

CO-ISOTOPIC SIGNATURE OF TWO MECHANISMS OF BASAL-ICE FORMATION IN ARCTIC OUTLET GLACIERS

by

R. Souchez, R. Lorrain and J.L. Tison

(Faculté des Sciences, CP 160, Université Libre de Bruxelles, 50 Avenue F. Roosevelt,
B-1050 Bruxelles, Belgium)

and

J. Jouzel

(Laboratoire de Géochimie Isotopique/LODYC, CEA/IRDI/DESICP,
DPC-CEN Saclay, 91191 Gif sur Yvette Cédex, France)

ABSTRACT

Basal ice is formed by regelation as a consequence of both pressure-melting and freezing-on at the glacier sole.

A study of D/H and $^{18}\text{O}/^{16}\text{O}$ ratios in basal ice from five Arctic outlet glaciers indicates that a co-isotopic signature exists for these two mechanisms of formation.

The dispersed and stratified facies of basal ice present in these glaciers are related respectively to the occurrence of regelation and to freezing-on at the glacier base. Their origin is tentatively connected with the onset of basal sliding and the zone of bed decoupling due to basal water pressure in these Arctic outlet glaciers.

INTRODUCTION

Basal ice has been studied for its composition in terms of stable isotopes but very often only the $^{18}\text{O}/^{16}\text{O}$ ratio has been considered. For example, Gow and others (1979) described an oxygen-isotope profile of the basal ice from the deep ice core from Byrd Station, in Antarctica. Lawson and Kulla (1978) conducted an oxygen-isotope investigation of the origin of the basal zone of Matanuska Glacier in Alaska, and Hooke and Clausen (1982) studied the bottom layers of the marginal zone of Barnes Ice Cap in Baffin Island.

Co-isotopic studies of basal ice involving both $^{18}\text{O}/^{16}\text{O}$ and D/H ratios are not numerous in the literature. However, the plotting of δD against $\delta^{18}\text{O}$ opens up the possibility of following the processes by which the two heavy components behave in different ways. Different patterns for basal ice are displayed on $\delta\text{D}-\delta^{18}\text{O}$ diagrams in connection with different situations at the ice-bedrock interface (Souchez and De Groot 1985). The present study considers how freezing-on and regelation lead to different co-isotopic signatures in basal ice.

Basal ice from five glaciers is considered here. Two of the glaciers drain ice from the western part of the Greenland ice sheet, whereas the three others are outlet glaciers from the ice cap on Bylot Island in Arctic Canada. Figure 1 gives the names and the location of these glaciers, and the sites where basal ice has been sampled. Sampling procedure and sample analysis are similar to those described in Souchez and De Groot (1985). However, for basal ice the procedure is adapted to the sampling of about 1 ml.

THE FREEZING EFFECT ON δD AND $\delta^{18}\text{O}$ IN ICE

Jouzel and Souchez (1982) have shown that samples of ice produced by the freezing of water are aligned in a $\delta\text{D}-\delta^{18}\text{O}$ diagram on a characteristic freezing slope, different from the meteoric-water line (slope of about 8) on which rain, snow and glacier ice samples are displayed.

In a closed system, a progressive impoverishment of

heavy isotopes in both the residual water and the ice occurs in the course of freezing, so that the first ice formed has the least negative values of δD and $\delta^{18}\text{O}$. At the end of the process, the last ice layers formed may exhibit a considerably more negative value δD and $\delta^{18}\text{O}$ than initial water. This freezing effect has also been studied theoretically for open systems and the model has been tested in experiments with initial waters which had quite different values of δD and $\delta^{18}\text{O}$ (Souchez and Jouzel 1984).

The freezing slope in an open system is the same as for a closed one when the input is not significantly different in its isotopic composition from that of the initial reservoir. If water which is impoverished in heavy isotopes is mixed as input with initial water in the course of freezing, the ice samples may be aligned on a steeper slope than the freezing slope (Souchez and De Groot 1985).

Thus basal ice can be studied for its isotopic composition in terms of stable isotopes in order to gain information about processes which occur at the glacier sole.

ISOTOPIC COMPOSITION OF BASAL ICE

Basal ice is well exposed at the five localities examined. A clear-cut distinction can be made between two facies. On the one hand, a banded sequence of ice and debris layers (sometimes with debris up to a few centimetres thick) occurs in the form of thinly laminated (1–5 mm) beds. The ice consists of layers of only one-crystal thickness; the ice crystals are up to 1 cm long, with no bubbles, although a few thicker layers may be bubbly. The debris consists mainly of fine sand with scattered coarser particles. This first type is equivalent to Lawson's stratified facies (Lawson 1979) and is present at Aktineq Glacier and at both West Greenland sites. On the other hand, basal ice which contains a suspension of fine debris particles appears uniformly at the decimetric level. Ice crystals are equi-granular, from 1 to 4 cm long. The fine debris, often aggregated, is either included in the crystals or at the crystal boundaries. This ice is probably equivalent to the dispersed facies (Lawson 1979) and is present at Russell Glacier and at the three Bylot Island sites.

Stratified facies

Samples of thinly laminated ice and dirt layers from the basal parts of Jakobshavn Isbræ and Aktineq Glacier display a characteristic freezing slope on $\delta\text{D}-\delta^{18}\text{O}$ diagrams, as seen from Figures 2 and 3. On these figures, glacier-ice samples are aligned on a meteoric-water line, the values of the slope being 7.95 and 7.63 for West Greenland and Bylot Island respectively. As indicated previously (Jouzel and Souchez 1982), the freezing slope for stratified basal ice from Aktineq Glacier (4.9) agrees well with a closed-system

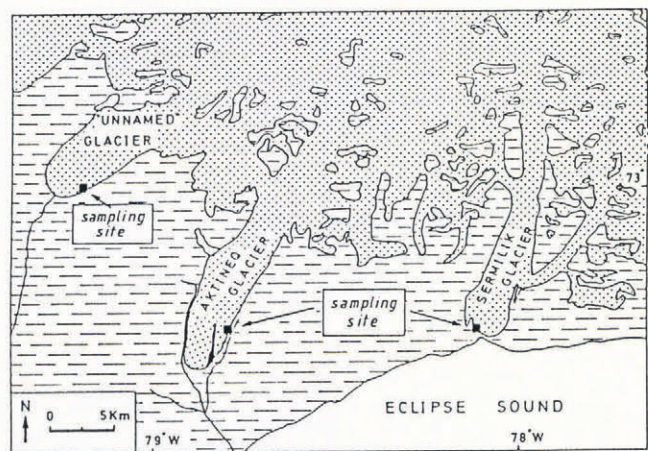
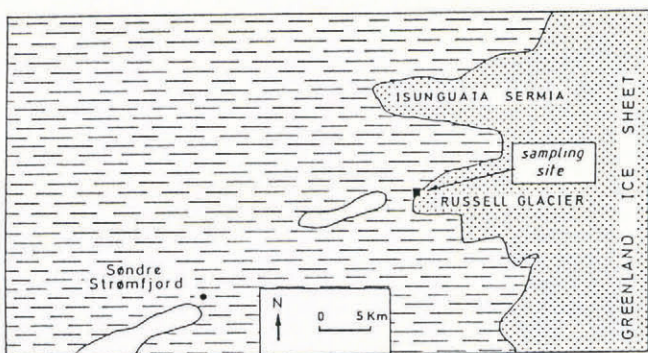
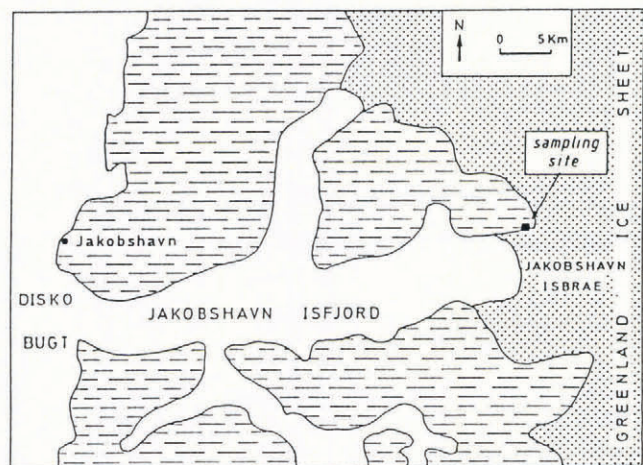


Fig.1. Location maps of the glaciers studied, showing the sampling sites.

model. The freezing slope at the Jakobshavn Isbrae site is a little too high: 6.3 instead of 5.5. Mixing in the course of freezing with water which is more depleted in heavy isotopes can explain the situation, but this water has not been sampled. If the two diagrams (Figs. 2 and 3) are compared, striking differences are noticeable. Let the isotopic values of the intersection point between the meteoric-water line and the freezing slope be those of the initial water. The range of δ values between the least negative sample on the freezing slope and initial water is, for Aktineq Glacier, less than the possible enrichment deduced from the equilibrium fractionation coefficients. However, for Jakobshavn Isbrae this range is over 28‰ in δD and over 4.5‰ in $\delta^{18}O$. Such a situation implies multiple melting-refreezing events: there is no fractionation on melting of the ice formed and enriched in a previous freezing step, and the melt water produced will be the initial liquid for the next freezing event. Therefore a higher range can be obtained. The thinly laminated ice and dirt layers from Jakobshavn Isbrae and from Aktineq Glacier can

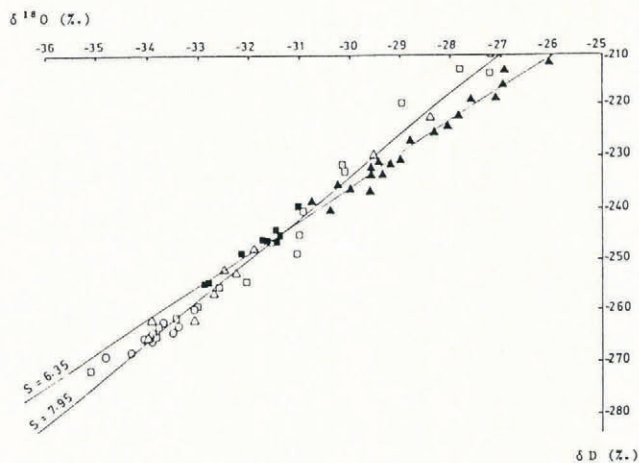


Fig.2. δD - $\delta^{18}O$ diagram for West Greenland ice samples. Squares represent glacier ice, triangles represent stratified basal ice, and circles represent ice with dispersed debris. Open symbols refer to Russell Glacier, black symbols refer to Jakobshavn Isbrae.

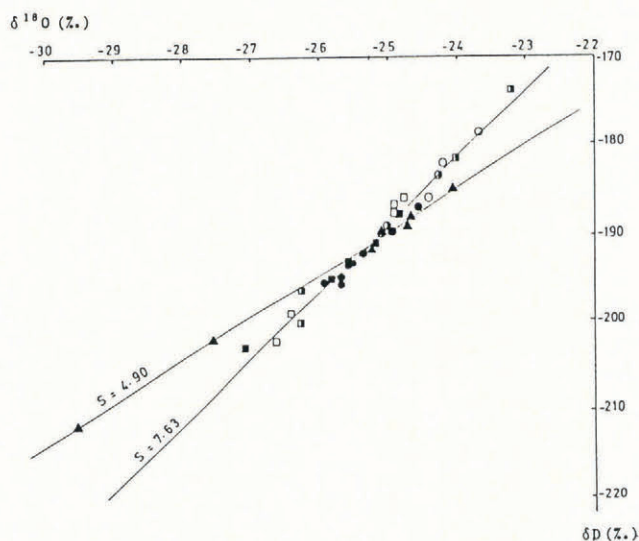


Fig.3. δD - $\delta^{18}O$ diagram for Bylot Island ice samples. Squares represent glacier ice, triangles represent stratified basal ice, and circles represent ice with dispersed debris. Open symbols refer to Sermilik Glacier, black symbols refer to Aktineq Glacier, and black-and-white symbols refer to the "unnamed" glacier.

thus be viewed, on isotopic grounds, as the result of freezing-on at the glacier sole from a limited water reservoir of specific isotopic composition, with (in some circumstances) the repeated occurrence of melting-refreezing processes.

At Russell Glacier, the distribution of points representing samples of thinly laminated ice and dirt layers is on the meteoric-water line. Such a situation can occur if the different ice layers originated from the freezing of different initial waters which had different compositions in terms of stable isotopes. In the Russell Glacier frontal zone, the range in δ values for glacier-ice samples is considerable: about 60‰ in δD and 7.5‰ in $\delta^{18}O$, which reflects the various origins of the ice from the West Greenland ice sheet. Because of folding, sequences are repeated and so melt water from various sources and of differing isotopic composition can be produced locally. A unique freezing slope can only be developed if the ice is formed from a single initial water source of limited extent, with a specific isotopic composition.

Dispersed facies

Ice with dispersed fine particles cannot be distinguished, by its composition in terms of stable isotopes, from glacier ice from the same locality. All points lie close to the meteoric-water line. Whereas for Sermilik and Aktineq glaciers the range in δ values in the dispersed facies is comparable with that in glacier ice, the situation appears to be different in Russell Glacier, of which the isotopic composition is very similar to the glacier ice with the most negative δ values (about -265‰ in δD and -34‰ in $\delta^{18}\text{O}$). The comparable range of δ values for ice with dispersed debris and for glacier ice at Sermilik and Aktineq glaciers is probably due to a less extensive sampling. Indeed, for logistic reasons glacier ice was taken only in the immediate vicinity of the basal sequence, but a considerable range of less negative values is likely to occur up-glacier near the centre line of these glaciers.

A regelation mechanism which occurs at the pressure melting-point around bed protuberances may explain these isotopic characteristics. Glacier sliding can occur by pressure-melting at the up-glacier side if the basal temperature is at the pressure-melting point and if a water film exists at the interface. The water film is likely to have an isotopic composition identical with that of the nearby glacier ice, as there is no fractionation on melting of glacier ice (Friedman and others 1964). As rates of freezing in the sliding process are equal to rates of melting, the amount of water present in the film remains constant. The freezing rate is equivalent to the sliding velocity, i.e. of the order 1 mm h^{-1} . Fractionation effects will not be detected if they occur only within individual frozen increments too small to be picked up at the scale of our sampling (1 ml) or, as is more likely, if depletion of heavy isotopes in the water film close to the freezing interface is not compensated for by the weak diffusion in the film. Indeed, in the last case, whereas the fractionation at the water-ice interface is always given by the equilibrium fractionation coefficient, the amount of observed fractionation is dependent on the ratio between the appropriate diffusion coefficient and the freezing rate (Souchez and others 1987). At a rate of 1 mm h^{-1} , after a short initial period of transition the isotopic distribution in the regelation ice will display a steady state: its ice composition will be identical with that of initial water. Regelation ice can thus have the same isotopic composition as the bulk water in the film and so it retains the original isotopic character of the parent ice. Differential movement of the ice around a bump, as suggested by Boulton (1975), allows the incorporation of the regelation ice which has been formed and the dispersal of debris which was originally located at the interface within the basal part of the glacier. This dispersal can also be considered as the result of a kind of diffusion mechanism in an ice matrix that undergoes simple shear deformation (Weertman 1968).

At the Russell Glacier site, regelation ice is produced in an interior zone where glacier ice with very negative δ values exists at the base at the pressure melting-point. This glacier ice is probably Pleistocene ice or ice formed at the ice divide, since it is only at Crête that present-day ice with a $\delta^{18}\text{O}$ of -35‰ has been found (Robin 1983).

More information can be obtained by comparing the results of fractional melting experiments conducted on glacier ice, on basal ice with dispersed debris and on banded basal ice. This is discussed below.

ISOTOPIC DISTRIBUTION IN BASAL ICE

As the diffusion coefficients of H_2O and H_2^{18}O molecules in ice are very low ($c. 10^{-11}\text{ cm}^2\text{ s}^{-1}$), melting of glacier ice occurs without isotopic fractionation. The situation appears to be different in a snow column when an isotopic change is measured during percolation of water (Arnason 1969), as a probable consequence of partial recrystallization (Búason 1972).

A fractional melting experiment performed on glacier ice from Russell Glacier confirms the absence of fractionation. In a suitable hot-water bath, a slice of about 125 ml from an ice core was locally heated at the base. As soon as it was produced, 25 ml of melt water was recovered with the aid of a polyethylene syringe and immediately press-filtered. Five successive 25 ml samples of filtered

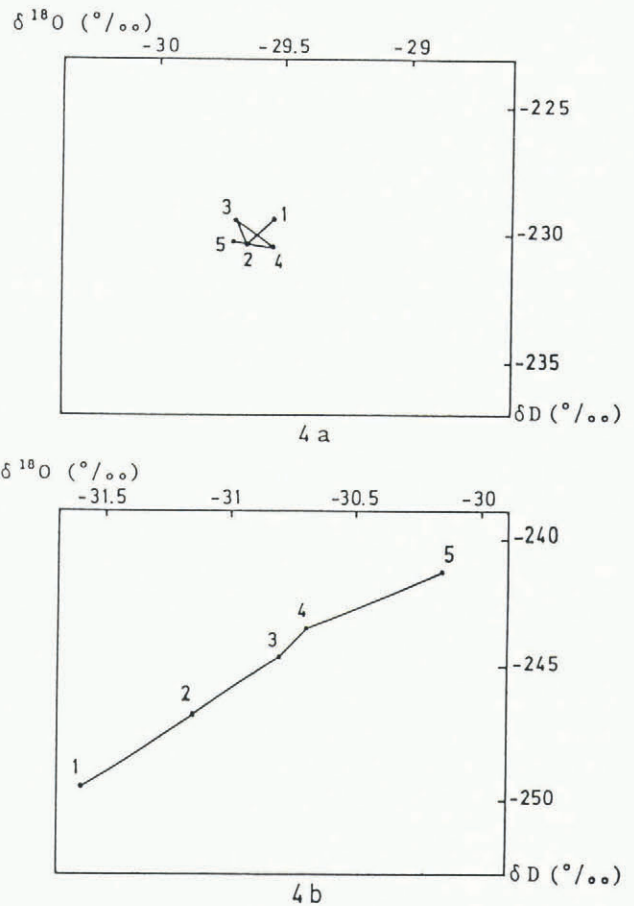


Fig. 4a and b. The two patterns obtained in fractional melting experiments on ice from Russell Glacier.

water were taken and their isotopic composition was analysed.

Figure 4a indicates that the points representing these five successive samples are clustered on a δD - $\delta^{18}\text{O}$ diagram. This implies not only the absence of fractionation during melting, but also (at the scale considered) the homogeneity of isotopic distribution in this glacier ice. A similar pattern is obtained for ice containing dispersed debris from the same glacier.

By contrast, in Figure 4b the points representing the five successive samples of ice from the banded sequences at the Russell site are aligned on a straight line. The slope of this line is about 5.8 and can thus be viewed as a freezing slope. The fractional melting experiment was conducted so that the first ice to melt was a debris-rich ice layer. Then the clear ice layer above was allowed to melt progressively. The first melt water obtained is the most depleted in heavy isotopes and a progressive change leads to the last melt water being the heaviest. Such isotopic distribution is not the consequence of fractionation during the experimental melting of this kind of ice but rather the influence of selective distribution during freezing. Therefore the experiment indicates a selective distribution of heavy stable isotopes in this basal ice which accords with the freezing-on process.

CONCLUSION

Because of folding of basal ice in the marginal zone of outlet glaciers, the morphology can be quite complicated. When the dispersed and the stratified facies are present together at the same site, there is evidence that ice with dispersed debris lies on top of the banded sequence and originates from a more distant source in the "interior" zone of the glaciers. Drewry (1986) recognized the widespread nature of this feature; the basal debris zone consists of a lower stratified and an upper dispersed region.

McIntyre (1985), in his study of the dynamics of ice-sheet outlets, considered that the origin of outlets

corresponds to the onset of basal sliding. Ice-sheet motion is dominated by internal deformation, whereas the steep gradients of subglacial valley heads initiate the basal sliding by pressure-melting and regelation. Down-glacier, as more water is present at the bed, higher basal velocities accompany a certain amount of decoupling. It is tempting to visualize the origin of the dispersed facies of basal ice, which has isotopic characteristics implying regelation, in that region of subglacial valley heads. When more water is present at the base and when, closer to the margin, the ice-bedrock interface temperature is below the pressure melting-point, the stratified facies – with isotopic characteristics which imply freezing-on – is formed by accretion below the dispersed facies. However, more information is clearly needed before this can be exposed to further critical tests.

ACKNOWLEDGEMENTS

A NATO grant for international collaboration in research and a Belgian National Science Foundation grant are gratefully acknowledged. They made possible the field trips to Greenland and to Arctic Canada respectively. The Geological Survey of Canada has also contributed support to the present study on Bylot Island, for which thanks are due to Dr Klassen.

REFERENCES

- Arnason B 1969 The exchange of hydrogen isotopes between ice and water in temperate glaciers. *Earth and Planetary Science Letters* 6(6): 423-430
- Boulton G S 1975 Processes and patterns of subglacial sedimentation: a theoretical approach. In Wright A E, Moseley F (eds) *Ice ages: ancient and modern*. Liverpool, Seel House Press: 7-42
- Büason T 1972 Equation of isotope fractionation between ice and water in a melting snow column with continuous rain and percolation. *Journal of Glaciology* 11(63): 387-405
- Drewry D 1986 *Glacial geologic processes*. London, Edward Arnold
- Friedman I, Redfield A C, Schoen B, Harris J 1964 The variations of the deuterium content of natural waters in the hydrologic cycle. *Reviews of Geophysics* 2(1): 177-189
- Gow A J, Epstein S, Sheehy W 1979 On the origin of stratified debris in ice cores from the bottom of the Antarctic ice sheet. *Journal of Glaciology* 23(89): 185-192
- Hooke, R L, Clausen H B 1982 Wisconsin and Holocene $\delta^{18}\text{O}$ variations, Barnes Ice Cap, Canada. *Geological Society of America. Bulletin* 93(8): 784-789
- Jouzel J, Souchez R A 1982 Melting-refreezing at the glacier sole and the isotopic composition of the ice. *Journal of Glaciology* 28(98): 35-42
- Lawson D E 1979 Sedimentological analysis of the western terminus region of the Matanuska Glacier, Alaska. *CRREL Report* 79-9
- Lawson D E, Kulla J B 1978 An oxygen isotope investigation of the origin of the basal zone of the Matanuska Glacier, Alaska. *Journal of Geology* 86(6): 673-685
- McIntyre N F 1985 The dynamics of ice-sheet outlets. *Journal of Glaciology* 31(108): 99-107
- Robin G de Q 1983 Profile data, Greenland region. In Robin G de Q (ed) *The climatic record in polar ice sheets*. Cambridge, Cambridge University Press: 98-112
- Souchez R A, De Groote J M 1985 δD - $\delta^{18}\text{O}$ relationships in ice formed by subglacial freezing: paleoclimatic implications. *Journal of Glaciology* 31(109): 229-232
- Souchez R A, Jouzel J 1984 On the isotopic composition in δD and $\delta^{18}\text{O}$ of water and ice during freezing. *Journal of Glaciology* 30(106): 369-372
- Souchez R A, Tison J-L, Jouzel J 1987 Freezing rate determination by the isotopic composition of the ice. *Geophysical Research Letters* 14(6): 509-602
- Weertman J 1968 Diffusion law for the dispersion of hard particles in an ice matrix that undergoes simple shear deformation. *Journal of Glaciology* 7(50): 161-165