Title: Lake Vanda: A Sentinel for Climate Change in the McMurdo Sound Region of Antarctica

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Lake Vanda: A Sentinel for Climate Change in the McMurdo Sound Region of Antarctica

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Abstract

Lake Vanda is a perennially ice-covered, meromictic, endorheic lake located in the McMurdo Dry Valleys of Antarctica, and an exceptional sentinel of climate change within the region. Lake levels rose 15 m over the past 68 years in response to climate-driven variability in ice-cover sublimation, meltwater production, and annual discharge of the Onyx River, the main source of water to the lake. Evidence from a new bathymetric map and water balance model combined with annual growth laminations in benthic mats suggest that the most recent filling trend began abruptly 80 years ago, in the early 1930's. This change increased lake volume by >50%, triggered the formation of a new, upper, thermohaline convection cell, and cooled the lower convection cell cooled by at least 2°C and the bottom-most waters by at >4°C. Additionally, the depth of the deep chlorophyll *a* maximum rose by >2 m, and deep-growing benthic algal mats declined while shallow benthic mats colonized freshly inundated areas. We attribute changes in hydrology to regional variations in air flow related to the strength and position of the Amundsen Sea Low (ASL) pressure system which have increased the frequency of down-valley, föhn winds associated with surface air temperature warming in the McMurdo Dry Valleys. The ASL has also been implicated in the recent warming of the Antarctic Peninsula, and provides a common link for climate-related change on opposite sides of the continent. If this trend persists, Lake Vanda should continue to rise and cool over the next 200 years until a new equilibrium lake level is achieved. Most likely, future lake rise will lead to isothermal conditions not conducive to thermohaline convection, resulting in a drastically different physical, biogeochemical, and biological structure than observed today.

1. Introduction

Lakes are widely regarded as sentinels of climate change (Williamson et al., 2009; Adrian et al., 2009; Schindler et al., 2009) as many of their physical, chemical and biological properties are sensitive to variations in meteorological and hydrological conditions. These include ice phenology (Magnuson et al., 2000; Dugway, 2006) and surface water temperature (Austin and Colman, 2007; Schneider et al., 2009; Schneider and Hook, 2010). The water level and salinity of endorheic (i.e., closed-basin, with no outflow) lakes are particularly sensitive to changes in the lake water balance (Adrian et al. 2009; Schneider et al., 2009). Lakes are the lowest points within watersheds, and as such, receive and integrate climate effects across landscapes, while the wide geographical distribution of lakes provides the ability to detect and monitor climate change at regional, continental, hemispheric, and global scales (Williamson et al., 2009).

Polar and sub-polar lakes in particular are considered some of the most sensitive to climate change (Vincent et al., 1998; Lyons et al., 2006; Adrian et al., 2009; Williamson et al., 2009; Vincent et al., 2008; Paquette et al., 2015). With a large volume of glacial ice and snow stored in watersheds, and summer air temperatures near the melting point of water, these lakes quickly respond to slight changes in the energy balance. Researchers have found evidence of warming trends in lakes in Alaska, USA (Engstrom et al., 2000; Surdu et al., 2014), Ellesmere Island, Canada (Mueller et al., 2009; Veillette et al., 2010; Paquette et al., 2015), Signy Island, Antarctica (Quayle et al., 2002), and the Antarctic Peninsula (Lyons et al., 2006). Perennially ice covered, meromictic lakes in the interior of Antarctica also show evidence of changing climate through changes in water balance leading to increases in lake level. Lake Untersee in the Schirmacher Oasis (71°S) showed little change from 1939 to 1981, and then increased in depth between 1981 and 1991 (Wand et al., 1997). The level of Lake Wilson in the Darwin Valley of Southern Victoria Land (80°S) increased 25 m between 1975 and 1993, a 54% increase in volume (Webster et al., 1996). Lake levels in the McMurdo Dry Valleys (77°S; MDV) of Southern Victoria Land, have risen substantially and steadily since the first observations were made by Sir Robert Falcon Scott's *Discovery Expedition* in 1903 (Chinn, 1993; Bombliet et al., 2001; Scott, 1905). Detailed records of level change in the MDV lakes have been made for nearly 50 years, through the efforts of the New Zealand Antarctic Programme (1970-present),

and the United States, McMurdo Dry Valleys, Long Term Ecological Research (MCM-LTER) Program (1993-present), and provide one of longest and richest environmental datasets on the continent.

Situated in the Wright Valley (Figure 1), Lake Vanda is the deepest and largest of the MDV lakes, has the most complete direct record of level and inflow volume. This paper seeks to summarize recent changes in this lake, place these changes in the context of historic changes, link these changes to local and regional drivers of climate change, and predict future changes to the lake. To achieve these goals, Section 2 describes the setting of Lake Vanda and Section 3 provides a detailed review of the lake level record over the past 25,000 years. In Section 4, we discuss changes to the physical characteristics of Lake Vanda which began in the early 1930's and involved the development a new thermohaline convection cell and reductions in ice thickness. Section 5 presents new bathymetric and water balance models for Lake Vanda, and uses these to argue that an abrupt change in Onyx River discharge and lake storage occurred 80 years ago. In Section 6, we discuss how lake level rise has impacted the microbiology of Lake Vanda. Section 7 summarize the current understanding of climate drivers for the region over the past 80 years, namely, the relationship between air pressures in the Amundsen Sea and the frequency of föhn winds in the MDV, and links changes in the MDV to observed changes on the Antarctic Peninsula. Finally, in Section 8, we predict the impact of future climate change on Lake Vanda, and conclude in Section 9 that Lake Vanda is one of the most valuable, indicative, and sensitive sentinels of climate change in the region.

2. Lake Setting

2.1 Geography

The MDV are situated in South Victoria Land, Antarctica, between the Transantarctic Mountains (~160°E) to the west and the Ross Sea (~164°E) to the East, and extend from Mt. Douglas (76° 31.25'N) in the north to Highway Ridge (78° 23.97'N) to the south. These valleys constitute the largest ice-free region on the continent of Antarctica. They exhibit a cold desert climate with mean annual temperatures ranging from -14.8 °C to -30.0 °C (Doran et al., 2008) and receive less than 50 mm of precipitation annually (Fountain et al., 2010). The MDV host 15 large, perennially-ice covered, meromictic lakes, and many smaller ponds. These include Webb

Lake in the Barwick Valley to the north; Lake Vida and Victoria Upper Lake in the Victoria Valley; lakes Brownworth and Vanda in the Wright Valley; Lake Joyce in the Pearse Valley; lakes Fryxell, Hoare, Chad, and Bonney (divided into an East and West Lobe) in the Taylor Valley; lakes Buddha and Garwood in the Garwood Valley; Lake Miers in the Miers Valley; and Trough Lake in the Trough Lake Catchment to the south. Between mid and late summer (December-January), several of these lakes tend to develop open-water channels along the edges of the lakes, called moats, that separate shorelines from perennial lake ice.

Lake Vanda (77° 31.8' S, 161° 34.6' E) occupies the lowest point (94.5 m amsl in January 2014) in Wright Valley between the Olympus Range (~1360 m amsl) to the north and Asgard Range (~1810 m amsl) to the south (Figure 1). This ice-free valley, trending predominantly in the East-West direction, is approximately 50 km long stretching from the Polar Plateau toward the Ross Sea (Hall and Denton, 2005) with the valley slopes being steeper on the northern side than on the southern side. To the west, Wright Upper Glacier, an outlet glacier from the Polar Plateau (Taylor Dome) and the entry point for föhn winds, drains seaward approximately six kilometers into the valley (McKelvey and Webb, 1962). At its eastern terminus, the valley is blocked by Wilson Piedmont Glacier, a maritime ice mass abutting the Ross Sea, with an outlet tongue called the Wright Lower Glacier, which flows westward into the valley (Hall and Denton, 2005). Morphologically, Lake Vanda is 35 meters deeper, 0.6 km longer, and 1.1 km wider than Lake Bonney, the second largest lake in the MDV (Spigel and Priscu, 1998), and thus, it stores the largest volume of liquid water in the region.

Figure 1. Location map showing the position of Lake Vanda between the Wright Upper Glacier (an outlet of East Antarctic Ice Sheet from the Polar Plateau) to the west and the Wright Lower Glacier (an outlet of the Wilson Piedmont Glacier) abutting the Ross Sea to the east. The Onyx River flows inland (westward) for 28 km, beginning at Lake Brownworth and ending at Lake Vanda.

2.2 Geology

The current geographical configuration of the MDV is a result of the geologic and glacial history of the region. Granitic rocks and the metamorphosed basement of Precambrian age were peneplained after the Ross Orogeny (early Paleozoic) and subsequently overlaid by a thick layer of sediment belonging to the Beacon Supergroup of Devonian to Triassic age (Gleadow and Fitzgerald, 1987; McKelvey and Webb, 1962). The basement complex and Beacon Supergroup were subsequently intruded by the massive Ferrar Dolorite (diabase) sills during the Jurassic period (McKelvey and Webb, 1962) and localized basaltic intrusions of Cenozoic volcanic activity (Hall and Denton, 2005). The main valleys were shaped by a combination of tectonic uplift (up to 5 km), and fluvial and glacial processes that began during the Cenozoic and essentially halted by 15 Ma.

Wright Valley and adjacent McKelvey Valley have been arid, free of valley-wide glaciers, and subject to minimal erosion for the past 14 Ma based on preserved ashfall deposits (Marchant et al., 1993a; 1993b) and lacustrine fossils (Lewis et al., 2008). This evidence also indicates that the East Antarctic Ice Sheet has been stable throughout glacial and interglacial cycles (Denton et al., 1993; Gleadow and Fitzgerald, 1987; Marchant et al., 1993a; Marchant et al., 1993b; Sugden et al., 1995; Summerfield et al., 1999). Seawater flooded the central Wright Valley in the late Miocene as evident by marine shells dated at 9 ± 1.5 Ma found in a glaciomarine diamicton (Prentice et al., 1993; Sugden et al., 1995).

3. The Lake Level Record

3.1 Last Glacial Maximum – Glacial Lake Wright

Large, proglacial lakes supposedly occupied most of the major valleys during the Last Glacial Maximum (LGM) (Hall et al., 2001; Hall et al., 2002; Hendy, 2000; Pewe, 1960; Stuiver et al., 1981). The Wilson Piedmont Glacier is thought to have expanded and blocked the mouth of the Wright Valley as a consequence of the westward advance of the Ross Ice Sheet and, according to Hall et al. (2001) and Hall and Denton (2005), this ice impoundment permitted the formation of a large, proglacial lake, called Glacial Lake Wright, which filled much of the valley. Hall et al. (2001) inferred the existence of this large lake from relict shorelines present at 566 m amsl, deltas located approximately 465 m amsl, and glaciolacustrine deposits present at 386 m amsl (for reference, Lake Vanda had a surface elevation of 94.7 m amsl on January 31, 2014; MCL-LTER database). Hall et al. (2001) dated the maximum lake level between ~14,500 to 15,300 ^{14}C yr BP using algal mats found in the high elevation deltas. More recent evidence suggests that the lake level fluctuated (filled and drained) every several thousands of years over the past 25 k years in response to climate fluctuations (Hall et al., 2010). At the 566 m elevation highstand, these authors estimate that Glacial Lake Wright would have been approximately 550 m deep with a surface area of 210 km² (Hall et al., 2001). Geomorphic evidence of large LGM lakes is found in the Taylor Valley (immediately south of, and parallel to, Wright Valley; Figure 1), which has abundant relic shorelines and perched deltas (e.g. Hall et al., 2000; Stuiver et al., 1981; Toner et al., 2013).

The existence of these large LGM lakes in the MDVs has, however, been questioned for at least a decade. Arguments against their existence are based around the absence of clear transitions in soil chemistry across proposed highstands (Bockheim et al., 2008), and on climate; the LGM is known to have been colder and dryer than present conditions, and as such, the mechanism for producing such a large volume of liquid water is still under debate. Recently, McGowan et al. (2014) suggested an alternate interpretation of topographic features previously thought to provide evidence of a large, proglacial lake in Victoria Valley (immediately north of, and parallel to, Wright Valley; Figure 1), and inferred that this casts doubt also on whether other large proglacial lakes existed as well. It is possible that paleo deltas and organic deposits could have been formed not by large lakes but by smaller lakes juxtaposed between the valley walls and intruded ice (Toner et al., 2013).

3.2 Holocene Lake Levels

More definitive evidence of higher lake levels, and level change comes from undisputed, Holocene paleo-shorelines clearly visible above Lake Vanda today at an elevation of 155 m amsl (Figure 2A). Smith and Friedman (1993) radiocarbon dated readily identifiable algal mats (Figure 2B) in the highest shorelines between ~ 2000 to 3000 ^{14}C yr BP. The age of the highest Holocene lake level coincides with a warm period in the Ross Sea inferred from the “Penguin Optimum,” when Adelie penguins reached their maximum Holocene southern distribution (Emslie et al, 2007).

Over the 1000 to 2000 years following the high stand, Lake Vanda is thought to have lost its ice cover and nearly completely desiccated (Wilson, 1964; Matsubaya et al., 1979). At its minimum extent, the lake had contracted to a shallow, ice-free, hypersaline pond some 4-5 m deep (Wilson, 1964). Lyons et al. (1998) report a similar desiccation event occurred at the same time in the neighboring Taylor Valley suggesting this was a regional shift toward a colder, dryer climate. Coincident with this desiccation period, Adelie penguins relocated to more northerly locations, presumably in search of warmer temperatures (Emslie et al., 2007).

Figure 2. (A) View of Lake Vanda looking west towards the Polar Plateau showing several prominent paleo-shorelines up to 130 m amsl, or 60 m above the January 2014 lake level. A New Zealand dive hut from December 2013 is shown in the center of the image. (B) Microbial mat buried under wind-blown sand at the highest Holocene paleo-shoreline (white arrow in A).

Identification of the Holocene low-stand and subsequent evolution of the lake is largely based on considerations of the current chemical meromixis in the lake. At present the bottom of the lake is hypersaline, overlain with layers of fresher water (Figure 3). The deep brine is calcium-chloride based, and evaporation models have shown that evaporating Onyx River water alone would not generate the calcium-rich brine observed at the lake bottom today (Matsubaya et al., 1979; Green and Lyons, 2009). Instead, during the low stand, Lake Vanda is thought to have received groundwater from a deep reservoir associated with the Don Juan Pond basin composed of a granitic, calcium-chloride-brine (Cartwright and Harris, 1981; Green and Canfield, 1984; Carlson et al., 1990; Takamatsu et al., 1998; Green and Lyons, 2009). Such brines are globally

unique to Lake Vanda, and the origin of the brine has not yet been determined (Lyons and Mayewski, 1993; Green and Lyons, 2009).

The current model of the generation of the modern profile requires a shift in climate that enhanced meltwater production and increased discharge in the Onyx River (Lyons et al., 1998). Canfield et al. (1995) demonstrated that the density contrast between influent fresh water from the Onyx River and the pre-existing brine would have been sufficient to result in a stable stratification. Once the surface of the lake was sufficiently fresh for ice formation to occur, this will have reinforced the stratification of the filling lake through the freeze-out of salts (Miller and Akin, 1996; Wait et al., 2006), and protection from wind-driven mixing.

The modern vertical profile of conductivity (Figure 3) thus shows a basal brine, the relic of the low-stand hypersaline pool, overlain by two layers of fresher water from the refilling. Using a model of two immiscible fluids, Onyx River water overlying a small brine pool, Wilson (1964) showed that it would take 1200 years to produce the observed salt profile through vertical molecular diffusion. This duration was later corrected to 1000 years by Roberts and Wilson (1965). A mixing model based on ^{36}Cl tracer by Carlson et al. (1990) supported Wilson's model that modern Lake Vanda was predominantly derived from these two sources. The estimated time of the start of refilling, 1000 years ago, would have required a change in climate, and coincides with the Medieval Warm Period (MWP - 1200 to 700 yr BP) in the Northern Hemisphere. While recent analyses suggest that the MWP did not extend to the entire southern hemisphere (Neukom et al., 2014), though Bertler et al. (2011) conclude from snow core analyses that the McMurdo Dry Valleys were 0.35°C warmer during the MWP.

Figure 3. Temperature and conductivity profiles from Lake Vanda collected on December 16, 2013, beginning at the lake surface (94.7 m amsl). Left diagram denotes the upper convection cell (UCC), upper-pycnocline (UP), lower convection cell (LCC) and diffusion zone (DZ). The right image enlarges the conductivity scale to highlight stepwise changes occurring with depth. Data provided by Hilary Dugan, Amy Chiuchiolo, John Priscu, and the MCM-LTER database.

3.3 Recent Lake level Change

Direct observations over the past 69 years show that lake level rise has steadily continued through to the present day (Figure 4). The US Geological Survey conducted aerial photographic surveys over Lake Vanda since 1947. Figure 5 shows a compilation of these images gathered from the Polar Geospatial Center (University of Minnesota), the Antarctica New Zealand, Digital Asset Manager (ADAM), and the Google Earth satellite image database. The largest increase in

surface area occurred between January 18, 1983 and January 16, 1985 when the Western Lobe was formed by overflow from the main basin. Evidence of overflowing is seen in images as early as the 1982/83 field season, a year which exhibited a series of flood events on the Onyx River that continued through to 1988 (Figure 4, 5).

Figure 4. Lake level recorded from the “Astro A” benchmark from 1961-2014 and interpolated to the 1947 lake level (see Bathymetry Section 4.3).

Figure 5. Aerial photographs of Lake Vanda, 1947-2013. Distances in captions are the elevation difference of the lake surface between the time the photo was taken and January, 2013.

4. Changes to Physical Characteristic

4.1 Position and Temperature of Convection Cells

Recent increases in the volume of the lake have almost entirely been through an increase in the thickness and volume of the upper convection cell (Hawes et al., 2013). During the 1962-63 field season, Ragotzkie and Likens (1964) described a near identical temperature-conductivity profile to that shown in Figure 3, except that the 20-m-thick lower convection cell was separated by a 1-2-m-thick upper pycnocline from a 4-m-thick upper convection cell (Figure 6). Examination of a time series of temperature profiles of the lake since 1960 (Figure 6) shows how change has had two significant impacts on lake structure. First, given the difficulties in comparing profiles across decades made with different instruments in a lake with varying water level, we conclude that the position of the diffusion zone, lower convection cell and upper pycnocline have been constant over time, but that the thickness of the upper convection has steadily quadrupled from 5 m to 20 m over this period (Figure 6). Secondly, the temperature of the lower convection cell has decreased by $\sim 2^{\circ}\text{C}$ and the bottom temperature by $\sim 4^{\circ}\text{C}$, while the upper convection cell has remained at close to or just above 4°C . To emphasize this point, Figure 7 shows trendlines in temperature for the same period recorded at consistent elevations within the upper convection cell (84 m elevation), the lower convection cell (54 m), and near the lake bottom (11 m). Heating of the deeper layers of the lake is solar (Ragoitzkie and Likens, 1964) and the decline in deeper temperatures is likely to be related to the increased adsorption of light associated with the upper convection cell.

4.2 Formation of the upper convection cell

An inference of observations that the upper convection cell has increased from <4 to >20 m thickness since earliest records is that it may be a relatively recent phenomenon, perhaps dating back to a few years or decades before 1960. Understanding of how the current structure may have evolved requires consideration of thermohaline convection. The convective cells in Lake Vanda are generated by the combined effects of total dissolved solids (i.e. salinity) and temperature on density (Turner, 1974; Kelley et al., 2003). What would otherwise be unstable temperature gradients – where water is warmer at depth – are density stabilized by salt. Heat conducts (diffuses) much faster than salt, which allows instabilities to develop at interfaces. For

instance, the diffusion of heat into the base of the lower convection zone at a faster rate than salt diffusion can create a package of warm, buoyant water that is displaced upwards until it either loses heat by conduction or reaches the top of the cell and water of lower density. Cooling at the upper boundary of the lower convection cell would conversely create a dense packet of water that would sink until it warmed or reached water of equal density.

Persistence of the upper convection cell depends on a density gradient generated by its lower salinity offsetting its lower temperature. Convection will occur though warming above 4°C at the base, and cooling back to 4°C at the surface. However, gradual salinization of the upper cell by salt exclusion during winter ice formation would, if the lake were stable or dropping in level, eventually create a condition whereby the upper cell would reach sufficient density to merge with the lower. The creation and persistence of an upper convection cell is therefore dependent on a period of rising lake level combined with continuous heating from below via solar radiation. The inception of the upper cell was dependent on a shift from a period of stable or falling levels to one of level rise, and the age of the upper pycnocline is likely to date that transition. Using the development of benthic vegetation along depth profiles, Hawes et al. (2013) argued that the inception of the upper convection cell could be dated to approximately 80 years before their publication, or the early 1930's.

Figure 6. Observed profiles of temperature from the water column of Lake Vanda. Panel A shows a compilation of data from 1962 to 1963 (Ragotzkie and Likens, 1964) and 1993 to 2007 (MCM-LTER database). The segments of relatively invariant temperatures represent the upper and lower convection cells. The relative position of these cells remains constant with time, despite substantial reduction in temperature. Panel B presents a simple comparison of the data collected the furthest apart in time, 1960 to 2010, and shows obvious shifts in temperature profiles (i.e. general cooling since 1960).

Figure 7. Compilation of temperatures reported at consistent elevations from 1960 to 2010 in (A) the upper convection cell at 10 m depth, (B) the lower convection cell at 40 m depth, and (C) near the lake bottom at 73 m depth. Depths are relative to 2014 lake levels. Trend lines, trend line equations, and correlation coefficients provided.

4.3 Reduction in ice thickness

The ice cover of Lake Vanda has also changed over the period of record. Between 1960 and 1975, ice thickness decreased from approximately 4 to 3 m (Figure 8). Since 1975, the ice cover has remained relatively stable with a mean thickness of 3.0 m and a standard deviation of ± 0.28 m. Long-term ice thickness changes on stratified perennially ice-covered lakes is a result of the thermal profile of the water column and climate variability (Obryk et al., Accepted).

Figure 8. Ice thickness on Lake Vanda from 1960 to present. Black circles represent individual measurements, the dashed line is a five year running average. Data are from published and unpublished sources.

Generally speaking, thin ice covers allow for a steep temperature gradient between the air/ice interface and the ice/water interface, promoting upward heat loss and basal freezing during winter. Thicker ice has the reverse effect of reducing the temperature gradient and decreasing new ice formation. These processes are primarily driven by the surface energy balance. At steady state, the net heat loss via ablation at the air/ice interface is balanced by latent heat released at the ice/water boundary which is conducted upward (McKay et al, 1985). However, new, bottom-down, ice growth can be hindered by heat from warmer water below (Obryk et al, Accepted). Thus, thin ice cover is most likely to be produced by high ice ablation rates (McKay et al., 1985), warm air temperature, low thermal conductivity of ice, and/or warm water close to the underside of the ice.

In the case of Lake Vanda, the ice thickness difference between 1960 and 2010 implies that the air temperature was colder earlier in the century, the ablation rate was lower, or the heat flux from the water column was less. Of these, the ablation rate appears to be rather constant over time, and air temperature has changed little (discussed in Section 5.2). However, during the 1960's, the temperature gradient immediately below the ice/water interface was less steep than

today (Figure 6B). This condition most likely produced a low heat flux across the ice/water interface and allowed for thicker ice formation between 1960 and 1975. After 1975, the temperature gradient below the ice/water interface became steeper as the upper convection cell grew in thickness (Figure 6B). This change will have increased the heat flux across the ice/water interface resulting in thinner ice conditions.

5. Changes in the Lake Water Balance

5.1 Onyx River Discharge

The New Zealand Hydrological Network recorded discharge of the Onyx River at the Vanda stream gauge above the East Lobe from 1969 until the early 1990's (Chinn, 1993; Chinn and Mason, 2015), and the MCM-LTER recorded discharge from 1993 to the present. This nearly-continuous, 45 year record shows a highly irregular flow regime with discharges ranging from 0 to 2.3×10^7 m³/yr (Figure 9), and an average annual discharge of 4.07×10^6 m³/yr (equivalent to ~2% of the lake's current estimated volume). Doran et al. (2008) attributed the variability of MDV stream discharge to the frequency of warm, down-valley föhn winds. Föhn winds are driven by the pressure gradient between the high air pressures above the pole and below-average air pressures in the Ross Sea, known as the Ross Sea Low (Spiers et al., 2013). These air currents warm adiabatically as they descend from the Polar Plateau and move toward the coast. Based on Doran et al.'s (2008) model, summers with frequent down-valley winds have warmer air temperatures, higher meltwater production, and higher stream discharge compared to summers with infrequent down-valley winds.

Figure 9. Forty five year flow record for the Onyx River at the Vanda stream gauge, 1969 to 2014. Mean annual discharge was $4.07 \times 10^6 \text{ m}^3/\text{yr}$ over this period. Values from the MCM-LTER database.

5.2 Interior Climate and Lake Water Losses

Water removed from Lake Vanda during winter months is only through ice sublimation and during summer months through a combination of sublimation, surface ice melt followed by evaporation, and direct evaporation of water at the marginal open-water “moat”. These losses are a function of local relative humidity, wind speed, air temperature, and solar radiation. Continuous measurements of meteorological conditions have been collected by the MCM-LTER since 1994 and show that the Vanda Basin is cold, windy and dry. Average air temperature and relative humidity are -19.4°C and 60.9%, with the latter split distinctly into summer (November through January) $40.9 \pm 3.9\%$ (mean $\pm$ s.d., $N = 20$) and winter (April to September) $70.0 \pm 3.4\%$. There are no significant linear trends over the two decade instrumented period. Wind velocity at Lake Vanda from 1994 to 2014 shows distinct summer and winter patterns. In summer, there is a diurnal variation in wind with a minimum at 06:00 of $3.8 \pm 2.9 \text{ m/s}$ and a maximum at 18:00 of $7.7 \pm 2.3 \text{ m/s}$ (local time). In winter, winds are more consistent averaging $3.6 \pm 0.1 \text{ m/s}$ with no diurnal pattern. Manual weather records collected at Lake Vanda during the

~1970 period (data mostly from 1969, 1970 and 1971) presented by Samson (1992) show only minor differences to more recent LTER records.

Based on the consistency of meteorological data, Lake Vanda's ablation rate has likely remained constant for at least the past two decades, and probably since at least 1969. Chinn (1981) measured an ablation rate of 0.32 m/yr between 1969 and 1970 which is consistent with predictions of a steady-state ice thickness model developed by McKay et al. (1985). This ablation rate is, however, significantly lower than values calculated for Lake Fryxell (0.76 m/yr), Lake Hoare (0.99 m/yr), and Lake Bonney (0.69 m/yr) (Dugan et al., 2013). In addition to slight differences in climate between the Wright and Taylor valleys, the difference in ablation rates is likely due to the smoother, relatively debris-free surface ice of Lake Vanda, which provides greater albedo, less surface melt around sediment deposits, and lower aerodynamic resistance (i.e. surface roughness) compared to lake ice in the Taylor Valley (Goldman et al., 1967; McKay et al., 1994; Dugan et al., 2013).

5.3 New Bathymetry and Water Storage Calculations

In order to quantify lake elevations and changes in lake storage over time, we developed a new bathymetric map of Lake Vanda using a GIS-based approach. Bathymetry was interpolated using a smoothed tension spline of points from drill hole surveys, aerial photography, and existing digital elevation models (DEMs). Drill hole points were taken from new data collected by the authors during the 2013 field season, located by commercial-grade GPS, as well as a georeferenced hand-drawn map from Nelson and Wilson (1972). Aerial photographs obtained when lake level was known (Figure 3) were georeferenced and their shorelines were converted into points with 20 m spacing. The elevation of each shoreline point was taken from the lake level as measured at the time that photography was undertaken. Depths were assigned to all points corresponding to the lake level relative to 2013 measurements. Overlapping shorelines were resolved based on the image with the lowest georeferenced total RMS error. Additional artificial points were added with depths interpolated manually from existing data in areas where the point spacing was sparse or where the interpolation model had trouble, which was mostly along the southern peninsula. The interpolated tension spline was

smoothed by applying a 5 cell radius zonal average and then averaging the results with the bathymetry output from a triangular irregular network of the points.

Figure 10 shows the new bathymetric model for Lake Vanda. Superimposed on the bathymetry data are the locations of (i) shorelines from the Holocene high-stand (3000 yr BP, Figure 5), (ii) the 1000 yr BP Holocene low-stand, (iii) the upper pycnocline, and (iv) the lake surface recorded on January 31, 2014 (surface elevation 94.7 m amsl). The new bathymetry suggests that at the high-stand Lake Vanda would have occupied a surface area of $15.6 \times 10^6 \text{ m}^2$ with a volume of $730 \times 10^6 \text{ m}^3$, and the low stand an estimated surface area of $0.3 \times 10^6 \text{ m}^2$ with a volume of $0.25 \times 10^6 \text{ m}^3$.

We estimated the 1947 lake level as 89.7 m amsl by comparing aerial imagery to the new bathymetric model. These results confirm that lake level has increased by 15 m in 68 years at an average rate of 22 cm/year (Figure 3). This agrees with previous average rates of change based on shorter time periods (Hawes et al., 2013) and is the largest change in lake levels reported among MDV lakes over that period (Bomblies et al., 2001; Doran et al., 2002; Lyons et al., 2006). The most dramatic increases occurred during 1971-1972, 1985-1992, 1998-2001, and 2008-2011. Most of these periods experienced higher Onyx River discharge, in particular, 1971, 1986, and 2001 (Figure 8).

5.4 New Water Balance Model

We used the new bathymetry to perform a water balance, which calculated the discharge of the Onyx River over time required to allow a given lake level to be sustained. In this model, the volume lost to ablation was equal to the ablation rate multiplied by the surface area of the lake. The model inherently assumed a constant ablation rate. As discussed above, meteorological variables affecting ablation appear to have been largely constant over the 1970 to 2014 period, though we have no method of confirming constant ablation rates outside of that time period. We calculated balances of Lake Vanda at four different lake levels: the Holocene high-stand, the Holocene low-stand, the lake surface immediately prior to the formation of the upper convection in approximately 1930, and the lake surface in 2014. Assuming an approximate hydrologic equilibrium occurred at each level, and an ablation rate equal to present values, we calculated the required discharge of the Onyx River to maintain each lake level. Finally, by

holding the 1970-2010 average Onyx River discharge and ablation rates constant, we calculated the volume and elevation of the future Lake Vanda at hydrologic equilibrium, and the time required to attain this lake level (Table 1). Of these, our assumptions are least likely correct for the low stand, which was hypersaline and probably ice-free for at least part of the year, but it is reasonable to assume that inflow was small at that time.

Table 1 summarizes the equilibrium water balance calculations for Lake Vanda for each scenario, and shows that the annual Onyx River input has fluctuated over two orders of magnitude over the past 3000 years. The present discharge rate is nearly four times larger than the discharge rate required for the equilibrium level at the base of the upper convection cell in approximately 1930, providing further evidence of recent climate change. However, current discharge is still roughly 20% lower than discharge rates responsible for the Holocene high-stand. Extrapolating current flow and ablation rates into the future, a new equilibrium lake level would occur in another 200 years, or roughly 2200, when current inputs will match annual loss to ablation at an elevation of approximately 115 m amsl, or 20 m above the 2014 lake level.

These data also provide the opportunity for an independent calculation of the start of the filling of the upper convection cell. From Table 1, we see that volume of Lake Vanda more than doubled as a function of the filling of the upper convection cell, from $9.77 \times 10^7 \text{ m}^3$ (lake volume below the upper pycnocline) to $2.10 \times 10^8 \text{ m}^3$ (lake volume in January 2014). This equates to a net increase of $1.12 \times 10^8 \text{ m}^3$. Assuming that the long term input from the Onyx River has been constant at $4.07 \times 10^6 \text{ m}^3/\text{yr}$, and that the long term output due to ablation has been constant at $2.60 \times 10^6 \text{ m}^3/\text{yr}$, then the volume of Lake Vanda has increased at an average rate of $1.47 \times 10^6 \text{ m}^3/\text{yr}$. By dividing the net volume increase by the rate of annual increase, we calculated a filling time of 76 years, which is consistent with age of the start of filling estimates from benthic vegetation (Hawes et al., 2013). This is an underestimation as initial rates of annual ablation loss would initially have been smaller due to the smaller surface area of the lake (Figure 4; Table 1).

Table 1. Equilibrium water balance calculations for past, present and future lake levels assuming a constant ablation rate with time. Measured values shown in normal text, whereas estimated and derived values shown in italics.

	3000 yr BP Steady State (High-stand)	1000 Yr BP Steady State (Low-stand)	80 Yr BP Steady State (~1930)	Current Conditions (2014)	Future Steady State (2200)
Lake elevation (m amsl)	148	22.6	74.3	94.6	~125
Surface area (m ²)	1.56×10^7	4.89×10^4	3.58×10^6	7.45×10^6	1.16×10^7
Volume (m ³)	7.33×10^8	9.63×10^4	9.77×10^7	2.10×10^8	4.41×10^8
Net volume change (m ³)	-	-7.33×10^8	$+9.76 \times 10^7$	$+1.12 \times 10^8$	$+2.31 \times 10^8$
Onyx River input (m ³ yr ⁻¹)	5.7×10^6	1.7×10^4	1.3×10^6	4.07×10^6	4.07×10^6
Ablation rate (m yr ⁻¹)	<i>0.35</i>	<i>0.35</i>	<i>0.35</i>	0.35	<i>0.35</i>
Ablation output (m ³ yr ⁻¹)	5.7×10^6	1.7×10^4	1.3×10^6	2.6×10^6	4.07×10^6

Figure 10. Bathymetry model for Lake Vanda. Colors reflect the depth in 10 m intervals from the January 2014 surface elevation of 94.7 m amsl. Also shown are paleo-shoreline levels from the 3000 yr BP high-stand, the 1000 yr BP low-stand, and the 80 yr BP lake level when the shoreline was at the base of the upper convection cell.

6. Change to the Lake Vanda Ecosystem

Lake Vanda supports a diverse biota, which, as with all MDV lakes, is primarily microbial. There are no macroinvertebrates, molluscs or fish (Goldman et al., 1967) and food chains tend to be short, with few predators and rotifers being the largest planktonic animals (James et al., 1998). It is convenient to separate the biota into a planktonic community that is distributed through all components of the water column (Vincent and Vincent, 1982) and benthic mats, which cover the lake floor to at least 45 m depth (Hawes and Schwarz, 2001). Both communities are profoundly influenced by the unique physical setting that the lake provides, and are vulnerable to change as lake level waxes and wanes.

6.1 Planktonic Communities

Early investigators of the lake identified the extraordinary oligotrophic nature and clarity of the upper part of Lake Vanda, the presence of planktonic phototrophs (flagellated algae and cyanobacteria), protists and rotifers to at least 60 m depth, and the absence of higher organisms including macroscopic zooplankton (Goldman et al., 1967). Later research confirmed this picture of a vertically stratified lake, with distinct floristic communities in the various hydraulic components of the lake and a phytoplankton biomass and activity maximum embedded in the deep diffusion zone. In 1980-81, close-interval sampling showed that this deep phytoplankton increase was a sharply delineated, deep chlorophyll maximum (DCM), just a few meters thick, formed primarily of filamentous cyanobacteria (*Leptolyngya*). The DCM was at 57.5 m depth in 1980-81, and was shown to coincide with a photosynthetic maximum and a corresponding localized peak in photosynthesis-derived dissolved oxygen (Vincent and Vincent, 1982).

The DCM has since proved to be a persistent feature of the lake (Burnett et al., 2006). In addition to an oxygen and chlorophyll peak, this zone hosts a local protozoan community (James et al., 1998), and is perched immediately above a steep oxycline where dissolved oxygen concentration declines to anoxia (Figure 11). This oxycline is accompanied by a rapid increase in the concentration of reduced nitrogen (ammonium) and dissolved reactive phosphorus (Vincent et al., 1981). The DCM is located at the depth to which sufficient light penetrates to allow phytoplankton growth, and where phosphate, the growth-limiting nutrient in Lake Vanda (Vincent and Vincent, 1982), diffuses vertically across the oxycline (Priscu, 1995). It follows

that the long term balance between downwards diffusion of oxygen from the DCM and oxygen consumption in deeper waters ultimately sets the depth of the oxycline (Vincent and Vincent, 1982; Burnett et al., 2006).

Figure 11. Profiles of temperature, pH, dissolved oxygen and chlorophyll fluorescence in Lake Vanda in December 2014. Data were obtained using a Yellow Springs Instruments 6600 Sonde lowered through the water column. DCM depth = 65 m; oxycline and nutricline depth = 67 m.

A suite of other microbial processes occur stratified within the oxycline-DCM complex. These include the nitrogen cycle, where the processes of ammonium oxidation, nitrification and denitrification straddle the oxycline (Vincent et al., 1981; Priscu 1997; Voytek et al., 1999), and sulfate reduction/sulfur oxidation (Purdy et al., 2001). The oxycline defines the location of not just microbial redox-related processes, but also geochemical processes that control the movement of metals and other redox-sensitive phases (Green et al., 1989; 1993; Webster 1994).

Assuming that the depth of the DCM and the oxycline are related to irradiance, the DCM may serve as a useful proxy for the stability of microbial stratification over time as a function of lake filling. As the lake has deepened over recent decades (Figure 7), it is inevitable that the quantity of irradiance reaching deep into the lake has declined, and we can hypothesize that the DCM should have gradually moved upwards through the water column. The magnitude of decline can be estimated from the attenuation rate for downwelling irradiance in Lake Vanda (0.05/m, Howard-Williams et al., 1998). This parameter indicates the irradiance at 57.5 m is reduced by 40% when 10 m of water is added to the lake.

The DCM is one of the features of the lake that can be traced over time, using available profiles of chlorophyll-a distribution. While no other early profiles reproduce the detail of Vincent and Vincent (1982), we examined other chlorophyll profiles that were obtained with a depth resolution of 2 m or less at the DCM to identify any change in its distance from the lake bottom. Since 1993, the depth of the DCM has tended to fluctuate around 63 m from the surface which, given ongoing lake level rise, suggests a gradual movement of the DCM away from the lake floor (Table 2). However, the apparent upward movement (4.5 m since 1980) is consistent with the depth of the DCM being affected by light, though the speed of rise has lagged the increase of the lake level. Such a slow change perhaps reflects a gradual, biologically-mediated response coupled to both available light and the upwards diffusion of nutrients.

Table 2. A synthesis of chlorophyll-a profiles with sufficient vertical resolution to be confident in describing the location of the deep chlorophyll maximum. Published depths have been converted to distance from the lake floor using the new bathymetric model. “Source” refers to the data source and the vertical resolution of sampling. “Extract” indicates water sample returned to surface, filtered and extracted in acetone prior to estimation of chlorophyll-a by fluorometry.

Year	Recorded DCM depth (m)	Distance from bottom (m)	Source (method, resolution)
1980	55.5	9.5	Vincent and Vincent 1982 (extract, 2.5 m))
1993	63.0	11.0	Authors’ unpublished (extract, 1 m))
1996	62.0	12.2	Authors’ unpublished (extract, 2 m and figure X)
1999	61.7	12.9	Authors’ unpublished (extract, 2m)
2004	62.0	13.3	LTER (http://www.mcmlter.org/) (extract 2 m)
2007	64.0	11.7	LTER (http://www.mcmlter.org/) (extract 2 m)
2014	63.6	14.0	Authors’ unpublished (Figure 15)

6.2 Benthic Communities

While less well documented than plankton communities, benthic communities probably dominate biological productivity in Lake Vanda (Hawes et al., 2001). In the euphotic, oxic part of the lake (i.e. the upper and lower convection cells and the upper part of the diffusion gradient), they take the form of microbial mats dominated by cyanobacteria and diatoms (Wharton, 1983; Hawes and Schwarz, 2001; Hawes et al., 2013). In the lower convection cell, a single moss species, *Bryum pseudotriquetum*, is also present, is not found in the upper convection cell (Kaspar et al., 1982). Preserved benthic mats are also found associated with the various paleo-shorelines perched above Lake Vanda (Figures 3a) where they are freeze-dried and buried under wind-blown sand (Figure 3b). This evidence shows that the mats have been a regular feature of Lake Vanda through filling and drying cycles. Recently, we have attempted to extract DNA

from these relict mats and preliminary results show that short sequences are consistent with species of *Leptolyngbya* and *Phormidium* currently found in the lake. The under-ice flora of Lake Vanda is dominated by filamentous cyanobacteria, mostly attributable to the genus *Leptolyngbya* (Zhang et al., 2015). Mats of these cyanobacteria have evolved a complex macroscopic morphology that takes the form of elaborate pinnacle structures (Figure 13).

Microbial mats can be expected to respond to lake level change at their upper and lower depth limits. At the upper limit, rising lake levels flood new areas of shoreline providing new littoral zone area for mat growth, while falling levels expose previously colonized regions and cause them to desiccate. Hawes et al. (2013) showed that mats appear to be capable of keeping pace with rising lake levels and their composition has not changed over time, and as discussed above, used features of mat morphology and stratification to estimate an age of the upper convection cell of approximately 80 years prior to their publication. Conversely, at the lower limit, it is expected that declining irradiance caused by lake level rise will impact on the photosynthetic capability of the algal mat, and lead to a shift in the maximum depth of colonization. The lower littoral area colonized by microbial mats show collapsing pinnacle structure (Figure 12, 55 m). This occurred at a depth of 55 m in 2013. In the absence of historical records from this depth it is impossible to be certain that this is the case. However, if the lower limit is set by light irradiance, then a gradual rise in lake level and associated reduction in irradiance would require the death of deepest microbial mats which is reflected by the decayed morphology of pinnacles.

Figure 12. The appearance of pinnacle mats from 17 m, 33 m, 41 m, and 55 m depth in Lake Vanda. The physical depth and optical depth (as % incident irradiance at noon) are indicated. Where visible, the red laser dots are 3 cm apart. Images obtained in November 2013 by Tyler Mackey.

7. Regional Climate Change over the past 80 years

Evidence cited here for a climate change at Lake Vanda beginning ~80 yr BP, or the early 1930's, includes the growth rate of algal mats in newly submerged littoral zone, and the time required to flood the upper convection cell to its present surface elevation. This change would have been large and rapid, given the four-fold increase in the flow of the Onyx River which is implied (Table 1). Could this change have also been part of a regional climate shift impacting the MDV as a whole?

In the 20th Century, all lakes in the MDV generally increased in elevation (Chinn, 1993; Doran et al., 2008). For Lake Bonney in the Taylor Valley, a 12 m lake level rise was documented between 1903 and 1970 (Chinn, 1993) based on the first lake level measurement by

Sir Robert Falcon Scott (Scott, 1905). Bombardieri et al. (2001) calculated an annual increase in inflows to Lake Bonney of 3% from 1903 and 1973 to account for this change. Lake Bonney experienced lake level drawdown between late 1980 and 1999, which was associated with air temperature cooling at 0.7°C per decade (Doran et al., 2002a). Following this cold period, lakes Vanda, Vida, Bonney, Hoare and Fryxell showed a rapid increase of 0.54 to 1.01 m during the anomalously warm summer of 2001-02 as a product of record stream discharge (Doran et al., 2008). This event changed the physical, chemical, and biological environment of the lakes by thinning perennial ice covers, increasing nutrient levels, and boosting primary productivity in most lakes (Foreman et al., 2004). As discussed above, Doran et al. (2008) attributed the increase in valley bottom temperatures to an increase in the frequency of föhn winds, and such wind are also associated with low cloud cover.

Recently, Hoffman et al., (2016) showed sensitivity of glacier melt to changes in surface albedo based on spatial and temporal model simulations. Albedo may change though increases in wind strength, which are responsible for aeolian sediment transport. This could either imply a pulse event or a gradual shift in climate, and point to a synergistic effect of increased katabatic conditions though increases in temperature, reduction in cloud cover and increases in dust deposition.

Worldwide, surface temperatures on Earth rose 0.3 to 0.6 °C during the past century (Cane et al., 1997). In Lower Victoria Land, evidence in support of an abrupt change in the last 100 years includes high temporal resolution records of changes in sea ice extent in the MDV region that have been inferred from ionic signatures in glacier ice cores (Sinclair et al., 2014). They record stable sea ice conditions up to the 1950's, after which sea ice extent first declined and then increased. Similarly, Curran et al. (2003) concluded that the total sea ice area around Antarctica has been declining since 1950's, based on methanesulphonic acid found in the Law Dome ice core record as a proxy for sea ice extent. Additionally, radiocarbon signature spanning the last 400 years from corals collected from the northern Ross Sea, shows an abrupt change ~80 yrs BP (Allinger unpublished– cited in Frank et al., 2014).

To date the fastest climate change in Antarctica is seen on the Antarctic Peninsula (Turner et al., 2005; Chapman and Walsh, 2007). The timing of a rapid increase in surface air temperature over the past 50 to 100 years in the Antarctic Peninsula (Jacobs and Comiso, 1997)

coincides with Lake Vanda lake level changes and a teleconnection between both locations may be sought. Warming temperatures on the Antarctic Peninsula are often associated with large scale changes in atmospheric circulation. In particular, the dynamics of the Amundsen Sea Low (ASL) is responsible for drawing warm air from the north (Jacobs and Comiso, 1997) due to the state of the phase of the Southern Hemisphere Annular Moment (SAM), which is currently exacerbated by the effects of the Ozone hole (Turner et al., 2005). Bertler et al (2004) argued that the linkage between phases of SAM, changes in the Western Antarctic Peninsula, and the McMurdo Sound Region, are driven by the circulation pattern around the ASL. Between 1957 and 2004, a negative phase of SAM was found to be correlated with higher surface air temperatures in Eastern Antarctica (Marshall, 2007). As a response to the SAM phase, a significant climatic change in the MDV occurred (Spiers et al., 2013). Specifically, this has resulted in an increase in the frequency of föhn winds flowing down valleys toward the sea. Such shifts are amongst those shown to correlate with increased flow in the McMurdo Dry Valley streams, including the Onyx River (Fountain et al., 2014). Correlations between a trend of increasing sea ice extent in the Ross Sea region during the second half of the 20th Century, the depth of the ASL (Turner et al., 2015), and increased northerly airflow over the western Ross Sea region (Coggins and McDonald, 2015) supports this overarching and persistent recent impact on the climate of the Ross Sea region.

8. Prediction for the future Lake Vanda

In their review on the impacts of climate change on Antarctic lakes, Lyons et al. (2006) postulated that increased summer warming will impact lakes as follows: (1) extend melting seasons, increase lake levels, and thin, and eventually remove, perennial ice cover; (2) increase nutrient loading and atmospheric gas exchange; (3) promote primary production; (4) strengthen stratification through the addition of fresh water above saline water, while enhancing wind-driven mixing of shallow layers; (5) increase the organic carbon flux to deep lake layers driving anoxia and modifications to biogeochemistry; and (6) cause an overall shift in biologic structure and biodiversity. Some of these changes have already been observed in Lake Vanda over the past 80 yr, and foreshadow additional changes resulting from continued climate change.

Assuming present discharge and ablation rates remain constant, lake levels will continue to rise over the next 200 years (Table 1). A new equilibrium lake level will form approximately 30 m above the 2014 level, or 124 m amsl, where net ablation loss will equal inflow. Should climate change increase present rates of meltwater production or decrease ablation rates, perhaps caused by an increase in present föhn wind frequency, there would be an acceleration in lake level rise leading to a higher equilibrium elevation.

We anticipate that increased lake levels will continue to cool temperatures in the lower convection cell and the lake bottom at a rate similar to the trend lines shown in Figure 7. Lake Vanda is not alone in this cooling trend among Antarctic lakes; Webster et al. (1996) reported that the 25 m increase in Lake Wilson between 1975 and 1993 caused slight decreases in the temperature both the surface and the deepest saline layers. Should the cooling trends shown in Figure 7 continue at an equal pace, the lower convection cell will have approximately the same temperature as the upper convection cell (4.5 °C) in approximately 50 years, or by 2065, and the entire water column will be isothermal at 4.5 °C in approximately 245 years, or roughly by 2260. This could ultimately lead to an isothermal water column. Given the dependency of thermohaline convection on a strong vertical temperature gradient, it is likely that thermohaline convection will halt at some point, and diffusion-dominated salt transport will prevail. The end result will be a salinity stratified lake where vertical changes in water chemistry are largely driven by molecular diffusion and downward sinking brines during ice formation (Miller and Aiken, 1996; Dowling et al., 2013), much like Lakes Fryxell, Bonney, and Joyce (e.g. Spigel and Priscu, 1998). Termination of convection would also impact the internal distribution of nutrients, such as nitrogen and carbon added from the Onyx River and presently homogenized throughout the upper convection cell, and phosphorous diffusing upwards across the chemocline and presently homogenized within the lower convection cell.

The effect of lake level rise on ice thickness is less clear. Should cooling begin to decrease temperature in the upper convection cell as a function of lake level rise, the thermal gradient below the ice-water interface may decrease and ice thickness could hypothetically increase. Alternatively, if the upper convection cell is the primary recipient of heat from lower lake layers, it will change its temperature more slowly than underlying layers as shown in Figure

7, perpetuating a thin ice cover. That said, both of these affect could be overwhelmed by warm air temperatures which can rapidly thin lake ice (Doran et al., 2008).

We expect the DCM will continue to rise in the water column as a function of lake level rise and light penetration. At present, the oxycline (67 m depth) and the nutricline are located deep within the diffusion zone (Figures 3 and 11), and the flux of recycled nutrients into shallower waters is limited by the rate of molecular diffusion and the consumption of nutrients at the DCM. Insofar as the location of these features are set by irradiance, if rising lake level moves these biogeochemical boundaries closer to the base of the lower convection cell, the potential flux of nutrients and hence phytoplankton growth through that layer is increased, which in turn has implications for absorption of radiation and temperature structure.

In the shallow regions of the lake, freshly inundated soils will be colonized by benthic mats whereas deep layers will become uninhabitable owing to lack of sunlight. The upper pycnocline will continue to act as colonization barrier separating organisms within these cells. This constrains the ability of organisms within to the lower convection cell to migrate into shallower water with more available light which could potential increase mortality rates as lake level rises. Finally, the impact on biological productivity will also depend on the endurance of thermohaline convection cells, as these cells presently homogenize nutrients. The removal of circulation could limit nitrogen distribution from the Onyx River to the upper few meters of the water column, and phosphorous distribution to the waters just above the chemocline.

Perhaps the singular, most significant potential change to Lake Vanda would be the complete loss of ice cover. Because this has not happened in the MDV for over 1000 years, we may refer to Arctic lakes on Northern Ellesmere Island, Canada for an indication of what to expect. Here, several, large, perennially ice-covered lakes lost their ice-cover within the past decade (Mueller et al., 2009; Veillette et al., 2010). In one case, Ward Hut Lake had 3.5 to 4.3 m thick ice between 1953 and 2007, and became ice free in 2011 (Paquette, et al., 2015). This example illustrates the rapidity with which perennial lake ice can be removed under favorable conditions. For Lake Vanda, such an event would subject the upper convection cell to strong winds for the first time in 1000 years, and potentially drive complete circulation between the upper and lower convection cells. Given the salinity of water below the chemocline (Figure 3), it

is likely that the lake would remain meromictic, with the upper-two thirds freely circulating each summer, and stratification returning during winter ice cover.

9. Conclusions

Using an extensive review of previously published data combined with new data, this paper has shown that multiple, significant climate-driven changes have occurred to Lake Vanda over the past 80 years, since the early 1930's. We hypothesize that following sequence of events occurred over this period:

1. Global climate change causes a deepening of Rose Sea Low.
2. An increased frequency of föhn winds between the polar plateau and the Ross Sea, indicative of regional climate change, brings warmth to the MDV.
3. Increased meltwater production.
4. A four-fold increase in the annual flow rate of the Onyx River while net ablation remained constant.
5. Lake level rise and development of the upper convection cell.
6. Cooling of water temperatures in the lower convection cell and lake bottom.
7. Thinning of lake ice due to a change in the shallow temperature profile.
8. Vertical rise of the DCM.
9. Colonization of newly submerged area by mats, coupled with increased mortality of at depth.

These examples showcase Lake Vanda as an exceptional natural laboratory, or sentinel, for the study of regional climate change. We predict Lake Vanda will continue to evolve in response to climate change in the near to distant future.

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