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STUDY OF AN ICE CORE TO THE BEDROCK IN THE ACCUMULATION ZONE OF AN ALPINE GLACIER

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Abstract. A water table appearing every summer where the ice begins, at a depth of approximately 30 m, accelerates the transformation of firn into ice during the summer (80% of the ice formed every year appears in less than 2 months). The ice formed in this way contains from 0 to 0.6% water. The average water content increases gradually with the depth because of the heat of deformation. But, near bedrock, between 180 and 187 m, the permeability of the blue ice is such that the water content drops (0.3% as compared to 1.3% between 160 and 180 m).

From a depth of 33 m, a foliation of sedimentary origin gradually develops in the ice. Its dip increases regularly to a depth of 145 m. At 145 m it jumps suddenly from 20° to 40°, then at 170 m from 40° to 65°, which can be explained by old modifications in the bergschrund. This foliation disappears near bedrock (180–187 m), where there are no bubbles in the ice.

The average size of an ice crystal increases slowly in the firn, shows seasonal fluctuations between 30 and 50 m, then jumps from a diameter of 1 or 2 mm to 10 or 20 mm between 50 and 80 m. Between 180 and 187 m, the ice is made of large crystals (3–10 cm diameter; the figure, however, is probably inexact due to a recrystallization of the samples).

The very strong sub-vertical orientation of the optic axes of the firn crystals disappears quickly, and from 66 m on, in ice with large crystals, a fabric of multiple maxima appears (generally, 3 or 4 directions, forming a triangle or a rhombus). On the other hand, in the small crystals that form bands parallel to the plane of foliation, only one direction of preferential orientation can be seen, or two close to one another. Crystals of intermediate size (10 to 50 mm) generally have two directions of preferred orientation at an angle of approximately 50° to one another. No matter how big the crystals are, the angle between the most common *c*-axis orientation and the vertical does not change from 60 to 170 m depth.

Résumé. *Étude d'une carotte de glace jusqu'au lit dans la zone d'accumulation d'un glacier tempéré.* Un niveau aquifère apparaissant chaque été au contact de la glace, vers 30 m de profondeur, accélère la transformation du névé en glace durant l'été (80% de la glace formée chaque année apparaît en 2 mois). La glace ainsi formée renferme de 0 à 0,6% d'eau liquide. La teneur moyenne en eau liquide augmente progressivement avec la profondeur, par suite de la chaleur de déformation. Mais au contact du lit, entre 180 et 187 m, la perméabilité de la glace bleue est telle que la teneur en eau devient faible (0,3% contre 1,3% entre 160 et 180 m).

Dans la glace, à partir de 33 m de profondeur, se développe progressivement une foliation d'origine sédimentaire. Son pendage croît régulièrement jusqu'à 145 m de profondeur. A 145 et 170 m, il passe brusquement de 20° à 40° puis de 40° à 65°, ce qu'on explique par d'anciennes modifications de la rimaye. Cette foliation disparaît au voisinage du lit (180 à 187 m) où la glace ne renferme aucune bulle.

La taille moyenne des cristaux augmente lentement dans le névé, montre des fluctuations saisonnières dans la glace entre 30 et 50 m, puis passe rapidement de 1 ou 2 mm à 10 ou 20 mm de diamètre entre 50 et 80 m. Entre 180 et 187 m la glace est formée de gros cristaux (3 à 10 cm de diamètre, valeur toutefois probablement faussée par une recristallisation des échantillons).

L'orientation subverticale très forte des axes optiques des cristaux de glace du névé disparaît rapidement et dès 66 m apparaît dans la glace à gros cristaux une texture à maximums multiples (généralement 3 ou 4 directions dessinant un triangle ou un losange). Par contre, les tout petits cristaux formant des bandes parallèles au plan de foliation montrent une seule direction d'orientation préférentielle, ou deux proches. Les cristaux de taille intermédiaire (10 à 50 mm) ont généralement deux directions d'orientation préférentielle à une cinquantaine de degrés l'une de l'autre. Quelle que soit la taille des cristaux, l'angle entre la direction la plus fréquente d'orientation des axes *c* et la direction verticale ne varie pas de 60 à 170 m de profondeur.

Zusammenfassung. *Untersuchung eines Eisbohrkernes bis zum anstehenden Fels in der Akkumulationszone eines alpinen Gletschers.* Eine wasserführende Schicht, die jeden Sommer in etwa 30 m Tiefe, dort wo das Eis anfängt, erscheint, beschleunigt die Umwandlung von Firn in Eis während des Sommers (80% des jährlich gebildeten Eises entsteht in weniger als zwei Monaten). Das auf diese Art gebildete Eis enthält zwischen 0 und 0,6% Wasser. Der durchschnittliche Wassergehalt nimmt wegen der Deformationswärme allmählich mit zunehmender Tiefe zu. In der Nähe des Felsuntergrundes hingegen, zwischen 180 und 187 m, ist die Durchlässigkeit des blauen Eises so hoch, dass der Wassergehalt abfällt (von 1,3% zwischen 160 und 180 m auf 0,3%).

Von einer Tiefe von 33 m an entwickelt sich im Eis allmählich eine Bänderung sedimentären Ursprungs. Ihre Neigung nimmt stetig bis zu einer Tiefe von 145 m zu. Bei 145 und 170 m springt sie plötzlich von 20° auf 40°, dann von 40° auf 65°, was durch ältere Veränderungen im Bergschrund erklärt werden kann. Diese Bänderung verschwindet in der Nähe des Felsuntergrundes (180–187 m), wo das Eis blasenfrei ist.

Die Durchschnittsgrösse eines Eiskristalls, die im Firn langsam zunimmt, fluktuiert jahreszeitlich zwischen 30 und 50 m, dann, bei 50–80 m, wächst der Durchmesser rasch von 1 oder 2 mm auf 10 oder 20 mm an. Zwischen 180 und 187 m besteht das Eis aus grossen Kristallen (3–10 cm Durchmesser; diese Zahl ist jedoch wegen einer Rekristallisation der Proben wahrscheinlich ungenau).

Die stark ausgeprägte subvertikale Orientierung der optischen Achsen der Firnkristalle verschwindet schnell und von einer Tiefe von 66 m an, im grosskristalligen Eis, ist eine Textur mit mehreren Maxima vorzufinden (im allgemeinen 3 oder 4 Richtungen, die ein Dreieck oder einen Rhombus bilden). Andererseits kann in den kleinen Kristallen, die Bänder parallel zur Bänderungsebene bilden, nur eine Hauptorientierungsrichtung festgestellt werden, oder aber zwei, die nahe beieinander liegen. Kristalle mittlerer Grösse (10–15 mm) bestizen im allgemeinen zwei bevorzugte Orientierungsrichtungen, die unter einem Winkel von 50° zueinander stehen. Unabhängig von der Grösse des Kristalls bleiben die Winkel zwischen den am häufigsten vorkommenden Richtungender *c*-Achse und den Vertikalen in der Tiefe von 60 bis 170 m unverändert.

1. Introduction

Though the texture of firn and of deep ice at the polar ice caps is fairly well known, thanks to the work of, for example, Schytt (1958), Langway (1967), and Gow (1970), the same is not true for the accumulation zone of temperate glaciers, except for the pioneeing work of Perutz and Seligman (1939). We have tried to fill this gap by drilling to bedrock in the accumulation zone of a temperate glacier.

The site chosen was the upper plateau of the Vallée Blanche in the Massif du Mont-Blanc, at an altitude of approximately 3 550 m. It has already been the site of several previous studies because of its accessibility thanks to the cable-car (téléphérique) of the Aiguille du Midi, and to the service cable-bucket of the Refuge des Cosmiques, belonging to the Laboratory. A seismic survey made a calculation of the thickness of the ice possible: it is a very uniform 145 m covered by 30 m of firn (Lliboutry and Vivet, 1961). A measure of the solid discharge coming out of this cirque led to the calculation of the average yearly balance at 2.70 m of water. In 1960, the firn was studied with pits and descents into crevasses. With a SIPRE coring drill, samplings were carried out to a depth of 55 m in 1963, 36 m in 1966 and 33 m in 1970 (Vallon, unpublished). In June and July 1971, using an electric drill developed by Gillet (1975), the first complete sampling of an alpine glacier in its accumulation zone was performed. 30 m of firn and 150 m of ice were found.

2. Stratigraphy of firn

In the winter, 6 to 8 m of light, cold, powder snow accumulates. Owing to the extremely low thermal conductivity of the snow, there is very little diffusion of the cold towards the older layers; in the spring, the snow is only really cold (−4°C to −10°C) in the first 5 to 7 m in the layer deposited during the winter; beyond 12 to 15 m, the firn from the previous years always stays at 0°C.

In the spring, the water that filters down will refreeze upon contact with the cold firn. The amount of cold stored in the snow will only refreeze a significant quantity of water if its temperature is below −10°C. The appearance of a layer containing a number of strata of ice, the density of which is about 10% higher than that of the surrounding snow, is thus seen at a depth of approximately 2 m, in the coldest zone.

In the summer, the precipitation, almost always in solid form, is destroyed by ablation, and the dust that it contains concentrates on the surface to form a grey cover.

In autumn, there is again diurnal thawing, but the humid firn refreezes every night near the surface, and a number of more or less thick strata of ice are formed.

Although the sequence is sometimes difficult to distinguish, when both the descriptions of the samples and their densities are available, interpretation is always possible. The study of four drillings (1963, 1966, 1970 and 1971) has made it possible for us to trace the annual layers back to the budget year 1954–55. However, the refreezing of the melt water in the spring ("internal accumulation") takes place in that year's layer, and in the two or three preceeding layers as well. The balance for the successive budget years cannot, therefore, be determined exactly. Only the average balance for the period 1954–71 is close to exact.

This average balance is 318±4.6 cm of water per year. Because every summer some 20 cm of water percolates to the bottom of the firn, the annual accumulation should be larger, close to 3.5 m of water per year. Several kilometres away, at the Col du Géant, in a location almost as high (3 370 m), but better protected from the western wind, a precipitation gauge has only captured 2.09 m of water per year between 1961 and 1970 (Bezinge and Bonvin, unpublished). At Chamonix (1 035 m), the average annual precipitation between 1 October 1954 and 1 October 1971 was 128.5 cm of water per year. These figures are similar to those measured in the Bernese Oberland.

3. Crystalline texture of the firn

The firn of one year is essentially made up of round crystals with 1 mm cross-sections. The growth of the crystals is very slow, and 6 years later, at a depth of 30 m, the average area of cross-section is about 2 mm, the largest crystals reaching 10 mm.

Unlike in cold firn (Schytt, 1958; Langway, 1962; Gow, 1963), the anisotropy here is very strong, and all the stereograms show a single sub-vertical preferential direction for the optic axes (Fig. 1). Though certain stereograms show that the preferential orientation is reinforced with depth between the surface and 15 m, they do not allow the determination of which orientations disappear through recrystallization. In Figure 2, the number of axes observed in zones 10° wide, centered on the vertical, is shown for the textures showing an axis of symmetry close to the vertical.

Assuming that snow crystals that settle are oriented at random, the histograms show that the crystals whose optic axes form an angle of 40° to 60° with the vertical are the first to disappear, leaving those whose optical axes are close to vertical. Recrystallization in the first few metres is certainly due to seasonal fluctuations of temperature, which create in the surface layer a fairly strong thermal gradient. In fact, two positions thermodynamically favorable to the growth of a crystal in a thermal gradient seem to exist: when the hexad axis, or one of the binary axes is parallel to the gradient (optic axis parallel or perpendicular to the gradient), the ice crystals grow fairly rapidly (Shumskiy, 1955). Basal ice from an ice core in Terre Adélie shows a similar fabric with two preferred orientations; in this case we are dealing with ice which probably is of marine origin (Lorius and Vallon, 1967); that ice recrystallizes in a thermal gradient and this recrystallization involves a strong desalination (actual sodium content 20×10^{-6}).

Lower, between 10 and 30 m below the surface, the crystals whose axes vary from the vertical by more than 30° are the first to disappear. In this zone, the temperature is almost constant, so another factor must enter, probably stresses. In fact, the stresses are not hydrostatic in the firn. Knowing the vertical and horizontal velocities we can use Haefeli's construction (Bader and others, 1939) to determine the directions of the stresses (this assumes no sliding, no thickness variation and equilibrium between stresses on the vertical faces of a prism). Knowing the surface slope we find the direction of compression is about 20° off the vertical, the maximal extension 70° (Petit, unpublished). Between 20 and 30 m, the preferential direction is no longer rigorously vertical, but is inclined from 10° to 20° (Fig. 1).

On passing through the water table, at a depth of approximately 33 m, the preferential orientation which had appeared, disappears somewhat. The tendency of the optic axes not to vary from the vertical by more than 40° continues, nevertheless, to be observed.

4. Variation of the density with the depth, and transformation of the firn into ice

In Figure 3, the densities observed for the different seasonal horizons are represented. The metamorphosis of the spring and autumn horizons is much faster than the metamorphosis of the winter horizons. Since they contain a number of ice strata, their permeability is lower.

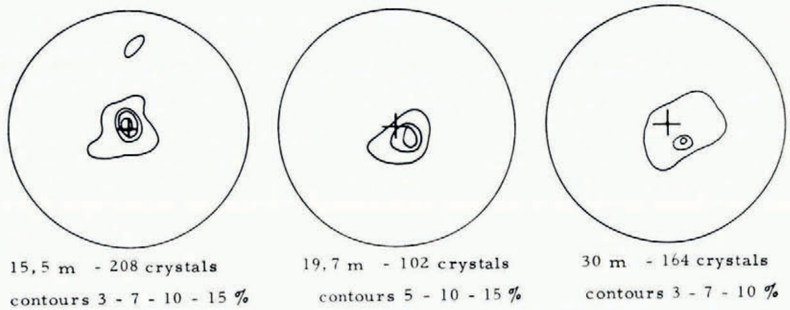

Fig. 1. Orientation of optic axes of the ice crystals in firn. Horizontal sections, equal-area projection on the upper hemisphere. The contour lines correspond to densities (percentage of points in 1% of the area) of 3, 7, 10 and 15%.

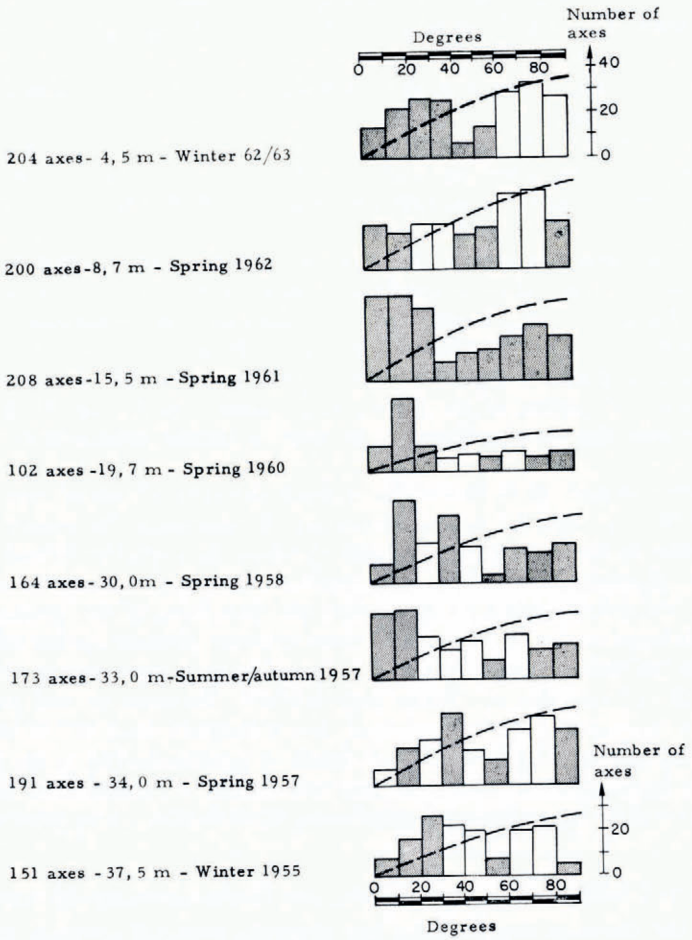

Fig. 2. Orientation of the optic axes versus depth. The hatched line represents the frequencies corresponding to a uniform density; frequencies that depart from this curve significantly (approaching 2%) are shaded.

They retain more water, settle more quickly, and normally reach the density of ice at a depth of approximately 28 m. Winter horizons, on the other hand, turn abruptly into ice at 32 m.

Every summer, a water table develops on the ice. The level of this water table has been studied during the summer of 1971, in the drill hole of June 1970, with the aid of a limnigraph. In addition, on 1 and 17 September, the exact depth and thickness of the aquifer were determined by noting the level of the water in the bore hole for different rates of pumping (Petit, unpublished).

In June, the firn was dry and we introduced some tens of litres of water into the hole to determine the impermeable level. Ablation began around 1 July, and the level of the water started to rise towards 15 July; around 15 August, the ablation season was over, but the water level continued to rise until 1 September. The water table being at a distance of about 30 m

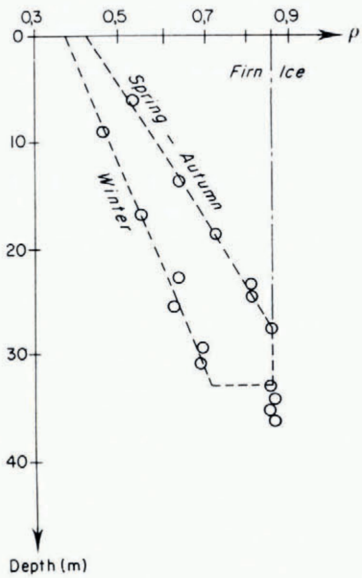

Fig. 3. Density versus depth.

from the surface, that corresponds to a speed of the water wave of 12 cm per hour, a speed comparable to that observed for the daily melt-water crest near the surface of the Seward Glacier by Sharp ([1952]).

The rise of the water, and of the impermeable level are reported in Figure 4. From 15 July to 15 September, the level of the ice rose 2.9 m. Since in 1963, 1966, 1970, and 1971, all the drillings hit ice at the same depth, the average annual speed of the rise of the impermeable level is almost the same as the vertical speed of the ice at the base of the firn: about 3.5 m per year. Thus, more than 80% of the ice is formed during the summer, during the time that the firn just above the firn–ice transition is saturated with water.

It is out of the question that there should be a simple closing-off of the pores at this time. While the firn at a depth of 30 m can contain 8 to 10% water, less than 1% remains in the impermeable ice (cf. below). The water present must intervene to promote the settling and the sintering of crystals (Wakahama, 1968).

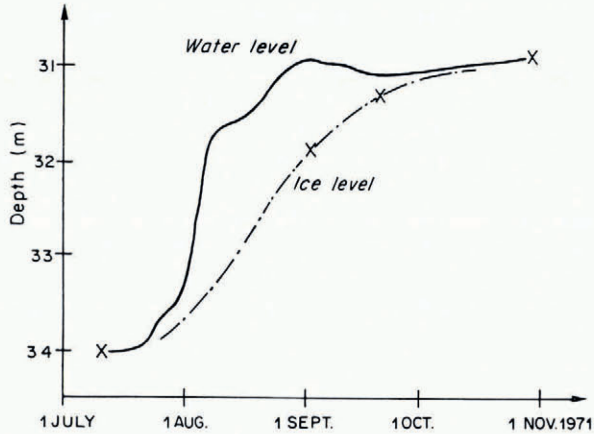

Fig. 4. Level of the impermeable horizon, and level of the water table during the summer of 1971.

5. Stratigraphy of the ice

We found three kinds of ice: full of bubbles, without foliation; foliated (with alternation between bands of blue ice without bubbles, and white ice with bubbles), and blue (without bubbles). The percentage of these three types of ice for successive lengths of 10 m is given in Figure 5.

Down to a depth of about 100 m, the percentage of foliated ice (20–35%) is almost the same as in the foliated layers of the spring and autumn firn, and one can guess that it comes directly from that. Yet, the distinction between foliated ice and non-foliated ice is not clear enough to be used to identify the annual layers of the ice from its stratified texture alone with certainty.

It is quite possibly for this same reason that the percentage of foliated ice increases with depth below 100 m. The vertical contraction of the ice, as it is buried, results in the classification as "entirely foliated", of a zone where both foliated and non-foliated sections would have been distinguished before contraction.

All of the rare layers of debris observed between 60 and 171 m are concordant with the foliation. We conclude from this that in the Vallée Blanche Supérieure, the foliation has a sedimentary, and not a tectonic origin. (The reader will find an exposé of this question, debated for over a century, in Lliboutry (1964–65, Tom. 2, p. 612–14).)

In Figure 5 the dip of the foliation is also given. At the start, at 30 m depth, it is equal to the slope of the surface (5 to 7%), which is to say that the blue bands are parallel to the surface. Then, this dip increases gradually, with two sharp discontinuities, at 147.7 m and 171.5 m (i.e. 39.3 m and 15.5 m from bedrock). Although the device for measuring the strike did not work, there is no reason to believe that the dip of the foliation changes its bearing with depth. It should always remain downwards.

Between 147.6 and 147.8 m there exists a layer of small crystals (1 to 5 mm^2 in cross-section) with silt in it. The dip of this layer is 20 to 30°. In the next sample, just below, the foliation is clear, with a 45° dip.

At 170 m the dip of the foliation is 45°. Between 170 and 171.2 m the ice was not kept for later study. The next sample showed however, towards 171.5 m, a bed of gravel concordant with a relatively discrete foliation, having a dip of 60° from the horizontal.

Near these two discontinuities, we see that the preferential orientation of the optic axis is very strong (15% of the points in 1% of the area). Once past the discontinuities, this

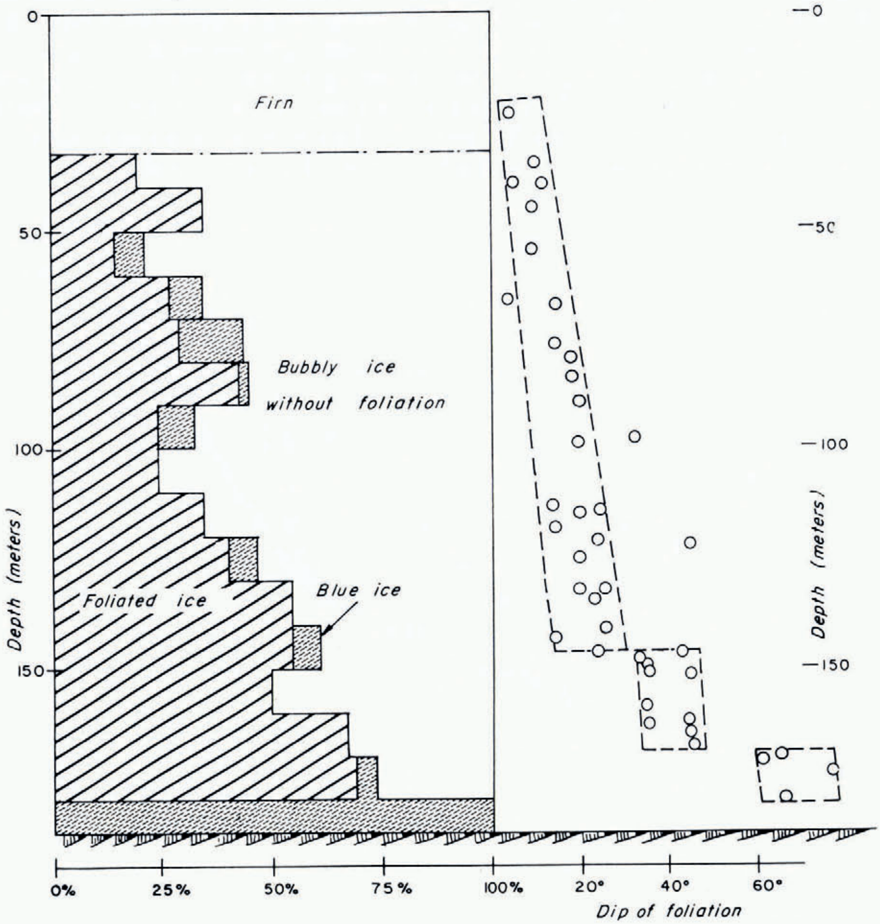

Fig. 5. Percentage of foliated, bubbly, and blue ice, and slope of the foliation as a function of the depth.

preferential direction maintains the same relation to the vertical, with no relation to the foliation, which confirms that the foliation has nothing to do with syntectonic recrystallization (Vallon, unpublished; Jonsson, 1970).

If we assume a uniform vertical contraction (Nye's flow, according to Lliboutry (1964–65, Tom. 2, p. 576)) and an average balance b without secular variation, the age t of a sheet located at a height z above bedrock is:

$$t = (h/b) \log (h/z).$$

30 m of firn equal 21 m of ice, so h will be $h = 178$ m, $b = 3.6$ m per year. It follows that the two discontinuities at 39.3 and 15.5 m from bedrock were formed 75 and 121 years ago (1896 and 1850), that is, at the time of the two principal retreats of the alpine glaciers. The ice flow velocity at the location of our drill hole being 8 m per year (Petit, unpublished), an estimation of the route followed in approximately 120 years leads us to the slopes of Mont Blanc du Tacul, near the bergschrund.

According to Professor Lliboutry, these two discontinuities are the traces of an old bergschrund pulled down-valley. Normally, the bergschrund separates a base of more or less motionless cold ice, joined to the wall of the cirque, from the temperate glacier. A long cold period should be able to make the cold ice base get bigger, with strata of very sharply inclined ice parallel to the slope. At the time of a warming of the climate, a new bergschrund would form further up-ice as shown in Figure 6. However this does not explain the slow increase in dip of stratification with the depth between the surface and 147 m.

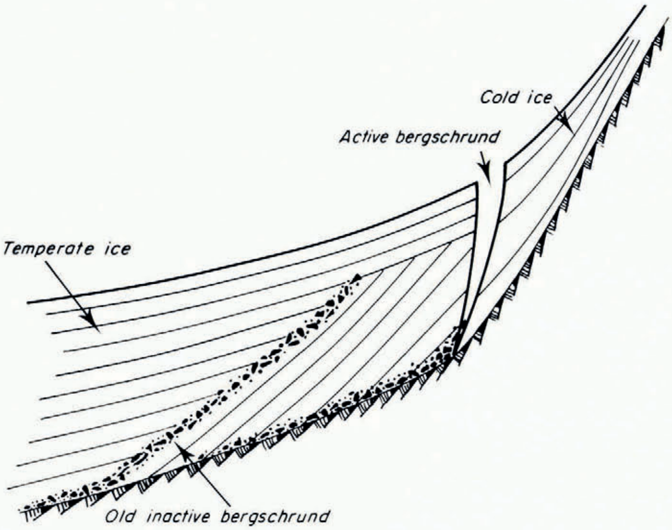

Fig. 6. Origin of the discontinuities observed in the deep foliation: the bergschrund might have shifted. (Sketch by Professor Lliboutry.)

6. Crystalline texture

(a) *The size of the crystals*

Between 30 and 50 m (Fig. 7), there is no evolution in the grain size but rather periodic fluctuations, proof of the stratification of the firn. After 10 years, even after passing through the water table, the summer and autumn horizons are made of much larger crystals than the winter horizons, which did not alternately freeze and unfreeze.

Then, the grain-size increases rapidly between 50 and 90 m (Fig. 8). It could be that this is a particularly active layer, as the very high degree of free water content, found in certain samples at 70 m suggests. Also, in this region numerous strata of blue ice (Fig. 5) can be seen, which could be blue tectonic bands, parallel to the plane of maximum shear stress. However we do not find in the blue bands any different fabric from the one in the surrounding ice; as a rule there is no difference either in crystal size or *c*-axis orientation. Only once near 66.7 m did a blue band show a slightly different fabric from the surrounding ice: smaller crystals (mean diameter 0.7 cm, cf. 0.9 cm in nearly bubbly ice) and a weaker fabric (3 maxima with maximum density 2 to 3%, cf. 4 maxima with 4 to 5%). Beyond 100 m, the size of the crystals does not increase, except very close to bedrock, where the last 7 m of blue ice are made of very large crystals (average cross-section about 10 cm). A very thick bed of blue ice of 5 to 20 m near the bedrock seems to be a common feature of temperate glaciers. We have found this in all drillings on the Glacier de Saint Sorlin and Kamb and Shreve (1963) report the same for the Blue Glacier.

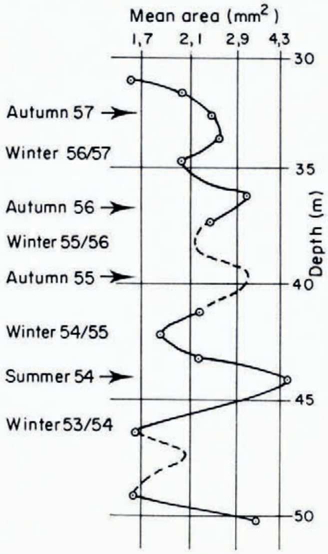

Fig. 7. Periodic variation of the crystal size (coring of 1963).

In polar ice caps, near bedrock, the size of the crystals decreases (Koerner, 1968). It is also small at the base of the tongues of active temperate glaciers, as we saw in our observations of the samples that were taken from beneath the Mer de Glace, against bedrock, in the E.D.F. galleries, and that were immediately immersed in liquid nitrogen.

If kept in the cold bath, crystals are small (mean section 5 mm²) and show a strong fabric (3 maxima with densities of 15 to 17% of points in 1% of area and one maximum of 6%). After 2 d at 0°C, the same sample shows mean section of crystals of 70 mm² and the same orientation of *c*-axes but with maximum densities of 4 to 9% (the same number of *c*-axes being used) (Fabre, unpublished).

The cores extracted from the Vallée Blanche Supérieure were not immediately immersed in liquid nitrogen, but only cooled to −5°C at the most, with the intention of measuring the free water content. During transport to our cold chamber in Chamonix, then from Chamonix to Grenoble, they warmed to 0°C. They must have undergone, therefore, some recrystallization and the size of the grains which is given may be overestimated.

(b) *Form of the crystals*

Initially round, the crystals rapidly acquire a complex form with numerous re-entrants, comparable to those seen in an ablation zone (Rigsby, 1968): the ratio between the area of the average quadrilateral or ellipsoid inscribed in the cross-sections of the crystals and their real mean area changes from 1 to 3 between 60 and 100 m, to more than 7 beyond 140 m. At the same time, the elongation of the inscribed quadrilateral increases with the depth: the ratio between these dimensions is calculated to be 1.1 and 1.4 between 60 and 150 m; 1.45 and 1.55 beyond 160 m. When the elongation is clear (ratio above to 1.3), the direction of the elongation is not linked to the direction of the foliation, but to the average orientation of the crystallographic axes: the crystals are elongated parallel to the trace of the basal planes (Fig. 9).

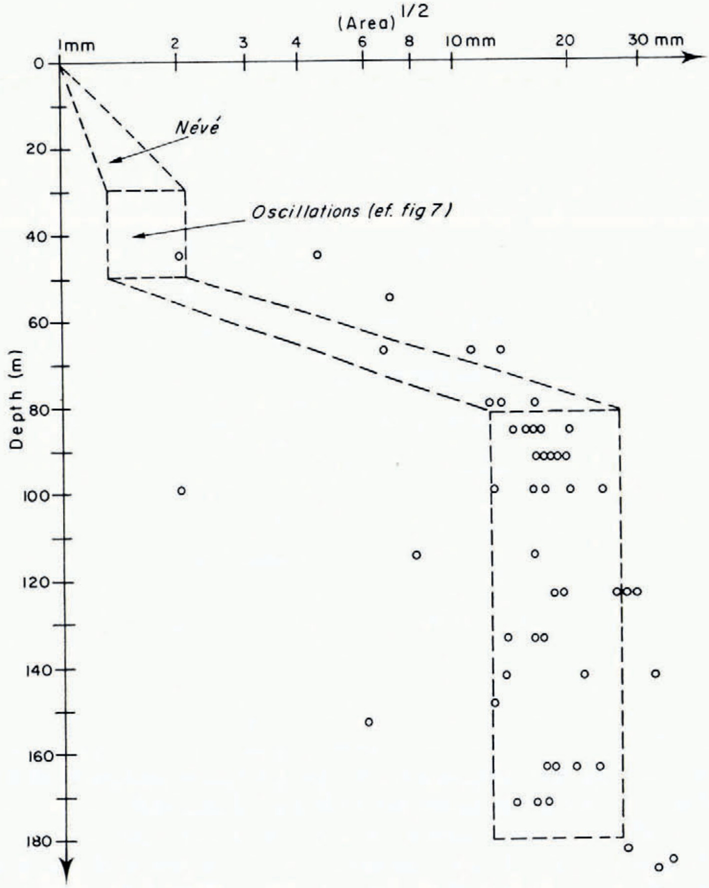

Fig. 8. Average size of crystals on thin horizontal sections (coring of 1971).

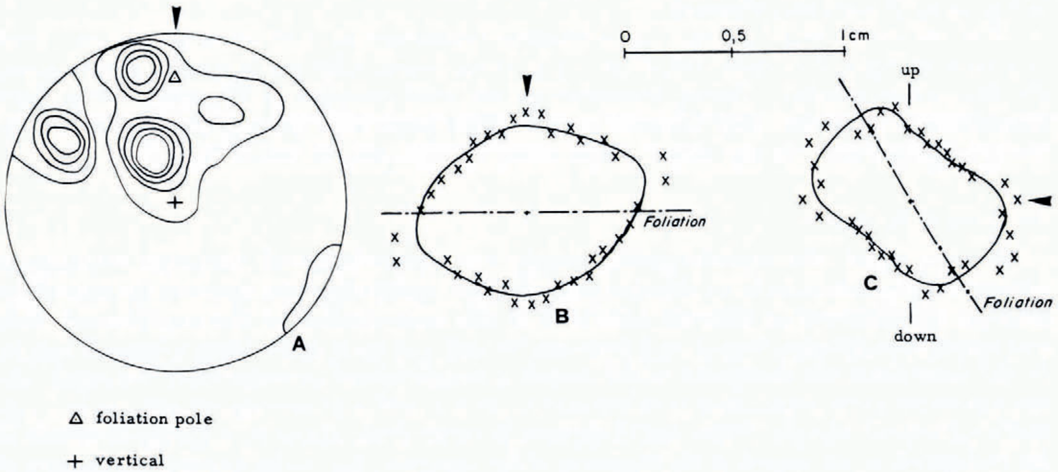

Fig. 9. Structure of the ice at 171 m. (a) Orientation of the optic axes in 3 sections (88 crystals). Contour lines for densities of 3, 6, 9, 12 and 15% of the points in 1% of the area. (b) and (c) Average size and form of the crystals. (b: in horizontal cross-section; c: in vertical cross-section.)

(c) *Preferential orientation of the optic axes*

The very strong subvertical preferential orientation of the crystals in the firn (maximum density of 10 to 15% of the points, in 1% of the area, between 15 and 33 m), is not found in ice. Between a depth of 34 and 60 m, the stereograms show one or several preferential directions, but they are no longer sub-vertical. The maximum densities are from 5 to 8% of the points in 1% of the area when a single preferential direction exists (perpendicular to the stratification), 3 to 7% when several more or less inclined preferential directions exist. From one slide to another, one notices a difference in ice fabric according to the size of the crystals: the small crystals (2 mm in diameter) show a single preferential direction, the larger crystals (4 to 5 mm) two directions, 40 to 50° from one another. The vertical plane passing through the line of the greatest slope of the foliation, is, roughly, the plane of symmetry of the texture.

Beyond a depth of 60 m, the crystals reach 1 cm in diameter, and the stereograms become comparable to those obtained in an ablation zone (Kamb, 1959; Rigsby, 1960; Vallon, unpublished; Jonsson, 1970) with four preferential orientations more or less placed at the corners of a rhombus. Between 66 and 171 m, the most important preferential direction is inclined from 30° to 45° in the same direction as the pole of the plane of foliation, that is, in the vertical plane of the line with the greatest slope, the top tilting downwards. But, though between 60 and 171 m the dip of the foliation increases from 15° to 65°, the preferred directions of the optic axes show no systematic variation with the depth. In the ablation zone of Blue Glacier, Kamb and Shreve (1963) on the other hand observed a shift of *c*-axis orientation with depth.

Regardless of the depth, the bands of small crystals inset into zones with predominantly larger crystals always show the same fabric: a single preferential direction for the small crystals (2 mm), and two directions for the intermediate sized crystals (3 to 8 mm). For the small crystals, the most significant preferential direction is the same as for the very large crystals (Fig. 10). The correlation between the size of the crystals and the number of preferential directions is much clearer than that observed by Kizaki (1969) in Antarctica.

The texture of the blue ice at bedrock is more difficult to study: the crystals are very large, there is no structure with which the elements of the core sample can be oriented to each other, and the number of orientations from which the stereograms can be established is

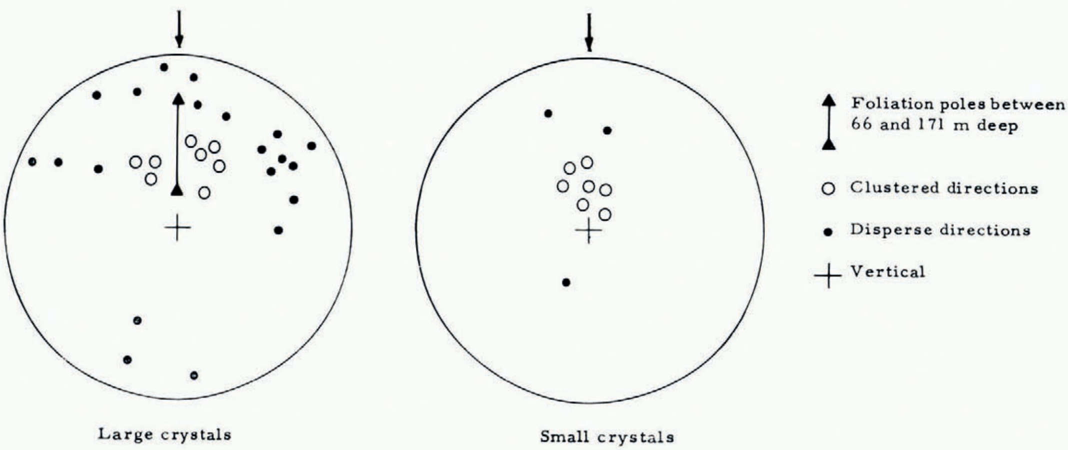

Fig. 10. Preferential direction of the optic axes in the ice (66–171 m) (Schmidt equal-area projection). It was assumed that the line of the greatest slope of foliation lies in the same vertical plane from the surface to the bedrock.

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very small. Nevertheless, with 48 crystals and a count of 7.5% of the total area we obtain a fabric identical to that of the adjacent ice: four preferential directions placed approximately at the corners of a rhombus (maximum density 2 to 5% in 1% of area).

7. Liquid-water content

Lliboutry (1971) defines temperate ice as "an ice containing inside it a liquid phase . . . and in local equilibrium with it". This water could have been imprisoned in the ice at the time of its formation, or it could be produced by the heat released at the time of deformation of the ice.

On the Vallée Blanche, between 34 and 54 m, Joubert (1963) observed a fluctuation in the water content from 0% for the summer and autumn horizons (transformed progressively into ice at a depth of approximately 28 m), to 0.7% for the winter horizons (transformed suddenly at 33 m, at the level of the aquiferous horizon). On the Glacier de Saint-Sorlin, Dupuy (unpublished) found 0% near the surface, 1% at 22 m and 1.2% at 55 m.

One of us (Petit, unpublished) has developed a portable calorimeter allowing the determination, by refreezing, of the water content in samples 10 cm in diameter and 5 cm long. The margin of error of the measurements is about 0.1%.

Even if the extracted cores are packed in melting firn, and stored in the dark, their degree of humidity is bound to change slowly because of a slight permeability, and because the adiabatic expansion that came with their extraction, has stored in them a degree of coldness. Therefore, we measured the water content soon after the cores were extracted. Because of this, it was only possible to perform one measurement for each 2.5 m core extracted.

In Figure 11 the results obtained by Joubert are only given as an indication, because his measuring technique was extremely defective (non-agitated refrigerating bath, measurements too long, imprecise calibrations). In 1971, a breakdown of the apparatus for recording the temperatures prevented us from making these measurements until after the coring had already reached 70 m. The degree of humidity of recent ice is therefore unknown. Not including two exceptionally high values between 70 and 80 m, the degree of humidity changes slowly with the depth: average content of 0.67% between 80 and 100 m, 0.70% between 100 and 120 m, 0.79% between 120 and 140 m, 1.25% between 140 and 160 m, 1.31% between 160 and 180 m. At bedrock, the water content is very low (0.32%).

As the calculations given in the Appendix show, the increase of the water content starting from an average value of 0.6 to 0.7% at the close-off, to a value twice that, can be explained by the heat of deformation. To explain why in the last 7 m, where one would expect a very high water content, and where there was, on the contrary, very little water, it must be assumed that the ice has become permeable to water.

The permeability of the ice at the melting point was studied recently by Lliboutry (1971) and Nye and Frank (1973). The permeability calculated by Nye, assuming that all the water is localized at the limits of three grains, is high as soon as the water content reaches 10^{-3}; for the degree of humidity measured, approximately 1%, it is completely unreasonable. Looking for the mechanism that rendered the ice impermable, Lliboutry (1971) thought at first of the obstacle created by the air bubbles, but concluded that with an air content of 1–2%, this mechanism is insufficient. But the air content used by Lliboutry was that of blue ice, without bubbles. Glacial ice normally contains much more air, about 10% for ice with bubbles. The obstacle created by the bubbles blocks the intercrystalline water channels, on the average, every $2D/3e$ (where D is the diameter of the bubbles, and e the air content of the ice) or 7 mm for bubbles with 1 mm diameter. When the crystals are close to 1 cm in size, the air bubbles do become an obstacle to the filtration of the water, and we can explain the observations reproduced in Figure 10.

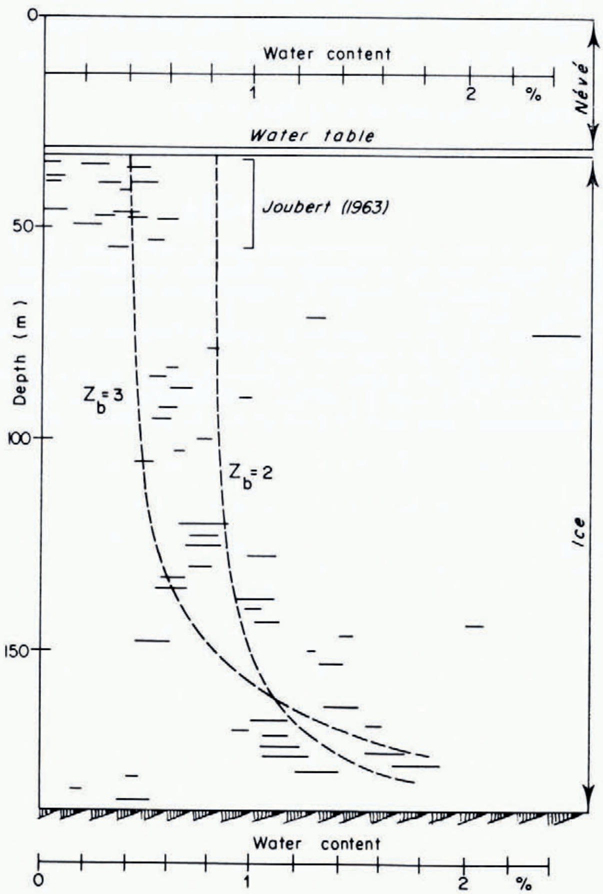

Fig. 11. Water content of the ice. The curves are theoretical increases due to the heat of deformation. $\mathcal{Z}_b = \rho gh \sin \alpha/2|a|B^{1/3} = 45\, B^{-1/3} \sin \alpha$, *B being expressed in* $\text{bar}^{-3}\,\text{year}^{-1}$.

But the bubbles will only stop the filtration if their average size is larger than the diameter of the channels. The radius of the channels R in Nye's model is $R = (fa^2/6\pi\sqrt{2})^{\frac{1}{2}}$ with f as the degree of humidity, and a the average diameter of the crystals (Frank, 1968). For the crystals 1 cm in diameter, the channels have a diameter of 0.40 mm when the water content is 1%, and 0.54 mm when the water content is 2%. One must expect, therefore, in the deep ice of the Vallée Blanche, where the average diameter of the bubbles is about 0.5 mm near a depth of 150 m, to observe a filtration when the water content reaches 1 to 2%.

Near bedrock, in blue ice, where the quantity of air is only around 0.5%, the filtration becomes possible because the average length of the channels between the bubbles is (again for bubbles 0.5 mm in diameter) around 7 cm, larger than the crystals. This filtration will take place if, at the ice–rock interface, there are open paths for the water.

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REFERENCES

Bader, H., *and others*. 1939. Der Schnee und seine metamorphose, von H. Bader, R. Haefeli, E. Bucher, J. Neher, O. Eckel, C. Thams, P. Niggli. *Beiträge zur Geologie der Schweiz. Geotechnische Serie. Hydrologie*, Lief. 3.

Bezinge, A., *and* Bonvin, G. Unpublished. Images du climat sur les Alpes. [Section Glaciologie de la Société Hydrotechnique de France, séance du 22–23 février 1973.]

Dupuy, D. Unpublished. Étude de la glace tempérée de glacier à l'aide de mesures thermiques et diélectriques. [Thesis, Doctorat d'État, Université de Grenoble, 1970.]

Fabre, B. Unpublished. Pétrographie structurale de la glace profonde. Vallée Blanche Supérieure, Massif du Mont-Blanc. [Thesis, 3e cycle, Université Scientifique et Médicale de Grenoble, 1973.]

Frank, F. C. 1968. Two-component flow model for convection in the Earth's upper mantle. *Nature*, Vol. 220, No. 5165, p. 350–52.

Gillet, F. 1975. Steam, hot-water and electrical thermal drills for temperate glaciers. *Journal of Glaciology*, Vol. 14, No. 70, p. 171–79.

Gow, A. J. 1963. Results of measurements in the 309 meter bore hole at Byrd station, Antarctica. *Journal of Glaciology*, Vol. 4, No. 36, p. 771–84.

Gow, A. J. 1970. Deep core studies of the crystal structure and fabrics of Antarctic glacier ice. *U.S. Cold Regions Research and Engineering Laboratory. Research Report* 282.

Jonsson, S. 1970. Strukturstudier av subpolar glaciaris från Kebnekaiseområdet. *Stockholms Universitet. Naturgeografiska Institutionen. Forskningsrapport* 8.

Joubert, J.-L. 1963. Stratigraphie de la glace tempérée à l'aide de la teneur en eau liquide. *Comptes Rendus Hebdomadaires des Séances de l'Académie des Sciences* (Paris), Tom. 257, No. 23, p. 3638–49.

Kamb, W. B. 1959. Ice petrofabric observations from Blue Glacier, Washington, in relation to theory and experiment. *Journal of Geophysical Research*, Vol. 64, No. 11, p. 1891–1909.

Kamb, W. B., *and* Shreve, R. L. 1963. Texture and fabric of ice at depth on a temperate glacier. *Transactions. American Geophysical Union*, Vol. 44, No. 1, p. 103.

Kizaki, K. 1969. Ice-fabric study of the Mawson region, east Antarctica. *Journal of Glaciology*, Vol. 8, No. 53, p. 253–76.

Koerner, R. M. 1968. Fabric analysis of a core from the Meighen ice cap, Northwest Territories, Canada. *Journal of Glaciology*, Vol. 7, No. 51, p. 421–30.

Langway, C. C., *jr*. 1962. Some physical and chemical investigations of a 411 meter deep Greenland ice core and their relationship to accumulation. *Union Géodésique et Géophysique Internationale. Association Internationale d'Hydrologie Scientifique. Commission des Neiges et des Glaces. Colloque d'Obergurgl, 10-9—18-9 1962*, p. 101–18.

Langway, C. C., *jr*. 1967. Stratigraphic analysis of a deep ice core from Greenland. *U.S. Cold Regions Research and Engineering Laboratory. Research Report* 77.

Lliboutry, L. A. 1964–65. *Traité de glaciologie.* Paris, Masson et Cie. 2 vols.

Lliboutry, L. A. 1971. Permeability, brine content and temperature of temperate ice. *Journal of Glaciology*, Vol. 10, No. 58, p. 15–29.

Lliboutry, L. A., *and* Vivet, R. 1961. Épaisseurs de glace et débit solide de la Vallée Blanche Supérieure (Massif du Mont-Blanc). *Comptes Rendus Hebdomadaires des Séances de l'Académie des Sciences* (Paris), Tom. 252, No. 15, p. 2274–76.

Lorius, C., *and* Vallon, M. 1967. Étude structurographique d'un glacier antarctique. *Comptes Rendus Hebdomadaires des Séances de l'Académie des Sciences* (Paris), Sér. D, Tom. 265, No. 4, p. 315–18.

Nye, J. F., *and* Frank, F. C. 1973. Hydrology of the intergranular veins in a temperate glacier. *Union Géodesique et Géophysique Internationale. Association Internationale d'Hydrologie Scientifique. Commission de Neiges et Glaces. Symposium on the Hydrology of Glaciers, Cambridge, 7–13 September 1969*, p. 157–61.

Perutz, M. F., *and* Seligman, G. 1939. A crystallographic investigation of glacier structure and the mechanism of glacier flow. *Proceedings of the Royal Society of London*, Ser. A, Vol. 172, No. 950, p. 335–60.

Petit, J. R. Unpublished. Bilans annuels, transformation du névé en glace, teneur en eau de la glace profonde (Vallée Blanche Supérieure, Massif du Mont-Blanc). [Thesis, 3e cycle, Université Scientifique et Médicale de Grenoble, 1972.]

Rigsby, G. P. 1960. Crystal orientation in glacier and in experimentally deformed ice. *Journal of Glaciology*, Vol. 3, No. 27, p. 589–606.

Rigsby, G. P. 1968. The complexities of the three-dimensional shape of individual crystals in glacier ice. *Journal of Glaciology*, Vol. 7, No. 50, p. 233–51.

Schytt, V. 1958. Glaciology. II. Snow studies at Maudheim. *Norwegian–British–Swedish Antarctic Expedition, 1949–52. Scientific Results*, Vol. 4, A.

Sharp, R. P. [1952.] Meltwater behavior in firn on upper Seward Glacier, St. Elias Mountains, Canada. *Union Géodésique et Géophysique Internationale. Association Internationale d'Hydrologie Scientifique. Assemblée générale de Bruxelles, 1951*, Tom. 1, p. 246–53.

Shumskiy, P. A. 1955. *Osnovy strukturnogo ledovedeniya. Petrografiya presnogo l'da kak metod glyatsiologicheskogo issledovaniya.* Moscow, Izdatel'stvo Akademii Nauk SSSR. [English translation: *Principles of structural glaciology: the petrography of fresh-water ice as a method of glaciological investigation. Translated from the Russian by David Kraus.* New York, Dover Publications, 1964.]

Vallon, M. Unpublished. Pétrographie du névé et de la glace, structure du glacier. [Thesis, Doctorat d'État, Université de Grenoble, 2e partie, 1967.]

Wakahama, G. 1968. The metamorphism of wet snow. *Union de Géodésie et Géophysique Internationale. Association Internationale d'Hydrologie Scientifique. Assemblée générale de Berne, 25 sept.–7 oct. 1967.* [*Commission de Neiges et Glaces.*] *Rapports et discussions*, p. 370–79.

APPENDIX

Amount of water produced in the ice by the heat of deformation since its formation

By L. Lliboutry

Given a glacier with a flat surface, let us take for coordinate axes: Ox along the steepest descending slope, Oy along a level curve, Oz towards the bottom. We call two-dimensional flow where stresses τ_{ij} and strain rates $\dot{\epsilon}_{ij}$ are independent of x Nye flow. Then, $\dot{\epsilon}_{xx} = a$, a constant. Taking as the flow law of ice $\dot{\gamma} = B\tau^3$, where $\dot{\gamma} = 2\sqrt{-\dot{J}_2}$ and $\tau = \sqrt{-I_2'}$, $\dot{J}_2$ and I_2' being the second invariants of the strain-rate and stress deviator tensors respectively (see Lliboutry, 1971, for further details), on the surface the viscosity η and τ are then equal to:

$$\eta_0 = (4a^2B)^{-1/3} \qquad \tau_0 = 2|a|\eta_0 = (2|a|/B)^{1/3}. \tag{A1}$$

Take α as the slope of the surface, and ρg the specific weight of the ice. (The layer of firn will be replaced by a layer of ice of the same weight.) Using the reduced variables:

$$\eta/\eta_0 = Y, \qquad \rho g z \sin\alpha/\tau_0 = Z, \tag{A2}$$

it was shown (Lliboutry, 1964–65, Tom. 2) that Y and Z were related by:

$$Z = \sqrt{\left(\frac{1}{Y} - Y^2\right)}. \tag{A3}$$

The force dissipated per unit of volume is:

$$\begin{aligned} \mathrm{d}W/\mathrm{d}t &= \eta\dot{\gamma}^2 = 4a^2\eta + (\rho g z \sin\alpha)^2/\eta \\ &= 4a^2\eta_0(Y + Z^2/Y) = 4a^2\eta_0 Y^{-2}. \end{aligned} \tag{A4}$$

The vertical velocity is

$$\mathrm{d}z/\mathrm{d}t = a(h-z) \tag{A5}$$

h being the thickness of the glacier.

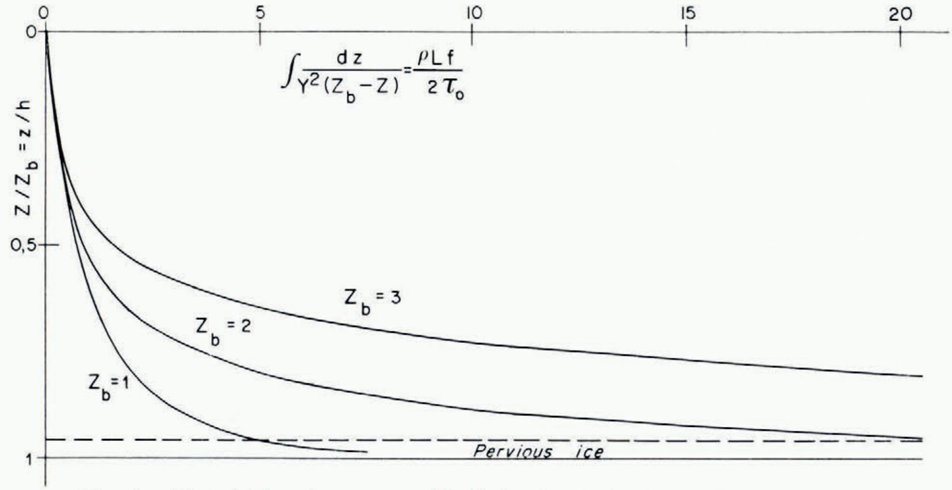

Fig. A1. Heat of deformation accumulated in the ice since its deposit (variables normalized).

The total amount of energy dissipated per unit of volume from the moment of its deposit is:

$$W = \int_0^t 4a^2\eta_0 Y^{-2}\,\mathrm{d}t = 4|a|\eta_0 \int_0^z Y^{-2}(h-z)^{-1}\,\mathrm{d}z. \tag{A6}$$

If the thickness of the glacier is assumed to be uniform, the friction on the bedrock equals $\rho gh \sin\alpha = \tau_b$. Let:

$$\tau_b/\tau_0 = Z_b \tag{A7}$$

then Equation (A6) can be written:

$$W = 2\tau_0 \int_0 \frac{\mathrm{d}Z}{Y^2(Z_b - Z)}. \tag{A8}$$

This integral is easily calculated if Y is given decreasing values starting from $Y = 1$, and if Z is calculated each time with Equation (A3). The results are shown in Figure A1. The quantity of water produced by the deformation is $W/\rho L$, L being the heat of fusion of the ice.

$$L = 0.90\ (\mathrm{g/cm}) \times 80\ (\mathrm{cal/g}) \times 4.18\ (\mathrm{J/cal}) = 300\ \mathrm{J/cm} = 3\ 000\ \mathrm{bar}.$$

In the case considered, the 30 m of firn equal about 20 m of ice, so $h = 177$ m. For this equivalent homogeneous glacier, the speed of sinking of the surface, equal to the annual balance, in metres of ice, is $ah = 3.6$ m/year. From which $a = 0.020$ year^{-1}, probably with a reasonably high precision because the secular variations of the climate are weak at high altitude. $B = 0.17$ year^{-1} bar^{-3} according to Lliboutry, and 0.25 year^{-1} bar^{-3} according to Haefeli. From which: $\tau_0 = 0.55$ to 0.62 bar. Moreover, the slope is 0.09 ± 0.02, so $\tau_b = 0.93$ at 1.54 bar. It follows that $Z_b = 1.5$ to 2.8.

If we assume that the ice is rigorously impermeable, we can calculate the theoretical increase in water content. Curves for $Z_b = 2$ and $Z_b = 3$ have been added to the experimental values in Figure 11. The magnitude is correct for $\tau_0 = 0.6$ bar and $\tau_b = 1.5$ bar ($Z_b = 2.5$), but the increase with the measured depth is more uniform, closer to linearity, than the theory predicts.

REFERENCES

Lliboutry, L. A. 1964–65. *Traité de glaciologie*. Paris, Masson et Cie. 2 vols.
Lliboutry, L. A. 1971. The glacier theory. *Advances in Hydroscience*, Vol. 7, p. 81–167.