Formation and structure of refrozen cracks in land-fast first-year sea ice

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This study characterizes the healing process and structure of undeformed, linear, parallel-sided, flooded cracks in land-fast sea ice. Field investigations and refreezing experiments were performed in McMurdo Sound, Ross Sea, Antarctica, between 1998 and 2002. Data from a two-dimensional thermistor array are used to show that the ice-water interface of freezing cracks is arch-shaped due to bidirectional heat flow to the surrounding ice cover and to the atmosphere. Ice growing laterally into the crack is found to desalinate over a prolonged period of time, until the isotherms are approximately horizontal. Superposition of heat flow to the atmosphere and to the host sea ice sheet allows the refreezing progress to be modeled analytically. Close to the ice–air interface, the salinity is higher at the sides of wide refrozen cracks than it is at the center. However, deeper down and in narrower cracks in general, the salinity is higher at the center than at the sides. A finite volume, computational fluid dynamics (CFD) model reproduces the generally arch-shaped alignment of brine pockets. This pattern is attributed to convection in the mushy layer. Crystals are found to grow upstream into the crack due to a salinity gradient in the buoyant convective flow.


1. Introduction

Cracks, open and refrozen, are ubiquitous features in sea ice. They affect operations on the ice, its dynamic behavior, the salt flux into the ocean, the break-up process, and wildlife. In spite of far-reaching implications, few studies have been devoted to the investigation of the evolution and structure of refrozen cracks in sea ice. In this paper, we investigate the development of the refrozen thickness of cracks, and their crystal, porosity and salinity structure.

Previous observations of freezing and refrozen parallel-sided cracks showed that the c-axes of crystals are generally horizontal in both, lake ice and in sea ice [Shackleton, 1909; David and Priestley, 1914; Taylor and Lyons, 1959; Weeks and Lee, 1958; Metge, 1976; Weeks and Ackley, 1986]. This alignment is consistent with bidirectional heat flux considerations [Weeks and Ackley, 1986]. Further, in lake ice, the grain boundaries of the crystals are approximately perpendicular to the interface [Metge, 1976]. The ice-ocean interface is arch-shaped (Figure 1), and empirical relationships between freezing time and refrozen thickness have been sought for both freshwater ice and sea ice [Metge, 1976; Christensen, 1986; Langhorne and Haskell, 2004]. Preliminary results from the present study indicated that in sea ice, the crystal boundaries deflect from the direction of heat flux. Further, the salinity was lowest where inclusion density was lowest, and an arch-shaped alignment of brine inclusions was observed [Petrich et al., 2003].

According to the definition of the WMO [1970], a crack is any fracture of fast ice, consolidated ice or a single floe which may have been followed by separation ranging from a few centimeters to 1 m. This paper presents field data and analysis of the freezing process and the structure of undeformed, linear, parallel-sided refrozen cracks in land-fast first-year sea ice. We will investigate refrozen cracks in three different ways: experimentally, in fluid dynamics simulations, and with an analytical model, and we will concentrate on growth rate, crystal fabric, and inclusion and salinity distribution. The rate of ice growth in the cracks gives an indication of the salt flux into the ocean [Gow et al., 1990; Wettlaufer et al., 2000], while refrozen thickness is an important parameter for the strength of the ice sheet [Langhorne and Haskell, 2004]. We develop an analytical model that describes refreezing and successfully interpolates experimental data and results from numerical simulations. Depending on the width of the crack and on the vertical position, growth is dominated by heat flux to either atmosphere or ice sheet. The direction of heat transport is preserved in the crystal structure. The crystal structure reflects the superposition of the direction of heat flow and fluid flow in the water at the time of growth [Langhorne and Robinson, 1986; Weeks and Ackley, 1986; Bergman et al., 2002]. The fabric may in turn affect the strength of the

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ice [Weeks and Ackley, 1986; Wu and Zhang, 1995; Cole, 2001]. We show that the crystal structure deviates from the structure expected due to heat flux alone, and that fluid motion at the ice–water interface can be used to explain the observed structure qualitatively. Fluid flow through the permeable part of the forming sea ice determines the inclusion and salinity distribution. The salinity and the distribution of inclusions affect the potential for mechanical failure [Petrich et al., 2000; Dempsey, 2001, 2002]. While the mean salinity of refrozen cracks is similar to the salinity of the surrounding sea ice, we show that the salinity is distributed unevenly. Toward the ice-air interface, the salinity of wider cracks is lower at the center than at the sides. However, deeper down and in narrower cracks in general, the salinity is higher at the center than at the sides. Inclusions throughout the crack are generally aligned as arches, which is reproduced in fluid dynamics simulations.

In the following, refrozen cracks are investigated experimentally, with fluid dynamics simulations, and with an analytical model. Since the methods are interlinked we start by introducing the CFD model in section 2. This model will be used to simulate the thickness and temperature evolution (section 4) and the salinity profile (section 6). In section 3 we describe the cracks investigated experimentally. The refrozen thickness is treated in section 4, an analytical model of crack refreezing is developed in section 4.2, and the analytical model is compared with experimental data and CFD simulations in section 4.3. The crystal structure is presented in section 5. Measured inclusion and salinity distributions are discussed in section 6 and compared with fluid dynamics simulations.

2. Computational Fluid Dynamics Simulations

Numerical fluid dynamics simulations of the refreezing of cracks are performed in order to assess whether the observed depth-dependent salinity pattern and the arch-shaped inclusion alignment could be a result of the fluid motion during crack refreezing alone, irrespective of details of the crystal fabric. Two-dimensional numerical fluid flow simulations of unidirectional sea ice growth have been performed previously by Medjani [1996] and by Oertling and Watts [2004], based on governing equations given by Bennon and Incropera [1987] and an empirical approach to flow resistance in sea ice. In this work we use a set of continuum governing equations for fluid flow through a porous medium that is derived from first principles [Ganesan and Poirier, 1990; Bear and Bachmat, 1991] and accounts for volume expansion during phase transition. The parameterization of flow resistance in sea ice is based on Petrich et al. [2006].

2.1. Governing Equations

A two-dimensional porous medium fluid dynamics simulation is performed, based on the finite volume method [Patankar, 1980; Ferziger and Perić, 2002]. Sea ice is treated as a stationary, locally homogeneous, porous medium that allows fluid motion against a flow resistance. The liquid volume is assumed to be interconnected, and the total porosity $f_t$ is defined as the fraction of the liquid volume $\delta V_t$ in the averaging volume $\delta V$,

$$ f_t = \frac{\delta V_t}{\delta V}. $$

Conservation equations of local average properties in a porous medium are volume averages of the Navier-Stokes equations with Boussinesq approximation, and of the heat and solute conservation equations. A first-order upwind scheme is used for the discretisation of the advection terms, and pressure-velocity coupling is attained by the SIMPLEx algorithm [Versteeg and Malalasekera, 1995].

The governing equations are expressed in terms of local, intrinsic volume averages of temperature, $T$, solute concentration in the liquid, $C$, and horizontal and vertical velocity components $u$ and $v$, respectively. Unless otherwise stated, all physical properties of liquid and solid, for example density and heat capacity, are assumed to be constant locally, i.e. within the averaging volume $\delta V$.

The volume-averaged mass conservation equation is

$$ \left[ 1 - \frac{\rho_s}{\rho_l} \right] \frac{\partial \rho_l}{\partial t} + \frac{\partial (\rho_l u)}{\partial x} + \frac{\partial (\rho_l v)}{\partial y} = 0, $$

where $u$ and $v$ are the fluid velocity components in the $x$ and $y$-directions, respectively (Figure 2), $f_t$ is the total porosity,
and \( \rho_l = 1000 \text{ kgm}^{-3} \) and \( \rho_s = 918 \text{ kgm}^{-3} \) are the constant densities of liquid and solid, respectively.

[10] The volume-averaged momentum conservation equations are

\[
\rho \left[ \frac{\partial (fu)}{\partial t} + \frac{\partial (fuu)}{\partial x} + \frac{\partial (fuv)}{\partial y} \right] = \mu \left[ \frac{\partial^2 (fu)}{\partial x^2} + \frac{\partial^2 (fu)}{\partial y^2} \right] - f_i \frac{\partial p}{\partial x} + f_i \frac{\mu}{\Pi} f_u, \tag{3}
\]

\[
\rho \left[ \frac{\partial (fv)}{\partial t} + \frac{\partial (fvy)}{\partial x} + \frac{\partial (fvy)}{\partial y} \right] = \mu \left[ \frac{\partial^2 (fv)}{\partial x^2} + \frac{\partial^2 (fv)}{\partial y^2} \right] - f_i \frac{\partial p}{\partial y} + f_i \rho g - f_i \frac{\mu}{\Pi} f_v, \tag{4}
\]

where \( \rho \) is the variable density of the liquid in the buoyancy term (third term on the right hand side of (4)), which is a function of brine salinity and temperature [UNESCO, 1981], \( \mu = 1.8 \times 10^{-3} \text{ kgm}^{-1} \text{s}^{-1} \) is the dynamic viscosity, \( p \) the pressure, \( \Pi \) is the permeability of the porous medium (see below), and \( g = -9.8 \text{ ms}^{-2} \) is the acceleration due to gravity, which is assumed to be parallel to the \( y \)-direction (positive \( y \)-direction is up, see Figure 2). Underlying the momentum equations are the assumptions of the Boussinesq approximations, i.e. a constant liquid density apart from the buoyant term, of a globally constant viscosity, \( \mu \), and that the fluid velocity is zero at the microscopic solid-liquid interface.

[11] The volume-averaged form of the energy conservation equation is

\[
\frac{\partial T}{\partial t} + \rho c_l \frac{\partial (fu T)}{\partial x} + \rho c_s \frac{\partial (f v T)}{\partial y} = \frac{\partial}{\partial x} \left[ \frac{k}{\mu} \frac{\partial T}{\partial x} \right] + \frac{\partial}{\partial y} \left[ \frac{k}{\mu} \frac{\partial T}{\partial y} \right] - [T \Delta (\rho c) + L \rho_a] \frac{\partial f_i}{\partial y}, \tag{5}
\]

where \( T \) is the temperature of solid and liquid in °C, \( L = 334 \times 10^3 \text{ Jkg}^{-1} \) is the latent heat of fusion at \( T = 0^\circ \text{C} \), and the average quantities in the porous medium are defined as

\[
\overline{\rho} = f_i \rho_l c_l + (1 - f_i) \rho_s c_s, \tag{6}
\]

\[
\Delta (\overline{\rho c}) = \rho_l c_l - \rho_s c_s, \tag{7}
\]

\[
\overline{k} = f_i k_l + (1 - f_i) k_s, \tag{8}
\]

with specific heat capacities \( c_l = 4.2 \text{ kJkg}^{-1} \text{K}^{-1} \) and \( c_s = 2.1 \text{ kJkg}^{-1} \text{K}^{-1} \), and thermal conductivities \( k_l = 0.56 \text{ Wm}^{-1} \text{K}^{-1} \) and \( k_s = 2.1 \text{ Wm}^{-1} \text{K}^{-1} \) of liquid and solid, respectively. Underlying (5) is the assumption of local thermal equilibrium between the liquid and the solid.

[12] The volume-averaged solute conservation equation is

\[
f_i \frac{\partial C}{\partial t} + \frac{\partial (f_i u C)}{\partial x} + \frac{\partial (f_i v C)}{\partial y} = \frac{\partial}{\partial x} \left[ f_i D \frac{\partial C}{\partial x} \right] + \frac{\partial}{\partial y} \left[ f_i D \frac{\partial C}{\partial y} \right] - C_i \frac{\partial f_i}{\partial t}, \tag{9}
\]

with the concentration \( C \) of solute in the liquid, and the solute diffusion coefficient \( D = 7 \times 10^{-10} \text{ m}^2\text{s}^{-1} \). In the solid, the solute concentration is zero, and at the microscopic solid-liquid interface, it is equal to the average solute concentration \( C \) in the liquid fraction of \( \delta V \).

[13] Finally, an equation is needed that governs phase transition. Supposing local thermodynamic equilibrium, in any volume \( \delta V \) the temperature \( T \) is equal to the freezing point \( T_F \) of the brine of concentration \( C \),

\[
T = T_F(C). \tag{10}
\]

With consideration given to latent heat release and solute partition at the microscopic interface, (10) constitutes a condition for the change of local liquid volume fraction required to establish thermodynamic equilibrium. Local thermodynamic equilibrium is enforced at every time step of the numerical simulation.

[14] The permeability of sea ice is described by the parameterisation [Petrich et al., 2004, 2006],

\[
\Pi(f_i) = 1 \times 10^{-10} \text{ m}^{-2} \cdot (f_i - 0.054)^{1.2}. \tag{11}
\]

The permeability, \( \Pi \), in (11) is isotropic, neglecting anisotropy that could be present due to the anisotropy of the sea ice matrix [Freitag and Eicken, 2003].

[15] Sea ice is assumed to be essentially impermeable at low total porosities [Golden et al., 1998]. This is simulated for \( f_i \leq 0.054 \) by removing all force terms from the momentum conservation equations and setting the velocity to zero. However, in order to be able to treat brine expulsion due to volume expansion during freezing, a permeability of \( \Pi = 1 \times 10^{-14} \text{ m}^2 \) is assumed in the pressure correction equation, i.e. when enforcing mass conservation. The numerical details are laid out elsewhere [Petrich, 2005].

### 2.2. Numerical Domain and Boundary Conditions

[16] The numerical domain is deeper than the ice sheet and spans half of the width of the crack for numerical
efficiency. No-slip boundary conditions are applied at the boundary to the air and to the host ice sheet, respectively. A vertical mirror boundary is used at the center of the crack, and the bottom boundary is open with fluid advected into the domain at prescribed temperature, $T_0$, and solute concentration, $C_0$. The heat transfer boundary conditions are borrowed from the analytical model described in section 4.2. They are $\frac{\partial T}{\partial y} = [T_a - T]_{\alpha_{ai}/k_i}$ and $\frac{\partial T}{\partial x} = [T(y) - T]$ $2/w_0$ at the air and host ice boundaries, respectively. The host ice boundary extends to $H = 2$ m, below which an adiabatic temperature boundary condition ($\frac{\partial T}{\partial x} = 0$) is used.

3. Experiments

Natural and man-made refrozen cracks were investigated in the sea ice of McMurdo Sound, Ross Sea, Antarctica, within 15 km of Ross Island. Visits to the area were made for this project in October and November of 1998, 1999, and 2001, and in September 2002. The refrozen cracks were selected to be of dimensions suitable for excavation (Figure 3 and Table 1). Samples of refrozen cracks were excavated either with a ditch digger [Haskell et al., 1996] and crane (for example, Figure 1), or with a chain saw with a long blade. Samples used to measure vertical salinity profiles were approximately 100 to 150 mm thick.

Linear, parallel-sided man-made cracks, hereafter called slots, were prepared with the ditch digger. After cutting a slot of 2 to 3 m length through the entire thickness of the ice sheet, care was taken to remove all ice bridges that occasionally remained at the bottom of the ice sheet. Slushy debris was removed from the seawater and a thermistor probe was installed before refreezing began. The removal of debris and installation of the probe took approximately 1 to 2 hours.

The thermistor probe in the slot consisted of three parallel thermistor strings, one close to one side of the slot, one in the center of the slot, and one in between (Figure 4). A fourth string was frozen into the host ice sheet in a hole drilled close to the slot. Each string consisted of a plastic conduit for support that was filled with potting compound. From this, the thermistors protruded 25 mm, held in place by 4 mm outer diameter plastic tubing. The thermistor beads were covered with an electrically insulating coating of nail varnish but were otherwise exposed directly to the environment.

An example of a natural, refrozen crack in September 2002 is shown in Figure 3. The linear shape and the smoothness of the surface of the crack are typical for the cracks examined. The crack harbored another crack, presumably a thermal crack, that narrowed with depth and probably did not extend to the bottom of the ice sheet [cf. Kingery and Coble, 1963].

Observations that are deemed representative of the refrozen cracks investigated are discussed in the following sections with further examples presented elsewhere [Petrich et al., 2003; Petrich, 2005].

4. Crack Thickness

4.1. Refrozen Thickness From Temperature Data

The presence of an arch-shaped interface (Figure 1) causes the temperature–time series obtained at the center of the slots, away from the ice–air interface, to differ qualitatively from unidirectional ice formation [cf. Lake and Lewis, 1970]. The temperature curves of slot 10 (Figure 5a) resemble temperature curves of unidirectional freezing in the upper 180 mm: the temperature in the water is constant at the freezing point and decreases rapidly once

<table>
<thead>
<tr>
<th>Table 1. Parameters Used for Model Calculations$^a$</th>
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<tbody>
<tr>
<td>Experiment</td>
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<td>----------</td>
</tr>
<tr>
<td>LH12d</td>
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<tr>
<td>LH7d</td>
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<tr>
<td>slot 1</td>
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<td>slot 2</td>
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<td>slot 10</td>
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<td>slot 12</td>
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<tr>
<td>M4a</td>
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<td>M4b</td>
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<td>D5</td>
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</table>

$^a$Experiments M4a and M4b are freshwater experiments. Estimated values are given in parentheses.
We conclude that ice began to form simultaneously at all depths between 320 and 500 mm at the center of slot 10, driven by approximately vertical isotherms which moved horizontally. This is consistent with the idea of a vertical freezing interface encroaching from the sides of the crack. The subsequent separation of temperature curves indicates that once the thermistors are in the ice the isotherms rotated toward the horizontal. The temperature change at a rate comparable to the one observed above 180 mm is consistent with the vertical movement of horizontal isotherms.

[24] The transition from vertical to horizontal isotherms at 415 mm is further illustrated by the dotted line in Figure 5b, which shows the angle that an isotherm through 415 mm and close to the center of the crack makes to the horizontal. Confirming the assessment above, the isotherm which is close to vertical initially, rotates when the temperature curve separates to become horizontal once the rate of temperature change is high.

[25] The initial formation of porous sea ice can be detected by a temperature threshold that is a few tenths of a degree below the freezing point [Wettlaufer et al., 2000]. It may be assumed that this method generally does not describe the formation of consolidated sea ice since such a temperature decrease has been detected simultaneously over depths from 320 to 500 mm in experiment slot 10 (Figure 5a, hour 50). In the absence of direct measurements, fluid dynamics simulations of the desalination of sea ice in cracks predict that the ice loses a significant portion of its salt over a period of time, from initial ice formation until isotherms become approximately horizontal (Figure 6).

[26] The time of formation of consolidated sea ice, i.e. the development of a relatively constant salinity, can be estimated from the time of maximum rate of temperature change. This is confirmed by Figure 6, which shows that the salinity is approximately constant from this time onwards. The time of maximum rate of temperature change is further very close (1–3 h) to the time of maximum vertical temperature gradient, which will be used in the data analyses since is less affected by temperature fluctuations during the experiments.

Figure 4. Thermistor probe in a refreezing slot with thermistor strings arranged as in experiments slot 1 and slot 2. The horizontal position of the strings varies between experiments. The vertical positions of the thermistors on the strings are drawn to scale. They are offset vertically to reduce the number of thermistors needed to cover the area spanned by the probe. Experiments slot 10 and slot 12 have the longest string at the center of the slot.

Figure 5. Temperature–time series at the center of slot 10; (a) thermistor measurements and (b) CFD simulation. The labels at the lines state the depth below the ice–air interface in mm. The dotted line in (b) shows the angle to the horizontal of the isotherm that passes through the center at 415 mm depth, 10 mm away from the center; 0° and 90° indicate that the isotherm is horizontal and vertical, respectively.
Figure 7 compares the porous and consolidated interfaces of experiments slot 1 and slot 10 at different times throughout the refreezing process. To generate the plots, the passage times of the interface at the thermistors were linearly interpolated throughout the area spanned by the probe. The interface marked by the heavy solid lines represents the onset of ice formation, which is defined here by a temperature decrease of 0.2°C below the water temperature [Wettlaufer et al., 2000]. The thin solid lines trace the consolidated ice at the same times, defined by the maximum vertical temperature gradient. The arch-shape of the interface is clearly reproduced by the temperature threshold approach. Owing to its relationship with the horizontal isotherms, the consolidated front is horizontal and propagates vertically downward. Both fronts coincide at the center throughout experiment slot 1, while in slot 10 they only coincide for thin ice. Once sufficient time has elapsed and lateral ice formation dominates, the porous front separates from the consolidated front. This implies that a large region exists, particularly at the sides of the crack, where active desalination can take place.

4.2. Analytical Model

Refreezing of cracks can be described as a two-dimensional heat transfer problem with phase transition. The solution does not yield itself to a simple expression. However, the present data can be approximated by a function with one free parameter, which is briefly derived next.

The refreezing process is considered to be a superposition of ice growth due to vertical heat transfer to the atmosphere and ice growth due to horizontal heat transfer to the host ice sheet. The ice sheet is assumed to be of constant thickness during the refreezing process.

We will define the freezing time as the time required for the ice at a given position to attain thermodynamic properties that are similar to the bulk of the desalinated, consolidated refrozen ice in the crack. The freezing time \( t_a \) of unidirectionally growing sea ice of negligible heat capacity and otherwise constant properties can be calculated from [Anderson, 1961]

\[
\frac{1}{t_a} = \frac{2 \rho_i L_i}{\rho_i L_i (h^2 + 2h \alpha_{ai}/k_i)}.
\]

Figure 7. Experimentally determined freezing front of experiments slot 1 and slot 10 defined by a temperature threshold of 0.2°C (thick lines) and by the maximum vertical component of the temperature gradient (thin lines), respectively. The contour lines follow the position of the freezing front at different times. The separation of contour lines is \( 5 \times 10^4 \) s (14 h) and \( 3 \times 10^4 \) s (8 h) for slot 1 and slot 10, respectively. 
is the heat transfer coefficient between air and ice, \( \rho_i = 920 \text{ kgm}^{-3} \) is the density of ice, \( L_i \) is the latent heat of fusion of sea ice, and \( k_i \) is the thermal conductivity. Equation (12) is formally identical to the quasi-steady solution of the Stefan problem [Carslaw and Jaeger, 1986] with the distance \( h_0 = \alpha_i/k_i \) being the displacement of the interface of constant temperature \( T_u \) from the actual ice-air interface (Figure 2), the material in between having a thermal conductivity of \( k_i \), and the initial condition \( h = 0 \) at \( t_u = 0 \). By analogy, the freezing time \( t_s \) of lateral solidification is

\[
\frac{1}{t_s} = \frac{2k_i(T_w - T_u)}{\rho_i L_i (w/2)^2 + 2(w/2)(w_0/2)},
\]

(13)

where \( T_u(y) \) is the temperature inside the ice sheet at the distance \( w_0/2 \) from the edge of the crack (Figure 2), and \( w/2 \) is half of the width of the crack. The temperature that drives ice growth from the side is assumed to be a linear function of depth and follows

\[
T_s(y) - T_w = \frac{H + y}{H + h_f + h_0} (T_u - T_w),
\]

(14)

where \( H \) is the draft of the ice sheet and \( h_f \) is the freeboard of the ice sheet, and \( y \leq 0 \). For a free-floating ice sheet, the freeboard is calculated from

\[
\frac{\rho_{sw}}{\rho_i} H = H + h_f,
\]

(15)

where \( \rho_{sw} \) and \( \rho_i \) are the densities of seawater and sea ice, respectively, and \( \rho_{sw}/\rho_i \approx 1.1 \). For simplicity, we treat vertical and lateral heat fluxes as independent of each other and the lateral heat flux as independent of depth. The time of consolidated ice formation is then

\[
\frac{1}{t} = \frac{1}{t_s} + \frac{1}{t_v}.
\]

(16)

The time of freezing at a given depth is written in dimensionless form

\[
\tau = \frac{z^2}{1 + \beta \frac{z}{d^2}(1 - z)},
\]

(17)

where the dimensionless time is

\[
\tau = t \frac{2k_i(T_u - T_w)}{\rho_i L_i H^2},
\]

(18)

and the dimensionless lengths are

\[
\begin{align*}
z &= \frac{h}{H}, & z_0 &= \frac{h_0}{H} &= \frac{k_i}{H \alpha_{sw}}, \\
\tilde{z}^2 &= z(z + 2z_0), \\
d &= \frac{w}{H}, & d_0 &= \frac{w_0}{H} &= \gamma, & \tilde{d}^2 &= d(d + 2d_0) \quad \text{where} \quad \gamma = \frac{D5}{V},
\end{align*}
\]

(19)

It is assumed that \( w_0 \) is proportional to \( H \), with an undetermined constant of proportionality \( \gamma \). According to equation (17), the refreezing process is not scale invariant since the heat transfer coefficient between atmosphere and ice does not scale with the draft of the ice sheet.

### 4.3. Comparison With Experiment and Simulations

The model of equation (17) is compared in Figure 8 with seawater data from the two direct refreezing experiments of 1999 (LH12d and LH7d, Langhorne and Haskell, 2004) and the thermistor array of Figure 4. In Figure 9, the model is compared with laboratory studies of refreezing cracks in saltwater (D5, Divett [2000]) and in freshwater (M4a and M4b, Metge [1976]). The comparison in freshwater is made with data of the two experiments that experienced the smallest growth of the surrounding ice sheet during refreezing (for freshwater: \( L_i = 334 \times 10^3 \text{ Jkg}^{-1}, k_i = 2.0 \text{ Wm}^{-1}\text{K}^{-1} \)). The boundary conditions and experimental conditions of ice growth are summarized in Table 1. Where possible, the heat transfer coefficient of field experiments was derived from the vertical growth velocity of thin ice \( v_0 \), according to \( v_0 = \rho_i L_i = \alpha_{sw}(T_u - T_w) \). The heat transfer coefficient of experiments M4a and M4b was determined by balancing the measured heat conducted...
through the ice sheet with the heat flux from the ice sheet to the atmosphere. Estimated values are given in brackets. The optimum value of $g$ for a given set of data depends on the choice of constants of proportionality between $t$ and $t_i$. In order to achieve consistency with the fluid dynamics simulations, the value for the latent heat of fusion of sea ice $L_i$ has been used to include sensible heat and porosity, similar to Yen [1981]. A salinity of $S_i = 7$ psu and the mean temperature of the ice have been used for each experiment. The representative thermal conductivity of sea ice is taken to be $k_i = 1.8 \text{ W m}^{-1}\text{K}^{-1}$ $C_0$. The calculations are presented for $g = 0.37$ and $g = 0.50$.

Figure 9. Comparison of model (solid line $\gamma = 0.37$, dashed line $\gamma = 0.50$) and data (markers) for the refrozen thickness of slots in laboratory experiments.

Figure 10. Comparison of the consolidated freezing front predicted from the analytical model (solid line; dashed line with $\gamma$ increased by 10%) and calculated from the CFD simulation (markers). Only every second point is shown for clarity.

5. Crystal Structure

[35] The texture shown in vertical thin sections of refrozen cracks differs markedly from the texture of sea ice formed by unidirectional freezing. The example of crack 5 in Figure 11 shows the vertical, columnar structure of the host ice sheet to the left and to the right of the crack. The crystal fabric in the crack is symmetrical. Granular crystals are visible immediately adjacent to the host ice sheet, with elongated crystals closer to the center. In order to describe their apparent alignment, we trace the crystals in their presumed direction of growth, which is toward the center of the crack. Crystals close to the sides of the crack are tilted upwards into the center of the crack (Figure 12). Crystals point downwards into the center from the ice–air interface. The black crystals at the center have their $c$-axes horizontal and parallel to the sides of the crack.

[36] The observed upward-tilt of crystals is contrary to observations in lake ice [Metge, 1976]. In lake ice, the crystal structure appears to be explained by geometric selection in the direction of heat flow [Weeks and Ackley, 1986]. In sea ice, the crystal structure may be affected by solute rejection during freezing: solute rejection would facilitate the development of a buoyancy driven, downwelling current at the curved ice–water interface (Figure 12). Upstream growth of the crystals is favored if the salinity increases as the water moves downward along the interface. Hence, the angle of growth would be determined by the superposition of geometric selection in the direction of heat flow and the growth advantage upstream. This concept is
illustrated in the cartoon in Figure 12. Upstream growth of crystals into a flowing melt is known to occur in alloys [Flemings et al., 1956; Flemings, 1974; Murakami et al., 1983; 1984], and it has been described in laboratory experiments for vertically growing saltwater ice in the presence of a forced current [Langhorne, 1983; Langhorne and Robinson, 1986] and noticed in laboratory experiments of horizontally growing saltwater ice in the presence of buoyancy driven convection [Bergman et al., 2002].

[37] The thin section of Figure 11 shows two vertical bands of granular ice in the top 80 mm close to either side of the crack. The location of the granular bands coincides with the location of an increased number of inclusions [Petrich et al., 2003]. They may indicate that this crack was narrower when it initially formed and underwent a fracture event after it refroze to a thickness of about 80 mm.

6. Inclusion and Salinity Structure

6.1. Distribution of Inclusions

[38] In refrozen cracks, the location of brine inclusions appears to be correlated with the crystal structure. The example of man-made crack slot 10, 180 mm wide, in Figure 13 shows a relatively clear (“V”-shape) region at the center of the thick section, where the corresponding thin section shows downward-aligned crystals. The inclusions are mostly in the region of tilted crystals and at the crack-host ice interface. Further, a vertical band of inclusions at the center of the crack is visible below approximately 190 mm. Correspondingly, the transition from vertical crystals at the center to tilted crystals takes place between 180 and 220 mm.

[39] The observation of few inclusions in regions of vertically elongated crystals and a band of inclusions where tilted crystals meet at the center of the crack is supported by findings in natural refrozen cracks. Figure 14 compares the vertical thick and thin sections of relatively narrow crack crack 20. The thick section shows a band of inclusions at the center of the crack throughout the entire sample height,
while the thin section shows that, apart from the upper 30 mm, tilted crystals meet at the center. The structure of this narrow crack hence conforms to the lower excavated part of slot 10 in Figure 13. A thick section of the relatively wide crack crack 5 of Figure 11 exhibits a center that is rather devoid of inclusions [Petrich et al., 2003]. The structure of this wide crack hence conforms to the upper part of slot 10 in Figure 13.

[40] Away from the center of the crack, the inclusions in slot 10 are aligned as arches. Likewise, inclusions not too far away from the center of crack 20 show arch-shaped alignment, as do inclusions in other cracks investigated [Petrich et al., 2003; Petrich, 2005]. Curved inclusion alignments have also been observed during alloy casting [Flemings, 1974]. This structure is in no obvious relationship to the crystal fabric, nor is it perpendicular to the freezing interface as observed in unidirectional freezing of sea ice [Weeks and Ackley, 1986]. However, brine channels have been observed to tilt toward the cold side in quasi-unidirectional ice growth [Niedrauer and Martin, 1979], which would be consistent with the alignment in refreezing cracks. We may postulate that the inclusion alignment is a result of solute redistribution in the skeletal layer during the initial stages of freezing. This hypothesis will be verified with the fluid dynamics simulations in section 6.3.

[41] The transition from a transparent center to a center marked by inclusions is probably related to the ratio of heat transfer to the top and to the sides. In the case of slot 10 this occurs around the point where \( t_a = t_s \), which is at \( h = 240 \) mm. The wide crack crack 5 should show this transition below the excavated depth, while refreezing of the narrow crack crack 20 appears to have been dominated by lateral heat flow soon after refreezing began.

[42] Granular ice visible at the sides of crack 20 and crack 5 could be incorporated frazil ice or the result of mechanical forces during refreezing.

6.2. Salinity Profile

[43] The salinity profile of slot 10 in Figure 15 shows that the vertical salinity profile changes with depth. Close to the ice–air interface, the salinity is lowest at the center, while below approximately 200 mm the salinity is high at the center and at the transition from slot to host ice. The salinities near the bottom of the sample may not have reached steady values at the time of excavation. However, the regions of high and low salinity correspond qualitatively to the regions of high and low inclusion volume, respectively, as seen in the vertical thick section.

[44] The positive correlation between apparent inclusion volume and sea ice salinity is also found in natural refrozen cracks. The salinity profile of crack 20 in Figure 14 shows an elevated salinity at the center of the crack, consistent with the presence of the band of inclusions. As in the case of slot 10, the salinity is also higher at the crack–host ice interface, where vertical bands of inclusions are visible in the thick section. Similarly, the salinity profile of crack 5
Petrich et al. [2003] exhibits a low salinity at the center of the crack, where few inclusions are visible. An increased salinity in the vicinity of brine channels in unidirectionally grown sea ice has previously been found in high resolution laboratory studies [Cottier et al., 1999; Tison and Verbeke, 2001].

### 6.3. Simulated Salinity Profile

The simulated salinity profile of the upper 560 mm of slot 10 is shown in Figure 16. The domain size is $0.09 \times 2.56$ m², divided into a $16 \times 256$ grid.

The salinity profile in Figures 16 reveals arches of high salinity. They are pathways of downward-flowing brine during freezing. The number of arches depends on grid size, more arches develop as the grid becomes finer. This is consistent with excavated cracks and slots that show general arch-shaped alignment of inclusions rather than alternating arches of high and low inclusion density, indicating that a characteristic separation does not exist. The arches are a desalination phenomenon and do not trace the freezing interface. They develop even in simulations where fluid motion is inhibited in purely liquid cells ($f_i = 1$). The arches further originate from a “V”-shaped outline at the upper part of the cracks, corresponding to the observation in Figure 13.

The average salinity of the upper 560 mm is, at 18 psu, considerably larger than the 7 psu measured in field experiments, which is probably due to the assumption of permeabilities II that are too low.

### 7. Summary and Conclusions

In this paper we have examined the development of the refrozen thickness, and the crystal, porosity and salinity structure of undeformed, linear cracks in sea ice. These cracks refreeze quasi two-dimensionally due to heat transfer to the atmosphere and to the adjacent host ice sheet, producing an arch-shaped freezing interface. The isotherms follow an arch shape at the ice-water interface, but are more horizontal higher up in the refrozen material, toward the ice–air interface. The rate of desalination is greatly reduced when the isotherms are horizontal.

Three modes of refreezing can be discriminated, each leading to a distinct crystal structure, inclusion structure and salinity distribution. The first two have been investigated in this study. In the first mode, seen in wide cracks, crystal growth is dominated by heat transfer to the atmosphere. Rejected brine at the center of the crack is readily removed, giving rise to a region of low inclusion density and low salinity there. The second mode is observed in wide cracks as refreezing progresses and in narrow cracks from the start of freezing. In this mode crystal growth is dominated by
heat transfer to the host ice at both sides. The center of the crack is filled with ice of high porosity, which impedes efficient solute removal and gives rise to a higher inclusion density and consequently high salinity at the center. In the final mode, freezing slows because the temperature of the host ice is close to the ocean temperature and not sufficient to enhance ice growth significantly.

[52] We have presented an analytical equation that can be applied to estimate the time needed to obtain a certain refrozen thickness. Since the strength of refrozen cracks increases with their thickness [Langhorne and Haskell, 2004], this time estimate could be useful for operations on the ice and to predict wave scattering of the ice field [Williams and Squire, 2006]. The strength of sea ice is also affected by crystal structure [Weeks and Ackley, 1986]. Consequently the differing fabrics of narrow cracks and the upper part of wide cracks should be considered in experimental work. The analytical equation can also be used to define a typical time scale for the refreezing process in models of ice flow dynamics [cf. Hopkins and Thordikne, 2006]. Further, the salt flux into the ocean can be estimated from the rate of refreezing.

[53] The position in the ice at which an isotherm becomes horizontal demarcates consolidated ice above, from the newly-formed, high porosity ice beneath. The arch-shaped distribution of brine inclusions observed in refrozen cracks documents the path of fluid convection in this porous region. We have demonstrated that these arch-shaped strings of inclusions are produced in fluid dynamics simulations as a result of solute redistribution and thermodynamics in the mushy layer. Consequently we prefer this convective explanation for the inclusion distribution over one in which brine channels are driven by thermal gradients toward the colder regions of the ice [Niedrauer and Martin, 1979]. Strings of inclusions that are tilted at an angle to the vertical are a ubiquitous feature of sea ice [Cole et al., 2004]. Convection in the porous ice above the ice—water interface may be responsible for the tilt. Inclusion distribution becomes especially important as the sea ice decays because meltwater percolation follows the pathways of highest inclusion density.

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