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TRAPPING OF AEOLIAN SEDIMENTS AND BUILD-UP OF THE ICE COVER OF A DRY-BASED ANTARCTIC LAKE

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ABSTRACT

Among the perennially frozen lakes of the Dry Valleys of South Victoria Land (Antarctica), some are dry-based, i.e. frozen to the bottom. One of these is studied here by a multiparametric investigation (isotopic composition in δD and δ^{18}O, ions, gas and ice texture analyses). A sediment layer about 10 cm thick appearing at a depth of 3·5 m is also studied by grain size, X-ray diffraction and scanning electron microscope analyses. The information retrieved indicates that this ice-block lake results from a build-up in two steps and explains how aeolian sediments were included as a layer into the ice. Copyright © 2002 John Wiley & Sons, Ltd.

KEY WORDS: lake ice; aeolian sediments; stable isotopes; dry valleys; Antarctica

INTRODUCTION

The Dry Valleys region of South Victoria Land in Antarctica is an unglaciated area which displays numerous perennially frozen lakes, either in land-locked basins or dammed by glaciers. Such lakes contain liquid water beneath the ice cover (wet-based) or are frozen through to the bed (dry-based). In the latter case, as indicated by Chinn (1993), the ice thickness may far exceed those of the wet-based lakes. Once a lake is frozen to its bed, there is no more latent heat released by freezing at the base of the ice cover and a greater thickness of ice can be maintained under a given climatic regime.

The Dry Valleys lakes have been studied by numerous authors from the point of view of physical limnology, water chemistry or microbial ecology (see for example Chinn, 1993; Simmons *et al.*, 1993, Spigel and Priscu, 1998; Priscu *et al.*, 1998). The ice cover itself was much less studied, the more recent investigation being that of Adams *et al.* (1998) conducted on Lake Bonney.

Here we focus on a dry-based lake to study how sediment becomes trapped in the lake ice. The study is based on data obtained not only from sedimentological investigations but also from ice composition analyses. Stable isotopes (δD and δ^{18}O), major ions and gas composition analyses were performed, together with textural and structural analyses. To our knowledge, such an investigation has not yet been published.

FIELD SITE AND METHODS

The dry-based lake investigated is Lake Popplewell (unofficial name), a small frozen lake of Taylor Valley (Figure 1a) about 350 m long and 150 m wide. This lake is dammed by Suess Glacier descending on the slope of the eastern Asgard Range into the Taylor Valley. Previous papers (Fitzsimons, 1996; Lorrain *et al.*, 1999) have shown that basal ice from the frontal zone of Suess Glacier has been partly formed by freezing

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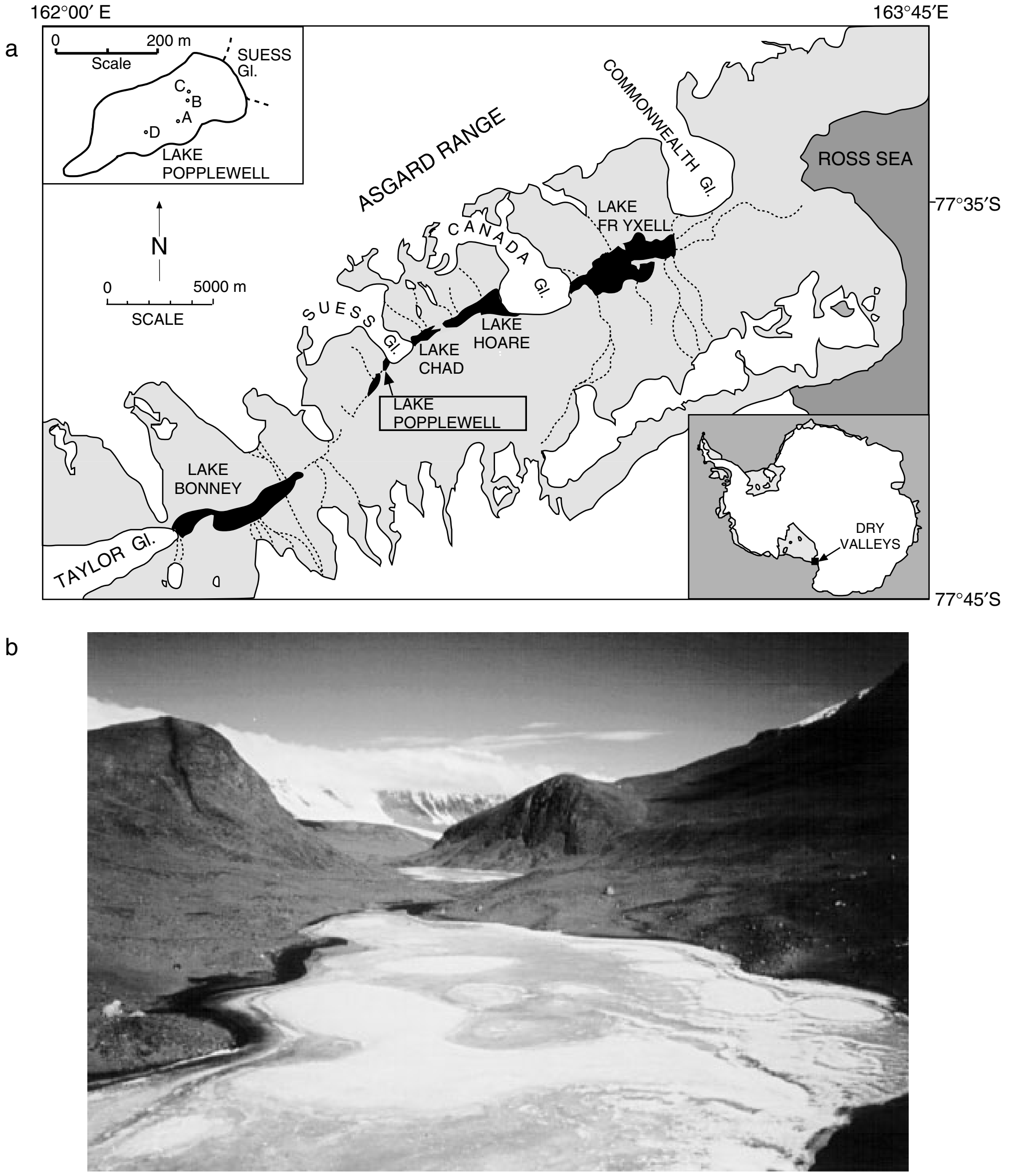

Figure 1. (a) Location map of the perennially frozen lakes of Taylor Valley. Inset: close-up of Lake Popplewell showing coring sites, and location of the Dry Valleys in Antarctica. (b) Westward view of Lake Popplewell from the top of the damming part of Suess Glacier

of water from the lake, this process leading to accretion of lake sediments at the base of the glacier. This suggests a wet-based Lake Popplewell in the past and progression of the glacier above the lake margins.

The surface of Lake Popplewell appears bluish and is very smooth except where a few ice mounds are present (Figure 1b). Four ice cores have been retrieved from the lake (points A to D in Figure 1a). They invariably show that the lake ice extends to the lake floor, indicating a dry-based lake. These cores were made during the summer but liquid water was not observed except in one case, in a decimetric pocket encountered at a depth of 307 cm. Thermistors put into the drilling holes indicated an isothermal situation at about −0·5 °C. The ice thickness is about 4 m. A narrow area of unfrozen water along the lake margin, called the moat, is usually present in summer. It is more developed along the southern lake margin, the northern one being more often in the shadow of a high slope dominating the valley.

Lake Popplewell is relatively young in the context of the late Pleistocene and Holocene history of Taylor Valley. During the Pleistocene, Taylor Valley lakes disappeared during warm climatic episodes because inland ice of the Taylor Glacier flowed further down. However, during the Pleistocene cold episodes of worldwide glaciations, the lower valley was invaded by Ross Sea ice which impounded large meltwater lakes. The last expansion of Taylor Glacier reached Lake Fryxell around 100 000 years BP while Ross Sea ice made an upvalley incursion in Late Wisconsin time (between 23 800 and 12 450 years BP) as far as Lake Fryxell (Denton *et al.*, 1989). It is only in the Holocene that grounded ice receded from the valley mouth and local glaciers like Suess Glacier expanded. As Lake Popplewell results from damming of the valley by Suess Glacier, it is only a few thousand years old at the most.

The ice cores retrieved from Lake Popplewell were transferred frozen to the cold room in Brussels, then sectioned and analysed. A crystallographic description was first made from vertical thin sections, followed by chemical and isotopic investigations. Chemical analyses were performed in Brussels using ion chromatography and atomic absorption spectrophotometry. Stable isotopes of hydrogen and oxygen were measured at the Centre d'Etudes Nucléaires de Saclay, France, by the usual methods of this laboratory (Lorrain *et al.*, 1999). Samples for sedimentological analyses were also taken in the cold room. These samples were analysed in the Blaustein Institute for Desert Research in Israël (Offer *et al.*, 1993): data were obtained from grain size and micromorphological analyses (Sedigraph and scanning electron microscope (SEM)), mineralogical investigations by X-ray diffraction and elemental chemistry (SEM). Furthermore, gas composition of the ice (CO_2, O_2 and N_2) was obtained in Brussels with the same techniques as those used for basal ice (Lorrain *et al.*, 1999).

RESULTS

Figure 2 gives the types of ice encountered in ice core A together with isotopic and chemical profiles. We concentrate our attention on this ice core because it contains well developed sediment layers within the ice in its deepest quarter (Figure 2). The lake bed is reached at a depth of 476 cm.

The upper part of the ice core down to a depth of 307 cm consists of lake ice with elongated crystals (about 40 cm long) having horizontal *c*-axes (Figure 3a). It is devoid of visible particles and contains vertical cylindrical bubbles about a few millimetres wide. The first 57 cm of ice from the surface were so brittle during sampling that they were not kept when packing the ice core in the field. At a depth of 347 cm, below a water pocket, there is a change in ice stratigraphy: a 5 cm thick ice layer with small crystals (a few millimetres long) and numerous small spherical bubbles is present, just above an ice-cemented sediment layer about 9 cm thick. From the base of this latter layer, sediments penetrate underlying ice within narrow vertical pipes which disappear below 365 cm (Figure 3b). From 361 cm to 423 cm, clear ice with clusters of particles is present. Ice crystals are still elongated but only a few centimetres long (Figure 3c). They show irregular boundaries and contain highly elongated bubbles. Below 423 cm, clear ice with spherical bubbles and sediment layers are interbedded until the lake bottom is reached at 476 cm. This ice shows polygonal crystals ranging from 0·1 to 1·3 cm in diameter (Figure 3d).

The isotopic and chemical profiles displayed in Figure 2 show striking features. The δD and $\delta^{18}O$ values become more negative downwards indicating an impoverishment in heavy isotopes down to 287 cm. This is followed by an enrichment between depths of 287 cm and 306 cm where impoverishment in heavy isotopes resumes, values as low as −286·7‰ in δD and −38·65‰ in $\delta^{18}O$ being reached at the lake floor. Concerning the chemical profiles, all exhibit low concentrations in Na^+, Ca^{2+}, Cl^- and $SO_4{}^{2-}$ down to a depth of 355 cm, i.e. at the level of the first sediment layer. Then the concentrations increase and reach the highest values for all the elements measured at 434 cm. The concentration increase reaches more than 300 times the values measured at the top of the core.

Sedimentological investigations produced the following results. Grain-size analysis of the sediment layer at a depth of 352 cm shows a bimodal distribution with one peak in the sand class and the other in the fine silt class (Figure 4). The intermediate class around 80 μm is quite weakly represented. Observation by scanning electron microscope shows that most of the particles exhibit features characteristic of aeolian facies. Some diatoms are present among the particles. X-ray diffraction shows that quartz, feldspars and calcite

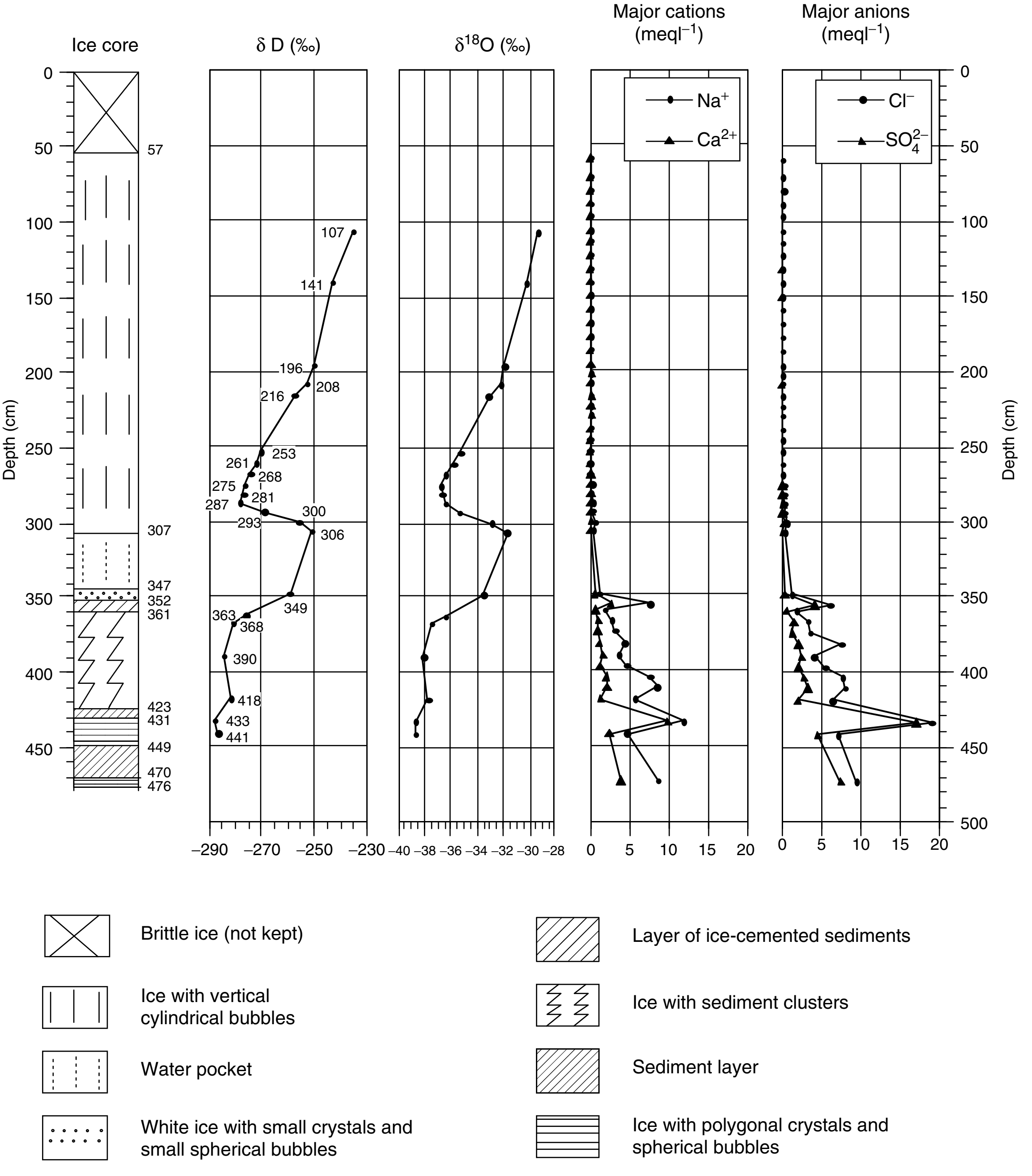

Figure 2. Main characteristics of core A: types of ice, stable isotopes and major ion profiles

are the dominant minerals with some kaolinite and montmorillonite in the finer fraction. Element chemical analyses (Table I) reveal the relatively high abundance of calcium (up to 36 per cent), iron (up to 13 per cent) and sulphur (up to 2 percent). Since sulphur-bearing inorganic minerals (pyrite, mirabilite, gypsum) are not present, it seems that sulphur is bound to organic compounds.

Algal mats are present on the lake floor and as traces in the sediment layers within the ice core. A microbial activity based on the sulphur cycle in an oxygen-depleted environment could have been present at the bottom

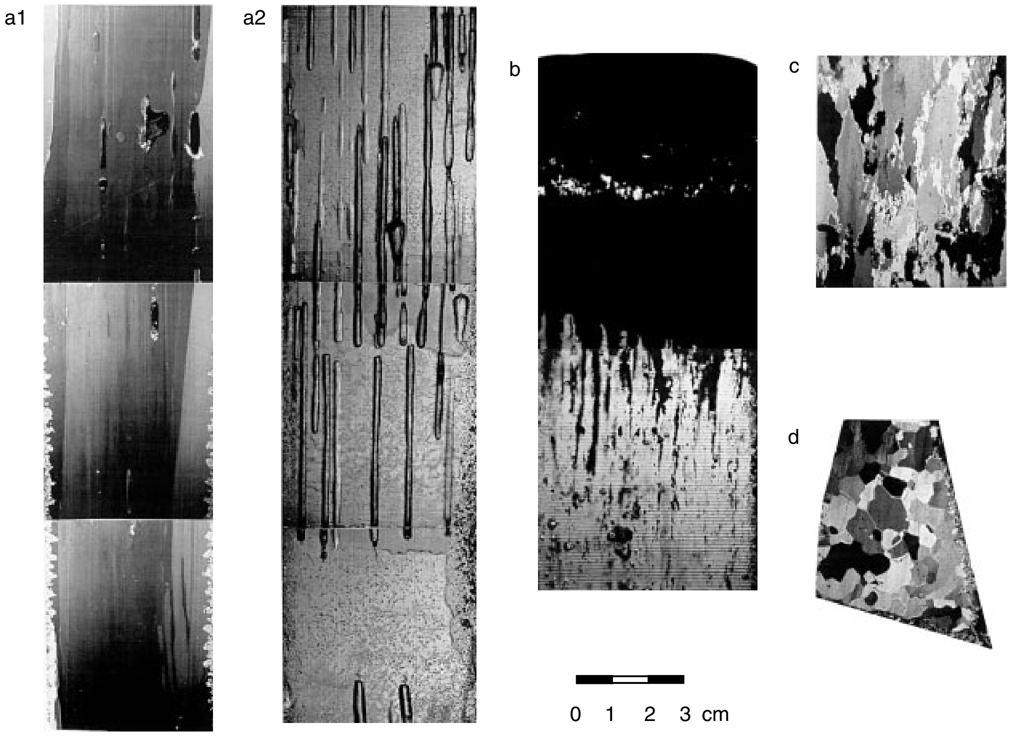

Figure 3. Different ice facies encountered in core A. (a1) Common elongated crystals (a1) thin sections between crossed polaroids; containing vertical cylindrical bubbles: (a2) thick section in normal light. (b) Sediment-filled vertical pipes below the sediment layer present at a depth of 352 cm (normal light). (c) Elongated crystals with irregular boundaries located at depths between 361 and 423 cm (thin section, polarized light). (d) Ice with polygonal crystals near the lake bottom (thin section, polarized light)

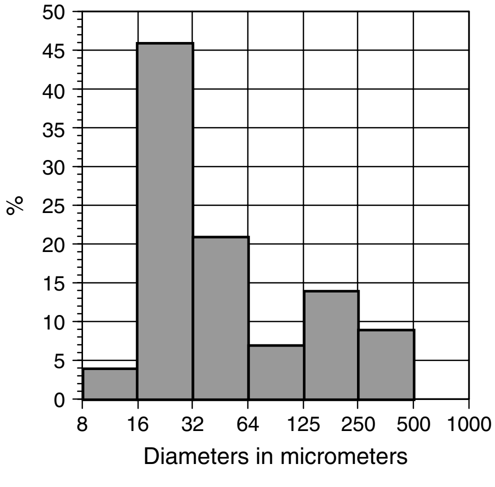

Figure 4. Grain-size distribution of the 9 cm thick sediment layer present at a depth of 352 cm

of the lake to explain the high sulphur and iron content of the sediments. Abundant granules of iron and sulphur compounds have been reported in similar mats of the neighbouring lakes Fryxell and Hoare (Simmons *et al.*, 1993). In the ice with cylindrical bubbles present from the top to the depth of 347 cm, gas analyses performed on two samples give values of CO_2 concentration ranging from 866 to 2250 ppmv and O_2/N_2

Table I. Results of chemical analyses of sediments from sediment layers or sediment clusters of the core

Depth (cm)	Na (wt %)	Mg (wt %)	Al (wt %)	Si (wt %)	P (wt %)	S (wt %)	Cl (wt %)	K (wt %)	Ca (wt %)	Ti (wt %)	Mn (wt %)	Fe (wt %)
349	2·80	3·85	12·11	49·24	0·16	0·42	0·14	3·99	16·57	0·80	0·40	9·89
355	1·89	3·55	7·76	35·52	0·30	0·86	0·80	3·39	36·22	1·00	0·33	9·10
363	2·17	4·02	9·70	43·16	0·16	0·39	0·20	3·91	22·65	1·10	0·28	12·39
425	1·52	3·66	9·74	36·06	0·47	1·34	0·50	3·99	29·01	1·10	0·30	12·69
433	1·68	3·47	7·48	31·93	0·47	1·36	0·19	3·56	37·39	1·00	0·25	11·25
441	2·11	3·57	8·30	37·83	0·30	1·00	0·07	3·83	31·07	0·80	0·30	11·04
450	3·65	3·49	10·34	42·89	0·10	1·11	0·55	4·43	22·21	0·90	0·08	10·25
476	2·69	4·13	10·92	49·90	0·22	1·58	0·12	3·89	15·86	1·00	0·06	9·47

ratios ranging from 0·40 to 0·64. By contrast, the ice with spherical bubbles situated at the bottom shows a CO_2 concentration of 252 076 ppmv and a O_2/N_2 ratio of only 0·12. This confirms the oxygen-depleted conditions at the bottom.

DISCUSSION

In a δD–$\delta^{18}O$ diagram, the ice samples are very well aligned on a straight line having the following equation: $\delta D = 5{\cdot}47\,\delta^{18}O - 6{\cdot}3$ (Figure 5). Such a slope was identified by Lorrain *et al.* (1999) as a freezing slope in this environment. The concept of the freezing slope developed by Jouzel and Souchez (1982; Souchez and Jouzel 1984) relies on the fact that freezing of liquid water has its own specific signature. This slope depends on the isotopic composition of the initial water (at the origin of freezing) and on the appropriate equilibrium fractionation coefficients between liquid water and ice for deuterium and oxygen-18. During progression of a freezing front into water, ice is enriched in heavy isotopes with regard to water. In a closed or quasi-closed system, the residual water is impoverished in deuterium and oxygen-18 in the course of freezing and the successive ice layers formed are therefore impoverished in heavy isotopes as well.

The impoverishment in heavy isotopes along the freezing slope observed here between depths of 107 and 287 cm (Figure 5) thus indicates that the lake has frozen as a closed water reservoir. The large elongated ice crystals with vertical cylindrical bubbles described from this part of the core (Figure 3a) are the fingerprint of that downward freezing process (Adams *et al.*, 1998). At first, the lake was wet-based. Progression of the freezing front in the liquid water body was possible because of surface ablation of the ice cover. The present-day ablation rate is about 30 cm a^{-1} in the Dry Valleys (Chinn, 1993). Such a situation can be in a steady state if the water losses are compensated by inflow of meltwater during the summer into the moat from surrounding areas of low albedo. However, if this water inflow markedly decreases, then the lake can become dry-based and, as indicated by Chinn (1993), cannot easily return to a wet-based situation.

Besides, there is an enrichment in heavy isotopes between 287 cm and 306 cm. This situation has also been observed in core D. It can be explained by mixing of the residual water remaining below the ice cover with water having a heavier isotopic composition. This composition must belong to the freezing slope cited above (Figure 5) because, if it were not the case, the ice formed after mixing would not plot on this slope. The moat is the most probable source of this water which results from melting during summer of local lake ice and of Suess Glacier ice. Eventually this water can make contact with the residual water trapped below the ice cover and diffusion of isotopes can take place (their diffusion coefficient in water is of the order of 10^{-5} cm^2 s^{-1}). The fact that there is again an impoverishment in heavy isotopes with depth below 306 cm suggests progressive closure of the residual water reservoir until the sediment layer present between the depths of 352 cm and 361 cm was reached. Isotopic and chemical evidence for this closure together with other developments are considered elsewhere (Lorrain *et al.*, in press).

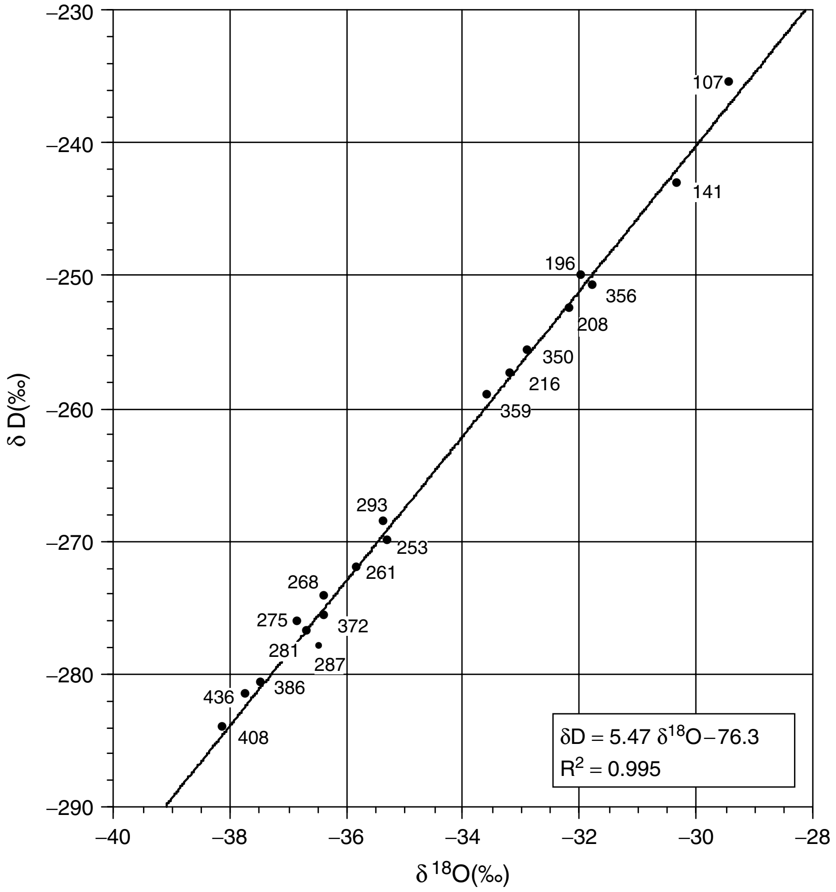

Figure 5. $\delta D/\delta^{18}O$ diagram showing the 21 isotopic samples taken from core A. Numbers refer to the depth (in cm) of each sample

The sediment layer located at a depth of 352 cm has most probably accumulated on top of a previous ice cover. It is unlikely that it was formed by progressive downward movement of particles into the ice as suggested by Nedell *et al.* (1987) and Squyres *et al.* (1991) for Lake Hoare. Indeed, there are no visible particles within the 3 m thick lake ice present on top of it. Furthermore, the bimodal character of the grain size distribution in the sediment layer together with a paucity of particles around 80 μm diameter is a feature well displayed in arid zones of the world. It is thought to be the consequence of saltation and sand movement by the wind (for an overview of some literature, see Pye, 1987; Offer and Azmon, 1992). Large sand particles can be pushed by the wind and roll on the smooth surface of the lake ice cover. During transport, the finer particles in saltation are trapped in the interstices between the large sand particles, with the larger particles in saltation bouncing, impacting the rolling particles and going farther. This results in the bimodal character observed. In such a scenario, the ice under the sediment layer represents an earlier ice-block phase of the lake which also resulted from complete freezing (as indicated by the isotopic impoverishment visible in Figure 2). The sediment layer can be considered as the record of strong winds in the history of the lake. Comparison with the other cores retrieved from the lake shows that the ice present under the sediment layer studied here is at a greater depth than the lake bottom encountered at these other drilling sites; this explains the unique characteristics of this core. These other cores show the same type of lake ice with cylindrical bubbles as in the top 3·5 m of the present core. They also indicate a closing reservoir effect as described above.

The chemical data confirm this interpretation. Concentrations in major elements in the lake ice remain quite low down to a few centimetres above the particle layer and show a peak within the layer. This profile illustrates the expulsion of most ions from the ice during downward freezing and their accumulation in the residual water leading to high ionic concentration in the last ice increments formed. The resulting peak appears at the level of the sediment layer. Below it, chemical concentrations rise again until they reach a maximum near the bottom of the core where the lowest values were detected in the isotopic profiles. It is possible that the sediments contain soluble precipitates which can influence ice chemistry in the laboratory when the ice samples are melted for analysis. This is suggested by the fact that calcium is quite high in the ice close to the sediments. Calcium carbonate precipitation is a normal feature of closing reservoirs at the end of freezing (Hallet, 1976; Killawee *et al.*, 1998) since its eutectic temperature is close to 0 °C. The chlorine content, however, exhibits the same profile, while the eutectic temperatures of chlorine salts are much lower, in the order of $-20\,^{\circ}$C. This last characteristic indicates that the increasing ionic concentration observed below 361 cm do not result from the sample treatment. The two concentration peaks appearing at the bottom of the core and at the level of the first sediment layer thus both indicate the closing reservoir effect detected through the isotopic study.

Sediments in lake ice of Dry Valleys lakes or on the lake floors are thought to be primarily of aeolian origin. Is the aeolian character displayed by some grains (roundness, pitted surfaces on electron microphotography) an acquisition during transport or is this character already present in the valley floor deposits? This question is difficult to answer. Two factors must be taken into account. Strong winds in the Dry Valleys were responsible for erosion and rounding of blocky material, development of stone pavements and presence of sand dunes (in Victoria Valley). At low temperatures, such as those prevailing in winter in the area ($-20\,^{\circ}$C and lower), ice crystals have the same hardness as orthoclase grains and they can readily be in suspension or in saltation during high wind periods. Therefore micromorphology of grains is perhaps a less efficient discriminating tool in such an environment.

Sediment layers within lake ice from the Dry Valleys are considered to represent a dynamic equilibrium between downward movement as a result of melting during the summer and upward movement of ice from ablation at the surface and freezing at the bottom (Priscu *et al.*, 1998). From different observations, it was also inferred that any sediment which was deposited on the ice cover could eventually make its way through the ice cover and settle in the water column below. Nedell *et al.* (1987) provided the first direct evidence of this by comparing the mineralogy and grain size distribution of sediment of different origins: from the surface of an ice cover, from the bottom of a lake and from adjacent melt streams. Developments along these lines by Wharton *et al.* (1989) and Squyres *et al.* (1991) put forward episodic point sources of sediment deposition through ice covers by way of cracks or gas bubble channels where ice covers are less than 3 m thick. More recently, Hendy (2000) proposed a model explaining how particles absorbing solar radiation can migrate through the ice cover or not, depending on their grain sizes. Such migration processes were probably effective in the lower part of the ice core beneath the sediment layer at a depth of 361 cm when the upper part of the lake ice was not yet present. This is indicated by penetration of particles into the ice from the base of the sediment layer (Figure 3c). The sediment layer itself cannot, however, be the result of such processes since the upper lake ice is devoid of visible particles and since isotopic and chemical data favour a build-up of lake ice in two steps separated by a ‘wind event’ responsible for the formation of the sediment layer.

Besides, Hendy *et al.* (2000) developed a conveyor belt model for Trough Lake, another perennially frozen lake in South Victoria Land. In this model, the floating ice cover is pushed by a glacier across a proglacial lake towards the marginal moat where melting occurs. Depending on their grain size, the sediments present on the ice surface differentially migrate through the ice cover and eventually settle on the lake floor along their path to the moat. This model is not applicable to the case studied here. Indeed, Lake Popplewell is presently dry-based, so that, due to the absence of liquid water at depth, the pressure exerted by glacier ice on the lake ice toward the moat does not exist. Such a situation could, however, have been present when the upper lake ice was constructed in wet-based conditions. But the absence of sediments in the upper lake ice rules out the set-up of the aeolian sediment layer by this process.

CONCLUSION

Progressive build-up of the ice cover of a dry-based lake is reconstructed thanks to a multiparametric study of the ice (isotopic composition of δD and δ^{18}O, ions and gas analyses, textural and structural investigations).

The information retrieved indicates that the present-day dry-based lake is the result of a complex series of events. This study puts forward how a layer of aeolian sediments was trapped in the ice. On top of a shallow ice block lake, wind-blown sediments are deposited at the surface until liquid water invades the lake basin allowing subsequent formation of an upper layer of lake ice by complete freezing of this water.

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REFERENCES

Adams EE, Priscu JC, Fritsen CH, Smith SR, Brackman SL. 1998. Permanent ice covers of the McMurdo Dry Valleys lakes, Antarctica: bubble formation and metamorphism. In Priscu JC (ed.), *Ecosystem Dynamics in a Polar Desert: The McMurdo Dry Valleys, Antarctica*. *Antarctic Research Series* 72. American Geophysical Union: Washington D.C.; 281–295.

Chinn TJ. 1993. Physical hydrology of the Dry Valleys Lakes. In Green WJ, Friedmann EI (eds), *Physical and Biogeochemical Processes in Antarctic Lakes*. *Antarctic Research Series* 59. American Geophysical Union: Washington D.C.; 1–51.

Denton GH, Brockheim JG, Wilson SC, Stuiver M. 1989. Late Wisconsin and early Holocene glacial history, Inner Ross embayement. *Quaernary Research* **31**: 151–182.

Fitzsimons SJ. 1996. Formation of thrust-block moraines at the margins of dry-based glaciers, South Victoria Land, Antarctica. *Annals of Glaciology Society* **22**: 68–74.

Hallet B. 1976. Deposits formed by subglacial precipitation. *Geological Society of America Bulletin* **87**: 1003–1015.

Hendy CH. 2000. The role of polar lake ice as filter for glacial lacustrine sediments. *Geografiska Annaler* **82A**(2–3): 271–274.

Hendy CH, Sadler AJ, Denton GH, Hall BL. 2000. Proglacial lake-ice conveyors: a new mechanism for deposition of drift in polar environments. *Geografiska Annaler* **82A**(2–3): 249–270.

Jouzel J, Souchez RA. 1982. Melting-refreezing at the glacier sole and the isotopic composition of the ice. *Journal of Glaciology* **28**(98): 35–42.

Killawee JA, Fairchild IJ, Tison JL, Janssens L, Lorrain R. 1998. Segregation of solutes and gases in experimental freezing of dilute solutions: Implications for natural glacial systems. *Geochimica and Cosmochimica Acta* **62**(23/24): 1637–3655.

Lorrain RD, Fitzsimons SJ, Vandergoes MJ, Stievenard M. 1999. Ice composition evidence for the formation of basal ice from lake water beneath a cold-based Antarctic glacier. *Annals of Glaciology* **28**: 277–281.

Lorrain R, Sleewaegen S, Fitzsimons S, Stiévenard M. In press. Ice formation in an Antarctic glacier-dammed lake and implications for glacier-lake interactions. *Arctic Antarctic and Alpine Research*.

Nedell SS, Anderson DW, Squyres SW, Love FG. 1987. Sedimentation in ice-covered Lake Hoare, Antarctica. *Sedimentology* **34**: 1093–1106.

Offer ZY, Azmon E. 1992. Chemistry and granulometry of settled dust in the northern Negev desert, Israel (1987–1988). *The Science of the Total Environment* **126**: 231–241.

Offer ZY, Zangvil A, Azmon E. 1993. Characterization of airborne dust in Sede Boqer area. *Israel Journal of Earth Sciences* **41**: 239–245.

Priscu JC, Fritsen CH, Adams EE, Giovannoni SJ, Paerl HW, McKay CP, Doran PT, Gordon DA, Lanoil BD, Pinckney JL. 1998. Perennial Antarctic lake ice: an oasis for life in a polar desert. *Science* **280**: 2095–2098.

Pye K. 1987. *Aeolian Dust and Dust Deposits*. Academic Press: London.

Simmons GM Jr, Vestal JR, Wharton RA. 1993. Environmental regulators of microbial activity in continental Antarctic lakes. In Green WJ, Friedmann EI (eds), *Physical and Biogeochemical Processes in Antarctic Lakes*. *Antarctic Research Series* 59. American Geophysical Union: Washington D.C.; 165–195.

Souchez RA, Jouzel J. 1984. On the isotopic composition in δD and δ^{18}O of water and ice during freezing. *Journal of Glaciology* **30**(106): 369–372.

Spigel RH, Priscu JC. 1998. Physical limnology of the McMurdo Dry Valley Lakes. In Priscu JC (ed.), *Ecosystem Dynamics in a Polar Desert: The McMurdo Dry Valleys, Antarctica*. *Antarctic Research Series* 72. American Geophysical Union: Washington D.C.; 153–187.

Squyres SW, Anderson SW, Nedell SS, Wharton RA. 1991. Lake Hoare, Antarctica: sedimentation through a thick perennial ice cover. *Sedimentology* **38**: 363–379.

Wharton RA Jr, McKay CP, Simmons GM Jr. 1989. Perennially ice covered Lake Hoare, Antarctica: physical environment, biology and sedimentation. *Hydrobiologia* **59**: 306–320.