BANDING IN MCMURDO FAST ICE

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ABSTRACT
We report results from field investigation of banded features with differing optical scattering density occurring in columnar-congelation first-year landfast sea ice samples from the Eastern side of McMurdo Sound. To decipher the origin of these features, an analysis of the bands’ spacing is combined with high-resolution measurements of the chlorinity (as a proxy to salinity), total gas content and gas composition (O₂, CO₂, N₂) in the white and dark portions of the bands at various levels in the core.

It is shown that the banding corresponds to a 24 hours cycle throughout the core. It cannot therefore be attributed to surface temperature-driven growth rate fluctuations, since those are unlikely to show a daily cycle, or to sustain significant contrasts below 0.5 m depth. Total gas content and chlorinity show sympathetic fluctuations between the dark and the white layers forming the banding, therefore precluding a biological origin for the banding cycle. Boundary layer fluctuations driven by tidal currents emerge as the best candidate to explain the observed features.

INTRODUCTION
In many parts of the Earth’s polar regions, columnar sea ice shows the formation of bands, on the scale of up to tens of mm, with differing optical scattering density. The bands are formed during “direct” freezing (congelation) at the ice-water interface (as opposed to “frazil” or “platelet” ice crystals formation), but their exact nature and their formation processes are as yet unclear. Banding results from the regular alternation of darker and whiter ice layers, that are obvious when observing “half cores” (Figure 1, left) or thick sections in transmitted light (Figure 1, middle). In the latter case, however, the contrast is inverted. The layers which are more strongly scattering, appearing white in reflection but dark in transmission, will be described below as the “white layers”, and we will refer to a “band” as a given dark-white layer association.

A common intuitive interpretation of this banding process from the sea ice research community is that it results from growth rates fluctuations and associated gas phase

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Figure 1: Structural features of core McM99 4C shown on two half cores (left), thick sections (middle) and thin sections viewed between crossed polarizers (right). Bottom right shows the reproducibility of features between cores McM99 4C and McM99 4D.
changes at the interface, the whiter layers reflecting a higher bubble content. However, little attention has been paid until now to the detailed processes involved. It is the aim of this paper to provide a geometrical, thermodynamical and physico-chemical data set to get a better insight into those processes.

**SAMPLE LOCATION AND ANALYTICAL PROCEDURE**

Four landfast first year sea ice cores have been collected in the framework of a study on biogeochemical properties of sea ice at Arrival Heights, on the Eastern side of McMurdo Sound (Figure 2). A fuller description of the ice at that location and season is given by Trodahl et al. (2001).

The cores were shipped back to Brussels in a reefer container equipped with a data-logger showing that the temperature was kept below –28 °C at all times, therefore preventing post-sampling relocation of impurities. Two of those cores have been sliced in two halves and thick and thin sections performed following the standard microtoming procedure of Langway (1958), in order to describe the structural and textural properties of the sea ice cover.

To achieve the high resolution imposed by the scale of the banded features, chlorinity has been measured as a proxy to the bulk salinity. Cl⁻ measurements were performed using a Dionex DX100 ion Chromatograph, a technique that only requires 0.5 ml of sample volume. Precision of the measurements was ±5 %.

Total gas volume (content in ml STP of gas per kg of ice) was measured using a melting-refreezing method coupled with a Toepler pump extraction (Blunier et al., 1993; Raynaud et al., 1988). The relative standard error was ±2–5 % (Martinerie et al., 1994). Gas composition was measured using the dry extraction technique (crushing at –55 °C in the cold room) described for glacial ice by Raynaud et al. (1982) and Barnola et al. (1983). Since the samples are crushed at very low temperature (–55 °C) and under vacuum (10⁻² Torr), the technique measures the gas composition both from gas bubbles and gas dissolved in the brines of the sea ice. The extraction line is used in conjunction with a Varian 3300 gas chromatograph. Oxygen and nitrogen were analyzed using a molecular sieve column and a thermal conductivity detector. Carbon dioxide was analyzed using an Haysep column and flame ionization detector following catalytic transformation into methane. Different standard mixtures of gases (Praxair) were used for calibration. Given the small amounts of samples and low total gas contents, the precision of the measurements was ±5 % of the measured CO₂ concentrations and ±0.8 % for O₂ and N₂.

Because of the relatively large amounts of ice necessary for the gas measurements (about 40 g for total gas content and 80 g for gas composition), we had to use two
different cores (McM99 4C and McM99 4D). These were collected about 25 cm from each other, and the excellent reproducibility of features occurring at identical depths (Fig. 1, bottom right corner) ensured that measurements were fully compatible between the two cores. Core McM99 4C was used for Cl⁻ and total gas content and core McM99 4D for O₂, N₂ and CO₂. However, even so and especially in the case of gas composition measurements, some (most) of the layers were too thin to achieve a reliable single layer measurement. Two subsequent layers of the same type were therefore analyzed simultaneously.

**BAND SPACING AND TIME CONSTRAINTS**

The regular spacing of the bands suggests the existence of some growth cycle. To check for this assumption, we will use a simple thermodynamical model in which the growth rate of the ice is related to the rate at which heat is extracted to the atmosphere:

\[
L \rho \frac{dz_0}{dt} = \kappa \frac{\partial \theta}{\partial z}
\]

where \(L \rho\) is the latent heat of freezing per unit volume, \(\kappa\) is the thermal conductivity of the sea ice cover and \(t\) is the time (Trodahl et al., 2000). The temperature of the ice at depth \(z\) below the surface is \(\theta\) and \(z_0\) is the thickness of the ice. The temperature gradient \(\partial \theta / \partial z\) can be approximated by the average temperature gradient through the ice \([\theta(0) - \theta(z_0)] / z_0\), in which case (1) integrates to:

\[
z_0^2 = \left\{ \frac{2 \kappa \theta(0)}{L \rho} \right\} t,
\]

where \((2\kappa / L \rho)\) is \(1.4 \times 10^{-6} \text{ m}^2\text{sec}^{-1}\text{K}^{-1}\). Thus the square of the ice depth should grow approximately linearly with time, showing only variations associated with the surface temperature.

Equation (1) relates the growth rate to the instantaneous temperature gradient at the base of the ice. It is however important to note that rapid variations in the surface temperature do not penetrate far into the ice. Indeed, only variations with periods \(P_0\) greater than (Trodahl et al., 2000):

\[
P_0 = \left( \frac{400 \text{ hr/m}^2}{} \right) z_0^2
\]

remain within a factor of three of their amplitude at the surface. To a very good approximation, the instantaneous surface temperature in equation (2) must be replaced with its average over the past \(P_0\) hours. For 0.5 m the average is four days and at 1 m a 16-day average is appropriate.

Five series of bands were selected in core McM99 4C, in the depth range 0.31 to 1.08 meters (Fig. 1), where they were the most obvious. In each series, the depth of occurrence of the band (using the same reference point in each band) was measured, squared, and plotted versus time in Figure 3, assuming a period of one day. Linear regression is very good in all cases \((r^2 \text{ always} > 0.98)\). The slope of the regression line, assuming a one-day period, is between 0.0133 and 0.0216 m² day⁻¹, which corresponds to a surface temperature between −12.5 and −20 °C in equation (2). These surface temperatures are just what is measured during the time of year when the ice thickness is nearing 1 meter (Trodahl et al., 2000). The assumed periodicity of the growth-driven
bANDING PERIOD INFLUENCES THE ESTIMATED SURFACE TEMPERATURE SIGNIFICANTLY; A 12-HOUR PERIOD WOULD LEAD TO ENTIRELY UNREASONABLE SURFACE TEMPERATURES OF –25 TO –38 °C. WE THEREFORE CONCLUDE THAT THE CONTROLLING MECHANISM HAS A PERIOD OF 24±5 HOURS.

CHEMICAL CONTRAST BETWEEN WHITE AND DARK LAYERS

Figure 4 shows the results from the chlorinity, total gas content and gas composition measurements in the McM99 cores. Chlorinity (Fig. 4 left panel, crosses) is expressed in 10^3 ppm concentration units, and total (Fig. 4 left panel, diamonds and solid circles) or individual (Fig. 4 middle panel) gas contents in ml per kg of ice (STP conditions). The O₂/N₂ ratio is also shown in the right panel of Figure 4. Salient features can be summarized as follows:

(a) Both chlorinity and total gas content show a local increasing trend in the upper 0.5 meter followed by a global decreasing trend with depth

(b) Chlorinity, total gas content and individual gas compositions all show a clear contrast between the dark and the white layers, with higher values in the white layers

(c) While the enrichment of the white layer relative to the neighboring dark layer doesn’t show any trend with depth for the total gas content, it is steadily decreasing for the salinity (72 % at 0.5 m, 38 % at 0.7 m and 18 % at 1.15 m). Both variables are thus contributing to the scattering contrast in the upper layers of the sea ice cover, while the total gas content takes over in the lower layers

(d) With the exception of a single dark layer (0.34), located close to a natural break in the core (possible bias from post-sampling exchange with the atmosphere), O₂/N₂ ratios are fairly constant, ranging between 0.40 and 0.44 with a mean value of 0.41, intermediate between sea water solubility and atmospheric ratios.

ORIGIN OF THE BANDING

INADEQUACY OF GROWTH RATE OR LIGHT-MEDIATED BIOLOGICAL CONTROLS

The data presented above are at odds with the classical explanation that the banding results from growth rate variations, with faster growth leading to more bubbles and stronger scattering. Firstly, bulk salinity is as much correlated with the scattering as is the gas content, at least in the upper half of the sea ice cover (Fig. 4). Secondly, the growth rate is determined by surface temperature with natural periods which are not locked to one day, in fact they have most of their weight at substantially larger time scales (Trodahl et al., 2000). Those natural variations cannot therefore be responsible
for the 24 hours cycle in the banding. Thirdly, any daily variation in the surface temperature will be severely attenuated for any ice thicker than 0.5 m (Eq. 3). One would therefore expect no growth rate fluctuation at all with that specific period, even if there were corresponding surface temperature variations.

In the search of a controlling variable that would result in a banding daily cycle throughout more than one meter of congelation sea ice, the light exposure emerges as one candidate with fluctuations that might survive to depths greater than 0.5 meters. One could therefore surmise that the variations in the gas content might result from light-level driven biological activity. Again, our data set indicates that the scattering contrast is as much an expression of salinity contrasts than gas content contrasts. This is difficult to conceive as a result of the biological activity at the ice-water interface, if ever present.

![Figure 4: Chlorinity and total gas content in section 0.28-1.20 m of core McM99 4C (left), gas composition in CO₂, O₂, N₂ (middle) and O₂/N₂ ratio (right) in section 0.18-0.62 m of core McM99 4D.](image-url)
Tidal flow

It has been suggested (Dave Cole, private communication) that tidal flow rates are correlated with the banding of sea ice. At first glance, it would appear that such a driving mechanism is unreasonable, given the 24-hour period determined above. The most common tidal currents peak approximately every six hours (two each, ebb and flow, in typically 23 hours), and the banding we measure is a factor of four too coarse for such a growth period. However, the tides in McMurdo Sound, in common with many high-latitude tides, are diurnal (Hodgeson, 1907; Doodson, 1924; Lewis and Perkin, 1985). Furthermore, although poorly characterised in detail, the current along the western shore of Ross Island is stronger in the ebb than in the flow (Lewis and Perkin, 1985; Barry and Dayton, 1988), so that the tidal current magnitude shows a 24-hour period. C-axes fabrics in the 0.20 to 0.65 m and 0.90 to 1.40 m depth ranges of core McM99 4C show a single maximum in a horizontal girdle in accordance with the existence of such currents in the sampling area. We thus consider whether our gas and chlorinity results are in agreement with a current-controlled mechanism.

The ice-water interface boundary layer control of the impurity content

Models for physico-chemical processes of impurity fractionation on freezing of water have been developed from equivalents in metallurgy. Strictly applicable to the progression of a plane freezing front in diluted waters, the “boundary layer model” has been shown to successfully apply for the fractionation of stable isotopes of oxygen and hydrogen, salt impurities and gases during sea water freezing (see, e.g. Weeks and Lofgren, 1967; Cox and Weeks, 1975; Nakawo and Sinha, 1981; Weeks and Ackley, 1986; Souchez et al., 1987; Souchez et al., 1988; Eicken, 1998; Tison et al., 2001; Tison et al., in press). During downward freezing, the layer close to the ice-water interface in which solute transport occurs only by diffusion is truncated by convection at its base and is reduced to a thickness of millimeters to form what is usually called a boundary layer. Beyond that layer, the reservoir is considered as homogenized at all times through compositional and mechanical convection. Eventually, a steady-state is reached when the incoming impurity flux associated with the downward movement of the boundary layer is compensated by the diffusional solutes evolved by the rejection at the ice-water interface. Fick’s first law for diffusion, in the case of moving coordinates applies:

$$\text{Flux} = D \frac{(C_{iw} - C_{bw})}{z_{bl}} - C_{iw} V = 0 \quad \text{or} \quad \frac{C_{iw}}{C_{bw}} = 1 + \frac{V z_{bl}}{D}$$

where $C_{bw}$ is the solute concentration in the bulk water reservoir, $C_{iw}$ the solute concentration at the ice-water interface, $V$ the linear rate of advance of the freezing front, $D$ the diffusion coefficient of the solute species (1.18, 1.61 and 2.08 $10^{-5}$ cm$^2$s$^{-1}$ for CO$_2$, N$_2$ and O$_2$ respectively (Stauffer et al., 1985)) and $z_{bl}$, the boundary layer thickness.

Equation (4) shows that the growth rate ($V$) and the boundary layer thickness ($z_{bl}$) are the two variables controlling the enrichment factor ($C_{iw}/C_{bw}$) at the ice-water interface. Although, as we have demonstrated above, the growth rate cannot be responsible for the daily alternation of the dark/white layers throughout the core, it is certainly responsible for the global decreasing trend of chlorinity and total gas content with depth and for the locally increasing trend (cooling event) in the upper 0.5 m (Figure 4). On the other hand, boundary layer thickness fluctuation emerges as a potential candidate to explain the layering contrast. In the case of gas fractionation, a further complication however
resides in the nucleation threshold, as demonstrated by the experimental basin study of Tison et al. (in press). Figure 5 shows the total gas content and O₂/N₂ profiles in basins C14 and Q14 of the INTERICE II experiment of young sea ice growth, at a constant air temperature of −15 °C. In basin Q14, no water circulation existed, whilst in basin C14 a constant water current of 0.065 to 0.085 m s⁻¹ was maintained, all other things being equal. As discussed by Tison et al. (in press), the drastic increase of the total gas content to more than 10 ml, and the decrease of the O₂/N₂ ratio towards atmospheric value (0.268) reveals active bubble nucleation in the Current basin C14.

Clearly, the ranges of total gas content (2–5.5 ml) and O₂/N₂ ratio (0.40–0.45) observed in this study indicate that no in situ interface bubble nucleation occurred on freezing (Figure 4). However, fluctuations of the boundary layer thickness could have easily resulted in the observed chemical contrast between the dark and the white layers, the latter reflecting higher enrichment in soluble gases and salts at the ice-water interface under lower current conditions, i.e. thicker boundary layer. Note that, if bubble nucleation had occurred, chlorinity and total gas content would have shown opposite behaviors, since nucleation is favored for lower boundary layer thickness, i.e. when salts enrichment is reduced at the interface. Finally, applying equation 4, with the appropriate values for V (depth interval 0.25–0.35 m), z₀ (0.29, Eicken et al., 1998) and D, gives enrichment factors of 1.40 and 1.51 in O₂ and N₂ respectively. This provides an O₂/N₂ ratio of 0.53 for the interface water (1.40/1.51 × sea water solubility ratio at −1.9 °C), which is still considerably higher than the observed range of 0.40 to 0.45 (Figure 4).

One must however remember that, after its initial entrapment in the intercrystalline platelets, brine concentration will rise under cooling, resulting in diffusion towards the ice-water interface, as long as the temperature remains higher than about −5 °C. This will result in selective diffusion of O₂ and N₂ in a “diffusion only” process, in which O₂/N₂ ratios will converge towards a value equal to 0.53 (the initial entrapment ratio calculated above) times the D_N2/D_O2 ratio, i.e. 0.41 (actually corresponding to the observed mean value in Figure 4).

CONCLUSION

Four sea ice cores taken from near Arrival Heights on the eastern side of McMurdo Sound show strong banding effects, with alternating strongly (white) and less strongly (dark) light scattering layers. The layering is shown to be consistent with a growth cycle of 24±5 hours. High resolution gas content and composition measurements were performed to determine the concentrations of Cl⁻, CO₂, O₂ and N₂ and their contrast.
between white and dark layers. All these concentrations are enhanced in the white layers.

We have sought to reconcile these observations with three different suggested mechanisms for the banding. A surface temperature-driven growth rate mechanism can be discarded on the basis of the depth to which the bands persist, and the chemical results are inconsistent with a biological mechanism. A tidal mechanism appears the most likely, with the 24-hr period relying first on the diurnal nature of tides in McMurdo Sound and second on the relatively stronger ebb than flow current along the eastern shore. We present a detailed model to explain the results, in which the thickness of an interface boundary layer at the growing surface is controlled by the current.

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