SNOW AVALANCHE FORMATION

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[1] Snow avalanches are a major natural hazard, endangering human life and infrastructure in mountainous areas throughout the world. In many countries with seasonally snow-covered mountains, avalanche-forecasting services reliably warn the public by issuing occurrence probabilities for a certain region. However, at present, a single avalanche event cannot be predicted in time and space. Much about the release process remains unknown, mainly because of the highly variable, layered character of the snowpack, a highly porous material that exists close to its melting point. The complex interaction between terrain, snowpack, and meteorological conditions leading to avalanche release is commonly described as avalanche formation. It is relevant to hazard mapping and essential to short-term forecasting, which involves weighting many contributory factors. Alternatively, the release process can be studied and modeled. This approach relies heavily on snow mechanics and snow properties, including texture. While the effect of meteorological conditions or changes on the deformational behavior of snow is known in qualitative or semiquantitative manner, the knowledge of the quantitative relation between snow texture and mechanical properties is limited, but promising developments are under way. Fracture mechanical models have been applied to explain the fracture propagation, and micromechanical models including the two competing processes (damage and sintering) have been applied to explain snow failure. There are knowledge gaps between the sequence of processes that lead to the release of the snow slab: snow deformation and failure, damage accumulation, fracture initiation, and fracture propagation. Simultaneously, the spatial variability that affects damage, fracture initiation, and fracture propagation has to be considered. This review focuses on dry snow slab avalanches and shows that dealing with a highly porous media close to its melting point and processes covering several orders of scale, from the size of a bond between snow grains to the size of a mountain slope, will continue to be very challenging. *Index Terms:* 1863 Hydrology: Snow and ice (1827); 1827 Hydrology: Glaciology (1863); 1899 Hydrology: General or miscellaneous; *Keywords:* snow, avalanche

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1. INTRODUCTION

[2] Snow avalanches are snow masses that rapidly descend steep slopes. They can contain rocks, soil, vegetation, or ice. There are two types of release: loose snow avalanches and slab avalanches. Loose snow avalanches start from a point in a relatively cohesionless surface layer of either dry or wet snow. Initial failure is analogous to the rotational slip of cohesionless sands or soil but occurs within a small volume (<1 m^3) in comparison to much larger initiation volumes in soil slides. Snow slab avalanches involve the release of a cohesive slab over an extended plane of weakness, analogous to the planar failure of rock slopes rather than to the rotational failure of soil slopes [*Perla*, 1980] (Figures 1–3). The observed ratio between width and thickness of the slab varies between 10 and 10^3. Slab thickness is usually <1 m [*Perla*, 1977; *Schweizer and Jamieson*, 2001]. As shear fracture below the slab spreads along the plane of weakness, high tensile stress and tensile fracture develops upslope. Accordingly, slab failure appears to begin with the propagation of tensile crown fracture across the slope. The triggering of a snow slab avalanche can occur by (1) localized rapid near-surface loading by, for example, people or explosives (called artificial triggering), (2) gradual uniform loading due to, for example, precipitation, or (3) a no-loading situation that changes snowpack properties, for example, surface warming (called natural triggering or spontaneous release). The main difference in triggering modes is the rate of loading, which is important since snow is highly rate-dependent.

[3] Slab avalanche release can be subclassified by the type of instability [*McClung*, 2002]. New snow instabilities are commonly the result of overloading due to rapid precipitation during storms. They lead to direct-action avalanches. Climax avalanches follow from old snow

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Figure 1. Crown fracture of a huge snow slab avalanche.

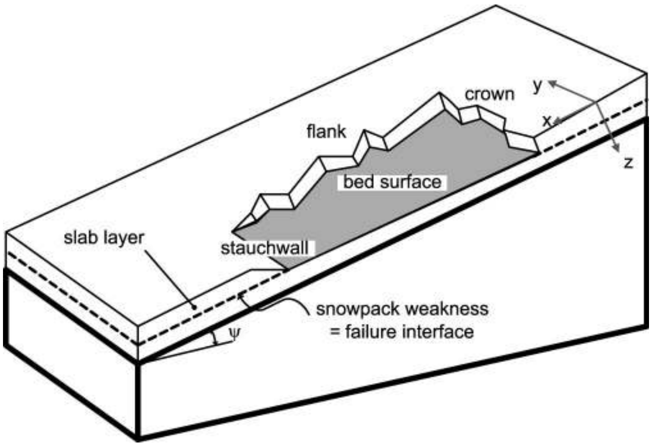

Figure 3. Slab avalanche nomenclature and coordinate system (adapted from *Perla* [1977] with permission of the National Research Council of Canada).

instabilities due to the failure of a buried layer of kinetic growth crystals. These crystal types include surface hoar, (near-surface) faceted crystals, and depth hoar, and they compose persistent weak layers [*Jamieson and Johnston*, 1992].

[4] Natural avalanches threaten residents and infrastructure, whereas human-triggered avalanches are the main threat to recreationists [*Jamieson and Geldsetzer*, 1996; *Tschirky et al*., 2001]. The average number of snow avalanche fatalities worldwide is estimated at 250 per year [*Meister*, 2002]. In the United States, snow avalanches cause on an average annual basis more fatalities than earthquakes or landslides [*Voight et al*., 1990].

Figure 2. Snowpack configuration at the crown of a slab avalanche: cohesive slab on top of a thin weak layer (buried surface hoar crystals).

[5] Avalanche formation is the complex interaction between terrain, snowpack, and meteorological conditions leading to avalanching. Avalanche formation can be broken down to the following questions: Where and when does what kind of avalanche occur? These are key questions for the backcountry traveler, the avalanche forecaster, and the snow safety manager alike. "How" and "why" are additional questions for the researcher. There is usually no clear answer to the questions of occurrence, and current science is not able to give definite answers on the processes involved. In fact, because of the stochastic nature of some of the meteorological processes acting on the snow cover, a purely deterministic approach to the questions of "where" and "when" will have limited success.

[6] The dynamics of avalanche motion and mitigation by planning and structures are not dealt with in this review. *Barbolini et al*. [2000] provide a recent overview on avalanche dynamics models.

[7] Avalanche formation can be approached in two ways: (1) The complex interaction between terrain, snowpack, and meteorological conditions can be explored by association or statistics, or (2) the physical and mechanical processes of avalanche formation can be studied and modeled. Whereas the latter approach is physical, the former is applied by most avalanche-forecasting services. By empirically weighting the influence of the contributory factors in a specific situation, the avalanche probability and characteristics are estimated, and a forecast is made. The contributory factors are known [*Atwater*, 1954; *de Quervain*, 1966; *Perla*; 1970] and have physical meanings that are related to the avalanche formation process.

[8] Snow slope failure has been studied with a strength-of-material approach. Snow stability S is calculated for a given time, depth within the snowpack, and location on the slope. Hence the avalanche problem is reduced to the question of balance between snow strength τ_f and stress (normal stress σ and shear stress τ) at time t and location $\mathbf{x}$:

$$S = \frac{\tau_f(\sigma,\mathbf{x},t)}{\tau(\mathbf{x},t)}. \tag{1}$$

Theoretically, unstable conditions will occur when the stability index S approaches 1. Since strength and load vary spatially and temporally within the snowpack, the application of this critical stress concept for snow slope failure is not straightforward, and snow stability depends on scale. Important parameters (strain and strain rate) and processes (fracture propagation) are not considered. While crack initiation will depend on stresses, the formation and propagation of cracks requires deformation energy [*Bazant and Planas*, 1998]. Therefore the failure of a snow slope needs to be considered from a fracture mechanical view focusing on three critical variables: stress, flaw size, and toughness [*Anderson*, 1995].

[9] The snow slope failure process involves a wide range of scales. Avalanche formation starts at the microscale, where single bonds break (10^{-4} m), and finishes with fracture propagation leading to the release of the slab (macroscale 10^1–10^2 m). The mesoscale is primarily the scale of the snowpack thickness (10^{-2}–10^1 m). Variability of the mechanical parameters due to varying meteorological conditions is inherent on all scales and is fundamental to both fracture initiation and fracture arrest.

[10] In the following we assess the state of knowledge on avalanche formation considering both the contributory factors and the failure mechanics at different scales. We focus on dry snow slab avalanches because they represent the major type of avalanche hazard.

[11] *Mellor*'s [1968] monograph on avalanches is a landmark, and *McClung and Schaerer* [1993] have comprehensively summarized the subject for a broader public. Our aim is a synthesis that steers the interested reader to the references that either represents recent developments or starting points for more detailed studies.

2. CONTRIBUTORY FACTORS

[12] Most contributory factors are related to either strength or load and their variation. *Atwater* [1954] proposed 10 weather and snow factors that contribute most to avalanche danger but did not consider terrain. We describe five essential factors: terrain, precipitation (especially new snow), wind, temperature (including radiation effects), and snowpack stratigraphy. If consistent avalanche occurrence data are available, the contribution of weather and snow variables can be determined quantitatively [e.g., *McClung and Tweedy*, 1993]. Assessing the avalanche release probability by estimating and weighting of each of the contributory factors in a given situation has been successfully applied because all relevant meteorological factors can be measured by automatic weather stations [*Gubler*, 1993].

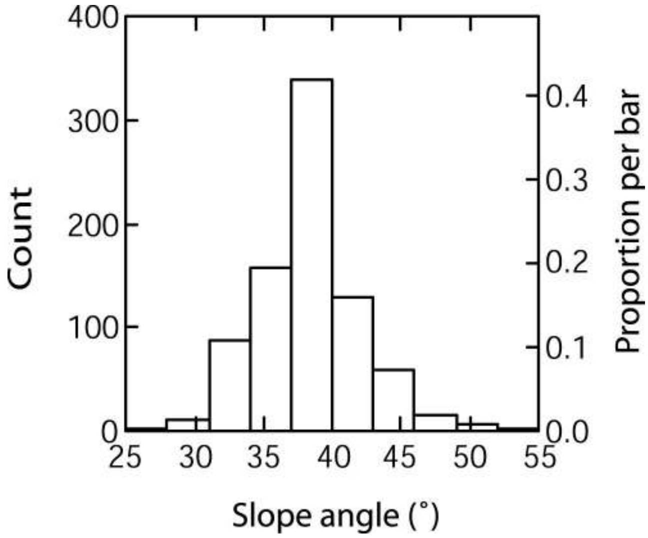

Figure 4. Slope angle in starting zone of human-triggered avalanches (N = 809, first quartile, 37°, with median of 39°; third quartile, 41°, with mean of 38.8° ± 3.8°). The mean thickness of the sampled slabs was 0.49 ± 0.22 m (reprinted from *Schweizer and Jamieson* [2001] with permission from Elsevier).

2.1. Terrain

[13] Terrain is an essential factor and the only factor that is constant over time. A slope angle of >30° is usually required for dry snow slab avalanches. However, there are different scales/ways to measure slope angle that affect the critical angle and hinder the comparison of data from different sources. For avalanche forecasting the critical slope is the steepest angle from the horizontal averaged over about 20 m in the starting zone. *Schweizer and Jamieson* [2001] analyzed a large data set of skier-triggered avalanches from Switzerland and Canada including data on aspect (compass direction that the slope faces) and slope angle (Figure 4). The results do not differ substantially from previous studies on natural avalanches or data sets of avalanches with different types of triggering [*Perla*, 1977]. Analysis of the catastrophic avalanches in the Alps during winter 1999 has shown that few avalanches released on terrain of <30° [*Ammann*, 2000].

[14] On some Swiss ski touring maps the terrain steeper than 30° is specially colored. *Munter* [1997] has proposed rules of terrain selection for recreationists by coupling slope angle to danger level (five-degree European avalanche danger scale: low, moderate, considerable, high, very high [*Meister*, 1995]). If, for example, the

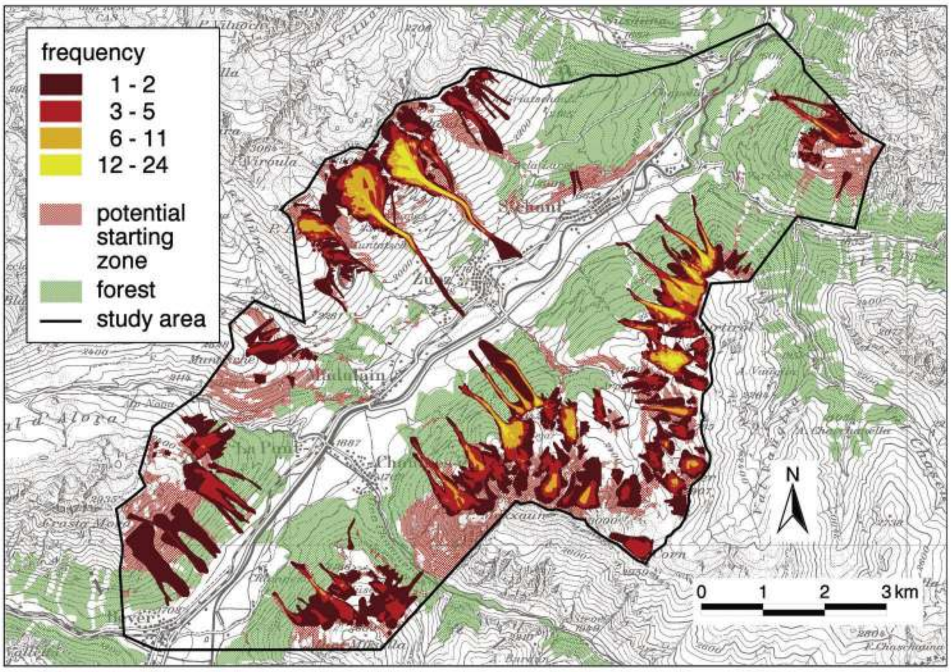

Figure 5. Spatial distribution of frequency of avalanche occurrence during the 14-year period of avalanche observations for the area of Zuoz, Engadine Valley, Switzerland (adapted from *Stoffel et al*. [1998], reprinted from the *Annals of Glaciology* with permission of the International Glaciological Society).

public avalanche bulletin rates the danger level as moderate, it is recommended not to ski slopes steeper than 40°. As there are no data on the frequency of skiing by slope angle, no real risk analysis can be done, and the true relation between slope angle and probability of triggering has not been established.

[15] There are no well-established rules on the effect of the microtopography of starting zones [e.g., *Bozhinskiy and Losyev*, 1998]. However, avalanches are more frequent in starting zones with concave cross-slope profiles [*Gleason*, 1995; *McClung*, 2001]. Slope angle variations (convex downslope curvature, e.g., a bump) provoke stress concentrations that favor avalanche formation [e.g., *Föhn et al*., 2002].

[16] With digital terrain models (DTM) and geographical information systems (GIS) [*Lied et al*., 1989], potential starting zones can be identified and their characteristics compared to avalanche occurrence [*Stoffel et al*., 1998] (Figure 5) or used as input for avalanche dynamics calculations [*Gruber et al*., 1998]. However, this approach is only reliable if the DTM resolution and accuracy are not larger than 20–30 m, and even then they do not indicate small-scale variations in slope angle that are of interest for avalanche forecasting. One of the first approaches along this line was the French expert system for avalanche starting zone path analysis [*Buisson and Charlier*, 1989]. So far, the systematic identification of starting zones has been restricted to slope angle (terrain between 30° and 50°, occasionally 60°) and vegetation (no forest). However, in order to improve hazard mapping the frequency of avalanching in a given potential starting zone needs to be known. The combined effect of topographic parameters (cross-slope curvature, slope, distance to the next ridge, and aspect) on the frequency of large avalanches has been analyzed using GIS and the extensive avalanche occurrence data from the region of Davos, Switzerland [*Maggioni and Gruber*, 2002]. It confirmed that highly concave cross-slope curvature in combination with a high mean slope angle (>36°) leads to high avalanche frequency but small release area compared to the potential starting zone area.

[17] Terrain roughness influences avalanche formation by hindering the formation of continuous weak snowpack layers. A minimal snow depth of 0.3–1 m is necessary to smooth out most terrain roughness. Therefore, when analyzing the effect of new snow loading, the snow depth prior to the snowfall is relevant, reducing the threshold value of the 3-day sum of new snow depth for a given avalanche probability [*Stoffel et al*., 1998]. Highly variable terrain roughness can influence the internal evolution of thin snowpacks. Rocky outcrops or slightly covered rocks may promote instability through growth of faceted crystals due to the higher temperature gradients where the snowpack is thinner [*Arons et al*., 1998; *Logan*, 1993].

[18] Forests inhibit avalanche formation. In dense forests (>200 trees of diameter >16 cm ha^{-1}) the snow cover is too irregular to produce avalanches [*Frey et al*., 1987; *Gubler and Rychetnik*, 1991; *Schneebeli and Meyer-Grass*, 1993]. In particular, snow interception modifies the old snow surface and hinders weak layer formation;

it also changes the distribution and accumulation rate of new snow during the storm. Extreme slab avalanches with fracture depth >1 m should only develop where a continuous weakness with no significant interruption exists within an open area of about 10 m width and 10–20 m length. Avalanches starting from clear-cut logging have destroyed timber and caused substantial losses in western Canada [*McClung*, 2001].

2.2. New Snow

[19] For large (catastrophic), new snow avalanches, precipitation is the strongest forecasting parameter [*Föhn et al.*, 2002] and is closely related to avalanche danger (Figure 6). Accumulation of a new snow depth of about 1 m within a storm is considered critical for the initiation of extreme avalanches; about 30–50 cm is critical for naturally released avalanches in general. However, even with large amounts of new snow, the combined release probability of a group of avalanche paths is frequently <50% [*Schaer*, 1995]. This shows that the new snow depth alone is not sufficient to explain avalanche activity.

[20] For natural releases during or shortly after storms the precipitation rate or loading rate can strongly influence the critical balance between stress and strength. If the new snow loading is rapid ($\geq$2.5 cm h^{-1}), the weak layer below the storm snow layer might not gain strength sufficiently quickly. The strength gain follows from the load of the overlaying slab. Hence there is a competition between the rate of loading from snowfall and the rate of strengthening of buried weaknesses. Forecasting models were developed based on this simple stress criterion (stability) [*Conway and Wilbour*, 1999; *Endo*, 1991]. Considering a planar snowpack inclined at angle ψ and neglecting longitudinal stresses, the stability index $S_z(t)$ for a weak layer at given depth z and time t is

$$S_z(t) = \frac{\tau_{fz}(t)}{\tau_{xz}(t)} = \frac{A\left(\frac{\rho_{\text{snow}\,z}(t)}{\rho_{\text{ice}}}\right)^2}{g\int \dot{P}_w \cos\psi \sin\psi \, dt}, \qquad (2)$$

where A is a constant, g is the gravitational acceleration, and $\dot{P}_w$ is the rate of loading [*Conway and Wilbour*, 1999]. The basal shear strength $\tau_{fz}(t)$ is estimated by fitting a power law relationship to measurements of shear strength and density [*Jamieson*, 1995]. In a maritime climate the model to predict direct-action avalanches during storms showed promising results, despite the fact it used an average strength value, which depends only on density.

[21] For skier-triggered avalanches, even lower critical values of new snow thickness were proposed, depending qualitatively on wind, temperature, surface conditions prior to the snowfall, and prior skiing activity on the slope [*Munter*, 1997]. The critical new snow values vary between 10–20 cm under unfavorable and 30–50 cm under favorable conditions, based on the above mentioned parameters.

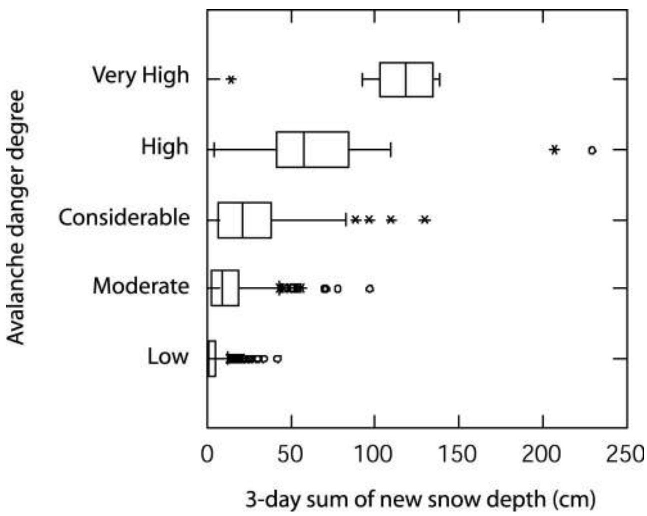

Figure 6. New snow depth during 3-day storm periods in relation to the verified degree of avalanche danger (five-degree unified European danger scale) for the region of Davos, Switzerland (N = 1512). Spearman rank-order correlation coefficient between 3-day sum of new snow depth and danger degree was 0.58 ($p < 0.001$). Boxes span the interquartile range from first to third quartile with a vertical line showing the median. Whiskers show the range of observed values that fall within 1.5 times the interquartile range above and below the interquartile range. Asterisks show outliers; circles show far outside values (adapted from *Schweizer and Föhn* [1996], reprinted from the *Journal of Glaciology* with permission of the International Glaciological Society).

[22] If new snow depth is not measured, it can be estimated from the increase in snow depth by taking into account settling of underlying layers [*Kominami et al.*, 1998]. This same approach is used operationally for the Swiss avalanche warning service based on continuous snow depth measurements from automatic weather stations [*Lehning et al.*, 1999].

[23] The density of new snow also affects avalanche formation. *Mueller* [2001] showed that decreasing density with depth (denser snow above less dense snow) is associated with increased avalanche activity.

2.3. Wind

[24] Wind contributes to loading and is often considered the most active contributing factor after new snow. Loading by wind-transported snow can be fast and produces irregular deposits with locally increased loading rates. Variations in wind speed and snow drift form layers of different density or hardness, creating stress concentrations within the layered snowpack. *de Quervain* [1966] proposed that the wind transforms the snow into a more brittle material and that wind-deposited snow layers are more prone to avalanching. *Meister* [1985] studied the relation between new snow density, air temperature, and wind speed and suggested that wind-hardened slabs have high viscosities. However, neither de

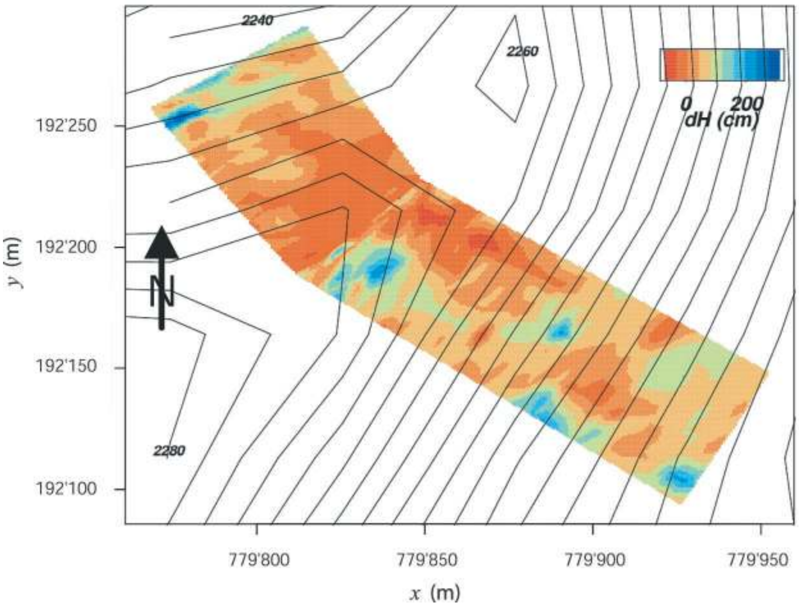

Figure 7. Snow distribution on windward and lee slopes of a mountain ridge with 5-m contour lines (Gaudergrat, near Davos, Switzerland) after a snowdrift period (26–31 January 1999). Main wind direction was northwest. The difference in snow depth *dH* compared to the snow depth prior to the snowdrift episode is shown. Negative values of *dH* indicate erosion [from *Doorschot et al.*, 2001] (reprinted from the *Annals of Glaciology* with permission of the International Glaciological Society).

Quervain nor Meister determined brittleness that is a fracture mechanical property. In addition, the characteristics of the weak layer below the wind-deposited layer are at least as important.

[25] *Gauer* [1999] measured the snow deposition (and erosion) pattern on both sides of a mountain ridge. The patterns were highly irregular, in particular on the lee slope, probably because of turbulent eddies. The eddies can be stationary for a certain wind speed and direction (dunes are formed), but they will change as wind speed or direction change. Averaged over the whole slope, 20–30% more snow was deposited compared to a level study plot, but locally much larger differences existed. *Doorschot et al*. [2001] reported up to fourfold increase in snow deposition in the lee area close to the ridge compared to the flat field (Figure 7). This indicates that fracture depth taken at the fracture line of large catastrophic avalanches might substantially overestimate the average slab thickness in the starting zone. In contrast to gentle terrain, in alpine terrain the transport by suspension is more important than transport by saltation. This is due to the higher turbulence in the mountains in general and particularly to the larger upward component of the flow velocity [*Gauer*, 1999].

[26] Numerical modeling of snow erosion and deposition has become increasingly sophisticated with increasing computing power [e.g., *Doorschot et al*., 2001; *Gauer*, 1999, 2001; *Greene et al*., 1999; *Guyomarc'h and Mérindol*, 1998; *Lehning et al*., 2000a, 2000b; *Naaim et al*., 1998]. When discussing the role of snow drift in avalanche formation, *Lehning et al*. [2000b] took four connected processes into account: (1) the wind field over steep topography, (2) the preferential deposition of snow during snowfall, (3) the possible redistribution of already deposited snow, and (4) the different snowpack conditions and development at the sites of erosion and deposition. Their model parameterized preferential deposition and redistribution by coupling the numerical model for drifting and blowing snow with the snow cover model called SNOWPACK [*Lehning et al*., 1999]. *Föhn* [1980] and *Meister* [1989] previously proposed an empirical formulation relating the additional snow depth H_{wind} deposited in lee slopes per day to the third power of the daily average wind speed $\bar{u}$ ($\bar{u} \leq 20$ m s^{-1}):

$$H_{\mathrm{wind}} = k\,\bar{u}^3\,, \tag{3}$$

where the coefficient $k = 8 \times 10^{-5}$ s^3 d^{-1} m^{-2} was determined empirically based on 3 years of mass balance measurements over a mountain ridge.

[27] *Gauer* [2001] showed that the initially rapid increase in mass flux with increasing wind speed decreased with further increase in wind speed. The power decreased from approximately 4 to 2. When only saltation was modeled, the power was about 3, corresponding to the empirical formulations. So far, it has been assumed that snow drift peaks at a wind speed of about 20–25 m s^{-1} and decreases with even higher wind speeds. Considering saltation and suspension, the calculations by *Doorschot et al*. [2001] suggest that snow drift might level off at lower wind speeds because of saturation. The saturation effect is particularly pronounced during snowfall where the amount of redistributed snow is limited and preferential deposition of the falling snow dominates snow loading. The numerical simulations repro-

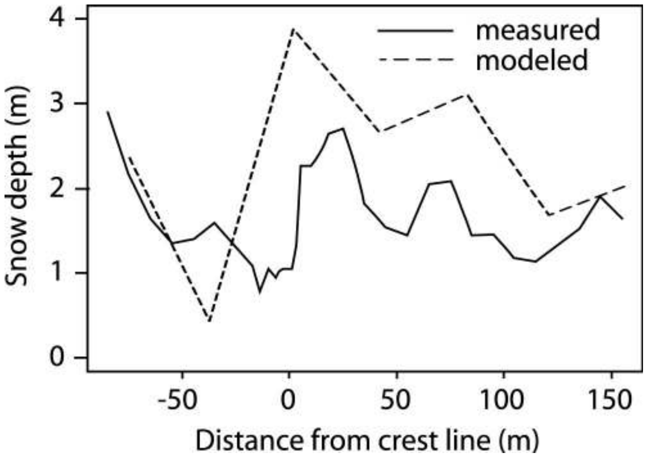

Figure 8. Comparison of modeled to measured snow distribution across a mountain ridge (Gaudergrat, near Davos, Switzerland) [from *Lehning et al.*, 2000b] (© Swets and Zeitlinger).

duce the general snow deposition patterns across a mountain ridge (Figure 8) but are limited to a single slope. In order to achieve useful results at the scale relevant for avalanche formation a high resolution (<20 m) of the wind field is needed, requiring extensive computing time. Application to whole mountain ranges for avalanche forecasting is presently not feasible.

[28] Alternatively, a more empirical approach can be followed using GIS [e.g., *Purves et al.*, 1998; *Mases et al.*, 1998], comparable to the approach by *Buisson and Charlier* [1989]. However, forecasters and hazard mapping engineers who assess the run out from extreme avalanches use judgment to assess the additional loading in starting zones. As a step toward applicability, *Lehning et al*. [2000a] developed the drift index for the additional amount of snow in a typical lee slope based on wind speed measured on a mountain crest and on modeled surface conditions (snow mobility) at the site of an automatic snow station. Snow mobility depended on grain size and bond size, parameters that were available from the numerical snow cover model.

[29] A new device to assess snow drift, FlowCapt, is used for local avalanche forecasting. This acoustic sensor determines wind velocity and particle flux based on the particle impacts on the sensor pipe [*Chritin et al.*, 1999].

2.4. Temperature

[30] Temperature is a decisive factor contributing to avalanche formation, particularly in situations without loading. Its effect on snow stability is complex since changes in air temperature affect snow stability in various ways. Again, the rate of change is important. Rising temperature during a storm and rapid temperature increase shortly after a storm contribute to instability. Changes in air temperature primarily affect surface layers, i.e., the slab, whereas the weak layer is relatively unaffected because of the generally low thermal conductivity of snow [*Sturm et al.*, 1997]. Although snow strength decreases with increasing snow temperature, instability after rapid warming does not develop from a weakening of the weak layer below the slab but from increased deformation of the surface layers of the slab, leading to increased strain and strain rates at the slab/weak layer interface.

[31] The mechanical properties of snow are highly temperature-dependent [*McClung*, 1996]. *McClung and Schweizer* [1999] reviewed temperature effects on snow hardness, failure toughness and shear strength. In general, there are two important groups of competing effects: (1) metamorphism (depending on temperature, temperature gradient, and other snow properties) and creep and (2) mechanical properties (excluding metamorphism effects) including snow hardness, fracture propagation potential (toughness), and strength (Table 1). Group 1 effects need more time, whereas group 2 effects change snow stability rapidly.

[32] *Schweizer* [1998] studied the shear strength of natural samples taken in a study plot at different temperatures in a cold laboratory. The stiffness (initial tangent modulus) was the most temperature-sensitive mechanical property of snow. *Camponovo and Schweizer*

Table 1. Snow Temperature Effects According to Time Required and Stability[a]

Snowpack Property or Process	*Change due to Warming*	*Response Time*	*Effect on Stability*
Stiffness or hardness of slab	significant decrease	immediate	decrease
Toughness	increase	immediate	decrease
Strength of weak layer/interface	slight decrease	immediate	decrease
Bond formation (metamorphism)	increase of bond formation rate and strength	delayed	increase
Creep	increase of creep rate causing settlement and densification hand hence increasing strength and hardness	delayed	increase
Snow temperature (temperature gradient)	usually decrease of temperature gradient causing change of crystal form and increasing strength	delayed	increase

[a]For warming, immediate effects promote instability; delayed effects promote stability. Under warming, instability is likely to come from immediate not delayed effects. Strength effects may be immediate (decrease) or delayed (time-dependent with increase) under warming, with the greatest strength changes being delayed [after *McClung and Schweizer*, 1997].

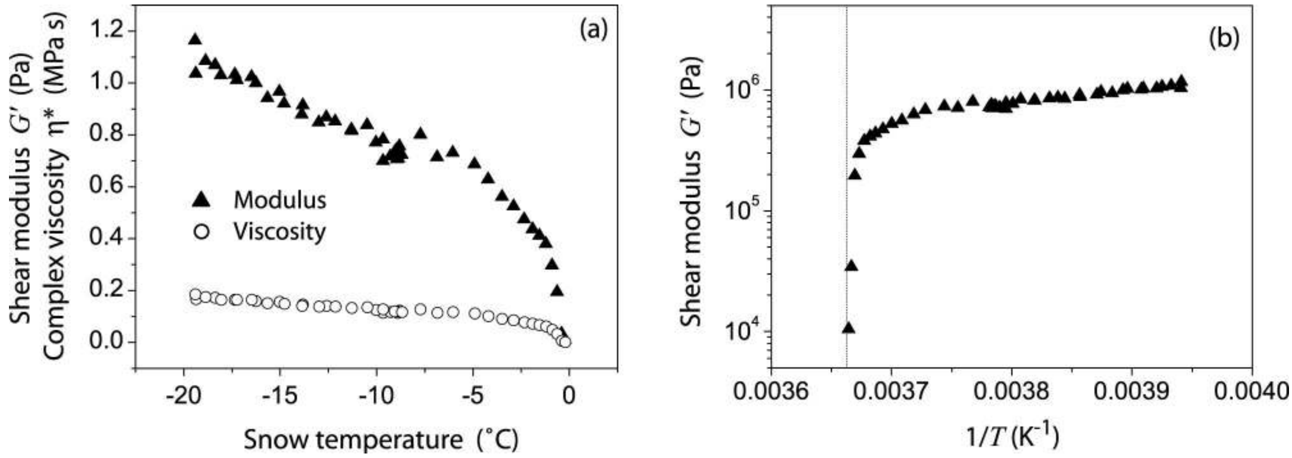

Figure 9. Effective elastic shear modulus G' versus temperature. Results are obtained with continuous oscillation measurements during 8 hours (torsional shear). Laboratory temperature changed from −7°C to −20°C and subsequently to −1°C and back to −7°C. Hardly any hysteresis is apparent, indicating that the experiment was performed in the linear viscoelastic range without textural changes happening during the experiment. (a) Dynamic shear modulus and complex viscosity versus temperature in °C. (b) Dynamic shear modulus versus inverse of absolute temperature to show linear behavior (Arrhenius relation) above about −6°C. Dashed vertical line indicates 0°C. Snow type tested consisted of small rounded grains and partly decomposing and fragmented precipitation particles, of size 0.25–0.5 mm, density 220 kg m^{-3}, and hand hardness index 2 (reprinted from *Schweizer and Camponovo* [2002] with permission from Elsevier).

[2001] recently measured the viscoelastic properties of snow by using a rheometer. By applying small deformations with a linear and recoverable response, continuous measurements, while changing the ambient temperature, yield the temperature dependence of the modulus [*Schweizer and Camponovo*, 2002]. The temperature dependence of the dynamic shear modulus G' follows an Arrhenius relation up to about −6°C, with a much accelerated decrease toward 0°C (Figure 9). The variation of the modulus is important for the penetration of load from the surface to the weak layer and also for fracture processes. Even if the strength in the weak layer decreases with increasing temperature, which is delayed and attenuated with respect to the surface perturbation, the fracture toughness increases. Because of that increase, fracture initiation and fracture propagation become less probable with increasing snow temperature. The high fracture toughness at temperatures close to the melting point also explains why triggering of moist or wet slab avalanches by localized rapid loading, e.g., explosives, is rare. On the other hand, rain on fresh snow triggers wet slabs by loading, by contraction-related stresses, and by changes in the properties of the surface layers [*Conway*, 1998].

[33] Radiation can reduce snow stability similarly to rapid warming but can be more effective. Snow cover modeling [*Bader and Weilenmann*, 1992; *Brun et al.*, 1989] as well as the high-quality input data that drive the models has substantially increased our understanding of the effects of radiation on the snowpack and has explained some of the processes leading to weak layer formation at the snow surface (see section 2.5). The energy balance for any aspect, slope angle, elevation, and time of the year can be calculated [*Durand et al.*, 1999]. However, there is little verification, and the effect of the surrounding terrain (e.g., shading effects, reflections, and emission from terrain) is not included or is only considered for the ablation period [*Fierz et al.*, 1997]. The energy balance can be dominated by the outgoing long-wavelength radiation during the winter months of December and January, particularly on shady slopes and in clear-sky conditions. This effect prevents an increase in air temperature (warming) from affecting the temperature of the snow and thereby its stability. However, this knowledge has not been applied systematically to stability evaluation or avalanche forecasting.

[34] Whereas there are many comprehensive measurements on snow albedo [*Sergent*, 1998], measurements of the radiation absorption in the snow cover suggest that this process in the uppermost 5–10 cm is not fully understood [*Gaia*, 1993]. Detailed temperature measurements in the snow cover have shown that daily variations affect the upper 10–20 cm. High-frequency radiation (blue and UV) might cause very minor effects on the snow temperature up to 50 cm depth but without substantial effect on snow stability.

2.5. Snow Cover Stratigraphy and Selected Snow Properties

[35] Snow cover stratigraphy is recognized as the key contributing factor for dry snow slab avalanche formation. Any loading by new or wind-driven snow or any temperature increase has no effect on snow stability if no weakness exists in either the old snow or at the old snow surface underlying the new snow. Therefore the weak layer or interface is a necessary prerequisite but not

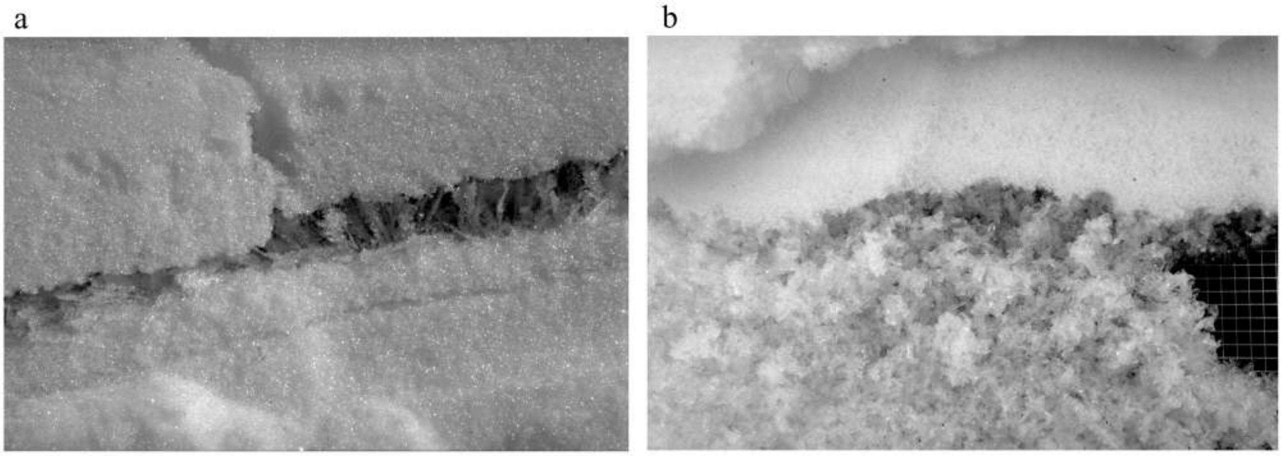

Figure 10. Snow stratigraphy. Weakness below the slab is required for dry snow slab avalanche formation. (a) A thin weak layer of buried surface hoar crystals that partly fractured (left) and that is still intact (right) [from *Jamieson and Schweizer*, 2000] (reprinted from the *Journal of Glaciology* with permission of the International Glaciological Society). Layer thickness of unfractured surface hoar is approximately 19 mm. (b) Weak interface between (below) a depth hoar layer and (above) the new snow layer. Scale of grid is 3 mm.

sufficient condition for avalanche formation (Figure 10). The properties of the overlying slab also have to be considered [*McClung and Schweizer*, 1999; *Schweizer*, 1993; *Schweizer et al*., 1998], particularly for fracture propagation. There are several studies on weak layer properties. *Föhn* [2001] and *Hachikubo* [2001] measured and modeled the development of surface hoar. Two other types of weaknesses are near-surface faceting [*Fierz*, 1998; *Fukuzawa and Akitaya*, 1993] and a poor bond to Sun-generated crusts [*Ozeki et al*., 1995]. The formation of near-surface faceted crystals is due to large temperature gradients near the snow surface resulting from the heat loss by outgoing long-wavelength radiation. At measured temperature gradients of 150 K m^{-1}, growth rates were of the order of 0.1 mm d^{-1}. *Birkeland* [1998] has comprehensively summarized the processes that lead to near-surface faceted crystals: radiation recrystallization, faceting adjacent to a wet layer, and diurnal recrystallization. Near-surface faceting is the most active process, besides surface hoar growth, that leads to weak layer formation. Faceting processes above crusts and wet layers, and to a lesser degree below, are the only efficient ways to form weak layers within the snowpack [*Colbeck and Jamieson*, 2001]. Although the large majority of avalanches during storms are probably released by nonpersistent weak layers, 70% of 186 skier-triggered avalanches were released by weak layers of persistent grain types (i.e., surface hoar, faceted crystals, and depth hoar) [*Schweizer and Jamieson*, 2001]. Analysis of fracture line profiles showed that the weak layer differs distinctly in grain size and hardness from the adjacent layers. These snowpack properties together with snowpack test results are the basis of five stability classes of snow profiles [*Schweizer and Wiesinger*, 2001]. When comparing stable with unstable profiles, the differences in grain size and hardness between the weak layer and the adjacent layer for the unstable profiles were significantly larger than for the stable profiles (Figure 11) [*Schweizer and Jamieson*, 2003b].

[36] *Jamieson and Johnston* [1999, 2001] made extensive measurements of weak layer strength and calculated a stability index, which related to skier-triggered avalanches. They provided shear strength for weak layers by density and grain type. Together with the data on tensile strength [*Jamieson and Johnston*, 1990], these are the most consistent set of brittle strength data for natural snow. *Jamieson and Schweizer* [2000] proposed a conceptual model to explain strength changes based on bonding and texture of buried surface hoar layers. *Jamieson and Johnston* [1999] reported that on average buried weak surface hoar layers gained strength at about 100 Pa d^{-1} (Figure 12). *Jamieson et al*. [2001] showed that in conditions characterized by a deep snowpack, the shear strength was best correlated with the overlaying load. They suggested loading was the main factor promoting increased strength through a pressure-sintering process [*Gubler*, 1982] by which the number and/or size of bonds progressively increases with time under load. Rapid loading produces instability more often than gradual loading, indicating that during rapid loading the strength of weak layers tends to lag behind the load on the order of days [*Chalmers*, 2001]. Such effects need to be included in models of weak layer strength and slab stability over time.

[37] The cantilever test of unnotched snow beams can be used to assess the slab properties in combination with a stability test of the weak layer [*Mears*, 1998; *Perla*, 1969]. *Johnson* [2000] improved the cantilever beam test and studied remotely triggered avalanches. Compared to avalanches triggered in steep starting zones, avalanches remotely triggered from low-angle terrain tended to have thicker, denser, and harder slabs.

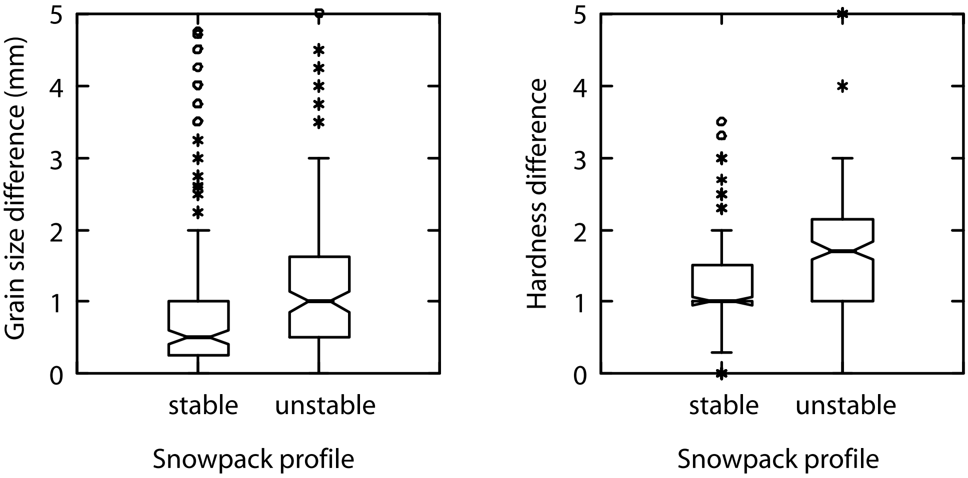

Figure 11. Comparison between 194 unstable (skier triggered) and 207 stable profiles. Statistically significant differences in (left) grain size (U test, $p < 0.001$) and hardness (U test, $p < 0.001$) across fracture interface are shown. Boxes span the interquartile range from first to third quartile with a horizontal line showing the median. Notches at the median indicate the confidence interval ($p < 0.05$). Whiskers show the range of observed values that fall within 1.5 times the interquartile range above and below the interquartile range. Asterisks show outliers; circles show far outside values (reprinted from *Schweizer and Jamieson* [2003b] with permission from Elsevier).

[38] Brittle fracture and fracture propagation are essential parts of snow slab release (see section 3.1). *Kirchner et al.* [2000] measured the fracture toughness of snow under tension. Notched cantilever beams of snow (20 cm $\times$ 10 cm $\times$ 50 cm in size) were broken under their own weight in the field at temperatures close to the melting point. Their data included newly fallen snow and snow consisting of melt-freeze grains. The critical stress intensity factor K_{Ic}, characteristic of brittle fracture, varied with relative density as

$$K_{Ic} = B\left(\frac{\rho_{\mathrm{snow}}}{\rho_{\mathrm{ice}}}\right)^{2.3} \tag{4}$$

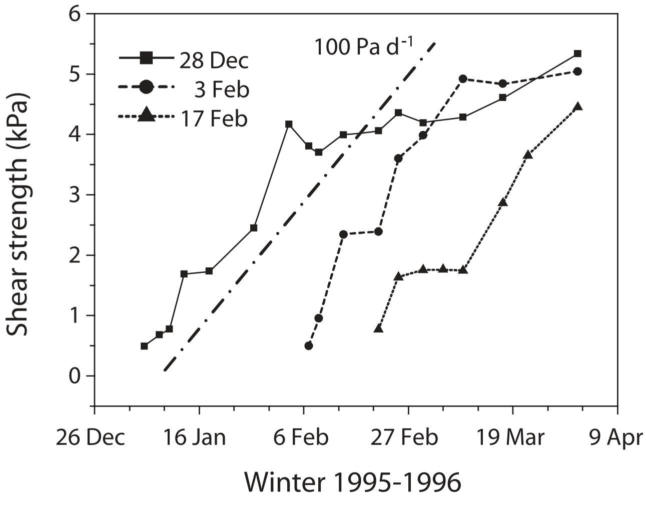

Figure 12. Strength changes of three surface hoar layers. Dates indicate day of burial at Mount Fidelity (Columbia Mountains, Canada). Dashed-dotted line indicates average rate of strength increase: about 100 Pa d^{-1}.

with $B = 7.84$ kPa m$^{1/2}$. *Kirchner et al.* [2000] suggested these extraordinarily low values of fracture toughness indicate snow is one of the most brittle materials known. *Kirchner et al.* [2002a, 2002b] assessed the effect of external loading and friction under shear. Homogeneous (nonlayered) snow samples under laboratory conditions were tested. However, before fracture toughness can become a relevant parameter for assessing snow slope stability, measurements for more different snow types are needed, and in particular, layered samples need to be tested. Furthermore, a field test must be designed to complement standard stability tests (see below).

[39] In view of the importance of snow cover stratigraphy for avalanche formation, numerical modeling of the snow cover evolution from meteorological measurements is a key research subject. The French model Crocus [*Durand et al.*, 1999] is part of the operational forecasting model chain SAFRAN-Crocus-MÉPRA, which includes stability evaluation and simplistically takes into account terrain (aspect and elevation). The Swiss model SNOWPACK [*Lehning et al.*, 1999] (Figure 13) is primarily microstructure-based and simulates the snow cover evolution in level study plots. It runs operationally to calculate parameters such as the new snow depth and drift index for the Swiss avalanche warning service (see section 2). A stability evaluation tool is under development [*Lehning et al.*, 2003]. At present, stability interpretation from manually observed snow profiles surpasses that from modeled or penetrometer profiles [*Schneebeli and Johnson*, 1998]. In general, snowpack evolution is well simulated with regard to snow depth, bulk density, and snow temperature [*Durand et al.*, 1999; *Lehning et al.*, 2001]. Verification of

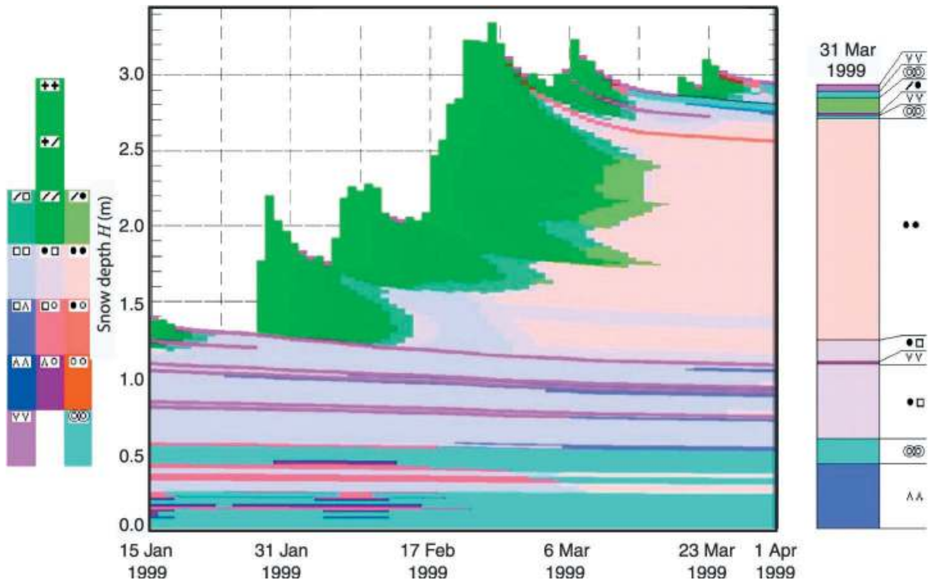

Figure 13. Simulation of snow cover evolution. Simulated grain type within snow cover during 2 and one-half months in winter 1999 for the Weissfluhjoch study plot 2540 m above sea level (Davos, Switzerland) are shown. On the right, snow stratigraphy from a manually observed profile from 31 March 1999 is given for comparison. During the 2- and one-half-month period, three major snowfalls occurred. Before the storm periods, during fair weather conditions, substantial parts of the snowpack had been unfavorably transformed to faceted grains because of increased temperature gradients within the shallow snowpack. During the subsequent storm periods several avalanche cycles caused numerous avalanche fatalities in the French, Swiss, and Austrian Alps (reprinted from *Lehning et al.* [2002] with permission from Elsevier).

snow stratigraphy is limited by the lack of objective snowpack data. Comparison with manual profiles showed that most layers were correctly modeled in terms of grain type and size. However, average stratigraphy might not be relevant for stability evaluation, since for avalanche formation, snowpack weaknesses are essential. Most of these result from the complex mass and energy balance at the snow surface and are more difficult to simulate.

[40] Other than avalanche occurrence data, snowpack stability tests and tests of weak layer strength provide the only direct information on snowpack characteristics relevant for avalanche formation. *McClung and Schweizer* [1999] briefly described the rutschblock test, the shovel shear test, and the shear frame test. Other relevant tests used are the compression test [*Jamieson*, 1999] and the stuff block test [*Birkeland and Johnson*, 1999] (Figure 14). Most tests identify potential weak layers or interfaces and give an index of snowpack stability, provided the effect of the slab properties is considered. In situ snow slope stability testing usually involves isolating a snow column and loading the surface of the column at specified steps. Therefore slab properties and weak layer properties are tested in combination. All common tests try to reproduce, to various degrees, the dynamic loading by a skier or snowboarder, and they cause brittle fracture in weak layers. Stability tests are most indicative on slopes, but some can be done in gentler terrain to avoid exposure to avalanche danger. Larger stability test areas provide more reliable results [*Föhn*, 1987a; *Tremper*, 2001]. None of the tests provides direct information on fracture propagation propensity. However, all methods test a size comparable to that required for self-propagating brittle fractures due to rapid loading (0.1–1 m) [*Schweizer*, 1999]. The type of failure (e.g., planar, partly nonplanar, etc.) as observed is essential and may serve as a first indicator of the propagation propensity [*Johnson and Birkeland*, 2002]. In particular, for the rutschblock test the type of release (whole block versus partial release) is also indicative of propagation propensity [*Schweizer*, 2002]. Besides mechanical stability information from stability tests, snow stratigraphical characteristics, such as changes in hardness or grain size at layer boundaries, proved predictive for evaluating snowpack stability with regard to skier loading [*Schweizer and Jamieson*, 2003b; *McCammon and Schweizer*, 2002].

3. MECHANICAL MODELS

[41] The dry snow slab avalanche is the result of four types of failures, leading to five fracture surfaces: one in tension at the top of the slab (crown), two lateral breaks on the sides of the slab (flanks) mostly in shear, one compressive failure at the lower end (stauchwall), and a failure between the slab and the supporting substratum [*de Quervain*, 1966] (Figure 3). It is clear that the primary failure is between the slab and the substratum, commonly in slope-parallel shear, but occasionally, compressive failure leads to the loss of shear support, com-

Figure 14. In situ snow slope stability testing usually involves isolating a snow block including a weak layer and loading the block at given steps: (a) rutschblock test, (b) compression test, and (c) stuff block test [from *Schweizer and Jamieson*, 2003a].

parable to adhesive versus cohesive fracture in layered materials [*Wei et al.*, 1996]. The compelling argument by *Perla and LaChapelle* [1970] that shear failure at the base of the slab precedes tensile fracture through the slab is based on observations that the tensile fracture surface is perpendicular to the slope.

3.1. Slab Release Models

[42] The simplest model compares the shear strength of the weak layer to the shear stress due to the overlaying slab and any artificial near-surface load (equation (1)). For release by localized rapid near-surface loading, measurements on the effect of the skier on snow stability showed the skier's impact strongly decreases with increasing snow depth and that the decrease depends on slab properties (stiffness) [*Schweizer et al.*, 1995]. These results agree with the simplified model for skier loading as a line load on an elastic half-space [*Föhn*, 1987b; *Schweizer*, 1993], which was introduced in the stability index. The concept of the stability index proved to be successful for the case of skier triggering [*Föhn*, 1987b; *Jamieson*, 1995], in part successful for storm snow avalanches [*Conway and Wilbour*, 1999] but in general not successful for most other natural avalanches, i.e., under gradual or no-loading conditions. This indicates that, in line with linear fracture mechanics, a simple stress criterion is insufficient or even inappropriate for natural avalanches. Failure starts at locations of below average strength (flaws). Macroscopic size effects are associated with such fracture initiation. There is a critical flaw size needed for catastrophic failure of the whole structure or snow slab.

[43] Snow slab failure models [e.g., *Bader and Salm*, 1990] consider a two-dimensional inclined snowpack with an assumed prior weakness existing in the otherwise homogeneous weak layer (Figure 15). In this deficit zone the shear stress from the overlaying slab is not supported

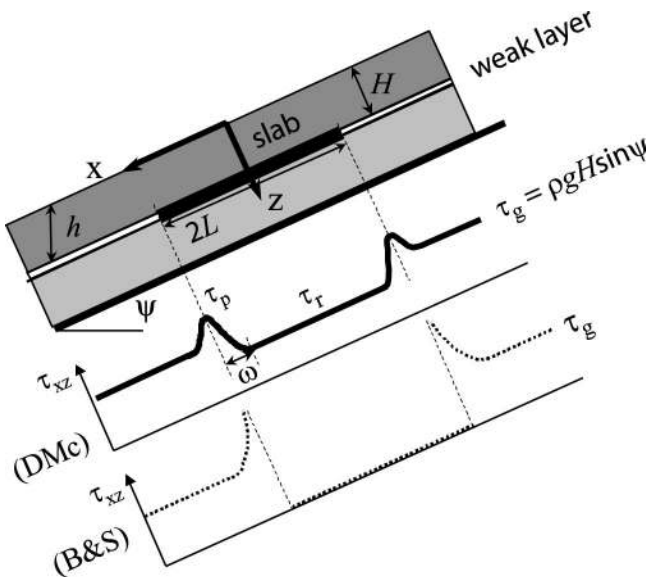

Figure 15. Snow slab release models with preexisting weakness (deficit zone or imperfection): two-dimensional inclined snowpack with slope angle ψ, slab thickness H, fracture depth h, and length of deficit zone or imperfection $2L$. Resulting distribution of slope-parallel shear stress τ_{xz} at weak layer depth z is shown for two models: *McClung* [1987] in the middle and *Bader and Salm* [1990] at bottom. For other parameters see text.

by the shear strength. Not much is known about the origin of this (assumed) preexisting deficit zone. There is no obvious process available for initiation of the stress concentration in the weak layer, except possibly strain softening during loading conditions. Conclusive experimental evidence for deficit zones is likewise lacking.

[44] *McClung* [1979, 1981, 1987] was the first to apply fracture mechanical principles. His work focused on ductile shear failure of the weak layer, followed by shear fracture and propagation, based on a model for the growth of a shear band (or slip surface) in an overconsolidated clay mass [*Palmer and Rice*, 1973]. A shear band is initiated at a stress concentration in the weak layer (Figure 15). Strain softening at the tip of the band follows. When a critical length L is reached, the band propagates rapidly. The approach is similar to a Griffith criterion, which leads to analogous results, with the difference that the stress at the tip of the band is nonsingular. The mode II propagation criterion is given by

$$\frac{(1-\nu)}{2G}K_{\mathrm{II}}^2 = \frac{H(1-\nu)}{4G}\left((\tau_g-\tau_r)\frac{L}{H}\right)^2 = (\tau_p-\tau_r)\delta\,, \tag{5}$$

where H is slab thickness perpendicular to slope, ν is Poisson's ratio, G is shear modulus, $\tau_g = \rho g H \sin\psi$, the shear stress due to the slab, where ρ is average slab density, g is acceleration due to gravity, ψ is slope angle, and δ is the displacement in the shear band from peak stress τ_p to residual stress τ_r. The two terms on the left are equivalent expressions for the driving term; the rightmost term provides the resistance to shear band extension. Assuming that the end zone length ω is small compared to the band extension L, the critical downslope length L_c for band extension can be given as

$$L_c = \frac{H}{\tau_g-\tau_r}\sqrt{\frac{4G}{H(1-\nu)}(\tau_p-\tau_r)\delta}\,. \tag{6}$$

The model can plausibly describe different slab release scenarios [*McClung*, 1979]. As for any slab release model involving a size effect, no verification data are available.

[45] On the basis of the energy balance approach, conditions for fracture propagation and hence for slab release can be calculated. The conditions primarily depend on the dimensions of the area over which the shear strength deficit exists. Accordingly, a critical size for fracture propagation can be given. In the model proposed by *McClung* [1979] (see above), the size of the end zone (or plastic zone) ω is considered as the minimal length to initiate any progressive failure process and is given by

$$\omega = \frac{9\pi G}{16(1-\nu)}\frac{\delta}{\tau_p-\tau_r}. \tag{7}$$

Using typical values of alpine snow combined with results from laboratory studies on shear failure [*McClung*, 1977; *Schweizer*, 1998] shows that the size of the end zone can be estimated to 0.2–2.2 m. The critical length of deficit zone for fracture propagation must be a multiple of the end zone size and decreases with increasing ratio of peak to residual stress and increasing loading rate. The rate dependence is consistent with results for concrete [*Bazant and Planas*, 1998] and is plausible in the light of skier triggering [*McClung and Schweizer*, 1999].

[46] In general, a critical length in the two-dimensional model between 0.1 and 10 m can be calculated, but this is only an estimate [*Schweizer*, 1999]. The lower range (0.1–1 m) is associated with slow growth; the higher is associated with fast growth (1–10 m). For the case of rapid loading the critical length for fast growth reduces to 0.1–1 m [*McClung and Schweizer*, 1999], i.e., to the order of the slab thickness. The spatial effect of a skier is of the same order, supporting the estimate of 0.1–1 m, since skiers frequently induce brittle, rapidly propagating fractures [*Schweizer and Camponovo*, 2001].

[47] Alternatively, critical crack size a_c can be estimated from preliminary measurements of the fracture toughness of snow in shear $K_{\mathrm{II}c}$ [*Kirchner et al*., 2002a; 2002b]:

$$a_c = \frac{1}{\pi}\left(\frac{K_{\mathrm{IIc}}}{\tau_g}\right)^2. \tag{8}$$

For typical values of $K_{\mathrm{II}c}$ and τ_g the critical crack size a_c is approximately 0.3–1 m. As these values are determined with linear fracture mechanics, they represent lower limits since energy dissipated because of the plasticity of the material is not included.

[48] The fact that the models assume a preexisting weakness of unknown origin raises the question of fracture nucleation [*Nye*, 1975]. *de Quervain* [1966] pointed out that to form the nucleus of fracture, stress must locally exceed strength. Accordingly, *Schweizer* [1999] argued that slab release should initially start with damage at the microscale rather than with a deficit zone at the mesoscale. He further proposed to model the failure process based on two fundamental processes at the microscale: bond fracturing and sintering (bond formation). By introducing variability in microstructure, damage could accumulate, leading to failure localization and finally fracture propagation. This leads to the unsolved question of how microstructure is related to strength (or the mechanical properties in general) (see section 3.2).

[49] The fracture mechanical approach has recently been revisited [*Louchet*, 2001b]. The stability of a basal crack was analyzed as a Griffith problem depending on whether shear crack propagation is quasi-static or unstable. In the case of quasi-static expansion the slab would first meet the stress instability criterion in tension. This release mode, called undercritical triggering by *Louchet* [2001b], assumed a slowly expanding shear failure up to the moment when tensile fracture takes place, as previously proposed by *Perla and LaChapelle* [1970]. The second scenario required that the basal crack meets conditions for unstable crack growth before the tensile

stress at the tip of the basal crack reaches the tensile strength. In that case the critical crack size can be estimated (equation (8)). *Louchet* [2001b] argued that independent of snow characteristics, the transition between the two triggering modes would occur at a critical slope angle of about 35°. If residual strength in the basal failure was considered, this critical slope angle increased. There are no observations or data available to support the proposed transition in triggering modes. In fact, assuming realistic values for density, slab thickness, tensile strength, and toughness, Louchet's model suggests triggering would be primarily of the undercritical type, which contradicts experimental evidence in which the tensile fracture through the slab thickness is often first observed well above the skier (trigger point).

[50] *Åström and Timonen* [2001] proposed a snow slab failure model based on statistical variation of strength. The slab was modeled as a two-dimensional square lattice of beams, connected to the substrate by other beams. These connecting "friction beams" modeled the static frictional contact between the substrate and the slab. Both types of beams failed at a threshold value. The fracture (slip) thresholds varied according to a statistical distribution such as the Weibull or modified Gumbel [*Duxbury*, 1990]. *Åström and Timonen* [2001] concluded that the grade of heterogeneity in the local fracture (slip) threshold and the ratio of the average substrate slip threshold to the average slab fracture threshold were decisive for fracture behavior. Although the model is neither realistic nor supported by data, consideration of strength variation is important.

[51] There are few attempts to model snow slab failure physically or numerically. Physical models using granular materials such as sand on inclined planes are so far of little relevance for avalanche formation [*Daudon and Louchet*, 2001], but they were successfully applied for modeling avalanche motion as granular flow [*Savage*, 1993]. Most numerical modeling has used finite element models to calculate stress, strain, and strain rates for complex geometry, including layering and irregularities in the snowpack [*Bader et al.*, 1989; *Bader and Salm*, 1990; *Schweizer*, 1993; *Wilson et al.*, 1999]. Other numerical models have used the finite difference method to solve the stress-strain relations for snow stability conditions using a simple stress failure criterion and a simplified concave slope [*Schillinger et al.*, 1998]. Implementing improved material properties and stability criteria will be the first step toward more realistic modeling of slab release [*Stoffel and Bartelt*, 2002]. However, numerical and physical modeling is presently of limited importance for the advancement of understanding and for practical applications.

3.2. Microstructure

[52] Snow is a porous material consisting of crystalline ice particles (or grains) welded together. The microstructure scale is of the order of 10^{-4} m. The microstructure describes the size, shape, and arrangement of grains that cannot be seen by the naked eye. Classical snow characterization [*Colbeck et al.*, 1990] using a 10-power hand lens focuses on grain type and size. The mechanical properties are determined by the arrangement of grains and particularly by the size and number of bonds. These characteristics, however, usually cannot be seen with a common hand lens. Whereas the term microstructure is commonly used in engineering, an alternative term (snow texture) is more commonly used in the geosciences [*Arons and Colbeck*, 1995].

[53] Snow can be considered as cellular solid rather than as a granular material. It is a sintered material, and for low densities has foam-like properties [*Kirchner et al.*, 2001]. However, as the microstructure changes with increasing density, the foam concept is probably not applicable for the whole density range (50–500 kg m^{-3}) of seasonal snow. The importance of microstructure (texture) to deformational processes has long been known, and the large scatter in plots of mechanical properties versus density [*Mellor*, 1975] has been attributed to the influence of texture. *Voitkovsky et al.* [1975] showed that cohesion correlated better with specific grain contact surface (total cross-sectional area of the bonds per unit bulk area) than with density. *Kry* [1975a, 1975b] and *Gubler* [1978b] first tried to relate mechanical properties to microstructure, in particular to bond size. For further details on microstructural studies the reader is referred to the work of *Shapiro et al.* [1997]. The bonds, or, more precisely, the size and number of bonds per unit volume and their orientation, should be related to strength. *Agrawal and Mittal* [1996] suggested that the degree of bonding between grains and the pore length (or the mean free distance between grains) should control the mechanical properties. Classical characterization of snow does not consider bonding, but size and shape of crystals/grains have some relation to bonding. Larger grains, in particular irregular angular (persistent) grains, usually have fewer bonds per grain (coordination number) as well as fewer bonds per unit volume. Accordingly, they form layers of low strength. *Jamieson and Schweizer* [2000] showed that shear strength of surface hoar increased with layer thinning, which they argued was associated with penetration of crystals into adjacent layers and increased bonding (Figure 16).

[54] Until recently, producing thin or surface sections and applying stereological methods to derive structural parameters has been the standard procedure for characterizing microstructure [*Dozier et al.*, 1987; *Good*, 1987]. *Schneebeli* [2001] improved the method by reconstructing three-dimensional (3-D) representations from serial sections (Figure 17). Alternatively, snow samples have been visualized by x-ray microtomography [*Coléou et al.*, 2001; *Schneebeli*, 2002], scanning electron microscope [*Adams et al.*, 2001], and nuclear magnetic resonance imaging [*Ozeki et al.*, 2002]. Visualizing the snow microstructure with modern imaging technology is only the first step. Quantitative description and interpretation of mechanical properties from 3-D images remains

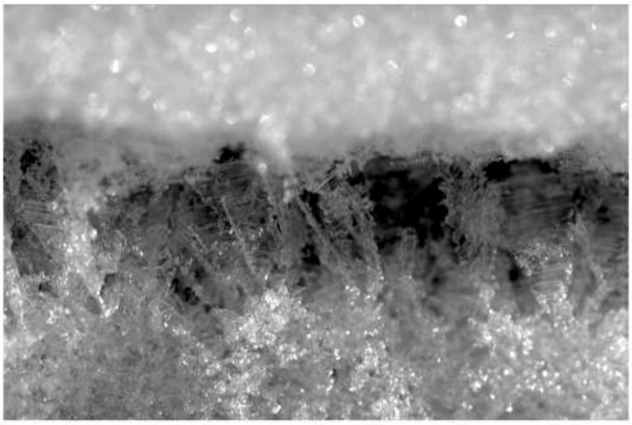

Figure 16. In situ photograph of buried surface hoar layer (layer thickness of 10 mm) sandwiched between two layers with different snow properties, exemplifying the importance of microstructure, especially bonding.

to be done. Local curvature and surface area, two essential microstructural parameters that influence metamorphism, can now be determined [*Brzoska et al.*, 2001; *Flin et al.*, 2001]. Microtomography during snow metamorphism or deformation experiments should improve our understanding of microstructural processes [*Schneebeli*, 2002]. During the deformation process, grains continuously rearrange accompanied by bond fracture and formation. Except for very small strains (about $\leq 10^{-4}$), texture changes [*Camponovo and Schweizer*, 2001], greatly complicating the development of constitutive equations [*Bartelt and von Moos*, 2000].

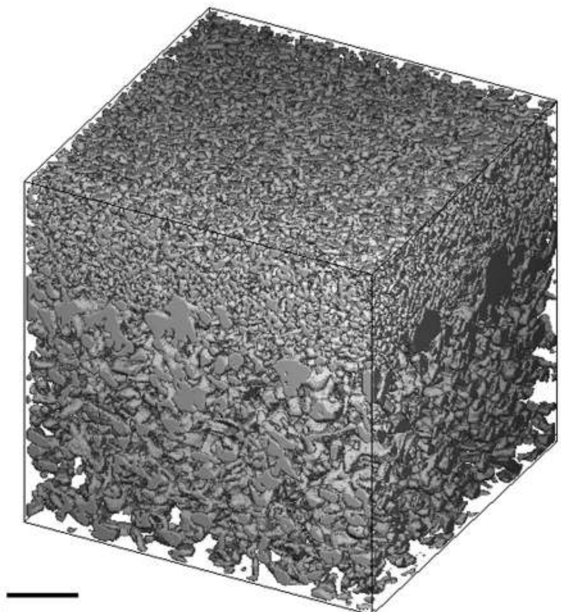

Figure 17. Three-dimensional image of snow microstructure reconstructed from surface sections. Vertical distance between surface sections is 15 μm. The sample taken from the snowpack consists of three snow layers: (top) small rounded grains, (middle) melt-freeze crystals, and (bottom) large faceted crystals. The potential fracture interface is below the melt-freeze layer at the transition to the large faceted crystals.

[55] *Mahajan and Brown* [1993] made one of the first attempts to include microstructure in their constitutive law for snow. After simplification it proved applicable for modeling snow cover evolution [*Lehning et al.*, 1999]. *Bartelt and von Moos* [2000] have used the same approach to interpret viscosity values from triaxial tests. To characterize bulk properties, they used ice properties and geometric parameters for microstructure, e.g., the reduced contact area (the total area of bonds per unit area or volume). Accordingly, the stresses in the bonds are many times the macroscopic stress, which explains why snow fails at much lower stress than ice despite the fact that it consists of the same material.

[56] *Shapiro et al.* [1997] recommended index properties since no instrument would be readily available to determine the required textural information. Recently, a high-resolution constant speed penetrometer was developed [*Schneebeli and Johnson*, 1998; *Schneebeli et al.*, 1999]. Characterizing snow as a foam [*Gibson and Ashby*, 1997], *Johnson and Schneebeli* [1999] developed a micromechanical theory of penetration and used the penetration force-distance signal to recover microstructural and micromechanical properties (Figure 18). The coefficient of variation of the high-resolution signal was related to the ratio of grain size to density (texture index). Probing the snowpack reveals highly complex force-distance signals including more information than from manually observed hardness profiles [*Pielmeier and Schneebeli*, 2002] (Figure 19). Although identifying layers and snowpack weaknesses is not straightforward, the resistance gradient is suggested to indicate the stiffness gradient and hence stress concentrations much better than in manual profiles.

3.3. Snow Failure

[57] Snow deformation and failure are highly rate- and temperature-dependent. The shear strength decreases with increasing strain rate and temperature. The ductile to brittle transition occurs at a strain rate of about 10^{-4}–10^{-3} s^{-1} [*Fukuzawa and Narita*, 1993; *McClung*, 1977; *Narita*, 1980; *Schweizer*, 1998] (Figure 20). The bulk rate and temperature dependence can be explained based on the properties of ice and an interpretation of deformation and failure at the scale of bonds in terms of the competing effects of damage (fracturing of bonds) and formation of bonds (sintering) [*Schweizer*, 1999]. The effect of sintering in shear deformation must decrease as deformation rate increases (since some time is needed for sintering) and as temperature decreases (since sintering is highly temperature-dependent, unlike fracturing). Accordingly, *Louchet* [2001a] developed a micromechanical model in which the weak layer was treated as open cell foam made of a network of ice bonds. These bonds were prone to break under stress, but broken bonds may reconstruct if in contact with each

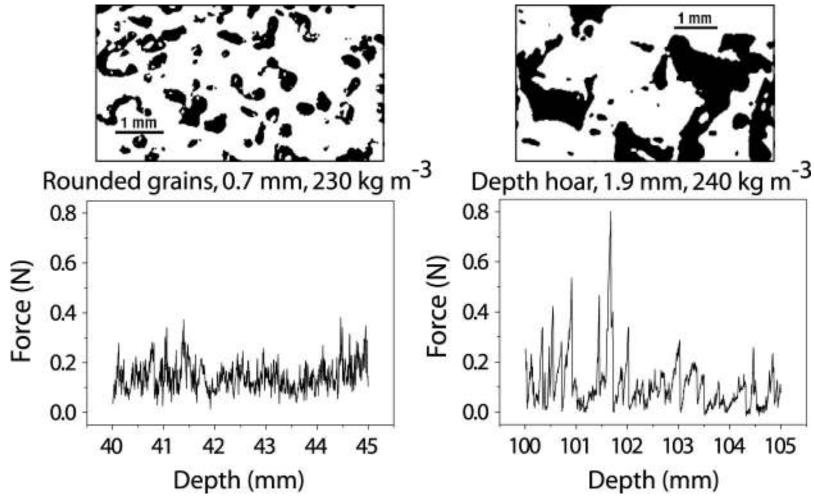

Figure 18. (bottom) High-resolution, constant speed penetrometer force-distance signals for two different snow types: (left) small rounded grains and (right) depth hoar. (top) Corresponding surface sections illustrating the microstructure of the two snow types. The variation in the force signal includes the textural information: The higher the variation, the larger is the grain size of the sample tested (reprinted from *Schneebeli et al*. [1999] with permission from Elsevier).

other for some time. Accordingly, *Louchet* [2001a] described creep in relation to the bond-breaking rate

$$\dot{n}^- = \alpha \frac{\tau}{n} n = \alpha\tau \tag{9}$$

and the bond-welding rate

$$\dot{n}^+ = \beta(1-n)^2 \frac{1}{\dot{\gamma}}, \tag{10}$$

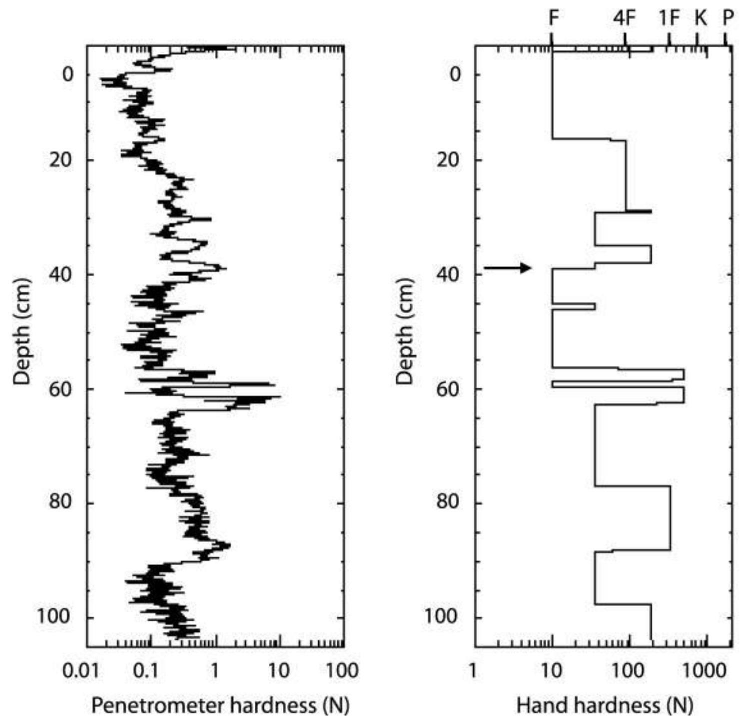

Figure 19. Comparison of (left) penetrometer hardness signal to (right) hand hardness index. Profiles were taken on 15 January 2002. Hand hardness index (F, fist; 4F, four finger; 1F, one finger; K, knife; and P, pencil) is given on top right horizontal axis. A rutschblock stability test revealed a not very critical weakness at 39 cm below the snow surface (arrow) [after *Pielmeier and Schneebeli*, 2002].

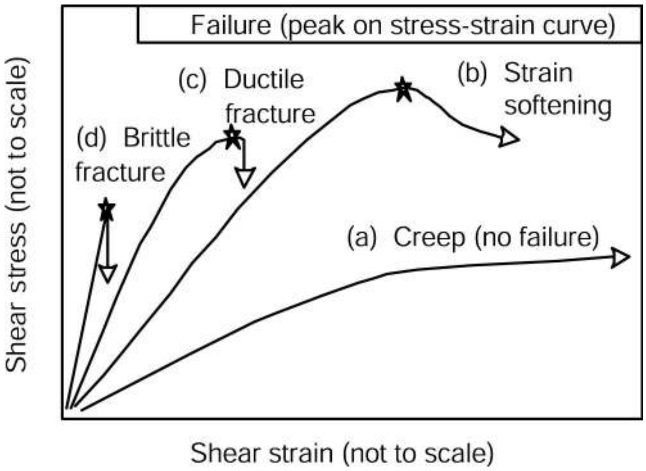

Figure 20. Schematic of deformation and failure/fracture for snow under shear loading. Strain rate increases in order for the curves labeled a, b, c, and d (based on data of *Fukuzawa and Narita* [1993], *McClung* [1977], *Narita* [1980], and *Schweizer* [1998]).

where n is the fraction of unbroken bonds, τ is the local shear stress, α is a factor accounting for the ice strength, β is a factor accounting for ice welding kinetics and including an Arrhenius-type temperature dependence, and $\dot{\gamma}$ is the shear strain rate. Sintering under compressive loading was neglected. The bond-breaking rate was independent of temperature and proportional to the number of unbroken bonds and the locally acting shear stress. The bond-welding rate was proportional to the square of the broken bond fraction, inversely proportional to the strain rate, and proportional to an Arrhenius factor for temperature dependence. Depending on the load, the net balance between breaking rate and welding rate should lead to stable, unstable, or critical conditions of slope stability. If the weak layer was not considered homogeneous at a scale larger than the porosity (bond spacing), the creep rate locally increased because of stress concentrations at the tip of areas of below average strength.

[58] *McClung* [1996] proposed a model of progressive microfracturing based on ice properties only, so the temperature effect on snow strength becomes plausible. However, the model did not include sintering. *Gubler and Bader* [1989] presented one of the first realistic models of snow failure based on microstructure. Assuming ductile failure and realistic microstructural parameters [*Gubler*, 1978a, 1978b], *Gubler and Bader* [1989] simulated initial stability during precipitation periods using a micromechanical model (including a statistical distribution of strength) for weak low-density snow combined with a relation describing the strength increase by settling and sintering as a function of snow temperature.

[59] Bond fracturing can be measured by counting the corresponding acoustic emissions [*McClung and Schaerer*, 1993; *Sinha*, 1996; *Sommerfeld and Gubler*, 1983]. Higher acoustic emission rates from natural snowpacks prior to avalanche release were suggested, but experimental problems and uncertainties hinder conclusions on the applicability for snow slope stability evaluation [*Sommerfeld*, 1982]. The other important bond-scale process involved in snow failure, sintering, is more difficult to quantify during a deformation experiment. However, with simultaneous x-ray microtomography this might become possible.

[60] Analyzing snow stratigraphy suggests that differences in grain size and hardness are indicators of failure probability [*Schweizer and Jamieson*, 2003b]. A large difference in grain size ($\geq$0.75 mm) between adjacent layers leads to a substantial difference in bond area per unit area within a short distance across the interface. In the coarser layer this causes stress concentrations in the spaced apart bonds near the interface. Hardness differences further promote this effect that leads to interface or subinterface failure, consistent with interfacial fracture mechanics [*Hutchinson et al.*, 1987]. The concentration of deformation in weak layers has been observed both for artificially made layered samples tested under shear [*Fukuzawa and Narita*, 1993] and on slopes based on observed downslope tilting of surface hoar crystals [*Jamieson and Schweizer*, 2000].

3.4. Fracture Propagation

[61] Fractures in snow are probably mixed mode fractures. At the scale of the slab thickness, shear is essential for fracture propagation in a layered sloping snowpack and is therefore assumed in modeling, consistent with field observation of fracture planes of snow slab avalanches. At the scale of the grains and bonds, fracture can be any mode.

[62] Prior to the release of a slab avalanche, conditions for fracture propagation in the weak layer/interface have to be met. Fracture propagation can be observed and frequently heard (the characteristic "whumpf" sound), and fracture speed has been measured [*Johnson*, 2000]. Remotely triggered avalanches are a special case in which fracture propagation is obvious. However, even after fracture propagation in the weak layer, the slab may not release since slab boundary strength may not be overcome and/or parts of the slab may not be inclined steeply enough to overcome bed surface friction. On the other hand, experience shows that frequently snowpack weaknesses are found, and stability tests suggest low stability, but no fracture propagation occurs, or locally induced fractures do not meet conditions for extensive fracture propagation. This indicates that traditional stability tests miss important information on fracture propagation potential and hence on slab release probability.

[63] Fracture propagation has been measured through weak layers on horizontal terrain [*Johnson et al.*, 2003]. A propagation velocity of about 20 m s^{-1} was measured with geophones. *Johnson et al.* [2003] concluded that compressive fracture (collapse) of the weak layer on low-angle terrain provided the energy needed for fracture propagation and that the velocity of the

resulting bending wave in the overlying slab depended on the stiffness of the slab. Accordingly, the bending wave induced a vibration in the air above the snow cover, thereby causing the "whumpf" sound. Modeling of fracture propagation in collapsing weak layers will require microstructural models rather than solid mechanics.

[64] *Bader and Salm* [1990] explored fracture propagation and concluded that once brittle fracture occurs, the speed of fracture would be of the order 100–1000 m s^{-1}, which is higher than observed in limited field measurements [*Johnson et al.*, 2003]. The extent of fracturing L_{slab} in their model is entirely controlled by the tensile strength σ_f of the overlying slab:

$$L_{slab} = \frac{\sigma_f}{\tau_g} H . \tag{11}$$

[65] On the basis of interfacial fracture mechanics, *Schweizer and Camponovo*, [2001] suggested that fracture propagation would depend on the difference in stiffness between the weak layer and the slab or, more precisely, the weak layer and the layer just above it. Interfacial fracture mechanics has been applied, for example, to characterize the failure of ice/substrate interfaces. For such bimaterial interface cracks, crack growth is always mixed mode [*Wei et al.*, 1996]. It is likely that interfacial fracture mechanics can be applied to snow slab release where fractures between layers of the same material but with different properties exist. The fracture resistance of bimaterial interfaces can be determined for beam specimens. According to *Wei et al.* [1996] the critical interface fracture energy G_c^{int} is given by

$$G_c^{int} = \frac{M_f(1-\nu_{WL}^2)}{2E_{WL}}\left(\frac{1}{I_{slab}} - \frac{\lambda}{I_c}\right) \quad \lambda = \frac{E_{WL}(1-\nu_{slab}^2)}{E_{slab}(1-\nu_{WL}^2)}, \tag{12}$$

where M_f is the applied bending moment at the initiation of the crack, I is the moment of inertia, and subscript c refers to the composite beam. Further evaluation of equation (12) reveals that the nondimensional interface fracture energy G_c^{int} increases monotonically with increase in relative slab thickness, H_{slab}/H_{WL}, and decreases as the relative modulus of the weak layer, E_{WL}/E_{slab}, increases. The smaller the interface fracture energy (because of a large difference in stiffness), the more likely fracture propagation becomes.

[66] Accordingly, and based on observation, it is proposed that fracture initiation at the base of a snow slab is an interface fracture. The fracture is expected to always be between two snow layers that are poorly connected. For a thin weak layer the fracture should start at the interface with the layer above or below. However, for fracture propagation the full thickness of the weak layer can be involved. Occasionally, cohesive fractures, such as the collapse of a thick weak layer of depth hoar, are observed.

4. SNOWPACK VARIABILITY

[67] As described in section 3.1, disorder is fundamental for the fracture process [*Herrmann and Roux*, 1990]. The behavior, especially failure, of a material cannot be based on its average properties. Variations are essential because they provide the nucleus of fracture and a necessary stabilizing mechanism to limit damage and inhibit fracture localization. Spatial variability of snowpack properties is believed to be a key issue for understanding avalanche formation. However, neither the scale of variation nor its effect on avalanche formation is clear. High spatial variability might offer numerous points of fracture initiation but might also limit fracture propagation. On the other hand, low variability seems favorable for fracture propagation, but in this case, failure initiation will depend on the level of stability. Intermediate snowpack stability with substantial variation at the scale of a few meters seems to be critical and hence challenging for the prediction of skier triggering [*Schweizer and Jamieson*, 2003a].

[68] There are several possibilities for variability in view of avalanche formation. There can be either variability of the slab properties, e.g., varying thickness, or variability of the weak layer properties. In particular, the weak layer may be discontinuous. For any case the scale of the variability is unknown so far.

[69] Variability is present at different scales in the snowpack. The scale depends on the specific conditions during formation: micrometeorology during deposition, wind effects due to small-scale topography (turbulence), large-scale topography, and climatic differences. Other sources of variability, particularly in shallow snowpacks, stem from variable properties of the underlying ground such as roughness and spatial variations in thermal conductivity. In the case of avalanches released by localized rapid surface loading, both snow stratigraphy and loading vary spatially. In general, when talking about variability the scale always needs to be mentioned.

[70] *Bloeschl* [1999] defined scale as a characteristic length of a natural process, e.g., the (unknown) correlation length of the spatial variation of the snowpack stability. Results derived from point measurements are strongly affected by the scales of sample size, sampling density, and areal coverage of the domain, which in the terminology of *Bloeschl* [1999] is the scale triplet of support, spacing, and extent.

[71] Whereas previous field studies [*Conway and Abrahamson*, 1984, 1988; *Föhn*, 1989; *Jamieson*, 1995] have not proven the existence of deficit zones (areas where the shear stress from the overlaying slab is not supported by the shear strength), they have clearly documented snowpack variability. It has been suggested that the snowpack variability at the scale of a region increases with increasing avalanche danger based on rutschblock scores from different aspects [*Munter*, 1997]. However, field studies on single slopes, as well as on the regional scale, have shown the opposite [*Jamieson*, 1995; *Kron-*

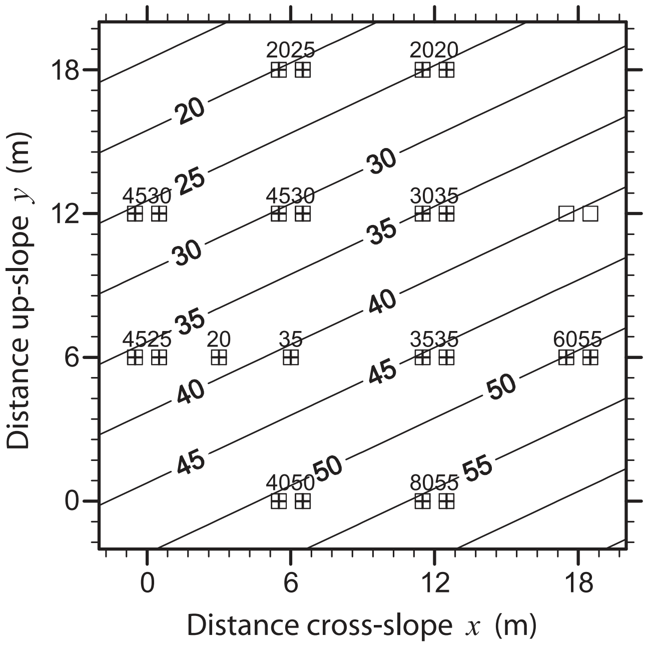

Figure 21. Results of stability measurements on a small avalanche slope. Stability test locations are marked by squares. A cross through a square marks a fracture at a specific weakness present at the date of measurement (14 February 2002). From the 24 stability tests that have been performed in pairs, 22 showed fractures at this weakness. The drop height (in cm) that is a measure of stability is shown above the test location. A linear slope stability trend was found and is indicated by 5-cm contours. The median drop height was 35 cm with a quartile coefficient of variation of 26% [from *Kronholm et al.*, 2002].

holm and Schweizer, 2003; *Schweizer et al*., 2003]. *Birkeland* [2001] conducted a large-scale study on two single days in an area of about 90 km^2 sampling data at over 70 sites. He found correlations between terrain (aspect and elevation) and stability, but the distance between sampling sites did not affect the correlation. *Kozak et al*. [2001] also considered differences at a relatively large scale, studying the effect of aspect and time on slab hardness. Measurements of snow stability on small avalanche slopes using a modified compression test showed that the relative variation of stability expressed as the quartile coefficient of variation was of the order of 50% and dropped to about 20% after removal of slope scale trends [*Kronholm and Schweizer*, 2003] (Figure 21). A similar study in a different snow climate found similar results [*Stewart*, 2002]. Areas of relatively low stability with dimensions from 60 cm up to 9 m were found. Most slopes with low median stability had low variability. *Landry* [2002] found more variation in a field study at a slightly smaller scale.

[72] Interpreting spatial variability in terms of fracture localization and propagation, *Kronholm and Schweizer* [2003] suggested that slope stability is controlled by the mean slope stability, by the spread of the stability on the slope, and by the scale of spatial patterns of strong and weak areas on the slope.

[73] Despite snowpack variability, weak layer formation is often quite consistent over whole mountain ranges [*Hägeli and McClung*, 2002]. This is supported by stability measurements in study plots that can be extrapolated and correlated with avalanche activity over considerable distances [*Jamieson*, 1995; *Chalmers and Jamieson*, 2003]. These findings are from a range with a deep-snow climate, the Columbia Mountains of western Canada, and from study sites near tree line. Above tree line, there is more wind effect and accordingly greater spatial variability.

[74] The importance of spatial variability is exemplified by the characteristics of avalanche initiation in forests. Beneath a tree the snowpack is highly variable, and weak layers are often discontinuous. However, in small clearings the variability is smaller than in the open field because of decreased wind effect, and avalanche initiation is likely, provided the clearing is large enough [*Gubler and Rychetnik*, 1991].

[75] Summing up, quantitative knowledge of the variability at different scales (simultaneously measured) and its relation to avalanche formation is still largely lacking. How large is the variability associated with certain meteorological and ground conditions, and what is the effect of that variability on snow slab stability? Variability at the scale of about 1–10 m may be important for fracture propagation since a weak layer or interface has to exist over a certain area so that a slab is released. Variability at the smaller scale ($\leq$10 cm) is probably related to failure initiation since it promotes local stress or strain rate concentration. The properties of the slab and the weak layer and their variation have to be considered in combination.

5. DISCUSSION AND CONCLUSIONS

[76] Our understanding of snow avalanche formation has significantly increased in the last 2 decades because of field studies and numerical modeling of the effect of the contributing factors: terrain, new snow, wind, temperature and radiation, and snow cover stratigraphy. Modeling the avalanche release mechanism (Figure 22) has proved to be more challenging. However, progress in characterizing the complex microstructure of snow, in particular the number and size of bonds, and field studies on spatial variability of the snowpack and its effects on avalanche formation will pave the way for detailed modeling of failure criteria for avalanche forecasting.

[77] The prerequisite for dry snow slab release is a snowpack weakness (thin weak layer or weak interface) below one or more slab layers. A distinctive difference in hardness and texture favors failure and causes stress and strain concentrations in the bonds connecting the two adjacent layers. Whereas failure usually starts at the interface between two adjacent layers, even when the weak layer is thin, fracture propagation can involve the full thickness of layers. If damage becomes localized, the

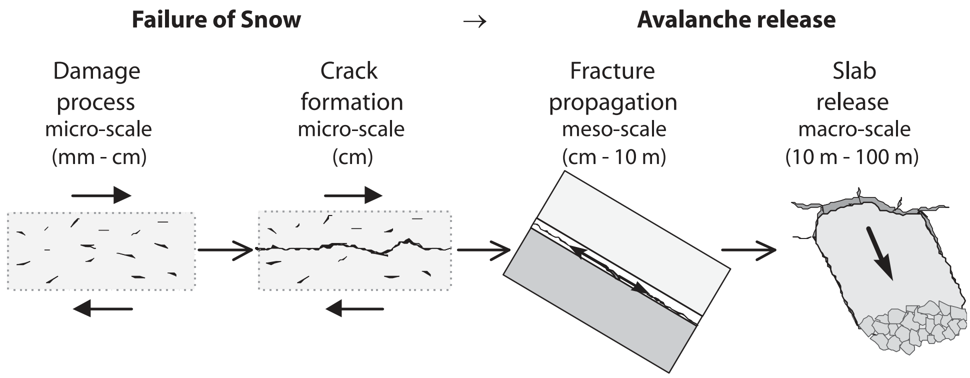

Figure 22. Conceptual model of dry snow slab avalanche release [from *Schweizer and Jamieson*, 2003a].

initial failure can propagate, and a slab may be released (Figure 22). The critical size for a self-propagating fracture is estimated to be of the order of 0.1–10 m, with the shorter lengths being critical for localized rapid surface loading. Whereas the snowpack configuration prone to failure is well established, less is known about conditions for fracture propagation that are strongly affected by, among other things, the spatial variability of the snowpack. Spatial variability of snow mechanical (and textural) properties is presently being investigated, and quantitative results are expected in the near future. Spatial variability provides failure initiation points and may arrest fractures. Surface warming of the sloping snowpack leads to increased deformation in the surface layer, thereby reducing the snow stability, provided the prerequisites for slab release are fulfilled. Although only mentioned briefly, wet snow avalanche formation is poorly understood. The textural properties may soon be characterized based on new means for 3-D imaging, with x-ray tomography being the most promising method. Texture can then be related to the mechanical properties. Present modeling at the microscale considers highly idealized structures but includes the two fundamental processes involved in snow failure: bond breaking and bond formation by sintering. Fracture mechanical models provide insight into the release condition at the mesoscale to macroscale. In situ measurements of fracture mechanical properties (such as toughness) should complete traditional stability evaluation. Modeling will need to close the gap between the microscale and the macroscale by linking improved micromechanical models (including statistical elements) to fracture mechanical models. Physical modeling, as well as numerical simulations of avalanche formation, lacks adequate parameterization of the physical processes, but promising attempts are under way.

[78] The quantitative understanding of snow avalanche phenomena is hindered by limitations common to other natural hazards. Avalanches are rare and are not reproducible events. Access to relevant study areas can be dangerous; however, safe study plots/slopes are helpful for operational forecasting. Field measurements are often not related closely enough to relevant processes or limited to a single process instead of covering all relevant parameters simultaneously. No remote device that directly indicates instability is available. Most field measurements include destructive methods thereby compromising studies of temporal evolution. Laboratory experiments with layered natural snow are difficult to perform because of the fragile nature of snowpack weaknesses and of low-density snow in general. Because of these limitations, numerical modeling is essential to advance our understanding but depends on data that are, as shown above, difficult to acquire.

[79] In order to approach a comprehensive understanding of snow avalanche formation we propose the following questions for future research for several areas, fracture propagation, spatial variability, effect of surface warming, wet snow avalanches, new snow stabilities, snow failure, and slab release. These areas are not ranked in order of importance, and the list is not exhaustive.

1. Which mechanical properties of which slab/weak layers describe the propensity for fracture propagation? How can practitioners test for propagation propensity?
2. What are the spatial scales of variability relevant to slab release as influenced by topography (aspect, inclination, distance to ridge, and ground cover), snow type, and meteorological conditions during/after deposition? In other words, what are the scale and distribution of areas of low stability and fracture tough areas and their effect on release probability?
3. How is snow stability affected (quantitatively) by surface warming?
4. How does failure initiation and fracture propagation occur for wet slabs?
5. What kind of slab and interface properties favor new snow avalanches? How can these properties be predicted by meteorological instruments?
6. How can the damage and sintering process leading to snow failure be measured and modeled? What should be measured? Is there a fatigue effect in snow damage/failure?
7. How can the sequence of events and processes leading to slab release (from the scale of 0.1 mm to 100 m) be measured, modeled, and verified? In partic-

ular, how can microstructural models be combined with fracture mechanical models?

[80] This review shows that dealing with an extraordinary material such as snow and a process that covers several orders of scale, from the size of a bond to the size of a mountain slope, will continue to be very challenging.

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