FORMATION OF REFROZEN SNOWPACK LAYERS AND THEIR ROLE IN SLAB AVALANCHE RELEASE

Bruce Jamieson

Received 16 June 2005; accepted 12 April 2006; published 17 June 2006.

[1] In a variety of snow climates, numerous slab avalanches release over crusts consisting of refrozen snow. Slab avalanches sometimes release in weak layers of faceted crystals that developed while underlying wet layers froze into crusts, often within a day. Weak layers of faceted crystals can also develop when less permeable and more conductive crusts alter the temperature and vapor pressure gradients. These processes create interfaces where differences in grain radii can contribute to weak bonding.

In western Canada, layers of faceted crystals (facet layers) on crusts are most common in early and late winter when thaws and rain are more frequent. Also, thin facet layers occur in spring when surface melting by solar radiation is common. Shear strength tests on facet layers show an initial strength loss during faceting followed by a slow strength increase. The spatial distribution of poorly bonded crusts can be interpreted from the interaction of terrain and meteorology that caused the antecedent wet layer.


1. INTRODUCTION

[2] Snow avalanches interrupt transportation and energy corridors, destroy buildings, and kill in excess of 150 people most years [Meister, 2001]. Most of these impacts are due to slab avalanches (Figure 1), which release cohesive snowpack layers when underlying weak layers or interfaces fail. In contrast to slab avalanches, loose snow avalanches start in relatively cohesionless snow and are typically smaller and less destructive. The snowpack layers essential to slab avalanches are formed by atmospheric and snowpack processes. Forecasting when and where destructive avalanches are likely, which is essential to reducing these impacts, is challenged by a limited understanding of these processes. This review focuses on the role of refrozen layers in slab avalanche formation.

[3] Wet layers on the snow surface that freeze and become melt-freeze crusts form the bed surface for many slab avalanches, including many difficult-to-forecast avalanches. Seligman [1936, pp. 308–310, 387] and Atwater [1954] stated that snow often fails to bond to a crust and thus can release slab avalanches. McClung and Schaerer [1993, p. 59] stated that “weak bonding of snow above crusts is the most important feature of crusts with respect to avalanche formation.” Because crusts are usually stiffer and harder than the overlying snow, they concentrate shear stress within the sloping snowpack and hence contribute to shear failure at the upper boundary of the crust [Schweizer and Jamieson, 2001, 2003]. Melt-freeze crusts can occur almost anywhere in mountainous snowpacks, but they are most often observed in maritime mountain ranges and early or late in the winter season when rain and/or surface melt are most likely. They have been reported and have been recognized as significant contributing factors to avalanche release in different areas [e.g., Barton and Wright, 2000, pp. 50–51; Irwin, 2004; Moore, 1982]. While the phenomenon is widespread in many mountain ranges, this review draws on observations in the Columbia Mountains of Canada (Figure 2) where many quantitative observations of this problem have been made. The Columbia Mountains are not maritime, but their climatic conditions are similar in many ways to maritime mountains such as the Cascade Mountains [Hägeli and McClung, 2003].

[4] On the basis of the winters from 1980/1981 to 2000/2001 at Mount Fidelity in the Columbia Mountains, Hägeli and McClung [2003] showed that rain occurs most years in October, November, and April and in March during a third of the winters. They report numerous rain events followed by dry snow that resulted in depth-averaged temperature gradients in the overlying dry snow that were sufficient for forming weak layers of faceted crystals (facets) and 18 events over the 21 winters with considerable cooling during or after the dry snowfall, which is very favorable to forming weak layers of facets in the dry snow above the wet layer. For spontaneous avalanching in the Columbia Mountains,
Hägeli and McClung [2003] reported that 17% of the releases were due to layers of facets on a crust and 7% were on crusts without a weak crystal type reported on the crust.

Jamieson et al. [2001] gave two case histories of facets on crusts that likely formed when dry snow fell on rain-wetted snow in November 1996 and November 1997. The first layer was widespread, releasing fatal slab avalanches in the Coast Mountains, the Columbia Mountains, and the Rocky Mountains of Canada. In the Columbia Mountains this layer of facets on a crust released over 700 large slab avalanches.

An understanding of the cause of the poor bond between a crust and the overlying snow is important to recognizing the distribution of poorly bonded crusts over terrain and hence to avalanche forecasting. Section 2 of this paper reviews the causes of melt-freeze crust formation including the effects of the terrain variables: aspect, elevation, and inclination. Detailed observations from recent slab avalanches on crusts in section 3 show that persistent grain types, notably faceted crystals and surface hoar, are found on almost all of the crusts in which the release layer on the crust was buried more than 3 days. Surface hoar layers consist of frost crystals, often 2 mm to over 10 mm in length. Faceted crystals, or facets, are angular crystals often with flat faces that result from kinetic growth in dry snow where the temperature gradient is sufficient. Faceted crystals are typically 0.3–3 mm in size. The advanced products of kinetic growth known as depth hoar crystals are typically larger. In contrast to faceting, grains in dry snow become rounder, i.e., evolve toward a spherical equilibrium form, when the temperature gradient is insufficient for faceting. During rounding, water vapor is deposited at the concave surfaces of the bonds [Colbeck, 1987] promoting bonding and strength gain faster than during faceting. Consequently, layers of faceted crystals are commonly weak snowpack layers, whereas layers of rounded grains rarely form weak layers.

Since most poorly bonded crusts in the Columbia Mountains, as summarized in section 3, involve faceted crystals above or below the crust, the various formative processes for faceting near crusts are reviewed in section 4. One of these processes, faceting of dry snow above a wet layer, forms an important portion of poorly bonded crusts in the Columbia Mountains. This paper includes an example of the spatially variable stability of a slab on a crust and proposes ideas for anticipating the formation and distribution of poorly bonded crusts in section 5.

A weak bond can occur between a crust and the snow above a crust, in the softer laminations within a crust, or below a crust (Figure 3). This paper focuses on the weak bond at the upper boundary of a crust, which is commonly the failure layer or interface for slab avalanches involving crusts. The difference in grain sizes between the crust and the facets that grow just over the crust can contribute to the poor bond between these layers. Colbeck [2001] showed that, if the ratio of grain sizes is greater than 1.57, the contact geometry is unfavorable to bonding.
In this review, all wet layers that freeze are referred to as crusts although these may include grain types 6a, 6b, 8a, 9b, 9c, and 9e [Colbeck et al., 1990]. The snow surface can become wet because of rain, wet snowfall, or if sufficient net energy flows into the snow causes melting at or near the snow surface. The crusts due to rain and sunshine observed in western Canada are rarely thin and transparent as required for the International Classification for Seasonal Snow on the Ground definition of rain crusts and sun crusts [Colbeck et al., 1990]. These are recorded as melt-freeze crusts (type 9e (CRmfc)); however, the labels rain crust and sun crust are used to identify the cause whenever weather data or field observations support the distinction.

2. FORMATION OF WET LAYERS AND CRUSTS

The sources of incoming thermal energy that form wet layers (rain, sensible heat, and solar radiation) are discussed separately. While one source often dominates heat input to the snow surface, secondary sources of incoming energy can be substantial and add complexity to the formation of wet layers at or near the snow surface. In addition to the heat sources discussed subsequently, condensation of water vapor or the deposition of water vapor as rime or surface hoar can add heat to the snow surface. By identifying the cause of the surface melting, as described in this section and summarized in Table 1, the distribution of crusts over the terrain can be better anticipated. If the cause of the antecedent wet layer is not considered, then the location of buried crusts, including poorly bonded crusts, may be difficult to anticipate as shown in the last row of Table 1.

2.1. Rain and Resulting Rain Crusts

Once wetted by rain, the snow surface can freeze into a crust. The freezing can occur before or after burial by subsequent snowfall. Using the Columbia Mountains from 1966 to 1986 to illustrate rain on snow, Figure 4 shows that at 1905 m on Mount Fidelity, rain usually occurred in all winter months, although the amount averaged 5.1 mm in November and was usually less than 3 mm per month in December through March, with about 12 mm in April and 50–60 mm in October and May.

However, since the mid 1990s, rain in several November [Jamieson et al., 2001; Hägeli and McClung, 2003] resulted in prominent rain crusts. Glacier National Park weather records show the November average from 1987 to 2004 at the same site was 15.8 mm, 3 times higher than in the preceding 20 years. If this warming trend continues, mountains which have historically been colder and drier will experience more refrozen crusts of the type reviewed here.

In November when incoming solar radiation is limited (Figure 4), wetting of the snow surface in the Columbia Mountains is typically due to rain. After being infrequent in December through March in the Columbia Mountains, rain becomes more abundant in April (Figure 4);

![Figure 2. Map of Columbia Mountains in western Canada showing study sites at Mount Fidelity in Glacier National Park and Mount St. Anne near Blue River, British Columbia, Canada.](image)

![Figure 3. Translucent profile of snowpack layers including a laminated crust, consisting of two bright layers between A and B. The poor bond can occur at the upper boundary of a crust (labeled A), which is the focus of this paper, in softer snow laminated within a crust (labeled I), or below a crust (labeled B). For scale each bright layer is approximately 2 cm thick. Reprinted from Jamieson [2004b] with permission of the Canadian Avalanche Association.](image)
hence rain crusts become more common, and some are thick. Elevation is not the only major terrain factor that affects the distribution of a rain crust. Since liquid water often penetrates less deeply on steep slopes than on gentle and flat slopes [Wankiewicz, 1979], the resulting crust may be deeper on less steep slopes. Also, during rain storms the windward slopes may receive more rain per unit area than leeward slopes, resulting in wetter or thicker wet layers on windward slopes than on leeward slopes.

2.2. Melting of the Snow Surface by Sensible Heat and Resulting “Temperature” Crusts

During October the daily maximum air temperature is often above freezing, and sensible heat is likely to melt the snow surface (Figure 4). However, in the Columbia Mountains the October snowpack is often thin or not contiguous. Ground roughness features often extend through the resulting crusts so that the crusts play a role in slab avalanche formation less frequently. From November to February the air temperature is typically below freezing, and the sensible heat transfer from the air is often insufficient for melting the snow surface. In March and April the number of days with maximum air temperatures above freezing increases, and surface melting by sensible heat transfer becomes more common. Once refrozen, field workers often refer to these layers as “temperature crusts.”

Usually, the air is warmer at lower elevations, and hence temperature crusts will be observed at lower elevations. The exchange of sensible heat increases with wind speed [Brun et al., 1989; Fierz et al., 2003], and thus wind can cause more surface melting on slopes that are locally windward. Slope angle will have little effect on the melting of the snow surface by sensible heat and therefore on the resulting crusts.

2.3. Melting by Solar Radiation and Resulting Sun Crusts

Approximately 85–95% of solar radiation (0.3–3 μm) reflects off the surface of fresh snow [Male and Gray, 1980]. Nevertheless, absorbed solar radiation can exceed outgoing long-wave radiation and cause surface melting [e.g., Ozeki et al., 1995]. The increase in hours of bright sunshine and solar radiation combined with the warm air temperature makes melting of the snow surface on sunny slopes likely in March and April (Figure 4). In late winter and spring the solar radiation that is absorbed, especially on sunny slopes tilted into the Sun, can be sufficient to melt the snow surface. Consequently, sun crusts are often spatially variable, conspicuous on steep slopes facing southeast to southwest, thinning where the slope angle is less or the aspect is less southerly, and often absent on north facing slopes, especially steep north facing slopes. These effects of aspect and inclination on sun crust formation exist over scales ranging from meters to many kilometers.

3. POORLY BONDED CRUSTS

Jamieson [2004a] summarized 70 snow profiles in the Columbia Mountains from 1990 to 2004 that were observed near slab avalanches or “whumpfs,” where a whumpf is a fracture in a weak layer under a snow slab that did not release an avalanche [Johnson et al., 2001], usually because of insufficiently steep terrain. Each of these slab releases had crusts as the bed surface, implying that the failure occurred in the overlying weak layer and likely at the interface of the crust and weak layer where shear stress was concentrated [Schweizer et al., 2003]. The grain type for the

![Figure 4](image-url)
weak layers was classified according to Colbeck et al. [1990]: 39 layers of faceted crystals (FC), 23 of frost crystals known as surface hoar (SH), three of rounded grains, three of decomposing and fragmented particles (DF), one of precipitation particles (PP), and one consisting of an advanced form of facets called depth hoar (DH). All of the three weak layers with a primary grain type of rounded grains had facets as their secondary grain type. Including these three layers with the other layers of facets or depth hoar, these kinetic growth forms were the most common primary grain types in weak layers and were found in at least 60% of the weak layers that released slabs on crusts.

[19] The age of these layers at the time of slab release, i.e., the number of days since they were buried, is plotted in Figure 5. Nonpersistent weak layers consisting of PP or DF are only a few days old when they release slabs on crusts [Jamieson, 2004a]. The interquartile range of the ages of surface hoar layers ranges from 8 to 17 days, compared to 15 to 27 days for layers of facets and depth hoar when they release slabs on crusts. Clearly, when overlying crusts, layers of facets and depth hoar are prone to releasing avalanches for longer than layers of surface hoar, both of which are prone to releasing avalanches longer than layers of PP and DF particles. In this paper, a poorly bonded crust is defined as a refrozen layer directly overlain by a persistent weak layer consisting of surface hoar, facets, or depth hoar [Jamieson and Johnston, 1992].

[20] The particle size also affects the persistence of surface hoar layers. However, to access a larger data set than shown in Figure 5, Jamieson [2004a] reported the age of surface hoar layers versus the average particle size when they fracture consistently in two or more adjacent compression tests within an hour or so. In each compression test the surface of 0.3 m by 0.3 m column of the snowpack is tapped until fractures occur in weak layers within the column [Greene et al., 2004, pp. 45–47; Canadian Avalanche Association, 2002, pp. 32–34]. The median age increases from 14 to 19 days as the average particle size increases from less than 2 mm to 7–8 mm. This is in spite of a tendency for the observed size of surface hoar particles to decrease over time because of fragmentation when manually extracted from the snowpack [Jamieson and Schweizer, 2000]. The median age declines for layers of crystals larger than 10 mm [Jamieson, 2004a], likely because such large crystals are often unbroken when extracted from the snowpack and are therefore from recently buried layers. When buried surface hoar overlies a crust, the surface hoar crystals will be slow to penetrate the crust, slowing an important mechanism for strength gain of a surface hoar layer [Jamieson and Schweizer, 2000]. Surface hoar layers consist of crystals, most of which extend from the top to the bottom of the layer, so the layer thickness and the size of the particles extracted from the layers are strongly correlated [Jamieson and Schweizer, 2000].

Figure 5. Number of days since burial (age) of release layers by grain type when they released slabs on crusts. The diamond indicates the median age of the layers. The box and whiskers indicate the interquartile range and full range, respectively, of layer age. Abbreviations are DF, decomposing and fragmented precipitation particles; DH, depth hoar; FC, solid faceted crystals; PP, precipitation particles; and SH, surface hoar. Reprinted from Jamieson [2004a] with permission of the Canadian Avalanche Association.

Figure 6. Age since burial of a facet layer when it fractured consistently in adjacent compression tests. For each range of particle size, e.g., >0.7 mm and ≤1.3 mm, the whisker shows the first and third quartiles of age. Squares and circles indicate the median ages of facets (types 4a and 4b) or depth hoar (type 5) of average size less than or equal to 0.7 mm [Jamieson, 2004a], which suggests that they do not persist as weak layers. Their median persistence increases to 67 days for crystals larger than 2.3 mm. Since small grains bond much faster than larger grains [Colbeck, 1998], layers consisting of large grains are expected to gain strength slower than layers of small grains (e.g., ≤0.7 mm). Also,
the spacing between larger grains dictates fewer bonds per unit area when the grains are bonded directly to a crust. For grain sizes between 0.7 and 1.7 mm, Figure 6 shows little difference in the persistence of facets and depth hoar compared to facets that exhibit signs of rounding (type 4c).

[22] Of 38 facet layers with measured thickness that released slab avalanches on crusts, 39% and 58% of the facet layers were less than 10 and 20 mm thick, respectively [Jamieson, 2004a], indicating the importance of thin facet layers for slab avalanche release in the Columbia Mountains and likely in other mountain ranges with a maritime influence. However, the data set of facet layers of slab releases on crusts was too small to show a change in thickness over the winter. To assess such a trend, Jamieson [2004a] analyzed the thicknesses of weak layers of facets on crusts that fractured in almost 300 sets of compression tests, in which each set of approximately three tests was at a different location and/or day. For weak layers less than or equal to 10 mm and 20 mm thick, the percentages were 22% and 34%, respectively, again showing the importance of thin weak layers of faceted crystals on crusts. While the percentage of facet layers on crusts more than 50 mm thick did not change systematically over the winter, the percentage of such layers less than or equal to 5 mm thickness increased monotonically from 0% to 54% from November to April. This is likely because thin wet layers caused by solar radiation become more frequent as the winter progresses and the limited latent heat in such layers is only favorable to growing thin layers of facets at the base of overlying dry snow.

### 4. FORMATION OF FACETS NEAR CRUSTS

[23] The basic ideas of kinetic growth (faceting) are outlined in section 4.1. The following four sections review faceting of thick layers (section 4.2), near-surface faceting (section 4.3), faceting of dry snow near crusts (section 4.4), and faceting of dry snow above wet layers (section 4.5). These processes are summarized in Table 2. While distinguishing the processes is useful for analysis, the processes often combine in the natural snowpack to promote faceting. Because spatial distribution of dry-on-wet faceting is important to avalanche forecasting, it is reviewed separately in section 4.5.

#### 4.1. Kinetic Growth of Crystals (Faceting)

[24] Faceted crystals and the advanced form of facets called depth hoar [Colbeck et al., 1990] form by the kinetic growth of crystals because of a temperature gradient moving water vapor across the pore space from grain to grain. Kinetic growth favors the growth of large grains at the expense of small grains. Also, kinetic growth does not, except perhaps at high snow density, favor bond growth. Hence during faceting, layers may lose strength or gain strength more slowly than during rounding.

[25] In this paper, \(dT/dz\) is defined as the temperature gradient on the scale of grains perpendicular to the slope, and \(\Delta T_{10}\) is defined as the temperature difference over 10 cm. Avalanche workers commonly measure \(\Delta T_{10}\) vertically during manual snow profiles [Greene et al., 2004, p. 28], even though the vapor flow is probably greatest perpendicular to the slope. The threshold magnitude of the temperature gradient for faceting, \(TG_F\), depends on temperature, snow density, and grain/pore size and is typically \(10^0 - 20^0\) C m\(^{-1}\) [Miller et al., 2003]. Faceting is faster for temperatures closer to freezing, lower densities, and larger pores/grains.

[26] Formation of recognizable facets in the snowpack is expected where \(dT/dz > TG_F\) is sustained for sufficient time. While a few hours are sufficient in low-density snow for gradients greater than or equal to 100 C m\(^{-1}\) [Fukuzawa and Akitaya, 1993; Birkeland et al., 1998; Jamieson and van Herwijnen, 2002], several days or longer are required for gradients close to 10 C m\(^{-1}\) [LaChapelle and Armstrong, 1977]. The time required also increases with

### Table 2. Processes for Formation of Facets on a Buried Crust

<table>
<thead>
<tr>
<th>Type of Faceting</th>
<th>Cause</th>
<th>Location of Most Advanced Faceting</th>
<th>Typical Time to First Observation of Facets</th>
<th>Selected References</th>
</tr>
</thead>
<tbody>
<tr>
<td>DW</td>
<td>Wet layer sustains (dT/dz &gt; TG_F) and warm subfreezing temperature at the base of the overlying dry snow.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>interface at base of dry layer</td>
<td>hours</td>
<td>Birkeland et al. [1998] and Colbeck and Jamieson [2001]</td>
</tr>
<tr>
<td>DC</td>
<td></td>
<td>uncertain</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>interface at base of low-density layer</td>
<td>probably days (unknown)</td>
<td>Adams and Brown [1983] and Colbeck [1991]</td>
</tr>
<tr>
<td>NS</td>
<td></td>
<td>near snow surface</td>
<td>hours</td>
<td>Birkeland [1998] and Birkeland et al. [1998]</td>
</tr>
<tr>
<td>NS</td>
<td></td>
<td>near snow surface</td>
<td>hours</td>
<td></td>
</tr>
<tr>
<td>TG</td>
<td></td>
<td>in less dense or warmer layers</td>
<td>days</td>
<td>Akitaya [1974] and Armstrong [1985]</td>
</tr>
</tbody>
</table>

*Abbreviations and text locations are DW, dry on wet (section 4.5); DC, dry on crust (section 4.4); NS, near surface (section 4.3); and TG, thick layers (section 4.2). Processes are not inherently near-crust processes, although they may cause faceting near crusts.*
initial snow density [Akitaya, 1974; Marbouty, 1980; Miller et al., 2003]. Also, fresh dry snow has a very high specific surface area, which probably leads to more rapid metamorphism into faceted crystals than would occur if the starting snow particles were initially rounded.

4.2. Formation of Thick Layers of Facets

[27] In many areas with a thin snowpack and cool surface temperatures, such as polar regions or the Rocky Mountains of North America, the temperature gradient is sufficient for faceting \( (\Delta T_{10} > T_{G}) \) throughout the snowpack or at least in thick layers of the snowpack [LaChapelle and Armstrong, 1977]. The thick layers of large faceted or depth hoar crystals are usually weaker than layers of rounded grains or crusts. If there are crusts in the snowpack when the temperature gradient is sufficient, facets will be first observed in the lower-density layers adjacent to the crusts because kinetic growth is faster in lower-density layers. Subsequently, the crust will become faceted and can disappear if the gradient is sustained for sufficient time, typically weeks or months [e.g., Colbeck, 1991]. Once the metamorphosed crust is no longer stiffer than adjacent layers, it ceases to concentrate shear stress at its upper and lower boundaries, and in this respect the crust no longer contributes to slab failure. However, other interfaces, including the ground-snow interface, may concentrate shear stress and/or provide a bed surface for avalanche motion.

4.3. Near-Surface Faceting

[28] Under relatively clear skies the upper part of the snowpack, especially the top few centimeters, is subjected to strong temperature gradients as the energy balance at the snow surface is either net incoming in the daytime to net outgoing at night [Birkeland, 1998; Birkeland et al., 1998]. While the resulting faceting, called diurnal recrystalization, is often concentrated in the upper few centimeters, at temperate latitudes it can extend to approximately 15 cm depth and form facets in the snow on crusts close to the snow surface [Birkeland, 1998].

[29] LaChapelle [1970], Armstrong [1985], and Birkeland [1998] also described a form of faceting called radiation recrystallization that is most common at, but not exclusive to, low latitudes and high altitudes. On sunny slopes, solar radiation penetrates and warms the upper snowpack, while long-wave radiation cools the surface. Maximum warming, sometimes resulting in melting, occurs a few centimeters below the snow surface resulting in strong near-surface temperature gradients sufficient for rapid faceting. This process can either form the crust and the overlying facets [Fukuzawa and Akitaya, 1993], or as with diurnal near-surface faceting the strong gradient can penetrate the low-density snow overlying near-surface crusts that already existed. Birkeland [1998] reported only occasional observations of radiation recrystallization in southwest Montana, and the process should be even less frequent in more northerly areas.

4.4. Faceting of Dry Snow Near Crusts

[30] Field workers often observe a layer of facets directly above or below a crust [e.g., Fierz, 1998; Greene and Johnson, 2002], even at the South Pole. Over a winter, but prior to the snow becoming isothermal in the Rocky Mountains of Montana, Adams and Brown [1989a] reported an apparent loss of strength and mass in low-density snow layers adjacent to dense layers such as a melt-freeze crust, even when the temperature gradient in the region was “insignificant.” In Antarctica, Alley [1988] showed that stratified firm was the result of faceting of low-density layers, while the denser layers resisted faceting.

[31] There have been theoretical approaches to explain the field observations of faceting above, below, and within a crust even when the magnitude of the depth-averaged temperature gradient is insufficient for faceting \( (\Delta T_{10} < T_{G}) \). On the basis of two-phase mixture theory, Adams and Brown [1989b] modeled mass transport from a low-density snow to an adjacent denser layer (e.g., crust). This would decrease the density of the low-density layer above a crust and increase the density of the underlying crust, sustaining a stratified snowpack of low- and high-density layers. In contrast to the theory of Adams and Brown [1983], Colbeck [1991] argued that the density change at the upper boundary of a crust would contribute to evaporation, potentially resulting in faceting. This is consistent with reports of faceting at the upper boundary of crusts in a maritime snowpack [e.g., Moore, 1982] where the magnitude of the depth-averaged temperature gradient is usually low and snow temperature is only a few degrees below freezing. However, laboratory observations of grain formation along with density and temperature measurements under steady thermal conditions are needed to test the theory of Adams and Brown [1989b] and ideas of Colbeck [1991]. Verification in the field will be difficult because near-surface temperature gradients vary within hours and typically change sign twice daily, but verification might be possible for a deeply buried crust.

[32] Condensation of the upward flowing water vapor at the lower surface of less permeable layers [Seligman, 1936, p. 70; Adams and Brown, 1983; Colbeck, 1991] can contribute to faceting below crusts [de Quervain, 1958; Adams and Brown, 1982; Moore, 1982; Palais, 1984; Alley, 1988; Fierz, 1998; Greene and Johnson, 2002]. In laboratory studies where the magnitude of the depth-averaged temperature gradients exceeded 30°C m⁻¹, Adams and Brown [1982] measured an increased temperature gradient and a strength loss over 4 to 9 days in low-density layers below hard layers, but a similar change was not observed above the crusts.

[33] In the Coast Mountains of North America, Moore [1982], C. Stethem (personal communication, 1999), and J. Hetherington (personal communication, 2004) reported a loss of hardness of crusts because of faceting when the snow temperature was often a few degrees below freezing. This may be the result of the below-crust faceting combined with above-crust faceting acting on the slight differences in density and permeability within a visually uniform crust (C. Stethem, personal communication, 1999). Once faceting has started, the characteristic small bonds and slow densification will limit conductivity [Adams and Brown, 1983],
increase vapor flux [Colbeck, 1991], and potentially contribute to further faceting.

[34] On the basis of observations in the Coast Mountains of western Canada, J. Hetherington (personal communication, 2004) reported that continuous areas of facets within rain crusts were often less than 1 m in length, presumably between percolation channels that froze. Occasionally, dry slab avalanches are reported on layers of facets within crusts (e.g., Figure 1), suggesting that areas of continuous faceting are sometimes large enough for fracture propagation.

[35] The phenomenon of within-crust faceting is usually reported in the upper meter of the snowpack; however, it is unclear if this is because manual snow profiles of depth 1–1.5 m are more frequent than deeper profiles or because the faceting process is inhibited at greater depths, perhaps because snow density increases with depth.

[36] Although there is no verified explanation for faceting above crusts when $|\Delta T_{10}| < TG_F$ (dry-on-crust (DC) faceting), such a process likely exists based on the unpublished observations of field workers and Moore [1982]. The limiting values of density, grain size, temperature, and temperature gradient for DC faceting are also unclear. In the deep snowpack of the Columbia Mountains, Jamieson et al. [2001] provided observations over the winter of 1996/1997 in which rounding of 1.5–2.5 mm facets above a crust were not observed although $|\Delta T_{10}|$ decreased from 7°C m$^{-1}$ to 1°C m$^{-1}$ over 3 months. Alternatively, as shown in section 4.5, there was a situation where rounding of 0.3–0.5 mm facets occurred above a crust while $|\Delta T_{10}|$ decreased from 5°C m$^{-1}$ to 2°C m$^{-1}$ between 17 and 20 January 2004.

4.5. Faceting of Dry Snow Above Wet Layers (Dry-on-Wet (DW) Faceting)

[37] Birkeland [1998] referred to faceting of dry snow above wet layers as “melt-layer recrystalization”; however, the terms “dry-on-wet faceting” or “near-crust faceting” are clearer because they specify that the metamorphosis of grains occurs in the dry snow above or adjacent to the wet layer, respectively. Also, while the term “recrystalization” is descriptive on the grain scale, it does not distinguish between faceting and rounding on the molecular scale since both of these processes involve the deposition of molecules that become aligned to the ice lattice.

[38] As an example of dry over wet snow layers favorable to DW faceting, Figure 7 shows a cross section of the snowpack from Jamieson and Langevin [2004]. Small facets developed in the dry snow just above the interface in a day.

[39] In a field study, Fukuzawa and Akitaya [1993] observed strong temperature gradients in a 2-cm-thick dry layer of density 80 kg m$^{-3}$ overlying a coarse-grained wet layer of density 320 kg m$^{-3}$ (Figure 8). At 1620 LT, 2 hours after the start of observations, the top of the wet snow layer froze. Freezing penetrated an additional 6 cm into the initially wet layer in the following 4 hours. Owing to radiant cooling of the snow surface and the latent heat flowing from the underlying wet snow, the temperature gradient in the top 2 cm of dry low-density snow averaged 143°C m$^{-1}$ overnight. By midnight, 0.3 mm facets were observed 1 cm beneath the snow surface.

[40] In two field studies in which snowfall, fog, and overcast skies limited radiant cooling of the snow surface, Jamieson and Langevin [2004] were able to isolate the effects of DW faceting and track the change in shear strength at the interface between the freezing wet layer and the overlying dry snow. The second of the two field studies is summarized in Figures 9 and 10. At 1600 m on Mount St. Anne at noon on 15 January 2004, 4 cm of low-density dry snow overlay 6 cm of moist snow. Additional snow fell in the afternoon and evening, and the magnitude of the temperature gradient reached a maximum of 91°C m$^{-1}$. The top of the initially moist layer froze 21 to 26 hours

![Figure 7](image-url) Vertical section of the upper snowpack showing the wet layer and overlying dry snow on 14 March 2003 on Mount Fidelity. For scale the wet layer was 1.8 cm thick. Reprinted from Jamieson and Langevin [2004] with permission of the Canadian Avalanche Association.
after it was buried. In the first 24 hours some 0.5 mm facets formed where the dry snow lay on the freezing wet layer. Also, the shear strength measured with a shear frame [see, e.g., Perla and Beck, 1983; Sommerfeld, 1984] dropped 50%. This decrease was attributed partly to a shear stress concentration resulting from the freezing and stiffening of the wet layer and partly to the formation of facets at the interface. The strength increased slightly in the following day after the wet layer froze into a crust and the temperature gradient in the overlying snow was reduced (Figure 9). In the following 6 days the small facets just above the interface showed evidence of rounding, and the shear strength steadily increased. At a site 300 m higher on the same mountain on 16 January 2004, the crust was only 1.6 cm thick, and no facets were found on the crust, probably because there was insufficient liquid water in the initially moist layer to sustain the temperature gradient in the overlying dry snow.

[41] Five days later and approximately 10 km to the south-southwest of the study site [Jamieson and Langevin, 2004], snow profiles at 1745 m and 1905 m showed 2 cm and 0.2 cm thick crusts, respectively. No facets were found on the crusts at these elevations probably because of less rain on the night of 14 January at these elevations and hence...

Figure 8. Snow temperature in the top 8 cm of the snowpack over time on 2–3 March 1990 on Hokkaido, Japan. The top 2-cm layer, that became faceted, had a density of 90 kg m$^{-3}$ and the underlying 320 kg m$^{-3}$ layer was initially wet. Reprinted from Fukuzawa and Akitaya [1993] in *Annals of Glaciology* with permission of the International Glaciological Society.

Figure 9. Five temperature profiles taken from the hourly profiles of the dry-on-wet layer combination buried 15 January 2004 at 1600 m on Mount St. Anne. Reprinted from Jamieson and Langevin [2004] with permission of the Canadian Avalanche Association.

Figure 10. Change in shear strength and type of grains at the upper boundary of the wet layer then crust buried on 15 January 2004 on Mount St. Anne. Reprinted from Jamieson and Langevin [2004] with permission of the Canadian Avalanche Association.
less freezing time to sustain the temperature gradient in the overlying dry snow. These observations at 1600 m and 1905 m bound the minimum conditions for forming a weak layer of facets, which are of interest to avalanche forecasters.

Two kilometers to the west and 100 m below the study site the rain crust and the facets on the rain crust were more developed. Twenty-three to 28 days after the wet layer was buried, Jamieson and Langevin [2004] report five profiles, 12 compression tests and 37 rutschblock stability tests, where a rutschblock test involves a 1.5 m by 2 m column of snow loaded in stages by a skier to cause fracture in weak layers. The facet layer that formed on 15 January produced fractures in all these tests. The median rutschblock stability score was midrange, or four, and the average compression score was moderate, or 19 taps. As further evidence of the instability at this elevation, while traveling on skis, the observers triggered two whumpfs where the compressions core was moderate, or 19 taps. As further evidence of the instability at this elevation, while traveling on skis, the observers triggered two whumpfs where the compressions core was moderate, or 19 taps.

Faceting occurring within an elevation band that is less than the difference in freezing levels in two adjacent storms. Reprinted from Jamieson and Langevin [2004] with permission of the Canadian Avalanche Association. See color version of this figure in the HTML.

Figure 11. Favorable conditions for dry-on-wet (DW) faceting occurring within an elevation band that is less than the difference in freezing levels of precipitation between two storms (Figure 11). From the freezing level of the first storm (level 1a in Figure 11) down to some elevation 1b, dry snow buries a moist layer, but there is insufficient latent heat in the moist snow to sustain DW faceting. At lower elevations, there is usually more latent heat stored in the wet layer and therefore more faceting of the dry snow below the interface with those modeled by SNOWPACK, which simulates the properties of snowpack layers including grain type based on continuous weather and radiation measurements [Bartelt and Lehning, 2002; Lehning et al., 2002a, 2002b].

Colbeck and Jamieson [2001] modeled the heat flow to show that a freezing wet layer can sustain a temperature gradient and warm subfreezing temperature in the overlying dry snow, which are favorable for faceting and grain growth within hours. Soon after the dry snow falls on the wet layer, the temperature gradient is concentrated at the interface explaining why facets are more developed in the dry snow just above the interface than higher in the overlying dry snow.

4.6. Effect of Terrain Factors on DW Faceting

4.6.1. DW Faceting Above Rain Crusts

When there is precipitation on a rain-wetted surface, DW faceting will only occur if the precipitation is dry snow and the underlying wet snow contains sufficient latent heat to sustain \( |dT/dZ| > T_{\text{G}} \) in the dry snow. As a consequence, DW faceting on a rain crust is sometimes confined to an elevation band [Jamieson, 2004b] that is less than the difference in freezing levels of precipitation between two storms (Figure 11). From the freezing level of the first storm (level 1a in Figure 11) down to some elevation 1b, dry snow buries a moist layer, but there is insufficient latent heat in the moist snow to sustain \( |dT/dZ| > T_{\text{G}} \) long enough for faceting to substantially affect the shape of the grains or the mechanical properties at the dry-on-wet interface.

For example, in one case of facets on crust, Jamieson et al. [2001] observed that the facets were more advanced at elevations near line than at higher elevations where the crust was thinner and less latent heat was available in the antecedent wet layer to drive DW faceting. At lower elevations, moist snow or rain fell on the already wet surface, inhibiting DW faceting. Also, during the attempts of Jamieson and Langevin [2004] to place thermistor arrays in rain-wetted snow before dry snow fell, the thermistor strings were often placed at the wrong elevation, illustrating that the elevation band for DW faceting is often narrow, at least in the Columbia Mountains.

4.6.2. DW Faceting Above Temperature Crusts

Using a model of a cold snow overlying a 1-m-thick layer of wet snow, Armstrong [1985] showed that the temperature gradient in the dry snow could be sustained for 6 days before the wet layer froze, which was ample time for facets to develop.

In seven cold laboratory experiments in which dry snow was sieved onto a wet layer, Jamieson and van Herwijnen [2002] observed faceting of crystals at the interface. Jamieson and Fierz [2004] derived an approximation for the freezing time and, for the same seven experiments, compared calculated freezing times with freezing times measured between 5 and 22 hours. The freezing times and observed formation of overlying faceted crystals compared well with those modeled by SNOWPACK, which simulates the properties of snowpack layers including grain type based on continuous weather and radiation measurements [Bartelt and Lehning, 2002; Lehning et al., 2002a, 2002b].

For example, in one case of facets on crust, Jamieson et al. [2001] observed that the facets were more advanced at elevations near line than at higher elevations where the crust was thinner and less latent heat was available in the antecedent wet layer to drive DW faceting. At lower elevations, moist snow or rain fell on the already wet surface, inhibiting DW faceting. Also, during the attempts of Jamieson and Langevin [2004] to place thermistor arrays in rain-wetted snow before dry snow fell, the thermistor strings were often placed at the wrong elevation, illustrating that the elevation band for DW faceting is often narrow, at least in the Columbia Mountains.

4.6.3. DW Faceting Above Sun Crusts

In the late winter and spring at temperate latitudes it is common for strong solar radiation and warm air to melt...
the snow surface on steep sunny slopes. Storms, including convective cells, may locally deposit dry snow, including on slopes where the wet layer can sustain \(dT/dZ > T_G\) at the dry-on-wet interface, resulting in areas of DW faceting and hence areas where the crust is poorly bonded. These areas can be on the scale of an avalanche starting zone and thus of sufficient area for slab avalanche release. While the wet surface layers may be thin, they sometimes have sufficient latent heat to create thin layers of facets.

\[50\] Campbell [2004, pp. 99–100] gave an example of the spatial variability of a poorly bonded sun crust. He performed an array of 23 closely spaced rutschblocks tests on the east and northeast sides of a knoll (Figure 12) on Mount Fidelity in Glacier National Park on 18 March 2004. Lower rutschblocks scores indicate lower stability and increased probability of skier triggering [e.g., Föhn, 1987]. Ignoring the bottom row of tests where slab thickness varied substantially, the rutschblocks scores on the more easterly and sunnier aspect of the knoll are substantially lower than on the shadier (right) side. Lower scores indicate lower stability for skier triggering. Reprinted from Campbell [2004] with permission of the author.

**Figure 12.** Array of closely spaced rutschblock tests on Mount Fidelity on 18 March 2004 that slid on facets on a sun crust. Except for the lower row where slab thickness varied substantially, the scores were mostly lower scores on the sunnier (left) side of the roll where the crust and facets were better developed than on the shadier (right) side. Lower scores indicate lower stability for skier triggering. Reprinted from Campbell [2004] with permission of the author.

5. **DISCUSSION**

\[51\] Having already identified various processes for the formation of layers poorly bonded to crusts or poorly bonded layers within crusts, it is important to know which processes are most likely to lead to these weaknesses in varied geographical locations and at different times. However, this discussion is limited to faceting processes which are common in the Columbia Mountains because that is where most of the field evidence has been gathered.

\[52\] As noted above, \(\Delta T_{10}\) in the Columbia Mountains is usually not sufficient over long enough periods of time to form thick layers of facets, although important exceptions occurred in the winter of 2000/2001 [Hägeli and McClung, 2003]. Certainly, diurnal near-surface faceting can and does extend through surface layers to facet the snow on crusts; however, the increase in thin layers of facets on crusts in late winter suggests either DW or DC faceting; both of these processes concentrate the temperature gradient in the dry snow just above the crust rather than at the snow surface. In the deep snowpack of the Columbia Mountains I have observed the rounding of various grain types including facets overlying crusts when \(\Delta T_{10} < T_{GF}\) but have no observations of DC faceting under this condition. These observations suggest but do not prove that DC faceting is rare in areas of deep snowpack. However, there are several factors and observations indicating that DW faceting is an important process leading to the formation of weak faceted layers on crusts:

\[53\] 1. Warm fronts capable of producing rain or melting of the snow surface below a specific elevation are often followed by cold fronts capable of precipitating dry snow in the same elevation range. A cold front followed by a warm front is characteristic of cyclonic systems at midlatitudes [Ahrens, 2000, pp. 332–333].

\[54\] 2. In the spring several hours of sunshine can warm and melt the snow surface and can be followed by convective snow showers. The wet layers are usually thin with a limited amount of liquid water and the facet layers on spring crusts are often thin.

\[55\] 3. Many of the facet layers on crusts in the Columbia Mountains are less than 1 cm thick, suggesting either DW or DC faceting. Although this paper, Jamieson et al. [2001], and Hägeli and McClung [2003] have identified cases of DW faceting in the Columbia Mountains, our literature review revealed no cases of DC faceting when \(\Delta T_{10} < T_{GF}\). Since it is slower and is likely to exhibit less spatial variability, DC faceting should be easier to observe, but when \(\Delta T_{10} < T_{GF}\) such observations are missing from the literature and my field experience.

\[56\] 4. Some thin layers of facets on crusts are found in a narrow elevation band, which is more likely for DW faceting than for DC faceting. (Figure 11)

\[57\] 5. Figure 12 shows an example of the resulting weakness when facets on a crust are better developed on the sunny aspects where the crust was thicker and hence where the antecedent wet layer would have sustained \(dT/dZ > T_{GF}\) for DW faceting. B. McMahon (personal communication, 2003) found that weaker bonding to crusts on sunny aspects is more common than on shady aspects.

\[58\] Typical growth patterns of facets in a temperate snowpack are shown in Figure 13. Lines A and E show growth of facets subject to a moderate temperature gradient (e.g., \(10^5–30^\circ\text{C m}^{-1}\)). Recognition of faceting under low
surface melting followed by refreezing results in melt-freeze avalanche formation in many mountainous areas. Snow poorly bonded to refrozen crusts is an important aspect of same rate as shown by line A. Grain growth in line Eh as subject to moderate gradients (e.g., 10

6. CONCLUSIONS

The formation of layers of faceted crystals that are poorly bonded to refrozen crusts is an important aspect of avalanche formation in many mountainous areas. Snow surface melting followed by refreezing results in melt-freeze crusts; the melting can be due to rain, warm air, solar radiation, or a combination of these factors. The cause of the melting is an important clue to the distribution of the crust: Rain crusts are usually found below a certain elevation band with little dependence on aspect or slope inclination; crusts caused primarily by sensible heat transfer from warm air are also found below a certain elevation; sun crusts are spatially variable because of the varied warming of the snow surface over the terrain and are especially affected by aspect and slope angle.

The process responsible for the formation of a crust is an important clue as to where it may be poorly bonded: Rain crusts and crusts formed by warm air may occur within a narrow elevation band where dry snow falls on a wet layer with sufficient liquid water content to sustain \( |dT/dZ| > T_{GF} \) and hence cause faceting at the dry-on-wet interface while the wet layer freezes. After melting by solar radiation the wet snow surface can be buried by dry snow, sometimes from a convective cell, resulting in DW faceting and hence a poor bond.

With typical air temperatures the magnitude of the temperature gradient at a dry-on-wet interface can exceed 50°C m⁻¹ for hours while much of the latent heat is drawn upward. Facets can then form within a day at the interface where dry snow overlies wet snow and the snow surface temperature is subfreezing. Facets can continue to grow after the wet layer freezes. The cause of the temperature gradient can vary, but the reduced area of bonds between pairs of facets and the tendency of facets to resist densification can contribute to additional faceting.

If a forecaster suspects a crust is poorly bonded because of DW faceting, then the bond of the overlying snow to the crust may be weak where the crust is relatively thick, as determined by manual probing. At these sites, further testing, including profiles, may be helpful.

Theory and observations indicate that the persistence of weak layers of facets increases with grain size. Surface hoar crystals are typically larger than facets and exhibit a similar trend up to approximately 8 mm. Many layers of facets, including rounded facets (FCmx (type 4c)) from 0.8 to 1.7 mm in size, fracture consistently in compression tests 2 to 4 weeks after burial. Many layers of surface hoar crystals of a size larger than 2 mm fracture consistently 1 to 4 weeks after burial. However, the crystal size of a persistent weak layer is just one of many observations helpful for avalanche forecasting.

In Columbia Mountain observations, facets and surface hoar are found at interfaces of many poorly bonded crusts, including most of those that release dry slab avalanches more than 3 days after the snow or surface hoar layer on the crust was buried. Facets that form at the base of dry snow overlying wet layers form an important portion of the poorly bonded crusts in the Columbia Mountains and should be found on any mountains with a similar climatic regime. These layers include poorly bonded rain crusts in early and late winter season and poorly bonded sun crusts in March and April. Under relatively clear skies, sun crusts can form on east, south, and west aspects in middle-to-late
winter season, especially on steep slopes. In late winter season, sun crusts can form on all aspects and for a wider range of slope angles. In the Columbia Mountains, thin facet layers (≤5 mm thick) are more common in March and April when sun crusts are more common.

[65] A number of other important conclusions can be drawn including what information or further research is needed to understand this phenomenon more completely. For example, the minimum requirements for faceting above a wet layer, such as thickness of the wet layer, density, and thickness of the dry layer, are important but remain unknown. The magnitude of these parameters can best be determined from simulations [e.g., Jamieson and Fierz, 2004] or cold laboratory studies.

[66] Over periods of several weeks or months in the Columbia Mountains I have observed facets below a buried crust retaining their flat faces and sharp edges while facets above the same crust become more rounded. This suggests that when \(\Delta T_{10} < T_{\text{GF}}\), as is common in mountains with a deep snowpack, \([dT/dZ]\) below a buried crust can be sufficient for faceting, while the gradient above the crust is not. Colbeck [1991] suggested cold laboratory experiments under steady thermal conditions to assess this idea, and they are still needed.

[67] Even in snow climates such as the Columbia and Coast mountains of western Canada where the temperature gradient below the top 30 cm is typically less than 10°C m\(^{-1}\) [Jamieson and Johnston, 1999], forecasters report the growth of weak laminations of faceted crystals within initially “uniform” crusts in the top meter of a mildly subfreezing snowpack over a period of weeks. These weak layers form from a multistep process that adds two layers of complexity to the DW processes described in section 4.5: As water enters thin layers of dry snow that consist of alternating layers of larger and smaller grains, the infiltrating water is preferentially absorbed by the fine-grained layers because they have a stronger capillary attraction [Wankiewicz, 1979]. Once freezing occurs, the preferentially wetted, fine-grain layers freeze into dense, icy layers of a higher thermal conductivity. Then if a temperature gradient is imposed on this new, multilayered crust, the temperature gradient is proportionately higher in the large-grained, low-density layers because of their lower thermal conductivity. The temperature gradient within the low-density layers can be high enough to produce faceting within the layer, and thus alternating layers of hard crust and faceted crystals develop from a single crust. Detailed studies of this phenomenon are needed in the laboratory or the field since work so far has concentrated on either faceting only above or below a crust.

NOTATION

- CRmfc: melt-freeze crust.
- \(dT/dZ\): temperature gradient on the scale of grains measured normal to the slope.
- DC: faceting of dry snow on a frozen crust.
- DF: decomposing and fragmented precipitation particles.
- DH: depth hoar; cup-shaped, striated crystals, usually hollow.
- DW: faceting of dry snow overlying a wet layer.
- FC: solid faceted crystals, usually angular crystals with flat faces.
- FCCa: faceted crystals.
- FCCmx: faceted crystal, mixed forms which show signs of rounding.
- K: knife hardness.
- P: pencil hardness.
- PP: precipitation particles, including dendritic forms with high surface to volume ratio.
- SH: surface hoar (frost crystals); striated, usually feathery crystals.
- TGf: threshold temperature gradient for kinetic growth (faceting).
- \(\Delta T_{10}\): temperature difference over 10 cm measured vertically or slope normal.
- 4F: four finger hardness.
- open square faceted crystals.

[68] ACKNOWLEDGMENTS. For many thoughtful suggestions on the manuscript I am grateful to S. C. Colbeck. Mark Moore, Chris Stethem, Ron Perla, John Hetherington, Bruce McMahon, and Ed LaChapelle provided insight and stimulating conversations on faceting near crusts. For financial support I thank HeliCat Canada, the Natural Sciences and Engineering Research Council of Canada, Mike Wiegele Helicopter Skiing, Canada West Ski Areas Association (CWSAA), and the Canadian Avalanche Association.

[69] The Editor responsible for this paper was Daniel Tartakovsky. He thanks the technical reviewers Samuel C. Colbeck and Ed Adams and one anonymous cross-disciplinary reviewer.

REFERENCES


B. Jamieson, Department of Civil Engineering, University of Calgary, 2500 University Drive NW, Calgary, Alberta, Canada T2N 1N4. (bruce.jamieson@ucalgary.ca)