An isotopic model for basal freeze-on associated with subglacial upward flow of pore water

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[1] Subglacial freezing in polar glaciers can have a significant dynamical effect. Recent studies have shown that freezing of pore water flowing upward through subglacial fine-grained sediments at the freezing interface and progression of this freezing front downward are responsible for fast ice flow stoppage in ice streams. The upward pore water flow leads to the formation of debris-bearing basal ice layers. A model for stable isotope composition, both in δD and $\delta^{18}O$, is developed for predicting the isotopic composition of the ice segregated by such a mechanism. The development of this isotopic model for water films present along the grains of the subglacial sediment predicts the absence of apparent fractionation for the ice formed. This prediction is tested against two East Antarctic outlet glaciers by studying the δD-$\delta^{18}O$ relationships in the basal ice layers of these glaciers. *INDEX TERMS:* 1040 Geochemistry: Isotopic composition/chemistry; 1827 Hydrology: Glaciology (1863); 9310 Information Related to Geographic Region: Antarctica; *KEYWORDS:* Isotopic composition, stable isotopes, basal ice, ice segregation. **Citation:** Souchez, R., D. Samyn, R. Lorrain, F. Pattyn, and S. Fitzsimons (2004), An isotopic model for basal freeze-on associated with subglacial upward flow of pore water, *Geophys. Res. Lett.*, *31*, L02401, doi:10.1029/2003GL018861.

1. Introduction

[2] There are numerous cold-based glaciers at temperatures several degrees C below the pressure-melting point in polar environments. Some of these glaciers are resting on subglacial sediments and exhibit stratified basal ice layers more or less loaded with debris. Such debris-bearing ice testifies for the action of basal freeze-on.

[3] Soft-bedded ice streams are the main control on ice discharge of some important sectors of ice sheets. Lubrication by build-up of pore-water pressure in subglacial sediments results in low effective stress and allows fast flow in these ice streams [*Tulaczyk et al.*, 2000]. Basal freeze-on can induce water-pressure changes that result in fast flow stoppage because of till consolidation. Basal freezing may result from climatic cooling, ice thinning or change in basal shear heating. *Christoffersen and Tulaczyk* [2003a] indicate that in such circumstances subglacial upward pore water flow occurs as a consequence of ice-water interface curva-

ture and surface tension effects. These effects and the progression of the freezing front downward lead to the construction of a stratified basal debris-bearing ice.

[4] In this paper, we investigate from an isotopic point of view the upward pore-water flow model and the consequent ice segregation giving rise to a basal ice accretion layer. An isotopic model for stable isotopes composition, both in δD and $\delta^{18}O$, is developed predicting the hydrogen and oxygen isotopic composition of the ice segregated by such a mechanism. Examples of isotopic composition of basal ice layers from some Antarctic glaciers matching the predictions of the model are then described. Such ice layers are supposed to be built by the subglacial process invoked by *Christoffersen and Tulaczyk* [2003a].

2. Freeze-on Thermodynamics

[5] The temperature of the freezing point is depressed by an increase in fluid pressure and by an increase in solute concentration. In some circumstances, ice-water interfacial effects arising from interface curvature have also to be taken into account. This is the case in pore spaces of fine-grained subglacial sediments. The pressure difference across the ice-water interface when supercooled liquid pore water is present gives rise to a lowering of the pore water pressure at the freezing interface. This drop of pressure induces an hydraulic gradient driving water flow toward the freezing interface where a layer of segregation ice is accreted. Building up on the analogy between frost heave phenomena and debris-bearing basal ice formation, *Christoffersen and Tulaczyk* [2003a] develop a high-resolution numerical model that is able to understand the freeze-on mechanism and to trigger ice stream stoppage.

[6] In their study, these authors consider either a closed water system where water driven to the freezing interface is only pore water of the subglacial sediment or an open system where water is supplied from beneath the subglacial sediment. In both situations the upward flow of pore water leads to ice formation. Since the freezing front is moving into the subglacial till, successive layers of segregation ice are produced and the observed stratification of debris-bearing basal ice is so developed. Figure 1 visualises the process.

3. The Isotopic Model

[7] The upward flow of pore water toward the freezing interface through the subglacial fine-grained sediment is considered to occur as water migration along thin liquid films surrounding the sediment particles. Let us consider a specific path and define the liquid film along this path. The same reasoning will apply to the other paths.

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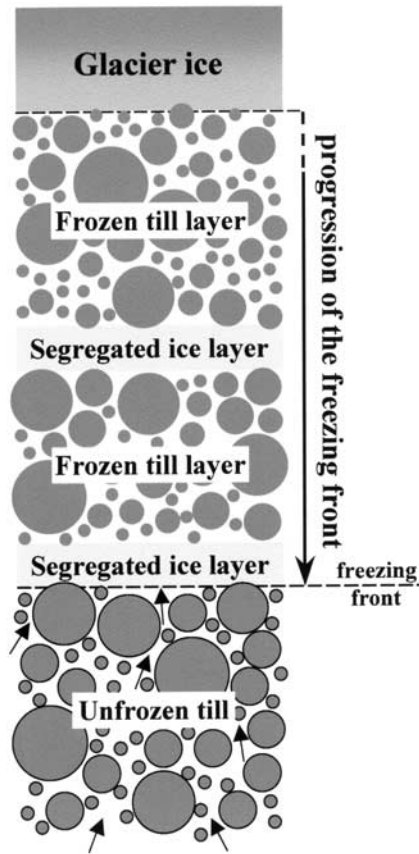[3]Department of Geography, University of Otago, Dunedin, New Zealand.

Figure 1. Schematic diagram showing pore water flowing within subglacial sediments towards the freezing front at the base of a glacier and producing accretion ice [modified from *Christoffersen and Tulaczyk*, 2003b]. Either frozen till or segregated ice layers are produced, depending on the physical conditions.

[8] Let N be the number of water molecules presenting a specific isotopic composition in $H_2^{18}O$ or HDO (D = deuterium), then

$$N = CV, \qquad (1)$$

where C is the concentration of such an isotopic water molecule in the liquid film and V, the volume of this film. It follows that

$$dN = d(CV) = CdV + VdC. \qquad (2)$$

[9] The volume of the water film is supposed to be constant in time, the amount of liquid water driven into the film at the base of the subglacial sediment being equal to the amount of film water frozen at the freezing front (open system). Thus, dV = 0.

[10] Therefore $dN = V\,dC = Q_{in}\,C_{in}\,dt - Q_{out}\,C_{out}\,dt$ where Q_{in} and Q_{out} are the water flux entering the film (with isotopic concentration C_{in}) and the water flux leaving the system by refreezing and ice export (with isotopic concentration C_{out}), respectively.

Since it is assumed that dV = 0, $Q_{in} = Q_{out} = Q$ and one obtains $Q(C_{in}\text{-}C_{out})dt = VdC$. Let T be the water film renewal time. It follows that V = Q T and

$$Q(C_{in} - C_{out})dt = QTdC \quad or \quad \frac{dt}{T} = \frac{dC}{(C_{in} - C_{out})} \qquad (3)$$

Let us consider now that $C_{out} = \alpha\ C$, where α is the equilibrium fractionation coefficient either for oxygen 18 or for deuterium. Whereas fractionation at the water-ice interface is given by the equilibrium fractionation coefficient, the amount of observed fractionation is dependent on the ratio between the appropriate diffusion coefficient in the water adjacent to the freezing front and the freezing rate [*Souchez et al.*, 1987]. This indeed controls the magnitude of the impoverishment in heavy isotopes of oxygen or hydrogen in the water adjacent to the progressing freezing front and so the amount of heavy isotopes introduced into the ice. Since $C_{out} = \alpha\ C$, integration of equation (3) gives

$$C = \frac{C_{in}}{\alpha} + \left(C_0 - \frac{C_{in}}{\alpha}\right)\exp\left(-\frac{\alpha t}{T}\right) \qquad (4)$$

where C_0 is C at t = 0.

[11] For long time scales (large t), it follows that $C = \frac{C_{in}}{\alpha}$ and $C_{out} \approx C_{in}$. Since α is close to unity, the exponential term of the equation becomes sufficiently small to consider $C_{in} \approx C_{out}$ when $t \geq 3\ T$. The composition of the ice formed by freezing is the same as the one of the water entering the pores of the fine-grained subglacial sediment (Figure 2).

[12] Considering that the water film has been renewed at least three times in the process of ice segregation is most probably correct in view of the micrometric thickness of the film and of the commonly observed thickness of segregation ice layers in debris-bearing basal ice and of the subglacial till.

4. Discussion

[13] The water present in the pores of the subglacial sediment has most probably a composition in stable iso-
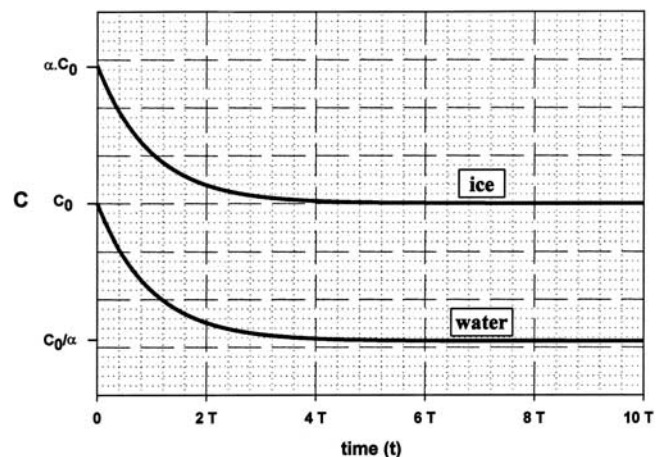
Figure 2. Evolution of the isotopic compositions (C) of the water film and of the ice segregated from it. T is the renewal time of the water films. C_0 is the concentration at time t = 0, and α is the equilibrium fractionation coefficient either for ^{18}O or for D.

topes very close to that of glacier ice. This is supported by a fractional melting experiment that has been performed on polar meteoric ice sample. In a δD-$\delta^{18}O$ diagram, points representing the successive melt samples taken during the experiment appear tightly clustered [*Souchez et al.*, 1988]. The extreme range is only 0.2‰ in $\delta^{18}O$. This implies, as suggested by *Moser and Stichler* [1980], the absence of fractionation during ice melting. It follows that the isotopic composition of segregation ice originating by the above-described process is also that of the overlying meteoric glacier ice. Since different meteoric ice samples are aligned in a δD-$\delta^{18}O$ diagram on a Meteoric Water Line with a slope close to 8, debris-bearing basal ice produced by upward pore water flow in subglacial sediment would also plot on the same line. This prediction can be tested and will be considered below.

[14] The study of fractionation by freezing has led to the freezing slope concept [*Souchez and Jouzel*, 1984] in closed as well as in open systems. Samples of ice due to freezing of water are aligned in a δD-$\delta^{18}O$ diagram on a slope lower than the one of the Meteoric Water Line. The value of the freezing slope can be calculated from knowledge of the isotopic composition of initial water at the onset of freezing and from the equilibrium fractionation coefficients between ice and water for deuterium and oxygen 18. Freezing slopes implying an inverse relationship between deuterium excess ($d = \delta D - 8\ \delta^{18}O$) and δ-values are displayed in lake ice [*Lorrain et al.*, 2002], in some basal stratified ice from polar glaciers [*Souchez et al.*, 1988; *Sleewaegen et al.*, 2004] and also in ice due to regelation past grains at the glacier bed [*Iverson and Souchez*, 1996]. By contrast, amber ice [*Cuffey et al.*, 1999; 2000] and basal ice with fine dispersed debris facies [*Knight*, 1997; *Souchez et al.*, 2002] do not exhibit freezing slopes but are produced by completely different mechanisms. Dealing with basal stratified ice, how can the apparent fractionation be explained, while in the isotopic model apparent fractionation does not occur at steady state?

5. Basal Stratified Ice With No Apparent Fractionation

[15] Figure 3 shows in different δD-$\delta^{18}O$ diagrams the position of samples of debris-bearing basal ice layers (black triangles) together with layers of glacier (meteoric) ice (open circles) from two Antarctic outlet glaciers. In Figure 3a are represented samples from Mackay glacier, an outlet glacier from the East Antarctic ice sheet crossing the Transantarctic Mountains and reaching the Ross Sea. The samples were taken from two overturned icebergs from that glacier, exhibiting a basal sequence tens of meters thick. In 1998, these icebergs were located at 76°57.5′S, 162°20.5′E and at 76°57.4′S, 162°20.1′E, respectively. In Figure 3b are represented samples from Taylor glacier, an outlet glacier from the East Antarctic ice sheet ending in the Dry Valleys area of South Victoria Land. The samples were taken in a tunnel excavated in the basal sequence in the left margin of the glacier, about 1 km from the snout.

[16] In both situations, samples from the debris-bearing ice are aligned on a Meteoric Water Line; the slope being 8.5 for Mackay glacier and 8.2 for Taylor glacier. If the middle point of the range of δ-values were considered as representative for glacier melt in the area, the calculated
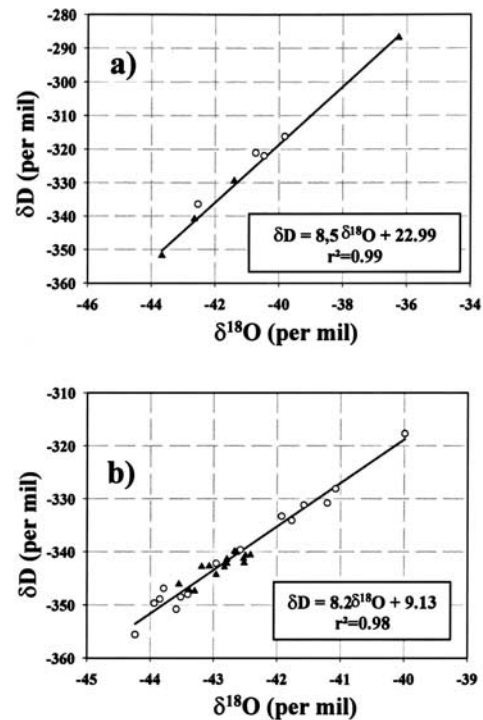

Figure 3. δD-$\delta^{18}O$ diagrams showing different samples from two Antarctic glaciers - Mackay (a) and Taylor (b) glaciers. Black triangles stand for debris-bearing basal ice layers. Open circles represent layers of glacier (meteoric) ice. The equations of the regression lines are given with their respective correlation coefficients.

freezing slope from equation (7) in *Souchez and Jouzel* [1984] would be 4.94 and 4.87 respectively. This is strikingly different from the values given above. There is thus no apparent fractionation in these debris-bearing ice layers. Thus, isotopic composition of the debris-bearing ice layers from these two Antarctic glaciers strongly suggests that they result from ice segregation by upward pore water flow in the subglacial fine-grained sediments.

6. Contrasting Basal Ice Isotopic Signatures

[17] While freezing slopes are displayed in a δD-$\delta^{18}O$ diagram for some debris-bearing basal ice layers of polar glaciers, none are present in the two examples given above. The question then arises: what are the conditions required for a freezing slope to be developed and to what extent are such conditions fulfilled.

[18] Since equilibrium fractionation always occurs at the ice-water interface, distribution of stable isotopes of oxygen or hydrogen in ice during water freezing is related to compositional variations in the liquid adjacent to the freezing front. These compositional variations depend on mixing in the liquid water that occurs by diffusion and by convection. A zone in which transport takes place by diffusion only always exists as a boundary layer adjacent to the ice-water interface. If the boundary layer is sufficiently thick so that the role of convection can be ruled out, a steady-state is developed in the ice in which the isotopic composition is the same as that of the initial water at the onset of freezing

[*Souchez et al.*, 1987]. The lower the isotopic diffusion constant in the water, the best developed is the steady state in the ice.

[19] Now, in the thin water films surrounding the grains of the subglacial sediment, diffusion must be particularly slow because of the tortuosity of the path and of the reduced thickness of the water film. Therefore, the isotopic model described above most probably applies and the debris-bearing ice shows no apparent fractionation. The suggested isotopic distribution in *Christoffersen and Tulaczyk* [2003b] is not very likely because diffusion cannot compensate sufficiently rapidly the impoverishment in heavy isotopes of the water adjacent to the ice-water interface.

[20] By contrast, if convection occurs in the water reservoir or if diffusion is facilitated, the ice will exhibit an isotopic composition different from that of the liquid water. Depending of the freezing rate, on trapping of unfractionated water pockets during ice accretion or on a reservoir effect - in a closed system, the enrichment of the ice produces an impoverishment of the residual water - the amount of observed fractionation is variable. This leads to the development of a freezing slope in a δD-$\delta^{18}O$ diagram. The same is true for regelation past grains in the subglacial sediment [*Iverson and Souchez*, 1996] in which multiple melting and freezing events and a loss of liquid water progressively produce an enrichment in heavy isotopes of the ice along the freezing slope.

[21] Varied isotopic responses result from different freeze-on processes in the subglacial environment. The isotopic composition can thus be an interesting tool for a better understanding of the mechanisms involved.

7. Conclusion

[22] Upward pore water flow in subglacial sediments as a consequence of ice-water interface curvature and surface tension effects can be at the origin of the formation of segregation ice at the glacier sole and of debris-bearing basal ice layers. A simple isotopic model indicates that no apparent fractionation will be detected in this case. Isotopic analyses, both in δD and in $\delta^{18}O$, of basal ice layers from two Antarctic outlet glaciers give weight to this assertion.

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