# EFFECTS OF FORESTS NEAR TIMBERLINE ON AVALANCHE FORMATION

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## ABSTRACT

In high alpine forests, extreme dry slab avalanches with fracture heights between 0.8 and 1.5m, corresponding to slope angles from 35 to 45° in the starting zone, and mean return periods of 30 to 300y, may start in openings with downslope lengths of as little as 30m and widths of 15m. Larch stands near timberline have an open structure with tree distances often exceeding 15m. Larch stands only slightly affect the formation of the snow cover. Additional support of the slab is limited to distortions of the stress field by the not very numerous stems. Therefore formation of extreme avalanches is not significantly hindered. Dense spruce stands, particularly if multiply layered, effectively prevent formation of extreme slab avalanches. If the avalanche flow is not braked on a distance downslope from the fracture line of 30 to 60m, depending on slope angle and fracture height, standard trees with diameters $\leq$ 0.3m will be broken. The minimum extents of weak and super-weak layers for slab formation ( 25 and 5m, respectively), as well as the ranges of typical supports to the new snow slab scale with slab thickness. Dense spruce stands significantly affect snow distribution and radiation balance, and therefore the formation of weak layers.

## INTRODUCTION

Large catastrophic avalanches rarely form within forested areas. More often large avalanches start above timberline or in only sparsely forested bowls and flow along well known, often channelled tracks through the forested area. A potential decay of high alpine forests may cause new, not yet recognized potential release zones. This work should help to identify within potential avalanche areas forests that may gradually lose their protective effect. By catastrophic or extreme events, we denote avalanches with mean return periods of at least several decades. They may build during extreme weather conditions and penetrate into landscape that is normally thought to be safe from avalanches. A majority of these events occur during extreme storms with high precipitation rates at low temperatures and heavy winds. An important additional condition is a weak base or interface layer between old and new snow because large avalanches release as slabs. In this paper we will concentrate on these most frequent events although rare extreme events also occur under different conditions.

Measurements and observations for the present investigation have been made in spruce and larch stands near timberline on a north facing slope at 2000 m.a.s.l. These are typical stands for formation of extreme dry slab avalanches (Imbeck, 1984; Meyer, 1987; Frey, 1990). To investigate the formation of large slab avalanches within forested areas, the following questions have to be answered: how do slab avalanches form, what are the minimum size conditions required by snow mechanics, and what is the critical avalanche size for extreme events (part 1). Compared to the open-field situation there are additional conditions that limit formation of slabs in forests: (1) distribution, shape and species of trees within the potential release zone, (2) grass, shrubs, stumps and micro-topography, and (3) formation and metamorphism of the snow cover under the modified conditions for mass and energy exchange with the atmosphere (part 2).

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## Failure Properties of Snow

Measurements of failure properties of snow by Narita (1983, 1987) as well as different types of field measurements and observations support the idea that snow will initially fail ductile. According to Narita, with some extrapolation to lower snow densities and temperatures typical of snow layers involved in direct-action avalanches, the mechanism that determines tensile and approximately also shear strength of snow as a function of strain $\epsilon$ and strain rate $\dot{\epsilon}$ may be subdivided into four distinct ranges: viscous behaviour without failure for $\dot{\epsilon} < 2 \cdot 10^{-6}$ $s^{-1}$, viscoplastic deformation with incomplete macroscopic failure at large strains for $\dot{\epsilon} = 2 \cdot 10^{-6}$-$8 \cdot 10^{-5} s^{-1}$, ductile failure for $\dot{\epsilon} = 8 \cdot 10^{-5}$-$3 \cdot 10^{-4} s^{-1}$ and at least partly brittle failure for higher values of $\dot{\epsilon}$. Corresponding minimum strain for brittle failure is 0.1 - 0.3%, and increases with decreasing $\dot{\epsilon}$ for ductile failure from 1% to above 10%. The ratio of maximum ductile strength to minimum brittle strengths is about 2. Strengths are of the order of $5 \cdot 10^2$ to $5 \cdot 10^3$Pa. Strain rates in inclined natural snow covers caused by the snow load itself hardly exceed $10^{-4} s^{-1}$, even if concentrated in weak layers or by obstacles. This finding is in correspondence with shear deformation measurements and model calculations. Therefore initial failure in natural slab releases will be ductile.

## Sequence of Events in Natural Slab Releases

The following sequence of events proved to be a plausible scenario for releases of direct-action avalanches. A necessary condition for slab fracture is the existence of a thin weak layer or a highly deformable layer interface of low cohesion between old and new snow. Shear strength and viscosity of this weak layer depend on the local structure of the snow. This snow has normally built the snow surface for some immediate past time. The interface between cold new snow and an old hard surface, buried surface hoar or windcrusts but also the fragile top boundary of a depth hoar layer are typical weak layers. The distribution of the new snow overburden, the slope angle, the local geometry of the old snow surface, and the deformability of the interface determine the stress and strain distribution along the weak layer. If the load from the overburden snow increases, strain starts to be concentrated within this weak layer. If local stresses and strains are high enough, ductile fracturing starts (Gubler, 1989). By this mechanism small, super-weak zones, that fail to transfer stresses from the top to the basal slab, form within the weak layer. Shear deformation concentrates in these super-weak zones. The high relative motion of the grains within the super-weak zone avoids a fast renewal of stress-conducting bonds. We call this first step of failure initial fracturing. Several nearby initial fractures may coalesce to a larger super-weak zone. Shear and tensile stresses concentrate at the lateral border of these super-weak zones. If critical, ductile-fracture strain (Bader, 1989) is attained, slow growth of super-weak areas by ductile fracture propagation is possible. This growth may be stopped by less deformable areas of higher strength within the weak layer or by stress redistribution. Eventually the super-weak zone may reach the critical size at which brittle shear fracture propagation starts. At its border, critical brittle strain corresponding to the higher brittle strain rate has to be attained (Bader, 1989).

As long as a well defined weak layer exists to guide shear fracture, brittle fracture may propagate through areas of higher initial stability (primary fracture). This follows because strength and critical strain are significantly reduced at the high strain rates that occur at the tip of a fast propagating fracture. Shear fracture usually propagates faster upslope than downslope. If the upslope propagating crack has reached some critical length, dependent on slab thickness and slope angle, principal stresses at the fracture tip turn nearly parallel to the shear-fracture plane. This is a necessary condition for slope-perpendicular tensile fracturing as observed. Any obstacle locally concentrating tensile stresses (rocks, trees, or an interruption of the weak layer) may start a tensile fracture perpendicular to the slope (secondary fracture). Only at this instant fracturing can be clearly recognized by an observer. The assumption that fracturing starts where you see it first, is wrong. For slab releases, the start of initial fractures is critical. They always happen downward from the tensile fracture and often far away from the observed secondary fractures. The slope-perpendicular fractures limit the size of the slab avalanche but are not responsible for its occurrence. Initial fracture and shear fracture propagation are the necessary steps for slab avalanche formation.

The size of a slab is limited by the extent of the weak zone, by the slope angle responsible for asymptotic shear stresses, and also by obstacles or snow patches of lower deformability that locally increase slope-parallel stresses and therefore support the slab. Packed snow under tree crowns and windpacked snow are examples for interruptions of weak layers or stress concentrators. Sometimes shear fracture propagates through areas where the slab does not move and connects separated, but simultaneously starting slabs.

There is one other important mechanism responsible for non-continuous fracture areas. Slab releases and flowing avalanches produce significant seismic energy in the underlying ground. These seismic waves may cause immediate initial fractures on neighbouring slopes.

Minimum-size Conditions for Slab Avalanches

In forests, trees disturb the snow stratification. The snow accumulation rate during the storm and snow metamorphism at the old snow surface differ significantly from free-field conditions. The local distribution of trees and other types of vegetation limit the potential fracture area for slabs. The required minimum size of an area of undisturbed snow accumulation and continuous weak layer formation will be determined from measurements and theoretical investigations.

Because we want to consider extreme avalanches only, we have to determine the corresponding extreme fracture heights. Extreme fracture heights for direct-action avalanches often result from new snow accumulations within up to 3 days. Typical fracture heights are of the order of 1 m depending on slope angle and return period. The values shown in Table 2 are determined on the base of extreme-value statistics on new-snow depth within 3 days and snow-mechanical facts (Bader, 1989).

Initial Fracture

In different field studies we have measured infra-sonic acoustic emission from snow (Gubler, 1979). Infra-sonic emissions in the frequency range of 10 to 70Hz result from initial fracturing in the snow cover. Fracturing in the structure of low density snow is equivalent to the rupture of individual ice bonds between ice grains. During ductile fracture and deformation, bond fractures are not correlated. This is exactly what happens in a stressed, weak shear layer. Ruptures of the weakest bonds decrease the load capacity of the weak layer if only fewer and smaller new bonds build simultaneously by compaction and sintering. A locally increasing strain rate, caused by decreasing viscosity, hinders the formation of new bonds. This is a slow process that typically goes on for hours. It involves large local strains but fairly low stresses. If in some small local zone the remaining bonds fail to transfer the stresses from the slab to the old snow underneath, the structure will locally collapse. Because at this stage of low stress and strain rates, the ice frame stores very little elastic energy , the fracture will not propagate significantly outside this weakening zone. These localized structural collapses produce measurable infra-sonic emissions. The fundamental frequency of the signals is approximately equal to $\nu = c/(4 \times l)$, where c is the shear wave propagation speed in snow, and l is the characteristic diameter of the super-weak zone being created. The sizes of the initial fractures are therefore 0.1 to 1 m. This characteristic length, typically smaller or equal to the slab thickness, defines the smallest length scale involved in natural slab releases. This scale is typical of the variability of stresses, but also of viscosity and strength in the weak layer.

Densification of the top layer occurring without affecting the weak layer or inhibited deformation within the old snow layer by stumps or rocks significantly increase strain and strain rate in the weak layer and may cause initial fracture. A local compaction of the surface slab by only 20% within an area of a width of 1 to 2D (D: thickness of the new-snow slab) causes a local increase of stress and strain rate by 10 to 20%. Local compaction may result from snow falling from tree branches saturated by intercepted snow during the storm.

The slab thickness and its deformability determine the length scale for the variability of the stresses. For hard slabs initial fracture zones may be much larger, but wind slabs will not build in forests. It is important to realize that initial fracturing is an unstable, positive-feedback process. Ductile weakening starts at a local minimum of stability or maximum of strain rate and gradually decreases strength and viscosity further.

Fracture Propagation

We have already seen that the same snow type, under the same stress state may or may not rupture under a certain stress, depending on whether or not a critical strain rate and a critical fracture strain are attained. This restricts the probability of fracture propagation considerably, and leads to the necessity of stress and strain rate concentrations. Such concentrations occur at the border line of super-weak zones created by initial fracturing. Thin weak layers existing in a snow pack are only a necessary, but not a sufficient condition for shear fracture propagation. To be sufficient, weak layers must be interrupted by super-weak zones of some minimum size, created by initial fracturing (Bader, 1989). From calculations of stress- and strain-rate distributions in the weak layer in the vicinity of super-weak zones, Bader and Salm determined minimum size conditions for the start of ductile- and brittle-fracture propagation. To start a propagating fracture, critical fracture strain that depends on the actual strain rate has to be reached under the stresses existing at the border of a weak zone. Depending on whether the critical fracture strain is reached at low strain rates ($10^{-4}s^{-1}$, ductile regime) or at high strain rates ($10^{-3}s^{-1}$, brittle regime), the rupture starts to propagate at speeds of the order of $0.1ms^{-1}$ or at some $100ms^{-1}$. We think that ruptures originating from the border of smaller super-weak zones will often propagate by ductile fracture, possibly connecting several super-weak zones. Eventually they may start to propagate at high speed from the border of the resulting larger super-weak zone. For typical values of snow viscosity, slope angle and layer thicknesses, minimum diameters of super-weak zones to start slow fracture propagation are of the order of several meters. Critical diameters for brittle fracture propagation are about 5 to 10 times larger. The critical downslope length L of a super-weak zone to reach critical shear strain rate is approximately equal to ( approximation of Bader's result, 1989)

$$L=cD\frac{\dot{\epsilon}_{cr}}{\dot{\epsilon}_g} \qquad (1)$$

c: constant depending on the Poisson ratio for the snow of the slab. c=3 is a typical value.
D: slab thickness
$\dot{\epsilon}_{cr}$: critical strain rate for ductile and brittle fracture, respectively

$$\dot{\epsilon}_g=\frac{1}{D}\sqrt{(\frac{D}{\mu_s}\frac{d}{\mu_w})}\tau \qquad (2)$$

d: thickness of weak layer
$\tau$: corresponding asymptotic shear stress at depth D
$\mu_s$: viscosity of slab
$\mu_w$: viscosity of weak layer

The square root is equal to the geometrical mean of the reciprocal specific viscosities, and therefore proportional to the geometrical mean of the deformation velocities of slab and weak layer. L scales with D. To get small critical L's, deformation speeds for the slab and the weak layer should be large. A very stiff slab (hard slab), as well as a high friction shear layer increases L significantly. A stiff slab reduces the ratios of peak to mean strain rate in the super-weak zone and of peak to mean shear stress at the border of the super-weak zone. High friction in the weak layer decreases strain rate at the boundary of the super-weak layer. For both cases L is very large and therefore fracture propagation in a forest is unlikely for these types of slabs. Slab formation in small areas is most likely for soft slabs on highly deformable weak layers. Because L is much larger than typical sizes of initial fractures, an additional necessary condition for fracture propagation to start is the coalescence of initial fractures to a super-weak zone of minimum size.

To fulfil the second condition of "critical strain", the weak layer has to be thin. Deformation speed of slab and weak layer are of the same order of magnitude. The shear strain-rate within the slab is of the order of $10^{-6}s^{-1}$. Critical shear strains are about $10^{-2}$ for critical strain rates of $10^{-4}s^{-1}$. Therefore the ratio of the thickness of the weak layer to the slab has to be smaller than $10^{-2}$. Accordingly, the width of weak layers should not exceed a few millimeters.

This part of the slab mechanical model defines a minimum size condition for continuous weak- and super-weak areas where initial fracturing may start fracture propagation.

Tensile Fracture at the Fracture Line

Tensile fracture may occur if the ratio of the downward length of the super-weak layer to slab thickness equals the ratio of tensile fracture strength of the slab to the asymptotic shear stress in the weak layer far away from the super-weak inclusion. This ratio will always be larger than 10. The above relation follows from statics as well as the fact, that tensile strength of the slab is always much larger than the shear strength of the undisturbed weak layer, and gives only a minimum condition. Dynamically, the brittle shear fracture may propagate further upslope. The fact that tensile fracture planes are always perpendicular to the slope within a few degrees confirms the number given above. The existence of a super-weak layer downslope from the location of tensile fracture is a necessary condition for a slope-perpendicular fracture plane. Finite-element (FE) calculations for typical material parameters result in ratios between 5 and 10 for maximum deviations of 5° to 3° from the slope perpendicular direction.

Width of the Weak Zone

So far, little work has been done to solve the 3-dimensional problem for fracture initialization. From experience we believe that the widths of the weak and super-weak zones outside the range of any lateral support have to be equal to at least 10 times the slab thickness.

Support of Surface Slab

Tree stems, tree branches reaching into the snow cover, compacted snow below certain tree crowns, bushes, shrubs, rocks, and small terraces may considerably hinder deformation of snow. They act as natural supporting structures. To determine the critical arrangement of natural supports that effectively avoids avalanche formation, the effects of the different types of disturbances on the creep field in the snow cover have to be known. We use FE codes to investigate the deformation and stress field around typical obstacles. Creep speed and stress are determined along weak layers of varying viscosity and super weak inclusions. Bader (1988) gives rules-of-thumb for the extent of the back pressure zone behind a wall and the zone of reduced deformation speed around stems. Any obstacle or compaction that reduces creep speed to approximately zero on a distance across the slope larger than about 3 times the slab thickness has a similar effect as a wall. Rows of bushes and close standing trees are examples of natural walls. Bader finds from FE calculations:

$$\frac{x}{D} = 3 + 2.6\left(\frac{\mu_s}{\mu_w}\frac{d}{D}\right)^{0.75} \tag{3}$$

with x : distance of influence (95% of asymptotic value)
D : slab thickness
d : thickness of weak layer

For typical ratios of the $\mu$'s and thicknesses, the distance of influence is about 3 to 4. For very weak or super-weak layers the second term gets significant, but for soft slabs the range is limited to about 10D. For isolated stems with diameters << D the zone of reduced deformation speed amounts to about 1.5D. For blocking obstacles of diameters <3D the back pressure zone reaches about 2.5D, for larger diameters an upper limit is given by the corresponding back pressure zone behind a wall. The lateral supporting range for single stems with diameters <3D is limited to 1.5D.

Berms and small terraces locally support the slab if the weak layer follows the shape of the terrace. The width of the berm should be larger than the thickness of the snow cover below the weak layer to have a significant effect. A terrace of a width L of about twice the slab thickness reduces the creep speed and the shear stress to about 1/3 of the corresponding asymptotic values for the weak layer. The length of the back-pressure zone is approximately equal D. Compaction at the upper end of the terrace increases material strength and viscosity.

Below tree crowns an increased compaction during the snow fall may result from intercepted snow falling off the branches (spruce trees). For isolated spruce trees with branches reaching close to the snow surface, the snow cascading from the branches concentrates at the edge of the crown projection and causes significant densification, even during the storm. The effect of this type of distortion depends on whether the weak layer is interrupted or not. If there is no weak layer within the densified cross section, these portions of the snow cover effectively reduce local deformation speeds. The reduction of deformation speed, compared to the asymptotic weak layer value, depends on densification and downslope width of the supporting zone. The corresponding stress peaks are situated well within the densified zones. Distribution of stress and deformation around this type of support (Table 1) make it very unlikely that initial fracturing happens near the distortion.

| compaction | downslope width of the zone | reduction of deformation speed at depth D | backpressure zone x/D |
|---|---|---|---|
| 20% | 0.5 - 4D | 70 - 85% | 2 - 2.7 |
| 50 - 80% | 4 D | 93 - 97% | 2.9 - 3 |

Table 1 Reduction of creep speed within zones of increased viscosity (weak layer interrupted), and range of backpressure (95% limit).

## Critical Avalanche Size

The basic question to be answered is: What is the critical size of an avalanche with extreme fracture height to be destructive to the forest and to eventually reach the valley floor? For the following discussion we will limit ourself to dense flowing avalanches. For snow stability reasons the extreme fracture height and the corresponding mean return period depend on the slope angle of the fracture zone. The snow mechanical parameters cohesion and internal friction that determine the above relationship depend on snow type and therefore for extreme situations also on regional climate. A typical relationship for the Swiss Alps for extreme events is given in Table 2.

Critical parameters that determine the reach and dynamic forces are flow height and flow speed. These parameters depend on mean fracture height, slope angle of the release zone, snow type, roughness of sliding plane, track geometry (confined or unconfined), and slab area. The Voellmy-Salm model (Salm, 1990) will be used to estimate the critical parameters for extreme events. One of the friction parameters depends strongly on the density of the forest stand in terms of mean number of trees per area, the mean stem diameter, and flow height.

| slope | fracture height [m] | return period [y] | critical length 120kNm [m] | 10kPa [m] |
|---|---|---|---|---|
| 45° | 1.1 | 300 | 30 | |
| | 0.9 | 100 | 35 | ≤10 |
| | 0.8 | 30 | 40 | |
| 40° | 1.3 | 300 | 40 | |
| | 1.1 | 100 | 40 | ≤10 |
| | 0.9 | 30 | 45 | |
| 35° | 1.5 | 300 | 50 | |
| | 1.3 | 100 | 55 | 10-15 |
| | 1.1 | 30 | 60 | |

Table 2 Relationship between slope angle, extreme fracture height, critical length of slab to cause a torque ≥ 120kNm or a dynamic pressure ≥ 10kPa, and mean return period for typical climate in the Swiss Alps.

As we have already seen, slab avalanches may start only on steep terrain in low density larch stands or forest openings of some minimum size. The minimum size is given in terms of the release height. The steeper the slope, the smaller the minimum length and width of the starting zone and, theoretically, the smaller the mean fracture height for extreme events. Because of possible partial support of the slab in the starting zone, extreme fracture heights for steep slopes may be larger compared to the open field situation. Observations show that in forested areas extreme avalanches start on steeper slopes than in the open field. Therefore calculations are for slope angles of 35° to 45°. We assume two standard forests, both with mean stem diameters of 0.3m (measured 1.3m above ground), and a bending yield strength for the stems of 50MPa (Frey, 1987). The mean distance between the stems is 5.5m and 11m for the high and low density stands, respectively. De Quervain (1979) proposed that dynamic pressure should not surpass 10kPa at a flow height of less than 3m to avoid damages to forests. This value is significantly lower compared to the values that result from the above assumption of maximum bending stress. The assumption of maximum bending stress implies that trees with stem diameters less than the design diameter of 0.3m break. Flexible shrubs and young trees often survive dry, surface-layer avalanche flows. In Table 2 estimates for critical distances between fracture line and the lower end of the opening, often identical with the Stauchwall, are given for different slope angles and fracture heights. An additional important parameter is the amount of confinement in the track. Confinement increases flow height and flow speed. Examples are given in Table 3.

Besides the question of damage to the forest, it is most important to know whether the avalanche will reach the valley floor and endanger houses, streets etc. For the results presented in Table 4 we assumed a gradually decreasing slope angle to about 10° for the lowest part of the slope. Extreme runout distance is tabulated for the two forest densities. A 50m long slope at less than 10°, covered with a dense forest, will stop any avalanche that did not harm similar type of forest further upslope. If an avalanche destroys forest, logs entrained into the flow increase the destructive power of the avalanche significantly. If forest within a track is damaged by an avalanche, further extreme avalanches may continue to destroy forest and eventually endanger zones to be protected. If dry avalanches are allowed to accelerate to speeds higher than $20ms^{-1}$, they may develop into powder snow avalanches. The dynamic pressure head of powder snow avalanches hits the trees at their crowns. The resulting high torques will break the trees.

| fracture height [m] | slope angle [°] | flow width, relative | forest density | flow height [m] | torque [kNm] |
|---|---|---|---|---|---|
| 1.5 | 30 | 1 | l | 1.6 | s |
| | 30 | 0.7 | l | 2.2 | c |
| | 35-40 | 1 | l | 1.4-1.5 | c |
| | 30-40 | 1 | h | 2-2.2 | s |
| | 40 | 0.7 | h | 2.7 | c |
| | 30-40 | 03-0.5 | h | 3.6-6.8 | d |
| 1.3 | 30-35 | 1 | l | 1.4-1.5 | s |
| | 35 | 0.7-1 | l | 1.3-1.9 | c |
| | 30 | 0.5 | l | 1.7 | d |
| | 30-40 | 1 | h | 1.8-2 | s |
| | 35-40 | 0.5 | h | 3.2-3.4 | c |
| | 35 | 0.3 | h | 5.5 | d |
| 1.1 | 30-40 | 1 | l | 1.1-1.3 | s |
| | 40 | 0.7 | l | 1.6 | c |
| | 35 | 0.5 | l | 2.1 | d |
| | 40 | 0.5 | h | 2.9 | c |
| | 35-40 | 0.3 | h | 4.5-4.8 | d |
| 0.9 | 40 | 0.5 | l | 1.9 | d |
| | 40 | 0.3 | h | >3 | c |
| 0.8 | 40 | .5 | h | 1.6 | c |

Table 3 Potential for damage from avalanches entering a forest canopy at a critical speed of $18m^{-1}$. At this speed no damage will occur at the line of entrance. Depending on type of forest, slope angle and confinement, the avalanche may cause critical torques to the trees a few tens of meters downhill. Forest density: mean tree distance 11m (l), 5.5m (h). Torque calculated for tree diameters of 0.3m and a height above ground of the sliding layer of 1m: small <100 kNm, no damage (s); critical, 100-140 kNm (c); large >140 kNm (l), trees knocked down (d).

| fracture height [m] | forest density | limit 30kPa [m] | run out distance [m] |
|---|---|---|---|
| 1.5 (300y) | l | 80 | 200 |
| | h | 15 | 50 |
| 0.8 (30y) | l | 40 | 80 |
| | h | 0 | 33 |

Table 4 Run-out distance on forested 10° slope connected to a previous 30° slope.

The following conclusions can be drawn from the examples presented in Tables 3 and 4: The critical length of the starting zone increases with decreasing slope angle. Dynamic pressures of slab avalanches with fracture heights ≥1m are always larger than 10kPa. Avalanches with speeds larger than $18ms^{-1}$ at the Stauchwall normally harm forests independent of slope angle of the starting zone (35-45°) and fracture height (0.8 to 1.5m).

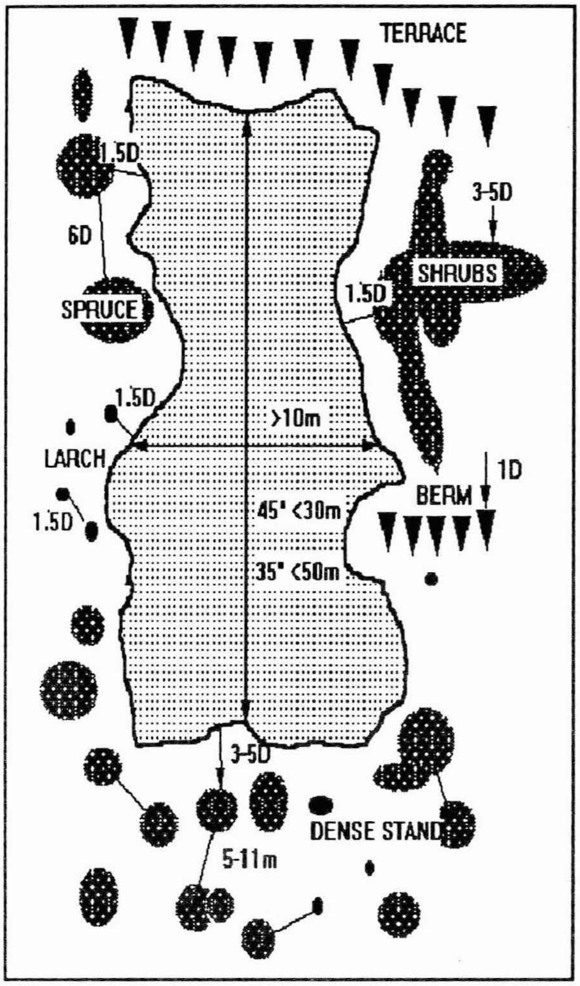



Fig. 1 Critical size of starting zone in an opening. Critical distances between trees and other types of distortions to the stress field. D: thickness of surface slab.

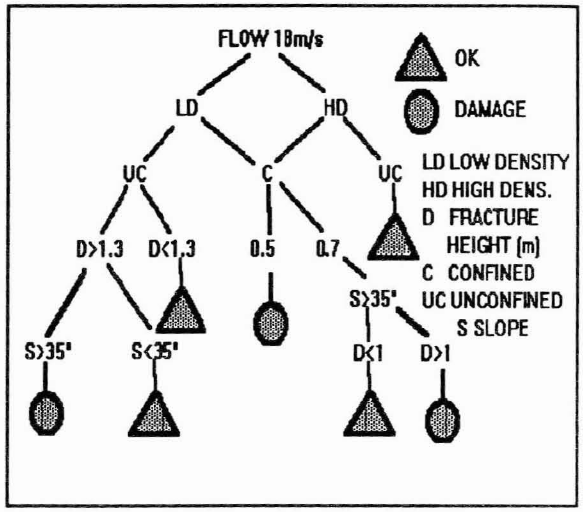



Fig. 2 Decision diagram for confined and unconfined flows starting from critical release zones for the assumptions described in the text.

If subcritical avalanches with flow heights <1.3m penetrate into a forest stand at slope angles ≤40° no damage will occur if the avalanche width is not reduced (unconfined avalanche). Avalanches with flow heights ≥1.3m flowing through low density forests on steep slopes (≥35°) may start to knock down trees a few tens of meters down from the beginning of the forest canopy. In high density forests, unharmful avalanches will not develop into harmful ones if they are not confined. Reduction of flow width in the track may increase speed and flow height to critical values. For flow heights at the Stauchwall of >1m, confinements to 70% of the original flow width are necessary to cause damage. If tree branches reach within the flow, the tree cross section enlarges significantly, making damage highly probable, even at lower speeds (≤10m/s). For flow heights >3m the combination of flow height and ascending height will reach the branches and often harm the tree.

In dense forests avalanches may loose moving material in forested tracks but normally dry avalanches with fracture heights ≥1m will flow through forests until they run out on slopes of lower inclination (≤10°).

Summary of Minimum Size Conditions for Harmful Dense Flow Avalanches in Forests

The main results are summarized in fig 1 and 2. Extreme slab avalanches with fracture heights >1m develop only if a continuous weak layer with no significant interruption exists within an area of 10 to 20 m in downward length, depending on slope angle, and at least 10m width without any significant support by vegetation or micro-topography.

Slabs starting in openings longer than 30 to 60 m, corresponding to slope angles of 45° to 35°, harm the forest immediately at the lower end of the opening. The fracture width of harmful confined avalanches is at least 20m.

## PART 2. MEASUREMENTS AND SIMULATION OF THE DEVELOPMENT OF DANGEROUS SNOW COVERS IN HIGH ALPINE FORESTS

Accumulation of snow into an undisturbed slab of some minimum size on a continuous, undisturbed weak layer during a single storm is a necessary condition for the formation of extreme direct-action slab avalanches. In forests the pattern of snow accumulation is modified by interception and dumping of intercepted snow by the tree branches. The development of the weak layer depends on the micro-climate altered by the forest canopy.

The effects of snow interception and snow fall from the trees (mechanical disturbance to the slab and the weak layer) and of the micro-climate (evolution of the weak layer) will be handled in separate chapters.

### Specifications of the Forest Stands under Investigation

Three typical forest stands and one open field plot on the same slope near timberline were chosen to investigate snow distribution and climatic conditions during the four winters 1986 to 1990. Tables 5 and 6 specify the four study plots.

| study plot | area # of courses | # of depth probings | elevat. m.a.s.l. range | aspect | slope angle [°] |
|---|---|---|---|---|---|
| open field (A) | $6 \cdot 26m^2$ 3 | 42 | 2064 2062-66 | NNW | 29 |
| larch disint(B) | $55 \cdot 58m^2$ 7 | 217 | 2003 1992-15 | NW-WNW | 23 |
| spruce disint(C) | $55 \cdot 58m^2$ 7 | 217 | 1994 1932-55 | NNW-NW | 22 |
| spruce dense (D) | 1 course | 21 | 1905 | WN | 24 |

Table 5 Stand types: (B) larch forest (Larix decidua MILL.), old, over-maturated, disintegrated, with some single spruce individuum, lack of regeneration because of grazing. (C) spruce stand (Picea abies Karst.), disintegrated, with many gaps, solitary spruce and spruce groups keeping their lower branches near the ground. Single larch trees are admixed.(D) spruce stand at its optimal-phase, dense canopy, the stems are free of branches (self-pruning).

### Effects and Mechanism of Snow Interception by Trees

Snow interception has two main effects on the formation of extreme avalanches: (1) Alteration of the old snow layers including the snow surface before the storm, and (2), modification of the distribution and rate of accumulation of new snow during the storm.

| species | fraction [%] of species in plots | | |
|---|---|---|---|
| | B | C | D |
| blanks | 65 | 62 | 5 |
| larch | 33 | 6 | 0 |
| spruce | 2 | 32 | 95 |

Table 6 Density and mixture of canopies.

Effects on Basal Layer

The mean water equivalent of the snow cover below trees is reduced by the amount of intercepted snow that sublimates. Intercepted snow sublimates faster than snow on the ground because of its higher surface-to-volume ratio, increased snow temperature caused by heat flow from snow free, short wave absorbing parts of the tree to the intercepted snow, and by increased ventilation. This leads to an uneven snow distribution below a forest canopy. In spruce stands, precipitation particles deflected by the tree branches, and intercepted snow cascading from the tree, increase the snow depth around the trees. This leads to a very bumpy snow distribution, especially in a typical high alpine spruce stand with openings between the tree crowns. The resulting depressions, terraces and berms support the new snow slab. Isolated groups of spruce have often branches reaching into the snow cover hindering downslope deformation of the snow layers completely. In deciduous larch stands, the effect of interception on the spatial distribution of snow depth is less significant.

The mean relative snow depths, determined from the snow depth courses (Table 5) that have been recorded in the larch and spruce stands during 4 winters, clearly show the effect of the different canopies on snow accumulation. The snow courses were performed after significant precipitation events at very different mean snow heights. The spacing between individual probings is 2m within the probe line. The results of the 13 snow courses performed are: plot A (opening) defined as 100%, B (larch stand) 78±13%, C (spruce, disintegrated) 69±7%, D (spruce, dense) 40±12%. Typical sections of spatial snow depth distributions in larch and spruce stands taken at maximum seasonal depth are shown in Fig. 3. The snow distributions are remarkably similar for the four winters independent of mean snow height. The significant difference of the snow distribution in the vicinity of isolated larch and spruce trees can easily be recognized. The depression below the larch trees is less pronounced. There is no rim around the tree because larch to not dump intercepted snow concentrated along the periphery of the crown. The large diameter of the area below a spruce tree indicated by significant variation of the surface slope angle, causes backpressure zones in the slab comparable to those of rigid walls (3 to 10 times the slab thickness D). The small slope angle deviation below larch trees limits the supporting zone to about 1.5D caused by the stem. Course D (Fig. 3) was measured in a dense, evenly aged spruce stand with a closed canopy consisting of crowns ending several meters above ground. The effect of single crowns is reduced in this case, but there remains the waviness of the ground, not being smoothed by the shallow snow cover (only 40% of the open field value!). The high number of stems per unit area causes significant support to the slab.

Not only the snow distribution, but also the layering, especially the surface of the snow cover are altered by interception and snow fall from the trees before the extreme storm. Snow falling from the trees locally densifies the layers near to the surface. Fragile surface layers are destroyed and interrupted. To investigate the areal influence of forest stands on the snow layering a sledge-mounted radar was pulled through the different types of canopies. The radar measures the electromagnetic distance between the snow surface and specular reflecting layer interfaces within the snow cover including the ground surface. The slope parallel and slope perpendicular spatial resolutions are 2 to 5cm. Profiles have been measured during 4 winters (Fig. 4). The results can be summarized as follows: Outside of forests, in the open field, blowing snow and snow drift cause a very fine layering. The waviness of the bare ground is smoothed already during the first snow accumulations in early winter. Snow is redistributed on a large scale, resulting in deep snow in gullies and on leeward slopes, and little snow on crests and windward slopes. The layering is uniform within large areas. In forest stands, including small clearings, snow redistribution by wind is not important with one exception on a larger scale: the mean water equivalent in clearings is increased compared to the free field by snow trapping (Schmidt, 1989; Imbeck, 1985). There is significantly less fine layering. The small-scale ground roughness is discernible in layers up to about 1m above ground. In dense spruce stands the characteristic layering disappears almost completely. In small openings between spruce, the layering is less pronounced compared to similar openings between larch trees. This results from further reduced wind and radiation influence in spruce stands compared to the deciduous larch stands. Below and at the border of isolated groups of spruces, snow depth varies significantly and the layering disappears completely. In larch stands continuous layering with some distortions below the crowns is still recognizable. The layering is similar to that in clearings. Only below very dense crowns and bushes fine layering disappears too. It is interesting to note that for small to medium new snow falls, lasting for less than one day, the resulting layers often remained undisturbed, even below the crowns, because in many cases the branches did not dump intercepted snow.

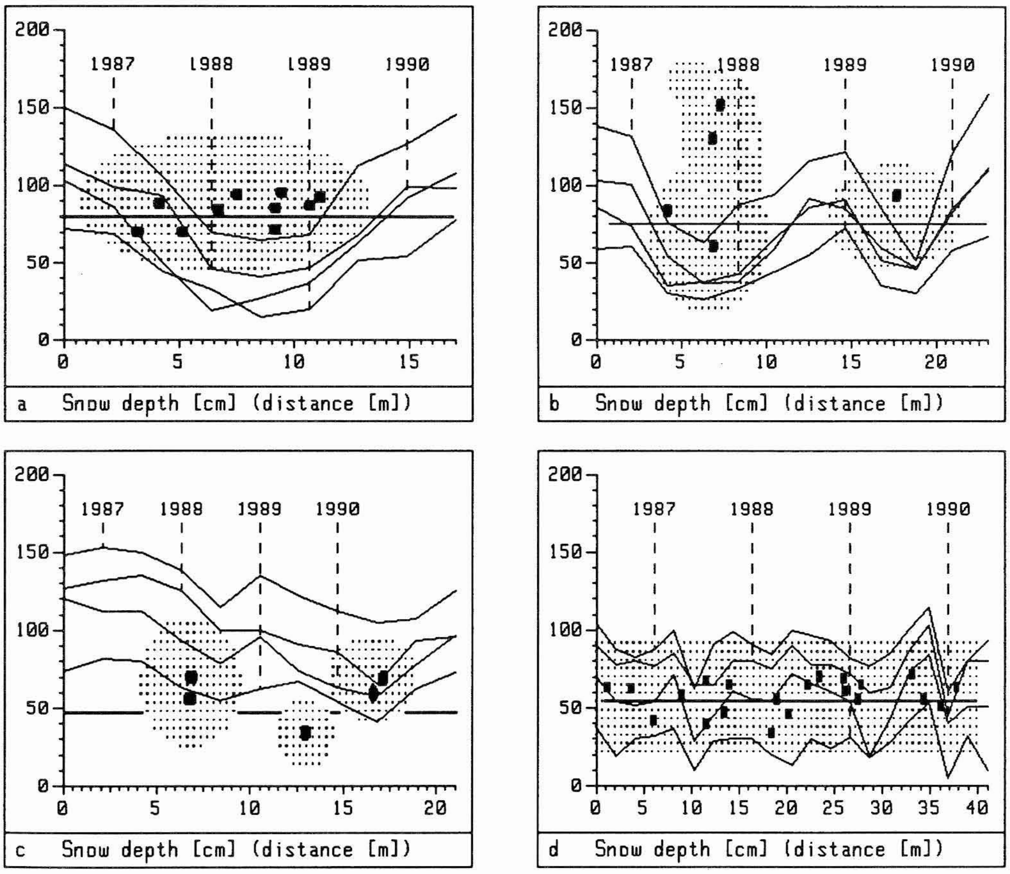



Fig. 3 Snow depth courses taken at maximum snow depth during the four winters 1987 to 1990 for typical snow distributions: (a) dense group of spruce, (b) single spruce, (c) dense group of larch, (d) optimal-phase single storied dense spruce stand. Crown projections are shaded, stems marked as dots.

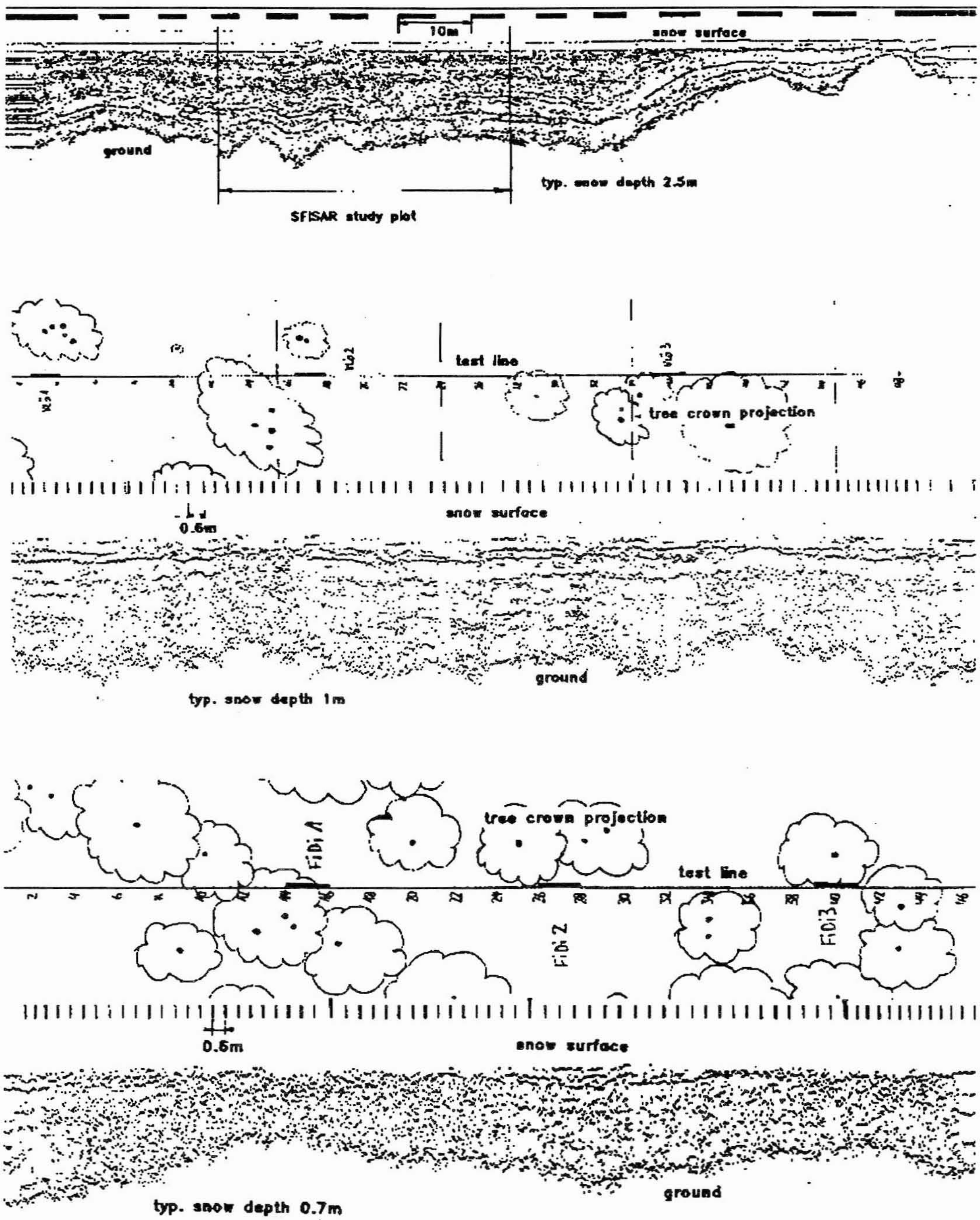



Fig. 4 Examples of microwave-radar profiles. Top to bottom: open field, larch stand, spru ce stand.

To summarize with respect to slab avalanche formation, it is important to state that continuous layering interrupted only by the tree stems is often found in larch stands, but never in spruce stands. In clearings and in larch stands, a larger snow-waterequivalent on the ground is necessary, compared to the free-field, to smooth the snow surface. Smoothing is a necessary condition to avoid local support of the new snow slab.

### Effects on New Snow Slab

Snow interception during the storm and its effect on avalanche formation depend on the amount and quality of the snow already accumulated on the branches at the beginning of the storm, tree species, air temperature and radiation.

Schmidt (1990) describes a sigmoidal growth curve for interception during snow falls at subfreezing temperatures. The main mechanism describing the growth curve are: In the beginning phase snow crystals rebound elastically from snow free parts of the branches. As crystals become lodged on the branches, the percentage of crystals falling on snow already held by the branches increases until the branch is completely covered. With growing interception, snow bridges neighbouring branches, thereby further increasing the intercepting surface. With increasing slope angles of the surfaces of the accumulated snow, more and more of the falling crystals that strike the intercepted snow bounce, often dragging off clusters of snow particles.

To further investigate the efficiency of interception and snow dumping of branches, we performed two experiments: Recording of sudden shaking of tree branches by snow falling off the tree branches, and measurement of the deflection of a tree branch caused by snow accumulation on the branch. Accumulation on the branches was compared to accumulation on the ground. The results can be summarized as follows: Efficiency of interception is largest at decreasing subfreezing temperatures. Interception during the storm is largest if the branches are initially covered by a few centimetres of well bonded old snow. We found that a maximum of 40 to 50cm of snow can be intercepted within 24 hours before significant snow fall from the branches starts. Snow dumping was always correlated with either wind gusts, increased short wave radiation (even during continuing precipitation), or sharp increases of air temperatures. Larch trees reach saturation earlier and tend to lose snow earlier than spruce.

The branch bending measurements did not show any dependence of the deflection at constant load on the branch temperature. This experiment is different from Schmidt's (1990), who measured bending resistance of conifer branches in function of the branch temperature. He applied bending forces for short periods of time, whereas we applied a constant load (snow and artificial weight). Schmidt measured the elastic response, we recorded long term creep deformation. Schmidt found a linear decrease of bending resistance with increasing temperature over a temperature range of -12 to 0°C. Schmidt proposes that unloading results from increasing branch bending at constant load with increasing temperature. We anticipate that snow strength decreases as snow temperature increases with increasing air temperature and solar radiation. It seems to us that the effect of energy input to the snow is more drastic than to the branch wood. Melting of ice crystals in the wood is proposed to be responsible for decreasing stiffness, consequently the necessary heat flow to the wood scales with the latent heat of fusion. To decrease strength of the intercepted snow, a small increase of the temperature of the well ventilated snow significantly increases the rate of "radius-of-curvature metamorphism" to decrease specific surface of the snow. This process involves only the heat capacity of snow that is about two orders of magnitude smaller than the latent heat of fusion. The metamorphism of the structure, that is well known to initially decrease snow strength, is driven by surface energy stored in the structure of new snow.

During extreme precipitation events, saturation will always be reached. The variation of air temperature, radiation, wind and the tree species determine if snow precipitation crystals just rebounce from the equilibrium shape of the snow or drag off small grain clusters, or if small lumps of snow fall down irregularly (larch) or snow cascades from the tree (spruce). Extreme direct-action avalanches occur during storms lasting for more than one day, therefore radiation or temperature induced unloading is likely. Spruce trees with sloping branches often cascade. The impact of the avalanching snow compacts and strengthens the new snow layer on the ground. This results in a considerable increase of the supporting area caused by a spruce tree (diameter of supporting zone equals at least the crown diameter). In the case of larch trees, cascading is unlikely, snow trickles down having a much smaller effect on the new snow slab ( support of the slab reduced to the stem area).

## Formation of Weak Layer and New snow Slab

A continuous weak layer is a necessary condition for initial fracture and fracture propagation. Local shear stress enhancement at the border line between weak and super-weak shear layer is proportional to the square root of the ratio of slab to weak layer deformation speed. The deformation speed is proportional to the ratio of layer thickness to corresponding layer viscosity. Stress enhancement is therefore smaller for hard slabs than for soft slabs. In forests formation of hard- or wind-compacted slabs is unlikely, because of generally reduced wind speeds. This fact increases the probability of slab releases with extreme fracture heights in small undisturbed forest openings.

Weak layer formation depends on the micro-climate at the snow surface. A weak layer is a low-friction interface between snow-layers. The thickness of the interface depends on its structure. To start a fracture, deformation speed within the interface has to reach at least the same order of magnitude as deformation speed of the slab. To reach critical shear strain locally, shear deformation has to be concentrated within the thin shear layer. Two types of interface layers are adequate: (a) thin layers of structures with very low viscosities, and, (b) interfaces with very weak bonding across. We limit the discussion to three classes of weak layers: surface hoar (type a), depth hoar and crusts (type b).

Surface hoar often consists of a monolayer of large featherlike crystals. It is very fragile. Typical layer thickness is a few millimeters. It grows on snow surfaces cooled below air temperature by long-wave cooling. The water vapour source is air humidity that reaches saturation at the supercooled crystal surface. Growth decreases or is stopped by surface heating from solar radiation and advection increased by wind.

Depth hoar grows at high temperature gradients in the snow cover. It has its weakest zone at the top of the layer where the local crystal growth rate is largest. High temperature gradients are maintained by radiation cooling, cold air temperatures in the absence of solar radiation and small snow depth. Depth hoar consists of a very fragile structure of large grains.

Crusts, in forests usually well settled surface layers or melt crusts, build during periods of high air temperature and short-wave irradiation. Cold, fluffy new snow develops very little bonding to a smooth, cold crust, as long as the snow temperature stays low.

## Micro-Climate in Forests

Short- and long-wave radiation, wind, air temperature, humidity and precipitation reaching the surface of the snow cover on the ground are modified by the canopy. Air temperature and air humidity are only slightly altered by a well ventilated forest stand near timberline. Its effect on incoming short-wave radiation depends on the aspect of the slope.

Measurements of wind, air temperature, humidity, global short-wave radiation and snow temperature (10cm resolution) have been performed in the open field, openings and dense stands during 4 years.

The results can be summarised as follows: In the forest the wind was mainly convective, downslope during the night and, depending on aspect, upslope during the day. Mean wind speeds, measured 2 m above ground, were very low, 0.5 to 0.9$ms^{-1}$ in the larch stands including the openings, 0.5$ms^{-1}$ in spruce openings and 0.1$ms^{-1}$ in dense spruce stands. Peaks above 2$ms^{-1}$ during heavy storms with open field wind speeds up to several 10$ms^{-1}$ were very rare. The mean short-wave radiation in dense larch stands was reduced by 20 to 30% compared to openings and open field. In dense spruce stands, solar radiation was reduced to about 10% of the open field values (diffuse radiation only).

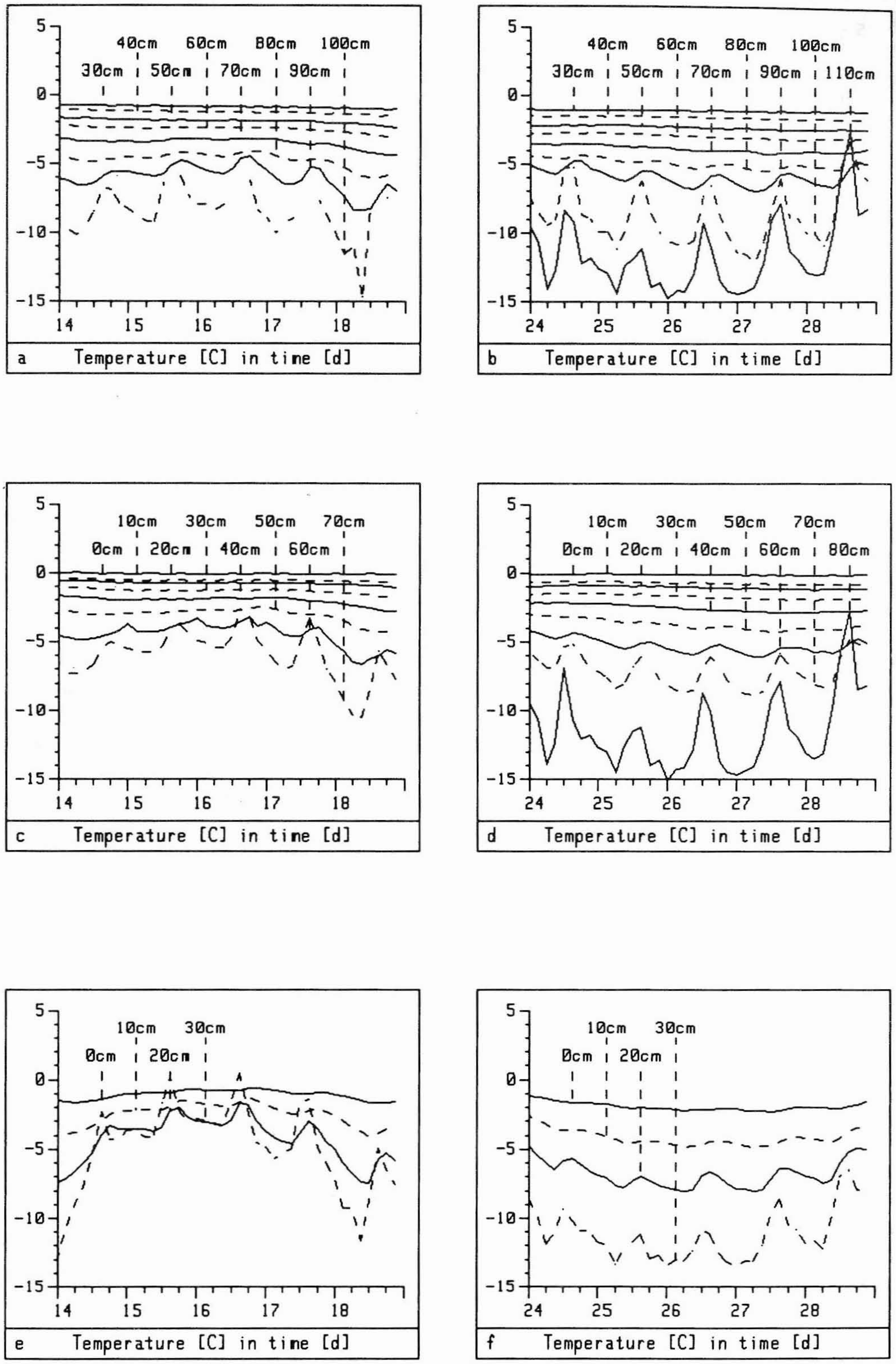
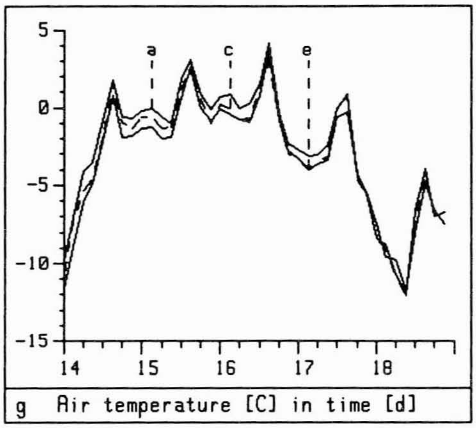



Fig. 5 continues next page



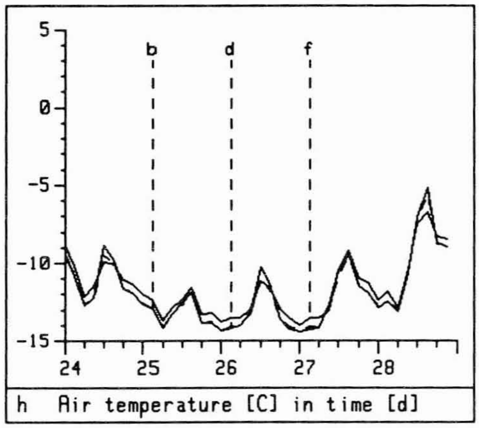




Fig. 5 Measured snow temperature profiles. 1st period (a,c,e): Feb. 14 to Feb. 18, 1988, clear sky, air humidity below 30%, air temperature increasing to the melting point; 2nd period (b,d,f): Feb. 24 to 28, 1988, overcast, air humidity 97%, air-temperature dropped to -18°C. Snow depth increased by 10 to 15cm between the two periods. (a,b) opening, (c,d) dense larch stand, (e,f) dense spruce stand. The numbers indicate the height of the sensor above ground. (g) air temperatures during the 1st period, and (h) for the 2nd period.

Mean air temperature at 2m was found to be the same for openings and stands. Under clear sky conditions, a significant temperature gradient was measured within the first meter above the snow surface at all test sites. This gradient disappears within about an hour if the sky gets 100% cloud-covered. The temperature gradient is caused by a stable stratification of the air above the radiation cooled snow surface.

Infrared measurements: Snow has a long-wave emissivity very close to 1. Clear sky has an effective emissivity with respect to air temperature (Gray, 1981) of 0.7, a completely overcast sky of 0.875. Larch stands have only a minor effect on the long wave balance. Larch stands are typically low-density stands, with only a small part of the cold sky covered up by the tree crowns. A 50% cover of the sky by trees reduces the negative long-wave radiation balance of the snow surface by only 30% (typically 20 $Wm^{-2}$). The modified long-wave balance ( -50$Wm^{-2}$) has to be compared to the shortwave balance and advection at the surface (+10$Wm^{-2}$, typically depending on aspect and date), and the energy flow balance at the snow-ground interface (+3$Wm^{-1}$). During the night an unshielded or only slightly shielded (larch stand) snow surface typically cools 10 to 20°C below air temperature under clear-sky conditions and absent direct short-wave radiation. If the surface is not screened from direct short-wave radiation during the day, the temperature difference reduces to a few degrees at noon. The top surfaces of snow covered spruces cool down very similarly to exposed snow surfaces. Bare spruce needles have also a very high long-wave emissivity (0.97), and therefore cool down very similarly to snow if not exposed to short-wave radiation (low short-wave albedo of spruce). In a fairly dense spruce stand snow surface temperature and spruce temperature are often found to be very similar. This temperature is significantly higher than in the open field, but the day-to-night variations are smaller. Temperature gradients in air measured underneath spruce for clear-sky conditions seem to result from stratified air masses slowly flowing from clearings and open field into the stand.

Typical measurements for clear and overcast conditions in openings, larch and spruce stands are shown in Fig. 5. Snow surface temperature has not been measured. The closest temperature sensor was a few centimetres below the surface. If the air temperature gradient disappears, snow surface temperature generally equals air temperature. This is normally the case under overcast conditions. A non-zero gradient is a good indicator of radiation cooling. Under clear-sky conditions temperature gradients of -1°$Ccm^{-1}$ are typical for the first few centimetres below the snow surface. Therefore the effective surface temperatures for the opening and larch stand were lower compared to the ones indicated in Fig. 5a,c,e. For clear-sky conditions all surfaces cooled below air temperature. The temperature differences were 7°C (10°C, estimate based on extrapolated surface temperature) for the opening, 5°C (7°C) for the larch stand and 1.5°C for the dense spruce stand. In dense stands the snow-to-air temperature difference is often small or insignificant depending on the air- flow pattern in the stand. During the overcast period (Fig. 5b,d,f) no radiation cooling of the surfaces could be observed, the surface temperature differed less than 0.5°C from the air temperature. The surface-to-ground temperature difference below the dense spruce canopy is roughly half the corresponding open-field value. At the end of the first period, cloudiness increased, and at the same time air temperature dropped. During this phase, the air-snow temperature difference disappeared quickly and the surface temperature was lowered by advection. During the second, overcast period, cold air lowered surface temperature by advection.

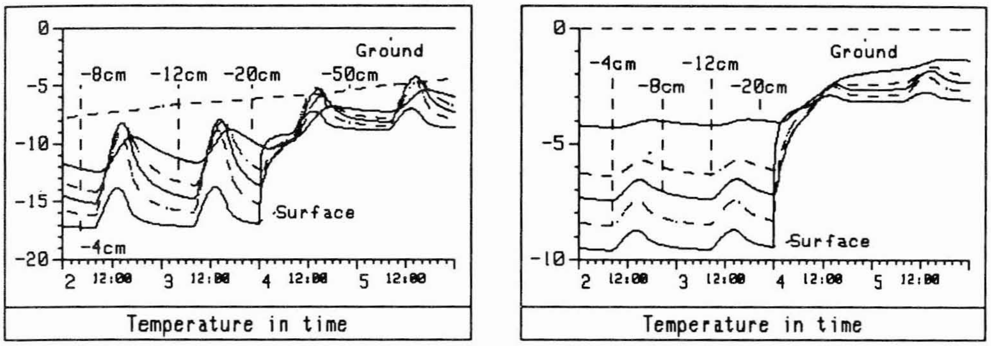



Fig. 6 Snow temperatures calculated from model at different depths measured from the snow surface. Left: open field, day 2,3 for clear sky, day 4,5 for overcast sky. Right: dense spruce stand, clear sky, day 2,3 spruce snow covered, day 4,5 bare spruce. Additional details in the text.

For the modelling calculations shown in Fig. 6 (Bader, 1990), the following assumptions were made: Sky emissivities with respect to air temperature for clear sky $\epsilon_L$=0.7, for overcast sky $\epsilon_L$=0.875; air temperature -3°C. For the case of dense forest the effective emissivity of trees and sky is mainly determined by the emissivity of the spruce (0.97) lowered to $\epsilon_{eff}$=0.943 by partial view to the clear sky. Four situations are compared to each other: open field and clear sky, open field and overcast sky, clear sky and snow covered spruce, clear sky and bare spruce. Tree temperature for the bare (snow-covered) case is assumed to equal air (snow-surface) temperature.

The calculations roughly confirm the measurements. Supercooling of the snow cover with respect to air temperature is smaller in a dense spruce stand compared to the open field. Diurnal variations of the surface temperature are reduced in the stand. Air-snow temperature differences are much smaller for overcast sky. The surface temperature modeled is too low compared to experimental results and therefore we may conclude that the estimate for $\epsilon_L$ for cloudy sky is too low.

## Effect of Micro-Climate on Weak Layer Formation

Surface temperature, temperature gradients in the snow and snow temperature are the important parameters for depth and surface hoar development and snow settling. Formation of depth hoar depends on the temperature gradient. The surface-to-ground temperature difference and the snow height are reduced by similar factors in a dense spruce stand compared to the open field. Therefore formation of depth hoar can be observed in forest stands too, but with large local variations. The very high near-surface gradients during periods of radiation cooling cause very fast crystal growth close to the surface (near-surface depth hoar). The formation of surface hoar depends on the snow-air temperature difference and air humidity. In dense spruce stands, tree, surface and air temperature are often close to equilibrium. Therefore development of surface hoar under spruce trees is very unlikely. In open forests, openings and along forest boundaries formation of surface hoar is often enhanced compared to the open field. The main reason is the partial shadowing from direct solar radiation, decreasing the warming of the surface during the day. The settling of surface layers and the formation of thermally induced surface crusts is reduced in dense, and to a lower degree also in open stands compared to the open field because of reduced direct short-wave irradiation.

The above discussion can be summarized as follows: Formation of depth hoar is not basically affected by forest stands but its spatial distribution varies strongly in spruce stands. Fragile near-surface layers of depth hoar develop only in open forests and openings. Surface hoar forms most effectively in places having a large solid angle of uncovered view to the sky, but shadowed from direct solar radiation. Surface crusts are more likely to form in open forests than in dense stands, at least as long as spruces stay snow covered and no melting occurs.

We have already seen that release of extreme dry slab avalanches from within dense spruce stands is not possible because of the effective support of the surface slab by tree stems and the bumpy snow surface. Continuous weak layer formation in spruce stands is unlikely. All three types of weak layers are likely to form in larch stands and openings. Therefore fracture propagation is very limited in spruce stands, but likely in larch stands.

## CONCLUSIONS

The existence of a continuous weak layer between basal and new snow slab is a necessary condition for dry slab avalanche formation and fracture propagation. The ranges of supporting effects of trees, berms, compacted parts of the snow cover scale with the surface slab thickness D and are of the order of 1 to 10D. The minimum extent of a weak layer outside any support of the surface slab for formation of surface slab avalanches is 5 to 10D. The maximum downslope distance between crown fracture and stand boundary should not exceed 30 to 60m depending on slope angle to hinder formation of destructive slabs. In spruce stands slabs form only in openings, fracture propagation is effectively hindered in dense parts of the stand. In open larch stands fractures are likely to propagate over areas only limited by the topography. Confinement of avalanche flow causes destructive avalanches if the width of the release zone is larger than about 10m. These facts show that the protective power of high alpine forests near timberline depends sensitively on the structure of the stands. Slab releases start in small openings. Multilayered and clustered stands normally stop fracture propagation contrary to single storied deciduous stands.

To preserve the optimum structure of a protective forest for all times, it is necessary to tend the stand. Human activities as grassing, commercial logging as well as diseases, avalanche formation in or above the stand may decrease its protective power. Natural mountain forests develop toward a uniform single storied structure and eventually disintegrate. Uniform or open stands with no regeneration are dangerous and should be avoided. The stand has to remain stable in all its components. The silvicultural goal is a clustered forest, sufficiently dense, stratified, mixed and unevenly aged with a vertically closed canopy. Clusters of different age and therefore different protective power should alternate in space. Other uses of the forest have to be subordinated to its main determination as a protective forest.

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