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JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 100, NO. C4, PAGES 7019–7025, APRIL 15, 1995

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Abstract. A better knowledge of boundary conditions near a grounding line is critical for understanding the behavior of ice shelves and floating glaciers. We show here that significant information can be gained from a high-resolution isotopic and textural study of marine ice accreted at the bottom of a floating glacier near its grounding line. Two different types of marine ice have been found. Type 1 is bubble- and debris-free ice with properties which, we believe, can be explained by intrusion of brackish water in open basal fissures. Closing of the fissures by progression of a freezing front from the sides is precluded, and filling by frazil ice is favored. Type 2 is made of thin, clear ice and debris layers which are thought to have formed when a subglacial water-filled sediment enters into contact with seawater and is subjected to freezing under a double-diffusion process. The paper also stresses that in a δD-$\delta^{18}O$ diagram the alignment of marine ice samples on a mixing line does not necessarily imply a mixture of continental water and seawater in varying proportions.

Introduction

Thermodynamic processes occurring at the base of an ice shelf control basal melting or accretion of marine ice. These thermodynamic processes are determined by the temperature of both ice and seawater near the interface and by the speed of the ocean currents.

Small Antarctic ice shelves or ice tongues, with an ice thickness at their front of approximately 100 m or less, are likely to react more rapidly than thicker shelves to a temperature change because their base can be affected by a temperature increase of oceanic waters. One critical area for the stability of the whole floating glacier is the grounding line.

It is generally assumed that if the melting point is reached at the glacier base in the coastal region, subglacial meltwater loaded with sediments discharges into the sea at the grounding line. *Zotikov* [1986] has pointed out that if this subglacial meltwater reaches the sea, a layer of relatively fresh water will exist above normal seawater beneath the ice shelf. Since the freezing point of fresh water is higher than the freezing point of seawater, bottom freezing will probably occur and be responsible for a thickening of the ice shelf at or near the grounding line. However, the only drilling made today of an ice shelf through its entire thickness (ice core J9 on the Ross Ice Shelf) showed no evidence of fresh water in accretion, but rather, adfreezing of sea ice [*Zotikov et al.*, 1980]. J9 is located along a flow line that connects with an ice stream. Accordingly, any fresh water flowing beneath the ice stream at the grounding line could potentially accrete as freshwater ice to the base of the Ross Ice Shelf. This, in fact, is not observed in the J9 core.

In order to preserve a body of relatively fresh water, the water column must be relatively calm and not influenced by tidal mixing which *MacAyeal* [1984] thinks should be efficient near a grounding line. The presence of an inverted topography such as crevasses, domes, or channels at the ice-ocean contact [*Orheim*, 1986; *Hellmer and Jacobs*, 1992; *Tison et al.*, 1993] is a way to achieve this sheltering effect.

Marine ice of substantial thickness has been found at the base of some ice shelves [*Oerter et al.*, 1992] and ice tongues [*Gow and Epstein*, 1972] or at coastal ice margins under an ice cap [*Goodwin*, 1993]. In order to better understand phase changes at the ice shelf–ocean interface, different modes of water circulation have been described and various models of ice shelf–ocean interactions have been proposed in the literature [*Jacobs et al.*, 1979; *Lewis and Perkin*, 1986; *Jenkins and Doake*, 1991; *Nicholls et al.*, 1991; *Hellmer and Jacobs*, 1992].

An insight into the problem of determining the conditions necessary for freezing at the ice shelf–ocean interface can be gained by studying the isotopic properties of the marine ice, both in δD and $\delta^{18}O$. With this last perspective in mind a field program was conducted in Terra Nova Bay area within the 1989–1990 Italian Antarctic Program in order to study outcropping sites of marine ice which has been accreted at the base of ice shelves or ice tongues. Two such sites have been sampled, one at Hell's Gate Ice Shelf in the south of Terra Nova Bay [*Souchez et al.*, 1991] and the other at the Campbell Glacier Tongue in the north. The ice cores collected at these sites were kept frozen at −25°C, transferred

Paper number 95JC00142.
0148-0227/95/95JC-00142$05.00

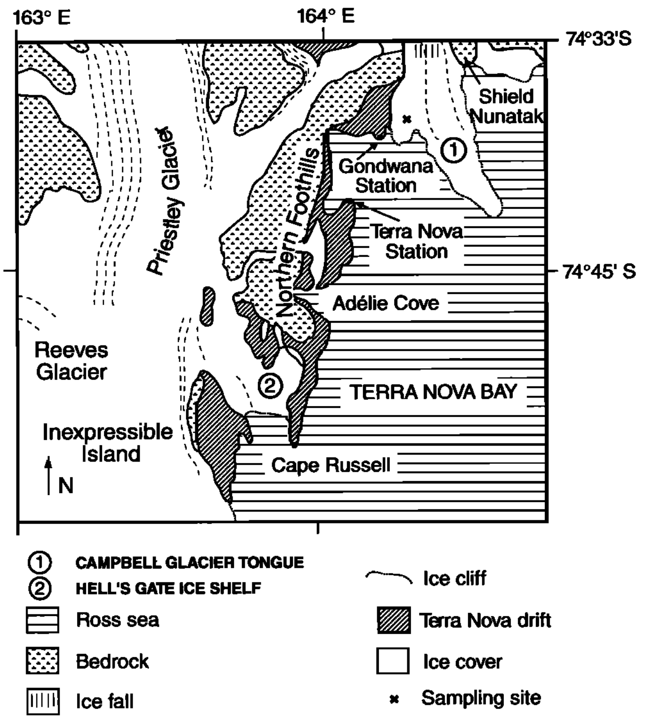

Figure 1. Location map of the Terra Nova Bay area.

to Brussels, and studied for their crystallographic and isotopic properties. The isotopic analyses were done at the Centre d'Etudes Nucléaires de Saclay in France with an accuracy of 0.5‰ for δD and of 0.1‰ for $\delta^{18}O$ on 2-mL samples.

Site Description

The Terra Nova Bay area is located in northern Victoria Land, along the western margin of the Ross Sea. It extends from Cape Washington in the north to the Drygalski Ice Tongue in the south.

Campbell Glacier has its accumulation zone in the Transantarctic Mountains; it flows in a NNW-SSE depression carved into the Precambrian and Paleozoic basement rocks [*Carmignani et al.*, 1987]. Being deviated by Mount Melbourne which is a Cenozoic volcano, it acquires a more or less north-south direction and shows a steeper gradient. The presence of small rock outcrops at this level indicates that the base of the glacier is above sea level. Then, Campbell Glacier reaches the sea, where it terminates as a protruding glacier tongue (Figure 1).

The grounding line, usually defined as a line across the glacier where it first goes afloat, is thought to be located just south of the zone with the steeper gradient [*Frezzoti*, 1993]. Radio echo sounding fails to detect it. The reflector is characterized by a weak energy contrast. This could be due to a change in the physical characteristics of the ice, the electromagnetic energy being absorbed because of the conductivity of a lower ice unit. Precise radio echo sounding is moreover difficult to perform in this crevassed area. Although the precise depth of the grounding line is not known, an approximate value can be given. A bathymetric survey in front of the ice tongue [*Angrisano*, 1989] gives a depth of about 160 m for the sea bottom. Overdeepening under the Campbell Glacier Tongue is a possibility, but the grounding line depth should be around this value. The sampling site is about 3 km downglacier from the presumed position of the grounding line.

In the present paper an extended definition of the grounding line is considered. In such a definition a grounding line is the limit between grounded ice and floating ice, either if the glacier goes afloat or becomes grounded again, as, for example, in the case of a pinning point.

Strong katabatic winds from the polar plateau characterize both the Reeves and Priestley glacial troughs, hence the name Hell's Gate Ice Shelf. As a result, extensive blue ice areas are developed at the surface of these glaciers, and extensive ice-free areas such as the southern part of the northern foothills and Inexpressible Island exist nearby. Because of the action of these katabatic winds, the surface of the Hell's Gate Ice Shelf is snow free and is losing ice, mostly by sublimation. Since top surface ablation is prevalent, an upward vertical velocity component exists and ice initially at depth is transferred along an inclined surface toward the ice-atmosphere interface in the downglacier part of the ice shelf. The marine ice formed at the ice shelf–ocean interface ultimately appears at the surface. The Hell's Gate Ice Shelf has been studied by *Baroni* [1988], *Baroni et al.* [1991], *Souchez et al.* [1991], and *Tison et al.* [1993]. In the Souchez et al. and Tison et al. papers, ice composition studies have shown that marine ice transfer occurs along the bottom of this ice shelf and that the frazil ice accreted at its base reflects contrasting depositional environments.

By contrast, the role of katabatic winds in the Campbell Glacier trough is much reduced. The winds are relatively weak so that the entire glacier is accumulating snow on its surface, and, even at the terminus of the floating tongue, a substantial part of the ice cliff is made of ice derived from snow deposited on the tongue. As a result, there is no upward movement of ice, and therefore marine ice, if present at the bottom, is not likely to outcrop at the terminus of the floating tongue. However, the southwestern part of the ice tongue near Gondwana Station has impinged on bedrock promontories or protuberances. Consequently, basal ice containing debris layers is visible and can be sampled. At the sampling site (Figure 1) a stacked sequence, a few meters thick, dipping 60° toward the center of the glacier tongue shows two distinctive types of ice interbedded with bubbly glacier ice. A first type, located in the upper part of the basal sequence, consists of bands of bubble-free ice with a thickness of the order of a few centimeters. The second type, located in the lower part of the basal sequence, shows bands of thin, clear, ice layers and layers of fine debris sometimes appearing folded with a few occasional pebbles. The proportion of bubbly glacier ice layers is much lower in this case. Debris consists mainly of quartz grains with rounded shapes and smooth edges (60%), volcanic glasses with elongated bubbles (30%), and lithic fragments. A few sponge spicules and shell fragments are also present. The heavy minerals mainly consist of pyroxenes (80%), indicating a major volcanic component and a few olivine, garnet, and epidote minerals. The scanning electron microscope (X ray energy dispersion probe) analysis of the volcanic glass gives a dispersion in a SiO_2 versus alkali diagram similar to that of rocks from the nearby Shield Nunatak complex [*Worner et al.*, 1989], thus suggesting a local origin.

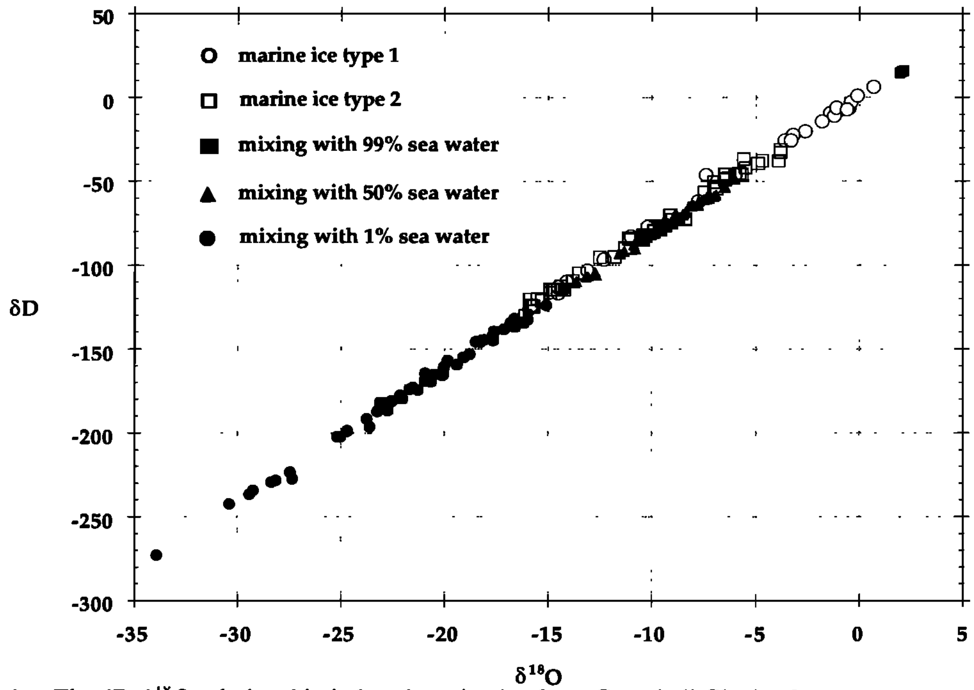

Figure 2. The δD-δ^{18}O relationship in basal marine ice from Campbell Glacier Tongue compared with a mixing computer simulation (see text).

The δD-δ^{18}O Characteristics of Basal Ice

An insight into the problem of the formation of these two ice types can be gained by a combined isotopic study of the ice, both in δD and δ^{18}O. Indeed, if only a single isotopic ratio is considered, the effects of freezing, isotopic exchange with clay minerals, or mixing cannot be distinguished.

Bubbly glacier ice derived from snow of coastal origin and snow fallen in the Terra Nova Bay area have δD values more negative than −130‰ and δ^{18}O values more negative than −17‰. Together with bubbly glacier ice from the continental interior with lower δ values, they are aligned, in a δD-δ^{18}O diagram, on a precipitation line with the equation δD = 7.92 δ^{18}O + 2.76; correlation coefficient r = 0.997 (n = 41 samples).

Basal ice with an isotopic composition less negative than −130‰ in δD and −17‰ in δ^{18}O cannot be considered as unmodified glacier ice. Basal ice samples have been plotted as open symbols on Figure 2. They fit quite well a straight line with equation δD = 7.86 δ^{18}O + 0.29; r = 0.998 (n = 71 samples). The large range of isotopic values reaching values close to those obtained by freezing normal seawater and the slope of the line preclude a simple freezing process. Indeed, if *Jouzel and Souchez* [1982, equation (1)] is used to compute the δ range between 1 and 99% freezing of a closed water reservoir, the range obtained is much less than the observed one. For an open reservoir the range in δ values in the ice would be even smaller. Enrichment by isotopic exchange with clay minerals is able to produce important isotopic shifts but on a much lower slope [*Souchez et al.*, 1990] and is therefore also precluded in this case.

A simulation has been conducted in order to test if this alignment of data of basal ice from the Campbell Glacier Tongue could correspond to a mixing line. The local seawater has been sampled at 30 m depth; its δD value is −3.23‰, and its δ^{18}O value is −0.69‰. In the simulation (solid symbols in Figure 2) this local seawater is mixed in varying proportions with the melt of each glacier ice sample collected in the area. It is assumed that no fractionation occurs during melting of ice. Three mixing ratios have been used as follows: 99% seawater (solid squares), 50% seawater (solid triangles), and 1% seawater (solid circles). The δ values of the ice resulting from freezing of the mixtures have been calculated for deuterium and for oxygen 18, using the equation $\delta_S = \alpha\ (1000 + \delta_w) - 1000$ where α is the equilibrium fractionation coefficient and δ_w is the δ value of the mixed water. The equilibrium fractionation coefficient for deuterium is taken as 1.0208 [*Arnason*, 1969], and for oxygen 18, as 1.003 [*O'Neil*, 1968]. The points representing these various computations have a distribution which is very close to the points representing the samples of basal ice. Therefore the straight line on which the basal ice samples are aligned can be considered as a mixing line and, consequently, the basal ice samples as marine ice. Let us note here that if a 2‰ higher equilibrium fractionation coefficient for deuterium is considered for seawater, a possibility that might exist [*Beck and Munnich*, 1988], the fit would be even better. As discussed below, the presence of a mixing line does not, however, necessarily imply a mixture of continental water and seawater in varying proportions beneath the floating glacier.

Texture-Isotope Relationships in the Marine Ice

Common features to all marine ice samples in this study can be deduced from a careful examination of Figure 3. Clear ice layers always display the less negative δ values, and a gradual change to values typical of glacier ice occurs as one proceeds toward the boundaries of these layers. Very often, transitional δ values occur within a single crystal layer which is bubble free in its inner part and contains bubbles in its

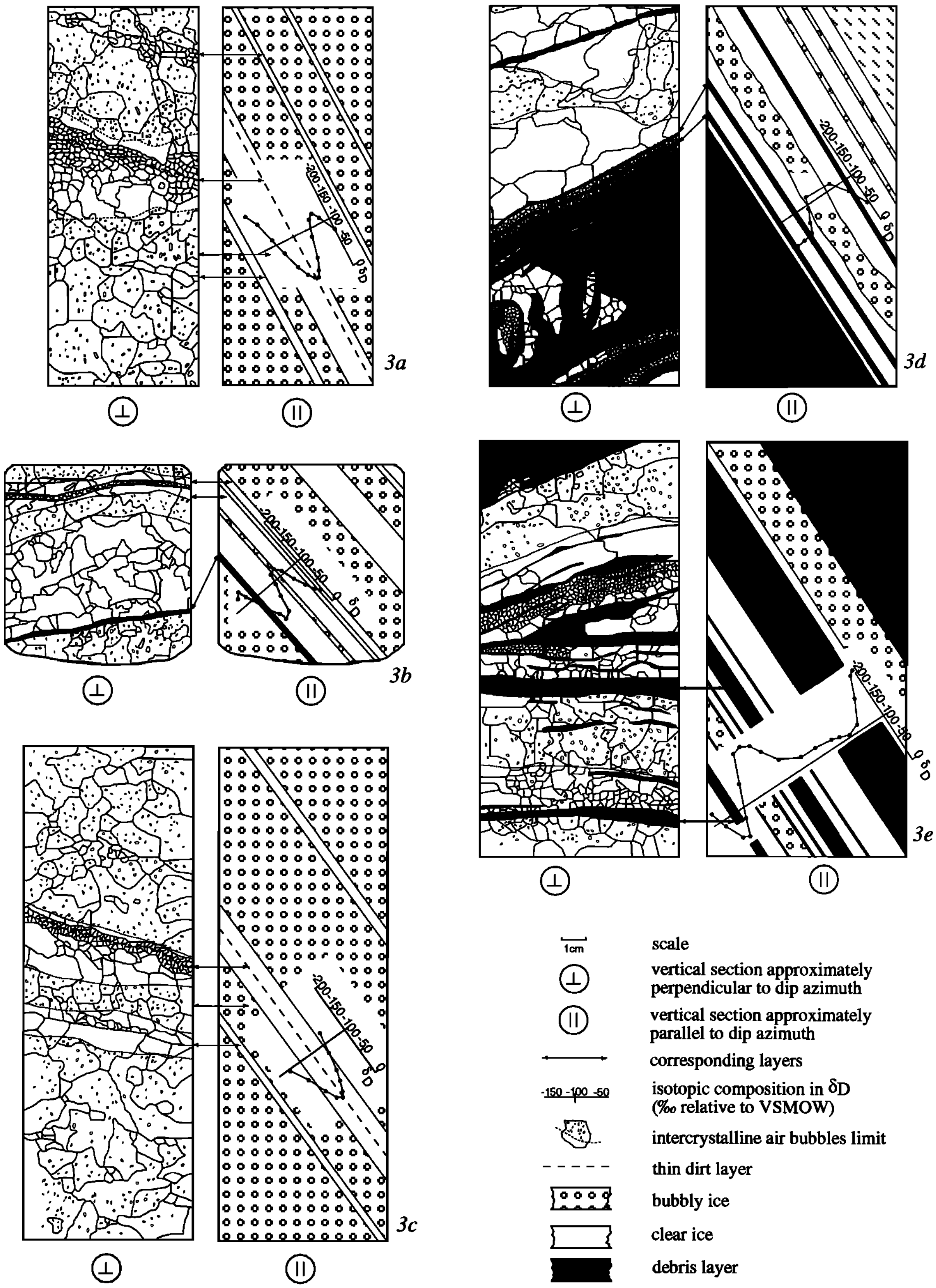

Figure 3. Detailed textural properties and δD profiles in selected ice cores from the (a)–(c) upper part, type 1, and (d)–(e) lower part, type 2, of the basal ice of Campbell Glacier Tongue.

outer part (Figure 3). The crystal size in this layer is similar to the one for unmodified glacier ice. Major differences, however, exist between the marine ice in the upper part of the basal sequence (type 1) and that occurring in the lower part (type 2).

In the upper part the center of the clear ice layers consists of very small crystals (sometimes with a thin line of tiny particles) where the maximum δD values vary between $-30‰$ and $+6‰$, thus approaching values observed in frozen seawater (Figures 3a–3c). They form the group of marine ice samples in the top right corner of Figure 2. As developed below, these characteristics can be understood if intrusion of brackish water has taken place in fissures freshly open at the base of the floating glacier, with subsequent ice formation.

The area at the bottom of a floating glacier or ice shelf close to the grounding line is a favorable site for the formation and opening of basal crevasses. Bottom crevasses have been recorded by *Orheim* [1986] on the Riiser-Larsenisen by radio echo sounding. *Hellmer and Jacobs* [1992] indicate that bottom crevasses are common near grounding lines where tidal bending occurs. They also explain the filling of bottom crevasses with marine ice by an ice pump mechanism driven by the pressure dependence of the freezing point. This ice pump mechanism could readily fill a 200-m-high crevasse with marine ice, a probable reason for the rapid disappearance of bottom crevasses. *Jezek and Bentley* [1983] believe that most bottom crevasse fields are associated with rapid grounding or ungrounding of ice. The crevasses reported by all these authors have dimensions of meters. The type 1 marine ice inclusions studied in this paper are only centimeters wide. Possibly, these fissures could exist in the same situations but remain undetected by radio echo sounding. They might have formed at a larger scale and subsequently been subjected to strain thinning.

Strain thinning could explain the presence of very small crystals in the center of the fissure as a consequence of grain fracturation. However, the δ values in the center of the fissure are so weakly negative that an overwhelming contribution of seawater is required. If present, the grain fracturation must thus have occurred after the formation of marine ice in the fissure.

This raises the possibility that larger crystals might have originally filled the fissure. Such large crystals could have been congelation ice due to ice growth from the sides of the fissure. Migration of a freezing front from the sides of the fissure toward its center as a consequence of the heat sink provided by the cold ice surrounding the crack must thus be investigated. There is no indication, in the large crystals present in the fissure, of the occurrence of brine layers which are a diagnostic feature for congelation ice. On the other hand, geometric selection of crystals during growth in the liquid phase cannot be detected in the thin sections examined. More importantly the isotopic distribution of δ values in the fissure is a strong argument against such a process. Indeed, if a freezing front is progressing into water, the first ice formed (in this case, the ice closest to the sides of the fissure) will have the less negative δ values since the maximum enrichment of heavy isotopes in the ice compared to the initial water is produced there. With the development of freezing the residual water becomes impoverished in heavy isotopes [*Souchez and Jouzel*, 1984]. The successive frozen layers will thus be more and more negative. Such an isotopic distribution is not present; a reverse distribution is displayed with less negative δ values in the center of the fissure and more negative δ values toward the sides.

Frazil ice formation processes yield other ways to produce ice crystals from brackish water [*Weeks and Ackley*, 1986]. Frazil ice can be produced by turbulence, but such a process is not likely to occur here in the confined space of the fissure. Frazil ice can also be produced by a mechanism of double diffusion (heat and salt) involving the existence of two water masses with contrasted temperatures and salinities, both at their pressure melting point. Clearly, this is not possible here. Finally, frazil ice crystals could be formed in adiabatically raising water as a consequence of the pressure dependence of the freezing point. Although we cannot prove that this process is effective here, we consider that the filling of the fissure by frazil ice is the best explanation with the information at hand.

The frazil ice crystals were perhaps originally bigger and their size reduced by grain fracturation. Figure 4a is a sketch illustrating the processes involved. Recrystallization must have occurred as indicated by the presence of ice crystals which are bubble free in their inner part and bubbly in their outer part at the glacier ice–frazil ice contact. Diffusion of isotopes along the pathway represented by water films at the ice crystal limits during recrystallization must certainly be taken into account. It probably explains the gradual decrease in δ values from the center of the fissure toward the sides in this type of marine ice. Clearly, in this case the alignment of the samples on a mixing line does not reflect the mixing of two water masses.

In the lower part of the basal sequence the type 2 marine ice with thin clear ice and debris layers (Figures 3d and 3e) exhibit a range in δD values of $-110‰$ to $-30‰$, indicating a greater influence of continental meltwater. These characteristics can be explained if a water-filled sediment at the glacier-ocean-rock contact is subjected to freezing because of the difference in freezing points between continental meltwater and seawater. In this process, thermal diffusivity being higher than salt diffusion, the water-filled sediment loses heat more rapidly than it gains salt. Moreover, heat can diffuse through both liquid and solid fractions, but salt can only through the liquid. The water within the sediment will have an isotopic composition dependent on the variable contribution of continental meltwater versus seawater. This is reflected in the shift in δ values of type 2 marine ice on the mixing line. Diffusion effects must also be considered for the transitional δ values between adjacent layers.

The presence of these thin clear ice layers between the debris layers with such a specific isotopic signature requires basal freezing. If basal freezing is to occur, the seawater must itself be at the freezing point which is not the case for surface waters reaching the bottom of ice shelves [*Lewis and Perkin*, 1986]. This implies melting and the removal of basal ice in other regions. Although we have no measurements of water temperatures and salinities near the grounding line, the complexity of this contact zone in three dimensions makes the process plausible. In the case of type 2 marine ice at Campbell Glacier, it is clear that it forms where the glacier runs aground farther downstream at shallower depths. Figure 4b is a sketch showing the processes involved. Mechanical incorporation of the ice and debris layers into the ice shelf is certainly a possibility at this location. An investigation of the marine ice properties, like the one conducted in

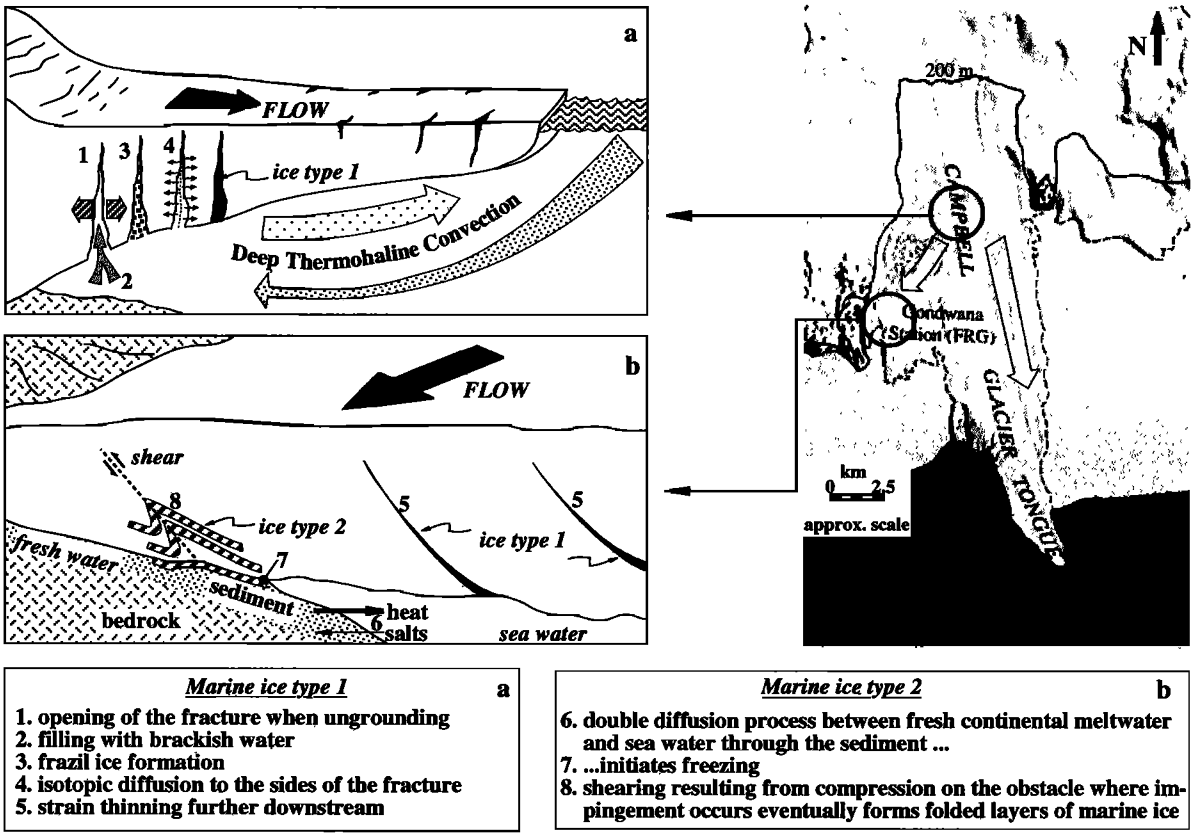

Figure 4. Sketches illustrating the processes involved in the formation of (a) type 1 marine ice and (b) type 2 marine ice. The scale of the features is enlarged for clarity.

this study, is able to give some clues to the formation processes but gives no indication of the transfer mechanisms. As shown in Figures 3d and 3e, small-scale folding can be invoked to explain the multiple-layered structure.

It is not known if the continental meltwater component has been produced at the glacier bed by melting due to geothermal and frictional heat or if it has been able to find its way to the bed from the surface. The geothermal heat flux does not seem to be abnormally high: geothermal anomalies have only been found in the top part of the Mount Melbourne volcanic edifice but not in the lower Campbell Glacier area [*Rossi*, 1991].

As the ice was raised to the surface when it impinged on bedrock promontories or protuberances, the possibility that pressure melting and regelation occurred should be discussed. Although it is not possible to prove that this process has not taken place, the type of debris included in the ice makes it unlikely as the initial accretion mechanism. It could have occurred as a postdepositional process. However, it has been shown earlier [*Souchez et al.*, 1988] that pressure melting and regelation does not significantly modify the isotopic properties of the ice submitted to this process.

The presence of shell fragments and sponge spicules in the debris layers raises the possibility that the debris originated from the Terra Nova drift which mantles the eastern flank of the northern foothills in the coastal region from Cape Russell in the south to the Campbell Glacier Tongue farther north [*Orombelli et al.*, 1990]. Shell fragments and worm tubes of mixed ages are common in Terra Nova drift but not sponge spicules. Near Adelie Cove (see Figure 1), this drift is commonly ice cored and hummocky. The ice cores are made of bubbly glacier ice with $\delta^{18}O$ values reaching −35‰ and bands of bubble-free ice reaching $\delta^{18}O$ values of +2‰. Such highly different values of $\delta^{18}O$ characterize samples collected in a vertical sequence of a few decimeters in the same ice core. The similarity with type 1 marine ice might indicate the presence of a former grounding line at proximity. In this hypothesis the buried ice could have been partially formed near a fossil grounding line of a late Wisconsin Campbell Glacier advancing onto the Ross Sea continental shelf. It is now above sea level because of isostatic rebound. The type 2 marine ice consisting of debris and clear ice layers with its specific isotopic signature has never been found in the ice cores from Terra Nova drift. It is thus quite unlikely that reworking of Terra Nova drift could explain the layered structure of type 2 marine ice at Campbell Glacier Tongue.

With the extended definition of the grounding line in mind, one can thus consider that the two types of marine ice here studied were formed near a grounding line. Type 1 was probably formed closer to the center line of the glacier tongue than type 2. Their superposition (type 1 over type 2) results from impingement on bedrock protuberances. Unfortunately, ice fabrics of the marine ice types do not give additional information on their formation processes. Indeed, on a Schmidt diagram, ice crystal *c* axes form a small girdle indicative of horizontal compression against an obstacle. The ice fabrics thus reflect the impingement on bedrock protuberances, and no remnant of the original fabric is preserved.

Conclusions

A combined isotopic study, both in δD and $\delta^{18}O$, is a powerful tool for understanding the mechanisms involved in marine ice formation. Two types of marine ice accreted at the base of a floating glacier near its grounding line have been distinguished. In both cases the ice samples are aligned in a δD-$\delta^{18}O$ diagram on a mixing line.

Freezing of brackish water in basal open fissures conducive to frazil ice production and consolidation followed by isotopic diffusion along grain boundaries is identified as the most plausible mechanism for formation of type 1 marine ice. Freezing of a water-filled sediment at the ice-ocean-bedrock contact as a consequence of double-diffusion effects is considered as the most likely process for formation of type 2 marine ice.

Acknowledgments. The "Programma Nazionale di Richerche in Antartide" (PNRA) is gratefully acknowledged for having supported the field work of the first author at Terra Nova Bay in 1989–1990. This paper is a contribution to the Belgian Antarctic Programme (Science Policy Office). J.-L. Tison is research associate at the National Fund for Scientific Research (Belgium).

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(Received April 2, 1993; revised July 6, 1994;
accepted December 14, 1994.)