Some Mechanical Aspects on the Formation of Avalanches

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Abstract

The mechanism of loose snow and snow slab avalanches is studied with regard to the metamorphosis of stresses. The latter is steered by a temporal change of the physical characteristics of the snow. For the loose-snow avalanches, the so-called critical inclination of the snow cover (critical slope) seems to be of great importance. For the formation of snow slab avalanches the fact is to be considered that the tensil strength of the snow cover as a whole is always smaller than the sum of the tensil strength of the single layers.

An enormous field of variations of the physical qualities of different forms of aggregate beginning with loose snow up to poreless ice is reflected in their plastic behaviour. This is demonstrated by the state of viscosity of the polar névé and of the alpine snow cover.

I. Introduction

Our paper deals mainly with the change of stress and strain in snow and firn layers in view to the formation of avalanches. May I, as an introduction, briefly mention some stages of the Swiss research on snow and avalanches under the auspices of which the present study has been carried out.

From 1935 to 1955 the activity of snow research in Switzerland was concentrated on the investigation of the alpine snow cover. Its chief purpose was to procure a scientific basis for the fight against avalanches. One of the most important problems was the investigation of the state of stress and the equilibrium of the inclined snow cover. Then the question about magnitude and characteristics of the creep pressure which the creeping and sliding snow is exerting on firm obstacles, as for example on avalanche construction work. This problem was solved by the development of a practicable approximate method (1937–1942). Therefore, a first manual for the design and calculation of avalanche construction work could be put at the disposal of the engineers (1961).

In 1956, the participation of Swiss snow scientists and glaciologists in the International Glaciological Expedition to Greenland, 1957–1960, and 1967–1968, made them to face with a new and common task (Haefeli, 1959).

Instead of the inclined snow cover and the steep alpine glacier, the flat ice sheet with its nearly horizontal firn and ice layers moved into the field of our vision. It offered an opportunity to study *in situ* the continual transition from snow over polar firn to the subjacent cold ice without contribution of the liquid phase. The state of stress and strain of the horizontal snow and firn layers was considered as a special case of the inclined snow and firn cover. Research on the viscosity conditions which is of great importance with regard to the formation of avalanches could be completed and rounded off by including the studies on polar firn.

II. On the Stress Metamorphosis

The stress metamorphosis we understand to be the continual change of the state of stress governed by the metamorphosis of the snow. Its simplest form occurs when a horizontal snow cover settles under its own weight.

1) *The horizontal snow cover*

The settling process of a homogeneous horizontal snow cover on a horizontal underlayer is, as it is known, connected with the following conditions:

a) Only perpendicular displacements occur, any transversal deformation being confined.
b) The direction of the principal stresses (perpendicular, or horizontal) remains unchanged during the settling process.
c) The magnitude of the main principal stress is constant and equal to the overburden pressure.
d) The second and third principal stress—equal to one another and named as pressure at rest—are slowly increasing.

The ratio between the minor and the main principal stress may be called the index of pressure at rest: the word "rest" is refered to the horizontal displacement which is equal to zero. This value is steadily increasing as the densification is going on until the snow transforms into ice, when the value shows its limit, unity (hydrostatic state of stress).

Under a risky assumption of the validity of the law of superposition and of an exponential relationship between stress and strain rate, the index of the pressure at rest ζ can be expressed as follows:

$$\zeta = \frac{\sigma_2}{\sigma_1} = \frac{1}{\sqrt[n]{m_2-1}} . \qquad (1)$$

The factor m_2 represents the ratio between longitudinal and transversal strain under uni-axial unconfined compression. The parameter n means the exponent of the potential dependence between strain and stress. This dependence being linear ($n=1$), we get the well known value for the index of pressure at rest:

$$\zeta = \frac{1}{m_2-1} . \qquad (2)$$

The ratio m_2 depends on the one hand on the density and on the other hand on the crystallographic and structural characteristics of the snow. An open aggregate can, particularly when its structure is feltlike (as for example new snow) be compressed in one direction without remarkable lateral expansion, even when the transvernal deformation is not confined. The m_2 value, *i.e.* is therefore the larger, the higher the porosity of the material is and the more its parts are felted. The ratio m_2 of highly porous plastic materials, for example, is extremely large. With increasing densification, m_2 steadily decreases until to attain the value of 2 as the border stage of pore-free ice, where the volume remains constant. Figure 1 a shows a number of equilateral hyperbolas as possible dependence of the ratio m_2 versus the relative density γ' or the porosity n'.

The distance c, fixing the position of the horizontal asymptote, may be considered as a structural parameter.

Figure 1 b, showing the calculated ratio m_2 as a function of the relative density for a model I with cubic pores, gives a relatively good agreement with the hyperbola for $c=-5$. Certainly m_2 does depend not only on the porosity n', but also on the form and the distribution of the pores.

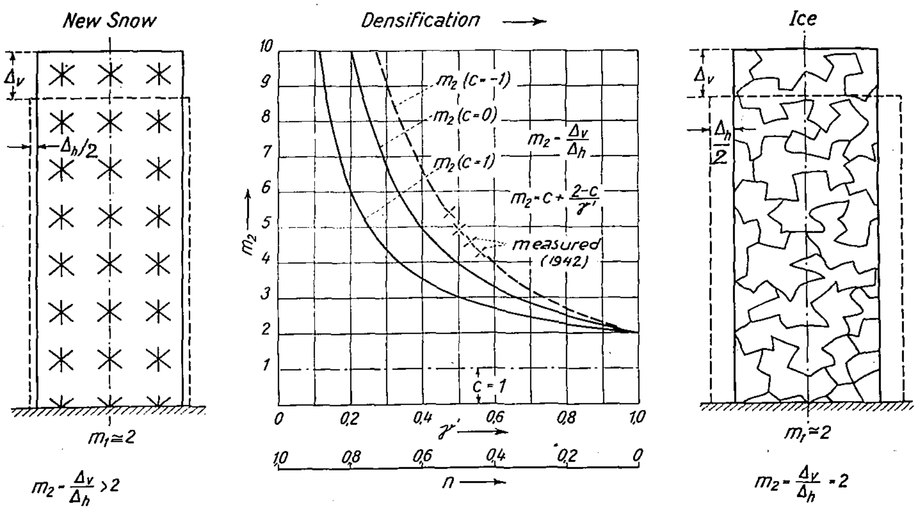


Fig. 1 a. Decrease of m_2 with increasing densification (snow-firn-ice)

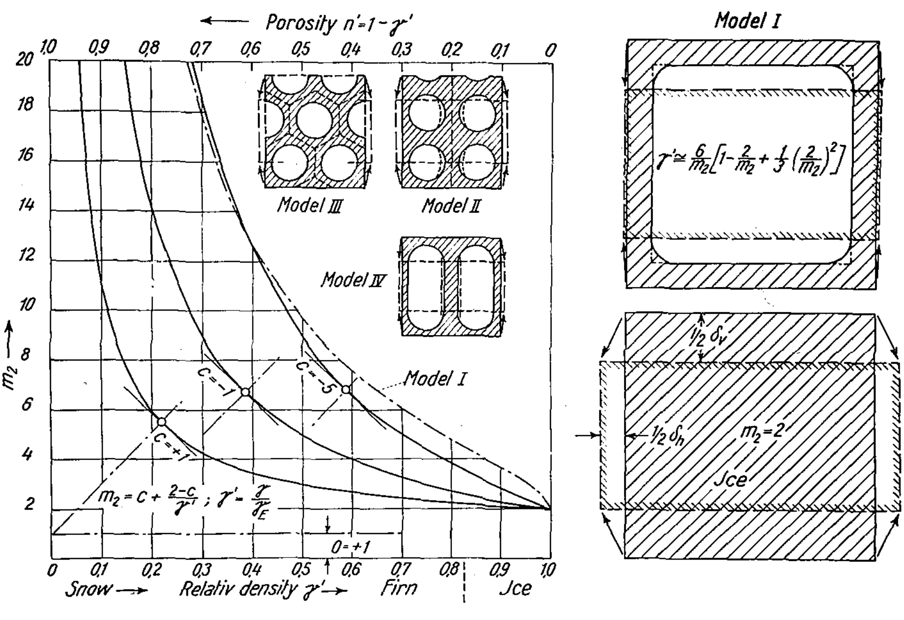


Fig. 1 b. Ratio m_2 for different models

Under the assumption of a symbolic relationship between m_2 and γ' we arrive at the following expressions:

$$m_2 = c + \frac{2-c}{\gamma'}, \tag{3}$$

$$\gamma' = \frac{\gamma}{\gamma_E} = 1 - n'. \tag{4}$$

$$\begin{aligned} \text{If} \quad c = 0, \qquad & \text{then} \quad m_2 = \frac{2}{\gamma'}, \\ c = 1, \qquad & m_2 = 1 + \frac{1}{\gamma'}, \\ c = -1, \qquad & m_2 = \frac{3}{\gamma'} - 1. \end{aligned} \tag{5}$$

where γ_E is the density of ice and n' is void ratio.

When inserting these values of m_2 in eq. (2), we obtain the following dependence of the pressure at rest on the relative density, or on the porosity n':

$$\begin{aligned} \text{If} \quad c = 0, \qquad & \text{then} \quad \zeta = \frac{\gamma}{2-\gamma'} = \frac{1-n'}{1+n'}, \\ c = 1, \qquad & \zeta = \gamma' = 1 - n', \\ c = -1, \qquad & \zeta = \frac{\gamma}{3-2\gamma'} = \frac{1-n'}{1+2n'}. \end{aligned} \tag{6}$$

These are only the possible solutions, the usefulness of which as practical hypothesis has to be examined more thoroughly.

Direct measurement of m_2 has repeatedly been tried (Haefeli, 1942; Landauer, 1955, 1958), but always encountered with great difficulties due on the one hand to the high anisotropism of the natural snow cover, and on the other hand to the lack of homogeneity perpendicular to the layers. Direct measurement of the value of the pressure at rest requires an extremely delicate experimental procedure (de Quervain, 1965; Salm, 1966). A relatively simple method to obtain the ratio m_2 consists in measuring the change of the density during unconfined compression tests. Then the value of m_2 can be calculated by the following equation (Haefeli, 1942):

$$m_2 = \frac{2}{1 - \frac{\Delta\gamma}{\gamma_m} \cdot \frac{h_m}{\Delta h}},$$

wherein:

$\Delta\gamma$ *is the change of density,*
γ_m the average density,
Δh the change of the height of the sample,
h_m the average height of the sample.

2) *The inclined snow cover*

The stress metamorphosis of the inclined snow cover which occurs in a neutral zone, *i. e.* in a zone without cumulative stresses along the slope, has a stabilizing effect. This process having been described in different publications (Haefeli, 1942–1966),

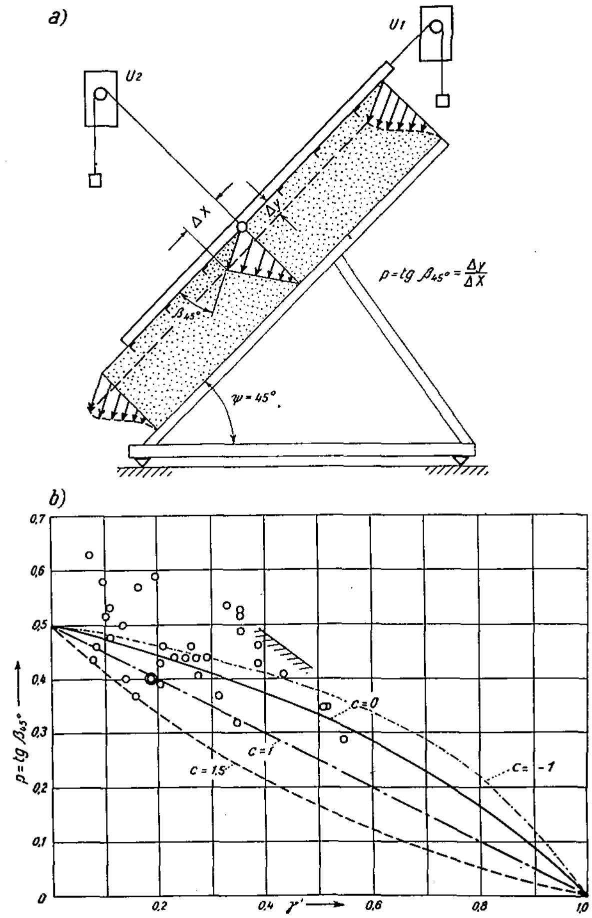


Fig. 2. a) Measurement of the creep parameter $p = \mathrm{tg}\,\beta_{45°}$,
b) Measured values p versus relative density γ'

we can confine ourselves here to the essential point.*

The state of stress and the creep process of an inclined planeparallel snow layer can approximately be reproduced in the laboratory with the equipment shown schematically in Fig. 2a. The creep angle β (between the snow surface and the creep vector) for a given inclination of the snow layer is measured. In order to compare the various snow types with regard to their creep angle, the experiments were carried out with a constant inclination of 45°. The measured values of the creep angles ($p = \mathrm{tg}\,\beta_{45°}$) were plotted in Fig. 2b as a function of the relative density (ratio between the densities of snow and ice). As it was to be expected the creep angle decreases with increasing density. At the final stage, ice as a border case of snow, $\beta_{45°}$ becomes equal to zero.

Former investigations showed the following theoretical approximate relationship between the ratio m_2 and the creep parameter p (Haefeli, 1966):

* We assume that the graphical method to obtain the directions of the principal stresses, based on the creep angle is known (Haefeli, 1939, 1942, 1966a).

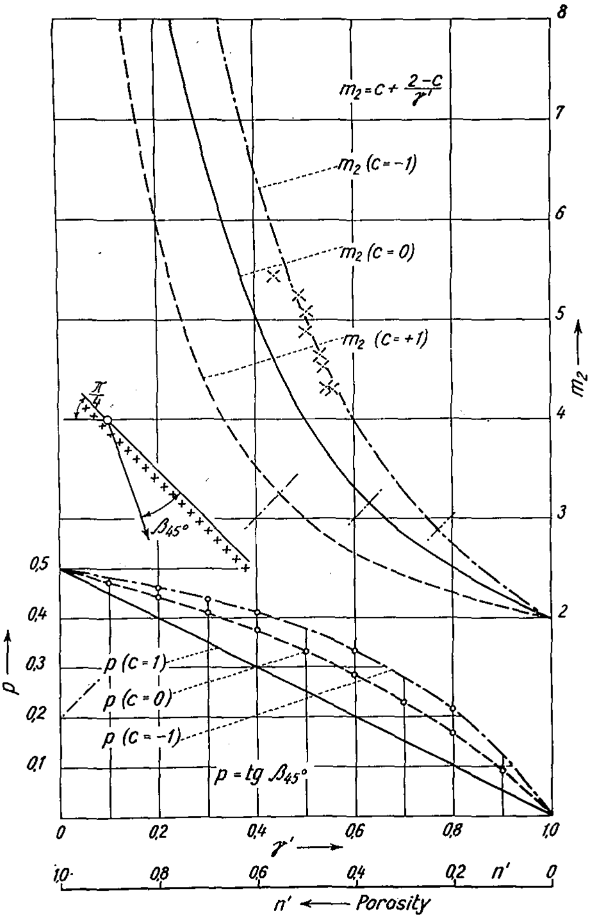


Fig. 3. Ratio m_2 and p as a function of the relative density, respectively of the porosity n' (working hypothesis)

$$m_2 = 2\,\frac{1-p}{1-2p}\,; \quad p = \operatorname{tg}\beta_{45°}\,,$$

$$p = \frac{1}{2}\,\frac{m_2-2}{m_2-1} = \frac{1}{2}\,\frac{1-\gamma'}{1-\gamma'\,\dfrac{1-c}{2-c}}\,. \tag{8}$$

As Fig. 3 shows, the ratio m_2 corresponding to a linear decrease of the creep parameter p with increasing density can be represented by an equilateral hyperbola. A hyperbola for $c=-1$ shows the best agreement with the m_2-values ascertained with uni-axial unconfined compression tests (Haefeli, 1942, p. 24, Table 2 a).

In order to represent the stress metamorphosis of the inclined snow cover, we may consider a simple example. We start from a snow layer with an inclination of 30° and a relative density at the time zero of 0.1 with the parameter $c=1$.

Figure 4, on the right-hand side, shows first of all the creep curve (hyperbola) characterized by a continuously decreasing creep angle β. As a consequence, the changing of directions of the principal stresses which show a counter-clockwise rotation are

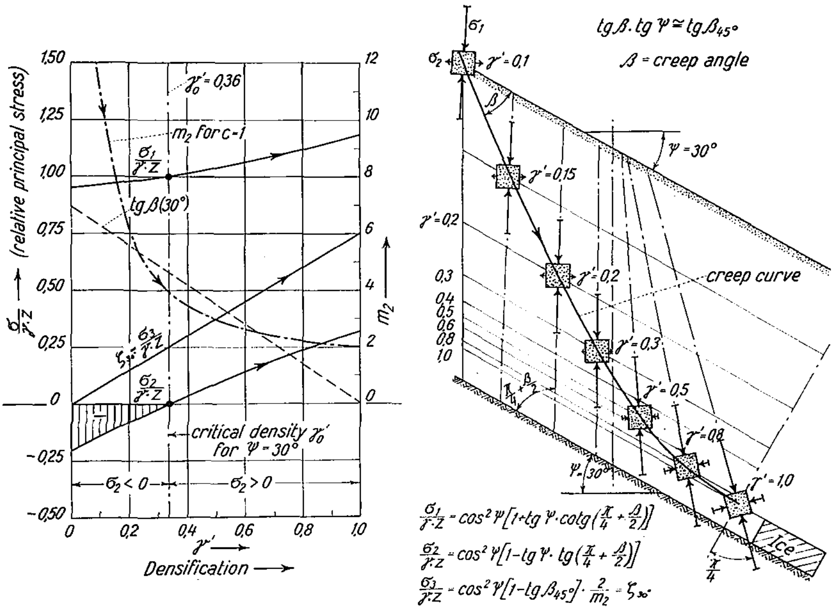


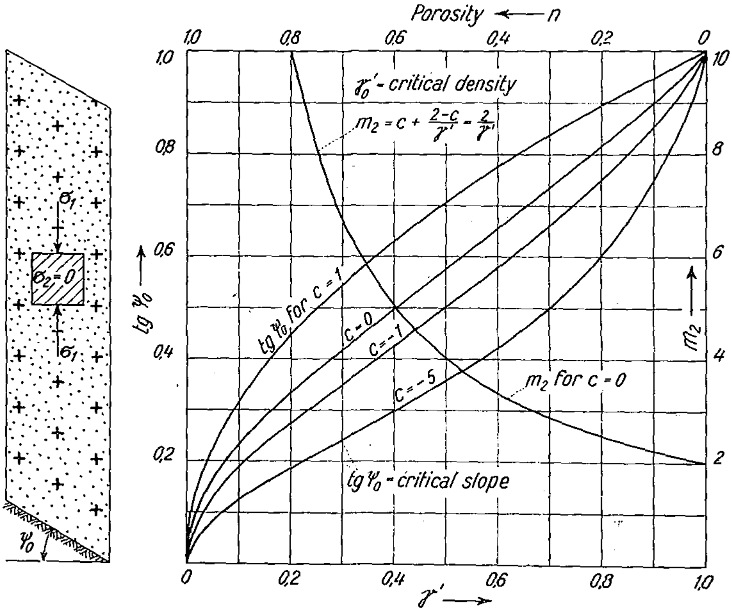
Fig. 4. Stress metamorphosis for a slope inclination of $\psi=30°$, creep curve and principal stresses as a function of the relative density for $c=1$

determined. The creep and densification process which takes place without sliding on the underlayer proceeds decreasing the rate of densification until the whole layer is transformed theoretically into ice. At this final stage, the first principal stress is inclined to the base at 45°. Particular attention has to be paid to the special stage of the consolidation process where the first principal stress is left vertical and the second principal stress is equal to zero, while the relative density reaches the critical value 0.33. This case may be considered as a critical state of stresses (Haefeli, 1966 a).

On the left side of Fig. 4, the ratio between each principal stress and the overburden pressure $\gamma \cdot z$ is expressed as a function of the relative density. Furthermore in Fig. 4 the corresponding fundamental equations for calculating the principal stresses are indicated (Haefeli, 1942). Whereas the main principal stress slowly increases and exceeds the overburden pressure only when reaching the critical density (0.33), the second principal stress is at the beginning negative (tension). At the same moment when the first principal stress σ_1 happens to be vertical and becomes identical with the overburden pressure, σ_2 passes through the value 0 and becomes positive. In this special case which occurs when the critical density is reached, the unstable equilibrium of the snow cover goes over in a somewhat stabler one where all tensile stresses are eliminated. In the same way as there exists for each angle of slope and snow type a critical density ($\sigma_2=0$), there is a corresponding critical slope inclination Ψ_0 for each given density and snow type. As formerly stated, the relation,

$$\mathrm{tg}\,\Psi_0 = \sqrt{\gamma'} = \sqrt{\frac{1}{m_2-1}} = \sqrt{1-2p}\,, \tag{9}$$

is valid here. The dependence of the critical slope inclination on the relative density



c	γ_0'
1	$tg^2\Psi$
0	$\frac{2}{1+cotg^2\Psi}$
-1	$\frac{3}{2+cotg^2\Psi}$
-5	$\frac{7}{6+cotg^2\Psi}$
c	$tg\,\Psi_0$
1	$\sqrt{\gamma'}$
0	$\sqrt{\frac{\gamma'}{2-\gamma'}}$
-1	$\sqrt{\frac{\gamma'}{3-2\gamma'}}$
-5	$\sqrt{\frac{\gamma'}{7-6\gamma'}}$

Fig. 5. Critical slope angle ($\mathrm{tg}\,\Psi_0$) versus relative density for different values of the parameter c

is shown in Fig. 5 for different parameters c (1, 0, -1 and -5) (Haefeli, 1966 a).

This form of stress metamorphosis distinguishes itself also with regard to the third principal stress which is identical with the pressure at rest. The latter increases linearly with the density, but always remains far behind the pressure at rest of the horizontal snow cover. For $c=1$ we obtain the following expression for the index of the pressure at rest of the inclined snow cover:

$$\zeta_\phi = \frac{\sigma_3}{\gamma \cdot z} = \gamma' \cos^2 \Psi_0 = \zeta \cos^2 \Psi_0 \,, \tag{10}$$

where ζ signifies the index of the pressure at rest of the horizontal snow cover (Haefeli, 1966).

The experiments conducted by Landauer (1957) showed that if the snow is not confined, essentially stronger forces are needed in order to obtain a determined densification than under confined transversal deformation or hydrostatic pressure. On a slope the snow is not completely confined, especially not when the angle of the slope is critical ($\sigma_2=0$). The steeper the slope, the more unfavourable are the conditions for the densification of the snow. This explains also the fact that under otherwise identical snow conditions, one sinks deeper in on a slope than on a horizontal plane.

III. On the Formation of Avalanches

Loose snow avalanches and snow slab avalanches form in every respect polar contrasts which does not exclude transition stages. We shall now try to comment this diametrically opposed behaviour.

1) *Loose snow avalanches*

They mainly form during or immediately after the snow fall. As shown before, every settling process in the neutral zone of the snow cover leads to a consolidation. But as long as the tensile stresses are not yet eliminated and the density not higher than the critical one, the situation remains unstable. At a given slope inclination the danger is the greater, the more intensive the snow fall, the lighter the snow and the lower its cohesion. On the other hand there exists for each snow type with a given density a critical slope inclination. The occurrence of loose snow avalanches is bound by this critical condition. This process does, however, depend not only on the density of the loose snow, but on its structure and the shapes of its grains which determine the degree of felting which has a favourable influence, producing a small cohesion. It depends on these small cohesive forces whether the tensile stresses appearing at the beginning of the stress metamorphosis can be absorbed or not.

The critical state of stress of an inclined snow cover shown in Fig. 6 is marked by the absence of the lateral pressure ($\sigma_2=0$). In this case an equilibrium is only possible when, according to Mohr's circle, a certain minimum cohesion c is effective. It follows that the known condition according to which the angle of friction of a slope-parallel discontinuity surface has to be greater than the slope angle does not give a sufficient criterion for the examination of loose snow avalanches. The question is here essentially one of the inner equilibrium. If the material were completely cohesionless the critical state of stress shown in Fig. 6 a had inevitably to lead to the collapse of the structure. Fortunately new snow is hardly ever cohesionless (except wild snow) and its cohesion increases with increase of density. On the other hand the destructive metamorphosis is often connected with a nearly complete loss of cohesion (swim-snow etc.).

If loose snow possesses an even very small cohesion c and a given angle of inner

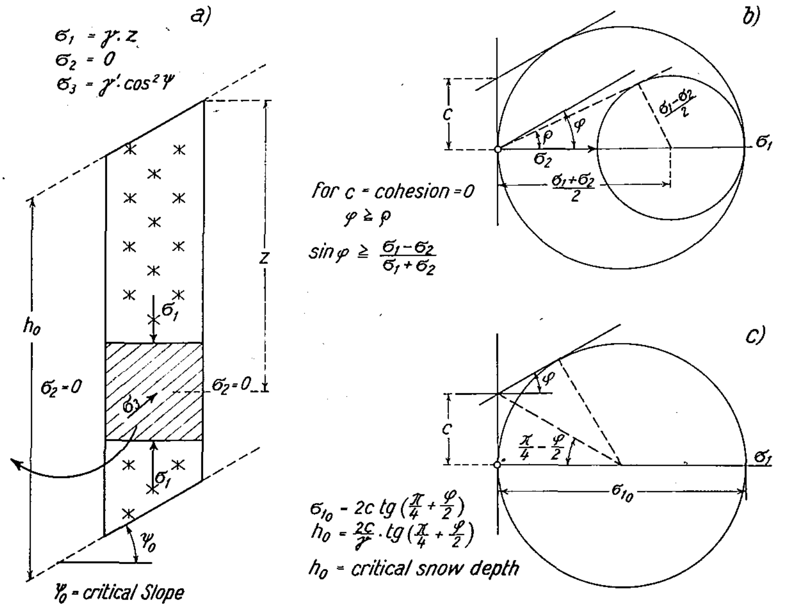


Fig. 6. Critical state of stress ($\sigma_2=0$) of the inclined snow cover (pattern)

friction, the equilibrium is only assured when h remains inferior to the critical value, *i.e.* when the height of the snow does not exceed a certain amount. This critical value h_0, or the critical height of snow are calculated according to Fig. 6 c as follows:

$$h_0 = \frac{2c}{\gamma} \cdot \mathrm{tg}\left(\frac{\pi}{4}+\frac{\varphi}{2}\right). \tag{11}$$

Experience shows that danger of loose snow avalanches increases with increase of height of snow until finally the critical height for a given intensity of snow fall is reached. As Fig. 6 a shows, the disturbance of the equilibrium starts at one point as a sudden local collapse of the loose structure, in connection with a lateral squeezing or flowing out of the disturbed material, preluding a chain reaction within an area of small angle directing downwards.

2) *Snow slab avalanches*

Contrary to the loose snow avalanches which can be traced back to a lack of cohesion, the snow slab avalanche occurs when individual layers have a surplus of cohesion and consequently a lack of plasticity. Apart from the opposed shape of rupture (failure at one point and broad sharp-rupture), there exists between the two types of avalanches also a marked difference in their temporal development. Whereas the loose snow avalanches occur mainly during or immediately after the snow fall, *i.e.* at the beginning of the metamorphosis (except swim snow and wet snow) the snow slab avalanches go during a slow maturing process which leads to that extremely labile situation where the lightest disturbance causes the rupture. LaChapelle (1965) does not, without good reason, distinguish between "direct-action avalanche" (during or after the snow storm) and "climax avalanche" (with a long development). The above mentioned maturing process of the snow slab avalanche consists in the fact that there occurs a kind of regression of the inner friction in the slope-parallel potential sliding layers as a result of the braking of the normal creeping process in the anchoring zone. This implies a gradual increase of the tensile force Z until this force finally overcomes the tensile strength of the breaking layers. This process forms the exact correlate to the slow increase of the creep pressure on avalanche construction work, but with negative signs (Haefeli, 1939, 1966) Fig. 7.

From the point of view of snow mechanics, the release of the snow slab avalanche depends—apart from the shear strength of the sliding surface—first of all on the viscosity and the tensile strength of the breaking layers. As a whole we have to deal with a progressive rupture in 3 different phases. As it is the case for any progressive rupture, the tensile strength of the breaking layer complex is always smaller than the sum of strengths of the individual layers. Besides, it has to be considered that the strains are very irregularly spread over the whole section, according to the position of the resultant Z and the viscosity conditions of the different layers.

In order to get an idea over the influence of the viscosity on the stress conditions in a very simple case, we assume in Fig. 7 that a relatively hard snow layer of the thickness b is sandwiched between two softer layers of the thickness $a/2$. This sequence of layers is exposed to uniaxial tension by a centric tensile force Z (anchoring force per 1 m rupture), the question being that of the stress distribution over the whole

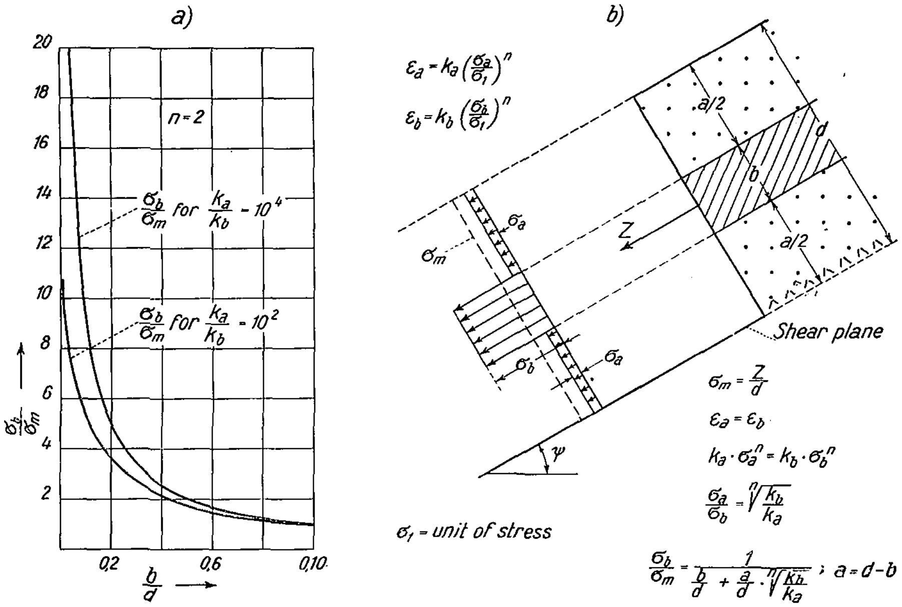


Fig. 7. Distribution of the tensile stresses in the critical cross-section of the tensile zone (pattern)

section (d).

Supposing that the strain rate $\dot{\varepsilon}$ increases with the n-th power of the tensile stress, whereas the sections remain plane, we get according to Fig. 7 the following approximate expression for the ratio between the greatest stress σ_b and the average tensile stress σ_m:

$$\frac{\sigma_b}{\sigma_m} \cong \frac{1}{\lambda+(1-\lambda)\sqrt[n]{\dfrac{k_b}{k_a}}}, \tag{12}$$

where

$$\sigma_m = \frac{Z}{d}, \quad \lambda = \frac{b}{d}.$$

Figure 7 a represents the above mentioned stress ratio for $n=2$ as a function of the ratio λ. We see that this quotient ($\sigma_b : \sigma_m$) attains higher values, the more the k-values, or the viscosities of the two types of snow differ from one another and the smaller b is compared with the thickness of the layer d. That means that the softer layers undertake the load only to a very low degree. The ideal case where each layer carries its part in proportion to its tensile strength is never fulfilled in nature, apart from the fact that, as a rule, the pressure Z acts in reality eccentric, contrary to Fig. 7.

Generally it can only be said that a snow layer resists rupture the better the greater the product of its k-value (or k_{11} in eq. (13)) and its strength. Besides, attention has to be paid to the fact that the danger of rupture depends in each layer on the continually changing relationship between stress and strength. As a consequence, a decrease of the temperature can have an unfavourable effect when the stress increases relatively more than the strength. At last it has to be considered that the hard layers in the tensile zone are subjected not only to pure tensile stresses, but simultaneously to shear stresses which we did not put into account (Jaccard, 1965). Concerning the viscosity, the difference of temperature between the individual snow layers must be considered.

IV. On the Measurement of the Viscosity of Snow and Firn

The viscosity conditions play a decisive role not only in the formation of snow slab avalanches but also in the breaking of loose snow and ice avalanches. It is therefore worth-while to consider the viscosity conditions within the whole range (from the lightest snow type to ice). The difficulty of such a comparison on a broad basis consists in the fact that the law of flow is not exactly the same for snow and ice. For ice the exponential law of Glen (1952) is valid, whereas for snow a linear dependence between stress and deformation rate may be assumed at least for small stresses (Landauer, 1957).

Our comparative viscosity data (reduced to $-10°C$) which have been collected partly on the alpine winter snow cover, partly on polar firn of the Greenland inland ice, are based on the following assumptions concerning the stress and strain rate:

$$\begin{aligned}
&\text{Snow and firn for small stresses}, \quad D = k_{11}\left(\frac{\tau}{\tau_1}\right)^n, \quad n=1\,.\\
&\text{Snow and firn for large stresses}, \quad D = k_1\left(\frac{\tau}{\tau_1}\right)^n, \quad n=3\,.\\
&\text{Ice for small and large stresses}, \quad D = k_1\left(\frac{\tau}{\tau_1}\right)^n, \quad n=3\,.
\end{aligned} \tag{13}$$

Between the viscosity η and the parameter k_{11} we have the following relationship:

$$\eta = \frac{\tau_1}{k_{11}} \quad \text{or} \quad k_{11} = \frac{\tau_1}{\eta}\,. \tag{14}$$

The measurements of viscosity in the firn mantle of the great ice sheets (Greenland and Antarctic) is especially informative. For a steady climate not only the density (Bader, 1954) but all characteristics of the firn in general (viscosity, strength etc.) are determined by the depth z under the firn surface. Making use of the supposed relationship between the pressure at rest and the density (eq. (6)), the pressure at rest for the firn can be calculated based on the density curve. This procedure was applied when evaluating the measurement of the International Glaciological Expedition to Greenland. The corresponding results are shown in Fig. 8, for $c=0$ and $c=1$ (Haefeli and Brandenberger, 1967).

When measuring *in situ* the average viscosity of the alpine winter snow cover we can use the clinometer shown schematically in Fig. 9. The change of inclination of a "Plexiglas" tube placed vertically in the snow cover is measured from time to time. The calculation of the k_{11} value, or of the viscosity η is based on the following approximate equation which gives exacter results, if the better the angle Ψ corresponds to the critical value Ψ_0 of the slope ($\sigma_2=0$):

$$k_{11} = m_2 \operatorname{tg}\Psi \frac{4\tau_1}{\gamma \cdot h} \frac{\Delta\alpha}{\Delta t}\,; \tag{15}$$

where $\tau_1 = 1\,\text{kg/cm}^2\,;\quad \Psi \cong \Psi_0\,;\quad t = \text{time}\,.$

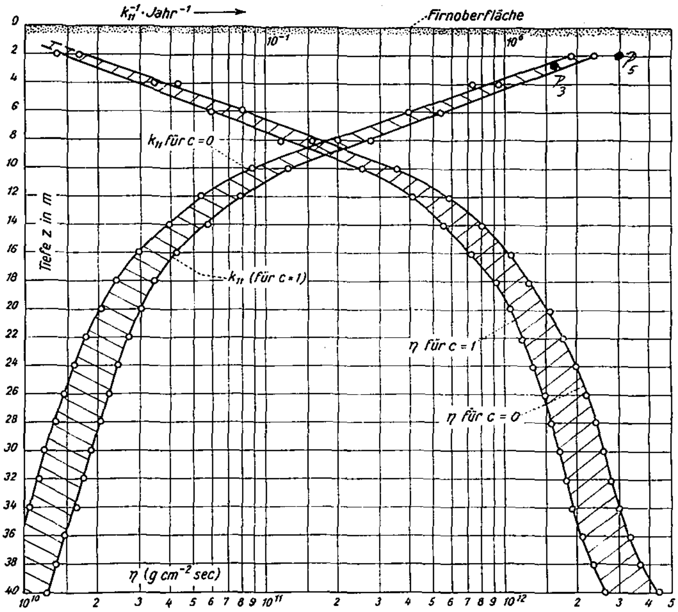


Fig. 8. Viscosity as a function of depth z (in m) at Jarl-Joset Station (International Glaciological Expedition to Greenland 1957–60) for $c=0$ and $c=1$

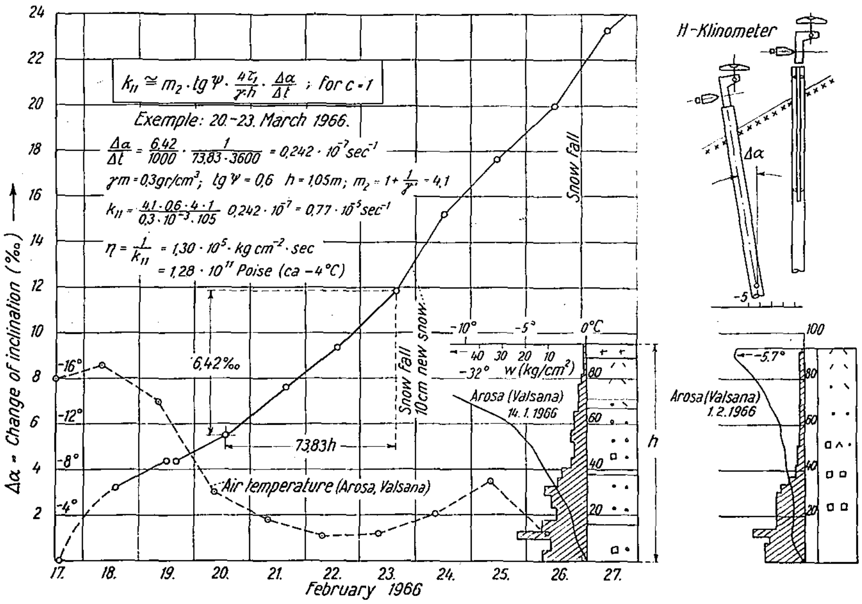


Fig. 9. Viscosity measurement *in situ* with H-Klinometer (Schamoin) Lenzerheide (January 1966)

The last Fig. 10 shows the viscosity values of snow and firn measured by different authors, all over the world. Attention may particularly be paid to the relatively good agreement between the values found in Japan by Yosida and his collaborators (1956) and those on the alpine winter snow cover in Switzerland measured by the author between 1936–1966.

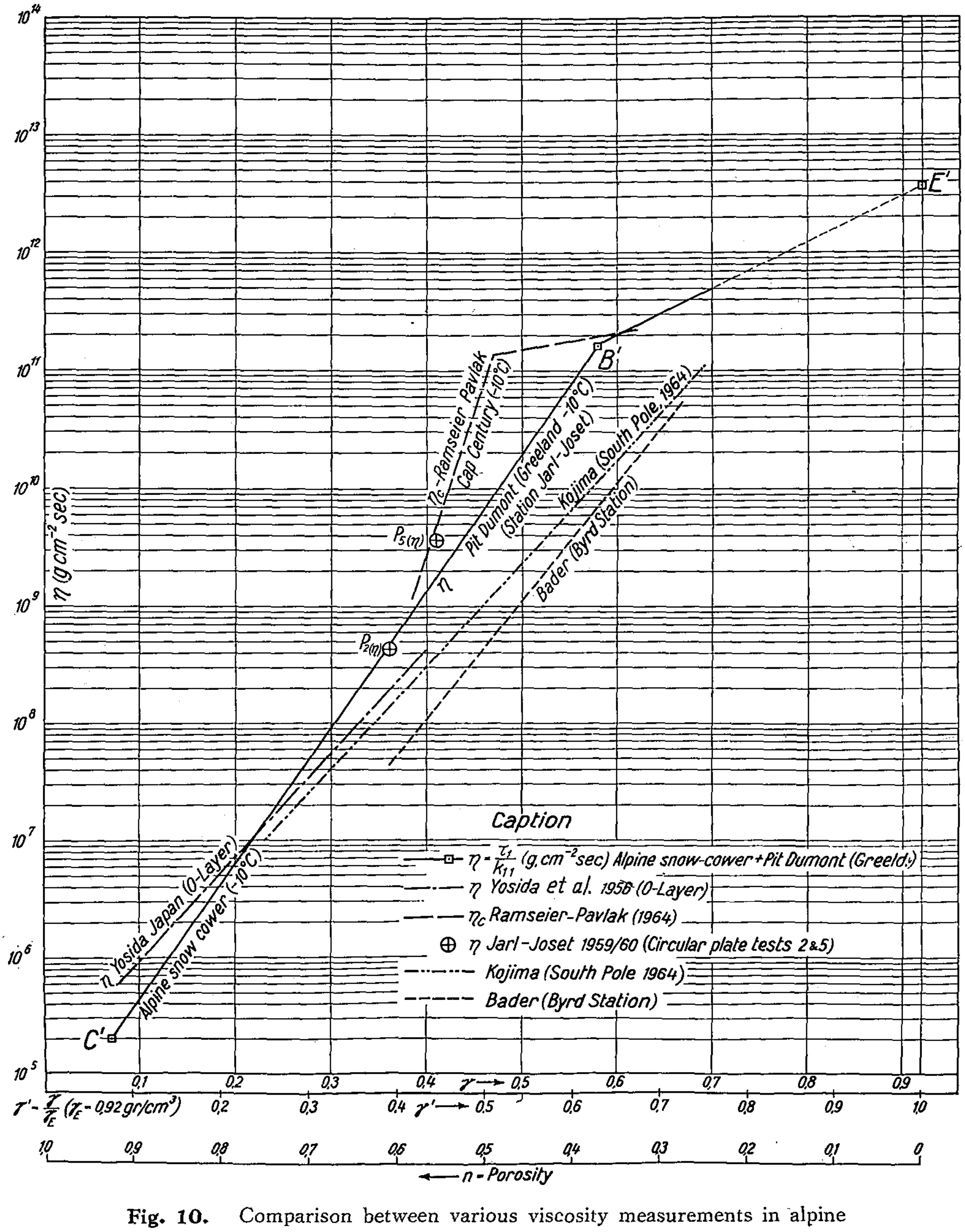


Fig. 10. Comparison between various viscosity measurements in alpine and polar regions

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