Hydrological and Biogeochemical Modeling of Taylor Valley Lakes, East Antarctica

BY

MACIEJ KRZYSZTOF OBRYK
B.S., University of Illinois at Chicago, 2008

THESIS
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Defense Committee:

Peter Doran, Chair and Advisor
Andrew Dombard
Fabien Kenig
Roy Plotnick
Chris McKay, NASA Ames Research Center
This thesis is dedicated to my family and the love of my life, without whom it would never have been accomplished.
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MKO
PREFACE

The motivation behind this thesis is to further the knowledge of the sensitivity of the Taylor Valley lakes with respect to changes in climate in the past, present, and future. Each chapter is formatted for a publication (exception is Chapter I and V, which are the introduction and conclusion, respectively), and each chapter provides germane background information for the reader, allowing each chapter to be considered as a stand-alone paper. Chapter II is focused on paleolimnology and the existence of large glacial lakes, Chapter III is dedicated to contemporary physical variations within the perennially ice-covered lake and its affects on its ecosystem, whereas Chapter IV is focused on understanding the drivers of the long-term ice thickness trends.
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<tr>
<td>AUV</td>
<td>Autonomous Underwater Vehicle</td>
</tr>
<tr>
<td>DDAF</td>
<td>Degree Days Above Freezing</td>
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<tr>
<td>DEM</td>
<td>Digital Elevation Model</td>
</tr>
<tr>
<td>ENDURANCE</td>
<td>Environmentally Non-Disturbing Under-ice Robotic AntarctiC Explorer</td>
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<tr>
<td>$K_{\text{PAR}}$</td>
<td>Diffuse Attenuation Coefficient</td>
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<td>LGM</td>
<td>Last Glacial Maximum</td>
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<tr>
<td>MDV</td>
<td>McMurdo Dry Valleys</td>
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<td>PCA</td>
<td>Principal Component Analysis</td>
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<td>(UW) PAR</td>
<td>(Under Water) Photosynthetically Active Radiation</td>
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<td>RMSE</td>
<td>Root Mean Square Error</td>
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<td>TD</td>
<td>Taylor Dome</td>
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<td>TV</td>
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<td>WLB</td>
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SUMMARY

This research is focused on better understanding of past and present hydrological and biological sensitivity of perennially ice-covered lakes in Taylor Valley, McMurdo Dry Valleys, East Antarctica, with respect to changes in climate, summarized in three distinct chapters.

Glacial Lake Washburn was present in Taylor Valley during the Last Glacial Maximum (LGM), despite a significantly cooler climate. Contemporary anomalous warm summer westerly winds are responsible for generating a large volume of melt water during the short austral summers by increasing the degree days above freezing. The high frequency of westerly winds during the LGM was responsible for the formation of large glacial lakes during the LGM. Data from a nearby Taylor Dome ice core record supports windier conditions and a link between Taylor Dome and Taylor Valley was established. However, the surface air temperature increase due to the westerly winds is not preserved in the annually averaged ice core record. Yet, it was the seasonal warming due to increased frequency of anomalous summer winds that contributed to the GLW formation. This analysis suggests that summer air temperature during the LGM were as warm as today.

An autonomous underwater vehicle was deployed under the thick ice cover of West Lobe of Lake Bonney, generating a high-resolution, spatially distributed biogeochemical and physical dataset. Ice thickness varied depending on shading from nearby mountains and sediment accumulation on the surface of the ice. Spatial ice thickness variations controlled available underwater photosynthetically active radiation (PAR) and chlorophyll-a distribution. PAR was negatively correlated with chlorophyll-a, which was attributed to short-term photoadaptation of phytoplanktonic communities.

Ice thicknesses of perennially ice-covered lakes in Taylor Valley were modeled utilizing a one-dimensioned heat equation coupled with the atmosphere and the underlying water column. The long-term ice thickness trends are strongly controlled by the heat content of a lake. Deep lakes with deep-water temperature maximum will either hinder or facilitate ice thickness growth or decay due to the heat flux from below. Shallow lakes are more responsive to climatic changes. Future ice thickness predictions suggest that ice covers of perennially ice-covered lakes can become seasonally ice-free within couple decades.
CHAPTER I
INTRODUCTION

A. Perennially Ice-Covered Lakes

Perennially ice-covered lakes are a common feature of high-latitude regions. These lakes are of interest due to their isolation from direct human contact and rapid response to small changes in climate (e.g., Barrett et al., 2008; Doran et al., 2008; Vincent and Laybourn-Parry, 2008). Arctic and Antarctic perennially ice-covered lakes have many similarities. The perennial ice covers reduce penetrating solar radiation and offer protection from wind-driven turbulent water mixing, allowing simple unique ecosystems to develop (Spigel and Priscu, 1998; Vincent et al., 2008). The most common feature of both northern and southern polar lakes is the dominance of cyanobacteria within the water column (Vincent and Laybourn-Parry, 2008); however, there are also striking differences. Arctic catchment areas are influenced by grazing of animals and presence of terrestrial plants, which can alter the landscapes and hydrologic regimes. Conversely, Antarctic basins are devoid of both terrestrial plants and animal grazing, reducing complexity of the system, as the landscapes are not altered by higher life forms (Gooseff et al., 2011). This prominent difference simplifies the hydrological cycle, which is ultimately controlled by the climate dynamics (Gooseff et al., 2011). Thus, the hydrology of Antarctic lakes provides a simplified case for studies related to response in climatic changes.

Perennially ice-covered lakes in Antarctica are located in ice-free regions; the largest of which is McMurdo Dry Valleys (Doran et al., 2010). The valleys are largely ice-free due to the Transantarctic Mountains, which block the ice sheet flow and decrease precipitation (Gooseff et al., 2011). The predominant features of the valleys are local alpine glaciers, perennially ice-
covered lakes, and dry soils. During the short austral summers, alpine glaciers melt and feed the perennally ice-covered lakes via ephemeral streams that transport the majority of nutrients, with a smaller contribution from shallow underground seeps and snowfall (Chinn, 1993; Gooseff et al., 2011; Levy et al., 2011). The streams mix with liquid water in the lakes due to the development of moats (small areas of open water around the perimeter of the lakes). This thesis is focused on perennally ice-covered lakes in the Taylor Valley, McMurdo Dry Valleys, adjacent to McMurdo Sound and Transantarctic Mountains.

Taylor Valley is a natural laboratory of terrestrial and aquatic ecosystems. The terrestrial habitat is limited to bare soils with invertebrates as the highest life forms, and the aquatic habitat is confined to perennally ice-covered lakes, ephemeral streams, and cryoconite holes inhabited by microbial communities (Barrett et al., 2007; Gooseff et al., 2011). All lakes in Taylor Valley are closed-basin (no outflow) and their ice thickness varies from 3 to 6 m (Doran et al., 2002b; McKay et al., 1985). Biological processes in Taylor Valley are limited by the availability of liquid water, and therefore, hydrologic processes that respond to climatic changes on seasonal and decadal scales are strong controlling factors in biological productivity (Fritsen and Priscu, 1999; Herbei et al., 2010).

**B. Purpose of Investigation**

Taylor Valley lakes have long been recognized to be sensitive to climatic changes. The unique characteristics of Taylor Valley lakes makes them an excellent study site for climate change, as the lakes respond rapidly to small climatic forcing on seasonal and decadal scales (Barrett et al., 2008; Doran et al., 2008; Fountain et al., 1999; Fritsen and Priscu, 1999). However, in order to envisage the future of the lakes’ response to a changing climate, their past
and present hydrologic and physical drivers have to be fully understood. Taylor Valley lakes have been extensively studied, beginning in 1960’s (e.g., Angino et al., 1962; Armitage and House, 1962; Shirtcliffe and Benseman, 1964), followed by the New Zealand Antarctic program starting in 1971 (Chinn, 1993), and then by the McMurdo Long Term Ecological Research project in 1993 (Barrett et al., 2008). However, several key past, present, and future hydrological and physical drivers are not fully understood and are addressed in this thesis.

Mainly, this thesis addresses: 1) a mechanism responsible for melt water production capable of sustaining large glacial lakes during the Last Glacial Maximum, 2) spatial variations of physical and biological properties of a lake, and 3) controlling drivers of the long-term ice thickness changes.

The history of Taylor Valley lakes is long and complex. The origin of the lakes has been suggested to date more than 4.6 Ma BP as a result of an uplift that trapped seawater in the already denuded valleys (Chinn, 1993; Summerfield et al., 1999). Since then, the lakes have been subjected to desiccation and refilling, depending on climatic conditions; however, each lake’s evolution differs, which results in unique characteristics of each lake. Modern geochemical stratification of the lakes is attributed to the drying and subsequent refilling of the basins with melt water (Lyons et al., 1997). In more recent past, during the Last Glacial Maximum, the entire TV was filled by a large glacial lake (Stuiver et al., 1981). Robust evidence for the existence of this large glacial lake exists (Hall and Denton, 2000; Hall et al., 2000; Stuiver et al., 1981; Toner et al., 2012). However, the source of melt water generation during a significantly colder climate than present time has been a conundrum. Contemporary strong summer westerly winds have been proposed to be responsible for sustaining large glacial lakes during the Last Glacial Maximum by prolonging the short melt season (Doran et al.,
Here, the effect of those winds on paleotemperatures and associated melt water production is investigated. The strong summer westerly winds had been extensively studied (e.g., Doran et al., 2002a; McKendry and Lewthwaite, 1990; Nylen et al., 2004; Speirs et al., 2010); however, their impact on the paleoclimate has been significantly underestimated.

Strong westerly winds are responsible for delivering sediment on the surface of the modern ice-covered lakes. By deploying an autonomous underwater vehicle under the perennially ice cover, the effects of the sediment distribution on the surface and within the ice on the biophysical properties of the underlying water column were investigated. Phytoplanktonic communities within the water column of Taylor Valley lakes are highly sensitive to changes in the solar radiation and can be used as an indicator of climatic changes (Fritsen and Priscu, 1999). Sediment accumulation on the surface and within the ice affects optical properties of the ice cover and, by association, penetrating solar radiation. However, until now, the effects of spatial heterogeneity of ice thickness on the photosynthetic communities have not been investigated. Surface sediment and shading from nearby mountains control spatial variations of the ice thickness, which governs penetrating solar radiation and primary productivity of lakes.

The ice thickness of the Taylor Valley lakes has been fluctuating. The thicknesses have been generally increasing since the 1900’s, followed by a decrease in early- to mid-2000’s. Those trends are assumed to reflect changes in the climate dynamic, as ice thicknesses are driven by surface energy balance (Launiainen and Cheng, 1998). However, the ice thickness of different lakes within Taylor Valley responded to climatic shifts at different times, indicating that surface energy balance is not the only controller of ice thickness. A numerical model was developed to understand the response of different ice thickness trends. The long-term ice
thickness trends are strongly controlled by the heat flux from the underlying water column. These findings have broad implications for using ice thickness trends of perennially ice-covered lakes as indicators of climatic change.
REFERENCES


CHAPTER II

THE INFLUENCE OF WESTERLY WINDS IN ANTARCTICA ON THE PRESENCE OF GLACIAL LAKE WASHBURN AND PALEOTEMPERATURES DURING THE LAST GLACIAL MAXIMUM

Chapter II in preparation for publication in Arctic, Antarctic, and Alpine Research as:

Obryk, M. K.\textsuperscript{1}, Doran, P. T.\textsuperscript{1}, McKay, C. P.\textsuperscript{2}, and Waddington, E. D.\textsuperscript{3}

\textsuperscript{1}Department of Earth and Environmental Sciences, University of Illinois at Chicago, Chicago, Illinois, USA

\textsuperscript{2}Space Science Division, NASA Ames Research Center, Moffett Field, California, USA

\textsuperscript{3}Earth and Space Sciences, University of Washington, Seattle, WA, USA
A. **Abstract**

Large glacial lakes, including Glacial Lake Washburn, were present in the McMurdo Dry Valleys, Antarctica during the last glacial maximum (LGM) despite a colder and drier climate. To address the mechanism capable of generating enough melt water to sustain these large lakes, a conceptual model was developed based on the warming potential of contemporary westerly winds that descend from the Antarctic plateau. The model suggests that westerly winds are capable of generating enough melt water to sustain large glacial lakes during the LGM by increasing degree days above freezing (DDAF), and prolonging the melt season. A present-day relationship between infrequent summer westerly winds and DDAF was established ($r = 0.84$, $p < 0.001$, $n = 17$). It is assumed that the Taylor Dome ice core record represents large-scale paleoclimatic variations for the McMurdo Dry Valleys region. A cross-correlation analysis of contemporary westerly winds and the associated adiabatic warming between Taylor Dome and McMurdo Dry Valleys supports this assumption and is presumed to have operated during the LGM. This analysis suggests that summer temperatures during the LGM were equivalent to present-day values due to the warming influence of the westerly winds that is not preserved in the annually averaged ice core record.
B. Introduction

The McMurdo Dry Valleys, in East Antarctica, is one of the most studied regions in Antarctica. The area, approximately 4800 km², is ice-free due to the Transantarctic Mountains, which effectively block the ice sheet flow and decrease precipitation (Doran et al., 2002a; Gooseff et al., 2011; Levy, 2013; Wharton et al., 1993). Common features in these valleys are perennially ice-covered lakes located in closed basins. Due to a lack of outflow, a delicate climatic balance regulates the presence of these lakes. The existence of fresh liquid water in a climate with mean annual valley-bottom temperatures ranging from -14.8°C to -30.0°C (Doran et al., 2002a) requires summer temperatures sufficiently high to induce glacial melt as the major source of liquid water, yet low enough to maintain the ice covers year round (Chinn, 1993). Cold winter temperatures allow for the replacement of the lake ice cover that is lost during the summer by evaporation and sublimation (Chinn, 1993).

During the last glacial maximum (LGM), the McMurdo Dry Valleys are hypothesized to have contained large glacial lakes, even though the climate was apparently cooler and more arid than today (Hall et al., 2000; Hall et al., 2002; Steig et al., 2000). Hyper-arid conditions during the LGM might appear to be inconsistent with the development of large glacial lakes, which presents a conundrum as to the mechanism responsible for generating enough melt water to sustain such large glacial lakes. Hall et al. (2010) and Hall and Denton (1996) suggested that hyper-arid conditions would expose more bare ice on glaciers, and decreased precipitation would expose more sediment, both acting to reduce albedo and enhance melt. Alternatively, Doran et al. (2002a) proposed warmer surface air temperatures to be responsible for increased melt. Hyper-arid conditions and warmer surface air temperatures are not mutually exclusive (Hall et al., 2010); however, a mechanism for increased air temperatures is required.
In this paper, I investigate surface air temperature warming by the anomalous summer westerly winds in the present-day climate and the potential role of warm westerly winds on the existence of Glacial Lake Washburn that was located in Taylor Valley, McMurdo Dry Valleys, East Antarctica during the LGM (Figure 1). I developed a conceptual model based on the impact of the anomalous westerly summer winds that can explain a mechanism responsible for generation of enough melt water during the LGM to sustain large glacial lakes. In addition, I examined the preservation potential of those winds in the Taylor Dome ice core record (the closest paleoclimatic proxy to McMurdo Dry Valleys) and the implication for interpretations for local paleotemperatures during the LGM.

Figure 1. Map of the McMurdo Dry Valleys region and approximate location of Taylor Dome. Contour lines represent elevation of Ross Sea Ice Sheet during the Last Glacial Maximum that dammed the mouth of Taylor (based on Denton and Marchant (2000)). The ice dam (not shown) allowed for filling Taylor Valley with melt water to form Glacial Lake Washburn.
C. Study Site

Taylor Valley has two main basins, the Fryxell basin to the east and Bonney basin to the west, which are separated by an elevation threshold of 156 m asl. Fryxell basin is characterized by a broad valley that opens to McMurdo Sound with a sill near the coast at 74 m asl. Lake Fryxell and Lake Hoare (current elevation 18 and 74 m asl, respectively) occupy this end of the valley and are separated by the Canada Glacier (Figure 2). Bonney basin is characterized by a narrow, steep-sided valley occupied by Lake Bonney (current elevation ~65 m asl), which is in contact on the western end with Taylor Glacier (Figure 2). Lakes are closed basin, where the only loss of water from the lakes is through sublimation of the ice covers and evaporation at moats, which are small areas of open water between the ice cover and the shore, which allows stream flow to mix with liquid water in the lake. All of the melt occurs in the summer months of December and January.

Figure 2. Map of Taylor Valley showing locations of 1) Lake Fryxell, 2) Lake Hoare, and 3) Lake Bonney.
During the LGM, Taylor Valley was filled by Glacial Lake Washburn, which is suggested to have reached a maximum elevation of 310 m asl (Stuiver et al., 1981), 336 m asl (Hall et al., 2000), and 300 m asl for western Taylor Valley and 120 m asl for the eastern Taylor Valley (Toner et al., 2012). The current 74 m asl spill point of Taylor Valley into McMurdo Sound indicates a requirement for an ice dam during the LGM in order to maintain a 300 m+ elevation for Glacial Lake Washburn (Hall et al., 2000; Stuiver et al., 1981; Toner et al., 2012). The ice dam was generated by a drop in sea level of 120 m (Hall and Denton, 2000; Hansen et al., 2007) that led to the expansion and grounding of the Ross Ice Sheet into Taylor Valley (Stuiver et al., 1981).

Grounding of the Ross Ice Sheet is inferred from the distribution of the Ross Sea drift, which is distinguished by the presence of kenyte and basalt erratics, normally not found in Taylor Valley (Hall et al., 2000; Stuiver et al., 1981). Kenyte is a volcanic mineral found only on Ross Island, located approximately 100 km east of Taylor Valley, and could only have been transported to Taylor Valley by a larger ice sheet (Figure 1) (Hall et al., 2000; Stuiver et al., 1981). Evidence from Ross Sea drift suggests that the Ross Ice Sheet abruptly shifted flow from northward toward the mouth of Taylor Valley to the west (Hall et al., 2000). The erratics are found at maximum elevation of 350 m asl, marking the elevation limit of Ross Ice Sheet in Taylor Valley, providing a constraint for the grounding of the Ross Ice Sheet (Hall et al., 2000; Stuiver et al., 1981). However, the extent of the ice dam has been debated. Stuiver et al. (1981) and Hall et al. (2000) suggested that Ross Ice Sheet terminated at the mouth of Taylor Valley. Whereas, Toner et al. (2012) proposed that Ross Ice Sheet advanced to the Lake Bonney basin, occupying the entire eastern Taylor Valley, followed by retreat of Ross Ice Sheet and formation of a smaller proglacial lake in the eastern Taylor Valley. The location of the Ross Ice Sheet
influences the size of Glacial Lake Washburn and therefore, the minimum and maximum Ross Ice Sheet extents will be considered in this paper; herein referred to as big (Hall et al., 2000) and small Glacial Lake Washburn (Toner et al., 2012), respectively.

Taylor Valley was dammed from ~23,000 to 8340 $^{14}$C yr BP (Hall and Denton, 2000). The timing of Glacial Lake Washburn high stands are inferred from the radiocarbon dating of benthic algal mats along ancient shorelines and deltas on the slopes of Taylor Valley, and soluble salt distributions (Hall and Denton, 2000; Toner et al., 2012). Of 232 radiocarbon dated algal samples, 119 samples require the existence of an ice dam (Hall and Denton, 2000). Soluble salts accumulate along the wetted margin of the lakes through evapoconcentration (Barrett et al., 2009) and are indicative of maximum paleolake level stands (Toner et al., 2012).

Geologic and geomorphic data provide robust evidence for the existence of Glacial Lake Washburn (Hall and Denton, 2000; Hall et al., 2000; Stuiver et al., 1981; Toner et al., 2012); however, the climate reconstructions propose that the McMurdo Dry Valleys experienced a colder, drier, and windier climate during the LGM. The $\delta^{18}$O record from Taylor Dome ice core at the head of the McMurdo Dry Valleys (Figure 1) indicates that the annual average temperature was 4-8 °C colder during the LGM than at present (Steig et al., 2000) (Figure 3). The expansion of the Ross Ice Sheet displaced the moisture source ~1000 km farther away than present (Hall et al., 2010; Hinkley and Matsumoto, 2001). This ice sheet expansion reorganized synoptic weather systems and as a result, changed trajectory of moisture bearing storms, based on Taylor Dome radar stratigraphy (Morse et al., 1998).
Figure 3. Taylor Dome $\delta^{18}$O record. The depletion of $\delta^{18}$O during the Last Glacial Maximum (LGM) indicates temperatures were 4-8 °C colder than present day (Steig et al., 2000).

Currently, the majority of the moisture received at Taylor Dome is from the southerly wind trajectory. However, during the LGM, this pathway was either reduced or eliminated, significantly decreasing precipitation (Morse et al., 1998). It is important to consider that low precipitation rates at Taylor Dome during the LGM are not necessarily indicative of more arid conditions in McMurdo Dry Valleys. Moisture source for local alpine glaciers in McMurdo Dry Valleys is derived from the Ross Sea, delivered by easterly winds (Fountain et al., 1999). The easterly wind pattern was most likely still active during the LGM, due to a thermal contrast between cold Ross Ice Sheet and relatively warm McMurdo Dry Valleys. Yet, during the LGM, an expanded Ross Ice Sheet would displace the moisture source significantly (Hall et al., 2010). Hence, despite the presence of easterly winds during the LGM, the McMurdo Dry Valleys could have received less moisture then present. The resulting hyper-arid climate likely led to the retreat of alpine glaciers located in the Taylor Valley, in comparison to their present day
position (Hall et al., 2000; Higgins et al., 2000). The Taylor Dome record also exhibits high levels of aerosol loading, such as dust (Ca) and salts (Na) from ~27 ka to 12 ka BP, during the highest stands of Glacial Lake Washburn (Grootes et al., 2001; Hinkley and Matsumoto, 2001; Steig et al., 2000), indicative of persistent windier conditions (Hinkley and Matsumoto, 2001; Steig et al., 2000). The dominance of dust over salts is consistent with expansion of Ross Ice Sheet (Hinkley and Matsumoto, 2001). Additionally, comparison of layer-thicknesses in the ice core vs. concentration of $^{10}$Be suggests significant wind scouring during the LGM, further supporting windier conditions (Morse et al., 2007).

It should be noted that Taylor Dome ice core record reflects large-scale wind patterns that are intricately linked with coastal near-surface westerly winds (discussed in next section). Models of large-scale meridional circulation show that the rising air north of Antarctica is transported southward (Parish and Bromwich, 2007). The air parcel reaching the plateau, subsides and feeds near-surface airflow northward, toward the coastline (Parish and Bromwich, 2007). The wind direction and speed is mainly controlled by the ice sheet topography. The strongest winds are experienced near the coastal regions due to steeper topography and confluence zones, where winds move downslope (Parish and Bromwich, 1987). It has been proposed that the confluence zones can enhance cyclone formation north of the continent, which further enhances the near-surface wind flow near the coasts (Parish and Bromwich, 1998). However, this positive feedback can only last for few days, until the source of the cold continental air is depleted (Parish and Bromwich, 1998). The persistent windier conditions suggested by the Taylor Dome ice core record during the LGM indicates that the replenishment of cold continental air masses was more frequent, resulting in more frequent and stronger down valley winds.
1. **Modern anomalous summer westerly winds**

The McMurdo Dry Valleys climate is punctuated by episodic events of strong westerly summer winds that are associated with adiabatic warming and low relative humidity (Doran et al., 2008; McKendry and Lewthwaite, 1990; Nylen et al., 2004; Speirs et al., 2010). The westerly winds originate on the Antarctic plateau where a high loss of long-wave radiation effectively cools near-surface air (Ishikawa et al., 1982; Nylen et al., 2004), creating an inversion. The cool and dense air masses undergo compressional warming (due to dry adiabatic laps rate) as they descend to the bottom of the valleys (Ahrens, 1994; Speirs et al., 2010). The descending air parcel can disturb the inversion, which further increases the air parcel’s temperature (Nylen et al., 2004). The occurrence and strength of the summer westerly winds are driven by a pressure gradient of low-pressure (cyclonic activity) due east of the McMurdo Dry Valleys over the Ross Sea, and a high-pressure system over the Antarctic Plateau due west of McMurdo Dry Valleys (Doran et al., 2008; Parish and Bromwich, 1998; Speirs et al., 2010). Strong summer westerly events cause an anomalous increase of valley-bottom temperatures, which usually persists for a few days following a wind event (Nylen et al., 2004). Based on a twelve-year study, the average summer temperature would be 0.1-0.4 °C cooler without westerly winds (Nylen et al., 2004). Enhanced glacial melt associated with higher temperatures results in high stream flow and an increase of lake levels. The summer westerly winds are rare, but their effect on local hydrological processes is profound.

The summer of 2001/2002 (herein referred to as the flood year) experienced an unusually large increase (up to ~1m) in Taylor Valley lake levels (Doran et al., 2008). The main discriminating factor between the flood year and the extremely cold and dry summer the previous year (non-flood year) was an increase in summer westerly wind frequency and a
resulting increase of mean summer temperatures by 2.4 °C (Doran et al., 2008). However, it has been suggested that the ozone hole contributes to seasonal melt (Jaros et al., in preparation). The ozone hole allows more solar radiation to penetrate deeper into the glaciers, allowing the sub-glacier melt water to reach the lakes earlier in the season (Jaros et al., in preparation). The combination of persistent westerly winds and ozone hole may be responsible for relatively large seasonal melt water inflow. The maximum summer temperatures have been recognized to increase degree days above freezing (DDAF), which extends the duration of the normally short time period when the temperatures are high enough to generate glacial melt that feeds into the lakes (Doran et al., 2008). The flood year experienced 56.8 DDAF, prolonging the melt season, which resulted in significant increase of lake levels (Doran et al., 2008). This compared to the previous non-flood year experienced only 5.4 DDAF and a very small lake level change (Doran et al., 2008). The two seasons are distinguished by the strength of the pressure gradient (large gradient during the flood year), which ultimately drives strong westerly winds from the plateau toward the coast (Doran et al., 2008; Speirs et al., 2010). I hypothesize that high frequency and seasonal persistence of summer westerly winds during the LGM (such as the flood year); a time when there was significantly more low elevation ice in the Taylor Valley than today (i.e., the Ross Ice Sheet damming the east end of Taylor Valley), is responsible for the increased melt water and for filling the Taylor Valley basin to Glacial Lake Washburn size.
D. Methods

Meteorological data (surface temperature and wind speed/direction) pertaining to this study were obtained from Lake Hoare, Lake Bonney, and Taylor Dome meteorological stations. Taylor Valley data used in this study are collected by Campbell Scientific dataloggers (CR10 and 21X) at 30 s sampling intervals and stored as averages between 15 to 20 minutes (Doran et al., 1995; Doran et al., 2002a). Air temperature is measured at 3 meter height using Campbell 207 and 107 temperature probes (Doran et al., 1995). Wind speed and direction is obtained with R. M. Young model 05103 wind monitors (Doran et al., 1995). Taylor Valley data is available at the McMurdo Long Term Ecological Research project database (www.mcmlter.org). Taylor Dome data were collected using Campbell Scientific CR10 dataloggers and averaged over 3 hr intervals. Air temperature was measured at 3 meters height using Thermometrics NTC glass-enclosed thermistors. Wind speed and direction was obtained with R. M. Young model 05103 wind monitor. Several meteorological stations were located across Taylor Dome and show distinct microclimates that cannot be explained by latitude and altitude alone (Waddington and Morse, 1994). However, due to limited overlapping data with Taylor Valley’s meteorological stations, Taylor Dome data used in this study is obtained from 20C weather station only (located near the center of Taylor Dome (Waddington and Morse, 1994)). Within Taylor Valley, Lake Hoare has the longest uninterrupted record, whereas the Lake Bonney station is located the farthest from the coast. Lake Bonney station is least influenced by the coastal winds due to its proximity to the Antarctic plateau and associated westerly winds, and topographic separation between eastern and western Taylor Valley (Doran et al., 2002a). As such, Lake Bonney station resembles the LGM configuration when sea ice extent was significantly larger. Summer westerly events are defined as westerly winds with wind speed exceeding 5 m s$^{-1}$, and lasting
longer than 6 hrs (Nylen et al., 2004; Speirs et al., 2010). DDAF and wind run (total amount or distance of the wind travel over a certain point for the time period in question) for summer westerly events only were calculated for each austral summer (DJF) from 1995 to 2012.

Years with a high frequency of summer westerly wind events are associated with large lake-level increases in Taylor Valley (Doran et al., 2008). To convert lake level increases to seasonal volumetric changes, high-resolution hypsometric data were generated for each basin in Taylor Valley. An integration of hypsometric data allowed for conversion of elevation (m asl) to volume. To reconstruct Glacial Lake Washburn, a Digital Elevation Model (DEM) of the entire Taylor Valley with 30-meter resolution was modified by incorporating an ice dam at a fixed elevation of 350 m asl to account for the Ross Ice Sheet expansion into Taylor Valley during the LGM. Big Glacial Lake Washburn was reconstructed by georeferencing schematic representing Ross Ice Sheet extent in Taylor Valley (obtained from Hall et al. (2000)) over the DEM of Taylor Valley. Small Glacial Lake Washburn boundary was set at the divide between eastern and western Taylor Valley (based on Toner et al. (2012)). The cumulative volume and surface area of the two Glacial Lake Washburn basins were calculated in ArcGIS 10.1.
E. Results and Discussion

1. Conceptual model

For the lake to be sustained, inflow has to be equal to outflow. Lake volume changes can be expressed as:

\[
\frac{dV}{dt} = GQ + Q_{str} + Q_{seep} + \left[ P - S - \dot{E} \right] A
\]

where \( \frac{dV}{dt} \) is the volume change rate, \( GQ \) is direct glacial melt, \( Q_{str} \) is inflowing stream discharge, \( Q_{seep} \) is shallow ground water recharge, \( P \) is precipitation, \( S \) is sublimation, \( \dot{E} \) represents evaporation, and \( A \) is surface area of the lake. Large lakes, such as Glacial Lake Washburn, require either an increase in melt water, or a decrease in the loss terms in equation 1, in comparison to todays levels. The flood year is an example of how melt water can increase significantly, due to an increased frequency/intensity of summer westerly winds. I use the flood year as an analog for conditions that may have sustained large glacial lakes during the LGM. In this model, it is assumed that outflow rates (\( S \) and \( \dot{E} \)) remain constant per unit area; hence the total outflow volume will vary with surface area of the lake. The surface area ratio of big and small Glacial Lake Washburn to the sum of surface area of modern lakes in Taylor Valley is 12:1 and 3:1, respectively. Hence, outflow volume for Glacial Lake Washburn is between 12 and 3 times larger than contemporary lakes, which represents the factors by which inflow would need to increase from today in order to keep lake levels stable at the higher elevations.

The ratio of volume of melt water generated during the flood year to non-flood year is ~3:1. During the non-flood year, lake levels remained largely unchanged, so the inflow balanced the outflow. So the flood year analog, if sustained over time, could create the smaller Glacial
Lake Washburn extent, because the flood year generated just enough water required to sustain Glacial Lake Washburn in a steady state. However, it is important to note that during the LGM there was significantly more low-elevation ice provided by the Ross Ice Sheet at the east end of the valley than today. The low-elevation ice contributes considerably to the seasonal melt (Doran et al., 2008), further supporting the conceptual model. The ice dam topography, as proposed by Denton and Marchant (2000), suggests that Ross Ice Sheet was sloping toward the Taylor Valley, significantly increasing the catchment area of Glacial Lake Washburn and melt water production potential (Figure 1). This additional melt could have potentially supported larger and/or small Glacial Lake Washburn with higher outflow rates.

2. **Numerical model**

The hydrologic budget of Taylor Valley lakes is difficult to determine because inflow and outflow variables cannot be constrain individually (based on current available data). However, equation (1) can be solved numerically using an iterative approach. Recognizing that surface area of a lake \( A \) at a specific height \( h \) times change in elevation equals to change in volume:

\[
dV = A(h)dh 
\]

\[
(2)
\]

and expressing equation (2) as a function of time:

\[
\frac{dV}{dt} = A(h) \frac{dh}{dt}
\]

\[
(3)
\]

allows equation (1) to be rewritten as:
\[ \frac{dh}{dt} = \frac{GQ + Q_{str} + Q_{seep}}{A(h)} + [P - S - E] \quad (3). \]

Given tabulated basin’s hypsometry (A) and assumed outflow rate (S and E), maximum elevation of Glacial Lake Washburn can be determined based on the inflow rate set to flood year volumetric change (GQ + Q_{str} + Q_{seep}). The numerical model is run over 12 thousand years; from 22 ka BP to 10 ka BP. Inflow is simulated using a transient spline function with maximum flow rates at 16 ka BP (flood year volume). Outflow rate (S and E) for the duration of the each model scenario is assumed constant. The lake level changes for small Glacial Lake Washburn are summarized in figure 4. Figure 4 shows high sensitivity of the lake level to the outflow rates because outflow rates are a function of surface area of the lake. Figure 4 validates the conceptual model and provides minimum boundary conditions for the existence of small Glacial Lake Washburn. In order to sustain small Glacial Lake Washburn with flood year inflow rate only, outflow rate needs to be equal or less than 0.5 m y^{-1}. Higher outflow rates would require additional melt water that can be provided from the Ross Ice Sheet.

The minimum time to fill Taylor Valley to small Glacial Lake Washburn can be calculated by setting an inflow rate as a constant instead of spline function. Assuming that flood-year like conditions were persistent each season, it would take approximately 900 years to flood Taylor Valley to the extent of the smaller Glacial Lake Washburn. However, considering that Glacial Lake Washburn existed for >10k years, this approximation is within a reasonable temporal range.
Figure 4. Small Glacial Lake Washburn elevation over time assuming flood-year like conditions during the Last Glacial Maximum. Inflow rate is simulated using a transient spline function with maximum inflow rate equal to flood-year volume change. Elevation of the lake is highly sensitive to outflow rates that are a function of surface area of a lake. Outflow rates are assumed constant for the duration of each model run.

3. **Taylor Dome and Taylor Valley climatic connection**

   a. **Contemporary westerly winds connection**

   The relationships between DDAF and maximum temperature, and westerly winds and temperature have been well documented (Doran et al., 2002a; Doran et al., 2008; McKendry and Lewthwaite, 1990; Nylen et al., 2004; Speirs et al., 2010). Yet, the relationship between summer westerly winds and DDAF has not been quantified despite the significant impact of DDAF on the local climate in Taylor Valley. Westerly winds are frequently experienced during the winter and can increase surface air temperature by more than 40 °C (Speirs et al., 2010) in several days. Although rare, summer westerly winds have a large impact on local hydrological and biological processes by increasing DDAF and by association
enhancing the melt season. A correlation between summer westerly winds at Lake Hoare meteorological station between 1995 and 2012, quantified as wind run, and DDAF was found ($r = 0.84, p < 0.001, n = 17$). The positive correlation indicates that westerly winds are associated with an increase of DDAF. The associated increase of DDAF is responsible for the increase of melt water that enters the lakes (Doran et al., 2008).

Persistence of summer westerly winds during the LGM would increased DDAF, and produced the required melt water for Glacial Lake Washburn. The ice core record obtained from Taylor Dome suggests high intensity/frequency of winds during the LGM based on the high aerosol loading (Ca and Na) and concentrations of $^{10}$Be (Hinkley and Matsumoto, 2001; Morse et al., 2007). It has been assumed that the Taylor Dome ice core record is representative of paleoclimate in McMurdo Dry Valleys during the LGM (Steig et al., 2000). Here, I test this assumption by comparing modern meteorological data from Taylor Dome and Taylor Valley as an analogue for conditions that might have existed during the LGM.

A case study of time series of wind speed and surface air temperature obtained from Taylor Dome and Taylor Valley is shown in Figure 5, as an example how westerly winds can increase temperature above freezing during the late spring. The temperature warming persists for several days following the first westerly wind event, lengthening the duration of the melt season. Cross correlation analysis of data used in Figure 5 indicates that the westerly winds experienced at Taylor Dome reach Taylor Valley approximately 21 hours later (Figure 6). In Figure 6, both temperature and wind speed show peaks at lag 7, which corresponds to a 21-hour delay between the two locations (1 lag = 3 hours). In addition, temperature also exhibits a strong peak at lag 0 indicating a diurnal cycle. The onset of westerly winds is associated with temperature increase (of the same magnitude) in both locations following a time lag (Figure 5
and 6). The above analysis suggests that the winds experienced at Taylor Dome and in Taylor Valley are of the same origin and their associated warming effect is experienced at both locations.

Figure 5. An example of late spring warming due to the westerly winds that descent from the Antarctic Plateau. Top panel: temperature at Taylor Dome (TD) and Lake Bonney, Taylor Valley (TV). Bottom panel: wind speed at Taylor Dome and Lake Bonney, Taylor Valley. Westerly wind speed increase raises the temperature at both locations.
Figure 6. Cross-correlation analysis of wind speed and surface air temperatures between Taylor Dome and Lake Bonney, Taylor Valley. Each lag corresponds to 3 hours. Top panel: wind speed increase experienced at Taylor Dome reaches Taylor Valley 21 hour later (lag 7). Bottom panel: temperature increase is observed at lag 7 (21 hour later) at both locations. In addition, strong temperature correlation at lag zero indicated diurnal cycle.

b. **Oxygen isotope modification**

Interpretation of the paleotemperature record from ice cores is complex. Paleotemperature reconstruction is based on a linear relationship between $\delta^{18}O$ and the mean annual surface air temperatures at the sampling location (Jouzel and Merlivat, 1984). This relationship is assumed to hold true spatially and temporally (Jouzel et al., 1997). Hence, the
Taylor Dome $\delta^{18}$O record will reflect paleotemperatures of high elevation (2450 m asl (Morse and Waddington, 1992)). Moreover, precipitation and oxygen isotope fractionation in Antarctica are assumed to be at the cloud base, above the inversion layer, and are a biased representation of surface temperatures (Jouzel et al., 1997; Jouzel and Merlivat, 1984; Robin, 1977). Westerly winds on the Antarctic Plateau and in the McMurdo Dry Valleys are boundary layer phenomena. Parish and Bromwich (2007) showed that westerly winds on the plateau extended vertically to 1000 m above the ground, with the strongest winds experienced at < 200 m above the ground. Bromley (1985) showed that in Wright Valley, McMurdo Dry Valleys, the vertical extent of westerly winds was confined to a layer 600 m thick. So during westerly wind events, the warming would not reach the cloud base, and would not be represented in the oxygen isotope record.

Oxygen isotope modification can arise during the fallout as the droplets interact with surrounding vapor (Jouzel and Merlivat, 1984). In Antarctica, this isotopic modification does not take place as the fallout is in the form of the ice crystals (Jouzel and Merlivat, 1984). Postdepositional alterations of stable oxygen isotopes, on the other hand, are known to occur in the Antarctic and Greenland ice cores (Town et al., 2008). Such alterations are influenced by wind speeds, snow accumulation rates, and temperatures (Town et al., 2008). Low accumulation rates during the LGM prolong exposure of surface snow to forced ventilation over the course of multiple summers, which results in $\delta^{18}$O enrichment during subsequent summers (Town et al., 2008). However, in Antarctica during the LGM, summer $\delta^{18}$O enrichment is offset by depletion during the following winter (Town et al., 2008). A combination of low temperatures and high winds speeds bring depleted winter moisture, modifying winter snow, which has $\delta^{18}$O values similar to the preceding summer (Town et al., 2008). Because the seasonal variability of $\delta^{18}$O in
Antarctica during the LGM is similar to present day (Town et al., 2008), it is reasonable to assume that seasonal climatic variations during the LGM were simply greater.

The Taylor Dome ice core record for the LGM is in annual averaged temporal scale (Steig et al., 2000), hence the surface air warming associated with low-level westerly winds during the austral summers is not distinguishable. This introduces an uncertainty in interpretation of paleotemperatures based on the $\delta^{18}$O record, as this record is not representative of seasonal climatic variations. However, I suggest that it is the seasonal climatic variation that is responsible for proglacial lake level formation. Based on the evidence presented above, greater amounts of summer westerly winds are responsible for an increase of DDAF and by association seasonal melt. Yet the signal from summer westerly winds is not preserved in the ice core record. The windier conditions based on the Taylor Dome ice core record increased DDAF enabling prolonged melt seasons that would generate enough melt water to fill Taylor Valley basin to Glacial Lake Washburn size, despite apparent colder average temperatures preserved in the ice core record.
F. Conclusions

The cold and hyper-arid climate during the LGM is inconsistent with the formation and sustainability of large glacial lakes. The closest paleoclimatic record for the McMurdo Dry Valleys is the Taylor Dome ice core record, which has been shown here to reflect climatic changes in the McMurdo Dry Valleys based on analysis of the contemporary westerly winds that descent from the Antarctic plateau. Cross correlation analysis of the summer westerly winds and temperature between Taylor Dome and Taylor Valley indicate that onset of westerly wind and associated temperature increase is experienced at Taylor Dome, followed by Taylor Valley 21 hours later. The warming effect due to westerly winds experienced in both locations, implies a regional climatic connection between Taylor Dome and Taylor Valley.

Annually averaged temperatures during the LGM appear to have been 4-8 °C colder than present, based on the Taylor Dome δ¹⁸O record. However, surficial mixing and post-depositional modification of δ¹⁸O in Antarctica can preclude preservation of seasonal temperature signals in areas such as Taylor Dome with very low accumulation rates. I suggest that warmer than average temperatures during the austral summer existed during the LGM due to an increase in frequency and intensity of westerly winds. The modern westerly winds have a profound effect on climate in Taylor Valley by prolonging the melt season due to an increase of DDAF, and the increased melt leads to an increase in lake levels. However, strong summer westerly winds are infrequent and short lived because they have limited air mass supply from the plateau. Climate during the LGM is inferred to be persistently windier. The windier conditions (meridional circulation) could replenish the source of air, permitting down-valley westerly winds to occur more frequently. Such conditions have a potential to intensify warming due to westerly winds and prolong the melt season during the LGM.
The colder temperatures during the LGM and persistence of large glacial lakes during the LGM cannot be explained by considering the isotopic record alone, because the annually averaged temperatures preserved in the ice cores record do not record the summer extremes. Yet, it is the summer temperature extreme caused by westerly winds that were able to sustain large proglacial lakes. I suggest that summer temperatures during the LGM could have been at least as warm as contemporary climate, driven by episodic westerly wind events. This work suggests that westerly winds were responsible for formation of large glacial lakes during the LGM by increasing the summer temperatures despite colder annual-average temperatures derived from the $\delta^{18}$O record.
G. Acknowledgments

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CHAPTER III

THE PERMANENT ICE-COVER OF LAKE BONNEY, ANTARCTICA: THE
INFLUENCE OF THICKNESS AND SEDIMENT DISTRIBUTION ON
PHOTOSYNTHETICALLY ACTIVE RADIATION AND CHLOROPHYLL-A
DISTRIBUTION IN THE UNDERLYING WATER COLUMN

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Obryk, M. K. 1, Doran, P. T. 1, and Priscu, J. C. 2

1 Department of Earth and Environmental Sciences, University of Illinois at Chicago, Chicago, USA

2 Department of Land Resources and Environmental Sciences, Montana State University, Bozeman, Montana, USA
A. Abstract

The thick permanent ice-cover on the lakes of the McMurdo Dry Valleys, Antarctica, shades the underlying water column to various degrees depending on the ice thickness and associated sediment load. The ice cover insulates the water column, resulting in a strong stratification. To overcome difficulties with sampling beneath the thick ice cover, an autonomous underwater vehicle (AUV) was deployed in Lake Bonney, Taylor Valley, to examine the spatial biogeochemical and physical properties of the lake. Measurements were obtained over the course of two years in a 100 x 100 meter horizontal sampling grid (at a 0.2 m vertical resolution). Additionally, the AUV measured the ice thickness (in water equivalent) and collected images looking up through the ice, which were used to quantify sediment distribution. Satellite imagery were used to qualitatively describe the surface sediment distribution on the ice cover. Here, the results of the investigation of the sediment distribution on the ice cover and its effects on biological processes are presented, with particular emphasis on the variation in the photosynthetically active radiation (PAR). The surface sediment is a secondary controller of the ice cover thickness, which in turn controls the depth-integrated PAR in the water column. This data revealed that depth-integrated PAR was negatively correlated with depth-integrated chlorophyll-a ($r = 0.88$, $p < 0.001$, $n = 83$), which appears to be related to short-term photoadaptation of phytoplanktonic communities to spatial and temporal variation in PAR within the water column.
B. Introduction

Taylor Valley, McMurdo Dry Valleys (MDVs), Antarctica, contains several perennially ice-covered lakes located in hydraulically closed basins. The MDVs are predominantly ice-free, as the Transantarctic Mountains effectively block the downvalley flow of the continental ice sheet. The main source of water for most lakes is from local alpine glaciers. The exception is Lake Bonney, which receives inflows from the Taylor Glacier, an outlet glacier from the Polar Plateau. These glaciers melt during the short austral summers, producing ephemeral streams that feed into the lakes. Due to logistical constraints and thick permanent ice covers, lake sampling in dry valley lakes is usually limited to only a few points a year. Owing to difficulties drilling the thick ice cover, sampling is performed at the deepest point in the lake, which is assumed to represent the entire lake. Spigel and Priscu (1998) have shown spatial and temporal stability of the water column based on conductivity and temperature profiles of Lake Bonney obtained over three-year sampling period in the center of the lake as well as horizontal transect. Distinct salinity gradients are responsible for controlling stratification of the water column, with temperature exerting very little influence on density (Spigel and Priscu, 1998). However, variations of underwater photosynthetically active radiation (UW PAR) and chlorophyll-a under the ice have not been previously documented in high spatial resolution. Here, I test whether UW PAR and chlorophyll-a are spatially stable using data collected by an autonomous underwater vehicle (AUV), which was deployed in the West Lobe of Lake Bonney to perform a high-resolution spatial and temporal biogeochemical and physical survey of the entire lake (Figure 1).

All Taylor Valley (TV) lakes have perennially frozen ice covers that prohibit wind-driven turbulence and its associated mixing of water (Spigel and Priscu, 1998). The ice cover thickness in all TV lakes fluctuates between 3 and 6 m (Doran et al., 2002b; McKay et al.,
The mean annual ablation of the ice cover in Taylor Valley lakes varies between 0.64 m to 0.99 m, as such the residence of lake ice ranges between 3 to 5 years (Dugan et al., 2013). The growth of the ice from the bottom down results in ice crystal growth with a vertical c-axis, permitting sufficient penetration of solar radiation to support phototrophic life (Chinn, 1993).

Figure 1. Top panel: Map of the lakes located in the Taylor Valley. Star indicates location of West Lobe of Lake Bonney. Bottom panel: satellite image of West Lobe of Lake Bonney with an overlaid 100 x 100 m grid. Black dots represent sampling locations during 2009/10 season. Satellite image was obtained from DigitalGlobe™ (shot on Dec 6th, 2008) and processed by the Polar Geospatial Center.
Due to the lack of mixing/overturning of the water column, the lakes exhibit a strong vertical gradient of salinity, a unique pattern of thermal stratification, and a deep chlorophyll maximum (Lizotte and Priscu, 1994; Spigel et al., 1991; Vincent et al., 2008). The source of heat trapped in the liquid water column comes from solar radiation through the ice cover during the austral summer, stream inflow, and latent heat of ice formation (M. N. Gooseff, unpublished). The typical profiles of the water column in the West Lobe of Lake Bonney (as shown in Figure 2) illustrate the sharp thermal and chemical stratification of this lake, which are spatially stable (Spigel and Priscu, 1998). The change in conductivity indicates a chemocline at a depth of approximately 15 m (Figure 2).

Lake Bonney is situated in the upper part of the Taylor Valley, 27 km from the coast (Figure 1). The lake is divided into two separate lobes (east and west) that are connected by a shallow, narrow channel. The West Lobe of Lake Bonney (WLB) is approximately 2.6 km in length and 0.9 km in width (Chinn, 1993). The WLB abuts Taylor Glacier on its westernmost side and connects with the East Lobe of Lake Bonney on the eastern side through a narrow channel (Figure 1). For a physical description of Lake Bonney, see Spigel and Priscu (1998).

In this paper, I present the results of an investigation by the Environmentally Non-Disturbing Under-ice Robotic ANtarctiC Explorer (ENDURANCE), which was deployed in the West Lobe of Lake Bonney during the 2008/09 and 2009/10 seasons with the objective of testing autonomous technologies for the under-ice exploration of icy moons such as Europa (Stone et al., 2010). Spatial patterns of sediment distribution, both on the surface and within the ice cover, and the effects of this sediment distribution on sub-ice processes are discussed. In particular, this paper focuses on the ice cover properties and their effect on photosynthetically active radiation and the spatial distribution of chlorophyll-a in the lake.
Figure 2. West Lobe of Lake Bonney, profiles in November 2009. Data obtained from www.mcmlter.org. Resolution of temperature and conductivity data was reduced to match resolution of chlorophyll-a measurements.
C. Methods


The AUV was designed to swim under the ice cover at a fixed shallow depth (generally 5 m below the piezometric water level). At predefined intervals, the AUV would stop, ascend, and prop itself against the bottom of the ice cover (a process herein termed ice-picking), at which time it would lower a sensor-equipped sonde. The sonde collected continuous data throughout the water column to the bottom of the lake. It profiled at 2 Hz with a typical decent speed of 0.4 m s\(^{-1}\), which translates into a measurement every 0.2 m. This design eliminated the need for the vehicle to swim through the entire volume of the lake to collect data thereby minimizing disturbance of the water column of Lake Bonney.

Navigation under a thick ice cover poses a challenge. The only access to the surface is via the entry melt hole in the lake ice. The ENDURANCE navigation system was composed of three hierarchical subsystems (dead-reckoning, acoustic beacon, and visual homing) to allow successful mission deployment and the return of the AUV to the melt hole (Stone et al., 2010). The main navigation was performed with a dead-reckoning subsystem that included a ring-laser gyroscope inertial unit for vehicle orientation, two pressure transducers for depth determination, and a Doppler velocity log for velocity detection through bottom tracking (Stone et al., 2010). The integration of these components allowed the AUV to determine its position based on its own velocity and orientation with high precision. This navigation system typically returned the vehicle directly under the deployment hole at the end of each mission. Once beneath the deployment hole, the vehicle would track a light shined down the hole and ascend to the surface. In case of a power failure, the AUV was equipped with a magnetic field generator, which allowed us to locate the vehicle under the ice using a phase-locked loop receiver (beaconing
device) (Stone et al., 2010). However, this recovery system proved to be a powerful tool in precisely georeferencing the location of the vehicle during ice-picking. Thus, each sampling location was recorded on the surface with a high-resolution real-time kinematic differential GPS unit. The accuracy of the positioning using this method is on the order of centimeters.

Measurements were taken in a 100 x 100 meter grid across the entire lake (higher-density sampling was performed near the snout of the Taylor Glacier), generating the highest spatially distributed data for this lake. During the 2008/09 season, 44 locations were sampled during 19 dives from Dec 5th to Dec 24th, covering the western half of the WLB. During the 2009/10 season, 83 locations (used in this study) were sampled over 11 days from Nov 6th to Nov 17th, representing the entire lake (Figure 1). After each “mission”, the AUV obtained measurements from one location (F6 – located near the deployment hole), which was designated to serve as a control point to observe temporal change in the lake. The temporal data were obtained from Nov 9th to Dec 2nd during the 2009/10 season, generating 12 profiles.

2. **Sonde and vehicle instruments**

The AUV was outfitted with scientific instruments that were attached to the sonde, which was attached to the vehicle by a cable wound on a spooler drum (Figure 3). The sonde was lowered to the bottom of the lake during ice-picking maneuvers. The sonde was equipped with sensors for depth (Digi-Quartz pressure transducer), conductivity, temperature, underwater photosynthetically active radiation, turbidity and chlorophyll-a fluorescence, colored dissolved organic matter (CDOM), pH, and reduction-oxidation (REDOX). Conductivity, temperature, and pressure were collected using a SEACAT CTD profiler (SBE 19 plus v2). Details are provided on UW PAR and chlorophyll-a sensors only, which generated data for this paper. Underwater PAR (400 nm to 700 nm) was collected using a Biospherical QSP-2300
spherical $3.7\pi$ steradians sensor mounted on top of the sonde. Chlorophyll-a was estimated using WETLABS combination fluorometer and turbidity FLNTU(RT) sensor. Chlorophyll-a fluorescence was measured via excitation at 470 nm and emission at 695 nm. In addition, the AUV collected thousands of images looking up through the ice using a Tritech Osprey high-resolution color underwater camera that was installed on top of the AUV (Stone et al., 2010). The upward looking images were recorded at ~1Hz, however, only images obtained during ice-picking were used in this analysis.

![Diagram](image)

Figure 3. Schematic representation of the Autonomous Underwater Vehicle (AUV) sampling (ice-picking) at West Lobe of Lake Bonney. During the ice-picking, the AUV was motionless, propping itself against the bottom of the ice, at which point it lowered a sensor-equipped sonde. Simultaneous downwelling and upwelling photosynthetically active radiation was obtained on the surface of the ice.
3. **Surface photosynthetically active radiation**

Surface PAR measurements (upwelling and downwelling) were obtained by people tracking the vehicle from above. While the vehicle was stationary and ice picking two LiCor Li-190 2π quantum PAR sensors were deployed, mounted on a frame with two horizontal outriggers, at the end of which sensors were located in a manner to avoid shading (Figure 3). However, only 54 surface PAR measurements were obtained during the 2009/10 season. The simultaneous measurement of surface and UW PAR permitted calculations of percentage of PAR transmittance through the ice cover as well as ice surface albedo. Percentage of PAR transmittance normalizes diurnal and seasonal climatic variations throughout the season such as cloudiness, shading by nearby mountains, and increased incident radiation over the two weeks when the data were collected.
D. Data processing

1. Ice thickness

When the AUV was motionless during ice-picking, the initial measurement obtained with the depth sensor (pressure transducer), minus the offset from the sensor’s location to the top of the AUV (bottom of the ice), was equal to the measurement of the ice thickness at that point with reference to the hydrostatic water surface. The hydrostatic level indicates where the water would be if a hole were drilled in the ice, and it is assumed to be the thickness of the ice in water equivalent. Sediment load both on the surface and within the ice, and air bubbles within the ice will account for some buoyancy variation in the ice cover.

2. Internal sediment

Images obtained using the upward-looking camera at georeferenced ice-picking stations were used to quantify sediment distribution within the ice cover. The angular view of the camera in the water is 65 by 50 degrees. Due to the wide field of view, the images were cropped, eliminating the optical distortion along the outer perimeter of the image. A method for batch image processing was developed using a custom written Matlab ® script to extract sediment data. The technique involved converting the images to gray scale, enhancing the image’s contrast by equalizing the histogram, selecting appropriate threshold values to enhance the sediment in the image, and, finally, converting it to a binary image (black and white) to count the pixels representing the sediment (Figure 4). Pixels were converted to square meters based on the field of view of the camera, which is a function of the ice thickness, i.e. image size is dependent on the thickness of the ice cover (thicker the ice cover, the lesser the resolution). The image processing only allowed quantification of sediment trapped within the ice cover; it does not include sediment on the surface.
Figure 4. An example of upward-looking image that was captured with a camera installed on the top of the autonomous underwater vehicle. The images were used to quantify internal sediment within the ice cover. A) Original image obtained with the camera. B) Image processing, which involved cropping the image, conversion to gray scale, enhancement of sediment (left panel), and conversion to black and white to count the pixels representing the sediment (right panel).

3. **Underwater PAR, chlorophyll-a, and diffuse attenuation coefficient**

Depth-integrated underwater PAR and chlorophyll-a were calculated over the 6 to 10 m depth interval. This depth range was selected because the AUV cast a shadow over the PAR sensor on the sonde up to a 6 m depth and to eliminate bathymetric influence on the results, providing a uniform distribution of the sampled data across the entire lake. The shading effect was determined by investigating UW PAR data in the water column based on all measured profiles. The observed increase of UW PAR, with depth, between the first measurement obtained during the ice-picking up to 6 m depth is indicative of the AUV casting a shadow on the sonde and its associated instruments.

Depth-integrated UW PAR was not corrected for the changes in the incident PAR due to limited overlapping surface PAR measurements. However, I found that % of PAR transmittance at 10 m depth (which is corrected for incident PAR) and depth-integrated UW PAR are
positively correlated \((r = 0.5438, p < 0.001, n = 54)\). The low correlation coefficient is a result of mountains casting a shadow on the northern side of the lake (half of the lake).

Diffuse attenuation coefficient \((K_{\text{PAR}})\) was calculated using a ratio of UW PAR irradiance at 6 and 10 m depth: \(K_{\text{PAR}} = \ln[E(z_6) \cdot E(z_{10})^{-1}] \cdot \Delta z^{-1}\), where \(E\) is measured UW PAR at depth \(z\).

4. **Statistical Analysis**

Principal component analysis (PCA) was employed due to the large number of variables used in this study. PCA was used to discern trends and patterns in the data, with the first components explaining the majority of the variance. The variables included in the PCA are: depth-integrated chlorophyll-a, depth-integrated UW PAR, ice thickness, and internal sediment. However, ice-picking locations with missing one or more datum were deleted in order to produce a gapless 74 x 4 matrix. Where, 74 is a number of ice-picking locations and 4 is number of variables. Because all variables are in different units, PCA was performed using the inverse variances of the data as the weights, in order to normalize the dataset. Upon interpretation of the PCs, Pearson’s linear correlation coefficients were determined between variables in question.
E. Results

1. Principal component analysis

The first three principal components explain 54.5, 24.9, and 15.1 % of variance, respectively (Table 1 and Figure 5). PC 1 describes a relationship between depth-integrated PAR and depth-integrated chlorophyll-a, and ice thickness, where PAR is negatively correlated with chlorophyll-a and ice thickness. PC 2 is concerned with the effect of internal sediment on chlorophyll-a. Finally, PC 3, shows a relationship between ice thickness and internal sediment. Scree plots and a biplots of PCA can be found in the Figure 5.

Statistically significant linear correlations between the trends observed by the PCA are explored below.

Figure 5. A) Scree plot of variance explained by the principal component analysis (PCA), and B) a biplot of loadings and scores for each variable, where Chl-a is depth-integrated chlorophyll-a, and UW PAR is depth-integrated underwater photosynthetically active radiation.
Table 1. Summary of Principal Component Analysis including loadings of the first three principal components (PC) and variance explained.

<table>
<thead>
<tr>
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<th>PC 1</th>
<th>PC 2</th>
<th>PC 3</th>
</tr>
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<tbody>
<tr>
<td>Depth-integrated chlorophyll-a</td>
<td>-0.57</td>
<td>-0.42</td>
<td>-0.11</td>
</tr>
<tr>
<td>Depth-integrated UW PAR</td>
<td>0.59</td>
<td>0.25</td>
<td>0.33</td>
</tr>
<tr>
<td>Internal sediment</td>
<td>-0.31</td>
<td>0.81</td>
<td>-0.46</td>
</tr>
<tr>
<td>Ice thickness (in 2009/10)</td>
<td>-0.48</td>
<td>0.30</td>
<td>0.82</td>
</tr>
<tr>
<td>Variance explained (%)</td>
<td>54.5</td>
<td>24.9</td>
<td>15.1</td>
</tr>
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2. Ice thickness

Data obtained from the ice-picking show large spatial variations in the ice thickness across the West Lobe (Figures 6A and 6B). The northern side of the lake shows pronounced thickening of the ice by about 70 cm compared with the southern side. In addition, the ice cover exhibits two distinct regions of ice thinning on the southwest side of the lake, where the lake abuts Taylor Glacier, and in the east central portion of the lake (Figure 6A). The two localized regions of thinner ice cover are associated with aeolian sediment deposition on the surface of the ice, as shown by a comparison with high-resolution satellite imagery (Figure 7).

A comparison of historical aerial and satellite imagery (dating to 1958) of Lake Bonney reveals that the distribution of the sediment on the surface of the ice cover is a persistent feature of this lake, especially the sediment on the west side along the glacier. All images show aeolian sediment deposition along the glacier face of the lake with small spatial variations (data obtained from Polar Geospatial Center: www.agic.umn.edu). The oval center-east region of the surface sediment accumulation appears to be relatively new and has been present for less than a decade.
Figure 6. Spatial ice thickness variations (in water equivalent) of West Lobe of Lake Bonney during the A) 2009/10 and B) 2008/09 seasons. C) Depth-integrated (from 6 to 10 m) underwater photosynthetically active radiation. D) Depth-integrated chlorophyll-a (from 6 to 10 m). All maps are projected in UTM coordinates. Black contour represents the West Lobe of Lake Bonney parameter obtained with high-resolution GPS. Dots represent locations of data collection.

The spatial ice thickness variations during the 2009/10 season are 1.4 m, with a maximum value of 4.4 m on the north side of the lake and a minimum value of 3.0 m by the glacial face where aeolian sediment accumulates on the surface of the ice cover ($\bar{x} = 3.7, \sigma = 0.23$) (Figure 6A). The ice thickness data from the 2008/09 season only cover the western half of the lake; however, the spatial ice thickness distribution was similar to that in the 2009/10 season (Figures 6A and 6B). Extensive thinning of the ice was observed along the glacier, where
aeolian sediment had accumulated, with a minimum value of 3.5 m. A temporal comparison of
the ice cover thickness between all overlapping data points (representing the western half of the
lake) for the two seasons shows an average ice cover loss of 0.22 m (σ = 0.23, min = -0.94, max
= 0.52) over a period of one year despite the decrease of winter temperature (March to
September) by 0.9 °C between the two seasons.

Figure 7. High-resolution satellite image of West Lobe of Lake Bonney. The circled regions
represent aeolian sediment accumulated on the surface of the ice cover, associated with thinner
ice cover. Quickbird satellite image was obtained from DigitalGlobe ™ (shot on Dec 6 th, 2008)
and processed by Polar Geospatial Center.
3. **Sediment within the ice**

Images obtained with the upward-looking camera show that sediment accumulates in small discrete patches or lenses. The surface area of entrapped sediment ranges from 0.00 to 0.12 m$^2$. Adams et al. (1998) reported sediment pockets that are several cm across, based on the ice cores recovered from WLB. However, the ice cores are only 10 cm in diameter (Adams et al., 1998); hence they will underestimate surface area calculations and cannot be compared with results presented in this paper. The spatial distribution of entrapped sediment patches varies widely across the lake (Figure 8A) and shows a poor, but statistically significant, linear correlation with ice thickness ($r = 0.3151, p = 0.005, n = 76$), depth-integrated UW PAR ($r = -0.2534, p = 0.029, n = 74$), and % of PAR transmittance at 10 m depth ($r = 0.2769, p = 0.041, n = 54$). Percentage of PAR transmittance at 10 m depth is shown in Figure 8B. It is noteworthy that although no images were recorded from the AUV’s forward-looking camera, sediment patches at the very bottom of the ice cover were observed in real-time. The existence of these patches is not explained by any ice/sediment models developed to date (e.g., Jepsen et al., 2010; Simmons et al., 1986).

4. **PAR and $K_{\text{PAR}}$**

Depth-integrated UW PAR (from a 6 to 10 m depth) shows two pronounced regions of greater light in the water column: along the glacier face (west side of the lake), extending roughly 300 m eastward from the glacier, and the center-east side of the lake (Figure 6C). The maximum values of depth-integrated UW PAR are found on the northwestern edge of the lake (425.8 µmol cm$^{-1}$ s$^{-1}$), with minimum values on the far eastern end of the lake (5.6 µmol cm$^{-1}$ s$^{-1}$). The relationship between ice thickness and depth-integrated UW PAR yields a poor
yet statistically significant negative linear correlation \( (r = -0.3959, p < 0.001, n = 83) \) (Figures 6A, 6C, and 9).

Diffuse attenuation coefficient positively correlates with depth-integrated chlorophyll-a \( (r = 0.3901, p < 0.001, n = 83) \) and negatively with depth-integrated UW PAR \( (r = -0.3296, p < 0.001, n = 83) \).

![Figure 8. A) Spatial variations of sediment incorporated within the ice and B) percentage of light transmittance at 10 m depth. Maps are projected in UTM coordinates. Black contour represents the West Lobe of Lake Bonney parameter. Dots represent locations of data collection.](image)

5. **Chlorophyll-a**

Depth-integrated chlorophyll-a (from 6 to 10 m) is the highest on the east side of the lake near the narrows (7.9 \( \mu g/m^2 \)), where the lake merges and connects with the east lobe of Lake Bonney (Figure 6D). The smallest chlorophyll-a concentration is along the glacier interface (1.6 \( \mu g/m^2 \)). The linear correlation coefficient was calculated between spatial log-transformed depth-integrated UW PAR and depth-integrated chlorophyll-a \( (r = -0.8846, p < \)
0.001, n = 83) (Figure 10) and % of PAR transmittance at 10 m depth and depth-integrated chlorophyll-a (r = -0.5015, p < 0.001, n = 54). The time series of depth-integrated UW PAR and depth-integrated chlorophyll-a also show a negative correlation (r = -0.8702, p < 0.001, n = 12) (Figure 11). In addition, ice thickness and depth-integrated chlorophyll-a show a statistically significant linear correlation (r = 0.4021, p < 0.001, n = 83).

Figure 9. Depth-integrated underwater photosynthetically active radiation vs. ice thickness (r = -0.3959, p < 0.001, n = 83).
Figure 10. Log-transformed underwater photosynthetically active radiation (UW PAR) vs. chlorophyll-a ($r = -0.8846$, $p < 0.001$, $n = 83$) for all sample locations during the 2009/10 season. Both UW PAR and chlorophyll-a are depth-integrated from 6 m to 10 m at each sample location.

Figure 11. Underwater photosynthetically active radiation (UW PAR) vs. chlorophyll-a ($r = -0.8702$, $p < 0.001$, $n = 12$) representing temporal data from Nov 9th, 2009, to Dec 2nd, 2009, obtained from F6 location. Both UW PAR and chlorophyll-a are depth-integrated from 6 m to 10 m at each sample location.
**F. Discussion**

Analyses of the high-resolution spatial data from the West Lobe of Lake Bonney revealed ice thinning along the Taylor Glacier face and the central east side of the lake. These two pronounced regions of thin ice cover are associated with the accumulation of aeolian sediment (Figures 6A and 7), which can enhance thinning due to the absorption of incident solar radiation by the surface sediment during the austral summer months (Jepsen et al., 2010).

Variations in spatial distribution of surface sediments have been attributed to the strength of the winter winds, during which most of the wind-blown sediment accumulates (Priscu et al., 1998). However, the analysis of historic aerial and satellite images reveals the apparent static nature of the sediment accumulation on the ice, especially along the glacier face, suggesting that the spatial pattern of the ice cover thickness of the West Lobe of Lake Bonney has remained the same for several decades. Conversely, the thicker ice cover on the north side of the lake presumably results from shading by mountains to the north of the lake, which is most pronounced at solar noon (Dana et al., 1998). Decrease in solar radiation will, in turn, decrease ablation during the summer months (Clow et al., 1988; Dana et al., 1998; McKay et al., 1985).

These data imply that accumulation of aeolian sediment on the surface of the ice cover and localized shading play a major role in controlling the ice cover thickness at small scales.

Ice cover thickness is positively correlated with sediment incorporated within the ice. This presents an apparent paradox as thinner ice cover contains a lesser amount of internal sediment, yet thinner ice is associated with aeolian sediment accumulation on the surface of the ice cover. Field studies and an energy balance model show that surface sediment can propagate through the ice to 2 m depth (Jepsen et al., 2010). The release of subsurface sediment from this layer occurs during the austral summers when the ice becomes unstable and permeable, forming vertical cracks that allow for downward movement of sediment (Jepsen et al., 2010; Nedell et
al., 1987; Squyres et al., 1991). However, Jepsen et al. (2010) proposed that the differences in the ice thickness can cause a rapid rebound of thinner ice that can generate cracks and fractures in the ice. The associated cracks become conduits for any subsurface sediment to be released from the ice into the water column (Jepsen et al., 2010; Nedell et al., 1987; Squyres et al., 1991). The weak positive relation between internal sediment and ice thickness is a result of longer residence of the internal sediment within thicker ice cover.

Water column, depth-integrated PAR is associated with variations of the ice thickness, i.e., thicker ice cover results in a lesser amount of depth-integrated PAR. The aeolian sediment accumulation on the surface of the ice cover has a pronounced effect on the thinning of the ice (up to 1 m) due to the absorption of the solar energy by the sediment, which increases ablation at the surface (Figures 6A and 7). Light transmittance through the ice has been characterized by dividing the ice into discrete horizontal layers that have distinct optical properties (Fritsen and Priscu, 1999; McKay et al., 1994). The majority of the light is absorbed and scattered at the very top layer of the ice (up to 1 m depth) due to morphologically distorted air bubbles, Tyndall figures (disks of liquid water resulting from the internal melting of ice), ice-whitening, and entrapped sediment (Fritsen and Priscu, 1999; McKay et al., 1994). The homogeneous clear bottom layer of the ice cover is assumed to be winter growth with vertical chains of undisturbed air bubbles (Adams et al., 1998; McKay et al., 1994). The subsurface sediment accumulates in small discrete patches at 2 m depth below morphologically disturbed ice (example Figure 4), as a result it has little impact on PAR transmittance (based on the decreasing diffuse attenuation coefficient with depth within the ice (Fritsen and Priscu, 1999)). Hence, the top meter of the ice alone plays a major role in controlling the total available PAR in the water column (Fritsen and Priscu, 1999). The surface sediment is responsible for the increased ablation of the ice (hence
thinner ice) by removing the morphologically disturbed layer (top 1 m) and allowing more light penetration. I suggest that the surface area of internal sediment patches is too small to interfere with the penetrating light. However, it is interesting to note that a linear correlation show a positive relationship between the subsurface sediment and % of PAR transmittance. This relationship does not demonstrate causation, as I have shown above that the internal sediment is associated with thicker ice cover, which attenuates penetrating light.

These data showed that depth-integrated UW PAR is negatively correlated with depth-integrated chlorophyll-a (Figures 6C and 6D, 10, 11, and Table 1). This negative correlation presents a paradox, as the long-term analysis (over the course of multiple austral summers) of PAR and chlorophyll-a in the WLB shows positive trends (Fritsen and Priscu, 1999). Increased phytoplankton biomass during early spring in WLB have been shown to be stimulated by an increase in incident irradiance (Lizotte et al., 1996). Chlorophyll fluorescence can be used as a measure of chlorophyll-a concentration and primary production (Lizotte and Priscu, 1992). However, the negative correlation is not necessarily indicative of biomass changes. Rather it reflects chlorophyll-a concentration of the phytoplankton. The negative relationship observed between chlorophyll-a and under ice PAR in WLB may be the result of short-term photo-adaptation of phytoplanktonic communities living in the water column. Microbial communities living in the Antarctic lakes are highly shade adapted, and they have a variety of mechanisms to cope with light deficiencies (Lizotte and Priscu, 1992; Morgan-Kiss et al., 2006; Neale and Priscu, 1995). One of the short-term techniques of photoadaptation is a change in light-harvesting antenna size with respect to available light (Morgan-Kiss et al., 2006). During low irradiance, microbial communities will increase chlorophyll-a to compensate for the lack of available light to meet their energy requirements (Miskiewicz et al., 2000; Morgan-Kiss et al.,
Such adaptations have been observed in fresh water lakes on the southern part of the Soya Coast, East Antarctica (Tanabe et al., 2008) and in Ariake Bay, Japan (Shibata et al., 2010) and have been discussed for Lake Bonney by Neale and Priscu (1995), Lizotte and Priscu (1992), and Kong et al. (In Review). These reports in concert with data presented in this paper indicate that cell specific changes in chlorophyll-a in response to variations in under-ice PAR may produce the trends shown in the ENDURANCE data (Figures 6D, 10, 11). Based on a temporal analysis, I suggest that the observed negative correlation persists throughout the season. These results suggest that spatial estimation of biomass based on chlorophyll-a fluorescence only might not accurately represent the biomass of an ecosystem.

The study of permanently ice-covered polar lakes is hindered by the lack of spatial under-ice data and high-density sampling over time. Use of an AUV to collect these data have allowed, for the first time, to examine the three-dimensional structure of selected physical and biological variables in an Antarctic lake. Together with satellite imagery, I show that sediment accumulation on the surface of the ice controls water column PAR, which in turn can influence the spatial distribution of chl-a within the near surface water column. In addition, this investigation shows that depth-integrated UW PAR is an alternative proxy for estimating chlorophyll-a within the water column.
G. Acknowledgments

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CHAPTER IV

MODELED RESPONSE OF TAYLOR VALLEY PERENNIAL LAKE ICE COVERS TO CHANGING CLIMATE

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Obryk, M. K.\textsuperscript{1}, Doran, P. T.\textsuperscript{1}, Hicks, J. A.\textsuperscript{2}, McKay, C. P.\textsuperscript{3}, and Priscu, J. C.\textsuperscript{4}

\textsuperscript{1}Department of Earth and Environmental Sciences, University of Illinois at Chicago, Chicago, Illinois, USA

\textsuperscript{2}Engineering Sciences and Applied Mathematics, Northwestern University, Evanston, Illinois

\textsuperscript{3}Space Science Division, NASA Ames Research Center, Moffett Field, California, USA

\textsuperscript{4}Department of Land Resources and Environmental Sciences, Montana State University, Bozeman, Montana, USA
A. Abstract

A one-dimensional ice cover model was developed in order to predict and constrain drivers of long-term ice thickness trends in Taylor Valley lakes. The physics-based model is driven by surface radiative heat fluxes and heat fluxes from the underlying water column. The model successfully reproduced 16 years of ice thickness changes for West Lobe of Lake Bonney (RMSE = 0.09 m) and Lake Fryxell (RMSE = 0.21 m). Long-term ice thickness trends cannot be modeled by surface radiative fluxes alone and require coupling with the thermal structure of the water column. The heat stored within the deep temperature maximum of a lake can either impede or facilitate ice thickness change depending of climatic trend. Ice thickness of shallow lakes, on the other hand, is more sensitive to climatic changes. As such, shallow perennially ice-covered lakes are a better indicator of climatic changes. Taylor Valley has been generally experiencing decreasing air temperature and increasing shortwave radiation over the past two decades, and occasional discrete warming events. The ice thickness trends of ice-covered lakes in Taylor Valley are not sensitive to discrete warming or cooling summer events. Rather, the ice thickness trends are a result of surface energy flux and heat flux from the deep temperature maximum. Future ice thickness prediction for Lake Bonney indicates that the lake can become seasonally ice-free within 20 years.
B. Introduction

Perennially ice-covered lakes in Antarctica have long been studied as novel habitats on Earth. Ice covers offer protection from wind-driven turbulence permitting stratification and development of supersaturated gas concentrations (Spigel and Priscu, 1998), and altered sediment pathways (Doran et al., 1994) among other things. Ice covers provide insulation from the atmosphere but can allow penetration of solar radiation, often producing thermal maxima in the water column that would not have existed otherwise (Vincent et al., 2008). These unique characteristics of perennially ice-covered lakes provide an opportunity to study the response of such lakes to climatic changes, which will change the thickness of lake ice. However, little attention has been paid to ice cover thickness changes in Taylor Valley lakes as an indicator of climate change.

Taylor Valley lakes are controlled by a delicate climatic balance of summer temperatures warm enough to generate melt from local glaciers, yet cold enough for the lake ice covers to persist through the summer (Chinn, 1993). The ice thickness in Taylor Valley lakes has been fluctuating between 3 and 6 m, depending on the lake (Doran et al., 2002b; McKay et al., 1985). Between 1986 and 1999 lake-ice thickness increased, which was associated with an annual temperature decrease of 0.7°C per decade (Doran et al., 2002b). This period of pronounced cooling was terminated by one of the warmest summers on record and years of rising lake levels (Barrett et al., 2008; Doran et al., 2008). Lake-ice thicknesses have been generally decreasing since early- to mid-2000 (unpublished data). In this paper, a physics-based lake ice model was developed in order to understand the sensitivity of Taylor Valley ice-covered lakes to climatic forcings. In addition, this model is used to model future trends in ice thickness and when we might expect the lakes to become seasonally ice-free.
C. Site Description

Lakes in Taylor Valley, Antarctica have been perennially ice-covered since at least 1903, when they were discovered by Robert Falcon Scott (Scott, 1905). McKay et al. (1985) developed a steady state energy-balance model that predicts ice cover thickness based on ablation (mass loss by all means) rates. Ice covers are primarily balanced by ablation at the surface and ice growth at the bottom (Chinn, 1993). Latent heat release, associated with ice growth, is conducted upwards and removed due to ablation (McKay et al., 1985). Because of the coupling between the lake ice and atmosphere, the ice covers can be used as indicators of local (and potentially global) climate change (Launiainen and Cheng, 1998; Reid and Crout, 2008; Wharton et al., 1992).

Thermodynamic modeling has been known since 1891 and general description and evolution of different ice models throughout time is summarized in Launiainen and Cheng (1998). However, no universal lake ice model is available, as each lake has different characteristics and unique adjustments are necessary (Reid and Crout, 2008). The model presented here was developed using generalized parameters to make it adaptable to other perennially ice-covered lakes. The model was tested against data obtained from West Lobe of Lake Bonney (WLB) and Lake Fryxell located at the western and eastern ends of Taylor Valley, respectively (Figure 1). The two lakes are in different microclimates in the valley; WLB is influenced by strong, dry winds from the Antarctic plateau, whereas Lake Fryxell is influenced by moist coastal winds (Doran et al., 2002a; Fountain et al., 1999). As a result, the surface ice cover morphology differs significantly, from relatively smooth (WLB) to highly disturbed (Lake Fryxell).
Figure 1. Map of Taylor Valley, Antarctica. Triangles indicate locations of meteorological stations.
D. Model Development

1. General Model Description

The model developed here is adapted from an approach by Launiainen and Cheng (1998) and Reid and Crout (2008). The model is focused on 1) the dynamic coupling of atmospheric processes with ice, 2) the dynamic coupling of the water column with ice, and 3) heat conduction within the ice and the water column (Figure 2). Three boundaries are defined in the model: atmosphere/ice, ice/water, and water/bedrock. The entire vertical temperature profile (including ice and water) is solved using a spectral method based on the determined state of the boundaries. The main advantage of the spectral method over the finite difference or the finite element method is a higher degree of accuracy (Trefethen, 2000).

The model is driven by surface radiative fluxes as depicted in Figure 2. The only input data in the model are shortwave radiation, surface air temperature, relative humidity, wind speed, and albedo. Remaining of the surface fluxes are calculated based on the published equations and are described in more detail in sections E.3, E.4, and E.5. Summation of the surface radiative fluxes yields temperature of the surface of the ice (see section E.1), which becomes the top boundary condition. The ice surface temperature is calculated iteratively. A positive heat flux on the ice surface is used to calculate ice removal (see section E.6). Known surface ice temperature allows for calculation of temperature profile of the ice as the temperature at the bottom of the ice is set to freezing point. During the same time step, water temperature profile is calculated that allows for properly constraining heat fluxes from the water column (conductive, latent, and sensible heat). The heat flux at the ice/water interface is used to determine growth or decay of the ice at the bottom. The model outputs are ice thickness changes.
over time, heat fluxes at the surface and bottom of the ice, and ice and water temperature profiles over time. The model was coded in Matlab® environment.

The model employs dimensional analysis permitting grouping of variables that describe physical parameters in a dimensionless form. Dimensional analysis is based on the Buckingham \(\pi\)-theorem, allowing for the reduction of independent variables by \(k\) dimensions, where \(k\) represents physical dimensions required to describe \(n\) variables used in the model (Fischer, 1979). Then, it follows that there will be \(\pi\) dimensionless groups based on \(n – k\) (Fischer, 1979). As a result, this method reduces the complexity of the model and, consequently, reduces the complexity of the solution to the heat equation.

A one-dimensional solution to the heat equation is valid because the physics in one dimension are preserved. Heat exchange within the ice occurs through diffusion based on a temperature gradient defined at the atmosphere/ice and ice/water boundaries. The heat within the ice diffuses vertically and laterally at rates defined by the thermal diffusivity of the ice and the temperature gradient. However, the thickness of the lake ice represents a small fraction of the surface area of the lake. Hence, the lateral thermal gradient can be considered negligible with regards to depth, allowing a one-dimensional heat equation to be employed because vertical processes dominate.
Figure 2. Schematic representation of heat fluxes and temperature profile for West Lobe of Lake Bonney. Numbers in boxes represents annually averaged fluxes (W m$^{-2}$).

Similar to the approach used for the ice, a one-dimensional heat equation is used to calculate the temperature profile of the water column in order to better constrain heat fluxes at the ice/water boundary. Taylor Valley lakes are meromictic (layers of water do not mix) and it is assumed here that the heat transfer in the lake is through a diffusive process. However, it is important to note that horizontal motion of the top most layer of water can occur during austral summers when streams enter the lakes (Spigel and Priscu, 1998).

2. **Input data**

Meteorological data used in the model include: air temperature, shortwave radiation, relative humidity, and wind speed. Longwave upwelling and downwelling radiation,
and turbulent sensible and latent heat fluxes were parameterized using equations adapted from Reid and Crout (2008), discussed in section E.3 and E.4, respectively. The albedo of the ice was simulated as a spline function based on a five-month data record (September to February) from WLB, obtained from Fritsen and Priscu (1999). Based on this record, the largest albedo change (increase) occurs during the summer months when ice morphology changes due to isothermal ice (Fritsen and Priscu, 1999). The missing albedo data for the remainder of the year (winter season) were extrapolated. Any error in this assumption is reduced by the diminished importance of radiation in the winter months.

The model was run with 16 years of meteorological data obtained from Lake Bonney’s meteorological weather station, located on the southern shore of East Lobe of Lake Bonney (Figure 1). Data were collected using Campbell Scientific dataloggers (CR10 and 21X) at 30 s sampling intervals and stored as 15 min averages (Doran et al., 1995; Doran et al., 2002a). Air temperature was measured using Campbell 207 and 107 temperature probes at a 3 m height (Doran et al., 1995). Shortwave radiation was obtained with Li-Cor LI200S and LI200X silicon pyranometer (Doran et al., 1995). Wind speed was obtained with R. M. Young model 05103 wind monitors (Doran et al., 1995). The model was run at 12-hour intervals, dynamically averaging high-resolution data (15 min intervals) during simulations.

Gaps in data occurred, ranging from hours to months, due to either failed sensors or failed dataloggers. Missing shortwave radiation was remedied by using well-known published equations (see section E.2) dynamically substituting gaps in data during model simulations. To determine missing air temperature values, daily average temperature was calculated based on 16 years of data. A function was derived based on those averages, substituting gaps in data. Missing relative humidity and wind speed data represent cumulative 6.6 and 4.6 % of the entire
dataset, respectively, and were determined by a random number generator (within the upper and lower limit of each variable). Random numbers generated in Matlab are uniformly distributed over the range specified by the variables in question.

3. Validation data

Ice thickness data was measured manually at the drill hole sites from the bottom of the ice to the piezometric water level, representing the ice thickness in water equivalent. Due to logistical constraints, the ice thickness data are only available during the austral summers when the ice is most dynamic, with the exception of an extended season in 2008, when data were obtained in March. Throughout every season, several ice thickness measurements were taken during the same day (within close proximity) or within few days of each other, and show a large variability, introducing a source of error during the validation procedure of the model.

Continuous ablation data were measured from suspended pressure transducers at the bottom of the ice (Dugan et al., 2013). Data were recorded at one-minute intervals (averaged over 20 min), using Druck PDCR 1830 or Keller Series 173 pressure transducer and stored on Campbell Scientific CR10X and CR1000 dataloggers (Dugan et al., 2013).
E. Heat Fluxes

The vertical temperature profile from the ice surface to the bottom of the lake and its evolution over time can be modeled utilizing a one-dimensional unsteady heat equation:

\[ p c \frac{\partial T(z,t)}{\partial t} = - \frac{\partial}{\partial z} \left( -k \frac{\partial T(z,t)}{\partial z} + I(z,t) \right) \]  

where \( p \) is the density of ice or water, \( c \) is the specific heat capacity of ice or water, \( T \) is temperature, \( z \) is depth, \( t \) is time, \( k \) is thermal conductivity, and \( I \) is the heat source term within the ice and the water column. All constants and variable units are summarized in Appendix A. Shortwave radiation is the only significant flux that contributes to the \( I \) term because ice is opaque in infrared and transparent in the visible spectrum (McKay et al., 1994). The transmittance of sunlight through the perennially ice-covered lakes in Taylor Valley can be approximated using the Lambert-Beer law (McKay et al., 1985), by defining a constant extinction coefficient. However, this technique does not account for the decreasing extinction coefficient with depth due to absorbed infrared radiation at the very surface of the ice (Bintanja and Vandenbroeke, 1995; Brandt and Warren, 1993). A method used in Bintanja and Vandenbroeke (1995) was adapted that parameterized solar energy by partitioning absorbed solar energy at the very surface and subsurface of the ice, expressing the heat source term as:

\[ I(z,t) = (1 - \chi)(1 - \alpha)Q_s e^{-\kappa z} \]  

where \( \chi \) accounts for a fraction of absorbed shortwave radiation at the surface, \( \alpha \) is albedo, \( Q_s \) is shortwave radiation, and \( \kappa \) is the extinction coefficient of the ice at depth \( z \).
1. **Surface Energy Balance**

Temperature at the atmosphere/ice boundary is calculated based on the summation of all surface radiative heat fluxes, and becomes a top boundary condition for Eq. (1):

\[ F(T_s) = \chi(1 - \alpha)Q_s + Q_b + Q_d + Q_h + Q_l + F_i \]

(3)

where \( T_s \) is ice surface temperature, \( Q_s \) is shortwave radiation, \( Q_b \) is longwave upwelling radiation, \( Q_d \) is longwave downwelling radiation, \( Q_h \) is sensible heat flux, \( Q_l \) is latent heat flux, and \( F_i \) is the heat conduction at the ice surface. However, \( T_s \) cannot be solved based on this equation alone because the longwave upwelling radiation, sensible, and conductive heat fluxes depend on the surface ice temperature (\( T_s \)). The Newtonian iterative method was implemented in the model to solve for \( T_s \). The surface ice temperature became the top boundary for Eq. (1) and was used to determine temperature gradient within the ice cover. Positive fluxes are toward the ice, negative are away from the ice.

2. **Shortwave Radiation**

Shortwave radiation (\( Q_s \)) for an atmosphere-less Earth is defined as:

\[ Q_s = S \cos Z \]

(4)

where \( S \) is a solar constant and \( Z \) is a solar zenith, derived using solar declination, solar angle, and latitude. Atmospheric attenuation of \( Q_s \) requires modification of Eq. (4) with empirical
parameters related to the study site and numerous equations are available in literature (Bennett, 1982; Lumb, 1964; Reid and Crout, 2008). However, parameterization of \( Q_s \) for an Antarctic site has been done by Reid and Crout (2008) and was adopted here.

3. **Longwave Radiation**

Longwave upwelling radiation (\( Q_b \)) was modeled based on the Stefan-Boltzmann law as a function of surface ice temperature and is expressed as:

\[
Q_b = -\varepsilon_i \sigma_s T_s^4
\]  

(5)

where \( \varepsilon_i \) represents dimensionless surface emissivity, \( \sigma_s \) is the Stefan-Boltzmann constant, and \( T_s \) is the surface ice temperature (Launiainen and Cheng, 1998).

Downwelling longwave radiation (\( Q_d \)), on the other hand, is mostly controlled by surface air temperature (\( T_a \)) and cloud cover (\( C \)) (Guest, 1998; Konig-Langlo and Augstein, 1994).

Cloud cover is expressed as a unitless value (from 0 to 1) representing the fraction of sky covered by clouds. However, cloud cover data for the WLB is unavailable. Missing cloud cover data can be treated as an adjustable parameter in models driven by surface energy balance (Liston et al., 1999). Here, I chose \( C \) to be generated by a random number generator at daily intervals in order to simplify the model by decreasing number of adjustable parameters.

Downwelling longwave radiation requires parameterization of atmospheric emissivity (\( \varepsilon_a \)), which accounts for \( C \), vapor distribution, and vertical temperature (Konig-Langlo and Augstein, 1994). The equation for \( Q_d \) was adopted from Konig-Langlo and Augstein (1994) because they parameterized \( \varepsilon_a \) based on Antarctic datasets, which is applicable for high-latitudes:
\[ Q_d = (0.765 + 0.22C^3) \sigma_s T_a^4 \] (6).

4. **Sensible and Latent Heat Flux**

Turbulent sensible \((Q_h)\) and latent heat \((Q_l)\) fluxes at the atmosphere/ice boundary are modeled after Launiainen and Cheng (1998) and are expressed as:

\[ Q_h = \rho_a c_a C_H(\Delta T)V \] (7)

\[ Q_l = \rho_a R_l C_E(\Delta q)V \] (8)

where \(\rho_a\) is air density, \(c_a\) is specific heat capacity of air, \(V\) is wind speed, \(R\) is enthalpy of vaporization (or sublimation if \(T_s < 273.15\)), \(\Delta T\) is the temperature difference between atmosphere and ice surface, \(\Delta q\) is specific humidity difference between atmosphere and ice surface, and \(C_H\) and \(C_E\) are dimensionless bulk transfer coefficients. Strictly speaking, \(C_H\) and \(C_E\) depend on atmospheric stability (Launiainen and Cheng, 1998); however, here they are treated as constants approximated based on Parkinson and Washington (1979).

5. **Conductive Heat Flux**

Conductive heat flux at the surface of the ice is defined as a temperature gradient between the air and the top layer of the ice:

\[ F_i = - k \left( \frac{dT}{\partial Z} \right) \] (9)
where $T$ represents the temperature difference defined by the depth $z$, and $k$ is the thermal conductivity of ice. Similarly, heat conduction between ice and the water takes the form of Eq. (9).

6. **Ice ablation and growth**

Ice removal from the top of the ice is approximated as a function of net surface radiative heat fluxes, similar to Eq (3). If the net radiative heat flux ($Q_{\text{net}}$) is greater than zero, the excess energy is used to remove the surface layer of the ice, and can be expressed as:

$$
\frac{d A_{\text{ABL}}}{dt} = -\left(\frac{1}{L_p}\right) \sum Q_{\text{net}}
$$

(10)

where $L$ is latent heat of freezing, $p$ is ice density. Such an approach does not distinguish between sublimation and ice melt; however, it sufficiently represents cumulative annual ice removal from the surface of the ice.

Ice growth is only considered at the bottom of the ice, which is controlled by sensible heat flux from the water, conduction of heat at the ice/water boundary, and latent heat flux associated with the phase change (Danard et al., 1984; Launiainen and Cheng, 1998). The temperature at the ice/water interface is assumed to be at the freezing point. However, heat fluxes at the ice/water boundary are difficult to determine because sensible and conductive heat fluxes from water are often unknown. Sensible heat from the water can be either approximated as a constant if water temperature beneath the ice is not known (Launiainen and Cheng, 1998) or calculated, if water temperature beneath ice is known, using modified Eq. (7) (Reid and Crout,
2008). However, for proper determination of fluxes from the water column, thermal structure of the water profile should be coupled with the ice (Launiainen and Cheng, 1998; Maykut and Untersteiner, 1971). Here, a second heat equation is solved (Eq. (1)) to calculate the water temperature profile permitting dynamic calculations of heat conduction between the water and the ice, as well as sensible and latent heat fluxes. The ice/water boundary is a moving Dirichlet boundary condition, which forces the solution of a heat equation to converge at the freezing point of the water (ice/water interface). This boundary incorporates latent heat fluxes associated with the phase transition (Gupta, 2003).
F. Initial Conditions

Initial conditions of the ice temperature profile can be approximated as a linear temperature gradient between the ice surface and the ice bottom because the system rapidly adjusts (Reid and Crout, 2008). However, when considering a unique temperature profile of the water column such as West Lobe of Lake Bonney’s, which takes decades to develop (Vincent et al., 2008), linear approximation is not appropriate when attempting to model long time series. The initial temperature profiles of the ice and the water column in this model were determined in two steps. First, the temperature gradient between the surface of the ice, ice/water interface, and bottom of the lake was set up as a linear function. The simulation was run until the water temperature profile developed a temperature maximum, similar to that of WLB. Second, the last calculated value, in the previous step, was used as an initial temperature profile for the ice and the water column condition.

The initial surface boundary condition is calculated as a step ‘0’ using the meteorological conditions that are used in step 1. Ice/water temperature was set at the freezing point, as a Dirichlet boundary condition, and the bottom of the lake temperature (water/bedrock) is fixed.
G. Model Validation

The model was validated using 16 years of seasonally measured ice thickness data and 6 years of continuous ablation data from WLB. In addition, modeled vs. measured temperature profiles from WLB are qualitatively compared. Versatility of the model was validated using 16 years of seasonally measured ice thickness data from nearby Lake Fryxell.

The model accurately predicts ice thickness trends for WLB (Figure 3), including the increasing ice thickness from 1996 to 2002, followed by the subsequent decrease after 2002. Ice thickness during March 2008 is grossly overestimated, suggesting that the seasonal amplitude of ice thickness variations could be poorly resolved by the model. However, measured vs. modeled thickness have a correlation coefficient of $r = 0.95$ ($p < 0.001$) and root-mean-square-error (RMSE) of 0.09 m suggesting that the model reasonably captures decadal ice thickness trend variability (Figure 4).

Despite the simplicity of the ablation calculation, cumulative annual ablation is well predicted (Figure 5) with a correlation coefficient of $r = 0.99$ ($p < 0.001$) and RMSE of 0.11 m between measured and modeled ablation. However, the significance of the fit should be interpreted with caution, as $r$ can be artificially inflated with a large number of data points ($n > 4200$). For this reason, RMSE is provided, as it represents an absolute error in units of the data (Willmott, 1981).

A typical WLB temperature profile (obtained in 2011) is shown in Figure 6 with temperature maximum at approximately 10 m depth. The model predicted a similar temperature profile; however, the modeled temperature maximum is approximately at 9 m depth (Figure 6). The most likely reason for this temperature offset is because the model does not account for biogeochemical properties of the water column. Modeling of an Arctic lake showed that
microbial communities have an effect on thermal structure of the water column (Vincent et al., 2008), which is not implement here.

For Lake Fryxell, the model was forced with meteorological data from Lake Fryxell basin (Figure 1). However, Lake Fryxell ice morphology (and ice thickness) is significantly different than of WLB and parameters related to optical properties of the ice needed adjustments. As such, $\kappa$ and $\chi$, were parameterized (increased) during the simulation. However, albedo data for Lake Fryxell is not available and values used for WLB simulations were incorporated. Predicted ice thickness shows a good measured vs. modeled correlation of $r = 0.90$ ($p < 0.001$) and RMSE of 0.21 m (Figure 7). The increase of RMSE between WLB and Lake Fryxell is most likely a result of poorly constrained albedo data.

Figure 3. Predicted ice thickness changes for West Lobe of Lake Bonney (solid line) and averaged measured ice thickness (circles) from 1996 to 2012. Numerous ice thickness measurements were obtained during the same day, showing large variability. Days with multiple ice thickness measurements were averaged. Error bars are shown as standard deviation.
Figure 4. Linear fit between 16 years of overlapping measured and predicted ice thickness changes for West Lobe of Lake Bonney. RMSE is Root-Mean-Square-Error. Dashed line is a reference line with a slope of 1.

Figure 5. Linear fit between 6 years of overlapping measured and predicted ice ablation for West Lobe of Lake Bonney. RMSE is Root-Mean-Square-Error.
Figure 6. A) A typical water temperature profile of West Lobe of Lake Bonney (data obtained on Nov 21st 2011). B) Modeled evolution of water temperature profile between 1996 and 2012.
Figure 7. Predicted ice thickness changes for Lake Fryxell (solid line) and averaged measured ice thickness (circles) from 1996 to 2012. Numerous ice thickness measurements were obtained during the same day, showing large variability. Days with multiple ice thickness measurements were averaged. Error bars are shown as standard deviation. RMSE – Root-Mean-Square-Error.
H. Model Sensitivity

A sensitivity experiment was designed by increasing/decreasing variables input by 10% of their nominal value, one at a time. The only exception is air temperature, as 10% change would create unrealistic conditions. Instead, the air temperature sensitivity experiment was an increase/decrease by 1°K. A simple modified sensitivity index ($S_i$) was adapted from Hoffmann and Gardner (1983):

$$S_i = 1 - \left( \frac{D_{\text{min}}}{D_{\text{max}}} \right)$$

where $D_{\text{min}}$ is model output (ice thickness) when a variable was decreased, and $D_{\text{max}}$ is the model output when a variable was increased. Sensitivity index values range from 0 to 1, where 1 represents high sensitivity to a change in a parameter; values less than 0.01 indicate no sensitivity (Hoffmann and Gardner, 1983). Each test was run over a 16-year period or until the ice cover melted, which ever came first. The sensitivity experiment was performed on shortwave radiation ($Q_s$), air temperature ($T_a$), bulk transfer coefficients ($C_E$ and $C_H$), albedo ($\alpha$), extinction coefficient ($\kappa$), and a parameter responsible for partitioning shortwave radiation at the surface of the ice ($\chi$). Sensitivity indices for each parameter are summarized in Table 1, from highest to lowest $S_i$, and selected parameters are briefly discussed below.

As expected with models driven by surface radiative heat fluxes, shortwave radiation is the largest contributor of energy at the very surface of the ice. As such, the output of the model will be highly sensitive to parameters controlling shortwave radiation, including absorption ($\chi$) and reflectance ($\alpha$) (Table 1). For example, an increase of $\chi$ or $\alpha$ decreases solar radiation that penetrates an ice cover, decreasing the overall energy of the system that otherwise would be
available for melt. This suggests that the model is highly sensitive to parameters controlling surface optical properties of the ice.

Q_s and T_a have an inverse effect on the ice cover thickness. An increase of Q_s and T_a significantly contributes to the surface ice temperature, as well as penetrating solar radiation, both of which will enhance melt.

The model is relatively insensitive to changes in bulk transfer coefficients (C_E and C_H, respectively). Changes in C_E create small variations in the model’s output, whereas, changes in C_H make virtually no difference to the output. Similar sensitivity results were found by (Vincent et al., 2008). As mentioned previously, C_E and C_H are determined based on atmospheric stability and are assumed constants in this model. The sensitivity test validates my original assumption of constant bulk transfer coefficients, reducing computational complexity of the model.

Table 1. Sensitivity Index (Si) results for parameters used in the model. χ – wavelength depended absorption of shortwave radiation at the surface of the ice, α – albedo, Q_s – shortwave radiation, T_a – air temperature, C_E – bulk transfer coefficient used in latent heat equation, C_H – bulk transfer coefficient used in the sensible heat equation, κ – extinction coefficient.

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I. Discussion

Lake ice thickness is controlled by surface energy balance; hence, it will reflect climatic changes (Reid and Crout, 2008; Wharton et al., 1992). Taylor Valley experienced a general cooling trend between 1986 and 2000, associated with increasing lake ice thickness and increasing shortwave radiation, based on meteorological data from Lake Hoare (Figure 1) (Doran et al., 2002b). Climatic trends between 2000 and 2012, for Lake Hoare, were determined using a method outlined in Doran et al. (2002b) and show a decrease of shortwave radiation by 3.0 W m$^{-2}$ per decade ($p = 0.24$) and increase of surface air temperature by 1.2 °C per decade ($p = 0.10$). Since early- to mid-2000, the increasing ice thickness trends reversed: WLB ice thickness started to decrease in 2002 (Figure 3) whereas ice thickness of Lake Fryxell started to decline in 2006 (Figure 7). The most notable change in the Taylor Valley climate during this time period was an exceptionally warm summer season during 2001/2002 (Barrett et al., 2008; Doran et al., 2008). However, it is not clear whether this discrete warming season was responsible for the trend reversal of the ice thickness.

An experiment was designed where the exceptionally warm season was substituted with meteorological data from an unusually cold season from the year prior (for WLB). The simulation results show virtually no changes in the ice thickness trend whatsoever ($r = 0.98$, $p < 0.001$). Conversely, the unusually cold season meteorological data was randomly forced in the model during the ice thickness decline to test whether it would influence the ice thickness trend. However, no changes in the trend were detected. This indicates that the ice thickness trends in Taylor Valley lakes are not particularly responsive to discrete warming/cooling events. This suggests that a secondary process that is driven on much longer timescale influences the long-
term trends of ice thickness. Since the model only considers atmosphere/ice and ice/water boundaries, the secondary process has to be related to heat fluxes from the water column.

In Antarctica’s perennially ice-covered lakes, heat accumulates in the water column due to the continuous influx of solar radiation during austral summers (Chinn, 1993; Shirtcliffe and Benseman, 1964). Ice covers insulate water from the atmosphere; as a result, heat is trapped and the only way to remove it is through upward conduction (McKay et al., 1985). A consequence is a unique thermal structure of the water column, often exhibiting deep temperature maximum in the water column (Vincent et al., 2008) (example Figure 6). Deep water temperature maximum is a result of a stable water column due to a salinity gradient, which allows the storage of solar energy at some depth (Shirtcliffe and Benseman, 1964; Spigel and Priscu, 1998). However, a temperature maximum takes decades to develop (Vincent et al., 2008); consequently, it varies over long time scales.

The modeled water temperature profile of WLB shows three distinct trends: a continuous increase of deep water temperature over time, a continuous decrease of temperature maximum within about the first 5 meters under the ice cover (Figure 6), and a shallow water temperature (immediately beneath the ice) that varies inversely with the ice thickness ($r = -0.85$, $p < 0.001$) (Figure 8). Annually measured temperature profiles of WLB show similar trends (unpublished data). The only exception in measured data is an increase of temperature maximum after 2009, which is not captured by the model. The cooling of shallow water is associated with an ice thickness increase from 1996 to the end of 2001 (Figure 8). Conversely, the warming of the shallow water is associated with the ice thickness decrease after 2001 (Figure 8). A similar trend of inversely proportional ice thickness and shallow water temperature is observed at Lake Fryxell; however, a trend reversal occurs in 2005 (Figure 9).
The difference in response time of the ice thickness in these two lakes is most likely a result of a different thermal structure of the water column.

Figure 8. Temperature of the shallow water (4 m depth) over time at West Lobe of Lake Bonney and ice thickness change over time. Shallow water temperature and ice thickness changes are inversely proportional.

 Thermal structures of water columns can influence long-term ice thickness trends and preclude a response of the ice thickness to climatic changes. The water column temperature in Taylor Valley lakes is controlled by penetrating solar radiation (Shirtcliffe and Benseman, 1964; Spigel and Priscu, 1998; Vincent et al., 2008). Penetrating solar radiation decays exponentially with depth, Eq. (2). As a result, shallow water receives more energy than deep water and it will be more responsive to seasonal variations of penetrating solar radiation. The energy stored at depth will affect shallow water temperature, to a certain extent, by upward heat conduction due
to the thermal gradient. As an example, for a deep lake with a temperature maximum at some depth, during a climate cooling trend (assuming constant shortwave radiation), shallow waters will cool down, increasing ice growth rates. Yet, shallow water will cool at rates slower than that of the air due to an influx of heat from the deep lake heat storage. The incoming heat from beneath will retard cooling and slow ice thickness growth (Figure 8). Conversely, for shallow lakes that do not have a deep temperature maximum, this process will be less pronounced and the shallow water temperature will more closely reflect climate cooling (Figure 9). The situation would reverse during climate warming; ice thickness of deep lakes with a well-developed temperature maximum would respond to surface air warming quicker due to the additional heat flux from beneath. For example, ice thickness of WLB responded earlier to the increase of surface air temperatures (Figures 3 and 8) between 2000 and 2012 in comparison with ice thickness of Lake Fryxell (Figures 7 and 9). This suggests that lakes with deep temperature maxima can reduce the response of ice thickness to climatic changes.

The observed ice thickness trends in both lakes are a result of the surface energy balance and thermal structure of the water column. The alteration of the ice thickness trend occurs at a threshold dictated by heat fluxes at the ice surface and shallow water. Shallow water heat content, on the other hand, is controlled by penetrating shortwave radiation and heat flux from deep water. As a result, lake ice thickness will respond to climate forcings on different time scales depending on the thermal structure of the lake in question.
Figure 9. Temperature of the shallow water (6 m depth) over time at Lake Fryxell and ice thickness change over time. Shallow water temperature and ice thickness changes are inversely proportional.
J. Model limitations

The one-dimensional thermodynamic model reasonably captures ice thickness trends and water temperature profiles for two different lakes. However, there are several limitations of this model:

1) The model does not account for internal ice melt during the austral summers. A consequence is an occasional internal ice temperature above freezing. The model accounts for the energy associated with the latent heat during the phase change in the internal ice. However, it does not consider further melt associated with isothermal ice. Physics and thermodynamics during an isothermal stage of the ice become complex, as the thermal and optical properties of the ice change. The slight internal temperature increase of the ice is believed to be caused by this issue.

2) The model does not account for lake level increase and associated heat influx from streams.

3) The one-dimensional heat equation captures heat transfer processes within the perennially ice-covered lakes and their stable water column well. The exception is during ice disintegration (Reid and Crout, 2008), which is not addressed in this model.

4) The model does not account for biogeochemical stratification of the water column, which influenced deep temperature maximum (Spigel and Priscu, 1998; Vincent et al., 2008). As a result, the temperature maximum is not correctly captured in the simulations.
K. Applying the Model to Predict Future Ice Cover Trends

I have shown that my model can reproduce 16 years of ice thickness data for two different lakes based on only four atmospheric variables (air temperature, shortwave radiation, relative humidity, and wind speed) and parameterization of only two variables ($\kappa$ and $\chi$). The remaining parameters are either set as constant, spline function (ex. albedo), or a random number generator (ex. cloud cover). This simplistic approach gives credibility for future ice thickness prediction because it does not introduce dependency on an excessive number of parameters. For example, cloud cover plays an important role in determining shortwave and downwelling longwave radiation. Yet, I have shown that cloud cover can be simply substituted by a random number generator.

Future ice thickness simulations were calculated using 24 h intervals for WLB. The initial condition for future ice thickness prediction was obtained from the last value calculated during the 16-year run used in the validation section. The model was forced with daily average meteorological data, calculated based on the data from 2002 to 2012. Results of future ice thickness prediction are shown in Figure 10. Ice thickness increases slightly in the first 10 years of future simulation, due to decreasing shortwave radiation trends based on the averaged data, follow by a sharp decline. The rapid ice thickness decline is associated with an increase of water temperature that has a positive feedback on the ice thickness. The ice thickness trend reversal is attributed to the increasing surface air temperatures. This analysis indicates that perennially ice-covered lakes have a potential to become seasonally ice free in only 20 years. However, it is important to note that without properly constrained surface radiative fluxes, accurate timing of seasonally ice-free lakes is impossible. Rather, this exercise only shows a potential of future ice thickness trend. Any deviations of climatic forcings from the averages used in this simulation
will change the trend of ice thickness. The lake ice thickness trends are already declining; any additional warming or increase of shortwave radiation will only contribute to a more rapid ice thickness decline.

Figure 10. Ice thickness scenario for West Lobe of Lake Bonney. Simulation after 2012 was forced with daily averaged meteorological data based on a ten-year record (from 2002 to 2012).
L. Conclusions

A thermodynamic model parameterized for ice cover of WLB was developed. The model is driven by surface radiative heat fluxes and heat fluxes from the underlying water column, coupled with a one-dimensional heat equation. The water temperature profile is calculated in order to properly constrain heat fluxes at the ice/water boundary. The model successfully calculates ice thickness changes as well as the temperature profile of the ice and the water column. Predicted ice thickness changes are in agreement with trends observed in measured ice thickness data. The predicted water temperature profile differs slightly from observed data because the model does not account for stratification of the water column. Vincent et al. (2008) found that the thermal structure of perennially ice-covered lakes is influenced by the absorption of solar radiation due to biogeochemical stratification of the water column. However, the model was still able to generate a deep temperature maximum (granted at slightly lower depth than observed) and showed its evolution over time.

Energy at the surface of the ice is the main driver of ice thickness; however, the response of ice thickness to climatic changes is strongly affected by the heat flux from the water column. Deep perennially ice-covered lakes with a deep-water temperature maximum will impede the response of ice thickness growth to surface air-cooling. The surface air-cooling is hindered by the heat flux from the water column. Conversely, during surface air warming, the water temperature maximum facilitates ice decay, due to an additional heat flux from below. Ice thickness of shallow lakes, on the other hand, will more accurately respond to climatic changes because of diminished effect of the water column temperature on the ice thickness. Hence, ice covers can be used as indicators of climate change; however, the ice thickness trends should be interpreted with caution due to the influence of thermal stratification of the water column. Ice
thickness models exclusively driven by the surface radiative fluxes might elude to the ice thickness evolution over long-time scales because thermal structure of the water column significantly contributes to the ice thickness evolution.

Preliminary lake ice thickness predictions based on historical averages can provide a valuable insight into the stability of a system and its potential evolution. The results indicate that the thickness of ice covers of Taylor Valley lakes is in fact declining and the lakes can become seasonally ice free in approximately 20 years.
M. Acknowledgments

This research was supported by the Office of Polar Programs (grants 9810219, 0096250, 0832755, 1041742, and 1115245). Logistical support was provided by the US Antarctic Program through funding from NSF.
# Appendix A

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<th>Variables</th>
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CHAPTER V
CONCLUSIONS AND FUTURE RESEARCH DIRECTIONS

The relatively pristine nature of Taylor Valley lakes provides an opportunity to study lake responses to climatic changes, past and present. Because this region of Antarctica is poised on a knife-edge between freezing and melting in the summer months, small changes in the climatic forcings deeply affect hydrologic processes, lake habitat, biodiversity, and primary production. Constraining past and present hydrological drivers is integral to understanding the future of these lakes.

In Chapter II, warm anomalous summer winds were investigated as a mechanism responsible for formation of large glacial lakes during the Last Glacial Maximum. Based on data from an ice core record obtained from Taylor Dome, the climate during the Last Glacial Maximum was colder, more arid, and windier than today. Such a climate is inconsistent with the formation of large lakes, however, robust geomorphic and geologic evidence points to their existence. In recent years, Taylor Valley had experienced an anomalously warm summer associated with anomalous warm winds and sharply rising lake levels. These winds also existed during the Last Glacial Maximum (LGM); however, they were more persistent, as suggested by aerosol loading and $^{10}$Be concentration obtained from the Taylor Dome ice core record. This thesis has postulated that the higher frequency of warm summer winds during the LGM is attributed to the existence and sustainability of large glacial lakes, despite colder mean annual temperatures. Models developed in this thesis show that the high frequency of warm summer winds could result in the filling of Taylor Valley by Glacial Lake Washburn in approximately 900 years.
Anomalous warm summer winds are not preserved in the $\delta^{18}O$ record obtained from the Taylor Dome ice core because: 1) the winds are low-elevation phenomena, whereas $\delta^{18}O$ is fixed at the bottom of the cloud which is at much higher elevation, 2) postdepositional alteration of $\delta^{18}O$ would mask any seasonal warming, and 3) ice core records are annually averaged. However, during the LGM those winds were more frequent and stronger, suggesting the summer season during the LGM could have been as warm as the present, generating a large volume of melt water.

Several questions remain unanswered: was Glacial Lake Washburn seasonally ice-free? A seasonally ice-free lake would be subjected to wind induced water mixing that would affect primary productivity of the lake and its legacy on contemporary ecology. A seasonally ice-free lake would also be subjected to higher evaporation rates, which could inhibit the lake’s growth/sustainability. Stronger winds would either delay seasonal ice growth or facilitate ice disintegration. The question then arises: what is the minimum ice thickness of Glacial Lake Washburn required to resist frequent strong winds? These questions can be answered by performing thermodynamic ice modeling using assumed climatic conditions derived from the Taylor Dome ice core record. The ice model would have to be coupled with the atmosphere and the underlying water column in order to account for heat fluxes from the water column. In addition, it would be necessary for the model to address the ice-free period.

In Chapter III, high-resolution spatial biogeochemical and physical processes of a perennially ice-covered lake were analyzed, for the first time. Limnological studies are usually limited to data obtained at the deepest point of the lake, which is assumed to represent an entire lake. This project has shown that perennially ice-covered lakes cannot be considered homogenous as the spatial ice thickness varies significantly. Ice thickness variations are due to
either shading by nearby mountains or surface sediment accumulation, which effectively changes the amount of penetrating solar radiation and chlorophyll-a distribution within the water column. Solar radiation and chlorophyll-a are negatively correlated. This paradox is as a result of short-term photoadaptation of phytoplankonic communities to the amount of available light by changing the size of light harvesting antenna size.

The physical and biological heterogeneity of West Lobe of Lake Bonney needs to be considered when estimating biomass of this ecosystem. The chlorophyll fluorescence does not accurately reflect spatial biomass distribution within the water column. Yet, it has been shown by Lizotte and Priscu (1992), that natural fluorescence can be used as an estimator of biomass and primary productivity. The findings of this chapter do not negate Lizotte and Priscu (1992) findings; rather this research poses a need to spatially calibrate chlorophyll-a concentration using standard water sampling and chlorophyll fluorescence. Such calibration would provide further confidence in using fluorometers.

In Chapter IV, long-term trends of ice covers were analyzed using a physics-based model. The model was driven by the surface radiative heat fluxes and heat fluxes from the underlying water column. Ice thickness changes were predicted for two different lakes to a high degree. The thicknesses of perennially ice-covered lakes reflect climatic changes. However, the response of ice thickness can be hindered or facilitated by the heat flux from the water column. This is especially evident in lakes with deep-water temperature maximum. As a result, ice covers of shallow lakes are more sensitive to climatic changes. This research emphasizes the importance of properly constraining ice cover models (coupling of the water column with the ice cover) for long-term ice thickness predictions (more than one year).
The results of this project can be used to identify the lakes most susceptible to climate change. However, each lake has unique thermal characteristics, which are driven by density stratification. The model would require the inclusion of water density stratification with depth for each lake in order to accurately determine thermal evolution of the water column. Inclusion of water density stratification and physics describing seasonally ice-free lakes in the model could also be used to predict future thermal evolution of the lakes during ice-free periods.

Ice thickness modeling, driven by surface radiative balance, is often limited by the available meteorological datasets. Cloud cover datasets are often unavailable, yet they play an important role in surface energy balance. Shortwave and downwelling longwave radiation are dependent on cloud cover, which often has to be estimated. In the model developed here, I used a random number generator for the cloud cover. Based on ten model runs for West Lobe of Lake Bonney, the ice thickness trends were accurately predicted with the largest RMSE = 0.13 m. However, the runs were determined to be statistically significantly different based on one-way ANOVA \[ F(9, 116790) = 136.4, p < 0.001 \] (Figure 1). Despite the statistical significance, the RMSE for all runs indicates a very good fit. The small variation of the model output (Figure 1) suggests that the random number generation for the cloud cover is an alternative substitution for missing data for models concerned with long-term ice thickness changes. However, for modelers concerned with fine-scale resolution (hours to months) lack of cloud cover data could yield erroneous surface energy balance. Comparison of model results with measured and artificially generated cloud cover data would profoundly benefit the modeling community and give confidence in using random number generators as an alternative option for missing cloud cover datasets. Similar to cloud cover data, ice thickness models rapidly respond to albedo changes due to its importance on the surface energy balance and penetration solar radiation.
Albedo of perennially ice-covered lakes varies on seasonal scales and its determination is imperative in ice thickness modeling. Collection of long time series of cloud cover and albedo might not be a requirement for modeling ice thickness based on surface radiative balance. However, such datasets would allow for proper calibration and development of empirical relationships that could be implemented in future models.

Figure 1. Box plot of ten simulations for West Lobe of Lake Bonney. Middle line in the box is a median; outer edges of the box are the 25th and 75th percentiles.

Warm summer westerly winds have been long recognized for their role in seasonal climate regime. However, the impact of these winds is more significant than previously thought; warm summer westerly winds were responsible for the sustainability of large glacial lakes during the LGM, despite an overall colder climate. Strong winds are responsible for delivering aeolian sediment to the surface of the ice cover, which decreases ice thickness. The ice thickness
controls the amount of penetrating solar radiation, which affects chlorophyll-a distribution and contributes to the development of deep-water temperature maxima within the water column. Deep-water temperature maxima contribute significantly to the ice cover thickness trends, which might confound interpretation of the ice thickness as a proxy for climatic changes.
VITA

EDUCATION

Ph.D., Earth and Environmental Sciences, 2014
University of Illinois at Chicago
Thesis: “Hydrological and Biogeochemical Modeling of Taylor Valley Lakes, East Antarctica”

B.S., Earth and Environmental Sciences, 2008
University of Illinois at Chicago

RESEARCH / EMPLOYMENT HISTORY

Research Specialist, 2008–present
University of Illinois at Chicago
• Analyze long-term limnological/hydrological datasets.
• Tabulate written and oral reports.
• Supervise undergraduate students’ projects.
• Planned and executed field work in McMurdo Dry Valleys, Antarctica for the Long Term Ecological Research (LTER) Project.
• Installed and maintained automated lake monitoring stations, collected and analyzed data, and collaborated with international scientific teams.
• Antarctic Field Team Leader from 2008 to 2012.
• Maintained lab equipment and scheduled lab use.
• Managed all experiments and implemented safety regulations.

Field Assistant, Summer 2007
Department of Energy (DoE), Waste Isolation Pilot Plant - New Mexico
• Assisted in extraction of gases out of deep aquifers.

Field Assistant, Austral Summer 2006-2007
Long Term Ecological Research Project - Antarctica
• Assisted in data collection and drilled ice holes.
• Upgraded and maintained scientific equipment.

PROGRAMMING/COMPUTER SKILLS

• Programming of Campbell Scientific dataloggers and sensors
• Proficient with Matlab and ArcGIS

PUBLICATIONS

PUBLICATIONS IN REVIEW / PREPARATION

Obryk, M. K., Doran, P. T., and Priscu, J. C., in prep. The permanent ice-cover of Lake Bonney, Antarctica: The influence of thickness and sediment distribution on photosynthetically available radiation and chlorophyll-a distribution in the underlying water column.


Spigel, R.H., Priscu, J.C., Stone, W., Obryk, M.K., Fabretti, A., and Doran, P.T., in prep. The Physical Limnology of a Permanently Ice-Covered and Chemically Stratified Antarctic Lake Using High Resolution Spatial Data from an Autonomous Underwater Vehicle

IN PROGRESS PROJECTS

- Decadal heat accumulation in ice-covered lakes of the McMurdo Dry Valley, Antarctica.
- Circadian response of microbial communities in ice-covered lakes in Antarctica.

ACADEMIC AWARDS

SCAR Travel Fellowship, 2011
GLEON Travel Fellowship, 2010
Illinois Space Grant Consortium Fellowship, 2009-2010
Graduate Citizenship Award, UIC, 2008-2009

LANGUAGES

Fluent in English and Polish