Ice Formation in an Antarctic Glacier-dammed Lake and Implications for Glacier-Lake Interactions

R. Lorrain,* S. Sleveerget,* S. Fitzsimmons,† and M. Stievenard†

*Département des Sciences de la Terre, Université Libre de Bruxelles, CP 160/03, B-1050 Brussels, Belgium. rlorrain@ulb.ac.be
†Département de Géographie, University of Otago, PO. Box 56, Dunedin, New Zealand.
‡Laboratoire des Sciences du Climat et de l’Environnement, Laboratoire CCA-CNRS-CE Saclay, F-91911 Gif-sur-Yvette Cedex, France.

Abstract

Perennially frozen lakes are common features in the McMurdo Dry Valleys of South Victoria Land in Antarctica. Some of them, called wet based, contain liquid water capped by a permanent ice cover between 3.5 and 6 m in thickness. The others, called dry based, contain ice-block lakes. The thickness of the latter may far exceed those of the former. Their level is rising from freezing of the surface flooding of summer meltwater. However, we show here for the first time, using isotopic analysis together with an isotopic and gas content and composition study, that the ice of one of these dry-based lakes has been formed by complete freezing from top to bottom of a closed water reservoir and not by successive layers of ice growing on top of each other. We show how this lake, dammed by a cold-based glacier, has contributed to the formation of the basal ice layer of this glacier.

Introduction

The McMurdo Dry Valleys, a small group of ice-free areas situated along the coastal regions of South Victoria Land (Antarctica), are the sources for a series of closed-basin, permanently ice-covered lakes, some of which are dammed by local "alpine" glaciers.

Recently, three of us (Lorrain et al., 1999) examined the possible relationships between one of these ice-dammed lakes and an adjacent cold-based damming glacier (Lake Poppewell and Snow Glacier situated in the Taylor Valley—Fig. 1). We concluded (Lorrain et al., 1999), on ice composition grounds, that water from the lake has played a major role in the formation of the started sequence of ice and sediment layers accreted at the glacier base. This conclusion was shown to be consistent with a previously proposed model of debris entrainment by cold-based glaciers (Brobeck into lakes (Fritsen et al., 1999). This model states that transport-related conditions occur as ice flows into the unfrozen sediments of the lake bottom, creating conditions favorable for the entrainment of sediments and ice accretion by water freezing. However, our new fieldwork has allowed us to note that the studied lake is frozen to its bed. The question of the origin of the liquid water considered in the model of debris entrainment has thus to be re-examined in this context. To this end, this paper focuses on the formation processes of the lake ice itself.

The Dry Valleys lakes have been the subject of many recently published papers. Extended reviews available are in two books, edited by Green and Francis (1993) and Byers (1998). However, very few of the papers (and those cited therein) in these books are devoted to the ice core; three exceptions are extensive studies by Charn (1993), Adams et al. (1998), and Fritsen et al. (1998). Only the first of these deals with dry-based lakes. In particular, Charn (1993) shows that dry-based lakes are the most common type found at higher altitudes than the "inter- regional snowline" or away from the arid Dry Valleys area. However, he points out that lake Volta (Victoria Valley) is an exception as it is well below the regional snowline. The present study is based on ice composition analysis. Stable isotopes (δD and δ18O), major ions, total gas content, and gas composition analyses were performed; ice texture and structural studies were also carried out. To our knowledge, such an investigation of lake ice from the Dry Valleys has never been reported.

Site Description and Methods

Lake Poppewell (cf. Fritsen et al., 1998) is a relatively small, perennially frozen lake, about 356 m long and 150 m wide, dammed by Taylor Valley and Snow Glacier which depend on the northern slope of the valley and flows across its floor (Fig. 1). The lake's surface, which is about 149 m a.s.l., appears smooth and very clean, except in a few places where ice mounds are present. The latter are less than 1 m high, have a diameter of about 10 m, and present some shallow radial cracks. During the summer, a narrow zone (0.5 m of mean width) of liquid water appears mainly along the southern shore. This zone is even narrower along the northern shore because this side is more frequently in the shadow of the steep slope which dominates the valley; sometimes, the zone does not form on the northern side. The ice lake was drilled during the summer (January 1998) at four locations with a SUPERicy ice auger; 7.5 cm in diameter (Fig. 1). All drilling was stopped when the auger was not able to go deeper. In each case, the ice was in contact with frozen sediments, which we consider to be the bottom sediments, yielding a dry-based lake. The ice thickness varies between 2.7 and 4.8 m. Isotopic temperature measured a few hours after coring, was never lower than −0.5°C. Liquid water was not encountered except in one case, in a packet about 40 cm thick located at about 230 cm deep. Similar water packets have been observed by Charn (1993) and Fritsen et al. (1998) in the ice covering neighboring lakes (Lake Bonney and Lake Hurley; see Fig. 1).

Temperature loggers were used to monitor the ice cores from their initial storage in the field to their laboratory storage in Brussels; the temperature of the samples did not rise above −15°C. Once in Brussels, the samples were kept at −8°C.
Close examination of the four cores was performed in the coal laboratory in Brussels. Longitudinal thin sections were cut along the cores and photographed between crossed polars, allowing ice texture observations. It clearly appeared that the four cores were very similar except the basal part of the core A (Fig. 1) which is the longest (c.175 cm). Core A is the only core containing sediments layers, a few centimeters thick interspersed with thicker ice bands. Because of this particular situation, this core will be the subject of another study devoted to sediment trapping by lake ice. In the present paper, we discuss the data obtained from the detailed sampling of core D.

The isotopic analyses were performed at the Nuclear Research Center of Sclayen in France. Results are expressed in δ-units normalized to SMOW (Standard Mean Ocean Water). Precision of the measurements is ±0.5% on δD and ±0.1% on δ18O. All the other analyses were performed in Brussels. Na+, K+, Ca2+, and Mg2+ were determined by atomic absorption spectrophotometry (Varian SpectraAA 300). With a precision of 3% for Ca and 5% for Na, K, and Mg, were determined by inductively coupled plasma (Thermo DX-100) with a compact mass spectrometer (Thermo X-100) with a precision of 3% for Cl and 5% for SO4.4-. The concentration range encountered: NO3- was undetectable in most of our samples (detection limit = 0.2 ppm). Total gas content was determined by a milling-refining method using a Tougaard pump. This method was fully described by Raymond et al. (1982) and Barandiaran et al. (1993). The relative precision is better than 5% (Mattia et al., 1994). Gas composition was measured using a dry extraction technique at ≤55°C in a cold room, using a Varian 3300 gas chromatograph. The procedure followed was fully described by Raymond et al. (1982) and Barandiaran et al. (1993). The precision of the measurements is 2.5% for CO2 and 0.4% for O2 and N2.

**Results**

Figure 2a shows the different ice types observed in one of the ice cores retrieved from the lake ice. This core, 331 cm long, is considered to be representative of the lake ice mass. It does not contain any water pockets or any visible mineral particles except at its very bottom where the ice auger was able to obtain 1 cm of frozen sediments. From the surface to 33 cm depth, the ice was extremely brittle so that it was not possible to obtain a useful sample. Below this depth, the core shows columnar crystals, a 0.5-mm diameter long (some of these reaching 1 m long), with vertical axes. This texture is described as "candle ice" by Clotton and Mazé (1983) is typical of lake ice. Cylindrical bubbles, 1 to 6 cm long and 1 to 3 mm wide, are present all along the core (Fig. 3a). These crystal and bubble characteristics indicate downward progression of a freezing front through a water mass (Gow and Langton, 1977). The bubbles are produced by occlusion of gas within the ice during freezing and are oriented along the direction of ice growth (Adams et al., 1994). Close observation of the core, in particular between 120 and 180 cm depths, reveals that the shapes of the bubbles are highly variable. Among others, one finds thin and sinusoidal cylindrical and sometimes domed bubbles less than 1 mm in diameters (Fig. 3b), thick and irregular cylindrical bubbles reaching 5 mm in diameter (Fig. 3c, d, and large, nearly spherical bubbles up to 10

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mm in diameter (Fig. 3b), similar to the "inverted teardrop" described by Adams et al. (1980).

In the bottom 30 cm of the core, the bubbles appear as vertically oriented filaments a few millimeters to 1 cm long and with a diameter of the order of a tenth of millimeters (Fig. 3c). The ice crystals are still vertically elongated, but with horizontal c-axes. However, in the last 10 cm, the ice crystals become relatively small (1 cm to a few centimeters long) with random c-axes orientation and with cylindrical bubbles situated along their boundaries (Fig. 3f, g). In horizontal thin section, each individual crystal shows irregular boundaries and evenly spaced ice plates and brine lamellae (Fig. 3h) which has been previously described in the case of brine types of sea ice in the "brine layering plate substructure." This has been clearly illustrated by Gwe et al. (1987). Moreover, most of the fragments are present within the last 1 cm of ice, and many in the frozen sediments which constitute the very end of the core.

The isotopic profile reveals clear trends (Fig. 2b, c). The δD and δ18O values become more negative with increasing depth indicating a depletion of heavy isotope down to 256 cm. Below this depth, an inverse trend takes place until 256 cm where a new depletion is again displayed until the sediment is reached. Values are -232.8% in δD and -38.76% in δ18O at the top of the profile and reach -290.6% and -39.3%, respectively, near the bottom. The inverse trend detected between 236 and 256 cm deep represents quite smaller shifts, between -270.2 and -240.5%, in δD and between -35.68 and -30.19%, in δ18O. The deuterium excess profile (δD - 8 δ18O) appears as a mirror image of the δ18O profile.

The major ions profiles are simpler (Fig. 2d, e). All of them have relatively low concentrations along the core except at its very bottom where they sharply rise, reaching higher values. Along the last 30 cm, chloride, the dominant ion, becomes 100 times more concentrated (from 0.39 to 38.49 meq l⁻¹) and sodium, the dominant cation, increases from 0.45 to 30.20 meq l⁻¹. The other ions analyzed exhibit the same trend but to a lesser extent.

Gas content and composition analyses were performed on six samples situated between 79 and 319 cm depth (Fig. 2a). Gas content is always low (it ranges from 0.019 to 0.080 cm³ g⁻¹). This range is clearly related to the bubble concentration, but it must be noted that some bubbles are so large that sampling
FIGURE 5. Bubbles shapes displayed in vertical sections along core D and substructure: (a) column vertical cylindrical bubble; (b) thin sinus, sometimes dendritic cylindrical bubbles and big nearly spherical bubble similar to an inverted teardrop bubble; (c) thick and irregular cylindrical bubbles; (e) very thin vertically sausaged bubble, appearing like plansters; (f) cylindrical bubbles sausaged along crystals boundaries of another crystals visible in vertical thin section on (g); (h) horizontal thin section of the basal part of the core showing crystals with sausaged surfaces similar to the "heave-layered plate substructure" developed in sea ice (Gou et al., 1987).
necessarily implies the cutting of thorns, and thus loss of gases. The CO₂ concentration is relatively low all along the core (between 364 and 457 ppmv) contrasting with nearly 700,000 ppmv near the bottom. The O₂/N₂ ratio gradually increases from top to bottom from 0.37 to 0.54.

**Discussion**

When plotted on a δD/δ18O diagram (Fig. 4), the 19 ice samples taken along all the ice core show a linear relationship expressed by the equation:

\[
\delta D = 5.4 \times \delta^{18}O - 77 \quad (r^2 = 0.99)
\]

(1)

The slope of this regression line (\(r = 2.4\)) called freezing slope, is an isotopic signature of ice formed by progressive freezing of a closed water reservoir, as explained by Jouzel and Souchez (1982) and Souchez and Jouzel (1986). As the freezing front moves, ice is enriched in heavy isotopes with respect to water. In a closed or quasi-closed reservoir, the residual water becomes more and more depleted in heavy isotopes and the successive ice layers formed are therefore depleted in heavy isotopes as well. The slope depends on the isotopic composition of the water at the beginning of freezing and on the ice-water equilibrium fractionation coefficient for D and \(\delta^{18}O\). The following equation from Souchez and Jouzel (1984) can be used to compute this slope:

\[
\delta D = \delta^{18}O - \delta^{18}O_{res} \left( \frac{1}{1 - \epsilon_{\delta D/\delta^{18}O}} \right)
\]

(2)

where \(\delta D\) is the δD composition of the parent water, \(\delta^{18}O\) is δ18O composition, and \(\delta^{18}O_{res}\) are the values of the ice-water equilibrium fraction coefficients for deuterium and for \(\delta^{18}O\), respectively (Souchez and Jouzel, 1984). The composition of the parent water, as suggested by Souchez and Jouzel (1984) is given by the intersection of the freezing slope with the local meteoric water line. The latter has been obtained by Lorrain et al. (1990) from 25 glacier ice samples from the surface of Saint Glacier. These 25 samples have isotopic values ranging from -39.55 to -30.79% in \(\delta^{18}O\) and from -307.4 to -248.6% in \(\delta D\). Very few measurements were previously published for this glacier: -30.00 and -220.60% (Merrit et al., 1970), -30.40 and -230.60, and -35.15% in \(\delta^{18}O\) (Soulier et al., 1982). Most of the values reported by Lorrain et al. (1990) are more negative. The meteoric water line they obtained has the following equation:

\[
\delta D = 8.1 \times \delta^{18}O - 9.4 \quad (r^2 = 0.99)
\]

(3)

The corresponding isotopic values of the parent water are 3% in δ18O = 33.3% and in δD = 253.89%. Using them as \(\delta D_{res}\) and \(\delta^{18}O_{res}\) in equation (2) with \(\alpha = 1.00\) and \(\beta = 1.00\), respectively (Souchez and Jouzel, 1984), we obtain a theoretical freezing slope of 5.5 which is in close agreement with the slope of 5.4 obtained from our samples.

It is also possible to estimate the isotopic composition of the ice formed when the freezing front has reached the bottom of the water reservoir. In a closed system, what mass is gained by the ice phase during freezing is lost by the liquid phase. Therefore, as demonstrated by Jouzel and Souchez (1982),

\[
\delta D = 1000 - \frac{N_y}{N_x} \delta^{18}O = 1000
\]

(4)

where \(N_x\) is the total number of moles in the liquid system or the number of moles of the liquid when freezing begins, \(N_y\) is the number of moles in the solid phase at time \(t\), \(\delta^{18}O\) is the δ18O value of the solid phase near the liquid-solid interface at time \(t\), \(\delta^{18}O\) is the δ18O value of the solution when freezing begins, and \(\alpha\) is the equilibrium fractionation coefficient between solid and liquid. From equation (4), we can estimate the δ18O values of the ice corresponding to a given frozen fraction \((\alpha / N_x / N_y)\). For instance, in 5% freezing, the corresponding isotopic composition is δ18O = -39.35%, and δD = 290.65%. These values are nearly the same as the ones of the deepest isotopic sample (see in Fig. 2, sample No. 10, 322 cm deep). δ18O = -39.35% and δD = 290.65%.

Thus, isotopic results confirm that the lake ice studied was formed by progressive and complete freezing of the whole liquid water mass from top to bottom. However, between 230 and 215 cm depth, an enrichment in heavy isotopes with depth has been detected. This can be explained by mixing of the residual liquid water of the lake with water having higher isotopic values that plot onto the freezing slope cited above. Only on the latter condition can the isotopic values of the ice formed after mixing plot onto the same freezing slope (e.g., point 12 in Fig. 4). We think this mixing occurred when, during summers, the residual water situated at the lake bottom eventually melted into contact with melt water.

**Chemical Data**

The chemical results confirm the interpretation of the isotopic data. The ionic concentrations profile shows low values all along the ice core except at its bottom that expresses the chemical effects of the migration of the freezing front. During downward freezing of the lake water, most ions are expelled from the ice and diffuse into the reservoir (Wright and Zill, 1970, Bullen, 1976). Near the end of the closed reservoir the ice concentration becomes very high and, as a result, ices in the ice as well. Finally as the freezing front becomes grounded, the last ice increments must include most of the expelled ions.

The high final ionic concentration of this ice, although weaker than in the case of sea ice, is responsible for the presence of the brine lens/freeze interface structure described above and which is typical in sea ice (Fig. 3b). Regular elongated proto-
berenices appear at the ice-water interface during ice growth and salts are present as liquid inclusions trapped along these sub-
structures (Lohmann and Weiss 1990). Comparison of isopic profiles with the isotopic ones, raises the question of the lack of symmetry between them. One could indeed, expect the latter, at least, to be the higher one, since on the one hand ice is expelled from the growing solid phase and on the other hand heavy isotopes are preferentially incor-
porated in the growing ice. In fact, we observe different situations. Firstly, the "end of reservoir" effect appears higher in the profile for isotopes than for ions. Secondly, the isopic enrich-
ment detected between 230 and 236 cm does not correspond to the same extent to a concentration of ions on the isopic profile. These differences can be explained as follows. The isopic profile mainly describes the isotopic com-
position of the ice laminate of the crystals whereas the isopic profile mainly describes the brines present at the grain boundaries. Be-
cause the ice crystals are large in the studied core, the samples taken for isotopic analyses have only a small number of crystal boundaries where the brines are located. On the con-
trary, within the deeper 10 cm as indicated above, the ice crys-
tals are much smaller and the crystal boundaries intercepted more numerous.

GAS DATA

Results of the gas analyses give support to the interpretation of the other data. The gas total content is very low compared to values typical of polar meteoric ice, which is about 0.1 cm³ g⁻¹ of ice at standard temperature and pressure conditions (Marti-
nette et al., 1992). The values obtained from our samples are around 0.02 cm³ g⁻¹ with one exception near the bottom of the core, where it is as high as 0.05 cm³ g⁻¹. These values are, however, higher than the normal gas content of bubble-free ice which usually is less than 0.0003 cm³ g⁻¹ (Berner et al., 1979). By comparison, the solubility of air in water at 0°C is about 0.03 cm³ g⁻¹. Experiments of downward growth of ice from dilute solutions in a closed reservoir were carried out by Killworth et al. (1998). Some of these, conducted at constant linear rates of about 0.8 mm y⁻¹ have lead to interesting conclu-
sions which are relevant to our results. For instance, they have produced total gas content ranging between 0.05 and 0.06 cm³ g⁻¹. In their later stages, i.e., in the ice formed when the freezing front was approaching the bottom of the reservoir, they are very similar to some observed in deep parts of our ice core (Fig. 3).

Our results thus again reflect the progressive freezing from top to bottom of the lake water mass. During this process, gases become supersaturated in the region immediately below until bubbles nucleate and form a gas phase. As freezing proceeds, gas is transferred from solution to the bubbles until finally the bubbles are entrapped in the ice (Craig et al., 1992).

The measured CO₂ concentration is higher than the present day atmospheric value all along the core (ranging from 564 to 4557 ppmv but at the bottom it reaches the exceptionally high level of 699,000 ppmv which to our knowledge is the highest value ever reported for natural ice. Until now, the highest pub-
lished values were 222,000 ppmv in the basal ice of Smeerenburg Glacier (Larose et al., 1999) and 135,000 ppmv in the basal ice of the GRIP core in Central Greenland (Schaerer et al., 1991). In the experiments of Killworth et al. (1998) cited above, CO₂ con-
tent commonly reached 20%. In one of them, where the rate of freezing was about 2.2 μm s⁻¹ and the calcium concentration of the ice from the last freezing stages was about 1.3 mmol L⁻¹, the CO₂ content peaked at about 85% (350,00 ppmv), which is close to our value of 699,00 ppmv obtained at the very bottom of the core. This experiment showed that this very high CO₂ concentration (much higher than the CO₂ concentration in air dissolved in water at 0°C and at atmospheric pressure) is ex-
plained by the addition of CO₂ produced by CaCO₃ precipitation (CaCO₃ + 2 HCO₃⁻ → CaCO₃ + H₂O + CO₂). And, indeed, calcium carbonates were found within the ice produced in the last freezing stages.

In our case, the calcium concentration of the deepest ice samples of the core (2.187 mmol L⁻¹) is even higher than the values of the experiment reported above (1.3 mmol L⁻¹). This process of CO₂ enrichment is thus highly probable in the con-
text of calcium precipitation in a calcium-carbonate-ice-saturated solu-
tion as described by Hallet (1976). The particle content of the ice from the very bottom of the core has been investigated by scanning electron microscopy. Calcium precipitates have been detected (Fig. 5) confirming the occurrence of the de-
scribed process. Concentration of solutes by freezing leads to a decrease in pCO₂ and supersaturation for calcium at the ice-water interface. When the solution becomes highly super-
saturated for calcium, precipitation occurs and is associated with degassing of bubbles containing over 50% CO₂ which become traped in the ice.

BUILD-UP OF THE ICE COVER

From the different aspects of the ice study developed above, it clearly appears that the ice retrieved from the "ice-
block" Lalor Peppermint has built up from a closed or quasi-
closed water reservoir by progressive downward freezing until the lake bottom has been reached. However, no water scale can be provided. It is well known that a water body in contact with the atmosphere at this locality can form an ice cover about 3 m thick over the first year of its existence (Christou, 2001), but this does not help in this case. Indeed, due to superficial ab-
lation, the 3.2 mm thick ice we have sampled does not corre-
spond necessarily to the whole thickness formed since the be-
inning of the process. The process may have lasted several

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FIGURE 5. SEM photographs of calcium carbonate precipitate from the particles in the ice from the very bottom of the studied core. Calcite crystals appear as scalenohedrons showing growth patterns, for instance in the upper right part of the picture;
years, so long as melt water has not been in contact with the water remaining at the bottom of the central part of the lake, as discussed above, when the ice between 230 and 256 cm deep formed. As pointed out by Chinn (1993), both the lake levels in the Dry Valleys and the equilibration thickness of the ice create difficulties in reading. We think that during the history of the glacier-dammed lake, a floating ice cover eventually grounded, except in the deeper central part of the lake and due to the dynamic conditions allowed thickness of the ice until the bottom was reached in that part. The ice studied here was formed after the grounding event, probably else when some of the lake ice crystals above, once the lake has completely frozen, there is no further heat from freezing at the base (the lake becomes dry based), it is not as critically. We think, small equigranular crystals that do not show discrete structure or c-axis orientation patterns (Pollard and Van Everdingen, 1982). Moreover, autochthon presents a specific isotopic signature; absence of correlation between δ and δtritium excess (d) values (Souchez et al., 2003). This is the opposite of what we observe here (see discussion above) and which range from 0.200 to 0.305 cm/s (Chinn, 1993). However, Lake Henderson Jones named Mummy Pond—see Fig. 1), which is located a few hundred meters westward of Lake Henderson and also dry based, has been reported as slightly rising (a few decimetres) between 1972 and 1990 (Chinn, 1993). Lake Pope- well, should also be described as a lake which more probably behaves like its close neighbor Lake Henderson. We think that autochthon is probably present on top of the studied mummified ice core, because it is the upper 35 cm which were discarded during sampling since it was extremely brittle, as indicated above.

GLACIER-LAKE INTERACTIONS

Chinn (1993) has extensively examined the possible interactions between Dry Valleys lakes and glaciers margins. He has considered configurations generated by equilibrium, advance or retreat of the glaciers and by static, rising or falling lake levels. Situations like lake ice moving over glacier ice or glacier ice extending or out of lake ice are possible and will contribute to the complexity of what can be found. As in glacier base. In the present case, basal ice of the dimenions of Sissu Glacier has almost been investigated. Thus, the glacier in total based (temperature lower than ~17°C), it shows some blocked of undistorted facies expands within its basal part. The lower contact occurs at ~2 m below the ice surface that the particle-size characteristics of the sediments are the same as those of the adjacent lacustrine environment (Prest- man, 1990). The width of the intercalation of the interbeds of the blocks cited above suggest the sediments were frozen during entrainment. The analysis of stable isotopes and of gas content and composition of this basal ice have indicated that these blocks were formed by the freezing of water. The 85 ice samples taken from this basal ice have isotopic values ranging from δ18O = -28.10 to -38.45%; δ18O = -22.85 to -28.60% in δD. In a δD-δ18O diagram, they show a regression line having the equation:

\[ \delta D = 5.4 \delta^18O - 9.74 \] (5)

which is practically the same as equation (1) obtained from the lake ice samples presented above. The gas analyses also have shown clear similarities with what is reported here for lake ice: low total gas content (0.000 to 0.083 cm³/g) high to very high CO₂ concentrations (446 to 2,220,000 ppmv) and O₂/N₂ ratios ranging from 0.25 to 0.52. From these results, Lorrain et al. (1999) concluded that the basal ice of Sissu Glacier has been formed by freezing of water from the lake. It is now established that Lake Popewell no longer exists (Chinn et al., 1998). As a glacial ice, basal ice investigated has partly formed by incorporation of lake ice at the glacier base, as in one of the situations envisaged by Chinn (1993). δ18O values of ice layers of isotopic composition present in the basal ice of Sissu Glacier confirm the local freezing of liquid water pockets. For instance, Figure 6 shows the variation of the isotopic composition in δD and δ18O of one of these ice layers about 13 cm thick situated at the bottom of the basal ice sequence. One can observe isotopic shifts from -29.32 to -37.57% in δ18O from and from -23.0 to -28.4% in δD. These shifts at very small thickness (about 12 cm) indicates that a water pocket of nearly the same isotopic composition as the lake water has frozen completely from both directions. This interpretation is reinforced by the fact that the CO₂ composition varies from 1177 ppmv at one side to 11.36 ppmv at the middle. The isotopic and gas characteristics are not compatible with what should be observed within a slab of lake ice about 12 cm thick incorporated solid at the glacier sole. In this case, indeed, only a small isotopic shift would be developed within a decimetre thick ice layer. Here the mean isotopic gradient in δD is about 8% cm⁻¹ whereas the highest gradient within the lake ice studied is about 0.6% cm⁻¹. Moreover, similar isotopic shifts and gradients have been encountered higher up in the basal ice sequence, 35, 270, and 290 cm above the lake ice layer described here in detail. The corresponding isotopic
Conclusion

Isotopic analyses in 8D and 18O of the ice of Lake Poppwell together with sonic and ice data strongly suggest that this ice-block lake has been formed by progressive freezing from ice to bottom of the closed water-trap remaining in the largest part of the lake after the grounding of a floating ice cover probably due to a lowering of the lake level. This quality confirms that the basal ice sequence visible at the front of Stairs Glacier, has been built up, at least partly, by freezing of bottom water from Lake Poppwell in a period during which the lake was wet-based.

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