:30
0
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
154
140
120
100
80
60
40
20
0
1.0 V
2.5 W
1.0
> 3.0 M
1.5 M
0.5 W
0.5 S
< 1.0 W
1.0 W
405
390.
450
410
350
415
I
K
P
1F
4F
F

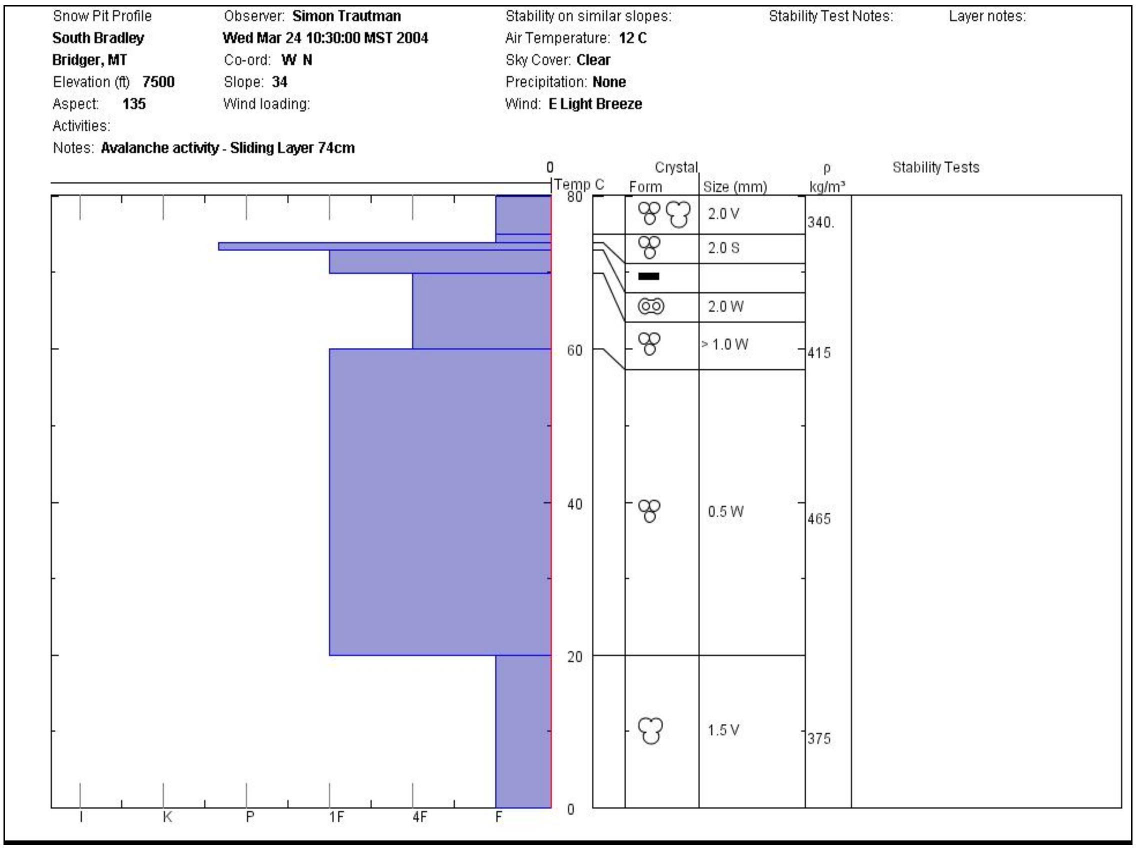

Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft)
Aspect: 90
Activities:
Notes: No avalanche activity
Observer: Simon Trautman
Fri Mar 12 11:40:00 MST 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 9.0 C
Sky Cover: Clear
Precipitation:
Wind:
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ kg/m³
Stability Tests
1.0 W
315
> 1.0 W
390
< 1.0 M
415
< 1.0 M
415
0.5 M
425
0.5 - 1.5 M
390
152
140
120
100
80
60
40
20
0
I
K
P
1F
4F
F

B5



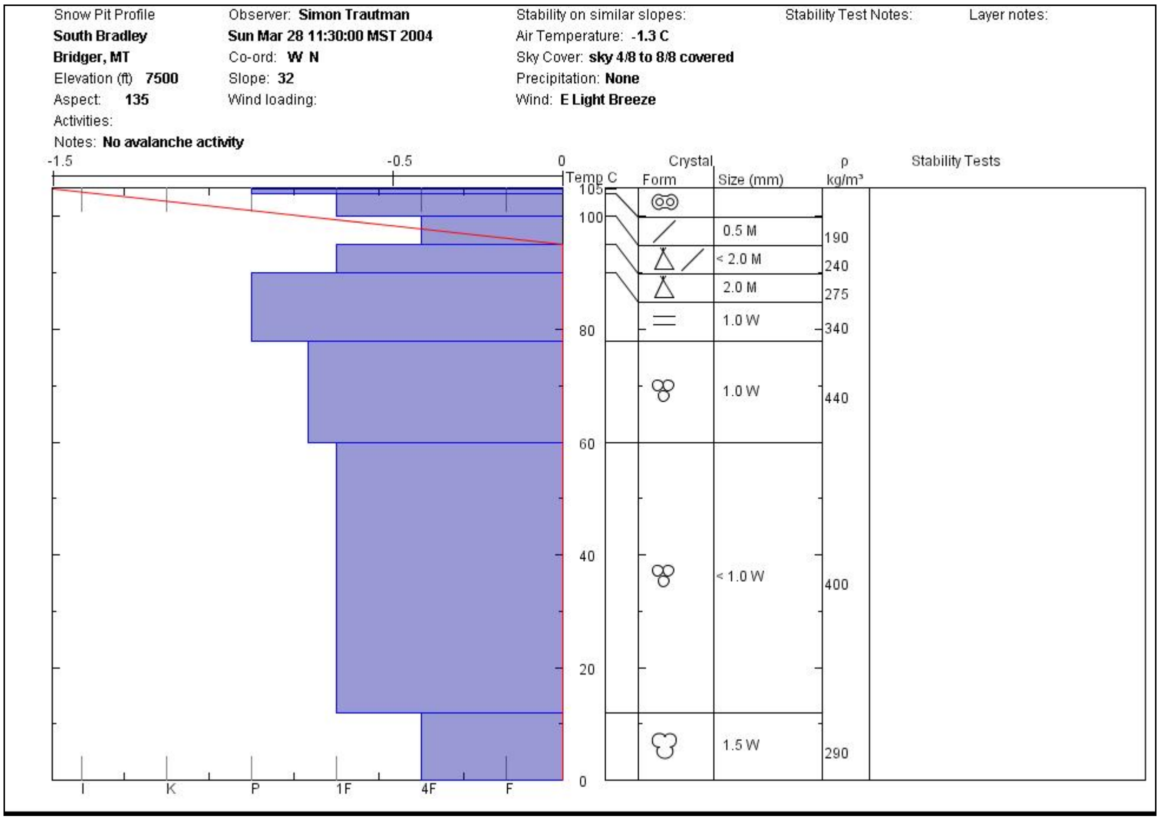
Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect: 135
Activities:
Notes: No avalanche activity
Observer: Simon Trautman
Fri Mar 19 12:42:00 MST 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 11 C
Sky Cover: Clear
Precipitation: None
Wind: Calm
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
1.0 V
375
> 1.0 W
440
1.5 W
460.
1.0 W
440
0.5 M
450
<
< 1.0 M
465
> 1.0 W
390
142
120
100
80
60
40
20
0
I
K
P
1F
4F
F

B6

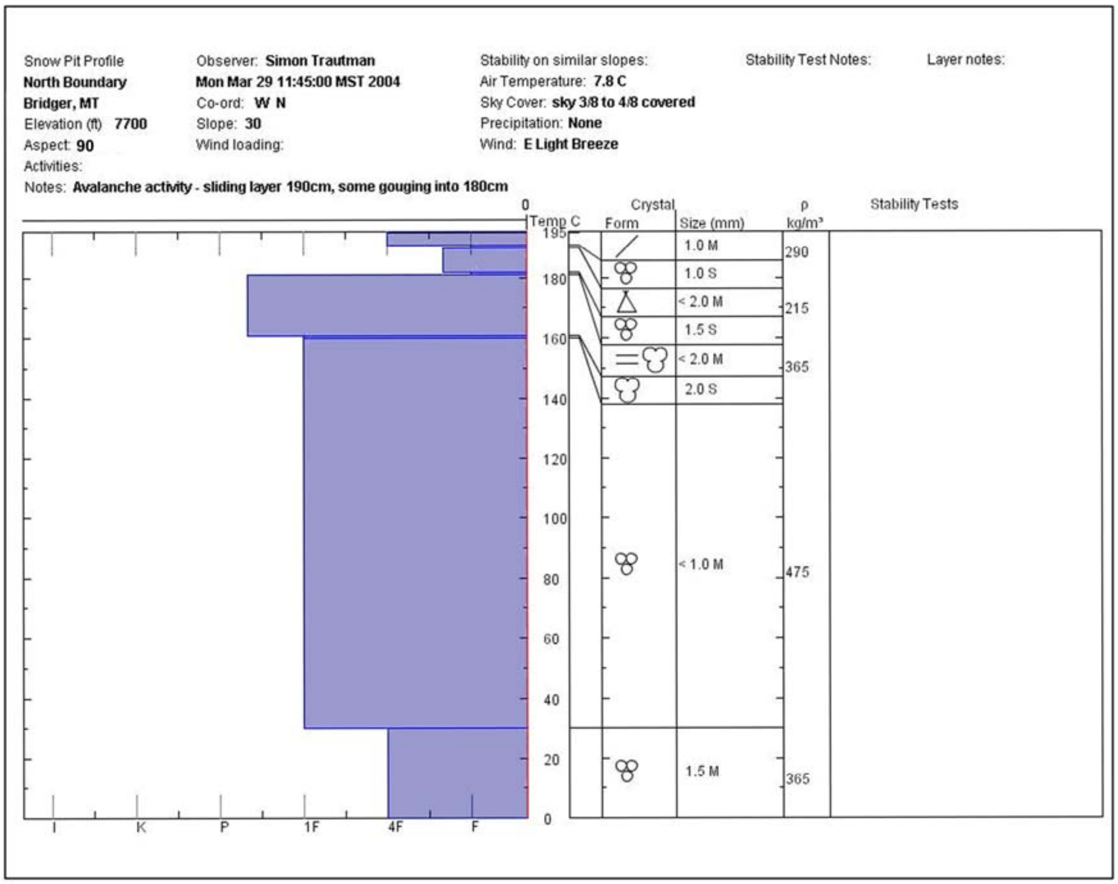


Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect: 135
Activities:
Notes: Avalanche activity - Sliding Layer 74cm
Observer: Simon Trautman
Wed Mar 24 10:30:00 MST 2004
Co-ord: W N
Slope: 34
Wind loading:
Stability on similar slopes:
Air Temperature: 12 C
Sky Cover: Clear
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
Temp C
2.0 V
2.0 S
2.0 W
> 1.0 W
0.5 W
1.5 V
340.
415
465
375
80
60
40
20
0
I
K
P
1F
4F
F

B7

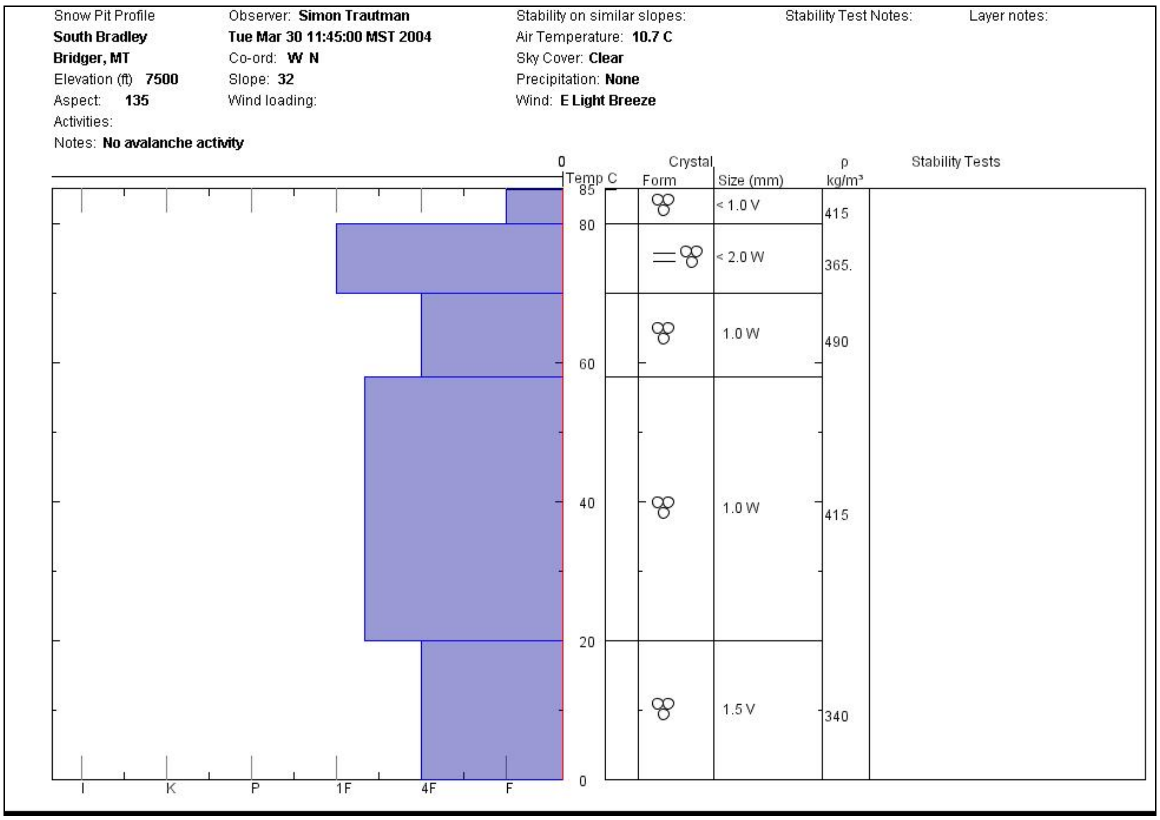

Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect: 135
Activities:
Notes: No avalanche activity
Observer: Simon Trautman
Sun Mar 28 11:30:00 MST 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: -1.3 C
Sky Cover: sky 4/8 to 8/8 covered
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
-1.5
-0.5
0
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
105
100
80
60
40
20
0
0.5 M
< 2.0 M
2.0 M
1.0 W
1.0 W
< 1.0 W
1.5 W
190
240
275
340
440
400
290
I
K
P
1F
4F
F

B8



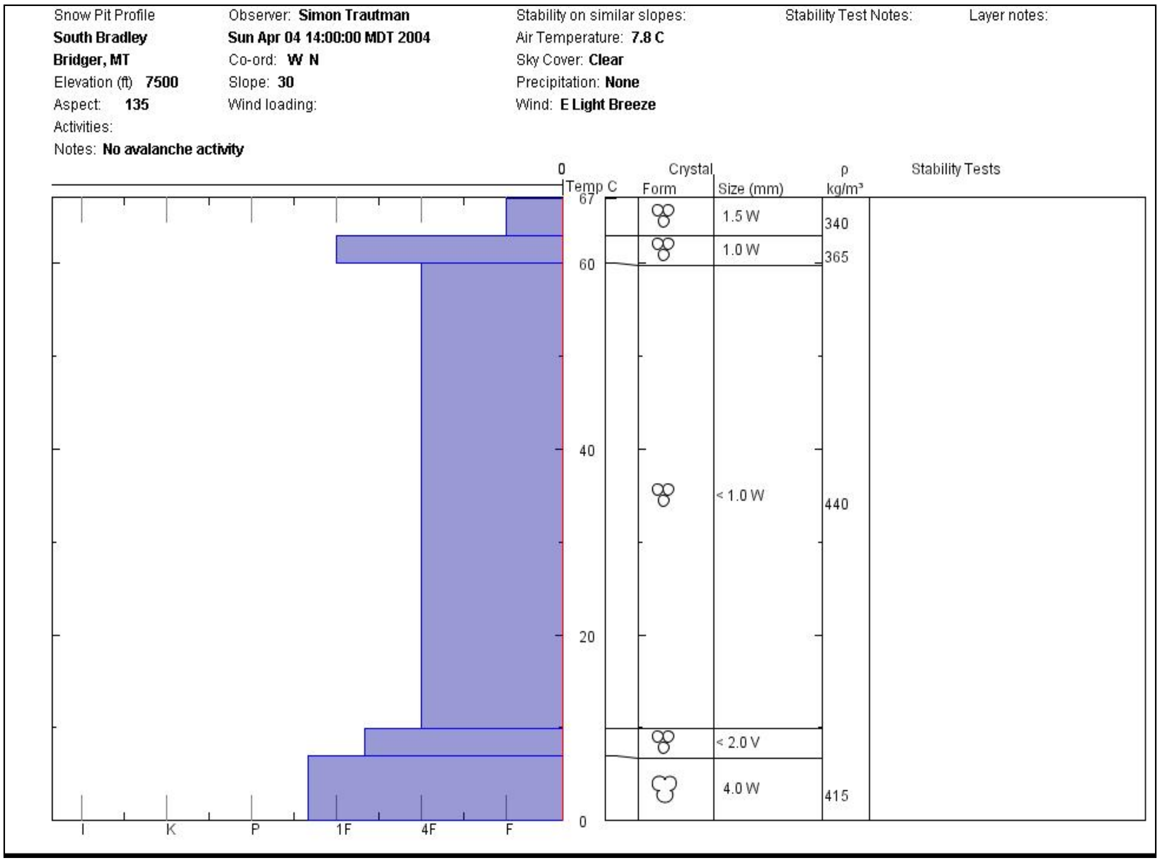
Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Observer: Simon Trautman
Mon Mar 29 11:45:00 MST 2004
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 7.8 C
Sky Cover: sky 3/8 to 4/8 covered
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
Notes: Avalanche activity - sliding layer 190cm, some gouging into 180cm
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
1.0 M
1.0 S
< 2.0 M
1.5 S
< 2.0 M
2.0 S
< 1.0 M
1.5 M
290
215
365
475
365
195
180
160
140
120
100
80
60
40
20
0
I
K
P
1F
4F
F

B9

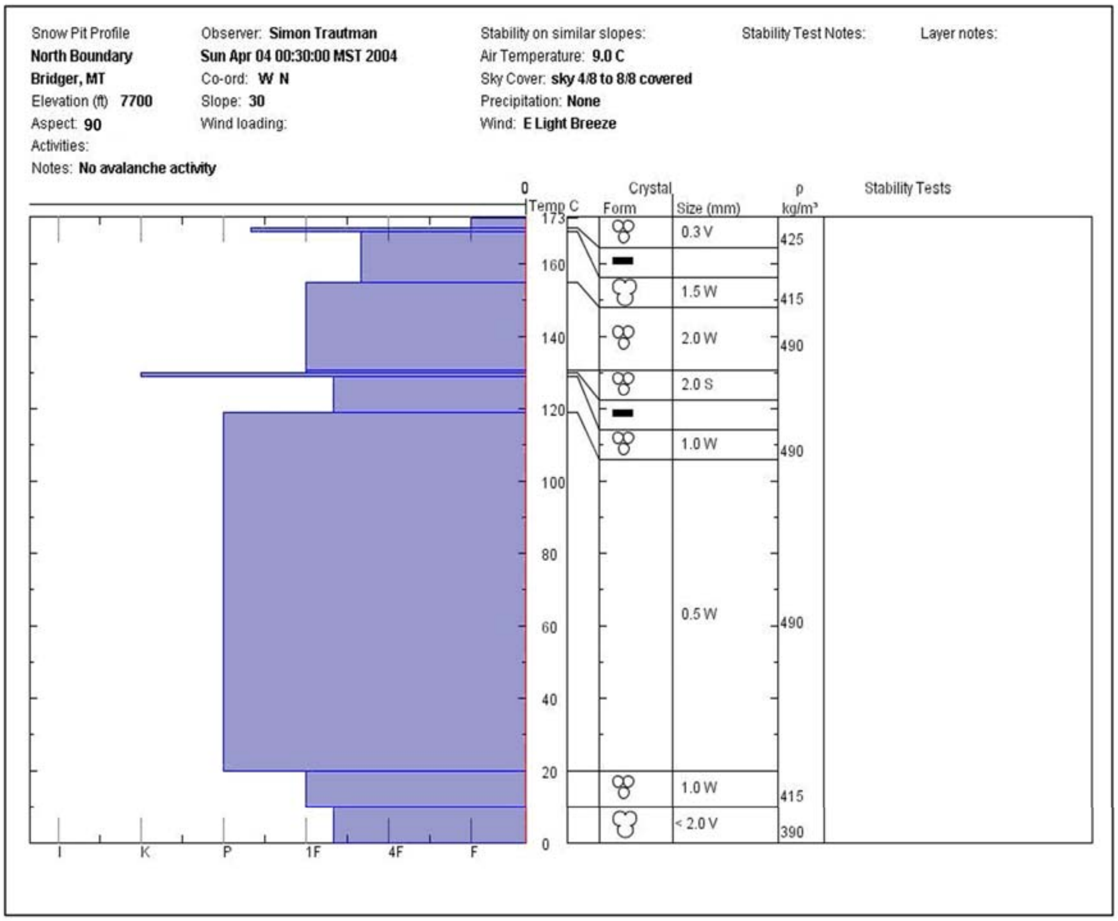


Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect: 135
Activities:
Notes: No avalanche activity
Observer: Simon Trautman
Tue Mar 30 11:45:00 MST 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 10.7 C
Sky Cover: Clear
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
0
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
85
80
60
40
20
0
< 1.0 V
< 2.0 W
1.0 W
1.0 W
1.5 V
415
365.
490
415
340
I
K
P
1F
4F
F

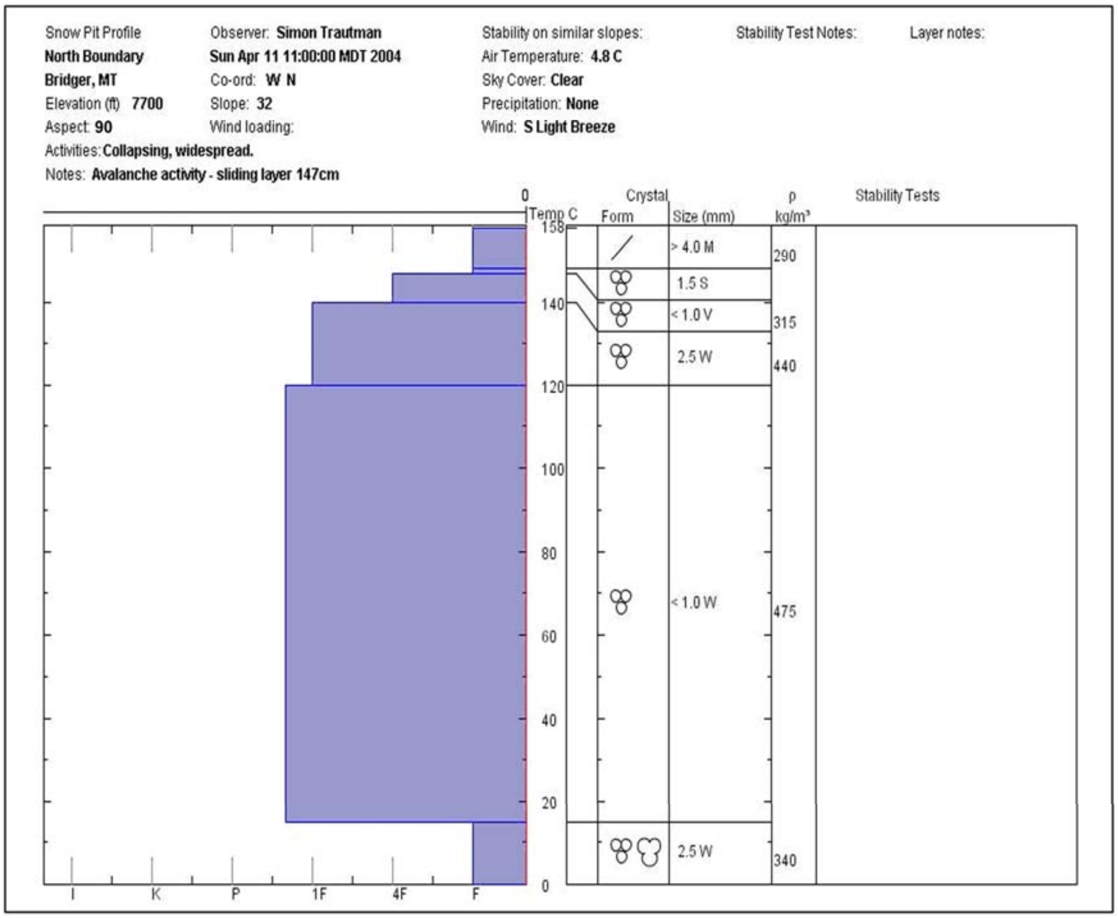
B10


Snow Pit Profile
South Bradley
Bridger, MT
Elevation (ft) 7500
Aspect: 135
Activities:
Notes: No avalanche activity
Observer: Simon Trautman
Sun Apr 04 14:00:00 MDT 2004
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 7.8 C
Sky Cover: Clear
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
67
60
40
20
0
1.5 W
1.0 W
< 1.0 W
< 2.0 V
4.0 W
340
365
440
415
I
K
P
1F
4F
F

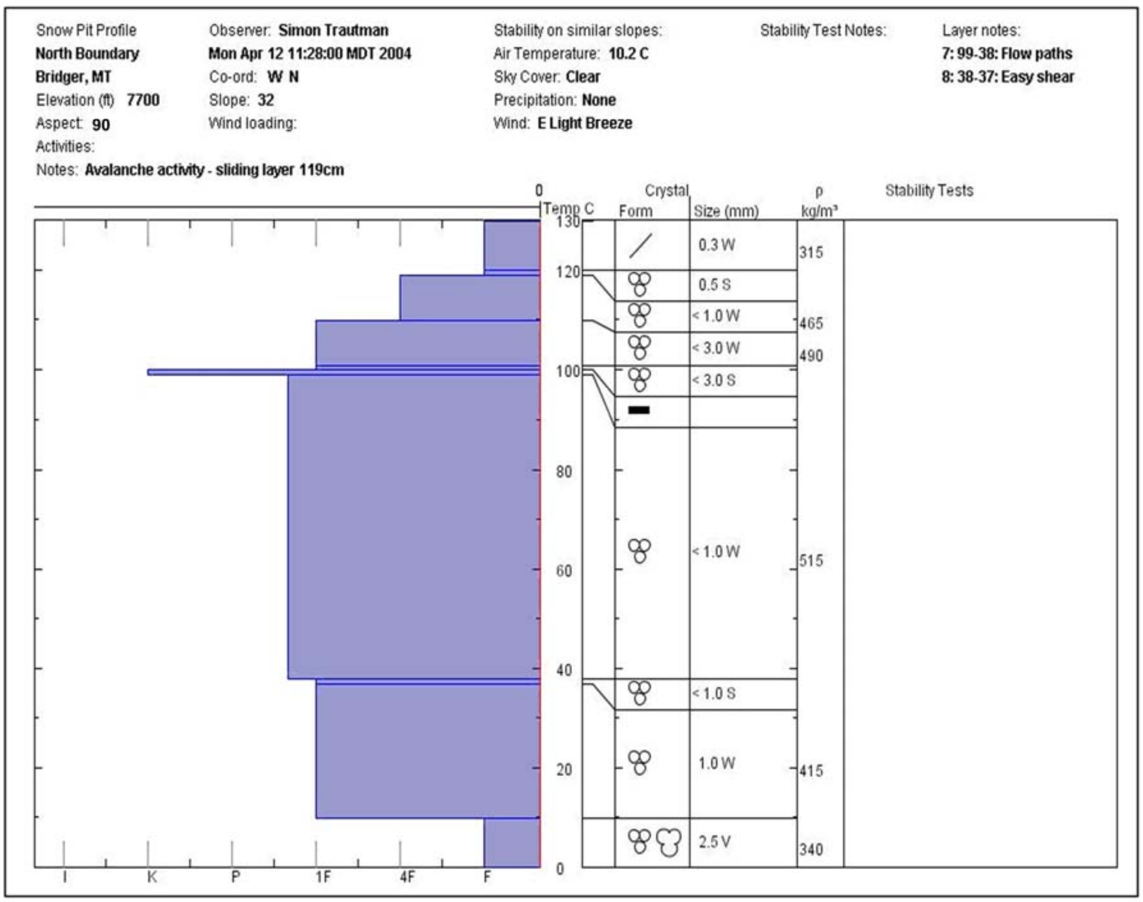
B11


Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: No avalanche activity
Observer: Simon Trautman
Sun Apr 04 00:30:00 MST 2004
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 9.0 C
Sky Cover: sky 4/8 to 8/8 covered
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ kg/m³
Stability Tests
0.3 V
1.5 W
2.0 W
2.0 S
1.0 W
0.5 W
1.0 W
< 2.0 V
425
415
490
490
490
415
390
I
K
P
1F
4F
F

B12

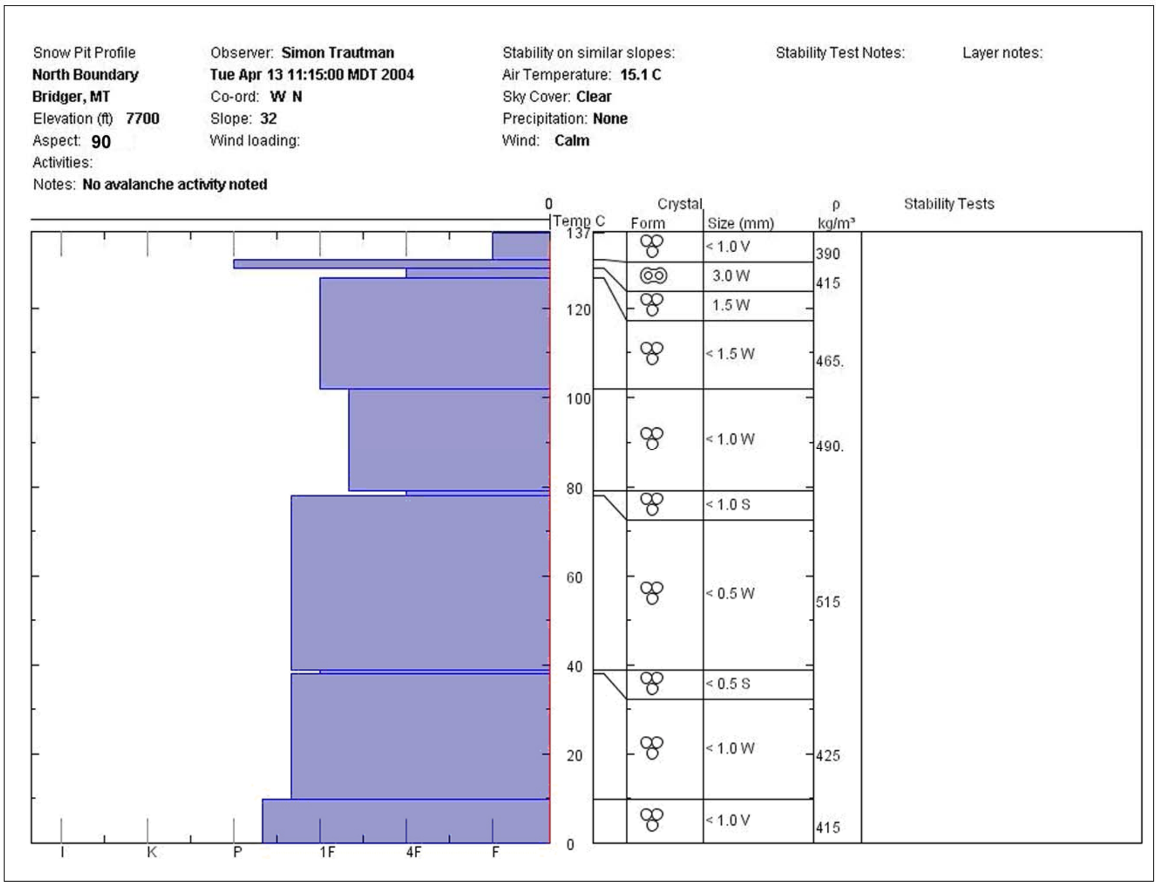

Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities: Collapsing, widespread.
Notes: Avalanche activity - sliding layer 147cm
Observer: Simon Trautman
Sun Apr 11 11:00:00 MDT 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 4.8 C
Sky Cover: Clear
Precipitation: None
Wind: S Light Breeze
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
> 4.0 M
1.5 S
< 1.0 V
2.5 W
< 1.0 W
2.5 W
290
315
440
475
340
158
140
120
100
80
60
40
20
0
I
K
P
1F
4F
F

B13



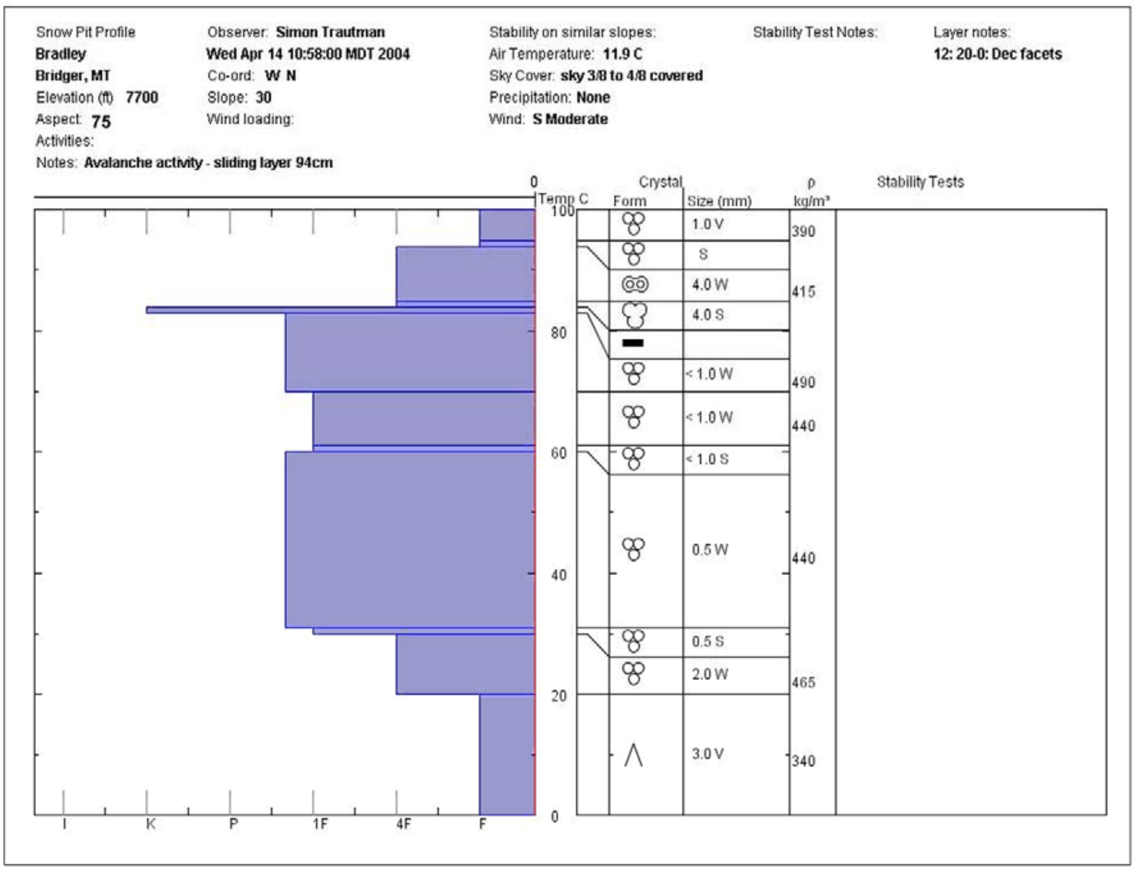
Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: Avalanche activity - sliding layer 119cm
Observer: Simon Trautman
Mon Apr 12 11:28:00 MDT 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 10.2 C
Sky Cover: Clear
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
7: 99-38: Flow paths
8: 38-37: Easy shear
Temp C
Crystal
Form
Size (mm)
ρ kg/m³
Stability Tests
0.3 W
0.5 S
< 1.0 W
< 3.0 W
< 3.0 S
< 1.0 W
< 1.0 S
1.0 W
2.5 V
315
465
490
515
415
340
130
120
100
80
60
40
20
0
I
K
P
1F
4F
F

B14

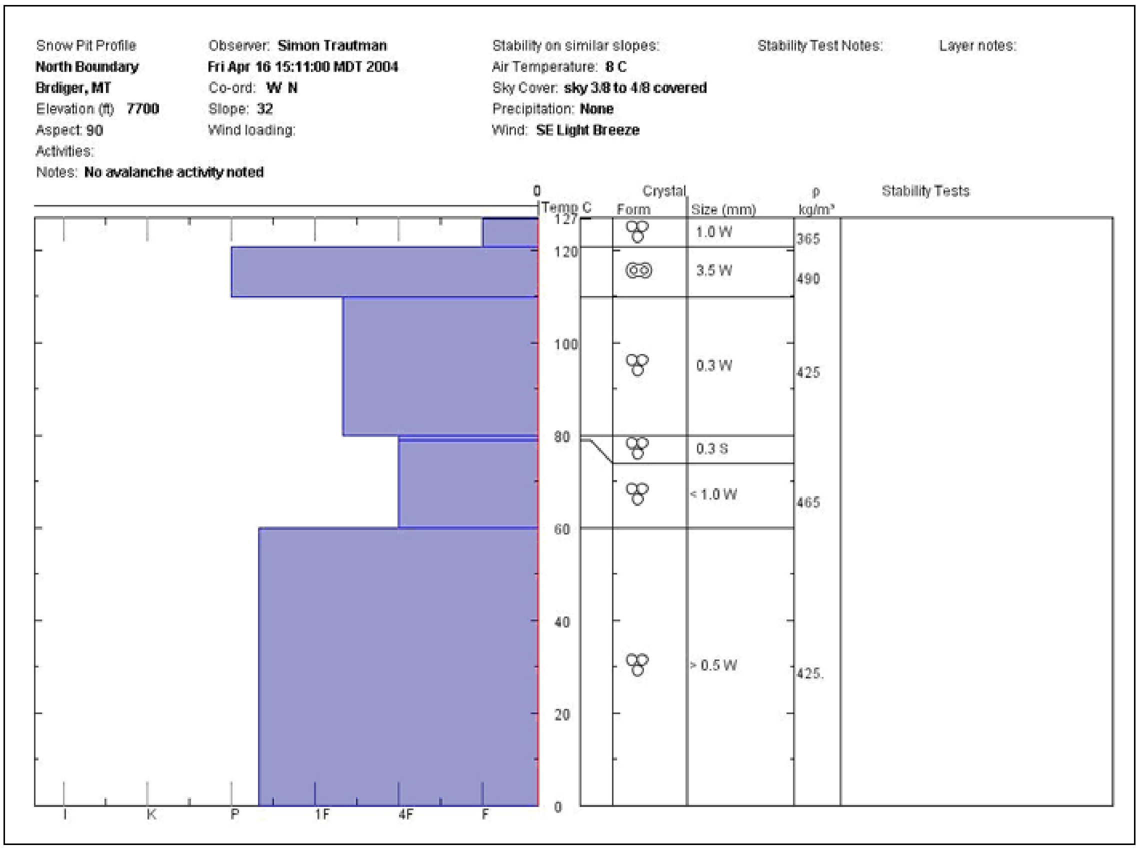


Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: No avalanche activity noted
Observer: Simon Trautman
Tue Apr 13 11:15:00 MDT 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 15.1 C
Sky Cover: Clear
Precipitation: None
Wind: Calm
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
137
120
100
80
60
40
20
0
< 1.0 V
3.0 W
1.5 W
< 1.5 W
< 1.0 W
< 1.0 S
< 0.5 W
< 0.5 S
< 1.0 W
< 1.0 V
390
415
465.
490.
515
425
415
I
K
P
1F
4F
F

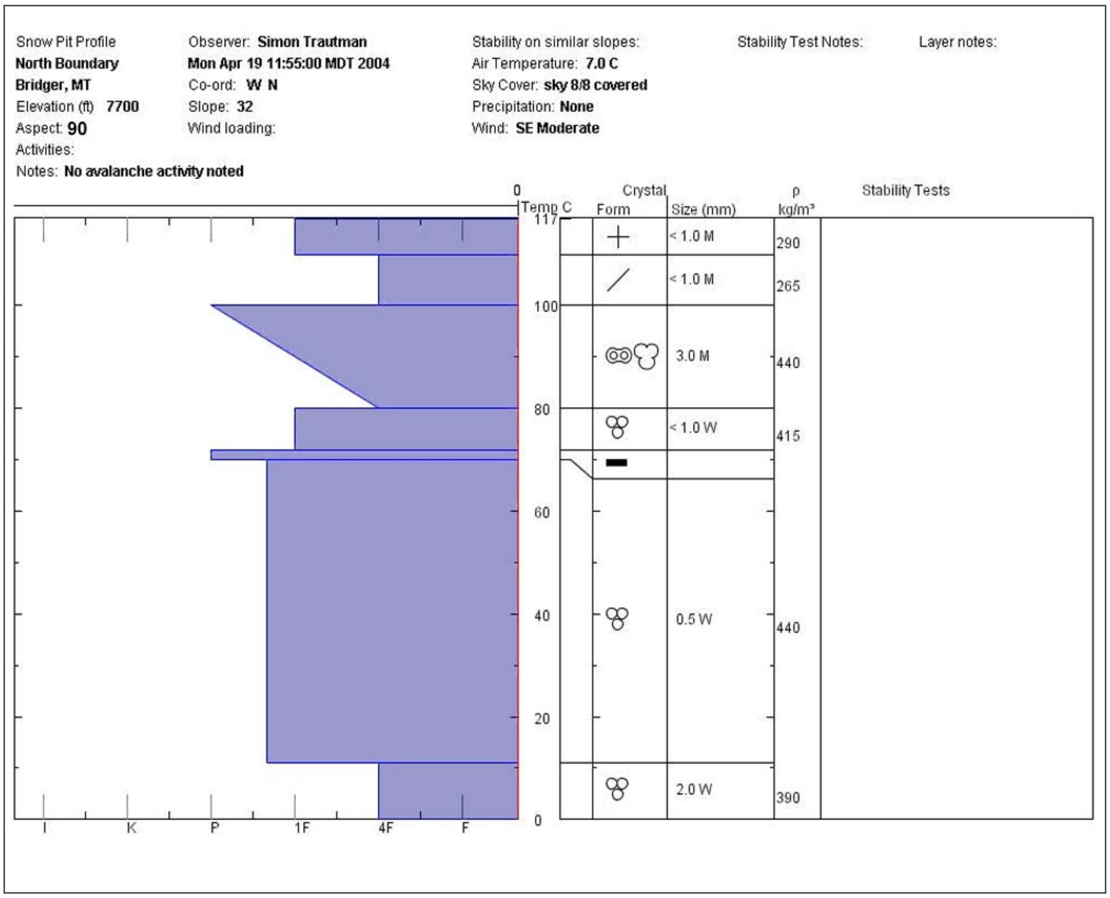
B15


Snow Pit Profile
Bradley
Bridger, MT
Elevation (ft) 7700
Aspect: 75
Activities:
Notes: Avalanche activity - sliding layer 94cm
Observer: Simon Trautman
Wed Apr 14 10:58:00 MDT 2004
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 11.9 C
Sky Cover: sky 3/8 to 4/8 covered
Precipitation: None
Wind: S Moderate
Stability Test Notes:
Layer notes:
12: 20-0: Dec facets
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
1.0 V
S
4.0 W
4.0 S
< 1.0 W
< 1.0 W
< 1.0 S
0.5 W
0.5 S
2.0 W
3.0 V
390
415
490
440
440
465
340
100
80
60
40
20
0
I
K
P
1F
4F
F

B16

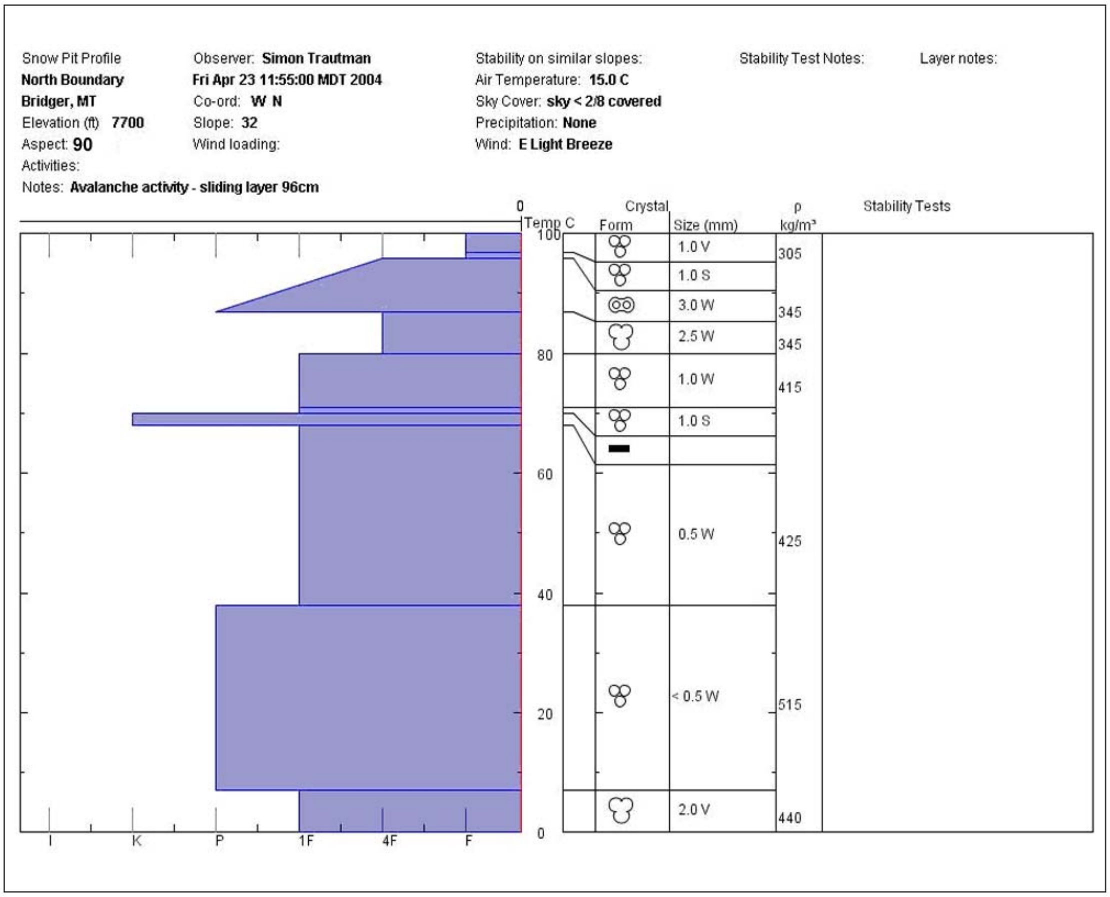

Snow Pit Profile
North Boundary
Brdiger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: No avalanche activity noted
Observer: Simon Trautman
Fri Apr 16 15:11:00 MDT 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 8 C
Sky Cover: sky 3/8 to 4/8 covered
Precipitation: None
Wind: SE Light Breeze
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
1.0 W
3.5 W
0.3 W
0.3 S
< 1.0 W
> 0.5 W
365
490
425
465
425.
127
120
100
80
60
40
20
0
I
K
P
1F
4F
F

B17

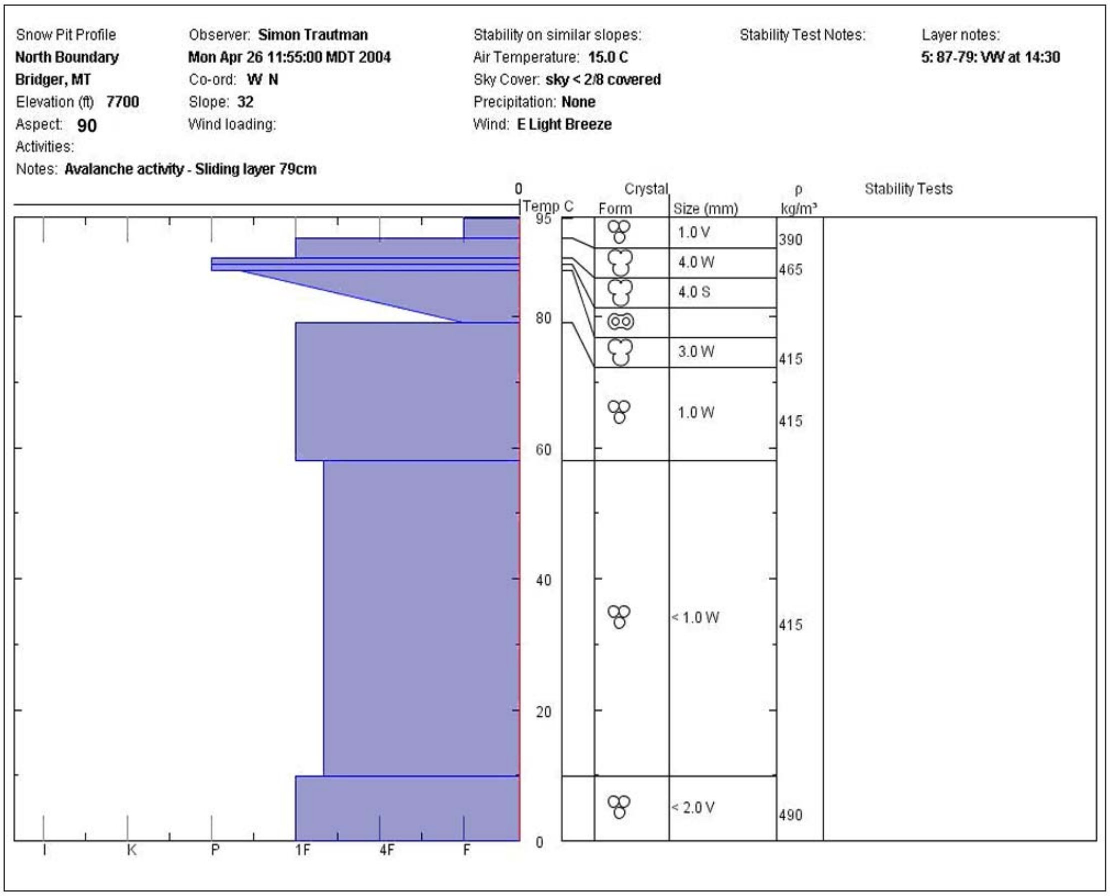


Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: No avalanche activity noted
Observer: Simon Trautman
Mon Apr 19 11:55:00 MDT 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 7.0 C
Sky Cover: sky 8/8 covered
Precipitation: None
Wind: SE Moderate
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
117
100
80
60
40
20
0
< 1.0 M
< 1.0 M
3.0 M
< 1.0 W
0.5 W
2.0 W
290
265
440
415
440
390
I
K
P
1F
4F
F

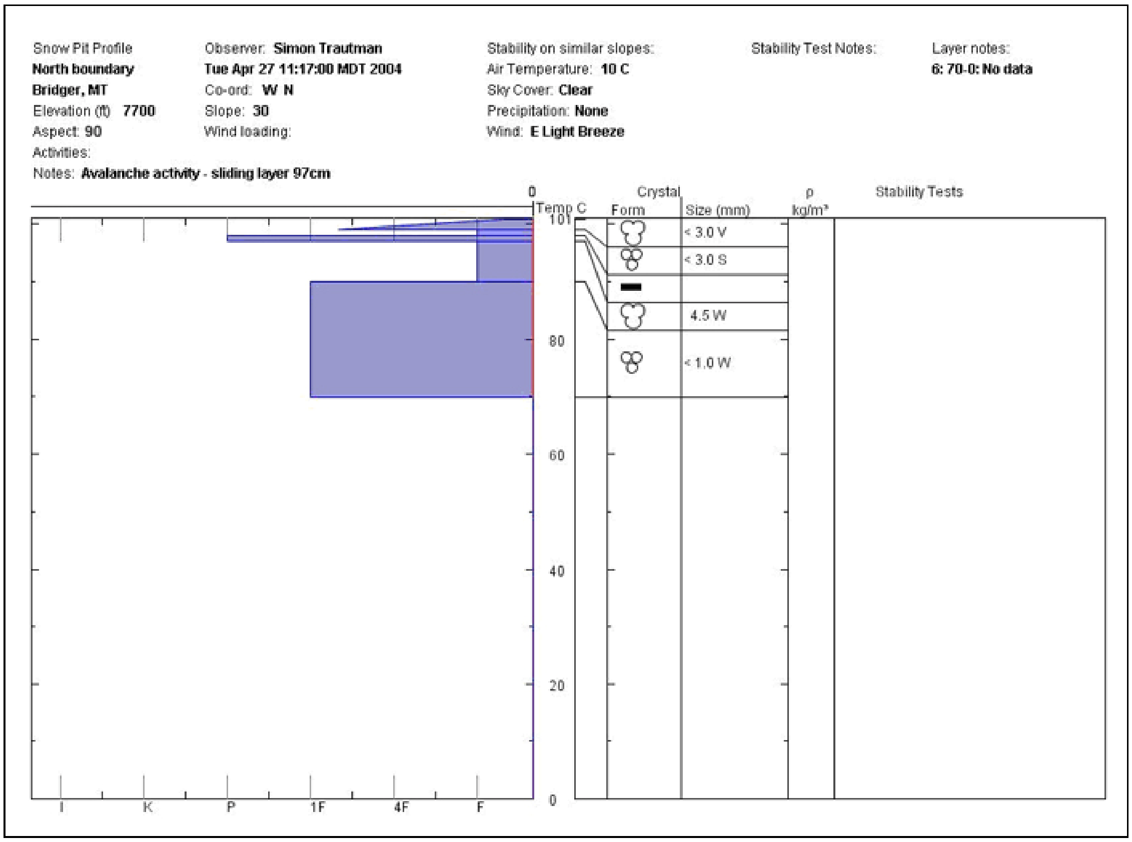
B18


Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: Avalanche activity - sliding layer 96cm
Observer: Simon Trautman
Fri Apr 23 11:55:00 MDT 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 15.0 C
Sky Cover: sky < 2/8 covered
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
1.0 V
1.0 S
3.0 W
2.5 W
1.0 W
1.0 S
0.5 W
< 0.5 W
2.0 V
305
345
345
415
425
515
440
100
80
60
40
20
0
I
K
P
1F
4F
F

B19


Snow Pit Profile
North Boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: Avalanche activity - Sliding layer 79cm
Observer: Simon Trautman
Mon Apr 26 11:55:00 MDT 2004
Co-ord: W N
Slope: 32
Wind loading:
Stability on similar slopes:
Air Temperature: 15.0 C
Sky Cover: sky < 2/8 covered
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
5: 87-79: VW at 14:30
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
Temp C
1.0 V
4.0 W
4.0 S
3.0 W
1.0 W
< 1.0 W
< 2.0 V
390
465
415
415
415
490
I
K
P
1F
4F
F

B20


Snow Pit Profile
North boundary
Bridger, MT
Elevation (ft) 7700
Aspect: 90
Activities:
Notes: Avalanche activity - sliding layer 97cm
Observer: Simon Trautman
Tue Apr 27 11:17:00 MDT 2004
Co-ord: W N
Slope: 30
Wind loading:
Stability on similar slopes:
Air Temperature: 10 C
Sky Cover: Clear
Precipitation: None
Wind: E Light Breeze
Stability Test Notes:
Layer notes:
6: 70-0: No data
Temp C
Crystal
Form
Size (mm)
ρ
kg/m³
Stability Tests
< 3.0 V
< 3.0 S
4.5 W
< 1.0 W
I
K
P
1F
4F
F

B21

APPENDIX C

SURFICIAL SHEAR STRENGTH DATA APRIL 2005 AND 2006

Shear Strength Data - Bridger Bowl, April 24, 2005

Time	Strength (kg)	τ_{250}	τ_{inf}
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
900	5	1961	1275
900	4.85	1902	1237
900	4	1569	1020
900	4.35	1706	1109
900	5	1961	1275
900	4	1569	1020
900	4.35	1706	1109
900	4.7	1844	1198
900	5	1961	1275
900	3.85	1510	982
900	4.7	1844	1198
900	4.75	1863	1211
1000	2.2	863	561
1000	2.2	863	561
1000	2.25	883	574
1000	2.4	941	612
1000	2.75	1079	701
1000	2.05	804	523
1000	2.3	902	586
1000	2.6	1020	663
1000	2.1	824	535
1000	2.75	1079	701
1000	2.15	843	548
1000	1.85	726	472
1100	1.75	686	446
1100	1.8	706	459
1100	1.85	726	472
1100	1.9	745	484
1100	2	785	510
1100	1.65	647	421
1100	2	785	510
1100	1.6	628	408

1100	1.6	628	408
1100	1.9	745	484
1100	2	785	510
1100	1.7	667	433
1200	2	785	510
1200	1.5	588	382
1200	1.7	667	433
1200	1.3	510	331
1200	1.4	549	357
1200	1.8	706	459
1200	1.25	490	319
1200	1.25	490	319
1200	2	785	510
1200	1.6	628	408
1200	1.75	686	446
1200	1.4	549	357
1300	1.5	588	382
1300	1.35	530	344
1300	1.35	530	344
1300	1.6	628	408
1300	1.65	647	421
1300	1.4	549	357
1300	1.4	549	357
1300	1.2	471	306
1300	1.35	530	344
1300	1.8	706	459
1300	1.3	510	331
1300	1.55	608	395
1400	1.1	431	280
1400	1.15	451	293
1400	1.2	471	306
1400	1.15	451	293
1400	1.25	490	319
1400	1.3	510	331
1400	1.75	686	446
1400	1.4	549	357
1400	1	392	255
1400	1.3	510	331
1400	1.4	549	357
1400	1.25	490	319
1500	1.3	510	331
1500	1.1	431	280
1500	0.85	333	217
1500	1	392	255
1500	1	392	255
1500	0.85	333	217
1500	1.1	431	280

1500	1	392	255
1500	1	392	255
1500	1.35	530	344
1500	1.15	451	293
1500	1.25	490	319
1600	0.9	353	229
1600	1.3	510	331
1600	1.1	431	280
1600	1.2	471	306
1600	1.25	490	319
1600	1.35	530	344
1600	1.25	490	319
1600	0.75	294	191
1600	1.5	588	382
1600	1.1	431	280
1600	1.1	431	280
1600	1.15	451	293
1700	1.25	490	319
1700	1.2	471	306
1700	1.2	471	306
1700	1.1	431	280
1700	0.85	333	217
1700	1.35	530	344
1700	1.05	412	268
1700	1.45	569	370
1700	1.1	431	280
1700	1.3	510	331
1700	1.15	451	293
1700	1.4	549	357
1800	1.8	706	459
1800	1.1	431	280
1800	1.2	471	306
1800	1.55	608	395
1800	1.45	569	370
1800	1.5	588	382
1800	2.25	883	574
1800	1.95	765	497
1800	1.5	588	382
1800	1.4	549	357
1800	1.45	569	370
1800	1.55	608	395

Shear Strength Data - Bridger Bowl, 25 April 2005

Time	Strength (kg)	τ_{250}	τ_{inf}
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
800	MAX	1961	1275
900	4.5	1636	1063
900	3.1	1127	732
900	3.25	1181	768
900	4.4	1599	1040
900	3	1090	709
900	4	1454	945
900	3.5	1272	827
900	4.05	1472	957
900	3.4	1236	803
900	3	1090	709
900	3	1090	709
900	3.95	1436	933
1000	2.9	1054	685
1000	2.35	854	555
1000	2.35	854	555
1000	2.55	927	602
1000	2.6	945	614
1000	2.1	763	496
1000	2.6	945	614
1000	2.1	763	496
1000	2.25	818	532
1000	1.8	654	425
1000	2.6	945	614
1000	2.3	836	543
1100	2.4	872	567
1100	1.8	654	425
1100	1.9	691	449
1100	2.05	745	484
1100	2.25	818	532
1100	1.75	636	413
1100	1.45	527	343

1100	1.8	654	425
1100	2.05	745	484
1100	1.85	672	437
1100	1.8	654	425
1100	1.65	600	390
1200	1.55	563	366
1200	1.55	563	366
1200	2	727	473
1200	1.8	654	425
1200	1.25	454	295
1200	1.4	509	331
1200	1.95	709	461
1200	1.3	473	307
1200	1.45	527	343
1200	1.5	545	354
1200	1.45	527	343
1200	1.6	582	378
1300	0.8	291	189
1300	1.25	454	295
1300	1.25	454	295
1300	1.1	400	260
1300	1.75	636	413
1300	1.75	636	413
1300	1.1	400	260
1300	1.35	491	319
1300	1.55	563	366
1300	1.7	618	402
1300	0.9	327	213
1300	1.1	400	260
1400	1.45	527	343
1400	1.45	527	343
1400	1.25	454	295
1400	1.3	473	307
1400	1.35	491	319
1400	1.45	527	343
1400	1.2	436	284
1400	1.4	509	331
1400	1.17	425	276
1400	1.35	491	319
1400	1.75	636	413
1400	1.5	545	354
1500	1.3	473	307
1500	1.1	400	260
1500	0.95	345	224
1500	1.05	382	248
1500	1.05	382	248
1500	1.2	436	284

1500	1.1	400	260
1500	1.1	400	260
1500	1.2	436	284
1500	1.35	491	319
1500	1	363	236
1500	1.3	473	307
1600	1.3	473	307
1600	1.3	473	307
1600	1.35	491	319
1600	1.2	436	284
1600	1.1	400	260
1600	1.4	509	331
1600	1.2	436	284
1600	1.6	582	378
1600	1.45	527	343
1600	1.5	545	354
1600	1.65	600	390
1600	1.5	545	354
1700	2.25	818	532
1700	2.35	854	555
1700	1.75	636	413
1700	2.1	763	496
1700	2.5	909	591
1700	1.65	600	390
1700	2.5	909	591
1700	2.3	836	543
1700	2.1	763	496
1700	1.95	709	461
1700	2.25	818	532
1700	1.6	582	378

Shear Strength Data - Bridger Bowl, 20 April 2006

Time	Strength (kg)	Strength (kg)	Strength (kg)	Average Strength	τ_{250}	τ_{inf}
9:55	MAX	MAX	4.55	4.55	1785.42	1160.523
10:06	MAX	MAX	4.45	4.45	1746.18	1135.017
10:11	MAX	MAX	MAX	MAX	2000	1300
10:21	3.81	MAX	MAX	?	?	#VALUE!
10:25	MAX	MAX	MAX	MAX	2000	1300
10:32	MAX	MAX	MAX	MAX	2000	1300
10:37	MAX	MAX	MAX	MAX	2000	1300
10:42	2.65	4.05	3.15	3.28	1288.38	837.447
10:43	2.65	2.76	3.15	2.85	1119.648	727.7712
10:49	2.61	2.1	2.25	2.32	910.368	591.7392
10:54	1.75	2	2.5	2.08	817.5	531.375
11:00	2.75	2.1	2.45	2.43	954.84	620.646
11:04	2.25	3	2.15	2.47	967.92	629.148
11:10	1.25	1.45	1.45	1.38	542.82	352.833
11:20	1.25	1.45	1.45	1.38	542.82	352.833
11:25	1.45	1.25	1.55	1.42	555.9	361.335
11:30	1.35	1.65	1.55	1.52	595.14	386.841
11:35	1.4	1.65	1.45	1.50	588.6	382.59
11:40	1.37	1.35	1.41	1.38	540.204	351.1326
11:45	1.13	1.16	1.19	1.16	455.184	295.8696
11:50	1.05	1.25	1.6	1.30	510.12	331.578
11:58	1.45	1.6	1.43	1.49	585.984	380.8896
12:05	1.23	1.18	1.2	1.20	472.188	306.9222
12:19	0.84	0.75	0.8	0.80	312.612	203.1978
12:24	0.95	1	0.98	0.98	383.244	249.1086
12:30	0.9	0.85	1.14	0.96	378.012	245.7078
12:35	1.1	0.8	1.05	0.98	385.86	250.809
12:45	1.1	0.8	1.05	0.98	385.86	250.809
12:50	1.12	1.03	1.24	1.13	443.412	288.2178
12:55	0.86	0.68	0.95	0.83	325.692	211.6998
13:00	1	0.98	0.85	0.94	370.164	240.6066
13:05	1	0.85	1.04	0.96	378.012	245.7078
13:12	1.05	1.03	1.25	1.11	435.564	283.1166
13:22	1.2	0.8	1.1	1.03	405.48	263.562
13:27	0.75	0.85	0.95	0.85	333.54	216.801
13:32	0.8	1	0.82	0.87	342.696	222.7524
13:38	1.1	1	1	1.03	405.48	263.562
13:45	1.05	0.8	0.98	0.94	370.164	240.6066
13:49	1.25	1.15	1.25	1.22	477.42	310.323
14:10	0.73	1.1	0.75	0.86	337.464	219.3516
14:17	0.8	0.8	0.96	0.85	334.848	217.6512
14:22	0.9	0.9	0.75	0.85	333.54	216.801

14:27	0.89	0.79	0.85	0.84	330.924	215.1006
14:37	1.1	1.12	1.4	1.21	473.496	307.7724
14:48	0.85	0.9	1.04	0.93	364.932	237.2058
14:55	0.9	0.8	0.89	0.86	338.772	220.2018
15:00	0.91	1.1	0.88	0.96	378.012	245.7078
15:05	1.1	1.25	1.1	1.15	451.26	293.319
15:11	0.35	1.3	1.2	0.95	372.78	242.307
15:21	0.9	1.05	1.05	1.00	392.4	255.06
15:27	0.93	0.9	0.97	0.93	366.24	238.056
15:32	0.89	0.85	0.95	0.90	351.852	228.7038
15:37	1.04	0.98	1	1.01	395.016	256.7604
15:43	1.1	0.9	1	1.00	392.4	255.06
16:14	0.9	1.02	1.13	1.02	398.94	259.311
16:20	0.91	1.05	1.08	1.01	397.632	258.4608
16:26	0.97	1.03	1.2	1.07	418.56	272.064
16:31	1.3	1.35	1.19	1.28	502.272	326.4768
16:36	1.25	1.4	1.1	1.25	490.5	318.825
16:42	0.9	1.1	1	1.00	392.4	255.06
16:47	1	1.32	1.25	1.19	466.956	303.5214
16:52	1.15	1	1.1	1.08	425.1	276.315
16:57	1.3	1.33	1.54	1.39	545.436	354.5334
17:02	1.54	1.2	1.24	1.33	520.584	338.3796
17:13	1.25	1.22	1.31	1.26	494.424	321.3756
17:18	1.15	1.1	1.25	1.17	457.8	297.57
17:22	1.3	1.3	1.2	1.27	497.04	323.076
17:26	1.25	1.6	1.55	1.47	575.52	374.088
17:32	1.56	1.5	1.57	1.54	605.604	393.6426
18:06	1.9	1.7	1.5	1.70	667.08	433.602
18:11	2	2.01	1.7	1.90	746.868	485.4642

Shear Strength Data - Moonlight Basin, 22 April 2006

Time	Strength (kg)	τ_{250}	τ_{inf}
945	1.75	636	413
945	2.9	1054	685
945	2.3	836	543
945	2.8	1018	662
945	2.2	800	520
945	1.8	654	425
945	2.85	1036	673
945	2.3	836	543
945	2.75	1000	650
945	2.15	781	508
1045	1.1	400	260
1045	1.18	429	279
1045	1.14	414	269
1045	1.28	465	302
1045	1.16	422	274
1045	1.07	389	253
1045	1.2	436	284
1045	1.26	458	298
1045	1.65	600	390
1045	1.4	509	331
1140	1.05	382	248
1140	0.9	327	213
1140	0.95	345	224
1140	0.85	309	201
1140	0.8	291	189
1140	0.85	309	201
1140	0.98	356	232
1140	0.88	320	208
1140	1.01	367	239
1140	0.75	273	177
1245	0.8	291	189
1245	0.7	254	165
1245	0.78	284	184
1245	0.85	309	201
1245	0.7	254	165
1245	0.73	265	172
1245	0.9	327	213
1245	0.95	345	224
1245	0.78	284	184
1245	0.95	345	224
1345	0.82	298	194
1345	0.85	309	201
1345	0.82	298	194

1345	0.9	327	213
1345	0.84	305	198
1345	0.7	254	165
1345	0.68	247	161
1345	0.8	291	189
1345	0.75	273	177
1345	0.78	284	184
1455	0.65	236	154
1455	0.78	284	184
1455	0.82	298	194
1455	0.84	305	198
1455	0.7	254	165
1455	0.9	327	213
1455	0.95	345	224
1455	0.75	273	177
1455	0.85	309	201
1455	0.85	309	201
1605	0.95	345	224
1605	1.25	454	295
1605	0.8	291	189
1605	0.75	273	177
1605	0.75	273	177
1605	0.83	302	196
1605	0.93	338	220
1605	0.93	338	220
1605	1	363	236
1605	0.9	327	213
1703	1.5	545	354
1703	1.05	382	248
1703	1	363	236
1703	1.08	393	255
1703	1.05	382	248
1703	1.37	498	324
1703	1.1	400	260
1703	1.08	393	255
1703	1.15	418	272
1703	1.12	407	265
1747	1.35	491	319
1747	1.34	487	317
1747	1.25	454	295
1747	1.55	563	366
1747	1.1	400	260
1747	1.45	527	343
1747	1.3	473	307
1747	1.32	480	312
1747	1.31	476	309
1747	1.2	436	284