Marine ice formation and deformation at the Southern McMurdo Ice Shelf, Antarctica

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Acknowledgements

Work for this thesis involved research stays on four different continents: my home base New Zealand, my fieldwork destination Antarctica, my primary laboratory work environment in Belgium and finally my adventurous new home Nepal, where I tied the different threads together in writing. Just before finishing I also spend a significant time on yet another continent, planet USA. Whilst the work there was not directly related to my thesis, the research stay likely gave me the needed confidence to prepare this piece of work for submission. Along my journey I met many souls that directly or indirectly supported me.

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Abstract

Marine ice accretes at the base of ice shelves often infilling open structural weaknesses and is thus thought to increase ice shelf stability. However, marine ice formation and deformation processes still remain poorly understood. Through measurements of marine ice properties, this study indirectly infers processes that occur during ice shelf flow and in the ice shelf cavity. Marine ice water isotope and solute chemistry are examined in ice cores from the Southern McMurdo Ice Shelf (SMIS) to derive marine ice source water composition and its origin. Marine ice microstructure (ice fabric, crystals size and shape) is also investigated in ice samples collected along an ice shelf flowline of increasing total strain to establish marine ice deformation in situ and compare it to deformation of ice formed from solid precipitation (meteoric ice).

The measured marine ice water isotope composition together with the output of a boundary-layer freezing model indicate a spatio-temporally varying water source of sea water and relatively fresher water, such as melted meteoric or marine ice. This is in agreement with the occurrence of primarily banded and granular ice crystal facies typical for frazil ice crystals that nucleate in a supercooled mixture of water masses. It is proposed that marine ice exposed at the surface of SMIS, which experiences summer melt, is routed to the ice shelf base via the tide crack. Here frazil crystals nucleate in a double diffusion mechanism of heat and salt between two water masses at their salinity-dependent freezing point and accrete at the ice shelf where they consolidate to marine ice. Recycling of previously formed marine ice facilitates ice shelf self-sustenance with increasing air temperatures.

Marine ice microstructure dynamically recrystallizes as a response to 20 - 25% total shear strain and vertical extension/horizontal compression. The marine ice extracted closer to shore develops a slightly less pointed anisotropic fabric, loses some of its horizontal shape preferred orientations (SPO) (with reference to vertical thin sections). Marine ice also adjusts its microstructure differentially downcore, indicating that it does not deform uniformly but shears in distinct planes. However, there is no evidence that SMIS marine ice deforms more
easily than meteoric ice. Even though total strains at the meteoric and marine ice core sites are not equal, annual strain rates are in the order of $x10^{-4}$ and the different ice types have similar minimum ages (of several thousand years). This makes their microstructural response to strain comparable. Meteoric ice shows stronger circle girdle fabrics, development of a vertical SPO and a decrease in its mean grain size with increasing total vertical extension and shear strain to 20% and 60% respectively downflow. The development of a circle girdle fabric and large increase in total strain downflow at the meteoric ice sites suggests that meteoric ice microstructure is preferentially oriented for horizontal compression as SMIS flows against shore. In contrast the marine ice microstructure is harder to deform in the ambient strain setting. The presence of marine ice thus could thus slow ice shelf dynamics and hence contribute to prolonging ice shelf life.

This study relates ice shelf surface melting to basal marine ice accretion in a cold-based ice shelf cavity and the presence of marine ice to decelerated shelf ice deformation. Thus, knowledge gained in this study contributes to a better assessment of the behaviour of heterogenic ice shelves. In a changing climate, ocean circulation patterns and atmospheric conditions will change and it is important to understand current ice shelf behaviour in order to make sound predictions of their future buffering capability of land ice.
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<th>Abbreviation</th>
<th>Definition</th>
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<tr>
<td>$A$</td>
<td>Constant in Glen’s flow law</td>
<td>-</td>
</tr>
<tr>
<td>$\alpha_{\text{eff}}$</td>
<td>Effective fractionation coefficient</td>
<td>‰</td>
</tr>
<tr>
<td>$\alpha_{\text{equ}}$</td>
<td>Equilibrium fractionation coefficient</td>
<td>‰</td>
</tr>
<tr>
<td>AVA</td>
<td>Achsenverteilungsanalyse</td>
<td>-</td>
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<tr>
<td>$c$</td>
<td>Isotopic concentration</td>
<td>‰</td>
</tr>
<tr>
<td>CDW</td>
<td>Circumpolar Deep Water</td>
<td>-</td>
</tr>
<tr>
<td>$D$</td>
<td>Diffusion coefficient</td>
<td>m² s⁻¹</td>
</tr>
<tr>
<td>$\dot{\varepsilon}$</td>
<td>Strain</td>
<td>-</td>
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<tr>
<td>EAD</td>
<td>Equal area dimension</td>
<td>-</td>
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<tr>
<td>FAME</td>
<td>Fabric Analyser based Microstructure Evaluation</td>
<td>-</td>
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<tr>
<td>GPR</td>
<td>Ground penetrating radar</td>
<td>-</td>
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<tr>
<td>HSSW</td>
<td>High Salinity Shelf Water</td>
<td>-</td>
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<td>ISW</td>
<td>Ice Shelf Water</td>
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<tr>
<td>IQR</td>
<td>Interquartile range</td>
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<tr>
<td>LED</td>
<td>Light emitting diodes</td>
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<td>MCDW</td>
<td>Modified Circumpolar Deep Water</td>
<td>-</td>
</tr>
<tr>
<td>MCM4</td>
<td>McMurdo Station</td>
<td>-</td>
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<tr>
<td>$n$</td>
<td>Constant in Glen’s flow law</td>
<td>-</td>
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<tr>
<td>NGG</td>
<td>Normal grain growth</td>
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<td>PISM</td>
<td>Parallel Ice Sheet Model</td>
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<td>RXX</td>
<td>Rotation recrystallization</td>
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<td>SMIS</td>
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<td>$T$</td>
<td>Time</td>
<td>S</td>
</tr>
<tr>
<td>$\tau$</td>
<td>Stress</td>
<td>Pa</td>
</tr>
<tr>
<td>TDS</td>
<td>Total dissolved solids</td>
<td>-</td>
</tr>
<tr>
<td>$V$</td>
<td>Growth rate</td>
<td>m s⁻¹</td>
</tr>
<tr>
<td>$V_f$</td>
<td>Freezing velocity</td>
<td>m s⁻¹</td>
</tr>
<tr>
<td>$Z$</td>
<td>Distance from the ice-water interface into the fluid</td>
<td>m</td>
</tr>
<tr>
<td>$z_{bl}$</td>
<td>Boundary layer thickness</td>
<td>m</td>
</tr>
</tbody>
</table>
1. Introduction: Significance of marine ice in ice shelves

This section provides a rationale for this study. It gives an overview of the importance of ice shelves in regulating the transfer of ice masses into the ocean contributing to sea level rise. It summarizes current knowledge on marine ice properties and formation mechanisms and highlights research gaps. Furthermore, it presents the current knowledge on marine ice influence on ice shelf stability. At the end of this chapter the thesis aims and objectives are introduced. The thesis outline and a statement of the contribution of individuals are also presented.

1.1. Introduction

The Antarctic continent comprises ice masses of around 27 million km$^3$ that could rise global sea level by a total of 58.3 m, whereby the unstable West Antarctic ice sheet holds a sea level equivalent of 4.3 m (Fretwell and others, 2013). Ice melt in Antarctica has already contributed ~6 mm in the last two decades (Vaughan and others, 2013). Most ice masses in Antarctica terminate in ice shelves (Bindschadler and others, 2011), which thus regulate the transfer of land ice into the ocean. Ice shelves are situated at the intersection of land, air and ocean. They occupy larger bays and are fed by ice streams and glaciers that flow into the ocean. Ice shelf behavior is thus influenced by local terrain, air temperature, precipitation and ocean circulation. Not all ice shelves are just made up of ice draining from land (i.e. meteoric ice) but some have a heterogeneous composition of meteoric ice and so-called marine ice (Tison and others, 1993, Treverrow and others, 2010), which is accreted from the ocean below. Formed from primarily ocean water rather than snow, marine ice has a different chemical composition and structure than meteoric ice (Tison and others, 1993). However, due to the difficulty in accessing this ice, few studies have investigated marine ice itself and thus little is known about how marine ice forms and behaves. Ice crystallographic alignment and temperature as well as chemical impurities (salt and dust) have been shown to influence ice flow on land (Morgan and others,
1998). However, little is known about marine ice properties and whether the presence of marine ice in ice shelves influences ice shelf deformation and stability.

### 1.2. Ice shelves in the Polar Regions

Ice shelves are thick floating masses of predominantly glacier ice attached to land. They are commonly several hundred metres thick, occupy embayments and occur at the fringes of the Antarctic continent (Bindschadler and others, 2011), and much less commonly in the Arctic (e.g. Williams and Dowdeswell, 2001, Jeffries, 2002). Ice shelves drain 74% of Antarctica’s grounded ice together with outlet glaciers (Bindschadler and others, 2011). They buffer the flow of glaciers and ice streams (Dupont and Alley, 2005, Gudmundsson, 2013) due to lateral friction from the embayment sides or local grounding points that create a backstress (Matsuoka and others, 2015, Berger and others, 2016). If an ice shelf disappears, the glaciers that originally fed the ice shelf speed up (e.g. De Angelis and Skvarca, 2003, Scambos and others, 2004). This causes an accelerated ice mass transfer from land to ocean and eventually sea level rise. Melt of floating ice shelves only has a negligible effect on sea level rise since they already displace ocean water (Jenkins and Holland, 2007).

Due to the enormous ice mass on the Antarctic continent, which feeds the ice shelves, and the low temperature of the surrounding oceans, which prevents the ice from melting, many Antarctic ice shelves are very large; the Ross Ice Shelf, for instance, has a surface area that approximates the size of France. Since ice shelves are floating on the ocean without any basal friction, their flow velocities are also very large, approaching 1000 m per year for many ice shelves including the large Ross and Filchner-Ronne Ice Shelves (Figure 1.1).
Figure 1.1: Flow velocities of Antarctic glacier ice (Rignot and others, 2011).

Antarctic ice shelves gain mass not only through dynamic inflow from glaciers and ice streams but also through local accumulation of snow on the ice shelf surface (Rignot and others, 2013). Ice shelves lose mass through sublimation and wind drift on its surface and ice berg calving and basal melting (Hooke, 2005, Cuffey and Paterson, 2010). Basal melting has recently been determined to be the dominant process of ablation of Antarctic ice shelves (Depoorter and others, 2013, Rignot and others, 2013). However, at ice shelves that experience enhanced basal melt, ice berg calving events also increase in frequency (Liu and others, 2015). Ice
shelves on the Antarctic Peninsula especially lose their mass through both, basal melting (Rignot and others, 2013) and ice berg calving (Liu and others, 2015) (Figure 1.2).

Ice shelves that lose mass by basal melting are reached by warmer ocean currents that enter into their cavities (Figure 1.3). Thus they are called warm-cavity ice shelves (Rignot and others, 2013). Most of these warm-cavity ice shelves can be found in the South East Pacific draining glaciers on the Antarctic Peninsula or the West Antarctic ice sheet (Rignot and others, 2013). Fewer but larger warm-based ice shelves can also be found in East Antarctica, such as the Totten Glacier (denoted by 'T' in Figure 1.3, Greenbaum and others, 2015).

Figure 1.2: Basal melt rate of Antarctic ice shelves and outlet glaciers and melt/calving percentages (Rignot and others, 2013).
Even if an ice shelf partially melts on its underside, ice shelves also frequently accrete basal mass locally (Fricker and others, 2001, Pritchard and others, 2012, Moholdt and others, 2014). Ice shelves that accrete mass in some areas and lose mass in others have been called cold-cavity ice shelves (Rignot and others, 2013). The accreting ice, which forms from a mixture of sea water and fresh water ice, is called marine ice (Tison and others, 1993). The location of marine ice accretion is dependent on local ocean circulation and the bottom topography of the ice shelf. Ocean circulation into ice shelf cavities is often not well understood due to a lack of observations below ice shelves including a detailed bathymetry. Small obstructions on the ocean floor (e.g. Greenbaum and others, 2015) could deflect dense, high salinity currents preventing them from reaching deep into the ice shelf cavity. Nonetheless, it is known that
marine ice can accrete in open basal structures, such as rifts and crevasses (Morgan, 1972, Souchez and others, 1991, Oerter and others, 1992, Tison and others, 1993, Eicken and others, 1994, Khazendar and others, 2001, Craven and others, 2004, 2009, Pattyn and others, 2012), suture zones (Jansen and others, 2013, Kulesza and others, 2014) or as a massive layer below meteoric shelf ice (Craven and others, 2009, Treverrow and others, 2010, Zotikov and others, 1980). Marine ice can thus locally reach a thickness of ~200m (e.g. Craven and others, 2009) (Figure 1.4).

1.3. Marine ice occurrence and detection

Marine ice was first observed in Antarctic ice shelves in deep or shallow ice cores (Zotikov and others, 1980, Souchez and others, 1991, Tison and others, 1993, Eicken and others, 1994, Khazendar and others, 2001, Craven and others, 2009, Pattyn and others, 2012, Dierckx and others, 2014) extracted from ice shelf rifts, or from the ice shelf surface where surface ablation processes caused erosion of the originally overlying meteoric ice layers. Rarely marine ice is extracted in deep ice shelf cores (e.g. Zotikov and others, 1980, Oerter and others 1992, Craven and others, 2004).

In recent years improved remote sensing methods have allowed for a remote detection of marine ice below entire ice shelves (Fricker and others, 2001, Joughin and Vaughan, 2004, McMahon and Lackie, 2006, Khazendar and others, 2009, Pattyn and others, 2012, Jansen and others, 2013). Generally a ground penetrating radar (GPR) or airborne radar system is employed that images the thickness of the meteoric ice layer whilst the radar signal gets lost in the marine ice layer of higher emissivity (due to its increased salinity/moisture). When the exact height of the ice shelf surface is also known (e.g. through GPS, laser or a detailed digital elevation model), the marine ice thickness can be calculated from the hydrostatic equilibrium (e.g. Fricker and others, 2001, Janssen and others, 2013). Like this, marine ice distribution in large rifts or suture zones of the Amery and Larsen C Ice Shelf was calculated (Fricker and others, 2001, Janssen and others, 2013). Despite these remote sensing methods, not all marine ice in Antarctica has been mapped.
1.4. Marine ice formation mechanisms

Due to the difficulty associated with accessing the ice shelf cavity, processes of marine ice formation, including timing, location and volume, remain largely unknown. Since marine ice can be several hundred meters thick (e.g. Amery Ice Shelf, Craven and others, 2009), fast forming frazil ice crystals are assumed to be mainly responsible for the generation of marine ice (Treverrow and others, 2010), allowing for marine ice accumulation rates of $> 1 \text{ m a}^{-1}$ at some ice shelves (Bomboch and Jenkins, 1995, Wen and others, 2010). Indeed, loose agglomerations of frazil ice crystals have been observed below ice shelves in borehole imagery (Craven and others, 2005, Hubbard and others, 2012). These small freely floating discoid ice crystals nucleate in supercooled water (Martin, 1981, Daly, 1984), which is water cooled below its in situ freezing point without changing state (Leonard and others, 2014). Supercooling of water masses below ice shelves can occur as a result of a change in the pressure-dependent freezing point due to adiabatically rising water masses (Foldvik and Kvinge, 1974) or double diffusion of heat and salt between water masses of different salinities at their freezing point (Souchez and others, 1998).
Both mechanisms generally involve mixing of sea water with fresher meltwater. The former process is part of a thermohaline circulation often referred to as the ‘ice pump’ mechanism (Lewis and Perkin, 1986) and has been widely associated with marine ice formation in thick layers (e.g. Galton-Fenzi and others, 2012). Hereby, continental fresh water ice is melted at depth close to the grounding line of an ice shelf by dense High Salinity Shelf Water (HSSW). This HSSW, which is generated during sea ice formation in winter (Figure 1.5), is warmer than the local pressure-dependent freezing point and warmer than the continental ice at the grounding line. Similarly, warmer Circumpolar Deep Water (CDW) or Modified Circumpolar Deep Water (MCDW) can enter the ice shelf cavity inducing melt at the ice shelf base (Jacobs and others, 1992). This meltwater then rises along the gradient of the ice shelf due to its buoyancy to shallower waters, where it becomes supercooled due to a rise in the pressure-dependent freezing point and frazil ice crystals nucleate (Galton-Fenzi and others, 2012, Figure 1.5).

Since water masses often become supercooled during uplift to shallower areas with a higher pressure-dependent freezing point, no marine ice has been observed (or modelled) to accrete close to ice shelf grounding lines. Marine ice generally accretes downstream in central or thinner frontal parts of the ice shelf (e.g. Filchner-Ronne and Ross Ice Shelves; Zotikov and others, 1980, Lambrecht and others, 2007), or in suture areas (e.g. Larsen C and Amery Ice Shelves; Fricker and others, 2001, Kulessa and others, 2014) or downstream of pinning points.
(e.g. Filchner-Ronne Ice Shelf; Lambrecht and others, 2007). Ice shelves with deeper grounding lines (and thus a steeper basal gradient) (e.g. Filchner-Ronne Ice Shelf) accrete more marine ice than ice shelves with shallower grounding lines (e.g. Ross Ice Shelf) (Depoorter and others, 2013).

If supercooled water extends to the ocean floor, frazil ice crystals can also flock together on the sea floor forming anchor ice (Mager and others, 2013) where the ocean floor is relatively shallow (a few tens of meters, Leonard and others, 2014). This ice can eventually lift off the ocean floor when its buoyancy is large enough and contributes to marine ice at the ice shelf base (e.g. Swithinbank, 1970).

Double diffusion-induced supercooling can occur close to the grounding line of shallower ice shelves, where surface meltwater of meteoric ice can percolate through sediment (Souchez and others, 1998) or drain through tide cracks (Gow and others, 1965, Gow and Epstein, 1972) to the ice-water interface. Heat diffuses faster than salt from the relatively warmer meltwater of the meteoric ice to the colder sea water (Martin and Kauffman, 1974). This results in rapid freezing of the less saline water mass, which then accretes at the base of the ice shelf. More recently McPhee and others (2013) suggested that two water masses near their salinity-dependent freezing point can become turbulently mixed by a shallow tidal circulation, which allows for frazil ice crystal nucleation in a double diffusion process. Over time frazil ice crystals gradually sinter together and adjust their shape to minimize the surface energy of the ice crystals (Martin, 1981) forming massive marine ice.

Due to difficulty in accessing the ice shelf cavity, none of these marine ice formation processes have been observed in situ. Whilst supercooled water has been observed below ice shelves and in front of them, the origin of water masses that mix to form this supercooled water is poorly understood. It could be made up of a mixture of sea water and meltwater from less saline meteoric or marine ice, which would have been melted at the underside of ice shelves by either MCDW, Shallow Ice Shelf Water (SISW), or a shallow tidal circulation. Surface meltwater could also drain to the underside of ice shelves.
1.5. Marine ice influence on ice shelf stability

Marine ice occurs primarily in cold-cavity ice shelves, which have a broad ice shelf area that can disintegrate without compensating for the buffering effect of the ice shelf, the so-called ‘ice shelf safety band’ (Furst and others, 2016). Not least by adding mass to the ice shelf and infilling weaknesses, the presence of marine ice likely influences ice shelf stability (e.g. Kulessa and others, 2014). Marine ice can thus prevent ice shelf disintegration. The cause for the sudden disintegration of entire ice shelves (e.g. Scambos and others, 2003, 2009, Cook and Vaughan, 2010) still remains poorly understood. Whilst rising air temperatures in the region could be the principle driver for disintegration (e.g. Vaughan and others, 2001, Domack and others, 2005), they do not explain its mechanics. Draining meltwater features on ice shelves may be the most important contributing factor to ice shelf disintegration through hydrofracture (e.g. Banwell and MacAyeal, 2015). Changes in the geometry of the ice front, especially the development of an arch that is bowed inward, may also contribute to ice shelf instability (Doake and others, 1998, Rack and others, 2000). Ocean-driven melting and associated ice shelf thinning may also make ice shelves more susceptible to collapse (Shepherd and others, 2003).

If all ice shelves, including the larger Ross and Filchner-Ronne Ice Shelf, disintegrated, the large Antarctic ice sheets would drain into the ocean much faster (Golledge and others, 2015) where they would melt and contribute to sea level rise. Of particular concern is the ice in West Antarctica, which is grounded below sea level due to low lying bedrock (Fretwell and others, 2013). The West Antarctic marine ice sheet is inherently unstable (Schoof, 2007) and fast retreat of ice shelf grounding lines, especially if resting on an upward sloping bed, could lead to significant sea level rise in a short amount of time (e.g. Favier and others, 2014). However, calculations have shown that the presence of ice shelves would significantly slow down this grounding line retreat (Gudmundsson, 2013). It remains unknown to what extent the presence of marine ice enhances ice shelf stability and prolongs ice shelf life, slowing down ice mass transfer to the ocean.

Ice shelves are only rarely well represented in projections of Antarctica’s future ice mass loss (e.g. Cornford and others, 2015, Feldmann and Levermann, 2015, Golledge and
Some melt prognoses of Antarctic ice consider the effect of ice shelves on ice drainage; nonetheless, often only thinning due to oceanic processes is considered (e.g. Cornford and others, 2015) and basal ice accretion is neglected. Some studies employing the Parallel Ice Sheet Model (PISM) also allow for a changing ice shelf geometry and local sub-ice shelf adfreezing dependent on a change in the pressure-dependent freezing point (Golledge and others, 2015, Feldmann and Levermann, 2015). However, the localised circulation below ice shelves is still poorly modelled, and hence marine ice accretion is still poorly represented. Furthermore it remains unknown what effect the presence of marine ice has on shelf ice deformation. Given that marine ice has microstructure that differs from meteoric ice and a relatively higher temperature (Treverrow and others, 2010), its presence likely influences ice shelf rheology (Khazendar and others, 2009). Considering differences in ice crystal orientation in meteoric ice improves the accuracy of ice flow simulations (Zwinger and others, 2014). However, how the presence of marine ice influences ice shelf flow and behaviour remains poorly known.

1.6. The need for further research

Whilst ice shelves play an essential role in regulating the transfer of land ice into the ocean, ice shelves are still poorly represented in prognoses of future Antarctic ice sheet evolution because current ice shelf behavior remains poorly understood. This is largely due to a lack of \textit{in situ} data that detail processes at the ice shelf boundaries (air-ice shelf and ice shelf-ocean). Processes occurring in the ice shelf cavity remain especially poorly understood due to the difficulty in accessing it. Ocean circulation exerts a large control on ice shelf health. Ice shelves that are cold-based show at least partial accretion of basal mass. This basally accreted marine ice has been observed in structural weaknesses and is thought to enhance ice shelf stability. Nonetheless, how marine ice influences ice shelf dynamics remains unknown to date. Furthermore it remains unknown whether marine ice will continue to form as ocean circulation and air temperature change. Through investigations of marine ice samples, this study will infer marine ice formation and deformation processes at the Southern McMurdo Ice Shelf (SMIS) in
Antarctica. The gained knowledge will contribute to better assessments of future ice shelf behavior.

### 1.6.1. Thesis aim and objectives

The aim of the thesis is to improve the understanding of marine ice formation processes and its behaviour during deformation. Marine ice chemistry and microstructure are investigated and compared to meteoric ice properties at SMIS. Both ice types crop out at the surface due to localized surface ablation, so ice extraction is possible along a flowline. Additionally, annual surface strain is measured and changes in ice microstructure are related to the changing strain regime downflow. The study site thus provides a unique opportunity to investigate and compare marine and meteoric ice deformation *in situ*. To meet the aim of the thesis the following research questions and objectives will be addressed:

1. **What is the chemical composition of source water masses that mix to form marine ice at SMIS?**
   1.1. Measure marine ice chemical composition (water isotopes and major ions) in shallow ice cores to establish whether there are spatiotemporal variations.
   1.2. Simulate marine ice isotopic composition running a frazil ice boundary layer freezing model with several different potential water sources and compare simulation results to the measured SMIS marine ice chemical composition to establish source water composition.

2. **What is the origin of water masses for marine ice formation at SMIS?**
   2.1. Investigate the derived SMIS marine ice source water composition in light of its geographical context to infer possible formation processes.

3. **How does marine ice microstructure become altered in a natural setting of increasing total strain?**
   3.1. Measure ice shelf surface strain and marine ice microstructure (ice crystal fabric, shape and size) in ice extracted along a flowline of increasing total strain. Relating the
microstructure to the strain regime investigate marine ice microstructure with increasing total strain.

4. Is there evidence that SMIS marine ice deforms more easily than meteoric ice?

4.1. Assess differences in marine ice and meteoric ice microstructure evolution with changes in total strain.

4.2. Relate the ice microstructure to the strain regime to determine whether marine and/or meteoric ice experience tertiary strain (steady state deformation).

5. Does the presence of marine ice promote ice shelf longevity at SMIS?

5.1. Discuss whether SMIS processes of marine ice formation are contributing to ice shelf sustenance.

5.2. Considering measured marine ice microstructure, discuss whether marine ice would slow down SMIS dynamics.

1.6.2. Thesis outline

Chapter 1 provides a literature background. In particular it presents the significance of ice shelves in Antarctica followed by a summary of literature on marine ice occurrence and detection. The current understanding of marine ice formation processes and the role of marine ice in ice shelves are presented. Thesis aims and objectives are stated, a thesis outline is given and contributors are listed.

Chapter 2 addresses research questions 1, 2 and 5. It reviews previous studies that measured marine ice chemical composition in ice samples. Subsequently marine ice chemical composition and crystal morphology are presented for data from three SMIS ice cores. Isotopic ice composition is compared to simulations of a theoretical frazil ice crystal freezing model. Possible marine ice formation processes are discussed in particular with regard to source water composition and origin. The possibility of local marine ice recycling (melting and refreezing) is evaluated.
Chapter 3 addresses research questions 3, 4 and 5. It reviews the influence of strain on ice deformation and how dynamic recrystallization influences ice microstructure (ice fabric, crystal size and shape). Previous studies on shelf ice microstructure are summarized for meteoric and marine ice. The measured surface strain regime at SMIS is presented. Meteoric and marine ice microstructure evolution along a flowline of increasing total strain is evaluated. Marine ice meteoric microstructure are compared to evaluate whether marine ice deforms more easily than meteoric ice and whether the presence of marine ice influences ice shelf dynamics.

Chapter 4 summarizes the findings of this study and states implications for SMIS behaviour and general ice shelf behavior. Suggestions for future research are given.

Appendix A shows the classified ice facies in thin section of all ice cores.

Appendix B shows the MATLAB code used to calculate accumulated strain and strain rates at the ice core sites.

Appendix C refers to existing and future publications and presentations resulting from this thesis. It includes an authorized reprint of an existing publication in the Journal of Glaciology on marine ice recycling at SMIS addressing the thesis research questions 1, 2, and 5 (Koch and others, 2015).

1.6.3. Statement of contributions of individuals

Inka Koch, University of Otago: Principle investigator and author. Responsible for research design, data acquisition through field work (in 2010 and 2011) and laboratory work, data analysis, manuscript design and writing.
**Sean Fitzsimons, University of Otago:** Primary supervisor, guidance and assistance regarding study design and execution, Principle Investigator for SMIS fieldwork in 2007, 2010 and 2011, general guidance and advice throughout the whole project, review of the thesis and manuscripts, grant application support for exchanges to Belgium, financial support for laboratory work, shipment of ice to Belgium and publication costs.

**Jean-Louis Tison, University of Brussels:** External advisor, assistance with study design, especially with regard to investigating the source water question for marine ice and application of the freezing model, facilitated and funded laboratory work in a walk in freezer in Belgium, guidance throughout the thin section analysis especially with regard to ice facies, contribution to and review of manuscript presented in chapter 2.

**Nicolas Cullen, University of Otago:** Secondary supervisor, guidance regarding study design especially with regard to framing research questions and ensuring consistency in written pieces, general support and motivation throughout the whole project, thesis and manuscript review, field assistance in 2010.

**Denis Samyn, Royal Museum of Central Africa:** Extracted the long marine ice core in the field in 2007, advice with regard to thin section processing, contributing author to chapter 2.

**Adam Treverrow, Antarctic Climate and Ecosystems Cooperative Research Centre:** Assisted with ice shelf strain calculations and gave advice regarding the interpretation of meteoric and marine ice microstructure, review of chapter 3.

**Andrew Clifford, University of Otago:** Gave advice with regard to fieldwork on SMIS. Allowed use of displacement stake data from field previous seasons (2002, 2003 and 2005).
2. Marine ice formation at SMIS

This chapter presents research published in slightly modified form in the Journal of Glaciology (Koch and others, 2015) which can be found in Appendix C. It addresses research questions 1, 2 and 5 (see section 1.6.1.). Previous research on marine ice composition is summarized initially. Subsequently, SMIS marine ice chemistry (water isotopes and major ions) is compared to the output of a frazil ice freezing model to determine marine ice source water composition. Possible origin of the source water and associated formation processes are discussed. For co-authors see section 1.6.3.

2.1. Introduction

Marine ice samples are characterized by an ice chemistry and crystal morphology that differs from that of meteoric ice and sea ice (Table 2.1). Marine ice is more enriched in heavy water isotopes than meteoric ice and slightly saline (e.g. Tison and others, 1993, Craven and others 2004, Pattyn and others, 2012). In comparison to sea ice, marine ice is one order of magnitude less saline (Tison and others, 1993) due to its different formation and consolidation process. Marine ice has an isotopic signature that can be more or less enriched than sea water (e.g. Souchez and others, 1995) dependent on its source water, whereas sea ice is generally more enriched than sea water (e.g. Smith and others, 2012). The first ever collected marine ice samples from below the Ross Ice Shelf showed vertically elongated columnar ice crystals (Zotikov and others, 1980), but so-called ‘banded’, ‘granular’ and ‘platelet’ ice crystal facies have also been observed (Table 2.1). Banded ice crystal facies are uniquely found in only marine ice (Tison and others, 1993, Treverrow and others, 2010), whereby columnar and platelet ice crystal facies also occur in sea ice (Gough and others, 2012, Langhorne and others, 2015) and granular ice crystal facies are also found in meteoric or sea ice (Montagnat and others, 2014a, Gough and others, 2012).
Table 2.1: Occurrence of ice crystal facies measured range in $\delta^{18}$O and salinity as well as co-isotopic slopes of marine ice in previous studies (gr = granular, bd = banded, clm = columnar, pl = platelet).

<table>
<thead>
<tr>
<th>Ice shelf</th>
<th>Ice crystal facies</th>
<th>$\delta^{18}$O (‰)</th>
<th>Salinity</th>
<th>Co-isotopic slope</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amery Ice Shelf</td>
<td>--</td>
<td>0.00 to 2.30</td>
<td>0.02</td>
<td>--</td>
<td>Morgan, 1972, cited in Goodwin, 1993</td>
</tr>
<tr>
<td>&quot;</td>
<td>gr, bd</td>
<td>-0.60 to 1.50</td>
<td>0.06 to 0.75</td>
<td>--</td>
<td>Craven and others, 2004</td>
</tr>
<tr>
<td>&quot;</td>
<td>gr, bd</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>Treverrow and others, 2010</td>
</tr>
<tr>
<td>Campbell Glacier Tongue</td>
<td>Gr</td>
<td>-17.00 to ~1.00</td>
<td>--</td>
<td>7.86 (r=0.998)</td>
<td>Souchez and others, 1995</td>
</tr>
<tr>
<td>Föhrner Ronne Ice Shelf</td>
<td>Gr</td>
<td>~2.00</td>
<td>&lt; 0.10</td>
<td>--</td>
<td>Oerter and others, 1992</td>
</tr>
<tr>
<td>&quot;</td>
<td>Gr</td>
<td>--</td>
<td>0.02 to 0.10</td>
<td>--</td>
<td>Eicken and others, 1994</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>clm, pl, gr</td>
<td>1.00 to 3.20</td>
<td>0.01 to 0.80 (Na content only)</td>
<td>6.60 (r=0.93, marine ice &amp; sea water)</td>
<td>Souchez and others, 1991</td>
</tr>
<tr>
<td>&quot;</td>
<td>clm, pl, gr, bd</td>
<td>1.14 to 3.26</td>
<td>0.03 to 1.42</td>
<td>--</td>
<td>Tison and others, 1993</td>
</tr>
<tr>
<td>&quot;</td>
<td>clm</td>
<td>1.14 to 2.02</td>
<td>1.35</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>&quot;</td>
<td>Pl</td>
<td>2.08 to 2.51</td>
<td>1.42</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>&quot;</td>
<td>Gr</td>
<td>1.82 to 3.26</td>
<td>0.03 to 0.14</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>&quot;</td>
<td>Bd</td>
<td>1.64 to 3.02</td>
<td>0.19 to 0.36</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>&quot;</td>
<td>--</td>
<td>~15.00 to ~3.00</td>
<td>--</td>
<td>7.71 (r=0.99)</td>
<td>Souchez and others, 1998</td>
</tr>
<tr>
<td>Koettlitz Ice Tongue</td>
<td>Pl</td>
<td>1.37 to 2.51</td>
<td>0.20 to 5.26</td>
<td>--</td>
<td>Gow and Epstein, 1972</td>
</tr>
<tr>
<td>Law Dome – type 1</td>
<td>clm</td>
<td>0.26 to 1.56</td>
<td>0.01 to 0.08 (Na content only)</td>
<td>8.10</td>
<td>Goodwin, 1993</td>
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<tr>
<td>Law Dome – type 2</td>
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<td>-16.93 to -2.57</td>
<td>&quot;</td>
<td>7.80</td>
<td>Goodwin, 1993</td>
</tr>
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<td>Northern McMurdo Ice Shelf</td>
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<td>--</td>
<td>~0.01 to ~0.90</td>
<td>--</td>
<td>Gow and others, 1965</td>
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<tr>
<td>Southern McMurdo Ice Shelf</td>
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<td>-2.8 to 2.9</td>
<td>--</td>
<td>--</td>
<td>Kellogg and others, 1991a</td>
</tr>
<tr>
<td>Southern McMurdo Ice Shelf</td>
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<td>~4.40 to ~2.50</td>
<td>--</td>
<td>7.82 (r=0.99, meteoric &amp; marine ice)</td>
<td>Fitzsimons and others, 2012</td>
</tr>
<tr>
<td>Nansen Ice Shelf</td>
<td>gr, bd</td>
<td>1.80 to 2.37</td>
<td>0.04 to 0.15</td>
<td>--</td>
<td>Khazendar and others, 2001</td>
</tr>
<tr>
<td>&quot;</td>
<td>gr, bd</td>
<td>--</td>
<td>0.03 to 0.23 (inferred from Cl measurements)</td>
<td>--</td>
<td>Dierckx and others, 2013, 2014</td>
</tr>
<tr>
<td>Ross Ice Shelf</td>
<td>clm</td>
<td>--</td>
<td>~2.00 to ~4.00</td>
<td>--</td>
<td>Zotikov and others, 1980</td>
</tr>
<tr>
<td>Roi Baudouin Ice Shelf rift</td>
<td>gr, bd</td>
<td>~0.00 to ~2.20</td>
<td>~0.03 to ~9.20</td>
<td>--</td>
<td>Pattyn and others, 2012</td>
</tr>
</tbody>
</table>
2.2. Chapter objectives

Inferences can be made about the source water composition of marine ice from its chemistry. Due to a lack of ice shelf cavity observations, water masses that mix to form marine ice have never been measured. In this chapter the ice crystal morphology and chemistry of marine ice from different sites was analyzed in conjunction with a boundary layer frazil ice freezing model to derive source water composition. The data are interpreted in light of the geographic setting to determine spatiotemporal variations and origin of the source water. Results help to better understand marine ice formation processes at SMIS.

2.3. Sampling and data analysis

In this section we describe the field site and give a summary of previous studies. We also detail marine ice extraction from the ice shelf and its laboratory analysis for water isotope and major ion composition. The frazil ice boundary layer freezing model, which is used to compare theoretical and measured water isotope signals of marine ice, is explained. Furthermore, thin sectioning of marine ice and the classification of ice crystal morphology into different ice facies are described.

2.3.1. Field site

SMIS is a small ice shelf (~30 x 35 km²) in Antarctica confined by Minna Bluff to the south and Black and White Islands to the north (Figure 2.1). SMIS is separated from the much larger and two magnitudes faster flowing Ross Ice Shelf by a rift zone (Figure 2.1). SMIS flows slowly at a rate of 0.4 to 7.3 m a⁻¹ in a WSW direction toward Minna Bluff in its southern parts and in a WNW direction in its north western parts (Clifford, 2005). The ice shelf floats on a water column of 300-400 m in its centre (Johnston and others, 2008). Ground penetrating and airborne radar investigations revealed that the ice shelf is thickest in the north (~180 m) and thins (to ~100 m) toward its eastern and southern margins (Swithinbank, 1970, Clifford, 2005), where the radar signal became lost 5 to 6 km from shore. This was ascribed to outcropping of slightly saline marine ice at the snow-free ice shelf surface close to the shore of Minna Bluff (Figure
The presence of marine ice was also detected in previous studies on ice composition (Kellogg and others, 1991a, Fitzsimons and others, 2012) and inferred from the presence of marine macrofossils at the ice shelf surface (Debenham, 1919, 1965, Gow and others, 1965, Swithinbank, 1970, Fitzsimons and others, 2012, Kellogg and others, 1990, Kellogg and others, 1991b) such as shells (Figure 2.2). Two shells were radiocarbon dated to 1230 ± 50 and 2850 ± 30 radiocarbon years BP, respectively (Denton and Marchant, 2000, Kellogg and others, 1990) using an Antarctic reservoir correction of 1300 years (Berkman and Forman, 1996). However, their age does not accurately reflect the date of marine ice accretion, since the time of death of the marine organisms is not necessarily related to their entrainment date (Fitzsimons, 1997).

Figure 2.1: Landsat image mosaic Antarctica showing the location of the Southern McMurdo Ice Shelf (SMIS) in McMurdo Sound, Antarctica. Inset shows a detailed view of the marine ice sampling sites close to Minna Bluff. Marine ice occurs in a zone of relatively darker ice at SMIS, which is separated from the lighter meteoric ice by a distinct boundary running parallel to shore (2 to 3.5 km away from shore) (see Fitzsimons and others, 2012, Fig. 1).
Surfacing of marine ice, which was originally accreted at SMIS’s base, is speculated to be a result of stripping of ice by katabatic winds (Clifford, 2005), like at the Hell’s Gate Ice Shelf (Tison and others, 1993). Modelled wind fields indeed show elevated wind speeds over the southern part of SMIS (Monaghan and others, 2005). Local summer surface melting also contributes to surface mass loss (>0.10 m, Clifford, 2005) in the snow-free band running parallel to Minna Bluff (Figure 2.1). Melting is likely partially induced by the locally high debris concentration on the ice shelf surface close to Minna Bluff (Denton and Marchant, 2000) and the lower surface albedo of the darker marine ice (Warren and others, 1997). Surface meltwater pools in several ice shelf surface lakes (Clifford, 2005), which are often elongated and oriented at right angle to shore (Figure 2.1) parallel to the prevailing wind direction (Swithinbank, 1970). These lakes were observed to form in summer (Figure 2.4), whilst the rest of the year these lakes were frozen over into “mirror-smooth” surfaces of ice (Swithinbank, 1970). The lakes are estimated to be ~1 m deep, similar to lakes observed on other ice shelves (e.g. Banwell and others, 2014), several tenths to hundreds of metres wide and up to 1 to 2 km long. Some lakes were observed to drain completely during the course of the melt season.
2.3.2. Ice sampling in the field

Marine ice was extracted in shallow cores from the apex of snow-free ~6 m high ice ridges that separate surficial lakes (Figure 2.5) and are oriented at right angles to shore, almost in a perfect north-south direction (Figure 2.3). Cores C5 and C9 were taken in November 2010 with a Kovacs corer (Figure 2.6). Denis Samyn and others extracted core C15 in December 2007 with a custom made SIPRE type coring auger. The non-successive labeling stems from the fact that series of ice core extraction attempts were made but not all were successful. The extracted three ice cores presented in this chapter (Figure 2.1) were 2.71 ± 0.01 m (C5), 3.04 ± 0.01 m (C9) and 9.49 m ± 0.01 m (C15) long. The freeboard level was not reached during ice core extraction since the ice shelf is ≥ 100 m thick (Clifford, 2005). The top 0.50 ± 0.01 m of every core were discarded to avoid the influence of potential surface melt (Tison and others, 1993). Ice cores were immediately put in a freezer and kept below -15 °C during transport and storage.
Figure 2.4: Field researcher ‘testing’ the depth of one surface lake in early December 2010. Lakes form between the ridges visible in Figure 2.3. Photo taken looking north-east.

Figure 2.5: Ice core extraction on the edge of a marine ice ridge. Distinct layers are visible in the marine ice. The extended rods measure around six meters.
2.3.3. Solutes and water isotopes

Ice cores were cut into $0.10 \pm 0.001$ m sections along the length of the cores and a $0.0075 \pm 0.0025$ m thick outside rim of the ice cores was cut off to avoid contamination. Subsequently the samples were allowed to melt in closed plastic containers at room temperature. Water samples were filtered under vacuum using MF-Millipore 0.45 µm membrane cellulose acetate and cellulose nitrate filters. The samples were analysed using a Dionex ICS-3000 ion chromatograph to determine the concentration of major cations and anions ($\text{Li}^+$, $\text{Na}^+$, $\text{Mg}^{2+}$, $\text{K}^+$, $\text{Ca}^{2+}$, $\text{NH}_4^+$ and $\text{Cl}^-$, $\text{NO}_2^-$, $\text{Br}^-$, $\text{NO}_3^-$, $\text{SO}_4^{2-}$). A carbonate removal device (CRD-200) was installed to remove the carbonate peak. Precision and accuracy of cation and anion concentrations are better than 5%. The sum of all cations and anions gives the total dissolved solids (TDS), which are quoted in parts per thousand (‰) to allow for easy comparison with salinity measurements of other studies. Ions were measured in mg l$^{-1}$ and are quoted in % with respect to Standard mean ocean water (SMOW) ion ratios according to Maus and others (2011) (e.g. $\Delta\text{Mg/Cl} = (\text{Mg/Cl} - (\text{Mg/Cl})_{\text{SMOW}})/(\text{Mg/Cl})_{\text{SMOW}}$), whereby the SMOW ratios were taken from Millero and others (2008). Since individual frazil ice crystals are thought to expel all
salt during their formation (Tison and others, 2001), salt could only be included in the pore spaces of the agglomeration or at grain boundaries. Tison and others (1993) suggest that ion fractionation in marine ice is further evidence for the presence of sea water filled pores, which consolidate slowly, allowing for the selective incorporation of ions. Previous studies have interpreted variations in ion ratios to result from changes in the consolidation rate of marine ice (Oerter and others, 1992, Tison and others, 1993, Moore and others, 1994).

The relative concentration of oxygen and deuterium isotopes ($\delta^{18}O$ and $\delta D$) was measured with a Picarro laser spectrometer relative to SMOW (e.g. $\delta^{18}O = ((^{18}O/^{16}O - (^{18}O/^{16}O)_{SMOW})/ (^{18}O/^{16}O)_{SMOW}) \times 1000$, quoted in ‰). The samples were repeat injected eight times whereby the first three injections from one sample were discarded due to carry-over from previous samples. The precision of the measurements was 0.07‰ for $\delta^{18}O$ and 0.40‰ for $\delta D$. In order to establish whether marine ice at SMIS was formed from a constant or a changing mixed water source, the regression slope of co-isotopic plots of $\delta^{18}O$ and $\delta D$ was established and compared to co-isotopic freezing (Souchez and Jouzel, 1984) and mixing models (Souchez and Groote, 1985). Regression slopes in mixing models are around 8 (Souchez and Groote, 1985) whereas freezing models have smaller slopes (Souchez and Jouzel, 1984) whereby the exact slope magnitude depends on the isotopic composition of the water source.

Data trends and interrelations are assessed with the Pearson’s product-moment correlation coefficient and are expressed in terms of $r$. All $r$ values quoted in this paper are significant at the 99% level.

### 2.3.4. Ice crystal facies

Thin sections of ice were prepared according to the method detailed by Durand and others (2006) vertically along the full length of the ice cores to a thickness of $\sim 0.5 \times 10^{-3}$ m using a conventional biological microtome in a cold room maintained at -15 °C. Thin sections were photographed between cross-polarized light and ice crystals were classified into different ice facies (granular, banded and platelet) based on their shape using the criteria described in Table 2.2 for approximately every 0.10 m downcore coinciding with the sampling for the ice
chemistry. Where 30% - 70% of both, banded and granular ice crystals, were present in a 0.10 m long section of thin sections, the ice facies were classified as a mixed ice facies. On the rare occasion when the ice crystal morphology did not fit any of the descriptions, they are logged as “other”. Classified ice crystal facies in thin sections of the three marine ice cores can be found in Appendix A. Ice crystal morphology, as apparent in thin section, is analysed in conjunction with ice chemistry in order to determine whether variations in marine ice source water lead to changes in the appearance of marine ice crystals.

Table 2.2: Criteria for the classification of ice crystal facies with examples.

<table>
<thead>
<tr>
<th>Ice crystal facies</th>
<th>Individual grain shape</th>
<th>Grain boundaries</th>
<th>Crystal size (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Banded</td>
<td>Elongated, often rectangular, sometimes acicular</td>
<td>Polygonal with square edges, often diffuse</td>
<td>0.5 – 30.0</td>
</tr>
<tr>
<td>Granular</td>
<td>Isometric and orbicular</td>
<td>Polygonal, interlobate</td>
<td>0.5 – 20.0</td>
</tr>
<tr>
<td>Platelet</td>
<td>Elongated, wider in the middle, thinner toward the edges, sometimes acicular</td>
<td>Interlobate</td>
<td>10.0 – 50.0</td>
</tr>
</tbody>
</table>

2.3.5. Freezing model for frazil ice crystals

The isotopic source water composition for marine ice predominantly formed by frazil ice crystals can be derived by applying a modelled effective fractionation coefficient ($\alpha_{ef}$) to the measured isotopic concentration of the ice. This coefficient was calculated using Tison and others (2001) boundary layer freezing model for individual marine ice frazil ice crystals. The model is based upon Burton and others (1953) boundary layer model originally developed for simulating solute diffusion into a liquid boundary layer during the solidification of metal (equation 2-1).
where the effective fractionation coefficient ($\alpha_{\text{eff}}$, ‰) is calculated from the equilibrium fractionation coefficient ($\alpha_{\text{equ}}$, ‰) as a function of the growth rate ($v$, m s$^{-1}$), diffusion coefficient ($D$, m$^2$ s$^{-1}$) and boundary layer thickness ($z_{\text{bl}}$, m). The isotopic concentration ($c$, ‰) in the boundary layer near the ice-water interface hereby changes over time ($t$, s) according to Fick’s law (equation 2-2) (Burton and others, 1953, Eicken, 1998).

\[
\frac{\partial c}{\partial t} = D \frac{\partial^2 c}{\partial z^2} + V \frac{\partial c}{\partial z} 
\]  (2-2)

where $z$ is the distance from the ice-water interface into the fluid (m), and $V$ is the freezing velocity of the ice front (m s$^{-1}$). The model is run with a Crank-Nicholson scheme to allow for simultaneous diffusion and fractionation during the advancement of a freezing front. In the model, water is frozen from a semi-infinite reservoir (basically assuming an open system). Tison and others (2001) used a simple rod geometry for frazil ice crystals, with a diameter of $1.0 \times 10^{-3}$ m and a boundary layer thickness half of the crystal size (i.e. $0.5 \times 10^{-3}$ m) (Daly, 1984). Equilibrium fractionation coefficients were taken from Lehmann and Siegenthaler (1991) ($\alpha_{18} = 21.2$‰ and $\beta_{18}^{18} = 2.91$‰) and diffusion coefficients from Ferrick and others (2002) ($1.06 \times 10^{-9}$ m$^2$ s$^{-1}$ for $^1$HD$^{18}$O and $1.21 \times 10^{-9}$ m$^2$ s$^{-1}$ for $^1$H$^2^{18}$O) scaled to 0°C with the Stokes-Einstein equation from Eicken (1998). Frazil ice crystals form in episodic bursts (Smith and others, 2012). Their freezing velocities were taken from Tison and others (2001) ($10^{-6}$ m s$^{-1}$ and $2.7 \times 10^{-6}$ m s$^{-1}$) and Smith and others (2012) ($0.3 \times 10^{-6}$ m s$^{-1}$ to $1.4 \times 10^{-6}$ m s$^{-1}$). Since the freezing speed (removing heavy isotopes from the liquid) is greater than the diffusion coefficient (replenishing the boundary layer with heavy isotopes), a temporary amplified depletion of heavy isotopes in the boundary layer occurs - the initial transient - (Tison and others, 2001) before the isotopic concentration equals steady state. This effect is taken into account by the model since it could be especially significant in the growth of small ice crystals (mm size). Lack of empirical data on frazil ice crystal isotopic composition makes it impossible to validate the model and quantify the error. However, the frazil ice boundary layer freezing
model is very sensitive to the assumed boundary layer thickness which in turn depends on the crystal size. Observed frazil ice crystal sizes vary between \(0.035 \times 10^{-3}\) m and \(4.5 \times 10^{-3}\) m (Morse and Richard, 2009). Since the size class distribution of marine ice frazil ice crystals is still poorly known (e.g. Galton-Fenzi and others, 2012), this model focuses on the typical size chosen by Tison and others (2001).

In this study, the model is run for two scenarios; marine ice formation from (a) frazil ice crystals only and (b) frazil ice crystals with an estimated 15% pore space on average, similar to the average pore space Tison and others (2001) calculated for the Ronne Ice Shelf (Oerter and others, 1992) based on heat conduction and accumulation rate. It remains uncertain whether marine ice is formed from frazil ice crystals only, that grow, settle and compact much like snow due to the buoyancy pressure, or whether ‘pore spaces’ between individual frazil ice crystals are filled by sea water, which freezes as a result of heat conduction through the ice shelf (Eicken, 1994, Tison and others, 2001).

The pore water is taken as pure sea water fractionating at equilibrium. Even though it remains unknown whether pore water would be merely trapped between frazil ice crystals or fractionate upon freezing to become solid marine ice, similar to river and lake ice (Ferrick and others, 2002), the modelled effective fractionation coefficient would decrease slightly, if the pore water was not fractionating (by 0.44‰ for \(\delta^{18}\)O and by 3.18‰ \(\delta\)D for marine ice with 15% sea water-filled pores). Previous boundary layer freezing modelling efforts of marine or sea ice have focused on the oxygen isotope and salinity signal (Eicken, 1998, Tison and others, 2001). However, reasons for variations in marine ice salinity are poorly understood (Tison and others, 2001). Even though \(\delta\)D is not an independent tracer, this study is modelling the composition of both water isotopes in marine ice to allow for comparison with the measured co-isotopic signal.
2.4. Results

In this section we first present the occurrence of banded, granular and mixed ice facies in marine ice from all three SMIS sites. Secondly, the variation of the $\delta^{18}$O and $\delta$D ratios in all marine ice cores is described and co-isotopic relationships are given. Subsequently, the calculated effective isotopic fractionation coefficients, that help to determine theoretical marine ice source water composition, are presented for different freezing speeds. Finally, we give marine ice salinity (TDS) and ion ratios ($\Delta$Mg/Cl and $\Delta$K/Mg) indicating ion fractionation during freezing of marine ice pores.

2.4.1. Ice crystal facies of marine ice

The marine ice cores are almost entirely made up of banded and granular ice facies, whereby a mixed ice facies of these two crystal types occurs in all three ice cores (Figure 2.7, Figure 2.8, Figure 2.9 and Table 2.3). Ice core C15 shows a high percentage of both, pure granular (32%) and pure banded ice crystals (24%) (Figure 2.9, Table 2.3). The two shorter ice cores (C5 and C9) however, show an exclusive preference for either pure banded ice crystals (45%) in the case of C5 (Figure 2.7, Table 2.3) or pure granular ice crystals (58%) in the case of C9 (Figure 2.8, Table 2.3). Platelet ice facies are uncommon in marine ice at SMIS and make up only less than 10% of the ice facies in C15 (Figure 2.9, Table 2.3).
Figure 2.7: Marine ice crystal facies, $\delta^{18}O$, salinity (TDS) and $\Delta$Mg/Cl ion ratio for the core C5.

Figure 2.8: Marine ice crystal facies, $\delta^{18}O$, salinity (TDS) and $\Delta$Mg/Cl ion ratio for the core C9. Legend in Figure 2.7.
Figure 2.9: Marine ice crystal facies, $\delta^{18}$O, salinity (TDS) and $\Delta$Mg/Cl ion ratio for the core C15. Legend in Figure 2.7.
2.4.2. Isotopic composition of marine ice

Isotopic values in all ice cores range from -0.43‰ to 2.29‰ δ¹⁸O and -3.80‰ to 17.61‰ δD (Figure 2.7, Figure 2.8, Figure 2.9 and Figure 2.10). The isotopic range increases with total core length (Figure 2.7, Figure 2.8 and Figure 2.9); in the shallowest core, C5 (2.65 m long), the δ¹⁸O range is 1.06‰, whilst δ¹⁸O ranges by 1.50‰ in C9 (3.04 m long) and 2.40‰ in the longest ice core, C15 (9.44 m long). In the two shallower cores C5 and C9 (Table 2.3), collected in vicinity to each other (Figure 2.1), the data are on average more enriched in heavy isotopes (> 1.6‰ δ¹⁸O) than in core C15 (= 0.47‰ δ¹⁸O) (Table 2.3). In cores C9 and C15 the δ¹⁸O signal shows a linear trend with depth, although the trend is positive (r=0.86) in C9 (Figure 2.8) and negative (r=-0.59) in C15 (Figure 2.9). The δ¹⁸O and δD signals are significantly correlated in all cores (r >0.93), with an overall slope of 8.68 ± 0.13 (r=0.97) in a co-isotopic plot (Figure 2.10). Furthermore, all marine ice samples at SMIS are enriched in heavy isotopes with respect to sea water and there is a wide overlapping range of isotope values for the different ice facies (Figure 2.10).

Table 2.3: Average marine ice chemical composition and occurrence of ice crystal facies (bd= banded, gr= granular, mx = mixed, pl = platelet, oth=other), in each SMIS marine ice core.

<table>
<thead>
<tr>
<th>ice core</th>
<th>depth m</th>
<th>n</th>
<th>δ¹⁸O %</th>
<th>δD %</th>
<th>TDS %</th>
<th>ΔMg/Cl %</th>
<th>ΔK/Mg %</th>
<th>bd %</th>
<th>gr %</th>
<th>mx %</th>
<th>pl %</th>
<th>oth %</th>
</tr>
</thead>
<tbody>
<tr>
<td>C15</td>
<td>9.44</td>
<td>89</td>
<td>0.47±0.48</td>
<td>2.42±3.88</td>
<td>0.29±0.18</td>
<td>-9.21±5.15</td>
<td>28.86±12.96</td>
<td>24</td>
<td>32</td>
<td>32</td>
<td>6</td>
<td>2</td>
</tr>
<tr>
<td>C9</td>
<td>3.04</td>
<td>23</td>
<td>1.64±0.43</td>
<td>12.43±3.38</td>
<td>0.20±0.15</td>
<td>-13.64±13.14</td>
<td>20.87±11.09</td>
<td>--</td>
<td>58</td>
<td>42</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>C5</td>
<td>2.65</td>
<td>22</td>
<td>1.63±0.24</td>
<td>13.90±1.37</td>
<td>0.26±0.11</td>
<td>-0.83±7.89</td>
<td>14.26±5.82</td>
<td>45</td>
<td>--</td>
<td>55</td>
<td>--</td>
<td>--</td>
</tr>
</tbody>
</table>

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2.4.3. Effective fractionation coefficients

Calculating the effective fractionation coefficients for frazil ice crystals using Tison and others’ (2001) model at variable freezing speeds shows that a change in freezing speed can theoretically cause an isotopic range of ≤1.55‰ $\delta^{18}$O (Table 2.4, Figure 2.11a). This range becomes slightly smaller when pores that fractionate at equilibrium are considered (Table 2.4, Figure 2.11b). In either case the calculated effective fractionation coefficient for $\delta^{18}$O ($\beta_{\text{eff}}$) is at least 0.6‰ lower than the equilibrium fractionation coefficient (Table 2.4).
Figure 2.11: $\delta^{18}$O and $\delta^D$ composition of measured marine ice samples plotted together with the calculated $\delta^{18}$O and $\delta^D$ composition of marine ice applying Tison and others' (2001) effective fractionation coefficients to a pure sea water source (dark grey line) or a mixed water source of sea water and melted meteoric ice (4% - black line) or marine ice (40% - light grey line) using a range of different freezing rates. The triangles within the grey shaded lines denote calculated marine ice isotopic compositions generated at different freezing rates ($2.7 \times 10^{-6}$ m s$^{-1}$, $1 \times 10^{-6}$ m s$^{-1}$ and $0.3 \times 10^{-6}$ m s$^{-1}$), whereby marine ice formed at a faster freezing rate remains isotopically more similar to the source water. Meteoric ice is taken as the average isotopic composition of meteoric ice samples collected from the surface of SMIS ($-30.00\%$ $\delta^{18}$O and $-238.27\%$ $\delta^D$, n=22) and marine ice is defined as the measured average isotopic composition of samples from the isotopically most enriched ice core C5. Figure a) assumes 100% frazil ice crystals, whilst Figure b) assumes 85% frazil ice crystals and 15% pores frozen from pure sea water at equilibrium fractionation.

Table 2.4: Calculated effective fractionation coefficients (quoted in ‰) for $\alpha^D$ and $\beta^{18}$O using Tison and others’ (2001) boundary layer freezing model considering different frazil ice freezing speeds (taken from Tison and others, 2001 and Smith and others, 2012). Two different scenarios are considered; marine ice formation from frazil ice crystals only (‘no pores’) and marine ice formation from 85% frazil ice crystals with 15% of pure sea water consolidating at equilibrium freezing speed in the remaining pore spaces (‘pores’).

<table>
<thead>
<tr>
<th>Freezing speed ($x 10^{-6}$ ms$^{-1}$)</th>
<th>$\beta$ ($\delta^{18}$O)</th>
<th>$\alpha$ ($\delta^D$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pores no pores pores no pores</td>
<td></td>
</tr>
<tr>
<td>equilibrium fractionation</td>
<td>2.91 21.20</td>
<td></td>
</tr>
<tr>
<td>0.3</td>
<td>2.32 2.22</td>
<td>16.24 15.37</td>
</tr>
<tr>
<td>1</td>
<td>1.57 1.33</td>
<td>10.62 8.76</td>
</tr>
<tr>
<td>1.4</td>
<td>1.34 1.07</td>
<td>9.09 6.95</td>
</tr>
<tr>
<td>2.7</td>
<td>1.01 0.67</td>
<td>6.90 4.37</td>
</tr>
</tbody>
</table>
2.4.4. Salinity and ion ratios of marine ice

Marine ice collected from SMIS is slightly saline (0.03 to 1.01‰ TDS) (Figure 2.7, Figure 2.8, and Figure 2.9) with an average of 0.25‰ TDS. Similar to the isotope record, the TDS record increases significantly with depth in C9 (r=0.83) and decreases with depth in C15 (r=0.48). Indeed, TDS and the oxygen isotope record are significantly positively correlated in C9 and C15 (r > 0.54). However, in C5 there is no significant linear relationship between isotopes and TDS. Samples of banded and granular ice facies occur within the whole range of measured salinities and isotopic compositions (Figure 2.12a). Nonetheless, there is an apparent difference in the mean isotopic composition and salinity of different ice facies in C15; banded ice facies are more isotopically enriched and more saline than granular and mixed ice facies (Figure 2.12b). Similarly, the granular ice facies in C9 are less saline than the mixed ice facies (Figure 2.12b). However, there is no significant difference between the chemical compositions of different ice facies; the standard deviations of mean chemical composition for different ice facies overlap (Figure 2.12b).

Figure 2.12: Total dissolved solids (TDS) and oxygen isotopes (δ¹⁸O) for the different ice crystal facies in the individual ice cores. Plot (a) shows individual samples from all cores and (b) shows averages for all cores ± one standard deviation. Note: same key for both diagrams.
The marine ice at SMIS is depleted in $\Delta$Mg/Cl (Table 3) and enriched in $\Delta$K/Mg (Table 2.3). Increases in the $\Delta$Mg/Cl signal coincide with increases in the TDS record in all cores (e.g. increase of ~0.20‰ TDS and ~0.30 ppm $\Delta$Mg/Cl at 2.5 m in C9, Figure 2.8). The $\Delta$Mg/Cl and TDS signal in C5 and C9 are positively correlated ($r > 0.64$). No significant relationship, however, exists between the $\Delta$Mg/Cl ratio and TDS in C15.

2.5. Discussion

Here the evidence for marine ice formation from frazil ice crystals that nucleate in supercooled water is presented. Applying Tison and others’ (2001) boundary layer freezing model, the isotopic composition of the marine ice water source is calculated using effective fractionation coefficients. Subsequently, the evidence for marine ice formation from a changing mixed water source rather than a constant water source with a changing freezing rate is evaluated. Finally, the possible geographic origin of the relatively fresh water needed in the process of frazil ice formation is discussed and what this implies with regard to marine ice formation mechanisms.

2.5.1. Marine ice formation from frazil ice crystals

The presence of predominantly banded and granular ice facies (Table 2.3) in marine ice at SMIS strongly suggests that the marine ice at SMIS was formed from predominantly frazil ice crystals (e.g. Tison and others, 1993, Treverrow and others, 2010) which nucleate from supercooled water. Marine ice formation solely from the advancement of a freezing front and associated formation of columnar ice crystals (e.g. Zotikov and others, 1980) is rare (Table 2.3) and no columnar ice crystal facies have been observed in thin sections from SMIS. Further evidence for SMIS frazil ice formation from supercooled water is a higher $\delta D/\delta^{18}O$ ratio in marine ice from C5 and C9 than what the output of Tison and others (2001) source model predicts (Figure 2.11a and Figure 2.11b). This enrichment in $\delta D$ could result from an enhanced diffusivity of deuterium in saline waters (Horita, 2009) together with a fast freezing speed, which is thought to amplify kinetic effects (Souchez and others, 2000). This is consistent with
Souchez and others (1995), who suggested using a 2‰ higher equilibrium fractionation coefficient to explain the higher enrichment of δD in marine ice.

Marine ice might not be made up of frazil ice crystals alone since these generally expel all salt during their formation (Tison and others, 1993) and SMIS marine ice is slightly saline (Table 2.3). Ion fractionation in SMIS marine ice samples (depletion in ΔMg/Cl and enrichment in ΔK/Mg, Table 2.3) also points toward the presence of sea water in pore spaces between frazil ice crystals. Nonetheless it is difficult to estimate the percentage of sea water-filled pores from its chemistry since marine ice pore water salinity and desalination processes are poorly understood; no brine channels have ever been observed in marine ice and evidence for brine pockets was only found in the lower tens of metres of the ~200 m thick marine ice layer at the Amery Ice Shelf that is still in hydraulic connection with the ocean (Craven and others, 2004, Treverrow and others, 2010).

2.5.2. Isotopic evidence of a changing water source

Whether marine ice was formed from pure frazil ice crystals only or from a combination of frazil ice crystals and frozen pore water, the isotopic record of marine ice predominantly reflects the source water composition of frazil ice crystals. The measured isotopic range and co-isotopic slope of all SMIS marine ice samples point toward frazil ice formation from a mixed water source that changes over time. Even though the measured δ¹⁸O range for samples from all ice cores at SMIS (2.71‰ δ¹⁸O and 21.41‰ δD, Figure 2.10) is similar to the theoretical maximum isotopic shift at equilibrium fractionation from a constant water source (Table 2.4), results from applying Tison and others’ (2001) boundary layer freezing model indicate that the isotopic composition of marine ice should only vary by ≤1.55‰ δ¹⁸O and ≤11.00‰ δD considering a range of measured freezing rates for frazil ice crystals and a constant water source (no pore water, Figure 2.11a and Table 2.4). If pores filled with sea water that fractionates upon consolidation would be present, the isotopic range of marine ice formed at variable typical freezing speeds would be even smaller (1.31‰ δ¹⁸O and 9.34‰ δD) (15% pore water, Figure 2.11b and Table 2.4). Hence, the measured isotopic range of marine ice
samples at SMIS points toward a source water composition that is likely not constant over time and in space assuming an open reservoir in the ice shelf cavity. This is consistent with a high co-isotopic regression slope of $8.68 \pm 0.13$ ($r=0.97$) (Figure 2.10), which is steeper than a freezing slope (i.e. water formed from the same water source with a variable freezing speed) in an open or closed system (Jouzel and Souchez, 1982, Souchez and Jouzel, 1984). However, since frazil ice crystals form in a water plume in the ice shelf cavity, closed system freezing is precluded. Nonetheless, the highest possible freezing slope in an open system would theoretically only be 7.38, applying Souchez and Jouzel's (1984) freezing model to a water source of pure sea water taken as SISW below the Ross Ice Shelf (Jacobs and others, 1985, Fitzsimons and others, 2012). Freezing slopes of sea water with a constant contribution of melted meteoric ice would be even lower. Co-isotopic mixing slopes for ice formed from a changeable water source, however, would be higher, around 8 (Souchez and Groote, 1985), which is more similar to the measured slope of all SMIS samples.

For individual ice cores the isotopic range is smaller and co-isotopic slopes are shallower (Figure 2.10). Whilst the co-isotopic slope of $7.92 \pm 0.17$ in C15 with an isotopic range of $2.4\% \delta^{18}O$ remains similar to a mixing slope, the two shallower ice cores C5 and C9 have a slope $\leq 7.69$ with a smaller isotopic range ($\leq 1.50\% \delta^{18}O$). This could suggest that the source water was similar during accretion events amounting up to $\sim 3$ m of marine ice at the core sites of the shallower ice cores, extracted in vicinity to each other (Figure 2.1), whilst the source water composition at the site of C15 could have changed over the course of accreting $\sim 9.5$ m of marine ice due to the co-isotopic slope close to 8 and an isotopic range ($> 1.55\% \delta^{18}O$). The low slope of $5.26 \pm 0.48$ in co-isotopic data from C5 is lower than a freezing slope, even if the source water was mainly glacial water derived from meteoric ice. The isotopic range of $1.06\% \delta^{18}O$ with most samples clustering tightly around $1.5\% \delta^{18}O$ (Figure 2.1) might be too small to allow for a representative freezing or mixing slope.

Since there is some evidence that marine ice at SMIS partially formed from the consolidation of sea water-filled pores, an increase in salinity of marine ice is consistent with slower consolidation rates of frazil ice crystals. A significant correlation between the TDS and
$\Delta$Mg/Cl signal in C5 and C9, could therefore suggests that sea water-filled pores consolidate more slowly (leading to higher ion fractionation) when salinity is high.

### 2.5.3. Source water composition

Since the crystallography provides evidence that marine ice is primarily composed of frazil ice crystals, which nucleate in supercooled water, the marine ice source water at SMIS is likely a mixture of local sea water and fresher and possibly warmer water. This combination of water masses supercools upon mixing (double diffusion mechanism) and/or adiabatic lifting (ice pump mechanism). The fresher-than-sea water could originate from melted meteoric ice (Souchez and others, 1995, 1998) and/or melted marine ice (Souchez and others, 1991, Tison and others, 1993).

Applying the calculated effective fractionation coefficients of Tison and others’ model (Table 2.4) to sea water, suggests that marine ice at SMIS with an isotopic concentration $\geq 1.70\%\delta^{18}$O (Figure 2.11b) could have formed from a water source with a component of recycled marine ice. The exact cut off threshold is dependent on the calculated fractionation coefficient and the assumed sea water composition. This study took SISW as measured below the Ross Ice Shelf (Jacobs and others, 1985, Fitzsimons and others, 2012) as the ambient sea water, but it remains unknown whether the pure sea water below SMIS could be more enriched in heavy isotopes (by $\sim 0.30\%\delta^{18}$O). The idea of marine ice formation from a water source with a component of melted and relatively fresher recycled marine ice was first developed by Souchez and others (1991) based on marine ice that was more enriched in heavy isotopes (up to $3.20\%\delta^{18}$O, Figure 2.1) than what could be explained by applying the equilibrium fractionation coefficient to sea water. However, equilibrium freezing during frazil ice formation is unlikely. Since the isotopic composition of marine ice in C5 and C9 is on average $\sim 2\%$ more enriched in $\delta^{18}$O than sea water, a high contribution of recycled marine ice ($\sim 40\%$ dependent on freezing speed) to the source water is necessary to explain all of the observed enriched marine ice samples (Figure 2.11a and Figure 2.11b). The exact percentage contribution of recycled marine ice, however, is difficult to determine, especially since marine
ice can be repeat recycled. Goodwin (1993) already considered the possibility that δ¹⁸O marine ice samples, which are not as enriched as in Souchez’s (1991) study (i.e. only up to 1.56‰ δ¹⁸O, Table 2.1), could have formed from repeat recycling of previously formed marine ice.

Less enriched (≤ 1.70‰ δ¹⁸O) marine ice samples could have formed with a small contribution of glacial water derived from meteoric ice to the marine ice water source (especially in C15, Figure 2.11a and Figure 2.11b). Since meteoric ice is very depleted in heavy isotopes (between -35‰ to -25‰ δ¹⁸O, Kellogg and others, 1991a), only a small glacial water contribution to the source water (up to a maximum of 4%) is sufficient to explain the isotopic composition of some of the less isotopically enriched and isotopically depleted ice samples in C15 (Figure 2.11a and Figure 2.11b). Previous studies interpreted the presence of a mixing slope together with a larger isotopic range (of up to 18.00‰ δ¹⁸O, Table 1) than observed at SMIS of isotopically depleted samples as the result of a mixed water source of sea water and varying proportions of melted meteoric ice (Goodwin, 1993, Souchez and others, 1995, 1998). The marine ice source water for all sample sites at SMIS, however, could be made up of a mixture of all three water sources; melted meteoric ice, melted marine ice and sea water, whereby melted meteoric ice would contribute the least to the mixture (in the order of 5%) due to its high isotopic depletion.

There is no evidence that different crystal morphologies are the result of marine formation from different water sources and/or proportions of pore water. There is no significant difference in the chemistry of the most common ice crystal facies (Figure 2.12a); the mean isotopic signature and salinity of banded ice crystals only differs slightly from that of granular ice facies in all cores whereby the standard deviations overlap (Figure 2.12b).

2.5.4. Processes of frazil ice formation at SMIS

In this section we discuss geophysical scenarios that can lead to the formation of marine ice source water as identified through the ice chemistry in the previous section. One possible mechanism for marine ice formation at SMIS could be the ice pump, delivering freshwater of meteoric origin to the ice shelf cavity. This process could explain marine ice formation at the
site of C15 where the ice likely formed from a mixed water source dominated by glacial and
sea water (Figure 2.11a and Figure 2.11b). Indeed, there is a thermohaline circulation driven
by either HSSW or MCDW in the cavity of the adjacent Ross Ice Shelf (Dinniman and others,
2007). Ocean circulation models reveal that there is an amplified basal melt rate at the eastern
margin of the Ross Ice Shelf that extends from the Ross Ice Shelf margin south past Minna
Bluff (Dinniman and others, 2007, Timmermann and others, 2012). And recent radar
investigations in the shear zone separating SMIS from the Ross Ice Shelf have revealed a
several tens of meters thick marine ice layer (Arcone and others, 2016). Thus, a buoyant
plume of meteoric meltwater and sea water generated below the Ross Ice Shelf could enter
the ice shelf cavity of the ~300 m thinner SMIS, similar to the vertical ascent of a water plume
from below the Ross Ice Shelf to the cavity of the Northern McMurdo Ice Shelf (Robinson and
others, 2014). Here the water in the plume would supercool due to a change in the pressure-
dependent freezing point in shallower waters and frazil ice crystals would nucleate ().
Potentially this water plume could also reach the ocean floor of the much shallower SMIS ice
shelf cavity with a maximum depth of ~400 m (Johnston and others, 2008) forming anchor ice
as suggested in a modelling effort by Leonard and others (2014) ()

Figure 2.13: Interpretive diagram of an ice pump originating below the Ross Ice Shelf (RIS), with MCDW
melting meteoric ice at the RIS base, which becomes supercooled as it lifts along the ice shelf gradient
and into the Southern McMurdo Ice Shelf (SMIS) cavity. Here frazil ice crystals form at the base of the
ice shelf from supercooled water, gradually compacting to form marine ice. Supercooled water also
could extend to the ocean floor where anchor ice would form. East is to the right, west is to the left of
the diagram. SMIS flows toward the reader and RIS flows away from the reader.
However, marine ice below the Ross Ice Shelf accumulates only in a thin layer (Zotikov and others, 1980, Timmermann and others, 2012). Hence the ice pump mechanism might be not very strong in the Ross Ice Shelf cavity. Also, this ice pump would only explain the delivery of freshwater of meteoric origin into the SMIS ice shelf cavity and not the recycling of marine ice.

A proportion of marine ice at SMIS shows an apparent fractionation in water isotopes that exceeds the modelled effective fractionation for freezing of frazil ice crystals from a pure sea water source. After Souchez and others (1991) and Tison and others (1993) marine ice could be recycled in a mechanism where a shallow thermohaline or tidal circulation melts the older marine ice at depth, which then becomes supercooled upon rising to shallower waters where frazil ice crystals nucleate. However, it is unlikely that a shallow tidal circulation would reach the SMIS ice shelf cavity, which is sheltered by Black and White Island and several tens of kilometres away from the open ocean (Figure 2.1). Alternatively, marine ice at SMIS could be recycled through melting of exposed marine ice at the ice shelf surface which would then be routed to the ice shelf base. Indeed, at SMIS surface melt was observed to temporarily pool in elongated lakes between ridges of exposed marine ice during the summers of 2007 and 2010 (Figure 2.4 and Figure 2.14).

![Surface lake in between SMIS marine ice ridges. Minna Bluff to the left, Mount Discovery in the background.](image)

Figure 2.14: Surface lake in between SMIS marine ice ridges. Minna Bluff to the left, Mount Discovery in the background.
Thus far surface lakes have been observed to appear seasonally on ice shelves of the Antarctic Peninsula (e.g. Luckman and others, 2014) but were thought to enhance a tensile stress field (Banwell and others, 2013) or through refreezing increase the weakness of the ice shelf eventually causing ice shelf weakening (Hubbard and others, 2016) and break up. Surface water on SMIS, however, is likely draining through an existing weakness in the ice shelf, the tide crack.

A tide crack was observed to run parallel to Minna Bluff in the marine ice zone close to shore, where the ice shelf becomes regrounded. Salt deposits along the tide crack indicate that it is actively connecting the ice shelf base with the ice shelf surface. However, surface water cannot percolate to the ice shelf base if cracks are filled with ocean water from the ice-ocean interface up to the height of the freeboard. Periodic tidal lifting could close off the bottom of the crack during low tide, allowing for water till fill the crack from the surface to the ice-ocean interface at the base of the ice shelf. During high tide the crack could then be opened allowing for the crevassse water to seep into the water cavity below the ice shelf. Here it would become supercooled in a double diffusion mechanism of heat and salt with the ambient sea water, whereby frazil ice crystals would nucleate (Figure 2.15). For this process to occur, both water masses, the sea water at the ice shelf base and the recycled marine ice, need to be at their salinity-dependent freezing point and would become mixed due to local turbulence (McPhee and others, 2013). Hence, the surface meltwater would need to be cooled during percolation.

During percolation some refreezing could occur. Indeed, solute chemistry and ice facies of the ~9.5 m long ice core C15 show evidence of local refreezing. Between 6.8 m and 7.2 m depth, for example, the marine ice salinity increases by almost one magnitude and the ice facies are predominantly large platelet type crystals (Figure 2.9). Similarly, platelet ice facies are present between 0.5 m and 1.8 m depth together with several peaks in salinity (Figure 2.9). This could be a result of surface meltwater refreezing at depth.

Recycling of ice melted at the ice shelf surface is not necessarily limited to marine ice only. Water from the nearby glaciers and snow patches on Minna Bluff close to the shore of the ice shelf (Figure 2.1) could find its way into the tide cracks of SMIS and hence to the ice shelf base. This process would be an alternative explanation for the small contribution of
glacial water to the source water at SMIS especially at the site of C15 (Figure 2.11a and Figure 2.11b).

Recycling surface meltwater in the process of marine ice formation would imply that the ice shelf could be sustaining itself to some extent, with most of the marine ice forming during the melt season in summer. Relating surface lakes to ground-based accretion of marine ice may change the way that ice shelves are modelled. If surface ablation and local accumulation of marine ice were in equilibrium (e.g. Kellogg and others, 1990) their rates would both be in the order of ~0.1 m a\(^{-1}\), as the measured surface ablation on SMIS (Clifford, 2005). Hence, the surface ice of the ~100 m thick SMIS would be ~1000 years old, similar to the youngest radiocarbon date of 1230 ± 50 years (Denton and Marchant, 2000).

Instead of recycling marine ice from the ice shelf surface, McPhee and others (2013) suggest that loose frazil ice crystals could also be recycled in ocean turbulence before they consolidate. In this process, frazil ice crystals would be moved to deeper waters where they melt due to the pressure dependence of the freezing point. Upon rising of the resulting more buoyant meltwater plume, frazil crystals would re-nucleate (McPhee and others, 2013). If recycling of surface meltwater or loose frazil crystals occurs below SMIS, turbulence must occur at the ice shelf base to allow for vertical mixing. In the absence of a thermohaline and tidal circulation, it still remains to be determined how this vertical turbulence would be initiated. In a marine environment with no atmospheric contact and thus no wind as a driver, tidal currents are generally the primary mechanism for mixing of water masses (Robinson, 2012). Tison and others (1998) suggested that a shallow tidal circulation would draw warmer waters below the front of the Hells Gate Ice Shelf in summer (e.g. Jacobs and others, 1992) where it would melt some of the loose frazil ice crystals upstream and thus cause some of the marine ice accretion downstream where the ice shelf is shallower. At SMIS, however, no Antarctic Surface Water is pushed under the ice shelf due to wind (e.g. Hattermann and others, 2014) and an aforementioned shallow tidal circulation is unlikely to reach the SMIS cavity since it is far away from the ice shelf ocean front and sheltered by Black and White Island. Tidal lifting of ice shelves, however, may cause an under-ice shelf water circulation that could cause localized turbulence. When different water masses are present, turbulent mixing could lead to
heat exchanges between water masses of different salinities and thus frazil ice formation (McPhee and others, 2016). Indeed, spring tidal lifting of ± 0.5 m has been measured in McMurdo Sound (McPhee and others, 2016) and results from tidal modelling using surface altimetry measurements over the adjacent Ross Ice Shelf estimate daily tidal cycles of up to ± 1.0 m in summer time (Padman and others, 2008).

There is little evidence that suggested different marine ice formation processes lead to different ice crystal facies at SMIS. Banded ice crystals are slightly more enriched in heavy isotopes and saline than the granular ice facies, especially in samples from C15 (Figure 2.12a and Figure 2.12b) but this difference is not statistically significant. Tison and others (1993) found a similarly weak chemical difference between the two ice facies but suggested that banded ice facies were generated by frazil ice crystals that aligned in a sub ice shelf current, whereas granular ice crystals facies were thought to result from fast frazil ice formation in a deeper thermohaline circulation further inland. In contrast, Treverrow and others (2010) found that banded ice crystal facies merely occurred in younger marine ice than granular ice crystals. This would suggest a change in crystal morphology over time. Indeed, ice deformation experiments showed that due to post-depositional ice growth and recrystallization processes associated with ice flow, the original ice crystal morphology is often altered (Wilson and others, 2014) and predominantly granular/less elongated ice crystal develop in marine ice (Dierckx and others, 2014).
Figure 2.15: Interpretive diagram of surface meltwater routing to the SMIS ice shelf base and associated formation of frazil ice crystals. Top: meltwater percolation below the ice shelf cavity, where heat diffuses faster than salt into the sea water below. Middle: the water becomes supercooled and frazil ice crystals form. Bottom: the frazil ice crystals accumulate at the base of the ice shelf and freeze/compact to form marine ice. North is to the right, south is to the left in the diagram with Minna Bluff as the shore. Ice flows from right to left.
2.6. Conclusion

The objectives of this chapter were to determine SMIS marine ice source water composition and origin in order to improve the understanding of marine ice formation processes. The analysis was focused on SMIS marine ice chemistry and crystal morphology and the application of a boundary layer freezing to derive the source water composition. Evidence for several possible marine ice formation processes was evaluated in light of the geographic setting. The specific conclusions are:

- The source water composition of SMIS marine ice is not constant but changes spatiotemporally. This is indicated by a co-isotopic mixing slope of ~8 and an isotopic range of 2.71‰ $\delta^{18}$O and 21.41‰ $\delta$D for all marine ice samples from all sites. This is almost double the isotopic range predicted by the frazil ice boundary layer freezing model formed from a constant water source at different freezing speeds. The use of the frazil ice freezing model is justified since SMIS marine ice is almost entirely made up of granular and banded ice crystal facies typical for consolidated frazil ice crystals.

- Marine ice samples, which are isotopically depleted in comparison to modelled marine frazil ice formed from pure sea water, have likely formed from a source water mixture of sea water and a relatively small proportion (~4%) of melted meteoric ice, which originates likely from below the ice shelf base of the adjacent ~300 m thicker Ross Ice Shelf. This would support the theory of vertically advected water in an ice pump mechanism.

- Marine ice samples, which are isotopically enriched in comparison to modelled marine frazil ice formed from pure sea water, have likely formed from a source water mixture of sea water and a large proportion (~40%) of isotopically enriched melted marine ice. The exact percentage varies with freezing speed and number of freeze-thaw-refreeze cycles. Melted marine ice at SMIS originates from a proposed recycling mechanism, where exposed marine ice melts at the ice shelf surface and drains through the tide crack to the ice shelf base. Here it becomes turbulently mixed with relatively colder and saltier sea water and frazil ice crystals nucleate in a double diffusion process.
Thus, summer surface melt could make a substantial contribution to basal ice shelf mass accretion particularly at the southern margin of SMIS. Consequently it can be concluded that investigations of marine ice chemistry can help to understand ice shelf basal boundary conditions. Results from this study suggest that formation of marine ice at the relatively thin SMIS is at least partially determined by surface meltwater generation. The proposed mechanism of marine ice recycling thus ties together surface and basal processes, and could potentially slow ice shelf disintegration.
3. Meteoric and marine ice deformation at SMIS

This chapter addresses research questions 3, 4 and 5 (see section 1.6.1.). Initially, literature regarding influence of strain on ice microstructure (ice crystal fabric, shape and size) is summarized and past measurements of meteoric and marine ice micromorphology are presented. The evolution of marine and meteoric ice micromorphology with increasing strain is investigated along a flow path. Marine and meteoric ice micromorphology are compared to evaluate whether marine ice deforms more easily than meteoric ice. It is discussed whether marine ice properties enhance ice shelf flow and what the implications are for ice shelf stability. Research presented in this chapter will be submitted to a scientific journal in slightly modified form. For co-authors see section 1.6.3..

3.1. Strain and ice deformation

Ice mass flow and deformation is largely controlled by the local terrain and thickness and size of the ice mass. The latter is related to ice mass input, which in the case of ice shelves, is inflow from ice streams and glaciers and local snow accumulation. Since ice shelves float on the ocean and thus experience no basal friction, the shelf ice deforms differently to ice resting on land. Figure 1.1 shows that ice shelves flow much faster than land ice. Whilst ice flow can happen without deformation of ice, generally internal deformation of ice, also often called creep, accompanies ice flow. During internal creep after an initial short period of elastic deformation, ice becomes strained as a power law function of stress ($\tau$) (Drewry and others, 1982, Budd and Jacka, 1989), where strain rate ($\dot{\varepsilon}$) is expressed in Glen’s Flow Law (Glen, 1955, Nye, 1957):

$$\dot{\varepsilon} = A\tau^n$$

(1)

Here, the constant $A$ represents hardness and depends largely on temperature but also crystal size and orientation, impurity content and water content (e.g. resulting from the pressure of the overlying ice). For glacier ice the exponent $n$ is a constant commonly around 3, which averages the ice flow behavior during different stages of deformation. Given that
marine ice is generally warmer than glacial ice due to its formation at ocean temperature (e.g. Treverrow and others, 2010), it is expected that it deforms more easily than glacier ice. Additionally the often higher impurities in marine ice in comparison to meteoric ice (e.g. Treverrow and others, 2010) promote deformation (Trickett and others, 2000; Hammonds and Baker, 2016). However, the crystal orientation and size of marine ice (e.g. Dierckx and Tison, 2013; Dierckx and others 2014) differ from that of meteoric ice and it remains unknown whether those properties outweigh temperature effects. Hence, this study will investigate changes in ice crystal orientation and size with increasing strain in a natural setting.

During deformation ice has to adjust its configuration of crystals since it is generally incompressible once ice has reached its maximum density of 917 kg m\(^{-3}\) (Cuffey and Paterson, 2010). Ice can be exposed to a couple of different stresses: normal stress (which acts perpendicular to the area it acts on) and shear stress (which acts tangentially to the area it acts on) (Nye, 1957). Normal stress can either be compressional or tensional (pure shear) and shear stress always involves the differential movement of two forces against a plane (shear or simple shear) (Figure 3.1). The strain regime of ice shelves is generally dominated by horizontal extension (Figure 3.1a) accompanied by ice shelf thinning. Localized horizontal shortening and thus vertical ice shelf thickening is possible, especially around islands (Figure 3.1b and Figure 3.2). Lateral shearing around the margins of ice shelves is also typical, resulting in shear strained ice (Figure 3.1c and Figure 3.2). Basal drag can occur over pinning points in ice shelves also resulting in shear strained ice (Figure 3.1d and Figure 3.2).
Figure 3.1: Typical strains in ice shelves simplified to the deformation of a square. Shelf ice becomes horizontally extended (a) or shortened (b) as a result of normal stress. Shelf ice can also be exposed to lateral friction which results in shearing and associated deformation (c) as a result of shear stress. Over pinning points, shelf ice can become deformed (d) as a result of simple shear.

Figure 3.2: Typical stresses acting on shelf ice. Simple shear due to sidewall or basal drag on topographic highs/pinning points and longitudinal extension due to ice shelf flow (modified from Paolo, 2015).
Ice deforms internally when exposed to stress. In terrestrial ice (type Ih), water molecules are held together by hydrogen bonds, in which the water molecules align favorably to the electrostatic attractions between hydrogen and oxygen atoms giving a tetrahedral structure (Figure 3.3a) with every oxygen atom surrounded by four hydrogen atoms. These tetrahedra arrange to form a hexagonal ice crystal lattice of water molecules (Figure 3.3b), which arrange into layers, so-called basal planes. Since the hydrogen bonds within the layers are much stronger than between layers, deformation of ice preferentially occurs along these basal planes during so-called ‘basal glide’ or ‘easy glide’ (Hooke, 2005). Deformation through dislocation climb between different basal planes or along prismatic or pyramidal planes of the ice lattice requires much more energy and is hence called ‘hard glide’. It does not contribute as significantly to bulk deformation of ice as it is nearly two orders of magnitude harder to deform ice in hard than in easy glide (Duval and others, 1983).

Furthermore ice almost always has small imperfections in its crystal structure, so-called ‘defects’, which allow for much easier deformation than if the ice lattice was perfect. The most relevant defects for allowing easier ice deformation are ‘line defects’ or dislocations in the crystal lattice, whereby entire rows of atoms are arranged anomalously (Hooke, 2005). Impurities weaken the ice lattice possibly through the formation of dislocations (e.g. Trickett and others, 2000). Thus, the relatively higher salt and debris content in marine ice in comparison to meteoric ice (e.g. Treverrow and others, 2010) contributes to weakening of the ice lattice.

![Diagram](image)

**Figure 3.3:** Typical tetrahedral structure of terrestrial ice. (a) The diagram shows one full water molecule and the hydrogen bonds (as dotted lines) to either the adjacent hydrogen (in white) or oxygen atoms (in grey) (modified from Bartels-Rausch and others, 2012). (b) A part of the ice crystal lattice, whereby oxygen atoms are denoted as round circles and hydrogen bonds as lines, and the corresponding hexagonal cell (from Thorsteinsson, 1996).
Ice does not deform uniformly when exposed to constant stress. Strain in ice follows a certain pattern, related to its internal energy. Commonly, changing patterns in the strain rate are referred to as primary, secondary and tertiary deformation (a). In isotropic ice, where ice crystals are oriented randomly, deformation is initially fast (during primary strain) as dislocations increase until the strain rate progressively decelerates mainly until it reaches minimum strain due to piling up of dislocations and so-called strain hardening. This stage is also called secondary strain. Subsequently, the strain rate accelerates and ice crystals start to split and recrystallize (Alley, 1992). The strain rate eventually becomes steady during tertiary creep (a), where it balances between strain hardening and recovery processes. Commonly the different stages of strain are shown in a diagram of log strain against log strain rate (b) as it illustrates the minimum strain rate during secondary strain and the constant tertiary strain rate more clearly (Budd and Jacka, 1989).

The initial orientation of ice crystals, especially with regard to the basal planes, influences how fast the ice deforms (Budd and Jacka, 1989). The ice crystal lattice adjusts further during creep. In nature ice is rarely exposed to constant stress. The stress regime changes with the ice flow according to changes in the topography. The type of stress the ice is exposed to influences which ice fabric the ice develops during secondary and tertiary strain. In turn, ice fabrics also influence the deformation of ice. Ice fabric is a term for the distribution of c-axes (or optical axes) in an ice crystal lattice. Routinely, the c-axes orientations of individual ice crystals with azimuth and dip are plotted on a Schmidt stereonet. Every c-axis is hereby represented as a dot on the surface of a hemisphere in equal area projection (assuming that each c-axis passes through the centre of the hemisphere) (Cuffey and Paterson, 2010) (Figure 3.5). Hence, ice fabrics can be characterized visually. Ice fabrics vary from isotropic (c-axes distribution is uniform in all directions) (Figure 3.5a) to anisotropic (strongly orientated c-axes pointing in a similar direction) (Figure 3.5b) and circle girdle fabrics (c-axes arranged in a circle) (Figure 3.5c).
In the primary stages of creep, there is very little change in the crystallographic preferred orientation of the ice crystals. The ice crystals start to align to the current strain regime during secondary creep when new crystals start to form. During tertiary creep, the crystal axes become aligned within the new kinematic reference frame (e.g. Gao and Jacka, 1987; Jacka and Budd, 1991).

When the ice lattice experiences horizontal uniaxial extension as during the regular flow of many ice shelves (e.g. Treverrow and others, 2010), c-axes of the ice crystals rotate away from the extension axis forming a small circle girdle or broad cone (Figure 3.6b). Laboratory experiments have shown that the small circle girdle ice fabric has a colatitude (i.e. 
dip or cone angle) of about 50° (Jacka and Maccagnan, 1984). When extension is confined, such as at ice divides, great circle girdle fabrics have been observed (Faria and others, 2014a). When the ice crystal lattice experiences uni-axial shortening as during compression experiments, a small circle girdle or tight cone fabric typically develops (Jacka and Budd, 1991, Vaughan, 2016) (Figure 3.6 and Figure 3.8). In polar ice this gives rise to small circle girdle fabrics with colatitudes of around 38° (Thorsteinsson and others 1997, De La Chapelle and others, 1998). If strain rates and temperatures are low, as typical in the top layers of polar ice, single maximum cluster fabrics may also develop as a result of compression (Alley, 1992, Durand and others, 2009, Faria and others, 2014a, Vaughan, 2016). Ice in simple shear, as typical around the margins of ice shelves (Figure 3.2), develops a single maximum or cluster fabric during secondary and tertiary strain (Gao and Jacka, 1987, Budd and Jacka, 1989) (Figure 3.6 and Figure 3.8).

Whilst the type of strain the ice is exposed to determines the evolution of a certain ice fabric, the initial ice fabric also influences the strain rate. In many flow conditions ice has been exposed to one kind of stress regime and is then exposed to a different stress regime due to terrain changes. If the ice fabric is favourably aligned to the strain regime, deformation happens faster and thus the state of tertiary strain is reached faster. In contrast if the ice crystals are unfavourably aligned for the stress regime, deformation is more difficult, lowering the strain rate even more than if the ice was initially isotropic. In contrast to undeformed meteoric ice, marine ice initially has an anisotropic fabric. Single maximum fabrics have been observed in marine ice (e.g. Treverrow and others, 2010), but it remains unclear whether this ice fabric is preferentially oriented to the strain regime of ice shelves. Laboratory deformation studies have indicated that the anisotropic marine ice is harder to deform than meteoric ice in some but not in all strain settings (Diercks and others, 2014). However, given the time and scale issues of laboratory experiments, it remains unsure how marine ice behaves in different strain regimes in a natural setting. This study therefore compares meteoric and marine ice samples in a natural setting of measured strain rates.
Figure 3.6: Alignment of the c-axis in a simplified ice crystal lattice depicting only the basal planes (playing card model) as a result of different strain regimes (modified from Paterson, 1994 and Thorsteinsson, 1996). The resulting ice fabric is shown as theoretical stereonet plots of horizontal and vertical thin sections.
3.2. Dynamic recrystallization of ice in response to strain

When ice flows it can experience primary, secondary and tertiary creep. Dependent on the creep stage, the ice lattice deforms during so-called dynamic recrystallization processes and adapts to the current strain regime changing its crystal orientation, size and possibly shape (Table 3.1). Dynamic recrystallization includes normal grain growth (NGG), rotation recrystallization (RRX) and migration recrystallization (SIBM) (Table 3.1), and refers to a re-orientation of the crystal lattice caused by grain boundary migration and/or the formation of new grains through either splitting of larger grains or nucleation of new grains (Faria and others, 2014b). Sometimes NGG is omitted in the definition of dynamic recrystallization since it only dominates when the strain rate is zero (Faria and others, 2014b) (Table 3.1), shows that dynamic recrystallization leads to an overall reduction of crystal size with increasing strain.

During primary creep, more dislocations form in ice crystals increasing the internal energy of the ice lattice (Poirier, 1985). This causes the strain rate to progressively decrease until it reaches minimum strain (b) because the dislocations pile up (so-called ‘strain hardening’) (Miguel and others, 2001). Some non-isotropic ice fabrics are more favorably oriented to the strain regime than isotropic fabrics, enhancing the strain rate during secondary strain (i.e. no strain hardening occurs) (Figure 3.8). For instance, a circle girdle or cone fabric is more favorably oriented for compression and a single maximum or cluster fabric is more preferentially oriented for shearing (Figure 3.8).

Table 3.1: Dynamic recrystallization processes in the crystal lattice and typical ice properties (after Cuffey and Paterson, 2010, Wilson and others, 2014 and Faria and others, 2014b).

<table>
<thead>
<tr>
<th>Process</th>
<th>Driving force</th>
<th>Ice fabric</th>
<th>Grain size</th>
<th>Grain shape</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Normal grain growth (NGG)</td>
<td>boundary curvature</td>
<td>No change</td>
<td>Continuous growth of ice crystals</td>
<td>Granular with curved grain boundaries</td>
</tr>
<tr>
<td>2. Polygonization/ Rotation recrystallization (RRX)</td>
<td>dislocation density, internal energy</td>
<td>Change</td>
<td>Near equal grain size</td>
<td>Granular</td>
</tr>
<tr>
<td>3. Migration recrystallization without nucleation (SIBM-O)</td>
<td>Major change</td>
<td>Unequal grain size</td>
<td>Granular with initially interlobate grain boundaries</td>
<td></td>
</tr>
<tr>
<td>Migration recrystallization with nucleation (SIBM-N)</td>
<td>Small, &lt;1mm</td>
<td>Granular</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Upon further straining, the dislocation pile up is eventually released and the ice experiences accelerating tertiary strain (b). During secondary and tertiary strain the ice experiences RRX (Hamann and others, 2007). Hereby the ice crystals initially bend, giving rise to so-called ‘undulose extinction’ in cross-polarized light (Wilson and others, 2014). As the ice crystal lattice adjusts to the new strain regime, more dislocations form, which align to form dislocation walls and eventually subgrain boundaries (Guilopé and Poirier, 1979, Weertman and Weertman, 1992, Durand and others, 2009, Weikusat and others, 2009). Eventually ice crystals are split into smaller crystals with almost the same orientation, theoretically resulting in a steady almost equal grain size if RRX is prevalent (Table 3.1, Figure 3.7) (Duval and Castelnau, 1995). Concurrently the crystal lattice rotates and starts to align favorably to the strain regime it experiences. This overall more preferential alignment of the ice crystals to the current strain regime causes an accelerated strain rate during secondary and tertiary strain (Figure 3.8).

Migration recrystallization (SIBM) also happens during secondary and tertiary strain. Unlike in normal grain growth, this SIBM is driven by the internal energy of the crystal lattice.
A net transfer of molecules from the grain with higher strain energy to the grain with lower grain energy allows for the growth of the ice crystals that are more preferentially oriented to the strain regime. This allows for initially irregular interlocking grain boundaries to form (so-called interlobate grain boundaries) (Urai and others, 1986, Duval and Castelnau, 1995) (Table 3.1). Simultaneously, new strain-free ice crystals can nucleate in a region in the crystal lattice with high stored energy (Faria and others, 2014b). The newly formed grains initially have very small grain sizes (<1 mm) and grow at the expense of the old ones (Alley, 1992, Cuffey and Paterson, 2010), thus the average grain size is initially smaller and the overall grain size is unequal (Table 3.1). These new grains nucleate predominantly close to grain boundaries and sub grain boundaries (Chauve and others, 2017). SIBM can occur with and without grain nucleation whereby higher strain rates favor the nucleation process (Faria and others, 2014b, Chauve and others, 2017). Temperature-activated diffusion then enhances the recrystallization rate (Montagnat and Duval, 2000). However, more recent work found that grain boundary migration is also taking place at much lower temperatures than previously thought (Faria and others, 2014b, Azuma and others, 2012). Overall SIBM also causes a preferential alignment of the ice crystals to the current strain regime and thus an accelerated strain rate during secondary and tertiary strain (Figure 3.8).

Irrespective of the starting ice fabric, the ice crystals will generally align in a fabric most favorable to the current strain regime during tertiary strain as they experience SIBM and/or RRX. Laboratory experiments have shown that a small circle girdle fabric eventually develops during tertiary strain in a stress regime of uniaxial compression even if the sample was not initially isotropic (Jacka, 1984, Jacka and Maccagnan, 1984, Gao and Jacka, 1987) (Figure 3.8).

RRX and SIBM are large driven by dislocation creep giving rise to relatively high strain rates (n = 4, Goldsby and Kohlstedt, 2001). At lower stresses grain boundary slip also becomes important, resulting in slightly lower strain rates (n = around 2, Goldsby and Kohlstedt, 2001). Grain boundary sliding is partially attributed to the presence of water at grain boundary triple junctions (Duval, 1977, Hobbs, 2000). Increased impurities at grain boundaries could lead to even more 'liquid' grain boundaries (Faria and others, 2014b).
Figure 3.8: Strain and strain rate for isotropic and anisotropic ice in uniaxial compression and simple shear. Top: Strain and strain rate for ice experiencing uniaxial compression at -3.3°C and 0.2 MPa (after Jacka and Budd, 1991). Three different initial fabrics were tested: ice with (A) a small circle girdle fabric; (B) near isotropic fabric and (C) an anisotropic fabric (single maximum fabric). All ice fabrics develop a stable new ice fabric of a small circle girdle at ~45° colatitude during tertiary strain. Fabric A shows a strain curve of easy glide, whereas fabric B and C deform in hard glide. Bottom: Strain and strain rate for ice experiencing simple shear (after Treverrow, 2009 based on experiments by Gao and Jacka, 1987). The single maximum fabric (A) deforms in easy glide whereas the isotropic fabric (B) deforms in hard glide. Both fabrics develop a single maximum fabric during tertiary strain. Stereonets are shown for theoretical horizontal thin sections.
When deformation of the ice stops, the ice still has an internal strain energy and becomes annealed. During this stress relaxation, ice crystals tend to grow into irregular interlocking (interlobate) ice crystals with typical multiple maxima ice fabrics (e.g. Gao, 1989).

### 3.3. Strain and ice fabric in ice shelves

A typical ice shelf is made up of three different zones that have different ice fabrics: (1) firn and ice formed from in situ accumulation, (2) meteoric ice that has been advected from glaciers that feed the ice shelf and (3) the marine ice zone (Wakahama and Budd, 1976, Craven and others, 2005, Rist and others, 2002). The top meters of an ice shelf characterized by local accumulation typically remain near isotropic (Figure 3.9, Table 3.2) (Gow, 1963, Eicken and others, 1994, Rist, 2002). Meteoric ice in in the top ~150 m (i.e. the second zone) develop small circle girdle fabrics due to horizontal extension (Figure 3.9, Table 3.2) (Gow, 1963, Eicken and others, 1994; Rist and others, 2002). Due to extensional flow of ice shelves constricted by land on the sides (e.g. Treverrow and others, 2010), large vertical circle fabrics would be expected to develop where the c-axes rotate toward a plane perpendicular to the tension axis (Van der Veen and Whillans, 1994) similar to ice divides (e.g. Wang and others, 2002). However, no study has yet reported such great circle girdle fabrics in meteoric ice from an ice shelf. This may be partially due to the complex strains that shelf ice can experience or due to the relatively higher strain rates in the floating ice in the order of $1 \times 10^{-2} \text{a}^{-1}$ to $1 \times 10^{-3} \text{a}^{-1}$ (e.g. Eicken and others, 1994, Treverrow and others, 2010) in comparison lower strain rates in land ice in the order of $1 \times 10^{-6} \text{a}^{-1}$ (e.g. Lipenkov and others, 1989) to of $1 \times 10^{-4} \text{a}^{-1}$ (e.g. Wang and others, 2002, Montagnat and others, 2014b). The presence of islands or ice rises within ice shelves (e.g. Treverrow and others, 2010, Matsuoka and others, 2015) could cause localized longitudinal compression of shelf ice and possibly lateral or basal shearing (.). Complex strain settings can also result in multiple maxima ice fabrics as observed below depths of ~150 m in several ice shelves (Table 3.2) (Gow, 1993, Eicken and others, 1994, Treverrow and others, 2010). These multiple maxima fabrics could also be a result of stress...
relaxation during flow stagnation (e.g. Russell-Head and Budd, 1979) and associated annealing.

Figure 3.9: Typical evolution of crystal size and ice fabric in horizontal thin sections with depth in a cold ice sheet to the left and an ice shelf to the right (modified from Alley, 1992), based on data summarized in Faria and others (2014a) for the inland ice and Table 2.1 for the ice shelf ice. Inland meteoric ice is initially isotropic after it has formed from snow. Vertical shortening allows for the formation of a weak single maximum fabric as a result of mainly rotation recrystallization (RRX) in cold ice. At depth strain intensifies since simple shear also occurs causing the development of stronger single maximum fabrics as a result of migration recrystallization (SIBM). Meteoric shelf ice is also initially isotropic toward the surface where it is nourished by local accumulation. Further within the ice shelf horizontal extension causes development of a circle girdle fabric during RRX and SIBM. The typical accretion fabric for marine ice is a single maximum fabric due to stacking of ice crystals and possibly vertical compression and simple shearing because of ocean currents. As marine ice gets older, it can get exposed to horizontal extension or shortening or a combination of both and then develops vertical small/great circle girdle fabrics during RRX/SIBM.
Unlike glacial ice, marine ice is naturally anisotropic. Discs of frazil ice crystals stack up due to their buoyancy and develop banded ice facies (Tison and others, 1993); these are frequently accompanied by strongly anisotropic fabrics (Tison and others, 1993, Treverrow and others, 2010, Dierckx and others, 2014) likely induced by the upward buoyancy pressure and laminar flow of ocean currents (Tison and others, 1993). Banded ice crystal facies and hence strongly anisotropic facies have only been observed in young marine ice closer to the ice-ocean interface as in the Amery Ice Shelf (Treverrow and others, 2010) or in marine ice exposed at the ice shelf surface due to locally high surface ablation as in the Hell’s Gate Ice Shelf (Souchez and others, 1991, Tison and others, 1993, Khazendar and others, 2001) or in ice shelf rifts as in the Nansen Ice Shelf (Dierckx and others, 2014) (Table 3.2).

Also common are multiple maxima ice fabrics in marine ice. They have been observed just above the ice-ocean interface (Rist and others, 2002, Eicken and others, 1994) (Table 3.2) similar to the multiple maxima ice fabrics in meteoric ice, where the strain is minimal and the temperatures are maximal, facilitating recrystallization in a low stress regime. Small or great circle girdle fabrics were observed in the middle and top of a marine ice layer in granular ice facies (Eicken and others, 1994, Rist and others, 2002, Treverrow and others, 2010, Dierckx and others, 2013), or in marine ice exposed at the ice shelf surface (Tison and others, 1993). Circle girdle fabrics likely develop in granular marine ice facies, since the ice is supposedly older and was affected by horizontal extension as a result of ice shelf flow and/or vertical compression due to continuous marine ice accretion.

Lateral shearing and compression due to flow against land can further influence the evolution of the marine ice fabric. Dierckx and others (2013) have determined from compression experiments that not considering the higher temperature of marine ice, the pure material properties of marine ice do not lend itself to faster deformation than meteoric ice since the anisotropy of marine ice is often not favorably aligned to the strain regime. However, laboratory experimental tests have to be conducted at much higher strain rates and shorter time frames than in real nature, and the ice crystal lattice responds differentially to different strain rates. Results thus need to be compared to real world examples of measured strain and marine ice crystallography.
Table 3.2: Meteoric and marine ice crystal properties in Antarctic Ice shelves. Ice crystal diameters are sometimes calculated from areas and rounded up or down to the nearest 0.5 mm.

<table>
<thead>
<tr>
<th>Ice Shelf</th>
<th>Core length (m)</th>
<th>Ice type</th>
<th>Depth (m)</th>
<th>Crystal size (diameter) (mm)</th>
<th>Ice fabric</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ross Ice Shelf (Little America V ice core)</td>
<td>~257</td>
<td>Meteoric</td>
<td>~50 to 61</td>
<td>~4.0</td>
<td>near isotropic</td>
<td>Gow, 1963</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>~110 to 150</td>
<td>~7.0</td>
<td>small circle girdle/ multiple maxima</td>
<td>Oerter and others, 1994</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>below 150</td>
<td>up to 11.5</td>
<td>multiple maxima</td>
<td></td>
</tr>
<tr>
<td>Füchser-Ronne Ice Shelf (B13)</td>
<td>215</td>
<td>Meteoric</td>
<td>0 to ~100</td>
<td>~0.5 to 5.5</td>
<td>--</td>
<td>Oerter and others, 1994</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>~100 to 153</td>
<td>5.5 to 7.0</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Marine</td>
<td>154 to ~190</td>
<td>1.0 to ~4.5</td>
<td>Close to isotropic</td>
<td>Oerter and others, 1992, 1994</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>190 to 215</td>
<td>~4.5 to ~7.0</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>215</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Marine</td>
<td>38</td>
<td>~1.5</td>
<td>--</td>
<td>Eicken and others, 1994</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>137</td>
<td>~6.5</td>
<td>Weak small circle girdle/ multiple maxima</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>154</td>
<td>~3.0</td>
<td>Weak small circle girdle</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>207</td>
<td>~7.0</td>
<td>multiple maxima</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(B15)</td>
<td>320</td>
<td>Meteoric</td>
<td>~50 to ~100</td>
<td>3.5 to ~5.5</td>
<td>--</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>100 to 153</td>
<td>~6.5 to 10.0</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Marine</td>
<td>15 to ~253</td>
<td>~3.5 to 7.0</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>~253 to 320</td>
<td>~8.0 to 13.0</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(B13/B14/B15/B24)</td>
<td>422</td>
<td>Meteoric</td>
<td>30 to 54</td>
<td>~1.0 to 4.0</td>
<td>Isotropic</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Marine</td>
<td>135</td>
<td>~9.0</td>
<td>Small circle girdle</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>162</td>
<td>~4.0 to 5.0</td>
<td>Great circle girdle</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>312</td>
<td>~12.0</td>
<td>Multiple maxima (weak)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Amery ice shelf (G1)</td>
<td>~320</td>
<td>Meteoric</td>
<td>70 to 270</td>
<td>~4.5 to 16.0</td>
<td>--</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>~270 to 320</td>
<td>~3.0 to 4.0</td>
<td>Great circle girdle</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(AM01/AM01b)</td>
<td>479</td>
<td>Meteoric</td>
<td>~255</td>
<td>~4.0 to 11.5</td>
<td>Multiple maxima</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Marine</td>
<td>~280</td>
<td>~1.5 to 3.0</td>
<td>Great circle girdle</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Marine</td>
<td>~280.5</td>
<td>~8.5</td>
<td>Small circle girdle</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Marine</td>
<td>390</td>
<td>~3.5</td>
<td>Single maximum</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(AM04)</td>
<td>603</td>
<td>Meteoric</td>
<td>~350</td>
<td>~10.5 to 13.0</td>
<td>Weak multiple maxima</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Marine</td>
<td>400 to 500</td>
<td>~1.5 to 4.0</td>
<td>Single maximum/ Great circle girdle</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Northern McMurdo Ice Shelf</td>
<td>~50</td>
<td>Freshwater</td>
<td>0 to 15.7</td>
<td>~10.0</td>
<td>Near isotropic with some planar c-axis distribution</td>
</tr>
<tr>
<td></td>
<td>Hell’s Gate Ice Shelf</td>
<td>1 to 1.5</td>
<td>Marine</td>
<td>0 to 1.5</td>
<td>0.5 to 20.0</td>
<td>Single maximum/ Small circle girdle</td>
</tr>
<tr>
<td></td>
<td>Nansen Ice Shelf</td>
<td>45</td>
<td>Marine</td>
<td>0 to 45</td>
<td>1.0 to 2.0</td>
<td>--</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Single maximum/ Weak small circle girdle</td>
<td>Dierckx and others, 2013</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Single maximum</td>
<td>Dierckx and others, 2014</td>
</tr>
</tbody>
</table>
3.4. Strain and ice crystal size and shape in ice shelves

Meteoric and marine ice crystals in ice shelves have distinctly different sizes and shapes. The mean ice crystal size of meteoric shelf ice crystals typically varies from 1 to 10 mm in diameter and increases with depth (Table 3.2) (Gow, 1963, Eicken and others, 1994, Rist and others, 2002) together with an increase in temperature (Rist and others, 2002, Treverrow and others, 2010). In the Ronne Ice Shelf meteoric ice crystals also become elongated with depth (Rist and others, 2002), a flattening likely induced by a strain increase in ice shelves. Ice crystals show an interlobate structure toward the base of the meteoric ice layer in the zone of multiple maxima fabrics of the Amery Ice Shelf (Treverrow and others, 2010). This phenomenon is typical for migration recrystallization likely induced by the higher ice temperatures closer to the ice-ocean interface and/or stress relaxation.

Marine ice crystals have crystal diameters smaller and larger than meteoric ice; they vary from 0.5mm to 20mm (Table 3.2). On average marine ice crystals are slightly smaller than meteoric ice crystals in ice shelves (Table 3.2). At the Amery Ice Shelf, for instance, the ice crystals in marine ice are almost one magnitude smaller than in meteoric ice (Figure 3.10). Smaller grain sizes in marine ice have been attributed to the presence of increased impurities due to the marine origin of the ice (e.g. Treverrow and others, 2010). Impurities such as dust particles or undissolved chemical impurities can pin the edges of the ice crystals and restrict them from growing larger (Faria and others, 2014b). Unlike meteoric ice, marine ice becomes younger with depth since it accretes at the ice shelf base. Nonetheless, the average ice crystal size in marine ice was also observed to increase close to the ice-ocean interface in several ice shelves, similar to meteoric ice (Table 3.2, Figure 3.6) (Oerter and others, 1992, Eicken and others, 1994, Rist and others, 2002). Crystal size is related to crystal shape; in general granular ice crystals are larger than elongated banded ice crystals. Banded ice crystals are common in marine ice and crystals are often elongated by a factor between 2 and 6 (Eicken and others, 1994, Khazendar and others, 2001). The crystal shape of marine ice has been related to the age of the ice crystals, whereby the banded ice crystals are supposedly younger, since they have been observed close to the ice-ocean interface in the Amery ice shelf,
whereas marine ice in upper layers is granular in shape (Treverrow and others, 2010). Some marine ice crystals also show interlobate crystal boundaries close to the ice-ocean interface indicative of SIBM.

When marine ice does not accumulate in layers but fills in an ice shelf rift, as at the Nansen Ice Shelf, its crystal shape and size remains relatively uniform with depth (Khazendar and others, 2001). Compression and shearing can lead to the development of folds within marine ice, which is especially discernible when the ice crystals are elongated (banded) (e.g. Tison and others, 1993, Khazendar and others, 2001, Dierckx and others, 2014).

Figure 3.10: Grain size evolution with depth in the Amery Ice Shelf. Data from the two different Amery Ice Shelf ice cores are plotted in different colours (from Treverrow and others, 2010).
3.5. Chapter objectives

Little is known about the effect of strain on the development of ice microstructure (ice fabric, crystal size and shape) in shelf ice, especially marine ice. Due to its consistently higher temperature and impurity content in comparison to meteoric ice, marine ice is often thought to deform more easily than meteoric ice (e.g. Khazendar and others, 2009). However, the initially strong anisotropic fabric of marine ice give it material properties that make it harder, especially during uniaxial compression (Dierckx and others, 2013) typical for ice shelf flow. In this chapter marine ice deformation is investigated in a natural strain setting. The Southern McMurdo Ice Shelf (SMIS) surface strain regime is measured and marine and meteoric ice samples were extracted along a flowline. Marine ice microstructure evolution with increasing total strain is investigated and compared to meteoric ice microstructure.

3.6. Methods

Marine and meteoric ice was extracted in shallow ice cores along a transect toward shore. Thin sections of ice were produced along the length of the collected ice cores and subsequently analyzed with an automatic ice fabric analyzer that generates digital pixel by pixel ice crystal c-axis data. These pixel data were processed with numerical codes to give orientation, size and shape of individual ice crystals. Full instrument and processing details are given in the following sections. Ice shelf surface strain was calculate from measured ice shelf velocity. Ice shelf surface ablation patterns were also measured.

3.6.1. Ice core sites

Two meteoric and two marine ice cores were collected from the surface of SMIS with increasing distance from shore (Figure 3.11, Table 3.3). The ice cores were extracted with a Kovacs corer in December 2010 and are 2.71m to 3.12m long (Table 3.3). The extracted marine ice cores C5 and C9 are identical to those analyzed in chapter 2 for ice chemistry. Marine ice cores were extracted from the top of ridges which show tilted layering of alternating darker and lighter marine ice (Figure 3.12). The ice shelf surface was mainly snow free at all meteoric and marine ice core sites (Table 3.3).
Figure 3.11: Study site showing marine ice and meteoric ice cores. The dotted line denotes the separation between the marine and meteoric ice zone after Fitzsimons and others, 2012. Note that C9 is closer to shore than C5 and C6 is closer to shore than C7. C5 and C9 are identical to cores C5 and C9 analyzed in chapter 2.

Figure 3.12: Details of the end of a marine ice ridge. Tilted layers of different colours in the marine ice are visible. Person for scale (photo credit: Michael Hambrey).
Table 3.3: Details of ice core sites.

<table>
<thead>
<tr>
<th>Core</th>
<th>Core Length (m)</th>
<th>Distance from shore (km)</th>
<th>Ice type</th>
<th>Ice shelf surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>C9</td>
<td>2.96</td>
<td>0.40</td>
<td>Marine ice</td>
<td>Patchy snow</td>
</tr>
<tr>
<td>C5</td>
<td>2.71</td>
<td>0.92</td>
<td>bare ice</td>
<td></td>
</tr>
<tr>
<td>C6</td>
<td>3.12</td>
<td>2.03</td>
<td>Meteoric ice</td>
<td></td>
</tr>
<tr>
<td>C7</td>
<td>3.03</td>
<td>3.55</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

3.6.2. Thin section analysis and digital crystal data processing

3.6.2.1. Thin section analysis

Thin sections were produced vertically parallel to the ice core axis. Care was taken to avoid rotation between individual core pieces, but sometimes the break in the ice was ambiguous. Hence, it is likely that the location of the vertical plane (for thin sectioning) varies along cores. Also, it was not possible to record the azimuth of the ice cores during extraction in the field. Thin sections were produced of a thickness between 0.3 and 0.5 mm with a conventional microtome in a walk in freezer at -15°C using the method detailed in Durand and others (2006). The thin sections were processed with a FA G50-white automatic fabric analyser manufactured by Russell-Head Instruments (Figure 3.13) (Russell-Head and Wilson, 2001, Wilson and others, 2003, 2007), which records the orientation of the optical axis of individual pixels of the ice crystals (43 µm in size) using nine white light emitting diodes (LEDs), of which one is oriented vertically and eight are aligned in a circle spaced by 45° and inclined at 20° from the vertical (note that Figure 3.14 shows an instrument with three vertical LEDs). Cross-polarizers (upper and lower polarizer) are rotated to get the orientation of the extinction plane for each pixel and each lamp (Figure 3.14). Combining the data, AVA (=’Achsenverteilungs-analyse’) images are generated, which are translated into ice crystal c-axis orientations using algorithms (Petenell and others, 2009, 2011, Russell-Head and Wilson, 2001, Wilson and others, 2003, 2007).
Figure 3.13: Schematic diagram of the automatic ice fabric analyser (Peternell and others, 2010).

Figure 3.14: Schematic diagram of the AVA image acquisition with the automatic ice fabric analyser (Peternell and others, 2009).
3.6.2.2. Digital crystal data processing

The digital pixel data were processed with FAME (Fabric Analyser based Microstructure Evaluation) (Peternell and others, 2014). These MATLAB codes combine the MTEX toolbox (Bachmann and others, 2010) and PolyLX toolbox (Lexa and others, 2005) to identify and delineate ice crystal boundaries. Hereby MTEX delineates individual ice crystals from the image data using quantitative texture analysis. The PolyLX toolbox facilitates a statistical and texture analysis of the ice crystal characteristics such as grain shape and grain size.

The data were reduced to every second row (pixel) to facilitate faster data processing. The FAME software defined culling criteria, geometric and retardation quality, were empirically set to larger than 85%. These parameters account for ambiguities and poor measurements due to impurities and overlapping ice crystals, especially at grain boundaries. Ice crystals were delineated automatically with the FAME codes, whereby the possible minimum crystal size is dependent on the thin section thickness and Fabric Analyser resolution. Hence, the minimum ice crystal diameter (=diameter of equal area (EAD) circle) was taken as 516 µm, which is the equivalent of twelve pixels (at 43 µm each) and is equal to the sample thickness of the thickest thin sections. An equal area dimension circle is a circle with an area equal to that of the polygon centered on the centroid. This small minimum crystal size was also chosen because small marine ice crystal sizes (0.5 to 1 mm) have been observed previously (Table 3.2). Also, a low misorientation angle of 3.5° was chosen to differentiate between grains. After Weikusat and others (2009), angle variations below ~3-4° are subgrain boundaries. Many studies choose a larger angle of ~5° to differentiate between ice crystals (e.g. Peternell and others, 2014) but because the frazil ice crystals in the banded ice facies of marine ice at SMIS have a very similar orientation, this study chooses a small angle to differentiate between ice crystals. This ensures that marine ice crystals with similar orientation in the banded ice facies are represented adequately (e.g. Figure 3.15). Nonetheless the crystal size results from the FAME processing are too dependent on the selection criteria (especially minimum crystal size). This study will thus not discuss crystal size results but moreover use the crystal size data as a selection criteria for the crystal shape analysis.
Figure 3.15: Examples of thin section processing for a thin section extracted at 0.62 m from C5. In a), c), d), and e) the ice crystals are coloured according to the trend flat only between 0° and 180. The thin section in b) shows the pixels of the ice crystals as recorded by the fabric analyzer. Ice crystal shape is calculated from ice crystals bigger than 0.9 mm as depicted in e).

For the ice crystal shape analysis, it was first determined whether there are one or two predominant grain size distributions by establishing the ratio of the mean area crystal size over the median ice crystal size. If this ratio is greater than 1.5, two different grain size distributions are predominant (Durand and others, 2006, Peternell and others, 2014). In this study, only the ice crystals larger than the EAD median, which has an average of around 0.9 mm for marine and meteoric ice, were analyzed for their shape. The smaller ice crystals have likely formed in the new strain regime and would thus always be granular. The ice crystal orientation is calculated as the shape preferred orientation (SPO) with the PolyLX toolbox and
displayed as a frequency rose histogram of the principal elongation azimuth of the ice grains (i.e. the average orientation of the longest axis). These rose diagrams are displayed with 10° bins. The radial length of each bar indicates the percentage of grains falling within than azimuthal bin. The bins of 180° to 360° are identical to the bins of 0° to 180° (Dierckx and others, 2014). The mean SPO is displayed in a thick black line with the circular standard deviation as the confidence interval (see Figure 3.16a). When the thin section does not have an SPO, no line is displayed.

The average grain c-axis azimuth (i.e. direction, 0° - 360°) and dip (i.e. co-latitude, 0° - 90°) is displayed on Schmidt equal area dimension stereonet plots (e.g. Figure 3.16b). Three eigenvalues (S1 ≥ S2 ≥ S3) calculated from the grain orientation tensor give an idea about fabric shape (Woodcock, 1977). Crystals in a perfectly isotropic thin section plotted as equal area stereonets have eigenvalues of equal size (and S1+ S2+ S3 = 1). In a single maximum cluster, S1 is much larger than the other eigenvalues, with S2 ≈ S3 < 1/3. For a great circle girdle fabric, where all c-axes are distributed in a plane, the first two eigenvalues approach the same value, S1 ≈ S2 > 1/3 (Durand and others, 2006). The isotropy (I = S3/S1) and elongation (E= 1 – (S2/S1)) of the fabric were calculated for each thin section after Benn (1994). As I increases to 1, the fabric becomes more isotropic. As E increases to 1, the fabric becomes more clustered. All data are displayed on a triangular diagram which graphically indicates whether the ice fabric is predominantly pointed, girdle-like (small or great circle) or isotropic (see Figure 3.17).

![Figure 3.16: Example ice crystal SPO and ice fabric diagrams](image)

*Figure 3.16: Example ice crystal SPO and ice fabric diagrams: (a) Example of a rose diagram showing the SPO of the ice crystals in one thin section. The mean orientation plus confidence intervals are displayed as thick black lines (b) Example of a Schmidt stereonet, lower hemisphere projection of vertical thin sections of ice crystal c-axis orientation.*
Ice fabrics are often extracted from horizontally cut thin sections. The azimuth and dip of the crystal orientation can, however, vary dependent on how the thin section was cut (Wilson and Peternell, 2011). Figure 3.18 theoretically illustrates how the ice crystal geometry for anisotropic and girdle fabrics (small and great circle) relate to the arrangement of the basal planes of the ice crystal lattice for horizontal and vertical thin sections. Similarly differences of horizontal and vertical thin sections were illustrated in Figure 3.6. In this study, orientation data are projected onto a vertical plane.

Figure 3.18: Typical ice fabrics (pointed, small circle and great circle girdle) plotted in a lower hemisphere Schmidt stereonet from horizontal and vertical thin sections.
3.6.3. Ice Velocity and surface ablation

A gridded network of velocity stakes approximately spaced 3 x 3 km apart was inserted on the surface of SMIS (n=22) with a steam drill, leaving ~0.6 m above ground (Figure 3.19). The stakes were made of wooden dowel and 1.2 m in length and 2.5 cm in width. Stake location was marked with a red marker flag on a bamboo stake. The exposed stake height above the ice shelf was repeat measured in November/December 2010 and November 2011 to get the total annual surface ablation. The horizontal GPS position of the stakes was also repeat-monitored in November 2010 and December 2011 with a Trimble R8 rover in fast static mode for 20 minutes. The rover was mounted on top of the stake in 2010 (Figure 3.19) to assure that the centre of the pole was measured. Some of the stakes were tilted during repeat measurement in 2011. Hence, a tripod was placed over the central point of the clearly visible original drill hole instead of the tilted stake. A base station was installed on the ice shelf and run for a minimum of 5 consecutive hours to link back to the permanent base station at Scott Base (SCTB). All data were postprocessed with 30 second GPS data from the permanent GPS station at Scott Base using the software Trimble Business Centre. Due to the large tidal movement of ~1.5 m of the ice shelf with only one high and low tide in a 24 hour period (Clifford, 2005), the vertical precision of the stake position measurement has a large error (~0.3 m) and has thus been discarded.

Data in this study were complemented by a data set of repeat stake measurements further afar on the ice shelf completed by Clifford (2005) during the seasons of 2002, 2003 and 2005 (Figure 3.20). In this study the base station was either located on Black Island (in 2002, 2005) or on the ice shelf (2003) dependent on the field camp location and its proximity to shore. The base station and rovers were Trimble 5700 receivers. The wooden dowels were 2.4m in length and 3.5cm in width and placed into holes hand-drilled by a Kovacs auger up to a depth of 1.2m. All data were postprocessed with 30 second GPS data from the permanent base station at McMurdo (MCM4) using Trimble Centre.
Figure 3.19: Marker pole with marker flag and mounted Trimble R8 on the surface of SMIS. Mount Discovery in the background.

Figure 3.20: Location of stakes on the ice shelf surface. Violet squares denote stakes surveyed in this study. Pink squares denote stakes surveyed by Clifford (2005). Marine ice cores are displayed as triangles, meteoric ice cores are displayed as circles. See Figure 3.11 for ice shelf location.
3.6.4. Strain rates

When ice is exposed to normal or shear stress (σ, the force that acts on a unit area) it becomes strained (i.e. deformed). Strain (ε) is defined as the ratio of the deformed material dimension to its original dimension. The rate at which this deformation occurs is called strain rate (\(\dot{\varepsilon}\)) and is calculated as the unit deformation over time (a\(^{-1}\)). Two dimensional surface strain rate in ice can be calculated from the measured ice surface velocity field (e.g. Treverrow and others 2010).

Deformation of ice is driven by stress, here illustrated as two dimensional stress vectors (\(\sigma_{ij}\)) that act on an ideal square, expressed in xy coordinates (\(\sigma_{xx}, \sigma_{yy}\), etc.). These are components of the stress tensor (\(\sigma\)) of equation 3-1, which is made up of normal stress vectors (\(\sigma_{xx}\) and \(\sigma_{yy}\)) and shear stress vectors (\(\sigma_{xy}\) and \(\sigma_{yx}\)).

\[
\sigma = \begin{bmatrix} \sigma_{xx} & \sigma_{xy} \\ \sigma_{yx} & \sigma_{yy} \end{bmatrix} \quad (3-1)
\]

Ice becomes strained as a response to stress. Typical strain regimes in ice shelves have been introduced in Figure 3.1 and can be summed up as extension, shortening and lateral shearing. The strain tensor is similar to the stress tensor in two dimensions (equation 3-2).

\[
\varepsilon = \begin{bmatrix} \varepsilon_{xx} & \varepsilon_{xy} \\ \varepsilon_{yx} & \varepsilon_{yy} \end{bmatrix} \quad (3-2)
\]

Figure 3.21 2D normal stress and shear stress vectors (\(\sigma_{ij}\)) acting on a square. Shear stress in italics.
Figure 3.22 simplifies shear deformation as typical around the margins of ice shelves. Displacement (U and V) results in deformation. Shear strain in ice is by definition symmetrical assuming volume conservation (i.e. \( \varepsilon_{xy} = \varepsilon_{yx} \)).

Figure 3.22: An initially square object (in a) experiences shear strain and displacement (U, V), this results in deformation (in b) and rigid body rotation (\( \omega \)). Modified from Lliboutry (1987).

Whilst strain can be calculated from displacements, strain rates (\( \dot{\varepsilon} \)) can be calculated from displacement rates, i.e. ice shelf flow velocity (e.g. Treverrow and others, 2010). The strain rate tensor in two dimensions is similar to the stress and strain tensors (equation 3-3).

\[
\dot{\varepsilon} = \begin{bmatrix}
\dot{\varepsilon}_{xx} & \dot{\varepsilon}_{xy} \\
\dot{\varepsilon}_{yx} & \dot{\varepsilon}_{yy}
\end{bmatrix}
\]  

(3-3)

SMIS surface strain rates (\( \dot{\varepsilon} \)) were thus calculated from an annual velocity field that was interpolated between the measured stakes at a resolution of 100 x 100 m using thin-plate smoothing splines in MATLAB. From the measured velocity of ice flow (\( \vec{u} \)), the extensional and compressional annual strain rates \( \dot{\varepsilon} \) in the xx and yy direction were calculated from the \( u_x \) and \( u_y \) displacement vectors, usually written as \( u \) and \( v \) (Cuffey and Paterson, 2010) (equation 3-4).
It is assumed that the difference in extension and compression between the $xx$ and $yy$ directions is balanced by vertical extension or compression (in the $zz$-direction) because ice in glaciers is nearly incompressible (volume conservation). Thus, the sum of vertical, horizontal and lateral strain rates equals zero (equation 3-5). Positive vertical strain rates indicate stretching and negative strain rates denote compression.

$$\dot{\varepsilon}_{xx} + \dot{\varepsilon}_{yy} + \dot{\varepsilon}_{zz} = 0$$

The surface shear strain rate ($\dot{\varepsilon}_{xy}$) was calculated using equation 3-6. Since no internal vertical displacement was measured as part of this study, it was not possible to calculate shearing in the vertical plane.

$$\dot{\varepsilon}_{xy} = \frac{1}{2} \left[ \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right]$$

Additionally rigid body rotation ($\omega$) can occur during ice flow (Figure 3.22, equation 3-7). This vorticity is by definition antisymmetric.

$$\omega = \frac{1}{2} \left[ \frac{\partial u}{\partial y} - \frac{\partial v}{\partial x} \right]$$

The calculated strain rates were rotated with respect to the local flow direction to get the strain tangent ($x$) and transverse ($y$) to the flowline after standard methods outlined by Cuffey and Paterson (2010) (MATLAB code in Appendix B).

### 3.7. Results

In this section annual ice shelf surface accumulation or ablation (expressed as a surface height change), flow velocity as well as vertical and surface strain rates are presented. A detailed account for the strain at every ice core site is given. Subsequently, the micromorphology of meteoric and marine ice is described.
3.7.1. Annual ice shelf surface accumulation/ablation, velocity and strain

At its surface SMIS loses mass (up to -0.31 m a\(^{-1}\)) in a ~5 km wide zone that runs parallel to Minna Bluff in the south of the ice shelf (Figure 3.23). In its north close to Black and White Islands snow or ice accumulates on the ice shelf surface adding up to 0.43 m a\(^{-1}\) (Figure 3.23). No assumptions were made regarding the density of the lost or gained mass, so the measurements quoted here are not equivalent to ice shelf surface mass balance.

SMIS flows from the northeast to the southwest, whereby its velocity decreases from 6 m a\(^{-1}\) to <1 m a\(^{-1}\) toward the shore of Minna Bluff (Figure 3.24). In its shear zone with the Ross Ice Shelf, the velocity of SMIS is about one magnitude higher and the ice shelf flows north at rate of ~44 m a\(^{-1}\) (Figure 3.24). The average horizontal precision of the measurement of the displacement stakes is ±0.05 m.

![Image: Measured and interpolated surface change (ablation and accumulation) on SMIS using data from the years 2002, 2003, 2005 (Clifford, 2015) and 2010 and 2011 (this study). See Figure 3.11 for ice shelf location. Marine ice core sites are denoted as triangles and meteoric ice core sites as circles.](image-url)
Toward shore the strain tangential to the flow direction increases, indicating shortening. This coincides with low strain normal to the flow direction. Thus, the data indicate a positive vertical strain rate and thickening toward shore (Figure 3.25) using equation 3-5. The shear strain rate at the ice core sites has a magnitude of around $-1 \times 10^{-4}$ a$^{-1}$ (Figure 3.26).

The strain rate at individual ice core sites is shown in equations 3-8 to 3-11 (expanding the 2D matrix in equation 3-3 to the zz dimension). The strain matrix are listed with decreasing distance from shore with the strain matrix for the ice core site furthest away from Minna Bluff (C7) in equation 3-8 and the strain for the ice core site closest to Minna Bluff (C9) in equation 3-11. The vertical strain ($\varepsilon_{zz}$) was calculated using equation 3-5. The maximum propagation error for all strain rates after Taylor (1997) is 0.1.

Figure 3.24: Measured and interpolated annual surface flow velocity, assuming constant flow. Black and green arrows are scaled to measured velocities from the more recent field seasons (2010 and 2011). The red arrows show scaled displacement measurements from field seasons 2002, 2003 and 2005 (Clifford, 2005). See Figure 3.11 for ice shelf location.
Figure 3.25: Ice shelf vertical strain rate per year transverse to the local flow direction (see Figure 3.24). Positive values (in blue) indicate thickening, negative values (in green) indicate thinning. See Figure 3.11 for ice shelf location.

Figure 3.26: Ice shelf surface shear strain rate per year transverse to the local flow direction (see Figure 3.24). See Figure 3.11 for ice shelf location.
Strain matrixes at individual ice core sites (equations 3-8 to 3-11) indicate that the strain regime is dominated by longitudinal compression (with a magnitude of $-3.14 \times 10^{-4} a^{-1}$ to $-3.69 \times 10^{-4} a^{-1}$). Compression is increasing toward shore for the meteoric ice core sites (equations 3-8 and 3-9), whereas compression decreases slightly for the marine ice cores (equations 3-10 and 3-11). Vertical strain is in the order of to $3.0 \times 10^{-4} a^{-1}$ and the shear strain is around $-1.32 \times 10^{-4} a^{-1}$ at the meteoric and marine ice core sites C6, C5 and C9 (equations 3-9 to 3-11) and only smaller at the meteoric core site of C7 (equation 3-8), which is furthest away from shore.

### 3.7.1.1. Meteoric ice age and accumulated strain

Due to the relatively low flow velocity at SMIS of just a few meters per year (Figure 3.24), SMIS ice is likely exposed to less accumulated strain than ice in faster flowing ice shelves, such as the Amery Ice Shelf, which flows several tens of meters per year (Treverrow and others, 2010). The difference in strain rate amounts to at least one order of magnitude. Nonetheless, slowly flowing ice can also deform substantially since strain accumulates over time and the ice fabric is strongly influenced by total strain (Montagnat and others, 2015). In
order to derive the minimum accumulated strain of SMIS shelf ice extracted at the ice core sites, strain was integrated along flow paths to the ice core sites. The minimum travel distance since accumulation for meteoric ice at site C7 was derived by following the ice flow lines back to the surface accumulation area of SMIS in its north (Figure 3.23). This equates to a travel distance of 5000 m to the ice core site (Figure 3.28), which corresponds to around 1300 years since deposition considering the annual flow rates. Ice extracted from the site of C6, which is closer to shore, would have thus experienced strain for around 2000 m further (equating to 700 more years) than ice from site C7 assuming a similar flow path (Figure 3.28).

Integrating the strain rates over the time it took the ice to travel to the core sites from the accumulation area gives around 20% accumulated vertical strain (mixture of shortening and extension) for ice extracted at C7 and around 12% accumulated vertical (extension only) strain for ice from C6. The accumulated shear strains along these flow paths are 63% for ice from site C6 and 44% for ice from site C7.
Figure 3.27: Flow paths of marine and meteoric ice at SMIS and ice shelf vertical strain rates (top panel) and surface shear strain rates (bottom panel).
Figure 3.28: Accumulated (a) vertical and (b) shear strain along a flowline toward the meteoric ice core site C6. The travelled time of 2000 years corresponds to a distance of 7000 m to the core site. The absolute accumulated strain is shown for both strain regimes. Hereby most of the vertical strain is shortening (until 750 years) and extension from 750 years onward (also refer to Figure 3.27). All of the cumulatively summed shear strain is negative.

Figure 3.29: Accumulated (a) vertical and (b) shear strain along a flowline toward the meteoric ice core site C7. The travelled time of 1300 years corresponds to a distance of 5000 m to the core site. The absolute accumulated strain is shown for both strain regimes. Hereby most of the vertical strain is negative (i.e. extensional) (also refer to Figure 3.27). All of the cumulatively summed shear strain is negative.
3.7.1.2. **Marine ice age and accumulated strain**

In order to estimate the accumulated strain of SMIS marine ice, the marine ice accretion area needs to be known. SMIS marine ice has likely formed more locally than SMIS meteoric ice, possibly partially from surface melt that is routed to the ice shelf base (see chapter 2). In this case marine ice now outcropping at the ice shelf surface could have once accreted at the ice shelf base within the marine ice zone of SMIS (Figure 3.11). Hereby the thickness of the ice shelf and ablation rate would determine how fast basally accreted marine ice would reach the ice shelf surface. Recent ice shelf surface radar investigations (Fitzsimons, personal communication) have shown that SMIS marine ice is at least 40 m thick (Figure 3.30).

![Figure 3.30: Radar profile along the intersection of marine and meteoric ice at SMIS. Locations A, B and C are marked in the inset map as well as ice core locations. The radar data was collected with 50MHz center frequency, 1000V pulse voltage and antenna separation of 2 m using a Pulse Ekko Pro system. The data was processed with a velocity of 0.1 m ns$^{-1}$ typical for ice to get the ice thickness. Typical isochrones within meteoric ice where traced in yellow and within marine ice traced in red. Additionally the marine ice basal ice shelf interface is tentatively traced in blue and continues outside the radargram according to data from nearby locations (Fitzsimons, personal communication).](image)

Thus considering an approximate ablation rate of 0.31 m a$^{-1}$ SMIS, marine ice must have a minimum age of about 130 years at site C9. The ablation rate at site C5 is slightly less (0.25 m a$^{-1}$) which means that the ice at this site would be slightly older than ice at C9, around 160 years. This assumes a quite localized accumulation from a frazil ice point source and a marine
ice thickness of 40m. However, it does not factor in that marine ice frazil ice crystals also take a while to consolidate.

Investigations of the chemical composition of marine ice in chapter 2 revealed that ice from sites C5 and C9 due, to the small isotopic range, could have formed from a mixture of sea water and recycled marine ice (Figure 2.11), whereby the water source (i.e. contribution of melted marine ice versus sea water) changes slightly over time as indicated by the co-isotopic mixing slope (Figure 2.10). Considering the observed tilt in the marine ice layers and the distance between core sites, a simple trigonometry calculation reveals that there could be an ice column of around 90m between ice layers extracted at site C5 and C9 (Figure 3.32) unless basal erosion occurs below the marine ice zone. It is thus likely that the marine ice layer is thicker than what was detected with the radar. Considering a marine ice layer thickness of >90m, it is also likely that the marine ice layer continues below the meteoric ice as depicted in Figure 3.32a. Thus, marine ice could also have accreted further down the flow line below the current layer of meteoric ice, especially should it have formed in an ice pump mechanism.

Consequently, the accretion location of marine ice could be determined from the difference in angle of the marine and meteoric ice isochrones at the meteoric – marine ice interface (location B in Figure 3.30). Overall the isochrones of meteoric ice show a steeper angle than those of marine ice and the angle of the meteoric ice isochrones becomes steeper closer to shore (Figure 3.30) consistent with a gradual increase in surface ablation (Figure 3.23). The marine ice layers have been measured to be inclined at around 11°, also visible in a photograph taken on the ice shelf surface (Figure 3.12). Scaling the marine ice isochrones to be inclined at 11° suggests that the meteoric ice layers close to B and site C6 in Figure 3.30 have an angle of around 28°. This means that at the time of accretion of the marine ice layers now outcropping at site C5 at the base of the SMIS meteoric ice, the meteoric ice layers must have had a tilt of around 17°. Assuming no significant change in the geometric configuration of SMIS over time, marine ice could have thus accreted close to site C7 where the meteoric ice layers show a tilt of tilt of around 17° (Figure 3.32a). Thus marine ice would have travelled around 3 km along its flow path (Figure 3.27: Flow paths of marine and meteoric ice at SMIS and ice shelf vertical strain rates (top panel) and surface shear strain rates (bottom panel).)
since accretion to the ice core sites, which would give an age of around 1200 years considering the flow rates (Figure 3.24). Ice extracted from site C5 would then be older than ice from C9. The relative age difference would depend on how long it would take to accumulate solid marine ice with a thickness of 90 m. If the marine ice accumulation rate is similar to the surface ablation rate, ice from C5 would be around 300 years older than ice from C9 (Figure 3.32a).

Figure 3.31: Two possible modes of marine ice accretion below meteoric ice: a) marine ice accretion in wide layers, where the difference in age between layer C5 and layer C9 is calculated from the distance between ice core sites and the dip of the ice layers using trigonometry b) marine ice accretion from different point sources, where the age difference between ice from different core sites calculated as the time that it takes to travel the distance between C5 and C9 (derived from the adjacent: 461 m) at the site of accretion (where the flow is around 1 m a\(^{-1}\)) minus the time it takes to accrete a marine ice layer of 90 m thickness. In both diagrams, the number subscripts next to C5 and C9 indicate how much younger that marine ice layer/site is in years in comparison to 0.
Alternatively marine ice could have also accreted at several point locations downstream (Figure 3.32b), whereby ice extracted from the different marine ice core sites would have accreted at different times and locations, and from different water source mixtures in line with the co-isotopic mixing slope of SMIS marine ice (Figure 2.10). Hereby the age difference between ice from C5 and C9 would be calculated as the time it takes to cover the 461 m distance between the two ice core sites as calculated by trigonometry from the surface distance between core sites and the inclination of the marine ice layers (Figure 3.31) minus the time it takes to accumulate an ice column of approximately 90 m of marine ice (estimated around 290 years). Since the flow velocity around C7 is around 2.3 m a\(^{-1}\) (Figure 3.24) this would take around 200 years. Hence, ice from C5 would still be older by about 90 years. However, given the uncertainty in accretion location and associated differences in flow speed along the flow path, ice from C9 could also be slightly older than ice from C5, if the ice would have been accumulated further downstream.

Neither above age estimate factors in the time it takes for the consolidation of marine ice frazil ice crystals. The relative age difference between ice extracted from C5 and C9 is also strongly dependent on the accumulation rate, which currently remains unknown. At the Amery ice shelf Craven and others (2009) detected a 37 – 103 m thick permeable marine ice layer and calculated a closure rate of the permeable layer of 0.32 m a\(^{-1}\) which equates to a quarter of their calculated marine ice accretion rate. Thus, dependent on the accretion rate of marine ice at SMIS, several meters to tens of meters unconsolidated marine ice are expected below SMIS. These are disregarded in the accumulated strain calculations that follow below.

The above consideration give an absolute minimum age of 130 years to the SMIS marine ice. However, considering the difference in in the dip of the accretions layers of meteoric and marine ice (Figure 3.30), it is quite likely that SMIS marine ice now outcropping at the ice shelf surface has once accumulated below the meteoric ice close to site C7. Additionally, since ice from C9 is now outcropping closer to shore and has thus experienced a slightly different strain history than ice from C5, the point accretion scenario (Figure 3.32b) is thus chosen to calculate accumulated strain.
Hence, marine ice from the site C5, which is closest to shore, would have accumulated 20% vertical strain (horizontal shortening) and 25% shear strain (Figure 3.32). Ice from sites C9 would have accumulated 21% vertical strain (horizontal shortening) and 20% shear strain. (Figure 3.33) (MATLAB code in Appendix B).

Figure 3.32: Accumulated (a) vertical and (b) shear strain along a flowline toward the marine ice core site C5. The travelled time of 1200 years corresponds to a distance of 3000 m to the core site.

Figure 3.33: Accumulated (a) vertical and (b) shear strain along a flowline toward the marine ice core site C9. The travelled time of 1100 years corresponds to a distance of 3000 m to the core site.
3.7.2. Micromorphology and ice fabric of meteoric and marine ice

3.7.2.1. Crystal size distribution

Crystal sizes (EADs) determined in the meteoric and marine ice thin sections using FAME are not robust, i.e. they depend largely on the selection criteria. Nonetheless a general trend can likely be observed comparing meteoric versus marine ice crystal sizes and size ranges (Table 3.4). Generally marine ice EAD are smaller than meteoric EAD. Hereby meteoric ice crystals have a larger interquartile range (IQR), whereas marine ice has a larger overall range in ice crystal sizes with a larger difference between minimum and maximum crystal size (Table 3.4). Indeed, the mean crystal EAD and IQR of meteoric and marine ice are significantly different (One-Way ANOVA, p=0.05).

Table 3.4: Meteoric and marine ice crystal properties at SMIS.

<table>
<thead>
<tr>
<th>Ice core</th>
<th>C7</th>
<th>C6</th>
<th>C9</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ice type</td>
<td>Meteoric ice</td>
<td>Marine ice</td>
<td></td>
</tr>
<tr>
<td>Distance to shore (km)</td>
<td>3.55</td>
<td>2.03</td>
<td>0.92</td>
</tr>
<tr>
<td>Mean crystal EAD (mm)</td>
<td>1.70 ± 0.12</td>
<td>1.66 ± 0.18</td>
<td>1.29 ± 0.10</td>
</tr>
<tr>
<td>Median crystal EAD (mm)</td>
<td>0.89 ± 0.04</td>
<td>0.95 ± 0.11</td>
<td>0.88 ± 0.06</td>
</tr>
<tr>
<td>IQR crystal EAD (mm)</td>
<td>1.46 ± 0.16</td>
<td>1.47 ± 0.36</td>
<td>0.79 ± 0.13</td>
</tr>
<tr>
<td>Min crystal EAD (mm)</td>
<td>0.52 ± 0.00</td>
<td>0.52 ± 0.00</td>
<td>0.52 ± 0.00</td>
</tr>
<tr>
<td>Max crystal EAD (mm)</td>
<td>11.00 ± 2.90</td>
<td>9.32 ± 1.12</td>
<td>18.39 ± 8.93</td>
</tr>
<tr>
<td>Mean crystal area (EAD) (mm²)</td>
<td>2.27 ± 0.30</td>
<td>2.18 ± 0.56</td>
<td>1.32 ± 0.21</td>
</tr>
<tr>
<td>Mean shape orientation azimuth (°)</td>
<td>--</td>
<td>4.92 ± 2.81 (n=2)</td>
<td>78.22 ± 22.08</td>
</tr>
<tr>
<td>Mean shape circular STD (°)</td>
<td>--</td>
<td>62.72 ± 1.03 (n=2)</td>
<td>53.82 ± 3.91</td>
</tr>
<tr>
<td>Mean colatitude (°)</td>
<td>67.75 ± 2.38</td>
<td>64.50 ± 5.09</td>
<td>68.31 ± 6.92</td>
</tr>
<tr>
<td>IQR colatitude (°)</td>
<td>21.00 ± 3.08</td>
<td>23.83 ± 5.49</td>
<td>19.38 ± 3.95</td>
</tr>
<tr>
<td>Mean eigenvalue S1</td>
<td>0.53 ± 0.06</td>
<td>0.49 ± 0.03</td>
<td>0.72 ± 0.09</td>
</tr>
<tr>
<td>Mean eigenvalue S2</td>
<td>0.33 ± 0.04</td>
<td>0.36 ± 0.02</td>
<td>0.18 ± 0.06</td>
</tr>
<tr>
<td>Mean eigenvalue S3</td>
<td>0.15 ± 0.02</td>
<td>0.15 ± 0.02</td>
<td>0.10 ± 0.03</td>
</tr>
<tr>
<td>Mean fabric isotropy (f)</td>
<td>0.28 ± 0.07</td>
<td>0.31 ± 0.06</td>
<td>0.15 ± 0.06</td>
</tr>
<tr>
<td>Mean fabric elongation (E)</td>
<td>0.37 ± 0.09</td>
<td>0.25 ± 0.14</td>
<td>0.74 ± 0.12</td>
</tr>
<tr>
<td>n (unless stated otherwise)</td>
<td>4</td>
<td>6</td>
<td>13</td>
</tr>
</tbody>
</table>

3.7.2.2. Crystal shape and shape orientation

Meteoric ice crystals in SMIS ice are typically equiaxed (a), whereas marine ice crystals are often elongated (b). Rose diagrams, displaying the shape orientation of all ice crystals in one thin section (e.g. Figure 3.16a), are often round for meteoric ice with radials pointing in all directions (Figure 3.36a and b). Rose diagrams from marine ice cores are often flattened and
elongated with radials pointing in opposing directions (Figure 3.36a and b). Most of the thin sections of meteoric ice show no shape preferred orientation (SPO). Indeed, in C7 no thin sections have an SPO (Figure 3.35a). However, 33% of the thin sections in C6, which was extracted from a site ~1.5 km closer to shore than C7, show ice crystals with a pronounced vertical SPO (Table 3.4, Figure 3.35b). Marine ice crystals show a SPO in almost all thin sections in both marine ice cores (Figure 3.36a and Figure 3.36b).

Thin sections from C5 and C9 show a near horizontal SPO with a slight tilt of 77° to 78° (Table 3.4) (90° would be perfectly horizontal). In C5, the thin section from a depth of ~1 m shows ice crystals that are oriented at a tilted angle, almost near vertical for the thin section collected at 0.93 m (Figure 3.36a). Thin sections from higher up and lower down in the core display a more horizontal SPO again (Figure 3.36a). In C9 the SPO of the ice crystals is more variable over the length of the core with a regular deviation from the horizontal orientation but also with no SPOs that are near vertical (Figure 3.36b). This is also portrayed in the slightly higher circular standard deviation than in C5 (Table 3.4). The SPO varies more all along C9 and some thin sections display no SPO at all (Figure 3.36b); only a total 62% of the thin sections in C9 have a significant preferred orientation.

Figure 3.34: Typical ice crystal shape in (a) meteoric ice cores as in C7 and (b) marine ice cores as in C5. The left image in each example shows ice crystal pixel data processed with FAME (Peternell and others, 2014) using a 3.5° angle to differentiate between grains, and only displaying ice crystals larger than the median (i.e. larger than 0.9 mm) that were used in the shape analysis. The ice crystals are coloured according to their trend flat only between 0° and 180°. The right image in each example shows ice crystals coloured according to their direction and dip orientation, with the same colour for every 180° (see small circle in bottom right corner). The numbers on the x and y axes denote the number of pixels, i.e. 200 pixels are 8.6 mm.
Figure 3.35: Ice crystal shape along the extracted meteoric ice cores in vertical thin sections; black = granular ice facies. Data for C7 are displayed in (a) and data for C6 in (b). The rose diagrams display the azimuth of the shape preferred orientation (SPO). The small numbers denote a ring around the circle that indicate percentages of radials pointing in a given direction.
Figure 3.36: Ice crystal shape along the extracted marine ice cores in vertical thin sections; black= granular, grey= mixed, striped= banded ice facies. Data for C5 are displayed in (a) and data for C9 in (b). The rose diagrams display the azimuth of the shape preferred orientation (SPO). The small numbers denote a ring around the circle that indicate percentages of radials pointing in a given direction.
3.7.2.3. **Folds**

Differential micro-movement within an ice crystal lattice can lead to apparent folding, which is visible in some SMIS thin sections. Folds are frequent in thin sections of banded marine ice crystals, especially in C5; around a depth of 1m the ice crystals are arranged with their long axes almost vertical (Figure 3.36a and Figure 3.37a), whilst ice crystals in thin sections above or below show a near horizontal elongationFigure 3.36a). Lower in the ice core, small scale (centimeters) folds are visible in the thin sections (e.g. Figure 3.37b). No folds or ice crystals elongated in the vertical direction are present in C9.

![Figure 3.37: Folds in ice core C5 in vertical thin sections.](image)

(a) 5.4 shows elongated ice crystals arranged diagonally, (b) 5.12 shows a small scale fold with two hinges. The left image in each example shows ice crystal pixel data processed with FAME (Peternell and others, 2014) using a 3.5° angle to differentiate between grains. The right image in each example shows ice crystals coloured according to their azimuth orientation. The numbers on the x and y axes denote the number of pixels.

3.7.2.4. **Ice crystal fabric**

In the meteoric ice cores great circle girdle fabrics are typical in both ice cores (Figure 3.38a and b). Between ~0.8 m and 1.8 m depth in C6 the circle girdle fabric is slightly rotated around the vertical axis (Figure 3.38b). Thus the visible pattern in the Schmidt stereonet changes from an almost equally distributed ring around the edge of the circle to a thick band of points through the middle of the circle. This might be an artefact of the thin sectioning process during which the pieces of the ice core could have been rotated (misaligned), leading to an intersection of
the fabric at a different angle. One thin section from C7 at ~ 2.7 m depth shows a weak cluster fabric with a predominant vertical c-axis azimuth but it still has a large colatitude and shows remains of a great circle girdle fabric pattern (Figure 3.38a). The mean colatitude of ice fabric in the meteoric ice cores at SMIS lies between 64.5° and 68.0° with a standard deviation between 21.0° and 24.0° (Table 3.4). The IQR of the colatitude is higher in C6 than in C7 (Table 3.4).

The ice crystals of the marine ice cores (C5 and C9) mainly show pronounced single maximum ice fabrics, with a vertical azimuth and a large colatitude of 63.5° to 68.5° (Figure 3.39a and b). The tilt of the c-axis (i.e. azimuth) becomes slightly subvertical along the length of the ice cores (Figure 3.39a and b). Ice fabric in a C5 thin sections from around 1 m depth (Figure 3.39a), for instance, show a single maximum at right angle to the elongation axis of the SPO (Figure 3.36a). In C9 the single maxima are not as pronounced as in C5 (Figure 3.39b). The mean colatitude of the ice fabric is around 5° smaller in C9 than in C5 with a 4° larger IQR (Table 3.4). Many thin sections in C9 (69%) show hints of a great circle girdle fabric (at 0.53 m, 0.88 m, 0.97 m, 1.15 m, 1.43 m, 1.58 m, 2.00 m, 2.64 m and 2.86 m depth, Figure 3.39b). In C5 only around 31% of thin sections show hints of a great circle girdle fabric (at 0.78 m, 1.07 m, 2.20 m and 2.29 m depth, Figure 3.39a). Even though the mean co-latitude and IQR of the ice fabric in the meteoric and marine ice cores is similar (Table 3.4), the marine ice cores show a much more pronounced cluster fabric than the meteoric ice cores, whilst the meteoric ice cores show are more pronounced small or great circle girdle fabric.
Figure 3.38: Ice crystal fabric per grain along the extracted meteoric ice cores in vertical thin sections. C7 is shown in (a) and C6 is shown in (b); black= granular ice facies. Breaks in the ice core are shown indicated by irregular white lines.
Figure 3.39: Ice fabric per grain along the extracted marine ice cores in vertical thin sections. C5 is shown in (a) and C9 in (b); black=granular, grey=mixed, striped=banded ice facies. Breaks in the ice core are indicated by irregular white lines.
The eigenvalues of the orientation tensor allow for a quantification of the anisotropy of the ice crystal fabric. In the meteoric ice cores on average the first eigenvalue, $S_1$, is close to 0.50, $S_2$ is close to 0.35 and $S_3$ is on average 0.15 (Table 3.4). The fabric of meteoric ice only has an isotropy factor of around 0.3 (Table 3.4), where 1 would be perfectly isotropic. However, the fabric of meteoric ice also only has a small elongation factor ($E=0.25$ and 0.37, Table 3.4), where 1 would denote a very elongated fabric. The eigenvalues of the ice fabric do not show a distinct trend or variation with depth (Figure 3.40). The meteoric ice core data plot quite centrally in the triangular eigenvalue diagram denoting the isotropy and elongation of the fabric (Figure 3.41). Thus, meteoric ice has weak great circle girdle fabric and is not strongly isotropic (Figure 3.41).

In the marine ice cores the maximum eigenvalue, $S_1$, is on average larger than that in the meteoric ice cores (between 0.63 and 0.72) (Table 3.4), which is indicative of a higher level of anisotropy, typical for single maximum fabrics. C5 shows on average higher $S_1$ eigenvalues (and thus more pronounced single maximum fabrics) than in C9 (Table 3.4). All marine ice fabrics show an eigenvalue pattern of $S_1>S_2>S_3$ (Figure 3.42). The eigenvalues become more similar to each other in C5 below a depth of 1.5 m, especially the eigenvalues $S_2$ and $S_3$ (Figure 3.42), indicating a weaker single maximum or possibly a weak great circle girdle ice fabric in the lower four thin sections (Figure 3.39a). The eigenvalues of ice fabric from C9 show no trend with depth (Figure 3.42). Overall, the isotropy and elongation triangle indicates that ice fabric data from C9 are more girdle-like than ice fabric data from C5 (Figure 3.43). At least the single maxima in C9 are much weaker than in C5. Nonetheless the great circle girdle fabric in the marine ice cores is not as strong as in the meteoric ice cores. The mean eigenvalue $S_3$ is on average smaller in the marine ice cores than in the meteoric ice cores (Table 3.4); this also means that the isotropy is smaller (Table 3.4, Figure 3.42). The ice fabric in the ice cores collected closer to shore appears more girdle-like for both, marine and meteoric ice (Figure 3.41 and Figure 3.43).
Figure 3.40: Principal eigenvalues (S1, S2, S3) in the meteoric ice cores and their evolution with depth. Data from the ice core extracted closer to the shore of Minna Bluff has black symbols.

Figure 3.41: Eigenvalue ratios for meteoric ice fabrics from sites C7 (diamonds) and C6 (circles). Isotropy and elongation are plotted on a triangular diagram to illustrate the nature of the ice fabric (isotropic, girdle or cluster) for meteoric ice crystals in thin sections. The data is colour coded according to depth (scale on the right in m).
Figure 3.42: Principal eigenvalues (S1, S2, S3) in the marine ice cores and their evolution with depth. Data from the ice core extracted closer to the shore of Minna Bluff has black symbols.

Figure 3.43: Eigenvalue ratios for marine ice fabrics from sites C5 (diamonds) and C6 (circles). Isotropy and elongation plotted on a triangular diagram to illustrate the nature of the ice fabric (isotropic, girdle or cluster) for marine ice crystals in thin sections. The data is colour coded according to depth (scale on the right in m).
3.8. Discussion

In this section meteoric ice and marine ice fabric and microstructure evolution are discussed in light of local accumulated strain of the ice shelf. Observed SMIS ice microstructure and fabric are placed in context with previous in situ ice shelf studies but are also compared to results from laboratory experiments. The difference between marine and meteoric ice deformation down a flow line at SMIS is discussed and it is evaluated whether there is enough evidence to determine which ice type deforms more easily in a given strain regime.

3.8.1. Strain regime and ice fabric of meteoric ice

Accumulated strains of ≥ 10% and ≤ 60% in SMIS meteoric ice (Figure 3.28 and Figure 3.29, Table 3.5) are likely large enough to alter the original ice crystal fabric and microstructure by causing a reorientation of the SMIS crystal lattice. After Jacka and Maccagnan (1984) tertiary strain may be reached at around 10% accumulated strain. Other studies suggest that a complete reorientation of the ice crystal lattice to the current strain regime only occurs at >35% total strain (Gao and Jacka, 1987, Wang, 1994, Wilson and others, 2014). Given that only the minimum accumulated strain was calculated in this study, the total strain could be much higher. Nonetheless, there is also the added complexity of a changing strain regime down the flow path at SMIS, which makes tertiary creep less likely. In any case, SMIS meteoric ice has been exposed to enough total strain to have an altered ice crystal fabric and microstructure.

Toward the top of ice shelves the ice fabric is usually isotropic (Table 3.2, Gow, 1963, Eicken and others, 1994, Rist and others, 2002) since the ice is young and has not experienced much strain yet. SMIS meteoric ice at the surface is old and crops out due to locally high surface ablation (Figure 3.23). As expected in a regime of longitudinal compression and vertical extension (Figure 3.25 and Figure 3.26) as SMIS flows toward land (Figure 3.24), SMIS meteoric shelf ice shows weak great circle girdle fabrics (Figure 3.38a and Figure 3.38b) with S1 and S2 eigenvalues that are almost equal (Table 3.5, Figure 3.40). Preferential fabric alignment into circle girdles promotes faster deformation of the ice in a tensile strain regime than if the ice was isotropic (Figure 3.8) since the basal planes of the ice lattice align preferentially to the current strain regime (e.g. Lipenkov and others, 1989). Indeed,
there is an acceleration of the vertical strain rate closer to Minna Bluff and the cores site of C6 in comparison to C7 (Figure 3.25) and the calculated total vertical strain is higher (Table 3.5). Furthermore, the ice fabric in C6 is overall more girdle-like than the ice fabric of C7 with a decrease in fabric elongation (Figure 3.37 and Figure 3.44). Nonetheless, the ice fabrics in neither C7 nor C6 are perfect circle girdle fabrics (i.e. S1 and S2 are not equal, Table 3.4) since SMIS meteoric ice is also experiencing significant shear strain (Table 3.5) which would actually encourage the development of an elongated or pointed fabric (Alley, 1992). Indeed, single maximum fabrics have been observed in other shelf ice downstream of an ice rise in the Filchner-Ronne Ice Shelf (Eiken and others, 1994), where the shelf ice was subjected to shearing along the margins of the ice rise (MacAyeal and others, 1998). At SMIS the shear strain also increase toward the shore of Minna Bluff (Figure 3.26). However, the meteoric shelf ice at SMIS does not show a very cluster-type ice fabric (Figure 3.41). Much like other ice shelves, SMIS shelf ice experiences a complex strain regime of shearing and compression, which encourages the development of small or great girdle and cluster fabrics in a secondary and tertiary strain regime. Given that the shear strain regime dominates with accumulated strains ≥ 45%, but great circle girdle fabrics are more frequent (Figure 3.44) clearly indicates that ice at either meteoric ice core site is not experiencing tertiary creep.

3.8.2. Strain regime and ice crystal shape in meteoric ice

The much smaller average grain size (diameter, EAD) of SMIS meteoric ice (~1.70 mm) (Table 3.5) than that at other ice shelves (between 1.5 and 16 mm) (Table 3.2) is likely due to the grain segmentation technique of the FAME codes. Meteoric ice crystal size is thus not further discussed.

The development of an SPO in many C6 thin sections in contrast to C7 (Table 3.5, Figure 3.36, Figure 3.44) further suggests that the ice closer to shore was subjected to more strain. Whether compressional or shear strain is responsible for the development of SPOs remains uncertain. At depth in the Ronne Ice Shelf (Rist and others, 2002), extensional strain was likely responsible for the development of elongated ice crystals in a horizontal direction, whereas simple shear caused the development of elongated grains at the Sorsdal Glacier
Since there is a much higher increase in the (minimum) accumulated shear strain at the SMIS core site of C6 (i.e. 60% instead of 45% at C7, Table 3.5, Figure 3.44), internal shearing could be responsible for the development of an SPO in SMIS meteoric ice. However, given that the SPO developed in an almost vertical direction (Figure 3.36b and Figure 3.44), it cannot be precluded that the compressional strain caused grain elongation along the vertical axis, even though the total vertical strain is much less.

Table 3.5: Summary of ice crystal properties and accumulated strain in meteoric ice.

<table>
<thead>
<tr>
<th>Core</th>
<th>Accumulated strain</th>
<th>Ice fabric</th>
<th>Crystal Shape/ folds</th>
</tr>
</thead>
<tbody>
<tr>
<td>C7</td>
<td>• 12% vertical strain • 45% shear strain</td>
<td>• great circle girdle fabrics, one cluster-type fabric • not very isotropic but slightly elongated fabric (E = 0.37) • S1 almost double of S2 which is almost double of S3</td>
<td>• no SPO – equiaxed crystals • no folds</td>
</tr>
<tr>
<td>C6</td>
<td>• 20% vertical strain • 60% shear strain</td>
<td>• great circle girdle fabrics • neither very isotropic nor elongated fabric • S1 and S2 are similar</td>
<td>• near vertical SPO in 40% of the thin sections, otherwise equiaxed crystals • no folds</td>
</tr>
</tbody>
</table>

Figure 3.44: Typical meteoric ice properties and total strain in each ice core. Typical thin sections of SMIS meteoric ice crystals (left), typical SPOs (left of the middle), typical ice fabrics with eigenvalues (right of the middle) and total strain for compression/vertical extension and shearing (right).
3.8.3. Strain regime and ice fabric of marine ice

Due to differential surface ablation at SMIS (Figure 3.23), marine ice shows tilted layers (Figure 3.12) – possibly accretion layers - and thus the single maximum fabric in marine ice also has a slight tilt and is not perfectly aligned to the vertical (Figure 3.39). The calculated minimum vertical and shear strain at the marine ice core sites are between 20 and 25% (Figure 3.32 and Figure 3.33, Table 3.6), whereby total shear and vertical strain at both ice core sites are almost equal. The older marine ice from C5 is dominated by total shear strain whereas the ice from C9 has experienced slightly more total vertical strain (Figure 3.34 and Figure 3.36).

In keeping with this, ice from the younger marine ice core site closer to Minna Bluff, C9, shows less pronounced single maxima as indicated by a lower elongation factor E and an a larger IQR of the ice fabric co-latitude (Table 3.4) and also tendencies of small or great circle girdle ice fabrics with the eigenvalues S1 and S2 becoming more similar to each other (Table 3.4, Figure 3.43, Figure 3.42 and Figure 3.45). Since the strain regime does not change significantly from C5 to C9 (see equations 3-10 and 3-11), the differences in ice fabric between cores could thus also be due to natural variability. Furthermore, weak multiple maxima ice fabrics are present in marine ice at C9 (Figure 3.39), similar to marine ice from the Amery Ice Shelf (Treverrow and others, 2010), possibly indicating adjustment of the crystal lattice to a complex strain regime or high temperature and little strain. SMIS marine ice could also have inherited strain from suggested previous grounding events of SMIS (e.g. Debenham, 1919). However, the large S1 eigenvalue in ice from both sites could also indicate that SMIS marine ice is still dominated by a single maximum fabric typical for undeformed marine ice (Table 3.4). The slight trends in eigenvalues with depth (Figure 3.42) could also be a relict from accretion and initial consolidation processes of marine frazil ice crystals; S1 is stronger at shallow depth, where marine ice is older, and weakens deeper within the ice shelf, where marine ice is relatively younger (Figure 3.42). Maybe the buoyancy pressure during the consolidation process caused strengthening of the anisotropic marine ice fabric in older layers.

Marine ice in the Amery and Filchner-Ronne Ice Shelves shows more pronounced great circle girdle fabric development than the marine ice extracted from both ice core sites at
SMIS (Treverrow and others, 2010, Rist and others, 2002). However, the Amery Ice Shelf and Filchner-Ronne ice shelves flow much faster than SMIS, which results in strain rates that are one magnitude higher (e.g. Treverrow and others, 2010). Furthermore, ice from these ice shelves was extracted where the ice experienced extensional and not compressional strain. Since surface shear and compression is occurring at SMIS’s ice margin close to Minna Bluff (Figure 3.25 and Figure 3.26), it is quite likely that subsurface internal shear is occurring also as the ice shelf flows toward shore (Figure 3.24). Internal shear could happen differentially in distinct planes of the ice shelf. Hulbe and Whillans (1997) suggested that the preferential orientation of ice crystals within ‘weak bands’ of an ice stream would lead to faster internal deformation of meteoric ice. There is evidence that SMIS marine ice also deforms non-uniformly in ice from C5. Selected thin sections show the development of great circle girdle fabrics, especially around a depth of -0.75 m, -1.10 m and between -2.20 m and -2.30 m (Figure 3.39a) mirrored by eigenvalue S1 becoming smaller at those depths (Figure 3.42). Given that these deformation ice fabrics predominantly occur in the granular ice facies in C5 (Figure 3.39a) suggests that only part of the ice column has experienced RRX and/or SIBM and thus nucleation of new (granular) grains. Further evidence for this is a tilt in the anisotropic fabric from the vertical to the subvertical at depths between -0.70 m and -1.20 m (Figure 3.39a) coinciding with the presence of folds in C5 (Figure 3.37). However, given that the accumulated strains are low, features observed as folds could also be due to actual differences in the orientation of the ice crystals as they accrete.

**Table 3.6: Summary of ice crystal properties and accumulated strain in marine ice.**

<table>
<thead>
<tr>
<th>Core</th>
<th>Accumulated strain</th>
<th>Ice fabric</th>
<th>Crystal Shape/ folds</th>
</tr>
</thead>
<tbody>
<tr>
<td>C5</td>
<td>20% vertical strain 26% shear strain</td>
<td>strongly anisotropic and cluster-type fabrics</td>
<td>horizontal or vertical SPO in 100% of thin sections</td>
</tr>
<tr>
<td></td>
<td></td>
<td>very strong fabric elongation (E = ~0.75)</td>
<td>folding apparent</td>
</tr>
<tr>
<td></td>
<td></td>
<td>strong S1, weak S2 and S3</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>hints of great circle girdle fabrics</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C9</td>
<td>22% vertical strain 21% shear strain</td>
<td>anisotropic, cluster-type fabrics</td>
<td>horizontal or vertical SPO in 62% of thin sections</td>
</tr>
<tr>
<td></td>
<td></td>
<td>strong fabric elongation (E = ~0.60)</td>
<td>mean shape circular standard deviation increases by ~5.1° in comparison to C5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Strong S1, weaker S2 and very weak S3</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>hints of great circle girdle fabrics</td>
<td>no folds</td>
</tr>
<tr>
<td></td>
<td></td>
<td>some multiple maxima</td>
<td></td>
</tr>
</tbody>
</table>
Overall, the slightly tilted marine ice fabrics at SMIS could be preferentially oriented for internal shearing but not uniaxial compression. Pimienta and others (1987) suggested that it is 25 times easier to deform an ice polycrystal with a single maximum fabric in shear perpendicular to the mean c-axis orientation than in uniaxial compression along the same orientation (Pimienta, 1987). Laboratory experiments by Dierckx and others (2014) corroborated that (folded) anisotropic marine ice generally slows deformation in uniaxial compression apart from when fold hinges are oriented at 45° to the compression axis. At SMIS the calculated accumulated strains in marine ice are low and the ice fabric seems to only partially align to the current strain regime. This study provides no evidence that the marine ice anisotropy enhances deformation in uniaxial compression and there is only weak evidence that marine ice anisotropy enhances deformation during internal shearing.
3.8.4. Strain regime and ice crystal shape in marine ice

Whilst newly accreted marine ice likely has an SPO due to the elongated nature of the highly anisotropic ice crystals (e.g. Treverrow and others, 2010), ice flow and deformation have been shown to modify a pre-existing SPO (Dierckx and others, 2014). Almost all thin sections of SMIS marine ice show an SPO at both ice core sites, many of which have an orientation near parallel to the ice shelf surface (Figure 3.36). However, many thin sections in both ice cores also show a subvertical SPO (Figure 3.36) a rotation cause by the strain regime of shear and compression. Only 63% of the thin sections in C9 have a significant SPO (Figure 3.36) whereas all of the thin sections in C5 have a significant SPO.

Marine ice at SMIS has a relatively small average crystal size of ~1.30 mm with a maximum measured crystal size of ~18.00 mm in C5 (Table 3.4). This is consistent with the range of crystal sizes measured in marine ice of other ice shelves (Table 3.2). In particular average marine ice crystal sizes in this study compare well with those of Dierckx and others (2013, 2014), who also used the FAME codes. In longer marine ice cores extracted from other ice shelves the crystal size roughly increases with depth (Eicken and others, 1994, Oerter and others, 1994, Rist and others, 2002), indicating that newly formed marine ice has larger ice crystals. Given that the FAME codes produce non-robust crystal size data, however, the ice crystal size data is not further discussed here.

3.8.5. Marine and meteoric ice microstructure compared

Even though strain regime of horizontal extension and lateral shear are typical for ice shelves (Figure 3.2), pinning points, which enhance the ice shelf buffering capability (Drews and others, 2015, Favier and others, 2016), also cause localized strain regimes similar to the ones that were observed at SMIS (e.g. Treverrow and others, 2010, Berger and others, 2016). Marine and meteoric ice both experience simple and pure shear at SMIS (Figure 3.25 and Figure 3.26). Hereby the vertical strain rate is highest at sites for both ice types (Figure 3.11 and Figure 3.12, equations 3-8 to 3-11). However, meteoric and marine ice was not exposed to similar total strains, which makes a comparison difficult. Whilst both ice types have experienced around 20% total accumulated vertical strain, meteoric ice has experienced at
least double the shear strain (i.e. ≥ 45%) of marine ice (Table 3.5 and Table 3.6). Nonetheless, all meteoric and marine ice analyzed in this study shows signs of increased dynamic recrystallization; with decreasing distance to shore the ice fabric loses its elongation and develops a slight great circle girdle, and an SPO develops in meteoric ice and ceases to exist in marine ice (Table 3.5, Table 3.6 and Figure 3.46).

There is no crystallographic evidence that marine ice deforms more easily than meteoric ice. Indeed, there is only evidence for the opposite scenario, albeit this evidence needs to be interpreted with caution since meteoric ice has experienced more than double of shear strain and consistently more vertical strain at both ice core sites (Table 3.5 and Table 3.6). And the calculated total strain was just a minimum estimate. Accordingly, girdle-type fabrics are more pronounced in meteoric ice than in marine ice; S1 and S2 are on average more similar in both meteoric ice cores than in the marine ice cores (Table 3.4). Due to its initially isotropic fabric (if formed locally in the north of SMIS) meteoric ice at SMIS may deform more easily in the SMIS strain regime than the non-preferentially oriented anisotropic marine ice. Considering the mechanical ice properties, ice fabric is the most important parameter in encouraging enhanced ice deformation followed by ice crystal size (Cuffey and Paterson, 2010). Additionally, experiments have shown that a second phase in ice, such as the impurities of marine ice (see chapter 2 for soluble impurities and Treverrow and others, 2010 for solid impurities in marine ice), also slows the development of deformation fabrics (Cyprych and others, 2016).

Marine ice is generally much warmer than glacial ice due to its formation from ocean water. Experimental studies have shown that ice at a higher temperature deforms faster than colder ice, progressively aligning its ice crystal fabric to the current strain regime (e.g. Budd and Jacka, 1989, Wilson and Peternell, 2012). Marine ice could have experienced relatively more internal shear (that was not measured in this study) since it is situated closer to the shore and the shear strain rates are thus higher than in meteoric ice on average (equations 3-8 to 3-11, Figure 3.46). This could have resulted in the subvertical anisotropic marine ice fabrics (Table 3.7). However, folds are also present in marine ice (Figure 3.37), and experiments have shown that they would slow down deformation (Dierckx and others, 2014).
Data from this study cannot confirm whether marine ice temperature or its crystallography primarily influence its behaviour. However, the marine ice in this study still shows original accretion features and there is no evidence that it deforms more easily than meteoric ice, supporting experiments which detected harder marine ice material properties (Dierckxls and others, 2014).

![Figure 3.46: Interpretive diagram of typical ice fabric and ice crystal shape (in vertical thin sections) at ice core sites investigated in this study (i.e. ice crystal shape and fabric evolution with proximity to shore). Strain rates are given in x 10^{-4}.](image)

**Table 3.7: Summary of ice crystal properties supporting certain types of deformation**

<table>
<thead>
<tr>
<th></th>
<th>Evidence for internal shearing in marine ice</th>
<th>Evidence for compression in marine ice</th>
<th>Evidence for easier deformation of meteoric ice versus marine ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>Strain</td>
<td>--</td>
<td>- -</td>
<td>Vertical strain rate increases at downstream meteoric ice site but not at marine ice core sites. Also, much higher total strains (vertical and shear) at the downstream meteoric ice core site.</td>
</tr>
<tr>
<td>Ice fabric</td>
<td>subvertical single-maximum fabrics</td>
<td>weak great circle girdle fabrics in ice from both marine ice core sites</td>
<td>More pronounced great circle girdle fabric development in meteoric ice versus marine ice</td>
</tr>
<tr>
<td>SPO</td>
<td>presence of subvertical SPO and folds in 100% of thin sections at the older marine ice core site</td>
<td>only 60% of SPOs in ice from C9 that experiences overall more total strain (sum of vertical and shear strain)</td>
<td>Downstream development of SPO in meteoric ice crystals versus loss of SPO in marine ice crystals</td>
</tr>
</tbody>
</table>
Since marine ice predominantly accretes where the ice shelf is thinner or where it is weak, a different rheology in comparison to meteoric ice may have an impact on ice shelf flow in zones that are more prominent to ice shelf disintegration than the rest of the ice shelf. When anisotropic marine ice occurs in layers in the middle on an ice shelf (e.g. at the Filchner-Ronne Ice Shelf, Lambrecht and others, 2007) it will be subjected to a tensile strain regime, but for a shorter amount of time than the overlying meteoric ice since it accreted downstream of the ice shelf grounding line. Thus, whilst the meteoric ice crystal structure may have had time to adjust its ice fabric to the strain regime, marine ice may be exposed to the strain regime for too little time to adjust its crystalline structure favourably. If the anisotropy of marine ice indeed outweighs temperature effects on ice rheology, marine ice would slow down flow and horizontal extension contributing to prolonging ice shelf live.

Marine ice accreted in areas of shear strain, such as the suture zone of the Larsen B ice shelf (Kulessa and others, 2014), would either enhance or slow down deformation, dependent on the direction of the shear strain. If the strain is lateral such as at SMIS, it acts at right angles to the basal planes of the ice crystals and deformation is slowed. If lateral shearing is accompanied by horizontal shearing of the internal layers, the anisotropy of marine ice could enhance deformation along internal planes. Thus, the effect of marine ice on ice shelf flow depends largely on the strain regime whereby its crystal lattice is not favourably oriented to the most common strain regimes (extension, compression, lateral shear).

3.9. Conclusion

In this chapter the development of marine ice microstructure along a flowline was assessed. Additionally marine and meteoric ice microstructure evolution with increasing total strain were compared. Major findings of this study are:

- Marine ice at SMIS is exposed to longitudinal compression as the ice flows toward Minna Bluff in the order of $3.5 \times 10^{-4}$ a$^{-1}$, accompanied by vertical thickening in the order of $3.0 \times 10^{-4}$ a$^{-1}$. Additionally, the ice is exposed to lateral shear in the order of $1.3 \times 10^{-4}$ a$^{-1}$ closer to shore. Whilst the vertical and shear strain rates stay constant,
the calculated total strain varies for the younger and older marine ice. Hereby the younger marine ice is currently closer to shore and experiences slightly more total shear strain and combined total vertical and shear strain (20% + 25%). Nonetheless, total accumulated strains in marine ice are still low and the microstructure only shows partial alignment to the ambient strain regime. Overall, SMIS marine ice microstructure shows deviations from its original accretion structure (of elongated ice crystals and a strongly anisotropic ice fabric) with weaker single maxima fabrics and subvertical SPOs at all ice core sites. With more combined total vertical and shear strain the marine ice fabric is becoming even less pointed and the ice crystals lose their SPO. The microstructure does not show evidence for preferential adjustment to a specific strain regime.

- The tilt in the orientation of single maxima fabrics of marine ice in granular facies together with tilted SPOs at certain depths in marine ice cores, suggest a preferential orientation of ice crystal planes for basal glide. Thus marine ice could deform in distinct planes rather than as a uniform mass.

- The strain rates measured in SMIS meteoric ice are comparable to those in marine ice. However, the vertical strain rate increases closer to shore at the meteoric ice core sites (from ~1.4 to ~3.0 x 10^{-4} a^{-1}). Even though SMIS strain rates are lower than at other ice shelves, the accumulated total strain in meteoric ice is ≥ 45%, enough to alter the meteoric ice microstructure; it shows small or great circle girdle fabrics which become stronger closer to shore and ice crystals develop a SPO.

- In the given strain setting at SMIS where marine and meteoric ice are exposed to slightly different strain regimes and their formation location remains uncertain, it remains difficult to say which ice type deforms more easily. Given that meteoric ice clearly develops a small or great circle girdle fabric downflow and marine ice only loses some strength in its single maximum fabric closer to shore, there is no evidence that marine ice deforms more easily than meteoric ice. Further study, for instance in a less complex strain setting, is necessary to compare marine and meteoric ice behaviour in a natural setting.
• Meteoric and marine ice at SMIS are likely not experiencing tertiary strain as the ice microstructure changes from one meteoric/marine ice core site to the other. An increased development of (great) circle girdle ice fabrics as well as small decreases in mean crystal size are good indicators that dynamic recrystallization is taking place in meteoric and marine ice. However, differences in the ice crystal properties at the different ice core sites could also be due to natural variability especially since the ice cores do not lie on the exactly the same flow line and thus have travelled through slightly different strain settings.

Microstructural analysis of marine and meteoric ice from SMIS indicates that the presence of marine ice does not necessarily accelerate ice shelf deformation in strain settings of horizontal compression and lateral shear despite its higher temperature in comparison to meteoric ice. Observed SMIS strain regimes are common in marine ice accretion areas of ice shelf suture zones, ice shelf margins and close to pinning points. However, the accumulated strain at the small SMIS is lower than at larger ice shelves and it remains difficult to generalize findings from this study.
4. Conclusions and further study

In this chapter the main findings of chapters 2 and 3 are summarized in relation to the thesis research objectives. The research presented in this thesis has significantly contributed to knowledge on marine ice composition, formation and behaviour during deformation. Since marine ice makes up a large portion of SMIS, knowledge gained in this study will help understanding the behaviour of SMIS. Even though SMIS is a small and unique ice shelf, some of the knowledge on its behaviour can be transferred to larger ice shelves. Research limitations and future research directions are suggested at the end of this chapter.

4.1. Summary of main findings

The aim of this study was to improve knowledge of marine ice formation processes and its behaviour during deformation through investigation of the chemical composition and microstructure of marine ice samples. Findings with regard to the five research questions introduced in chapter 1 are summarized below:

(1) In order to determine the chemical composition of source water masses that mix to form marine ice, measured marine ice chemical composition was compared to simulation results of a frazil ice freezing model in chapter 2. The presence of banded and granular marine ice facies in SMIS marine ice warrants the use of this model since they are indicative of marine ice formation from frazil ice crystals. Analysis of the isotopic record of SMIS marine ice, which has a co-isotopic slope of around 8, suggest that the ice forms from a spatiotemporally changing water source. Result of the numerical simulation indicate that marine ice was formed from at least three different original water sources: sea water mixed with fresher melted meteoric ice or melted marine ice. Some SMIS marine ice is isotopically more enriched than what can be explained by fractionation from a source water mixture of just sea water and melted meteoric ice. Thus, this marine ice goes through at least one freeze-thaw-refreeze cycle.

(2) The origin of water sources that mix to form marine ice and likely formation processes were evaluated in chapter 2 in light of the geographic setting at SMIS. Two main processes of SMIS
marine ice formation were discussed: (1) percolation of surface meltwater to the ice shelf base where the water mixes turbulently with saltier and colder sea water, and (2) uplift of basal ice shelf meltwater from below RIS into the higher SMIS cavity in an ice pump mechanism accompanied by anchor ice formation. In the former process water from summer ice shelf surface lakes within outcropped SMIS marine ice drains to the ice shelf base via the tide crack. In the SMIS cavity tidal lifting causes turbulence, which facilitates mixing of the more buoyant melted marine ice with sea water. Thus, frazil ice crystals nucleate in a double diffusion mechanism of heat and salt between the two water masses at their salinity-dependent freezing point. Given that SMIS is also adjacent to the much larger and thicker RIS, some marine ice at SMIS could also have formed by the latter ice pump mechanism. Hereby basal ice shelf meltwater rises turbulently from the RIS into the shallower waters of the SMIS cavity, where it reaches its pressure-dependent freezing point. The water becomes supercooled and frazil ice crystals precipitate. Sponges in marine ice on the ice shelf surface suggest that supercooled water also reaches to the ocean floor where it initially forms anchor ice.

(3) Marine ice microstructure evolution was investigated with increasing strain in chapter 3. Vertical and shear strain rates are low at the marine ice core sites (3.0 and 1.3 x 10^{-4} respectively) and total strains are in the order of ~20%. Nonetheless, the ice shows evidence of dynamic recrystallization. The microstructure of marine ice is altered from its typical formation structure at both SMIS core sites; the strong anisotropic ice fabric is tilted from the vertical and ice crystals are not consistently elongated. Furthermore folds are visible. With increasing total combined vertical and shear strain marine ice fabric further decreases its anisotropy and shows weak great circle girdle fabrics and ice crystals lose their horizontal SPO. There is evidence that marine ice adjusts its microstructure more in distinct planes rather than as uniform mass; the ice fabric and SPOs are more tilted at certain depths in the marine ice and folds are present in some thin sections.

(4) Marine and meteoric ice microstructures were compared in chapter 3 to evaluate whether one ice type deforms more easily. Furthermore, it was investigated whether marine and/or
meteoric ice experience tertiary strain. Both meteoric and marine ice experience dynamic recrystallization and progressively adjust their microstructure to increasing total strain closer to shore. Unfortunately neither the strain rate nor the calculated minimum total strain are equal at the marine and meteoric ice core sites. As a result meteoric ice shows increased adjustment downflow; meteoric ice circle girdle fabrics become stronger and crystals develop a SPO. Marine ice microstructure does not adjust as strongly to the vertical extension/longitudinal compression. It keeps a strongly anisotropic fabric at both ice core sites and some of its horizontal SPO typical for marine ice accretion from frazil ice crystals. Thus, results from this study suggest that marine ice does not deform more easily than meteoric ice. Given that marine and meteoric ice change their microstructure between sites as a response to a slight change in the strain regime but both ice types keep some accretion structures, there is evidence that do not experience tertiary strain.

(5) The fifth objective of this thesis was to evaluate whether the presence of marine promotes ice shelf longevity. Results from this study show that marine ice does not only increase ice shelf stability due to adding mass but also due to its different rheology. In chapter 2 a scenario of marine ice melting and refreezing from surface meltwater was proposed where basal mass can accrete at the ice shelf base even when increased air temperatures cause surface melting. Results on marine ice microstructure evolution presented in chapter 3 suggest that the presence of marine ice decelerates shelf ice flow and deformation and thus slows ice shelf ductile deformation.

4.2. Significant contributions to knowledge on marine ice

This thesis provides valuable in situ data on marine ice chemical composition and ice crystallography. Due to the difficulty in accessing marine ice, few studies to date have measured marine ice chemical composition, ice facies (Table 2.1) and ice microstructure (Table 3.2). At SMIS marine ice can be readily accessed from the ice shelf surface. More measurements from different geographical settings help to constrain processes of marine ice formation and deformation. In situ marine ice samples basically provide a geological record
from which processes can be inferred that are too hard to observe in nature, either due to the long time scales associated with it (marine ice deformation) or due to the difficulty in accessing the SMIS cavity and deploying instruments into supercooled water without losing them in the newly formed ice (marine ice formation).

The isotopically enriched composition of marine ice in comparison to sea water at SMIS is consistent with measurements of marine ice from the Hell’s Gate and Nansen Ice Shelves (Tison and others, 1991, Souchez and others, 1991, 1998, 2001, Khazendar and others, 2001, Dierckx and others, 2013, 2014). These ice shelves are small and thin and marine ice also crops out at the ice shelf surface due to erosion from katabatic winds, similar to SMIS. Based on very isotopically enriched data from the Hell’s Gate Ice Shelf, Souchez and others (1991) initially proposed a marine ice freeze-thaw-refreeze cycle. However, at the Hell’s Gate Ice Shelf marine ice melting was thought to occur at the ice shelf base as a result of a shallow tidal circulation, and not at the surface as the presence of surface lakes suggest on SMIS.

This study is the first to link air-ice shelf surface boundary conditions with ocean-ice shelf base boundary conditions, i.e. basal accretion of marine ice. Souchez and others (1998) already suggested that - in absence of a relatively warm sub-ice cavity circulation - surface meltwater could be routed to the ice shelf grounding line through sediment. Here double diffusion of heat and salt between meltwater and sea water occurs, and thus sediment-rich ice accretes. This study, however, suggests direct channeling of meltwater to the ice shelf base via the tide crack and double diffusion of heat and salt between two water masses at their salinity-dependent freezing point as they mix turbulently, considering recent studies of tidally induced turbulence below ice (McPhee and others 2013, 2016). The proposed novel process is an alternative to the ice pump mechanism. Nonetheless, some of SMIS marine ice may still have formed in an ice pump mechanism with water originating from below RIS which becomes supercooled as it enters the SMIS cavity as a result of a change in the pressure-dependent freezing point. Understanding all possible marine ice formation processes will help to map marine ice occurrence in a changing climate.
The marine ice chemical composition presented in this study does not corroborate the marine ice formation theory of closed ice shelf cavity freezing as suggested by Fitzsimons and others (2012). Instead, this study found that the composition of marine ice source water varies spatio-temporally as evident by the co-isotopic mixing slope of ~8 in marine ice.

Once it has been determined where marine ice accretes, it is important to assess how the presence of marine ice influences ice shelf dynamics. SMIS provides a unique opportunity of a ‘natural laboratory’ where the marine ice microstructure evolution could be sampled along a flowline of increasing total strain. Whilst it is possible to measure the anisotropy of meteoric ice remotely with radar (e.g. Eisen and others, 2007, Drews and others, 2012), this is not yet possible in the more conductive marine ice and investigations rely on in situ samples. Due to the difficulty of access, few previous studies have taken marine ice samples at more than one location within an ice shelf (e.g. Tison and others, 1993, Treverrow and others, 2010) to link marine ice microstructure to ice shelf strain (e.g. Treverrow and others, 2010). Since SMIS is accessible and small, in situ surface strain measurements were practicable and the marine ice microstructure could be related to it. Given that SMIS is a small ice shelf, strain rates are also lower than at other ice shelves. Nonetheless the total calculated accumulated strain (of ~ 20%) would be high enough to alter its microstructure. Whilst the accumulated strain was calculated based on the flow path and estimated thickness of the ice shelf in chapter 3, it remains difficult to estimate the marine ice age accurately. The distribution and thickness of marine ice at SMIS needs to be confirmed by, for instance, a marine ice thickness map derived from a detailed radar survey and ice shelf freeboard measurements (e.g. Fricker and others, 2001).

SMIS marine ice shows single maxima and weak (great) circle girdle fabrics consistent with observations at the large Filchner-Ronne and Amery Ice Shelves (Eicken and others, 1993, Rist and others, 2002, Treverrow and others, 2010) and the smaller Hell’s Gate and Nansen Ice Shelves (Tison and others, 1993, Dierckx and others, 2013, 2014) (Table 3.2). However, at SMIS there is no evidence that SMIS marine ice deforms more easily than meteoric ice, which is consistent with laboratory experiments by Dierckx and others (2013, 2014). The anisotropic fabric of marine ice is likely not preferentially oriented for lateral
compression/vertical extension with the ice basal planes largely parallel to the ice-ocean interface. Equally SMIS marine ice microstructure is not preferentially oriented for lateral shearing. Not only the crystal orientation but also the relatively small crystal size of marine ice - likely a result of restriction by a second phase (e.g. Cyprych and others, 2016) - could be responsible for slowing down ice deformation. The observed marine ice microstructure at SMIS could thus also be an accretion structure. However, this marine ice microstructure may support sub-horizontal internal shearing in distinct planes with more preferential deformation at certain depths. An improved understanding of changes in marine ice microstructure as a result of ice shelf flow informs about SMIS behavior and more generally ice shelf dynamics if some marine ice is present.

4.3. Marine ice and SMIS behaviour

SMIS has been called ‘the strangest ice shelf in the world’ (Debenham, 1965), partially due to the presence of macrofossils on the ice shelf surface and partially due to the fact that it is possibly fed by the much larger RIS rather than by ice streams and glaciers. However, SMIS may also sustain itself by surface accumulation in its north and surface ablation in its south (Clifford, 2005). The only indicator for inflow from RIS into SMIS is an accelerated flow rate at SMIS’s eastern flank close to the shear zone with RIS (Figure 3.24).

SMIS is a heterogeneous ice shelf, made up of meteoric ice and a large portion of marine ice. Marine ice makes up the entire ice shelf thickness in most of SMIS’s southern ablation area (Clifford, 2005). Such massive marine ice is rare and has only been observed at other thin ice shelves (e.g. Hell’s Gate Ice Shelf; Tison and others, 1993). Marine ice is thus significant for SMIS ice shelf dynamics and stability.

All evidence presented in this study points toward local marine ice accretion either from a mixture of surface meltwater or basal RIS water with seawater. The mere presence of marine ice at SMIS thus indicates that a cold ice shelf cavity exists below the ice shelf that facilitates this basal mass accretion, which contributes to a positive basal ice shelf mass balance. Since relatively warm water influx from the open ocean into the SMIS ice shelf cavity is likely blocked by Black and White Islands, no basal melt is likely occurring. Given that no basal accumulation
rates were measured in this study though, it remains unknown whether these rates. Thus, SMIS may continue to exist in its present configuration well into the future even though it is relatively thin.

Results from this study have shown that marine ice does not accelerate ice shelf deformation. Instead, marine ice slows down lateral shearing and ice shelf thickening in SMIS’ south. These slow ice shelf dynamics encourage predominantly ductile deformation observed in SMIS marine ice. Given that SMIS marine ice microstructure shows no evidence that the ice experienced tertiary strain, SMIS is not in dynamic steady state. Research presented in this thesis provided no evidence that SMIS would suffer brittle disintegration in the near future.

4.4. SMIS marine ice and behaviour of Antarctic ice shelves

Whilst ice shelves play an essential role regulating the transfer of land ice into the ocean, they are still poorly represented in prognoses of future Antarctic ice sheet evolution because current ice shelf behavior remains poorly understood. Ice shelf behaviour is largely determined by mass influx from ice streams and glaciers, local snowfall, ocean circulation and air temperature. SMIS, however, is not fed by glaciers and ice streams and hence does not buffer these either. Its disintegration would therefore not promote accelerated sea level rise. SMIS is not a typical ice shelf; it is thin and not very dynamically active. Additionally massive marine ice makes up a significant portion of SMIS. In larger ice shelves (e.g. Amery, Larsen C and Filchner-Ronne), that buffer ice streams behind them, marine ice often accretes in suture zones (Kulessa and others 2014, Treverrow and others, 2010, Lambrecht and others, 2007), where lateral shearing is prevalent. Some marine ice can accrete close to pinning points or along ice shelf margins (e.g. Fricker and others, 2001), which experience lateral shearing and compression. Results from this study show that marine ice does not deform more easily than meteoric ice in strain settings of lateral shearing and compression, likely because of its strong isotropic fabric. Thus, the presence of marine ice in the suture zones of these larger ice shelves or along the ice shelf margins will likely not accelerate ice shelf flow.

The improved knowledge of meteoric and marine ice shelf microstructure in relationship to shelf ice deformation generated in this study, will help improving simulations of
ice shelf dynamics. Since anisotropy influences ice flow significantly (e.g. Zwinger and others, 2014), Ma and others (2010) developed enhancement factors for ice shelf flow simulations based on typical meteoric shelf ice fabrics. However, to date numerical simulations of ice shelves do not consider heterogeneity of ice shelves well. Ice shelves are generally treated as a uniform mass (e.g. Golledge and others, 2015, Feldmann and Levermann, 2015). When marine ice is included, only a temperature-dependent rheology is considered (e.g. Khazendar and others, 2009), but so far not its anisotropy. In order to model the role marine ice has on ice shelf behaviour well, the processes in the ice shelf cavity need to be better represented also, not least because in comparison to rising air temperatures, ocean warming causes a larger and faster ice sheet response (Golledge and others, 2015).

The chemical composition of SMIS marine ice revealed that some marine ice has undergone at least one freeze-thaw-refreeze cycle. This study discussed the possibility of meltwater and sea water mixing turbulently at their salinity-dependent freezing point, whereby tidal lifting may cause the turbulence. Not only surface meltwater but also basal meltwater can enter ice shelf cavities (e.g. Marsh and others, 2016). If turbulent mixing with the ambient sea water is facilitated, this meltwater may also contribute to frazil ice precipitation in a double-diffusion mechanism of heat and salt. Thus, if the cavity on an ice shelf stays cold-based, addition of surface or basal meltwater may encourage marine ice formation.

4.5. Limitations and future research

The novel explanation of this study for marine ice recycling by channelling surface meltwater to the ice shelf base via the tide crack, where it mixes turbulently, has two limitations: (1) turbulence required to mix different waters in the absence of a thermohaline circulation is poorly understood below ice shelves and (2) SMIS ice chemistry provides no evidence that marine ice recycles several times.

As discussed in chapter 2, tidal lifting could cause turbulence in the SMIS cavity, facilitating a mixture of different water masses and double diffusion processes of heat and salt between these water masses. However, only a few sub-ice shelf oceanic measurements exist (e.g. Herraiz-Borreguero and others, 2016) because of the difficulty associated with accessing
the ice shelf cavity. Temperature and salinity measurements by autonomous underwater vehicles (AUV) (e.g. Dowdeswell and others, 2008) would help to constrain these basal ice shelf processes and allow their spatio-temporal variation to be assessed. However, the SMIS ice shelf cavity does not open up to the open ocean and access is thus difficult.

SMIS marine ice chemistry provides no evidence that marine ice recycles more than once. The ice stays always less enriched than 2.5‰. Thus, either the marine ice at SMIS is young and has only undergone one cycle of recycling or the source water mixture is made up of more than two different water sources including melted meteoric ice. Whilst this melted meteoric ice could come from melting glaciers on Minna Bluff, the ice pump mechanism routing ice shelf water from below the Ross Ice Shelf cavity may play a more significant role at SMIS. Mapping the complete marine ice extent at SMIS with ice radar would help to remove uncertainties regarding marine ice formation processes/ location at SMIS. This would also allow for better marine ice age and thus also total accumulated strain calculations. Arcone and others (2016), for instance, found massive layers of marine ice below meteoric ice in the shear zone of SMIS, for which formation processes remain unknown.

Comparing the evolution of marine and meteoric ice microstructure along a flowline at SMIS in chapter 3 showed no evidence for easier deformation of marine ice. This is in line with the detection of harder marine ice material properties in an experimental study of uniaxial compression (Dierckxs and others, 2014). However the total strain in SMIS marine ice is low due to the low and complex strain rates of the ice shelf (in the order of $x10^{-4}$). Thus, the marine ice microstructure only partially adjusts to the new strain regime. Hence, additional investigations are necessary where marine ice would be exposed to a larger and less complex strain regime. Since marine ice often accretes close to ice margins and in suture zones, it is also exposed to a complex strain regime at other ice shelves (e.g. Treverrow and others, 2010, Berger and others, 2016). However, for simplified ice shelf modelling scenarios it is important to know how marine ice responds to specific strain regimes. Experimental studies provide a means to address one strain regime at a time (e.g. uniaxial compression in Dierckx and others, 2014). Nonetheless, experimental studies have scale issues. For instance much higher
stresses than in nature are needed to facilitate deformation. Alternatively, though, it may not be practical to extract marine ice along a flowline of a faster flowing and much thicker ice shelf. This study could not confirm that the higher temperature of marine ice makes it mechanically soft (e.g. Kulessa and others, 2014). Whilst a temperature-dependent rheology for modelling marine ice (e.g. Khazendar and others, 2009) is likely important, it remains to be evaluated whether marine ice temperature is more important than the ice microstructure. Comparing marine and meteoric ice in deformation experiments at different temperatures would help to answer that question.

Marine ice deformation in this study was limited to ductile deformation as a result of ice shelf flow. However, shelf ice also experiences brittle deformation as crevasses and fractures form especially during higher tensile strain rates. Experimental tests reveal that fractures do not propagate as easily in marine ice as they do in meteoric shelf ice (Rist and others, 2002). However, in the heavily crevassed shear zone of SMIS crevasses also propagate into the marine ice (Arcone and others, 2016). It thus remains unclear whether the presence of marine ice prevents brittle ice shelf disintegration.

Even though the marine ice formation process proposed in this thesis would allow for a longer ice shelf life in a climate of increasing air temperatures, marine ice occurrence may still largely be determined by ocean circulation. Ice shelves do not usually melt and thin from the surface down but from the bottom up as a result of increasing ocean temperatures and marine ice accretion has not been observed in warm water ice shelf cavities. Thus, whether an ice shelf disintegrates or accretes marine ice as a result of surface water percolation may largely be determined by its basal ocean temperature. Most ice shelves that have been observed to disintegrate rapidly (Scambos and others, 2003, 2009) were on the Antarctic Peninsular and thus warm-based. The presence of marine ice could also thus largely be an indicator of a ‘healthy cold-based ice shelf’.
References


Clifford, A.E. 2005. Physiography, flow characteristics and vulnerability of the Southern McMurdo Ice shelf, Antarctica. (MSc University of Otago.)


Daly, S.F. 1984. Frazil Ice Dynamics. Cold Region Research and Engineering Lab


Gao, X.J. 1989. Laboratory studies of the development of anisotropic crystal structure and the flow properties of ice. (PhD University of Melbourne.)


Hughes, K.G. 2013. Propagation of an Ice Shelf Water Plume beneath Sea Ice in McMurdo Sound, Antarctica. (MSc University of Otago.)


Paolo, F.S. 2015. Interannual and decadal variations of Antarctic ice shelves using multi-mission satellite radar altimetry, and links with oceanic and atmospheric forcings. (PhD University of California, San Diego.)


Robinson, N.J. 2012. Circulation, mixing and interactions in the ocean near the Ross Ice Shelf, Antarctica. (PhD University of Otago.)


Treverrow, A. 2009. The Flow of Polycrystalline Anisotropic Ice: Laboratory and Model Studies. (PhD University of Tasmania.)


Vaughan, M.J.H. 2016. The creep behaviour, and elastic and anelastic properties of polycrystalline ice. (PhD University of Otago.)


Wang, W.L. 1994. Laboratory studies of flow properties and associated crystal structure in Holocene and Wisconsin ice. (MSc University of Tasmania.)


Appendix A

This appendix contains the classified ice crystal facies presented in chapter 2.
Figure A-1: Thin sections of ice crystals in C5. The colours on the left denote the classified ice crystal facies (red = banded, green = mixed).
Figure A-2: Thin sections of ice crystals in C9. The colours on the left denote the classified ice crystal facies (green = mixed, blue = granular).
Figure A-3: Thin sections of ice crystals in C15 for the top six sections. The colours on the left denote the classified ice crystal facies (red = banded, green = mixed, blue = granular, violet = platelet).
Figure A-4: Thin sections of ice crystals in C15 for the middle six sections. The colours on the left denote the classified ice crystal facies (red = banded, green = mixed, blue = granular).
Figure A-5: Thin sections of ice crystals in C15 for the lowermost seven sections. The colours on the left denote the classified ice crystal facies (red = banded, green = mixed, blue = granular, violet = platelet).
Appendix B
MATLAB code used for strain calculations:

\[
x = [7065613:100:7085113]; \quad \% \text{easting}
\]
\[
y = [4939227:100:4967727]; \quad \% \text{northing}
\]
\[
y = \text{flipud}(y);
y = y';
[x\_mesh, y\_mesh] = \text{meshgrid}(x,y);
easting = x\_mesh;
northing = y\_mesh;
\]

% load ice\_grid\_x and ice\_grid\_y
load('D:\H-drive_Feb_2015\My_Ph.D\Strain\Tangent, normal and vertical strain\FINAL_ice_grid_x\&\_y.mat')
vel\_E = ice\_grid\_x;
vel\_N = ice\_grid\_y;

% C9
% start\_x = 7077002;
% start\_y = 4943912;

% C5
% start\_x = 7077439;
% start\_y = 4943963;

% C6
start\_x = 7078185;
start\_y = 4945005;

% C7
% start\_x = 7078507;
% start\_y = 4946934;

% somewhere else
% start\_x = 7068129;
% start\_y = 4950973;

% get the streamline coordinates
stream\_coords = \text{stream2}(x\_mesh, y\_mesh, vel\_E, vel\_N, start\_x, start\_y);
% easting, northing: polar stereographic grid spacing
% vel_E, vel_N: gridded velocities (polar stereographic or other spacing)-ve velocities used to get the
flowline heading 'upstream'.
% start_x, start_y: coordinates from where you want the streamline to start.

stream_xy = stream_coords{1};  % create a double array of flowline coordinates. Use the function
'streamline' if you want to plot the streamline as a check.

% calculate the flow speed
speed = sqrt(vel_E.^2 + vel_N.^2);

% calculate flow speed for points along the flowline by interpolation.
speed_stream = interp2(x_mesh,y_mesh,speed,stream_xy(:,1),stream_xy(:,2));

% calculate the difference in distance between points on the streamline
del_stream_x = diff(stream_xy(:,1));
del_stream_y = diff(stream_xy(:,2));

del_stream = sqrt(del_stream_x.^2 + del_stream_y.^2);

% get the distance upstream from the point of interest (e.g. core site).
% (m)
dist_stream = [0;cumsum(del_stream)];

% calculate the speed at midpoints of the grid and corresponding time
% intervals (from point to point). (m per year)
speed_mid = (speed_stream(1:end-1) + speed_stream(2:end))./2;

% the time between points on the flowline (years)
del_time = [(del_stream./speed_mid);0];

time_along = cumsum(del_time);

% Strain rate components are calculated in 1) the polar stereographic projection coordinate
% system and 2) with respect to the local flow direction:
%
% 1.
% exx - strain rate (x direction - polar stereographic northing)
% eyy - strain rate (y direction - polar stereographic easting)
% exy - shear strain rate - polar stereographic coordinates
%
% 2.
% ett - strain rate component (parallel to local flow direction)
% enn - strain rate component (normal to local flow direction)
% etn - shear strain rate (across the plane containing the flow direction vector).

% Adam Treverrow 16/05/2013
%\% use the gradient function to get the strain rate components

% you may need to transpose the velocity components - I did for the subset
% of Measures data about the Amery that I pinched from Ben. You may or may not need to do this.
U = ice_grid_x;
V = ice_grid_y;
delta=100; % grid spacing (m)
[dU_dx,dU_dy] = gradient(U,delta); % calculate gradient
[dV_dx,dV_dy] = gradient(V,delta);

% define the strain rate components in polar stereographic coordinates
exx = dU_dx; % longitudinal
eyy = dV_dy; % transverse
exy = (dU_dy + dV_dx)./2; % shear

% determine the strain rate components in local coordinates, i.e with respect to the local flow direction
% - using the standard
% methods for rotation of coordinates axes e.g. Appendix A - Cuffey and Paterson (2010).
% theta is the orientation (rotation) of local flow direction relative to polar stereographic
% coords.
costheta = U./(sqrt(U.^2 + V.^2)+eps);
sintheta = V./(sqrt(U.^2 + V.^2)+eps);

% shear strain rate
etn = (-exx + eyy).*costheta.*sintheta + exy.*(costheta.^2 - sintheta.^2);
etn2 = (-eyy + exx).*sintheta.*costheta + exy.*(sintheta.^2 - costheta.^2);

% strain rate tangent to local flow direction
ett = exx.*costheta.*2 + eyy.*sintheta.^2 + 2.*exy.*sintheta.*costheta;

% strain rate normal to local flow direction.
enn = exx.*sintheta.^2 + eyy.*costheta.^2 - 2.*exy.*sintheta.*costheta;

% vertical strain
ezz = (-ett-enn);
% Plot the shear strain rates, with shear transverse to the local flow direction.

% you can get the strain rate components along the flowline by interpolation.
E_xx_stream = interp2(easting,northing,exx,stream_xy(:,1),stream_xy(:,2));
E_yy_stream = interp2(easting,northing,eyy,stream_xy(:,1),stream_xy(:,2));
E_xy_stream = interp2(easting,northing,exy,stream_xy(:,1),stream_xy(:,2));
E_zz_stream = interp2(easting,northing,ezz,stream_xy(:,1),stream_xy(:,2));

% total accumulated vertical strain:
acc_vert_strain = cumtrapz(time_along,abs(E_zz_stream));
figure
subplot(1,2,1);
plot(time_along,acc_vert_strain);
xlim([0 1200])
ylim([0 1.0])
%figure,plot(time_along,acc_vert_strain);
title('vertical strain');
xlabel('Time (years)');
ylabel('accumulated strain');

% total accumulated shear strain:
acc_shr_strain = cumtrapz(time_along,abs(E_xy_stream));
subplot(1,2,2);
plot(time_along,acc_shr_strain);
%figure,plot(time_along,acc_shr_strain);
xlim([0 1200])
ylim([0 1.0])
title('shear strain');
xlabel('Time (years)');
ylabel('accumulated strain');

figure, plot(time_along,E_zz_stream);
%figure,plot(time_along,acc_shr_strain);
title('vertical strain rate');
xlabel('Time (years)');
ylabel('strain rate (a-1)');
figure, plot(time_along,E_xy_stream);
%figure,plot(time_along,acc_shr_strain);
title('shear strain rate');
xlabel('Time (years)');
ylabel('strain rate (a-1)');
Appendix C

This appendix contains one peer reviewed publication. It also refers to one publication in preparation and international conference presentations that were based on work presented in this thesis.

Published article:


This article was presented in Chapter 2 with slight modifications to remove redundancy, consider recent literature published in 2016 and ensure consistency with the thesis. It is reprinted here in full with kind permission of the International Glaciological Society.

Article in preparation:


This article is based on data presented in Chapter 3.

International conference presentations:

Koch, I., S. Fitzsimons, J.L. Tison, A. Treverrow, N. Cullen, 2016: Deformation of meteoric and marine ice at the southern McMurdo Ice Shelf, Antarctica. *Oral presentation* at the International Symposium on Interactions of Ice Sheets and Glaciers with the Ocean, 10–15 July 2016, La Jolla, California, USA.


Fitzsimons, S., M. Sharp, M. Hambrey, S. MacDonell, and I. Koch, 2011. Ice accretion and structural development of the Southern McMurdo Ice Shelf, Antarctica. *Oral presentation* at the International Symposium on Interactions of Ice Sheets and Glaciers with the Ocean, 5–10 June, La Jolla, California, USA.
Marine ice recycling at the southern McMurdo Ice Shelf, Antarctica

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ABSTRACT. Marine ice accretes at the base of ice shelves, often infilling open structural weaknesses and increasing ice-shelf stability. However, the timing and location of marine ice formation remain poorly understood. This study determines marine ice source water composition and origin by examining marine ice crystal morphology, water isotope and solute chemistry in ice samples collected from the southern McMurdo Ice Shelf (SMIS), Antarctica. The measured co-isotopic input of a freezing model for frazil crystals indicate a spatio-temporally varying source water of sea water and relatively fresher water, such as melted meteoric or marine ice. This is in agreement with the occurrence of primarily banded and granular ice crystal facies typical for frazil crystals that nucleate in a supercooled mixture of water masses. We propose that marine ice exposed at the surface of SMIS, which experiences summer melt, is routed to the ice-shelf base via the tide crack. Here frazil crystals nucleate in a double diffusion mechanism of heat and salt between two water masses at their salinity-dependent freezing point. Recycling of previously formed marine ice facilitates ice-shelf self-sustenance in a warming climate.

KEYWORDS: Antarctic glaciology, ice chemistry, ice crystal studies, ice shelves, ice/ocean interactions

INTRODUCTION

Marine ice forms from a mixture of sea and fresh water and adds mass to the underside of ice shelves. It can accrete in open basal structures, such as rifts and crevasses (Morgan, 1972; Souchez and others, 1991; Cerit and others, 1992; Tison and others, 1993; Eicken, 1994; Khazendar and others, 2001; Craven and others, 2004, 2009; Pattyn and others, 2012; Jansen and others, 2013). Presence influences the rheology and stability of ice shelves (Khazendar and others, 2009; Kussela and others, 2014) or as a massive layer below meteoric ice-shelf ice (Zotkov and others, 1980; Craven and others, 2009; Treverrow and others, 2010). Its presence influences the rheology and stability of ice shelves (Khazendar and others, 2009; Kussela and others, 2014), which drains 74% of Antarctica’s grounded ice together with outlet glaciers (Bindschadler and others, 2011). Ice shelves also buttress most ice streams (Dupont and Alley, 2005). Marine ice has been increasingly observed in Antarctic ice shelves, either directly in ice-core samples (Zotkov and others, 1980; Souchez and others, 1991; Tison and others, 1993; Eicken and others, 1994; Khazendar and others, 2001; Craven and others, 2009; Pattyn and others, 2012; Dierckx and others, 2014) or remotely using geophysical methods (Fricker and others, 2001; Joughin and Vaughan, 2004; McMahon and Luckie, 2006; Khazendar and others, 2009; Pattyn and others, 2012; Jansen and others, 2013).

Due to the difficulty associated with accessing the ice-shelf cavity, processes of marine ice formation, including timing, location and volume, remain largely unknown. Since marine ice can be several hundred metres thick (e.g. Amery Ice Shelf, East Antarctica; Craven and others, 2009), fast-forming frazil ice crystals are assumed to be mainly responsible for the generation of marine ice (Treverrow and others, 2010), allowing for marine ice accumulation rates of >1 m a−1 at some ice shelves (Bombosch and Jenkins, 1995; Wen and others, 2010). Indeed, loose agglomerates of frazil ice crystals have been observed below ice shelves in borehole imagery (Craven and others, 2005; Hubbard and others, 2012). These small freely floating discoid ice crystals nucleate in supercooled water (Martin, 1981; Daly, 1984), which is water cooled below its in situ freezing point without changing state (Leonard and others, 2014). Supercooling of water masses below ice shelves can occur as a result of a change in the pressure-dependent freezing point due to adiabatically rising water masses (Fjeldvik and Kvinge, 1974) or double diffusion of heat and salt between water masses of different salinities at their freezing point (Souchez and others, 1998). Both mechanisms generally involve mixing of sea water with fresher water. The former process is part of a thermohaline circulation often referred to as the ‘ice pump’ mechanism (Lewis and Perkin, 1986) and has been widely associated with marine ice formation in thick layers (e.g. Galton-Fenzi and others, 2012). Hereby, continental freshwater ice is melted at depth close to the grounding line of an ice shelf by denser High Salinity Shelf Water (HSSW), which is generated during sea-ice formation in winter. Similarly, warmer Circumpolar Deep Water (CDW) can enter the ice-shelf cavity inducing melt at the ice-shelf base (Jacobs and others, 1992). This meltwater then rises along the gradient of the ice shelf due to its buoyancy to shallower waters, where it becomes supercooled due to a change in the pressure-dependent freezing point and frazil ice crystals nucleate (Galton-Fenzi and others, 2012). Double diffusion-induced supercooling could occur close to the grounding line of shallower ice shelves, where surface meltwater can percolate through sediment (Souchez and others, 1998) or drain through tide cracks (Gow and others, 1965; Gow and Epstein, 1972) to the ice/ water interface. Heat diffuses faster than salt from relatively
Table 1. Occurrence of ice crystal facies (gr: granular; bd: banded; clm: columnar; pl: platelet), measured range in δ18O and salinity, and co-isotopic slopes of marine ice in previous studies

<table>
<thead>
<tr>
<th>Ice shelf</th>
<th>Ice crystals</th>
<th>δ18O °/oo</th>
<th>Salinity °/oo</th>
<th>Co-isotopic slope</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amery Ice Shelf</td>
<td>–</td>
<td>0.00 to 2.30</td>
<td>&lt;0.02</td>
<td>–</td>
<td>Morgan (1972), cited in Goodwin (1993)</td>
</tr>
<tr>
<td>Amery Ice Shelf</td>
<td>gr, bd</td>
<td>-0.60 to 1.50</td>
<td>0.06 to 0.75</td>
<td>–</td>
<td>Craven and others (2004)</td>
</tr>
<tr>
<td>Amery Ice Shelf</td>
<td>gr, bd</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>Tremain and others (2010)</td>
</tr>
<tr>
<td>Campbell Glacier Tongue</td>
<td>gr</td>
<td>-17.00 to -1.00</td>
<td>–</td>
<td>7.86 (r = 0.980)</td>
<td>Scott and others (1995)</td>
</tr>
<tr>
<td>Fächer-Ronne Ice Shelf</td>
<td>gr</td>
<td>-2.00</td>
<td>&lt;0.10</td>
<td>–</td>
<td>Oeter and others (1992)</td>
</tr>
<tr>
<td>Fächer-Ronne Ice Shelf</td>
<td>gr</td>
<td>0.02 to 0.10</td>
<td>0.01 to 0.80</td>
<td>6.60 (r = 0.93, marine ice + sea-water samples)</td>
<td>Eicken (1994)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>clm, pl, gr</td>
<td>1.00 to 3.20</td>
<td>–</td>
<td>–</td>
<td>Scott and others (1991)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>clm, pl, gr, bd</td>
<td>1.14 to 3.26</td>
<td>0.03 to 1.42</td>
<td>–</td>
<td>Tison and others (1993)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>clm</td>
<td>1.14 to 2.02</td>
<td>1.35</td>
<td>–</td>
<td>Tison and others (1993)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>pl</td>
<td>2.00 to 2.51</td>
<td>1.42</td>
<td>–</td>
<td>Tison and others (1993)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>pl</td>
<td>1.62 to 3.26</td>
<td>0.03 to 1.54</td>
<td>–</td>
<td>Tison and others (1993)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>pl</td>
<td>1.64 to 3.02</td>
<td>0.19 to 0.36</td>
<td>–</td>
<td>Tison and others (1993)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>pl</td>
<td>~1.50 to ~3.00</td>
<td>7.71 (r = 0.099)</td>
<td>–</td>
<td>Scott and others (1996)</td>
</tr>
<tr>
<td>Koeltzke Ice Tongue</td>
<td>pl</td>
<td>1.37 to 2.51</td>
<td>0.20 to 5.26</td>
<td>8.10 (Na content only)</td>
<td>Goodwin (1993)</td>
</tr>
<tr>
<td>Law Dome</td>
<td>clm</td>
<td>0.26 to 1.56</td>
<td>0.01 to 0.08</td>
<td>–</td>
<td>Goodwin (1993)</td>
</tr>
<tr>
<td>type 1 marine ice</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>Goodwin (1993)</td>
</tr>
<tr>
<td>Law Dome</td>
<td>gr</td>
<td>-16.93 to -2.57</td>
<td>0.01 to 0.08</td>
<td>7.80 (Na content only)</td>
<td>Goodwin (1993)</td>
</tr>
<tr>
<td>Northern McMurdo Ice Shelf</td>
<td>gr</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>Gow and others (1965)</td>
</tr>
<tr>
<td>Southern McMurdo Ice Shelf</td>
<td>gr</td>
<td>-2.80 to 2.90</td>
<td>–</td>
<td>–</td>
<td>Kellogg and others (1991a)</td>
</tr>
<tr>
<td>Southern McMurdo Ice Shelf</td>
<td>–</td>
<td>~4.40 to ~2.50</td>
<td>7.82 (r = 0.99, mix of meteoric and marine ice)</td>
<td>Fitzsimons and others (2012)</td>
<td></td>
</tr>
<tr>
<td>Nansen Ice Shelf</td>
<td>gr, bd</td>
<td>1.80 to 2.37</td>
<td>0.04 to 0.15</td>
<td>–</td>
<td>Khazendar and others (2001)</td>
</tr>
<tr>
<td>Hell’s Gate Ice Shelf</td>
<td>gr, bd</td>
<td>–</td>
<td>0.03 to 0.23</td>
<td>–</td>
<td>Dierckx and others (2013, 2014)</td>
</tr>
<tr>
<td>Ross Ice Shelf</td>
<td>clm</td>
<td>–</td>
<td>–</td>
<td>0.00 to 0.00 (Cl measurements)</td>
<td>Scott and others (1993)</td>
</tr>
<tr>
<td>Roi Baudoin Ice Shelf</td>
<td>gr, bd</td>
<td>~0.00 to ~2.20</td>
<td>~0.03 to ~9.20</td>
<td>–</td>
<td>Zottl and others (1980)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Pattyn and others (2012)</td>
</tr>
</tbody>
</table>

warmer meltwater to colder sea water (Martin and Kaufman, 1974). This results in rapid freezing of the less saline water mass. If both water masses are near their salinity-dependent freezing point and become turbulently mixed, frazil ice crystals might nucleate in a double diffusion process (McPhee and others, 2013). This could be driven by a shallow tidal circulation. Supercooled water extends to the ocean floor, frazil ice crystals can also flock together on the sea floor, forming anchor ice (Mager and others, 2013). This ice could eventually lift off the ocean floor when its buoyancy is large enough and contribute to marine ice at the ice-shelf base (e.g. Swithinbank, 1970).

Over time, frazil ice crystals gradually sinter together and adjust their shape to minimize the surface energy of the ice crystals (Martin, 1981). However, it remains uncertain whether marine ice is formed from frazil ice crystals only, that grow, settle and compact much like snow due to the buoyancy pressure, or whether 'pore spaces' between individual frazil ice crystals are filled by sea water, which freezes as a result of heat conduction through the ice shelf (Eicken, 1994; Tison and others, 2001). In either case, the bulk of marine ice (70-95%) would be made up of frazil ice crystals due to their high buoyancy pressure (Tison and others, 2001). Once consolidated, frazil ice crystals in marine ice form granular, banded or platelet ice facies. Granular ice facies have been observed most frequently in marine ice formed from primarily frazil ice crystals (Oeter and others, 1992; Tison and others, 1993; Eicken and others, 1994; Treverrow and others, 2010), followed by banded ice facies (Khazendar and others, 2001; Craven and others, 2004; Tison and others, 2010; Treverrow and others, 2010; Pattyn and others, 2012) (Table 1). Large platelet ice facies are less common in marine ice (Gow and Epstein, 1972; Tison and others, 1993) (Table 1).

A better understanding of marine ice formation processes outlined above contributes to the assessment of the stability of ice shelves, many of which are experiencing rapid change as a result of increased ice-shelf basal melt rates due to oceanic erosion (e.g. Depoorter and others, 2013) and surface melting (e.g. Banwell and others, 2014). The aim of this study is to determine the composition and origin of water masses that freeze to form marine ice at the southern McMurdo Ice Shelf (SMIS), Antarctica (Fig. 1) and whether there are spatio-temporal variations. To accomplish this, marine ice samples were collected from different SMIS sites, and the ice crystal morphology and chemistry was analysed in conjunction with a boundary layer freezing model. The data are interpreted in the light of the geographic setting to determine whether marine ice at SMIS was formed as a result of a deep thermohaline circulation, or in a double diffusion mechanism of heat and salt during turbulent mixing of sea water with fresher surface meltwater.

**SAMPLING AND DATA ANALYSIS**

In this section we describe the field site and give a summary of previous studies. We also detail marine ice extraction from the ice shelf and its laboratory analysis for water
isotope and major ion composition. The frazil ice boundary-layer freezing model, which is used to compare theoretical and measured water isotope signals of marine ice, is explained below. Thin sectioning of marine ice and the classification of ice crystal morphology into different ice facies are also described.

**Field site**

SMIS is a small ice shelf (~30 x 35 km²) in Antarctica confined by Minna Bluff to the south and Black and White Islands to the north (Fig. 1). It is separated from a much larger and two magnitudes faster-flowing Ross Ice Shelf by a rift zone (Fig. 1). SMIS flows slowly at a rate of 0.4–7.3 m a⁻¹ in a west-southwesterly direction toward Minna Bluff in its southern parts and in a west-northwesterly direction in its northwestern parts (Clifford, 2005). The ice shelf floats on a water column of 300–400 m in its centre (Johnston and others, 2008). Ground-penetrating and airborne radar investigations have revealed that the ice shelf is thickest in the north (~180 m) and thins (to ~100 m) toward its eastern and southern margins (Swithinbank, 1970; Clifford, 2005), where the radar signal becomes lost 5–6 km from shore. This was ascribed to outcropping of slightly saline marine ice at the snow-free ice-shelf surface close to the shore of Minna Bluff (Fig. 1) (Clifford, 2005). The presence of marine ice was also detected in previous studies on ice composition (Kellogg and others, 1991a; Fitzsimons and others, 2012) and inferred from the presence of marine macrofossils at the ice-shelf surface (Debenham, 1919; Gow and others, 1965; Swithinbank, 1970; Kellogg and others, 1990). Two of these shells were radiocarbon-dated to 1230 ± 50 and 2650 ± 30 radiocarbon years BP (Kellogg and others, 1990; Denton and Marchant, 2000) using an Antarctic reservoir correction of 1300 years (Berkman and Forman, 1996). However, their age does not necessarily reflect the date of marine ice accretion, since the time of death of the marine organisms is necessarily related to their entrainment date (Fitzsimons, 1997). Surfacing of marine ice, which was originally accreted at SMIS’s base, is speculated to be a result of stripping by katabatic winds (Clifford, 2005), as at the Hell’s Gate Ice Shelf (Tison and others, 1993). Modelled wind fields indeed show elevated wind speeds over the southern part of SMIS (Monaghan and others, 2005). Local summer surface melting also contributes to the observed negative surface mass balance (>0.10 m; Clifford, 2005) in the snow-free band running parallel to Minna Bluff (Fig. 1). Melting is likely partially induced by the locally high debris concentration on the ice-shelf surface close to Minna Bluff (Denton and Marchant, 2000) and the lower surface albedo of the darker marine ice (Warren and others, 1997). Surface meltwater pools in several ice-shelf surface lakes (Clifford, 2005), which are often elongated and oriented at right angles to shore (Fig. 2) parallel to the prevailing wind direction (Swithinbank, 1970). These lakes were observed to form in summer, while the rest of the year they were frozen over into ‘mirror-smooth’ surfaces of ice (Swithinbank, 1970). The lakes are estimated to be ~1 m deep, similar to lakes observed on other ice shelves (e.g., Banwell and others, 2014), several tens to hundreds of metres wide and up to 1–2 km long. Some lakes were observed to drain completely during the course of the melt season.

**Ice sampling in the field**

Marine ice was extracted in shallow cores from the apex of snow-free ~6 m high ice ridges that separate surficial lakes and are oriented at right angles to shore, almost in a perfect north–south direction (Fig. 2). Cores were taken in November 2010 with a Kovacs corer, and in December 2007 with a custom-made SIPRE-type coring auger. The extracted three ice cores (C5, C9, and C15; Fig. 1) were 2.71 ± 0.01 m, 3.04 ± 0.01 m and 9.49 ± 0.01 m long. The freeboard level was not reached during ice-core extraction since the ice shelf is ≥100 m thick (Clifford, 2005). The top 0.50 ± 0.01 m of every core was discarded to avoid the influence of potential surface melt (Tison and others, 1993). Ice cores were immediately put in a freezer and kept below −15°C during transport and storage.
Solute and water isotopes

Ice cores were cut into 0.10 ± 0.001 m sections along the length of the cores and a 0.0075 ± 0.0025 m thick outside rim of the ice cores was cut off to avoid contamination. Subsequently the samples were allowed to melt in closed plastic containers at room temperature. Water samples were filtered under vacuum using MF-Millipore 0.45 µm membrane cellulose acetate and cellulose nitrate filters. The samples were analysed using a Dionex ICS-3000 ion chromatograph to determine the concentration of major cations and anions (Li⁺, Na⁺, Mg²⁺, K⁺, Ca²⁺, NH₄⁺ and Cl⁻, NO₃⁻, Br⁻, NO₂⁻, SO₄²⁻). A carbonate removal device (CRD-200) was installed to remove the carbonate peak. Precision and accuracy of cation and anion concentrations are better than 5%. The sum of all cations and anions gives the total dissolved solids (TDS), which are quoted in parts per thousand (‰) to allow for easy comparison with salinity measurements of other studies. Ions were measured in mg L⁻¹ and are quoted with respect to Standard Mean Ocean Water (SMOW) ion ratios according to Maus and others (2011) (e.g., ΔMg/Cl = (Mg/Cl - (Mg/Cl)SMOW)/(Mg/Cl)SMOW), in which the SMOW ratios were taken from Millero and others (2008). Since individual frazil ice crystals are thought to expel all salt during their formation (Tison and others, 2001), salt could only be included in the pore spaces of the agglomeration or at grain boundaries. Tison and others (1993) suggest that ion fractionation in marine ice is further evidence of the presence of sea-water filled pores, which consolidate slowly, allowing for the selective incorporation of ions. Previous studies have interpreted variations in ion ratios to result from changes in the consolidation rate of marine ice (Oerter and others, 1992; Tison and others, 1993; Moore and others, 1994).

The relative concentration of oxygen and deuterium isotopes (δ¹⁸O and δD) was measured with a Picarro laser spectrometer relative to SMOW (e.g., δ¹⁸O = ((¹⁸O/¹⁶O)SMOW) / ((¹⁸O/¹⁶O)SMOW) × 1000). The samples were repeat-injected eight times whereby the first three injections from one sample were discarded due to carry-over from previous samples. The measurement precision was 0.07‰ for δ¹⁸O and 0.40‰ for δD. In order to establish whether marine ice at SMIS was formed from a constant or a changing mixed water source, the regression slope of co-isotopic plots of δ¹⁸O and δD was established and compared to co-isotopic freezing slopes (Souchez and Jouzel, 1984) and mixing slopes (Souchez and Groote, 1985) models. Regression slopes in mixing models are ~8 (Souchez and Groote, 1985), whereas freezing models have smaller slopes (Souchez and Jouzel, 1984) whereby the exact slope magnitude depends on the isotopic composition of the water source. Data trends and interrelations are assessed with the Pearson’s product-moment correlation coefficient and are expressed in terms of r. All r values quoted in this paper are significant at the 99% level.

Ice crystal facies

Thin sections of ice were prepared according to the method detailed by Durand and others (2006) vertically along the full length of the ice cores to a thickness of ~0.5 × 10⁻³ m using a conventional biological microscope in a cold room maintained at ~15°C. Thin sections were photographed between cross-polarized light, and ice crystals were classified into different ice facies (granular, banded and platelet) based on their shape (Table 2) for every 0.10 m down-core coinciding with the sampling for the ice chemistry. Where 30–70% of both banded and granular ice crystals were present in a 0.10 m long section of thin sections, the ice facies were classified as a mixed ice facies. On the rare occasions when the ice crystal morphology did not fit any of the descriptions, they were logged as ‘other’. Ice crystal morphology, as apparent in thin section, is analysed in conjunction with ice chemistry in order to determine whether variations in marine ice source water lead to changes in the appearance of marine ice crystals.

Freezing model for frazil ice crystals

The isotopic source water composition for marine ice predominantly formed by frazil ice crystals can be derived by applying a modelled effective fractionation coefficient (αₑ) to the measured isotopic concentration of the ice. This coefficient was calculated using Tison and others’ (2001) boundary layer freezing model for individual marine ice
Table 2. Criteria for the classification of ice crystal facies with examples

<table>
<thead>
<tr>
<th>Ice crystal facies</th>
<th>Individual grain shape</th>
<th>Grain boundaries</th>
<th>Crystal size (approx.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Banded</td>
<td>Elongated, often rectangular, sometimes acicular</td>
<td>Polygon with square edges, often diffuse</td>
<td>0.5–20.0 mm</td>
</tr>
<tr>
<td>Granular</td>
<td>Isometric and orbicular</td>
<td>Polygonal, interlobate</td>
<td>0.5–20.0 mm</td>
</tr>
<tr>
<td>Platelet</td>
<td>Elongated, wider in the middle, thinner toward the edges, sometimes acicular</td>
<td>Interlobate</td>
<td>10.0–50.0 mm</td>
</tr>
</tbody>
</table>

Banded ice crystals
Granular ice crystals
Platelet ice crystals

frazil ice crystals. The model is based upon Burton and others’ (1953) boundary layer model originally developed for simulating solute diffusion into a liquid boundary layer during the solidification of metal:

\[
\alpha_{eff} = \frac{\alpha_{equ}}{1 - \exp\left(-\frac{z_0 \nu}{D}\right)}
\]

where the effective fractionation coefficient \(\alpha_{eff}\) (unitless) is calculated from the equilibrium fractionation coefficient \(\alpha_{equ}\) (unitless) as a function of the growth rate \(\nu\) (m s\(^{-1}\)), diffusion coefficient \(D\) (m\(^2\) s\(^{-1}\)) and boundary layer thickness \(z_0\) (m). The isotopic concentration \(c\) (unitless) in the boundary layer near the ice/water interface hereby changes over time \(t\) (s) according to Fick’s law (Burton and others, 1953; Eicken, 1998):

\[
\frac{\partial c}{\partial t} = D \frac{\partial^2 c}{\partial z^2} + \nu \frac{\partial c}{\partial z}
\]

where \(z\) is the distance from the ice/water interface into the fluid (m) and \(V\) is the freezing velocity of the ice front (m s\(^{-1}\)). The model is run with a Crank–Nicolson scheme to allow for simultaneous diffusion and fractionation during the advancement of a freezing front. In the model, water is frozen from a semi-infinite reservoir (basically assuming an open system).

Tison and others (2001) used a simple rod geometry for frazil ice crystals, with a diameter of 1.0 × 10\(^{-3}\) m and a boundary layer thickness of half the crystal size (0.5 × 10\(^{-3}\) m) (Daly, 1984). Equilibrium fractionation coefficients were taken from Lehmann and Siegenthaler (1981) (\(\alpha_{H2O} = 1.0022\) and \(\alpha_{I2O} = 1.0029\)), and diffusion coefficients from Ferrick and others (2002) (1.06 × 10\(^{-5}\) m\(^2\) s\(^{-1}\) for \(H\), 0.1 m\(^2\) s\(^{-1}\) and 1.21 × 10\(^{-8}\) m\(^2\) s\(^{-1}\) for \(H\)).

RESULTS

In this study we first present the occurrence of banded, granular and mixed ice facies in marine ice from all three SMIS sites. Secondly, the variation of the \(^{8}O\) and \(^{8}D\) ratios in all marine ice cores is described.
relationships are given. Subsequently, the calculated effective isotopic fractionation coefficients, which help to determine theoretical marine ice source water composition, are presented for different freezing speeds. Finally, we give marine ice salinity (TDS) and ion ratios (ΔMg/Cl and ΔK/Mg) indicating ion fractionation during freezing of marine ice pores.

Ice crystal facies of marine ice

The marine ice cores are almost entirely made up of banded and granular ice facies, including a mixed ice facies of these two crystal types that occurs in all three ice cores (Table 3; Figs 3–5). Ice core C5 shows a high percentage of both pure granular (32%) and pure banded ice crystals (24%) (Fig. 5; Table 3). The two shorter ice cores (C3 and C9), however, show an exclusive preference for either pure banded ice crystals (45%) in the case of C3 (Fig. 3; Table 3) or pure granular ice crystals (58%) in the case of C9 (Fig. 4; Table 3). Platelet ice facies are uncommon in marine ice at SMIS and make up <10% of the ice facies in C15 (Fig. 5; Table 3).

Isotopic composition of marine ice

Isotopic values in all the ice cores range from -0.43% to 2.29%, δ18O and -3.80% to 17.61%, δD (Figs 3–6). The isotopic range increases with total core length (Figs 3–5) in the shallowest core, C5 (2.65 m long), the δ18O range is 1.06‰, while δ18O ranges are 1.60‰ in C9 (3.04 m long) and 2.40‰ in the longest ice core, C15 (9.44 m long). In the two shallower cores C5 and C9 (Table 3), collected close to each other (Fig. 1), the data are on average more enriched in heavy isotopes (>1.6%; δ18O) than in core C15 (0.47%; δ18O) (Table 3). In C9 and C15 the δ18O signal shows a linear trend with depth, although the trend is positive in C9 (r = 0.86; Fig. 4) and negative in C15 (r = -0.59; Fig. 5). The δ18O and δD signals are significantly correlated in all cores (r > 0.93), with an overall slope of 8.66 ± 0.13 (r = 0.97) in a co-isotopic plot (Fig. 6). Furthermore, all marine ice samples at SMIS are enriched in heavy isotopes with respect to sea water and there is a wide overlapping range of isotope values for the different ice facies (Fig. 6).

Effective fractionation coefficients

Calculating the effective fractionation coefficients for frazil ice crystals using Tison and others’ (2001) model at variable freezing speeds shows that a change in freezing speed can theoretically cause an isotopic range of 1.55‰ δ18O (Table 4; r = 0.2). This range becomes slightly smaller when pores that fractionate at equilibrium are considered (Table 4; Fig. 7b). In either case the calculated effective fractionation coefficient for oxygen isotopes is at least 0.60‰ lower than the equilibrium fractionation coefficient (Table 4).

Salinity and ion ratios of marine ice

Marine ice collected from SMIS is slightly saline (0.03–1.01% TDS; Figs 3–5), with an average of 0.25% TDS. Similar to the isotope record, the TDS record increases

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**Table 3.** Average marine ice chemical composition and occurrence of ice crystal facies (bd: banded; gr: granular; mx: mixed; pl: platelet; oth: other) in each SMIS marine ice core.

<table>
<thead>
<tr>
<th>Ice core</th>
<th>Length</th>
<th>n</th>
<th>δ18O (%)</th>
<th>δD (%)</th>
<th>TDS (%)</th>
<th>ΔMg/Cl (%)</th>
<th>ΔK/Mg (%)</th>
<th>bd</th>
<th>gr</th>
<th>mx</th>
<th>pl</th>
<th>oth</th>
</tr>
</thead>
<tbody>
<tr>
<td>C5</td>
<td>9.44</td>
<td>89</td>
<td>0.47 ± 0.48</td>
<td>2.42 ± 3.88</td>
<td>0.29 ± 0.18</td>
<td>-9.21 ± 5.15</td>
<td>28.86 ± 12.96</td>
<td>24</td>
<td>32</td>
<td>32</td>
<td>6</td>
<td>2</td>
</tr>
<tr>
<td>C9</td>
<td>3.04</td>
<td>23</td>
<td>1.64 ± 0.43</td>
<td>12.43 ± 3.38</td>
<td>0.20 ± 0.15</td>
<td>-13.64 ± 13.14</td>
<td>20.87 ± 11.09</td>
<td>24</td>
<td>58</td>
<td>42</td>
<td>2</td>
<td>7</td>
</tr>
<tr>
<td>C3</td>
<td>2.65</td>
<td>22</td>
<td>1.63 ± 0.24</td>
<td>13.90 ± 1.37</td>
<td>0.26 ± 0.11</td>
<td>-0.83 ± 7.89</td>
<td>14.29 ± 5.82</td>
<td>43</td>
<td>35</td>
<td>35</td>
<td>2</td>
<td>2</td>
</tr>
</tbody>
</table>
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Fig. 5. Marine ice crystal facies, $\delta^{18}$O, salinity (TDS) and $\Delta$Mg/Cl ratio for core C15. Legend in Figure 3.

Table 4. Calculated effective fractionation coefficients (quoted as ice-water fractionation constants in %, i.e. $\alpha - 1$) using Tisson and others' (2001) boundary layer freezing model considering different frazil ice freezing speeds (taken from Tisson and others, 2001; Smith and others, 2012). Two different scenarios are considered: marine ice formation from frazil ice crystals only ('no pores') and marine ice formation from 85% frazil ice crystals with 15% of pure sea water condensing at equilibrium freezing speed in the remaining pore spaces ('pores').

<table>
<thead>
<tr>
<th>Freezing speed $\times 10^{-6}$ m s$^{-1}$</th>
<th>$\alpha_{H_{2}O}$</th>
<th>$\alpha_{O}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>pores no pores</td>
<td>pores no pores</td>
</tr>
<tr>
<td>equilibrium fractionation</td>
<td>2.91</td>
<td>21.20</td>
</tr>
<tr>
<td>0.3</td>
<td>2.32 2.22</td>
<td>16.24 15.37</td>
</tr>
<tr>
<td>1</td>
<td>1.57 1.33</td>
<td>10.62 8.76</td>
</tr>
<tr>
<td>1.4</td>
<td>1.34 1.07</td>
<td>9.09 6.95</td>
</tr>
<tr>
<td>2.7</td>
<td>1.01 0.67</td>
<td>6.90 4.37</td>
</tr>
</tbody>
</table>

Fig. 6. Co-isotopic diagram for marine ice samples classified by ice facies from all SMIS sampling sites, plotted together with local sea water (taken as Shallow Ice Shelf Water from below the Ross Ice Shelf (Jacobs and others, 1985; Fitzsimons and others, 2012)). The slope for the linear regression for isotopic data from all ice cores together is 8.68 $\pm$ 0.13 ($r = 0.97$). Individually data from C15 have a slope of 7.92 $\pm$ 0.17 ($r = 0.98$), from C9 have a slope of 7.69 $\pm$ 0.29 ($r = 0.99$) and from C5 have a slope of 5.26 $\pm$ 0.48 ($r = 0.93$).

significantly with depth in C9 ($r = 0.83$) and decreases with depth in C15 ($r = 0.48$). Indeed, TDS and the oxygen isotope record are significantly positively correlated in C9 and C15 ($r > 0.54$). However, in C5 there is no significant linear relationship between isotopes and TDS. Samples of banded and granular ice facies occur within the whole range of measured salinities and isotopic compositions (Fig. 8a).

Nonetheless, there is an apparent difference in the mean isotopic composition and salinity of different ice facies in C15; banded ice facies are more isotopically enriched and more saline than granular and mixed ice facies (Fig. 8b). Similarly, the granular ice facies in C9 are less saline than the mixed ice facies (Fig. 8b). However, there is no significant difference between the chemical compositions of different ice facies; the standard deviations of mean chemical composition for different ice facies overlap (Fig. 8b).

The marine ice at SMIS is depleted in $\Delta$Mg/Cl and enriched in $\Delta$K/Mg (Table 3). Increases in the $\Delta$Mg/Cl signal coincide with increases in the TDS record in all cores (e.g. increase of $\sim$0.20% TDS and $\sim$0.30 $\Delta$Mg/Cl at 2.5 m in C9; Fig. 4). The $\Delta$Mg/Cl and TDS signal in C5 and C9 are positively correlated ($r > 0.64$). No significant relationship, however, exists between the $\Delta$Mg/Cl ratio and TDS in C15.

**DISCUSSION**

Here we first present the evidence for marine ice formation from frazil ice crystals that nucleate in supercooled water. Applying Tisson and others' (2001) boundary layer freezing model, the isotopic composition of the marine ice water source is calculated using effective fractionation coefficients. Subsequently, we evaluate the evidence for marine ice formation from a changing mixed water source rather than a constant water source with a changing freezing rate. Finally, we discuss the possible geographic origin of the relatively fresher water needed in the frazil ice formation.
formation solely from the advancement of a freezing front and associated formation of columnar ice crystals (e.g., Zotikov and others, 1980) is indeed rare (Table 1) and no columnar ice crystal facies have been observed in thin sections from SMIS. Further evidence for SMIS frazil ice formation from supercooled water is provided by a higher $\delta^{18}$O/$\delta$D ratio in marine ice from C5 and C9 than is predicted by the output of Tison and others’ (2001) source model (Fig. 7a and b). This enrichment in $\delta$D could result from an enhanced diffusivity of deuterium in saline waters (Horita, 2009) together with a fast freezing speed, which is thought to amplify kinetic effects (Souchez and others, 2000). This is in keeping with Souchez and others (1995), who suggested using a 2% higher equilibrium fractionation coefficient to explain the higher enrichment of $\delta$D in marine ice.

Marine ice might not be made up of frazil ice crystals alone since these generally expel all salt during their formation (Tison and others, 1993) and SMIS marine ice is slightly saline (Table 2). Ion fractionation in SMIS marine ice samples (depletion in $\Delta$Mg/Cl and enrichment in $\Delta$K/Mg: Table 3) also suggests the presence of sea water in pore spaces between frazil ice crystals. Nonetheless it is difficult to estimate the percentage of sea-water filled pores from its chemistry since marine ice pore water salinity and desalination processes are poorly understood. No brine channels have ever been observed in marine ice, and evidence of brine pockets was only found in the lower tens of metres of the ~200 m thick marine ice layer (which is still in hydraulic connection with the ocean) at the Amery Ice Shelf (Craven and others, 2004; Treverrow and others, 2010).

**Isotopic evidence of a changing water source**

Whether marine ice was formed from pure frazil ice crystals only or from a combination of frazil ice crystals and frozen pore water, the isotopic record of marine ice predominantly reflects the source water composition of frazil ice crystals. The measured isotopic range and co-isotopic slope of all SMIS marine ice samples indicate frazil ice formation from a mixed water source that changes over time. Even though the measured $\delta^{18}$O range for samples from all ice cores at SMIS (2.71% $\delta^{18}$O and 21.41% $\delta$D; Fig. 6) is similar to the theoretical maximum isotope shift at equilibrium fractionation from a constant water source (Table 4), results obtained by applying Tison and others’ (2001) boundary layer freezing model indicate that the isotopic composition of marine ice should only vary by ±1.55‰ $\delta^{18}$O and ±11.00‰ $\delta$D considering a range of measured freezing rates for frazil ice crystals and a constant water source (no pore water; Fig. 7a; Table 4). If pores filled with sea water that fractionates upon consolidation are present, the isotopic range of marine ice formed at variable typical freezing speeds will be even smaller (1.31% $\delta^{18}$O and 9.34% $\delta$D; 15% pore water; Fig. 7b; Table 4). Hence, the measured isotopic range of marine ice samples at SMIS indicates a source water composition that is likely not constant over time and in space, assuming an open reservoir in the ice-shelf cavity. This idea is further supported by a high co-isotopic regression slope of 8.66 ± 0.13 (r = 0.97) (Fig. 6), which is steeper than a freezing slop (i.e. water formed from the same water source with a variable freezing speed) in an open or closed system (Jouzel and Souchez, 1982; Souchez and Jouzel, 1984). However, since frazil ice crystals form in a water plume in the ice-shelf cavity, closed system freezing is precluded. Nonetheless, the highest process and what this implies with regard to marine ice formation mechanisms.

**Marine ice formation from frazil ice crystals**

The presence of predominantly banded and granular ice facies (Table 1) in marine ice at SMIS strongly suggests that this marine ice was formed from predominantly frazil ice crystals (e.g., Tison and others, 1993; Treverrow and others, 2010) which nucleate from supercooled water. Marine ice
possible freezing slope in an open system is theoretically only 7.38, applying Souchez and Jouzel's (1984) freezing model to a water source of pure sea water taken as Shallow Ice Shelf Water (SISW) below the Ross Ice Shelf (Jacobs and others, 1985; Fitzsimons and others, 2012). Freezing slopes of sea water with a constant contribution of melted meteoric ice would be even lower. Co-isotopic mixing slopes for ice formed from a changeable water source, however, would be higher, ~8 (Souchez and Grotte, 1985), which is more similar to the measured slope of all SMIS samples (Fig. 6).

For individual ice cores the isotopic range is smaller and co-isotopic slopes are shallower (Fig. 6). While the co-isotopic slope of 7.92 ± 0.17 in C15 with an isotopic range of 2.4% δ18O remains similar to a mixing slope, the two shallower ice cores C5 and C9 have a slope ≤ 7.69 with a smaller isotopic range ≤ 1.50% δ18O. This could suggest that the source water was similar during accretion events, the core of which marine ice at the core sites of the shallower Ice cores, extracted closer to each other (Fig. 1), while the source water composition at the C15 site could have changed over the course of accreting ~9.5 m of marine ice due to the co-isotopic slope close to 8 and an isotopic range ≥ 3.5% δ18O. The low slope of 5.2 ± 0.48 in co-isotopic data from C5 is lower than a freezing slope, even if the source water was mainly glacial water derived from meteoric ice. The isotopic range of 1.06% δ18O with most samples clustering tightly around 1.5% δ18O (Fig. 6) might be too small to allow for a representative freezing or mixing slope.

Since there is some evidence that marine ice at SMIS partially formed from consolidation of sea-water filled pores, an increase in marine ice salinity is consistent with slower consolidation rates of frazil ice crystals. A significant correlation between the TDS and ΔMg/Cl signal in C5 and C9 could therefore suggest that sea-water filled pores consolidate more slowly (leading to higher ion fractionation) when salinity is high.

**Source water composition**

Since the crystallography provides evidence that marine ice is primarily composed of frazil ice crystals, which nucleate in supercooled water, the marine ice source water at SMIS is likely a mixture of local sea water and fresher and possibly warmer water. This combination of water masses supercools upon mixing (double diffusion mechanism) and adiabatic lifting (ice pump mechanism). The fresher-than-sea water could originate from melted meteoric ice (Souchez and others, 1995, 1998) and/or melted marine ice (Souchez and others, 1991; Tison and others, 1993).

Applying the calculated effective fractionation coefficients of Tison and others’ (1993) model (Table 4) to sea water suggests that marine ice at SMIS with an isotopic concentration ≥ 1.70% δ18O (Fig. 7b) could have formed from a water source with a component of recycled marine ice. The exact cut-off threshold depends on the calculated fractionation coefficient and the assumed sea-water composition. This study took SISW as measured below the Ross Ice Shelf (Jacobs and others, 1985; Fitzsimons and others, 2012) as the ambient sea water, but it remains unknown whether the pure sea water below SMIS could be more enriched in heavy isotopes by ~0.30% δ18O. The idea of marine ice formation from a water source with a component of melted and relatively fresher recycled marine ice was first developed by Souchez and others (1991) based on marine ice that was more enriched in heavy isotopes (up to 3.20% δ18O: Table 1) than could be explained by applying the equilibrium fractionation coefficient to sea water. However, equilibrium freezing during frazil ice formation is unlikely. Since the isotopic composition of marine ice in C5 and C9 is on average ~2% more enriched in δ18O than sea water, a high contribution of recycled marine ice (~40% dependent on freezing speed) to the source water is necessary to explain all the observed enriched marine ice samples (Fig. 7a and b). The exact percentage contribution of recycled marine ice, however, is difficult to determine, especially since marine ice can be repeat-recycled. Goodwin (1993) considered the possibility that δ18O marine ice samples, which are not as enriched as in Souchez and others’ (1991) study (i.e. only up to 1.56% δ18O: Table 1), could have formed from repeat recycling of previously formed marine ice.

Less enriched (~1.70% δ18O) marine ice samples could have formed with a small contribution of glacial water derived from meteoric ice to the marine ice water source (especially in C15; Fig. 7a and b). Since meteoric ice is very

---

**Fig. 8.** Total dissolved solids (TDS) and oxygen isotopes (δ18O) for the different ice crystal facies in the individual ice cores. (a) Individual samples from all cores; (b) averages for all cores ± one standard deviation. Note: same key for both diagrams.
depleted in heavy isotopes (–35% to –25% δ¹⁸O; Kellogg and others, 1991a,b), a small glacial water contribution to the source water (up to a maximum of 4%) is sufficient to explain the isotopic composition of some of the less isotopically enriched and isotopically depleted ice samples in C15 (Fig. 7a and b). Previous studies interpreted the presence of a mixing slope, together with a larger isotopic range (of up to 18.00% δ¹⁸O; Table 1) than observed at SMIS of isotopically depleted samples, as the result of a mixed water source of sea water and varying proportions of melted meteoric ice (Goodwin, 1993; Souchez and others, 1995, 1998). The marine ice source water for all sample sites at SMIS, however, could consist of a mixture of all three water sources – melted meteoric ice, melted marine ice and sea water – whereby melted meteoric ice would contribute least to the mixture (in the order of 5%) due to its high isotopic depletion.

There is no evidence that different crystal morphologies are the result of marine formation from different water sources and/or proportions of pore water. There is no significant difference in the chemistry of the most common ice crystal facies (Fig. 8a); the mean isotopic signature and salinity of banded ice crystals differs only slightly from that of granular ice facies in all cores in which the standard deviations overlap (Fig. 8b).

**Processes of frazil ice formation at SMIS**

Here we discuss geophysical scenarios that can lead to the formation of marine ice source water as identified through the ice chemistry in the previous subsection. One possible mechanism for marine ice formation at SMIS could be the ice pump, delivering fresh water of meteoric origin to the ice-shelf cavity. This process could explain marine ice formation at site C15 where the ice likely formed from a mixed water source dominated by glacial and sea water (Fig. 7a and b). Indeed, there is a thermohaline circulation driven by either HSSW or modified Circumpolar Deep Water (mCDW) in the cavity of the adjacent Ross Ice Shelf (Dinniman and others, 2007). Ocean circulation models reveal that there is an amplified basal melt rate at the eastern margin of the Ross Ice Shelf that extends from the Ross Ice Shelf margin south past Minna Bluff (Linnemann and others, 2007; Timmermann and others, 2012). A buoyant plume of meteoric meltwater and sea water generated below the Ross Ice Shelf could enter the ice-shelf cavity of the ~300 m thinner SMIS, similar to the vertical ascent of a water plume from below the Ross Ice Shelf to the cavity of the northern McMurdo Ice Shelf (Robinson and others, 2014). Here the water in the plume would supercool due to a change in the pressure-dependent freezing point in shallower waters, and frazil ice crystals would nucleate. Potentially this water plume could also reach the ocean floor of the much shallower SMIS cavity with a maximum depth of ~400 m (Johnston and others, 2008), forming anchor ice as suggested in a modelling effort by Leonard and others (2014). However, marine ice below the Ross Ice Shelf accumulates only in a thin layer (Zotikov and others, 1980; Timmermann and others, 2012). Hence the ice pump mechanism might be not very strong in the Ross Ice Shelf cavity. Also, this ice pump would only explain the delivery of fresh water of meteoric origin into the SMIS cavity and not the recycling of marine ice.

A proportion of marine ice at SMIS shows an apparent fractionation in water isotopes that exceeds the modelled effective fractionation for freezing of frazil ice crystals from a pure sea-water source. After Souchez and others (1991) and Tison and others (1993), marine ice could be recycled in a mechanism where a shallow thermohaline or tidal circulation melts the older marine ice at depth, which then becomes supercooled upon rising to shallower waters where frazil ice crystals nucleate. However, it is unlikely that a shallow tidal circulation would reach the SMIS cavity, which is sheltered by Black and White Islands and is several tens of kilometres from the open ocean (Fig. 1).

Alternatively, marine ice at SMIS could be recycled through melting of exposed marine ice at the ice-shelf surface which would then be routed to the ice-shelf base through tide cracks. Indeed, surface melt at SMIS was observed to temporarily pool in elongated lakes between ridges of exposed marine ice during the summers of 2007 and 2010. In Figure 2 such lakes occur between the clearly distinguishable ice ridges but are frozen since the picture was taken at the start of the season. Also, a tide crack was observed to run parallel to Minna Bluff in the marine ice zone close to shore, where the ice shelf becomes regrounded. Salt deposits along the tide crack indicate that it actively connects the ice-shelf base with the ice-shelf surface. As surface meltwater would drain through the tide crack, it could become supercooled in a double-diffusion mechanism of heat and salt with the ambient sea water, whereby frazil ice crystals would nucleate. For this process to occur, both water masses, the sea water at the ice-shelf base and the recycled marine ice, need to be at their salinity-dependent freezing point and would become mixed due to local turbulence (McPhee and others, 2013). Hence, the surface meltwater would need to be cooled during percolation. In this process some refreezing could occur. Indeed, solute chemistry and ice facies of the ~9.5 m long ice core C15 show evidence of local refreezing. Between 6.8 and 7.2 m depth, for example, the marine ice salinity increases by almost an order of magnitude and the ice facies are predominantly large platelet-type crystals (Fig. 5). Similarly, platelet ice facies are present between 0.5 and 1.8 m depth together with several peaks in salinity (Fig. 5). This could be a result of surface meltwater refreezing at depth.

Recycling of ice melted at the ice-shelf surface, however, is not necessarily limited to marine ice. Water from the nearby glaciers and snowpatches on Minna Bluff close to the shore of the ice shelf (Fig. 1) could find its way into the tide cracks of SMIS and hence to the ice-shelf base. This process would be an alternative explanation for the small contribution of glacial water to the source water at SMIS, especially at site C15 (Fig. 7a and b). Recycling surface meltwater in the process of marine ice formation would imply that the ice shelf could be sustaining itself to some extent, with most of the marine ice forming during the melt season in summer. Surface ablation and local accumulation of marine ice were in equilibrium (e.g. Kellogg and others, 1990) their rates would both be in the order of ~0.1 m a⁻¹, as the measured surface ablation on SMIS (Clifford, 2005). Hence, the surface ice of the ~100 m thick SMIS would be ~1000 years old, similar to the youngest radiocarbon date of 1230 ± 50 years BP (Denton and Marchant, 2000).

Instead of recycling marine ice from the ice-shelf surface, McPhee and others (2013) suggests that loose frazil ice crystals could also be recycled in ocean turbulence before they consolidate. In this process, frazil ice crystals would be moved to deeper waters where they melt due to the pressure
dependence of the freezing point. Upon rising of the resulting more buoyant meltwater plume, frazil crystals would re-ripenate (McPhee and others, 2013). If recycling of surface meltwater or loose frazil crystals occurs below SMIS, turbulence must occur at the ice-shelf base to allow for vertical mixing. In the absence of a thermohaline and tidal circulation, it still remains to be determined how this vertical turbulence would be initiated.

There is little evidence that suggests different marine ice formation processes lead to different ice crystal facies at SMIS. Banded ice crystals are slightly more enriched in heavy isotopes and saline than the granular ice facies, especially in samples from C15 (Fig. 8a and b), but this difference is not statistically significant. Tison and others (1993) found a similarly weak chemical difference between the two ice facies but suggested that banded ice facies were generated by frazil ice crystals that aligned in a sub-ice-shelf current, whereas granular ice crystal facies were thought to result from fast frazil ice formation in a deeper thermohaline circulation further inland. In contrast, Treverrow and others (2010) found that banded ice crystal facies merely occurred in younger marine ice than granular ice crystals. This would suggest a change in crystal morphology over time. Indeed, ice deformation experiments showed that due to post-depositional ice growth and recrystallization processes associated with ice flow, the original ice crystal morphology is often altered (Wilson and others, 2014) and predominantly granular/less elongated ice crystals develop in marine ice (Dierckxsens and others, 2014).

CONCLUSION

The aims of this study were to determine the composition and origin of water masses of marine ice at SMIS. We have focused our analysis on the crystal morphology and the chemistry of the ice and used a boundary layer freezing model to derive the source water composition. The specific conclusions are:

SMIS marine ice was mainly formed by frazil ice crystals since the ice is almost entirely composed of granular and banded ice crystal facies.

The source water composition of SMIS marine ice is not constant but changes spatio-temporally. This is indicated by a co-isotopic mixing slope of \( \sim 8 \) and an isotopic range of 2.7\% \( \delta^18O \) and 21.4\% \( \delta^13C \) for all marine ice samples from all sites, which is almost double the isotopic range predicted by the boundary layer freezing model for marine frazil ice formed from a constant water source at different freezing speeds.

Marine ice samples, which are isotopically depleted as compared to modelled marine frazil ice formed from pure sea water, have likely formed from a source water mixture of sea water and a relatively small proportion (~4\%) of melted meteoric ice. This isotopically depleted melted meteoric ice could have been vertically advected from the ice-shelf base of the adjacent ~300 m thicker Ross Ice Shelf in an ice pump mechanism.

Marine ice samples, which are isotopically enriched compared to modelled marine frazil ice formed from pure sea water, have likely formed from a source water mixture of sea water and a large proportion (~40\%) of isotopically enriched melted marine ice. The exact percentage varies with freezing speed and the number of freeze–thaw–refreeze cycles. In a proposed recycling mechanism, marine ice, which is exposed at the ice-shelf surface, melts and drains through the tide crack to the ice-shelf base, where it becomes turbulently mixed with relatively colder and saltier sea water and frazil ice crystals nucleate in a double diffusion process. Thus, summer surface melt could make a substantial contribution to basal ice-shelf mass accretion particularly at the southern margin of SMIS.

Consequently we conclude that basal boundary conditions and the formation of marine ice at the relatively thin SMIS are at least partially determined by surface meltwater generation. The proposed mechanism of marine ice recycling thus ties together surface and basal processes, and could potentially slow ice-shelf disintegration.

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REFERENCES


Clifford AE (2005) Physiography, flow characteristics and vulnerability of the Southern McMurdo Ice shelf, Antarctica. (MSc thesis, University of Otago)
Koch and others. Marine ice recycling at southern McMurdo Ice Shelf


Koch and others: Marine ice recycling at southern McMurdo ice shelf

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formation sites in coastal Antarctic Waters. J. Spatial Sci., 59(2),
297–312 (doi: 10.1007/s11280-014-9132-727)

Lewis EL and Perkin RG (1986) Ice pumps and their rates.
JC091iC10p1756

Mager S, Smith IJ, Kempema EW, Thomson BJ and Leonard GH
468–483 (doi: 10.1177/0309133313479815)

002115)

ponds, or double diffusion at the freezing point. J. Fluid Mech.,
64, 307–327 (doi: 10.1017/S0022112074002527)

Maus S and 7 others (2011) Ion fractionation in young sea ice from
Kongsfjorden, Svalbard. Ann. Glaciol., 52(57), 301–310 (doi:
10.3189/026030611798931804)

McMahon KL and Lackie MA (2006) Seismic reflection studies of
the Amery Ice Shelf, East Antarctica: delineating meteoric and
j.1365-246X.2006.03043.x)

McPhee MG, Skogseth R, Nilsen F and Smelshus LH (2013)
Creation and tidal advection of a cold salinity front in
Storfjorden: 2. Supercooling induced by turbulent mixing of
cold water. J. Geophys. Res.: Oceans, 118(3), 3737–3751 (doi:
10.1002/jgrc.20161)

compositional standard of seawater and the definition of the
50–72 (doi: 10.1016/j.dsr.2007.10.001)

Monaghan AJ, Bromwich DH, Powers JG and Manning KW
(2005) The climate of the McMurdo, Antarctica, region as
represented by one year of forecasts from the Antarctic
Mesoscale Prediction System. J. Climate, 18(8), 1174–1189
(doi: 10.1175/JCLI3336.1)

Moore JC, Reid AP and Kipfstuhl J (1994) Microstructure and
electrical-properties of marine ice and its relationship to
meteoric ice and sea-ice. J. Geophys. Res., 99(C3), 5171–5180 (doi:
10.1029/93JC02832)

Morgan VI (1972) Oxygen isotope evidence for bottom freezing on
Amery Ice Shelf. Nature, 238(5364), 393–394 (doi: 10.1038/
238393a0)

Morse B and Richard M (2009) A field study of suspended frazil ice
j.coldregions.2008.03.004)

Ooster H and 6 others (1992) Evidence for basal marine ice in the
Filchner-Ronne Ice Shelf. Nature, 358(6385), 395–401 (doi:
10.1038/358395a0)

Pattyn F and 8 others (2012) Melting and refreezing beneath Roi
Baudouin Ice Shelf (East Antarctica) inferred from radar, GFS,
and ice core data. J. Geophys. Res.: Earth Surf., 117(F4), F04008
(doi: 10.1029/2012JF002154)

Robinson NJ, Williams MM, Stevens CL, Langhome PJ and Haskell
TG (2014) Evolution of a supercooled Ice Shelf Water plume with
an actively growing subice platelet matrix. J. Geophys.

83–84, 57–70 (doi: 10.1016/j.coldregions.2012.06.005)

Sauzeau RA and Grooto JM (1985) δ18O-δ16O relationships in ice
formed by subglacial freezing: paleoclimatic implications. J.
Glaciol., 31(109), 229–232

and 88O of water and ice during freezing. J. Glaciol., 30(106),
369–372

Sauzeau R and 8 others (1991) δ13C composition evidence of marine
ice transfer along the bottom of a small Antarctic ice shelf.

Sauzeau R and 6 others (1995) Investigating processes of marine ice
formation in a floating ice tongue by a high-resolution isotopic
94JC01429)

at the grounding line in East Antarctica: possible implications for

Sauzeau R, Jouzel J, Lorain R, Skewes D, Stevenud M and
(doi: 10.1029/2000GL006103)

Swifthank N (1970) Ice movement in the McMurdo Sound area
of Antarctica. IAHS Publ. 86 (Symposium at Hanover 1968 –
Antarctic Glaciological Exploration (SAGE)), 472–487

Timmermann R, Wang Q and Hellmer HH (2012) Ice-shelf basal
melting in a global finite-element sea-ice/cryosphere model.

generation at the base of a small Antarctic ice shelf. Antarct.

to the simulation of the combined isotope/salinity signal of
10.1029/2000JC000207)

and marine ice crystal orientation fabrics from the Amery Ice
Shelf, East Antarctica, J. Glaciol., 56(199), 877–890 (doi:
10.3109/00221431307941457353)

processes in the East Antarctic sea ice zone. Antarct. J. US, 32,
185–187

(2010) Basal melting and freezing under the Amery Ice Shelf,
East Antarctica, J. Glaciol., 56(195), 01–90 (doi: 10.3109/
002214313079198020)

Wilson CJL, Petemell M, Piazolo S and Luzin V (2014) Micro-
structure and fabric development in ice: lessons learned from
in situ experiments and implications for understanding rock
jsug.2013.05.006)

Zotikov VA, Zagorodnov VS and Raikovskiy IV (1980) Core drilling
through the Ross Ice Shelf (Antarctica) confirmed basal freezing.
4438.1463)

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