

PHYSICAL PROCESSES IN TEMPERATE GLACIERS

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ABSTRACT. In a temperate glacier ice the water content results from the past history of the ice, and the temperature adjusts itself to the appropriate value. Results obtained by the Laboratoire de Glaciologie are summarized. Several facts seem to prove some migration of the liquid phase and a coalescence of liquid inclusions. The enthalpy flux is calculated, but it cannot be proved that a spontaneous gathering together of liquid inclusions could occur. The water-rich, small grain-size layers should come from the lowering of the viscosity when the water content increases, as discovered by Duval. The form of Weertman's sliding law without cavitation can be derived from simple dimensional considerations. Numerically, it cannot account for high sliding velocities, for which subglacial cavitation must be introduced. The cavities are infilled with more stagnant regelation ice than water. Two subglacial hydraulic regimes are distinguished: autonomous and interconnected. The effect which the salt content may have on cavitation is estimated.

RÉSUMÉ. *Physique des glaciers tempérés.* Dans la glace des glaciers tempérés, la teneur en eau dépend de l'histoire passée de la glace, tandis que la température s'ajuste en conséquence. On résume les résultats obtenus par le Laboratoire de Glaciologie dans ce domaine. Divers faits tendent à prouver une légère migration de la phase liquide et une coalescence des inclusions. On exprime le flux d'enthalpie dans la glace, mais on ne peut prouver qu'une concentration spontanée des inclusions liquides soit possible. Les couches à forte teneur en eau et à petits grains doivent provenir plutôt de l'abaissement de la viscosité pour des teneurs en eau croissantes, découvert par Duval. La forme de la loi de glissement sans cavitation de Weertman peut être retrouvée par de simples considérations dimensionnelles. Numériquement, elle ne permet pas d'expliquer les fortes vitesses de glissement, pour lesquelles il faut faire appel aux cavités sous-glaciaires. Ces cavités sont remplies de plus de glace de regel immobile que d'eau. On distingue deux régimes hydrauliques sous-glaciaires, autonome et interconnecté. On estime l'effet que peut avoir la teneur en sels sur la cavitation.

ZUSAMMENFASSUNG. *Physikalische Vorgänge in temperierten Gletschern.* Der Wassergehalt im Eis temperierter Gletscher entstammt früheren Zuständen des Eises; seine Temperatur stellt sich von selbst auf den angemessenen Wert ein. Es wird eine Zusammenfassung von Ergebnissen aus dem Laboratoire de Glaciologie vorgelegt. Einige Tatsachen scheinen eine Wanderung der flüssigen Phase und eine Vereinigung flüssiger Einschlüsse zu erweisen. Der Enthalpie-Fluss wird berechnet, doch lässt sich nicht beweisen, dass spontane Vereinigungen flüssiger Einschlüsse auftreten können. Die wasserreichen, feinkörnigen Schichten sollten von einer Abnahme der Viskosität bei Zunahme des Wassergehaltes herrühren, entsprechend einer Entdeckung von Duval. Die Form von Weertmans Gleitgesetz ohne Kavitation kann aus einfachen Dimensionsbetrachtungen hergeleitet werden. Numerisch ergeben sich dabei jedoch keine hohen Gleitgeschwindigkeiten, für die vielmehr subglaziale Kavitation anzunehmen ist. Die Hohlräume sind mit mehr stagnierendem Regelationeis als Wasser erfüllt. Es werden zwei subglaziale hydraulische Systeme unterschieden: das selbständige und das verbundene. Der Einfluss des Salzgehaltes auf die Kavitation wird abgeschätzt.

WATER CONTENT OF TEMPERATE ICE

While in cold ice temperatures are an important source of information, in temperate ice it is the amount of liquid water w (which can be expressed in kilogrammes of liquid phase per kilogramme of ice). It will be seen that, depending on dissolved substances, temperatures can fluctuate over some hundredths of a degree with respect to the melting point under the pertinent pressure. The corresponding sensible heat is 10 to 100 times lower than the latent heat of the liquid phase. Thus in a temperate glacier w comes from the past history of ice, and the temperature adjusts itself to the appropriate value.

At the Laboratoire de Glaciologie of the CNRS we measure the water content by retrieving ice cores and freezing them at once in a very accurate calorimeter. The adiabatic decompression of the core cools it. α being the linear thermal dilatation of the sample ($\alpha = 5.25 \times 10^{-5} \text{ deg}^{-1}$), T its absolute temperature, C_p its thermal capacity per unit volume ($C_p = 1.90 \times 10^6 \text{ J m}^{-3} \text{ deg}^{-1} = 19.0 \text{ bar deg}^{-1}$):

$$dT/dp = 3\alpha T/C_p = 0.00226 \text{ deg bar}^{-1}, \quad (1)$$

Nevertheless during this fast cooling, the moisture w does not change. On the contrary if the coring rate is too low, or the measurement too belated, heat can enter the sample because its surface is washed by practically pure water. Then the moisture w increases continuously.

On the Vallée Blanche, at 3 550 m, the amount of liquid water, between 34 and 140 m deep, fluctuates between 0.2 and 1%. It rises to 1.3 to 1.8%, between 140 m and 180 m, the crystal average size being always 2 cm. In the last seven meters (180–187 m in depth), crystals are exceptionally large (10 cm) and the liquid water content exceptionally low (0.1–0.2%) (Vallon and others, 1976).

On the Glacier de Saint-Sorlin, at 2 750 m, Dupuy's (unpublished) measurements, when corrected from a systematic error resulting from the stocking time, lead to $w = 0.0\%$ at 5.5 m deep, 0.4% at 22 m and 0.7% at 55 m. (At 5.5 m the cold winter wave should enter, even if in August, when the measurement was made, the ice surface temperature had risen to melting point again.)

The amount of liquid is high in layers with small-sized crystals. Unfortunately measurements were not made on a scale minute enough to establish a quantitative correlation.

ORIGIN OF LIQUID WATER

The liquid water existing in temperate glacier ice has two origins.

(1) *Water produced by frictional heating.* Let τ be the effective shear stress ($2\tau^2 = \sum_{i,j} (\tau_{ij}')^2$, denoting by τ_{ij}' the deviatoric stresses) and $\dot{\gamma}$ the effective shear strain-rate ($2(\dot{\gamma}/2)^2 = \sum_{i,j} (\dot{\epsilon}_{ij})^2$ denoting by $\dot{\epsilon}_{ij}$ the strain-rates). The frictional heat per unit volume and unit time is $\tau\dot{\gamma}$, and then the rate of water production:

$$dw/dt = \tau\dot{\gamma}/L\rho \quad (2)$$

where $L\rho$ is the latent heat per unit volume ($L\rho = 3.00 \times 10^8 \text{ J m}^{-3} = 3 \text{ 000 bar}$). In the bottom layers of temperate mountain glaciers τ is of the order of 1 bar and $\dot{\gamma}$ of 0.3 year⁻¹. The rate of water production is then of the order of 1% per century.

(2) *Primeval water*, trapped during the close-off. In the accumulation area of alpine glaciers the transformation of firn into ice happens swiftly at a fixed depth (33–34 m in the upper Vallée Blanche) during the time when the percolating melt water has formed an aquifer layer at the bottom of the firn (15 July–31 August) (Vallon and others, 1976). This process has been checked by laboratory experiments (La Casinière and Heyraud, unpublished).

It is then a puzzling fact that only 0.2% of water is found in young ice between 34 and 50 m deep. This may be explained, as well as the higher values found between 80 and 150 m, which the frictional heating cannot explain, by assuming that the permeability of ice comes from its deformation and recrystallization. The newly formed ice should be very impervious, since stresses, strains, and recrystallization are low at this level. This impervious zone should allow the ice below to drain gradually.

LOCALIZATION OF THE LIQUID PHASE IN TEMPERATE ICE

In Nye and Frank's (1973) model, all the water is found in capillary pipes, $2r$ in diameter, at the junction of three crystals. With this model, $r \approx 0.2a\sqrt{w}$, a denoting the average crystal size. Common values are $a = 2 \text{ cm}$, $w = 0.8\%$, whence $r = 0.36 \text{ mm}$, which is obviously wrong. The negligible permeability of ice would lead to $r < 5 \text{ }\mu\text{m}$ (Llibouty, 1971).

Actually: (1) these pipes are cut by the strains, and (2) all water is not at the junction of three crystals.

(a) Nye and Mae (1972) have observed the production by stress of micro-lenses of water (diameter 20–40 μm), without vapour bubbles (the pressure in water should be negative,

because vaporization nuclei are lacking). The plane of these lenses is normal to the maximum pressure, as predicted by theory (Lliboutry, 1971).

(b) Around a salt inclusion, a small disk of brine should exist, within a basal crystallographic plane. It may be similar to Tyndall flowers (2–4 mm in size), which appear by irradiating ice with infrared rays. (This last phenomenon has nothing to do with salt impurities however. According to Klinger and others (unpublished) Tyndall flowers are linked to the production of ionic defect pairs when ice selectively absorbs radiations of 12.4 μm).

(c) Water films should cover the walls of air bubbles.

(d) In old active ice, large air and water bubbles of an irregular form exist, as well as small water pockets of several cubic centimetres. We observed them some years ago under the tongue of the Mer de Glace, which could be reached through the tunnels dug in the rock to tap the subglacial stream (today this catchment is no longer subglacial).

Our opinion is that the continuous strain and recrystallization of ice allows liquid inclusions to coalesce and to migrate downwards. It also allows the air bubbles to coalesce and migrate upwards. The layers of bubbly ice (which form foliation) are divided into vertical ribbons with cross-sections about 10 cm × 30 cm with large bubbles.

TEMPERATURE OF TEMPERATE ICE

When local equilibrium is reached, the temperature of ice is (Lliboutry, 1971):

$$T = T_0 - \beta' p_m - A'n/w + A''(1/r_1 + 1/r_2) \tag{3}$$

where T_0 is the melting point under normal pressure: $T_0 = 273.15$ K, p_m is the maximum compressive stress, since the liquid inclusions grow in planes perpendicular to this stress, and the pressure involved is the stress normal to the liquid–solid interface. (A quick variation of the liquid inclusion size generates very local stresses, owing to the change in volume, but in temperate ice such stresses due to a fixed strain relax in less than one hour). As variation of melting point per unit pressure, we should take in a temperate glacier $\beta' = 0.0098$ deg bar⁻¹ instead of the well-known value $\beta = 0.00742$ deg bar⁻¹, because the liquid phase should be air saturated. During its life, an inclusion migrates and probably comes in contact with air bubbles many times. At least a value between β and β' is needed. The value β' is the correct one for temperatures in bore holes full of air-saturated water (Harrison 1972). n is the number of salt equivalents per unit mass of ice and $c = n/w$ is the salt concentration of the liquid phase. $A' = 1.86$ deg mol⁻¹ kg.

According to Boutron and others (1972), $n = (3 \text{ to } 8) \times 10^{-6}$ for east Antarctica samples (the ice there is cold, but a bottom layer of temperate ice should exist in many areas, specially under the outlets). For the glaciers of temperate regions of the northern hemisphere $n = (10 \text{ to } 30) \times 10^{-6}$. For instance according to Duval (unpublished) in the ice of the upper Vallée Blanche (Mont Blanc area), $n = 17 \times 10^{-6}$.

Then $A'n = -\theta_m \approx 10^{-5}$ deg for polar ices and $\approx 3 \times 10^{-5}$ deg for “alpine” ices (and not $(3 \text{ to } 55) \times 10^{-5}$ as assumed by Harrison (1972)).

The last term in Equation (3) proceeds from capillarity. r_1 and r_2 are the principal radii of curvature of the interface and $A'' = 0.0212$ deg μm. In general this term is negligible compared with the one proceeding from brine inclusions, and we may write the Celsius temperature:

$$T - T_0 \equiv \theta = -\beta' p_m - \theta_m/w. \tag{4}$$

The temperature in a temperate glacier has been estimated by Harrison (1972) from the refreezing of water in a bore hole: $\theta = -0.03^\circ\text{C}$. In our view, one half only of this lowering comes from the salt content. The other half comes from a maximum compressive stress $p_m \approx 1.5$ bar.

ENTHALPY FLUX IN TEMPERATE ICE

The enthalpy per unit volume is:

$$H = L\rho w + C\rho(T - T_0) = L\rho(w - p_m/p_t - w_t^2/w^2) \quad (5)$$

with $p_t = L/C\beta' = 16\,330$ bar and $w_t^2 = -\theta_m C/L$. w_t is the moisture corresponding to the transition temperature θ_t introduced by Harrison.

For polar ices	$w_t \approx 0.025\%$,	$\theta_t \approx -0.04^\circ\text{C}$.
For alpine ices	$w_t \approx 0.043\%$,	$\theta_t \approx -0.07^\circ\text{C}$.
In temperate ice	$w \gg w_t$	and then $H \approx L\rho w$.

Nevertheless other terms must be introduced to calculate the enthalpy flux ϕ .

$$\phi = H\mathbf{u} + L\rho w\mathbf{v} - K\nabla T = L\rho \left[w(\mathbf{u} + \mathbf{v}) - h \frac{w_t^2}{w^2} \nabla w + h\nabla \left\{ \frac{p_m}{p_t} \right\} \right] \quad (6)$$

\mathbf{u} being the velocity of ice, \mathbf{v} the migration velocity of brine inclusions with respect to ice, and $h = K/C\rho$ the thermal diffusivity of ice ($h \approx 38$ m² year⁻¹).

If the migration velocity were controlled only by the diffusion rate of salts from the cold side of the inclusion (where water freezes) to the warm one (where ice melts), its value would be (Hockstra and others, 1965):

$$\mathbf{v} = -(D/c)\nabla c \quad (7)$$

with $D \approx 0.047$ m² year⁻¹. Since the salt concentration of liquid inclusions is $c = n/w$, it becomes:

$$\mathbf{v} = (D/w)\nabla w. \quad (8)$$

The enthalpy flux, in a reference frame motionless with respect to ice would be, if the variations of p_m were negligible:

$$\begin{aligned} \phi &= L\rho D\nabla w - K\nabla(-\theta_m/w) \\ &= (L\rho D + K\theta_m/w^2)\nabla w \\ &= L\rho(D - hw_t^2/w^2)\nabla w. \end{aligned} \quad (9)$$

If $w > w_t(\sqrt{h/D})^{\frac{1}{2}} \approx 0.7-1.2\%$, the enthalpy flux produced by the migration of the brine inclusions would be larger than that produced by the thermal conductivity, the enthalpy would diffuse in the same direction as ∇w , and an instability would occur (spontaneous concentration of the liquid inclusions).

Actually Equation (7) is wrong (Tiller, 1969). The migration velocity of *intracrystalline* brine inclusions, of size δ , is controlled by:

- (1) either the two-dimensional nucleation on the wall where freezing proceeds, which leads to:

$$\mathbf{v} = \frac{D}{T_m - T} \left(|\nabla T| - \frac{\Gamma_1}{\delta} \right) \quad (10)$$

where T_m is the melting point for pure ice under the same pressure and Γ_1 a constant of the order 0.001 deg. Since in temperate glaciers $|\nabla T| < \Gamma_1/\delta$, there would be no migration at all.

- (2) or by the adsorption of impurities which impedes crystal growth on the rear side.

What actually happens for *intercrystalline* inclusions, in active ice where strain and recrystallization favours the mobility of water, is unknown. Maybe the suggested instability occurs, but the law for \mathbf{v} is unknown. We are not able therefore to use the following heat equation (valid for slow processes, when the local equilibrium between both phases is reached at any instant):

$$\partial H/\partial t + \nabla \cdot \phi = \dot{\gamma}\tau. \quad (11)$$

STRAIN-RATE ENHANCEMENT BY THE WATER CONTENT

Duval (in press) has shown that for a given stress, the strain-rate of ice rises consistently with the water content. For an increase of w from 0.1% to 1% the strain-rate gets higher by a full order of magnitude. This effect must have the same influence on stability as the variation of viscosity with temperature in cold glaciers.

For a given effective strain-rate $\dot{\gamma}$, the viscosity $\eta = \tau/\dot{\gamma}$ is lower where the water content is high. There, the energy dissipation $\tau\dot{\gamma} = \eta\dot{\gamma}^2$, and therefore the rate of water production, are lower. Since its strain-rate is imposed by the surrounding ice, a small volume with a higher content of primeval water than the average falls into this category. With time this local perturbation of the water content smoothes out.

For a given effective stress τ , the reverse is true. The energy dissipation $\tau\dot{\gamma} = \tau^2/\eta$ and the rate of water production is higher where the water content is higher. This condition prevails when a very large layer with a higher water content is present. In this case the stresses at the boundaries of the layer are imposed, since they are due to the weight of the overlying ice mass. The water production in such a layer must increase at an accelerating rate.

A high strain within a layer produces extensive recrystallization and minute grains, a mylonitization as petrographers say. This should explain the correlation between high water content and small-sized crystals noted by Vallon and others (1976).

A place where such an instability may be expected is at the bottom of glaciers. Nevertheless this is not always the case. Our explanation is that the water can exude from the ice at the ice-bedrock interface.

GLACIER SLIDING WITHOUT CAVITATION

In order that a cold or temperate glacier can slide on its bed, the ice-rock interface must reach melting point, allowing the minute rugosities of the bedrock to be overcome by a melting-refreezing process (Nye, 1970). The thickness of the water film which flows from the melting side to the refreezing side must be about 1 μm (Nye, 1973). Only colloidal particles (the well-known glacial flour) can transit in it.

The mathematical formulation of this process involves geometrical parameters, the stress field, and, assuming the thermal conductivities of ice and rock to be the same, a single physical parameter $\Gamma = L\rho/4CK$ (adopting Kamb's notation). $L\rho$ is the latent heat of ice per unit volume, C the coefficient by which the melting point is lowered by pressure and K the thermal conductivity of ice and bedrock, assumed to be the same. It may be expressed in units of bar year m^{-2} . On the other hand, assuming Glen's ice flow law:

$$\dot{\epsilon}_{ij} = (B/2)\tau^{n-1}\tau'_{ij} \quad (12)$$

we have another single dimensional parameter, B , which may be expressed in units of bar^{- n} year⁻¹.

The relation between the sliding velocity U (in m year⁻¹) and the friction over the bedrock τ_b (in bars), for a "clean" glacier (no bottom moraine, no debris scratching the bedrock), has been sought by Kamb (1970). He uses as model a bedrock with a white roughness, also introduced by Nye (1970). A white roughness bedrock is a bedrock which has the same aspect at all sizes, when these sizes are very small compared with some smoothing length Λ used to separate what is roughness from what is the general shape of the bedrock. Since the influence of Λ on the friction law is negligible, we can say that the model is a *non-dimensional bedrock*. Weertman's model of protuberances of the same shape and different sizes decreasing in geometrical progression, and our model of superimposed sine profiles of the same shape and sizes decreasing in geometrical progression are approximations to this non-dimensional bedrock.

When the glacier is everywhere in contact with the bedrock, the only dimensional parameters entering into the model are U , τ_b , Γ and B . The expression $U^{-1}(B/\Gamma)^{1/2}\tau_b^{(n+1)/2}$ is dimensionless, and then the friction law must be Weertman's one:

$$U = C(B/\Gamma)^{1/2}\tau_b^{(n+1)/2} \quad (13)$$

where C is a dimensionless parameter depending on the dimensionless parameters which define the model of the bedrock. The problem is to find realistic values for C .

A very full study of the problem (Lliboutry, 1975) gives numerically, n being equal to 3 (Duval, unpublished):

$$U = 0.016T_b^2m_*^{-3} \quad (14)$$

where m_* is the mean quadratic slope, in the direction of flow of a conveniently filtered bedrock. Direct and indirect field measurements give $0.11 \leq m_* \leq 0.27$ and thus only sliding velocities lower than about 10 m year⁻¹ can be explained in this way.

GLACIER SLIDING WITH CAVITATION

To obtain higher sliding velocities, we must introduce cavities which may exist in the lee of some obstacles of the bedrock. (In his last article on this subject, Weertman (1972) speaks of a "very thick water film" at the interface, which does not exist on the up-stream side of the obstacles. Do not quibble about names.) The sites where cavitation appears first must be in the lee of the obstacles where the local stresses are higher, that is those which have the controlling obstacle size as introduced in former theories. According to our estimate the controlling wavelength for most glaciers should be some decimetres, and the corresponding size of the bumps 2–10 cm.

These cavities are infilled, not only with water, but also with regelation ice which is stuck to the bedrock. (This corresponds to the input of cold through the bedrock; the input of cold through the glacier ice forms regelation ice stuck to the moving roof of the cavity.) A simple calculation shows that in the lee of a controlling-size obstacle the rate of production of motionless regelation ice is one order of magnitude larger than the water produced by geothermal heat (the latter being about 0.5 cm per year).

Two limiting cases are then possible, as well as all the intermediate situations.

(1) The cavities are isolated as regards water circulation, a situation which I call an *autonomous hydraulic regime*. A boundary condition is then that some imposed volume of water and motionless ice, which increases with time, exists between the moving glacier and the bedrock. The pressures within the cavities adjust themselves in order to minimize the friction for a given sliding velocity.

(2) The cavities succeed in communicating with waterways at a given pressure, a situation which I call a *connected hydraulic regime*. In this case, since the outflow is very small, we may assume that there is no Darcy loss of head in the conduits (contrarily to what was assumed in Lliboutry, 1968). We can then define a piezometric surface for all the interconnected cavities and waterways.

Now there must be a cyclic evolution between both cases. When a cavity is interconnected (i.e. has an outlet), its volume cannot vary very much. With time it gets infilled with motionless regelation ice. Then this ice fuses with the moving glacier ice and a new cavity appears close to the bedrock, in the lee of the obstacle. This new cavity is at first isolated (autonomous regime). It increases with time, until an outlet (or another cavity nearby) is reached. Little by little all the subglacial cavities become interconnected, and the cycle starts again.

In the interconnected situation, a new parameter enters into the problem: $N = p_i - p$, where p_i is the mean pressure of ice at the bottom and p the pressure of water and motionless ice within the cavities. (N is Terzaghi's effective pressure.) Thus we have two dimensionless

numbers: the one needed to establish Weertman's law and τ_b/N . The law of friction must be some relation linking these two numbers to the geometrical, dimensionless parameters which define a dimensionless bedrock.

This time Weertman's model of the bedrock and ours do not lead to the same friction law. With the superimposed sine-curve model, at high sliding velocities, the friction due to melting-refreezing processes becomes negligible, and then the physical parameter Γ has no influence. The friction law must then be

$$\tau_b/N = f \quad (15)$$

where f is a dimensionless parameter depending on the bedrock shape.

The autonomous regime is less simple. There is a new parameter, the volume trapped beneath the glacier per unit area (that is a length), but it is in turn a function of time. The cycles of the different cavities may be incoherent or synchronized. More specifications are needed about the model. These questions will be dealt with in another article.

INFLUENCE OF THE SALT CONTENT ON SLIDING THEORY

Is it necessary to re-examine the whole sliding theory to take into account the temperature-lowering introduced by the salt content? In a large cavity under Glacier d'Argentière, on the lee side of a *Riegel*, (which is at atmospheric pressure and may be visited). Souchez and others (1973) have measured the ionic content at $n = 24$ to 35×10^{-6} equivalents/kg for the glacier ice. This rather high value comes from the existence of some morainic material within the ice (Ca^{++} is abundant). There were also spicules and crushed regelation ice, with a very high ionic content ($n = 90$ to 242×10^{-6}) which formed, at least in part, from brine inclusions which have extruded from the ice.

In the lee of an obstacle, the decompression favours the enlargement of those water lenses and films which are perpendicular to the ice surface at the expense of the others. This brine can then reach the surface and drain off. The fact that this brine freezes rather than the purer water in the cavity shows that the raising of the melting temperature caused by the decrease of p_m prevails over its lowering by dissolved salts. It follows that the essentials of the sliding theory remain valid.

Nevertheless when a cavity is isolated (autonomous regime), its water becomes more and more salty. Its temperature will be lowered, making the melting-refreezing process of sliding (and therefore the whole sliding) more and more difficult. On the other hand, when superficial and purer water enters within the cavity, its temperature is raised and the melting process enhanced.

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DISCUSSION

C. F. RAYMOND: Could you clarify your interpretation of the effect of adiabatic decompression of the core on its temperature in view of the existence of liquid–solid interfaces in the sample and change of pressure in the liquid?

L. LLIBOUTRY: The adiabatic decompression cools both the ice and the water (but not by the same amount). This temperature difference must be added to the temperature difference between the final melting point temperature (at atmospheric pressure) and the previous one (under pressure). Both are eliminated by the refreezing of a very small amount of liquid inclusions. For instance, for a decompression of 15 bars, the adiabatic cooling is about -0.04 deg, the change in melting point about -0.12 deg, and a change of the liquid water content of -0.1% is needed to restore the equilibrium.

M. M. MILLER: You note that in the accumulation area the transformation of firn into ice occurs rather suddenly at a fixed depth (on the Vallée Blanche at about 34 m), and that this relates in the spring to a firn-base aquifer horizon. Furthermore, you indicate that this has been checked by laboratory experiments. First, did the lab tests sufficiently model the age relationship in natural firn? Secondly, is your consideration compatible with the more gradual density transition in the firn-ice zone (say of a density 0.74 to 0.88 Mg m^{-3}) that has been reported from some other temperate glaciers at and above the water table? Is this indeed a sudden density change between true firn and the underlying glacier ice or is it to be properly described as between the firn-ice zone and the bubbly glacier ice?

LLIBOUTRY: First, the laboratory experiments (A. de La Casinière and M. Heyraud, paper submitted to *Journal of Glaciology*) were conducted on artificial firn made with crushed and sieved ice, with different grain sizes. Secondly, the annual accumulation includes two zones with regelation ice layers, and thus a high average density, which corresponds to spring and autumn, and two zones without ice layers, of lower mean density. The former zones turn progressively into ice. Thus thick and impervious layers of ice are formed, which allow the formation of a water table. Within this water table the lighter firn of the latter zones turns very swiftly into ice during summer.