Winter Dynamics in an Epishelf Lake

by

Jérémie Bonneau

B.Ing., Université Laval, 2017

A THESIS SUBMITTED IN PARTIAL FULFILLMENT
OF THE REQUIREMENTS FOR THE DEGREE OF

Master of Applied Science

in

THE FACULTY OF GRADUATE AND POSTDOCTORAL
STUDIES
(Civil Engineering)

The University of British Columbia
(Vancouver)

August 2020

© Jérémie Bonneau, 2020
The following individuals certify that they have read, and recommend to the Faculty of Graduate and Postdoctoral Studies for acceptance, the thesis entitled:

**Winter Dynamics in an Epishelf Lake**

submitted by **Jérémie Bonneau** in partial fulfillment of the requirements for the degree of **Master of Applied Science in Civil Engineering**.

**Examinining Committee:**

Bernard Laval, Civil Engineering, The University of British Columbia  
*Co-supervisor*

Derek Mueller, Geography and Environmental Studies, Carleton University  
*Co-supervisor*
Abstract

The last intact ice shelf in the Canadian Arctic is located at the mouth of Milne Fiord (82.6°N, 81.0°W), on Ellesmere Island, Nunavut. During melt season, the ice shelf acts as a dam preventing meltwater from flowing freely to the ocean. This results in a permanent layer of freshwater that "floats" on top of the seawater of the fjord. This layer of freshwater is called an epishelf lake. The winter data from a mooring installed in Milne Fiord epishelf lake (2011-2019) is analysed in the framework of a one dimensional model in order to study 1) mixing in the upper water column and 2) the evolution of a basal channel in the ice shelf. The results show that vertical mixing is surprisingly higher in the epishelf lake than in the seawater underneath. Estimation of the Richardson number using geostrophic balance indicates that enhanced mixing in the epishelf lake is associated with horizontal temperature gradients. Moreover, the analysis suggests that all of the freshwater reaching the ocean travels through a single basal channel in the ice shelf. The model did not detect significant variation in outflow characteristics over the eights years of study, implying that the basal channel area is in ice mass balance.
Lay Summary

At the mouth of Milne Fiord, in Nunavut, is the last intact ice shelf in Canada. An ice shelf is a thick floating sheet of ice attached to the land. Because Milne Ice Shelf is attached to the land on both sides of the fjord, it acts like a dam preventing fresh meltwater from the watershed to directly reach the ocean. This layer of freshwater floating on top of the seawater is called an epishelf lake. This study uses field observations and a numerical model to conclude that there is more mixing in the epishelf lake than in the seawater below. This is surprising because Milne Fiord epishelf lake is ice-covered year round. Moreover, this study suggests that most of the water flowing out of the epishelf lake follows a channel under the ice shelf and that this channel is not evolving rapidly.
Preface

This is a continuation of the work done Andrew Hamilton in Milne Fiord from 2009 to 2016. The data analysis and the writing was realized by the author under the supervision of Dr. Laval (University of British Columbia) and Dr. Mueller (Carleton University). The data analysed was collected by the Environmental Fluid Mechanics lab (EFM) from the University of British Columbia (which the author is part of), the Water and Ice Research Laboratory (WIRL) from Carleton University and the Tahoe Environmental Research Center (TERC) from the University of California at Davis.

A version of chapter 3 of this thesis is being prepared for submission to a peer-reviewed journal as *Winter Dynamics in the Last Epishelf Lake in the Canadian Arctic, Milne Fiord, Nunavut, Canada* by J. Bonneau, B. E. Laval, D. Mueller, A. K. Hamilton, A. M. Friedrichs and A. L. Forrest. For this chapter, the data analysis and the writing was realized by the author under the supervision of Dr. Laval (University of British Columbia) and Dr. Mueller (Carleton University). Dr. Hamilton provided mooring and CTD data (before 2017) as well as comments and insights on the work. A.M. Friedrichs and Dr. Forrest provided the ADCP data.
# Table of Contents

Abstract ................................................................. iii

Lay Summary ........................................................... iv

Preface ................................................................. v

Table of Contents ...................................................... vi

List of Tables .......................................................... ix

List of Figures ........................................................ x

Acknowledgments ....................................................... xvi

1 Introduction ......................................................... 1
   1.1 Arctic Climate Change and Ice Shelf Loss ...................... 1
   1.2 Milne Fiord .................................................. 2
   1.3 Objectives ................................................... 4
   1.4 Motivation and Implications .................................. 5
   1.5 Thesis Structure ............................................ 6

2 Literature Review .................................................. 7
   2.1 Milne Fiord .................................................. 7
       2.1.1 Milne Glacier .......................................... 8
       2.1.2 Milne Ice Shelf ....................................... 10
       2.1.3 Milne Fiord Epishelf Lake ......................... 10
2.1.4 Below the Epishelf Lake .............................................. 11
2.2 Mixing, Energy, Geostrophy ........................................... 13
  2.2.1 Eddy Viscosity ..................................................... 13
  2.2.2 Stability ............................................................ 14
  2.2.3 Energy and Mixing ................................................ 14
  2.2.4 Geostrophic Scale ............................................... 16
2.3 Ice-Covered Lakes ....................................................... 17
  2.3.1 Seiche ............................................................. 18
  2.3.2 Radiative Mixing ................................................ 18
  2.3.3 Inflows ............................................................ 19
  2.3.4 Ice Walls .......................................................... 19
  2.3.5 Diffusive Convection ............................................. 19
  2.3.6 Earth Rotation .................................................... 20
  2.3.7 Tides ............................................................... 20
2.4 Arctic Fjords .......................................................... 20
  2.4.1 Wind ............................................................... 21
  2.4.2 Meltwater Plumes ................................................ 21
  2.4.3 Tides ............................................................... 22
  2.4.4 Earth Rotation .................................................... 23
2.5 Basal Channels ........................................................ 23
  2.5.1 Formation .......................................................... 24
  2.5.2 Evolution .......................................................... 24
  2.5.3 Melting ............................................................ 25
  2.5.4 Impacts of Basal Channels ...................................... 26
3 Winter Dynamics in the Last Epishelf Lake in the Canadian Arctic,
  Milne Fiord, Nunavut, Canada ........................................ 27
  3.1 Key Points ........................................................... 27
  3.2 Introduction .......................................................... 28
  3.3 Geophysical Setting and Study Area Background ............... 31
    3.3.1 The Fjord ......................................................... 31
    3.3.2 The Glacier ..................................................... 32
    3.3.3 The Ice Shelf .................................................. 32
List of Tables

Table C.1 Model evaluation statistics . . . . . . . . . . . . . . . . . . . . 77
## List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Schematic of Milne Fiord. The ice shelf acts like a dam preventing the freshwater (epishelf lake) to flush in the ocean. In <em>italic</em> are possible processes affecting the ice shelf-epishelf lake-glacier system.</td>
<td>2</td>
</tr>
<tr>
<td>1.2</td>
<td>Geographical location of Milne Fiord.</td>
<td>3</td>
</tr>
<tr>
<td>2.1</td>
<td>Meteorological data from Purple Valley weather station (Figure 2.2). A) Hourly air temperature at 2 m from the ground. Horizontal dashed line is 0°C. B) Hourly downwelling solar radiation.</td>
<td>8</td>
</tr>
<tr>
<td>2.2</td>
<td>Satellite image, bathymetry and ice draft map of Milne Fiord. A) ASTER image of Milne Fiord (July 21st 2009). The glacier grounding line is in red. MIS: Milne Ice Shelf. MEL: Milne Fiord Epishelf Lake. MGT: Milne Glacier Tongue. MLSI: Multi-year land-fast sea ice. The red triangle is MEL mooring and the yellow is the weather station. B) Bathymetry map. Note: shallow section in the red circle is in fact not shallow, it is ~300 m. Circles refer to CTD cast locations. The bathymetry data comes from lead line sounding and CTD casts. C) Ice draft measurements from ice penetrating radar, airborne radar and remote sensing. D) Modeled ice draft with measurement from C. This figure comes from Hamilton (2016).</td>
<td>9</td>
</tr>
</tbody>
</table>
Figure 2.3 CTD water profile taken in May 2011 at the location of MEL mooring (Figure 2.2). A) Conservative temperature (Θ), with the different water masses labeled. MEL: Milne epishelf lake, freshwater, 0 to 4°C. FMW: Fjord modified water, 5 to 30 g kg⁻¹, 0 to -1.5°C. PW: Polar water, 30 to 34.4 g kg⁻¹, temperature lower than 0°C. AW: Atlantic water, >34.4 g kg⁻¹, ~0°C. DW: deep water, as AW but generally slightly fresher and colder. B) Salinity profile. C) Stability profile, smoothed with a 20 point moving average; note the logarithmic transformation.

Figure 3.1 A) Location of Milne Fiord (82.6°N, 81.0°W). B) Landsat 8 image of Milne Fiord taken in September 2018. MEL: Milne Epishelf Lake, MIS: Milne Ice Shelf, MGT: Milne Glacier Tongue, MG: Milne Glacier.

Figure 3.2 Schematic of Milne Fiord including Milne Ice Shelf (MIS), Milne Epishelf Lake (MEL) and Milne Glacier Tongue (MGT). A) During the melt season (summer) (~June 1st to ~September 1st), runoff drives the deepening of MEL. B) During the remainder of the year, the thickness of MEL decreases slowly toward the minimum draft of the MIS.

Figure 3.3 Schematic of the one-dimensional model. Typical absolute salinity and stability profiles are on the right. The data from the uppermost and the 25 m thermistor are used as boundary conditions for temperature and the no flux boundary conditions are used for salinity. Mixing coefficients of the top freshwater layer (K<sub>top</sub>) and the bottom seawater layer (K<sub>bot</sub>) are parameters of the model; only molecular mixing is considered for the halocline layer. The outflow layer is between the minimum draft of the ice shelf (h₀) and the bottom of the halocline layer (z). The top and bottom dashed lines show the top and bottom boundaries of the model. The two middle dashed lines are the top and bottom of the halocline layer (molecular diffusivity only).
Figure 3.4  Schematic of the outflow of the lake through the basal channel of the ice shelf. A modified weir equation using a two layer simplification (equation 3.3) is used to constrain the number of parameters related to the outflow. A) Top view. B) Along fjord section. C) Across fjord section through MIS. Note: not to scale .......................... ........................................ 40

Figure 3.5  Once the initial conditions (IC) and the boundary conditions (BC) are implemented into the mesh, the model loops through the possible outflow coefficients ($C_e b$ and $h_0$) for every winter and through the possible mixing coefficients for every winter days. (See text for detailed explanation) ...................... 42

Figure 3.6  Temperature timeseries of the top of the water column in Milne Fiord. A) Conservative temperature timeseries of the model results. B) Conservative temperature timeseries from the mooring (linearly interpolated). C) Difference between the model and the mooring data, positive values mean the model temperature is higher than the mooring temperature. Triangles show the location of the thermistors on the mooring line and the squares show the location of the conductivity instruments. 44

Figure 3.7  Salinity data from the model (solid line) at the depth of the conductivity instruments on the mooring (dashed line). Labels are the depth of the instruments, in meters. ......................... 45

Figure 3.8  A) Depth of MEL (maximum $N^2$ value) given by the model for each winter. The staircase effect is due to the vertical discretization of the model (10 cm). Outflow parameters $h_0$ and $C_e b$ for each winter are in the legend. B) Outflow through the basal channel in MIS. Dotted lines are the model estimation using the rectangular weir equation and an estimated lake area of 71.2 km$^2$. Solid lines are the ADCP estimation using the channel morphology data from Rajewicz (2017). .................. 47
Figure 3.9  A) Daily mixing coefficients from the model, for the heat and salt transport equations for the top and bottom layers (dots). Solid lines are the 30 day averaged quantities for the top (blue) and bottom (red/orange) coefficients of the heat equation. Minimum possible values of daily mixing coefficients are the molecular diffusivities; $1.4 \times 10^{-7}$ m$^2$ s$^{-1}$ for heat and $1.4 \times 10^{-9}$ m$^2$ s$^{-1}$ for salt. B) 30 day averaged mixing coefficient of the model (dark blue) and inverse of the Richardson number (light blue) computed according to equation 3.6. $N^2$ (dotted line) computed by the model and 30 day maximum density difference attributed to the temperature oscillations $\Delta \rho_{30}$ (dashed line) are used in equation 3.6. Enhanced mixing (dark blue line) is linked to stronger eddy activity (dashed line) hence to a lower Richardson number (light blue line).

Figure 3.10  30 day averaged mixing coefficient of the model ($K_{TOP}^\Theta$) as function of the Richardson number estimated by equation 3.6. Commonly used parameterizations by Peters et al. (1988) (green) and Pacanowski and Philander (1981) (cyan).

Figure A.1  Parameterization of $K_{SA}$ as function of $K_{\Theta}$.
Figure B.1 RADARSAT-2 Fine Quad image taken on the 27th of April 2013 showing a dark disc in the middle of the Milne Fiord epishelf lake (Courtesy of the Canadian Ice Service, Environment and Climate Change Canada). Panels VV, VH and HH are three different polarizations (VV: vertical-vertical, VH: vertical-horizontal and HH: horizontal-horizontal) and the RGB panel is the combined Pauli decomposition (HH: red, VV: green and VH/HV: blue). The disc is visible in all polarizations. Color and gray scales are qualitative with brighter tones indicating higher backscatter. RADARSAT-2 Data and Products © MacDonald, Dettwiler and Associates Ltd. (2013) – All Rights Reserved, RADARSAT is an official mark of the Canadian Space Agency.

Figure D.1 Getting dropped off at Purple Valley campsite by a Kenn Borek Twin Otter. Maximum 2500 pounds, crew included... Photo: Jérémie Bonneau

Figure D.2 Milne Fiord, looking downfjord. The glacier grounding line area is visible on the right of the icefall and Purple Valley, 75% on the right. The ice-covered ocean on the horizon. Photo: Jérémie Bonneau

Figure D.3 Northern Ellesmere icefields, looking upfjord from the same location as Figure D.2, ~ 5000 feet up. Photo: Jérémie Bonneau

Figure D.4 Derek Mueller and Drew Friedrichs looking at Milne Ice Shelf, easily distinguishable by its rolling topography with the through filled with melt water. Photo: Jérémie Bonneau

Figure D.5 Moving the camp from Purple Valley to the ice shelf. Drew Friedrichs signaling “upward” to the helicopter pilot. Photo: Jérémie Bonneau

Figure D.6 Drew Friedrichs doing a CTD profile after pulling the epishelf lake mooring out. Looking upfjord. Photo: Jérémie Bonneau

Figure D.7 CTD profile in a crack on the margin of the glacier tongue. Photo: Drew Friedrichs
Figure D.8  Pulling the ice penetrating radar around the grounding line of the glacier. Photo: Drew Friedrichs . . . . . . . . . . . . . . 86
Figure D.9  Drew Friedrichs "hole melting" to get the ice shelf channel mooring back. Photo: Jérémie Bonneau . . . . . . . . . . . . 87
Figure D.10  Picking up a camera from the summit of Mega Nunatak (unofficial name). Photo: Derek Mueller . . . . . . . . . . . . 88
Figure D.11  26-Resolute, 26-Resolute, this is Purple Valley, Purple Valley. Drew Friedrichs and Yulia Antropova broadcasting (blurred). Photo: Jérémie Bonneau . . . . . . . . . . . . . . 89
Figure D.12  Looking up fjord, with Derek Mueller. Mega Nunatak (unofficial name) on the left of the fjord. Photo: Jérémie Bonneau . . 90
Acknowledgments

I would like to express my gratitude towards many people and organizations;

• The National Sciences and Engineering Research Council of Canada (NSERC), The University of British Columbia (UBC) and the Fond de Recherche du Québec Nature et Technologie (FRQNT) for providing the daily bread.

• The Northern Scientific Training Program (NSTP), by Polar Knowledge Canada, for fieldwork monetary assistance.

• The Polar Continental Shelf Program (PCSP) for providing founding and logistic assistance necessary for fieldwork...and killer Nanaimo bars.

• Dr. Bernard Laval for giving me the opportunity to go check out some of the most remote and beautiful "corners" of this planet, as well as for the freedom and guidance that led to these pages.

• Dr. Derek Mueller for teaching me the fieldwork ropes, improving my rickety english writing and keeping calm when I entangled 800 m of CTD line on my first day.

• Drew Friedrichs and Yulia Antropova for help and great company in the field.

• Dr. Andrew Hamilton for the amazing work you have done in Milne Fiord and for the feedback and insights. I often feel like I stole your project.

• All the people of the EFM lab for the great discussions and the lovely Deeks hike; the hut is rusty, but only on the outside.
- Mes camarades PERY-BON-Kiens pour le divertissement quotidien, me rapprochant de la maison.

- Mes parents, Annie et Vincent, et toute ma famille pour le soutien inconditionné.
Chapter 1

Introduction

1.1 Arctic Climate Change and Ice Shelf Loss

The area of the planet where climate change is the most pronounced is its northernmost part: the Arctic. Conversely, not many people live there and there is little news coverage when a glacier disappears or when an iceberg many times the size of Manhattan breaks off in Greenland. Arctic sea ice has shown accelerated loss in extent and thickness from 1990 to 2008 (Lang et al., 2017; Maslowski et al., 2012), which plays a major role in Arctic temperature regulation (Anisimov et al., 2007; Lang et al., 2017). Air temperature near the surface has increased more than two times faster than the global average (Screen and Simmonds, 2010). In some areas, air temperature rose by as much as 5°C in the last century (Anisimov et al., 2007). In the Canadian High Arctic, every major glacier of the Queen Elizabeth Islands is showing a decrease in surface elevation (Mortimer et al., 2018). Changes are unmistakable on the northern coast of Ellesmere Island, where the ice shelf extent has been decreasing episodically over the last century (Mueller et al., 2017) This has been linked to higher air temperatures (Copland et al., 2007).

Ice shelves are floating masses of ice attached to the coast. They are formed by the extension of glaciers over the ocean, by in situ accumulation of snow and by aggregation of sea ice (Mueller et al., 2017). Ice shelves can be many hundreds of meters thick and are much larger in Antarctica than in the Canadian Arctic. On the northern coast of Ellesmere Island, there used to be an ice shelf at the mouth of
Figure 1.1: Schematic of Milne Fiord. The ice shelf acts like a dam preventing the freshwater (epishelf lake) to flush in the ocean. In italic are possible processes affecting the ice shelf-epishelf lake-glacier system.

every fjord (Mueller et al., 2017). When an ice shelf is grounded on both sides of a fjord and is not connected with the upstream glacier, it creates a dam trapping fresh meltwater from the fjord watershed (Laybourn-Parry and Wadham, 2014). This layer of freshwater floating atop the seawater forms what is called an epishelf lake (Figure 1.1). Milne Fiord epishelf lake depends entirely on the ice shelf blocking the inlet of the fjord; if it fractures completely from north to south, the freshwater will simply disperse into the Arctic Ocean. The recent fracture and break up of ice shelves in the Canadian Arctic has left one last intact ice shelf in Milne Fiord and thus, the last remaining epishelf lake of the Canadian Arctic (down from \(\sim 17\) in 1906 (Laybourn-Parry and Wadham, 2014; Veillette et al., 2008)).

1.2 Milne Fiord

Milne Fiord is located on the northern coast of Ellesmere Island, Canada’s northernmost island (Figure 1.2). As depicted in Figure 1.1, the fjord system is composed of the glacier, the glacier tongue (floating part of the glacier), the epishelf lake and the ice shelf, going downfjord. All of the above are in contact with the
Figure 1.2: Geographical location of Milne Fiord.

seawater, emphasizing their interconnection.

The most striking feature of the water column in Milne Fiord is the very sharp salinity gradient at the halocline (∼10 g kg⁻¹ m⁻¹). Hamilton et al. (2017) defined the bottom of the epishelf lake as the point where the density (salinity) gradient is the highest. The epishelf lake experiences an annual cycle of deepening and shoaling (Hamilton et al., 2017). During the summer, freshwater runoff pushes the halocline down, increasing the thickness of the lake. Then, when air temperature drops below zero, melting stops and the lake thickness decreases until the next summer. The lake is approximately 10 m deep and the difference between the start and the end of the melt season is around 2.5 m.

Milne Fiord is perennially ice-covered, which prevents the direct impact of wind. Moreover, tides are very limited (15 cm amplitude). Current measurements in the top 100 m of Milne Fiord by Hamilton (2016) have shown horizontal velocities around 1 cm s⁻¹ that only rarely exceed 5 cm s⁻¹. This calm environment with low tidal amplitudes and no wind implies there is little circulation and mixing in the fjord, especially in winter, when there is no surface runoff and likely no subglacial discharge.

Similar to all major glaciers of the Queen Elizabeth Islands, Milne Ice Shelf is in negative mass balance (i.e. it is thinning) (Mortimer et al., 2012, 2018). Its
thickness varies between \(\sim 8 \text{ m}\) and \(\sim 95 \text{ m}\). The thinnest areas are at its margins and along an east-west depression across the middle of the ice shelf. CTD and ice-penetrating radar measurements suggest that this east-west depression is the surface expression of a basal channel where the epishelf lake water exits the fjord (Hamilton, 2016; Rajewicz, 2017). This basal channel is analogous to those found under ice shelves in Greenland and Antarctica and thought to be related to ice shelf stability (Dow et al., 2018; Rignot and Steffen, 2008). Melting rates over 10 m a\(^{-1}\) are often observed in Antarctica and Greenland basal channels (Rignot and Steffen, 2008; Stanton et al., 2013). This is certainly not the case for the Milne Ice Shelf basal channel, since at that rate it would have melted through already. However, it is not known if the channel is evolving or if it is stable.

1.3 Objectives

A mooring was deployed in the center of Milne Fiord epishelf lake in May 2011 and has been recording continuously until present. The mooring data from 2011 to 2014 was used by Hamilton (2016) to study the annual cycle of the epishelf lake, but it has not been analyzed since. The eight years of mooring data available are one of the most extensive oceanographic data sets for an Arctic fjord. The objective of this study is to use the mooring data in order to:

1. Quantify the mixing occurring in the epishelf lake and associate it with forcing mechanisms.

2. Confirm that the channel is the main outflow path for the epishelf lake and to get information on the evolution of its morphology.

In order to use most of the mooring data in a simple and meaningful way, a one-dimensional model is used. This allows focus on the most important physical parameters (mixing and outflow through the channel). This study has two main limitations, one spatial and one temporal. Because instruments on the mooring line become very sparse below 25 m (3-4 instruments from 30 m to 350 m), this study focuses on the upper water column above 25 m. The second limitation is related to the different processes affecting the water column during the melt season (\(\sim\)mid-June to \(\sim\)early-August). During that period of the year, surface runoff, solar
radiation and melt water plumes can have a significant impact on the upper water column (Hamilton, 2016; Keys, 1977; Straneo and Cenedese, 2015). Because these processes add extra complexities in the mooring analysis, the melt season period is not considered in this study.

1.4 Motivation and Implications

In addition to other epishelf lakes in Antarctica, Milne Fiord shares physical characteristics with ice-covered lakes and fjords. In these isolated water bodies, relatively uncommon circulation and mixing mechanisms can have a disproportionate impact. This is because the ice-cover suppresses the effect of wind and solar radiation, when capped with snow (Keys, 1977; Perkin and Lewis, 1978). It then appears that Milne Fiord (ice-covered all year) is an ideal setting to study these more obscure processes. Objectives 1 and 2 are in line with this and it is anticipated that new knowledge acquired here will improve our understanding of Arctic fjords and ice-covered lakes.

As mentioned above, the basal channel in Milne Ice Shelf is very similar to those found on glacier tongues and ice shelves of Antarctica and Greenland. Hence, knowledge acquired here can be transferred to these systems. Accordingly, it is anticipated that the third objective will lead to an increased comprehension of the role basal channels play in the stability of ice shelves and in ice-ocean interactions. At a more local scale, due to the strong interconnection between epishelf lakes and their environment, they are considered to be ”sentinel ecosystem” (Veillette et al., 2008). For example, the depth of Milne Fiord epishelf lake varies throughout the year due to runoff from the watershed and outflow under the ice shelf, hence, studying the fluctuation of the epishelf lake thickness will also increase the knowledge of the hydrology and glaciology of Milne Fiord. The realization of each objective will lead to a improved map of the interactions between ice and water in Milne Fiord.

From a biological point of view, Milne Fiord is remarkable because the presence of the epishelf lake allows for the presence of freshwater organisms just above marine organisms (Jungblut et al., 2017; Veillette et al., 2008). Analysis of the microbiota in Milne Fiord (Veillette et al., 2011) showed that the composition of the phy-
toplankton and bacteria communities in the freshwater layer and in the seawater below were absolutely distinct. Even though Milne Fiord epishelf lake will likely disappear in the next decade or two, the study of this unique system is important on a local scale, to monitor the rapid environmental change in the Canadian High Arctic, and on a global scale, to provide insights on the evolution of other similar systems in Greenland and Antarctica.

1.5 Thesis Structure

The core of this thesis is chapter 3, which will be submitted to a peer-reviewed journal in the near future. Chapter 2 is an overview of the study site and a literature review on physical processes in ice-covered lakes and Arctic fjords as well as on basal channels found under glacier tongues and ice shelves. Appendices A, B and C are supplementary material to chapter 3. Finally, Appendix D is a fieldwork photo gallery.
Chapter 2

Literature Review

This literature review is divided in five parts. The first is a overview the study site and associated literature that points out the main features of Milne Fiord and gives the reader a good understanding of the Milne Fiord system. It is an extended version of the geophysical setting and background of study section of Chapter 3. The second part reviews some physical oceanography and limnology concepts that are relied on in this study. Finally, sections 3-4 are reviews of the relevant physical processes of ice-covered lakes, Arctic fjords and ice basal channels, respectively.

2.1 Milne Fiord

Milne Fiord (82.6°N, 81.0°W) is situated on the north coast of Ellesmere Island (Figure 1.2). The closest permanently occupied settlements are Alert, a military base 260 km to the east and Eureka, a weather station to the 300 km south. Milne Fiord is in the Arctic climate zone (Figure 2.1). Data from a weather station in a bay of Milne Fiord (Figure 2.2, data courtesy: Luke Copland) shows an annual average air temperature of -18.6°C with a 10 year maximum of 19°C and a minimum of -55°C. Because of its high latitude (82.5°N), the sun stays below the horizon from mid-October to early March. The melt season is typically from early-June to mid-August, however, freezing temperatures are possible at any time of the year. The area experiences between 100 and 300 positive degree days per year (Hamilton, 2016).
The work done by Hamilton (2016) describes the site in detail (Figure 2.2). The fjord is 44 km long from the outermost extent of the ice shelf to the grounding line of the glacier. From the innermost extent of the ice shelf to the grounding line, its width is approximately 6 km. Alike Greenlandic fjords, there is a sill in Milne Fiord. It is just downfjord of the epishelf lake and its depth is \( \sim 220 \) m. The maximum depth below the epishelf lake is \( \sim 400 \) m. From this point, the seafloor climbs upfjord to reach \( \sim 150 \) m at the grounding line (Hamilton, 2016).

### 2.1.1 Milne Glacier

Milne Glacier (MG) is over 50 km long and is 4 to 5 km wide. It has a thickness of approximately 150 m at the grounding line (Hamilton, 2016) but becomes much thicker upglacier (Narod et al., 1988). From the grounding line, The Milne Glacier Tongue (MGT) extends 15 km down the fjord. Its thickness decreases rapidly moving away from the grounding line and it is less than 10 m on its margins. MGT broke away from the glacier in 2009 but has not moved significantly since.
Figure 2.2: Satellite image, bathymetry and ice draft map of Milne Fiord. A) ASTER image of Milne Fiord (July 21st 2009). The glacier grounding line is in red. MIS: Milne Ice Shelf. MEL: Milne Fiord Epishelf Lake. MGT: Milne Glacier Tongue. MLSI: Multi-year land-fast sea ice. The red triangle is MEL mooring and the yellow is the weather station. B) Bathymetry map. Note: shallow section in the red circle is in fact not shallow, it is $\sim$300 m. Circles refer to CTD cast locations. The bathymetry data comes from lead line sounding and CTD casts. C) Ice draft measurements from ice penetrating radar, airborne radar and remote sensing. D) Modeled ice draft with measurement from C. This figure comes from Hamilton (2016).
The watershed of MG is 1500 km$^2$ and has a glaciated area of approximately 1100 km$^2$ (Hamilton, 2016). It is classified as a possible surge-type glacier (Van Wychen et al., 2016), meaning MG could possibly have fast flowing episodes where it would move faster than its typical flow ($\sim$100 m a$^{-1}$ from 2000 to 2015 (Van Wychen et al., 2016)). From 1999 to 2016, MG lost up to 10% of its surface area (White and Copland, 2018).

### 2.1.2 Milne Ice Shelf

Milne Ice Shelf (MIS) occupies $\sim$205 km$^2$ at the mouth of the fjord (Mortimer et al., 2012) and is attached to the land on both sides. The latest estimated mean ice thickness is 47 m with a minimum and a maximum around 8 m and 94 m, respectively (Hamilton, 2016). The thinnest area is along a basal channel that runs westward from the east shore (Hamilton et al., 2017; Mortimer et al., 2012; Rajewicz, 2017) (Figure 2.2). Ice thickness data is from hand and snowmobile-towed ice-penetrating radar (Hamilton, 2016; Mortimer et al., 2012) Figure D.8, aerial radar (Leuschen et al., 2016), and remote sensing surveys from 2008 to 2014 (Hamilton, 2016; Mortimer et al., 2012). Therefore, changes certainly occurred, although the ice shelf is still relatively intact relative to other ice shelves in Canada. Mortimer et al. (2012) estimated an average thinning of 8.1 m and a 29% area reduction from 1950 to 2009. Before 1950, MIS and MGT were one single ice structure (Jeffries, 1985) and it is the melting of the area between them that created the epishelf lake. (Mortimer et al., 2012)

The lake ice can be distinguished from the ice shelf ice and the glacier ice because it is absolutely flat (no topography) and displays a light gray tone (Mortimer et al., 2012) on aerial photos. Moreover, in Synthetic Aperture Radar imagery (SAR), the epishelf lake area has a higher backscatter due to the freshwater underneath (Veillette et al., 2008); Hamilton (2016) used a threshold of $>$-6dB to distinguish the ice types.

### 2.1.3 Milne Fiord Epishelf Lake

It is estimated that Milne Fiord Epishelf Lake (MEL) surface area was 71.2 km$^2$ in 2015 (Hamilton, 2016). Comparison with more recent satellite images does not
show a drastic change in MEL surface area since then. MEL experiences an annual cycle of deepening and shoaling (Hamilton et al., 2017). During the summer, when snow and ice are melting, water from surface runoff flows into the lake and causes the halocline to deepen. Meanwhile, lake water deeper than the minimum draft of the ice shelf flows out of the fjord to the Arctic Ocean (Figure 1.1). When the melt season is over, surface runoff stops and the lake slowly shoals until the next melt season. It is thought that the flow to the ocean occurs within the ice shelf basal channel and is hydraulically controlled by its dimensions (Hamilton et al., 2017). Hamilton et al. (2017) determined that the annual minimum thickness of MEL has significantly decreased since 2004 to reach a minimum of 7.9 m in 2013. The only data available before 2004 are water bottle measurements from 1983 that suggests the halocline was at 17.7 m depth, if linearly interpolated (Jeffries, 1985). As a result of short summers and cold long winters, the lake is permanently ice-covered (Hamilton, 2016). The minimum ice thickness observed was 0.65 m in July 2010 (Hamilton, 2016) and the maximum was 3.19 m in May 1983 (Jeffries, 1985). The latest ice thickness measurement available is 1.7 m, from July 2019. Keys (1977) proposed that formation of frazil ice by heat transfer at the halocline contributes significantly to the conservation of the ice-cover during summers. As freshwater loses heat at the interface between the cold ocean and the relatively warm lake, frazil is formed and floats upward to acrete to the ice cover. The water in the epishelf lake is fresh (<0.5 g kg\(^{-1}\)) and cold (<4 °C) (Figure 2.3). The maximum lake water temperature is normally recorded in late August when ice cover is minimal and accumulation of heat from solar radiation is maximal. Figure 2.3 shows a water profile taken in May 2011 at the location of MEL mooring (Figure 2.2).

2.1.4 Below the Epishelf Lake

Below the epishelf lake freshwater, a layer of fjord modified water (FMW) of approximately 30 m shows intermediate salinity (5 to 30 g kg\(^{-1}\)) and temperature from 0°C to -1.5°C (Figure 2.3). This layer of relatively fresh water, very close to the freezing point, is linked to submarine melting from the glacier and ice shelf (Hamilton, 2016). Under the FMW, polar water (PW), of salinity 30 to 34.4 g kg\(^{-1}\) and temperature lower than 0°C spans from ~50 m to ~250 m. Below PW is At-
Figure 2.3: CTD water profile taken in May 2011 at the location of MEL mooring (Figure 2.2). A) Conservative temperature ($\Theta$), with the different water masses labeled. MEL: Milne epishelf lake, freshwater, 0 to 4°C. FMW: Fjord modified water, 5 to 30 g kg$^{-1}$, 0 to -1.5°C. PW: Polar water, 30 to 34.4 g kg$^{-1}$, temperature lower than 0°C. AW: Atlantic water, >34.4 g kg$^{-1}$, ~0°C. DW: deep water, as AW but generally slightly fresher and colder. B) Salinity profile. C) Stability profile, smoothed with a 20 point moving average; note the logarithmic transformation. Atlantic water (AW) of salinity >34.4 g kg$^{-1}$ and a temperature ~0°C. Water deeper than the depth of the sill (deep water, DW) is slightly colder and fresher inside the fjord than offshore. Water velocities were measured by Hamilton (2016) with an ADCP from 0 m to 50 m in May 2011 (4 days), from 0 m to 100 m in July 2012 (6 days), and in July 2013 (10 days). The results show that the currents were very weak (<5 cm s$^{-1}$), especially in the top 25 m of the water column (<2 cm s$^{-1}$). Most of the data acquired was below the instrument accuracy (0.5 cm s$^{-1}$), which makes it difficult to infer any circulation patterns. Tidal harmonic analysis (Pawlowicz et al., 2002) was carried out using the pressure record from a bottom anchored instrument from September 2016 to July 2017. The results show that tides are semidiurnal and very limited in amplitude. The main components are M2 (~6.1 cm), K1 (~4.5 cm), S2 (~2.5 cm) and O1 (~2.4 cm). Hamilton (2016) also reported small baroclinicity linked to the tidal cycle.
2.2 Mixing, Energy, Geostrophy

Some important concepts are introduced here in order to better outline studies of ice-covered lakes and polar fjords related to this thesis. This background is used to describe earlier studies as well as the methods, results and implication of this research. For further reading, a more thorough treatment of these concepts can be found in Kundu et al. (2012), Cushman-Roisin and Beckers (2011) or Pedlosky (2013).

2.2.1 Eddy Viscosity

The most common way to take turbulence into account when investigating geophysical flow is to use time-averaged equations and to account for the effect of small fluctuations (turbulence) by the way of the eddy viscosity approximation (Kundu et al., 2012).

\[
\frac{\partial \rho u}{\partial t} + \rho u \frac{\partial u}{\partial x} + \rho v \frac{\partial u}{\partial y} + \rho w \frac{\partial u}{\partial z} - f v = -\frac{1}{\rho_0} \frac{\partial p}{\partial x} - \frac{\partial (u'w')}{\partial x} - \frac{\partial (u'w')}{\partial y} - \frac{\partial (u'w')}{\partial z} + \nu \frac{\partial^2 u}{\partial x^2} + \nu \frac{\partial^2 u}{\partial y^2} + \nu \frac{\partial^2 u}{\partial z^2} \quad (2.1)
\]

This is the averaged equation for momentum in the x direction (\(\rho u\)) (Kundu et al., 2012). The overbars denote the time averaging and the primes denote the fluctuation quantities. \(u, v\) and \(w\) are the velocities in the \(x, y\) and \(z\) directions. \(f\) is the Coriolis frequency, \(\rho_0\) is a reference density, \(p\) is the pressure and \(\nu\) is the kinematic viscosity. Using the eddy viscosity approximation, the fluctuation terms are written with respect to the averaged velocity and a turbulent mixing coefficient \(K_t\) \([\text{m}^2 \text{s}^{-1}]\) is introduced, for example:

\[
\overline{u'v'} = -K_t \frac{\partial \overline{u}}{\partial y} \quad \Rightarrow \quad -\frac{\partial (\overline{u'v'})}{\partial y} = K_t \frac{\partial^2 \overline{u}}{\partial y^2} \quad (2.2)
\]

Because the fluctuation terms are now in the same form as the viscosity terms, these can be combined and a combined (turbulent and molecular) mixing coefficient can therefore be used (\(K_t + \nu = K\)). Moreover, because mixing is normally similar in \(x\) and \(y\) (Kundu et al., 2012), the same mixing coefficient can be used in the horizontal
directions ($K_H$). Using these simplifications, equation 2.1 becomes (Kundu et al., 2012):

$$\frac{\partial \bar{u}}{\partial t} + \bar{u} \frac{\partial \bar{u}}{\partial x} + \bar{v} \frac{\partial \bar{u}}{\partial y} + \bar{w} \frac{\partial \bar{u}}{\partial z} - f \bar{v} = - \frac{1}{\rho_0} \frac{\partial p}{\partial x} + K_H \left[ \frac{\partial^2 \bar{u}}{\partial x^2} + \frac{\partial^2 \bar{u}}{\partial y^2} \right] + K_V \frac{\partial^2 \bar{u}}{\partial z^2}$$

(2.3)

Where the effect of turbulent fluctuations and viscosity is included in $K_H$ and $K_V$ (i.e. $K_H = K_t + \nu$ for the $x$ and $y$ directions and $K_V = K_t + \nu$ in $z$).

### 2.2.2 Stability

A water column is said to be stable if the water potential density increases from the surface to the bottom. In that configuration, no vertical movement will arise unless a force is applied (Kundu et al., 2012). For the same forcing, a water column where the density increases slowly with depth will mix more than a water column where density increases rapidly. If, for some reason, a water parcel has higher density than the water parcel below it, the water profile is said to be unstable. The water parcel will sink until it reaches the location where it is less dense than the water below, and denser than the water above; no external forcing is needed for movement in the water column in that case. The common measure of the water column stability is the Brunt-Väisälä frequency, $N$ (Kundu et al., 2012):

$$N^2 = -\frac{g}{\rho_0} \frac{\partial \rho}{\partial z}$$

(2.4)

Where $g$ is the gravitational acceleration, $\rho$ is the density and $\rho_0$ is a reference density ($\sim$1000 kg m$^{-3}$). $N$ is the (angular) frequency at which a water parcel will oscillate if it is displaced upward or downward by a $\partial z$ distance. Its units are radians per s and are often simplified as $s^{-1}$. A negative $N^2$ value means the profile is unstable.

### 2.2.3 Energy and Mixing

Roughly, kinetic energy in a fluid moves from large scales to smaller and smaller scales until it reaches a scale were it is dissipated by viscosity. The large scale can be seen as the mean flow ($\bar{u}$, $\bar{v}$, $\bar{w}$) and the small scales as the turbulent flow ($u'$, $v'$, $w'$, and $\bar{u}'$, $\bar{v}'$, $\bar{w}'$).
\( v', w' \). It is at the small (turbulent) scale that mixing happens. A simplified steady state budget for turbulent kinetic energy (TKE) of the following form is typical in oceanography (Osborn, 1980):

\[
P = \varepsilon - B
\]  

(2.5)

Where \( P \) is the TKE production by shear, \( \varepsilon \) is the TKE dissipation by viscosity and \( B \) is the buoyancy term. \( B \) is negative when the water profile is stable \((N^2 > 0)\) and positive when the water column is unstable \((N^2 < 0)\). In other words, if the water column is stable, a fraction of \( P \) will go to potential energy and if the water column is unstable, then this extra potential energy will increase the amount of TKE dissipated by viscosity \((\varepsilon)\). A higher \( \varepsilon \) means more turbulence. In order to link turbulence to mixing, Osborn (1980) proposed a semi-empirical relationship between the vertical eddy coefficient \((K_z \text{ in the } z \text{ direction})\), \( \varepsilon \) and \( N \):

\[
K_z \leq 0.2 \frac{\varepsilon}{N^2}
\]  

(2.6)

A newer version of this relation is found in Shih et al. (2005), but it is quite similar to equation 2.6. Equation 2.6 shows that even though there is a lot of kinetic energy in a system, if the water column is highly stratified, mixing will be limited. Inversely, if the water column is close to neutral stability \((N^2 = 0)\), then a small amount of energy injected to the system (potential or kinetic) can lead to a lot of mixing. This is what happens during lake turnover when the water column is no longer stratified and wind is able mix the whole water column.

A common way of measuring \( \varepsilon \) is to use microstructure profilers (Osborn, 1974). These instruments are equipped with small shear probes that detect the small velocity fluctuations \((u', v', w')\). Microstructure instruments can be lowered through the water column or mounted on gliders (Peterson and Fer, 2014). Many other techniques to estimate \( \varepsilon \) also exists, such as finescale parameterization (Polzin et al., 2014), the structure function method (Wiles et al., 2006) or turbulent instrument clusters (McPhee, 2008).

For context, in the Arctic Ocean, a value of \( \varepsilon \) below \(10^{-10} \text{ W kg}^{-1}\) is considered calm and a value above \(10^{-8} \text{ W kg}^{-1}\) is considered energetic (Chanona et al., 2018).
2.2.4 Geostrophic Scale

Equation 2.3 can be scaled by the characteristic dimensions of its terms. Using $U$ as the scale of horizontal velocities $u$ and $v$, $W$ for vertical velocity, $L$ and $D$ as the horizontal and vertical length scales and $T$ for the time scale, 2.3 can be written:

$$\frac{U}{T} + U \frac{U}{L} + U \frac{W}{D} - fU = -\frac{1}{\rho_0} \frac{\Delta P}{L} + K_H \left[ \frac{U}{L^2} + \frac{U}{L^2} \right] + K_V \frac{U}{D^2}$$  \hspace{1cm} (2.7)

Length, velocity and time scales are dependent on each other ($U = L/T$, $W = D/T$); the first four terms then become $4U^2/L$. Dividing all terms by the scale for the Coriolis term ($fU$), equation 2.7 becomes:

$$4 \frac{U}{fL} = 1 - \frac{1}{\rho_0} \frac{\Delta P}{fUL} + 2\frac{K_H}{fL^2} + \frac{K_V}{fD^2}$$  \hspace{1cm} (2.8)

The terms in equation 2.8 do not have any dimensions. The term on the left side is often named the Rossby number ($Ro$) and the two terms at the far right with the mixing coefficients are called Ekman numbers ($Ek$):

$$4Ro = 1 - \frac{1}{\rho_0} \frac{\Delta P}{fUL} + 2Ek_H + Ek_V$$  \hspace{1cm} (2.9)

The Rossby number is commonly seen as the ratio of inertial forces to the Coriolis force. If $Ro \ll 1$ then the influence of the Earth’s rotation has to be taken into account. The Ekman number is the ratio of viscous forces to the Coriolis force. Close to boundaries, the mixing coefficient increases and the scale of the flow decreases; hence $Ek \gg 1$. If the Coriolis force is much more important than the inertial forces (i.e. $Ro \ll 1$) and the effect of boundaries can be neglected (i.e. $Ek_H \ll 1$, $Ek_V \ll 1$), then equation 2.9 collapses to (Kundu et al., 2012):

$$1 = \frac{1}{\rho_0} \frac{\Delta P}{fUL} \Rightarrow f\overline{v} = \frac{1}{\rho_0} \frac{\partial \overline{P}}{\partial x}$$  \hspace{1cm} (2.10)

This balance between the Coriolis force and the horizontal pressure gradient is called the geostrophic balance. The equation for $u$ is:

$$f\overline{u} = -\frac{1}{\rho_0} \frac{\partial \overline{P}}{\partial y}$$  \hspace{1cm} (2.11)
The solution to equations 2.10 and 2.11 in the Northern Hemisphere is a clockwise flow around a high pressure center, or an anti-clockwise flow, around a low pressure center. Horizontal pressure gradients can be set up by external forces that induce current (e.g. wind, inflows) or difference in water properties (salinity, temperature).

Dimensional quantities are also related to the Rossby and Ekman numbers. The Rossby radius of deformation \( R_L = ND/f \pi \) is the length scale at which rotation becomes important for phenomena in a fluid (Cushman-Roisin and Beckers, 2011). For example, using the epishelf lake (82.5°N) average stratification of \( \sim 10^{-2} \) s\(^{-1}\) and a \( \sim 10 \) m vertical scale, the Rossby radius is around 200 m. This means that phenomena in the epishelf lake that have a horizontal scale at or over 200 m will be affected by the Earth’s rotation. In the 390 m water column below the epishelf lake (similar averaged stratification), the Rossby radius is \( \sim 7800 \) m, which is close to the width of the fjord. The Rossby radius of deformation \( R_L \) is the horizontal scale at which the Rossby number \( (Ro) \) is 1. A phenomenon that has a length scale longer than \( R_L \) will be affected by rotation \( (Ro < 1) \).

A water column in geostrophic balance, away from lateral boundaries is similar to a cylinder of spinning fluid. The only thing that can slow the rotation of this fluid is the friction of the fluid column at the top or bottom. The decay time scale of kinetic energy of this fluid cylinder is the Ekman spin-down time \( t_E = D/\sqrt{2K_v f} \) (Pedlosky, 2013). For example, if we consider a geostrophic flow in the epishelf lake \( (D=10 \) m) and estimate the mixing coefficient to be \( \sim 10^{-6} \) m\(^2\) s\(^{-1}\) (Chapter 3), this gives a spin-down time of \( 10^6 \) s, or \( \sim 5 \) days. Hence, if no energy is added in the system, the water velocities will be significantly reduced after 5 days because of the friction on the ice (assuming the seawater below does not exerts a stress on the freshwater layer).

### 2.3 Ice-Covered Lakes

Lakes at latitude over 40° commonly experience ice cover for some time of the year (Kirillin et al., 2012). Therefore, the understanding of mechanisms controlling the physical regime under ice is essential to predict the future of these lakes. If the bottom boundary is neglected, epishelf lakes and ice-covered lakes are similar in
many ways. Several mechanisms can influence the circulation and mixing in these water bodies. The most important are described here, it is however not possible to rank them by importance as it can change greatly from a lake to another.

2.3.1 Seiche

Seiches are standing wave oscillations found in enclosed basins. The oscillation can be at the surface (surface waves; barotropic) or at density discontinuities in the water column (internal waves; baroclinic), or both. Strong winds have been shown to tilt lake ice covers, resulting in significant baroclinic seiches (Bengtsson, 1996; Malm et al., 1998). These internal waves create horizontal circulation that can lead to shear instabilities and hence mixing. The breaking of these waves on shore slopes can also generate significant mixing (Boegman et al., 2005). Kirillin et al. (2018) measured TKE dissipation rates higher than $2 \times 10^{-8}$ W kg$^{-1}$ under the ice during seiching events on Lake Kilpisjärvi in Finland. Simpson et al. (2011) even recorded values up to $10^{-7}$ W kg$^{-1}$ on the bottom boundary layer in Lake Bala in Wales. This illustrates that seiches can significantly contribute to mixing even in ice-covered lakes.

2.3.2 Radiative Mixing

When the ice cover is free of snow, solar radiation will penetrate into the water below. Because the temperature of maximum density ($T_{md}$) is above the freezing temperature in water with salinity below 24 g kg$^{-1}$, heating of water that is below $T_{md}$ increases its density. This radiative heating creates convection in the upper water layer where a heated parcel of water sinks down and is replaced by colder (less dense) water (Bouffard et al., 2019; Farmer, 1975; Forrest et al., 2008). Jonas et al. (2003) and Volkov et al. (2019) reported $\varepsilon$ values of $\sim 8 \times 10^{-10}$ and $\sim 2.5 \times 10^{-9}$ W kg$^{-1}$ during radiative mixing at the daily radiation peak in spring. Radiative mixing can be a very important mechanism for temperate lake as it starts the spring mixing before the wind as direct contact with the water (Bouffard et al., 2016). However, as discussed in the Chapter 1, the present study focuses on the winter period so radiative mixing will be ignored.
2.3.3 Inflows

Inflows from streams and rivers can drive circulation and mix the water column of ice-covered lakes. The extent of mixing depends on the characteristics of the inflow (density, velocity and cross sectional area) and on the stratification of the lake (Stigebrandt, 1978). If the conditions at the inlet are subcritical (laminar flow), the inflow will enter the lake without creating much disturbance and spread out at the depth of neutral buoyancy. If the velocity is high enough, this will produce shear instabilities and there will be mixing at the inflow/lake interface. Mixing and circulation resulting from inflows will be neglected because this is negligible outside of the melt season.

2.3.4 Ice Walls

If one of the boundaries of the lake is a glacier (e.g. Lake Untarsee, Antarctica (Steel et al., 2015)) melting of the ice at the ice-lake interface produces buoyant freshwater that induces upward circulation at the glacier boundary. Epishelf lakes are, by definition, in contact with ice walls. Half of the perimeter of Milne Fiord epishelf lake is ice (Figure 2.2) and there are many independent ice pieces of ice shelf and glacier tongue that increase the surface of the ice-water interface.

2.3.5 Diffusive Convection

Double diffusion is a phenomenon created by the difference in molecular diffusivity of heat ($\sim 10^{-7} \text{ m}^2 \text{ s}^{-1}$) and salt ($\sim 10^{-9} \text{ m}^2 \text{ s}^{-1}$). One type of double diffusion is diffusive convection. Diffusive convection can develop when water with lower temperature is above water with higher salinity and temperature (Ruddick and Gargett, 2003; Schmitt, 1994). In diffusive convection conditions, a relatively cold and fresh water parcel that is displaced downward absorbs heat much faster than it looses salt and therefore becomes less dense than it was initially. This is because this water parcel loses density faster due to the increase temperature than it gains density due to salinity increase. This phenomenon creates a growing oscillation that manifests itself by creating vertical convective cells that result in staircase features in the temperature and salinity profiles. This process can significantly increase heat and salt fluxes in lakes, as the gradients between the convective cells
can be very sharp (e.g. Lake Vanda, Antarctica (Hoare, 1966; Huppert and Turner, 1972)).

2.3.6 Earth Rotation

The rotation of planet Earth can influence physical processes that have a time scale of the order of the inertial period (~12 h) or longer. In geophysical flow, the inertial period is the time scale for which the fluid inertia is perfectly balanced by the Coriolis force. In the Northern Hemisphere, the Earth’s rotation has the effect of ”bending” the flow to the right (Coriolis effect). Away from boundaries, geostrophic flow (balance between Coriolis force and pressure gradient) can develop when the Rossby radius is smaller than the horizontal scale. In lakes, this is often observed as cyclonic or anti-cyclonic gyres (Forrest et al., 2008; Graves, 2015; Rizk et al., 2014).

2.3.7 Tides

Epishelf lakes are hydraulically linked to the ocean, therefore, they experience tides. The back and forth of water in the conduit between the lakes and the ocean (water jet during flood) is thought to be an important mixing mechanism for some epishelf lake in Antarctica (Galton-Fenzi et al., 2012; Gibson and Andersen, 2002) where tidal amplitude is around 1.3 m. The situation of these epishelf lakes is different than MEL; the connection to the open ocean is more restrained and the tides are stronger. However, as Milne Fiord is a quiet environment, tides do not have to be huge to have a significant impact.

2.4 Arctic Fjords

In recent years, there has been a surge of scientific research on fjords in Greenland. These fjords, where tidewater glaciers terminate, are positioned at the nexus of the Greenland Ice Sheet and the Arctic Ocean (Straneo and Cenedese, 2015). As 50% of the mass loss of the Greenland Ice Sheet is linked to ice discharge (melting and calving), Greenland fjords are key to understand and predict the effect of climate change. The fjords that have been studied to date (e.g. Petermann (Johnson et al., 2011), Sermilik (Jackson et al., 2014b), Kangerdlugssaq (Sutherland et al., 2014))
are similar to those of the Canadian High Arctic. It is therefore interesting to look at the work done in Greenland to get a better understanding of the circulation in Milne Fiord. The main circulation mechanisms in Greenlandic fjords are wind, meltwater plumes, tides and Coriolis force. Fjords in Svalbard are also similar as they become ice covered in winter.

2.4.1 Wind

Alongshore wind episodes can generate coastal upwelling or downwelling, increasing the water exchange between the fjord and the coastal shelf substantially. Upwelling can be observed as a lifting of isotherms, generated as deep denser water enters the fjord and pushes shallower water out. Conversely, downwelling can be observed as a deepening of the isotherms as less dense water from the upper water column enters the fjord and pushes deep water out. This circulation driven by density differences arising outside the fjord is called intermediary circulation (Sutherland et al., 2014). In Greenland, strong along-shore wind episodes were linked to rapid water exchange in glacial fjords during both summer (Straneo et al., 2010) and winter (Jackson et al., 2014b). Enhanced water renewal is linked tide-water glacier melt rates. Even though the ocean is ice-covered most of the time offshore of Milne Fiord, wind can still generate up- or down-welling (Dmitrenko et al., 2016; Kirillov et al., 2016; Williams et al., 2006). This is because ice movement on the ocean is dictated by the wind so the wind stress can still be transferred to the water through the sea ice.

Along-fjord wind can also significantly increase the circulation (Carroll et al., 2017) in fjords. However, because all Milne Fiord ice (ice shelf, glacier tongue, epishelf lake) is effectively immobile (does not move with the wind) year round, it does not transfer momentum to the water in the fjord.

2.4.2 Meltwater Plumes

Plumes of relatively fresh water are released at the base of glaciers via subglacial discharge or are formed from the melting of the ice in contact with the fjord water (Figure 1.1). These plumes alter the fjord circulation by moving water upward close to the glacier and entraining ambient water until this combined water mass
reaches neutral buoyancy. This leads to a 2-dimensional circulation pattern where fresher water in the upper water column flows downfjord and more saline water below this flows upfjord (Carroll et al., 2017; Sciascia et al., 2013; Straneo and Cenedese, 2015; Xu et al., 2012). This circulation driven by density difference arising in the fjord is a form of estuarine circulation (Sutherland et al., 2014). Meltwater plumes from subglacial discharge normally have a bigger impact than those from direct melting (Straneo and Cenedese, 2015). Sutherland et al. (2014) estimated that intermediary circulation was an order of magnitude greater than estuarine circulation in summer for two Greenland fjords (Sermilik and Kangerdlugssuaq). On the other hand, Mortensen et al. (2014) showed that estuarine circulation can also play an important role in Godthåbsfjord (Greenland). It is also noteworthy to mention that residual meltwater was observed in Petermann Fjord months after the melt season was over (Washam et al., 2019) and subglacial discharge was discovered in the middle of the winter under a glacier in the Yukon (Schoof et al., 2014). This implies that meltwater plumes can potentially have an impact on the fjord circulation even after the end of the melt season.

2.4.3 Tides

The intensity (currents and amplitude) of the tides and the height of sills determine the amount of energy available for tidal mixing in fjords. If the tides are weak and the sill is not very prominent, tides will likely not have a significant effect. As the intensity of the tides and prominence of the sill increases, baroclinic currents will also increase. If the baroclinic currents are sufficiently strong, shear instabilities can develop and lead to turbulent mixing. In cases where the water depth at the sill is very small, the tide can generate a jet (Stoylen and Fer, 2014). In that case, turbulence level can be quite high; Fer and Widell (2007) measured an average \( \varepsilon \) value of \( 1.1 \times 10^{-7} \) W kg\(^{-1}\) in ice-covered Van Mijenfjorden (Svalbard).

The interaction of the tides with the sill can also generate internal waves of all sorts which can be an important mixing mechanism in Arctic fjords, especially when they are ice-covered and with limited inflow in winter (Perkin and Lewis, 1978). As previously mentioned, tidal amplitude and currents are small in Milne Fiord. In addition, the sill is not substantially prominent (depth \( \sim 220 \) m). This
suggests tidal mixing is not important in this context.

2.4.4 Earth Rotation

In cases where the fjord is wide enough (>Rossby radius), geostrophic flow (the balance between Coriolis force and pressure gradient) can develop (Inall et al., 2014; Johnson et al., 2011). This can significantly increase exchange with shelf water and accelerate subglacial melting and freshwater export (Carroll et al., 2017) by increasing the water renewal in the fjord.

Despite numerous recent studies on circulation in Greenland fjords, no turbulence data has been published yet, which leaves a knowledge gap to be filled in the years to come. However, turbulent kinetic energy dissipation data from Andvord Bay on the Antarctic Peninsula shows that polar fjords can be quiet environments, even in the summer ($\varepsilon < 10^{-9}$ W kg$^{-1}$ below the surface layer) (Lundesgaard et al., 2020).

It is also accepted that the level of turbulence in the Arctic Ocean is quite low (Chanona et al., 2018). There are many explanations for this low turbulent kinetic energy. The first is that the Arctic Ocean is above the turning latitude where linear internal waves generated by semidiurnal tides cannot propagate (Rippeth et al., 2017). The second is that the ice cover reduces the effect of wind and damps internal waves in the upper water column (Guthrie et al., 2013). Taking this into account and keeping in mind that tides are very small around Milne Fiord, it appears that this is a low energy area.

2.5 Basal Channels

Ice shelves in Antarctica are very important to the global climate as they act to buttress terrestrial glaciers. The disintegration of ice shelves therefore causes an acceleration of glacier velocity (Seehaus et al., 2016; Wuite et al., 2015) resulting in an increase of ice flux to the ocean, with its associated consequences (e.g. sea level rise, sea ice increase (Richardson et al., 2005) and possible global circulation alteration (Park and Latif, 2019)). Channels under ice shelves are a key feature to understand in order to predict the stability of these ice structures (e.g. Dow et al. (2018); Drews (2015)). Several studies of these basal channels were carried out in the last 10 years in order to understand the physical processes concerning their
formation, evolution and the consequences of their presence on ice shelves.

2.5.1 Formation

The formation of ice shelf basal channels is linked to the path of the meltwater plumes. Some channels are directly related to subglacial discharge at the grounding line. The discharged water moves upward along the underside of the ice shelf where it melts ice above it preferentially as it rises, creating a basal channel (Alley et al., 2016; Le Brocq et al., 2013). On the other hand, subglacial discharge is not necessarily needed for these structures to appear. They can spontaneously form along the bottom of ice shelves, where any kind of irregularity focusses meltwater flow, melting that area faster and eventually creating a channel (Alley et al., 2016; Sergienko, 2013). Irregularities under ice shelves can be created by the presence of bathymetric features close to the grounding line (Gladish et al., 2012). The melting potential of seawater depends on its temperature and salinity. An increase of temperature or salinity means an increase of melting potential (Jenkins et al., 2010). Therefore, the tidal glaciers in contact with water that has a higher melting potential are more likely to display basal channels. Indeed, Alley et al. (2016) showed that the area where the density of basal channels is the highest in Antarctica is where the water reaching the grounding line has the highest melting potential. In Antarctica, the circumpolar deep water (CDW) is of special interest as it has a high melting potential Alley et al. (2016). The Arctic counterpart of the CDW is Atlantic water (Straneo et al., 2012).

2.5.2 Evolution

Numerical studies (Gladish et al., 2012; Millgate et al., 2013) and observations (Alley et al., 2016; Dutrieux et al., 2013) showed that melting in basal channels is higher close to the grounding line and diminishes in the seaward direction. As water moves toward the ocean in the channels, the freshwater content increases and its melting potential decreases (Gladish et al., 2012; Millgate et al., 2013). This is why channels are normally narrow and deep close to the grounding line but become flatter and wider as the distance from the grounding line increases. One exception to this is the Dotson Ice Shelf, where it appears that melting is sustained from the
grounding line to the calving front, likely because of the presence of warm water relatively high in the water column (Gourmelen et al., 2017). Ice shelves and glacier tongues are normally in hydrostatic equilibrium (Alley et al., 2016; Vaughan et al., 2012). The consequence of this is that the relief underneath the ice is reproduced at the surface and this allows for the use of remote sensing to study the basal channels. In Milne Fiord, Rajewicz (2017) used ice penetrating radar and water measurements (CTD and current meter) to describe the channel and path of the epishelf lake water under MIS. In summer, the depressions at the surface above basal channels can fill with meltwater and form streams and rivers (Dow et al., 2018; Rajewicz, 2017).

2.5.3 Melting

Numerical modeling (Gladish et al., 2012; Millgate et al., 2013) and observations (Dutrieux et al., 2013; Malm et al., 1998) also confirm that melting is much higher inside the basal channels than outside. Melting is also much higher in summer than in the winter. For example, Washam et al. (2019) found basal channel vertical melt rates of 80 m a\(^{-1}\) and 2 m a\(^{-1}\) in summer and winter, respectively, 16 km from the grounding line of Petermann Glacier, Greenland. In Antarctica, summer melting rates of \(\sim 22 \text{ m a}^{-1}\) were observed in channels under the Pine Island (Stanton et al., 2013) and Ross (Marsh et al., 2016) ice shelves. This is higher than the annual average basal melt rate for Ross (\(\sim 5 \text{ m a}^{-1}\) (Marsh et al., 2016)) Filchner-Ronne (\(\sim 5 \text{ m a}^{-1}\), (Le Brocq et al., 2013)) and Getz (8.8-14.7 m a\(^{-1}\),(Alley et al., 2016)) ice shelves. Hamilton (2016) calculated that melt rates were only significant at the very top of MIS channel because the epishelf lake water has a higher melting potential. However, if the melt rates are estimated using the offshore water profile, there is no melting happening in the MIS channel. The reality must be in between, where melt rates at the apex of MIS channel can be significant in the upfjord part of the channel, but decreases as the epishelf lake water loses heat by melting the ice.
2.5.4 Impacts of Basal Channels

Even though the presence of basal channels reduces the overall amount of melting (Gladish et al., 2012), it is thought that these area of reduced ice thickness jeopardize the mechanical stability of ice shelves and glaciers and therefore can increase the ice flux to the oceans (Alley et al., 2016; Rignot and Steffen, 2008). Vaughan et al. (2012) showed that on Pine Island Glacier Ice Shelf, hydrostatic readjustment of the ice in response to channelized melting has formed longitudinal crevasses at the apex of the basal channels (under the ice) and between the channels (at the surface). In addition to parallel crevasses, Dow et al. (2018) showed that high melting in basal channels can also cause transverse fractures, increasing vulnerability to calving.

The main difference between the channel in MIS and others found in Greenland and Antarctica is that the water flowing through MIS channel is much fresher at the top of the channel (as it comes from the epishelf lake). Indeed, the water found in channels outside of Milne Fiord is seawater that is slightly fresher than the ambient water offshore. Other than this peculiarity associated with the epishelf lake, comparison of MIS channel and other basal channels is straightforward, reinforcing the relevance of this study.
Chapter 3

Winter Dynamics in the Last Epishelf Lake in the Canadian Arctic, Milne Fiord, Nunavut, Canada

A version of this chapter is being prepared for submission to a peer-reviewed journal as Winter Dynamics in the Last Epishelf Lake in the Canadian Arctic, Milne Fiord, Nunavut, Canada by J. Bonneau, B. E. Laval, D. Mueller, A. K. Hamilton, A. M. Friedrichs and A. L. Forrest.

3.1 Key Points

- One-dimensional model to analyse eight years of winter mooring data of the upper water column in an perennially ice-covered fjord.

- Mixing is more pronounced in the isolated freshwater layer (epishelf lake) than in the seawater below.

- The morphology of the basal channel under the ice shelf is apparently stable.
3.2 Introduction

Global climate change effects are the most pronounced in the Arctic (IPCC, 2013). In the Canadian High Arctic, all major glaciers of the Queen Elizabeth Islands are showing a decrease in surface elevation (Mortimer et al., 2018). Climate-related changes are unmistakable on the northern coast of Ellesmere Island where the ice shelf extent has decreased over the last century (Mueller et al., 2017). In this region, where historically there was an ice shelf at the mouth of every fjord (Vincent et al., 2001), the continuous fracture and break up of ice shelves has left only one relatively intact, in Milne Fiord. This platform of ice is grounded on both sides of the fjord and acts like a dam, preventing summer freshwater runoff from dispersing freely into the ocean. This perennial layer of freshwater overlaying to Arctic seawater behind the ice shelf is called an epishelf lake. Since Milne Ice Shelf (MIS) is the only ice shelf in Canada that has not recently calved, Milne Fiord epishelf lake (MEL) is the last epishelf lake along its coast. Figure 3.1 is a satellite image of Milne Fiord showing the downstream part of Milne Glacier (MG), the glacier tongue (MGT), the epishelf lake (MEL) and the ice shelf (MIS). MG, MGT, MEL and MIS form a strongly interconnected system not yet fully understood (Hamilton et al., 2017). Studying one of the pieces of this puzzle also improves knowledge of the other parts. Moreover, even though MEL is the last epishelf lake in Canada, several can be found in Antarctica (Gibson and Andersen, 2002; Laybourn-Parry and Wadham, 2014) and one in Greenland (Bennike and Weidick, 2001). Hence, knowledge acquired in the study of Milne Fiord can be transferred to the other sites in the Polar Regions.

It is known that MEL experiences annual cycles of deepening due to meltwater runoff in summer and shoaling due to outflow below MIS (Hamilton et al., 2017) (Figure 3.2). It is thought that epishelf lake outflow is through a basal channel under the ice shelf, analogous to those found beneath Petermann Glacier in Greenland (Rignot and Steffen, 2008) or Pine Island Glacier Ice Shelf (Dutrieux et al., 2013; Stanton et al., 2013) in Antarctica. Basal channels have attracted attention in recent years as meltwater concentrates in these structures (Millgate et al., 2013) (Gladish et al., 2012) and increases melt rates (Alley et al., 2016; Dutrieux et al., 2013; Stanton et al., 2013). Even though they reduce the overall melting of ice shelves
Figure 3.1: A) Location of Milne Fiord (82.6°N, 81.0°W). B) Landsat 8 image of Milne Fiord taken in September 2018. MEL: Milne Epishelf Lake, MIS: Milne Ice Shelf, MGT: Milne Glacier Tongue, MG: Milne Glacier
Figure 3.2: Schematic of Milne Fiord including Milne Ice Shelf (MIS), Milne Epishef Lake (MEL) and Milne Glacier Tongue (MGT). A) During the melt season (summer) (∼June 1st to ∼September 1st), runoff drives the deepening of MEL. B) During the remainder of the year, the thickness of MEL decreases slowly toward the minimum draft of the MIS. (Gladish et al., 2012; Millgate et al., 2013), it is thought that the localized strong melting within channels will lead to faster breakup (Dow et al., 2018; Gourmelen et al., 2017; Rignot and Steffen, 2008).

From an oceanographic perspective, the physical structure of the water column in Milne Fiord is well known during summer (Hamilton et al., 2017), but less during winter. Because the fjord is ice-covered all year long, no wind-induced circulation or mixing occurs. Moreover, the tidal amplitudes are very small (∼10 cm). Wind and tides are the two main forcing mechanisms in an Arctic fjord (Keys, 1977). Nonetheless, other mechanisms of lower importance can modify the water column of ice-covered fjords. Sciascia et al. (2013) found that even during the winter, meltwater plumes influenced the circulation in Sermilik Fjord, Greenland. Also in Sermilik, Jackson et al. (2014b) and Straneo et al. (2010) showed that water properties changed in response to along-shore wind episodes creating downwelling. In Cambridge Bay, Canada, Perkin and Lewis (1978) concluded that during winter, breaking of internal waves on the shore was the main mixing mechanism. The Earth’s rotation can also alter the circulation when the fjord is sufficiently wide (Straneo and Cenedese, 2015). Due to the presence of the ice shelf, the upper water column in Milne Fiord has similarities with ice-covered lakes where the influence of the Earth’s rotation has also been reported (Bengtsson, 1996; Forrest...
et al., 2013; Huttula et al., 2010; Rizk et al., 2014; Steel et al., 2015). Regarding mixing mechanisms, it is noteworthy to mention that solar radiation does not play important role outside of the melt season as the ice is quickly covered by snow. Circulation and mixing processes have implications for the location of the halocline; important to understand the outflow in the channel and the amount of surface runoff. Furthermore, a better understanding of mixing processes in the fjord would improve insight on the fate of the ice shelf and the glacier. Finally, in addition to other epishelf lakes in Antarctica, physical processes in MEL can be similar to those in ice-covered lakes, where external forcing is limited, especially when snow cover significantly reduces solar radiation reaching the water. With that perspective, MEL is a great laboratory to study these lakes since it is ice-covered all year long.

Extensive (relative to Arctic research) field work has been carried out in Milne Fiord over the last 15 years and a mooring deployed in the lake in May 2011 has been recording continuously to present. This study has two main objectives. The first is to quantify the mixing occurring in the epishelf lake and associate it to forcing mechanism(s). The second is to confirm that the channel is the main outflow path for the epishelf lake and to get information on the evolution of its morphology. In order to do this, a one-dimensional model was calibrated with the mooring data (July 2011 to July 2019) in order to estimate the mixing in the top of the water column and the discharge through the channel from the end of one melting season to the start of the next one.

The next section (2) describes the important geophysical features of Milne Fiord. It is followed by the description of the field data and the numerical model (section 3). Section 4 goes over model validation and results. Section 5 is the discussion of the main results and their implications. Finally, section 6 is the conclusion.

### 3.3 Geophysical Setting and Study Area Background

#### 3.3.1 The Fjord

Milne Fiord (82.6°N, 81.0°W) (Figure 3.1) is 40 km long from the glacier grounding line to the outer edge of the ice shelf. Its width is 6 km from the glacier ground-
ing line to the epishelf lake and then becomes wider downfjord. The bathymetry of the fjord, inferred from CTD casts and depth soundings (Hamilton et al., 2017) exhibits a U-shape profile with a maximal depth of 436 m. A sill with a maximum depth of 220 m is present 22 km upfjord from the tip of the ice shelf, just downfjord of the epishelf lake. The maximum fjord depth below the epishelf lake is approximately 400 m and the depth near the grounding line of Milne Glacier is 150 m Figure 2.2. Tides in Milne Fiord are semidiurnal with an amplitude around 10 cm. Tidal baroclinicity was found from current measurements but was too low to produce shear mixing (Hamilton et al., 2017). The average annual air temperature at sea level is -19°C, with the number of positive degree days between 100 and 300 from June 1st to September 1st (Hamilton, 2016). At this high latitude, there is no direct solar radiation from mid-October to the beginning of March.

3.3.2 The Glacier

MG is over 50 km long, 4 to 5 km wide, has a thickness of approximately 150 m at the terminus (Hamilton, 2016) and becomes thicker going upglacier. The watershed of this tidewater glacier is 1500 km² and has a glaciated area of approximately 1100 km² (Hamilton, 2016). MG is classified as a possible surge glacier (Van Wychen et al., 2016), meaning MG could possibly have fast flowing episodes where it would move at velocities many times the normal 100 m/year observed lately (Van Wychen et al., 2016). Downfjord from the grounding line, MGT extends 15 km. The thickness of the MGT decreases rapidly and the ice thickness is less than 10 m at its margins. It broke away from the glacier in 2009 but did not move significantly since.

3.3.3 The Ice Shelf

MIS occupies 200 km² at the mouth of Milne Fiord and is attached to land on both sides (Figure D.4). The estimated mean ice thickness is 47 m with a maximum and a minimum around 94 m and 8 m, respectively (Hamilton, 2016). The thinnest area is along a basal channel that runs westward from the east shore (Hamilton et al., 2017; Mortimer et al., 2012; Rajewicz, 2017) (Figure 3.1).
3.3.4 The Epishelf Lake

From a physical perspective, the most striking feature in MEL is the extremely sharp salinity interface between the freshwater and the ocean below (Figure 3.3). The salinity gradient, over 6 g kg\(^{-1}\) m\(^{-1}\), is well above what is typically found year-round in an Arctic fjord. The depth of the halocline, taken as the depth of maximum stratification, is used to mark the depth of the epishelf lake (Hamilton et al., 2017). MEL experiences an annual cycle of deepening and shoaling. During the summer, when snow and ice are melting, water from surface runoff flows into the lake deepening the freshwater layer (Figure 3.2A). Meanwhile, water deeper than the minimum draft of the ice shelf flows to the ocean. When the melting season is over, surface runoff stops and the lake slowly shoals until summer (Figure 3.2B). It is thought that the flow to the ocean is exclusively along the ice shelf basal channel and is hydraulically controlled by its dimensions; this study tests this hypothesis by modeling the outflow during the winter period. Hamilton et al. (2017) showed that the annual minimum thickness of the epishelf lake significantly decreased from 2004 to reach a minimum of 7.9 m in 2013. The only data available before 2004 are water bottle measurements from 1983 that, if linearly interpolated, suggests the halocline was at 17.7 m (Jeffries, 1985). Using satellite imagery and aerial photos, it is estimated that the MEL surface area went from 13.5 km\(^2\) in 1959 to 71.2 km\(^2\) in 2015 (Hamilton, 2016). Comparison with more recent satellite images does not show a drastic change in MEL surface area since 2015 so 71.2 km\(^2\) will later be used for hydrological analysis. As a result of short summers, cold long winters and a freshwater cap, the lake is permanently ice-covered. The minimum ice thickness observed was 0.65 m in July 2010 (Hamilton, 2016) and a maximum of 3.19 m was observed in May 1983 (Jeffries, 1985).

3.3.5 Offshore Oceanography

Oceanographic measurements offshore from Ellesmere Island are sparse. Jackson et al. (2014a) used mooring, CTD data and numerical modeling to state that currents are weak (less than 10 cm s\(^{-1}\)) and directed westwards on the North Ellesmere continental shelf. This is in agreement with other numerical experiments (Aksenov et al., 2011) and iceberg drift paths (Copland et al., 2007; Garbo). CTD profiles
offshore of Milne Fiord every year since 2011 show a mixed layer of 30 to 60 m on top of water gradually transitioning from -1.6°C and 31 g kg⁻¹ at the base of the mixed layer to 0.3°C and 35 g kg⁻¹ at 250 m. Water properties lower than 250 m do not change greatly.

### 3.4 Data and Methods

#### 3.4.1 Epishelf Lake Mooring

An ice tethered mooring was deployed in the center of MEL in May 2011 and has been recording since then (Figure D.6). To the authors knowledge, this is the longest deployment of a mooring in a Canadian High Arctic fjord or lake. The data from the original deployment to July 2019 are analyzed here. The only time gaps are during fieldwork when the mooring was serviced. Over the years, the mooring configuration changed substantially and different types of instrument were used. The mooring line configuration for each deployment is included in Figure 3.6. Since the focus here is on annual and interannual variations, data from all the mooring instruments were averaged daily for the following analysis. When pressure data was available (6 out of 8 years), instrument depth was corrected for shifts in elevation due to ice formation. All temperature and salinity data in this study were converted to the TEOS-10 standard as conservative temperature, Θ [°C] and absolute salinity, S_A [g kg⁻¹] using the GSW oceanographic toolbox (McDougall and Barker, 2011).

#### 3.4.2 CTD Casts

Water profiles have been taken annually in MEL since 2009. Instruments used for profiling are an Idronaut Ocean304plus (2015-2016) and a RBR XR-620 (2009-2014, 2017-2019). The 2010 CTD profile is from NEIGE (2017). The profiles were taken in natural holes made with a power auger or hand tools. Profiles were taken at recovery and deployment of the mooring to crosscheck the mooring instruments and get a full vertical resolution.
3.4.3 Weather Data

A weather station installed next to a small bay in Purple Valley (Figure 3.1) has been recording hourly data continuously since 2009 (data courtesy: Luke Copland). A weather station was also installed on the ice shelf from July 2016 to July 2018. Even though the Purple Valley station is much more sheltered than the ice shelf station, decomposed N-E-S-W wind at ~2 m show similar trends. Wind data can be compared to the model to investigate forcing mechanisms.

3.4.4 Ice Shelf Channel ADCP

An ADCP moored at the top of the ice shelf basal channel (green circle in Figure 3.1) recorded from July 2017 to July 2019. Using the July 2016 channel morphology data from Rajewicz (2017) and the approximated 71.2 km$^2$ lake area, the data from the ADCP and the model can be compared in order to evaluate the model outflow over the 2016-2017 and 2017-2018 deployments.

3.4.5 Model Formulation

A one-dimensional model was used to analyze the mooring data outside of the melting season (i.e. winter). It is emphasized here that the model was used as a diagnostic tool to examine mixing and the outflow, not a prognostic one. It was designed to relate all the available winter data from the mooring together and analyze these data in a simplified context with a small number of free parameters. The model works to estimate the vertical mixing in the upper water column and the outflow through the basal channel. In order to do this, the mooring data was employed to determine the parameters of the model using an iterative method.

To model the transport of heat and salt, the Reynolds-averaged transport equation for scalar properties was used (Kundu et al., 2012):

$$ \frac{\partial \bar{\psi}}{\partial t} + \bar{u}_j \frac{\partial \bar{\psi}}{\partial x_j} + \frac{\partial (\bar{u}' \bar{\psi}')}{\partial x_j} = K_m \frac{\partial^2 \bar{\psi}}{\partial x_j^2} $$  (3.1)

Where $\bar{\psi}$ is a Reynolds-averaged scalar (e.g. conservative temperature or absolute salinity), $\bar{u}_j$ is the Reynolds-averaged velocity vector and $K_m$ is the molecular diffusivity. $t$ is the time and $x$ is the dimension ($[x_1 x_2 x_3] = [x y z]$, $x_3$ ($z$) being in
Figure 3.3: Schematic of the one-dimensional model. Typical absolute salinity and stability profiles are on the right. The data from the uppermost and the 25 m thermistor are used as boundary conditions for temperature and the no flux boundary conditions are used for salinity. Mixing coefficients of the top freshwater layer ($K^{top}$) and the bottom seawater layer ($K^{bot}$) are parameters of the model; only molecular mixing is considered for the halocline layer. The outflow layer is between the minimum draft of the ice shelf ($h_0$) and the bottom of the halocline layer ($z$). The top and bottom dashed lines show the top and bottom boundaries of the model. The two middle dashed lines are the top and bottom of the halocline layer (molecular diffusivity only).

The following simplifying assumptions were made:

- The average vertical velocity $\overline{u_3}$ is nil ($\overline{u_3} = 0$)
- The horizontal gradients are negligible ($\overline{u_1} \frac{\partial \psi}{\partial x_1} \approx \overline{u_2} \frac{\partial \psi}{\partial x_2} \approx \frac{\partial^2 \psi}{\partial x_1^2} \approx \frac{\partial^2 \psi}{\partial x_2^2} \approx 0$)
- Eddy diffusivity can be used to estimate turbulence ($\overline{u_3' \psi'} \approx -K_{t} \frac{\partial \psi}{\partial x_3}$)
This leads to:

\[
\frac{\partial \phi}{\partial t} = K \frac{\partial^2 \phi}{\partial x^2}
\]  

(3.2)

where \(K\) is the combined \((K_t + K_m)\) mixing coefficient.

In order to take into account the outflow of the epishelf lake (Figure 3.3), the basal channel was simplified as a rectangular weir (Figure 3.4), which allowed the outflow to be described with two annual parameters \((C_r, h_0)\), see below. The model works to estimate four different parameters: two mixing coefficients \((K^{\text{top}}, K^{\text{bot}})\), as in section 3.2, see below) for every day and two annual outflow parameters \((C_r, h_0)\), see below). To find the value of these parameters, these quantities were constrained to a number of possible values and then the best fit was found using the daily averaged mooring data as the evaluation data set.

Equation 3.2 was solved on a 10 cm by 30 minute grid using a Crank-Nicolson finite difference scheme. Grid space and time independence was verified using higher and lower order of magnitude meshes.

**Boundary conditions**

For temperature, the daily averaged data from the uppermost unfrozen thermistor and the thermistor at 25 m depth were used as Dirichlet boundary conditions at the top and bottom nodes. For salinity, a no flux \((\frac{\partial S_A}{\partial z} = 0)\) Neumann boundary condition was used. Since mixing is very limited (demonstrated below), a null salt flux at 25 m is a reasonable assumption even though gradients exist at that depth.

**Initial conditions**

For temperature, initial conditions were given by a linear interpolation between the mooring instruments. Initial salinity conditions were obtained by using the last CTD cast taken during fieldwork and fitting this profile to the mooring salinity data by shifting it vertically to get the best fit (minimum RMSE).

**Mixing**

In order to account for mixing, the water column was divided into three layers (freshwater, halocline and seawater, Figure 3.3), each with different mixing coeffi-
The layer boundaries were defined as the points where the squared Brunt-Väisälä frequency equals $10^{-2}$ s$^{-2}$. This demarcation is supported by the minimum Richardson number ($Ri = N^2/(\partial u/\partial z)^2$) found in the water profile. An observed velocity gradient around 0.01 s$^{-1}$ from Hamilton (2016) and induction current meter measurements (not shown) in combination with perpetual high stratification ($>10^{-2}$ s$^{-2}$) made it possible to rule-out turbulent mixing in the halocline. However, within the top freshwater and bottom seawater layers, CTD measurements indicate stratification is not strong enough to preclude turbulent mixing. The threshold of $10^{-2}$ s$^{-2}$ was arbitrarily chosen as it delineates the region of high salinity gradient with good precision in all CTD profiles (Figure 3.3). To summarize:

- The top layer, from the top boundary to $N^2 = 10^{-2}$ s$^{-2}$ had mixing coefficients for heat ($K_{\Theta}^{\text{top}}$) and salt ($K_{SA}^{\text{top}}$).
- The halocline layer, where $N^2 > 10^{-2}$ s$^{-2}$, only had molecular diffusion.
- The bottom layer, from $N^2 = 10^{-2}$ s$^{-2}$ to the bottom boundary, had mixing coefficients for heat ($K_{\Theta}^{\text{bot}}$) and salt ($K_{SA}^{\text{bot}}$).

Values of $1.4 \times 10^{-7}$ m$^2$ s$^{-1}$ and $1.4 \times 10^{-9}$ m$^2$ s$^{-1}$ were employed for molecular diffusivities of heat and salt, respectively (Jackson and Rehmann, 2014). Possible $K_{\Theta}$ coefficients were: $[1.4, 2, 4, 8, 16, 32, 64, 128, 256, 512, 1024] \times 10^{-7}$ m$^2$ s$^{-1}$. The power of two increment was chosen to reasonably cover orders of magnitude from $10^{-7}$ to $10^{-4}$ for $K_{\Theta}$ while keeping the required computational power reasonably low (normal laptop). The upper limit of $1024 \times 10^{-7}$ m$^2$ s$^{-1}$ was chosen as $K_{\Theta}$ only very rarely reaches this value (as will be shown later).

As first pointed out by Turner (1968), when turbulence is weak, the mixing of heat and salt is not the same (i.e. the turbulent Lewis number ($Le = \frac{K_{\Theta}}{K_{SA}}$) is not one). Differential diffusion ($Le \neq 1$) has been observed in oceanic measurements and demonstrated in laboratory and numerical experiments (Gargett, 2003). Because Milne Fiord is a quiet environment, it is important to take into account differential diffusion in order to model the water column properly. To account for this phenomenon, the parameterization of Jackson and Rehmann (2014) was applied to link $K_{\Theta}^{\text{top}}$ to $K_{SA}^{\text{top}}$ and $K_{\Theta}^{\text{bot}}$ to $K_{SA}^{\text{bot}}$. This related the mixing parameters in the top and bottom layers, thereby reducing the number of unknowns. Using the
ratio determined by Jackson and Rehmann (2014), possible $K_{SA}$ coefficients were: 
$[0.0014, 0.002, 0.004, 1.5, 4.3, 12, 33, 81, 191, 425, 915] \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$.

**Outflow**

Assuming most of the water outflow is through the basal channel in the ice shelf and that it can be simplified as a rectangular channel (Hamilton et al., 2017), the outflow was modelled based on an inverse rectangular weir equation (Kindsvater and Carter, 1959) (Figure 3.4):

$$\frac{d(z-h_0)}{dt} = \frac{2}{3A_{lake}} \sqrt{2g'Ce'b(z-h_0)^{2/3}}$$

(3.3)

Where $h_0$ is the depth of the minimum draft of the ice shelf, $C_e$ is a friction coefficient, $b$ is the width of the rectangular channel, $z$ is the depth at the bottom of the outflow layer and $A_{lake}$ is the area of the lake ($\sim 71.2 \times 10^3$ m$^2$). $g'$ is the reduced gravity ($\frac{g\Delta \rho}{\rho} \approx 0.25$ m s$^{-2}$), where $g$ is the gravitational acceleration, $\Delta \rho$ is the density difference between the freshwater and the seawater (25 kg m$^{-3}$) and $\rho$ is a reference density (1000 kg m$^{-3}$). $Ce'b$ and $h_0$ are two unknown parameters that were assumed constant outside of the melt season. $\frac{d(z-h_0)}{dt}$ was computed every day and the amount of water flowing out to the ocean was modeled by shrinking the outflow layer (i.e. $z-h_0$ was reduced). This assumes that the outflow velocity is the same everywhere in the outflow layer. The above simplifications were derived from current measurements and ice penetrating radar measurements over the channel (Rajewicz, 2017). In order to compensate for the outflow, water was added at the bottom of the model using the properties of the bottom node. Using equation 3.3, possible values for $h_0$ ranged from 3 to 10 m. The possible values for $Ce'b$ ranged from 2 to 10 m, which gives a minimum channel width from 2.5 to 18 m using common rectangular weir coefficients (0.55 to 0.8) (Hamilton et al., 2017).

**3.4.6 Model Fitting**

In order to find the optimal model parameters, an iterative scheme was employed using a custom coefficient of agreement ($C_a$). All mooring instruments in the top 25 m were used for the calibration. For the temperature calibration, the model
Figure 3.4: Schematic of the outflow of the lake through the basal channel of the ice shelf. A modified weir equation using a two layer simplification (equation 3.3) is used to constrain the number of parameters related to the outflow. A) Top view. B) Along fjord section. C) Across fjord section through MIS. Note: not to scale

output was compared to the linearly interpolated data at every grid node and the root mean squared error ($\text{RMSE}_{\theta,\text{model}}$) was computed and then normalized by the standard deviation of daily averaged and linearly interpolated mooring data ($\text{STD}_{\theta,\text{mooring}}$). For the salinity data, the model output was linearly interpolated to the precise depth of each salinity instrument, then the root mean squared error ($\text{RMSE}_{SA,\text{model}}$) was computed and normalized by the standard deviation of the salinity data ($\text{STD}_{SA,\text{mooring}}$). The temperature score ($\frac{\text{RMSE}_{\theta,\text{model}}}{\text{STD}_{\theta,\text{mooring}}}$) and the salinity score ($\frac{\text{RMSE}_{SA,\text{model}}}{\text{STD}_{SA,\text{mooring}}}$) were weighted to take into account the number of temperature measurements ($n_T$) and conductivity measurements ($n_C$). For example, if there were 10 temperature data points and four conductivity data points on the mooring line, the temperature score was weighted by $10/(10+4)$ and the salinity score was weighted by $4/(10+4)$. To summarize, the $C_o$ was computed with the following
The normalization by the standard deviation of the mooring data allowed combination of temperature and salinity data by placing them on a similar scale. The weights allowed combination of these evaluation scores by adjusting their importance according the respective number of measurements. The $-1$ exponent was used to yield a positive relationship between $C_a$ and the model skill (i.e. a high $C_a$ means a good agreement). A value around 1 would mean that the model did poorly, since the $RMSE$ would have the same magnitude as the $STD$ (basically random). On the other hand, a value of 10 would mean the model performed well since the $RMSE$ would be 10 times smaller than the $STD$.

Figure 3.5 is a schematic of the model iterative calibration workflow. The first step was to write the boundary and initial conditions into the model mesh (1). Then, a pair of outflow coefficients ($C_e b$ and $h_0$) was selected in order to calculate the daily outflow throughout the deployment (2). Next, the mixing coefficients returning the highest $C_a$ were found for day 1 (3-6). Steps 4-6 were repeated for every consecutive day. Once the daily mixing coefficients were determined, the $C_a$ for the whole deployment was computed (7) and steps 2-7 were repeated, narrowing down on the optimal $C_e / h_0$ pair. Finally, the pair of outflow coefficient returning the highest $C_a$ for the whole deployment was selected as the best fitting parameters. This procedure was repeated for the eight winters of mooring data (2011-2019). The model version with the best fitting parameters was used to generate the output in the Results section.

3.5 Model Results and Validation

3.5.1 Temperature

Temperature output from the model (Figure 3.6A) agrees well with the temperature data from the mooring (Figure 3.6B). The average RMSE for the whole time-series is 0.19°C, the standard deviation is 0.17°C, the $R^2$ is 0.97 and the bias is...
0.067°C. The $C_a$ is 5.6 for the 8 years. Figure 3.6C shows the difference between the model and the linearly interpolated mooring data. The model output appears as a smoothed version of the mooring data, which signifies that the main seasonal physical characteristics are reproduced. Three main differences can be observed between the model and the mooring data. First, the temperature oscillations above the halocline present in the mooring data are not reproduced in the model. Since there is no significant addition of heat during the winter, these oscillations must
be the result of horizontal advection, which are not considered in the model. The second major disparity is found near the halocline where slight deviations in the modeled halocline depth result in substantial temperature errors because of the sharp gradient at this location. The blue line at 7.5 m in fall 2014 (Figure 3.6C) is the best example of this. This type of error is due to the simplified consideration of the outflow in the model, which is held constant for the whole winter. The last major difference is the inflow of cold water between 15 m and 22 m in the mooring data of 2015-2016 and 2016-2017 (Figure 3.6B). This increased the error for the bottom part of the water column for many months. These main differences between the model and the mooring data are the result of the simplifications made in the model (one-dimension, no advection, rectangular weir outflow). However, considering the visual and statistical agreement, the model is considered appropriate for the study of the full winter timeseries.
Figure 3.6: Temperature timeseries of the top of the water column in Milne Fiord. A) Conservative temperature timeseries of the model results. B) Conservative temperature timeseries from the mooring (linearly interpolated). C) Difference between the model and the mooring data, positive values mean the model temperature is higher than the mooring temperature. Triangles show the location of the thermistors on the mooring line and the squares show the location of the conductivity instruments.
Figure 3.7: Salinity data from the model (solid line) at the depth of the conductivity instruments on the mooring (dashed line). Labels are the depth of the instruments, in meters.

3.5.2 Salinity

The agreement between the model and the mooring salinity data is inferior to that for temperature but the model still fits the main characteristics of the salinity profile in the fjord (Figure 3.7). The average RMSE is 2.0 g kg\(^{-1}\), the standard deviation is 1.7 g kg\(^{-1}\) and the bias is 1.1 g kg\(^{-1}\). The main discrepancies are found in the top of the water column where the model tends to diffuse more salt into the freshwater than the mooring instruments indicate. This is likely because the simulated halocline is compressed in the outflow process, increasing the salinity gradient to values higher than observed in the field. Since the salt flux is proportional to the gradient, more salt makes its way to the top layer as a result of this artifact. Nonetheless, the model shows the main features of the epishelf lake which are a freshwater layer atop a sharp halocline that moves upwards during the winter.

3.5.3 Outflow

The depth of the halocline (bottom of the epishelf lake) returned by the model is shown is Figure 3.8A, as well as the outflow parameters for each year (legend). The trajectory of the halocline from 2014 to 2019 are closely grouped but the first
two years of the record (2011-2012 and 2012-2013) exhibit a more pronounced shoaling. The minimum draft of the ice shelf \((h_0)\) returned by the model is similar for every year (between 6.7 and 7.9 m) except for the 2015-2016 year (5.3 m). The outflow friction-width coefficient \((C_e)\) varies more but stays within the range of realistic values. There is no perceptible trends in either \(C_e\) or \(h_0\).

Figure 3.8B shows the approximated flow rate through the channel according to the model (dashed lines) and the moored ADCP (solid lines). The total discharge of \(2.0 \times 10^8\) (2017-2018) and \(2.3 \times 10^8\) m\(^3\) (2018-2019) for the model and \(1.4 \times 10^8\) (2017-2018) and \(0.9 \times 10^8\) m\(^3\) (2018-2019) for the ADCP are similar.

### 3.5.4 Mixing

The daily mixing coefficients returned by the model are shown by the dots in Figure 3.9. The solid lines represent a 30 day moving average of apparent thermal diffusivity. The top layer of freshwater above the halocline experiences more mixing than the seawater below except during the first months (September and October) after the melt season. The total amount of mixing in the top of the water column differs greatly from year to year, with a minimum winter average of \(2.5 \times 10^{-7}\) m\(^2\) s\(^{-1}\) in 2015-2016 and maximum \(1.2 \times 10^{-5}\) m\(^2\) s\(^{-1}\) in 2018-2019. The mixing below the halocline is more uniform, spanning from \(1.8 \times 10^{-7}\) m\(^2\) s\(^{-1}\) in 2015-2016 to \(2.3 \times 10^{-6}\) m\(^2\) s\(^{-1}\) in 2013-2014.

If the model would be over mixing, the heat content of the freshwater layer would be lower in the model than in the mooring data, as a higher \(K_{\Theta}^{\text{top}}\) means a higher heat flux out of the top layer. Analysis of the heat content in the top freshwater layer shows that the model does not over- or under-mix, increasing the confidence in the results.
Figure 3.8: A) Depth of MEL (maximum $N^2$ value) given by the model for each winter. The staircase effect is due to the vertical discretization of the model (10 cm). Outflow parameters $h_0$ and $C_e b$ for each winter are in the legend. B) Outflow through the basal channel in MIS. Dotted lines are the model estimation using the rectangular weir equation and an estimated lake area of 71.2 km$^2$. Solid lines are the ADCP estimation using the channel morphology data from Rajewicz (2017).
Figure 3.9: A) Daily mixing coefficients from the model, for the heat and salt transport equations for the top and bottom layers (dots). Solid lines are the 30 day averaged quantities for the top (blue) and bottom (red/orange) coefficients of the heat equation. Minimum possible values of daily mixing coefficients are the molecular diffusivities; $1.4 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$ for heat and $1.4 \times 10^{-9} \text{ m}^2 \text{ s}^{-1}$ for salt. B) 30 day averaged mixing coefficient of the model (dark blue) and inverse of the Richardson number (light blue) computed according to equation 3.6. $N^2$ (dotted line) computed by the model and 30 day maximum density difference attributed to the temperature oscillations $\Delta \rho_{30}$ (dashed line) are used in equation 3.6. Enhanced mixing (dark blue line) is linked to stronger eddy activity (dashed line) hence to a lower Richardson number (light blue line).
3.6 Discussion

3.6.1 Numerical Model

The one dimensional model employed in this study was shown to compared fairly well with the winter mooring data in Milne Fiord. It enables the description of the main physical mechanisms using only four different parameters. Consequently, these parameters can be examined in order to get a better understanding of Milne Fiord system.

3.6.2 Outflow

Considering the simplifications (rectangular weir equation with $h_0$ and $C_e b$ constant throughout the winter) and estimations (channel cross section, lake area), it is reasonable to state that the basal channel is the main outflow path for the epishelf lake water. Moreover, since ice penetrating radar measurements did not detect another probable path for outflow, the basal channel taken into account in this study is most likely the only outflow for the epishelf lake water.

There are yet some differences in trends between the model output and the ADCP data. For example, the flow rate increase at the end of December 2018. Possible explanations are 1) an abrupt (7 day) change in the channel morphology at the ADCP location and 2) a mixing event disrupting the halocline. No mixing was inferred by the model in November 2017 which indicates the first process is more likely. On the other hand, $h_0$ and $C_e b$ are very similar from July 2016 to 2019 which suggests channel thinning may have occurred in the area of the ADCP mooring but not in the area of the principal channel constriction (where weir equation parameters come from). In other words, if the cross sectional area decreases (increases) at the ADCP location, the outflow rate inferred will be higher (lower) than reality because the velocity recorded by the ADCP would be greater (smaller). Moreover, the model $C_a$ is no worse for 2017-2019 than for the other years, suggesting the outflow increase, as estimated by the ADCP, is not directly linked to the epishelf lake and that the constant outflow parameter assumption holds.

Channels under ice shelves and glacier tongues have attracted attention in recent years (e.g. Alley et al. (2016); Dow et al. (2018); Le Brocq et al. (2013)). Buoy-
ant water from subglacial melt and discharge converge in these longitudinal ice depressions, concentrating the melting in the channel apex (Millgate et al., 2013; Rignot and Steffen, 2008). Many recent studies estimated basal channel melt rates (Alley et al., 2016; Dutrieux et al., 2013; Gourmelen et al., 2017; Stanton et al., 2013), all agreeing that these locations experience enhanced melt (negative ice mass balance). For the present study, the outflow parameters of the model lead to the conclusion that the minimum draft of the ice shelf has not experienced a major change since September 2011. Indeed, if the melting was continuous in the channel, there would be a trend in the outflow coefficients, but this is not the case. The fact that $h_0$ went from 7.0 m (2014-2015) to 5.2 m (2015-2016) then up to 7.9 m (2016-2017) implies that the apex of the channel at the main constriction is in neutral mass balance over a multi-year time scale. Likewise, the fact that $C_b$ does not show any trend and experiences only small variations ($\pm 50\%$) also points towards a multi-year equilibrium in ice mass along the channel. This agrees with the insignificant melting rate obtained by Hamilton (2016) in the Milne channel area and the evidence past ice accretion under Ward Hunt Ice Shelf (100 km to the east) (Jeffries, 1985). Moreover, CTD casts in the channel (not shown here) also support this hypothesis, showing that the water flowing in the channel at the ADCP mooring location is generally within 0.02°C of its freezing temperature during the summer. This result shows that the MIS basal channel is not necessarily an area of concentrated melt, even in the summer. This is in marked contrast to what is found in the literature at other locations (Alley et al., 2016; Dutrieux et al., 2013; Rignot and Steffen, 2008; Stanton et al., 2013).

The present study can only make a statement on the channel area of the ice shelf and it is important to keep in mind that the ice shelf in general has a negative mass balance (Hamilton, 2016; Mortimer et al., 2012). Moreover, even though the channel area is in mass balance, $\sim 8$ m of ice is not a lot and a sudden break up can happen any time, has it happened for Ward Hunt Ice Shelf, which had a minimum ice thickness of 30 m (Mueller et al., 2003). This is in line with studies forecasting enhanced fracture and break up of ice shelves in Antarctica because of channeling (Dow et al., 2018; Gourmelen et al., 2017; Rignot and Steffen, 2008).
3.6.3 Mixing

The most surprising result of this study is that mixing is higher in the epishelf lake than below the halocline. MEL is constantly covered with ice which limits wind mixing and radiative convection (as soon as snow covers the ice). This suggests the epishelf lake should be mixing less, being isolated from the seawater by a sharp halocline.

Mixing during the first part of the winter (September-November) occurs both above and below the halocline which suggests this is linked to coastal upwelling (c.f. Jackson et al. (2014b)). Wind data from the Purple Valley weather station (2011 to present) and from a weather station on the ice shelf (2016-2018) shows that along-coast (NE) winds are low during the whole year except from July to September. If a shore lead is present or if the wind stress drives sea ice movement (Williams et al., 2006), coastal upwelling is expected to happen during these NE wind episodes. Examination of the 50, 150 and 300 m temperatures shows an upward deflection of isotherms during these periods (not shown). Another possible mechanism influencing summer circulation is subglacial discharge. Recent studies identified subglacial discharge as an important circulation driver in Greenland fjords (Carroll et al., 2015; Mortensen et al., 2014; Sciascia et al., 2013; Straneo and Cenedese, 2015; Xu et al., 2012). We then suggest that mixing occurring just after the end of the melt season is linked to coastal upwelling and possible residual subglacial melting and discharge; these two processes generating mixing by increasing the circulation (energy) in the fjord. It is noteworthy to mention that residual meltwater was observed in Petermann Fjord months after the melt season was over (Washam et al., 2019) and subglacial discharge was discovered in the middle of the winter under a glacier in Yukon (Schoof et al., 2014).

Aside from a peak in January 2012 and April 2019, mixing below the halocline is at molecular levels ($K_\Theta = 1.4 \times 10^{-7} \text{m}^2 \text{s}^{-1}$) for the entire time after the residual summer mixing vanishes. Without wind or solar radiation, with minimal tides and a solid sea ice cover, the processes leading to enhanced mixing above the halocline in winter are very limited. Analysis of available meteorological data found no correlation with mixing. However, examination of the water temperature timeseries suggests that enhanced mixing is linked to the presence of 7 to 30 day oscillations.
in the temperature signal (Figure 3.6B). Because there is no addition of heat to the lake during winter, these temperature oscillations can only be due to advection of horizontal temperature gradients in MEL. Considering that forcing mechanisms are weak and the mooring is in the middle of the lake (i.e. away from lateral boundaries), geostrophic equilibrium can be assumed. Horizontal eddies, created by the balance between the Coriolis force and horizontal pressure gradients are relatively common in ice-covered lakes (Forrest et al., 2013; Rizk et al., 2014) and oceans (Hunkins, 1974; Timmermans et al., 2008). These coherent rotary structures have recently been observed in Lake Baikal as they create circles of thinner ice that appear darker from above (Kouraev et al., 2016). An image from Milne Fiord taken in 2013 shows a black disc similar to those in Lake Baikal, which may indicate the presence of eddies that year (see Appendix B).

In order to link the temperature oscillations seen by the mooring and the possible presence of horizontal eddies, the thermal wind equation is used to calculate the vertical shear from the horizontal density gradient (Kundu et al., 2012):

\[
\frac{\partial u_1}{\partial x_3} = \frac{g}{\rho_0 f} \frac{\partial \rho}{\partial x_2}
\]  

(3.5)

Where \( f \) is the Coriolis frequency (1.45x10^{-4} s^{-1} at 82.6°N). The thermal wind equation is employed to scale the Richardson number:

\[
Ri = \left( \frac{N^2}{\frac{\partial u_1}{\partial x_3}} \right)^2 \approx \left( \frac{N^2}{\frac{g}{\rho_0 f} \frac{\partial \rho}{\partial x_2}} \right)^2 \approx \left( \frac{N^2}{\frac{g}{\rho_0 f} \frac{\Delta \rho_{30}}{R_L}} \right)^2 = \left( \frac{N^2 \rho_0 H}{g \Delta \rho_{30} \pi} \right)^2
\]  

(3.6)

Where the Rossby radius of deformation (\( R_L = NH/f \pi \)) is used to scale the horizontal span of the eddies. \( \Delta \rho_{30} \) is the 30 day maximum density variation related to the temperature oscillations. It is calculated using the mooring temperature data, assuming that the mooring captures both the center and the periphery of the temperature anomalies (eddies) during a 30 day interval. The temperature oscillations (period of 7 to 30 days) from the mooring data have a greater amplitude than the 30 day temperature difference given by the model (i.e. \( \Delta T_{30}^{mooring} > \Delta T_{30}^{model} \)) which gives confidence that \( \Delta \rho_{30} \) is attributable to horizontal movements (eddies moving around). Moreover, it is acknowledged that the model has higher mixing coeffi-
cient when it encounters temperature oscillations because of the fitting method employed. However, consistently high mixing coefficients (30 day average) are undeniably related to increased vertical mixing. $N^2$ is computed with the model results and the vertical height scale is taken as the distance between the bottom of the ice cover and the point where $N^2$ becomes larger than 0.01 s$^{-2}$, which is larger than the top layer of the model because its top boundary is at the first unfrozen thermistor. The ice thickness is approximated to increase linearly from 0.5 m at the end of the melt season to 1.5 m at the beginning of the next one. This rough approximation is derived from the small number of data available (14 data points in 8 years from direct measurement and moored thermistor freezing). Figure 3.9B shows the average Richardson number in the epishelf lake as well as vertically averaged quantities used for its calculation ($N^2$ and $\Delta \rho_{30}$) and the 30 day averaged mixing coefficient for heat in the top layer returned by the model ($K_{top}^{\Theta}$). Peaks in the Richardson number (and $\Delta \rho_{30}$) are definitely correlated with increased mixing. This supports the hypothesis that vertical mixing in MEL during the winter is linked to one or more geostrophic wave-like structures. Comparing results with common $K \propto Ri$ parameterizations (Figure 3.10) (Pacanowski and Philander, 1981; Peters et al., 1988), the mixing in the epishelf lake is lower than what would be expected for a traditional shear flow with a similar Richardson number. Considering that the top of the water column in Milne Fiord is very calm (maximum 30 day averaged mixing is 3.6x10$^{-5}$ m$^2$ s$^{-1}$) in comparison with the studies parameterizing mixing in the ocean (Pacanowski and Philander, 1981; Peters et al., 1988), this is not surprising. Moreover, also taking into account our hypothesis that the mixing is linked to the temperature oscillations with long periods, we do not expect small Richardson numbers (large shear values) but rather a laminar environment with enhanced molecular diffusion modulated by the strength of the eddies.

Equation 3.6 can also be used to estimate horizontal velocities. Using average values of $\approx 10^{-4}$ s$^{-2}$ for $N^2$, $\approx 10^2$ for $Ri$ and $\approx 8$ m for the vertical scale, a horizontal velocity of $\approx 1$ cm s$^{-1}$ is obtained. This is similar to that reported in ice-covered lakes (Bengtsson, 1996; Forrest et al., 2013; Huttula et al., 2010) and what ADCP measured in MEL (Hamilton, 2016).

As previously mentioned, sources of energy in the epishelf lake are very limited during winter. Because mixing does not show any tendency to decrease after the
end of the melt season, something has to be energizing the lake motion throughout
the winter. Estimation of the Ekman spin down time $t_E = D/(\sqrt{2KF})$ (Pedlosky,
2013) with the height of the eddies $D \approx 8$ m and the eddy viscosity $K \approx 10^{-6}$ m$^2$
s$^{-1}$, gives a time scale around 5 days. This is obviously too short to attribute the
existence of the eddies to residual energy from summer processes. Even using
only molecular diffusivity, the spin down time is too small (50 days). Numerical
modeling of Lake Untersee (Antarctica) has shown that the presence of an ice wall
created a gyre in winter due to the change of water properties following ice-water
interactions Steel et al. (2015). Since the main body of MEL is bordered by ice
upfjord and downfjord (Figure 3.1), cooling of epishelf lake water due to ice melt-
ing could possibly create a clockwise gyre. This could explain the presence of the eddies (and enhanced mixing) throughout winter.

3.7 Conclusion

Here we have used a one-dimensional model to analyze the winter mooring data of the top water column in Milne Fiord from 2011 to 2019. Three major results stand out from the analysis.

1. The model outflow rates, compared with two years of ADCP data, show that the main outflow path for the epishelf lake water is likely through a basal channel under the ice shelf. This can be exploited in hydrological studies of Milne Fiord watershed.

2. The model outflow coefficients indicate that the channel area is in ice mass equilibrium over a multi-year time scale. This implies that Milne Ice Shelf has been stable in the last 8 years but could suddenly break at any time, similar to the Ward Hunt Ice Shelf in 2001/2002.

3. The model mixing coefficients reveal that mixing is greater in the epishelf lake than in the seawater below the halocline. Moreover, estimation of the Richardson number shows that enhanced mixing in the epishelf lake is linked to one or more geostrophic wave-like structures (eddies). This demonstrates that even though a body of water is ice- and snow-covered, it is not necessarily quiescent and processes of lower importance can play a significant role. As the climate continues to change, a better knowledge of ice-covered water bodies is important to predict their evolution.
Chapter 4

Conclusion

4.1 Summary

The goal of this study was to further the work done in Milne Epishelf Lake (Hamilton, 2016) and analyze the mooring data more in depth. Eight years of mooring data outside of the melt season (~September to ~May) were used, together with a one-dimensional model, to study the mixing in the upper water column and the hydraulic characteristics of the ice shelf basal channel. The mooring data and model were used in an inverse fashion in order to find four key parameters: 1) the daily mixing coefficient in the freshwater above the halocline (~3 to ~10 m), 2) the daily mixing coefficient in the seawater below the halocline (~14 to 25 m), 3) the annual minimum ice shelf draft in the ice shelf basal channel and 4) a width-friction coefficient for the flow in the channel. An iterative calibration procedure was used to find the best fitting mixing coefficients that matched the mooring data. As the model validation indicates, these four parameters are sufficient to describe the main physical processes in the upper water column of Milne Fiord during winter.

The results show the upper water column of the fjord is very calm, which is expected, but, surprisingly, mixing is more pronounced in the freshwater layer (mean $K_\Theta=2.5\times10^{-6}$ m$^2$ s$^{-1}$) above the halocline than in the seawater below (mean $K_\Theta=1.0\times10^{-6}$ m$^2$ s$^{-1}$). This is the first quantitative estimation of mixing in an epishelf lake. Using geostrophic balance to scale the Richardson number, it appears that high mixing episodes from November to May are related to horizontal
density gradients caused by temperature anomalies. It is suggested that these gradients are associated with horizontal eddies similar to those observed in Lake Baikal (Kouraev et al., 2016). These findings have implications for ice-covered lakes, epishelf lakes and polar fjords where the dynamics of these systems are poorly understood. They demonstrate that even though a water body seems fairly isolated, its physical structure can be dynamic. Understanding the underlying physics in these remote and vulnerable systems is key to predict the future of these water bodies in regards to climate change.

ADCP measurements in the Milne Ice Shelf channel from July 2017 to July 2019 show an outflow of the same order of magnitude as calculated by the model. This, combined with ice penetrating radar measurements on the ice shelf, implies that this channel is likely the only outflow pathway for the epishelf lake water. The weir equation parameters of the model (minimum draft of the ice shelf, \( h_0 \), and width-friction parameter, \( C_e b \)) do not exhibit any trend over the eight years of the study. If melt was occurring at a similar rate to what is reported for many channels under Antarctic ice shelves (e.g. Alley et al. (2016); Dutrieux et al. (2013); Rignot and Steffen (2008); Stanton et al. (2013)), there would be a consistent decrease in the minimum draft and an increase in the channel width. Since the model indicates these parameters do not change appreciably, it is then suggested that Milne Ice Shelf basal channel is in ice mass balance on a multi year time scale. This is in marked contrast to what is found in the literature at other locations (e.g. Alley et al. (2016); Dutrieux et al. (2013); Rignot and Steffen (2008); Stanton et al. (2013)). This implies melting and refreezing can have similar magnitudes even in basal channels. However, this does not preclude a catastrophic failure as the ice thickness along the channel is relatively thin. Therefore, a break up like the one on the Ward Hunt Ice Shelf in 2002 (Mueller et al., 2003), could happen at any time.

### 4.2 Future Work

Even though this study helps understanding Milne Fiord system a little better, it also leaves many possible research topics that would be interesting to elucidate before the ice shelf breaks up.
A natural continuation of this study would be to confirm the presence of eddies in Milne Fjord epishelf lake by direct measurement and explain how they form. In order to do this, better horizontal coverage is needed during the winter period. This could be achieved in different ways. Adding several shallow moorings or carrying out an intensive CTD profiling campaign would clarify this topic. The problem with adding supplemental moorings is that the actual extent and structure of the eddies is unknown and could be missed or the acquired information maybe insufficient. A CTD campaign would be more flexible in that it can be adjusted as the data are acquired, but the temporal resolution would be limited and it would present a logistical challenge to sample closely and rapidly over kilometers. The best method to improve horizontal coverage would be to use an autonomous underwater vehicle (AUV). This would give the best data but would also entail a substantial risk due to the inferior reliability of AUVs, especially under ice (Kaminski et al., 2010). Indeed, the deployment of an AUV under ice involves additional risks since the vehicle cannot surface if a problem arises. In addition to the increased risks, an AUV survey is a high cost operation with complex logistics. However, at the present time, no AUV has been deployed in a Canadian or Greenlandic fjord (for science purposes), which would definitely make the experience more rewarding.

The further constraint of the basal channel dimensions would definitely help improve the usefulness of the ice shelf mooring data. This could be achieved by deploying a remotely operated vehicle (ROV) in the channel, with mapping a sonar. This would enable a more precise calculation of the freshwater outflow and a better estimation of the heat and mass fluxes in the channel, the latter achieved by comparing the data from the epishelf lake and the ice shelf moorings. This would give great insights on processes happening in other channels in Antarctica and Greenland (e.g. (Washam et al., 2019)).

The present study used the first 25 m of the available mooring data, but other instruments also recorded temperature at lower depths (50 m, 150 m, 315 m and sometimes 260 m). Although most likely not enough on its own, this data could be combined with the bank of over 100 summer CTD profiles or more extensive winter data to make a macroscopic heat, salt and freshwater budget, as it is done for Greenlandic fjords (Jackson and Straneo, 2016). This would allow a quantification of the circulation mechanisms in the fjord and a further constraint of the melt
estimates by Hamilton (2016).
Milne Fiord epishelf lake is the last remnant of what used to be a common hydrographic system on the northern coast of Ellesmere Island. Yet, it is realistic to say that it will drastically change in the next decade or two, when Milne Ice Shelf collapses and the epishelf disappears. The study of this unique site is ideal to understand the effect of climate change in the Canadian High Arctic and advance the knowledge on ice-covered lakes, glacial fjords and ice shelves.
Bibliography


IPCC. The physical science basis. Contribution of working group I to the fifth assessment report of the Intergovernmental Panel on Climate Change. 2013. → page 28


T. J. McDougall and P. M. Barker. Getting started with TEOS-10 and the Gibbs seawater (GSW) oceanographic toolbox. SCOR/IAPSO WG127, 2011. → page 34


NEIGE. Water column physico-chemical profiles of lakes and fiords along the northern coastline of Ellesmere Island v.1.1 (1954-2016), 2017. → page 34


67


<table>
<thead>
<tr>
<th>Author(s)</th>
<th>Title</th>
<th>Journal/Conference, Volume:Issue/Pages, Year</th>
<th>Pages</th>
</tr>
</thead>
</table>


Appendix A

Turbulent Lewis Number

In order to reduce the number of unknown parameters of the model, the mixing coefficients for heat and salt in the top layer (epishelf lake) are linked together and the mixing coefficients for heat and salt in the bottom layer are also linked together (section 3.5.4). This reduces the number of daily mixing parameters from four to two. In order to link together heat and salt mixing coefficients, the parameterization by Jackson and Rehmann (2014) based on laboratory experiments is employed. This is done in two steps. The first is to estimate the $\frac{\varepsilon}{\nu N^2}$, where $\nu$ is the kinematic viscosity, $\varepsilon$ is the rate of dissipation of turbulent kinetic energy and $N$ is the buoyancy frequency. This is done using the following equation (see table 2 of Jackson and Rehmann (2014)):

$$K_{CT} \nu = 0.31 \left( \frac{\varepsilon/\nu N^2}{0.36} \right)^{0.16} \text{ for } \varepsilon/\nu N^2 < 0.36$$

(A.1)

$$K_{CT} \nu = 0.31 \left( \frac{\varepsilon/\nu N^2}{0.36} \right)^{1.06} \text{ for } \varepsilon/\nu N^2 > 0.36$$

(A.2)

Once $\varepsilon/\nu N^2$ is found, $K_{SA}$ is computed with the following equation:

$$\frac{K_{CT}}{K_{SA}} = \frac{1 + R}{2} + \left( \frac{1 - R}{2} \right) \cdot tanh \left( 0.92 \left[ log_{10} \left( \frac{\varepsilon}{\nu N^2} \right) - 0.6 \right] \right)$$

(A.3)

Where $R$ is the ratio of molecular diffusivity for salt over molecular diffusivity for heat.
Equation A.1 and A.2 introduce a steep jump around the critical value $\varepsilon/\nu N^2 = 0.36$, which is the cause of the greater discrepancy between $K_{SA}$ and $K_{\Theta}$ when $K_{\Theta} < 5.4 \times 10^{-7}$. From a physical point of view, it means that when $K_{\Theta} < 5.4 \times 10^{-7}$, heat diffuses much more rapidly than salt. From $K_{\Theta} = 5.4 \times 10^{-7}$ upward, the difference between $K_{SA}$ and $K_{\Theta}$ becomes smaller as turbulent mixing dominates. This parameterization discrepancy definitely has an impact on daily values, but it is not significant here as 30 day average values are used for the analysis.
Appendix B

Ice Circle in Milne Fiord

Giant ice rings have been observed in Lake Baikal and Lake Hovsgol (Mongolia) from 1974 to 2014 (Kouraev et al., 2016). Using optical imagery, Kouraev et al. (2016) identified 45 dark rings created by circles (not discs) of thinner ice. The diameter of these rings ranged from 2.2 to 8.2 km and had a dimension similar to the Rossby radius \( R_L = ND/f \pi \). The formation of these ice rings is attributed to the presence of anti-cyclonic eddies. These rotary structures are thought to increase the heat exchange between the water and the ice, resulting in a thinner ice at the periphery of eddies.

A RADARSAT-2 image from Milne Fiord in April 2013 (Courtesy of the Canadian Ice Service, Environment and Climate Change Canada) shows a similar feature in Milne Fiord epishelf lake with diameter around 3000 m (Figure B.1). However, Figure B.1 is a radar image (not optical) and ice cores suggest that the disc-shaped reduction of backscatter in the epishelf lake is due to a decrease of the amount of bubbles in the ice (McCallum, 2015). Clear ice has a lower microwave backscatter than ice with bubbles, which results in a darker tone in Figure B.1. Thinner ice would also decrease the amount of backscatter, but there is no evidence that the ice was thinner under the dark disc. What caused the ice to have less bubbles might be linked to snow melt, which in turn could be linked to water properties.
Figure B.1: RADARSAT-2 Fine Quad image taken on the 27th of April 2013 showing a dark disc in the middle of the Milne Fiord epishelf lake (Courtesy of the Canadian Ice Service, Environment and Climate Change Canada). Panels VV, VH and HH are three different polarizations (VV: vertical-vertical, VH: vertical-horizontal and HH: horizontal-horizontal) and the RGB panel is the combined Pauli decomposition (HH: red, VV: green and VH/HV: blue). The disc is visible in all polarizations. Color and gray scales are qualitative with brighter tones indicating higher backscatter. RADARSAT-2 Data and Products © MacDONALD, DETTWILER AND ASSOCIATES LTD. (2013) – All Rights Reserved, RADARSAT is an official mark of the Canadian Space Agency.
Appendix C

Model Evaluation

This is a more extensive explanation of the evaluation of the model. The root mean square error (\(RMSE\)), the standard deviation (\(STD\)) are calculated for conservative temperature (CT) and Salinity (SA). The coefficient of determination (\(R^2\)) and the bias are also shown for conservative temperature. In addition, the custom coefficient of agreement (\(C_a\), see equation 3.4) is presented.
Table C.1: Model evaluation statistics

<table>
<thead>
<tr>
<th>Year / Simulation</th>
<th>CT RMSE</th>
<th>CT STD</th>
<th>CT $R^2$</th>
<th>CT bias</th>
<th>SA RMSE</th>
<th>SA STD</th>
<th>$C_a$</th>
</tr>
</thead>
<tbody>
<tr>
<td>2011-2012</td>
<td>0.21</td>
<td>0.21</td>
<td>0.96</td>
<td>-0.008</td>
<td>3.7</td>
<td>3.3</td>
<td>4.2</td>
</tr>
<tr>
<td>2012-2013</td>
<td>0.19</td>
<td>0.19</td>
<td>0.98</td>
<td>0.050</td>
<td>1.9</td>
<td>0.7</td>
<td>6.0</td>
</tr>
<tr>
<td>2013-2014</td>
<td>0.13</td>
<td>0.12</td>
<td>0.98</td>
<td>0.048</td>
<td>0.3</td>
<td>0.2</td>
<td>8.7</td>
</tr>
<tr>
<td>2014-2015</td>
<td>0.23</td>
<td>0.22</td>
<td>0.96</td>
<td>0.08</td>
<td>0.2</td>
<td>0.18</td>
<td>5.0</td>
</tr>
<tr>
<td>2015-2016</td>
<td>0.19</td>
<td>0.15</td>
<td>0.98</td>
<td>0.11</td>
<td>1.2</td>
<td>0.94</td>
<td>6.6</td>
</tr>
<tr>
<td>2016-2017</td>
<td>0.29</td>
<td>0.22</td>
<td>0.94</td>
<td>0.19</td>
<td>2.6</td>
<td>2.3</td>
<td>3.7</td>
</tr>
<tr>
<td>2017-2018</td>
<td>0.16</td>
<td>0.15</td>
<td>0.97</td>
<td>0.055</td>
<td>2.8</td>
<td>2.2</td>
<td>4.9</td>
</tr>
<tr>
<td>2018-2019</td>
<td>0.13</td>
<td>0.13</td>
<td>0.98</td>
<td>0.004</td>
<td>3.2</td>
<td>2.4</td>
<td>6.2</td>
</tr>
<tr>
<td>Average</td>
<td>0.19</td>
<td>0.17</td>
<td>0.97</td>
<td>0.067</td>
<td>2.0</td>
<td>1.7</td>
<td>5.6</td>
</tr>
</tbody>
</table>
Appendix D

Photo Gallery
Figure D.1: Getting dropped off at Purple Valley campsite by a Kenn Borek Twin Otter. Maximum 2500 pounds, crew included... Photo: Jérémie Bonneau
Figure D.2: Milne Fiord, looking downfjord. The glacier grounding line area is visible on the right of the icefall and Purple Valley, 75% on the right. The ice-covered ocean on the horizon. Photo: Jérémie Bonneau
Figure D.3: Northern Ellesmere icefields, looking upfjord from the same location as Figure D.2, ∼ 5000 feet up.
Photo: Jérémie Bonneau
Figure D.4: Derek Mueller and Drew Friedrichs looking at Milne Ice Shelf, easily distinguishable by its rolling topography with the trough filled with melt water. Photo: Jérémie Bonneau
Figure D.5: Moving the camp from Purple Valley to the ice shelf. Drew Friedrichs signaling "upward" to the helicopter pilot. Photo: Jérémie Bonneau
**Figure D.6:** Drew Friedrichs doing a CTD profile after pulling the epishelf lake mooring out. Looking upfjord. Photo: Jérémie Bonneau
Figure D.7: CTD profile in a crack on the margin of the glacier tongue. Photo: Drew Friedrichs
Figure D.8: Pulling the ice penetrating radar around the grounding line of the glacier. Photo: Drew Friedrichs
Figure D.9: Drew Friedrichs "hole melting" to get the ice shelf channel mooring back. Photo: Jérémie Bonneau
Figure D.10: Picking up a camera from the summit of Mega Nunatak (unofficial name). Photo: Derek Mueller
Figure D.11: 26-Resolute, 26-Resolute, this is Purple Valley, Purple Valley. Drew Friedrichs and Yulia Antropova broadcasting (blurred). Photo: Jérémie Bonneau
Figure D.12: Looking upfjord, with Derek Mueller. Mega Nunatak (unofficial name) on the left of the fjord. Photo: Jérémie Bonneau