Geophysics, Water Balance, and History of Thick Perennial Ice Covers on Antarctic Lakes

BY

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THESIS

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This is dedicated to all those who shared a smile with me over the last four years.

Keep smiling.

And remember, “Many a fine theory has been punctured by a drill hole.” – FJ Pettijohn (1956)

In defense of outdoor geology, *Bull. Am. Assoc. Pet. Geol.*, *40*, 1455–1461.

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The McMurdo Dry Valleys quickly felt like a second home thanks to the wonderful people I was fortunate enough to live with over the years. Peter Glenday, Ema Kuhn, Maciek Obryk, Kyle Cronin, and Luke Winslow put up with me day and night for too many months with too few showers. Amy Chiuchiolo, Chris Jaros, Thomas Nylen and Rae Spain tirelessly answered all of my questions. Jill Mikucki, Slawek Tulaczyk, and Steve Arcone were knowledgeable and supportive mentors at different stages of this winding journey. Thank you.

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HD

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LIST OF ABBREVIATIONS

AEM	Airborne Transient Electromagnetic
asl	Above Sea Level
DOI	Depth of Investigation
DVDP	Dry Valley Drilling Project
FIR	Finite Impulse Response
GIS	Geographic Information Systems
GMWL	Global Meteoric Water Line
GPR	Ground Penetrating Radar
GPS	Global Positioning System
LMWL	Local Meteoric Water Line
LTER	Long Term Ecological Research
MCM	McMurdo
NSERC	Natural Sciences and Engineering Research Council of Canada
NSF	National Science Foundation
OSL	Optically Stimulated Luminescence
PAR	Photosynthetically Active Radiation
PVC	Polyvinyl chloride
TC	Total Carbon
TDEM	Time Domain Electromagnetic
TIC	Total Inorganic Carbon

TOC	Total Organic Carbon
UIC	University of Illinois at Chicago
VSMOW	Vienna Standard Mean Ocean Water
w.e.	Water Equivalent
yr BP	Years Before Present

SUMMARY

This dissertation aims to better understand physical hydrological processes in the McMurdo Dry Valley lakes of East Antarctica. The overarching hypothesis is that **a better understanding of hydrologic and sediment transport processes associated with lake ice formation and preservation can be used to reveal past climatic changes, and further our awareness of current changes in climate and water balance in the Dry Valleys**. The two main goals of this dissertation are given below.

1. **To understand water loss from closed basin lakes in Taylor Valley, Antarctica:** Permanently ice-covered lakes are in a delicate hydrologic balance with climate. Warmer temperatures will lead to ice melt and eventually seasonally ice-free lakes, just as colder temperatures will lead to ice cover thickening. In the closed basin lakes of the McMurdo Dry Valleys, water is lost solely through sublimation and some evaporation. The second chapter of this dissertation details the ablation (loss of lake ice mass through sublimation and melt) record measured from 2000-2010, and the production of the first long term empirical record of ice ablation on a lake. It is hoped that this research will contribute to future sublimation models for the Taylor Valley lakes, and attempts to model the requisite conditions for complete ice loss.

2. **To reconstruct the formation of Lake Vida, Victoria Valley, Antarctica:** Lake Vida has a 27+ m ice cover, which has sealed in and isolated liquid brine at depth creating an extreme and unique aquatic habitat. In 2010, our research team recovered two long ice cores (27 and 20 m) as well as a suite of brine samples from Lake Vida. This builds on previous field campaigns in 1996 and 2005 [*Doran et al.*, 2003; *Murray et al.*, 2012]. The third chapter of this dissertation examines ice chemistry and ground penetrating radar (GPR) imagery to understand how the ice cover formed, and discusses how distinct sediment layers in the ice cover may reveal climatic variability in Victoria Valley. The fourth chapter examines GPR profiles and a helicopter borne time-domain electromagnetic (TDEM) survey to map the location and volume of the highly saline brine beneath Lake Vida.

1: INTRODUCTION

Antarctic lakes are considered extreme systems from almost every perspective. Biologically, some of the lakes host cold and dark adapted microbes and support limited trophic levels [*Fritsen and Priscu*, 1998; *Morgan-Kiss et al.*, 2006], and chemically, some boast higher concentrations of dissolved solids and gases than seen in temperate lakes [*Priscu et al.*, 1996; *Samarkin et al.*, 2010; *Murray et al.*, 2012]. There is also significant variation in physical characteristics between the lakes. Within the McMurdo Dry Valleys, lakes are found that range from shallow ponds to 70 m deep, freshwater to hypersaline, shallow groundwater to surface water fed, and endorheic (closed-basin) to exorheic (open-basin). Furthermore, most are permanently ice-covered, which restricts gas exchange, light penetration, and mixing by wind. In Antarctica, there are less than a dozen locations where lakes are documented to have retained an ice-cover through most summers, these include the Bunger Hills, Vestfold Hills and Schirmacher Hills [*Doran et al.*, 2004]. In the Arctic, only a handful of lakes in Greenland, the Canadian High Arctic, and Franz Josef Land have been documented to be permanently ice covered most years [*Doran et al.*, 2004; *Vincent et al.*, 2008].

The Dry Valley lakes are well studied, in part due to the inclusion of Taylor Valley as a National Science Foundation (NSF) Long Term Ecological Research (LTER) site in 1993, and its retention of this designation through to the present. Yet, the lakes remain poorly understood for a variety of reasons: i) research on the lakes began relatively recently, especially when compared to lakes in the Northern Hemisphere where records stretch back into the 1800s [*Magnuson et al.*, 2000]. It was not until 1903 that the Taylor Valley lakes were first encountered by Robert Falcon Scott during the Discovery expedition [*Scott*, 1905], and the

1960s when more rigorous scientific research began [*Armitage and House*, 1962; *Angino and Armitage*, 1963; *Wilson*, 1964; *Goldman et al.*, 1967]; ii) the inhospitable environment limits site access to the austral summer and restricts what equipment can be easily deployed; iii) the lakes are such end-member ecosystems that there are almost no analogous lakes in other environments for comparison.

1.1 The McMurdo Dry Valley lakes

The McMurdo Dry Valleys (77.5°S, 162°E) are the largest ice-free region of Antarctica (Figure 1). Taylor, Wright, and Victoria Valleys cover most of the central region of the valleys, which have a total ice free area of 3000 km^2 [*Levy*, 2013]. In Taylor Valley, three large lakes, Lake Bonney, Lake Hoare, and Lake Fryxell, are situated from west to east along the valley floor, and in Wright Valley, Lake Vanda is situated in the center of the valley. All lakes are permanently covered with 3-7 m of ice, and have no outlets [*Priscu et al.*, 1998; *Doran et al.*, 2002a]. The lake ice acts as an upward conveyor belt through time, as bottom ice slowly becomes surface ice as the top-most layers ablate from the floating ice cover. In the summer, ice covers typically thin by less than 1.5 m [*J. Priscu, unpublished data*], and a narrow moat of liquid water develops around the edge of the lakes; however, open water accounts for only a small percentage of the total surface area. During the spring melt, streams flow into the lakes and beneath the ice cover. However, atmospheric exchange is limited, as evidenced by disequilibrium between surface water and atmospheric gases concentrations [*Wharton Jr. et al.*, 1993; *Hood et al.*, 1998; *Neumann et al.*, 2001].

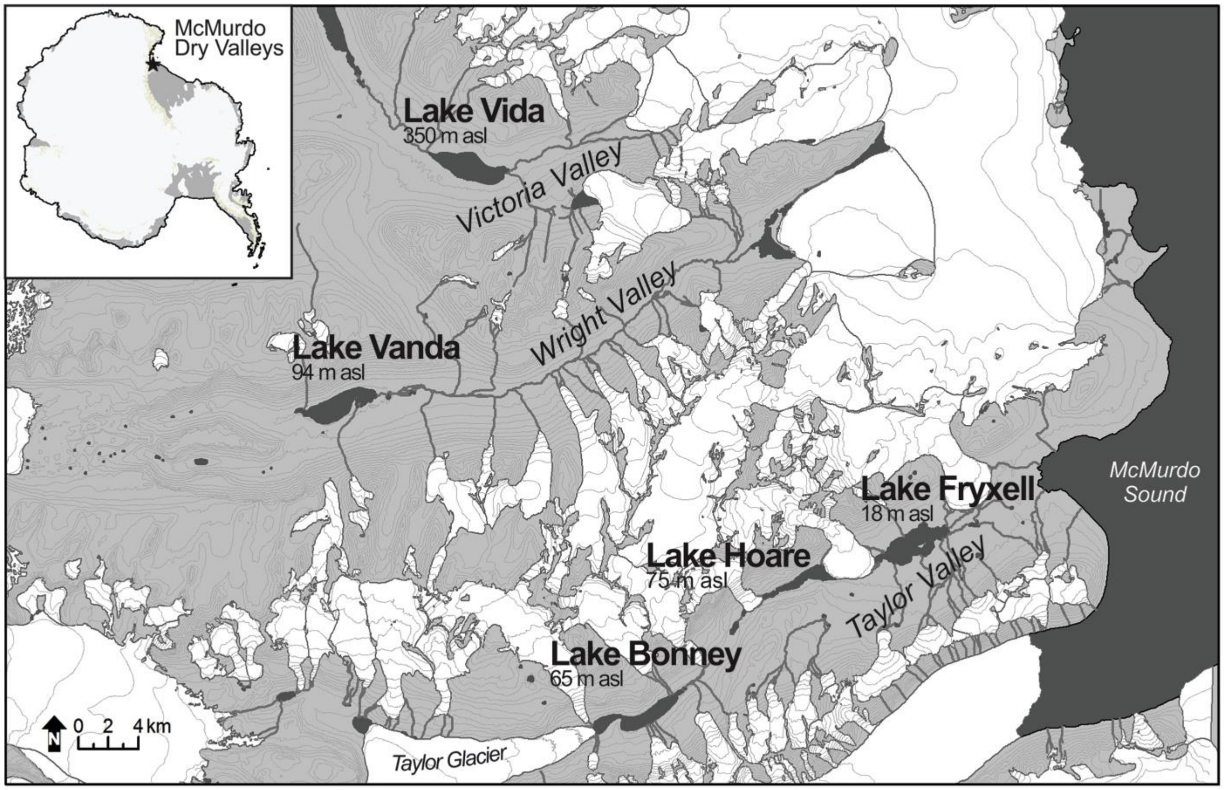

Figure 1. The central McMurdo Dry Valleys, Antarctica.

Lake Vida is located at 350 m above sea level (asl) in the center of Victoria Valley, and has the second largest surface area of all of the lakes in the McMurdo Dry Valleys (Figure 1). The lake is endorheic, and receives inflow via streams originating from Victoria Upper, Victoria Lower, and Clark Glaciers. The mean valley bottom temperature of Victoria Valley is noticeably different from Taylor and Wright Valleys [TABLE I, *Doran et al.*, 2002b]. This variation is largely driven by the lack of strong winter winds in the Victoria Valley that warm the Taylor and Wright Valleys.

TABLE I. MEAN AND ABSOLUTE MINIMUM AIR TEMPERATURES AT SIX LAKES IN THE MCMURDO DRY VALLEYS [*Doran et al.*, 2002b].

	Taylor Valley			**Wright Valley**		**Victoria Valley**
	L. Fryxell	**L. Hoare**	**L. Bonney**	**L. Brownworth**	**L. Vanda**	**L. Vida**
Mean temp (^oC)	-20.2	-17.7	-17.9	-20.9	-19.3	-27.4
Min temp (^oC)	-60.2	-45.4	-47.9	-51.9	-53.7	-65.7

From the surface, Lake Vida does not appear much different than the lakes in Taylor Valley, where liquid water is present below 3-7 m of ice. However, even with dynamite, the first attempts to penetrate Lake Vida were thwarted. Upon reaching an impenetrable layer of sand at 11.5 m, it was hypothesized that Lake Vida was most likely frozen to its base [*Calkin and Bull*, 1967]. In spite of this, a resistivity survey was conducted in the 1970s. The study found high seismic velocities and low resistivities at 40 m below the center of the lake, which are indicative of unfrozen saline water or sediment [*McGinnis et al.*, 1973]. In 1995, ground penetrating radar (GPR) transects revealed a strong reflector at ~20 m depth, which was hypothesized to be liquid brine [*Doran et al.*, 2003]. In 1996 and 2005, Dr. Peter Doran led the first deep drilling expeditions in an attempt to understand the formation and character of Lake Vida. In 1996, wet ice was encountered at 15.8 m, and ice coring was halted to prevent possible contamination. Ice cores were collected, and a thermistor string was left in the borehole. In 2005, brine was sampled at 16 m [*Doran et al.*, 2008a], but drilling was halted due to an early spring melt that threatened to flood the surface of the ice with river water. From these recent research projects, we have learned a great deal about the chemical conditions and microbial populations in the Lake Vida brine [*Mosier et al.*, 2007; *Mondino et al.*, 2009; *Murray et al.*, 2009, 2012]. However, we still know little about the physical hydrology of the lake, and what processes led to the formation of the 27+ m ice cover.

1.2 Outline

The following chapters focus on physical hydrological processes in the McMurdo Dry Valley lakes, with an overarching hypothesis **that an understanding of hydrologic and sediment transport processes associated with lake ice formation and preservation can be used to reveal past climatic changes, and further our awareness of current changes in climate and water balance in the Dry Valleys**.

Chapter 2 examines the rates of water loss from closed basin lakes in Taylor Valley. In the Dry Valleys, permanently ice-covered lakes are in a hydrologic balance with climate. Warmer temperatures will lead to higher river inflows and lake ice melt, and will result in higher lake levels (as there is no outflow) and thinner ice covers. It is not implausible that a warm summer could lead to ice-free conditions on a Taylor Valley lake. Seasonally ice-free conditions have never been observed in the McMurdo Dry Valleys. Conversely, colder temperatures will lead to ice cover thickening. In these closed basin lakes, water is lost solely through sublimation and some evaporation. This chapter documents the ablation (loss of lake ice mass through sublimation and melt) record measured by the LTER from 2000-2010. This is the first long-term empirical record of ice ablation on a lake, and has been published in the *Journal of Glaciology*.

Chapter 3 focuses on the processes that formed a thick ice cover on Lake Vida. Lake Vida has a 27+ m ice cover, which is the thickest ice cover of any surface lake on Earth, and is unlike any seen in the Dry Valleys. The ice cover has sealed in and isolated liquid-brine at depth, which has been considered one of the most extreme and unique aquatic habitats on Earth. This chapter presents data and analysis on the geochemistry of ice and sediment from two long ice cores to understand the processes that allowed the ice cover to thicken over time. Radiocarbon and optically stimulated luminescence (OSL) dates are used to reconstruct the timing of lake level fluctuations.

Chapter 4 follows on the results of Chapter 3, and describes a detailed geophysical study using GPR and a helicopter borne TDEM survey to interpret the extent and volume of highly saline brine beneath Lake Vida. This work provides an innovative view of subsurface hydrology, and is a primary step in modeling hydrogeological processes in unglaciated regions of Antarctica.

Chapter 5 concludes this dissertation, assesses the contributions made by this research, and suggests research questions that should be the focus of future work.

2: LAKE ICE ABLATION RATES FROM PERMANENTLY ICE COVERED ANTARCTIC LAKES

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2.1 Abstract

In the McMurdo Dry Valleys of Antarctica, three large, permanently ice-covered, closed-basin lakes exist along the floor of Taylor Valley. Lake ice ablation (loss of ice mass) is calculated as the sum of sublimation and surface melt, and is the driver of ice cover turnover in these systems. In Taylor Valley, both manual and automated lake ice ablation rates have been calculated from 2001 to 2011. Results indicate relatively consistent winter ablation of 0.07-0.21 m of ice loss (0.2-0.7 mm w.e. d^{-1}). Summer ablation of lake ice is more variable and ranges from 0.25-1.62 m (5-31 mm w.e. d^{-1}) over an average 51 day period. Previous to this study, ablation rates have been cited as 0.35 m yr^{-1} in the Dry Valleys from sublimation modeling based on meteorological variables. We show that this value has significantly underestimated mean ablation and ice cover turnover on the Taylor Valley lakes.

2.2 Introduction

Throughout the winter, most lakes located at high latitudes and altitudes develop a seasonal ice cover. During this period, and in the absence of an outflow, loss of water from the lakes is governed by ice sublimation, the phase change of ice directly to water vapor. Sublimation, like evaporation, is difficult to empirically quantify [*Box and Steffen*, 2001], and is often neglected in hydrological budgets, or assumed to be a very small fraction of overall water loss [*Woo*, 1980]. Modeling sublimation is possible, but necessitates the measurement of many climate variables and local ice conditions. Often, surface roughness and surface temperature must themselves be modeled, and albedo estimated, which compounds the associated error on final sublimation rates [*McKay et al.*, 1985; *Clow et al.*, 1988; *Brock and Arnold*, 2000; *Bliss et al.*, 2011]. Furthermore, the transience of most lake ice prevents the year-round installation of mass balance stakes, which are typically used in glacial settings to manually record ice loss [*Heron and Woo*, 1994].

In the McMurdo Dry Valleys of Antarctica, three large, permanently ice-covered, closed-basin lakes are located along the floor of Taylor Valley. The persistence of a year-round ice cover provides a stable platform and an ideal environment for empirically measuring the overall rate of ablation (ice loss), and sublimation (the transformation of ice to water vapor, which is lost from the system). In this extreme desert ecosystem, lake volumes are regulated almost solely by gain from glacially fed streams and direct glacial melt, and mass loss through sublimation [*Chinn*, 1993; *Doran et al.*, 1994]. Gain from snowfall and loss through the evaporation of liquid water at the lake edges are minor components. The only assessment of groundwater in the region is in association with the 1972-1974 Dry Valleys Drilling Project.

Boreholes at Lake Hoare and to the east of Lake Fryxell did not exhibit water level fluctuations or geothermal displacement of drilling fluid, which suggests an absence of subsurface flow [*Cartwright and Harris*, 1981]. During the dark austral winter, the negative water balance of the lakes reflects the sublimation of lake ice. In the summer, ablation of ice results from both surface sublimation and surface melt. It is important to note that ice melt, where water is assumed to drain downward into the lake (and not remain as inclusions in the ice), does not alter lake volume, as no water is lost from the system.

In this paper, we discuss ten years of winter sublimation and annual ablation rates from three large lakes in the McMurdo Dry Valleys. From 2001 to 2011, the levels of Lake Fryxell, Lake Hoare, and Lake Bonney, have risen rapidly [*Barrett et al.*, 2008]. Accurately quantifying ablation and sublimation rates is critical for closing the hydrological balance in the Dry Valleys, where sublimation is the fundamental control on water loss in this desert ecosystem. The longevity of this dataset is unequaled, and results should be used to corroborate modeled sublimation in this and other lacustrine settings.

2.3 Site Description

The McMurdo Dry Valleys receive only 3-50 mm of annual snowfall due to a precipitation shadow effect produced along the Transantarctic Mountains to the south, and as a result, are one of the few unglaciated regions of Antarctica [*Monaghan et al.*, 2004; *Fountain et al.*, 2010]. In Taylor Valley, three large lakes, Lake Bonney (4.6 km^2), Lake Hoare (2.3 km^2), and Lake Fryxell (7.4 km^2), are situated from west to east along the valley floor (Figure 2). Lake Bonney is separated into two basins by a shallow sill, and is commonly referred to as two bodies of water: East Lobe and West Lobe Bonney.

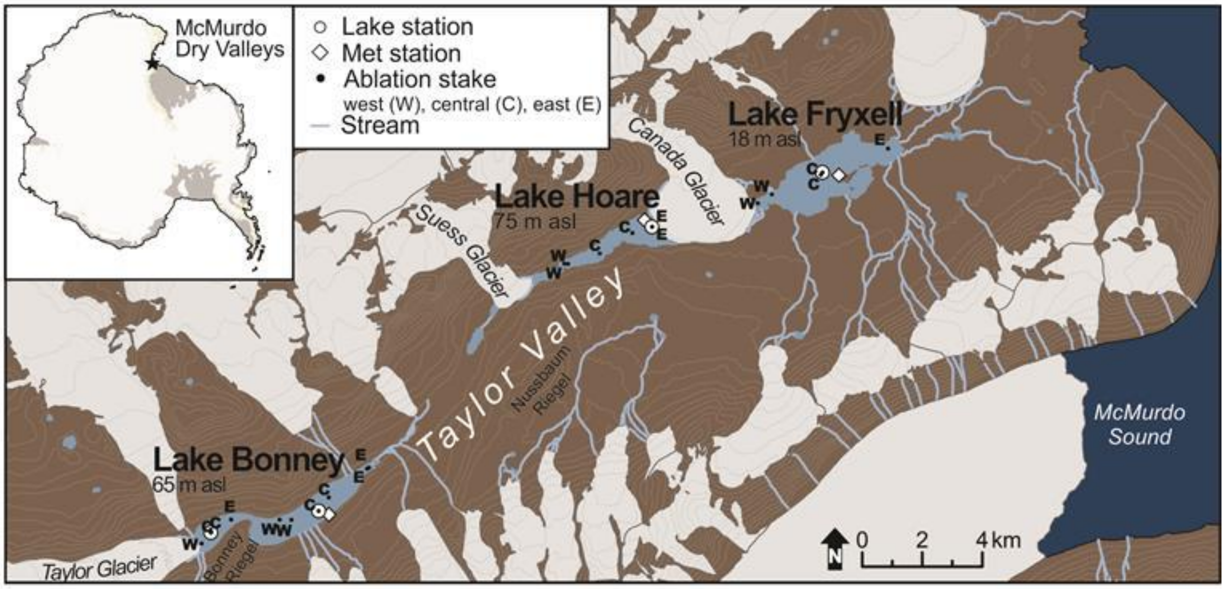

Figure 2. The location of ablation stakes, lake stations, and meteorological stations at Lake Bonney, Lake Hoare, and Lake Fryxell, in Taylor Valley, Antarctica. Contour interval is 100 m.

The lakes have distinct climates as a result of their position in the valley. Lake Fryxell, which lies adjacent to the coast, has the lowest mean annual temperature and highest annual precipitation [*Doran et al.*, 2002b; *Fountain et al.*, 2010] (TABLE II). Lake Hoare has the lowest mean annual wind speeds, as it is relatively sheltered from easterly coastal winds by the Canada Glacier, and westerly winds by the Seuss Glacier and Nussbaum Riegel [*Doran et al.*, 2002b; *Monaghan et al.*, 2004] (Figure 2). Lake Bonney is the warmest and windiest of the three lakes, with a climate that is largely moderated by westerly winds which surge off Taylor Glacier [*Nylen et al.*, 2004]. These winds warm air temperatures throughout the valley, but are especially noticeable and frequent at Lake Bonney, which is closest to the polar plateau.

TABLE II. MEAN ANNUAL, SUMMER (DEC-JAN), AND WINTER (FEB-NOV) AIR TEMPERATURE, WIND SPEED, INCOMING SOLAR RADIATION AND DEGREE DAYS ABOVE FREEZING AT THE LAKE FRYXELL (LF), LAKE HOARE (LH) AND LAKE BONNEY (LB) METEOROLOGICAL STATIONS FOR 2010 (www.mcmlter.org).

	2010 Mean Annual			**Dec-Jan Mean**			**Feb-Nov Mean**		
	LF	LH	LB	LF	LH	LB	LF	LH	LB
Air temperature (^{o}C)	-20.8	-18.2	-17.5	-2.0	-1.9	-0.3	-24.6	-21.5	-21.0
Incoming shortwave radiation (W m^{-2})	112.3	93.9	98.4	315.8	293.2	289.5	70.7	53.8	59.3
Wind speed (m s^{-1})	2.8	2.5	3.7	4.0	2.8	4.6	2.5	2.4	3.5
Relative humidity (%)	77.9	72.2	64.1	68.2	67.3	52.2	79.9	73.1	66.5
Degree days above freezing	18.6	21.1	67.0						

All lakes are permanently covered with 3-6 m of ice, and have no outlets [*Priscu et al.*, 1998; *Doran et al.*, 2002a]. The lake ice acts as an upward conveyor belt through time, as bottom ice slowly becomes surface ice as the top-most layers ablate. During the summer, ice covers typically lose less than 1.5 m of ice [*P. Doran, J. Priscu, unpublished data*], and a narrow moat of liquid water develops around the edge of the lakes; however, open water accounts for only a small percentage of the total surface area. Sediment is frequently blown onto the lake surfaces by strong winds, and can accumulate in hollows. The enhanced melt associated with sediment pockets is likely the origin of the rough surface ice (Figure 3) [*Simmons et al.*, 1986; *Squyres et al.*, 1991; *Jepsen et al.*, 2010].

2.4 Methods

Automated lake stations are situated in the center of Lake Fryxell, East and West Lobe Bonney, and the eastern side of Lake Hoare (Figure 2). A datalogger housed on the ice surface records the depth of two pressure transducers. A moored pressure transducer is used to calculate stage, and is affixed to a floating buoy weighted to the lake bottom (Figure 4a). The second pressure transducer is suspended from the ice on a cable, which is affixed with short sections of PVC to act as ice anchors, and moves upward through the water column as the ice ablates (Figure 4a). Ablation values are calculated as the change in depth of the hanging pressure transducer. As the depth is relative, ablation values are reset to a zero baseline following an extended sensor failure (Figure 4).

Figure 3. Heavily ablated surface ice at Lake Hoare, Dec. 2011. Note the dark sediment deposited in the hollows of the ice.

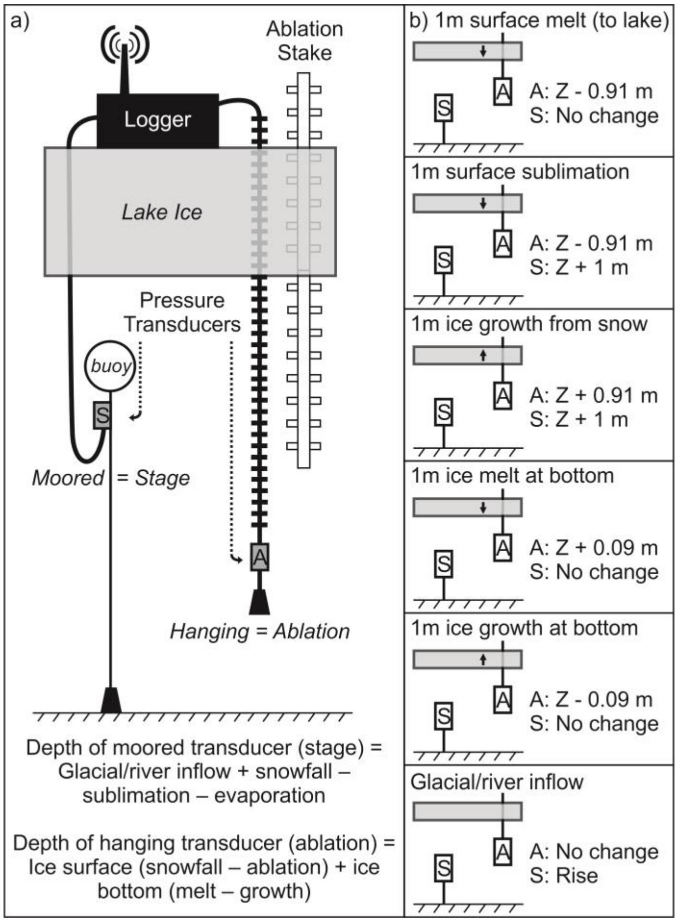

Figure 4. a) Setup of datalogger and sensors housed on the ice surface of each lake. The moored and hanging pressure transducers are used to calculate changes in stage and surface ice ablation, respectively, based on the equations given. b) The depth output (z).

The depths of the moored and hanging transducers are affected by all processes that alter the volume of the lake, and the thickness of the surface ice, respectively (Figure 4b).

Depth of moored transducer (stage) = Glacial/river inflow + snowfall – sublimation – evaporation

Depth of hanging transducer (ablation) = Ice surface (snowfall – ablation) + ice bottom (melt – growth)

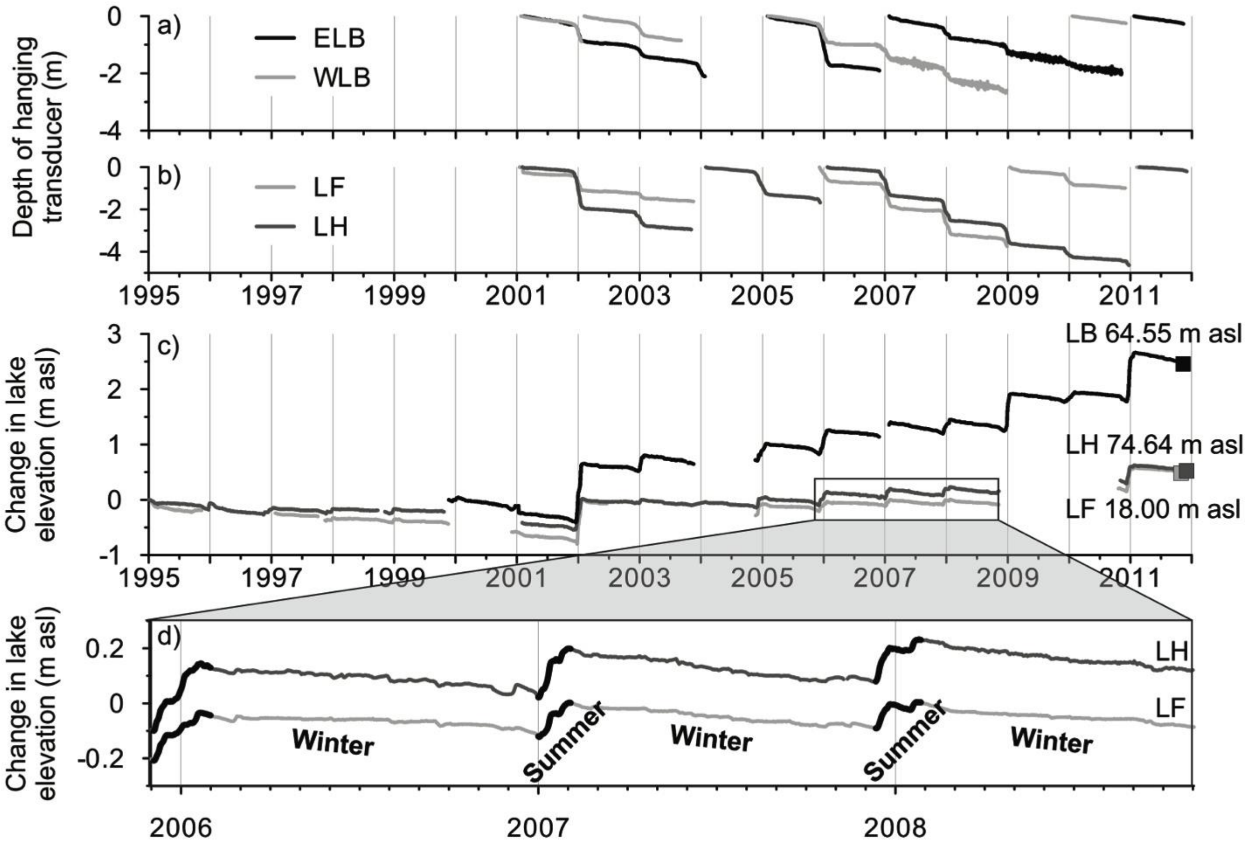

Figure 5. a,b) The relative change in depth of the hanging pressure transducer used to record surface ice ablation at Lake Fryxell (LF), Lake Hoare (LH), and Lake Bonney (LB). Ablation values are reset to a zero baseline following an extended sensor failure. The data becomes progressively noisier due to aging sensors that were replaced in 2010 and 2011. c) Lake levels based on water surface elevation. East Lobe and West Lobe Bonney are connected by an 11 m deep sill, and therefore register equivalent stage. Data is referenced against surveyed benchmarks and 2011 lake elevations are presented as square symbols. The slight dips in elevation directly prior to annual lake level rise represent increased summer sublimation that precedes river discharge. d) The inset depicts the demarcation of summer and winter seasons. Some of the small scale variation noticeable in the winter may be result from snow accumulation events.

When ice accretes onto the bottom of the ice cover during the winter, the ice cover is buoyantly displaced upwards by a factor proportional to the thickness of new ice multiplied by the density contrast between the ice and water. This artificially raises ablation values as the sensor is moved higher in the water column. To avoid erroneously high winter values, stage is used in lieu of ablation during the winter when calculating absolute rates of sublimation. The limited snowfall in Taylor Valley does add surface mass to the lake ice covers, but is extremely heterogeneous and tends to only accumulate in hollows on the lake surfaces. We do not have data to correlate the accumulation of snow at shoreline meteorological stations versus the middle of the lakes, and therefore do not include snowfall in our sublimation/ ablation calculations. Snowfall likely has a minor impact on the long term stage record at Lake Hoare and Lake Bonney due to mean precipitation rates of < 0.008 m w.e. yr^{-1} (Fountain and others, 2010). At Lake Fryxell, our calculations may underestimate sublimation/ablation by ~0.03 m yr^{-1} based on annual snow accumulation [*Fountain et al.*, 2010]. Lastly, melt at the bottom of the ice cover will increase the pressure reading of the hanging sensor as thinning ice is displaced downwards. In an Arctic lake, bottom melt has been shown to become significant only near the end of the melt season when the ice thins to <1 m [*Heron and Woo*, 1994]. Due to the thick ice cover on the Taylor Valley lakes, and limited heat transport by glacial/river inflow, we assume that bottom melt is negligible.

This paper presents ablation values, or the amount of ice that is lost from the surface of the lake ice throughout the year. In the winter, ablation is solely a result of sublimation, and therefore the two variables are equal. In the summer, ablation represents the sum of sublimation and any surface melt, which is assumed to flow into the lake. All sublimation/ablation values are presented in w.e. With simplifying assumptions and an ice to water density ratio of 0.91, our equations become:

During winter: Depth of moored transducer (stage) = -Sublimation

During summer: Depth of hanging transducer (ablation) = -Ablation / 0.91

To compare seasonal trends, annual records were partitioned as summer or winter based on stage data. Using the longest continuous data set, the onset of summer and winter was defined as the days of the year corresponding to the local minimum and maximum (turning points) in stage values at Lake Bonney (Figure 5c). For example, the summer of 2007-08 is demarcated as Dec 12th to Jan 27th, and the summer of 2008-09 as Dec 9th to Jan 30st (Figure 5d). From 2001-2010, the summer seasons range from 33 to 66 days in length (TABLE III). In two instances (2003-04, 2007), stage values were not available from Lake Bonney, and seasonal endpoints were based on data from Lake Hoare and Lake Fryxell. Daily sublimation rates from March to October were combined for all years and statistically compared between lakes using two-tailed t-tests for equal variance.

TABLE III. TOTAL SUMMER ABLATION AS CALCULATED BY A HANGING PRESSURE SENSOR INSTALLED IN THE ICE COVER, AND TOTAL WINTER SUBLIMATION AS CALCULATED BY THE DIFFERENCE IN MAXIMUM STAGE AT THE BEGINNING OF WINTER AND MINIMUM STAGE AT THE END OF WINTER.

	Summer Ablation (m)					**Winter Sublimation (m)**		
Summer (days)	**LF**	**LH**	**ELB**	**WLB**	*Winter (days)*	**LF**	**LH**	**LB**
2000-01 (n/a)					2001 (317)	0.13	0.12	0.17
2001-02 (52)	0.67	1.62	0.58	0.60	2002 (321)		0.07	0.14
2002-03 (38)	0.25	0.49	0.27	0.31	2003 (292)		0.10	0.16
2003-04 (66)			0.51		2004 (313)		0.09	
2004-05 (52)		0.93			2005 (315)	0.13	0.14	0.19
2005-06 (58)	0.70		1.41	0.68	2006 (335)	0.08	0.11	0.12
2006-07 (33)	0.81	0.78		0.37	2007 (312)	0.09	0.12	0.21
2007-08 (46)	1.10	0.95	0.38	0.38	2008 (312)	0.09	0.07	0.21
2008-09 (52)	0.51	0.99	0.27		2009 (306)			0.15
2009-10 (63)	0.54	0.45			2010 (309)			0.17
Mean (2001-2010)	**0.65**	**0.89**	**0.57**	**0.47**	**Mean (2001-2010)**	**0.10**	**0.10**	**0.17**

Mean annual ablation: LF 0.76 m, LH 0.99 m, ELB 0.74 m, WLB 0.64 m

Manual ablation stakes were installed on the east, center, and west sides of each lake in 1999. No more than two stakes existed at a given site concurrently. The stakes are vertically oriented PVC pipes with short 5 cm sections of plexiglass rod mounted horizontally at 1 m intervals along the length of the stake to anchor it in the ice cover. Manual measurements, recorded as the annual difference in stake height, represent a combined measurement of winter and summer ablation, and were not recorded on the same day every season. To facilitate comparison of manual measurements to automated data, the measurement interval was subdivided into the winter and summer seasons based on the previously defined dates. Summer ablation totals were calculated by applying winter ablation rates of 0.33, 0.33, and 0.53 mm d^{-1} for Lake Fryxell, Lake Hoare, and Lake Bonney, respectively, based on the low seasonal variability of the long term stage record (TABLE III). In an attempt to compare the manual ablation record over a ten year period, we calculated the z-scores at each stake in each given year. The standardized values for each stake were then averaged over the length of the dataset.

The long term stage record spans 1995-2011 for Lake Hoare and Lake Fryxell, and 2001-2011 for East Lobe Bonney. Automated ablation values are available from 2001-2011. Pressure data was recorded every minute and averaged at 20 minute intervals. In 2009, Campbell Scientific CR10X dataloggers were replaced with CR1000 dataloggers. Prior to 2010, ablation and stage were measured with Druck PDCR 1830 pressure transducers (accuracy ±0.06%) or Keller Series 173 pressure transducers (at Lake Hoare, accuracy ±0.1%). In 2010, pressure transducers on Lake Fryxell, Lake Hoare, and East Lobe Bonney were replaced with Campbell Scientific CS455 pressure transducers (accuracy ±0.1%). Stage at West Lobe Bonney is equivalent to that of East Lobe Bonney, and all pressure transducers are vented to the surface. In order to calibrate the automated stage measurements, lake levels were annually surveyed prior to, and near the end of, the summer melt, and are presented in meters above sea level (asl). Due to the discrepancy between GPS measurements recorded as geoid elevations and elevations optically surveyed from the ocean, an offset is applied to current GPS data to match historic optical surveys transects of Taylor Valley conducted by the New Zealand Antarctic Program in the 1960s [*Chinn*, 1993].

Meteorological data from 2010 are presented as an example of climate variability between the three lakes. Mean annual air temperature, wind speed, and incoming solar radiation are calculated from 15-min averages from meteorological stations which are located on the shorelines of Lake Fryxell, Lake Hoare, and East Lobe Bonney, and have been in operation since 1987, 1985 and 1993, respectively (Figure 2). A complete description of sensors and long term averages are available in Doran et al [2002b].

2.5 Results

Between Jan 1, 2001 and Jan 1, 2011, the levels of Lake Bonney, Lake Hoare, and Lake Fryxell, have risen 2.6 m, 1.0 m and 1.0 m, respectively (Figure 5). The majority of annual variation in the water balance of the Taylor Valley lakes occurs during the summer as a result of gain from river and glacial input and loss from sublimation. During the winter, mean sublimation values are relatively constant from 2000-2010 at 0.33±0.08 mm d^{-1} for Lake Fryxell, 0.33±0.08 mm d^{-1} for Lake Hoare, and 0.53±0.11 mm d^{-1} for Lake Bonney (± 1 s.d.). This amounts to 0.10 m, 0.10 m, and 0.17 m of average winter sublimation for the three lakes (TABLE III). Lake Bonney has a significantly higher daily sublimation rate from March to October than either Lake Hoare or Lake Fryxell (t-test, $p<0.01$, Figure 6a). Winter sublimation rates are compared to those modeled by water vapor mass flux equations at Lake Hoare, and plotted against mean wind speed (Figure 6b, Clow and others, 1988). Overall, the average March to October sublimation rate of 0.24 mm d^{-1} at Lake Hoare is lower than the modeled rate of 0.52 mm d^{-1} [*Clow et al.*, 1988].

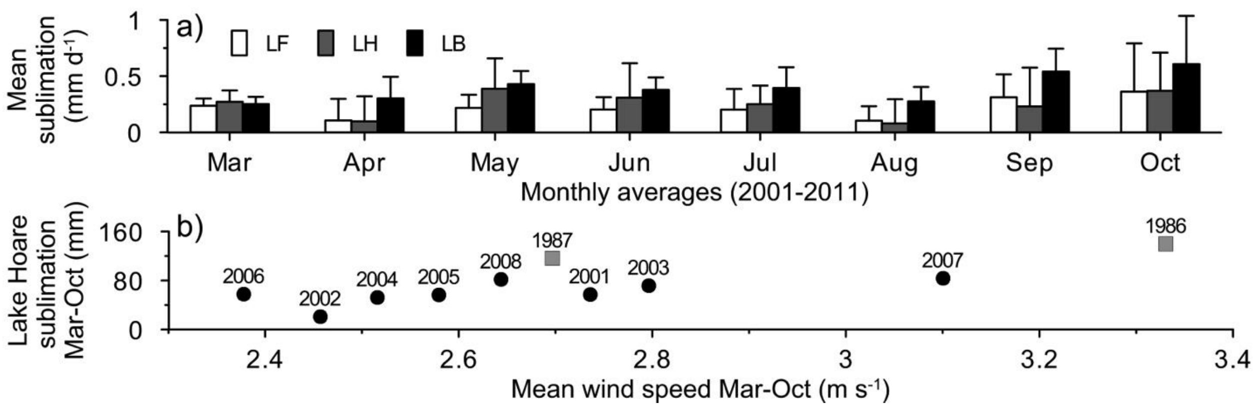

Figure 6. Sublimation from March to October at Lake Hoare plotted against mean wind speed. Grey squares represent modeled sublimation estimates at Lake Hoare from Clow and others (1988) for 1986 and 1987.

Summer ablation values are much more variable, and range from 0.25 to 1.62 m (5-31 mm d^{-1}) over an approximate two month period (TABLE III). In 2001/02 summer ablation at Lake Hoare was substantially higher than the other lakes, and likewise in 2005/06 at East Lobe Bonney. The lake stations are located adjacent to the central ablation stakes at Lake Fryxell and Lake Bonney, and adjacent to the east ablation stake at Lake Hoare (Figure 2). Manual ablation readings match or underestimate total summer ablation as recorded by the hanging pressure transducer, and in no year does the manual ablation reading markedly exceed the automated value (for example, see Figure 7a). We hypothesize that our underestimation results from stakes periodically slipping downward in the ice. Any downward movement of the hanging transducer would be clearly noticeable in the pressure data, and therefore we take the automated data as the more accurate of the two measurements. Due to errors and missing data in the manual record, only data from 2001/02 is plotted with absolute values of ablation (Figure 7a). Along the length of the valley, there is no significant spatial trend in summer ablation as recorded by the ablation stakes; nevertheless, it is important to note that on average, the highest summer ablation occurs at Lake Hoare, and on the west end of Lake Bonney which abuts Taylor Glacier (Figure 7b).

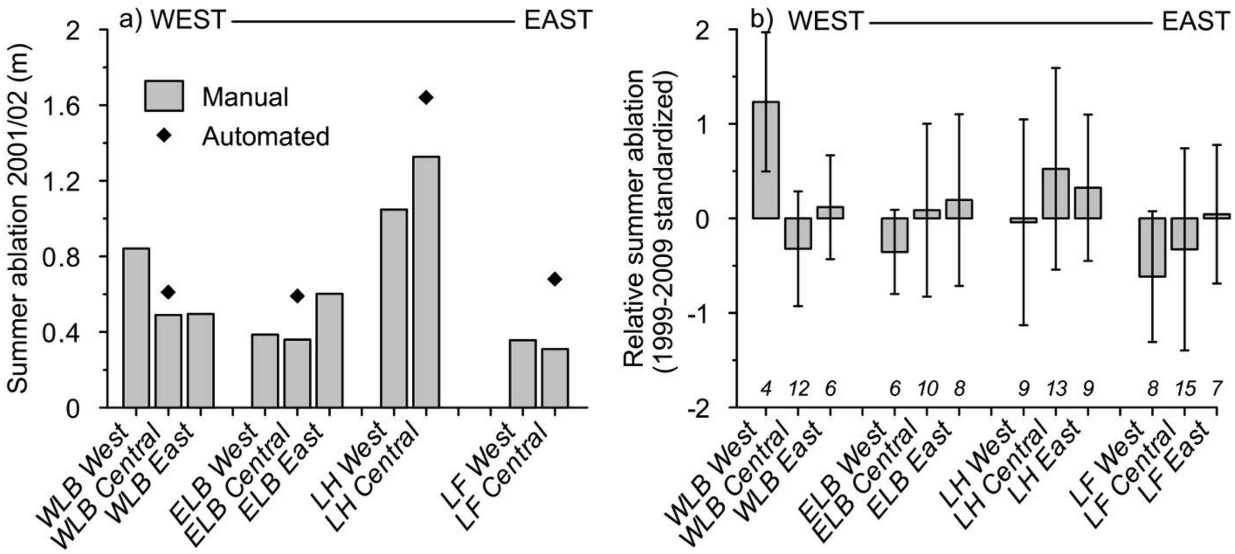

Figure 7. a) 2001-02 summer ablation recorded at individual ablation stakes in the western, central and eastern portions of the three lakes. Automated data collected at stations adjacent to 'central' stakes are plotted as black triangles. b) Spatial variation in summer ablation measurements at manual ablation stakes from 1999 to 2009. To standardize, Z-scores of summer ablation values were calculated for each stake in each given year, and then averaged over the length of the dataset. Error bars correspond to the standard deviation of all data at one location. Values along the x-axis signify the number of data points included in the average, as some years had two ablation stakes present, while other years are missing data.

2.6 Discussion

Sublimation rates can be modeled by the latent heat flux emanating from the ice cover into the atmosphere. Two critical variables in this calculation are wind speed, and the vapor pressure difference above and at the ice surface, which corresponds to air and ice surface temperature [*Brock and Arnold*, 2000; *Bliss et al.*, 2011]. In Taylor Valley, climate is significantly influenced by westerly winds that descend from Taylor Dome. This is especially evident between June and August, when high wind speeds result in average temperature increases of ~ 17^{o}C and lowering of relative humidity by 25% during katabatic events [*Nylen et al.*, 2004]. Consequently, it was expected that Lake Bonney, which is situated at the base of Taylor Glacier and experiences a higher mean annual wind speed (TABLE II) and frequency of winter winds > 4 m s^{-1} [*Doran et al.*, 2002b] than either Lake Hoare or Lake Fryxell, would have the highest rates of winter sublimation. Results confirm that winter sublimation averages 0.17 m at Lake Bonney and only 0.10 m at the other two lakes (TABLE III).

The major difficulty in elucidating the driver of summer ablation is predicting the effect of incoming solar radiation on the surface temperature of the ice cover, as the spatial heterogeneity of surface albedo unequally warms the ice surface. Patches of sediment deposited on the ice cover during windy periods warm the surface by absorbing a much higher proportion of solar radiation than that of a clean ice surface. Beach sand deposited on the lakes has an albedo of only 0.1-0.15 [*McKay et al.*, 1994; *Jepsen et al.*, 2010], which is much lower than that of smooth ice, measured at Lake Hoare and nearby Lake Vanda at 0.2-0.4, and that of angular late-summer ice, which can have an albedo of 0.6 [*Goldman et al.*, 1967; *McKay et al.*, 1994]. Elevated surface temperatures permit a larger vapor pressure difference and therefore, rate of ablation. Moreover, this process is a positive feedback. Variable ablation due to heterogeneous sediment deposition produces hollows in the ice cover which in turn can alter wind and temperature patterns at the ice surface and promote greater ablation [*Jepsen et al.*, 2010].

Our study found that total summer ablation, while extremely variable, was often the greatest at Lake Hoare. There is no singular climatological factor that accounts for this trend, unlike the winter season, when katabatic events strongly control the rate of sublimation. It is hypothesized that the high rates of summer ablation at Lake Hoare are a result of sediment loading on the ice cover, and not due to enhanced solar radiation (TABLE II). The Lake Hoare basin is the smallest and narrowest of the three basins, and is enclosed by the Seuss and Canada glaciers. Aeolian sedimentation input is highest at Lake Hoare [*Šabacká et al.*, 2011], and it may be that once sediment is deposited on the ice cover it is less mobile than at the other lakes where wide basin floors do not restrict near-surface winds. Likewise, Lake Bonney typically has the lowest summer ablation while averaging the highest number of degree days above freezing (TABLE II). It may be that high winds are able to effectively redistribute sediment and smooth the ice surface, which reduces the energy flux to the ice and lowers the surface temperature. On West Lobe Bonney, the higher rates of summer ablation on the western edge may originate from high winds descending immediately off the face of Taylor Glacier. Also, the western edge of the lake typically has a larger volume of surface sediment [*M. Obryk personal communication*], which would act to decrease albedo and warm the surface layer. Similarly, the western edge of East Lobe Bonney, which has lower rates of summer ablation, is sheltered in the lee of Bonney Riegel, a large bedrock feature that likely impedes westerly winds. Unfortunately, no estimates of percent debris cover exist for the three lakes. Field observations and satellite imagery of the lakes corroborate that Lake Hoare, in most years, has the roughest and most sediment-laden ice cover, whereas Lake Bonney has the flattest, cleanest ice [*P. Doran, personal communication*].

Near the terminus of Taylor Glacier, annual ablation (in w.e.) has been measured at 0.44 m [*Robinson*, 1984], 0.34 m [*Fountain et al.*, 2006], 0.18 m [*Hoffman et al.*, 2008] and 0.20 m [*Bliss et al.*, 2011]. Winter and summer ablation rates were found to be 0.17 mm d^{-1} and 1.09 mm d^{-1}, respectively, using an estimate of latent heat flux based on meteorological variables [*Bliss et al.*, 2011]. These values are lower than those measured at Lake Bonney, which is expected as the lake is situated 267 m lower in elevation and averages more positive degree days [1993-2000 avg. = 34.3, *Doran et al.*, 2002b] than those recorded at the meteorological station on Taylor Glacier [1995-2006 avg. = 4.7, *Hoffman et al.*, 2008].

In an environment where temperatures exceed 0^{o}C for less than two months, atypical meteorological conditions can lead to extreme hydrological responses [*Barrett et al.*, 2008; *Doran et al.*, 2008b]. The highest summer ablation of 1.62 m was recorded during the anomalously warm summer of 2001/02 [*Doran et al.*, 2008b]; yet, only at Lake Hoare was ablation markedly high. The annual ablation values recorded from 2001-2010 are in general much higher than the commonly quoted value of 0.3-0.4 m yr^{-1} for Taylor Valley, which was derived from physical measurements of the ice ablation at Lake Fryxell [*Henderson et al.*, 1966], and from a sublimation model based on the climate record at Lake Hoare in 1986 and 1987 [*Clow et al.*, 1988]. Our annual ablation values only approach 0.3 m yr^{-1} in the lowest years, and never at Lake Hoare. Our data better corresponds to the 1.5 m yr^{-1} documented at Lake Hoare based on the migration time of five corks released below the ice cover and allowed to ablate to the surface [*Simmons et al.*, 1986].

Ablation values are often incorporated into ice thickness models [*McKay et al.*, 1985], and used to determine the residence times of water and gases in the Dry Valley lakes [*Wharton Jr. et al.*, 1989]. In Taylor Valley, where ice thicknesses range from 3-6 m, our data set proves that a 3 m ice cover would turnover in three to five years if average ablation is quoted between 0.6-1.0 m yr^{-1}. This is twice the previously assumed rate [*Clow et al.*, 1988], and has important ramifications for sediment dynamics in the ice cover, and gas and solute exclusion processes. In the permanent ice covers, a microbial ecosystem exists in association with sediment particles [*Fritsen et al.*, 1998; *Paerl and Priscu*, 1998; *Priscu et al.*, 1998, 2005]. In Lake Bonney and Lake Hoare, a dominant peak in sediment concentration and chlorophyll *a* exists at 2 m depth, and ranges from 0.5-3 m across the Dry Valley lakes, as sediment migrates downward in the ice cover [*Fritsen and Priscu*, 1998; *Priscu et al.*, 1998]. Modeled cycles of sediment movement show a decrease in maximum depth and an increase in annual depth variance of the sediment peaks with increased ablation [*Jepsen et al.*, 2010]. This repetitive cycle of upward ablation of the ice cover and descent through ice melt may promote the coalescing of sediment into larger pockets and enhance melt migration. Higher ablation rates would also more quickly replenish surface ice that has undergone morphological changes, such as 'ice-whitening', which limits the transmission of photosynthetically active radiation (PAR) through the ice cover [*Fritsen and Priscu*, 1998].

2.7 Conclusions

To our knowledge, this dataset is the first long term ablation record from lake ice. In the past, empirical measurements were limited to annual or summer values. We find that summer ablation is the major driver of ice loss and can vary from 0.25 m to 1.62 m in warm years, whereas winter sublimation accounts for only 0.07-0.21 m of ice loss. Along Taylor Valley, annual ablation is highest at Lake Hoare, which we hypothesize is a result of sediment loading on the ice cover. Sublimation is the main mechanism of water loss in closed basin, perennially ice covered lakes. With accurate ablation data we are one step closer towards closing the water balance of the Dry Valley lakes and understanding the climatic variables that control water routing. The next objective will be to use the high resolution pressure data to validate summer sublimation models, such as that used by Clow and others [1988], in order to calculate total water loss from the lakes. Ultimately, sublimation values will be input into a whole-lake water balance to establish the direct glacial discharge entering the lakes, which is currently unknown.

2.8 Acknowledgements

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3: 27 m OF LAKE ICE ON AN ANTARCTIC LAKE REVEALS PAST HYDROLOGIC VARIABILITY

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3.1 Abstract

Lake Vida, located in Victoria Valley, is one of the largest lakes in the McMurdo Dry Valleys. Unlike other lakes in the region, the surface ice extends at least 27 m, which has created an extreme and unique habitat by isolating a liquid-brine with salinity of 195 g L^{-1}. Below 21 m, the ice is marked by well-sorted sand layers up to 20 cm thick, within a matrix of salty ice. From ice chemistry, isotopic abundances of ^{18}O and ^{2}H, ground penetrating radar

profiles, and mineralogy, we conclude that the entire 27 m of ice formed from surface runoff, and the sediment layers represent the accumulation of fluvial and aeolian deposits. Radiocarbon and optically stimulated luminescence dating limit the maximum age of the lower ice to 6300 ^{14}C yr BP. As the ice cover ablated downwards during periods of low surface inflow, progressive accumulation of sediment layers insulated and preserved the ice and brine beneath, analogous to the processes that preserve shallow ground ice. The repetition of these sediment layers reveals climatic variability in Victoria Valley during the mid to late Holocene. Lake Vida is an excellent Mars analogue for understanding the preservation of subsurface brine, ice and sediment in a cold desert environment.

3.2 Introduction

Little is known about the habitability of cold liquid environments sealed off from the atmosphere, such as the subglacial lakes of Antarctica [*Siegert et al.*, 2013], or beneath the icy shell of Europa [*McKay*, 2011]. Located in the McMurdo Dry Valleys of Antarctica, Lake Vida has the thickest ice of any subaerial lake on Earth and is one of the few “ice-sealed” ecosystems known to support a diverse and active microbial population in a cold, anoxic, aphotic brine [*Murray et al.*, 2012]. The brine was first discovered 16 m below the surface of Lake Vida, and is hypothesized to have been sealed from the atmosphere for several millennia [*Doran et al.*, 2003]. At present, the lake level (i.e. the ice surface) of Lake Vida is rising, which implies that the brine is progressively getting farther from the surface. In this paper, we present evidence for how and when Lake Vida formed to further understand the structure and evolution of the existing brine system beneath Lake Vida.

On most lakes in the Dry Valleys, the thickness of ice ranges from 3-6 m [*Wharton Jr. et al.*, 1992; *J. Priscu unpublished data*]. This thickness is maintained by energy loss at the surface (conduction and ablation) and energy gained at the bottom of the ice cover (freezing) [*McKay et al.*, 1985]. Constant ablation at the ice surface, and freezing at the ice bottom of these floating ice covers limits the maximum age of the surface ice to ~ 5 years [*Dugan et al.*, 2013]. However, the Lake Vida ice cover is at least partially grounded [*Doran et al.*, 2003], so the ice does not turn over in the same way. Water that is flowing to the lake is trapped on the surface of the ice where it freezes and is later ablated or buried by subsequent ice buildup. In this way, the thick ice on Lake Vida may record past hydrological changes, similar to a glacier; but, unlike a glacier, intermittent accumulation may lead to large discontinuities in the ice cover during prolonged cold/dry periods.

In Victoria Valley, the Lake Vida basin was occupied by a 200 m deep glacial lake 8600 ^{14}C yr BP [*Hall et al.*, 2002], after which lake levels began to decline. It is unlikely that any of the observed ice existed during this time, which implies that the entire 27 m of ice on Lake Vida was formed during the mid- to late Holocene. It is also likely that the formation of Lake Vida was influenced by events similar to the repeated lake level drawdowns and complete desiccation events recorded in lacustrine sediment cores and geochemical diffusion profiles of the large lakes of Taylor and Wright Valleys. For instance, it is speculated that Taylor Valley underwent a valley wide desiccation event at 1000-1200 yr BP [*Lyons et al.*, 1998a]; Lake Fryxell had low-stands at 6400, 4700, 3800 and around 1600 yr BP [*Wagner et al.*, 2006; *Whittaker et al.*, 2008]; and, Lake Bonney was lower than present at 1500, 800, and 400 yr BP [*Croall*, 2005]. Lake Vanda, in Wright Valley, underwent a low-stand at 1200 yr BP [*Wilson*, 1964], or prior to 2000 yr BP [*Gumbley et al.*, 1974]. Some or all of these events may be recorded in the Lake Vida ice cover.

The overall aim of this study was to reconstruct the history of the ice cover on Lake Vida. We examined the isotopic and chemical composition, texture, mineral characteristics, and diatom composition of a 27 m ice core, as well as used ground penetrating radar (GPR) profiles to map the strata observed in single cores. In order to determine the provenance of sediment onto the lake, we compared mineralogy and granulometry of the sediment layers with those of sand samples collected along the floor and stream beds of Victoria Valley. Both radiocarbon dating and optically stimulated luminescence (OSL) dating were employed to establish the time of deposition of sediment layers, and ice core stratigraphy was used as a means of establishing periods of lake level drawdown.

3.3 Study Site

Lake Vida (77°23'S, 161°56'E), situated in Victoria Valley, Antarctica, is one of the largest (6.8 km^2) and highest (340 m above sea level) lakes in the McMurdo Dry Valleys (Figure 2). The lake is endorheic (closed basin), and receives inflow via streams originating from Victoria Upper, Victoria Lower, and Clark Glaciers. Lake Vida occupies a unique climatological niche where summer temperatures can rise slightly above 0°C to generate stream flow, yet unusually cold winters (compared to the other major valleys in the region) maintain a thick ice cover on the lake. From 1995 to 2000, the mean annual air temperature at Lake Vida (-27.4°C) was 7°C to 10°C lower than valley bottom temperatures in Taylor and Wright Valleys, but mean summer temperatures were similar [*Doran et al.*, 2002b].

From the drilling expedition in 2010, it is known that the ice on Lake Vida extends to at least 27 m [*Murray et al.*, 2012]. A unique feature of Lake Vida is the presence of liquid brine within the ice cover, which infiltrates the drill-hole at approximately 16 m and rises to 10.5 m below the surface. The brine is anoxic, with salinity of 195 g L^{-1} and temperature of -13.4°C [*Murray et al.*, 2012]. It is hypothesized that the brine is contained within small fractures or channels in the ice, and rises to 10.5 m when the confining layer of freshwater ice in the upper 16 m is breached.

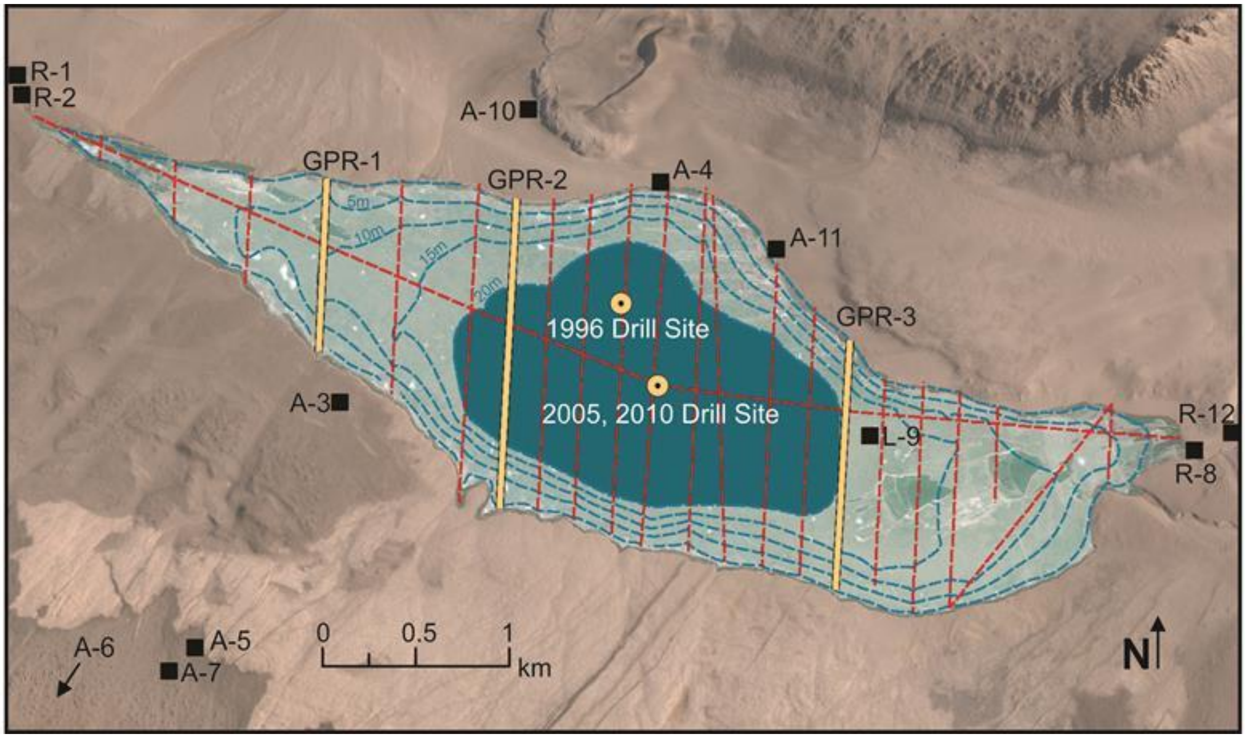

Figure 8. Location of drill sites and GPR transects (red dashed line) on Lake Vida in central Victoria Valley ©DigitalGlobe, Inc. (2011). Highlighted GPR transects 1-3 are presented in the manuscript. Bathymetric lines 0-20 m were digitized and interpolated from GPR profiles. The dark blue area below 20 m is of unknown depth. Locations of catchment sediment samples are labeled based on source (R = riverine, A = aeolian, L = lake surface).

3.4 Methods

An electric 15 cm diameter SideWinder drill [*Kyne and McConnell*, 2007] was used to retrieve two ice cores, one 27 m and the other 20 m long, located 6 m apart in the center of Lake Vida in November 2010. The 27 m ice core was split in half for archival purposes, and subsampled into 5 cm lengths. Where recovery was incomplete for the 27 m core (between 16 and 20 m), the 20 m ice core was subsampled. Longitudinal thick-sections (approximately 0.5 cm thick) were cut from the ice core face and viewed under cross-polarized light for ice crystal fabric analyses. Subsamples were washed with deionized MilliQ water to remove possible brine contamination and allowed to completely melt for processing on a Dionex 1500 ion chromatograph for major ion analysis, and a Los Gatos Research Liquid Water Isotope analyzer for isotopic abundances of ^2H and ^{18}O. Salinity is reported as the sum of concentration of total ions. Isotopic values are reported with respect to the VSMOW international standard. Deterium excess is calculated as $\delta^2H - (8 \times \delta^{18}O)$.

Sediment layers in the 27 m ice core were subsampled in duplicate 1 cm segments, which were freeze-dried or allowed to melt in order to extract pore water by centrifugation. For grain size analyses, 2 g samples were sieved through a 1000 µm sieve, and pretreated with 30% H_2O_2 for 18 hours in a 50°C water bath. Following pretreatment, samples were shaken following the addition of 1 mL of 30 mg L^{-1} Graham's salt ($Na_4P_2O_7$) as a dispersant, and analyzed on a Micromeritics Saturn Digisizer 5200 particle size analyzer (detection limit 0.1-1000 µm). Sand/silt classifications are based on the Udden-Wenworth scale [*Wentworth*, 1922]. Selected samples were treated with 10% HCl and photographed under a CamScan CS 44 scanning electron microscope (SEM). Sediment layers were also subsampled to evaluate diatom assemblages and absolute abundance (valves/g dry weight) via light microscopy. Preparation methods followed standard techniques [*Scherer*, 1994]. A known mass of freeze-dried sediment was reacted with 10% HCl and 10% H_2O_2 to remove carbonates and organics. Abundance per gram was extrapolated from diatom counts on coverslips in a beaker of known area.

Freeze dried samples were analyzed for total carbon (TC) and total inorganic carbon (TIC) with an elemental analyzer (Dimatec Co.). Total organic carbon (TOC) was calculated from the difference in TC and TIC. Based on carbon content, six samples were chosen for radiocarbon dating of the organic fraction. Samples were prepared by separating the humic and fulvic acid fractions by acid-alkali-acid extraction prior to graphite conversion in an automated graphitization system [*Grootes et al.*, 2004]. This required an overnight treatment with 1% HCl to remove carbonates, followed by 4-hour humic acid extraction with 1% NaOH, and a secondary overnight treatment of 1% HCl to eliminate any CO_2 that may have been absorbed during the NaOH treatment [*Grootes et al.*, 2004]. Two samples at 23.90 and 26.43 m with high TIC were selected for radiocarbon dating of carbonates and were treated with phosphoric acid. Two 5 cm long sediment sections at 21.51 and 25.54 m were cut by band saw under red-light for optically stimulated luminescence (OSL) dating (UIC Luminescence Dating Research Laboratory), and remained frozen until analysis. The advantage of frozen sediment for OSL dating is that the water content is expected to have remained constant since freezing took place [*Demuro et al.*, 2008; *Arnold and Roberts*, 2011]. OSL dating of quartz grains from sediment layers was performed using single aliquot regeneration protocols [*Murray and Wintle*, 2003].

Catchment samples were obtained from twelve areas surrounding Lake Vida (Figure 8) to examine sediment provenance. Four samples were fluvial bedload deposits from river channels (streams were not flowing at the time), seven were from aeolian sources outside of stream channels, and one was collected on the ice surface. Grain size was evaluated on a Malvern particle size analyzer. Mineralogy of sediments in all of the catchment and nine lake samples was identified via binocular optical microscopy. Approximately 300-800 grains were counted per sample.

To confirm the continuity of horizons and sediment layers noted in the ice cores, 55 km of GPR transects were recorded over the surface of Lake Vida in 2010 (Figure 8), using a GSSI

SIR-3000 acquisition unit equipped with a 400 MHz antenna. Transects were recorded at 400 ns time range and 2048 16-bit samples per trace, with five manual gain points at -20, 0, 25, 30, and 50. A dielectric constant of 3.15 was initially chosen for depth calibration, but this was altered based on known characteristics of the ice cores. In post-processing, radar profiles were triple stacked and passed through a 200 and 500 MHz triangle finite impulse response (FIR) filter to remove high and low frequency noise.

Lake levels were annually surveyed from benchmarks tied into historical optical survey transects conducted by the New Zealand Antarctic Program and were recorded in meters above sea level (m asl). Meteorological data were obtained from a meteorological station at the western edge of Lake Vida operated by the McMurdo Long Term Ecological Research site (*mcmlter.org*).

3.5 Results

A water column of brine was not encountered 20 m below the surface of Lake Vida as previously hypothesized [*Doran et al.*, 2003]. Rather, wet ice and sediment continued below this depth. After four thick (> 10 cm) sediment layers were encountered below 21 m, the drill became lodged in what was almost certainly a sediment-rich layer (based on the slow progress of drilling) at 27.01 m. The last sample obtained was an ice layer from 26.62 to 26.81 m.

There was an overall trend of increasing salinity in the ice core with depth, as well as a shift at 23 m from Cl^- to SO_4^{2-} as the dominant anion and an accompanying decrease in Mg^{2+} (Figure 9). In the upper 15 m, salinity was less than 0.1 g L^{-1}, with the exception of one sample at 12.69-12.81 m that had salinity of 1.5 g L^{-1}. From 15 to 20 m, salinity increased, but remained less than 4 g L^{-1}. Below 20 m, the ice salinity was variable from less than 1 g L^{-1} to 34 g L^{-1}.

Although brine inundated the borehole up to 10.5 m depth, and all of the cores below 16 m depth were in contact with the brine while being extracted, it appears from the ratios of major anions to cations (Figure 9), that the ice was not substantially contaminated by brine in the hole. The pH of the brine was 6.2 [*Murray et al.*, 2012] with an ionic strength of 4.4 M calculated from FREZCHEM [*Marion et al.*, 2010].

The fabric of the ice changed with depth, from large individual ice crystals with c-axes oriented upwards, which is typical of freshwater ice, to ice composed of randomly oriented small crystals that appear to have recrystallized over time (Figure 10a).

Throughout the cores, there are many small pockets of sediments and thin sediment layers (Figure 10b, 11). All sediment samples from the ice core were predominately sand (grain size: 62.5 to 2000 μm), with only 4 out of 79 samples having a mean grain size <100 μm or a percentage of silt and clay (<62.5 μm) > 6% of the total volume. In the 27 m core, the four thick sediment layers below 21 m will be referred to as SL21.62, SL22.88, SL25.59, and SL26.28 (Figure 9, 10). These layers had water contents < 10 % and total thicknesses of 19, 15, 11, and 19 cm, respectively. At the base of SL26.28, mean grain size began to decrease with a concomitant increase in the total carbon as percentage of mass (Figure 12). Microscopy of SL26.28 revealed abundant diatom frustules of the genera *Pinnulaira* and *Luticola.*

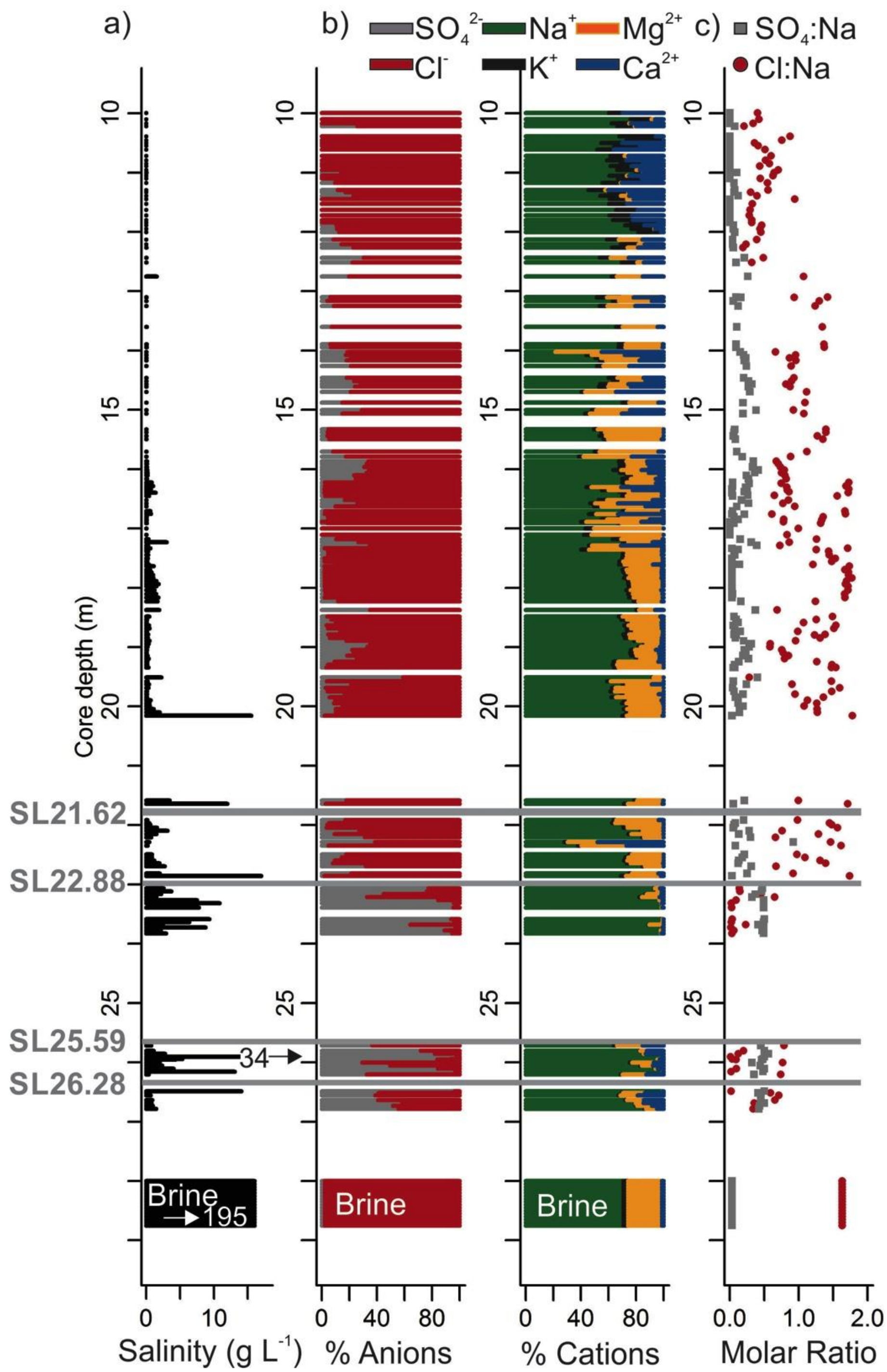

Figure 9. a) Salinity of ice samples below 10 m sub-sampled from the Lake Vida ice cores. Salinities ranged from < 1 to 34 g L^{-1}. Brine salinity is 195 g L^{-1}. SL21.62, SL22.88, SL25.59, and SL26.28 represent sediment layers > 10 cm thick, present in the 27 m ice core. b) Total molar percentage of major anions and cations present below 10 m in the Lake Vida ice cores. The lengths of the colored bars represent the percentage of the given ion relative to the other anion or cations present in the ice or brine. c) Ratios of Cl^{-1}:Na^{+} to SO_4^{2-}:Na^{+} (mol:mol).

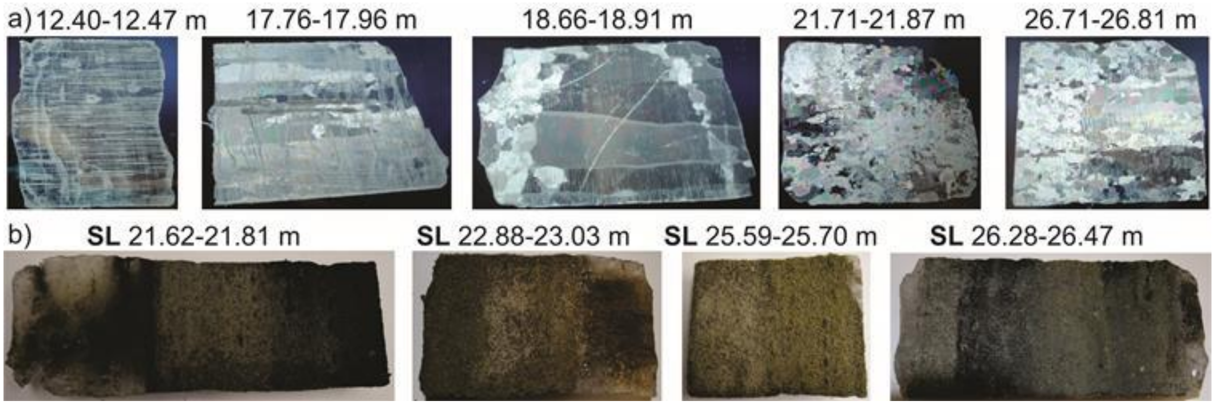

Figure 10. a) Images of thick sections of Lake Vida ice between two sheets of polarized film. At 12.40-12.47 m, the entire section is a single ice crystal. With depth, the average grain size of ice crystals decreases to less than 1 cm. b) Photographs of sediment sections SL21.62, SL22.88, SL25.59, and SL26.28.

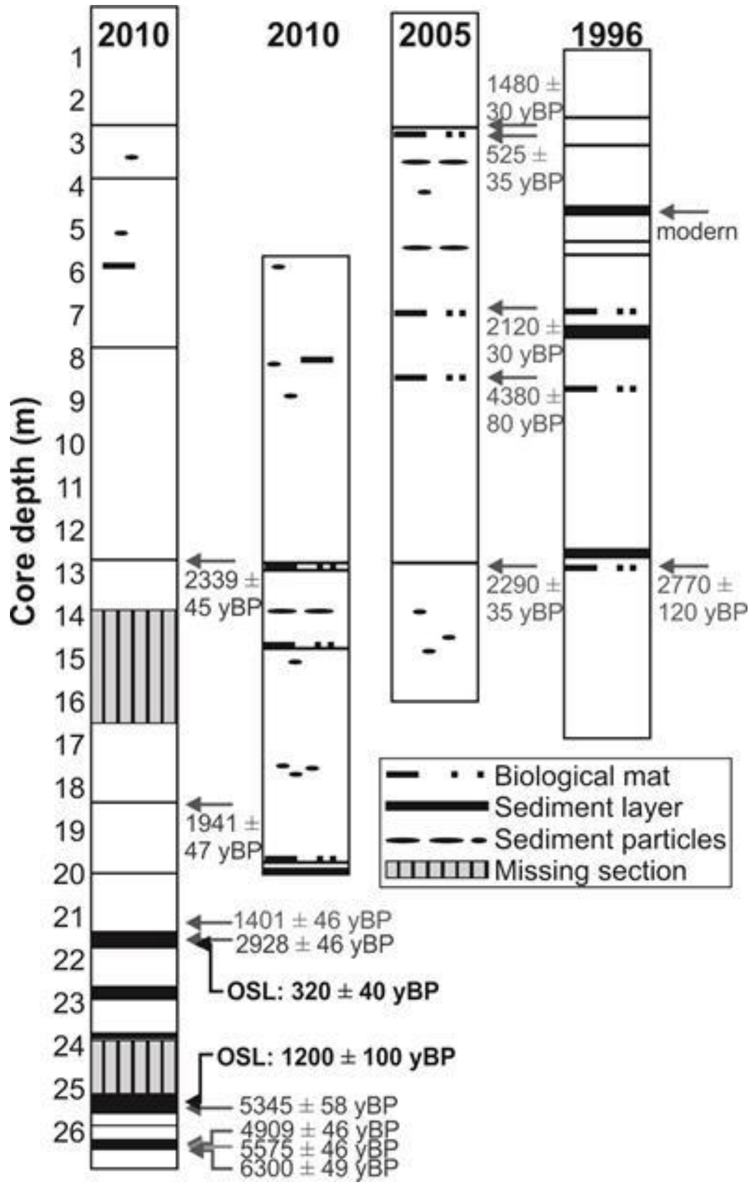

Figure 11. Lake Vida ice core logs from 1996, 2005 and 2010. Two ice cores were drilled in 2010. For core locations in 1996 and 2005, see Doran et al (2003) and Murray et al (2012). Surface heights are adjusted to the 2010 lake elevation.

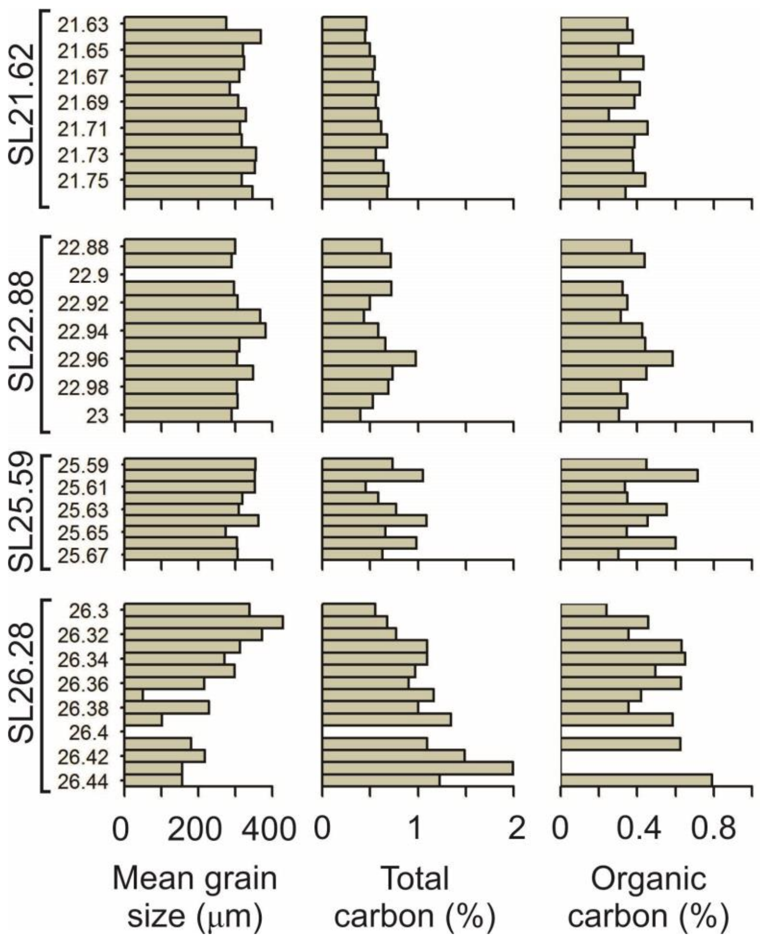

Figure 12. Mean grain size (μm), percentage of total carbon by mass (%), and percentage of organic carbon by mass (%) in sediment sections removed from the 27 m ice core from Lake Vida.

There was a higher percentage of quartz and plagioclase feldspar in the ice core sediment layers than in Victoria Valley samples, but overall the dominant minerals and rock fragments were similar in all samples (Figure 13a) The grain size distribution of the ice core sediments was more aligned with stream bed sediments than aeolian deposits (Figure 13b). In comparison, the sole sediment sample collected on the surface of the lake was coarser than ice core sediments.

SEM images from the lower sediment layers show no differentiation in microtextures between layers. Grains surfaces, especially those of quartz, ranged from well-rounded to heavily abraded and showed evidence of glacial, aeolian, and fluvial transport (Figure 14). Deeply imbedded features, such as conchoidal fractures and multiple striations may suggest glacial transport (Figure 14a,b,e); v-shaped percussion cracks are indicative of fluvial transport (Figure 14c,d), and rounded grains with surface craters due to collisions may result from aeolian transport (Figure 14f,g,h) [*Mahaney*, 2011].

The upper sediment layers noted in two ice cores retrieved in 2010 correspond to those cored in 1996 [*Doran et al.*, 2003] and 2005 [*Taylor*, 2009] (Figure 11). The radiocarbon dates vary significantly, especially in the upper ice where there is no correlation between age and depth. The dates from SL26.28 do show increasing age with depth (4909 ± 46 to 6300 ± 49 ^{14}C yr BP) and provide a maximum age of the lower ice. A portion of SL21.62 and SL25.59 showed no evidence of exposure to sunlight during collection, and returned OSL dates of 320 ± 40 and 1200 ± 100 yr BP, respectively; dates that are younger than the radiocarbon ages from the respective horizons. It is noted that ^{13}C values are unreliable and not reported due to high fractionation during AMS measurement.

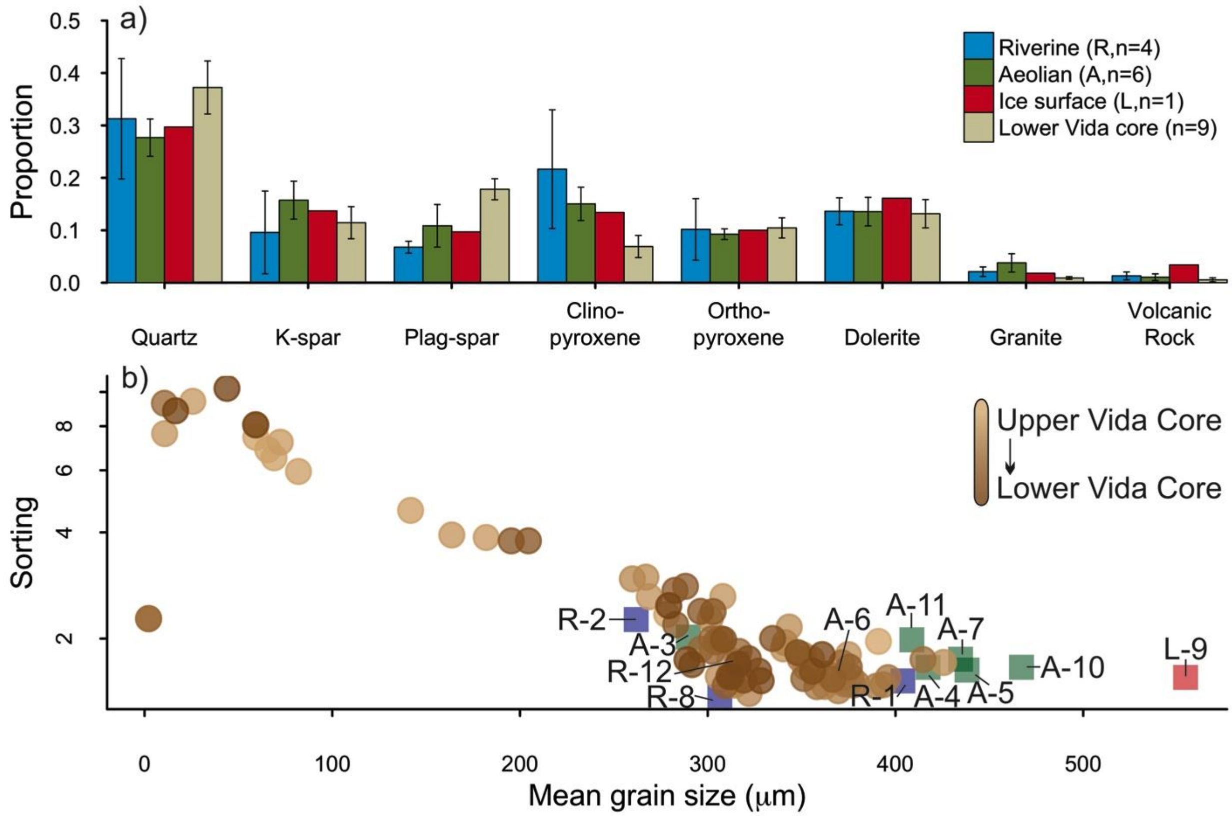

Figure 13. a) Proportion of minerals and rock fragments in sediment samples from Victoria Valley and the Lake Vida ice core from point counting. Fluvial samples were collected within the stream channels and likely represent bedload. Aeolian samples were outside the channels and the ice surface sample was sampled near the middle of the lake (Figure 8). The ice core samples were obtained from sediment layers within the ice core. b) Mean grain size and sorting of the samples represented in a). The darker shading of the brown circles represents a lower depth in the Vida ice core.

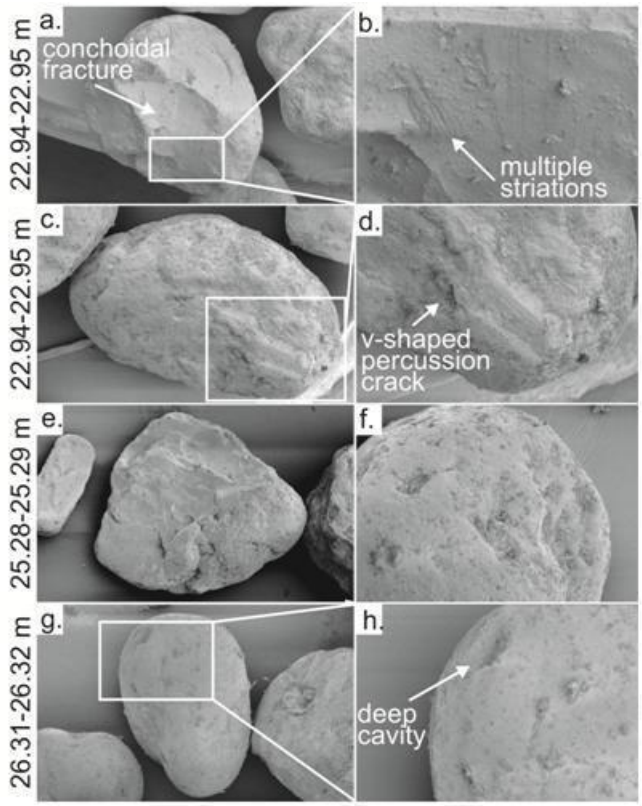

Figure 14. SEM images of select grains from sediment layers in Lake Vida. a,b) Grain shows a large conchoidal fracture and multiple striations. c,d) Sub-rounded grain with abraded surface. Micro-features include a v-shaped percussion crack. e) Grain with high relief features indicative of glacial transport. f) Rounded grain with multiple collision features. g,h) Rounded grain with several deep cavities.

The $\delta^{18}O$ and $\delta^{2}H$ values of both the ice and sediment pore water fell below the local meteoric water line (LMWL) of glacial water [*Gooseff et al.*, 2006] (Figure 15a), but on a slope consistent with the sublimation of ice [*Sokratov and Golubev*, 2009; *Hagedorn et al.*, 2010; *Lacelle et al.*, 2011, 2013] and isotopic values reported around the Dry Valleys [*Gooseff et al.*, 2006; *Harris et al.*, 2007]. Most of the ice and sediment pore water samples have $\delta^{18}O$ between -25‰ and -32‰. There are, however, two notable exceptions. Firstly, the sample at

12.69-12.81 has $\delta^{18}O$ of -34.6‰, which is substantially depleted in comparison to the nearby ice. Secondly, the $\delta^{18}O$ of pore water from the four thick sediment layers below 21 m ranges from -22‰ to -25‰, which is relatively enriched when compared to the rest of the ice core (Figure 15a). The median deuterium excess of the sediment layers is -22.7‰. This is lower than the median deuterium excess of all other groups, which range from -4.2‰ to -11.5‰.

GPR profiles distinctly map the edge of the lake basin until an impenetrable basal reflector around 21 m (Figure 16). Along the basin edges there are features resembling ancient terraces, especially around 8 m (Figure 16). From 8 to 12 m, synchronous wavy reflectors are spaced approximately 1 m apart. In all transects, ice and sediment layers appear to be continuous across the lake.

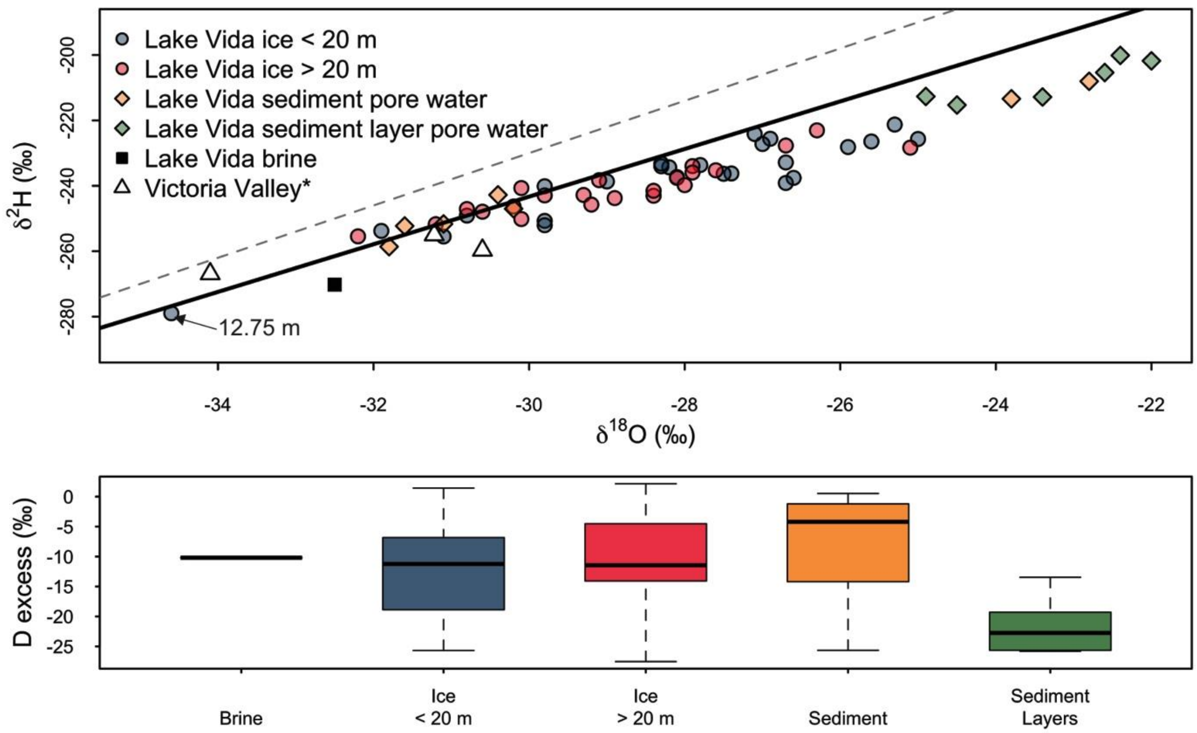

*Figure 15. a) Stable isotope concentrations for deuterium (δ^2H) and δ $\delta^{18}O$ in Lake Vida ice, sediment pore water, and brine. Published values for Victoria Valley surface water and snow are denoted by black triangles (*Hagedorn et al, 2010). The local meteoric water line of glacial water (solid line [Gooseff et al. 2006]) and global meteoric water line (dashed line) are plotted for reference. b) Box and whisker plot of deuterium excess values for brine (n=1), ice samples above 20 m (n=25), ice samples below 20 m (n=26), thin sediment pockets (n=7) and sediment layers (n=6). The thick black line represents the median values, and box edges represent the 25th and 75th quantiles. Whisker lines extend to the extreme values.*

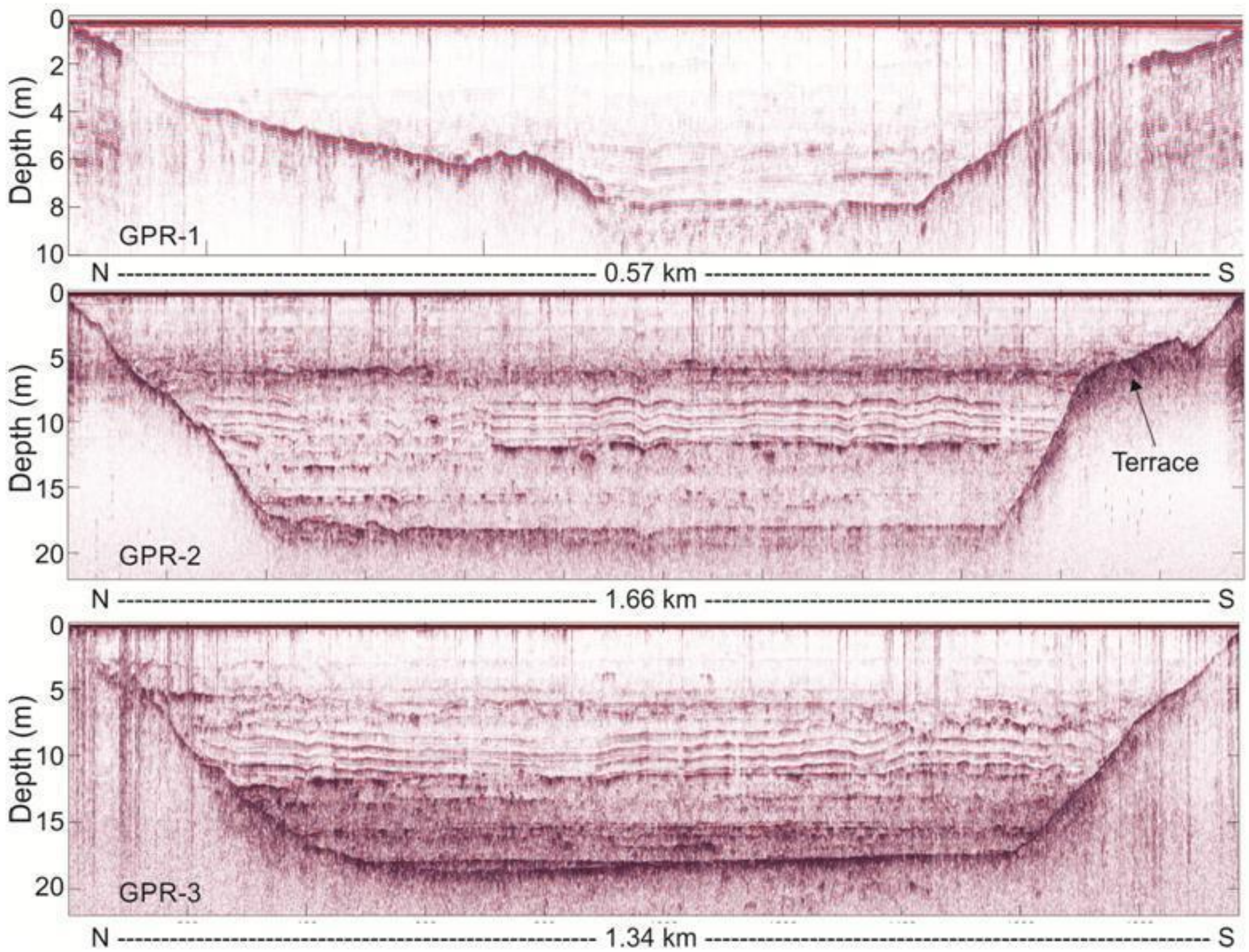

Figure 16. GPR transects recorded north to south across the surface of Lake Vida (Figure 8).

3.6 Discussion

The low ion concentrations, absence of large sediment layers and clear GPR returns in the top 8 m of ice suggest that the upper ice has formed recently under a positive water balance. The level of Lake Vida has risen 3.5 m in the last 40 years, and has a hydrologic history similar to Lake Bonney (Figure 17), which has been documented to have risen ~16 m from 1903 to 2010 [*Chinn*, 1993; *Bomblies et al.*, 2001; *P. Doran unpublished data*]. If a linear extrapolation is applied to the Vida record based on the correlation of volumetric change to Lake Bonney, the surface of Lake Vida would have risen approximately 7.7 m from 1903 to 2010 when our cores were collected.

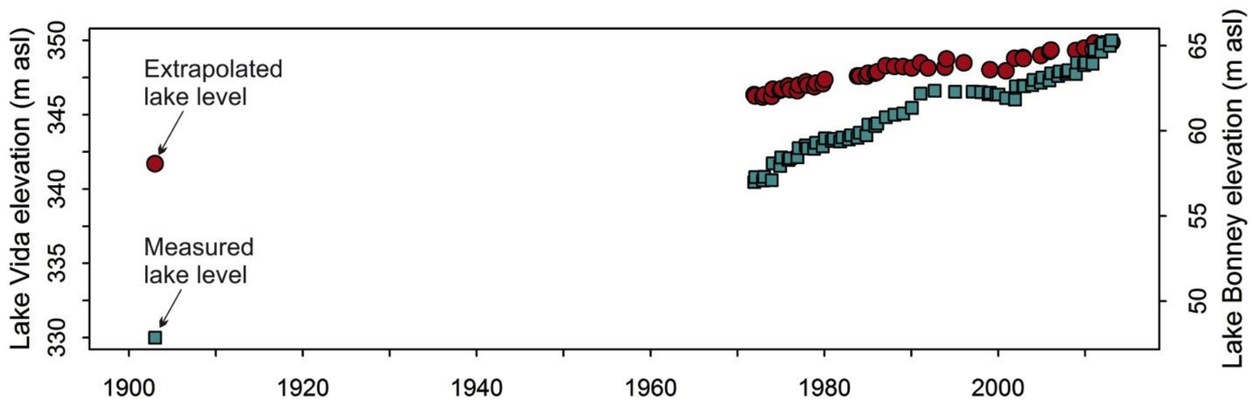

Figure 17. Lake Vida and Lake Bonney surface elevations. The 1903 Lake Bonney elevation is inferred from a measurement at Lake Bonney narrows by Robert Falcon Scott (Chinn, 1993). The 1903 Lake Vida elevation is extrapolated from the correlation between the two lakes for 1971-2010.

Therefore, the 8 m contour in Figure 8 may be an approximate representation of the lake shore in 1903. This is an indication of the rapidity with which the level of Lake Vida can change over time.

The ice between 9 and 13 m contains almost no sediment. In addition, the ice sample at 12.75 m (average depth of 12.69-12.81 m) has a salinity of 3.5 g L^{-1}, and is heavily depleted in ^{18}O and ^{2}H versus all other ice samples (Figure 15). Both an increase in salinity and the depletion of heavy isotopes are a signature of freshwater freezing. During freezing at ice-water interface, equilibrium isotope fractionation preferentially retains heavy isotopes in the ice, which leads to a depletion in heavy isotopes in the unfrozen water [*Gallagher et al.*, 1989; *Horita*, 2009]. The isotopic value for the fraction of water that remains unfrozen during the freezing of ice downward (δ_f) can be approximated by Rayleigh distillation as shown in equation (1), where α is the fractionation factor, f is the fraction of water that remains unfrozen, δ_o is the original isotopic value of the water [*Miller and Aiken*, 1996; *Kendall and McDonnell*, 1998].

$$\delta_f - \delta_o \cong 10^3 (\alpha - 1) \cdot \ln f \qquad \text{eq. (1)}$$

As a simple model for the depletion of stable isotopes during freezing, we employ equation (1) with $\delta_o{}^{18}O$ = -25.0‰ (isotopic value of ice at 10.5 m) and α = 1.0029 [average value from *Horita*, 2009] to test if the high salinity and low isotopic values at 12.75 m could have resulted from the downward freezing of 3 to 4 m of ice above. With these parameters and f = 0.04, $\delta^{18}O$ = -34.3‰. This fits the observed $\delta^{18}O$ at 12.75 m of -34.6 ‰, and along with the high salinity and lack of sediment particles, indicates a high likelihood of downward freezing in this section of the ice core. Unfortunately, we do not have high-resolution sampling at this depth, which prevents a detailed analysis of Rayleigh fractionation in this system. From this model, we posit a 3 to 4 m layer of water on the surface of the lake could easily result from the combination of a large surface flood and the melt generated at the water/ice contact. This record may be affirmation that anomalous warming events, such as the flood year of 2001/02 [*Barrett et al.*, 2008; *Doran et al.*, 2008b], are not unprecedented in the Dry Valleys.

Below 16 m in the ice core, salinity increases and the ice appears to have recrystallized. Recrystallization is induced by temperature changes, stress or strain on the ice, and/or the presence of debris [*Samyn et al.*, 2008]. There are two processes by which the lower ice may have formed: 1) the freezing of surface water, where salts were concentrated through evaporation/sublimation, or 2) the freezing of brine from beneath. Three lines of evidence support that the entirety of the ice core was formed from surface inflow rather than brine freezing.

1) Sediment layers found in the lower portion of the core (21 to 26 m) must have been deposited on the surface of the ice. These sediment layers are underlain by ice which is too thick to be segregation ice (ice lenses formed horizontally between layers of soil/sediments), as ice lenses formed within basal sediments beneath glaciers are typically thin at 0.01-0.1 m and are spaced more than 0.5 m apart [*Christoffersen and Tulaczyk*, 2003].
2) The ice in SL26.28 contains abundant diatom frustules, which are photosynthetic organisms. All genera are freshwater diatoms associated with Dry Valley streams [*McKnight et al.*, 1999; *Warnock and Doran*, 2013].
3) If the ice formed from brine freezing, the lower ice would be expected to have similar ionic ratios to the underlying brine, yet these are clearly distinct (Figure 9). This final argument may not hold if the brine underwent significant mineral precipitation following freezing that altered the relative proportions of ions in the brine.

Based on these observations, our interpretation is that the entire 27 m of ice cored on Lake Vida has formed from freshwater freezing at the surface, rather than brine freezing downward as discussed in Doran et al. [2003].

The sediment layers in the lower ice are also unusual, given that we hypothesize the ice formed from surface inflow. In January 2002, anomalously high stream discharge flooded the surface of Lake Vida with turbid water; yet, only a thin band of sediment < 1 cm is evident in the upper 3 m of the ice core. Therefore, it is improbable that sediment layers 20 cm in thickness formed from a single surface flooding event. Additionally, it has been noted in Taylor Valley (where lake ice is formed at the bottom of the ice cover, not the surface), that during the austral summer the low albedo of surface sediment can cause it to warm and move downward in the ice cover [*Hendy*, 2010]. This movement tends to aggregate sediment into layers and pockets approximately 2 m below the surface [*Priscu et al.*, 1998]. From this, it is hypothesized that the thick sediment layers in Lake Vida formed from repeated deposition rather than individual events, and were amassed during periods having a negative water balance as the ice cover ablated.

Our hypothesis that sediment layers were formed from long-term evaporation and sublimation is further supported by the isotopic enrichment and low deuterium excess of water contained within the sediment layers (Figure 15). When water evaporates from a water body to the atmosphere, the molecules are depleted in the heavy isotopes of oxygen and hydrogen, which leaves the remaining water enriched [*Horita*, 2009]. The same has been shown for the sublimation of snow and ice [*Neumann et al.*, 2008; *Sokratov and Golubev*, 2009; *Lacelle et al.*, 2011]. Likewise, negative deuterium excess values suggest that evaporative fractionation has more strongly modified oxygen isotopes over hydrogen isotopes (this is due to the composition of water, where there are two hydrogen atoms to every one oxygen atom, and therefore any given oxygen atom is more readily influenced).

From the overall mineralogy of stream and aeolian samples it was not possible to ascertain the provenance of the sediments in the ice core; however, the majority of sediment contained within the ice is had a mean grain size smaller than sediment collected from the ice surface and most of the aeolian deposits collected on the valley floor (Figure 13). This may be the result of smaller particles preferentially melting through the ice cover [*Hendy*, 2010] or the prevalence of riverine input. Field observations support that during high flow years, turbid river water flows over the surface of the lake. A saturated lake surface provides a mechanism for sediment to both infiltrate cracks in the ice, as well as become frozen beneath a new layer of water. Aeolian transport of sediment is common in Victoria Valley [*Speirs et al.*, 2008]; however, aeolian deposition onto a dry, flat ice cover has a high probability of further redistribution by wind and may not be readily entrained into the ice column. This conclusion is supported by field observations that windblown sediment largely does not get trapped on the frozen ice surface of Lake Vida, but saltates across the lake.

During periods of dry climatic conditions in Victoria Valley, the sediment layers may have been visible near the surface of the lake, and would have been analogous to the ice-cemented soils found at higher elevations in the Victoria Valley. At 450 m asl in Victoria Valley, the ice-rich permafrost in a 1.6 m soil profile had similar sediment characteristics to the lower sediment layers in Lake Vida, with a median grain size range of 357-510 µm, and water content <13% [*Hagedorn et al.*, 2007]. The dry permafrost/ice-cemented contact was found at 22 cm below the surface, which was also the approximate maximum depth of the 0°C isotherm during the three years of study [*Hagedorn et al.*, 2007]. At the edge of Lake Vida, soil temperatures at 10 cm depth, rise only slightly above 0°C during the short summer (MCM LTER). On Lake Vida, we propose that similar thermal conditions preserved the ice found below and between the thick sediment layers. As sediment layers on the ice thickened to almost 20 cm, the amassed sediment provided insulation for the ice beneath, and allowed the ice to remain below freezing during the summer.

Only in SL26.28, does the mean grain size significantly decrease toward the base of the core (Figure 12). The occurrence of silt is not common in the Dry Valleys, but silts are found in the sediment beneath ice covered lakes [*Wagner et al.*, 2006, 2011]. The increase in total carbon concurrent with the decrease in grain size, as well as the presence of freshwater diatom frustules, points to the possible occurrence of more open water conditions during this time.

In all GPR transects, the radar signal is attenuated below 21 m. Doran et al. [2003] interpreted this horizontal reflector as the top of a large brine body. From the drilling detailed in this study, we now interpret this impenetrable basal reflector to be the SL21.62 sediment layer, as the thickness and salt content likely inhibited radar penetration [*Frolov*, 2003]. Above SL21.62, the continuity of the horizontal reflectors across the lake validates the extrapolation of ice core records to the entire lake body. The noticeable undulations in the reflectors between 8 to 12 m are interpreted as density contrasts in ice layers [*Arcone and Kreutz*, 2009], which may have been formed from 3 to 4 meters of liquid water freezing downwards, as discussed previously based on isotopic composition. Initially horizontal, these bands were later forced into their present configuration by pressure due to freezing below. Also evident in the radar profiles is a preserved paleo-terrace at 8 m depth along the south end of the lake (Figure 16). This may be evidence that Lake Vida maintained an elevation of 8 m below present for a prolonged period. The lack of downcutting along the lake margins below 8 m (Figure 16a) further suggests the lake level has mostly risen since this time, which based on our previous calculation encompasses the past 100 years.

Isolating individual dates of deposition or burial of the sediment layers is challenging. Radiocarbon dates in the Dry Valleys can often be erroneously old due to a reservoir effect [*Doran et al.*, 1999; *Berger and Doran*, 2001], where an inherited age can result from the direct input of old carbon into lakes [*Doran et al.*, 2014]. A residence age can similarly result from limited atmospheric exchange of lake water resulting from permanent ice covers or strong salinity gradients, or the inclusion of old organic material reworked into modern stream water [*Doran et al.*, 1999; *Hendy and Hall*, 2006]. In Victoria Valley, water travels more than 1 km from glacial sources to Lake Vida, which should allow waters to equilibrate with modern atmosphere before reaching the lake. Moat waters too should contain mostly modern carbon, although Doran et al. [2014] found moat waters in Taylor Valley lakes with apparent DIC ages as high as 3790 ^{14}C yr BP. Only lakes with large open water moats like Lake Fryxell seem to have modern DIC. Hall and Henderson (2001) found that at Lake Vida, lacustrine carbonates along the shorelines with uranium/thorium ages of 9600 yr BP had a ^{14}C reservoir age of +3600 yr BP. They concluded that old CO_2 was likely input into the system with meltwater from the Ross Ice Sheet.

Of the eight radiocarbon samples taken from the upper 13 m of ice in Lake Vida, only one sample returned a concentration indicative of modern carbon (Figure 11). All indications point to erroneously old carbon dates as we assume that at least the upper 7.7 m of ice formed during the last century. However, the radiocarbon dates impart a maximum age constraint on the ice, and indicate the top 27 m of ice formed after 6300 ^{14}C yr BP. This aligns with the geomorphic reconstructions that indicate that Victoria Valley was filled with a deep (>200 m) glacial lake prior to 8600 yr ^{14}C BP [*Hall et al.*, 2002, 2010], and the Ross Ice Sheet retreated from the mouth of the Dry Valleys between 6500 and 8340 ^{14}C yr BP [*Hall and Denton*, 2000]. Furthermore, it may be that the ice cover originated much later, as radiocarbon dates of the dissolved organic carbon fractions in the Lake Vida brine date between 2955 and 4150 ^{14}C yr BP [*Murray et al.*, 2012].

OSL dates at SL21.62 and SL25.59 represent the date at which the minerals in the sediment layers were last exposed to solar radiation, or more specifically, an interval when the ice cover was thin enough to allow sunlight to penetrate to the dated sediment layer, followed by a period of ice growth or further sediment burial which extinguished the light source to the sediment layer. Therefore, the OSL dates indicate a lake level drawdown and rebound at 1200 and 320 yr BP. A lowering at 1200 years matches previous paleolimnological studies of lake levels in Taylor and Wright Valleys [*Wilson*, 1964; *Lyons et al.*, 1998a]. Also, if our previous interpolation of a 7.7 m lake level rise in 103 years is further extrapolated, it suggests that a 21.62 m lake level rise is not improbable over 320 years.

The discrepancy between OSL and radiocarbon dating techniques has been documented in the Dry Valleys [*Berger et al.*, 2010, 2013]. Contaminated OSL samples yield artificially young dates, whereas contaminated radiocarbon samples tend to yield artificially old dates. Together, the two dating techniques constrain evaporation events, and suggest that the current Lake Vida system is a few millennia in age.

The capacity of Lake Vida to integrate watershed process presents a fundamental framework for the understanding of hydrological and climatological shifts over time. With the limited reliable dates available in this study, there is no discernible correlation between lake highstands/lowstands and the temperature proxy record from neighboring Taylor Dome [*Mayewski et al.*, 1996; *Stager and Mayewski*, 1997; *Steig et al.*, 2000]. This lack of synchronicity between runoff/lake level and temperature has been noted in other lake level reconstructions [*Whittaker et al.*, 2008] and throughout the instrumental climate record [*Levy et al.*, 2013].

3.7 Conclusions

The inconsistency in radiocarbon dates makes a full reconstruction of the history of the Lake Vida ice cover challenging. However, several important conclusions are gained from this ice/sediment record:

1) A hydrologically variable climate is not unique to recent times. Lake Vida has experienced at least four major drawdowns that led to the accumulation of thick sediment layers in the lower ice cover. These drawdowns may have occurred as early as 6300 ^{14}C yr BP, but OSL ages and a presumed reservoir effect in radiocarbon ages suggest these events were likely constrained to the last 2-3 millennia.
2) The accumulation of discrete sediment pockets into ~20 cm thick layers insulates the ice beneath. This is a significant process for the preservation of ancient ice

beneath sediment in cold and dry climates, and may have consequences for the preservation of ancient ice and brine on other icy planets.

3) The entire 27 m of ice was produced from glacial streams flooding the ice surface at lower lake levels than present. In the ice collected to date, there is no freezing from the bottom of the ice cover downwards, as suggested by Doran et al. (2003).

4) There is a common brine pool that is hydrologically connected between the holes drilled in 2005 and 2010 (the 1996 hole did not penetrate this depth). Lake Vida is an ice aquifer with a piezometric surface approximately 10 m below the current ice surface.

5) The source of sediment in the ice appears to be dominantly from suspended sediment in stream flow.

Lake Vida represents a unique lacustrine system, which during wet periods, records a hydrologic history in the growing ice cover. As climate is projected to warm [*Shindell and Schmidt*, 2004; *Arblaster and Meehl*, 2006], Lake Vida may provide an ideal environment for tracking the influence of climate on hydrology in the Dry Valleys.

3.8 Acknowledgments

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4: SUBSURFACE IMAGING REVEALS A CONFINED AQUIFER BENEATH AN ICE-SEALED ANTARCTIC LAKE

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4.1 Abstract

Liquid water oases are rare under extreme cold desert conditions found in the Antarctic McMurdo Dry Valleys and the long-term sustainability of near surface water bodies is not assured. In areas where the surface is frozen, lenses of water present in the subsurface may act as microbial refugia. Here, we report geophysical results that indicate that Lake Vida, one of the five major lakes in the region, is nearly frozen, and cryoconcentrated brine remains in a confined aquifer situated beneath the lake. A ground penetrating radar survey penetrated 20 m into lake ice and facilitated bathymetric mapping of the grounded ice. An airborne transient electromagnetic survey revealed a low resistivity zone 30-100 m beneath the lake basin. Based on previous knowledge of brine chemistry and local geology, we interpret this zone to be an isolated and confined aquifer situated in unconsolidated sediments with a porosity of 23-42%. Our results facilitate the modeling of hydrologic processes in unglaciated regions of Antarctica, provide a better understanding of historic lake conditions in the Dry Valleys, and reveal a system where liquid water may act as microbial refugia beneath a cold, desert environment.

4.2 Introduction

The McMurdo Dry Valleys of East Antarctica are one of the coldest unglaciated environments on Earth, and a possible planetary analogue to cold, desert environments of Mars and other similar planetary bodies [*Doran et al.*, 2010]. In this habitat, terrestrial hydrologic inputs are limited to summertime snow and glacial melt feeding surface streams and lakes. In the valleys, deep glacial lakes once covered areas that are currently exposed sediments. Hall et al. [2010] proposed that the lake levels in the Dry Valleys have fluctuated 100s of meters throughout the last 20,000 years. If water infiltrated into the subsurface during these times, then it may still exist beneath the valley floors, providing hidden microbial refugia [*Mikucki et al.*, in review].

One major challenge in understanding the hydrogeology of cold polar deserts is to assess the volume of liquid water in the subsurface [*Ireson et al.*, 2012]. For groundwater to remain liquid below 0°C, the freezing point must be depressed by increased salinity. This can occur through cryoconcentration, whereby solutes are excluded during the freezing process and concentrated in the remaining liquid. Even in continuous permafrost, saline groundwater may pervade through tortuous pathways in rocks and sediments with low hydraulic conductivities [*Seyfried and Murdock*, 1997; *Dickinson and Rosen*, 2003].

In the Dry Valleys, only Lakes Bonney, Vanda and Vida [*Chinn*, 1993] are known to contain concentrated brine at depth. Geochemical data at Lake Bonney indicate that this brine results largely from cryoconcentration events [*Lyons et al.*, 2005]. Resistivity measurements from the Dry Valleys Drilling Project (DVDP) in the 1970s indicated that Lake Vanda and Lake Bonney were not underlain by frozen ground, and authors suggested that a hydrological connection between the lakes and a deep groundwater source was possible [*McGinnis and Jensen*, 1971]. However, there has been little evidence to support a groundwater flux beyond a weak hydraulic gradient in the bottom sediments of Lake Vanda [*Cartwright and Harris*, 1981]. More recent studies on brine geochemistry have also found little evidence of a deep groundwater source to most of the lakes [*Lyons et al.*, 1998b; *Witherow et al.*, 2010]. Consequently research has shifted towards seasonal shallow subsurface water [*Levy et al.*, 2011; *Dickson et al.*, 2013; *Gooseff et al.*, 2013].

Lake Vida is unique among all Antarctic oasis lakes investigated to date, because of the presence of an isolated, pressurized brine system within and beneath approximately 27 m of lake ice [*Murray et al.*, 2012]. The brine was sampled via surface drilling in 1996, 2005 and 2010, but the extent of lake ice and volume of brine beyond 27 m remained undefined.

Here we mapped the ice stratigraphy and ice-bed contact of Lake Vida with ground penetrating radar (GPR) and constrained the volume and spatial extent of brine in and beneath the lake, and the potential for groundwater in this system, with an airborne transient electromagnetic (AEM) survey.

4.3 Study Site

Lake Vida, at 6.8 km^2, is one of the largest lakes in the McMurdo Dry Valleys, and is located in the middle of Victoria Valley at 340 m above sea level (asl) (77°23′S, 161°56′E). The lake receives inflow from three glacial streams, and loses water entirely through sublimation and evaporation. Unlike other lakes in the Dry Valleys, summer stream flow pools on the surface of the unusually thick ice cover where it eventually freezes. Victoria Valley summer temperatures are similar to those in the other valleys where melt water is present, but winter temperatures are much colder, which enables the thick ice cover to build up [*Doran et al.*, 2002b].

Lake Vida is the only lake known to contain pressurized brine confined beneath ice and sediment layers. In 2010, 27 m of ice and sediment cores were collected. The ice below 20 m was atypical of lake ice, in that it contained multiple thick (10-20 cm) sediment layers [*Dugan et al.*, 2014]. During both the 2005 and 2010 coring campaigns, brine entered the borehole below 16 m and rose to 10.5 m below the surface [*Murray et al.*, 2012]. The brine was -13.4°C with a salinity of 195 g L^{-1} [*Murray et al.*, 2012].

The only lateral constraint on the extent of the brine system is provided by the DVDP-6 borehole, which was drilled on the northwestern shore of the lake in 1973, when the lake level was 3.6 m lower than present. The borehole penetrated 306 m and reached crystalline basement rock after only 10.5 m of frozen sand and gravel [*Kurasawa et al.*, 1974]. Notably, drilling fluid left in the borehole was observed multiple times to have been displaced upwards [*Cartwright and Harris*, 1981; Doran pers comm], and DVDP-6 was the only borehole to exhibit temperature fluctuations [*Decker and Bucher*, 1982], which could indicate groundwater seepage. The DVDP borehole temperature was -24 to -25°C at depths corresponding to where Murray et al. [2012] encountered the -13.4°C brine in the center of the lake.

4.4 Methods

GPR profiles were recorded across Lake Vida with a GSSI SIR-3000 acquisition unit and 400 and 900 MHz antenna units in Nov. 2010 and 2011, respectively. In 2010, over 50 km of transects were collected and tracked with standard GPS (Figure 18). We recorded 2048 16-bit samples per trace at 400 MHz with a 400 ns two-way travel time, and 100 MHz high-pass and

700 MHz low-pass finite impulse response (FIR) filters. In 2011, half of the transects were repeated at 900 MHz, with a 300 ns time range and 400 MHz high-pass and 1200 MHz low-pass IR filters.

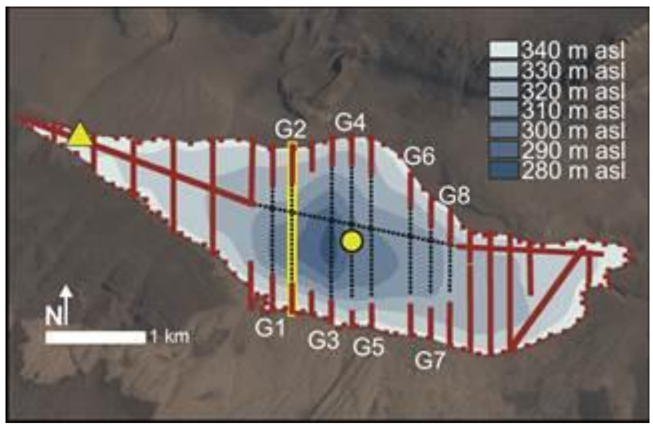

Figure 18. Lake Vida bathymetry acquired from interpolation of ground penetrating radar (GPR) transects. This is an estimate of maximum potential ice thickness. Thick red squares near the edges of the lake represent known lake bottom depths from GPR data, whereas black dots in the center represent the eight transects where data were estimated by spline interpolation. The yellow dot corresponds to the 2010 Lake Vida borehole, and the yellow triangle locates DVDP-6. The highlighted yellow transect is shown in Figure 19.

Bathymetry was mapped from the ice/basin contact depths resolved from the GPR transects (Figure 19). In general, this was possible down to 20 m depth before the signal attenuated and no lake bed reflection was returned. For transects G1 to G8, the ice/bed contact below 20 m was approximated by a fitting a cubic spline function using R [R *R Core Team*, 2012, Version 2.15.2] Fitted points were merged with the GPR dataset and spatially interpolated into a bathymetric map using natural neighbor interpolation in ESRI ArcGIS 10.1.

The AEM survey was performed using a SkyTEM504 system flown underneath a Bell 212 helicopter [*Sørensen and Auken*, 2004]. The system records laser altimetry, transmitter frame pitch and roll, and GPS positions, together with the time derivative of the decaying magnetic field in the subsurface. The maximum transmitter moment of the system was 165,000 Amp·m^2 and data were usable from 12 µs from the start of the turn off ramp until the noise level was reached at 1-8 ms. In total, 70 km of data were collected over Lake Vida. Of these, there were only electromagnetic signals in soundings above brine, as next to no signal was produced from permafrost or crystalline bedrock. In total, 727 soundings, covering 16.7 km, were left after processing, split between five flights lines (ST1 to ST5 – Figure 20,21). The soundings were inverted using a multilayer inversion approach discretizing the models in 28 depth logarithmic distributed bins from 0 to 600 m below the surface [*Auken and Christiansen*, 2004; *Viezzoli et al.*, 2008], although in approximately half of the locations, the depth of investigation (DOI) determined during data inversion was limited to 150 m below the surface [*Christiansen and Auken*, 2012].

Total brine volume was estimated using a horizontal ~40x40 m x,y-grid bounded by the flight lines, over which resistivity values were interpolated using a natural neighbor method at a 1 m depth interval. The volume fraction of subsurface brine in each bin was estimated from resistivities using a simplified Archie's Law [*Archie*, 1950]:

$$\phi = \rho_o/\rho_b^{(-1/m)} \qquad \text{Eq. (1)}$$

where ϕ is liquid brine fraction, ρ_o is the bulk resistivity determined from AEM, ρ_b is the resistivity of the brine, and m is the cementation factor. We assume that brine resistivity is equal to that recovered from the Lake Vida drill holes at 6.5 S m^{-1}.

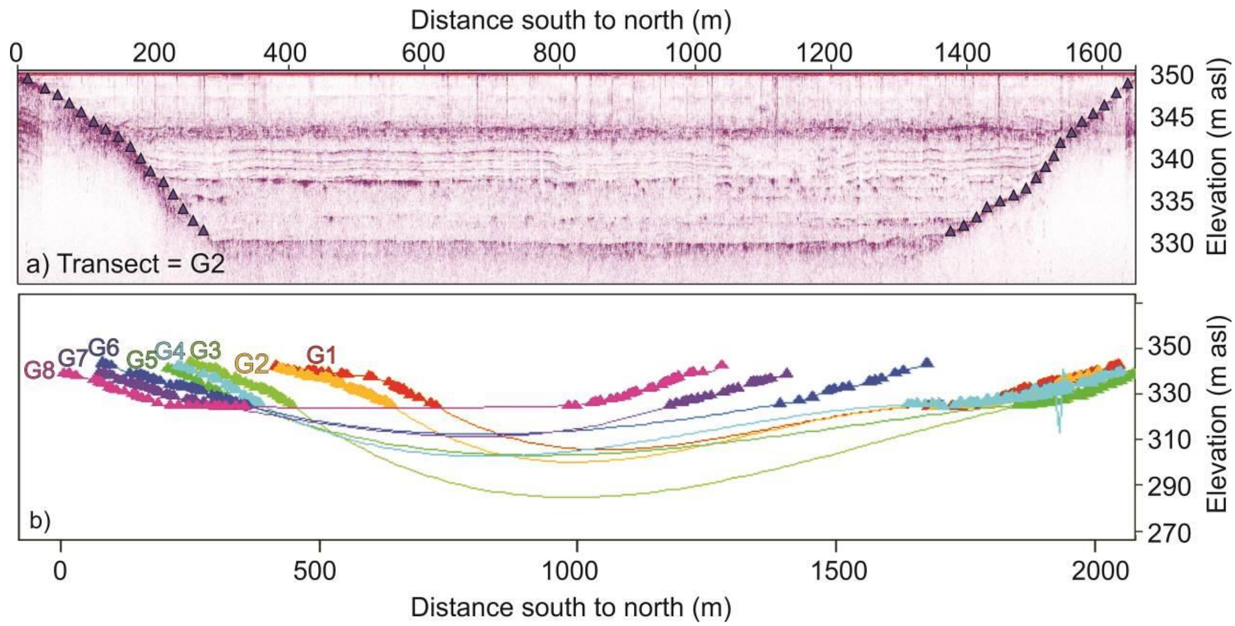

Figure 19. a) GPR profile across the middle of Lake Vida. The location of transect G2 is shown in Figure 18. The ice/basin contact, marked by triangles, was digitized for all transects. At 330 m asl, the basin contact is lost when the GPR cannot penetrate a strong horizontal reflector. b) Cubic spline interpolation across the eight transects where the GPR did not return the ice/basin contact below 330 m asl.

The cementation factors in permafrost are not available, but published values typically range from 1.4 to 2.8, with highly fractured rock on the lower end and rock with spherical pores or increased tortuosity on the higher end. To investigate the sensitivity of our results to uncertainties in this poorly constrained parameter, we performed calculations using m equal to 1.5, 2, and 2.5.

4.5 ***Results***

The GPR data show the lakebed where ice is less than ~20 m thick (Figure 19a), providing bathymetry for 69% of the area (Figure 18). For the eight north-south GPR transects which crossed the center of the lake, spline fitting returned maximum depths of 40, 41, 60, 42, 42, 33, 34, and 21 m, from west to east (Figure 19b). Resistivity values obtained from the sensor range from 1.3 to 20,000 Ω•m. The minimum in the first 20 m beneath the surface is 84 Ω•m. The largest anomaly lies 30 to 100 m beneath the center of Lake Vida (Figure 20), where values approach 1.3 Ω•m at 33-47 m depth. The low resistivities begin to dissipate near the north and south edges of the lake. Flight line ST5 did not register any values below 250 Ω•m, indicating the absence of brine at this line. It must be noted that weakening resistivity at the upper or lower corners of a strong brine signal may be exaggerated as the inversion is based on local 1-D models [*Goldman et al.*, 1994; *Auken et al.*, 2008].

The brine beneath Lake Vida has a resistivity (ρ_b) equal to 0.15 Ω•m, determined from the conductivity measured at the in-situ temperature of -13°C. We assume that all pore space is occupied either by this brine, or by ice (~10,000 Ω•m or higher in our AEM data). Therefore, the maximum brine fraction ranges between 23-42% (Figure 21). The total brine volume beneath Lake Vida is 5.3×10^6 m^3, 1.58×10^7 m^3, 3.23×10^7 m^3, when m = 1.5, 2, and 2.5, respectively.

All AEM flight lines were completely within the footprint of Lake Vida, with the exception of the edges of ST1 and the eastern half of ST2. Based on our estimated bathymetry, the high concentration of brine resides immediately below the modeled lake bottom in ST1, ST2 and ST3. Only in ST4, does it appear that a large fraction of brine resides in lake ice (Figure 21). However, if our modeled bathymetry is an overestimate, the majority of brine is stored beneath the basin of Lake Vida.

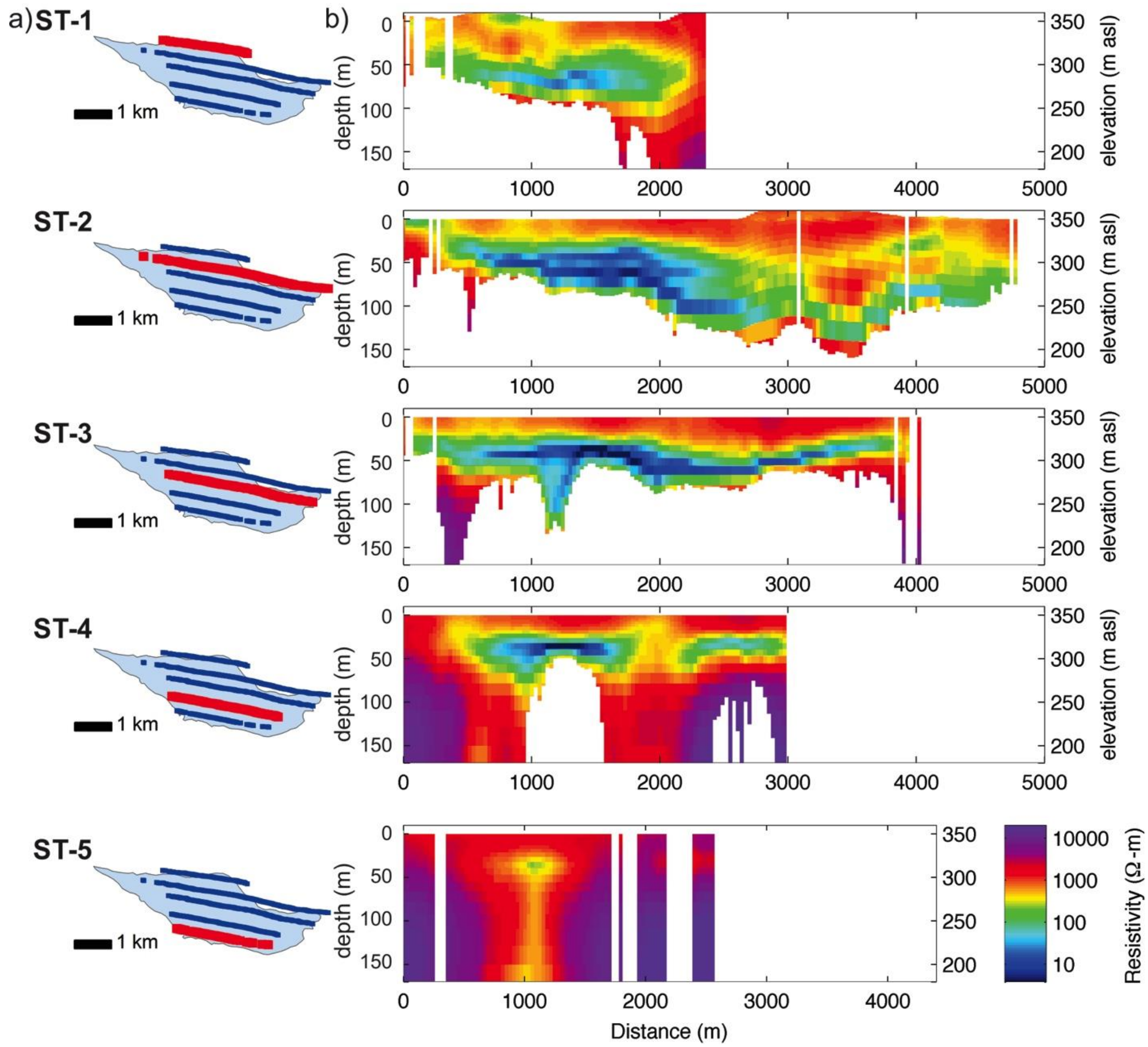

Figure 20. a) Location of AEM flight lines. b) Resistivity cross-sections returned from the AEM flight lines over Lake Vida.

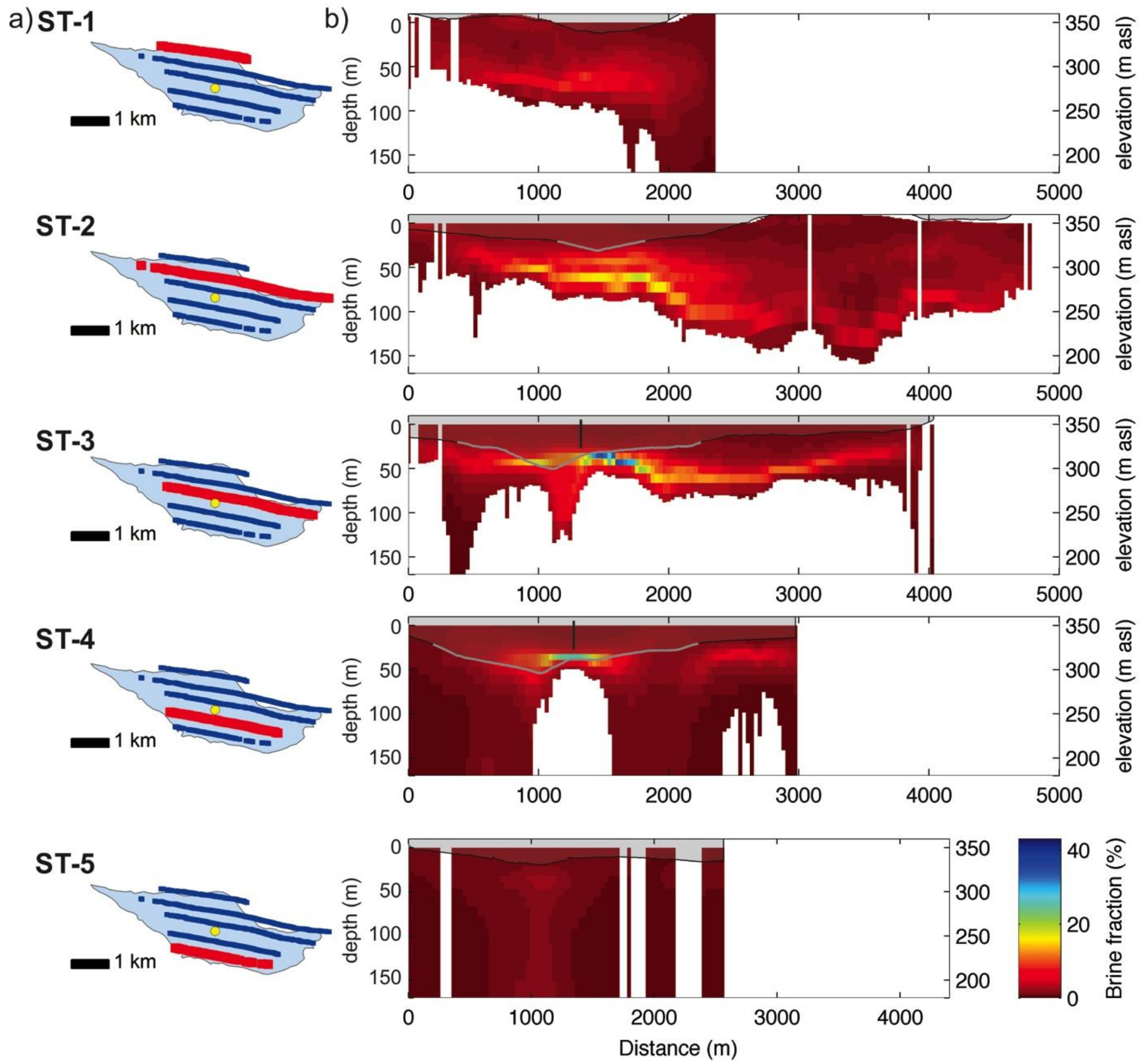

Figure 21. a) Location of AEM flight lines. The 2010 Lake Vida drill site is shown as a yellow dot between ST3 and ST4. b) Brine fraction estimates under Lake Vida estimated from Archie's Law with a cementation factor of m=2. Black lines and grey lines superimposed on the profiles are known and estimated bathymetric depths, respectively. The area above this line represents lake ice. The grey rectangle in ST-3 and ST-4 is the nearest approximate location and depth of the 2010 Lake Vida drill hole.

4.6 *Discussion*

Natural freshwaters have variable resistivities, 5-100 Ω•m, depending largely on solute content. Typical seawater at its freezing point has resistivity slightly higher than 0.3 Ω•m. When water freezes, the resistivity of the remaining liquid gradually decreases while the resulting ice has high resistivity. Insulating materials, such as unweathered rock and ice, have resistivity values > 1000 Ω•m and upwards of 100,000 Ω•m (TABLE IV). The only geologic deposits with resistivities <10 Ω•m are extremely weathered rocks, clays, or economical mining targets, such as sulfide deposits [*Palacky*, 1988].

TABLE IV. RESISTIVITY OF COMMON SUBSTRATE AND SURFICIAL WATERS IN THE MCMURDO DRY VALLEYS

Substrate	Resistivity (Ω•m)
Lake Vanda (Wright Valley) surface water[1]	111, 180
Lake Vanda ice 0.1 m[1]	19,500
Lake Vanda ice 0.9 m[1]	18,690
Wright Valley unsorted soils[1]	90-413
Beacon sandstone[2]	103,800
Granite Gneiss (21°C)[2]	100,360
Diorite (21°C/− 25°C)[2]	102,200/452,000
Diorite (21°C/− 25°C)[2]	33,100/27,700
Lake Vida brine[3]	0.15

[1]Field measurements with Wheatstone bridge [*McGinnis and Jensen*, 1971]. [2]Laboratory measurements [*McGinnis and Jensen*, 1971]. [3]Inverse of brine conductivity measured in 2010

Beneath Lake Vida, the resistivity minimum of 1.3 Ω•m indicates only a fraction of the subsurface volume consists of the highly concentrated 0.15 Ω•m brine observed in drill holes in the ice. We do not consider it plausible that 1.3 Ω•m signifies open water, as water with a resistivity of 1.3 Ω•m would have an *in situ* salinity half that of seawater and could not underlie the brine without inducing an unstable density stratification.

From resistivity values alone, it is difficult to distinguish if the brine exists in a network of bedrock, ice, or permafrost, as these are all highly insulating materials. Here, we examine each of these possibilities in detail.

5.1 Brine in bedrock

The calculated maximum brine fraction of 23-42% beneath Lake Vida indicates porosity much higher than typically found in the massive biotite gneiss documented in the near surface of the DVDP-6 core. Therefore, we infer that the valley continues to have a steep relief beneath Lake Vida, and the entire low resistivity anomaly below Lake Vida is situated above the local bedrock. In areas where resistivity is > 500 Ω•m, brine may be present in the cracks of the bedrock. At 1% porosity, the overall resistivity would equal 1500 Ω•m (m=2), which is well within the range returned by the AEM sensor, despite the poor resolution of the method at high resistivities. In DVDP-6, thermal fluctuations were attributed to groundwater intrusion into the borehole at 230-250 m depth (100-120 m asl) in an area of massive biotite-granite [*Cartwright and Harris*, 1981]. In most areas covered by the AEM survey, the depth of investigation was less than 200 m, and brine at 230-250 m cannot be confirmed.

5.2 Brine in lake ice

The GPR performed on Lake Vida was unable to penetrate beneath a reflector in the lake at 20 m depth (Figure 19a). Initially assumed to be the top of a liquid brine body [*Doran et al.*, 2003], Dugan et al. [in review] have since shown that this reflector is the first in a series of thick sediment layers present in the lower ice. Below 20 m, our extrapolated bathymetry reaches a maximum depth of 60 m (transect G3, Figure 18,19). All other extrapolated GPR transects do not attain depths greater than 42 m. Based on these constraints, Lake Vida is likely 40-45 m deep, with a much lower possibility of depths reaching 60 m. If lake ice does not extend below 45 m, the majority of brine is below the lake bottom. However, considering the lowest resistivity was recorded in ST3 at 33-47 m depth, we cannot rule out the possibility of a high concentration of brine in an ice matrix at the very bottom of the lake.

5.3 Brine in unconsolidated sediments

Our calculations indicate that the brine network is focused between 30 and 100 m depth (250 to 320 m asl), and largely within the footprint of the lake surface. Our preferred scenario is that below 350 m asl (2010 lake level), the deep valley bottom is filled with porous unconsolidated sediments. The calculated brine fractions are comparable to measured soil porosities of 20% [*Hindmarsh et al.*, 1998] and 40% [*McKay et al.*, 1998] in the Dry Valleys.

We propose that the brine is mainly limited to voids in lacustrine sediments, in effect creating a confined aquifer, or large cryopeg (lens of unfrozen brine surrounded by permafrost). This aquifer beneath Lake Vida contains between 15 $x10^6$ m^3 to 32 $x10^6$ m^3 of brine. In comparison to Taylor Valley lakes, this is roughly 2-5 times the volume of liquid beneath the chemocline in brackish Lake Fryxell (6.7 x10^6 m^3) and 1-2x that beneath the chemocline in hyper-saline East Lake Bonney (1.9 x10^7 m^3) [*Obryk et a*l., *in prep*]. The source of the Lake Vida brine must stem from either external input or evaporation/freeze concentration of surface lake water. Since Victoria Valley has not been inundated by the ocean since the Miocene [*Chinn*, 1993], and there is little evidence of extensive groundwater connectivity beyond the lake, the brine presumably formed from the concentration of lake water that permeated the sediments. The majority of research in Taylor Valley supports repeated evaporation events in both Lakes Fryxell and Bonney [*Lyons et al.*, 1998a; *Poreda et al.*, 2004; *Whittaker et al.*, 2008], and repeated lake level fluctuations have been documented in the Vida basin [*Hall et al.*, 2010; *Dugan et al.*, 2014].

Brine age is 2800-4000 ^{14}C yr BP at 16 m, based on radiocarbon dating of isolated organic carbon fractions of recovered brine [*Murray et al.*, 2012]. This should indicate the last time the uppermost brine was exposed to the atmosphere. At this time, there is no estimate as to the formation age of the brine; however, the brine may have continued to concentrate over time. The brine is -13°C, while the mean annual surface air temperature varied between -30.0°C and -25.4°C during 1995 to 2000 [*Doran et al.*, 2002b], and the ground temperature immediately adjacent the lake is near mean annual temperature at the surface and increases to approximately -22°C at 100 m depth [*Decker and Bucher*, 1982]. The surface ice on Lake Vida is on average 7°C colder than the brine [*Doran et al.*, 2003] and contains sediment layers that prevent the transmission of light and solar radiation through the ice cover. Consequently, with no heat source, the relatively warm brine is a shrinking remnant of a larger historic lake. The remaining brine is kept unfrozen by a combination of ongoing cryoconcentration, which would increase the salinity of the brine and depress the freezing point, and latent heat production during the freezing of brine and any mineral precipitation [*Doran et al.*, 2003], which would slow the rate of cooling.

Two further observations support the idea that the Vida aquifer has cooled and concentrated over time. Firstly, the unstructured shape of the aquifer suggests heterogeneous freezing of the surrounding permafrost inward [*Gaidos and Marion*, 2003]. Secondly, in aquifers confined by basement rock and frozen sediments, the volumetric expansion of ice during freezing can increase the hydrostatic pressure [*Gaidos*, 2001; *Gaidos and Marion*, 2003]. One hole in 2005 and two holes in 2010 were drilled past 16 m in Lake Vida, and in all cases, the brine rose to a potentiometric level of 10.5 m (relative to 2010 ice level). If all 27 m of ice formed after brine formation [*Dugan et al.*, 2014], then the potentiometric level at the time of ice formation must have been below 27 m depth, and water pressure must have increased over time. Furthermore, the inferred absence of a free brine body even in the deepest part of the lake basin negates the previous interpretation that the potentiometric level resulted from a partial floatation of the ice cover [*Doran et al.*, 2003].

4.7 ***Conclusions***

Data collected using an airborne AEM survey of central Victoria Valley allowed us to define the limits of a confined aquifer beneath Lake Vida, and estimate the porosity at 23-42% and the brine volume at 15 to 32 million m^3. These findings are distinct from those in neighboring Taylor Valley, where an extensive groundwater network is present valley-wide [*Mikucki et al.*, in review], and reflect the unique microclimates and hydrologic histories of each valley.

The lack of a strong hydrologic connection beyond the edges of the lake to a larger groundwater system in Victoria Valley is an important insight into the regional hydrogeology, the legacy of past shifts in climate on the present-day ecosystem [*Moorhead et al.*, 1999], and role of Antarctic Dry Valleys as astrobiological analogues [*Doran et al.*, 2010]. Our results

- Do not support a large valley-wide groundwater network in Victoria Valley, at least in the top 300 m. This allows hydrologic processes at a lake ecosystem scale to be studied solely as surface/active layer fluxes in Victoria Valley.
- Provide an estimate of the volume of brine beneath Lake Vida, thereby providing a basis for reconstructions of historic lake levels and climate.
- Present a real world analogue for the preservation of water in a cold, desert environment analogous to Mars [*Gaidos and Marion*, 2003], and provide boundaries on the brine ecosystem, which has been shown to harbor a diverse and metabolically active microbial population [*Murray et al.*, 2012].

Our airborne sensor survey has provided a means of quantifying the spatial extent of brine in a remote environment. This has given us a new outlook on the potential for subsurface habitats in areas often considered devoid of life. Future research should focus on potential habitability of this system and the prevalence of groundwater in other polar deserts.

4.8 Acknowledgments

This work was supported by the US National Science Foundation Office of Polar Programs award # 1043618 to JAM and RAV, #1344349 to ST, and #0739698 to PTD. Raytheon Polar Services and Petroleum Helicopters Inc coordinated field logistics. NSERC and UIC provided graduate student support.

5: CONCLUSIONS

This dissertation addressed three major questions: can we better quantify ice loss through ablation in Taylor Valley lakes, how did the thick ice cover on Lake Vida form, and what is the extent of the brine system beneath Lake Vida? In answering these questions, we further the understanding of hydrologic and sediment transport processes associated with lake ice formation and preservation.

5.1 Contributions made by this research and future research questions

The chapters in this dissertation build on a history of research in the McMurdo Dry Valleys, which started with early geophysical and limnological studies in 1960s [*Henderson et al.*, 1966; *Calkin and Bull*, 1967; *Kurasawa et al.*, 1974]. It is hoped that the research discussed herein is a stepping-stone toward more fully understanding the functioning of Antarctic lakes. Discussed below are only a handful of questions that remain to be studied.

5.2 Taylor Valley ablation rates

Prior to the collection of the long term ablation record in Taylor Valley, the most commonly cited estimate for lake ice ablation in Antarctica (and elsewhere) was 0.35 cm yr^{-1}, based on output from a vertical water vapor mass flux model which used meteorological data from Lake Hoare to estimate sublimation [*Clow et al.*, 1988].

The ablation results presented here, suggest that 0.35 m yr^{-1} is likely an underestimate of true sublimation rates. A starting point for future work would be to re-run the sublimation model used in Clow et al [1988]. Since 1988, the sensor network has expanded in Taylor Valley and sensor accuracies and resolutions have improved [*Winslow et al.*, 2014]. Dataloggers that draw minimal power can sample at frequencies < 1 min year-round, which can enable a more accurate representation of fluxes.

As high resolution satellite imagery becomes more accessible, spatial analyses of both lakes and landscapes become possible [*Eveland et al.*, 2012; *Obryk et al.*, 2014]. Using imagery we can better estimate lake ice albedo by reclassifying reflectance values as either sediment or ice (Figure 22). If spectrophotometry was conducted on small patches of lake ice, albedo results could be extrapolated based on percent sediment cover, as sediment mineralogy and morphometry appears to be consistent across the ice. Additionally, the imagery can be used to estimate the percentage of open water during the summer as a gauge of potential evaporation.

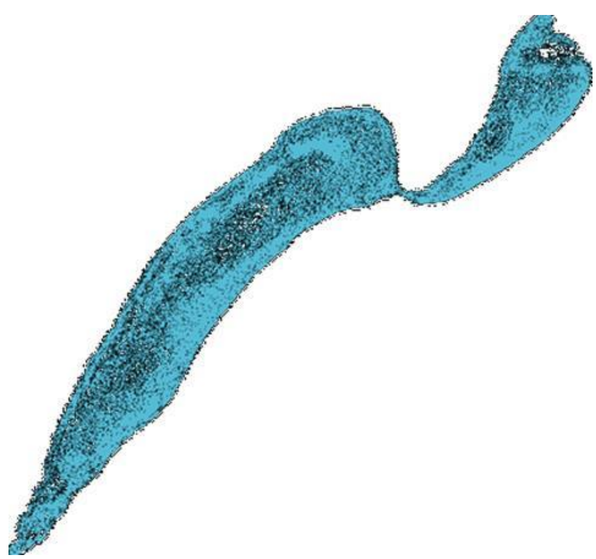

Figure 22. Reclassified raster of Lake Bonney. Black represents surface sediment and blue represents ice.

The model could be re-run at Lake Hoare, and also applied at Lake Fryxell and Lake Bonney. These are all sites with long-term meteorological and lake station records. The empirical ablation data obtained from the lake stations [*Dugan et al.*, 2013] are only representative of sublimation during the winter months. In the summer, ablation is the sum of sublimation and any melt. By rerunning the Clow model with climate data for years that overlap with the lake station data, it will be possible to separate sublimation and calculate ice melt during the summer. Other sublimation models, such as that used by Bliss et al. [2011] should also be compared. For best results, adding a second anemometer to meteorological stations at 1 m above the surface and an ice surface temperature sensor, preferably an infrared radiometer, at the lake stations should be considered.

Quantifying sublimation year-round, as well as possible evaporation, will help close the hydrologic balance in the Dry Valley lakes. Gauging rivers is a notoriously difficult task, and recording flow through water tracks and via direct glacial melt is near to impossible. If sublimation rates and lake levels are both quantified, it may be possible to accurately estimate inflow into the lakes without recording the terrestrial components.

Sublimation data can also be used to refine ice cover models, such as in McKay et al. [1985]. More robust ice cover models will allow us to describe conditions necessary for permanent ice covers to remain in equilibrium (such as those in Taylor Valley) or for the sustained growth of ice covers (such as Lake Vida), or if/when ice covers might be completely lost. This information is advantageous for investigating ice growth on other planets, or predicting ecosystem change under future warming scenarios on Earth.

5.3 Formation of Lake Vida

Data from Chapter 3 significantly update our understanding of the structure and formation of the Lake Vida ice cover since Doran et al. [2003] (Figure 23). However, the ice coring in 2010 did not hit a brine body nor end in bottom lake sediments at 27 m. The final core retrieved was lake ice, which has resulted in further speculation as to the extent of the lake ice/brine system beneath Lake Vida.

The AEM dataset detailed in Chapter 4 is the first geophysical survey that reveals the nature of the subsurface beneath the brine, and presents the first robust estimates as to the extent and volume of brine in the Vida basin. From the results in Chapter 4, it appears that the ice cover on Lake Vida may reach 45 m in depth. If future exploration of the brine system is planned for this lake, 45 m should be the target depth. This could only be accomplished using a drill capable of penetrating frozen sand layers > 20 cm thick [such as *Juck et al.*, 2005]. Drilling beyond 45 m would answer questions such as: Does the chemistry of the brine change with depth? Does the microbial community change? Are sediment layers present beneath 27 m that represent lake drawdown events? What age are the sediments beneath Lake Vida? Are physical or chemical indicators present in the sediments that could be used as proxies for understanding historic climate and ecosystem shifts?

The AEM survey should also be used to identify optimal drill sites based on the estimated quantity and location of brine. A future drilling campaign would clearly benefit from drilling in a location where the brine is thought to be most abundant. Alternatively, it would be worthwhile to select a drill site near the edge of the lake that is underlain by brine to confirm that brine geochemistry is uniform across the lake.

Future drilling should also be combined with further geophysical soundings. Resistivity data from the AEM survey does not extend beyond the edges of the lake. A ground based electromagnetic survey may provide a better means of collecting data above the permafrost that borders the lake. Employing a ground based method would also allow higher resolution data to be collected in areas of interest, such as the deepest brine pocket in the center of the lake.

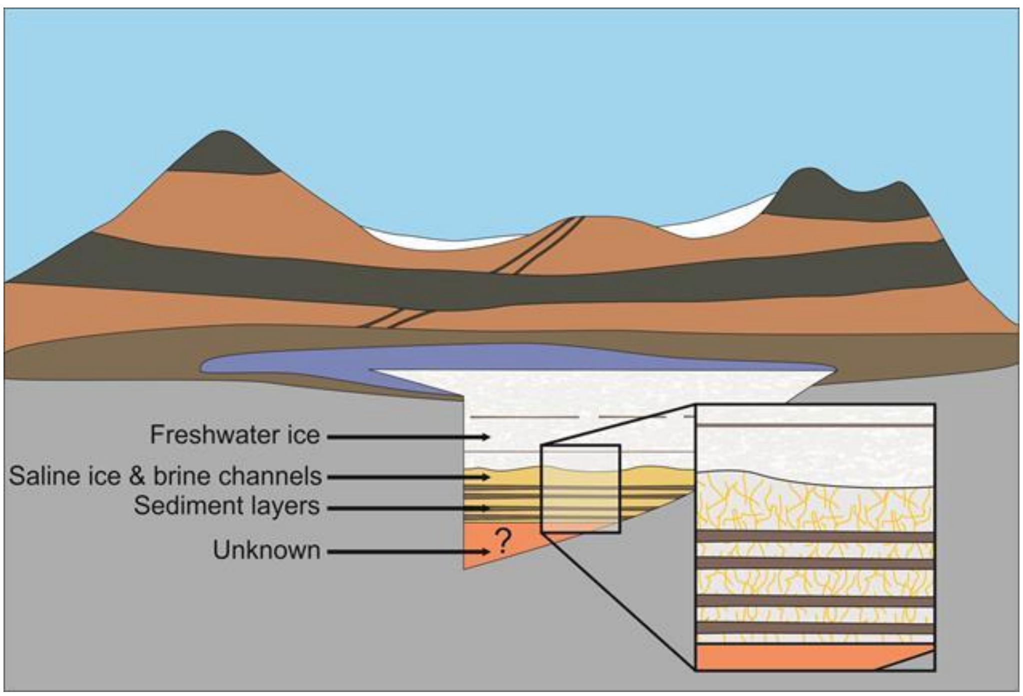

Figure 23. A depiction of our current understanding of the Lake Vida system. Freshwater ice (white) interspersed with thin sediment layers is capping briny ice (yellow) interbedded with thick sediment layers and thin brine channels.

The results presented in both Chapters 3 and 4 combine to transform our understanding of the brine system beneath Lake Vida. Future research should use the conclusions presented herein to investigate the age of the brine system and the processes by which it formed. This analysis could have implications for our interpretation of the history of subsurface brines elsewhere in our solar system, such as Mars.

APPENDIX

Data collected on samples from Lake Vida, 2010.

TABLE V. ISOTOPIC ABUNDANCES OF ^{2}H AND ^{18}O IN SELECT SAMPLES FROM THE 2010 LAKE VIDA ICE CORE. BOTTOM DEPTH IS NOT INCLUDED FOR SEDIMENT SAMPLES $\leq$1 CM THICK.

Top depth (cm)	Bottom depth (cm)	δ^2H	δ^{18}O	Sample type
268		-251.66	-31.09	Sediment pore water
395		-252.28	-31.59	Sediment pore water
785		-247.00	-30.19	Sediment pore water
1005	1015	-236.22	-27.38	Ice
1052	1057	-225.68	-25.03	Ice
1103	1108	-237.57	-26.56	Ice
1150	1155	-239.13	-26.70	Ice
1208	1218	-250.72	-29.75	Ice
1224	1229	-252.04	-29.81	Ice
1269	1281	-278.95	-34.55	Ice
1314	1319	-226.44	-25.59	Ice
1620	1625	-221.27	-25.28	Ice
1631	1636	-228.14	-25.90	Ice
1673	1678	-255.49	-31.07	Ice
1731	1736	-238.67	-29.04	Ice
1753	1758	-249.09	-30.78	Ice
1767	1772	-225.61	-26.88	Ice
1786	1791	-227.21	-26.99	Ice
1800	1808	-234.33	-28.15	Ice
1820	1826	-253.78	-31.87	Ice
1840		-213.39	-23.80	Sediment pore water
1851	1856	-232.85	-26.67	Ice
1866	1871	-236.32	-27.45	Ice
1877	1882	-233.64	-27.77	Ice
1892	1897	-224.15	-27.12	Ice
1912	1917	-240.13	-29.82	Ice
1927	1932	-234.03	-28.28	Ice
1947	1955	-233.24	-28.34	Ice
1997	2002	-237.32	-28.06	Ice

2007	2013	-238.27	-29.13	Ice
2018		-208.05	-22.78	Sediment pore water
2156		-242.77	-30.36	Sediment pore water
2156	2161	-223.04	-26.32	Ice
2163		-200.13	-22.45	Sediment pore water
2187		-201.80	-21.96	Sediment pore water
2197	2202	-227.68	-26.73	Ice
2218	2223	-240.68	-30.06	Ice
2257	2262	-242.80	-29.27	Ice
2267	2272	-237.57	-28.14	Ice
2283	2288	-233.98	-27.90	Ice
2297		-205.36	-22.65	Sediment pore water
2298		-212.84	-23.45	Sediment pore water
2300		-212.67	-24.85	Sediment pore water
2304	2309	-255.45	-32.24	Ice
2319	2324	-250.16	-30.10	Ice
2324	2329	-247.10	-30.78	Ice
2334	2344	-246.32	-30.22	Ice
2366	2370	-242.92	-29.80	Ice
2380	2385	-239.79	-28.04	Ice
2570	2573	-235.25	-27.57	Ice
2583	2588	-228.33	-25.14	Ice
2593	2596	-243.75	-28.90	Ice
2596		-258.60	-31.85	Sediment pore water
2603	2608	-251.73	-31.16	Ice
2608	2613	-245.73	-29.17	Ice
2618	2623	-243.03	-28.42	Ice
2644		-215.29	-24.46	Sediment pore water
2653	2658	-235.92	-27.88	Ice
2666	2671	-241.52	-28.40	Ice
2676	2681	-247.91	-30.57	Ice

TABLE VI. ANION AND CATION CONCENTRATIONS OF LAKE VIDA ICE SAMPLES 2010

Top depth (cm)	Bottom depth (cm)	Chloride (µM)	Sulfate (µM)	Sodium (µM)	Potassium (µM)	Magnesium (µM)	Calcium (µM)
995	1004	0.34	0.00	0.85	0.14	0.00	0.42
1005	1015	0.36	0.00	0.85	0.17	0.03	0.07
1015	1019	0.36	0.00	1.07	0.22	0.02	0.41
1019	1025	0.26	0.09	1.26	0.24	0.02	0.38
1035	1043	0.20	0.00	0.23	0.09	0.00	0.02
1043	1048	0.27	0.00	0.35	0.13	0.00	0.00
1048	1052	0.16	0.00	0.43	0.12	0.00	0.24
1052	1057	0.25	0.00	0.62	0.18	0.00	0.17
1057	1065	0.31	0.00	0.60	0.14	0.00	0.42
1068	1076	0.52	0.00	0.88	0.15	0.04	0.37
1076	1083	0.34	0.00	0.67	0.16	0.00	0.31
1083	1087	0.44	0.00	0.77	0.18	0.00	0.17
1087	1091	0.33	0.00	0.76	0.19	0.00	0.29
1094	1097	0.72	0.00	1.03	0.22	0.04	0.27
1097	1103	2.90	0.47	4.52	0.30	0.02	0.83
1103	1108	0.88	0.00	1.40	0.29	0.04	0.58
1108	1112	0.55	0.00	1.24	0.32	0.00	0.30
1112	1123	0.77	0.08	1.42	0.34	0.02	0.78
1128	1131	0.85	0.11	1.53	0.41	0.07	1.36
1131	1135	0.39	0.08	1.28	0.29	0.02	0.53
1135	1143	0.77	0.23	1.94	0.54	0.22	0.90
1143	1147	1.08	0.00	1.14	0.24	0.09	0.41
1150	1155	0.29	0.00	0.88	0.19	0.00	0.39
1160	1166	0.28	0.00	0.96	0.23	0.00	0.28
1170	1174	0.35	0.00	1.23	0.33	0.00	0.47
1178	1182	0.43	0.00	1.35	0.31	0.00	0.48
1182	1186	0.42	0.00	1.32	0.34	0.00	0.07
1186	1192	0.77	0.09	1.66	0.32	0.00	0.33
1192	1198	0.74	0.09	1.69	0.37	0.01	0.14
1198	1203	0.70	0.10	1.55	0.31	0.00	0.04
1208	1218	1.10	0.10	2.82	0.42	0.83	0.74
1218	1224	0.67	0.11	2.87	0.54	0.18	0.58
1224	1229	0.74	0.22	3.94	0.72	0.48	1.18
1240	1247	3.15	1.37	6.49	0.95	1.62	3.09
1247	1256	1.65	0.47	5.22	0.84	0.37	1.08
1269	1281	15450.48	3810.99	14458.89	905.09	4373.78	3564.11
1307	1314	12.73	2.14	13.61	1.97	5.18	5.81
1314	1319	22.12	0.92	17.05	1.22	6.44	2.36
1322	1328	34.63	3.29	27.96	2.91	10.63	10.76
1308	1313	575.88	37.08	405.83	26.89	174.19	19.28
1355	1365	774.62	59.88	577.27	42.02	223.50	39.75
1387	1392	680.52	44.35	497.56	38.11	199.69	4.67

1392	1397	872.37	60.07	636.44	47.06	257.02	7.41
1400	1405	13.61	3.08	20.58	0.00	32.63	43.52
1405	1410	17.97	3.78	18.82	1.58	10.31	12.11
1410	1415	24.51	6.57	28.33	2.31	16.59	8.89
1415	1419	31.09	6.79	32.46	2.64	14.62	10.62
1423	1429	37.13	9.99	41.70	3.76	15.27	20.34
1443	1449	85.19	18.55	91.49	10.05	28.81	22.54
1449	1454	168.25	50.87	188.12	18.90	51.84	39.80
1454	1459	121.95	47.82	150.01	18.73	38.74	71.71
1459	1463	224.19	70.50	256.58	33.32	79.14	116.40
1467	1472	33.42	8.82	29.97	2.02	16.66	25.83
1486	1490	469.62	82.41	430.96	29.92	130.73	23.62
1498	1503	60.39	24.65	65.45	4.18	33.42	33.50
1503	1510	30.08	5.25	27.97	2.24	15.22	19.87
1531	1535	29.39	1.45	21.06	1.18	20.45	0.00
1535	1541	31.84	1.38	22.82	1.22	17.84	0.00
1541	1547	25.64	1.05	20.17	1.69	15.95	0.00
1547	1552	37.54	2.13	27.82	1.55	19.27	0.72
1567	1574	15.74	1.37	14.07	0.98	12.11	1.08
1576	1582	8.02	1.68	9.07	0.80	8.31	5.50
1584	1589	620.78	314.82	921.43	77.32	204.60	164.64
1589	1593	674.74	324.72	961.60	83.44	205.87	165.69
1594	1599	886.22	391.38	1203.93	101.84	283.10	184.36
1599	1604	1162.85	600.16	1474.80	132.95	346.07	373.67
1604	1608	1245.57	595.89	1659.13	133.44	381.56	263.09
1608	1613	1803.98	550.53	2255.60	190.45	545.96	101.05
1613	1620	979.66	319.53	1197.33	108.22	310.57	99.71
1620	1628	1051.37	395.62	1408.44	114.13	345.99	113.77
1628	1633	13.65	3.84	16.53	1.35	8.16	11.66
1633	1638	42.09	12.64	50.40	4.25	16.23	20.45
1635	1641	2542.69	617.15	2962.00	285.47	1073.65	48.11
1641	1647	2668.79	1085.97	4119.00	308.93	1038.09	47.59
1649	1655	7.55	1.50	8.88	1.04	5.58	1.36
1655	1660	12.30	4.24	15.98	1.24	5.70	11.41
1660	1666	8.47	0.91	8.97	0.96	5.81	2.42
1673	1679	7.76	2.69	12.70	1.43	4.43	9.24
1708	1713	8.96	0.00	10.78	1.39	13.45	0.00
1715	1721	3835.43	442.27	3045.08	299.85	1450.66	39.16
1721	1725	26701.81	9642.62	30948.79	1743.57	7757.67	3548.82
1725	1732	1483.81	787.54	2039.91	98.69	934.25	1391.36
1732	1739	641.03	60.57	511.37	75.35	696.99	0.00
1682	1687	4.49	0.71	5.75	0.42	3.24	1.06
1687	1691	3.01	0.00	3.86	0.36	3.86	1.58
1697	1703	7.45	0.00	7.50	1.07	8.79	0.00
1620	1625	13854.46	285.56	8027.83	361.66	2900.59	187.80
1626	1631	17331.09	384.50	10269.06	448.28	3112.09	179.89
1631	1636	11555.66	205.13	6691.76	297.33	2445.52	128.93
1637	1642	24384.46	407.27	14194.10	624.37	4230.20	280.78

1642	1648	6547.34	152.22	4204.28	194.61	1530.62	87.80
1668	1673	9867.85	227.71	5907.08	254.58	2160.95	110.30
1673	1678	11782.17	218.84	7023.96	298.21	2470.35	174.98
1668	1673	1316.43	53.74	959.64	40.56	401.02	23.69
1673	1678	764.23	31.52	552.30	23.01	230.93	1.43
1678	1683	717.66	32.43	530.42	24.08	239.71	0.00
1683	1688	1118.16	53.09	837.92	36.60	356.03	9.67
1688	1694	714.74	32.84	541.59	21.90	241.12	5.73
1731	1736	10310.17	203.55	6030.92	280.98	2216.91	118.12
1736	1741	4391.52	197.71	3064.26	129.10	1108.00	68.43
1741	1748	4238.19	193.03	2940.05	129.48	1065.69	63.20
1748	1753	3779.09	69.09	2469.34	108.08	936.87	31.58
1753	1758	2108.24	40.22	1424.86	60.46	600.17	18.25
1758	1763	5120.92	577.86	4248.78	249.46	1296.30	500.52
1761	1766	18998.51	321.42	10945.01	482.76	3437.89	237.82
1767	1772	9296.91	176.29	5572.16	237.96	2050.45	110.41
1771	1776	13393.73	226.87	7774.65	341.98	2700.71	180.60
1776	1781	17667.23	343.18	10262.70	459.14	3275.15	214.88
1781	1786	18331.98	347.65	10363.23	461.07	3504.10	230.45
1786	1791	26142.31	503.42	15415.24	689.38	4571.19	188.37
1791	1796	32147.92	617.72	19163.44	834.10	5576.44	246.65
1795	1800	26524.90	502.71	15474.47	692.11	4412.58	199.14
1800	1808	21182.07	423.01	12361.84	547.09	3806.49	302.75
1808	1813	25144.32	620.51	15074.69	666.18	4367.32	207.67
1813	1820	31537.66	638.03	18867.45	850.37	5299.16	222.22
1820	1826	22420.06	2968.13	18022.10	819.88	3918.32	334.04
1835	1840	14248.08	7665.95	20808.33	525.01	3042.38	1618.50
1846	1851	6435.27	259.44	4322.96	181.94	1756.58	159.35
1851	1856	1205.49	85.34	959.77	57.81	353.94	17.53
1856	1861	552.56	60.81	515.55	24.22	152.41	11.57
1861	1866	6553.07	221.98	4263.37	184.52	1435.79	100.02
1866	1871	7827.03	355.16	5178.41	233.56	1584.45	103.43
1871	1876	1303.95	200.71	1323.70	105.33	361.80	13.85
1876	1881	3980.18	186.67	2878.79	184.32	984.96	39.78
1877	1882	2817.28	208.73	2290.97	149.84	771.59	25.91
1882	1887	3064.00	184.74	2338.74	124.24	756.02	45.61
1887	1892	5522.94	1181.67	5797.12	184.56	1264.64	128.79
1892	1897	2561.36	1368.69	4357.54	82.16	562.74	82.09
1897	1902	1444.56	705.47	2480.62	50.55	424.92	83.44
1902	1907	1020.50	287.51	1354.64	56.23	270.71	99.52
1907	1912	653.82	192.52	842.82	45.82	189.79	37.88
1912	1917	1854.79	415.20	2180.26	112.78	474.76	87.77
1917	1922	3177.47	1075.44	3999.05	163.74	696.38	156.55
1922	1927	1620.95	161.72	1284.23	66.21	590.91	78.29
1922	1947	4987.25	160.45	3240.61	136.26	1656.12	0.00
1922	1947	1810.05	83.97	1226.65	62.45	627.07	0.00
1927	1932	2091.25	92.86	1411.15	70.96	758.53	0.00
1947	1955	8465.16	12110.69	29929.59	234.57	2417.11	0.00

1955	1960	2527.18	89.23	1713.16	76.89	1119.52	0.00
1960	1965	3727.89	998.98	4139.30	173.55	1055.46	509.94
1965	1972	4648.91	130.00	2919.31	121.17	1640.98	0.00
1972	1977	1680.10	73.42	1137.76	57.51	658.71	0.00
1977	1982	1299.26	256.09	1372.57	72.56	462.90	71.27
1982	1987	1278.87	74.99	940.78	51.71	520.11	5.00
1987	1992	8118.28	858.93	7216.27	330.43	2399.68	167.41
1992	1997	6294.61	598.42	4986.55	220.06	1888.39	53.96
1997	2002	6143.87	1076.25	5690.04	175.47	1900.47	0.00
2002	2007	15025.37	1642.37	11864.95	415.98	4451.64	0.00
2007	2013	27489.20	2996.63	21636.63	775.42	7214.57	44.52
2013	2018	273800.86	5145.89	154057.02	6800.32	57525.58	1737.93
2156	2161	39705.69	8510.61	39984.95	1127.71	9537.38	279.62
2161	2167	207127.94	5772.64	121101.99	4834.08	42702.93	1540.93
2189	2194	1361.62	276.37	1384.16	36.74	670.99	77.84
2194	2199	6318.68	346.59	4372.54	159.22	2469.82	0.00
2197	2202	9747.85	395.76	6588.39	304.36	2507.37	0.00
2202	2207	25945.20	883.85	16618.93	637.01	6086.92	60.41
2207	2212	28690.09	10671.91	37742.34	729.00	6812.31	97.91
2212	2218	21007.12	2083.20	16351.64	504.52	5120.64	51.68
2218	2223	5901.73	2677.33	8901.03	155.26	1485.86	10.59
2247	2252	9662.77	2042.87	9921.01	228.91	3053.30	0.00
2252	2257	10140.90	1701.49	9303.81	274.41	3325.94	0.00
2257	2262	16169.37	1647.82	12326.34	419.22	4595.16	0.00
2262	2267	29084.56	2487.19	20926.84	786.44	6805.42	110.23
2267	2272	22428.19	10374.25	33421.43	606.12	5369.04	16.21
2280	2283	20528.86	5553.74	22384.89	599.19	5739.13	1162.29
2283	2288	296302.73	5041.35	170307.55	8669.36	62457.80	5348.11
2304	2309	4630.79	15741.76	33587.20	205.51	1391.18	954.94
2309	2314	7184.59	23000.80	49610.23	219.87	1689.74	798.93
2314	2319	12475.97	10098.26	27664.07	297.18	3348.55	541.92
2319	2324	11620.74	5630.61	17873.73	376.86	2793.95	675.02
2324	2329	9092.48	49236.99	100341.36	387.33	2503.78	812.94
2329	2334	3349.57	74204.39	150729.81	120.04	655.57	1896.34
2334	2344	2656.17	53147.05	108164.53	111.67	615.66	1202.24
2356	2361	4299.46	64349.33	128605.48	141.65	1125.27	16.39
2361	2366	1814.52	43614.44	88955.76	62.44	934.79	0.00
2366	2370	7723.57	13773.79	33703.72	290.93	3940.27	0.00
2370	2375	1903.63	60264.67	123782.71	89.65	285.63	974.43
2375	2380	1771.90	14292.25	30178.21	49.08	515.84	165.76
2380	2385	1135.89	19941.96	40509.90	19.82	150.70	682.81
2226	2231	2441.27	1528.38	1668.30	139.51	1292.06	2863.90
2231	2238	1886.01	92.43	1167.48	78.39	1009.99	834.04
2570	2573	4932.99	2838.46	6290.06	235.97	1845.49	1500.13
2578	2583	2450.52	6226.36	12389.13	58.51	695.21	1854.83
2583	2588	3859.21	17539.74	31188.09	145.63	567.79	4807.16
2588	2593	5169.43	242540.09	487073.24	183.24	768.02	4467.31
2593	2596	2452.09	35687.28	73694.09	57.86	495.76	2121.02

2593	2598	5370.87	27550.88	59606.39	186.91	1650.07	1240.29
2598	2603	16520.68	6848.38	21559.15	523.20	4469.28	2078.06
2603	2608	11919.52	11707.08	26895.02	403.00	2578.00	2553.98
2608	2613	4955.52	26181.04	52304.25	149.64	1066.61	2369.67
2613	2618	3067.35	89855.79	185732.41	175.46	789.01	848.61
2618	2623	2521.53	1214.26	3421.92	61.57	855.93	635.57
2647	2650	3781.35	97419.51	196541.01	124.17	788.59	760.77
2650	2653	2775.26	1933.43	4707.93	68.68	1068.26	920.28
2653	2658	4087.07	2576.09	5755.68	111.94	1076.70	1618.67
2661	2666	5591.56	3827.52	8547.21	144.34	1447.09	1721.55
2666	2671	2900.88	4044.55	8158.04	57.56	592.70	1510.80
2671	2676	4476.24	4745.70	11132.36	116.83	1487.26	1192.64
2676	2681	6040.59	7529.34	17773.47	203.70	1673.83	995.21

TABLE VII. GRAIN SIZE, WATER CONTENT AND CARBON CONTENT OF SEDIMENT LAYERS IN THE LAKE VIDA ICE CORE, 2010. WATER PERCENTAGE INCLUDES THE ICE CONTENT SURROUNDING THE SEDIMENT LAYER, AND IS NOT NECESSARILY EQUIVALENT TO POROSITY. TC = TOTAL CARBON, TIC = TOTAL INORGANIC CARBON, TOC = TOTAL ORGANIC CARBON (CALCULATED AS THE DIFFERENCE OF TC AND TIC).

Depth (cm)	Mean Grain Size (µm)	Median Grain Size (µm)	Water %	TC (%)	TIC (%)	TOC (%)
64	317.7	300.2	11.41			
268	290.2	299.3	49.54	0.49	0.07	0.42
395	409.9	375.1	28.69	0.50	0.12	0.38
785	314.8	287.4	49.51	0.45	0.07	0.38
1840	289.5	306.3	38.24	0.70	0.23	0.47
2018	170.4	197.6	26.74	0.92	0.41	0.50
2156	77.5	16.0	59.15	1.56	0.39	1.17
2163	274.9	280.3	18.79	0.46	0.11	0.35
2164	368.7	329.9	10.80	0.45	0.07	0.38
2165	321.2	318.0	8.35	0.50	0.20	0.30
2166	323.2	299.7	9.96	0.55	0.12	0.43
2167	311.0	298.3	8.89	0.53	0.22	0.31
2168	285.9	281.9	8.17	0.58	0.17	0.41
2169	308.4	302.8	8.92	0.56	0.17	0.38
2170	328.9	312.1	8.21	0.59	0.34	0.25
2171	313.0	307.2	8.79	0.62	0.16	0.45
2172	318.1	309.5	9.61	0.68	0.29	0.39
2173	356.1	341.6	9.27	0.56	0.18	0.38
2174	351.9	339.2	9.82	0.64	0.26	0.38
2175	317.9	314.2	6.40	0.69	0.25	0.44
2176	345.7	328.3	6.55	0.68	0.34	0.34
2182	245.4	268.4	9.53	0.63	0.24	0.38
2185	299.8	300.5	9.55	1.05	0.42	0.63
2187	256.7	271.9	12.61	0.60	0.30	0.30
2288	299.0	314.9	8.37	0.62	0.25	0.37
2289	289.5	303.1	9.27	0.72	0.28	0.44
2291	296.1	309.3	9.43	0.73	0.40	0.33
2292	306.5	296.6	10.62	0.50	0.15	0.35
2293	367.4	342.4	9.74	0.44	0.12	0.32
2294	381.2	350.6	9.61	0.59	0.16	0.43
2295	310.4	305.5	9.36	0.66	0.22	0.44
2296	304.9	305.7	10.84	0.98	0.39	0.59
2297	347.1	342.0	21.69	0.74	0.29	0.45
2298	304.8	293.0	32.15	0.69	0.38	0.31
2299	305.7	315.8	39.30	0.53	0.19	0.35

2300	289.8	297.7	31.81	0.40	0.09	0.31
2528	438.7	403.8	3.68	0.46	0.11	0.35
2529	382.6	367.1	4.89	0.60	0.21	0.39
2530	373.8	374.7	6.46	0.47	0.17	0.30
2531	367.9	362.9	6.52	1.18	0.72	0.46
2532	384.2	361.7	7.10	0.48	0.12	0.36
2533	379.9	359.9	8.14	0.67	0.19	0.48
2534	372.4	361.3	7.89	0.82	0.31	0.50
2535	408.7	380.9	8.13	0.42	0.19	0.23
2536	360.4	346.7	7.26	0.46	0.10	0.35
2537	371.6	367.1	7.88	0.51	0.30	0.21
2538	393.9	377.8	8.51	0.42	0.13	0.29
2539	379.4	361.9	7.02	0.46	0.16	0.30
2540	397.8	379.8	7.08	0.43	0.08	0.35
2541	365.4	360.6	4.97	0.40	0.11	0.28
2548	354.9	351.7	12.65	0.46	0.12	0.34
2554	346.0	340.1	8.30	0.68	0.44	0.24
2559	354.0	321.7	8.05	0.74	0.29	0.45
2560	352.6	350.5	9.52	1.05	0.33	0.72
2561	352.1	342.6	8.98	0.45	0.12	0.34
2562	318.2	310.5	9.55	0.59	0.24	0.35
2563	309.8	301.7	8.38	0.77	0.22	0.55
2564	362.3	326.5	9.86	1.09	0.64	0.45
2565	274.0	289.0	8.49	0.66	0.31	0.34
2566	303.9	304.1	6.99	0.98	0.38	0.60
2567	305.8	318.2	8.84	0.63	0.33	0.30
2596	341.9	332.4	47.77	0.59	0.15	0.44
2628	288.9	304.9	39.15	0.56	0.19	0.37
2630	338.0	345.3	8.16	0.55	0.31	0.24
2631	428.2	417.4	7.89	0.68	0.22	0.46
2632	372.0	367.3	7.62	0.77	0.41	0.36
2633	311.7	310.7	8.80	1.09	0.46	0.63
2634	270.5	284.7	9.90	1.10	0.45	0.65
2635	298.2	312.4	9.62	0.97	0.48	0.50
2636	215.2	238.6	10.89	0.90	0.27	0.63
2637	50.0	7.9	8.12	1.17	0.74	0.42
2638	228.6	252.6	9.90	1.00	0.65	0.35
2639	102.2	52.3	9.32	1.34	0.76	0.59
2641	180.5	202.8	10.14	1.09	0.47	0.63
2642	217.0	240.1	11.12	1.49	1.88	-0.39
2643	156.9	157.1	24.58	1.99	2.30	-0.31
2644	156.7	169.0	31.48	1.23	0.44	0.79

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VITA

Hilary Dugan
University of Illinois at Chicago, Dept. of Earth and Environmental Sciences
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EDUCATION

PhD:	University of Illinois at Chicago. Chicago, IL, USA Department of Earth and Environmental Sciences	2010-2014
MSc:	Queen's University. Kingston, Ontario, Canada Department of Geography	2008-2010
BSc(H):	Queen's University. Kingston, Ontario, Canada Physical geography major, biology minor	2004-2008

ACADEMIC APPOINTMENTS

Postdoctoral Associate: Center for Limnology at the University of Wisconsin – Madison Cary Institute of Ecosystem Studies. Millbrook NY	2014-Present

ACADEMIC AWARDS

National and International

GLEON Graduate Student Fellowship	2013-2014
Antarctic Science International Bursary	2012
NSERC Alexander Graham Bell Canadian Graduate Scholarship (Doctoral) -Declined for NSERC Postgraduate Scholarship	2010-2012
ArcticNet Training Fund Award	2010
NSERC Alexander Graham Bell Canadian Graduate Scholarship (Master's)	2008-2010
Ontario Graduate Scholarship (Declined)	2008-2009
Nortek Student Equipment Grant	2008
NSERC Undergraduate Student Research Award (USRA)	2007

Institutional

University of Illinois at Chicago University Fellowship	2010-2014
Chicago Consular Corps Scholarship	
Queen's University Tri-Council Award	2008-2009
Queen's University Medal in Geography -Top standing in department	2008
Queen's University Julian Szeicz Memorial Prize	2008
Queen's University Roscoe R. Miller Prize in Geography	2008
Canadian Association of Geographers Prize	2008
Queen's University Dean's Honour List	2004-2008
Queen's University Thomas Allardyce Brough Prize in Geography	2005-2006

Queen’s University Annie Bentley Lillie Prize in Calculus	2004-2005
Queen’s University Alban H. Norton Entrance Scholarship	2004

Conferences

XXXII SCAR Meeting – Poster Award	2012
Canadian Geophysical Union Annual Meeting - Runner-up Best Student Talk	2010
Canadian Geophysical Union Annual Meeting - Campbell Scientific Best Poster Award	2008

PUBLICATIONS

Dugan, HA, MK Obryk, PT Doran. 2013 Lake ice ablation rates from permanently ice covered Antarctic lakes. *Journal of Glaciology* 59: 491-498.

Dugan, HA, SF Lamoureux, T Lewis, MJ Lafrenière. 2012. The impact of permafrost disturbances and sediment loading on the limnological characteristics of two High Arctic lakes. *Permafrost and Periglacial Processes.* DOI: 10.1002/ppp.173

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Dugan, HA, PT Doran, B Wagner, F Kenig, CH Fritsen, SA Arcone, E Kuhn, NE Ostrom, J Warnock, AE Murray. 2014. 27 m of lake ice on an Antarctic lake reveals past hydrologic variability. *The Cryosphere Discussions* 8: 4127-4158.

Winslow, LA, **HA Dugan**, H Buelow, KD Cronin, JC Priscu, C Vesbach, PT Doran. (In Revision). Autonomous year-round sampling and sensing to explore the physical and biological habitability of permanently ice-covered Antarctic lakes. *Marine Technology Society.*

Mikucki, J, E Auken, S Tulaczyk, RA Virginia, C Schamper, KI Sorensen, PT Doran, **HA Dugan,** N Foley. (In Revision). Ancient groundwater and potential deep subsurface habitats beneath an Antarctic dry valley. *Nature Communications.*

Dugan, HA, AB Santoso, JR Corman, A Jaimes, ER Nodie, VP Patil, RI Woolway, JA Zwart, JA Bentrup, AL Hetherington, SK Oliver, JS Read, KM Winters, PC Hanson, EK Read, LA Winslow, KC Weathers. (In Review). Consequences of gas flux model choice on the interpretation of metabolic balance across 15 lakes. *Limnology and Oceanography: Methods*.

PUBLICATIONS IN PREP

Dugan, HA, PT Doran, S Tulaczyk, JA Mikucki, SA Arcone, E Aucken, C Schamper. (In prep). Subsurface imaging reveals aquifer beneath an ice-sealed Antarctic lake. Will be submitted to *Geophysical Research Letters.*

Obryk, MO , S Niebuhr, B Herried, PT Doran, and **HA Dugan**. (In prep). Hypsometry of Taylor Valley, McMurdo Dry Valleys, Antarctica: Predicting closed-basin spill and merge events. Will be submitted to *Antarctic Science*.

Castendyk, D, H Gallagher, B Lyons, J Priscu, **HA Dugan**, P Doran. (In prep). The origin of shallow currents in a perennially ice-covered lake, Antarctica. *Hydrological Processes.*

CONFERENCE PRESENTATIONS

Dugan, H, AB Santoso, JR Corman, A Jaimes, ER Nodie, VP Patil, RI Woolway, JA Zwart, JA Bentrup, AL Hetherington, SK Oliver, JS Read, KM Winters, PC Hanson, EK Read, LA Winslow, KC Weathers. "Consequences of gas flux model choice on the interpretation of metabolic balance across 15 lakes" 2014 Joint Aquatic Sciences Meeting, Portland, Oregon, May 18-23, 2014 (PAPER).

Dugan, H, MK Obryk, PT Doran. "High-resolution monitoring of long term changes in physical limnology, McMurdo Dry Valleys, Antarctica" 2012 LTER ASM, Estes Park, Colorado, Sep 10-13, 2012. (POSTER)

Dugan, H, MK Obryk, PT Doran. "Ablation rates of permanently ice-covered Antarctic lakes" XXXII SCAR Meeting, Portland, Oregon, July 16-19, 2012. (POSTER)

Kuhn, E, A Murray, **H Dugan**, AS Ichimura, R Edwards, V Peng, C Fritsen, F Kenig, S Young, P Doran. "Microbial life in the iron-rich, anoxic cryobrine of Lake Vida, Antarctica" XXXII SCAR Meeting, Portland, Oregon, July 16-19, 2012. (POSTER *presented by E. Kuhn*)

Dugan, H, PT Doran, B Wagner, S Arcone, C Fritsen, A Murray. "Exposing Lake Vida, Antarctica" NASA Astrobiology Conference, Atlanta, Georgia, Apr 16-20, 2012. (PAPER)

Dugan, H, PT Doran, C Fritsen, F Kenig, A Murray, S Arcone. "A 26 m ice cover on Lake Vida, Antarctica" AGU Fall Meeting, San Francisco, California, Dec 5-9, 2011. (POSTER *presented by P. Doran*).

Dugan, H, PT Doran, C Fritsen, F Kenig, A Murray. "The formation of a 26 m ice cover on Lake Vida, Antarctica" 11th International Symposium on Antarctic Earth Sciences, Edinburgh, Scotland, July 10-16, 2011. (PAPER)

Dugan, H, S Lamoureux, M Lafrenière, T Lewis. "The impact of permafrost disturbances and sediment loading on the seasonal mixing of two High Arctic lakes" The 36th Annual Meeting of the Canadian Geophysical Union, Ottawa, Ontario, May 31-June 4, 2010. (PAPER)

Dugan, H, and S Lamoureux. "The chemical evolution of a hypersaline coastal lake in the High Arctic" The 40th International Arctic Workshop, Winter Park, Colorado, Mar 10-12, 2010. (PAPER)

Dugan, H, T Lewis, S Lamoureux, M Lafrenière. “Investigating the Formation of High Conductivity Bottom Water in a Freshwater High Arctic Lake” Arctic Change 2008, Quebec City, Quebec, Dec 9-12, 2008. (POSTER)

Dugan, H, S Lamoureux, M Lafrenière, T Lewis. “Hydrological and Sediment Yield Response to Rainfall in a High Arctic Watershed” The 34th Annual Meeting of the Canadian Geophysical Union, Banff, Alberta, May 10-14, 2008. (POSTER)

INVITED TALKS

“The formation of Lake Vida” U.S. Army Cold Regions Research and Engineering Laboratory (CRREL), Hanover, NH, April 6, 2011.

THESES

M.Sc. Thesis: Long-term development and recent dynamics of High Arctic coastal basins.
- Queen’s University, Kingston, Ontario. p. 159.

BSc(H) Thesis: Hydrological and sediment yield response to summer rainfall events in a small high arctic watershed.
- Queen’s University, Kingston, Ontario. p. 43.

TEACHING ASSISSTANT

Awards

Department TA of the year – UIC	2012-2013
Department TA of the year – Queen’s University	2008-2009

Courses taught

EAES 111 – Earth, Energy, and the Environment	2013 Winter
EAES 400 – Field Experience in Earth Sciences (Field course to Churchill, MB)	2012 Summer
GPHY 102 – Earth System Science	2010 Winter
GPHY 312 – Watershed Hydrology	2009 Winter
GPHY 209 – Principals of Hydroclimatology	2009 Winter
GPHY 314 – Climate Change	2008 Fall

WORKSHOPS

International Workshop on Tracer Applications of Noble Gas Radionuclides — 2012 June
- Argonne National Laboratory, Illinois

Data Acquisition from Remote Locations Workshop — 2012 June
- University of New Mexico Sevilleta Field Station

Campbell Scientific CR1000/LoggerNet Training — 2011 May
- Logan, Utah

IPY International Polar Field School 2010 July

- The University Centre in Svalbard (UNIS), Norway

FIELD WORK

Taylor Valley, McMurdo Dry Valleys, Antarctica 2013 Oct-Dec
2012 Oct-Dec
2011 Oct-Dec

- In association with the McMurdo Long Term Ecological Research site
- Wrote supported information packages (SIPs) for all field work
- Constructed and maintained limnological stations designed to continuously record temporal variability of Taylor Valley lakes

Lake Vida, McMurdo Dry Valleys, Antarctica 2010 Oct-Dec

- Helped organize logistics for the implementation of a ten-person field camp at Lake Vida
- Recovered two 20 m ice cores from the surface of Lake Vida
- Sampled saline brine using clean access procedures developed to investigate the biological diversity, adaptation, and evolutionary processes of the ecosystem

Shellabear Point, Melville Island, Arctic Canada 2009 June-July

- Arranged research licenses, field equipment and other logistics for remote field camp
- Established and maintained a two-person field camp for five weeks

Cape Bounty, Melville Island, Arctic Canada 2008 May-Aug

- Successfully carried out a 12 week intensive limnological sampling program
- Helped maintain an11 person field operation

Tromsø and Finnmark District, Norway 2008 April

- Assisted Norwegian researcher studying the paleo-lake sediment record along the northern Norwegian coast.
- Sediment cores were removed from several lake basins using piston coring and Russian peat coring techniques.

Cape Bounty, Melville Island, Arctic Canada 2007 May-July

- Involved in the collection of hydroclimatological and biogeochemical data, as well as the monitoring of watershed sediment and nutrient fluxes, over a nine week period
- Knowledge of field skills include: river rating, water sampling, water chemistry analyses, sediment coring, installing, maintaining and downloading data loggers, water column profiling, and remote field logistics and field safety procedures

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20 February 2014

Dear Hilary

We are pleased to give you permission to use a version of your paper in your upcoming PhD thesis. The permission covers this publication, in paper or in such electronic forms as are now known, and in future reprints and all languages.

Dugan HA, Obryk MO, and Doran PT (2013) Lake ice ablation rates from permanently ice covered Antarctic lakes. *J. Glaciol.*, **59**(215), 491–498.

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I would like to congratulate you on the completion of your dissertation.

Sincerely

Magnús Már Magnússon
Secretary General