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Isotopic, chemical and crystallographic characteristics of first-year sea ice from Breid Bay (Princess Ragnhild Coast — Antarctica)

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Abstract: A detailed profile of the ice fabric, the deuterium content and the sodium concentration of a 1.64 m long, first-year sea-ice core from Breid Bay, near Syowa station (Antarctica), is described. The core consists mainly of frazil ice (77%), a common feature observed in recent extended studies of the first-year sea-ice cover in the Weddell Sea area. Neither snow ice, nor platelet ice is present. The remainder of the core consists of congelation ice. The typical substructure of ice plates/brine lamellae occurs only at the bottom of the core. Otherwise fine-grained congelation ice is 'sandwiched' between layers of frazil. It lacks the intra-crystalline substructure but shows a strong textural elongation and c-axis clustering in the horizontal plane. The evolution processes of the first-year sea-ice cover in Breid Bay are analysed. The dynamical component, demonstrated to play a major role in the eastern Weddell Sea, seems to be of minor importance in this area, where thermodynamics satisfactorily explains the isotopic, chemical and textural characteristics of the core. It is proposed that the topmost part of the core consists of frazil ice produced by wind- and wave-induced turbulence. Once a consolidated ice cover is provided, the growth proceeds at a slower rate, through congelation ice formation and frazil ice production, initiated by thermohaline convection processes in the water column. The lower alternate layers of fine grained congelation ice and frazil ice could result from cyclic thermal and salinity regimes at the ice-water interface, connected with the major meteorological events of the year.

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Introduction

The growth, structure and properties of sea ice have been permanent subjects of interest in the last decades. Recent work has provided a good overview of the topic (Weeks & Ackley 1986) and made field observations from the Arctic and the Antarctic more widely known (Gow *et al.* 1987*a, b*, Lange 1988, Lange *et al.* 1989). Results from extensive investigations of sea ice properties during the Winter Weddell Sea Project 1986, in the central deep ocean and eastern coastal regimes of the Weddell Sea (Antarctica) stressed the importance of dynamic against thermodynamic processes in forming the first-year sea-ice cover. Lange *et al.* (1989) suggested that most of the sea-ice growth occurs at the 'open water' growth rate during pancake-ice formation, thereby substantially reducing the contribution of other mechanisms previously invoked (Weeks & Ackley 1986, Gow *et al.* 1987*a*).

Deuterium profiles in sea-ice cores are rarely described in the literature. Friedman *et al.* (1961) explained the negative δD values (–2 to –14‰) of ice samples collected on Ice Island T3 and on Drift Station Alpha by the presence of a relatively thin layer of deuterium-depleted water on parts of the surface of the Arctic Ocean during the summer. More recently, Gow & Epstein (1972) reported on deuterium concentrations connected with a special sea-ice growth situation in McMurdo Sound, Antarctica.

Profiles of total salinity are more widespread in the literature. In some cases they are related to ice thickness measurements and used to evaluate growth rates of the sea-ice cover (Weeks & Lofgren 1967, Cox & Weeks 1975, Nakawo & Sinha 1981).

In this paper, we study a first-year sea-ice core from Breid Bay (East Antarctica), using the crystalline structure (texture and c-axis fabrics) and the combined deuterium and sodium profiles to distinguish the different types of ice and to analyse the processes of their formation.

Location and analytical techniques

Four sea-ice cores were sampled at Breid Bay, in December 1986, as part of a joint Belgian–Japanese project during the Japanese JARE 28 Antarctic programme. Breid Bay, close to the Japanese station, Syowa, in East Antarctica, was totally ice-free during the previous summer so that the ice cores retrieved are undoubtedly first-year sea ice. The sampling site (70°13'S, 23°47'E) was located in a small bay approximately 50 km offshore from the ice shelf edge (Fig.

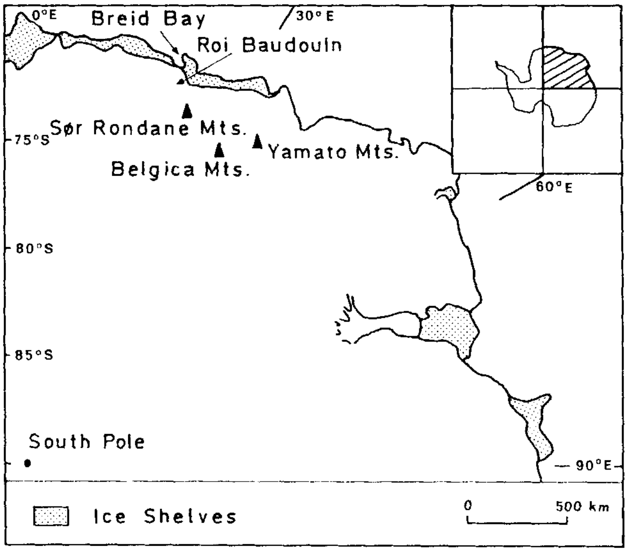

Fig. 1. Location of Breid Bay and Roi Baudouin station in East Antarctica.

1). The sea-ice cover was perfectly smooth, lacking the small-scale roughness, typical of rafting processes in early growth history, or any sign of deformation due to pressure ridging (H. Decleir, personal communication 1988; Fig. 2). This situation seems to be quite common in the area and has already been reported in the past (Wakatsuchi 1977). There was a snow cover of about 40 cm on the sea ice at the period of sampling but it can be considered as a late addition.

The cores were sampled with a SIPRE ice auger, transferred into plastic bags and kept in cylindrical insulated containers at –25°C during transport aboard RS *Shirase*, and temporarily (approximately 6 hours) at –15°C during transport to Belgium where they were stored at –28°C before analysis. As the cores were stored horizontally, and only briefly between –25°C and –15°C, this should have prevented any alteration of their vertical chemical profile (Cox & Weeks 1986). Each core was broken in several places during sampling. As already underlined by Gow *et al.* (1987*a, b*), these natural breaks appeared to coincide with structural discontinuities; in most cases the transition from congelation ice to frazil ice in the core. The four cores were similar in length (between 164.5 and 167.5 cm) and structure, so that only one of them will be described here in full detail.

Fig. 2. The first-year sea-ice cover in Breid Bay near the sampling area.

The following analytical techniques were performed on the cores in the cold laboratory (–25°C).

Crystalline texture

Mean crystal diameter and mean crystal elongation were measured on vertical thin sections of the cores at regularly spaced intervals and in a few selected horizontal thin sections. The crystal diameter is defined as $d = \sqrt{(Lw)}$ where L is the maximum dimension of the crystal and w the value measured at right angles to L (Weeks & Ackley 1986). The crystal elongation is the L/w ratio. In all cases, measurements were made under a 16× magnification binocular microscope and taken on at least 40 individual crystals to ensure a statistically valid mean value.

Ice fabric

C-axis (optic axes) orientations were measured in horizontal thin-sections at regular intervals of approximately 5 cm and in a few vertical thin sections, using a four-axis universal stage.

Sodium profile

One half of the core was cut into 1 cm-thick slices along the whole profile, using standard methods to reduce contamination to the noise level of the Perkin-Elmer atomic absorption spectrophotometer. The Na and Sr content was measured in each sample. Since these two ions behaved in exactly the same way, only the sodium profile will be discussed here.

Deuterium profile

The other half of the core was sampled in selected spots for δD analysis. The small amounts of ice necessary for the mass-spectrometer measurements (0.3 ml) were collected using a microtome. This technique allowed precise location of the samples and a detailed profile to be obtained where required. The mass-spectrometer analyses were performed at the 'Laboratoire de Geochimie Isotopique, Centre d'Etudes Nucléaires de Saclay, France'.

HDO concentrations will be denoted hereafter in units with respect to VSMOW (Vienna Standard Mean Ocean Water) expressed per mil i.e.:

$$\delta D = 1000 \times \frac{R_{sample} - R_{SMOW}}{R_{SMOW}}$$

where R_{sample} and R_{SMOW} are the deuterium/hydrogen ratios of the sample and of the SMOW respectively. The accuracy of the δD measurements is ± 0.5‰.

Results and discussion

Sea-ice types in the core

A comprehensive classification of sea-ice types has been developed by Weeks & Ackley (1986), Gow *et al.* (1987*a, b*) and Lange *et al.* (1989), among others. The latter authors stressed the importance of differentiating between textural and genetic terms in qualifying the various existing ice types.

Fig. 3 gives the stratigraphy of one of the cores from Breid Bay. A major part of the profile consists of granular sea ice, with a mean crystal diameter ranging from 2.8 mm to 1 mm, and showing a general decrease with depth. In this textural unit, the mean crystal elongation fluctuates between 1.5 and 2.5 and indicates a slight lengthening in the vertical plane (Fig. 4a). This probably reflects the effect of recrystallization when individual frazil or soaked snow crystals freeze together in a vertical thermal gradient. Indeed, when measured in the horizontal plane at different heights in the core, the mean crystal elongation is very close to 1 (equi-dimensional grains, Fig. 4b). Another reason for this lengthening in the vertical plane could be the tendency for the individual discoids of frazil ice nuclei to be stacked with their long axes vertical by the movement of slush pancakes against one another, as suggested by Weeks & Ackley (1986). However, this should lead to at least a partial clustering of c-axes in the horizontal plane, which is not observed (thin-sections 1 and 2 in Fig. 3).

The ice of the upper 32 cm of the core differs from the lower granular sea ice mainly because it has a higher bubble content, clearly visible in transmitted light. These closely

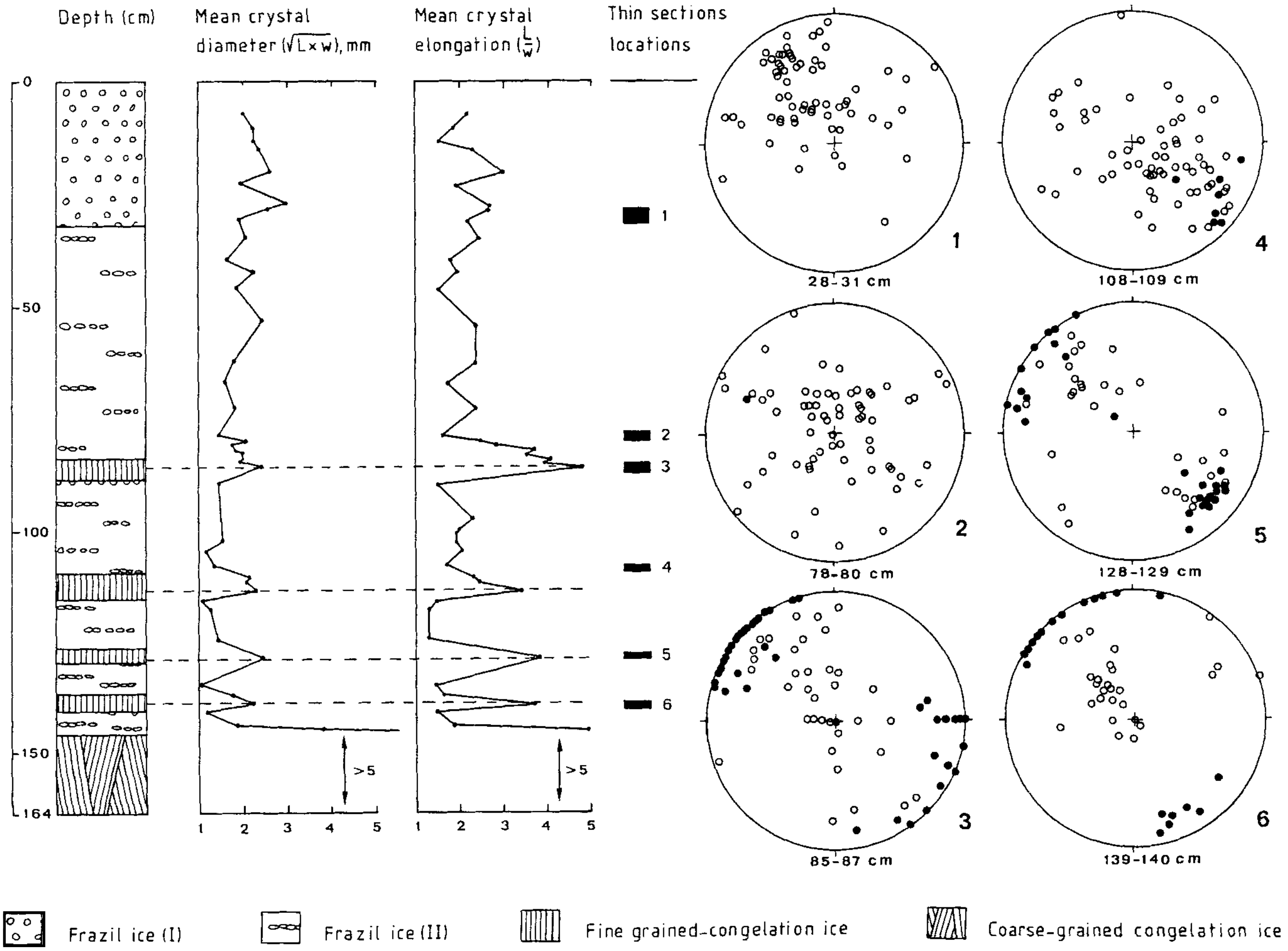

Fig. 3. The stratigraphy of a first-year sea-ice core from Breid Bay. The mean crystal diameter and the mean crystal elongation are measured in the vertical plane. The fabric diagrams are plotted in the horizontal plane; O = equi-granular crystals and ● = elongated crystals (partly from Souchez *et al.* 1988).

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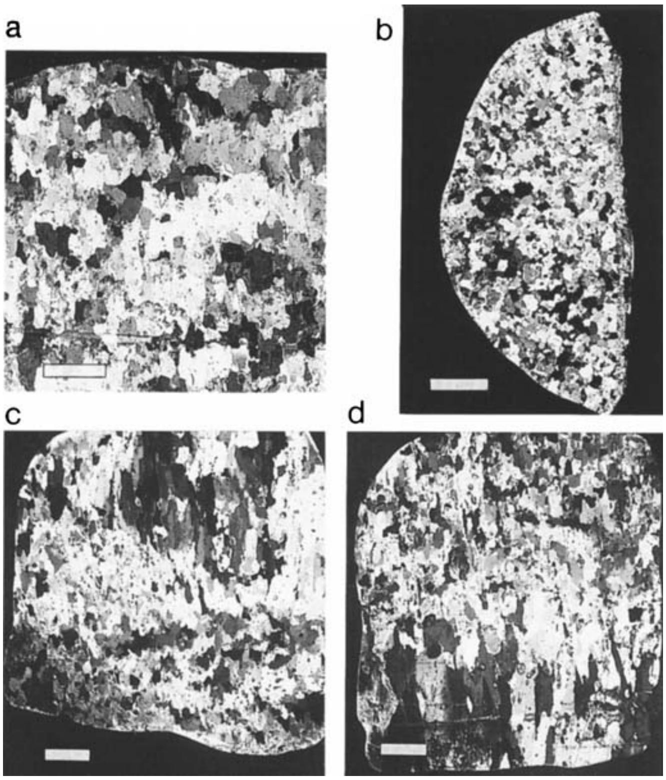

Fig. 4. Thin section photographs from Breid Bay ice core; scale bars 1 cm. **a.** Frazil ice in vertical section (32–36 cm). Note the slight elongation in the vertical plane and the W-shaped crystals. **b.** Frazil ice in horizontal section (79 cm). Note the equi-granular structure and the smaller mean crystal diameter. **c.** Transition zone between fine-grained congelation ice and frazil ice in vertical section (85–90 cm). **d.** Transition zone between frazil ice and coarse-grained congelation ice in vertical section (144–149 cm).

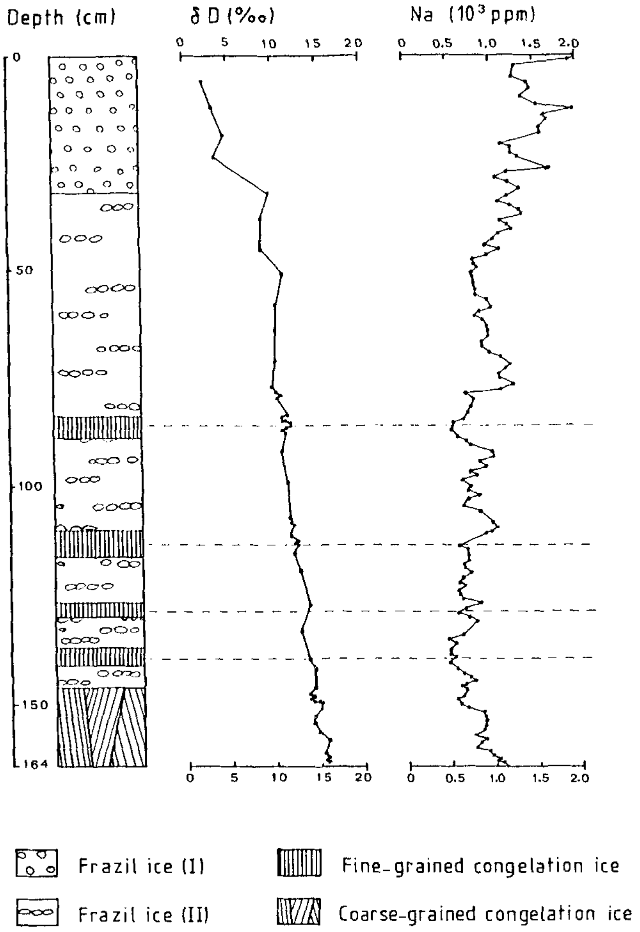

Fig. 5. The deuterium and sodium profiles in the Breid Bay core (partly from Souchez *et al.* 1988).

spaced spherical bubbles, with diameters of between 0.1 and 1.5 mm, do not show any preferential spatial distribution. The higher bubble content and the position of this 32 cm layer at the top of the core might be the distinctive characteristics of snow ice (Gow *et al.* 1987*a*). However, the isotopic profile makes this proposition highly improbable. Fig. 5 gives the detailed deuterium and sodium profile in the Breid Bay core. The δD values in the top 32 cm of the profile range between 2.3‰ and 4.5‰. The deuterium concentration for snow in the area is about –160‰. It is thus impossible to form any kind of infiltration ice with a positive deuterium value, whatever the interstitial water available in the environment. The whole unit of granular sea ice in the core can therefore be considered genetically as frazil ice.

The lower 18 cm section of the core shows a marked increase in the mean crystal diameter and elongation, corresponding to the typical columnar-shaped crystals. Fig. 4d clearly illustrates the transition zone between the lower granular sea ice layer and this columnar sea ice. Crystals several cm in length are present, and the substructure of ice plates and brine lamellae is commonly present in this bottom layer. The c-axes are invariably located in the horizontal plane, a signature of congelation ice.

At four different levels in the profile, the mean crystal elongation shows a marked two- to three-fold increase. There is a corresponding moderate increase in the mean crystal diameter, which nevertheless stays much less than the values observed in the congelation ice at the bottom of the core. Fig. 4c illustrates the transition zone between one of these peculiar textural units (above) and granular ice (below) between 85 and 92 cm depth. It clearly shows that, at a given level, a mixture of both ice types occurs. This fact, combined with the observation that the elongated crystals do not display the typical ice plate/brine-lamella substructures of coarse-grained columnar sea ice, could be the signature of platelet ice. However, in this case, the crystals are strictly elongated vertically and they show a strong bias of their c-axes to the horizontal plane (black dots in thin sections 3, 5 and 6 in Fig. 3). We are thus dealing with a fine-grained columnar ice showing the genetic characteristics of congelation ice. Similar units of fine-grained congelation ice have been observed in Weddell Sea floes by Gow *et al.* (1987*a*,

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e.g. fig. 10).

Thus, from the crystalline structure data summarized in Figs 3 and 4 and from the deuterium profile in Fig. 5, we can establish the vertical section drawn on the left side of Figs 3 and 5: the core consists mainly of frazil ice (77%), the topmost part (32 cm) being more bubbly. Another 11% consists of coarse-grained congelation ice located at the base of the core, and the remaining 12% consists of fine-grained congelation ice layers dispersed in the lower part of the frazil ice. This 'sandwiching' of frazil between layers of congelation ice has also been found in Weddell Sea floes by Gow *et al.* (1987*a*) and Lange *et al.* (1989).

A proportion of the 77% frazil vs. 23% congelation ice corresponds to the 'mainly frazil' to 'predominantly frazil' ice classes defined by Lange *et al.* (1989). More than 60% of frazil has been observed in 48% of the cores analysed on leg 1 of WWSP 86 by these authors. Amongst the 71 cores examined for quantitative textural composition by Gow *et al.* (1987*a*), such high proportions of frazil ice are more typical of multi-year sea-ice floes. Nevertheless, 17% of the first-year sea-ice floes considered contained more than 70% of frazil ice. Thus we consider that the structure of the four cores from Breid Bay is in good agreement with the many results obtained in the Weddell Sea area. We did not, however, detect the existence of platelet ice reported by Lange *et al.* (1989) at the bottom of 30% of their cores, obtained from the coastal area of the eastern Weddell Sea. This could be explained by the fact that the ice shelf near Breid Bay is restricted to a very narrow fringe. Melting of ice at its bottom surface is thus probably more restricted than below large ice shelves like Filchner and Ross and the supercooling of water due to this process is insufficient to produce ice platelets in significant amounts.

Processes of evolution of the sea-ice cover

The total length of the core described in Figs 3 and 5 is 164.5 cm. Such a value has been observed for first-year sea ice in 22% of the 58 profiles described by Gow *et al.* (1987*a*). However, Wadhams *et al.* (1987) show that most of the first-year sea-ice thicknesses measured during the WWSP 86 project fell in a narrow range of 40 to 60 cm. Mean ice thicknesses of more than 70 cm were always encountered under deformed ridged areas. Furthermore, Lange *et al.* (1989) stressed the major role of the 'pancake cycle' in the building of the first-year sea-ice cover in the Weddell Sea. In this process, consolidated pancake ice close to the ice edge aggregates to form a continuous ice cover, up to some tens of centimetres in thickness. Wind and wave action give rise to divergence and lead to rafting of parts of the ice cover. When a mean thickness of 50 to 60 cm is reached, the wave energy is damped out by the existing ice cover. Further consolidation of the sea-ice cover significantly reduces the growth rate to a mean value estimated at 0.4 cm day^{-1}, for subsequent congelation ice formation (Wadhams *et al.* 1987). In the eastern coastal area of the Weddell Sea, characterized by the frequent formation and disappearance of polynyas, grease and pancake ice or congelation ice form depending on the wind and wave activity at the time. Changes from offshore winds to onshore winds lead to the closing of the polynyas and usually cause rafting and ridging of the ice. The whole process gives floes of mixed frazil/congelation ice between 20 and 80 cm thick. In addition platelet ice accretes at the base of the ice cover, representing 10 to 50% of the total ice thickness (Lange *et al.* 1989).

Souchez *et al.* (1988) used the isotopic profile through the congelation ice layers in the Breid Bay core (Fig. 5) to determine the growth rates at their respective levels. The values were compared with predictions from a box diffusion model in which the initial sea-water δD value is held constant in the course of freezing, outside a boundary layer developed close to the ice-water interface. The modelled growth rates were in good agreement with those predicted by incorporating climatological data collected at the Antarctic station Roi Baudouin into a simple thermodynamical model, similar to the one used by Nakawo & Sinha (1981). These results suggest that the evolution of the sea-ice cover in the Breid Bay area was controlled by thermodynamic rather than dynamic processes for at least the lower half of the ice cover, a situation different from the one demonstrated by Wadhams *et al.* (1987) and Lange *et al.* (1989) in the Weddell Sea.

Besides the good fit of the calculated growth rates to the climatological data, there are several indications that preclude any significant role of the dynamics in the formation of the ice cover studied in Breid Bay.

a. There was no morphological evidence of rafting or deformation in the sampling area; the sea-ice surface was perfectly smooth with no stony field ice or visible ridges.
b. The texture of the cores shows no sign of deformation or rafting. The limits between the textural units are parallel to each other, sub-horizontal, with clear contacts.
c. No voids or mylonitization zones were observed (see Lange *et al.* 1989, fig. 5).
d. The alternation of frazil and congelation ice, that could result from rafting in the coastal area, only occurs between 84 and 164 cm depth. Rafting of floes of more than a few tens of centimetre thickness is difficult to understand (see fig. 4 in Lange *et al.* 1989) and it would have to occur several times in the same year if dynamics alone were responsible for the recurrence of the congelation ice observed in the core.
e. The repeated opening of polynyas necessary to achieve the dynamical build-up of the ice cover has not been reported in the Breid Bay area (Wakatsuchi 1977).
f. The deuterium profile of the core (Fig. 5) does not show any significant discordance that should result from the difference between the isotopic signal of ice formed in

'open water' conditions and the isotopic signal of the ice, formed at depth, that overrides it.

Thus, in this case, it appears that dynamical effects must be considerably reduced and that the growth of a significant part of the sea-ice cover is thermodynamically controlled.

A closer examination of the deuterium and sodium profiles in the core (Fig. 5) sheds some light on the processes involved. Considering the deuterium value of the sea-water at Breid Bay (–4.8‰) and the Na concentration (10 300 ppm) derived from the measured salinity of sea-water (33 600 ppm), the general trend of the two curves indicates a decreasing freezing rate with depth due to the damping effect of the increasing thickness of the ice sheet on the heat transfer. Indeed, the fractionation effect occurring during phase change from water to ice enriches the ice in heavy isotopes and depletes it in salts. The slower the freezing rate is, the more efficient the fractionation will be (i.e. closer to the equilibrium value). There is, however, an exception in the lower 14 cm of the core where the generally decreasing sodium profile tends to rise again. This c-shaped curve is common for salinity profiles in sea ice. Nakawo & Sinha (1981) explained this trend by the fact that the freshly frozen sea ice at the bottom of the core has not yet undergone the rapid desalinization which often occurs in the first few weeks after initial freezing. It should be noted that this effect of the brine volume does not affect significantly the deuterium signal which mainly reflects the characteristics of the ice crystals when they were formed. This is confirmed by the following simple calculation based on the deuterium and sodium values observed in the first congelation ice layer of the profile (86 cm level). Using the total salinity of the layer in the equations developed by Cox & Weeks (1983) the relative brine volume can be estimated as 4.3% at –2°C. If we consider further that the measured δD value of the layer (+11.8‰) is the result of the mixing with the δD value of sea water in the brine (–4.8‰), we obtain a δD value for the ice itself of +12.5‰. This implies an error of 0.70‰ on the δD value, which is very close to the accuracy of the measurement (± 0.5‰). A similar calculation for other ice layers in the core gives similar results.

As mentioned above, the bubble-rich sea ice forming the top 32 cm of the profile has to be considered as frazil ice on isotopic grounds. A marked shift of about 6‰ in deuterium occurs at the level of the discontinuity between bubble-rich and bubble-poor frazil ice that might be explained by the existence of a small amount of melted snow or shelf ice mixed with open-sea water during the summer season. In that case, the first frazil ice formed would show a slightly depleted δD value compared to the frazil formed later in the season. However, as this input of fresh water is likely to be very limited (negative mean daily air temperature all year round and narrow ice shelf), the observed shift could be the tracer of the transition from grease or pancake ice to a consolidated sea-ice cover. Indeed, as underlined by Wadhams *et al.* 1987, as soon as the open water surface is cut off by the cementation of the frazil into a continuous cover, the growth rate drops radically to a value thermodynamically appropriate to the thickness of the ice cover. This will correspond to a substantial positive shift in the deuterium values of the sea ice newly formed. Furthermore, frazil ice generated by wind- and wave-induced turbulence is very likely to display a higher density of bubbles as observed.

Three main mechanisms could be proposed to explain the substantial amount of frazil occurring below 32 cm in our core (Weeks & Ackley 1986, Lange *et al.* 1989).

Water with a depressed freezing point due to pressure at depth. This could produce frazil ice, if suddenly brought up to the surface in adiabatic conditions. These water masses exist beneath large ice shelves, and the ascent occurs near the front. However, no correlation has been established between higher frazil percentages and the proximity of the shelf border for ice shelves such as the Filchner–Ronne, which discharges into the Weddell Sea (Gow *et al.* 1987*a*). Results from the WWSP 86 led to the same conclusions, the 'predominantly frazil' ice class (more than 80%) being mostly observed away from the coast (Lange *et al.* 1989). Moreover, the platelet ice, described by these authors in the eastern coastal area of the Weddell Sea and related to the processes occurring beneath large ice shelves, was not observed in our area, probably because the shelf is restricted to a narrow coastal fringe. Finally, it is difficult to explain the slow but regular δD-enrichment of frazil ice with depth by a mechanism that does not imply heat transfer through the pre-existing sea-ice cover.

Contact between two water masses of significantly different salinities, but both at their freezing point. This process requires an input of lighter fresh water, on top of the sea-water either by river or glacier drainage into the ocean, or because of a significant amount of surface melting. Neither situation is likely to occur in the area studied.

Thermohaline convection related to surface freezing. Freezing at the bottom of the sea-ice cover results in the formation of cold brine plumes. These denser plumes gain heat and lose salts as they descend in the water column. However, the diffusion rate for salt loss is much lower than the diffusion rate for heat gain. Hence, the adjacent waters are cooled to temperatures below the freezing point at their slowly-changing salinity. Frazil ice crystals are then nucleated at the interface between the brine plumes and the adjacent waters, and subsequently float upwards. This process is the most likely to have occurred in the present case, as it allows congelation ice to be sandwiched between frazil layers which may be formed in considerable numbers and at considerable depth.

It is striking how, in Fig. 5, the δD values of frazil ice fit well the general trend of increasing δD with depth observed in the successive congelation layers. As the mechanism of frazil generation through thermohaline convection does not imply the migration of a well defined freezing front, the model of Souchez *et al.* (1988) cannot be applied *sensu-stricto*. However, in this process the primary cold source is still the heat sink through the sea-ice cover, and the frazil crystals nucleate at the interface between the brine plumes and the adjacent water, the heat transfer being perpendicular to this interface. A quasi steady-state value of the isotopic signal, dependent on the freezing rate, could thus be reached in the frazil ice crystal in a way similar to that observed in congelation ice. This implies that each individual frazil crystal is thick enough for the isotopic signal to reach its quasi steady-state value at a given freezing rate (Souchez *et al.* 1987). There is some indication that this might be the case in the core. Indeed, any influence of the initial transient on the isotopic signal response would imply a higher δD value in frazil ice than in congelation ice at a given depth, which is not the case in the profile.

Small-scale fluctuations in the lower part of the two profiles in Fig. 5 can be related to the different types of ice. The fine-grained congelation ice layers occurring at 84–89 cm, 110–116 cm, 126–129 cm and 137–141 cm in the core show a higher δD value and a lower salinity than the surrounding frazil ice. If we assume that the relation between the isotopic signal and the growth rate is also valid for the frazil ice layers, then the freezing rate slightly increases at the beginning of each new frazil stage. It is tempting, in this case, to correlate the frazil and congelation ice episodes in the core with the few cycles of very cold periods (down to –40°C) and of milder temperatures (–10°C to –15°C) that characterize the temperature profiles for the second half of the year in the Breid Bay area (Souchez *et al.* 1988). Congelation ice would then form at the end of a 'cold' cycle, at a slower freezing rate; and the efficient salt rejection in the water at the interface, together with the heat loss connected with the new cold wave, would then favour the frazil ice generation by thermohaline convection at a slightly higher growth rate.

Conclusion

A comparative analysis of the isotopic, chemical and crystallographic properties of sea ice allows a better understanding of the processes at work. Textural parameters like the mean crystal diameter and the mean crystal elongation, together with the fabric diagrams, characterize granular ice or columnar ice and enable us to detect fine-grained congelation ice which lacks the typical ice plate/brine lamella sub-structure. The isotopic and chemical profiles help us to discriminate between snow ice and frazil ice, to estimate the growth rates in congelation ice, and to choose between the different mechanisms of frazil ice generation involved. The discrepancy between the δD and the Na profiles at the bottom of the core confirms that salts are much more sensitive to brine migration and drainage. Isotopic determination of the growth rates is thus likely to be more accurate than a determination based on the salinity profile, as most of the stable isotope signal arises from isotopes in the crystal lattice itself.

The relative proportions of frazil and congelation ice in the cores from Breid Bay are in good agreement with previous observations in the Weddell Sea (Gow *et al.* 1987*a*, Lange *et al.* 1989). However, the evolution processes of the sea-ice cover seem to differ significantly from those demonstrated by Wadhams *et al.* (1987) and Lange *et al.* (1989). Morphological, textural, mechanical and isotopic evidence does not support this predominance of dynamical over thermodynamical processes in the Breid Bay cores. No platelet ice was observed, probably due to the limited extension of the ice shelf in the sampling area, and the accretion of the lower half of the core is satisfactorily explained in terms of thermodynamics.

It is suggested that the alternative layers of frazil ice formed by thermohaline convection and of fine-grained congelation ice in the core reflect the influence of the major meteorological events of the year on the thermal and salinity profiles at the ice–water interface. However this hypothesis clearly needs to be exposed to further critical tests. A detailed, synchronous survey of the meteorological data at the sampling site needs to be undertaken for future cores.

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