Brine distribution in young sea ice

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A dissertation submitted for the degree of Doctor of Philosophy

Jesus College
March 1999
Declaration

I declare that this dissertation is the result of my own original work and includes nothing which is the outcome of work done in collaboration. The research was part of an interdisciplinary project and in those instances where I have availed myself of the work of others, this is clearly indicated and acknowledged. The dissertation does not exceed the regulations on length and has not been submitted to any other university, or similar institution, for any degree or similar qualification.

[Signature]

Finlo Cottier
March 1999

Publications arising from this research:


For my family

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Brine distribution in young sea ice

Finlo Cottier

Many properties of sea ice are influenced by its salt content. Salt, in the form of brine, is incorporated in the solid ice matrix during ice growth. This research represents a systematic study of the distribution of salt, principally in the form of brine, in young sea ice, according to the thermal, structural and crystallographic environment.

Sea-ice processes are influenced by the existence of brine drainage channels. These have been propounded as the cause of horizontal variability in brine distribution though no formal correlation has yet been established. A nomenclature is defined for the brine features found in sea ice, according their simple geometric characteristics, and a physical chronology of the formation and evolution of brine drainage channels is presented based on previously independent observations.

The data were derived from sea ice grown in a large ice tank. A novel technique for sampling was employed in which the retention of brine is maximised and its spatial arrangement in the sample is preserved. Using this method, the brine distribution was determined to a high spatial resolution and the location of the brine structures in the ice was recorded.

The brine distribution was compared to the occurrence the brine channels, and the subsequent redistribution of brine was investigated with respect to changes in the ice temperature. The research has demonstrated conclusively that variability in the brine distribution of cold, congelation ice can be correlated to the existence of brine drainage channels; in warmer ice the brine is more uniformly distributed. By simulating melting and refreezing, mechanisms for brine redistribution through the ice are proposed based on temperature induced changes in the porosity and permeability of the ice.

The brine distribution in new sea ice was investigated and shown to be homogeneous in the initial skim. As the ice thickens the lateral distribution becomes increasingly variable, coinciding with the emergence and development of brine channels. It is hypothesised that the spatial density of the brine channels is determined by their efficiency as sinks for brine. Immediately after ice formation brine is rejected from new ice. The disparity between this and recently published observations on the formation of sea ice as a mushy layer is resolved on the grounds of ice growth rate, concluding that mushy layers are not representative of sea ice in general.

The research examined the lateral distribution of brine in a granular ice type and it was shown to be uniform on the scale of the measurements. The ice was devoid of brine channels and migration of brine was mediated by an amorphous intergranular network. It is hypothesised that the absence of brine channels is a consequence of the ice growth mechanism which impedes the formation of brine channels.
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Although there is only one name on the cover of this dissertation, it is the product of input from a great many people. I would like to extend my appreciation to all those with whom I have shared the last few years.

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# Contents

Title ........................................................................................................... i
Declaration .................................................................................................. ii
Dedication .................................................................................................... iii
Abstract ...................................................................................................... iv
Acknowledgements .................................................................................... v
Contents ...................................................................................................... vi
Figures ........................................................................................................ xii
Tables .......................................................................................................... xvii

## Chapter 1 Introduction

1.1 Sea ice in the polar environment ..................................................... 1
1.2 Young sea ice .................................................................................... 3
1.3 Brine distribution and variability .................................................... 4
   1.3.1 Observations ............................................................................ 5
      1.3.1.1 Metre scale ....................................................................... 5
      1.3.1.2 Micro scale ....................................................................... 6
1.4 Thesis objectives ............................................................................. 7
1.5 Thesis structure ............................................................................. 8

## Chapter 2 Sea ice

2.1 Introduction ..................................................................................... 10
2.2 Fundamentals .................................................................................. 10
   2.2.1 Ice ......................................................................................... 11
   2.2.2 Phase relations ....................................................................... 12
Chapter 3 Brine channels: structure, formation and processes

3.1 Introduction .................................................. 32
3.2 Visualising brine structures ........................................ 33
   3.2.1 Two-dimensional imaging ..................................... 33
   3.2.2 Three-dimensional imaging .................................... 34
3.3 Nomenclature .................................................. 36
   3.3.1 Brine structures ............................................. 37
       3.3.1.1 Brine channels ......................................... 37
       3.3.1.2 Brine tubes ........................................... 39
       3.3.1.3 Salient characteristics ............................... 39
   3.3.2 Classification schemes ...................................... 40
       3.3.2.1 Generations ........................................... 41
       3.3.2.2 Pore space ........................................... 42
   3.3.3 Definition of terms ........................................ 43
3.4 Formation and evolution of brine channels

3.4.1 Brine tube formation

3.4.2 Brine channel formation

3.4.2.1 Location

3.4.2.2 Spatial distribution

3.4.3 Brine channel evolution

3.4.4 Brine channel geometry

3.5 Brine channels and sea-ice processes

3.5.1 Mass transport

3.5.2 Thermal transport

3.5.3 Optics

3.5.4 Biology

3.5.5 Mechanics

3.6 Summary

Chapter 4 Experimental conditions and methods

4.1 Introduction

4.2 Ice tank studies

4.3 The experimental facility and projects

4.3.1 The ice tank facility

4.3.2 The Interice projects

4.3.2.1 Interice I

4.3.2.2 Interice II

4.4 Bulk parameters

4.4.1 Air temperature

4.4.1.1 January 1997

4.4.1.2 November 1998

4.4.2 CTD data

4.4.2.1 January 1997

4.4.2.2 November 1998

4.4.3 Ice thickness

4.4.3.1 January 1997
Chapter 5 Brine redistribution during thermal forcing

5.1 Introduction ............................................................................................................................................ 91
5.2 Melt .......................................................................................................................................................... 92
  5.2.1 Cold ice .............................................................................................................................................. 93
    5.2.1.1 Bulk profiles .............................................................................................................................. 93
    5.2.1.2 Bulk salinity ............................................................................................................................ 95
    5.2.1.3 Brine structures ......................................................................................................................... 97
  5.2.2 Warm ice ........................................................................................................................................... 100
    5.2.2.1 Bulk profiles .......................................................................................................................... 100
    5.2.2.2 Bulk salinity .......................................................................................................................... 102
    5.2.2.3 Brine structures ...................................................................................................................... 105
5.3 Discussion ............................................................................................................................................... 107
  5.3.1 Justification of the methods ............................................................................................................. 107
    5.3.1.1 Distortion of the brine distribution ........................................................................................ 108
    5.3.1.2 Sampling resolution .............................................................................................................. 108
    5.3.1.3 Interpolation of raw data ...................................................................................................... 109
  5.3.2 Brine distribution ............................................................................................................................... 110
    5.3.2.1 Characteristics of the brine distribution .................................................................................. 111
    5.3.2.2 Mechanisms for brine redistribution ..................................................................................... 114
5.4 Refreeze ................................................................. 118
   5.4.1 Warm ice ......................................................... 118
      5.4.1.1 Bulk profiles ........................................... 118
      5.4.1.2 Bulk salinity .......................................... 119
   5.4.2 Cold ice .......................................................... 121
      5.4.2.1 Bulk profiles ........................................... 121
      5.4.2.2 Bulk salinity .......................................... 122
   5.4.3 Brine structures in warm and cold ice .................... 124
   5.4.4 Discussion ......................................................... 127
5.5 Summary ............................................................. 128

Chapter 6 Evolution of the brine distribution in new ice

6.1 Introduction .......................................................... 130
6.2 Initial ice skim ................................................................ 131
   6.2.1 Results ............................................................. 132
   6.2.2 Discussion .......................................................... 136
6.3 Time series ice sampling ................................................ 138
   6.3.1 Results ............................................................. 138
      6.3.1.1 Bulk profiles ........................................... 139
      6.3.1.2 Day 1 .......................................................... 141
      6.3.1.3 Day 2 .......................................................... 144
      6.3.1.4 Day 3 .......................................................... 147
   6.3.2 Discussion .......................................................... 151
      6.3.2.1 Desalination of new ice .................................. 151
      6.3.2.2 Brine distribution ........................................ 155
      6.3.2.3 Brine channel distribution ................................ 156
      6.3.2.4 Development of brine channel morphology .......... 158
      6.3.2.5 Synthesis ....................................................... 159
6.4 Summary ................................................................. 160
Chapter 7  Aspects of brine distribution in frazil ice

7.1 Introduction .......................................................... 161
7.2 Frazil formation .................................................. 162
7.3 Results ................................................................. 163
    7.3.1 Textural analysis .............................................. 164
    7.3.2 Bulk profiles .................................................... 165
    7.3.3 Bulk salinity .................................................... 167
    7.3.4 Brine structures ................................................. 169
7.4 Discussion ............................................................. 171
7.5 Summary ............................................................... 176

Chapter 8  Summary of conclusions ........................................... 177

8.1 Brine distribution and its variation with ice temperature .......... 179
8.2 Evolution of brine distribution and brine channel morphology .. 181
8.3 Contrasts of the brine distribution associated with ice texture .. 182
8.4 Future directions ...................................................... 183

References ................................................................. 185
Figures

Figure 2.1: The crystal lattice of ice I(h) showing the hexagonal symmetry of the structure and the crystal axes. ................................................................. 11

Figure 2.2: Equilibrium phase diagram for sea ice.................................................................................................................................................. 13

Figure 2.3: The temperature of maximum density ($T_m$) and the freezing point ($T_f$) of water as a function of its salinity. ................................................................................................................................. 15

Figure 2.4: A schematic representation of the mismatch in temperature profiles which gives rise to constitutional supercooling at the ice-water interface. ........................................................................ 19

Figure 2.5: A simplification of the skeletal layer showing the platelet and groove structure of two columnar crystals. The large arrow indicates the direction of ice growth. ................................................................. 20

Figure 2.6: Bulk salinity as a function of ice thickness in Arctic first-year ice showing the slope break in ice salinity at a thickness of about 30 cm. .................................................................................................. 30

Figure 3.1: Schematic representation of a brine drainage channel showing the starburst pattern in horizontal section................................................................. 34

Figure 3.2: Resin casts of the brine-filled spaces in (a) granular ice and (b) congelation ice showing the variability of orientation and branching......................................................... 36

Figure 3.3: Hierarchical diagram showing the relation between the different terms used to describe brine features in sea ice........................................................................................................ 47

Figure 3.4: An illustrative comparison of the characteristics of brine exchange at the ice-water interface from the observations of (a) Niedrauer and Martin (1979) and (b) Worster and Wettlaufer (1997). The arrows in (a) indicate direction of brine flow and the lines in (b) are streamlines of brine flow. ........................................................................................................ 49
Figure 3.5: A schematic drawing in the vertical plane of two possible orientations of inclined platelets at a vertical grain boundary (the c-axis is perpendicular to the platelets). The brine pockets in case A migrate towards the grain boundary whilst in case B they move away.

Figure 4.1: A floor plan of the Environment Test Basin at HSVA.

Figure 4.2: The air temperature in the ice tank - January 1997.

Figure 4.3: The air temperature in the ice tank - November 1998.

Figure 4.4: The water salinity time series - January 1997.

Figure 4.5: The water temperature time series - January 1997.

Figure 4.6: The water temperature time series - November 1998.

Figure 4.7: The water salinity time series - November 1998.

Figure 4.9: Ice thickness measured in the current zone - January 1997.

Figure 4.10: Ice thickness versus square root of time for January 1997 (filled circles) and November 1998 (open circles) showing the linear relation which confirms ice growth under a uniaxial heat flux.

Figure 4.11: The ice surface with brine channels identifiable as the bright points; remnants of frost flowers are also visible.

Figure 4.12: Spatial density $D_c$ (number of channels per 100 cm$^2$) of brine channels as a function of initial ice growth rate.

Figure 4.13: Vertical thin section of the congelation ice fabric seen through crossed polarisers.

Figure 4.14: A series of schematic diagrams illustrating the steps taken to retrieve ice samples from the tank.

Figure 4.15: The subdivisions of each sample to obtain 2 cm$^3$ subsamples.

Figure 4.16: A typical image of a horizontal plate of ice showing the brine channels as bright, star-shaped features.

Figure 4.17: A typical reconstruction of a vertical thick section clearly showing three brine drainage channels.
Figure 4.18: Calibration curve for the salinometer ........................................................... 88

Figure 5.1: Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for cold ice sampled on day 8 of Interice I, January 1997 ................................................. 94

Figure 5.2: Raw and interpolated forms of the bulk salinity data for cold ice on day 8 of the January 1997 period; x and y positions in the horizontal plane ........................................ 96

Figure 5.3: Composite images of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for cold ice sampled on day 8 of the January 1997 period ........................................................................ 98

Figure 5.4: Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for warm ice sampled on day 24 of Interice I, January 1997 ......................................... 100

Figure 5.5: Raw and interpolated forms of the bulk salinity data for warm ice on day 24 of the January 1997 period; x and y positions in the horizontal plane ......................... 103

Figure 5.6: Composite images of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for warm ice sampled on day 24 of the January 1997 period ................................................................. 106

Figure 5.7: Standard deviation in bulk salinity for each horizontal layer of the cold (filled circles) and warm (open circles) ice collected during the January 1997 period ........................................................................ 114

Figure 5.8: Pore space as determined from horizontal thin sections (20 mm horizontal width) for (a) cold ice on day 8 and (b) warm ice on day 17 of the January 1997 period ........................................................................ 115

Figure 5.9: Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for warm ice sampled on day 8 of the November 1998 period ........................................ 119

Figure 5.10: Raw and interpolated forms of the bulk salinity data for warm ice on day 8 of the November 1998 period; x and y positions lie in the horizontal plane ............... 120

Figure 5.11: Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for cold ice sampled on day 10 of the November 1998 period ........................................ 122

Figure 5.12: Raw and interpolated forms of the bulk salinity data for cold ice on day 10 of the November 1998 period; x and y positions lie in the horizontal plane ............... 123
Figure 5.13: Composite images of the horizontal sections showing the brine structures in the ice with the isohalines for (a) warm ice and (b) cold ice obtained on days 8 and 10 of the November 1998 period respectively.

Figure 6.1: Raw and interpolated forms of the bulk salinity data in the horizontal plane for the skim of new ice sampled in the calm zone on day 2 of the November 1998 period.

Figure 6.2: Composite image of the horizontal section showing the brine structures in the ice with the isohalines for the skim of new ice sampled in the calm zone on day 2 of the November 1998 period.

Figure 6.3: Composite image of two vertical sections showing the brine structures in the ice with the bulk salinity for the skim of new ice sampled in the calm zone on day 2 of the November 1998 period.

Figure 6.4: Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for four samples of new ice collected during the November 1998 period.

Figure 6.5: Raw and interpolated forms of the bulk salinity data in the horizontal plane for new ice sampled in the current zone on day 1 of the November 1998 period.

Figure 6.6: Composite image of the horizontal section showing the brine structures in the ice with the isohalines for new ice sampled in the current zone on day 1 of the November 1998 period.

Figure 6.7: Composite image a vertical section showing the brine structures in the ice with the bulk salinity for new ice sampled in the current zone on day 1 of the November 1998 period.

Figure 6.8: Raw and interpolated forms of the bulk salinity data for new ice sampled in the current zone on day 2 of the November 1998 period; x and y positions lie in the horizontal plane.

Figure 6.9: Composite image of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for new ice sampled in the current zone on day 2 of the November 1998 period.
Figure 6.10: Raw and interpolated forms of the bulk salinity data for new ice sampled in the current zone on day 3 of the November 1998 period; x and y positions lie in the horizontal plane. ................................................................. 148

Figure 6.11: Composite image of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for new ice sampled in the current zone on day 3 of the November 1998 period ................................................................. 150

Figure 6.12: Water salinity in the current zone of the tank during the first four days of the November 1998 period. The open circles correspond to the time of ice sampling .................................................................................................................. 152

Figure 7.1: A cross section through the ice cover illustrating the formation of a frazil layer within the congelation ice ............................................................................................................................................. 162

Figure 7.2: Vertical thin section of the frazil ice fabric seen through crossed polarisers ..................................................................................................................................................... 164

Figure 7.3: Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for frazil ice sampled on day 7 of Interice II, November 1998 ................................................................. 166

Figure 7.4: Raw and interpolated forms of the bulk salinity data for frazil ice on day 7 of the November 1998 period ............................................................................................................................. 168

Figure 7.5: Composite images of the (a) horizontal and (b) vertical sections showing the structure of the ice with the isohalines for frazil ice sampled on day 7 of the November 1998 period ............................................................................................................................. 170

Figure 7.6: Profiles of bulk salinity found in the frazil sample formed in the ice tank (solid line) and two samples of pancake ice (dotted and dashed lines) recovered from the Odden region of the Greenland Sea ........................................................................................................................................ 174
Tables

Table 3.1: A summary of the comparable and contrasting features which define brine channels and brine tubes.................................................................................................................. 40

Table 3.2: A summary of the terms commonly found in sea-ice literature which are used to describe brine-filled spaces within the ice................................................................. 41

Table 4.1: Ice growth rates in the current zone during the first 24 and 120 hours of each experimental period........................................................................................................... 78

Table 4.2: The spatial density of brine channels derived from observations of the ice surface, and the growth rate calculated at three ice thicknesses for both experimental periods................................................................. 80

Table 6.1: Standard deviation in the bulk salinity of new ice at selected depth intervals. The lowest layer (which includes the skeletal layer) corresponds to 4.2–6.3 cm for day 2 and 6.3–8.5 cm for day 3........................................................................................................... 155

Table 6.2: Nearest neighbour separation of brine channels and their spatial density at a depth of 2 cm in new ice........................................................................................................... 157
Chapter 1
Introduction

1.1 Sea ice in the polar environment

Sea ice is a dynamic and transient natural material which covers vast tracts of both polar oceans. By area, sea ice accounts for two thirds of the Earth’s permanent ice cover but its contribution to the total ice volume is only about 0.2% (Fizharris et al., 1996). Therefore, sea ice may be regarded as a thin veneer forming a physical barrier at the interface between the cold polar atmosphere and the relatively warm polar oceans. The existence of sea ice modifies the energy and mass exchanges occurring at this interface and therefore can influence, and respond to, changes in these two fluid masses. The formation of sea ice from freezing of the ocean surface results in the incorporation of salt, in the form of brine, into the ice. To a large extent, it is the continuously varying salt content of sea ice that determines its physical properties which in turn influence oceanic and atmospheric interactions. Drainage of brine from sea ice alters not only the properties of the ice but also changes the density structure of the underlying ocean. Therefore, furthering our understanding of the material properties of sea ice, particularly the distribution and transport of salt within it, is fundamental to understanding of how sea ice interacts with, and influences, elements of the polar environment.

The existence of sea ice in the polar regions substantially alters the surface properties of the ocean particularly its surface temperature and albedo. Consequently, sea ice is an important component in the energy balances which regulate regional and global climate (Maykut, 1978). Additionally, because the ice exists in dynamic equilibrium with the atmosphere and ocean, and will grow or decay depending on the
thermodynamic balance of the energy fluxes, it is one of the most responsive components of the cryosphere with respect to climate change. As such, relatively small changes in some climatic parameters may be accompanied by large changes in the ice cover. The thermal conductivity of sea ice, which influences the magnitude of the energy fluxes and the degree of thermodynamic interaction, is determined by the brine content of the ice (Maykut, 1986). Recently, the mobility of brine in sea ice has been implicated in the modification of simple conductive models of sea ice (Lytle and Ackley, 1996; McGuinness et al., 1998).

The formation and growth of sea ice is an important factor influencing oceanic thermohaline circulation (Aagaard and Carmack, 1994). With seasonal differences in surface cover of 8 million km² in the Arctic and 18 million km² in the Antarctic, the salt fluxes accompanying ice growth have a substantial effect on the vertical circulation of the ocean and the formation of deep and bottom waters which spread throughout the world's oceans. After formation, desalination of sea ice generates cold, saline water which contributes an additional salt flux to the underlying ocean, further effecting the stability of the oceanic mixed layer. Brine drainage from sea ice occurs through vertical conduits, called brine channels, which form in the very early stages of ice growth and persist until the eventual decay of the ice. Furthermore, these drainage channels not only provide a path for the transport of brine from the ice interior to the ice-water interface, they also allow the migration of nutrients, pollutants and light through the ice.

The theme of this thesis is the distribution of brine in sea ice and how it is associated to the existence of brine drainage channels. Brine distribution in sea ice is a fundamental parameter in understanding the development and physical characteristics of the ice. However, it is only the vertical distribution of brine that has received close attention whilst no detailed description of brine distribution in the horizontal plane exists, other than cursory attempts. Therefore, the aim of this thesis is to determine how brine is distributed throughout young sea ice according to the thermal, structural and crystallographic environment of the ice. The remainder of this chapter will consider the significance of young ice in the polar environments and aspects of the variability in sea ice salinity and brine distribution. The chapter will conclude with a statement of the specific thesis objectives and a description of its structure.
1.2 Young sea ice

Young sea ice is defined as being, *ice in the transition stage between nilas and first-year ice, 10–30 cm in thickness*, (WMO, 1970). As an amendment to this, the transition is taken as being from new ice to first-year ice (Armstrong *et al.*, 1973), where new ice is a generic term for sea ice which has recently formed. A description of the formation and characteristics of new ice and nilas is given in chapter 2, and an additional means of classifying young ice, based on differences in the bulk salinity, is given in section 2.6.5. For now, the essential point of the definition for the thesis content is that the research concerns the distribution of salt in thin sea ice, that which is less than 30 cm thick. It should be explicitly stated that extensively ablated sea ice which has decayed such that its thickness is less than 30 cm can not be classified as young ice.

The justification for investigating young sea ice can be found in its effects on the polar environment where it influences the large scale heat and mass balance at the surface of polar oceans significantly (Makshtas, 1984; Smith *et al.*, 1990). Areas of open water within the winter ice pack, for example Arctic leads and polynyas, are regions of rapid ice formation and a major source of new ice. Leads are linear breaks in the sea ice cover, with widths of tens to hundreds of metres. They expose the relatively warm ocean water to the cold air creating one of the steepest natural temperature gradients on the planet. Estimates of ice growth indicate that the total annual ice production associated with open leads and young ice is nearly twice that of thicker ice (Maykut, 1982). Further, it has been demonstrated, from both observation and modelling of energy exchanges in the Arctic, that the heat loss through thin ice growing in leads can be up to three orders of magnitude greater than for thicker (3–4 m) multiyear ice (Maykut, 1978; Gow *et al.*, 1990; Morison *et al.*, 1993), thereby accounting for roughly half of the total oceanic heat loss.

Likewise the salt input into the upper ocean from young sea ice production far exceeds that for thicker ice to the extent that to model the temporal changes in the oceanic mixed layer fully, it is necessary to include a young ice component (Maykut, 1982). Brine rejection during the formation and growth of young sea ice causes
significant modification of the upper ocean density structure which drives convection leading to ventilation and oceanic mixing. Therefore, the heat and salt fluxes which accompany the formation of young sea ice play an important role in the development of atmospheric and oceanic boundary layers.

Given the significance of young sea ice in determining the regional and basin wide processes in the polar oceans, being able to monitor its distribution remotely is of great value. Detection of young sea ice and determining its areal fraction constitutes a major effort in microwave satellite remote sensing observations of the polar oceans (Wensnahan et al., 1993; Tadross, 1997). The microwave signature of sea ice is linked to the microstructural properties of the ice, particularly at the upper surface (Lin et al., 1988; Winebrenner et al., 1992; Shokr and Sinha, 1994). Of particular importance, with respect to the dielectric properties of the ice, is the volume and distribution of brine found in its upper layers (Drinkwater and Crocker, 1988; Perovich and Richter-Menge, 1994). Therefore, the precise interpretation of microwave signatures of young sea ice relies on developing a clearer concept of how salt becomes distributed through young sea ice and how this distribution evolves.

1.3 Brine distribution and variability

Sea ice exists in a dynamic environment which ensures that its distribution, thickness and physical properties vary both spatially and temporally (Maykut et al., 1992). When discussing variability, it is essential to consider the scale on which it is reported. Clearly spatial scale can vary over many orders of magnitude from regional (hundreds of kilometres) down to microscale ($10^{-4}$–$10^{-6}$ m) and at each scale, the same material may be observed and described from a different perspective. A property of the ice, which appears homogeneous at one particular scale, will often reveal itself to be highly variable as the scale of observation is shifted. This phenomenon of the degree of variability being dependent on scale is particularly evident when studying brine distribution in sea ice.
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1.3.1 Observations

Vertical variability of brine distribution in sea ice is manifest in the bulk salinity profiles which are central to much of sea ice research. Although the ice volume contained in a single floe will probably have experienced the same thermal forcing conditions, it does not necessarily follow that ice core samples from that floe will exhibit identical profiles of bulk salinity. It is recognised that the variability in brine distribution is both spatial (10^2–10^9 m) and temporal (days–seasons) and has attracted attention from both direct field observations (Nakawo and Sinha, 1981; Weeks and Ackley, 1986; Eicken et al., 1991) and in efforts at modelling vertical salinity profiles (Cox and Weeks, 1988; Eicken, 1992b). This aspect of brine distribution will be discussed in more depth in section 2.5.3.

In contrast, lateral variability in brine distribution has not been the subject of such extensive investigation despite the fact that it may be indicative of horizontal brine transport processes which will ultimately determine the vertical variability observed in bulk salinity profiles. Of the studies that have been conducted on lateral variability, the scales of measurement fall into two distinct ranges; metre and submillimetre. A summary of the principal observations on each of these scales will be presented.

1.3.1.1 Metre scale

The field campaigns which have investigated lateral variability in bulk salinity have shown that in general, closely spaced samples of ice in a level, undeformed floe will yield bulk salinity profiles that are similar in form but variable in salt content. In the Arctic, early attempts to quantify the degree of variability in first year ice were made by Weeks and Lee (1962), where the minimum spacing between samples was 0.6 m, and similarly for multiyear ice by Untersteiner (1968) with a minimum sample spacing of 1.0 m. In both cases, they observe significant lateral variability in the bulk salinity and state that the degree of correlation between the bulk salinity values of closely spaced ice cores is no greater than for randomly distributed samples. Untersteiner takes this statement a step further by concluding that the salinity profile from just one core has little significance in terms of being representative of the floe as a whole.
Chapter 1: Introduction

A similar conclusion was reached by Eicken et al. (1991) who made bulk salinity measurements on three floes in the Weddell Sea, Antarctica. The floes were of level second-year ice with similar stratigraphy, the spacing between core samples ranged from 0.25 m to 20 m. Two important conclusions emerge from this work; first that there was found to be no correlation between ice crystal texture and bulk salinity, and second that the horizontal salinity variations between cores are of the same magnitude irrespective of sample spacing. At a slightly higher spatial resolution, Tucker (1984) compared values of bulk salinity from corresponding horizontal layers between cores of first-year Arctic ice which were spaced 0.38 m between centres. Differences in bulk salinity amounting to an increase of 50% were observed between corresponding depth intervals.

All the investigations conclude that the observed variance in bulk salinity is linked to the distribution of brine on spatial scales smaller than the cores. This statement is developed further by suggesting that it is the existence of brine drainage channels which generates the variability in salinity rather than it being an inherent property from the initial formation of the ice. Therefore, the horizontal variability in bulk salinity which is observed in the core samples of sea ice is controlled by desalination processes and attributed to differential brine drainage across the sample (Tucker III et al., 1984; Eicken et al., 1991).

1.3.1.2 Micro scale

Sea ice salinity and its variations are determined by the distribution of brine within the solid matrix of ice. Practically, the microscale variations in brine distribution determine the electromagnetic properties of sea ice, particularly the permittivity. Therefore the importance of understanding and specifying the microscale anisotropy of sea ice in a quantitative way has been recognised by the microwave remote sensing community (Winebrenner et al., 1992; Shokr and Sinha, 1994).

The most detailed observations of the anisotropy in the distribution of brine on the microscale have been made by Perovich (1991; and 1996). Even at this scale in relatively homogeneous young sea ice variability in the brine distribution is evident and the degree of variability tends to increase with thicker first year ice and multiyear ice.
Chapter 1: Introduction

The degree of anisotropy in horizontal thin section samples was determined from correlation lengths of the ice fabric and the size distribution of the brine inclusions. By virtue of the distribution of brine in sea ice, which is linked to the crystallographic organisation of the ice, differences in the correlation lengths exist over small spatial scales (<1 mm) even in relatively homogeneous ice.

The most significant aspect of the observations of Perovich (1996), with respect to this thesis, concerns the effect of temperature on the size and distribution of brine inclusions. In general, most of the geometric variability in the brine inclusions is temperature related and hence linked to brine volume. With increasing brine volume, the number of inclusions decreases whilst their mean size increases rapidly; this is observational and quantitative evidence for the interconnection and coalescence of brine inclusions in response to increases in the ice temperature. Allied to this, correlation lengths also increase with temperature in response to the increase in the inclusion size (Perovich and Gow, 1991).

1.4 Thesis objectives

Throughout the summary of observations presented in section 1.3, lateral variability in the distribution of brine in sea ice is a recurrent theme. It would appear that an apparently homogeneous sheet of sea ice has an inherent heterogeneity in the distribution of salt on all spatial scales. Although the lateral variability in bulk salinity is assumed to be the result of the migration of brine towards brine drainage channels, there has been no direct observation of this. A detailed study of brine distribution at a spatial resolution which is comparable to the spacing of brine drainage channels would reveal whether brine is structured according to the existence of these features. This defines the aim of the thesis - to examine the distribution of brine in young sea ice and determine to what extent it is influenced by the presence of brine drainage channels.
The specific objectives of the thesis are:

- Review the formation process and characteristics of brine drainage channels.
- Determine to what extent brine distribution in young sea ice is correlated with the occurrence of brine drainage channels.
- Investigate the redistribution of brine according to changes in the ice temperature and consequent ice porosity.
- Establish at what stage in sea ice growth the structuring of the brine distribution occurs and how this is linked to the formation of brine drainage channels.
- Contrast the brine distribution and morphology of brine structures in sea ice having different crystal textures.

1.5 Thesis structure

To examine the link between brine distribution and brine drainage channels, a basis must be established for the development of the analysis and discussion. Chapter 2 describes and reviews the pertinent physics of sea ice. This includes a discussion of the mechanism for incorporation of brine into the ice and subsequent desalination; both these processes contribute to shaping the vertical profiles of bulk salinity. Chapter 3 is a detailed discussion of brine drainage channels. In it, a scheme of classification is proposed and a nomenclature is defined. This is followed by an extensive analytical review and synthesis of the literature which describes the formation and evolution of the channels. Chapter 3 concludes by examining how brine drainage channels influence the properties and processes of sea ice.

The vast majority of the data on which this thesis is based were obtained from sea ice prepared in a tank. Chapter 4 commences with a brief review of ice tank research before presenting the methods by which the data were gathered and processed. Within this chapter, a novel technique of sample acquisition and analysis will be described which facilitates attaining the required accuracy necessary for high resolution studies of
brine distribution. The data acquired using this technique represent a new aspect on the variation of brine distribution in sea ice.

Three contrasting regimes of sea ice development will be considered in the subsequent chapters. In chapter 5, the redistribution of brine in sea ice which has undergone significant temperature cycling will be investigated. Two cases are presented, melt and refreeze, during which changes in the air temperature provide the required thermal forcing. The characteristics and evolution of the brine distribution are described and interpreted according to the occurrence of brine channels and the thermally induced changes in the porosity of the ice. Mechanisms for brine redistribution are proposed. In addition to the analysis of the observations, a discussion of the methods is presented with reference to data from exemplary samples.

The evolution of the brine distribution in new ice, and the associated development of brine channels, form the subject of chapter 6. Ice forming from a homogeneous body of water undergoes substantial transition in terms of the salt content and its distribution. Further, during the initial development of the ice cover, brine channels are observed to form in the ice matrix. From a time series of samples, an association is established between the redistribution of brine and the changing morphology of the brine channels. Prompted by these observations, recently published interpretations of the evolving brine content of sea ice are questioned and re-examined.

Chapter 7 concentrates on brine distribution in frazil ice. This granular ice type provides a textural contrast which complements the preceding discussions on columnar ice. From the homogeneous nature of the brine distribution, and the structural form of the ice, mechanisms of brine migration are suggested which may account for the distribution observed. Further, the processes that control the formation of brine structures in sea ice with differing crystal texture are discussed. Finally, in chapter 8, the results of the thesis are summarised and the principal conclusions are stated with potential directions for future research.
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Chapter 2
Sea ice

2.1 Introduction

Before commencing a discussion of the spatial distribution of brine in young sea ice, it is necessary to introduce some basic concepts. The object of this chapter is to present a comprehensive description of sea ice, including its formation, growth and temporal evolution, which will then be elaborated in subsequent chapters. Particular emphasis is placed upon the phase relations between the solid ice and the liquid brine, the morphology of the skeletal layer, brine incorporation and the desalination mechanisms which modify the salt content of sea ice. A more extensive exposition of the material which follows may be found in the many excellent reviews of sea ice (Maykut, 1985; Weeks and Ackley, 1986; Gow and Tucker III, 1990; Ackley, 1996; Leppäranta, 1998).

2.2 Fundamentals

Sea ice is a porous solid. It is composed of a pure ice matrix with inclusions of brine and gas. The ice and brine exist in dynamic phase equilibrium making the microstructure of sea ice sensitive to small changes in temperature. Variations in the structural and phase state of sea ice results in changes in its physical properties. In this section, the solid and liquid components of the sea-ice system, and the equilibrium between them, will be described.
2.2.1 Ice

The most prevalent form of ice found on earth is ice I(h). Although a variety of more exotic polymorphs are known to form under extremes of temperature and pressure, attention will focus on the I(h) type exclusively. The structure of ice I(h), shown in Figure 2.1, illustrates the basic arrangement of the crystal lattice with the oxygen atoms of the H₂O molecule located at the apices of a tetrahedron giving the structure a hexagonal symmetry. The regular lattice of oxygen atoms is held together with highly directional hydrogen bonds resulting in an open structure of low density. This is an unusual characteristic for solids but one which has important geophysical implications.

![Figure 2.1: The crystal lattice of ice I(h) showing the hexagonal symmetry of the structure and the crystal axes (Weeks and Ackley, 1986).](image)

An important feature of the crystal structure is that the oxygen atoms are located close to a series of parallel planes. These hexagonally symmetric planes are referred to as the basal planes with the perpendicular vector defined as the c-axis or optic axis. The c-axis is fundamental for describing the crystallography of sea ice as it is the reference
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axis for defining the crystal orientation. An important property of the basal planes is that they are the planes of easiest cleavage, that is they are held together by fewer bonds than any other crystal plane. Three equivalent a-axes, which lie perpendicular to the c-axis, define the basal plane, and it is along these axes that the dendritic arms of growing sea ice crystals and snowflakes form.

An important aspect of ice crystal growth is that the growth rate along the a-axes is as much as two orders of magnitude greater than along the c-axis (Hallett, 1960). This anisotropy in the kinetics of crystal growth gives rise to the formation of discs of ice, rather than spheres, with the c-axis normal to the plane of the disc. A final point to note is that the ice can be considered a pure phase. The substitution of the major sea salt ions into the crystal lattice of the ice is negligible with partition coefficients estimated at less than $10^{-4}$. A full discussion of crystal growth phenomena may be found in Hobbs (1974) and Wettlaufer (1998).

### 2.2.2 Phase relations

Neglecting the gas content, sea ice is composed of pure ice, brine and solid salts. The equilibrium phase relations between these components allow the relative amount of each constituent to be computed for a given temperature and bulk composition. As a consequence, the salinity of brine which resides within the sea ice is independent of the salinity of the seawater from which it formed, but depends on the temperature of the brine alone. When a pocket of brine is in thermal equilibrium with its surroundings it must be at the salinity determined freezing point of the solution ($T_f$). As the temperature of sea ice varies, phase equilibrium is attained by mass changes which take place at the walls of the brine pocket. If the temperature of the brine ($T_b$) is greater than $T_f$, melting will occur at the walls of the pocket, diluting the brine and reducing its salinity until equilibrium is restored when $T_b = T_f$ and vice versa.

Seawater is a complex solution of dissolved salts and gases and though the concentration of salt may vary, the ratios of the constituent ions relative to each other may be regarded as constant. Further, as the salinity of the brine in sea ice changes, the ratio of the major salt ions in the overall sea-ice system remains the same as the ratio in
the seawater from which it formed. Enrichment of the minor anions in brine has been recorded (Meese, 1989) though Figure 2.2 shows their contribution to the overall salt content to be negligible. As the ice cools, solid salts begin to precipitate from the brine solution, causing the ice to assume a milky appearance. The two most abundant salts in seawater are mirabilite (Na₂SO₄.10H₂O) and hydrohalite (NaCl.2H₂O) which precipitate from the brine at temperatures of −8.2°C and −22.9°C respectively. Precipitation of salts alters the concentration of the various ions in the brine and thus the freezing point of the solution.

**Figure 2.2:** Equilibrium phase diagram for sea ice (Weeks and Ackley, 1986).

Phase relationships between the ice, brine and solid salts are governed by chemical thermodynamics which can be represented more simply by the phase diagram in Figure 2.2, which was constructed by Assur (1958) and later verified by the measurements of Richardson and Keller (1966). It can be seen from Figure 2.2 that the phase behaviour of sea ice is dominated by its sodium chloride content. The phase diagram, with knowledge of the ice temperature, is often sufficient to specify the sea-ice system according to its constituents when they are at equilibrium.
2.3 Formation of sea ice

The freezing of the polar oceans is controlled by the salinity of the water in two ways. First, the salt content of the water depresses the freezing point ($T_{fsw}$) of the water at the ocean surface. Without the need to consider pressure effects, the freezing point can be expressed according to (Neumann and Pierson, 1966):

$$T_{fsw} = -0.003 - 0.0527S_w - 0.00004S_w^2$$

(2.1)

which can be approximated by (Maykut, 1985):

$$T_{fsw} \approx -0.055S_w$$

(2.2)

Here, $S_w$ is the salinity of the water in parts per thousand (‰). In conformity with convention, salinity values will be expressed according to the current accepted international standard, the practical salinity unit (psu) (UNESCO, 1981). For the purposes of this thesis the two units, ‰ and psu, may be regarded as equivalent. The second effect on freezing is that salt reduces the temperature of maximum density ($T_m$) but at a more rapid rate than $T_f$, see Figure 2.3, such that for seawater with salinities greater than 24.7 psu, the temperature of maximum density is lower than the freezing point. This defines the transition between brackish and true seawater.

Atmospheric cooling of the ocean increases the density of the surface layer creating a vertical density instability and convective overturning which forces deeper, warmer water to the surface. The ocean surface will not normally freeze until the entire underlying column of water that is undergoing convection reaches its salinity determined freezing point. However, the stable density structure of the ocean limits the depth to which cooling must occur prior to the onset of surface freezing and this is typically 10–40 m. Additional heat loss from the surface then produces a slight supercooling of the water and ice formation. The degree of supercooling required to initiate ice formation is dependent on the availability of freezing nuclei (Katsaros and Liu, 1974) but
is not usually more than a few tenths of a degree Celsius because of the abundance of snow crystals and biogenic material in the water (Weeks and Ackley, 1986). With sufficient cooling, a skim of ice crystals forms consisting of small disks, needles and platelets of ice called frazil.

![Temperature of maximum density (Tm) and freezing point (Tf) of water as a function of its salinity.](image)

**Figure 2.3:** The temperature of maximum density ($T_m$) and the freezing point ($T_f$) of water as a function of its salinity.

### 2.3.1 Young ice types

There are many forms of sea ice which can develop from the great variety of dynamic, thermodynamic and oceanographic conditions that exist. Each type of ice has a characteristic crystal texture from which the nature and formation history of the ice may be derived because a change in the texture signifies a change in the growth mechanism (Lange and Eicken, 1991). As a result, there can be great variability in the textural characteristics and properties of sea ice which grows in the same location but under significantly different environmental conditions (Gow et al., 1990). For a clear presentation and description of the nomenclature of crystal types and their genesis see Eicken and Lange (1989) and Tison et al. (1998, pp. 376-381).
Chapter 2: Sea ice

The continued production of frazil ice results in a soupy, unconsolidated suspension of crystals called grease ice - which comes under the general description of new ice. Capillary waves are damped by the grease ice giving the water surface the appearance of a smooth, oily slick (Martin and Kauffman, 1981; Wadhams and Holt, 1991). Under quiescent conditions, the frazil crystals will tend to coalesce and freeze together forming a continuous, solid sheet of ice 1–10 cm thick; this stage of ice growth is referred to as nilas, which is typical of ice growing in the relatively sheltered open water areas of the Central Arctic. Nilas is initially slightly transparent and referred to as dark nilas, with a thickness between 1–5 cm; it then develops into the thicker, more opaque white nilas with thickness ranging between 5–10 cm. Frost flowers are transient features which will often develop on the surface of nilas (Drinkwater and Crocker, 1988) and can be highly saline with bulk salinities of the order 100 psu (Martin, 1979; Perovich and Richter-Menge, 1994).

Under turbulent conditions, wind and wave action will inhibit the consolidation of a solid ice cover and frazil formation is rapid (Ushio and Wakatsuchi, 1993; Brandon and Wadhams, 1996). Under sustained wave action, the thick suspension of frazil crystals will then tend to form pancake ice with a granular crystal texture. Individual pancakes are a roughly circular mass of ice in varying states of consolidation with diameters ranging from 20 cm up to 3 m. The continuous colliding together of the pancakes splashes frazil crystals onto their upper surface to create a raised perimeter. As the wave energy attenuates, the individual pancakes will consolidate and nilas may grow between them forming a continuous solid sheet. Areas of high frazil production include the turbulent waters around Antarctica (Wadhams et al., 1987; Lange et al., 1989) and the Arctic Marginal Ice Zones (Wadhams et al., 1996).

Once the ice has consolidated and a continuous sheet of sea ice has formed, the relatively warm ocean becomes insulated from the cold polar air; further, the ocean is protected from wind-generated turbulence. Under these quiet conditions, continued ice growth occurs thermodynamically by congelation growth at the ice-water interface forming vertically elongated columnar crystals. In areas of pancake formation, ice thickness can increase rapidly by the rafting of pancakes onto one another (Lange et al., 1989). Arctic sea ice has been found to be composed predominantly of congelation ice
with a columnar crystal texture (Weeks and Ackley, 1986; Gow et al., 1987b; Eicken et al., 1995) whilst in Antarctic sea ice, columnar and granular textures occur in varying amounts (Gow et al., 1987a; Lange and Eicken, 1991; Worby et al., 1998).

2.3.2 Crystal orientation

At the upper surface of the ice sheet, where the original layer of grease ice formed, the crystal orientation is random (Weeks and Wettlaufer, 1996), though in the first skim, crystals with a vertical c-axis prevail. Once a continuous solid sheet has formed, the ice then develops a highly oriented crystal arrangement by a process called geometric selection. Crystals which are oriented favourably with respect to the growth direction will eliminate crystals with an unfavoured orientation due to the anisotropic kinetics of crystal growth, noted in section 2.2.1. Preferential growth occurs when the c-axis of the crystal is oriented parallel to the ice-water interface such that growth occurs along the basal planes, parallel to the direction of heat flow. There are two reasons why growth parallel to the basal planes is favoured. First, there is no kinetic barrier to limit growth in this direction and second, the thermal conductivity of the ice/brine composite is greater perpendicular to the c-axis due to the fact that brine has a lower thermal conductivity than pure ice. Crystals which are oriented in this manner will grow slightly faster, wedging out the more slowly growing crystals with unfavourable orientation. The effects of crystal orientation on geometric selection have been clearly demonstrated in a series of controlled experiments using large single crystals (Kawamura, 1987).

The continuation of geometric selection culminates in the growing interface becoming almost completely dominated by crystals with a horizontal c-axis. Further, there is a tendency for crystal size to increase with depth producing strong vertical elongation of the crystals parallel to the direction of heat flow (Weeks and Ackley, 1986). Under certain conditions a further degree of orientation may develop within the columnar zone. In addition to horizontal orientation, the c-axes can also adopt an azimuthal alignment (Gow et al., 1990; Weeks and Wettlaufer, 1996). This mode of orientation is attributed to the movement of water at the ice-water interface by shear currents (Langhorne, 1982; Langhorne, 1983; Langhorne and Robinson, 1986). Crystals with their c-axis parallel to the direction of flow will experience a greater relative growth
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rate caused by enhanced solute diffusion induced by turbulence. Crystal orientation is known to influence the properties of sea ice and further alignment of the crystals generates anisotropic behaviour in the physical and electromagnetic properties of the ice (Kovacs and Morey, 1978; Morey et al., 1984; Reisemann, 1998).

The initial stages of sea ice growth and its evolution have been presented. To summarise, the slush of frazil ice which forms under dynamic conditions at the ocean surface consolidates to form a solid ice cover. This layer of granular ice may then increase in thickness either by rafting or by congelation growth at the ice-water interface. In the latter case, there will be a transition from sea ice with a granular texture to one with columnar texture during which the c-axis of the ice crystals become aligned in the horizontal direction, parallel to the ice-water interface.

2.4 Brine incorporation

The advancing interface of growing sea ice consists of a delicate layer of ice platelets, sometimes referred to as ice lamellae. Each crystal, or grain, is composed of a series of these pure ice platelets, which are arranged parallel to each other and which lie perpendicularly to the c-axis. This layer is called the skeletal layer and its dendritic structure determines the mechanism for the incorporation of significant amounts of brine into the ice. Without the entrapment of brine, there would be little to distinguish sea ice from freshwater ice and the notable absence of a skeletal layer at the growing interface of freshwater ice suggests that its existence in sea ice is associated with the presence of salts in seawater.

2.4.1 Constitutional supercooling

The development of a dendritic interface is explained by a condition referred to as constitutional supercooling which was originally described by Tiller et al. (1953) and later applied to sea ice by Harrison and Tiller (1963). Constitutional supercooling in the boundary layer adjacent to the growing interface is caused by the different rates of
diffusion for heat and dissolved salts. When seawater freezes, salt is rejected into the water ahead of the advancing interface creating a highly saline boundary layer. Therefore, the layer adjacent to the growing interface, being at its salinity determined freezing temperature, is colder than the seawater reservoir below. As the diffusion of heat to the interface is more rapid than the diffusion of salt through the liquid, the temperature profile in the boundary layer is linear whilst the concentration profile has an exponential form. This mismatch in profiles causes the brine within the boundary layer to be at a temperature below its salinity determined freezing temperature, as illustrated in Figure 2.4. A more complex situation exists in reality due to convection currents at the interface and a more comprehensive discussion of this phenomenon can be found in Wettlaufer (1998).

![Figure 2.4](image)

**Figure 2.4:** *A schematic representation of the mismatch in temperature profiles which gives rise to constitutional supercooling at the ice-water interface (Maykut, 1985).*

The consequence of constitutional supercooling is that any slight protrusion on the interface will be bathed in supercooled brine and thus will tend to grow more quickly than at a planar interface. A dendritic interface then develops which is composed of an array of elongated platelets projecting into the supercooled liquid and separated by grooves of enriched brine, the basic structure of which is presented in Figure 2.5. It is at this delicate ice-water interface that brine is systematically incorporated into the ice to create a laminate of pure ice platelets separated by pockets of brine.
Figure 2.5: A simplification of the skeletal layer showing the platelet and groove structure of two columnar crystals. The large arrow indicates the direction of ice growth (Gow and Tucker III, 1990).

2.4.2 Skeletal layer

Sea ice structure is characterised not only by the columnar crystal texture which forms by congelation growth but also by series of liquid inclusions located at the platelet and grain boundaries. The first observations that sea ice is composed of pure ice and pockets of trapped brine were made by Buchanan (1887) and was further investigated for Arctic sea ice by Malmgren (1927). Brine incorporation occurs within the dendritic skeletal layer, between the ice platelets which extend down into the water. At the tips of the platelets, brine can be transported away from the interface, but higher in the groove formed between the platelets it becomes trapped. As the platelets lengthen they also widen and periodic lateral connections are made between them trapping pockets of residual brine. These isolated brine inclusions are generally elliptical and have dimensions of tenths of millimetres, although the precise geometry of inclusions varies with crystal type (Eicken and Lange, 1989).

The thickness of the skeletal layer in growing sea ice is about 1–3 cm whilst the spacing of the platelets is typically about 0.4–1 mm. In common with many of the properties of sea ice microstructure, the platelet width is not constant but is a function of the ice growth rate (Nakawo and Sinha, 1984). At faster growth rates the plate spacing becomes narrower, enabling more brine to be trapped and thus the ice becomes saltier. As the ice thickens, the growth rate decreases and the plate spacing increases, reducing
the volume of brine which is incorporated and therefore the ice is less saline. Any subsequent changes in the ice substructure occur mainly in response to temperature changes in the ice (Perovich and Gow, 1991; Perovich and Gow, 1996). During periods of prolonged warming the shapes of the inclusions will change. The pockets tend to elongate along the linear boundaries between platelets where they ultimately coalesce forming brine layers. This evolution and interconnection of the inclusions leads to significant changes in the geometry and volume of the liquid phase.

2.5 Bulk properties

Much of our knowledge of sea ice, its behaviour and physical evolution has been derived from extensive investigation into its bulk properties. Ideally, a bulk quantity will give a representative measure of a particular parameter by averaging over large sample volumes. In most cases these samples are ice cores drilled in representative floes. Three of the primary sea ice parameters are considered here.

2.5.1 Bulk salinity

Of all sea ice parameters, bulk salinity is one of the most frequently measured and its influence on sea-ice and oceanographic properties and processes is manifold. The bulk salinity of sea ice \((S_i)\) is essentially the volume averaged salt content of the ice and can be defined as:

\[
S_i = \frac{\text{mass of salt}}{\text{mass of ice} + \text{mass of brine}} \cdot 10^3
\]  

(2.3)

The process of salt entrapment, described in section 2.4, determines the initial bulk salinity of the ice. In simplistic terms the mass of salt trapped in the solid ice matrix depends on both the salinity of the seawater and the rate of ice growth. Brine entrapment
can be regarded as a segregation process where the initial bulk salinity is expressed in terms of an effective segregation or distribution coefficient, $K_{\text{eff}}$.

$$S_i = K_{\text{eff}} \cdot S_w$$ (2.4)

Quite clearly, the bulk salinity of the ice is proportional to the salinity of the seawater. Further, the value of $K_{\text{eff}}$ is found to be a function of the ice growth rate and the thickness of the interfacial boundary layer (Weeks and Ackley, 1986; Eicken, 1998). It is a non-trivial problem to determine the absolute value of $K_{\text{eff}}$, but quantitative expressions have been formulated which are dependent on growth rate alone (Cox and Weeks, 1975).

### 2.5.2 Porosity

Sea ice, being a solid matrix interspersed with inclusions of brine and some gases, has an inherent porosity. The degree of porosity is changeable as the ice/brine composite maintains phase equilibrium in a continuously variable temperature field. Total porosity $(V_r)$ can be expressed as simply the sum of brine volume $(V_b)$ and gas volume $(V_g)$ normalised by the sample volume $(V)$:

$$\frac{V_r}{V} = \frac{V_b}{V} + \frac{V_g}{V}$$ (2.5)

It is possible to determine the porosity of sea ice at any particular moment from measurements of bulk salinity, temperature and density. From these variables, the relative fractions of brine and gas volumes can be calculated. For a first approximation, gas volume is ignored and porosity is equated with brine volume which can be calculated from simple numerical relations which are independent of ice density (Frankenstein and Garner, 1967). A more sophisticated approach (Cox and Weeks, 1983) includes the sea ice density as a means of determining the contribution of gas volume to the total porosity of the ice and which can be applied to any solute system (Häusler, 1989). For sea ice
that is warm (> -2°C) and fresh, for example multiyear ice in summer or sea ice forming in the brackish Baltic Sea, a slightly modified set of expressions need to be used which converge correctly towards the freshwater boundary condition (Leppäranta and Manninen, 1988).

In young sea ice, brine volume is the dominant contribution to porosity (Weeks and Ackley, 1986; Cox and Weeks, 1988). Typical brine volumes in the skeletal layer will vary between 8 and 40% (Maykut, 1985) whilst in the majority of the ice profile the gas volume is less than 2% (Matsuo and Miyake, 1966; Nakawo, 1983).

2.5.3 Salinity profiles

If the bulk salinity of sea ice were determined simply by brine entrapment according to equation 2.4 then calculating vertical salinity profiles would be possible from knowledge of the growth rate alone. In reality, a more complex situation exists because salinity profiles are highly time dependent, suggesting vertical movement of brine. If brine did not move one would expect to observe static salinity profiles in sea ice with the bulk salinity having a maximum value at the surface decreasing with depth as the growth rate of the ice decreases. Because vertical salinity profiles reflect the thermal and dynamic history of the ice, interpreting them can be a complex problem (Eicken, 1992b). However, there are some well defined trends which can be identified in the salinity profiles of first-year sea ice.

When sea ice forms from seawater which has a salinity of about 34 psu, its initial bulk salinity will be typically 10–15 psu decreasing to 4–6 psu by the start of the spring melt (Weeks, 1994). In addition to temporal variations in the total bulk salinity of the sea ice cover, the salinity profiles of first-year ice display consistent characteristics. During growth, young ice exhibits a well defined C-shaped salinity distribution with high bulk salinities at the top and bottom of the ice. The high bulk salinity in the upper portion is primarily attributed to the rapid initial growth rate and to brine expulsion towards the surface. In the lower part of the ice, the occurrence of high salinities results from both the retention of brine in the skeletal layer by capillary effects and the downward migration of brine through the ice. Observations of very high bulk salinities
which have been made in the skeletal layer of growing sea ice (Nakawo and Sinha, 1981; Gow et al., 1990) may represent a transient situation only, as much of this brine will not be incorporated into the ice.

Further desalination reduces the mean bulk salinity of the ice though the C-shape still persists, albeit less pronounced. As the ice thickens, the salinity of the upper layers gradually decreases, indicating vertical migration of the brine towards the ice-water interface which retains elevated bulk salinities. With the onset of melting, the mean bulk salinity decreases quickly with the salinity of the upper layers approaching zero. The most comprehensive record of the characteristics and evolution of bulk salinity profiles in first-year ice was compiled from field measurements by Nakawo and Sinha (1981). Their results show the modification of the C-shaped profile by thermal forcing and changes in the ice growth rate. Comparisons between the salinity profiles of samples of ice which were of either predominantly congelation ice or predominantly frazil ice show no systematic differences in the salinity profiles. It is speculated that sea ice of differing crystal textures will have similar fractionation and desalination mechanisms (Eicken, 1992b). The results from this thesis indicate that this assumption may not be valid.

2.6 Desalination

The phenomenon of the desalination of sea ice has been known since the last century (Nansen, 1897) and early Arctic explorers were aware that the surface layers of multiyear floes were potable when melted. After initial formation, when the bulk salinity of the ice may be as high as 20 psu, salt is continuously lost from the ice to the ocean, a concept originally described by Malmgren (1927). Desalination is an important process with respect to the associated changes in the bulk salinity and hence changes in the physical properties of sea ice, for example in determining its mechanical, electromagnetic and thermal properties. Further, the flux of brine to the upper ocean resulting from desalination is an important component of ice-ocean interaction; it modifies the local physical oceanography of the underlying seawater which can induce thermohaline convection (Carmack, 1986).
Immediately after brine becomes incorporated in the ice it begins to drain out by a number of mechanisms thus modifying the initial salinity profile. An early attempt to quantify the process of desalination was made by Untersteiner (1968) in which the basic mechanisms that had been proposed for desalination were presented and assessed. Four mechanisms are commonly regarded as being potentially responsible for desalination: (1) brine pocket migration, (2) brine expulsion, (3) gravity drainage and (4) flushing. It is likely that each of these mechanisms will have a role in the transport of brine through sea ice causing its salinity profile to vary continuously. However, the relative contribution that each mechanism makes to desalination will alter according to the thickness, porosity and temperature of the ice. This aspect will be commented on for each.

2.6.1 Brine pocket migration

Historically, brine pocket migration was the first desalination mechanism to receive attention (Whitman, 1926). As a consequence of the temperature gradient in sea ice, brine pockets tend to migrate in the direction of higher temperatures. Any brine pocket in the ice will have a cold end and a warm end which, via the phase relation, produces a difference in brine salinity across the pocket. Salt then diffuses through the pocket to minimise the concentration gradient and restore equilibrium. Associated with the diffusion of salt and the necessity for phase equilibrium, freezing occurs at the cold end of the pocket as the salinity decreases whilst ice melts at the warm end as the brine concentration increases. The net effect is for the pocket to migrate towards the warm side of the ice, that is downwards to the unfrozen sea.

For typical temperature gradients found in sea ice, this mechanism has a negligible effect on desalination and the migration rates derived from both observation and theoretical formulations are too slow to account for the observed rates of desalination (Kingery and Goodnow, 1963; Harrison, 1965; Hoekstra et al., 1965; Untersteiner, 1968; Uusitalo, 1983). In small pockets, the rate of migration is limited by the rate of salt diffusion; for larger pockets (diameter greater than 1 mm) migration may be significantly faster as the effective diffusivity of salt can be enhanced by convection processes within the pocket.
2.6.2 Brine expulsion

The phenomenon of expulsion, due to volumetric changes accompanying ice formation, was originally described by Nakaya (1956) though it was Bennington (1963) who first observed it in sea ice and proposed it as a desalination mechanism. As a pocket of brine begins to cool, phase equilibrium is maintained by freezing on the interior of the pocket thus concentrating the brine within it. Ice occupies a greater volume (approximately 10%) than the water from which it formed resulting in an increase of pressure and internal stress within the pocket. Boundaries between the grains and platelets act as natural lines of weakness in the ice structure and the pressure in the brine pockets is released by failure of the ice along these crystallographic planes allowing the brine to escape into a region of lower pressure. The mechanics behind the process, in terms of the internal stresses created within a pocket during phase changes, have been investigated (McKittrick and Brown, 1996) and it was concluded that brine expulsion is a physically sound model for desalination.

It is possible that the mechanism utilises liquid connections that are known to exist at the grain junctions of freshwater ice (Nye, 1989) providing a natural microscopic permeable network. By this process, it is possible to provide hydrostatic communication within the ice matrix and make expulsion a realistic means of desalination. However, from estimates of desalination derived from numerical models of the process (Untersteiner, 1968; Cox and Weeks, 1975), it has been concluded that whilst expulsion may be a significant mechanism in thin sea ice (< 25 cm thick), where it contributes to the formation of a high salinity surface layer (Perovich and Richter-Menge, 1994), it plays only a minor role in the desalination of thicker ice.

2.6.3 Gravity drainage

Kingery and Goodnow (1963) determined that of all the possible desalination mechanisms, gravity drainage of brine through sea ice predominates. Support for gravity drainage as the primary desalination mechanism also came from the experiments and observations of Cox and Weeks (1975) and Eide and Martin (1975) although it has been
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Chapter 2: Sea ice

suggested that it is limited to the case of warm porous ice only (Gow et al., 1990). Gravity influences desalination in two circumstances. First, as an ice sheet thickens it rises above sea level to maintain isostatic equilibrium thus increasing the freeboard. This generates a hydrostatic head which can drive the mobile brine out of the ice through an interconnected brine network (Eide and Martin, 1975). This process, combined with flushing, probably accounts for the very low salinities found in the upper layers of thick multiyear ice (Cox and Weeks, 1974).

The second gravity driven process results from the unstable vertical density distribution of the brine which develops by virtue of the temperature gradient in a growing ice sheet and the requirement for phase equilibrium. It is the density gradient between the cold brine high in the ice and the warmer, less dense brine lower down which provides the potential energy to drive convective overturning of the interstitial brine and fluid exchange with the underlying seawater; a process sometimes called buoyancy driven convection. In experiments by Cox and Weeks (1975), using NaCl ice doped with the radioactive isotope $^{22}$Na, an in situ study of desalination determined some key aspects of gravity drainage. First, the rate of brine drainage increases with an increase in the temperature gradient as it is this which determines the density gradient in the brine. Second, gravity drainage is a function of brine volume, and therefore permeability, and is inhibited when the brine volume falls below 5%.

Desalination via gravity drainage is mediated by the permeable brine channels which penetrate vertically through the ice and are open at the ice-water interface. A thorough discourse on brine drainage channels will be presented in chapter 3 and so for now the discussion here will be limited to the hydrodynamics which influence the desalination process. The brine flow in the channels is usually oscillatory (Eide and Martin, 1975), though simultaneous bi-directional flow has been observed in sloping channels (Niedrauer and Martin, 1979). The period of oscillation observed in ice which was 10 cm thick was 1 hour, comprising an 8–15 minute inflow and a 45 minute outflow. A comparable oscillation was implied from measurements of brine and nutrient concentrations in the skeletal layer of 28 cm thick Antarctic lead ice (Melnikov, 1995), though the interpretation is less conclusive.
A theoretical framework for explaining the observed brine oscillations has been developed by Martin (1970; 1974). The physical basis of the oscillatory motion lies with the density difference which exists between the brine in the ice and the seawater. This creates a positive buoyancy force which acts to oppose the pressure gradient of the more dense brine in the ice. A sensitive balance between these two forces is controlled by the diameter of the channel at the ice-water interface, which constricts to form a neck (Eide and Martin, 1975). This narrowing of the channel is fundamental to the hydrodynamic arguments which allow the oscillatory motion of brine. Two states of hydrostatic equilibrium exist, depending on whether the channel contains cold, saline brine or warmer, fresher seawater. Any small mass perturbations will disturb this balance and brine exchange is initiated as the system tries to re-establish hydrostatic stability.

In the field, gravity drainage has been observed by divers under Antarctic leads (Melnikov, 1995) and it has also been interpreted from variations in the salinity of seawater under very porous ice (Hudier et al., 1995). Enhanced gravity drainage is predicted from mathematical models of shear flow under ice keels (Feltham, 1997), again a process which is found to be dependent on ice porosity and density gradients. In near isothermal sea ice, the effectiveness of gravity drainage wanes as the density gradient of the brine decreases impeding convective overturning within the ice, a situation which has been observed in field studies of ice during summer months (Hudier and Ingram, 1994; Eicken et al., 1995).

Although a complete numerical model of gravity drainage does not exist, experimental data have provided useful quantitative estimates for the rate of desalination (Cox and Weeks, 1975). The expressions developed for gravity drainage and brine expulsion have been used in coupled thermodynamic-desalination models to simulate salinity profiles of first-year ice which agree well with field data (Cox and Weeks, 1988; Cox, 1990; Eicken, 1992b). However, the expression for gravity drainage is empirical and takes no account of the physical analysis of the drainage process investigated by Martin (1974).
2.6.4 Flushing

The final mechanism to be considered is flushing by melt water. It is essentially a specific form of gravity drainage and dominates desalination during the melt period. During late spring and summer when air and ice temperatures begin to increase, a hydrostatic head of freshwater forms from the melting of surface ice and snow. Brine volume, and therefore the permeability of the warm sea ice, increases with the elevated ice temperatures (Saeki et al., 1986). Thus, if all levels of the ice are permeable then meltwater will percolate through the interconnected network of channels, driven by the hydrostatic pressure, resulting in significant desalination (Nakawo and Sinha, 1981; Wadhams and Martin, 1990; Hudier et al., 1995). The input of meltwater at the surface of the ice, and the negative temperature gradient through it, causes stable stratification of the brine in summer sea ice (Eicken et al., 1995). Untersteiner (1968) formulated a simple model of desalination and estimated that the greatest loss of salt occurs during this process of vigorous flushing, which he also recognised to be a function of brine volume. Like Cox and Weeks (1975), Untersteiner derived a cutoff point for brine movement when the ice becomes impermeable at brine volumes less than 5%. However, for the purposes of flushing, it is implicit that the ice is warm, melting and therefore permeable.

2.6.5 Defining young sea ice by desalination

Further to the definition of young sea ice given in section 1.2, a distinction can be made between young ice and first-year ice based upon observations of bulk salinity and an understanding of desalination mechanisms. A study comparing the average bulk salinity of Arctic first-year ice with ice thickness was made by Cox and Weeks (1974). The results of this comparison, reproduced in Figure 2.6, show a number of important properties of the data set.
First, although the data were obtained from a wide variety of sources, they show very minimal scatter indicating consistency between observations. Second, it is evident that a rapid decrease in the total salt content of the ice occurs in the early stages of growth. At an ice thickness of about 30 cm, there is a break in the slope which implies a decrease in the rate of change in salt content with thickness. The reasons for the slope break are not completely understood but it is possible that it represents a shift in the dominant desalination mechanism from brine expulsion to gravity drainage. Another explanation may be found in the salt segregation, or fractionation, occurring at the growing interface, section 2.5.1, which is itself a function of growth rate and so by implication ice thickness. Whatever the absolute reasons, the slope break at a thickness of 30 cm provides a physical basis for differentiating between young ice and first-year ice. The change in the desalination rate has also been reported in more recent field observations of young ice in a newly formed lead (Gow et al., 1990); rapid desalination in the early days of young ice growth was followed by less rapid desalination with the transition occurring at a thickness of approximately 35 cm.
2.7 Summary

This chapter has presented the basic concepts and processes necessary for the discussion of brine distribution in sea ice and how it is influenced by the existence of brine drainage channels. Although it started at the most fundamental level with respect to the crystal structure of ice, it was seen that many of the subsequent processes and properties were linked to the anisotropy of the ice crystal. Likewise, the phase relationship that exists between ice, brine and solid salts occupies much of the later evolution and structuring that occurs in sea ice.

The central theme of this chapter has been the fate of brine in sea ice from its initial partitioning from the seawater, the entrapment in the skeletal layer and its distribution through the ice culminating in desalination, principally via brine drainage channels. It has been emphasised that bulk salinity, and therefore brine volume, controls many of the properties of sea ice. It is the distribution of brine and the evolution of that distribution as a function of structure and temperature, which will form the core of this thesis.
Chapter 3
Brine channels: structure, formation and processes

3.1 Introduction

This chapter describes the structural characteristics of brine drainage channels and how they influence the properties and processes of sea ice. Although a great variety of different brine structures exist in sea ice, this thesis is concerned with the brine drainage channels which are ubiquitous in sea ice of all types and ages. To describe brine channels adequately, they must first be visualised physically; this is the rationale of section 3.2. Given that a great variety of brine structures can be identified in sea ice, this leads to potential problems with their nomenclature. Therefore, the characteristics of the brine structures that will be discussed later are described and a nomenclature defined, concluding with a systematic scheme of classification. On that foundation, section 3.4 presents a complete synthesis of the formation and evolution of brine drainage channels. This provides a full discussion of their formation in which previously unrelated observations are linked to assemble a physical chronology of brine channel evolution. The chapter concludes with a review of five properties of sea ice in which brine channels have a significant controlling influence.
3.2 Visualising brine structures

The internal structure of sea ice is a complex, evolving collection of air- and brine-filled spaces. These range from isolated, spherical pockets of brine less than a millimetre in diameter to large, anastomosing, or interconnected, structures which may extend many tens of centimetres through the ice. To understand fully how these structures form, and how they then influence the properties and processes in sea ice, it is essential to have a clear physical concept of their nature and morphology. In this section, various techniques for obtaining images of brine structures in sea ice are described.

3.2.1 Two-dimensional imaging

Traditionally, the microstructure of sea ice has been studied using thin section techniques. An advantage of thin sections is that the crystal structure of the ice and the distribution of brine can be viewed simultaneously at very high resolution. However, the method is destructive, restricted to two dimensions and to small sample sizes, often only a few square centimetres. Early examples of this technique revealed the brine pockets distributed between the ice platelets (Tabata and Ono, 1957; Weeks and Assur, 1967; Sinha, 1977) whilst later work (Perovich and Hirai, 1988; Perovich and Gow, 1991; Eicken, 1993; Perovich and Gow, 1996; Cole and Shapiro, 1998) incorporated the use of image processing systems to generate quantitative descriptions of the brine inclusions.

Thick sections of sea ice enable larger samples of ice to be examined but at the expense of crystallographic detail. Horizontal sections of sea ice have been used to study the distribution of brine drainage channels over areas of many hundreds of square centimetres (Wakatsuchi and Saito, 1985). The ultimate extension of the thick section technique was achieved by Cole and Shapiro (1998) during field campaigns in the Arctic. They were able to prepare 7 cm thick plates of ice which extended through the entire thickness of the floes. From these plates, an extensive picture of brine drainage channels could be observed, including information on their spacing, size, symmetry, continuity and a variety of orientations and geometry in channels which have not previously been described. This novel method of sample acquisition has provided valuable information on the nature of brine channels forming under natural conditions.
Vertical, two-dimensional views of brine structures have also been viewed in situ within saline ice grown in narrow Perspex tanks. The first attempt at this was by Bennington (1963) and later by Eide and Martin (1975) and Niedrauer and Martin (Niedrauer, 1977; Niedrauer and Martin, 1979). In each case, dye was introduced into the water to obtain a greater contrast for viewing the vertical, two-dimensional structures and the brine exchange processes. In addition to clarifying some of the mechanisms for the formation and structural evolution of brine structures, observations from these experiments have provided numerical estimates for the rates of brine transport during convective exchange at the ice-water interface and brine drainage.

3.2.2 Three-dimensional imaging

Lake and Lewis (1970) made the first attempt at a three-dimensional representation of brine drainage channels. They were able to construct a generic picture of a brine channel by examining successive two-dimensional, horizontal sections of sea ice. In horizontal section the channel comprises a starburst pattern surrounding a central drainage area. Their classic picture of the three dimensional structure of a brine drainage channel is reproduced in Figure 3.1 and it remains consistent with recent observations.

![Figure 3.1: Schematic representation of a brine drainage channel showing the starburst pattern in horizontal section (Lake and Lewis, 1970).](image-url)
Imaging techniques used in medicine and industry have proved successful in visualising brine-filled spaces non-destructively. These new methods overcome one of the principal restrictions of the traditional methods in that a complete three-dimensional image of the sample can be generated. Nuclear Magnetic Resonance (NMR) utilises the different resonance properties of the hydrogen atom when it exists in either the liquid or solid phase to differentiate between signals from the brine or the ice. Early applications of this method enabled quantitative measurements on the thermodynamic phase relationships in sea ice to be made (Richardson and Keller, 1966; Richardson, 1976) and later, measurements on brine volume (Melnichenko et al., 1979). Edelstein and Schulson (1991) made the first observations of brine-filled spaces in doped saline ice using NMR methods. Direct non-destructive imaging of the internal brine volume at varying sample temperatures revealed enlargement of the brine spaces at higher temperatures. Recently, NMR imaging has been used successfully to study the structure and evolution of brine features down to a resolution of 0.2 mm (Eicken, pers. comm., 1998) and the distribution of hydrocarbons within sea ice contaminated with crude oil (Lange, pers. comm., 1998).

Another imaging technique which has been successfully applied to studying glacier and sea ice is X-ray computed tomography (CT) (Kawamura, 1988b; Kawamura, 1990). The CT technique exploits the difference in X-ray absorption properties of the ice and the brine to generate a three-dimensional picture. Kawamura (1988b) was able to observe both brine layers and brine structures to a resolution of 2 mm using this method. These new imaging techniques seem to offer the best hope for visualising the internal architecture of sea ice. With new technology and improved temperature regulation it appears entirely feasible to map the distribution and evolution of brine structures, as a function of temperature, at high spatial resolution.

Finally, an elegant method for the direct visualisation of brine structures was developed by Weissenberger (1992). The technique involved creating resin casts of the porous network inside centrifuged sections of ice cores. Following sublimation of the surrounding ice, the cast was then examined using a scanning electron microscope. The intricate three-dimensional structure of the brine channels could then be visualised at scales ranging from greater than 30 mm down to a few micrometers. The images in
Figure 3.2 clearly indicate that the nature of the brine structures changes dramatically according to the crystal texture. Granular ice gives rise to a complex, highly branched network of channels contrasting with the channels found in congelation ice which are long and straight with much less branching. Although the method is ultimately destructive and size limited, it has provided perhaps the clearest and most detailed images of the internal organisation of the interconnected brine spaces in sea ice.

![Figure 3.2: Resin casts of the brine-filled spaces in (a) granular ice and (b) congelation ice showing the variability of orientation and branching (Weissenberger et al., 1992). In each case, scale graduations are 1 mm.](image)

3.3 Nomenclature

In many areas of sea-ice research, the presence of brine structures in the ice, and their influence on ice processes, is described and qualified. Because these structures have been known for many years a nomenclature has evolved in an effort to describe their different origins and geometry. However, there is no current complete scheme of
classification or nomenclature which can be applied to these structures, though there are a number of frequently used terms. This ultimately leads to confusion when making comparisons in the literature as authors will commonly use different terms for the same feature or similar terms for quite different features. Frequently the context of the description clarifies the meaning but inconsistencies in terminology persist.

To resolve this deficiency, a discussion of the principal brine structures found in sea ice will be followed by a summary of the various terms that can be found in the literature with some examples of terminology and classification that have been used previously. From this, the terms which will be used throughout this thesis will be listed and defined, and their mutual relationship described. This nomenclature will be used to ensure consistency of meaning and clarity of description throughout the thesis. Without such definitions some degree of confusion is ineluctable.

3.3.1 Brine structures

The basic description of brine drainage channels is that they take the form of a tree; the channels consist of large vertical tubes attended by smaller tributary tubes (Wakatsuchi and Kawamura, 1987). The original observation and conception of brine drainage trees can probably be ascribed to Bennington (1963) though much of the pioneering work in elucidating their structure was made by Lake and Lewis (1970). In vertical sections through the ice, they observed large drainage channels extending up into the ice and many small vertical tubes at the ice-water interface. From these observations we can immediately make a distinction between two classes of brine structure which coexist in the ice. First there are the large brine channels which extend up into the interior of the ice and are surrounded by feeder channels. The second class comprises the small, vertical brine tubes which are found within the skeletal layer at the ice-water interface.

3.3.1.1 Brine channels

Brine channels are the most frequently observed and described brine structures. The network of channels has been described as a combination labyrinth of the inclined feeder channels and larger vertical drainage ports (Criminale and LeLong, 1984). The general
form, represented in Figure 3.1, is of a central channel with a typical diameter of a few millimetres (Eide and Martin, 1975; Martin, 1979) fed by associated small feeder channels. From examining the distribution and geometry of mature brine drainage channels in first-year ice, Cole and Shapiro (1998) concluded that the form of brine drainage channels can vary significantly depending on whether they develop under quiescent or dynamic conditions. Cole advocates that the brine drainage features, with long, straight central channels and shorter, diagonal side branches, generally form under quiescent conditions and in laboratory experiments.

A widely reported characteristic for describing brine drainage channels is their horizontal distribution and length. Although this will be discussed more fully later in the chapter, it is an important element in defining them. In the horizontal plane, the nearest neighbour spacing of drainage channels is typically 10 cm (Morey et al., 1984; Wakatsuchi and Saito, 1985) though their size and spatial density (number per unit area) are dependent on the thickness and age of the ice (Martin, 1979; Shapiro and Weeks, 1993). Vertical slabs of first-year ice show channel spacings of between 5 and 10 cm at the top becoming more widely spaced towards the bottom with channel lengths typically 0.3–0.5 m (Cole and Shapiro, 1998). In thicker sea ice, the distribution of channels is more sparse. For example, in 160 cm thick ice, Lake and Lewis (1970) reported the mean separation of brine channels to be 13.4 cm which was later modified to 10 cm (Martin, 1979) with the channels extending 0.1 to 0.5 m up into the ice.

A common element of many descriptions of brine drainage channels, whether from ice formed artificially or naturally, is the occurrence of inclined feeder channels. These feeder channels radiate outwards like the limbs of a tree from larger vertical drainage channels or tubes (Morey et al., 1984). The angle of the feeder channels from the vertical is a quantity in which there is considerable agreement both from observation and theory. Measurements on saline ice grown in a small, narrow tank put this quantity in the range 30°–60° with a mean of 45° (Niedrauer and Martin, 1979) which compares well to measurements of between 40° and 54° in natural ice (Lake and Lewis, 1970). Numerical analysis of the angle of slope has shown this range to be optimal for brine drainage (Criminale and LeLong, 1984).
3.3.1.2 Brine tubes

Although brine tubes are not the principal theme of the thesis, they are influential in brine channel formation and are frequently referred to as brine channels. Their inclusion and description at this point is therefore beneficial to the continuity of the discussion. From the limited commentary in literature, it is evident that the brine tubes observed at the ice-water interface form an entirely different class of brine structure from the brine channels discussed above. Again, Lake and Lewis (1970) provide the initial discourse and describe the tubes as cylindrical with a median diameter of 0.4 mm, 2–3 cm in length and located close to the interface. The spatial density of the tubes was very high at 42 tubes cm\(^{-2}\) giving a coverage of approximately 5% of the interface surface. From this description it is clear that they have little resemblance to the much larger and more widely spaced brine drainage channels.

Two other reports of these interfacial tubes also imply that they have an entirely different characteristic geometry to brine channels. Niedrauer and Martin (1979) observe cusps within the porous skeletal layer from which there is an outflow of brine, though no indication of diameter or length of the tube is given. Small tubes, 1–2 mm in diameter with a circular cross section were observed by Wettlaufer et al. (1997a) in the skeletal layer of growing ice. In both cases, these features had a typical nearest neighbour spacing of 10 mm. It is interesting to note that despite the obvious differences in size, distribution and geometry of brine tubes, they are at times also referred to as brine channels (Wettlaufer et al., 1997a).

3.3.1.3 Salient characteristics

From the descriptions and discussion of brine structures it is possible to identify some basic defining characteristics of brine drainage channels and brine tubes. Although there are fewer descriptions of brine tubes found in the literature, it is clear that their geometry is fundamentally different from that of brine drainage channels. A summary and comparison of the salient characteristics of each is presented in Table 3.1.
Near vertical channels of high aspect ratio.

Channel diameters greater than a few millimetres but variable according to the ice temperature.

Channel lengths greater than a few centimetres depending on ice thickness.

Located at the ice-water interface and penetrate beyond the skeletal layer. They can also be found in the ice interior.

Horizontal spacing usually greater than 5 cm - sparse horizontal distribution.

Often surrounded by inclined feeder channels.

Vertical channels of high aspect ratio.

Channel diameters generally less than 1–2 mm.

Channel lengths no more than 2–3 cm.

Open at the ice-water interface and are generally confined to the skeletal layer.

Horizontal spacing of a few millimetres or less - dense horizontal distribution.

Independent structure with no attendant network.

Table 3.1: A summary of the comparable and contrasting features which define brine channels and brine tubes.

3.3.2 Classification schemes

To illustrate the need for a systematic classification with a refined nomenclature, and to summarise some of the terminology in current use, Table 3.2 contains a selection of terms that appear in the literature. These terms, in various combinations, are the most commonly used though without any strict or consistent application. Some of the terms are more generic in nature than others, for example channels, inclusions and pores, and are the most frequently used.
Table 3.2: A summary of the terms commonly found in sea-ice literature which are used to describe brine-filled spaces within the ice.

<table>
<thead>
<tr>
<th>Brine Spaces</th>
<th>Brine Structures</th>
<th>Qualifying Terms</th>
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<td>Pores</td>
<td>Channels</td>
<td>Drainage</td>
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<td>Pockets</td>
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<td>Inclusions</td>
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Deriving a scheme for classifying the brine structures occurring in sea ice would seem to be an obvious step given the volume of literature either devoted to or describing them. However, despite the numerous terms used to describe brine structures in sea ice, there are few such schemes for systematic classification. In this section, two of these schemes will be described, and their merits and limitations discussed.

3.3.2.1 Generations

Bennington (1967) made the first attempts at systematic classification of brine structures. He classified brine structures according to generations; first generation being those which form during ice growth and the second generation those which form during ice deterioration. Of the first generation structures, there exist two types; the straight, vertical channels which are not consistently associated with the crystal boundaries (cf. Wakatsuchi and Saito, 1985) and the interconnecting network of brine pockets which appear as cone-shaped networks leading into the larger channels (cf. Lake and Lewis, 1970). The second generation features, which are formed during warming and melting of the ice, are subdivided into channels which form at the surface and melt through the ice, and pockets of brine which migrate vertically along the temperature gradient.
Though Bennington’s scheme encompasses many of the features that can be observed in sea ice, classification of an individual brine structure requires the observer to either know or postulate the circumstances of its formation. Ideally, a classification scheme for brine structures should be based on their size and geometry, remaining independent of the mode of their formation, which may not be known. Additionally, Bennington’s classification is independent of size and the term drainage channel may include features of quite different magnitudes. Whilst this approach simplifies the scheme it permits no differentiation between the quite distinct brine channels and brine tubes discussed above.

3.3.2.2 Pore space

Another classification scheme for brine structures, developed by Eicken (Eicken et al., 1991; Eicken et al., 1995), has the basis that all spaces in sea ice, whether they contain air or brine, come under the generic description of pores. The terms small-scale primary and large-scale secondary are adopted to describe the pore space (Eicken et al., 1991). Primary pores are the brine pockets which form between ice platelets during congelation growth. These pockets are distributed throughout the ice and may be viewed easily in thin section analysis (Sinha, 1977; Eicken, 1991; Perovich and Gow, 1991; Perovich and Gow, 1996). In general, they are smaller than the ice grains, though their size and distribution is dependent on both the ice temperature and age (Perovich and Gow, 1996).

Secondary pore spaces evolve from the primary pores, and have dimensions similar to, or greater than, the crystal size. Their geometry, interconnection and size undergo continuous evolution as the ice cover ages to the extent that the pores may become so extensive that their geometry is often not apparent from ice cores alone. Eicken et al. (1991) supports the assumption that there will be a gradual transition in the ice from a primary pore regime to one of secondary pores (Wakatsuchi and Ono, 1983), with secondary pore development occurring until complete melting of the floe (Martin, 1979; Cox and Schultz, 1980). A variation of this scheme, used for summer, multiyear ice, is based upon the size, aspect ratio and spacing of the pores (Eicken et al., 1995).

The classification of brine features by Eicken is essentially based upon geometry, rather than the age and formation approach used by Bennington. This method of
classifying by geometry takes into account the dimensions of the features and their relationship to the associated crystal structure, and is independent of the precise mode of formation. It is acknowledged that the use of pores as a generic term to describe brine features is appropriate and relevant in the context of microstructural analyses of sea ice. However, it is often not always clear from the appearance of a brine feature whether it formed during ice growth or whether it developed at a later time. In this respect, the transition from primary to secondary could render the classification of an individual structure somewhat arbitrary. For example, at what stage does the interconnection of brine pockets between platelets (primary pores) constitute a mature, inclined feeder channel (secondary pore)? Additionally, there is no possibility for distinguishing quite different structures which would be in the same class of primary or secondary porosity leaving the scheme open to arbitrary terminology.

3.3.3 Definition of terms

From the analysis of brine structures which has been presented, current terminology and previous schemes of classification, it is practicable to identify individual classes of brine features and to define them accordingly. The following terms and definitions will be applied throughout the thesis.

**Brine Features:** This generic term encompasses all the brine-filled spaces which are present within a solid ice matrix.

**Brine Inclusions:** This is another generic term defining a class of brine-filled spaces which are located between platelets. In this respect, most of the brine features resulting from the entrapment of brine at the growing ice interface will be classified as brine inclusions. Essentially the class includes all features described by Eicken *et al.* (1991) as the small-scale, primary porosity with the exception that it is subdivided into *brine pockets* and *brine layers*. 
**Brine Pockets:** This category of brine inclusion applies to any feature contained between the ice platelets, generally smaller than the grain size and with no obvious interconnection. The shape of brine pockets is typically spherical or ellipsoidal with a relatively low aspect ratio (Kovacs and Morey, 1978). A spherical inclusion much larger than the grain size which might form in warm, summer sea ice (Eicken et al., 1995) would also be classified as a brine pocket by virtue of its shape and isolation from other brine features.

**Brine Layers:** As sea ice warms, the brine pockets between the platelets will enlarge and tend to coalesce (Perovich and Gow, 1996), this results in the formation of brine layers. This category of brine inclusion differs from brine pockets by the high degree of interconnection between the platelets which results from the merging of previously isolated brine pockets. The platelets continue to influence the dimensions of the brine layers so in this respect they are smaller in size than the grains.

**Brine Structures:** Whereas brine inclusions are quite passive features, brine structures are more active, particularly in the movement of brine. The principal difference between brine structures and brine inclusions is the dynamic nature of the brine structures. In general, brine structures will be of high aspect ratio and will be open at the ice-water interface where they contribute to convective or desalination processes.
Brine Tubes: Brine tubes are the small-scale brine structures with diameters of approximately 1 mm. There are two possibilities for brine tubes depending on where they are found in the ice. Interfacial brine tubes are those within the skeletal layer at the ice-water interface whereas internal tubes are located away from the skeletal layer. Interfacial tubes with circular cross section (Lake and Lewis, 1970; Wettlaufer et al., 1997a) are typically found between the platelets or at grain boundaries. They are closely spaced with a high spatial density in the horizontal plane. Internal tubes, unlike brine inclusions, will be independent of crystal boundaries, penetrating through platelets and grains and generally larger than the grain dimensions; they have been described previously and called tubular pores by Eicken et al. (1995).

Brine Channel: Large-scale brine structures in sea ice are classified as brine channels, and will vary in terms of complexity and organisation. A brine channel is generally a combination of two components, the central channel, with a diameter greater than a few millimetres, surrounded by the inclined feeder channels. The development of brine channels in sea ice is equivalent to the development of the large-scale secondary pores described by Eicken et al. (1991). Brine channels are a network of interconnected brine features which make a significant contribution to the movement of brine though the ice, hence the more complete, though synonymous term, brine drainage channel. Their distribution is more sparse than for brine tubes and they extend into the ice beyond the skeletal layer. In some cases, not exclusively determined by the crystallography, a brine channel will not be attended by feeder channels and will consist solely of a central brine channel (Cole and Shapiro, 1998).
Chapter 3: Brine channels: structure, formation and processes

The proposed definitions of the terms has a number of advantages over existing classification schemes:

- Independent of the growth history - often the exact time or mechanism of formation of a particular feature is not known.

- Classification by simple comparative observation - it should be possible to classify or describe a particular feature without resorting to a detailed quantitative analysis.

- Based on the structural characteristics - the dimensions, aspect ratio, distribution and degree of complexity or interconnection are all characteristics which may be used to define a range of brine features.

- Simple subdivision of large classes - broad classes of structures are subdivided once, removing ambiguity, but without resorting to multiple levels of subdivision based on subtle structural differences.

- Includes generic terms - this property of the definitions enables brine features in the ice to be described easily and referred to in general terms without resorting to a more complex or specialised terminology.

Even though the nomenclature above is based upon commonly used terminology which is entirely compatible with current ideas of brine structures, it has given a potentially confusing and inconsistent nomenclature a more systematic and formal basis. To make the relation between different terms clearer, the nomenclature as defined may be represented in a hierarchical form, Figure 3.3. This is the structure for a systematic classification for brine features in sea ice.
Chapter 3: Brine channels: structure, formation and processes

![Hierarchical diagram showing the relation between the different terms used to describe brine features in sea ice.](image)

**3.4 Formation and evolution of brine channels**

A fundamental question when discussing brine drainage channels is simply, how do they form? In the most simplistic sense, they form by the interconnection of the brine inclusions between ice platelets and grains to create a desalination path through the ice. Although this is essentially correct, it is by no means a complete answer to the question. In this section the formation, evolution and geometry of brine drainage channels will be described and discussed, based on the available literature and the author's observations.

**3.4.1 Brine tube formation**

Recent work by Wettlaufer, Worster and Huppert (Wettlaufer et al., 1997b; Wettlaufer et al., 1997a; Worster, 1997), based on simple laboratory experiments and fluid dynamic theory, has provided new ideas for the formation process of young sea ice and
Chapter 3: Brine channels: structure, formation and processes

particularly the origin of brine drainage channels. Because sea ice may be considered as the solidification of a multicomponent melt, or alloy, much of their interpretation and theory is based on analogous metallurgical and geological solidification processes. For example, analogous channels which form in sea ice have also been observed in castings of metallic alloys where they are called chimneys (Hellawell et al., 1993). As a consequence, they adopt the term mushy layer, meaning the 2-phase region of pure ice crystals and liquid brine, when referring to sea ice (Worster, 1992; Worster, 1997). In the interests of consistency, the term sea ice will be retained during this discussion rather than mushy layer.

The mechanism of channel formation is linked to the fate of the brine formed during initial ice growth. During the initial stages of the solidification process, the liquid fraction, which is trapped within the solid matrix, becomes enriched in salt. With continued growth, the density of this brine increases both for thermal and compositional reasons. The dense, enriched liquid becomes trapped within the matrix because it has insufficient negative buoyancy to overcome mechanical resistance imposed by the aggregation of the solid crystals. At this time, observations indicate that weak compositional convection occurs at the interface (Wettlaufer et al., 1997a; Wettlaufer et al., 1997b) whilst the concentration in the underlying water remains essentially constant\footnote{The concentration of the liquid beneath the ice remained \textit{constant to within the resolution of the measurements} (Worster and Wettlaufer, 1997), which in this case were made by refractometer. Additionally, the water in the tank was stagnant and any increase in salinity that may have occurred may not have been resolved due to the inefficiency of mixing.}. Convection at the ice-water interface is induced by density instabilities in the boundary layer caused by differences in temperature and composition, this is termed \textit{interfacially driven convection}. This mode of convection operates on small length scales with very minimal penetration into the porous ice, and with a small associated salt flux.
The ice continues to grow, and on attaining a critical thickness, there is an abrupt onset of brine drainage with the appearance of strong convective plumes and a sudden increase in the salt flux. This second mode of convection, *internally driven convection*, is driven by the compositional density gradients within the growing sea ice which convey a negative buoyancy to the dense, enriched internal brine. As the ice thickens, the brine density and ice permeability increase sufficiently to force brine through the ice structure. In contrast to *interfacial convection*, this mode of convection penetrates the porous ice and appears as plumes of dense brine, constituting a large salt flux, which emanate from brine tubes at the ice-water interface. A similar sudden onset of brine drainage is observed from leads where the maximum brine flux was observed within about 6 hours of lead formation (Morison and McPhee, 1998).

Careful observation of the brine tubes in the skeletal layer at the ice-water interface reveal that they have a diameter of 1–2mm with a circular cross section which is unaffected by the platelet structure of the crystals (Wetlaufer et al., 1997a; Worster and Wetlaufer, 1997). Further, they are distributed randomly across the interface with a mean separation of approximately 10 mm. It is remarkable to compare this with the observations of brine cusps forming at a growing ice-water interface which also have a mean spacing of 10 mm (Niedrauer and Martin, 1979). The diagrams representing the flow characteristics of these independent observations are presented in Figure 3.4. In each case the flow lines penetrate the ice so that the point of brine outflow is flanked by a region of brine inflow.

![Diagram](image)

**Figure 3.4:** An illustrative comparison of the characteristics of brine exchange at the ice-water interface from the observations of (a) Niedrauer and Martin (1979) and (b) Worster and Wetlaufer (1997). The arrows in (a) indicate direction of brine flow and the lines in (b) are streamlines of brine flow.
Internally driven convection allows an exchange of brine between the ice and the underlying water. This exchange of brine, which has been inferred from temperature fluctuations at the growing interface (Lake and Lewis, 1970; Niedrauer and Martin, 1979), mediates the development of the tubes as the brine tries to maintain phase equilibrium. Because the rate of thermal diffusion is approximately 250 times greater than the rate of diffusion for salt, equilibrium is re-established by phase transition rather than solute migration. The outflow of cold, saline brine from the ice results in the dissolution of the solid phase at the tube walls, increasing permeability and focusing the flow into narrow channels. This is replaced by the inflow of warmer, fresher brine which causes accretion of ice within the porous new ice increasing its solid fraction and reducing permeability (Worster and Wettlaufer, 1997; Wettlaufer, 1998, (p.185)).

Later, in chapter 6, it is argued that mushy layers are not representative of sea ice in general. However, it is likely that the basic mechanisms governing brine tube formation, as derived from mushy layer analysis, are still valid. There are many other accounts which describe plumes of brine descending from discrete points in the ice-water interface (Bennington, 1963; Foster, 1969; Farhadieh and Tankin, 1972; Eide and Martin, 1975; Wakatsuchi, 1977). The plumes, or streamers, emanate from brine tubes and it is these which may act as a seed from which brine channels will later form.

### 3.4.2 Brine channel formation

The precise transition from brine tubes to brine channels is a moot point. Certainly brine channels are spaced more sparsely than the streamers which mark the tubes, so the transition must entail some element of competition and preferential selection. In some cases stronger plumes, representing a larger salt flux, can be identified (Foster, 1969; Wakatsuchi, 1977; Wettlaufer et al., 1997b), so it is possible that it is these tubes, which show a greater propensity for brine transport, which will then develop into brine drainage channels. What is certain is that brine drainage channels form in the very early stages of ice growth when the ice is only a few centimetres thick (Bennington, 1963; Eide and Martin, 1975; Wakatsuchi and Saito, 1985). Once a brine channel has formed at the porous interface it is then likely to develop into the branched structure comprising a central channel and feeder channels.
In terms of brine channel development, the downward movement of cold, dense brine through the ice has two positive feedback effects. First, as the brine attempts to maintain phase equilibrium it causes melting at the channel wall, increasing the radius of the channel. With a larger channel radius, the velocity of the brine increases which tends to reduce the pressure field in the channel. The second positive effect of brine movement is that it cools the ice around the channel through which it moves. Brine expulsion is enhanced by this cooling and brine is forced along cracks from a region of high pressure (the brine pocket), along the feeder channels, to one of low pressure (the brine channel) which is a point of natural weakness. The fluid dynamics of this process are very sensitive to the channel radius (in fact it has an $r^4$ dependence (Martin, 1970)), therefore, small variations in channel diameter will produce channels of variable efficiency. For the complete analysis of this effect, see (Eide and Martin, 1975).

Brine drainage channels which can be identified in relatively thin ice (Bennington, 1963; Eide and Martin, 1975), generally have a homogenous distribution in the horizontal plane. Observations of drainage channels in both natural and tank-grown sea ice have independently demonstrated this homogenous nature of their distribution (Saito and Ono, 1980; Wakatsuchi and Saito, 1985, Cottier, pers. obs., 1997 and 1998). There are two aspects of brine channel distribution which need to be considered; the location of the channels and their spacing or spatial distribution. These will now be discussed in relation to brine channel formation processes.

### 3.4.2.1 Location

What are the factors that determine the location in the ice where a brine drainage channel will form? In the early stages of ice growth, their location is determined by the crystal structure. It has been shown experimentally that brine channels will form preferentially at the intersection of a grain boundary (Wakatsuchi and Kawamura, 1987; Kawamura, 1988a). In columnar ice the single crystals (or grains) of ice, which are composed from pure ice platelets, form an interlocking structure with many grain boundaries. From microscopic observations (Wakatsuchi and Saito, 1985), it has been determined that brine channels form at the intersection of two or more grain boundaries and not within the grains themselves. However, lower in the ice, brine channels will grow through the
interior of grains. This implies that once a brine channel has formed, it will advance in a direction towards the ice-water interface independent of the crystal structure.

The answer to the question of brine channel location may lie in the relative orientations of the crystal c-axes (Wakatsuchi and Kawamura, 1987; Kawamura, 1988a). The link between brine channel formation and crystal orientation was investigated systematically by growing sea ice from a seed composed of single crystals with predetermined c-axis orientations (Kawamura, 1986). The boundaries between the single crystals with different relative c-axis orientations show different propensities for brine channel formation. This effect will be illustrated with reference to Figure 3.5. Although the inclination of c-axes is in these examples is somewhat extreme compared to that encountered in a natural situation, the results demonstrate the clear influence of crystal structure on brine channel location.

![Figure 3.5: A schematic drawing in the vertical plane of two possible orientations of inclined platelets at a vertical grain boundary (the c-axis is perpendicular to the platelets). The brine pockets in case A migrate towards the grain boundary whilst in case B they move away.](image)

At a grain boundary where the c-axes on each side of the boundary are angled so that the platelets are growing towards the boundary, case A, brine channels form with a regular spacing and are symmetrical. In contrast, brine channels do not form at the boundaries where the c-axes on each side are angled such that the platelets grow away from the boundary, case B. In the intermediate case where only one side of the grain boundary has an angled c-axis, with platelets growing towards the boundary, brine channels form at regular intervals but the channels are asymmetric. Finally, brine channels do not form at boundaries where the c-axes and platelet orientations on each side of the boundary are identical.
From these observations, it is concluded that the source of brine for the formation of a brine channel comes from the movement of brine along the platelet boundaries towards the grain boundary (Wakatsuchi and Kawamura, 1987). This movement of brine is probably by brine expulsion occurring preferentially along the inclined platelet boundaries. In sea ice with a steep temperature gradient, the expulsion of brine from fractures along platelet boundaries is the most likely process for the formation of feeder channels (Eide and Martin, 1975). As the brine reaches the grain boundary, gravity drainage will dominate and the brine channel will enhance the desalination process.

To summarise, brine drainage channels tend to form between crystal boundaries with their feeder channels forming between platelets (Lake and Lewis, 1970; Wakatsuchi and Saito, 1985; Wakatsuchi and Kawamura, 1987). The formation of drainage channels depends on the relative crystal orientations at grain boundaries where brine expulsion and gravity drainage contribute to the development of the brine drainage channels.

3.4.2.2 Spatial distribution

From systematic observations of brine channel distribution there is a clear link between initial ice growth rate and the spatial distribution of brine channels which is quantified by their spatial density. In both artificially (Saito and Ono, 1980; Kawamura, 1988a) and naturally formed sea ice (Wakatsuchi and Saito, 1985), the spatial density of brine channels increases with growth rate. At the limit of slowest growth rates, discrete drainage channels do not form, instead brine drainage occurs uniformly across the whole growing interface (Kawamura, 1988a).

One explanation for this result is that the highest probability of occurrence of the optimal crystal orientation is during the initial stages of ice growth where the c-axis orientation is random (Weeks and Wettlaufer, 1996). Additionally, grain sizes are generally smaller at rapid growth rates (Wakatsuchi, 1974) giving rise to a greater frequency of crystal intersections per unit area. Therefore, on the basis of probabilities, the likelihood of attaining a preferential orientation of c-axes at a grain boundary increases with growth rate leading to a greater spatial density of brine drainage channels.
Arguments against this explanation can be found in later work in which sea ice was grown from a composite seed crystal. The relative orientation of the component crystals was optimal, as described in section 3.4.2.1, and their intersection frequency was fixed (Wakatsuchi and Kawamura, 1987; Kawamura, 1988a). In this arrangement, the spatial density of brine channels remained a function of growth rate even though the variability of the crystal intersection frequency had been removed. From this result it can be concluded that the relationship between brine channel distribution and ice growth rate is not determined by the random nature of c-axis orientation. Rather, for a given optimal relative orientation of the c-axes at a grain boundary, the spatial density of brine channels is influenced by another growth rate dependent property of sea ice.

As mentioned previously, grain size is growth rate dependent though no clear link has been found between this crystallographic property and brine channel distribution (Kawamura, 1988a). Therefore, it is proposed that the bulk salinity of the ice is the controlling factor in determining the spatial density of brine drainage channels. This explanation is based on the assumption that an active channel will require a minimum throughput of brine for its continued existence.

At rapid ice growth rates, more brine is trapped and incorporated in the skeletal layer leading to greater bulk salinities - as discussed in section 2.5.1. If the bulk salinity increases, a greater number of brine channels can be sustained per unit volume of ice. With more channels in the ice, they become necessarily more closely spaced. At slower growth rates, larger volumes of more saline brine are excluded from the ice at the advancing interface reducing the salt content of the ice (Wakatsuchi and Ono, 1983). In this case, fewer, more sparsely distributed brine channels will develop. The mechanisms by which brine is incorporated into sea ice may provide a link between spatial density of brine channels and the ice growth rate by virtue of its salt content.

In support of this proposed mechanism, it is noted that the angle of inclination of the c-axes from the horizontal also influences the spacing of brine channels. In general, for a fixed growth rate, the greater the inclination of the c-axis the more widely spaced are the drainage channels (Kawamura, 1988a). From Figure 3.5 it is clear that as the angle of inclination of c-axis increases, the migration of brine will become less efficient.
It is postulated that the spacing of brine channels is dependent on the salt content of the ice and the efficiency with which this brine can migrate towards a grain boundary.

3.4.3 Brine channel evolution

As stated in section 3.3.1.1, the nearest neighbour separation of brine channels is about 5–10 cm (Morey et al., 1984; Wakatsuchi and Saito, 1985; Cole and Shapiro, 1998). In the early stages of ice growth, the horizontal distribution of brine channels remains quite constant, certainly up to ice thicknesses of about 25 cm (Shapiro and Weeks, 1993; Cole and Shapiro, 1998), though this changes with ice thickness and age (Martin, 1979; Shapiro and Weeks, 1993). As the ice grows and thickens, discontinuity occurs in the brine channels and their spatial density decreases with depth (Shapiro and Weeks, 1993; Cole and Shapiro, 1998). Discontinuities occur at different depths where some drainage channels will terminate and others form, though some channels will pass through these horizons unperturbed. Remnants of previously active brine channels can be found in the upper part of the ice (Lake and Lewis, 1970; Cole and Shapiro, 1998) with active channels located at the ice-water interface. This suggests that brine channels tend to grow with the ice and then become extinct, being superseded by the formation of a new channel, rather than migrate downwards with the advancing ice-water interface.

An additional characteristic of brine drainage channels is that the central channel and the feeder channels often contain fine-grained ice (Lake and Lewis, 1970; Cole and Shapiro, 1998). The origin of this porous polycrystalline mass is attributed to the ingress of warmer, less saline water into the relatively cold channels which then freezes to attain equilibrium. It could also be an artefact of sampling as the brine within the channel freezes whilst the extracted samples stand in the air at a low temperature.

Prior to melt, brine channels rarely extend through the entire thickness of the ice. In late spring and summer, as the air and ice temperatures increase, the diameter and length of the channels increases (Eicken, 1994; Werner and Lindemann, 1997) and complete vertical penetration through the ice is possible, dramatically increasing the permeability of the ice (Martin, 1979; Fritsen et al., 1994; Lytle and Ackley, 1996). This process of melting through the ice is enhanced by the downward movement of brine.
from upper layers and by solar radiation being absorbed by the liquid fraction within the brine channels (Wadhams and Martin, 1990). By early summer, brine volume and interconnection of brine inclusions has increased to the extent that the ice is highly permeable both in the brine channels and between grain boundaries (Martin, 1979).

The final stage of the evolution of a brine channel is the continual process of enlargement and interconnection as the ice deteriorates by rotting rather than ablation. Brine channels at an advanced stage of warming may be considered to be like a river or arboreal system. In models of sea ice decay (Cox and Schultz, 1980) the diameter of the brine channel has a dependence on the thawing-degree time, a direct inversion of simple ice growth models (Maykut, 1986). Eventually, the ice becomes so porous and weak that it disintegrates when subject to ocean waves.

### 3.4.4 Brine channel geometry

What are the physical and environmental factors which influence the development of brine drainage channels? The effect of the degree of c-axis alignment on the geometry of brine channels is not clear. Brine channels may be larger in ice with an aligned c-axis (Martin, 1979) but this was a single observation and may be affected by the natural variability in brine channel sizes. However, the symmetry aspects of brine channels are noted to be independent of any c-axis alignment (Cole and Shapiro, 1998).

The influence of crystal structure on the geometry of brine channels is not fully understood. Observations of channels indicate that crystal structure is not the dominant influence on their geometry (Cole and Shapiro, 1998) although the specific connectivity within a brine channel will in general be different for ice with different growth histories. This is consistent with the observations of the interconnection within brine structures which is shown to be highly dependent on crystal structure (Weissenberger et al., 1992), see Figure 3.2. Under dynamic conditions, the structure of brine channels is found to be more variable and asymmetric (Cole and Shapiro, 1998). Therefore, although the classical idealised picture may not be valid for all cases, it may illustrate the processes which occur in an atypical situation.
3.5 Brine channels and sea-ice processes

As discussed in chapter 2, the fundamental difference between sea ice and freshwater ice is the labile and mobile brine that exists in sea ice. The distribution of this liquid fraction and the geometry of the brine channels, with the hydraulic link they establish with the ocean, have an acknowledged and important role in many sea-ice processes. In this section, five sea-ice processes that are directly influenced by the presence of brine channels will be considered. Each topic presents an extensive research field in its own right and thus the discussion will be restricted to the role that brine channels exert in controlling or modifying these processes.

3.5.1 Mass transport

In contrast to the comparatively impermeable river and lake ice, the enhanced permeability possessed by sea ice, because of the occurrence of connected internal brine structures, provides a path for the movement and exchange of brine and gas as well as nutrients and pollutants. The movement of fluid in sea ice is confined to two locations. First, within the dendritic layer at the ice-water interface and second, through the interconnected network constituting a brine drainage channel. This ability of the brine channels to allow mass transport into, through and within the sea ice, influences the salt flux to the oceans, the transport of nutrients and pollutants, and the evolution of algal communities. The transport properties of sea ice are in general determined by the variable and strongly anisotropic nature of its permeability which is much greater in the vertical axis than in the horizontal plane (Gosink et al., 1976).

Reeburgh (1984) presents a detailed review of brine movement in sea ice using data obtained in both laboratory and field studies. A distinction is made between transport within the sea ice by the brine drainage channels and the exchange of water at the skeletal layer. The movement of brine within brine channels in growing sea ice is reported to be of either an oscillatory nature (Eide and Martin, 1975) or via simultaneous bi-directional flow (Niedrauer and Martin, 1979). Within the skeletal layer, brine exchange has been inferred from temperature oscillations with periodic fluctuation (Lake and Lewis, 1970; Niedrauer and Martin, 1979) and brine movement has been observed
using dyes with brine exiting the ice in discrete plumes (Niedrauer and Martin, 1979), though direct measurements of brine fluxes from sea ice are few (Wakatsuchi, 1977; Wakatsuchi, 1983; Wakatsuchi and Ono, 1983).

In the spring, as the ice begins to warm, brine channels will enlarge thus enhancing its vertical permeability (Martin, 1979; Hudier and Ingram, 1994). The vertical movement of liquid through the ice is caused either by surface loading (upward movement) or by surface melting (downward movement). The ability for sea ice to convey brine through its entire thickness is dependent on the condition that the ice sheet is permeable at all depths. In general this requires that the brine volume is greater than the critical value of 5% identified by Cox and Weeks (1975).

Surface flooding of sea ice is prevalent in the Antarctic and can affect the microwave signature of the ice (Drinkwater and Lytle, 1997). Following severe snow fall, depression of the ice causes upward flushing of the brine channels by seawater. Complete penetration of the brine will be prevented if a single impermeable layer is present in the ice (Jeffries et al., 1998). Flushing of sea ice by surface meltwater will usually occur only under warm conditions when the ice is highly porous (Hudier et al., 1995). This is a time of significant desalination of the ice through the creation of a hydrostatic head at the surface (Hudier and Ingram, 1994).

The transport of nutrients into sea ice is an important process for influencing the development of the biological communities (Gradinger et al., 1992). In young sea ice, nutrient concentrations will follow the salinity distribution (Gradinger and Ikävalko, 1998), therefore the movement of nutrients is allied to the movement of brine. Brine exchange between the ice and the water column is one of a number of possible mechanisms for the transport of nutrients into sea ice (Meguro et al., 1967; Melnikov, 1995). Under some circumstances this process will supply only a fraction of the nutrient demand during an algal bloom (Cota et al., 1987) but in other cases the flow of brine and nutrients through brine channels is the trigger for algal blooms (Fritsen et al., 1994). Other field studies have indicated the importance of brine drainage in controlling the nutrient status of young ice (Melnikov, 1995). In summer sea ice the temperature gradients decrease and there is stable stratification of the brine within the ice resulting in a cessation of convective exchange (Hudier and Ingram, 1994; Eicken et al., 1995).
A direct environmental consequence of the permeability of sea ice concerns the migration of oil through the ice following an oil spill beneath it, a real possibility because of current transport and drilling in the Arctic Ocean. Any spill occurring under the ice during winter will be trapped there as a lens or pocket and will become incorporated in the skeletal layer by percolation into this porous region of the ice. Continued growth of the ice beyond the oil ensures that it becomes encapsulated within the ice mass. In the spring, as the ice begins to warm and become porous, the oil will migrate towards the surface (Martin, 1979). This migration occurs along the path of the enlarged brine drainage channels and is enhanced by its positive buoyancy. The first appearance of oil on the ice surface occurs in early summer where it promotes melt pond formation.

3.5.2 Thermal transport

In many thermodynamic models of sea ice, conduction is considered as the primary heat transfer mechanism (Maykut, 1986; Wettlaufer et al., 1990; Wettlaufer, 1991). Heat flux is calculated using the thermal conductivity of the ice and the temperature gradient within the ice. Brine pockets behave as thermal buffers in sea ice by the release of latent heat, thus models of thermal conductivity in sea ice vary in their treatment of the geometry of brine distribution (Schwerdtfeger, 1963; Yen, 1981; Weeks and Ackley, 1986). Of recent interest, however is the effect of a mobile brine fraction which is not included in these early models.

The contribution of convection to the heat transfer properties of growing sea ice is estimated to be about 1% (Niedrauer and Martin, 1979). However, convection events have been detected in isothermal sea ice leading to the conclusion that in some instances, conduction may not always be the principal heat transfer mechanism. The observations were made in sea ice of the Weddell Sea, Antarctica, and convection was interpreted from time series profiles of salinity and nutrients (Fritsen et al., 1994), and salinity and temperature (Lytle and Ackley, 1996). During the passage of a freezing front through the ice, density instabilities are generated within the brine channels initiating convection; this causes cold, dense brine to flow downwards through the ice to be replaced by the relatively warmer, fresher seawater. Brine channels extending through the full depth of
the ice act as conduits for the exchange and movement of these two water masses. Biological data indicated that the brine in the ice may be replaced many times (Fritsen et al., 1994). The net result is an overall upward heat flux from ocean to atmosphere decreasing the effective thermal insulation of the sea ice.

A more detailed study of a long time series of thermistor measurements in Antarctic sea ice presents additional evidence for the possibility of brine movement modifying the simple concept of conductive heat fluxes (McGuinness et al., 1998). Their results indicate that a transport mechanism, other than conduction, is contributing to the heat flow and the possibilities of brine convection or migration are considered as a second order effect. The precise nature of this brine movement is not conclusive; however, brine expulsion is postulated as the primary mechanism in cold ice whilst in warmer ice, convection is proposed. The recent interest in the role that brine structures have in modifying the thermal properties of sea ice has led to the conclusion that convection in brine channels has the potential to reduce the effective thermal insulation of the sea ice cover thus increasing the net ocean-atmosphere heat flux.

3.5.3 Optics

The interaction of visible light with sea ice is reviewed thoroughly by Perovich (Perovich, 1996; Perovich, 1998). The main considerations of brine inclusions on the optical properties of sea ice are in the absorption and scattering processes. Of the two processes, absorption is the more straightforward to describe physically. The total absorption coefficient for sea ice is expressed as a simple addition of the relative contributions to absorption by the ice, brine, solid salts and impurities (Grenfell, 1991). Scattering however, requires a more thorough knowledge of the number and size distribution of the brine inclusions. In most cases the inclusions are treated as spherical or ellipsoidal brine pockets uniformly distributed throughout the solid ice matrix, though some models have attempted to incorporate the distribution of the inclusions according to the growth history and salinity of the ice (Grenfell, 1983). In more sophisticated models, any degree of vertical complexity may be applied with the distribution, size and scattering coefficient of the brine pockets varying as a function of temperature and bulk salinity (Grenfell, 1991).
Chapter 3: Brine channels: structure, formation and processes

The effect of vertically oriented brine inclusions on the characteristics of scattering in sea ice has been investigated in field experiments. Congelation ice acts as a weakly channelling medium for light passing through it due to anisotropic scattering (Buckley and Trodahl, 1987a). The angular distribution of the light path is peaked in the vertical direction and independent of the angle of incidence. The anisotropic nature of the radiance has been attributed to the brine layers and channels (Trodahl et al., 1989) and the bulk scattering properties of the ice are dominated by the geometrical arrangement and volume of the brine channels. Additionally, brine drainage from the upper portion of the ice during warming changes the optical properties of the ice surface with an increase in the air fraction leading to an increase in the amount of scatter (Buckley and Trodahl, 1987b).

Larger brine structures are likely to have a more significant effect on the optical transmission properties of sea ice. Stalactites which often form below large brine channels (Dayton and Martin, 1971; Lewis and Milne, 1977; Perovich et al., 1995) commonly act as a point source for light (Lewis and Milne, 1977, Lewis, pers. comm., 1998) The possibility of brine channels having a waveguide effect on light was first proposed by Wadhams (1980) and by Wadhams and Martin (1990) as a means for brine channel enlargement by the channelling of solar radiation. This is sometimes referred to as the fibre optic effect though calculations of the total internal reflection condition makes this a physically unrealistic mechanism for light propagation through sea ice.

Enhancement of the light field in the region of brine channels has been observed at the ice-water interface of sea ice grown in a large tank (Cottier, pers. obs., 1997). In other studies of light enhancement effects, no increase in light intensity was observed in ice which contained small brine channels only (Krembs, pers. comm., 1998). Warm sea ice has a more complex internal brine structure and as brine inclusions become interconnected, it is unclear how the new geometry of the brine will influence the propagation of light. It is anticipated that as brine inclusions coalesce and connectivity increases, there will be a corresponding decrease in scattering leading to enhanced transmission, a situation which is supported by field observations (Light, 1995).
3.5.4 Biology

The occurrence of biota in sea ice has been known for many years (Horner, 1985b). Sea ice has been shown to be inhabited by a diverse community of organisms (Horner, 1985c; Melnikov, 1997) and these are recognised as a major source of primary production in the polar oceans (Legendre et al., 1992; Spindler, 1994). The characteristics of the communities vary significantly between the two polar regions, a consequence of differing mechanisms for ice growth (Spindler, 1990), and the spatial and temporal distribution of these organisms is controlled by both physico-chemical and biological factors (Eicken, 1992a). Brine channels play a major role in sustaining sea ice biota by providing an optimised habitat in which the communities can develop in the ice (Thomas, 1996). The ecology of the organisms is, to a large extent, governed by environmental factors (Horner, 1985a; Weissenberger, 1992) and the habitable space (Weissenberger et al., 1992) which can act to structure the communities according to the species size (Gleitz and Thomas, 1993).

As a habitat, brine channels present many advantages to biota residing within them. They provide a substrate onto which they can attach themselves close to the ice-water interface; here they are regularly supplied with nutrients and oxygen via the exchange of brine with the underlying water column. Further, they are held at the top of the water column where the light intensity is greatest, they are not subjected to downward mixing and by living inside the brine channels they are protected from predation by the organisms present in the upper levels of the water column. The predators and grazers capable of living in the ice benefit by being effectively surrounded by their captive prey.

However, compared to the pelagic environment, the sea-ice environment undergoes dramatic changes in its physical, chemical and biotic conditions. Substantial increases in the brine salinity and a decrease in the temperature can put the organisms under tremendous osmotic and other physiological strain. During freezing, the space available for habitation decreases rapidly and reduced permeability decreases the potential for brine exchange with a resultant decrease in nutrient concentrations and an increase in waste toxin concentration. Given that sea ice constitutes a significant portion of the biological cycle in the polar oceans, attempts have been made to model examples
of this habitat (Arrigo et al., 1993). Amongst the factors that need to be considered is the ever changing nature of the brine structures and salinity distribution in the ice.

3.5.5 Mechanics

The mechanical properties of sea ice is an area of research that has received much attention (Mellor, 1986). There are numerous relations that have been developed to link the mechanical behaviour of the ice to its brine content and more specifically its brine volume. The tensile strength of first-year sea ice was shown to have a temperature dependence (Richter-Menge and Jones, 1993) attributable to the changes which occur in brine volume and distribution. It has been known for some time that the failure plane of sea ice during flexural tests is along the line of the brine inclusions between the ice platelets, the c-axis fabric defining this obvious line of weakness (Shapiro and Weeks, 1993).

Like the brine inclusions, brine channels also act as flaw structures in sea ice; they are important structural elements of the ice. Therefore they are also a factor in determining the flexural strength of a sea ice cover (Dempsey, 1996). Whilst the c-axis fabric determines the strong and weak directions of the ice, this is further influenced by the occurrence of brine channels where they will often act as the point of nucleation for a fracture (Shapiro and Weeks, 1995). Furthermore, the brine channels may act to moderate the fracture velocity by dissipating energy to the mobile brine (Dempsey et al., 1999) and even change the fracture direction (Weeks, 1998). Even at low ice temperatures brine channels can influence the fracture properties of the ice. Later in the year, as the ice warms and the brine channels enlarge, sea ice becomes weaker thus accelerating its break-up (Shapiro and Weeks, 1993). The effect of brine channels on the strength of sea ice is an emerging field of ice-engineering research in which the scale effects of the various structural elements are being formalised.
3.6 Summary

This chapter has reviewed brine drainage channels in sea ice. Their static structural characteristics are adequately known following numerous investigations which have employed a range of imaging techniques. However, the precise structural evolution of brine channels which occurs under a varying temperature field remains unresolved, though state of the art 3-D imaging techniques may prove valuable in this area of research. Although brine channels have been studied for many years, no established conventions in terminology exists. To resolve this, a review of current terminology was undertaken and a systematic classification scheme and nomenclature for brine features in sea ice, based on simple geometric contrasts, was proposed.

Although many of the individual steps and variables which control brine channel development have been described previously, there has been no formal attempt at collating this into a life history of brine drainage channels. A novel aspect of this chapter was the development of the hypothesis that it is the salt content of sea ice which controls their spatial distribution; this will be pursued in subsequent chapters. The chapter concluded with a discussion of the sea-ice processes in which brine channels have a controlling influence. In view of the effects that brine channels have on these important processes, it is necessary to clarify the understanding of the fundamental influence brine channels have on the brine distribution in sea ice. This provides the justification for investigating a more detailed analysis of this aspect of young sea ice.
Chapter 4
Experimental conditions and methods

4.1 Introduction

Chapter 4 describes the two experimental periods during which the data for this thesis were collected. The experiments were performed in a large ice tank as part of an interdisciplinary European project. To make a detailed study of brine distribution it was recognised that an ice tank, although an artificial environment, provides the necessary control over the boundary conditions of ice development. A brief review of the value of ice tank studies for sea-ice research is given in section 4.2 whilst section 4.3 gives a more specific account of the tank which was used and of the two Interice projects, of which this work formed a part. During the experimental periods the environmental conditions and the bulk properties of the ice were measured and these are described in section 4.4. It is important to establish that the ice which forms in the tank is representative of naturally occurring sea ice. Arguments are given to support this, based on the properties of ice formation during each experimental period.

Hitherto, any study of bulk salinity of sea ice has been hampered by the inevitable loss of brine from the ice during sampling. During these experiments a novel, yet simple, means of sample acquisition was developed which minimises brine loss and preserves its spatial distribution; a detailed methodology is presented in section 4.5. All the bulk salinity data that will be presented in subsequent chapters are derived from this methodology and therefore a description of the data processing and presentation that has been adopted is given.
4.2 Ice tank studies

Ice tanks have been used for the simulation and observation of many sea-ice processes and have had an important role in the development of the understanding of sea ice, its properties and its interaction with the natural environment. Areas of research which have utilised ice tanks include mechanical deformation testing (Tuhkuri and Lensu, 1998), crystal growth (Langhorne, 1983; Kawamura, 1987), oil spill technology (Løset and Timco, 1992), wave propagation (Newyear and Martin, 1997), sea-ice biology (Grossmann and Gleitz, 1993; Weissenberger, 1998; Weissenberger and Grossmann, 1998) satellite remote sensing (Swift et al., 1992) and sedimentology (Reimnitz et al., 1993; Ackermann et al., 1994). Ice tank studies of brine drainage processes were pioneered by Cox (1975) and later by Martin and his associates (Eide and Martin, 1975; Niedrauer and Martin, 1979). In the 1980s, Japanese scientists were at the forefront of utilising tanks for studying the microstructure of sea ice, particularly brine expulsion (Wakatsuchi and Ono, 1983), brine channel formation (Wakatsuchi and Kawamura, 1987; Kawamura, 1988a) and their structure (Kawamura, 1988b). More recently, the mechanisms which control the fate of interstitial brine during freeze-up and subsequent desalination of sea ice have been investigated using tanks (Wettlaufer et al., 1997a).

The raison d'être for using tanks to study sea-ice processes, is to mimic the natural situation as closely as practicable whilst maintaining a close control on the environmental conditions and therefore growth history; this simplifies the physical interactions and hence facilitates interpretation. Regulation of the air temperature, shear current velocities, wind fields, wave spectra and in some cases the crystal orientation (Kawamura, 1986), allows a wide range of naturally occurring ice types and growth conditions to be simulated. Inherent to having this control, and in contrast to field sampling, the formation history of the ice is fully documented and repeatable. Furthermore, the versatility and accessibility afforded by tank experiments allows for the installation of sensors and sampling devices prior to freeze-up, a situation which is usually precluded in field experiments.
Chapter 4: Experimental conditions and methods

The physical environment within any tank is affected by the fact that it is necessarily a closed system. Thus, ice tanks have a number of serious disadvantages which limit their use. Of these, salination of the water in the tank is perhaps the most important, also unnatural stresses in the ice, horizontal heat fluxes and the loss of isostasy; though these problems may be eliminated or minimised in a carefully designed tank. A fundamental limitation of ice tanks is that they are constrained in scale, both spatial and temporal. The finite volume of ice available restricts the frequency and size of samples which can be obtained whilst the duration of experiments is relatively short compared to natural seasonal periods.

4.3 The experimental facility and projects

This section gives a description of the experimental facility and the details of two Interice projects during which the data for this thesis were gathered. The research was conducted in the Arctic Environmental Test Basin (AETB) at the Hamburgische Schiffbau- und Versuchsanstalt (HSVA), Hamburg, Germany; this will be referred to as the tank. The Interice projects included several interdisciplinary studies: sea-ice physics and biology, sedimentology and physical and chemical oceanography. A principal aim of the Interice projects was to establish the feasibility of using a large ice tank to undertake fundamental research which might decouple the various interactions occurring in a natural sea-ice system.

4.3.1 The ice tank facility

The arrangement of the tank during the Interice projects is shown in Figure 4.1 and was virtually identical for both experimental periods. The tank is 30 m long and 6 m wide with a water depth of 1 m. Tank coordinates were x along its length, y across its breadth and z for depth. The origin is located in the bottom left hand corner of Figure 4.1 and the positive z direction is downwards. The tank was divided into two compartments, the calm zone and the current zone, by a full-depth partition wall at x = 8.5 m. The current
Chapter 4: Experimental conditions and methods

zone was further subdivided along the centre line with a partition wall from \( x = 13 \) m to \( x = 27 \) m. The temperature and salinity of the water were measured with a mini conductivity-temperature-depth (CTD) instrument. A set of curved baffles located downstream of the impellers created a laminar shear current in the lower portion of the current zone in Figure 4.1 (the instrument side of the partition wall). Current velocity was recorded with an ultrasonic current meter (UCM). Air temperature was regulated by overhead chilling units. Additional facilities included a wave machine and a wind generator for use in experiments requiring dynamic ice growth, though these were not used during the sampling periods pertinent to this thesis.

![Diagram](image)

**Figure 4.1:** *A floor plan of the Environment Test Basin at HSVA.*

Ice growth was initiated, or seeded, with ice fog (Evers and Jochmann, 1993) which produced an initial ice cover of uniform crystal size. Once a congealed sheet of ice had formed, with a thickness of approximately 3 cm, the impellers were operated. This avoided disintegration of the fragile ice skim during the first hours of ice growth. The relatively large surface area of ice in the tank permitted numerous samples to be collected without significant perturbation of the vertical heat flux. As with all closed system experiments, salination of the water was active as the ice thickness increased though this was not compensated for. The ice sheet adhered to the walls of the tank so it was not free to float according to its natural isostasy. Therefore, as the thickness of the sheet increased pressure would develop underneath it causing internal flooding of the ice. To prevent this, and to maintain a natural isostasy, water was pumped out of the tank regularly. The volume of water removed was equivalent to 10% of the mean increase in ice thickness, according to the density difference between seawater and ice.
4.3.2 The Interice projects

The Interice project was a joint programme of experiments between several European institutions. The project was financed by the Commission of the European Communities through the Large-Scale Facility Programme.

4.3.2.1 Interice I

The Interice I project ran from November 1996 to February 1997 and was arranged into three experimental phases; sea-ice physics, physical-biological interactions and dynamic-sedimentology processes (Eicken et al., 1998). Part of the data set for this thesis was collected during the physical-biological interactions phase between 7 and 31 January 1997, this will be referred to as the January 1997 experimental period. The biological activities during this period required that the water in the tank closely represented natural seawater and contained the same essential salts and nutrients. Therefore, the water was prepared from tap water and synthetic sea salt (Instant Ocean, Aquarium Systems Ltd.) with an initial salinity of approximately 32 psu. The experiment attempted to simulate the life cycle of an ice sheet beginning with a winter growth period followed by a spring stagnation period with no change in ice thickness and finally concluding with a summer melt phase. All samples were obtained from the ice growing in the current zone of the tank (shear velocity = 4 cm s⁻¹).

4.3.2.2 Interice II

The Interice II project ran from November 1998 to December 1998 and the remainder of the data for this thesis were collected between 9 and 24 November. This phase of experiments was primarily concerned with sea-ice physics and will be referred to as the November 1998 experimental period. Without the biological requirements for the salt composition of the water, the tank was filled with a solution of tap water and NaCl salts which forms a suitable approximation for seawater. The temperature cycle for November 1998 was intended to simulate a refreeze, typical of that occurring in late summer or early autumn following a brief spell of increased air temperatures. An initial period of ice growth was followed by a short melt phase during which time the ice
became isothermal. Refreeze was initiated with low air temperatures during which a temperature gradient in the ice was re-established and ice growth continued. Most of the samples were obtained from the ice in the current zone (shear velocity = 7 cm s\(^{-1}\)) whilst the remainder were obtained from ice grown in the calm zone.

### 4.4 Bulk parameters

Throughout each experimental period, the environmental conditions of the air-ice-water system were monitored continuously. These measurements define the physical conditions under which the ice formed and evolved and establish a basis for more detailed investigation of brine distribution. From these bulk parameters, it was possible to establish that the sea ice forming in the tank has similar growth and structural characteristics to natural sea ice.

#### 4.4.1 Air temperature

The air temperature was measured from sensors suspended along the tank; their precise number and location varied between the two experimental phases. Air temperatures were recorded as an analogue output which was then digitised. The accuracy of the measurements was better than ±0.5°C which is less than typical fluctuations caused by occupation of the tank during sampling.

##### 4.4.1.1 January 1997

The air temperature time series for the experimental period in January 1997 is shown in Figure 4.2 and plotted against the number of days after initial seeding of the ice. The temperature data has a temporal resolution of 3 hours and are the mean of four sensors spaced equally along the length of the tank and suspended 1.5 m above the ice surface. The air temperature during the period of maximum cooling in the growth phase was approximately \(-18°C\) for a duration of 8 days. In the stagnation phase the temperature was adjusted from \(-4°C\) to \(-6°C\) to maintain a near constant ice thickness. The cooling
system was switched off after 16 days after which the air temperature increased substantially to initiate the melt phase. Each phase of the experiment, as described in section 4.3.2.1, is demarcated in Figure 4.2 with vertical dotted lines.

![Figure 4.2: The air temperature in the ice tank - January 1997.](image)

### 4.4.1.2 November 1998

The air temperature time series for the experimental phase in November 1998 is shown in Figure 4.3 and plotted against the number of days after the initial seeding of ice in the current zone. The temperature data have a temporal resolution of 2 hours and were derived from a single sensor suspended 1.0 metre above the ice surface in the current zone at \( x = 15 \) m. The air temperature of \(-15^\circ C\) during the first growth phase was warmer than for the corresponding phase in January 1997. This resulted in a slightly slower growth rate which will be discussed in section 4.4.4. The cooling was switched off after 6 days to initiate a melt phase which lasted for just over 2 days. Before recommencement of cooling, warm brine was circulated in the chiller elements to remove frost which had accumulated on them, and thus improve their efficiency; this accounts for the high temperature spike. Much colder conditions were experienced in the second growth phase with the restoration of efficient cooling when air temperatures went below \(-20^\circ C\). The experiment ended on day 14. Again, the three phases of the experiment, as described in section 4.3.2.2, are demarcated with vertical dotted lines.
Figure 4.3: The air temperature in the ice tank - November 1998.

4.4.2 CTD data

CTD measurements of salinity and temperature at a fixed depth in the tank were made every minute from which hourly means were calculated. The water column in the current zone was well mixed at all depths and can be regarded as a single, slowly varying, homogeneous mass. The instrument had a precision of ± 0.002 psu for salinity and ± 0.002°C for temperature.

4.4.2.1 January 1997

Figure 4.5 shows the water temperature as a function of the number of days after the initial seeding of ice. The solid line is the CTD measured data and the dash-dot line is a linear extrapolation of the data to fill the gap in the record between days 22 and 24. This is justified because the melting during this period proceeded without disturbance. A rapid warming of the water is seen at the beginning of the melt phase with the increase in air temperature. The step in water temperature to −1.65°C on day 24 resulted from refilling the tank with warm water at the end of the experiment.
Chapter 4: Experimental conditions and methods

Figure 4.5: The water temperature time series - January 1997.

The salinity time series is shown in Figure 4.4 as a solid line with interpolated and extrapolated data as a dash-dot line. Transient features in the CTD record between days 5 and 15 were most probably the result of frazil crystals becoming trapped in the conductivity cell leading to a decrease in the measured conductivity and therefore lower salinity. The crystals then either melted or were flushed out of the cell yielding a resumption of valid salinity measurements. Between days 5 and 11, a second order polynomial was fitted between those data points plotted with ‘o’ which represent

Figure 4.4: The water salinity time series - January 1997. Data points used for interpolation from day 5 to 14 are marked as ‘o’.
maxima in the original data. Between days 11 and 15, in the stagnation phase, a linear interpolation was calculated. This approach is valid as no major changes in the CTD data are expected in this phase of the experiment which was confirmed by the temperature data for the same period. Like the temperature data, the salinity data were linearly extrapolated between days 22 and 24.

4.4.2.2 November 1998

Figure 4.6 shows a time series of the water temperature as a function of the number of days after the initial seeding of ice in the current zone. The rapid increase in water temperature during the melt phase is in response to the increase in air temperature at that time and is similar to that observed in Figure 4.5 after day 15. With the resumption of freezing, the water temperature then decreased throughout the remainder of the experiment.

Figure 4.6: The water temperature time series - November 1998.

The salinity time series is shown in Figure 4.7. There is a steady increase in salinity during both growth phases interrupted by a slight decrease in salinity during the melt phase indicating that ablation of the ice occurred. As anticipated, the general form of the salinity data is comparable to that of the ice thickness data presented later in Figure 4.9. The dip at the end of the time series is the result of fresh water being pumped into the tank.
4.4.3 Ice thickness

During both experimental periods, ice thickness measurements were made on a near daily basis in both the current and calm zones. The thickness transects involved drilling holes of 10 mm diameter and then measuring the ice thickness using a rigid rule. It was found that there were significant differences in the ice thickness on each side of the partition wall in the current zone. The data presented here will be restricted to measurements on the instrument side only as this is where ice samples were obtained. Mean values are plotted with 95% confidence limits.

4.4.3.1 January 1997

The ice thickness transect was located at $x = 20.5$ m and the data consist of 5 measurements spaced at 0.5 m intervals from $y = 0.5$ m to $y = 2.5$ m. The results of the measurements during the experiment are shown in Figure 4.8. Initially, the ice growth was rapid and the thickness reached a maximum of 22.3 cm about 10 days after seeding. During stagnation and melt, the ice thickness decreased slowly and uneven ablation, due to variable heat fluxes throughout the tank, led to an increase in the standard deviation associated with the mean thickness.

Figure 4.7: The water salinity time series - November 1998.
4.4.3.2 November 1998

The ice thickness data consist of 2 transects located at $x = 15.5$ m and $x = 21.25$ m. Each transect comprised 3 measurements at 1.0 m intervals from $y = 0.5$ m to $y = 2.5$ m. The results of the measurements during the experiment are shown in Figure 4.9.
Chapter 4: Experimental conditions and methods

Growth rates during the initial ice formation were slightly slower than for January 1997; this is attributable to the warmer air during this phase. Ice growth persists into the melt phase for 1 day as the temperature gradient in the ice reaches equilibrium. This is followed by two days of ablation and then rapid growth for the remainder of the experiment in the refreeze phase. As for January 1997, standard deviations increase as the experiment progressed and with the onset of melt.

4.4.3.3 Comparison of observation and theory

Because the primary data on brine distribution are obtained from sea ice that formed in an artificial environment, it is important to establish whether the kinetics of ice growth in the tank are comparable to those found in nature. At the margins of the ice sheet, up to 30 cm from the tank walls, it was significantly thicker than the bulk of the sheet due to enhanced heat fluxes through the walls. Otherwise, ice thickness in the current zone was quite uniform, particularly during the early days of growth.

The thickness of sea ice growing under natural conditions may be linked empirically to the duration of growth where the time coordinate is often taken to be the cumulative number of freezing-degree days (Maykut, 1986). As the ice grown in the tank has no snow cover, it is of little value to compare the tank data with field data directly. A number of ice growth models exist (Leppäranta, 1993) but although the tank represents a simplified system, accurate modelling of the ice growth is not straightforward because of the complex nature of the heat budget in the tank and the phase changes occurring within the thin ice.

However, the basic kinetics of uniaxial ice growth are described by a simple analytical relationship from which the ice thickness is found to be proportional to the square root of growth period (Ashton, 1989). Using the thickness data from the initial growth phase for each experiment, Figure 4.10 shows that the ice growth in the tank follows the $t^{1/2}$ relation and therefore results from a uniaxial heat flux.
4.4.4 Ice growth rate

The growth rate of sea ice is influential in determining the physical properties of the ice including its salinity and crystal size. As the ice thickens and the temperature gradient in the ice decreases, the growth rate also decreases; this was observed during each experiment and is summarised in Table 4.1. The more rapid growth rate in January 1997 is attributed to the lower air temperatures whilst at 120 hours the lower growth rate in January is a consequence of a thicker ice cover (14.8 cm compared with 12.9 cm in November 1998).

<table>
<thead>
<tr>
<th>Time after seeding</th>
<th>January 1997</th>
<th>November 1998</th>
</tr>
</thead>
<tbody>
<tr>
<td>24 hours</td>
<td>0.22</td>
<td>0.18</td>
</tr>
<tr>
<td>120 hours</td>
<td>0.05</td>
<td>0.07</td>
</tr>
</tbody>
</table>

Table 4.1: Ice growth rates in the current zone during the first 24 and 120 hours of each experimental period.
4.4.4.1 Ice growth rate and brine channel distribution

The spatial distribution of brine channels in sea ice was discussed fully in section 3.4.2.2. Brine channel spacing, and hence their spatial density in the horizontal plane, has been linked to the initial growth rate of the ice (Wakatsuchi and Saito, 1985). To confirm that the brine channels within the ice grown in the tank are comparable to those in sea ice grown under natural conditions, a comparison of data concerning the spatial density of brine channels will be presented.

During each experimental period, the spatial density of brine channels, \( D_e \), was calculated from observations of the ice surface in the current zone. In January 1997, two areas of ice were examined and photographed from which \( D_e \) was calculated. The first area consisted of a slab of ice 1.5 m long, 30 cm wide and 20 cm thick which had been cut from the ice on day 10 and raised 5 cm to form a ridge. The enhanced hydrostatic head resulting from this procedure, and elevated air temperatures, increased brine drainage from the channels thus improving the contrast for their observation. The spatial density of the brine channels was then calculated by subdividing the slab and counting the channels in each subsample. A second area of ice was protected from disturbance throughout the experiment and during the melt phase brine channels were clearly visible from the surface as brighter points contrasted against the darker ice. Quantitative measurements of spatial density were made with 25 x 25 cm grids as illustrated in Figure 4.11. In November 1998, a similar method was adopted and measurements were made on day 8 when the ice had warmed sufficiently to reveal the brine channels.

![Image of ice surface with brine channels](image)

**Figure 4.11:** The ice surface with brine channels identifiable as the bright points; remnants of frost flowers are also visible. Grid squares have 25 cm sides.
To compare these observations with the work of Wakatsuchi and Saito (1985), determination of the initial growth rate of the ice was required. In his field measurements, Wakatsuchi calculated growth rate by simply dividing the thickness of new ice grown overnight (typically between 3 and 6 cm) by the duration of ice growth giving a mean growth rate over that period. Similarly, a mean growth rate was calculated for the ice grown in the current zone for a median thickness of 4.5 cm and also for 3.0 cm and 6.0 cm to give upper and lower bounds. A summary of the spatial density measurements and the growth rates is given in Table 4.2 and these data (indicated by □) are presented in Figure 4.12 with those from previous observations.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>January 1997</th>
<th>November 1998</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spatial density of brine channels</td>
<td>6.5 ± 0.5</td>
<td>4.9 ± 0.4</td>
</tr>
<tr>
<td>(number per 100 cm²)</td>
<td>(n = 18)</td>
<td>(n = 5)</td>
</tr>
<tr>
<td>Growth rate at 3.0 cm (× 10⁻⁵ cm s⁻¹)</td>
<td>6.48</td>
<td>5.41</td>
</tr>
<tr>
<td>Growth rate at 4.5 cm (× 10⁻⁵ cm s⁻¹)</td>
<td>5.80</td>
<td>4.80</td>
</tr>
<tr>
<td>Growth rate at 6.0 cm (× 10⁻⁵ cm s⁻¹)</td>
<td>5.25</td>
<td>4.32</td>
</tr>
</tbody>
</table>

Table 4.2: The spatial density of brine channels derived from observations of the ice surface, and the growth rate calculated at three ice thicknesses for both experimental periods.

The curve in Figure 4.12 is derived from the field data of Wakatsuchi and Saito (1985) and from NaCl ice grown in a tank which is small compared to the HSVA tank (Saito and Ono, 1980). It is encouraging to note that the data from the observations made in January 1997 and November 1998 lie on this curve. Although the relation between spatial density of brine channels and growth rate is purely empirical, it is significant to note that the three independent data sets show the same trend. The important conclusion from this comparative exercise is the spatial density of brine
channels in sea ice which formed in the tank during the two experimental periods follows a similar dependence on initial growth rate as that found in sea ice growing naturally. Therefore, in terms of brine channel formation mechanisms, the sea ice from the tank is a good proxy for natural ice.

![Graph showing spatial density of brine channels as a function of initial ice growth rate](Image)

**Figure 4.12:** Spatial density $D_c$ (number of channels per 100 cm$^2$) of brine channels as a function of initial ice growth rate (adapted from Wakatsuchi and Saito, 1985, figure 3). □ Interice data, ○ (Wakatsuchi and Saito, 1985) and ● (Saito and Ono, 1980).

4.4.5 Textural analysis

An example of the crystal texture typical of the ice grown in the current zone during both experimental periods is shown in Figure 4.13. The texture is dominated by large columnar crystals throughout the entire thickness except for the uppermost millimetres which had fine grained crystals originating from the ice fog used for seeding. The strong crystal elongation in the vertical direction provides further confirmation that a uniaxial heat flux existed through the majority of the ice sheet (Perey and Pounder, 1958). The crystal c-axis showed alignment with the current direction in accordance
with the observations and theory described by Langhorne and Robinson (1986). The skeletal layer is well developed and in the current zone was typically 2 cm thick. During the melt phases of both experimental periods, the skeletal layer was ablated.

Figure 4.13: Vertical thin section of the congelation ice fabric seen through crossed polarisers. Image courtesy of C. Haas, AWI. The scale graduations in the top left are at 1 mm.

4.5 Experimental methods

In this section, a detailed description of the methods used for obtaining the data will be described. The results presented in chapters 5, 6 and 7 were obtained using the techniques described here. The aim was to measure the bulk salinity of each sample at a spatial resolution adequate to compare the brine distribution with occurrence of brine drainage channels within the ice. The principal difficulty in any study of the bulk salinity of sea ice is retention of the brine in the sample with its in situ distribution. To this end, a novel method of sample retrieval and preparation was developed which minimised brine loss and enabled detailed examination and processing of the sample.
4.5.1 Sample retrieval

In the majority of cases, sampling was confined to the current zone of the tank whilst the remaining samples were retrieved from ice in the calm zone. Care was taken when collecting samples to avoid extensive perforation of the ice sheet which might have distorted the simple one-dimensional nature of the heat flux. Therefore, samples were taken from a portion of the ice which was at least 0.5 m away from previous areas of open water. The samples were large blocks of ice with variable dimensions; the surface area of the blocks ranged from 200 cm² to 400 cm² with thicknesses from 3 cm to 21 cm.

The technique for sample retrieval will be described with reference to Figure 4.14. To obtain the samples, a plastic box was submerged under the ice via an access hole (A). The sample was then cut and released from the ice sheet and floated into open water until it was directly over the box. The box was then lifted from the tank with the sample floating inside it (B). Once clear of the tank, the water surrounding the sample was carefully drained out until the ice was nearly grounded on the bottom of the box. In effect, the ice remained floating in its own melt without the introduction of an artificial hydrostatic head which would have caused a loss of brine from the sample. The box, with its contents, was then transferred to a freezer and shock-frozen at −40°C, solidifying both the brine within the sample and the surrounding water (C).

![Diagram](image)

**Figure 4.14:** A series of schematic diagrams illustrating the steps taken to retrieve ice samples from the tank.
All subsequent preparation and handling of the frozen sample was conducted below -25°C to ensure retention of the \textit{in situ} brine. The entire frozen mass was removed from the plastic box and the outer solid portion (the frozen water) was cut away (D) leaving the sample of ice with a higher percentage of retained brine than one can expect from traditional retrieval methods. Care was taken to ensure that all the frozen water was removed, particularly traces on the bottom surface. Further, the exposed sides of the sample were trimmed substantially in case of horizontal seepage of water into the sample, though this will be minimal as the lateral permeability of sea ice is very low. Whilst the technique permits a detailed, quantitative study of the sample to be made, it destroys the \textit{in situ} composition of the brine channels. Therefore, any investigation of the changes occurring in the microstructural characteristics of the channels during thermal evolution of the ice is precluded.

Although the rapid, shock-freezing process clearly destroys the original \textit{in situ} temperature gradient and subjects the sample to an extreme temperature cycle, it is likely that the spatial distribution of brine is not altered significantly. Any relocation of brine by expulsion will occur as the cold front penetrates through the sample. However, microstructural analysis of similarly shock-frozen core samples shows no direct evidence of significant brine relocation (Eicken, \textit{pers. comm.}, 1998). As will be described in the next section, each sample is ultimately subdivided into cubic subsamples of side 2 cm. Assuming that brine expulsion is not occurring on scales significantly greater than 2 cm then brine will remain within an individual subsample. This assumption is reasonable given that brine expulsion is a mechanism which usually occurs on a scale comparable to the crystal size in a slowly varying one-dimensional temperature gradient. Therefore, any relocation of brine that is confined within an individual subsample may be disregarded for the purposes of this study. Further, any expulsion of brine from a subsample is likely to be compensated for by an equivalent expulsion of brine into the subsample yielding a negligible net brine loss. The data presented in chapter 5 will substantiate these arguments relating to the minimal distortion of the \textit{in situ} brine distribution by shock-freezing.
Another potential source of distortion to the brine distribution resulting from this method relates to the small amount of water from the tank which occasionally spilled onto the upper surface of the ice during the removal of samples. For most samples, the ice at the upper surface was very cold and therefore of low permeability restricting the potential for penetration of water into the sample. However, even in the case of warm, porous ice this source of contamination only affects the upper few millimetres of the ice before complete solidification of the brine terminates further percolation into the sample; this will be illustrated in chapter 5.

4.5.2 Sample processing

To link the distribution of brine with the occurrence of brine drainage channels it was necessary to observe the channels in the ice samples, and record their geometry, prior to melting. Additionally, it was advantageous to obtain structural information about the brine channels in both the horizontal and vertical planes. To achieve this, the following processing method was employed in which the sample was systematically subdivided. For clarification of the following description of the subdivision process, refer to Figure 4.15.

![Diagram of sample subdivision](Image)

**Figure 4.15:** *The subdivisions of each sample to obtain 2 cm³ subsamples.*

The recovered and prepared blocks of ice were cut into a series of horizontal plates, 2 cm thick, using a band saw. Each plate constituted a large, thick section in which brine channels were clearly visible. To see the brine channels with sufficient contrast, the plates of ice were placed on a black background (cf. Wakatsuchi and Saito,
1985), illuminated and photographed under incident light. This method allowed the brine channels to be observed clearly as bright areas against the much darker solid ice matrix, such as those seen in Figure 4.16.

![Figure 4.16: A typical image of a horizontal plate of ice showing the brine channels as bright, star-shaped features. Horizontal width of the image is 22 cm.](image)

Each horizontal plate was then cut into rows 2 cm wide. The corresponding rows from adjacent depths in the ice were arranged sequentially so as to reconstruct a series of vertical thick sections which show the brine channels and the associated feeder channels, an example is reproduced in Figure 4.17. An obvious limitation of the processing method is that the vertical thick sections for each sample can be reconstructed in only one plane. Each row was further subdivided into cubic subsamples with sides of 2 cm. The subsamples were melted individually and the salinity of each was measured with a salinometer to obtain their bulk salinity.

![Figure 4.17: A typical reconstruction of a vertical thick section clearly showing three brine drainage channels. Horizontal width of the image 26 cm.](image)
Chapter 4: Experimental conditions and methods

The choice of the spatial resolution for subdivision was strongly influenced by two conflicting factors. First, the subsampling necessarily had to be at a sufficiently high resolution to enable any variations in bulk salinity, associated with the brine channels, to be discerned. Therefore, the subdivision had to be on a scale comparable to the spacing of the brine channels. The second, and opposing factor, was the large number of subsamples which would be created if the resolution was set too high. Increasing the spatial resolution from 2 cm to 1 cm produces an eight-fold increase in the number of subsamples. Since considerable manual handling was involved in their preparation and measurement, it was important to find a compromise between resolution and sample number. The choice of a 2 cm grid as the optimum sampling resolution for the purposes of this work is justified in section 5.3.1.2.

Closely allied to the choice of sampling resolution was the sampling strategy. Again a balance had to be struck between obtaining sufficient samples to generate representative data yet the practicalities of sampling an entire slab precludes this. The approach adopted was to ensure that a complete record of bulk salinity was acquired in each of three orthogonal planes. Horizontal sections were selected at a consistent depth interval in the ice to allow comparisons to be made between the samples. The selection of the two orthogonal vertical sections was made by examining the horizontal plates prior to their subdivision. Using a transparent grid it was possible to map the positions of the brine channels in each plate and from the map determine which vertical planes would contain brine channels thereby ensuring that the most relevant data were obtained. The remainder of the subsamples which did not lie in any of these planes were combined and melted according to their depth in the ice. From these, the true bulk properties of the ice were determined.

4.5.3 Measurement of bulk salinity

The bulk salinity of each subsample was measured using a conductiometric probe with automatic correction to a reference temperature of 20°C. The same instrument and probe combination was used during both experimental periods and was calibrated against a sample of standard seawater with a guaranteed salinity of 34.996‰ (which for practical purposes can be taken as 35 psu). The standard was diluted in aliquots with de-ionised
water and the salinity was calculated and measured at each dilution. The results of this
dilution method are presented in Figure 4.18 which shows a linear response from the
instrument.

![Figure 4.18](calibration_curve.png)

**Figure 4.18:** Calibration curve for the salinometer. The dashed line is a least squares
fit to the data of the form \( y = 1.063 \times \) with \( R^2 = 0.9995 \).

This study is based upon observed differences in bulk salinity throughout a
particular sample of ice, therefore absolute values of bulk salinity are not critical.
However, the data were corrected accordingly and it was assumed that the response of
the instrument was similar during each period of use.

### 4.5.4 Data processing

The majority of the data which will be presented in the following chapters follows
similar data processing and presentation methods. Therefore, at this juncture it is useful
to explain the data processing and presentation that have been adopted.

In subsequent chapters, the salinity data for each sample will be presented in two
ways. First, raw bulk salinity data will be presented as a simple matrix comprising the
individual salinity measurements of the subsamples in the three orthogonal planes, as
described in 4.5.2. Second, adjacent to the raw data, an interpolated form of the bulk salinity measurements will be shown, overlaid by the same grid used for the matrix of raw data. Interpolation of the raw data was performed using a simple cubic interpolation routine which increased the spatial resolution of the bulk salinity matrix from 2.0 cm to 0.1 cm. The interpolation method maintains a continuously varying value for the salinity and the rate of change in salinity. Whilst it is appreciated that interpolation of the data will not necessarily produce an accurate representation of the real changes in salinity, the principal advantage is that it produces a more graphic representation of the bulk salinity data. This is valuable for observing and comparing the variation of salinity in the sections more easily. Implicit to interpolation is the generation of values which have not been measured physically, therefore any statistics are calculated from the raw data alone. Further discussion of the validity of this method of data processing, with reference to the data sets, is presented in section 5.3.1.3.

The photographs of the relevant sections of ice were scanned and modified by image processing software to improve contrast. For clarity, the brine drainage channels were highlighted where required in images which displayed insufficient contrast. To make the essential comparison between bulk salinity and the occurrence of brine drainage channels, isohalines were generated from the interpolated bulk salinity data. These contours of equal salinity were then superimposed onto the corresponding photographic images of the ice. During the preparation of the samples, 1 mm of ice was lost in the saw cut which was compensated for by shifting the length scale of the salinity data. The result is an effective sample side of 2.1 cm which becomes significant for samples located far from the origin where the discrepancy becomes greater than 1 cm. This correction ensured that the origins of the salinity data and photographic data coincided and that their length scales corresponded. By utilising this technique of superimposing the data sets it is possible to compare the salinity distribution to the incidence of brine channels structures directly.
4.6 Summary

This chapter has given a complete account of the experimental conditions encountered during two periods of the *Interice* projects when data were collected. It has been established that the sea ice which forms in the tank is representative of sea ice growing under natural conditions in its growth kinetics, crystal structure and brine channel distribution. This is an essential element of the work on which the validity of the data depends. A novel method of sample retrieval was described, which minimises brine loss from the ice and preserves its spatial distribution. Under appropriate conditions, the sample can then be processed with ease to acquire the maximum amount of pertinent data. Some aspects of the method require certain assumptions to be made. Although these assumptions were touched upon in section 4.5, a more complete discussion of their validity will be given in chapter 5 where reference will be made to the actual data sets.
Chapter 5
Brine redistribution during thermal forcing

5.1 Introduction

Thermal forcing is the term used to indicate the response of sea ice to substantial changes in the air temperature. As discussed in section 2.2.2, varying the temperature of sea ice shifts the phase equilibrium of the ice/brine composite which, in turn, changes the total porosity of the ice. It is also known that changes in porosity will modify the permeability of the ice and thus its capability of transporting brine. Therefore, it is anticipated that thermally induced changes in the ice porosity will be accompanied by a redistribution of brine throughout the ice. The aim of this chapter is to determine how the salt content of young sea ice is distributed in samples having different temperature profiles and thermal histories. In section 3.5 it was stated that brine channels control many sea ice processes. It is hypothesised that the channels will also influence the spatial distribution of salt in sea ice. At the in situ temperatures of the ice sampled during these investigations the majority of the salt content was brine.

To investigate the potential redistribution of brine occurring in young sea ice as it is subjected to thermal forcing, two regimes were considered. First a melt was simulated in which actively growing sea ice underwent substantial warming to the point where ablation became dominant (Cottier et al., 1999). This forms the first part of the data set for this chapter and is presented in section 5.2. From the results of this investigation, a more complete justification of the sampling methods and data processing is offered in section 5.3.1. The significance of the results, in both the evolution of the distribution of brine and its migration through the ice, is discussed in section 5.3.2.
Chapter 5: Brine redistribution during thermal forcing

The second part of the data set, presented in section 5.4, considers refreezing, the reverse thermal forcing process to melting. In a similar manner to that of the melt phase, two samples of ice were obtained, separated by a period of low air temperatures. Ultimately, the redistribution of the brine occurring in response to varying air temperatures might be explained by mechanisms of brine migration, a consequence of changes in the ice porosity.

5.2 Melt

A melt phase in young sea ice would typically be encountered in polar regions during late spring. Whilst air temperatures are low, advection of the ice pack by wind and currents, and the formation of leads, would allow ice to form in the open water. Soon afterwards, a period of higher air temperature would terminate growth and induce melting of the young ice. This is the natural situation which was simulated in the final stages of the January 1997 period during Interice I. The two samples of ice which will be examined were collected from the current zone on days 8 and 24 of the experiment. They were separated by a stagnation phase and a melt phase defined by the changes in air temperature, indicated in Figure 4.2.

To ensure maximum contrast in the brine distribution, samples were taken from the ice cover when it was in quite different physical states. For ease of identification, the two samples described will be referred to as cold ice and warm ice according to their vertical temperature profiles. Cold ice is defined for this purpose as ice with a positive temperature gradient (temperature increasing towards the ice-water interface) resulting in a negative heat flux (directed upwards towards the air-ice interface). Therefore, the ice is growing and possesses a well developed skeletal layer. In contrast, the term warm ice is used to describe ice which is either isothermal or has a negative temperature gradient. In this case, the heat flux is towards the ice-water interface where ablation is active. Warm ice is melting and thus devoid of a dendritic skeletal layer.
Chapter 5: Brine redistribution during thermal forcing

Three types of data will be presented for each sample of ice. First, the bulk parameters which define the physical condition of the ice will be described. These include profiles of temperature and bulk salinity from which the brine volume can be calculated. In many ways, these bulk profiles are similar to the profiles of natural sea ice derived from core segments. Second, matrices of the raw bulk salinity values of each subsample in three orthogonal planes will be presented with the interpolated forms of the data adjacent to them. Third, composite images showing the photographed sections of ice and the internal brine structures, superimposed with the bulk salinity data, will be presented. When describing the features in these data, they will be referred to using the coordinate system of each figure axis.

5.2.1 Cold ice

The sample of cold ice was obtained at the end of the growth phase on day 8 of the January 1997 period when the ice sheet was 21.5 cm thick. The vertical profiles of the bulk parameters which characterise this sample are given in Figure 5.1.

5.2.1.1 Bulk profiles

The temperature profile shown in Figure 5.1(a), was determined from measurements made at 25 mm intervals using a digital thermometer with a resolution of ± 0.1°C. The profile is virtually linear indicating that the ice is in steady thermal equilibrium with the tank environment. Ice temperatures increase from −7.3°C just below the air-ice interface to −1.9°C close to the ice-water interface. This slope represents a mean temperature gradient in the growing ice of +0.27°C cm⁻¹ producing a negative heat flux through the ice and therefore ice accretion at the ice-water interface. Observation of the interface confirmed that a well developed skeletal layer was present.

The bulk salinity profile shown in Figure 5.1(b) follows the C-shaped form reported in previous investigations of salinity profiles of first year ice (Nakawo and Sinha, 1981) and which was discussed in section 2.5.3. The high bulk salinity found in the surface layer is attributed to the natural increase in salinity in the upper part of young sea ice caused by a combination of higher growth rates and brine expulsion.
Chapter 5: Brine redistribution during thermal forcing

**Figure 5.1:** Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for cold ice sampled on day 8 of Interice I, January 1997.

Furthermore, during the course of the experiments, frost flowers were abundant, covering much of the ice surface. Through a process of wicking, frost flowers may draw brine to the upper surface and thus increase its bulk salinity (Perovich and Richter-Menge, 1994); it is likely that this is also occurring in the laboratory-grown sea ice. Between depths of 3.5 cm and 18 cm the bulk salinity remains between 5.6 and 7.8 psu and the entire profile is consistent with field observations of bulk salinity in growing leads (Gow et al., 1990).

The vertical profile of brine volume, Figure 5.1(c), was calculated using the equations of Cox and Weeks (1983) adopting the assumption that the gas content of the ice was negligible (Eicken, pers. comm., 1998). This profile indicates that the brine volume in the upper half of the ice is consistently about 5% with a sharp increase at the surface. As the ice temperature increases with depth so the brine volume steadily increases to a maximum of 23% at the porous skeletal layer. At depths from 14 cm to the ice-water interface, the brine volume is greater than the critical value of 5% identified by Cox and Weeks (1975); thus ensuring that the ice is permeable and capable of undergoing gravity drainage as described in section 2.6.3.
5.2.1.2 Bulk salinity

The raw bulk salinity data for the cold ice are shown in Figure 5.2. The horizontal section, Figure 5.2(a), is a 2 cm thick plate of ice located at depth interval between 8.5 and 10.5 cm, (H--H) in Figure 5.2(c) and (e). The mean value of bulk salinity for this layer is 5.7 ± 1.3 psu with a maximum of 9.6 psu and a minimum of 3.4 psu. A feature of this section is that the subsamples with highest bulk salinity tend to be grouped together and are bordered by subsamples having a significantly lower salinity. In some instances the difference in bulk salinity between adjacent samples can be up to 6 psu. In general, the data display a degree of clustering and therefore a heterogeneous distribution of brine in the horizontal plane on the scale of the measurements.

More detailed examination of the horizontal section in Figure 5.2(a) shows two distinct regions of high bulk salinity located at coordinates (7,6) and (12,11). Further areas of relatively enhanced bulk salinity, become more clearly visible in the interpolated form of the data, Figure 5.2(b), and are located at coordinates (16,16), (16,4), (7,17) and around (3,13). All these areas have bulk salinities greater than the mean for the section. The distance between the centres of these discrete regions is quite uniform, with the nearest neighbour separation being consistent at around 7 cm. These areas of ice with enhanced bulk salinity, which are 2–3 standard deviations greater than the mean salinity, are demarcated by regions of ice which have a bulk salinity more than one standard deviation lower than the mean bulk salinity for the section. In general, the distribution of the regions of ice showing maximum salt depletion is midway between the areas of high bulk salinity.

The bulk salinity data from the two vertical sections shown in Figure 5.2(c) and Figure 5.2(e) exhibit similar characteristics. First, like the vertical bulk salinity profile in Figure 5.1(b), and for the same reasons, the subsamples in the surface layer of the vertical sections have consistently high values of bulk salinity. Additionally, both sections have distinct vertical bands of ice with high bulk salinity which extend through the ice sheet. These bands can be discerned in the raw data but are more clearly illustrated in the figures which show the interpolated form of the data, Figure 5.2(d) and (f). Adjacent to these vertical bands are areas of ice which have significantly lower values of bulk salinity where the salinity minima lie at a point approximately midway
Figure 5.2: Raw and interpolated forms of the bulk salinity data for cold ice on day 8 of the January 1997 period; x and y positions in the horizontal plane. NB. Colour scaling for (c) and (e) differs from (a). Depth interval for (a) indicated by line H--H in (c) and (e). Location of the vertical sections (c) and (e) are indicated in (a) by lines A--A and B--B respectively.
between the regions of high salinity. Large differences in bulk salinity are observed between adjacent samples at the same depth, attaining a maximum difference of 6.1 psu.

Figure 5.2(c) shows the raw bulk salinity data for the vertical section of ice located along the $y = 11$ cm coordinate, (A--A) in Figure 5.2(a). The most significant feature in this section is the vertical band of ice with high bulk salinity centred at $x = 11$ cm penetrating through the entire thickness of the ice. Accompanying this feature is another vertical band of ice, centred at $x = 3$ cm, of similar geometric form but of slightly lesser bulk salinity. This secondary vertical band is present in the upper 12 cm of the ice only. The lateral spacing between these two features is 8 cm.

Figure 5.2(e) is the vertical section of ice located along the $x = 7$ cm coordinate, (B--B) in Figure 5.2(a). Only one significant vertical band of high bulk salinity, centred at $y = 7$ cm, can be seen in this section. This band extends into the ice as a distinct feature from the upper surface down to a depth of 12 cm. Below this depth there are no clearly defined vertical bands of high bulk salinity in this section. Within the lowest layer of the ice in Figure 5.2(e) there are three pockets of ice with high bulk salinity that are located at $y = 1$ cm, $y = 9$ cm and $y = 15$ cm, giving a mean nearest neighbour spacing for these features of 7 cm.

5.2.1.3 Brine structures

Using the interpolated form of the raw data in Figure 5.1(b) and (d), a series of isohalines was generated which were superimposed onto the corresponding photographic images of the horizontal and vertical sections. Justification for using interpolated data is given in section 5.3.1.3. The composite images of the ice and brine structures, with the values of bulk salinity for cold ice, are shown in Figure 5.3. For clarity, the grey scale in the images has been reversed so that the brine channels within the ice show as dark areas contrasting with the lighter solid phase

In both Figure 5.3(a) and Figure 5.3(b) it is immediately apparent that in cold, growing ice there is a strong link between the distribution of salt and the presence of brine channels. A closer examination of the two figures demonstrates that the areas of ice with a high salt content coincide exactly with the location of brine channels. In the horizontal plane, Figure 5.3(a), the isohalines are plotted at 1 psu intervals whilst in the
Figure 5.3: Composite images of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for cold ice sampled on day 8 of the January 1997 period.
vertical section Figure 5.3(b), they are at 2 psu intervals. The most significant features in Figure 5.3(a) are the two areas of increased bulk salinity at coordinates (7,6) and (12,11). There is a very clear correspondence between these areas and the two brine channels centred at those coordinates. The brine distribution around the two channels changes rapidly over short distances with salinity gradients as great as 3 psu cm⁻¹. Other areas of enhanced bulk salinity can be seen in the figure and, without exception, they are all in a region of ice through which a brine channel passes. The areas which show significant depletion of salt are found approximately midway between the brine channels.

The principle feature of Figure 5.3(b) is the large channel at x = 11 cm running vertically through the ice. Again, the contours of high bulk salinity coincide with this dominant feature confirming that the band of high bulk salinity, identified in Figure 5.2(c) and (d), is due to the brine channel. A secondary, structure just visible at x = 3 cm and confined to the upper half of the section, can also be correlated with an enhancement in the bulk salinity of the ice at that location. Similarly to the horizontal section, these regions are typified by steep salinity gradients which can be as great as 4 psu cm⁻¹. Between the two vertical channels the bulk salinity reaches a minimum in an area of ice totally devoid of brine channels although small brine pockets can be seen in the image.

In many instances, particularly in the horizontal section, the form of the isohalines follows the morphology of the brine channels. For example, in Figure 5.3(a), the brine channel centred at (12,12) has a feeder channel extending away from the central channel towards the upper-right corner of the image. The distortion of the 6 and 7 psu contours along the line of the feeder channel corresponds to this extension of the brine channel. Further, the brine channel at (2,14) can be traced from y = 11 to y = 16. Similarly, the shape of the 7 psu contour in this region is also orientated in the same direction.

Quite clearly, Figure 5.3(a) and (b) further illustrate that in cold ice there is structuring of the salt distribution in both the horizontal and vertical planes. It is also significant to note that the vertical image shows the inclined feeder channels of the large brine channel on the right of the sample. This feature of brine channel geometry, and its influence on desalination, were discussed in section 3.3.1.1. In this context it reaffirms that the brine channels which form in the laboratory-grown ice are comparable to those forming under natural conditions (Lake and Lewis, 1970; Cole and Shapiro, 1998).
5.2.2 Warm ice

Day 24 of the January 1997 period marked the end of the melt phase when a sample of warm ice was collected from the current zone of the tank. The ice was 16.5 cm thick which is slightly thinner than indicated in Figure 4.8 and results from differences in ablation rates across the tank leading to increased variability in the ice thickness.

5.2.2.1 Bulk profiles

The vertical profiles of the bulk parameters which characterise this sample are shown in Figure 5.4. These profiles confirm that the physical character of the warm ice is quite different from that of the cold ice on day 8. The temperature profile, Figure 5.4(a), was also determined using a digital thermometer and shows a reverse gradient to that in Figure 5.1(a). The warmest part of the ice, at −0.9°C, is towards the air-ice interface and

![Figure 5.4](image_url)

**Figure 5.4:** Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for warm ice sampled on day 24 of Interice I, January 1997.
the temperature decreases steadily, towards the ice-water interface, to a minimum of \(-1.7^\circ\text{C}\). The negative temperature gradient in the sample of warm ice has a mean value of \(-0.05^\circ\text{C cm}^{-1}\) producing a positive heat flux from the upper to lower surface. At the time of sampling, the ice was ablatting, readily confirmed by the marked difference in ice thickness (5 cm) between the warm and cold samples, and the absence of a skeletal layer.

The bulk salinity profile for the warm ice, Figure 5.4(b) is similar to that of the cold ice insomuch that it also exhibits a very high salinity in the surface layer. At this stage in the experiment there were no frost flowers. Furthermore, there was significant freshwater flooding of the ice surface caused by the melting and precipitation of frost which had accumulated on the chilling units and on the surface of the ice. A likely explanation for the high salinity is that it results from water from the tank being spilled onto the top surface of the ice cover during the course of the experiment and whilst sampling. This explanation is reinforced by measurements of stable oxygen isotopes (\(^{18}\text{O}\)) made on cores obtained on days 8 and 21 of the experiment which indicate that the surface layer of the ice became tainted with tank water (Cottier et al., 1999). A contrast between the bulk salinity profiles of the warm and cold ice is found in the layer of ice immediately below the surface, the subsurface layer. In the warm ice the bulk salinity in this layer is nearly half that measured in the cold ice, 3.7 psu compared to 6.7 psu. The reduced salt content in this layer is most likely the result of freshwater flushing during the melt phase.

Due to the relatively high ice temperatures, the brine volume was calculated using the equations of Leppäranta and Manninen (1988). There is a significant difference in the porosity of the warm ice compared to that of the cold ice with the brine volume increasing to about 20% throughout the majority of the ice. Other than the surface layer, brine volume is consistently between 16% and 25%, very much greater than the 5% required for brine drainage (Cox and Weeks, 1975). Due to the coalescence of brine pockets as the temperature of the ice increases, a phenomenon noted in section 2.4.2, it is likely that the enhanced porosity is complemented by an increase in permeability.
5.2.2.2 Bulk salinity

The raw bulk salinity data for the warm ice sampled on day 24 of the January 1997 experimental period are shown in Figure 5.5. The horizontal section, Figure 5.5(a), is a 2 cm thick plate of ice located at a depth interval between 4 and 6 cm, (H--H) in Figure 5.5(c) and (e). In Figure 5.5(a) the positions of the two vertical sections are indicated by the lines marked as (A--A) and (B--B). The horizontal section has a mean bulk salinity of 5.5 ± 0.9 psu with a maximum 7.9 psu and a minimum of 3.3 psu. The salinity scale is identical to that in Figure 5.2(a) and (b). In contrast to the horizontal section for cold ice, Figure 5.2(a), there is no obvious clustering of those subsamples with relatively higher bulk salinity. Further, areas of higher bulk salinity are not consistently bordered by areas of ice with significantly lower bulk salinity.

The horizontal section, showing the raw bulk salinity data for the sample of warm ice, is presented with its interpolated form in Figure 5.5(a) and (b) respectively. A diffuse region with a bulk salinity greater than the mean is situated around coordinates (15,3). From the raw data in Figure 5.5(a) it is difficult to identify other distinct areas of higher bulk salinity, but interpolation of the data reveals more clearly three other maxima in the section, at coordinates (11,15), (3,13) and (7,4). Although these areas are not well defined, they all have bulk salinities greater than the mean. In addition, there are three smaller points which also have a bulk salinity greater than the mean. These are clustered at coordinates (9,10), (14,10) and (11,7). Unlike the horizontal section of cold ice in Figure 5.2(a) and (b), the spacing of these areas which have enhanced bulk salinity is variable, with nearest neighbour distances ranging from 3 to 10 cm; there is no regular organisation in the distribution of these regions. Further, their areal extent is very variable with no consistent form to their boundaries.

The bulk salinity data for the two orthogonal, vertical sections of warm ice are shown in Figure 5.5(c) and (e). These sections are similar in that they both have a very high bulk salinity surface layer which is approximately 5–6 psu greater than the bulk salinity in the rest of the section. Explanations for this enhancement have been given in section 5.2.2.1 during the discussion of the bulk salinity profile. Another common feature between the salinity values in the vertical sections is found in the subsurface
Chapter 5: Brine redistribution during thermal forcing

Figure 5.5: Raw and interpolated forms of the bulk salinity data for warm ice on day 24 of the January 1997 period; x and y positions in the horizontal plane. NB. Colour scaling for (c) and (e) differs from (a). Depth interval for (a) indicated by line H--H in (c) and (e). Location of the vertical sections (c) and (e) are indicated in (a) by lines A--A and B--B respectively.
Chapter 5: Brine redistribution during thermal forcing

layer. At a depth of 3 cm, both sections have a horizontal band of very low bulk salinity of about 3–4 psu, which is attributed to natural desalination and freshwater flushing during the melt phase. Below this band of low bulk salinity there is minimal horizontal variability and the distribution of brine is very much more diffuse than that seen in the vertical sections of cold ice, Figure 5.2(c) and (e).

Figure 5.5(d), shows the interpolated form of the data for the vertical section located along the y = 13 cm coordinate, (A--A) in Figure 5.5(a). Here, there are two very indistinct vertical bands of ice with bulk salinities slightly greater than the surrounding ice. These two features are located in the lower half of the ice at x = 4 cm and x = 11 cm. Although their boundaries are not well defined spatially they are clearly elongated in the vertical direction. The spacing between these features is approximately 7 cm. These structures are really observable only in the interpolated plot of the data, but once identified, the individual subsamples from which they are derived can be found in the raw data, albeit less clearly. In general, areas of higher bulk salinity in the warm ice sampled on day 24 of the January 1998 experimental period are not readily observed. The data do not indicate significant differences in bulk salinity between adjacent subsamples.

Figure 5.5(e) and (f) are located in the plane extending along x = 11 cm, (B--B) in Figure 5.5(a). In this section there are two regions of significantly higher bulk salinity. One of these is a pocket centred at y = 4 cm and at a depth of 12 cm, the other being a much larger area at y = 14 cm and extending from a depth of 9 cm to the ice water interface at 16.5 cm. From the photographs of the vertical sections it is possible to explain the origin of both these high salinity features which dominate the brine distribution. In both instances, the photographs show large brine channels coincident at these locations. Specifically, there is a large and extensive brine channel at x = 11 cm and from y = 16 to 18 cm. The diameter of this channel, close to the ice-water interface, is approximately 6 mm. These regions of the ice, by virtue of the brine structures, have unusually high liquid fractions. This causes the high bulk salinity of the corresponding subsamples.
5.2.2.3 Brine structures

The composite images showing the ice and brine structures in warm ice with the isohalines of bulk salinity, derived from Figure 5.5(b) and (d), are shown in Figure 5.6. Both the horizontal and vertical sections indicate that the coupling between the distribution of brine and the position of the brine channels has become weaker compared with that for cold ice. In the horizontal image, Figure 5.6(a), the salinity contours are spaced at 1 psu intervals. There are two extensive brine channels which can be clearly identified in this image; they are located around (3,13) and (11,16). In the upper part of the image there is an area which appears to be an agglomeration of brine structures without the distinctive branching pattern characteristic of a well defined individual brine channel. Photographs of the vertical sections, which contain this collection of brine structures, show it to be an extensive mass of interconnected channels which penetrate through virtually the full depth of the sample.

In the horizontal plane, Figure 5.6(a), a direct coincidence between high bulk salinity and brine structures occurs at one location only, (11,6). Here the bulk salinity maximum is approximately 1.5 psu greater than the mean for the section. Conversely, other regions of the ice which also contain brine channels do not show any increase in their bulk salinity. Furthermore, there are areas of higher bulk salinity which were identified in Figure 5.5(b), most notably at (13,10) and approximately (15,3), that are located in areas of the ice devoid of any visible brine channels. It would appear from Figure 5.6(a), in contrast to cold ice, that no strong correlation exists in the horizontal plane between the distribution of salt and the occurrence of brine channels in warm sea ice. A further contrast is found in the form of the isohalines. Where previously, in cold ice, the shape of the isohalines was matched closely to the morphology of the brine channels and their feeder channels, in general the form of the isohalines in warm ice is found to be independent of the morphology of the brine channels.

The composite image in Figure 5.6(b) shows the vertical section of ice located along the y = 13 cm coordinate, (A--A) in Figure 5.5(a); the isohalines are spaced at 2 psu intervals. A striking aspect of Figure 5.6(b) is the abundance of brine channels in the lower half of the ice, particularly in the lowest 4 cm. Of these, the most prominent channel is that located at x = 3 cm and which penetrates almost entirely through the ice.
Chapter 5: Brine redistribution during thermal forcing

Figure 5.6: Composite images of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for warm ice sampled on day 24 of the January 1997 period.
Chapter 5: Brine redistribution during thermal forcing

The brine distribution for this section, represented by the isohalines, has a very different character to that of the vertical section of cold ice shown in Figure 5.3(b). Other than in the surface layers of the sample, there are no steep gradients in the salt content of the ice which is seen to vary smoothly throughout the sample. In this case, the isohalines run laterally across the section with the exception of the 7 psu isohaline which deviates at approximately $x = 4$ cm and $x = 10$ cm corresponding to the slight enhancement in bulk salinity which was identified in Figure 5.5(d). The location of this enhancement coincides with the brine channels found between $x = 3$ and $x = 6$ cm, and similarly in the region between $x = 10$ and $x = 13$ cm where two other brine channels can be identified. However, in general there is no strong correlation between the form of the distribution of brine in warm sea ice and the existence of brine channels.

5.3 Discussion

In sections 5.2.1 and 5.2.2, the bulk salinity data from samples of ice at different stages of thermal evolution have been presented. The data were obtained at a spatial resolution comparable to that of the lateral dimensions of brine channels, and were superimposed onto images showing the associated brine channel structures. Superimposition of these two complementary data sets allows them to be compared directly. The discussion will comprise two parts. First, the data that have been presented are suitable for justifying specific aspects of the experimental methods and sampling processing. It is important to establish at this juncture their validity before proceeding with a comparative analysis of the brine distribution. This analysis will occupy the second part of the discussion.

5.3.1 Justification of the methods

In section 4.5, a number of issues concerning the experimental methods were raised. Having presented two contrasting data sets, derived from samples from the same ice sheet when it was in two disparate physical states, it is now useful to address these issues with reference to the data sets.
5.3.1.1 *Distortion of the brine distribution*

Brine relocation during shock-freezing of the samples was identified in section 4.5.1 as a potential source of distortion to its distribution. Although this processes of modification cannot be completely eliminated, the two samples of ice, which were obtained and processed using identical methods, have revealed highly contrasting forms of the brine distribution. Therefore, it is very likely that what is observed in the measurements is indeed representative of the *in situ* brine distribution rather than an artefact of the method. Second, the water which was spilled onto the surface of the warm ice during sampling, causing a high salinity surface layer, did not percolate into the adjacent subsurface layer before shock-freezing despite the high porosity of the ice in this region. Two important conclusions can be drawn from this; first, spillage of water onto the samples will not distort the brine distribution deeper within the ice and second, the anisotropic nature of sea ice permeability (Gosink *et al.*, 1976) implies that distortion due to lateral percolation of water through the less permeable sides of a sample can also be regarded as negligible.

5.3.1.2 *Sampling resolution*

In section 4.5.2, two competing factors were cited as the justification for adopting the spatial resolution used for this study. First, the resolution necessarily had to be sufficient to discern the effect of brine channels on the brine distribution in the ice whilst second, too high a resolution and the number of subsamples acquired increases beyond the limits of practicality. The composite images of cold ice in Figure 5.3 demonstrate that by working at a spatial resolution of 2 cm an instructive comparison may be made between the bulk salinity data and the photographic record of brine structures in a sample. Therefore, a compromise between the spatial resolution and the quantity of subsamples has been successfully achieved.

At 2 cm, the spatial resolution of the subsamples enabled individual channels in the ice to be *captured* because their nearest neighbour spacing is typically between 6 and 8 cm. With a lesser subsample resolution, gradients in the salt distribution would have been minimised as the influence of a brine channel on the salt content of a subsample
would have become smeared throughout the larger volume element. A reduction in the salinity gradients would then have reduced the significance of the observed correlation between the distribution of salt and the locations of brine channels.

Conversely, increasing the size of the subsamples avoided the possibility of the brine distribution responding to variations in bulk salinity from brine features on the millimetre to submillimetre scale. By obtaining measurements of bulk salinity, at a spatial resolution which is comparable to the lateral dimensions of the brine channels, the variations in the brine distribution can be attributed and described with direct reference to both the location and morphology of the channels. In demonstrating this point for cold ice, interpretation of the contrasting brine distribution in warm ice can be made with confidence, knowing that the subsample resolution is optimal.

5.3.1.3 Interpolation of raw data

The raw data presented in sections 5.2.1.2 and 5.2.2.2 are accompanied by an interpolated form of the same data. In section 4.5.4, the reasons for processing the data in this way were given, namely that it provides a more graphic representation of how the salt content of the solid ice matrix varies within it spatially and from the results of the interpolation a set of isohalines could be generated. However, it was also acknowledged that this method may not represent the real, physical variation in the salt content of the ice accurately. The validity of the interpolation method, with respect to its representation of the physical environment of the ice, will be considered.

This discussion is based on the composite images of cold ice, section 5.2.1.3, where the interpolated bulk salinity data are seen to vary smoothly from the low salinity areas surrounding each channel increasing towards a brine-rich core within the brine channel. At this point it is important to recall the scale on which the variations in bulk salinity are being studied. In the areas of cold ice which do not contain brine channels, the salt is located in the small brine inclusions. It is reasonable to assume that on the length scale under consideration (2 cm) these pockets will be distributed uniformly throughout the ice. Each subsample will contain many brine pockets, thus the salinity gradients in these relatively homogeneous areas of ice will be small. Therefore, the
minimal variation in bulk salinity, generated by the interpolation routine, is entirely consistent with the physical reality.

In the vicinity of brine channels, a different concept of the physical environment of the ice is needed. From the images and observations of brine channels, it is anticipated that their in situ composition will be very different compared with that of the surrounding ice. Therefore, both the solid fraction and the brine content within the channels will be significantly different to that outside. If it were possible to measure the salinity changes at the boundary of a brine channel it is anticipated that there would be a very rapid increase in salinity inside the channel. The interpolated form of the raw data and the resulting isohalines, do indeed reproduce this rapid increase in the salt content of the ice in the vicinity of a brine channel. Furthermore, the morphology of the isohalines follows that of the brine channels indicating that the interpolated data resemble the fundamental structure of the material. However, the isohalines will not necessarily represent absolute changes in the salt content of the ice. It should be noted that these statements are purely conjectural and no specific investigation of the brine channel microstructure has been made with which to support them. However, the form of the salt distribution derived from interpolation of the raw bulk salinity data does conform to the nature of the ice thus providing a valuable and reliable adjunct to the primary data set.

### 5.3.2 Brine distribution

In this section the characteristics of the brine distribution, observed in the two contrasting samples of ice, and the changes which it undergoes during thermal forcing, will be discussed. These characteristics of the distribution will be explained by analysis of the brine structures present in the ice and of the physical parameters of the ice. Finally, an argument for thermally induced redistribution of brine in young sea ice is presented, based on the interpretation of the bulk salinity and structural data.

Before discussing the results it is appropriate to clarify what the bulk salinity measurements represent. Each subsample comprises a solid and liquid fraction so that the salinity of a melted subsample is a measure of the bulk salinity, as defined in section 2.5.1, rather than the brine salinity which is purely a function of temperature. These
measurements of bulk salinity quantify the relative salt content of each subsample which is determined by the distribution of salt throughout the sample, usually in the form of brine. Though the approach used in this work is somewhat contrary to the traditional aims of making bulk measurements, i.e. to obtain a representative measure of the sample's gross properties, it has provided a unique depiction of brine distribution in young sea ice.

5.3.2.1 Characteristics of the brine distribution

In previous sections, frequent reference has been made, with respect to the cold ice, to the greater, or enhanced, bulk salinity of the ice in the vicinity of the brine channels. Although the meaning of this description of the salt distribution appears quite obvious in its context, it is important to recall the relevant scale when referring to greater or lesser bulk salinities. Depending on the resolution achieved by the study methods, two extreme cases can be distinguished.

At the lowest resolution (corresponding to an integration over the largest sample volume), brine features on all spatial scales are present within the samples in sufficient numbers such that the spatial variability of salinity measurements on parallel samples is reduced to a minimum. Ideally, such a measurement is required for a determination of the true bulk salinity of a sample, as assumed in most studies. Because samples are usually smaller than ideally required, the corresponding bulk salinity measurements are affected by the distribution of large-scale brine structures (Tucker III, Eicken et al., 1984; Eicken et al., 1991), as discussed in section 1.3.1.1. Studies of salt segregation and brine drainage are mostly based on a true bulk-salinity approach (Cox and Weeks, 1988) and so disregard the contribution of the different populations of brine features to the integrated bulk salinity and its evolution.

At the other extreme, the sampling resolution may be sufficiently high to determine the distribution of salt at the submillimetre level, e.g. within brine pockets and brine layers between the ice crystal platelets. Ideally, at such a resolution, the bulk salinity of a sample can be derived by determining the in situ brine salinity (e.g. from temperature measurements) and then integrating over the entire volume fraction of brine determined, for example, through thin section studies such as those shown in Figure 5.8.
Chapter 5: Brine redistribution during thermal forcing

The horizontal and vertical sections of both the cold and warm ice demonstrate that the distribution of salt in young sea ice is heterogeneous at the resolution of the measurements. This heterogeneity in the distribution is illustrated most clearly in the cold ice sampled on day 8 of the January 1997 period. At this stage of the experiment, the ice has a strongly positive temperature gradient and is actively growing. Figure 5.3(a) and (b) indicate that in all instances the areas of ice with enhanced bulk salinity coincide exactly with the positions of the brine channels.

Furthermore, the isohalines closely follow the morphology of the brine channels in the horizontal and vertical planes. In the horizontal plane, the salinity distribution follows the direction and extent of the feeder channels. The vertical plane shows that the brine channels are bounded by a volume of ice with a greater salt content. In addition to these, the regions of ice which surround the brine channels have a significantly lower salt content. In this respect the ice may be roughly partitioned into two types; ice volumes which show an enhanced salt content and coincide exactly with brine channels, and ice volumes which show a relatively depleted salt content and are devoid of brine channels.

There are two further significant observations from the data of the cold ice with respect to brine channels and their formation. First, the distribution of brine channels in the horizontal plane, and the corresponding areas of greater bulk salinity, have an apparently regular spacing of 7 cm. This is consistent with observations from sea ice forming under natural conditions, section 3.3.1.1. Second, those areas of greater bulk salinity are demarcated by regions of ice with a lesser bulk salinity. In section 3.4.2 the mechanisms and conditions regulating the formation of brine channels were discussed with reference to experimental evidence. The spacing of brine channels in the horizontal plane was found to be dependent on ice growth rate and that this may be due to the differences in the salt content of the ice. It was proposed that during the development of a brine channel it required a minimum amount of salt for it to persist which would then initiate a situation whereby the channels and their catchment areas became distributed uniformly. The segregated nature of the brine distribution, which was observed in cold ice, is empirical evidence supporting this hypothesis. The brine channels define the centre of a brine rich volume of ice surrounded by ice relatively depleted in salt, creating a regular pattern of brine distribution compatible with the idea of a catchment area.
The brine distribution within the warm, melting ice, sampled on day 24 of the January 1997 period, shows a significant change as the temperature gradient through the ice is reversed. In this case, although the brine distribution is also variable, the degree of variability observed in the horizontal and particularly in the vertical section is much reduced compared to that of the cold ice. Even though structures, which resemble brine channels, can still be clearly observed in the images of both the horizontal and the vertical sections, Figure 5.6(a) and (b), significant decoupling of the brine distribution and the brine channels has occurred. The brine in the porous ice has become more uniformly distributed throughout the entire volume irrespective of the numerous brine channels still present. There appears to be no macroscopic segregation of the salt into discrete areas of the ice.

It is possible to quantify and compare the variability in the horizontal distribution of brine in each sample by calculating the standard deviation of the bulk salinity for the subsamples at each depth. If brine pervades the ice matrix uniformly, so there is no variability in the bulk salinity, then the standard deviation of a layer is zero. A small standard deviation represents a situation where there is minimal segregation of the brine and hence very small salinity gradients. Conversely, a large standard deviation in the bulk salinity is indicative of spatial segregation of the brine within a layer containing volumes of ice with an enhanced salt content and others with salt depletion. The standard deviation in salinity for both samples, as a function of ice depth, is shown in Figure 5.7 which illustrates the changes in variability which occur as the ice undergoes a prolonged period of warming and equilibration.

The standard deviations of bulk salinity in cold ice show no clear trend; in general the values lie between 1.5 and 2.0 psu. As expected, the greatest variability in the brine distribution for cold ice is found in the deepest layer because the brine channels at the ice-water interface will have a much greater liquid fraction compared to that of the surrounding ice. During the transition from cold to warm ice the variability decreases by a factor of approximately two. The relatively large value in the surface layer is likely to be caused by uneven spillage of water onto the surface of the ice. Below the surface, standard deviations are smaller than for the corresponding values in cold ice, but steadily increase to similar values towards the ice-water interface. Whereas in the upper part of
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Chapter 5: Brine redistribution during thermal forcing

Figure 5.7: Standard deviation in bulk salinity for each horizontal layer of the cold (filled circles) and warm (open circles) ice collected during the January 1997 period.

As the warm ice the brine is uniformly distributed, it becomes increasingly localised at the ice-water interface. The greater volume fraction of the brine channels and the large pockets with high bulk salinity found in the lower part of the warm ice, Figure 5.5(f), will account for the increase in the standard deviation with depth.

5.3.2.2 Mechanisms for brine redistribution

So far, simple descriptive and comparative observations have been made between the characteristics of the brine distribution in cold ice and warm ice. To provide an explanation for the changing characteristics of the distribution, it is necessary to invoke mechanisms which permit the redistribution of brine. The case of cold ice will be discussed first.

It is assumed that during initial ice growth, prior to the development of mature brine channels, the distribution of salt throughout the ice will have been more uniform on the scale of the current measurements; an assumption which will be assessed in chapter 6. From the bulk salinity data in Figure 5.2, it is postulated that due to the varying thermal field, further evolution of the cold, growing ice cover is accompanied by horizontal migration of brine. This process of brine relocation creates areas within the ice which have low bulk salinity adjacent to areas of much greater bulk salinity. It is
likely that horizontal migration of brine is via brine expulsion, initiated by the changing temperature gradient in the growing ice sheet. As the channels are attended by a network of inclined feeder channels it is these which are the most likely path for migration. The increase in the salt content of the ice in the vicinity of the brine channels indicates that the movement of brine is towards these areas of relatively higher porosity. In this respect the results substantiate the long-standing proposition by Lake and Lewis (1970) that brine channels act as a sink for brine residing within a surrounding catchment area.

In the case of warm, melting ice sampled on day 24 of the January 1997 period, a contrasting salt distribution within the ice matrix is revealed. The redistribution of brine in the ice is a consequence of the changing temperature field. During the stagnation and melt phases, the increasing air temperature forces the temperature gradient in the ice to shift from positive through isothermal to negative. As the ice temperature evolves the brine within the solid matrix reacts to maintain ambient thermodynamic equilibrium, a concept discussed in section 2.2.2. This equilibrium is achieved by dilution of the brine by melting of the walls of the brine pockets and channels. The difference in density between ice and brine creates a differential increase in the total porosity of the sheet.

Figure 5.8(a) and (b) show the pore space as determined by analysis of horizontal thin sections of samples of cold and warm ice respectively. In this instance, the generic term pore space has been adopted as this is appropriate when discussing microstructural observations. The section of cold ice is from day 8 of the January 1997 period and

Figure 5.8: Pore space as determined from horizontal thin sections (20 mm horizontal width) for (a) cold ice on day 8 and (b) warm ice on day 17 of the January 1997 period. Data courtesy of H. Eicken, Geophysical Institute, University of Alaska, Fairbanks.
Chapter 5: Brine redistribution during thermal forcing

corresponds to a position in the ice 118 mm above the ice-water interface. The section of warm ice is taken on day 17 of the same period and is 129 mm from the ice-water interface. Total porosity of the ice, calculated from the thin sections, increases from 2.3% to 8.0% with the mean pore area increasing from 0.016 mm² to 0.039 mm². This is direct corroboration of the increase in porosity demonstrated in the vertical profile of brine volume in Figure 5.4(c) when compared with Figure 5.1(c). It should be noted that the porosity derived from the thin sections is less than the calculated brine volume. This is due to the lack of large volume brine structures present in the thin sections; they consist entirely of brine pockets and layers.

In sea ice, porosity and permeability are associative and hence the permeability of the sheet also increases as it warms. Enhancement of the permeability of the ice, by coalescence and interconnection of brine pockets and layers (Perovich and Gow, 1996), increases the propensity for brine migration within the ice matrix with less reliance on the presence of macroscopic paths, such as brine channels. It is physically plausible for permeation of brine to occur in both vertical and lateral directions, creating the possibility for redistribution of the previously confined brine.

A comparison of the salinity profiles of the cold and warm ice samples, Figure 5.1(a) and Figure 5.4(a), demonstrates how warming affects, and completely modifies, the salinity structure of sea ice. This was further illustrated in the comparison of the profiles of standard deviation in bulk salinity of each sample, Figure 5.7. One mode of migration, vertical percolation, is a potential candidate for rationalising the redistribution of brine. With warming, the differential increase in the pore volume has to be compensated by an influx of brine or meltwater from above or below (cf. suction process of Cox, 1990). In the upper layers of the ice, i.e. down to 10 or even 15 cm depth a proportion of the influx may be the meltwater which covered the ice surface during the melt phase. The meltwater would tend to percolate downward in an analogous manner to the flushing process proposed and studied by Untersteiner (1968), section 2.6.4. In this situation, brine channels act as the principal conduits for flushing because of their greater permeability. This mechanism would generate a differential decrease in brine channel salinities over the ice matrix salinities since the former will experience enhanced meltwater fluxes over the latter. Therefore, vertical flushing of
brine through the more permeable brine channels will have the effect of equalising the horizontal salt distribution. Furthermore, it would explain the gradual increase with depth of bulk salinity from comparatively low values at the surface.

Vertical redistribution of the brine by flushing may also be accompanied by lateral migration. Although the permeability of sea ice in the horizontal direction is less than that in the vertical, perhaps by an order of magnitude or more (Freitag, *pers. comm.*, 1998), migration of brine in this direction remains a potential mechanism to effect the equalisation of the salt content in the ice. Although no microstructural data were obtained from which paths for lateral migration might be identified, Figure 5.8 shows how interconnection of brine layers and pockets will occur, thereby establishing such paths in the horizontal direction. Perhaps more convincing evidence for the likelihood of this processes is provided by field observations. Core holes drilled part way through summer sea ice rapidly fill with brine depending on the porosity of the ice. This technique is often used for sampling the chemical properties of the brine or for making lateral percolation studies in the field (Eicken *et al.*, 1995). Taken together, the two modes of migration, vertical and horizontal, would provide a viable mechanism for equalising the brine distribution in sea ice as its porosity increases. Finally, a stable density stratification associated with the negative temperature gradient of the warm ice, inhibits convective brine exchange and the associated effects on brine channel evolution and salinity structure.

To summarise, the change in porosity, resulting in the free movement of brine within the ice matrix, explains the break down in the salinity-structure link discussed in section 5.3.2.1. The segregated brine in the cold ice, Figure 5.3, permeates the warm ice, Figure 5.6, regardless of the presence of brine channels. The increasing porosity and permeability that occur during warming enables the redistribution of brine through the ice. Brine is flushed, or able to flow, from the areas of the ice which are salt rich into the adjacent areas which were previously depleted of salt. The consequence of this is a gradual equalisation of the salt content of the ice and hence a decrease in the variability of the brine distribution. Brine channels that are observed in the warm ice do not now act as sinks for brine. Although they may have an enhanced vertical permeability, their salt content is now comparable to that of the surrounding ice.
Chapter 5: Brine redistribution during thermal forcing

5.4 Refreeze

The discussion in section 5.3.2 has analysed the mechanisms which regulate the distribution of brine in young sea ice and its redistribution as the ice is forced into a melting state by inversion of the temperature gradient. Refreeze is the reverse process and is therefore complementary to the observations and interpretations made during the melt phase. An attempt to simulate a refreeze was undertaken during the November 1998 period of *Interice II*. During this time, two samples of ice were recovered from the current zone and processed according to the methods described in section 4.5. Again, applying the same terminology, one sample was of warm ice and the other was cold ice, but separated by a period of intense cooling rather than warming. For each sample, the results will be presented in a similar fashion as for the melt phase of section 5.2.

5.4.1 Warm ice

The warm ice was sampled on day 8 of the experiment when it was 14.5 cm thick. The ice had then experienced 6 days of active growth followed immediately by 2 days of thermal forcing with increased air temperatures, see Figure 4.3.

5.4.1.1 Bulk profiles

Vertical profiles of temperature, bulk salinity and brine volume for the sample of warm ice are given in Figure 5.9. Measurements of the ice temperature were made to a resolution of ±0.01°C using a thermistor string frozen into the ice cover. The profile shows that, at the time of sampling, the ice was isothermal. On day 8 of the November 1998 period, the ice had been subjected to only 2 days of warming and from Figure 4.9 it can been seen that the ice was still growing 1 day after the air temperature was increased. Therefore, the warm ice in this example is not as porous in comparison to that of the January 1997 period.

Further evidence that the ice was not in an advanced state of deterioration is provided by the brine volume profile, Figure 5.9(c), which shows that the ice porosity never exceed 20%. In the previous example of warm ice the brine volume was mostly
greater than 20% throughout the ice, Figure 5.4(c). However, at all depths, brine volume is found to be greater than the critical value of 5% required for vertical movement of brine (Cox and Weeks, 1975). Figure 5.9(b) shows a steady increase of bulk salinity with depth towards the ice-water interface. Absence of the characteristic C-shaped profile of bulk salinity in this case indicates some brine redistribution has occurred. This is similar to the change in salinity profile during the summer transition from first-year to multiyear ice.

5.4.1.2 Bulk salinity

The raw bulk salinity data for the warm, isothermal ice are shown in Figure 5.10. The horizontal section, Figure 5.10(a) is a plate of ice, 2 cm thick, located at a depth interval between 2 and 4 cm, (H-H) in Figure 5.10(c) and (e). Mean bulk salinity for the section is 5.7 ± 0.6 psu with a maximum of 8.2 psu and minimum of 4.9 psu. Other than the point of enhanced bulk salinity at (7,3) the subsamples are all of similar bulk salinity, as implied by the low standard deviation. The largest difference between adjacent
Figure 5.10: Raw and interpolated forms of the bulk salinity data for warm ice on day 8 of the November 1998 period; x and y positions lie in the horizontal plane. NB. Colour scaling for (c) and (e) differs from (a). Depth interval for (a) indicated by line H--H in (c) and (e). Location of the vertical sections (c) and (e) are indicated in (a) by lines A--A and B--B respectively.
Chapter 5: Brine redistribution during thermal forcing

subsamples is 2.2 psu but this is in the seemingly atypical portion of the ice at (7,3). Only in the interpolated plot, Figure 5.10(b), can two other areas, with bulk salinities greater than the mean, be identified; these are at (9,11) and (3,10). However, these areas have a salinity less than two standard deviations from the mean, therefore the distribution of salt can be considered effectively uniform.

In the vertical sections, Figure 5.10(c) and (e), located along (A--A) and (B--B) respectively in Figure 5.10(a), the character of the brine distribution is seen to change noticeably with depth. The ice has become partitioned into an upper part (0–6 cm) and a lower part (6–14.5 cm). In the upper part, the bulk salinity is quite uniform across the sample and there are no distinct features of increased bulk salinity. In contrast, in the lower part of the ice, a number of distinct features can be identified. A large area of enhanced bulk salinity centred at x = 4 cm dominates the data in Figure 5.10(c) and (d). Additionally, a vertical feature at x = 9 cm penetrates the ice from the ice-water interface to a depth of 6 cm where it becomes more diffuse. Only one dominant feature can be discerned in the raw data, Figure 5.10(e), but is seen more clearly in Figure 5.10(f). This feature, a vertical band of high bulk salinity at y = 11 cm, penetrates through the ice from the ice-water interface again to a depth of approximately 5–6 cm.

5.4.2 Cold ice

The cold ice was sampled on day 10 of the November 1998 period, 2 days after sampling the warm ice during which the air temperature had decreased rapidly to below -20°C. Ice thickness at the time of sampling had increased to 16 cm.

5.4.2.1 Bulk profiles

The temperature profile of the ice following refreeze, Figure 5.11(a), demonstrates that the ice sheet was actively growing at the time of sampling. A linear temperature gradient of +0.5°C cm⁻¹ had been established in the ice giving a strongly negative heat flux through the sheet. An interesting feature in the bulk salinity profile Figure 5.11(b) is that it has a shape like a question mark, a ‘?-profile’ similar to that identified by Eicken (1992b). This difference in the salinity profile compared to that in Figure 5.9(b) is itself
evidence that brine has become redistributed as the temperature field has changed. The very high value of 10.8 psu at the ice-water interface can be associated with the establishment of a dendritic skeletal layer with a high brine content. Predictably, the skeletal layer has a high brine volume of approximately 16%. The profile of brine volume in Figure 5.11(c) shows that porosity decreases rapidly from the skeletal layer to values close to the critical 5% whilst at the surface, in the coldest part of the ice, it can be regarded as impermeable as the brine volume falls below 4%.

### 5.4.2.2 Bulk salinity

The raw bulk salinity data for the cold ice are shown in Figure 5.12. The salinity scaling is identical to that used in Figure 5.10. The horizontal section, Figure 5.12(a) is a plate of ice, 2 cm thick, which is also located at a depth interval of 2–4 cm, (H--H) in Figure 5.12(c) and (e). Mean bulk salinity for the section is 6.7 ± 1.1 psu with a maximum of 9.5 psu and minimum of 4.9 psu. It is quite apparent from the salinity data in the horizontal plane that a significant change in the brine distribution has occurred during
Figure 5.12: Raw and interpolated forms of the bulk salinity data for cold ice on day 10 of the November 1998 period; x and y positions lie in the horizontal plane. NB. Colour scaling for (c) and (e) differs from (a). Depth interval for (a) indicated by line H--H in (c) and (e). Location of the vertical sections (c) and (e) are indicated in (a) by lines A--A and B--B respectively.
refreeze. Areas of ice with a bulk salinity much greater than the mean can be identified in both the raw and interpolated forms of the data. The most notable are at (11,12), (9,3) and (3,3) which extends towards another maximum at (4,7). Evidently, the distribution of salt in the cold ice is more clustered compared to that in the warm ice. The maximum difference between adjacent subsamples has increased from 2.2 psu to 3.4 psu. Refreezing of the ice has increased the salinity gradients found within the ice matrix.

The vertical sections, Figure 5.12(c) and (e), are located along (A--A) and (B--B) respectively in Figure 5.12(a). Whereas previously the brine distribution in the warm ice differed between the upper and lower portions, in the cold ice this partitioning is not so evident. The interpolated data depicts the variations in bulk salinity more effectively; features with high bulk salinity can be identified in both sections. Most notable are the vertical bands at x = 3 cm in Figure 5.12(d), and also at y = 3 cm and y = 12 cm in Figure 5.12(f). These are all in the upper portion of the ice (0–6 cm depth) which in the warm ice was devoid of significant salinity variation. In the lowest layer of the ice the high bulk salinity of the skeletal layer is clearly defined in both vertical sections below about 13 cm. Finally, a volume of ice of high bulk salinity is centred at y = 8 cm in Figure 5.12(f). In this location the photographic images of the sections show a very well developed brine channel extending from the ice-water interface to a depth of 7 cm.

5.4.3 Brine structures in warm and cold ice

From the bulk salinity data described in the two preceding sections it was found that, other than the upper part of the ice, there is minimal contrast in the brine distribution between the warm and cold ice. Therefore, for the purpose of comparing the distribution with the occurrence of brine structures in the ice, only the horizontal sections will be considered. The images of the two sections are both of ice at a depth of 2 cm and it has been shown in Figure 5.10(a) and Figure 5.12(a) that the distribution of brine in the layer between 2–4 cm is significantly different within each. Like the previous examples of composite images of brine structures and bulk salinity, e.g. Figure 5.3(a), the photographs of the sections will be overlaid with isohalines of equal bulk salinity. These composite images are presented in Figure 5.13. To illustrate the contrasts more clearly the isohalines in each of the figures are plotted at the same salinity interval, 1 psu steps.
Figure 5.13: Composite images of the horizontal sections showing the brine structures in the ice with the isohalines for (a) warm ice and (b) cold ice obtained on days 8 and 10 of the November 1998 period respectively.
Chapter 5: Brine redistribution during thermal forcing

Considering first the case of the warm ice, Figure 5.13(a), correlation between areas of enhanced bulk salinity and the brine channels is still evident. At (7,3) the unusually high bulk salinity of this area is precisely located in a volume of ice containing a well developed and branched brine channel. This is the place of steepest salinity gradients of approximately 1 psu cm\(^{-1}\). However, two brine channels which are of comparable dimensions can be seen at (10,10) and (2,11). Here, the occurrence of only a single isohaline indicates that even in these regions the brine is dispersed quite homogeneously through the ice matrix reducing gradients of salinity to a minimum.

Furthermore, there are clearly defined brine channels located at (16,1), (13,13), (14,6), (6,14) and (2,1). In these areas, there is so little variation in salinity that the channels are not attended by isohalines. Each lies in a volume of ice with a bulk salinity between 5 and 6 psu which corresponds with the mean value of 5.7 psu. As a final observation, where isohalines coincide with brine channels they tend to follow the general morphology of the channel and its feeder channels. Other than the single, clearly defined feature, this section shows a remarkably homogeneous distribution of brine.

The composite image of cold ice in Figure 5.13(b) is very similar to the composite image of cold ice in Figure 5.3(a). This observation is founded on three comparable characteristics of the two images. First, the isohalines indicate that the brine distribution is heterogeneous and thus there is a continuous and substantial variation in the salt content of the ice. Consequently, bulk salinity gradients are generally much greater than those in Figure 5.13(a). Second, these areas of ice with high bulk salinity are associated exclusively with regions of ice which contain a brine drainage channel. The well-established link between salinity and structure, identified in section 5.2.1.3, is maintained in this sample. Those areas of high bulk salinity are surrounded by ice having significantly lower salinity with minima at (10,7), (8,5) and (6,12). The final characteristic to note is that the shape of the isohalines closely follows the morphology of the channels. A prime example of this is in the case of the channel centred at (4,6). This channel has a feeder channel which extends through the ice to approximately (6,9); the 7 psu isohaline conforms to this aspect of the channel morphology. In all aspects the cold ice represented in Figure 5.13(b) shows that brine has become redistributed during refreeze and segregated according to the location of the brine channels.
5.4.4 Discussion

In the earlier discussion of brine redistribution in ice undergoing melt, section 5.3.2, mechanisms for brine migration were suggested to account for the observed changes in distribution. By thermally forcing the ice through a refreeze phase, brine redistribution is also induced and similar mechanisms might be usefully employed to explain these. An important point to emphasise at this stage is that the warm ice in the refreeze phase was subjected to 2 days of elevated air temperatures only. In the warm ice investigated during the melt phase it had experienced 8 days of stagnation and 8 days of elevated air temperatures before sampling. Therefore, although they are both referred to as warm ice, their thermal histories are quite different.

The first aspect of brine migration to discuss relates to the profiles of bulk salinity. In the case of warm ice, the profile in Figure 5.9(b) shows an increase in salinity with depth. Assuming that the profile was originally C-shaped, then brine has migrated downwards. In the previous example of melting ice this vertical migration was accompanied by a flushing mechanism resulting from the accumulation of meltwater on the upper surface of the ice. During the refreeze phase, care was taken to protect the ice surface from precipitation from the chilling units so the efficiency of this mechanism was potentially reduced.

After refreeze, the form of the salinity profile for the cold ice, Figure 5.11(b), has evolved from that observed for warm ice. This modification of the profile is a clear indicator that during refreeze vertical redistribution of brine has occurred. The greatest difference in the profiles is the apparent loss of brine in the middle depths of the ice and the large increase at the ice-water interface. During refreeze, a positive temperature gradient is re-established when the cold front passes through the ice. As the ice cools, its porosity will decrease and this will have the effect of driving brine towards the areas of higher porosity. The effect is a brine expulsion process which is directed toward the more porous, lower portion of the ice.

Looking now at the two composite images of brine distribution illustrated in Figure 5.13 the mechanisms causing them can be considered. During the discussion of the variation in bulk salinity it was noted that the sample of warm ice exhibited different characteristics in its upper and lower portion. The lateral distribution of brine in the
upper part, of which Figure 5.13(a) is an example, is seen to be quite uniform and in general unperturbed by the presence of brine channels. Here, the mechanism of flushing, which would have been active at this time, may be invoked to account for the equalisation of the brine distribution. However, as stated earlier, the efficiency of flushing and the time for which it acted was minimal with the effect that brine redistribution was restricted to the upper part of the sample only. Deeper in the ice, a segregated distribution has been retained. Additionally, because the ice temperature never increased above -2°C, the flushing could not have constituted a significant addition of freshwater. The cumulative effect is one of brine redistribution both vertically and laterally rather than extensive desalination.

Once the ice undergoes refreeze, the nature of the brine distribution changes. Figure 5.13(b) shows that in this state the brine becomes preferentially located in the brine channels. The degree of variability in the bulk salinity of the subsamples increases by a factor of two and the salinity gradients across the section increase correspondingly. The changes that have occurred are clearly induced by the thermal forcing. This indicates that phase equilibrium is active and therefore expulsion becomes the most likely mechanism for brine redistribution. Expulsion will be directed along the path of least resistance, i.e. greatest porosity. Thus the brine channels act as a focal point to which the brine migrates. Hence the bulk salinity increases in those volumes of ice containing brine channels and increasing the variability in salt content across the sample.

5.5 Summary

In this chapter the distribution of brine in samples of young sea ice with contrasting temperature profiles has been analysed. Two thermal forcing situations were considered; melting and refreeze. In each case, samples of ice were obtained from the ice sheet prior to forcing and then after a period of thermal evolution. During this period the porosity of the ice had changed and this has been implicated for the observed redistribution of the brine. The comparison that has been made between cold and warm ice demonstrates that thermal forcing significantly modifies the salinity structure of sea ice.
Chapter 5: Brine redistribution during thermal forcing

It has been demonstrated that the horizontal and vertical distribution of salt within young sea ice is heterogeneous at the spatial resolution of the measurements. The variability in the salt content through the samples is linked to the porosity of the ice by both the brine channels and the brine inclusions. In the introduction to this chapter a hypothesis was proposed that brine channels would be influential on the distribution of salt in sea ice. This hypothesis has been supported by the results obtained for cold ice in both melt and refreeze conditions where the nature of the distribution of brine is linked to the location and morphology of the brine channels within the ice.

The present study of brine channels demonstrates that variability in their shape and distribution may be responsible for a major part of the variability commonly observed in measurements of bulk salinity. Previous studies (Tucker III et al., 1984; Eicken et al., 1991) have commented on this, but were not able to provide proof. This situation is evident in Figure 5.3(a), which by cross-sectional area corresponds to 4.5 standard core samples, any of which may or may not include high salinity brine channels. It is acknowledged that this result may be intuitive since it is generally accepted that brine channels act as conduits for brine. Therefore, it is not unexpected that brine channels will exhibit a significantly higher bulk salinity by focusing the flow of brine. However, the novel method used in this study to generate the data has demonstrated this link conclusively and enabled the variations in the distribution of salt to be attributed directly to the occurrence of brine channels in young sea ice.

During thermal forcing, the brine undergoes considerable redistribution. In the case of warm ice, two mechanisms have been proposed to facilitate this. First, as the ice temperature increases the permeability of the brine channels increases differentially with respect to the ice matrix. With an influx of surface meltwater, this allows percolation processes to equalise the lateral brine distribution. Second, this effect is enhanced by the increase in connectivity of the brine inclusions allows brine to pervade the sample by lateral migration. During refreeze, it was found that brine is forced towards the areas of high porosity, the brine channels, and the brine becomes localised. It was suggested that the driving force for brine redistribution in this instance was provided by thermally induced brine expulsion process. In conclusion, temperature cycling of a sea ice cover has shown that brine in the ice is highly mobile and its distribution is closely linked to the location of brine channels.
Chapter 6
Evolution of the brine distribution in new ice

6.1 Introduction

The object of this chapter is to investigate the evolving nature of the brine distribution in new ice, the term used to describe recently formed sea ice. In the previous chapter the distribution in cold sea ice was shown to be highly variable at the spatial resolution of the measurements. The distribution was structured with respect to the location of the brine drainage channels which act as a sink for brine. In section 5.3.2.2 it was assumed that the lateral distribution of brine in the initial skim of ice would be more homogeneous than that occurring in thicker ice and can be regarded as the basis from which all subsequent evolution of the distribution occurs. It was proposed that during subsequent ice growth and brine channel development, redistribution of the brine, by migration along the inclined feeder channels, would cause a more heterogeneous brine distribution corresponding to the location of the brine channels.

In section 6.2, the assumption of homogeneity in the initial brine distribution during ice formation will be investigated. Then in section 6.3, two aspects of the evolving brine distribution will be discussed. First, will be the redistribution of brine in new ice which will include the issue of brine loss during the early stages of ice growth and the ensuing change in variability of the brine distribution as the mobile brine becomes structured within the ice matrix. Second, it will be shown that the changes in morphology and lateral spacing of the brine channels are closely linked to the brine distribution during the first days of ice growth.
To date, the formation and development of brine drainage channels in sea ice has not been investigated by experimentation in large tanks. Previous studies have been restricted to quasi two-dimensional systems (Eide and Martin, 1975; Niedrauer and Martin, 1979) or conducted in very small tanks (Wettlaufer et al., 1997a). This is the first systematic investigation of the temporal evolution of brine channels, and the associated brine distribution, in a three-dimensional system with an ice cover that can be regarded as infinite compared with the scale of the structures.

To investigate these specific aspects of new ice it was planned to obtain a time series of samples of increasing thickness from the current zone, see Figure 4.1. However, owing to the scheduling of the freeze-up and access to the tank, the first sample obtained from the current zone had a thickness of 4.5 cm. This was considered too thick to establish the nature of the brine distribution after the formation of the initial skim of ice. Therefore, thinner ice from the calm zone was collected to ascertain the brine distribution at an earlier stage in the development of the ice. The growth kinetics of the ice in these two areas of the tank were slightly different, and although the data from the calm zone establish the initial condition, they cannot be compared directly to subsequent data from the current zone. Data for the initial skim of ice and the time series will be presented and discussed separately.

6.2 Initial ice skim

The raison d'être for investigating the brine distribution in the thin skim of new ice was to ascertain its initial state. It has been assumed, (Cottier et al., 1999) and section 5.3.2.2, that the distribution of brine in the initial skim of ice would be quite homogeneous with minimal lateral variation. Simply stated, the distribution of brine changes from fully homogeneous, as found in the surface water, to the heterogeneous state observed in cold sea ice in section 5.2.1. This section assesses that assumption.
6.2.1 Results

A sample of new ice was obtained from the calm zone, 2 days after seeding, during the November 1998 experimental period of Interise II. At this time the mean ice thickness in the current zone was 6.6 cm whilst in the calm zone it was 1.6 cm. Owing to the air and water temperature increase from the partition wall at \( x = 8.5 \) m towards the tank boundary at \( x = 0 \) m, ice formation in the calm zone commenced approximately 24 hours later than in the current zone (refer back to Figure 4.1 which illustrates the coordinate system of the tank). Further, the ice thickness in the calm zone on day 2 of the experiment ranged from open water at \( x = 0 \) to 3.5 cm at \( x = 7.5 \) m. The sample of ice obtained from the calm zone had a thickness of 2.6 cm with vertically oriented columnar crystals and a well established skeletal layer.

Measurements of ice temperature in the calm zone were made using a thermistor string with a resolution of ±0.01°C. At the time of sampling, the temperature at a depth of 11.5 mm in the ice was -4.34°C. Assuming (a) that the temperature at the ice-water interface was at the freezing point of the water (-2.0°C at that time) and (b) that the gradient in the ice was linear, then the temperature gradient at the time of sampling was +1.6°C cm\(^{-1}\). Therefore the ice temperature, for a median depth of 1.3 cm, was -4.10°C. The large, positive temperature gradient classifies the ice as being cold ice, as defined in section 5.2.

Because of the fragile and porous nature of the ice skim, sampling for bulk salinity by standard methods is inherently fraught with problems of brine drainage. The method of sample collection employed throughout this research proved ideal for obtaining reliable measurements of bulk salinity in such ice. Because the ice was so thin, a single horizontal plate of ice was prepared and subsampled. From these melted subsamples, the mean bulk salinity of the plate was found to be 12.7 psu with a maximum of 13.8 psu and a minimum of 11.3 psu.

Using the mean values of temperature and salinity, and the formulae of Cox and Weeks (1983), the brine volume for the skim is calculated to be 15.5%. Additionally, using the potential upper and lower limits of ice temperature (-6.16°C at the air-ice interface and -2.0°C at the ice-water interface) and bulk salinity (13.8 and 11.3 psu),
gives possible magnitudes of brine volume from 9.4% up to 35.0%. This calculated range concurs with estimates of brine volume in the skeletal layer of between 8 and 40% (Maykut, 1985).

The results of the bulk salinity measurements are shown in Figure 6.1. The salinity scale has been set to a range of 5 psu which is comparable to that of the horizontal section of cold ice in Figure 5.12(a) and (b). By doing this, the uniformity of the salt content of the ice is more clearly illustrated. Quantitatively, the standard deviation of the bulk salinity in the sample is calculated to be 0.5 psu. The high degree of uniformity is, in itself, supporting evidence that brine drainage from the sample is not significant. Although there are no specific points of enhanced bulk salinity, there is a general trend of decreasing salinity from the lower left corner of the image (2,16) towards the upper right corner (18,1). Within this trend there are three identifiable points of reduced bulk salinity compared to that of the surrounding ice; these are at (12,10), (9,6) and (5,3).

Figure 6.1: Raw and interpolated forms of the bulk salinity data in the horizontal plane for the skim of new ice sampled in the calm zone on day 2 of the November 1998 period. Lines A--A and B--B in (a) indicate the location of the vertical sections presented in Figure 6.3 (a) and (b) respectively.

The photographic image of the horizontal section with the superimposed isohalines of bulk salinity is shown in Figure 6.2; the grey scaling has been reversed. As expected, the isohalines indicate that there is no systematic structure to the brine distribution, which is characterised by small salinity gradients. The large-scale lateral variation in
grey scale intensity in the image is caused by uneven illumination of the slab. However, within this variation a few structural features can be identified. The clearest brine structure which can be identified is that at (13,10). It is approximately 1 cm across and forms a branched pattern. Others which appear quite distinctly on the original photograph, but which have not rendered well in the image, are located at (12.5,12.5), (9,14.5) and (5,5). These are slightly smaller and although not as obviously branched they do not have a simple circular cross section. Other features in the image are the small dots, though it is difficult to ascertain whether these are very small channels, or brine- or air-filled pockets.

![Composite image of the horizontal section showing the brine structures in the ice with the isohalines for the skim of new ice sampled in the calm zone on day 2 of the November 1998 period.](image)

**Figure 6.2:** *Composite image of the horizontal section showing the brine structures in the ice with the isohalines for the skim of new ice sampled in the calm zone on day 2 of the November 1998 period.*

A variation on the standard composite image is presented in Figure 6.3. Here, two vertical sections of the sample with the corresponding interpolated values of bulk salinity are shown. It should be stressed that the red data line shows the variation in bulk salinity rather than it being an isohaline of equal bulk salinity. The left *y*-axis gives the depth of
the ice whilst the right y-axis indicates the bulk salinity. The x-axis is identical for each data set and gives the lateral coordinate across the sample. Figure 6.3(a) corresponds to the subsamples located along the dashed line (A--A) in Figure 6.1(a), and Figure 6.3(b) corresponds to the subsamples located along the line (B--B) also in Figure 6.1(a). These two sections were chosen because they contain readily identifiable brine structures which are seen as light coloured structures contrasting with the darker ice.

In Figure 6.3(a), the largest brine structure, though not particularly clear, is the broad, vertical structure at \(x = 13.5\) cm (bounded by red markers). Another is located at \(x = 5\) cm and a third at \(x = 1\) cm. Within the image there are numerous other vertical streaks which are possibly very narrow brine structures within the ice and therefore very poorly defined. An example of this is found at \(x = 8.75\) cm which may be a brine structure penetrating the entire thickness depth of the ice. Bulk salinity in this section varies between a maximum of 13.2 psu and a minimum of 11.4 psu. The greatest difference in bulk salinity between adjacent samples is 1 psu which represents a maximum salinity gradient of 0.5 psu cm\(^{-1}\).

**Figure 6.3:** Composite image of two vertical sections showing the brine structures in the ice with the bulk salinity for the skim of new ice sampled in the calm zone on day 2 of the November 1998 period.
Chapter 6: Evolution of the brine distribution in new ice

The branched brine structure identified in Figure 6.2(b) at (12.5,12.5) is found in Figure 6.3(b) at \( x = 12.5 \) cm penetrating the entire thickness of ice. Other brine structures can be identified at \( x = 9 \) cm and \( x = 1.5 \) cm. Like Figure 6.3(a), the image includes many vertical streaks which cannot be positively identified as discrete brine structures. The bulk salinity in this section varies between a maximum of 13.3 psu and a minimum of 11.9 psu; the maximum salinity gradient is also 0.5 psu cm\(^{-1}\).

Even at this early stage in the ice development there may be some association between the variation in bulk salinity and the brine structures. For example, in Figure 6.3(a) the salinity maxima are located at \( x = 5 \) cm and \( x = 9 \) cm, close to where brine structures are also located. Further, in neither the image nor the original photograph can any features be identified in the ice at \( x = 12 \) cm, where the bulk salinity minimum occurs. Contrary to this is the variation in bulk salinity plotted in Figure 6.3(b). The maximum at \( x = 5 \) cm corresponds to a volume of ice which appears from the image to be devoid of any brine structures. Even at \( x = 12.5 \) cm, the well defined structure induces no deviation in the bulk salinity. Therefore, correlation between bulk salinity and the immature brine structures is either very weak or nil.

6.2.2 Discussion

From these data, it has been demonstrated conclusively that the assumption of homogeneity of the brine distribution in the initial skim of new ice is found to be valid. The points of reduced bulk salinity identified in Figure 6.1 should be treated with caution. Where previously, these features in the brine distribution have indicated areas of significant brine depletion, here they represent a decrease in bulk salinity of less than 1 psu compared to that of the adjacent subsamples. In much thicker, cold ice, differences between adjacent subsamples were measured to be more than 3 psu.

Another important observation to make with the mean bulk salinity of the ice is that at 12.7 psu it is approximately one third of the salinity of the water from which it formed. Therefore, there has been significant brine loss from the ice, either by rejection at the growing interface or by drainage or expulsion from the ice interior. This appears to contrast with observations by Wettlaufer et al. (Wettlaufer et al., 1997a; Wettlaufer et
al., 1997b; Worster and Wettlaufer, 1997) who predict minimal brine loss from new ice. Further discussion of this apparent inconsistency will be given in section 6.3.2.1.

Quantitatively, the standard deviation of bulk salinity for the layer is 0.5 psu. In reality, it is probably lower because with such thin ice any slight perturbation in the brine system, such as that caused by imperfect levelling of the sample during shock-freezing and the accuracy of sawing, will manifest as a gradual variation in the bulk salinity of the ice as seen in Figure 6.1(b). Simply recalculating the standard deviation for the subsamples in the lower left section of Figure 6.1(a) and in the upper right part of the same figure yields values of 0.4 psu and 0.3 psu respectively. Therefore, for the in situ condition, the standard deviation in bulk salinity is likely to be less than the measured 0.5 psu. It is important to note that the recalculated standard deviation of bulk salinity is still greater than the resolution of the conductivity meter of 0.1 psu.

The principal observation from the composite image in Figure 6.2 is that, at this stage in ice growth, there are no clearly identifiable brine channels. However, structures which appear to show the characteristics of brine channels occur, that is they are discrete, slightly branched structures. However, if the structures identified in Figure 6.2 are indeed embryonic brine channels then they are too small and without a sufficiently enhanced brine content to have a significant influence on the brine distribution at the spatial resolution of the measurements.

This interpretation is confirmed from the vertical sections in Figure 6.3. There are brine structures in this figure which have been identified as potential brine channels and which may be correlated with small variations in the bulk salinity of the ice at their location. However, there are others, notably the structure at x = 12.5 cm in Figure 6.3(b), which are found in areas of ice having no relative enhancement of the bulk salinity. A possible explanation for this is that the structures have been bisected during sawing and thus their effect on the brine distribution is divided between subsamples.

From these sections it is difficult to positively identify sufficient brine channels to get an accurate quantitative estimate of their lateral distribution. However, it appears that they are more closely spaced than the mature channels seen in the composite figures in chapter 5. It would seem that their spacing is more comparable to that of the brine tubes described in section 3.3.1.2 than the typical brine channel nearest neighbour
Chapter 6: Evolution of the brine distribution in new ice

spacing of 5–10 cm (Morey et al., 1984; Wakatsuchi and Saito, 1985; Cole and Shapiro, 1998), discussed in section 3.3.1.1. This is the first indication that the nearest neighbour spacing of brine channels increases as they mature and become more highly branched.

To summarise, in the initial skim of new ice, with a thickness of less than 3 cm, the distribution of brine measured at a spatial scale of 2 cm is found to be homogeneous, with the mean bulk salinity less than the salinity of the water from which it formed. Further, the brine structures found in the ice at this time are closely spaced vertical channels with a low degree of branching. These channels are seen to extend through the full thickness of the ice. Although in some instances there is correlation between the small variation in bulk salinity and the position of the brine structures, this is by no means consistent throughout the ice. There is minimal redistribution of brine occurring in the initial skim which has undeveloped brine structures.

6.3 Time series ice sampling

To follow the evolution of the brine distribution and the brine channels during ice growth, a time series of three samples was obtained during the November 1998 period of Interice II. All samples were collected from the current zone, 1 day, 2 days and 3 days after seeding. During this period the air temperature was approximately −15°C, see Figure 4.3. For the first 24 hours, there was no shear current but once the ice had congealed the impellers were operated to produce a current of 7 cm s⁻¹.

6.3.1 Results

All samples were collected in the region of ice between \( x = 17 \) m and \( x = 19 \) m (refer to Figure 4.1). The ice thickness at the time of sampling on each day was: day 1 = 4.5 cm, day 2 = 6.3 cm, day 3 = 8.5 cm. The results will include the combined bulk profiles for the three samples, followed by the matrices of bulk salinity and composite images presented chronologically for each sample to give a complete description of the ice and the brine distribution on each day.
6.3.1.1 Bulk profiles

The vertical profiles of bulk parameters for each sample of ice are shown in Figure 6.4. For each parameter, four data sets are presented with a consistent key. Also included in each figure is the single data point (star) for the initial skim of new ice described in section 6.2. Although it has been noted that the ice from the calm and current zones were not suitable for direct quantitative comparison, including the additional data points here may give some indication of the properties of the initial skim of ice which formed in the current zone.

![Figure 6.4](image-url)

**Figure 6.4:** Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for four samples of new ice collected during the November 1998 period of Interice II. In each case the key is: day 1 (□ - □), day 2 (● - ●), day 3 (○ - ○) and the initial skim described in 6.2 (*).

The temperature profiles, Figure 6.4(a), were obtained using a thermistor string embedded in the ice. Measurements were made to a resolution of ±0.01°C at intervals of 20 mm. Each sample has a positive, linear, temperature gradient with a consistent magnitude of approximately +0.6°C cm⁻¹. Therefore, the ice comes under the definition of cold ice as described in section 5.2. As expected, the temperature of the ice at a fixed
depth decreases as the ice thickens, indicating that the growing ice is in a continuous state of phase change as it attains equilibrium in the varying temperature field. All the profiles tend towards the salinity determined freezing point of the water, between -2.0 and -2.15°C.

The profiles of bulk salinity are shown in Figure 6.4(b). During processing it was necessary to square each sample to obtain true surfaces at both the air-ice and ice-water interface. This required paring 2–3 mm from the air-ice interface which consisted of frozen frost flowers. Naturally formed frost flowers are known to be highly saline (Perovich and Richter-Menge, 1994) thus their removal from the sample potentially decreased the total bulk salinity of the surface layer which shows a trend of decreasing bulk salinity from day 1 to day 3. It is possible that brine from within this layer has been expelled onto the ice surface and is then wicked upwards by surface tension, increasing the salinity of the frost flowers which were then smoothed off and discarded. As the ice thickens and temperature in the upper layer decreases, more brine is forced up to the surface (Martin et al., 1995). The net effect is a steady decline in the bulk salinity in the upper layer.

Immediately beneath the upper layer, the subsurface layer, only two reliable measurements of bulk salinity are available. The subsurface layer on day 1 (open square) includes the ice-water interface and has a mean bulk salinity of 20.5 psu. The water which is frozen to the bottom of this layer was not completely removed prior to subdivision of the sample, therefore the measured values of bulk salinity are very high, being dominated by the highly saline water. The two complete profiles from day 2 and 3 are C-shaped in form and have a similar bulk salinity in the deepest layer, which contains the skeletal layer. The form of the profiles and the bulk salinity values are comparable with those obtained from young sea ice forming in an Arctic lead (Gow et al., 1990).

The time series of brine volume profiles were calculated using the formulae of Cox and Weeks (1983) and are shown in Figure 6.4(c). As both the temperature and the bulk salinity in the upper layer of the ice decreases with time, the brine volume also decreases. The porosity of the upper layer decreases by a factor of two from day 1 to day 3 which will tend to force brine through the ice matrix by expulsion. Due to the erroneously high bulk salinity in the subsurface layer on day 1, the calculated brine volume of 37.7% is
unreliable. For days 2 and 3 the brine volume increases steadily with depth to approximately 20%. These values of brine volume for ice which hosts the skeletal layer lie within the typical range of 8 to 40% (Maykut, 1985). In all cases, the brine volume is greater than the critical minimum value of 5% required for brine drainage (Cox and Weeks, 1975).

6.3.1.2 Day 1

The raw and interpolated data for bulk salinity measurements in the horizontal plane for new ice sampled on day 1 are seen in Figure 6.5. The ice thickness is 4.5 cm. The section is the layer at the top of the sample, i.e. it is from the air-ice interface to a depth of 2 cm. The mean bulk salinity for this section is 10.9 psu with a maximum of 12.9 psu and minimum of 9.4 psu. The data from the subsurface layer have been rejected for the reasons outlined in section 6.3.1.1. From Figure 6.5(a) and (b), it may be seen that the distribution of brine is heterogeneous. In sea ice which is 4.5 cm thick, there is redistribution of brine in the surface layer.

Figure 6.5: Raw and interpolated forms of the bulk salinity data in the horizontal plane for new ice sampled in the current zone on day 1 of the November 1998 period.

The section is located at a depth interval of 0–2 cm. Line A--A in (a) indicates the location of the vertical section presented in Figure 6.7.
The redistribution causes variability in the bulk salinity values of the subsamples which have a standard deviation of 0.8 psu; approximately twice that calculated for the initial skim. In both the raw and interpolated forms of the data, there are discrete regions of ice which have an increased bulk salinity with respect to the mean salinity of the layer. Specifically, these are at (3,11), (3,7), (7,5), (9,9), (14,9), (14,3), (16,7) and approximately (19,5). In most cases, these areas are clearly demarcated by regions of ice of relatively reduced brine content. The maximum difference in bulk salinity between adjacent subsamples is 2.4 psu, equivalent to a bulk salinity gradient of 1.2 psu cm\(^{-1}\); more than twice that measured for the initial skim in section 6.2.1. The slight trend of increasing bulk salinities towards the right of the figure is most likely attributable to inaccurate sectioning creating samples with oblique horizontal interfaces.

A composite image of ice structure and isohalines for the horizontal section is presented in Figure 6.6. Because the air-ice interface has a granular texture, it is not always suited for photography of brine structures in the ice. Therefore, the image is of the ice at a depth of 2 cm with the bulk salinity data from the layer in the depth interval 0–2 cm, Figure 6.5(a). The isohalines, generated from the interpolated data, are plotted at 1 psu intervals. The grey scaling of the image has been reversed making the brine channels appear as dark structures against the lighter solid phase.

The brine channels in the image are relatively immature compared with the extensive structures seen in chapter 5. However, in most cases, they can be identified as brine channels, having a branched appearance rather than as simple cylindrical tubes. Immediately it can be seen that the areas of ice with enhanced bulk salinity, identified in Figure 6.5, are correlated with individual brine channels. Furthermore, the areas with reduced bulk salinity correspond to areas of ice which are either devoid of brine channels or contain only small channels. A clear example of this is the triangular shaped isohaline centred at (5,5). It is also significant to note that there are numerous brine channels in the image which are located in areas of ice having a bulk salinity which shows no enhancement.
Chapter 6: Evolution of the brine distribution in new ice

Figure 6.6: Composite image of the horizontal section showing the brine structures in the ice with the isohalines for new ice sampled in the current zone on day 1 of the November 1998 period.

A vertical section of the sample, located along (A--A) in Figure 6.5(a), is shown in Figure 6.7. Superimposed on this is the interpolated bulk salinity data for the surface layer of the section, depth interval 0–2 cm. The red data line, with the scale on the right-hand y-axis of the figure, indicates the variation in bulk salinity across the section; it is not an isohaline. Grey scaling has again been reversed, thus brine structures show as dark features. Two such structures are clearly identified at x = 9.5 cm and x = 14 cm. They penetrate the entire thickness of the ice and are branched. There is a clear correspondence between these features and the bulk salinity which peaks precisely at these features. The maximum salinity gradient across the sample is 0.8 psu cm⁻¹. From a homogeneous initial skim, brine redistribution is occurring in conjunction with brine channel development.
Figure 6.7: Composite image a vertical section showing the brine structures in the ice with the bulk salinity for new ice sampled in the current zone on day 1 of the November 1998 period.

6.3.1.3 Day 2

Figure 6.8 shows the raw and interpolated forms of the bulk salinity from the ice sampled on day 2, 48 hours after seeding. The horizontal section, Figure 6.8(a) and (b), is the layer at the depth interval between 2–4 cm, that is the subsurface layer, (H--H) in Figure 6.8(c) and (e). The mean bulk salinity for the layer is 7.6 psu with a maximum and minimum of 9.0 psu and 6.5 psu respectively. Other than a slight increase in the bulk salinity in the region of ice around (10,12), the distribution of brine is quite uniform which is confirmed qualitatively with a standard deviation in bulk salinity of 0.6 psu. This indicates that the variability in brine distribution throughout the subsurface layer is less than that found in the surface layer of ice sampled on day 1. Further, the maximum salinity gradient is only 0.6 psu cm\(^{-1}\).

Vertical sections of bulk salinity for ice sampled on day 2 are also presented in Figure 6.8. Figure 6.8(c) and (d) are the raw and interpolated forms of the data located along y = 1 cm, (A--A) in Figure 6.8(a), and Figure 6.8(e) and (f) show the bulk salinity data for the vertical section located along x = 5 cm, (B--B) in Figure 6.8(a). Both vertical sections have similar characteristics. Variation in bulk salinity is least in the middle depth interval, 2–4 cm. In the surface layer, there are small pockets of increased bulk salinity in Figure 6.8(d) at x = 5 cm and x = 11 cm, and in Figure 6.8(f) at x = 9 cm with a maximum lateral gradient of bulk salinity of 1.5 psu cm\(^{-1}\). In the lowest depth interval the distribution of brine is essentially uniform.
Chapter 6: Evolution of the brine distribution in new ice

Figure 6.8: Raw and interpolated forms of the bulk salinity data for new ice sampled in the current zone on day 2 of the November 1998 period; x and y positions lie in the horizontal plane.

NB. Colour scaling for (c) and (e) differs from (a). Depth interval for (a) indicated by line H--H in (c) and (e). Location of the vertical sections (c) and (e) are indicated in (a) by lines A--A and B--B respectively.
Figure 6.9 shows the composite images of ice structure and the isohalines of bulk salinity for the subsurface layer. The image in the horizontal section, Figure 6.9(a), is from ice at a depth of 2 cm, similar to Figure 6.6. The isohalines are generated from the interpolated bulk salinity data for the subsurface layer, Figure 6.8(b), and plotted at 1 psu.

**Figure 6.9:** Composite image of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for new ice sampled in the current zone on day 2 of the November 1998 period.
Chapter 6: Evolution of the brine distribution in new ice

intervals. Two aspects of the image are noteworthy. First, the brine channels are now more developed being more extensively branched. The typical tip-to-tip distance of the branches, which are the feeder channels, is now approximately 1.5 cm compared to less than 1 cm for the embryonic channels on day 1. The second aspect is that there is no systematic structuring of the brine distribution which can be linked to the position of the brine channels. As observed earlier, and illustrated in Figure 6.8(a) and (b), the lateral distribution of brine in the subsurface layer at this time is heterogeneous.

Figure 6.9(b) is a composite image of ice structure and isohalines of bulk salinity for the vertical section located along (A--A) in Figure 6.8(a) with the isohalines at 1 psu intervals. Here three channels may be identified at \( x = 3 \) cm, \( x = 7 \) cm and \( x = 11.5 \) cm. All penetrate through the entire thickness of the ice and all are more extensively branched than those seen in Figure 6.7. In general, the lateral brine distribution is homogeneous with no correlation between the distribution and the brine channels. One minor exception to this is the isohalines in the surface layer which show a slight deviation at \( x = 11.5 \) cm. Here the salinity gradient attains a maximum of 1.0 psu cm\(^{-1}\). The steepest vertical salinity gradients are those at the ice-water interface.

6.3.1.4 Day 3

Raw and interpolated forms of the bulk salinity data for ice sampled on day 3 are shown in Figure 6.10. The horizontal section, Figure 6.10(a) and (b) is located in the depth interval between 2–4 cm, again it is the subsurface layer, (H--H) in Figure 6.10(c) and (e). The mean bulk salinity of the layer is 7.2 psu with a maximum of 9.4 psu and a minimum of 5.5 psu. Here, the scaling as been set to match the range of the data and the same scaling was used in Figure 6.8(a) and (b) to facilitate the comparison of the data.

A striking aspect of the data, which is more apparent in the interpolated form, is that the areas of greatest bulk salinity are restricted to the lower and left portions of the figures. It is here also that the steepest horizontal salinity gradients are found with a maximum of 1.4 psu cm\(^{-1}\), more than double that found in the sample from day 2. In contrast, the brine distribution in the upper right portion of the figure is considerably more uniform. Overall, the standard deviation in bulk salinity at this depth interval is 0.9 psu, 50% greater than for the same interval 24 hours earlier.
Figure 6.10: Raw and interpolated forms of the bulk salinity data for new ice sampled in the current zone on day 3 of the November 1998 period; x and y positions lie in the horizontal plane.

NB. Colour scaling for (c) and (e) differs from (a). Depth interval for (a) indicated by line H--H in (c) and (e). Location of the vertical sections (c) and (e) are indicated in (a) by lines A--A and B--B respectively.
The vertical sections of bulk salinity for the ice sampled on day 3 are also presented in Figure 6.10 in both the raw and interpolated forms. Figure 6.10(c) and (d) are the vertical section located along \( y = 9 \) cm, (A--A) in Figure 6.10(a), and Figure 6.10(e) and (f) correspond to the section located along \( x = 5 \) cm, (B--B) in Figure 6.10(a). Again, for the purpose of comparison, the scaling of bulk salinity in these figures is identical to that used in Figure 6.8.

A significant feature of the vertical sections is that the regions of increased bulk salinity in the ice are seen to extend vertically through it. No longer is there simple lateral zonation of the brine distribution according to depth. This is seen most clearly in Figure 6.10(d) where there are three instances of these features, at \( x = 5 \) cm, \( x = 12 \) cm and \( x = 16 \) cm. This is less obvious in Figure 6.10(f) but it is still apparent that the distribution of brine in the middle layers of the ice is more heterogeneous than that for day 2. These differences will be discussed in a more quantitative manner in a subsequent section.

The composite image of ice structure and the isohalines of bulk salinity is shown in Figure 6.11. Like the horizontal images for days 1 and 2, the image in Figure 6.11(a) is from ice at a depth of 2 cm with isohalines plotted at 1 psu intervals. A number of very distinct brine channels can be seen in this figure, such as those at (10,11), (4,10) and (6,6). Like those in Figure 6.9, the channels are highly branched with tip-to-tip distances of approximately 2 cm. In the upper part of Figure 6.11(a), there is a region of ice which, other than a small channel at (10,4), is devoid of brine channels. Significantly, it is this area which shows the most heterogeneous distribution of brine contrasting with the lower part of the figure where the distribution becomes increasingly variable.

In general, the areas of ice which contain a brine channel are those which also have increased bulk salinity. Further, the areas of reduced bulk salinity, which can be identified in Figure 6.10(b) and from the isohalines in Figure 6.11(a), correspond to those regions of the ice without brine channels. It is surprising that the distinctive channel at (6,6) is not associated with closely spaced isohalines. However, the channel lies precisely at the junction of four subsamples. Any enhancement associated with this channel will have been divided between these subsamples thus lessening the measured variation of brine in those volumes.
Chapter 6: Evolution of the brine distribution in new ice

The composite image of the vertical section showing ice structure with the isohalines of bulk salinity is given in Figure 6.11(b). The section is located along (A--A) in Figure 6.10(a) and the isohalines are plotted at 1 psu intervals. Now the spatial structuring of brine with respect to the brine channels is clearly evident. The large brine

(a)

(b)

Figure 6.11: Composite image of (a) the horizontal and (b) vertical sections showing the brine structures in the ice with the isohalines for new ice sampled in the current zone on day 3 of the November 1998 period.
Chapter 6: Evolution of the brine distribution in new ice

channel at $x = 5$ cm, which is well developed with extensive branching, is bounded by the 8 psu isohaline. The nature of the brine distribution is now comparable to that seen in the cold ice of section 5.2.1.3 where a volume of ice, with a bulk salinity greater than the surrounding ice, extends vertically through the section. Two other channels can be discerned in Figure 6.11(b) at $x = 11$ cm and $x = 14$ cm. Although these channels appear less well developed than that at $x = 5$ cm, they still influence the brine distribution, again evident from the curvature of the 8 psu isohaline.

6.3.2 Discussion

A great amount of data has been presented in the preceding sections; the discussion seeks to establish the links between them. First, the change in the bulk salinity of new ice by desalination is discussed with reference to other published experimental data, which then leads to a more detailed and quantitative examination of the brine distribution in the three samples of ice. From here, a link is made between the distribution of brine and the changing distribution of the brine channels. Finally, the evolution of the brine channels and their morphology is discussed with reference to the redistribution of the brine and ultimately the desalination of the ice. To conclude the discussion, the key points from each subsection are presented in a synthesis of brine distribution and brine channel formation during the growth of new ice.

6.3.2.1 Desalination of new ice

From the profiles of bulk salinity in Figure 6.4(b), it can be seen that the mean salinity of new ice on each of the three days is considerably lower than the salinity of the water from which it formed. Therefore, it would appear that brine loss, either by rejection at the interface or by drainage, is active within the first 24 hours of ice formation. Because the tank is a closed system, desalination of the ice may be detected from the increase in salinity of the underlying water.

Figure 6.12 shows the water salinity data recorded with a CTD for the first 4 days of ice growth. The CTD sampling rate was 1 per minute and the data are constructed from hourly means. From an initial salinity of 34.8 psu, the salinity of the water
increases steadily to approximately 37.4 psu. The slight kink in the data 1 day after seeding corresponds to the impellers being operated causing full mixing of the water. Open circles indicate the time when samples of ice were obtained. Clearly, salination of the water, which results from desalination of the ice, is active from the first hour of the ice growth. At this time the ice thickness would have been approximately 2 mm.

These observations conflict directly with those reported by Wettlaufer and Worster (Wettlaufer et al., 1997b; Wettlaufer et al., 1997a; Worster and Wettlaufer, 1997) who regard sea ice as a mushy layer, see section 3.4.1. Their experiments on mushy layer formation were conducted using an NaCl solution in a small, insulated ice tank with dimensions 20 × 20 × 40 cm, and cooled from above by a brass cooling plate. With a salt solution of 35.5 psu and a plate temperature of −20°C, they reported no measurable salination of the tank until the mushy layer attained a critical thickness of 6.7 cm, 4.25 hours from the commencement of freezing (Wettlaufer et al., 1997b). This represents an extremely rapid rate of growth compared to that experienced in the Hamburg tank where, with an air temperature of −15°C, it required more than 48 hours to achieve the same ice thickness. It is likely that the ice-surface temperature in the Hamburg tank was greater than −10°C. By using a cold plate at the air-ice interface it is certain that a much colder ice-surface temperature was achieved by Wettlaufer et al.
Chapter 6: Evolution of the brine distribution in new ice

An explanation for the apparent incompatibility of these observations may lie in the vast difference between the ice growth rates for each case. In Arctic field experiments during September, the mean initial ice growth rate in a freezing lead was measured to be 3.7 cm day\(^{-1}\) (Gow et al., 1990). Additionally, the initial growth rate of ice forming in an Antarctic lead in May was measured to be 9.1 cm day\(^{-1}\) (Melnikov, 1995). It is significant to note that these growth rates compare more favourably with the mean growth rate of new ice in the HSVA tank of 4.4 cm day\(^{-1}\) than with the mean growth rate for the mushy layer of 37.8 cm day\(^{-1}\), nearly an order of magnitude greater.

In section 2.5.1 it was explained that the bulk salinity of young sea ice is, to a large extent, controlled by the segregation of brine at the advancing ice-water interface. The amount of brine incorporated into the ice is determined by the effective segregation coefficient, \(K_{\text{eff}}\), in equation 2.4. Using the respective mean growth rates and the published empirical expressions for \(K_{\text{eff}}\) (Cox and Weeks, 1988), its value was calculated to be 0.91 for the mushy layer and 0.30 for new ice grown during day 1. The value of 0.30 concurs with the measured bulk salinity of the ice, being approximately one third of the water salinity from which it formed. In the experiments by Wettlaufer et al., the rapidity at which the ice forms can almost be regarded as a flash-freeze event with an associated high degree of brine entrapment. Therefore, the results of their experiment may only relate to the specific case of very intense freezing in winter rather than the freezing occurring during the late summer and autumn.

The hypothesis proposed to explain the behaviour of the brine in a mushy layer assumes that the ice is a simple porous medium. Therefore, drainage of brine from the mushy layer is determined by a porous-medium Rayleigh number:

\[
Ra = \frac{g\beta\Delta C\Pi(\phi)h}{K\nu}
\]  

(6.1)

In this definition, \(g\) is the acceleration due to gravity and \(\beta\Delta C\) is the brine density difference across the mushy layer with \(\beta\) the solutal expansion coefficient and \(\Delta C\) the concentration difference between the brine in the tank and the brine in the mushy layer, \(h\) is the ice thickness and \(\Pi\) is the permeability of the mushy layer which is a function of
the solid fraction $\phi$. In the denominator, $\kappa$ and $\nu$ are the thermal diffusivity and kinematic viscosity of the brine. So here the Rayleigh number is a measure of the driving buoyancy of the enriched interstitial brine relative to the resistance to fluid flow of the solid ice matrix. It is from this formulation that the concept of a critical ice thickness for brine drainage $h_c$ is derived; this is when a critical value of $Ra$ is exceeded and the brine overcomes the resistance of the ice. The experimental results gathered by Wettlaufer, et al., are found to support this hypothesis (Wettlaufer et al., 1997a).

Although mushy layer theory has been confirmed by experiment, there still remains the disparity between those experiments and the results presented in this chapter. In light of the above discussion, it is proposed that during the formation of a mushy layer ice growth may be so rapid that the natural kinetics of solute diffusion at an advancing, dendritic interface are impeded. A quantitative justification of this interpretation relies heavily on boundary layer theory (Eicken, 1998) and is beyond the scope of this thesis. However, it may be further speculated that the concept of a critical ice thickness for brine drainage is an artefact of the growth mode. Once the ice thickness has increased to this critical point, the growth rate becomes sufficiently slow for standard boundary layer processes to be re-established and consequently for brine drainage to become active. This interpretation is also borne out in mushy layer experiments performed at varying surface temperatures. As the surface temperature is increased, which decreases the growth rate, the critical thickness $h_c$ also decreases (Wettlaufer et al., 1997a). This was interpreted by Wettlaufer, et al., as being the result of differences in the magnitude of the solid fraction according to the thermal driving. However, it could also be argued that with an increase in the surface temperature of the ice the growth will slow sufficiently for brine drainage to become active after a thinner layer of ice has formed.

It can be concluded from this comparative analysis that sea ice forming at growth rates typical of autumn freeze-up should not be regarded as a simple porous medium. The theory of mushy layers is not generally applicable to the formation of sea ice in the polar oceans yet may be a more relevant when describing the phase evolution of sea ice forming under very intense freezing conditions.
6.3.2.2 Brine distribution

The bulk salinity profiles in Figure 6.4(b), established that the distribution of brine in the vertical plane is nonuniform and evolves with time. In addition, the distribution of brine in the horizontal plane has also been shown to be variable. Further, the variability in the bulk salinity at different depth intervals in the ice also changes with time. The degree of variability has been quantified by calculating the standard deviation in bulk salinity of the subsamples from each depth interval. This gives a measure of the changing variability of the brine distribution in the horizontal plane as the ice thickens. The data have been summarised in Table 6.1 for the three samples at three different depth intervals. The term lowest layer applies to the horizontal plate of ice which includes the skeletal layer at the ice-water interface; it occurs at a different depth for days 2 and 3 and has been excluded for day 1 for the reasons given in section 6.3.1.1.

<table>
<thead>
<tr>
<th>Depth interval (cm)</th>
<th>Standard deviation in bulk salinity (psu)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>day 1</td>
</tr>
<tr>
<td>0-2</td>
<td>0.8</td>
</tr>
<tr>
<td>2-4</td>
<td>×</td>
</tr>
<tr>
<td>lowest layer</td>
<td>×</td>
</tr>
</tbody>
</table>

Table 6.1: Standard deviation in the bulk salinity of new ice at selected depth intervals. The lowest layer (which includes the skeletal layer) corresponds to 4.2–6.3 cm for day 2 and 6.3–8.5 cm for day 3.

The depth interval 0–2 cm corresponds to the surface layer of ice at the air-ice interface. Here, although the mean bulk salinity of the layer decreases, as seen in Figure 6.4(b), the variability of bulk salinity in it remains quite constant as the ice grows. This is in contrast to the deeper layers in the ice. The variability in the layer adjacent to the surface layer, the subsurface layer, shows an increase of 50% in a 24 hour period from day 2 to day 3. Likewise, the bulk salinity of the ice in the lowest layer begins to show slightly greater variability as the ice thickens and the brine distribution evolves. It is
likely that the increased variability in the brine distribution is caused by brine relocation through lateral migration within the ice matrix.

It was shown in section 6.2 that the brine distribution in the initial skim of new ice is very much more homogeneous than observed at other times in its later development. Now it is found that the brine distribution in the surface layer of 4.5 cm thick new ice is more heterogeneous than in the initial skim. It is at this point that clearly identifiable brine channels can be seen in the ice. Additionally, Figure 6.6 and Figure 6.7 indicate that the redistribution of the brine is centred on those areas of the ice which contain brine channels. The redistribution of brine in the surface layer of new ice occurs in association with the location and development of the brine channels.

6.3.2.3 Brine channel distribution

Another parameter of the brine distribution that can be quantified relates to the distance between those areas with enhanced bulk salinity. This nearest neighbour analysis is particularly applicable to the surface layer of the ice from day 1. A number of these areas of enhanced bulk salinity can identified in Figure 6.5(b) and were specified in section 6.3.1.2. It is found that the nearest neighbour separation between the centres of these areas is $5.0 \pm 1.0$ cm. However, a close examination of the composite image of the horizontal section in Figure 6.6 reveals that whilst all the areas of enhanced bulk salinity correspond to the location of a brine channel, not all the brine channels are located in areas with enhanced bulk salinity. In effect there are more brine channels per unit area than there are areas of enhanced bulk salinity.

A similar nearest neighbour analysis on the spacing of the brine channels in Figure 6.6 yields a separation of $3.6 \pm 0.6$ cm. Comparing the two distances of nearest neighbour separation confirms that in Figure 6.6 the areas of ice with increased bulk salinity are more widely spaced than the brine channels. It is unlikely that this is a consequence of the limited spatial resolution of the salinity measurements because it is possible to resolve areas of enhanced bulk salinity, found at (14,9) and (16,7) in Figure 6.5(b), which have a spacing of 3.5 cm. Therefore the mismatch between the nearest
neighbour spacing of the brine channels and of those areas of enhanced bulk salinity is not an artefact of the sample processing.

It is possible to take this nearest neighbour analysis further because all the composite images of the horizontal sections are derived from photographs of ice at the same depth, 2 cm. The pertinent images are in Figure 6.6, Figure 6.9(a) and Figure 6.11(a); a summary of the nearest neighbour separation of brine channels in each is given in Table 6.2. Additionally, it is also possible to calculate the spatial density, $D_c$, of the brine channels in each horizontal section. This parameter of brine channel distribution was discussed in sections 3.4.2.2 and 4.4.4.1 and the same units will be adopted here; number of channels per 100 cm$^2$. Because only one sample is available in each case, no standard deviation has been calculated.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>day 1</th>
<th>day 2</th>
<th>day 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nearest neighbour separation of</td>
<td>3.6 ± 0.6</td>
<td>4.2 ± 0.8</td>
<td>4.6 ± 0.9</td>
</tr>
<tr>
<td>brine channels (cm)</td>
<td>(n = 54)</td>
<td>(n = 35)</td>
<td>(n = 21)</td>
</tr>
<tr>
<td>Spatial density of brine channels</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(number per 100 cm$^2$)</td>
<td>8.5</td>
<td>5.8</td>
<td>4.2</td>
</tr>
</tbody>
</table>

Table 6.2: Nearest neighbour separation of brine channels and their spatial density at a depth of 2 cm in new ice; $n$ = number of nearest neighbour pairs.

It can be seen that as the ice cover thickens and develops, the nearest neighbour spacing of the brine channels decreases. However, by day 3 the spacing of the channels is comparable to that on day 2. It would appear from these observations that after 72 hours, the spacing of the channels may approach a steady state. It should be noted that the spacing of the channels, at approximately 5 cm, is entirely consistent with that observed in other examples of young sea ice (Wakatsuchi and Saito, 1985).

Furthermore, the increase in the nearest neighbour spacing is manifest in the spatial density of the channels which is found to decrease with time. Over a period of 48 hours, the spatial density decreases by a factor of two from 8.5 per 100 cm$^2$ to 4.2 per
100 cm$^2$. However, the spatial distribution of channels in the sample from day 3, Figure 6.11(a), is highly variable with a large area of ice being devoid of channels. This has the effect of decreasing the spatial density of brine channels to a value which is perhaps unrepresentative of the ice at that time. Notwithstanding this, the spatial density of channels in the ice on day 2 and 3 is comparable to the upper and lower bounds of the spatial density calculated on day 8 of the same experiment which was summarised in Table 4.2.

6.3.2.4 Development of brine channel morphology

Concurrent with the increase in the nearest neighbour spacing of the brine channels there is also an observable change in the morphology of the channels themselves. In the vertical axis, the majority of channels penetrate the entire thickness of the ice. This confirms previous interpretations of brine channel evolution, namely that once a channel becomes established then its vertical development matches that of the advancing interface. Along the horizontal axis, the lateral extent of the channels increases with time. The increase in the effective diameter of the structure (rather than the diameter of the central drainage channel) is by the increase in branching with development of feeder channels. This was quantified with the typical tip-to-tip distance measured from the composite images of horizontal sections, Figure 6.6, Figure 6.9(a) and Figure 6.11(a). This quantity increased from less than 1 cm on day 1 to approximately 2 cm by day 3.

As a brine drainage channel develops a system of feeder channels it becomes a more effective sink for brine. The increase in branching allows brine from a wider catchment area, or a greater volume of ice, to migrate along the feeder channels towards the central channel. It should be noted that the feeder channels are inclined towards the central channel, a characteristic which has been widely reported (Lake and Lewis, 1970; Niedrauer and Martin, 1979; Morey et al., 1984; Cole and Shapiro, 1998). The inclination of these channels gives a horizontal component to the direction of brine migration along them. Therefore, it is likely that the existence of the inclined feeder channels facilitates the increasing lateral variability in the brine distribution. Brine from the surrounding ice volume migrates toward a brine channel from which it is ultimately drained from the ice.
6.3.2.5 Synthesis

Following from these analyses of the lateral distribution of brine, and the distribution and morphology of brine drainage channels, it is now possible to form a clearer concept of brine distribution and brine channel development during the growth of new ice. It was shown that the distribution of brine in the initial skim of ice is relatively homogeneous. As the ice grows there is redistribution of the brine, increasing the lateral variability of the bulk salinity. It was seen from Figure 6.6 and Figure 6.7 that redistribution coincides with the development of feeder channels.

One of the principal features observed in the horizontal section of ice from day 1 is the mismatch in the distribution of brine in the surface layer and the nearest neighbour spacing of the channels. The spacing between areas of increased bulk salinity was calculated at 5.0 ± 1.0 cm. There was found to be an inequality between this quantity and the spacing of the brine channels at a depth of 2 cm which was 3.6 ± 0.6 cm. During a 48 hour period, when the ice thickness increased from 4.5 cm to 8.5 cm, the brine channel spacing at the same depth had increased to 4.6 ± 0.9 cm. There is now a good agreement between the distribution of brine in the surface layer on day 1 and the nearest neighbour spacing of the brine channels on day 3.

Therefore, it is postulated that the change in brine channel spacing occurs by a process involving both brine enrichment and depletion. Up to day 1, closely spaced channels form in the ice. Of these channels, some will be more efficient sinks for brine from the surrounding ice volume. These channels become enriched in brine creating areas of ice registering an increased bulk salinity. Less efficient channels will not have an abundant supply of brine and thus will be in an area of ice without any increase in the bulk salinity. Therefore, the brine channels which form in the initial stages of ice growth can be categorised into those which are enriched in brine and those which are relatively depleted of brine.

Without a source of brine, a brine channel cannot persist. Therefore, those channels which are relatively depleted of brine will decay, and may freeze and close entirely. Conversely, those which are enriched in brine will continue to develop. The formation of feeder channels, most likely by fracturing at grain boundaries, increases the effectiveness of the central channel as a sink for brine.

159
6.4 Summary

In this chapter the evolution of the brine distribution and the structure and spacing of brine channels in new ice has been analysed. Two sets of data were obtained, one from a thin skim of ice and the other consisting of a time series of samples. In previous discussions of brine distribution in sea ice, chapter 5 and (Cottier et al., 1999), it was assumed that the distribution in initial skim was homogeneous. From the measurements of bulk salinity in a sample of new ice 2.6 cm thick, this assumption was shown to be valid. The variability in the bulk salinity was estimated to be less than 0.5 psu. Allied to this is the lack of developed brine structures in the ice. Although some vertical channels could be identified, there was no definite correlation between them and the small variations observed in brine distribution. However, it is important to indicate that brine channels in their most basic form may be found in ice with a thickness of less than 3 cm.

The time series data provided valuable indications of the redistribution of brine as the new ice evolved. It was seen from CTD data that brine is rejected from the ice immediately after its formation. The conflict existing between these observations and those for a mushy layer was resolved by acknowledging the effect of ice growth rate on solute diffusion at the ice-water interface. It was also noted that redistribution of brine was evident in the surface layer of 4.5 cm thick ice. Further, as the ice thickened, the variability in the brine distribution in all depth intervals increased. The brine distribution in new ice was found to be correlated with the location of the brine channels.

On day 1, in new ice 4.5 cm thick, well defined brine channels are present. They extend through the entire thickness of the ice and are closely spaced, with a small nearest neighbour separation. However, a mismatch exists between the distribution of brine and the spacing of these immature channels. As the ice thickens the spacing between the channels increases and they become more highly branched, developing a network of feeder channels. The effective diameter of the brine drainage channel increases thus increasing its potential catchment area for brine. The nearest neighbour spacing of the channels on day 3 was equated with the distribution of brine in the surface layer of ice on day 1. Further, it was proposed that brine channels in young sea ice survive or decay depending on their efficiency as a sink for brine, a condition which is determined during the first centimetres of ice growth.
Chapter 7

Aspects of brine distribution in frazil ice

7.1 Introduction

The essence of this chapter is the distribution of brine observed in sea ice which has a predominantly granular texture. In chapters 5 and 6, the observations of brine distribution were restricted to samples of congelation ice with a columnar crystal texture. In section 2.3.1, the mechanisms of formation, and the geographical distribution, of sea ice with these two distinct crystal types were described. The essential difference between them is that the columnar texture is the product of a thermodynamic, or congelation, growth mode whilst frazil ice, having a granular crystal texture, forms under dynamic conditions where the water column is turbulent.

The term frazil ice has been adopted for this chapter to conform to the nomenclature used for natural sea ice which has a granular crystal texture (Tison et al., 1998). The method used to prepare the ice sample is described in section 7.2. Although the method of ice production used is not perfectly representative of the process by which frazil ice forms naturally, that is, it does not utilise wave and wind action in situ, it produces an ice type with a vastly different crystal texture to that investigated in the preceding chapters. The extent to which the sample is comparable to natural frazil ice will be assessed in section 7.3.1. The nature of the brine distribution will be investigated and discussed, with reference to field samples, in section 7.4. In addition, the geometry of the brine structures found in ice of granular texture will be inferred and contrasted with those in ice having a columnar texture. To explain these contrasts, arguments based on the differences in the ice formation mechanisms and crystal structure are presented.
Chapter 7: Aspects of brine distribution in frazil ice

7.2 Frazil formation

The investigation of frazil ice was conducted during the November 1998 period of Interice II. At this time, the temperature regulation and environmental boundary conditions in the tank ensured that all ice formation was by quiescent, thermodynamic growth. To prepare a sample of frazil within the sheet of congelation ice the following method was adopted which had proved successful in previous experiments requiring the formation of a granular ice cover (Lindemann, 1998).

A separate, small basin (2 × 1 × 1 m) was filled with tap water and Instant Ocean synthetic sea salt added. The salinity of the water in the basin was initially 32 psu and it was circulated using pumps to maintain turbulence in the water. Furthermore, a fan was directed to blow air across the basin to increase the turbulence and the heat flux from the water surface. Using this arrangement, a steady production of individual crystals was maintained. The crystals were plate-like with a typical cross section between 2 and 3 mm. As the quantity of crystals within the basin increased it was possible to gather them using a sieve. These frazil crystals were then transferred to the main experimental tank.

On day 4 of the November 1998 period an area of open water, with dimensions approximately 0.5 × 0.5 m, was made in the congelation ice cover. The open water was just upstream of the impellers at position (12,5) in Figure 4.1 where ice thickness was approximately 13 cm. The frazil crystals, transferred from the small basin, were then

![Figure 7.1: A cross section through the ice cover illustrating the formation of a frazil layer within the congelation ice.](image-url)
decanted into the open water and dispersed by hand to ensure their even distribution through the upper part of the water column. More crystals were added to form a loose agglomeration which became natural isostatic. The intention was to simulate a situation where frazil crystals in a turbulent water column rise to the surface as conditions become more quiescent forming a layer of crystals typically 1–10 cm thick (Weeks and Ackley, 1986). To achieve this, care was taken to avoid compacting the ice excessively and thus maintain the uncongealed nature of the surface layer. Using this method, a layer of suspended frazil crystals with a thickness of approximately 11 cm was produced in the tank. This arrangement is clarified in Figure 7.1

At this time in the experiment the air temperature was approximately −15°C and remained so for a period of approximately 48 hours, see Figure 4.3. The growing conditions allowed the ice to consolidate and form a solid mass as the interstitial liquid froze. Because the frazil was bounded on all sides by solid ice, the vertical component of the heat flux through the frazil/brine mixture was the most significant. On day 6, the cooling was stopped and the air temperature increased rapidly to approximately 0°C. Sampling of the ice was done on day 7 and followed exactly the same method as in section 4.5.

### 7.3 Results

All the results in this section are based on the analysis of a single sample of frazil ice. Unlike the results in chapters 5 and 6, neither thermal forcing nor time series investigations were undertaken. Similarly to previous chapters, the results include bulk profiles, bulk salinity measurements and a direct comparison of the brine distribution with brine structures in the ice. In addition to these data, a description of the textural properties of the sample will be given; this establishes the distinction upon which this chapter is based. The frazil ice was sampled on day 7 of the November 1998 experimental period. The ice thickness, measured from an ice core, was then 11.5 cm, with a variability resulting from the undulating topography of the ice-water interface estimated at approximately ±1.0 cm.
7.3.1 Textural analysis

Thin section analysis of the frazil ice, Figure 7.2, reveals that it is composed almost entirely of granular ice. In this example of the crystal structure, the upper 11 cm are seen to be wholly granular in texture whilst in the lowest 5 mm a delicate layer containing larger, columnar crystals is evident. Grain size in the granular region is typically 3 mm and the grains have a random orientation. This arrangement of crystals gives the ice an isotropic character. The occurrence of columnar crystals at the ice-water interface indicates that during the consolidation period of the loose agglomeration of crystals, thermal equilibrium of the ice-brine composite was attained and a positive temperature gradient established. Consequently, the heat flux along this gradient initiated congelation growth at the interface; a situation that is also observed in the lower layers of pancake ice.

Figure 7.2: Vertical thin section of the frazil ice fabric seen through crossed polarisers. Image courtesy of C. Haas, AWI. The scale graduations in the top left are at 1 mm.
Chapter 7: Aspects of brine distribution in frazil ice

The fabric analysis has substantiated two important requirements of the method for forming frazil ice. First, the crystal texture of the ice is comparable to that which forms under natural and turbulent conditions. Frazil ice crystals have sections which are generally less than 5 mm (Eicken et al., 1995; Weeks, 1998) and in the pancake ice commonly found in the Greenland Sea, typical crystal size is approximately 1–2 mm (Wadhams and Wilkinson, 1996). Second, a simple comparison of Figure 7.2 and Figure 4.13 confirms that the fabric of the two ice types formed in the tank are vastly different both in the crystal size and shape, and their orientation. Therefore, this chapter is concerned with an ice type which has a very different textural character to those studied in chapters 5 and 6.

7.3.2 Bulk profiles

The bulk salinity and brine volume profiles are determined from the block of frazil ice retrieved for subsampling. Because of the unevenness of both the air-ice and ice-water interfaces, these had to be trimmed to facilitate accurate sawing of the plates. The trimming process reduced the thickness of the sample to 10.5 cm. Thus, all the profiles presented in Figure 7.3 are for a sample 10.5 cm thick.

Ice temperature was measured on a core that was extracted from the frazil area prior to obtaining the block sample. The temperature profile, Figure 7.3(a), comprises measurements made at 20 mm intervals using a digital thermometer with a resolution of ±0.1°C. In the lowest 6 cm of the ice the temperature gradient is positive with a mean value of +0.08°C cm⁻¹. However, the single measurement at a depth of 2 cm indicates that in the upper 4 cm of the sample the gradient had reversed and become negative, estimated at −0.05°C cm⁻¹. At the time of sampling, the air temperature had been elevated for approximately 26 hours and during this the ice cover was undergoing active re-equilibration. Although the positive temperature gradient is of small magnitude, the small skeletal layer at the ice-water interface indicates that during consolidation the ice had attained thermal equilibrium and established a negative heat flux resulting in active ice growth at the interface.
Chapter 7: Aspects of brine distribution in frazil ice

Figure 7.3: Vertical profiles of (a) temperature, (b) bulk salinity and (c) brine volume for frazil ice sampled on day 7 of Interice II, November 1998.

The profile of bulk salinity for the frazil sample, Figure 7.3(b) has a C-shaped form. This is an important observation because it confirms that the brine in the frazil sample was able to move rather than it being a stagnant ice/brine composite. One of the propositions for the origins of a C-shaped profile, which is commonly observed in young sea ice, is that the efficiency of brine rejection at the growing interface increases as the growth rate decreases with increasing ice thickness. However, because the frazil sample was formed from a suspension of crystals, the growth rate argument is not applicable. Assuming that the composition of ice crystals and brine was uniform throughout the initial suspension, i.e. the ratio of solid to liquid was constant, then the appearance of the curved profile indicates that brine migrated during consolidation. Further discussion of this feature will be given in section 7.4.

The vertical profile of brine volume, Figure 7.3(c), was calculated using the equations of Cox and Weeks (1983). As no measurements of ice density were made, it was necessary to assume that the gas volume in the ice was negligible. From the preparation, and the appearance of the ice, it was clear that the gas content of the ice was
greater than for samples of columnar ice. However, it is apparent from the brine volume that, in this instance, the brine content will determine the major part of the porosity. Although a critical value for brine volume to allow vertical migration has been calculated at 5% (Cox and Weeks, 1975), it was derived from measurements on congelation ice. It is not known whether this criterion for permeability also holds for purely granular ice but it is anticipated that, in this instance, the sample will be permeable at all depths.

7.3.3 Bulk salinity

The raw bulk salinity data for the three orthogonal sections of frazil ice are presented in Figure 7.4. The data for the horizontal section of the sample are shown in Figure 7.4(a) and (b), the latter being the interpolated form. In this instance, the plate of ice is 2 cm thick from a depth interval between 4–6 cm, at the midpoint of the sample, (H--H) in Figure 7.4(c) and (e). Mean bulk salinity for the layer is 8.5 ± 0.4 psu with a maximum of 9.3 psu and a minimum of 7.5 psu. This restricted spread of bulk salinity values is indicated in the scaling of Figure 7.4(a) and (b), the range extending over 4 psu only (compared to a range of 7 psu for Figure 5.2(a)).

It is apparent from both the figures, and the standard deviation of the mean bulk salinity, that in the horizontal plane the distribution of brine is very homogeneous at the spatial resolution of the measurements. Furthermore, the maximum difference in bulk salinity between adjacent samples is only 1.2 psu; three standard deviations from the mean. Therefore salinity gradients across the section are minimal in comparison to those of samples of cold ice described in chapters 5 and 6. An area of slightly enhanced bulk salinity can be found in the centre of the sample around x = 6–10 cm and y = 3–9 cm but this enhancement is less than 2 standard deviations above the mean bulk salinity.

Raw bulk salinity data for the two vertical sections of the frazil sample are presented in Figure 7.4(c) and (e) together with their respective interpolated forms in Figure 7.4(d) and (f). The vertical sections are located along (A--A) and (B--B) in Figure 7.4(a) which correspond to Figure 7.4(c) and Figure 7.4(e) respectively. In each case, the lowest layer of the section is thicker than the others, an arrangement dictated by
Figure 7.4: Raw and interpolated forms of the bulk salinity data for frazil ice on day 7 of the November 1998 period. NB. Colour scaling for (c) and (e) differs from (a). Depth interval for (a) indicated by line H--H in (c) and (e). Location of the vertical sections (c) and (e) are indicated in (a) by lines A--A and B--B respectively.
the practicalities of sawing the ice. In each case, this thick layer of ice shows the greatest bulk salinities, consistently above 10 psu.

The distribution of the bulk salinity values in both the raw and the interpolated data indicates that within each layer the bulk salinity is relatively uniform; the standard deviation for each horizontal plane further confirm this. For all layers, other than the surface layer, standard deviation in the bulk salinity is 0.4 psu. In the surface layer this increases to 1.0 psu, the most likely cause being the pocket of relatively high bulk salinity (12.3 psu) at y = 3 cm in Figure 7.4(e). Treating this subsample as an outlier and discarding it, the standard deviation for the surface layer decreases to 0.5 psu which concurs with that of the other three layers. The high bulk salinity at that point is consistent with splashing of water from the tank onto the surface during sampling.

7.3.4 Brine structures

The composite images presented in Figure 7.5 are similar to previous images in that they show both the ice structure overlaid with isohalines of constant bulk salinity. However, in this case, the grey scaling has not been reversed, therefore areas of ice which have a high proportion of precipitated salts appear lighter than those with a higher ice fraction. Variation in the brightness across the images is more an effect of uneven illumination of the ice than of variation in its salt content. The horizontal section, Figure 7.5(a), corresponds to the ice at a depth of 4 cm and the salinity data are that from Figure 7.4(a). It is evident that the nature of the granular ice structure is significantly different from that observed in samples of congelation ice. There is very little contrast in the image and no branched brine channels can be identified. Small points of brightness can be seen in the figure, for example at (14,6) and (2,4.5); these are spherical pockets of air trapped in the solid matrix. The isohalines have been plotted at 1 psu intervals which emphasises the homogeneous nature of the brine distribution. There is no observable link between the isohalines and the ice structure in the horizontal plane.

Figure 7.5(b) represents the vertical section of frazil ice located along (A--A) in Figure 7.4(a) and the superimposed isohalines are derived from the bulk salinity data shown in Figure 7.4(c). Between the depths of 2.2 and 4.2 cm a row of ice is missing
Figure 7.5: Composite images of the (a) horizontal and (b) vertical sections showing the structure of the ice with the isohalines for frazil ice sampled on day 7 of the November 1998 period.
from the image because of an error in the sawing process. However, the three included rows indicate that throughout the ice, there is no incidence of vertical brine channels and that its structure is uniform. Furthermore, there was little difference between the structural appearance of the four horizontal plates, such as that shown in Figure 7.5(a). Therefore it is assumed that the ice at the missing depth interval in Figure 7.5(b) is essentially similar in nature to the other three. Like the horizontal section, many small bright points can be seen in Figure 7.5(b) which are air pockets trapped in the ice. The isohalines are plotted at 1 psu intervals; they run laterally across the sample with minimal vertical deviation. In none of the photographs of vertical or horizontal sections of frazil ice could clearly identifiable brine channels be found. The ice is observed to be entirely devoid of these structures, having a very uniform appearance which corresponds to in the homogeneous nature of the brine distribution.

7.4 Discussion

The data presented in section 7.3 are from a single sample of frazil ice formed in the tank. The ice was allowed to consolidate from a suspension of preformed crystals and was sampled whilst the ice still had a predominantly positive temperature gradient and possessed a thin, dendritic skeletal layer at the ice-water interface. Therefore, according to the terminology defined in section 5.2, the ice can be regarded as cold ice. This is an important and necessary requirement to enable the discussion to focus on the textural contrast between sea ice samples in comparable thermal regimes.

All aspects of the bulk salinity and ice structure data for frazil ice have illustrated a common theme; homogeneity. The vertical thin section of the frazil ice in Figure 7.2 showed that the ice is isotropic with respect to crystal orientation and dimensions. In contrast to congelation ice, there is no predominant orientation of the crystals and there is no elongation of the crystals in the direction of the heat flux. This is the result of the formation mechanism by which a suspension of uniform plate-like crystals consolidates by freezing of the interstitial brine. This indicates the important distinction between brine entrapment in congelation and in frazil ice. Brine incorporation into congelation
Chapter 7: Aspects of brine distribution in frazil ice

ice is by a steady process of advancement of a dendritic interface within which brine is retained and trapped as pockets between individual crystals and crystal platelets. In frazil ice formation, the ice crystals in suspension are surrounded by interstitial brine. During consolidation of the suspended material there is no systematic incorporation of brine, it is retained en masse at the crystal boundaries.

The matrices of bulk salinity illustrate the homogeneity of frazil ice further. Brine distribution in frazil ice is characterised by low variability in the salt content of the solid ice matrix. This is evident in the horizontal section, Figure 7.4(a) and (b), which shows minimal gradients in the bulk salinity in all directions across it. Furthermore, it has been calculated that at all the depth intervals of the ice, standard deviation in the bulk salinity of the individual subsamples is consistent and of low magnitude. This is in significant contrast to the distribution of brine in samples of cold ice with columnar crystal structure which is typically highly variable with steep gradients in bulk salinity. Arguably the clearest indication of the homogeneous nature of frazil ice is seen in Figure 7.5. The ice shows no evidence of brine structures within the solid matrix. No features resembling the vertically oriented, branched structure, which is now the well established signature of a brine channel, appear in any of the photographs of the ice. The superimposed isohalines reaffirm the theme of uniformity of salinity and structure which has been demonstrated to occur in frazil ice.

The absence of brine channels in the frazil sample is significant with regard to the behaviour and properties of this ice type compared to those of congelation ice. Brine channels are considered as ubiquitous features in sea ice so it is important to establish whether the frazil sample formed in the tank is representative of natural sea ice with a granular crystal texture. There are few description of brine structures in purely frazil ice. Eicken et al., (1995) has presented profiles of stratigraphy and brine features for cores of summer sea ice in the Arctic. From these observations it would seem that the occurrence of brine channels is more closely correlated with ice of columnar texture rather than with granular texture. However, there are regions of granular ice through which brine channels penetrate though these are found to lie beneath a layer of columnar ice which also contains channels. Possibly these channels, although not normally associated with granular ice, will continue through the granular region to maintain a continuous
hydraulic connection. Observations of Antarctic sea ice also indicate that brine channels are more closely associated with congelation ice having columnar crystal texture (Eicken, *pers. comm.*, 1999). The conclusion here is that the absence of brine channels in frazil ice formed in the tank is entirely consistent with field observations.

Simply considering the difference in the formation processes of frazil ice and congelation ice provides some explanation for the absence of brine channels in the frazil sample. In section 3.4.2.1, a detailed description of the crystallographic factors which affect brine channel formation was presented. It was stated that their formation required favourable orientation of the crystals to allow brine to migrate along well defined paths located at the crystal and platelet boundaries. With the systematic incorporation of brine at the skeletal layer, and the steady advancement of the interface, there is considerable organisation of the brine within the ice matrix. This facilitates the controlled development and maturing of brine drainage channels, with brine migrating towards them along inclined grain boundaries. In contrast, the formation process for frazil ice is rapid, generating random orientation in the crystal structure. Thus there is no organisation in the system which becomes highly isotropic. The crystallographic environment is not conducive to brine channel development with brine residing in the ice as irregular pockets at the grain boundaries (Eicken and Lange, 1989). It would appear that microstructural control is important in initiating the formation of brine channels.

Whilst it has been demonstrated that there is lateral uniformity of the brine distribution, the vertical distribution has been shown to vary, causing the development of a C-shaped profile. Given the uniformity of the crystal texture it seems reasonable to assume that, prior to consolidation, the suspended mass of crystals established a very consistent composition of solid and liquid fractions. The fact that a C-shaped profile exists implies that brine has migrated through the ice during the period between formation of the suspension and subsequent sampling. Vertical redistribution of the brine is occurring despite the absence of brine channels which would normally act as the conduits.

It is useful to compare the bulk salinity profiles observed in the frazil sample with profiles found in a natural frazil system. For this, data obtained from pancake ice during a field cruise to the Odden region of the Greenland Sea (Wadhams *et al.*, 1997) are
Chapter 7: Aspects of brine distribution in frazil ice

presented with the tank-grown frazil in Figure 7.6. The profile for the tank-grown frazil (solid line) is identical to that shown in Figure 7.3(b). Profiles for the two pancake samples are shown as the dotted and dashed lines. In both cases the pancake ice is thicker than that sampled in the tank; 15 cm and 19 cm compared to 10.5 cm. Air temperature during the days when the pancakes were sampled was approximately −5°C. Therefore, ice temperatures were likely to have been high with a weak temperature gradient through the ice. In both cases the pancakes were reported to be soft which indicates that the ice was highly porous.

![Graph showing bulk salinity profiles](image-url)

**Figure 7.6:** Profiles of bulk salinity found in the frazil sample formed in the ice tank (solid line) and two samples of pancake ice (dotted and dashed lines) recovered from the Odden region of the Greenland Sea.

Although the bulk salinities of the pancake ice are generally lower than those of the tank ice they all show a C-shaped form to the profile. Clearly, what has been observed in the vertical brine distribution of frazil formed in the tank, is not unlike the natural situation. This confers validity to the method and subsequent evolution of the ice and brine. From the comparison presented, it would appear that the vertical redistribution of brine in frazil ice to create a C-shaped profile is a natural phenomenon.
The differences in bulk salinity could be due to three factors. First, the processing methods to obtain the bulk salinity data are quite different. For the tank ice, the method is designed to minimise brine drainage compared to that used on the ship whereby there is an appreciable delay between the ice being removed from the water and the subsampling being completed. Therefore the propensity for draining is significantly different in the two samples. Second, the two pancakes will perhaps have been subject to freshwater precipitation which will tend to flush brine through the ice. Third, variations in the age and thickness of the pancake ice and the tank ice ensures that their histories of brine redistribution and drainage will be different.

From the data in this chapter and observations of frazil ice forming in the polar regions, the processes of brine channel formation and brine drainage in this ice type are seen to be quite different to that in congelation ice. The form of the bulk salinity profiles have indicated that vertical redistribution of brine is an active process. However, the images in Figure 7.5 indicate that the paths available for brine migration are not well defined. To account for this apparent inconsistency in observation it should be remembered that, during consolidation of the crystals, brine will be retained within the ice matrix at the boundary of the individual crystals. Given that the distribution of these crystals is quite uniform throughout the ice then the brine filled boundaries may form an isotropic porous network. Therefore, in the randomly oriented granular ice, brine will tend to move through the brine networks (similar to those seen in Figure 3.2(a)) rather than through vertical brine channels in the case of well ordered columnar ice. During consolidation of the crystal suspension, the increasing temperature gradient will have the effect of driving brine through the interconnected network by expulsion and percolation. In this way the brine becomes redistributed causing the C-shaped profile. However, the lack of brine channels to act as foci for lateral brine migration maintains the lateral uniformity in brine distribution.
7.5 Summary

The aim of this chapter has been to determine how brine is distributed in sea ice which has a granular crystal texture; to provide a textural contrast with the samples of ice which had been studied in preceding chapters. A method of producing granular ice was described and by thin section analysis it was shown to produce ice of quite different textural character to that of congelation ice. Further, the crystal structure is comparable to that observed in naturally occurring frazil ice. The matrices of bulk salinity data have shown that in frazil ice, the lateral distribution of brine is highly uniform at all depths. There are found to be no significant salinity gradients across the ice. Photographic images of the ice have confirmed that no brine channels formed in the frazil ice and that the ice structure was uniform throughout.

The absence of brine channels, which are regarded as ubiquitous in sea ice, is consistent with the sparse observations of field samples. It would appear that the formation process of frazil ice is not conducive to the development of these structures and indicates the importance of the crystallographic organisation required for their existence. The profile of bulk salinity indicated that, during consolidation and thermal equilibration, vertical brine migration was active. By comparing this profile with the salinity profile from two samples of pancake ice from the Greenland Sea it was argued that vertical brine migration in frazil ice is natural. With the absence of a clearly defined vertical path for migration, the concept of brine networks between crystals was introduced. These networks provide an interconnected system through which the brine can move either by expulsion, driven by the change in ice porosity, or by percolation. This concept retains the lateral uniformity observed in the brine distribution whilst permitting vertical brine migration to account for the development of a C-shaped profile.
Chapter 8
Summary of conclusions

One of the most fundamental properties of sea ice is its salt content. Salt, in the form of brine, is incorporated in the solid ice matrix during its growth. As the ice ages the brine is lost by desalination processes. Many of the properties of sea ice will vary according to its brine content and the brine which is lost during desalination influences oceanic processes. Brine is fundamental to the interactions of sea ice within the polar environment; thus the aim of this thesis was to advance the knowledge of brine distribution in young sea ice.

The salt content of sea ice is known to vary vertically; this is manifest in the frequently obtained bulk salinity profiles. However, the salt content is also observed to vary in the horizontal plane which might be indicative of lateral brine migration. Allied to the form of the brine distribution is the occurrence and location of brine drainage channels in sea ice. Before reviewing these structures a scheme of classification for the brine features found in sea ice was proposed, based on their simple geometric characteristics, and a nomenclature defined. The scheme proposed has proved successful throughout the thesis, providing consistency and clarity to the discussions. A thorough review of the current ideas on the formation and evolution of brine channels was made. This synthesis of the pertinent literature linked previously independent observations to develop a life history of brine drainage channels.

To investigate the nature of the brine distribution in young sea ice, and to determine the influence of brine channels on it, experimental work was conducted in a large ice tank. Using the facility, sea ice could be formed in a controlled environment which allowed ready access to the ice. Traditional methods of sampling sea ice are
fraught with the problem of brine loss from the sample by drainage. To overcome this, a novel technique for sampling ice has been described which was successfully employed to maximise the retention of brine and preserve its spatial arrangement in the sample. The technique, although very simple, provides the means for a very detailed study of brine processes in young sea ice. A fundamental question when using ice tanks is whether the ice forming under artificial conditions is representative of natural sea ice. By considering the growth kinetics, crystal structure and brine structures of the ice which formed in the tank it has been demonstrated that it is a suitable proxy.

The investigation of brine distribution in young sea ice focused on three distinct areas. First; the changes in the brine distribution and its correlation with the brine channels with respect to the ice temperature and thus ice porosity. The brine content of cold sea ice was found to be heterogeneous in the horizontal plane with areas of enhanced bulk salinity being collocated in areas of the ice containing brine channels. In warmer ice with greater porosity the distribution of brine was found to be more uniform and less dependent on the location of brine channels. Refreezing of ice to induce a positive temperature gradient restored the heterogeneous nature of the brine distribution. From these observations, mechanisms for the redistribution of brine within the ice were proposed.

Second; the evolution of the brine distribution and the concurrent development of brine channels in new ice. The bulk salinity of new ice was found to be less than the salinity of the water from which it formed, indicating that under these growth conditions there was immediate rejection of brine from the ice. Further evolution of the brine distribution was seen to coincide with the maturation of embryonic brine channels. It was hypothesised that the spatial density of the brine channels was linked to the efficiency of individual channels to act as a sink for brine. This generated an increasingly variable lateral distribution of brine.

Third; aspects of the brine distribution in a granular ice type which extended the application of this research to sea ice of different textures. It was found that frazil ice, with a granular texture, was devoid of distinct brine channels and that the brine distribution was uniform on the scale of the measurements. It was hypothesised that the
growth mechanism for frazil ice impeded the formation of brine drainage channels, and that migration of brine was mediated by an interconnected brine network.

The evolution of sea ice on the regional or meso-scale is controlled by physical processes which occur on a much smaller scale. The research described in this thesis is the first attempt to studying both the brine and the structures in sea ice simultaneously and at a scale on which these processes operate. The mechanisms that ultimately influence the large-scale evolution of sea ice have been elucidated and described.

8.1 Brine distribution and its variation with ice temperature

One of the objectives, stated in the introduction of the thesis, was to determine to what extent brine distribution in young sea ice is correlated with the occurrence of brine drainage channels. Lateral variability in the bulk salinity of sea ice, derived from field measurements, was reviewed in section 1.3.1.1. There, brine channels were propounded to be the source of this variability but without direct supportive evidence this could remain as speculation only. This research represents a systematic study of the distribution of salt, principally in the form of brine, in relation to the occurrence of brine drainage channels in young sea ice. The results have demonstrated conclusively that the variability of brine distribution in congelation ice can be explicitly linked to the occurrence of brine drainage channels.

Using a technique for retaining brine in a sample and by subsampling the ice, on a scale comparable with the lateral dimensions of the brine channels, is was possible to investigate the correlation between the bulk salinity and the channels. The distribution of brine within sea ice is found to be variable at the centimetre scale. Furthermore, it has been demonstrated that the distribution can be closely correlated with the location and morphology of individual brine drainage channels. However, it was shown that the degree of variability in the brine distribution is dependent on the physical environment of the ice, its thickness, temperature and crystal structure.

A major component of the thesis concerned the redistribution of brine in sea ice during thermal forcing, thus causing a reversal in the temperature gradient through it.
With a change in ice temperature there was an accompanying change in its porosity and permeability. In actively growing sea ice, with a positive temperature gradient and low ice temperatures, brine distribution is very heterogeneous. Regions of ice containing a brine channel show an enhancement in their salt content compared with regions devoid of brine channels which are relatively depleted in brine. In contrast, the lateral brine distribution in warm, ablating ice with a negative, or isothermal temperature gradient, shows a greater degree of uniformity. There is no clear association between the minimal variability observed in the distribution and the brine channels. That is, the uniformity in the salt content of warm sea ice is independent of the location of brine channels. These observations fulfilled another of the initial objectives, namely to investigate the redistribution of brine according to changes in the ice temperature.

Evidently, changing the temperature gradient through the ice sheet induces redistribution of the brine. During the current investigation, two different regimes were simulated; melt and refreeze, both yielding a considerable change in the observed brine distribution. Founded on these observations, mechanisms causing the redistribution of brine, linked to changes in the ice porosity, were proposed. During a melt phase, with the porosity and permeability of the ice increasing, it was concluded that vertical percolation of surface meltwater through the brine channels acts to flush brine through the ice. Additionally, the enhanced connectivity of the brine inclusions facilitates lateral migration of the brine such that it permeates throughout the ice. Both mechanisms, acting together, have the effect of reducing the lateral variability of brine yielding a more uniform distribution.

In the reverse process, refreeze, the uniformly distributed brine becomes relocated within the ice matrix. It was proposed that the brine migrates towards the brine channels as the freezing front passes through the ice. Brine expulsion was cited as the driving force for migration. As the porosity within the ice decreases, brine is forced along a path of least resistance towards those areas of the ice with greater porosity, namely the brine drainage channels. In conclusion, as the ice cools, the initially uniform brine distribution observed in warm sea ice becomes increasingly heterogeneous and collocated in those areas containing brine channels.
8.2 Evolution of brine distribution and brine channel morphology

Chapter 6 reported the observations of brine distribution and structures in new ice. This is the first investigation, using a large, three-dimensional ice cover, of how these two associated aspects of new ice are linked and evolve. One of the aims of the experiments was to determine the nature of the brine distribution in the first skim of ice. In 2.6 cm thick sea ice, the distribution was found to be relatively homogeneous in comparison to thicker ice where it was found to be highly variable. It was demonstrated that during ice growth the distribution evolves by lateral migration of the brine, thus increasing its variability.

A seemingly obvious, yet crucial observation of the formation of new ice is that the bulk salinity of the initial skim was less than the salinity of the water from which it formed. Furthermore, the salinity of the water in the closed tank system was increasing within an hour of seeding. This leads to the clear conclusion that ice begins to reject brine at the initial stage of its formation. Given the uniformity of the brine distribution and the absence of mature brine channels in the first skim of new ice it is probable that the brine is lost at the dendritic ice-water interface.

These conclusions are found to be inconsistent with those from a similar tank experiment (Wettlaufer et al., 1997a; Wettlaufer et al., 1997b). The ice which formed in that experiment was termed a mushy layer and the brine was observed to be initially confined within the solid ice matrix before the attainment a critical ice thickness for the onset of brine drainage. In chapter 6, it was argued that the conditions under which the mushy layer formed corresponded to an extremely rapid growth rate; one which would not generally be experienced in nature except perhaps at the surface of a lead in midwinter. Consequently, the dynamics of brine incorporation at a dendritic interface may be exceptional in that case. Therefore, it was concluded that the theory of phase evolution which has been developed for mushy layers in general is not necessarily applicable to the case of sea ice forming with more typical growth rates.

The development of brine channels as distinctive features within the ice matrix occurs at an early stage in the formation process. It has been shown that such structures may form in sea ice less than 3 cm thick. Once brine channels become established, the
variability in the lateral distribution of brine is observed to increase substantially. This evolution of the brine distribution is consistent with the development of the embryonic brine channels into mature brine drainage channels which are uniformly spaced and have attendant feeder channels. As the channels become more branched they become increasingly efficient sinks for brine. Furthermore, their nearest neighbour spacing increases towards a constant value. It was proposed that the change in the spatial density of the channels is linked to differences in their efficiency as sinks for brine, with some channels maturing and others eventually decaying.

8.3 Contrasts of the brine distribution associated with ice texture

The scope of the research included measurements of the brine distribution in sea ice with a granular crystal texture. From the observations described in chapter 7 it was shown that the spatial arrangement of brine in cold frazil ice is vastly different from that found in cold congelation ice. In frazil ice, the lateral distribution of brine is homogeneous at all depths and the ice is devoid of large brine channels with a branched structure. However, it was argued that vertical redistribution of brine through the ice was active thus necessitating a path for its migration.

Comparing the observations of brine structures in cold sea ice from chapters 5 and 7 allows two contrasting morphological arrangements of the liquid fraction to be identified or inferred. First, in congelation ice with a columnar crystal texture, the brine is in vertical channels which, in young ice, may penetrate through the entire thickness. This work has shown conclusively that brine is located preferentially in these structures which are the primary path for vertical migration. Second; when the crystal texture changes to a granular form, the morphology of the brine volume changes to one of an amorphous brine network. The principal conclusion is that brine migration in frazil ice occurs by vertical percolation through the intergranular network. Further, it was concluded that the absence of brine channels in frazil ice is a consequence of the ice forming under turbulent conditions and is therefore characteristic of a granular ice type.

Regarding the nomenclature described in section 3.3.3, the description of an intergranular brine network does not fit easily into any of the terms defined.
completeness, the nomenclature should be extended to include this arrangement of brine. A brine network can not be classed as a collection of brine layers as this term is reserved exclusively for the brine located between platelets in congelation ice. It would be logical to include brine networks as a subclass of brine structures which in general contribute to the migration of brine and desalination of the ice. Therefore, it is proposed that intergranular brine network be incorporated into the nomenclature as a new class of brine structure.

8.4 Future directions

Central to the research has been the novel method of ice acquisition and sampling. It has been demonstrated that this method is versatile, simple and may be used to study sea ice of varying thickness and temperature. Whilst this research has concentrated solely on the distribution of brine in the ice, there is scope to extend the application to other constituents of the ice. As discussed in chapter 3, brine channels have a central role in the transport and biological properties of the ice. To understand the brine redistribution process, and thus the transport properties, of the ice more clearly, measurements of the stable isotope $^{18}$O at a high spatial resolution would be valuable. This would enable the origins of the liquid phase to be established and therefore determine the extent to which vertical flushing of melt water, lateral permeation and sucking (section 5.3.2.2) contribute to the restructuring of the brine. Further, ice tanks have been used for culturing ice organisms to accumulate a substantial amount of biomass. Similar high resolution studies of the nutrient status and biota of the ice in association with the development of brine channels may provide new insights into how a sea ice community develops during ice growth.

One of the most fundamental aspects of sea ice formation is the fate of the interstitial brine. It would appear from the disparity in observations that no clear consensus exists. New theories of treating sea ice as a mushy layer, whose fluid dynamics can be expressed via a Rayleigh number, seem to conflict with established conventions that brine is lost from sea ice from the commencement of its formation. The
conclusions of this research point in favour of the latter although it was acknowledged that under extreme freezing conditions a different mechanism for brine loss may be active. Clearly this is a key area of sea-ice research and one which demands attention. To resolve the disparity, both practical and theoretical advances will be required. Ice tanks offer great advantages for pursuing experimental work in this field.

Much of the understanding about desalination processes in sea ice is confined to congelation ice. As this work has concluded, the distribution of the brine and the morphology of the brine structures found in sea ice having a granular texture is quite different. It is likely that the kinetics and mechanisms of desalination in frazil ice will be different from those of congelation ice. Areas in the Marginal Ice Zone where frazil ice production is greatest have been intensively studied in recent years (Wadhams, 1996; Wadhams et al., 1996). A central feature of these areas is the formation of pancake ice and their contribution to the oceanic salt flux. A constructive area of future research would be to develop a quantitative approach to desalination that was specific to ice of frazil origin. Pancake ice has been grown successfully in the HSVA tank and it is under these controlled conditions that a careful study of its formation, evolution and desalination might be achieved.

Finally, it should be noted that all the areas nominated for potential future research are aimed at clarifying the mechanisms for the evolution of the brine content of sea ice. It is the mobile, labile nature of brine which, to a large extent, determines the properties and interactions of sea ice. This research has extended the knowledge of this influential component of sea ice.
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196


