Creep and Failure of Alpine Snow: Measurements and Observations

H. Conway[1], S. Breyfogle[2], J.B. Johnson[3] and C. Wilbour[2]
[1] University of Washington, Geophysics, Box 351650, Seattle, WA 98195
[2] Washington State Department of Transportation, Box 1008, Snoqualmie Pass, WA 98068
[3] US Army Cold Regions Research and Engineering Laboratory, Ft Wainwright, AK 99703

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ABSTRACT

We investigated the creep behavior of alpine snow in an effort to help understand and predict the timing of avalanche release. Measurements of motion of glide shoes buried within a natural snowpack show strains within low density snow are typically large (often exceeding 70%). The rate of deformation increases with temperature and is especially rapid in the presence of liquid water. Creep rates decrease rapidly as the snow densifies. The slope-parallel shearing component of motion is much smaller than expected from the usual constitutive assumptions for snow. Even when snow is first wetted and on slopes up to 36°, the resultant direction of motion is typically close to vertical. We explain this apparently anomalous behavior by considering the effects of metamorphic processes and "capillary strain" (when liquid water is present) which cause deformation independently of gravity. It is well known that avalanche activity usually increases at the onset of rain, long before liquid water has penetrated to depth. We discuss how capillary induced shrinkage at the surface might alter the distribution of stress through the slab sufficiently to cause existing zones of deficit (or "super weak spots") to extend in length. A rain induced surface alteration occurs rapidly over a wide region and has the potential to perturb all existing zones of deficit simultaneously, thereby increasing the possibility of slope failure. The analysis predicts slope failure is more likely if the overlying slab is thin and the stability is already close to critical. Field observations of behavior at the onset of rain support this prediction.

BACKGROUND

Insitu measurements of creep on slopes have been used by others to establish constitutive laws for snow. Constitutive laws are needed to describe stress-strain relationships and to formulate models for predicting snow slope stability. Past measurements have been made by monitoring the motion of light-weight tracers such as ping pong balls (Perla, 1971) or changes in tilt of poles placed in snow (McClung, 1974). Most measurements have been made in well settled snow, and here we present high resolution measurements of deformation within low density snow. We are particularly interested in the effects of wetting - some measurements were made during rain-on-snow.

Models of dry snow slope failure suggest that two conditions must be satisfied before avalanches will release: the downslope component of the weight of the slab must be close to the average shear strength of a buried weak layer; the rate of deformation within the buried weak layer must be sufficiently high to cause failure (McClung, 1979, 1981; Bader and Salm, 1990). McClung (1981) analyzed two extreme types of avalanche release mechanisms: (I) - where increasing stresses caused by loading exceed the peak shear strength of a basal weak layer over some critical area; (II) - where conditions somehow become favorable to cause an existing flaw in the basal layer to propagate without additional loading. McClung pointed out that in reality, the likely mechanism of release for most dry snow avalanches probably lies between these two extreme scenarios.

Observations indicate that most natural dry snow slab avalanches release during storms as a result of rapid loading from snowfall (McClung and Schaerer, 1993). These avalanches are thought to be examples of type I behavior. Observations also indicate that avalanche activity usually increases within a few minutes of the onset of rain-on-snow. These avalanches typically release as slabs several hours before liquid water has penetrated to the sliding layer; the shear failure at the basal layer is within dry snow (Conway and Raymond, 1993; Heywood, 1988; Conway et al., 1988). The release mechanism is clearly different from the more commonly reported scenario for wet slab release where it is thought that infiltrating water has the effect of increasing the stress and weakening a sub-surface layer (eg. McClung and Schaerer, 1993). Further, the increase in gravitational loading from the additional weight of rain is usually small at the time of release; the type I contribution is small.

The mechanism of release of these immediate type avalanches has been a puzzle; it is not clear how the snow at the shear plane at depth knows that it is raining at the surface. From an operational perspective it is important to understand that these avalanches release much sooner than would be expected if loading and/or lubrication caused the failure. In principle it should be possible to predict the time of avalanching to within a few minutes by predicting the timing of transition from snow to rain. In practice, forecasting meteorological conditions in the start zones of avalanches is not always straightforward.

In this paper we present and discuss observations of snow stability and measurements of snow creep. We discuss the observed macro-scale behavior in context of grain-scale processes.

EXPERIMENTAL METHODS

Observations of weather, avalanche activity, snow stratigraphy and deformation were made in the Cascade mountains near Snoqualmie Pass, Washington. The terrain near Snoqualmie Pass lies between 900 m and 1700 m and mid-winter rain is common at these elevations. A typical snowpack in the region contains a relatively homogeneous base 2 to 3 m deep that has settled and grain-coarsened during one or more episodes of rain. Storms typically deposit up to 1 m or more new snow and subsequent rain often causes some or most of the new snow to avalanche.

Figure 1 shows the experimental setup used to measure snow deformation profile on slopes. Measurements were made near the middle of a 200 m long slope inclined at an angle of 30-40°. Snow depths of 2 to 3 m covered alder brush that had been bent down slope by the weight of the overburden. No avalanches released on the study slope during the measurement period. The vertical and slope parallel components of motion were measured separately using pairs of shoes made from light-weight aluminium screening. The screening minimized interruptions to water infiltration. Pairs of shoes were placed sequentially at the surface after 10 to 30 cm of snow had accumulated. One shoe was constrained to slide vertically down a fixed pole. A sliding contact mounted on the shoe made electrical contact with a resistance wire strung along the length of the pole. We configured this as a voltage divider circuit and shoe position could be calculated to an accuracy of ± 2 mm. The non-vertical component was measured by running a cord from a second shoe upslope to a 10-turn rotary potentiometer mounted on the vertical velocity shoe (Figure 1). Although the rotary potentiometers could resolve motions as small as 0.2 mm, additional uncertainties arise from the experimental setup. We estimate the overall accuracy of the measurement of downslope position to be ± 2 mm. Measurements, made at 5 minute intervals were recorded using a data logger.

FIELD OBSERVATIONS AND MEASUREMENTS

Below we present observations and measurements during two case histories to illustrate some salient points of rain-on-snow avalanche activity.

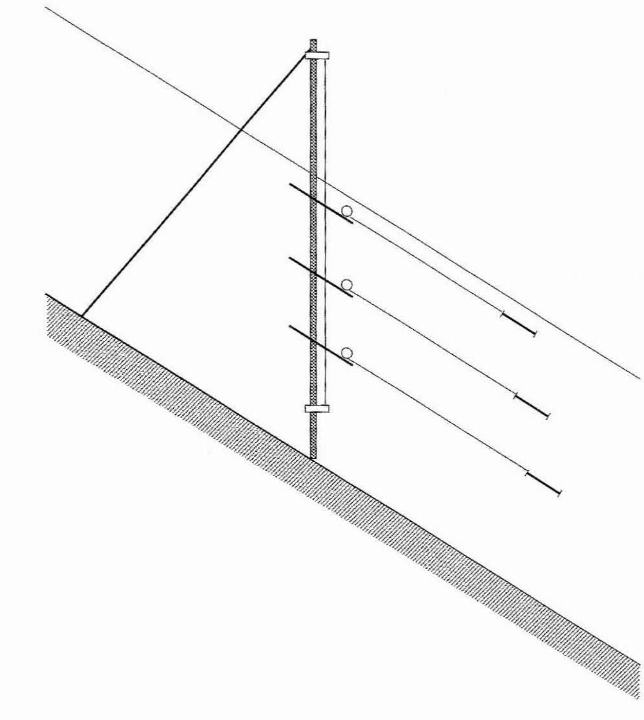

Fig. 1 Experimental setup used to measure creep profiles on slopes. Pairs of glide shoes were used to measure motion of a layer. One shoe was constrained to slide vertically down a fixed pole. A sliding contact on the shoe made electrical contact with a resistance wire strung along the length of the pole. Downslope motion was measured by running a cord from a second shoe upslope to a rotary potentionmeter mounted on the vertical velocity shoe. Measurements, made at 5 minute intervals, were recorded on a data logger.

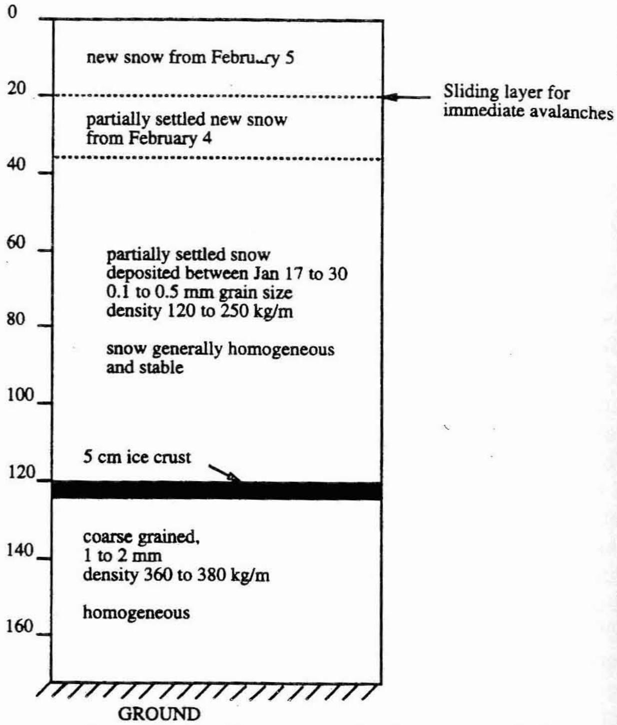

Figure 2. Snow stratigraphy in the study plot at Snoqualmie Pass just prior to rain on February 5,1996.

(i) Case Study February 5, 1996

The combined effects of intense snowfall followed by a classic warm-up and snow/rain transition on February 5, 1996 resulted in a major avalanche cycle at Snoqualmie Pass that closed the highway for 44 hours. Just prior to the onset of rain, cars and people were caught and buried by avalanches. One traveller who was out of his car was buried about 2 m below the surface for 29 minutes before being found by probing and dug out alive. Avalanches also released at the onset of rain. Slides hit and partially buried a snow blower and a vehicle occupied by two avalanche technicians who were spotting for the blower. In all, 11 vehicles and at least 20 people were hit by avalanches; more were dusted. None of these avalanches were large - the maximum depth of fracture was 20 cm.

Figure 2 shows the snow stratigraphy just prior to rain. Freezing rain on January 6-7 had formed a thick (5 cm) ice crust and rain on January 15 added to the crust thickness. A total of 263 cm of snow fell between January 17 to 30. Temperatures were generally cool and consistent throughout the period and the snowpack was remarkably homogeneous and stable. Conditions were clear and cold (-15 to -20°C) during for the next few days and by February 3 the new snow had settled to 95 cm. The general stability increased with time. Numerous snow pits and shear tests throughout the region indicated a strong bond between the new snow and the ice layer; shear tests through the full depth of the snowpack yielded either moderate to hard shears, or no distinct shear failure.

By morning on February 4, about 12 cm of new snow had accumulated. It cleared in the afternoon but snow showers started again on the morning of February 5. Snow continued to accumulate (10 cm by 1300; 19 cm by 1620 on February 5) and heavy snowfall continued through the evening. Natural avalanches first started to run at 1730, by which time just over 20 cm had accumulated. Snow changed to rain just before 1900, at which time avalanche activity increased again. These hazardous avalanches that released both before, and at the onset of rain were slabs 20 to 25 cm deep; only the snowfall from the past 15 hours avalanched at this time.

Although avalanche activity decreased considerably, continued rain during the next three days resulted in numerous natural and controlled avalanches. Most of these delayed avalanches released as wet loose slides. On several paths, control with explosives produced avalanches 38 hours after rain first started. These paths had been controlled (resulting in size 3-4 avalanches) 33 hours earlier; we think that liquid water had only penetrated the upper portion of the snowpack when the paths were first controlled. The heavy rain (400 mm in 3 days) produced the second 100 year flood in two months, and yet only one slab (triggered by explosives) was observed to release at the thick ice crust. A commonly cited mechanism for wet slab release is water lubrication on top of an impermeable ice layer. Our observations during this storm, and many others, indicate that this is rare in the Snoqualmie Pass region.

This storm cycle clearly illustrates a continuum of avalanche activity. The first avalanches released before warming, probably as a result of increasing stresses caused by rapid loading. Avalanche activity increased again at the onset of rain. In this case it is possible that the additional loading contributed to the instability but experience from other avalanche cycles indicates that other factors (discussed later) are also important. Although the frequency decreased, avalanches continued to release many hours after rain started. These wet snow avalanches are difficult to predict because it is difficult to define the evolving snow stratigraphy and pattern of infiltration. For practical purposes we have found it preferable to do control work at the onset of rain when the slope stability is most sensitive. If we wait, many paths will have released naturally and in addition, control becomes less effective after rain has started. Wet snow responds poorly to standard hand charges, and in our case, 105 mm rifle fire (~2.5 kg TNT), but we have had success using 10-20 kg bags of ammonium nitrate/fuel oil (ANFO) elevated a few meters above the surface.

(ii) Case Study January 24, 1993

Precipitation at Pass level changed to rain at 1500 on January 24. Rain, first on December 22, 1992 and again on January 3 had caused the snow deeper than 60 cm to settle and become relatively homogeneous. Just prior to rain about 55 cm of snow had accumulated on a thin layer of faceted crystals that had formed during cold weather earlier. A number of slabs released at the faceted layer, especially on steeper slopes, during the cold part of the storm. However activity increased considerably at the onset of rain and many slabs up to 60 cm deep released both naturally and artificially at this time.

Figure 3a shows the evolution of creep profiles that have been constructed from measurements of shoe positions on a 36° slope. Material lines have been drawn at hourly intervals to show the evolution of deformation in the snowpack over three hours centered around the onset of rain. The shoes deepest in the snowpack were set above a rain crust on December 22nd. The other shoes were placed sequentially after snow accumulated on January 3rd, 20th, 21st and 24th. The shoe near the surface was placed two hours before rain started.

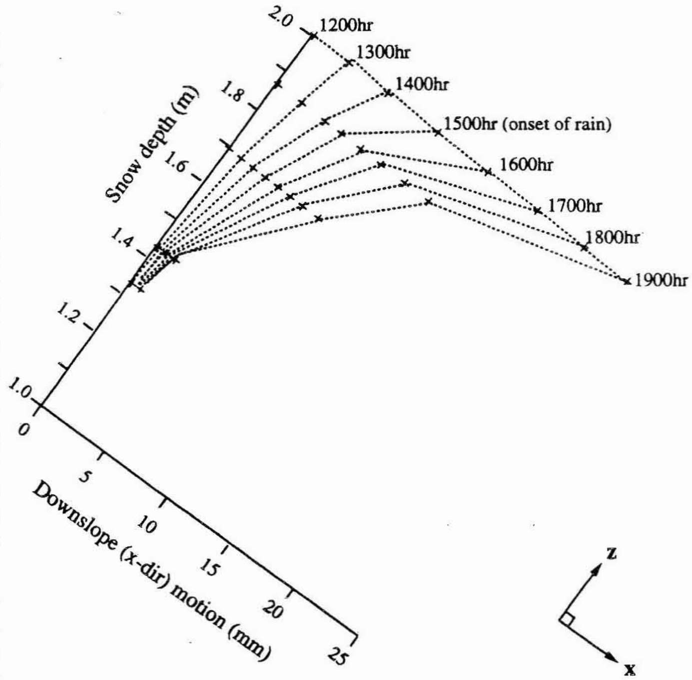

Fig. 3a Creep profiles at 1 hour intervals, starting at midday on January 24, 1993. Rain started at 1500hrs; many avalanches released as slabs at a depth of 1.45m soon after rain started.

The creep velocity at the surface more than doubled (from about 1 to 2.5 x10^{-6} m s^{-1}) when rain started. This is not unexpected because it is well known that introduction of liquid water causes rapid settlement of snow. Much more remarkable is the high creep angle observed for low density snow, even during first wetting. Note the scales in the x and z directions plotted in fig. 3a are different and in fact, the downslope motion was less than that in the z direction. In the near-surface snow, the direction of motion was within a few degrees of vertical. Figure 3b is a sketch of the direction motion of a particle P. Measurements during this and other events, indicate that high angles of creep are common in low density snow, even when liquid water is not present.

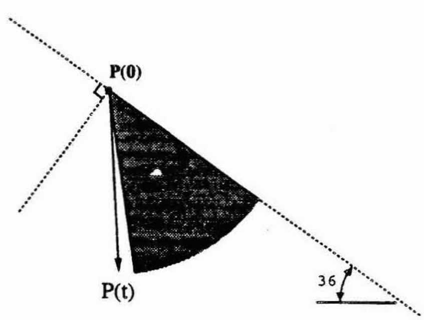

Fig. 3b The direction of motion was within a few degrees of vertical (shown above as motion of partial P). Also shown shaded is the range of motion expected using the usual constitutive assumptions for snow on a 36 degree slope. We think the apparently anomalous motion is a result of metamorphic and/or capillary strain acting independently of gravity.

SYNTHESIS AND DISCUSSION OF OBSERVATIONS

We suspect the apparently anomalous behavior discussed above is caused by metamorphic processes. Although metamorphism is fairly well understood on the grain scale (eg. Arons and Colbeck, 1995; Dash et al, 1996), the impact on macroscale behavior is less well understood. Below we discuss observed macroscale behavior in context of grain scale processes.

SHRINKAGE AND SINTERING

Thermodynamic processes act to minimize the surface energy of snow grains. The preferred shape is a sphere, hence grains become more rounded with time. Concurrently, adjacent particles sinter bond which reduces the surface energy by removing free surfaces. In dry, natural snowpacks the rate of rounding and sintering depends mainly on microscale temperature gradients (Colbeck, 1980, 1983). The saturation vapor pressure over ice increases exponentially with temperature which means the sintering rate should increase rapidly with temperature. Experimental evidence (eg. Gubler, 1982; de Montmollin, 1982; Dash et al, 1996) confirms that sintering rates increase rapidly as temperatures increase above about -10 °C.

Introduction of liquid water not only enhances the rate of mass transport on the grain scale but also forms liquid bridges between grains resulting in a "capillary strain" (Hwang et al., 1987). Capillary forces in snow in excess of 4000 Pa have been measured in partially saturated snow (Colbeck, 1974; Wankiewicz, 1979). Such forces are much higher than gravitational forces expected in the near-surface snow and cause rapid densification independently of gravity. The high capillary pressures cause grains to cluster (Colbeck, 1982). We have conducted experiments to investigate rates of shrinkage when water is first introduced to snow. Figure 4a shows an experiment in which snow (ρ = 90 kg m^{-3}) was collected in a can 140 mm high and 100 mm diameter. The sample was kept at -4 °C for 20 hours during which time the snow settled 1 cm. Introduction of liquid water (300 ml at 0 °C) caused the sample to shrink both vertically and from the sides (figure 4a). We interpret the shrinkage, in particular the lateral shrinkage, to be evidence of capillary strain. The rate of densification increased by four orders of magnitude at the onset of warming and wetting (figure 4b). The rate decreased rapidly as the density approached the final (dry) density of about 360 kg m^{-3}; the rate of densification depends on density. Highest rates are expected during first wetting of very low density snow. The experiment clearly illustrates that snow can undergo substantial deformation without significant changes in the gravitational stress.

CREEP BEHAVIOR OF ALPINE SNOW

The usual approach to modelling creep behavior is to apply a constitutive law that relates the bulk response of the snow to an applied stress. Following the usual assumptions, for snow on a 36° slope (see Perla (1980) for a review of previous work) we expect the shearing component of motion would be about 50% greater than the compressive component (in the range of the shaded area of figure 3b). This contrasts with our measurements and some by Perla (1971) that indicate the shearing component is often less than the compressive motion (on a 36° slope) and the resultant direction of motion is close to vertical. The component of motion in the downslope direction depends on the slope angle. Most of the curvature in the creep profile is a result of changes in the rate of (vertical) settlement through the depth of the snowpack. The rate of motion decreases rapidly with density; large gradients in the velocity profile occur because of contrasts in layer properties.

The result suggests that the usual assumptions used to describe the mechanical behavior of snow are incomplete. Following on from our previous discussion, we believe that metamorphic and/or capillary strain contribute to the observed pattern of deformation and, as a first approximation, we rewrite the creep equation:

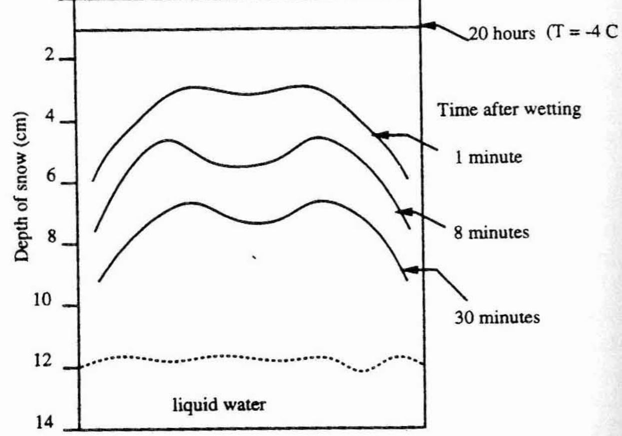

Fig. 4a Snow of relative density 0.09 was collected in a can and kept at -4° C for 20 hours. The snow settled 1cm. Cold water was then introduced and the evolution of the shape is sketched after 1, 8 and 30 minutes. We interpret the rapid shrinkage to be evidence of capillary strain acting independently of gravity.

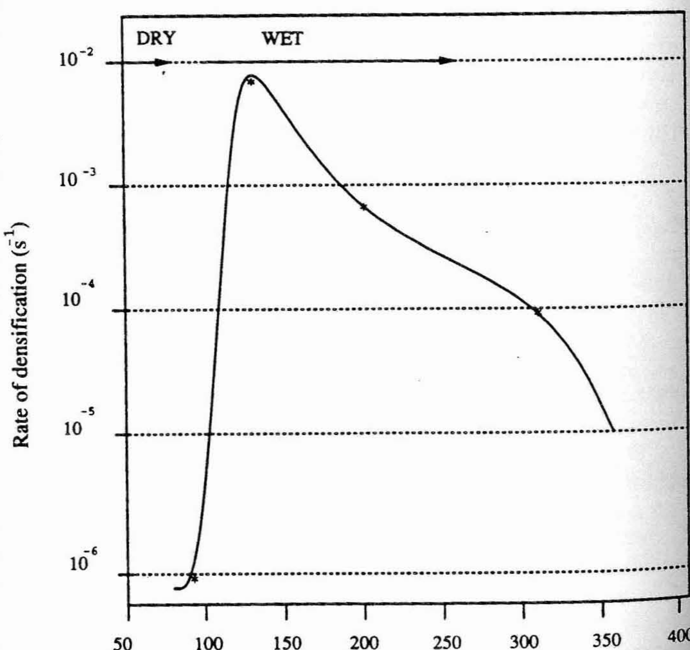

Fig. 4b The rate of densification increases almost 4 orders of magnitude with introduction of liquid water and then decreased rapidly as densification proceeded.

$$e_{ij} = A\,\delta_{ij} + B\,\sigma_{ij} + C\,\sigma_{ij}\sigma_{kj} + \delta_{ij}\,em_{ij} \quad (1)$$

where e_{ij} is the deformation rate tensor, em_{ij} is the metamorphic (or capillary) strain rate, σ_{ij} is the stress tensor, δ_{ij} is the Kronecker delta and A, B, and C are scalar functions of the stress invariants.

EVOLUTION OF MECHANICAL PROPERTIES

Mechanical properties such as strength and ductility undergo significant improvement as a result of densification and sintering. For example experimental data indicate the shear strength of snow increases rapidly with density (eg. Perla et al, 1982; Jamieson, 1995). Their results predict the resistance to shear (other things being equal) will increase by more than an order of magnitude as the snow density increases from 60 kg m^{-3} to 180 kg m^{-3} (typical for partly settled snow). It is also well known that the shear strength of snow depends on temperature, the rate of shear as well as the density. For example measurements of shear behavior in the laboratory (McClung, 1977; de Montmollin, 1982; Fukuzawa and Narita, 1992) indicate that snow does not show a distinct failure at low rates of shear; the strength increases with strain (strain-hardening). On the otherhand at high rates, the stress increases to a peak and then decreases to a residual after displacements of just a few millimeters (strain-softening). We think that sintering processes contribute to this behavior. At warm temperatures and low rates of deformation, grain bonds may form faster than they break, but as the rate increases (or the temperature decreases), bonds tend to break faster than they form. We expect the critical strain rate at which behavior changes from strain-hardening to strain-softening depends on both temperature and snow texture.

SLAB FAILURE AT THE ONSET OF RAIN

Recent models of snow slope stability calculate the energy needed to cause a zone of basal weakness (where the stresses from the overburden are not fully supported) to propagate. Figure 5a shows a physical picture of this condition and McClung (1979) showed that a basal weakness of length 2 L overlain by a slab of height H would propagate if:

$$\frac{H}{2E'}\left[\frac{(\tau_g - \tau_r)L}{H}\right]^2 > (\tau_p - \tau_r)\varpi \quad (2)$$

where E′ is the viscoelastic modulus of the slab, τ_g is the body weight shear stress, τ_p is the peak shear stress near the tip of the weak zone, τ_r is the residual shear stress (after softening) which is attained after a characteristic distance ϖ.

McClung considered the length L to be the only free parameter but in some cases the effective depth H may also be a free parameter. For example avalanche controllers often successfully "ski-cut" slopes by cutting across the bottom and then the top of a slab. If such cuts were say 25 cm deep and the basal weak zone was 50 cm below the surface, the energy available to drive propagation (the LHS of eqn. 2) would double (even without accounting for the additional stress imposed by the skier).

Capillary induced shrinkage at the surface at the onset of rain also reduces the effective depth of a slab. Measurements indicate shrinkage in the z direction (normal to the slope) typically reduces the slab thickness by 5 to 10 cm almost instantaneously (fig. 4). Longitudinal stresses can not be supported in the wetted zone which effectively reduces the depth of slab that can support longitudinal stress. Any pre-existing stresses must then be redistributed through the remaining slab and the weak layer. Following eqn. 2, the energy available to drive propagation increases as the effective depth of the slab decreases (other things being equal). Figure 5b shows the increase in available energy as a function of initial slab depth for cases where the effective slab depth decreases by 5, 10 and 20 cm. The analysis predicts that the increase in available energy is greatest in thin slabs; slope failure is more likely if the overlying slab is thin and the stability is already close to critical.

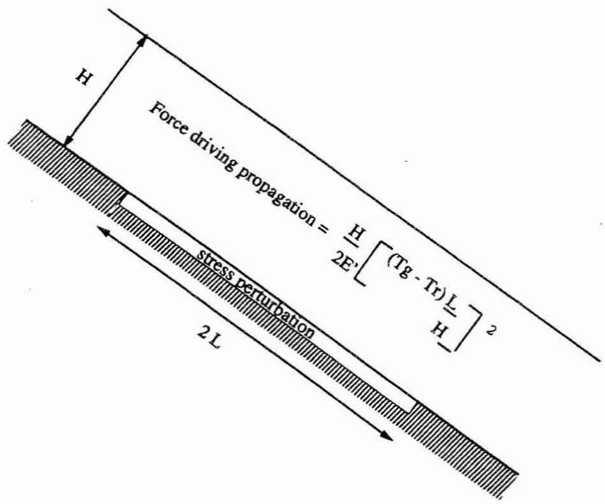

Fig. 5a Model of bed surface perturbation. The force driving the failure at the bed depends on the slab depth and the length of the perturbation. Wetting at the surface effectively reduces the depth of slab that supports longitudinal stresses. The energy driving the basal failure increases.

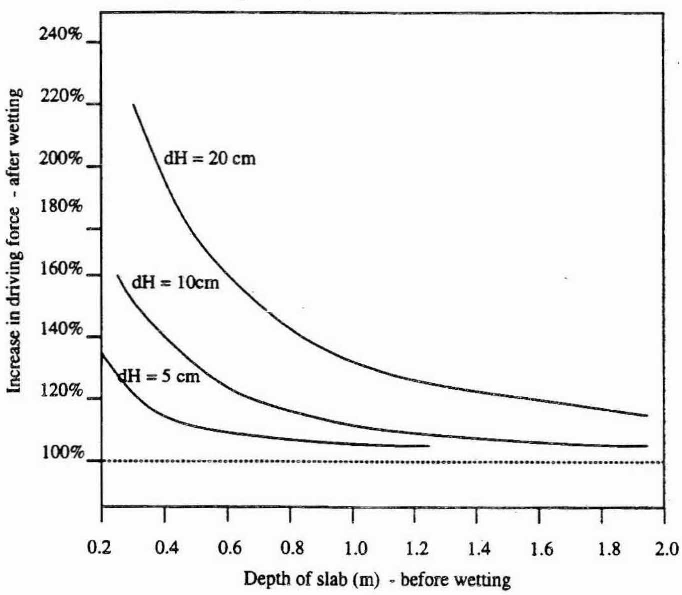

Fig. 5b Increase in available energy as a function of initial slab thickness for cases where the effective slab depth decreases by 5, 10, and 20 cm.

To summarize, we envisage the following sequence of events:

- before rain starts the downslope stresses from the weight of the slab are close to the average shear strength of a buried weak layer. At some locations, stresses from the overburden exceed the basal shear strength (a "deficit zone" or "super weak spot"); the slab is in tension at the top end of such a deficit zone.
- rain induced shrinkage occurring almost instantaneously at the surface alters the distribution of stresses through the snowpack. This effectively increases the energy available to extend the length of existing zones of deficit. In addition, wetting occurs over the entire slope and perturbs all existing zones of deficit simultaneously.
- slope failure at the onset of rain is more likely in cases where the slab is less than ~ 1 m thick, and when the transition from snow to rain is rapid. Rapid strain rates needed to cause strain-softening and propagation of a zone of deficit are more likely when the transition is rapid. When warm-ups are slow, grains are more likely to sinter bond and the associated densification causes the snow strength to increase rapidly.

CONCLUSIONS

Avalanche activity often increases at the onset of rain and decreases with continued rain. The avalanches that release immediately usually release as slabs well before liquid water has penetrated and lubricated the basal layer. To be effective with control work we have found it preferable to control slopes at the onset of rain when the slope stability is most sensitive. If we wait, many paths will have released naturally. Furthermore, control becomes less effective after rain has penetrated the snowpack, although we have had success by using 10 to 20 kg charges elevated a few meters above slopes.

Creep behavior of alpine snow is strongly influenced by metamorphic processes and capillary forces (when liquid water is present). Both processes cause the snow to shrink independently of gravity, and the rate of shrinkage increases with temperature and is particularly rapid in the presence of liquid water. The rate of deformation decreases rapidly as the snow densifies. It is likely that the rapid shrinkage of the surface snow during first wetting contributes to the instability. The capillary induced strain reduces the depth of slab that can support longitudinal stresses and effectively increases the energy available to drive an existing shear band to instability. A rain induced surface alteration occurs rapidly over a wide region and perturbs all existing zones of deficit simultaneously, thus increasing the possibility of slope failure.

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