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Stratigraphy of Lake Vida, Antarctica: hydrologic implications of 27 m of ice

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Abstract. Lake Vida, located in Victoria Valley, is one of the largest lakes in the McMurdo dry valleys and is known to contain hypersaline liquid brine sealed below 16 m of freshwater ice. For the first time, Lake Vida was drilled to a depth of 27 m. Below 21 m the ice is marked by well-sorted sand layers up to 20 cm thick within a matrix of salty ice. From ice chemistry, isotopic composition of $\delta^{18}O$ and $\delta^{2}H$, and ground penetrating radar profiles, we conclude that the entire 27 m of ice formed from surface runoff and the sediment layers represent the accumulation of surface deposits. Radiocarbon and optically stimulated luminescence dating limit the maximum age of the lower ice to 6300 ^{14}C yr BP. As the ice cover ablated downwards during periods of low surface inflow, progressive accumulation of sediment layers insulated and preserved the ice and brine beneath, analogous to the processes that preserve shallow ground ice. The repetition of these sediment layers reveals hydrologic variability in Victoria Valley during the mid- to late Holocene. Lake Vida is an exemplar site for understanding the preservation of subsurface brine, ice, and sediment in a cold desert environment.

1 Introduction

Little is known about the habitability of cold liquid environments sealed off from the atmosphere, be it the subglacial lakes of Antarctica (Siegert et al., 2013) or beneath the icy shell of Europa (McKay, 2011). Located in the McMurdo dry valleys of Antarctica, Lake Vida has the thickest ice of any subaerial lake on Earth and is one of the few "ice-sealed" ecosystems known to support a diverse and active microbial population in a cold, anoxic, aphotic brine (Murray et al., 2012; Dugan et al., 2015). The brine was first discovered 16 m below the surface of Lake Vida and is hypothesized to have been sealed from the atmosphere for several millennia (Doran et al., 2003). At present, the lake level (i.e., the ice surface) of Lake Vida is rising, which implies that the brine is progressively getting farther from the surface. In this paper, we present evidence for how and when Lake Vida formed to further understand the structure and evolution of the existing brine system beneath Lake Vida.

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

In Victoria Valley, it is theorized that the Lake Vida basin was occupied by a 200 m deep glacial lake 8600 ^{14}C yr BP (Hall et al., 2002), after which lake levels began to decline. This lake is inferred to have had a water column (likely with an ice cover) that permitted sedimentation and led to the lacustrine deposits seen on the landscape today. A deep water column would preclude the presence of bottom ice, and therefore it is improbable that any of the observed ice existed during this time. Therefore, the 27 m of ice currently on Lake Vida was formed subsequent to 8600 yr BP.

During mid- to late Holocene, it is possible that Lake Vida was influenced by events similar to the repeated lake level drawdowns and complete desiccation events recorded in lacustrine sediment cores and geochemical diffusion profiles of the large lakes of Taylor and Wright valleys. For instance, it is speculated that Taylor Valley underwent a valley-wide desiccation event at 1000–1200 yr BP (Lyons et al., 1998), and Lake Fryxell had lowstands at 6400, 4700, 3800, and around 1600 yr BP (Wagner et al., 2006; Whittaker et al., 2008). Lake Vanda, in Wright Valley, underwent a lowstand at 1200 yr BP (Wilson, 1964), or prior to 2000 yr BP (Gumbley et al., 1974). In these closed basin lakes, desiccation is thought to be the result of climatic changes and not a result of large drainage events. If Lake Vida was influenced by similar climatic patterns, some or all of these events may be recorded in the ice cover.

The aim of this study is to reconstruct the history of the ice cover on Lake Vida. We examine the isotopic and ion geochemistry, sediment characteristics, and diatom composition of a 27 m ice core, as well as ground penetrating radar (GPR) profiles, to extrapolate the strata in single cores. Both radiocarbon dating and optically stimulated luminescence (OSL) dating are employed to establish the time of deposition of sediment layers, and ice core stratigraphy is used as a means of establishing periods of lake level drawdown.

2 Study site

Lake Vida (77°23′ S, 161°56′ E), situated in Victoria Valley, Antarctica, is one of the largest (6.8 km^2) and highest (350 m a.s.l. – above sea level) lakes in the McMurdo dry valleys (Fig. 1). The lake is endorheic (closed basin) and receives inflow via streams originating from the Victoria Upper, Victoria Lower, and Clark glaciers. Annual precipitation is < 35 mm (Fountain et al., 2010). Lake Vida occupies a unique climatological niche where summer temperatures can rise slightly above 0 °C to generate stream flow, yet unusually cold winters (compared to the other major valleys in the region) maintain a thick ice cover on the lake. From 1995 to 2000, the mean annual air temperature at Lake Vida (−27.4 °C) was 7 to 10 °C lower than valley bottom temperatures in Taylor and Wright Valleys, but mean summer temperatures were similar (Doran et al., 2002).

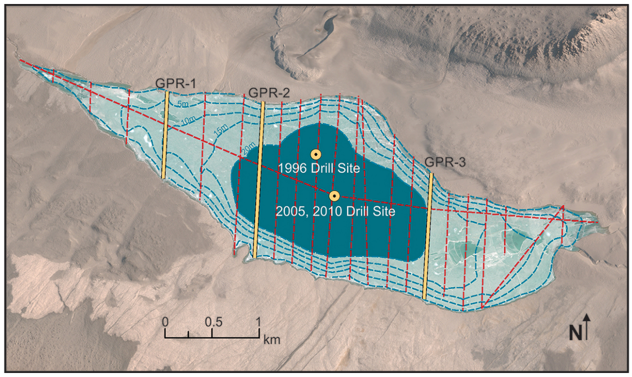

Figure 1. Location of drill sites and GPR transects (red dashed line) on Lake Vida in central Victoria Valley ©DigitalGlobe (2011). Highlighted GPR transects 1–3 are presented in the paper. Bathymetric lines 0–20 m were digitized and interpolated from GPR profiles. The dark blue area below 20 m is of unknown depth.

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

3 Methods

An electric 15 cm diameter SideWinder drill (Kyne and McConnell, 2007) was used to retrieve a 27 m and a 20 m long ice core located 6 m apart in the center of Lake Vida in November 2010. The 27 m ice core was split in half for archival purposes and subsampled into 5 cm lengths. Where recovery was incomplete for the 27 m core (between 16 and 20 m), the 20 m ice core was subsampled. Longitudinal thick-sections (∼ 0.5 cm thick) were cut from the ice core face and viewed under cross-polarized light for ice crystal fabric analyses. Subsamples were washed with deionized MilliQ water to remove possible brine contamination and allowed to completely melt for processing on a Dionex 1500 ion chromatograph for major ion analysis and a Los Gatos Research

Table 1. Summary of analyses and analytical errors.

Analysis	Material	Analytical error
TC and TIC	Ice	±0.16 %
Anions	Ice	±2.8 µm (SO_4^{2-}), ±0.7 µm (Cl^-)
Cations	Ice	±1.7 µm (Na^+), ±0.4 µm (K^+)
		±1.0 µm (Mg^{2+}), ±1.0 mum (Ca^{2+})
δ^2H	Ice/sediment pore water	±0.8 ‰
δ^2H	Brine	±2 ‰
$\delta^{18}O$	Ice/sediment pore water	±0.1 ‰
$\delta^{18}O$	Brine	±0.2 ‰
OSL	Bulk sediment	$\pm \leq$ 100 yr BP
^{14}C	Sediment (insoluble organic fraction)	$\pm \leq$ 58 yr BP
^{14}C	Sediment (carbonate)	$\pm \leq$ 49 yr BP

liquid water isotope analyzer for isotopic composition of ^2H and ^{18}O (analytical errors provided in Table 1). When necessary, samples were diluted to near seawater salinity prior to analysis. Isotopic values are reported with respect to the VSMOW international standard. Salinity is reported as the sum of concentration of total ions.

Sediment layers in the 27 m ice core were subsampled in duplicate 1 cm segments, which were freeze-dried or allowed to melt in order to extract pore water by centrifugation. For grain size analyses, 2 g samples were sieved through a 1000 µm sieve and pretreated with 30 % H_2O_2 for 18 h in a 50 °C water bath. Following pretreatment, samples were shaken following the addition of 1 mL of 30 mg L^{-1} Graham's salt ($Na_4P_2O_7$) as a dispersant and analyzed on a Micromeritics Saturn DigiSizer 5200 particle size analyzer (detection limit 0.1–1000 µm). Sand/silt classifications are based on the Udden–Wentworth scale (Wentworth, 1922). Sediment layers were also subsampled to evaluate diatom assemblages and absolute abundance (valves/g dry weight) via light microscopy. Preparation methods followed standard techniques (Warnock and Scherer, 2015). A known mass of freeze-dried sediment was reacted with 10 % HCl and 10 % H_2O_2 to remove carbonates and organics. Abundance per gram was extrapolated from diatom counts on coverslips in a beaker of known area.

Freeze-dried samples were analyzed for total carbon (TC) and total inorganic carbon (TIC) with an elemental analyzer (Dimatec Co.). Total organic carbon was calculated from the difference in TC and TIC. Six samples with the highest organic carbon content (0.4–2.1 %) were chosen for radiocarbon dating of the organic fraction. Samples were prepared by removing carbonates and humic acids by acid–alkali–acid extraction prior to graphite conversion in an automated graphitization system. This required an overnight treatment with 1 % HCl to remove carbonates, followed by 4-hour humic acid extraction with 1 % NaOH and a secondary overnight treatment of 1 % HCl to eliminate any CO_2 that may have been absorbed during the NaOH treatment (Grootes et al., 2004). Two samples at 23.90 and 26.43 m with 1.3–2.3 % TIC were selected for radiocarbon dating of carbonates and were treated with dilute sulphuric acid. Two 5 cm long sediment sections at 21.51 and 25.54 m were cut by band saw under red light for OSL dating (UIC Luminescence Dating Research Laboratory) and remained frozen until analysis. The advantage of frozen sediment for OSL dating is that the water content is expected to have remained constant since freezing took place (Demuro et al., 2008; Arnold and Roberts, 2011). OSL dating of quartz grains from sediment layers was performed using single aliquot regeneration protocols (Murray and Wintle, 2003).

To confirm the continuity of horizons and sediment layers noted in the ice cores, 55 km of GPR transects was recorded over the surface of Lake Vida in 2010 (Fig. 1) using a GSSI SIR-3000 acquisition unit equipped with a 400 MHz antenna. Transects were recorded at 400 ns time range and 2048 16-bit samples per trace, with five manual gain points at −20, 0, 25, 30, and 50. A dielectric constant of 3.15 was initially chosen for depth calibration, but this was altered based on known characteristics of the ice cores. In post-processing, radar profiles were triple stacked and passed through a 200 and 500 MHz triangle FIR filter to remove high- and low-frequency noise. Lake levels were annually surveyed from benchmarks tied into historical optical survey transects conducted by New Zealand Antarctic Program and were recorded in m a.s.l.

4 Results

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

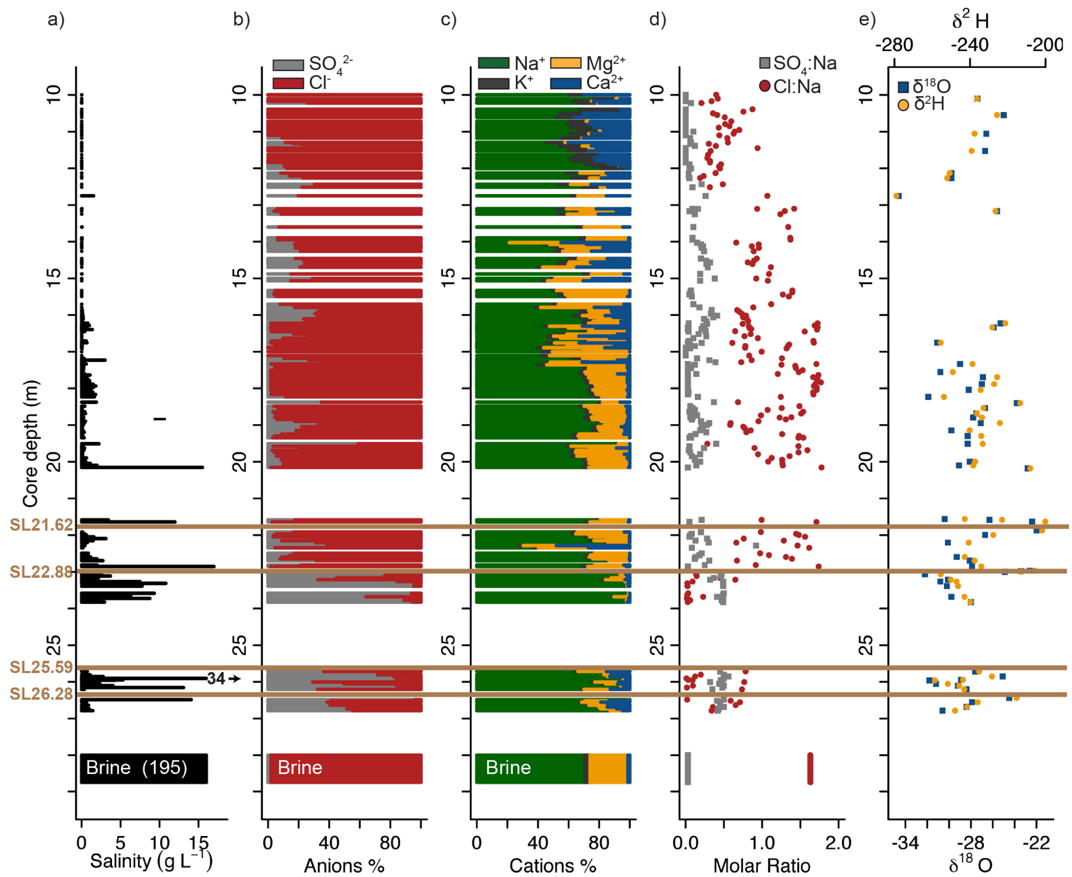

Figure 2. (a) Salinity of ice samples below 10 m subsampled from the Lake Vida ice cores. Salinities ranged from < 1 to $34\,\mathrm{g\,L^{-1}}$. Brine salinity is $195\,\mathrm{g\,L^{-1}}$. SL21.62, SL22.88, SL25.59, and SL26.28 represent sediment layers $>$ 10 cm thick present in the 27 m ice core. **(b, c)** Total percentage of major anions and cations. Ion percentages in the brine are noted at the base of the figure. **(c)** Ratio of $Na^{+} : Cl^{-}$ and $Na^{+} : SO_4^{2-}$ (mol : mol). **(d)** Stable isotope composition (δ^2H and δ^{18}O) of ice samples and sediment pore water.

There was an overall trend of increasing salinity in the ice core with depth as well as a shift at 23 m from Cl^{-} to SO_4^{2-} as the dominant anion and an accompanying decrease in Mg^{2+} (Fig. 2). Below 21 m, the ice salinity was variable from < 1 to $34\,\mathrm{g\,L^{-1}}$. During extraction, all cores below 16 m depth were in contact with the brine. If samples were contaminated by brine in the drill hole, we would expect the percentage of major anions/cations and the ratio of $Na^{+} : Cl^{-}$ and $Na^{+} : SO_4^{2-}$ in the ice to be similar to the brine (Fig. 2). As the ice chemistry is distinct from the brine, we conclude the ice was not substantially contaminated by brine in the drill hole. The fabric of the ice also changed with depth, from large individual ice crystals with c-axes oriented upwards, which is typical of freshwater ice, to ice composed of randomly oriented small crystals that appear to have recrystallized over time (Fig. 3).

Throughout the cores, there are many small pockets of sediments and thin sediment layers (Fig. 4). All sediments sampled from the ice core were predominately sand (grain size: 62.5 to 2000 µm), with only 4 out of 79 samples having a mean grain size $<$ 100 µm or a percentage of silt and clay ($<$ 62.5 µm) $>$ 6 % of the total volume. In the 27 m core, the four thick sediment layers below 21 m will be referred to by their depth in the core: SL21.62, SL22.88, SL25.59, and SL26.28 (Fig. 3). These layers had water contents $<$ 10 % and total thicknesses of 19, 15, 11, and 19 cm, respectively. At the base of SL26.28, mean grain size began to decrease with a concomitant increase in TC (Fig. 5). Microscopy of the lower sediment layers and ice revealed abundant diatom frustules of *Luticola gaussi* and the genera *Pinnularia*, specifically *P. deltaica* and *P. quaternaria*, only in SL26.28. These species are commonplace in the sediments and water column of Taylor Valley lakes, streams, and cryoconite holes (Stanish et al., 2012, 2013). *L. gaussi* is considered a freshwater species and cosmopolitan to the Antarctic diatom flora (Kopalová et al., 2013), as are *P. deltaica* and *P. quaternaria*. However, no studies have been done to specifically test the salinity tolerance of these species.

The upper sediment layers noted in two ice cores retrieved in 2010 correspond to those cored in 1996 (Doran et al., 2003) and 2005 (Taylor, 2009) (Fig. 4). The radiocarbon

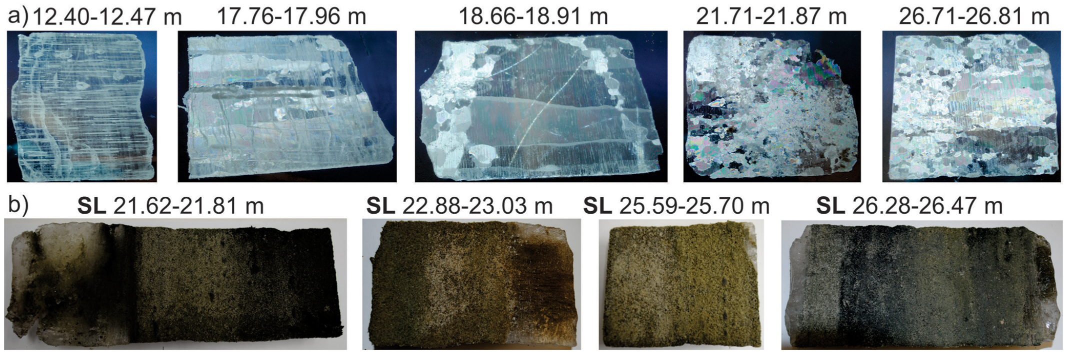

Figure 3. (a) Images of thick sections of Lake Vida ice between two sheets of polarized film. At 12.40–12.47 m, the entire section is a single ice crystal. With depth the average grain size of ice crystals decreases to < 1 cm. **(b)** Photographs of sediment sections SL21.62, SL22.88, SL25.59, and SL26.28.

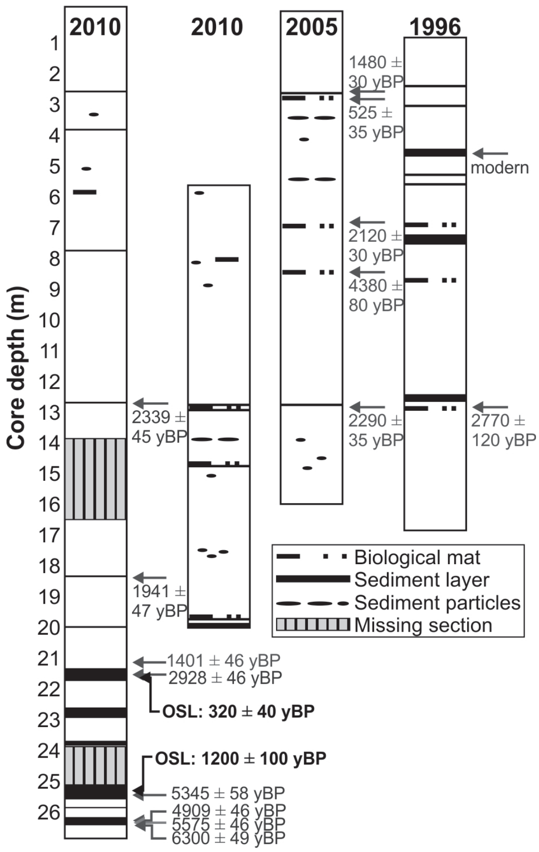

Figure 4. Lake Vida ice core logs from 1996, 2005 and 2010. Two ice cores were drilled in 2010. Where samples were missing in the 27 m core (between 16 and 20 m), the 20 m ice core was subsampled. For core locations in 1996 and 2005, see Doran et al. (2003) and Murray et al. (2012). Surface heights are adjusted to the 2010 lake elevation.

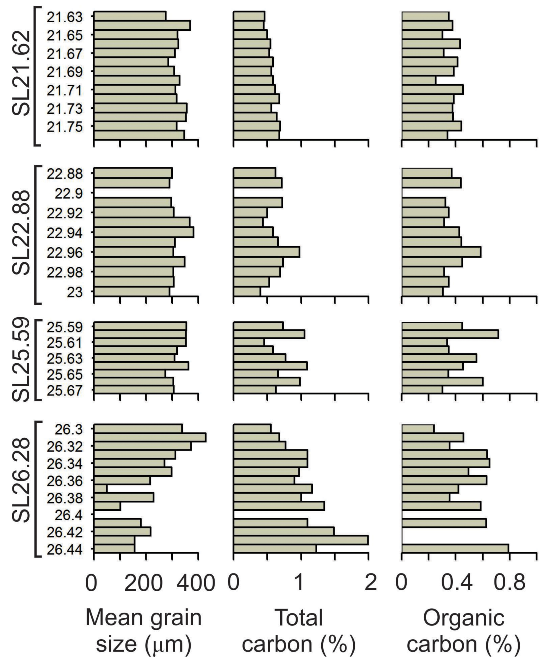

Figure 5. Mean grain size (μm), percentage of total carbon (%), and percentage of organic carbon (%) in sediment sections removed from the 27 m ice core from Lake Vida.

dates vary significantly, especially in the upper ice where there is no correlation between age and depth. The dates from SL26.28 do show increasing age with depth (4909 ± 46 to 6300 ± 49 ^{14}C yr BP). It is noted that δ^{13}C values are unreliable and not reported due to high fractionation during accelerator mass spectrometry measurements.

OSL samples showed no evidence of exposure to sunlight during collection and returned dates of 320 ± 40 and 1200 ± 100 yr BP for SL21.62 and SL25.59, respectively, dates which are younger than the radiocarbon ages from the respective horizons.

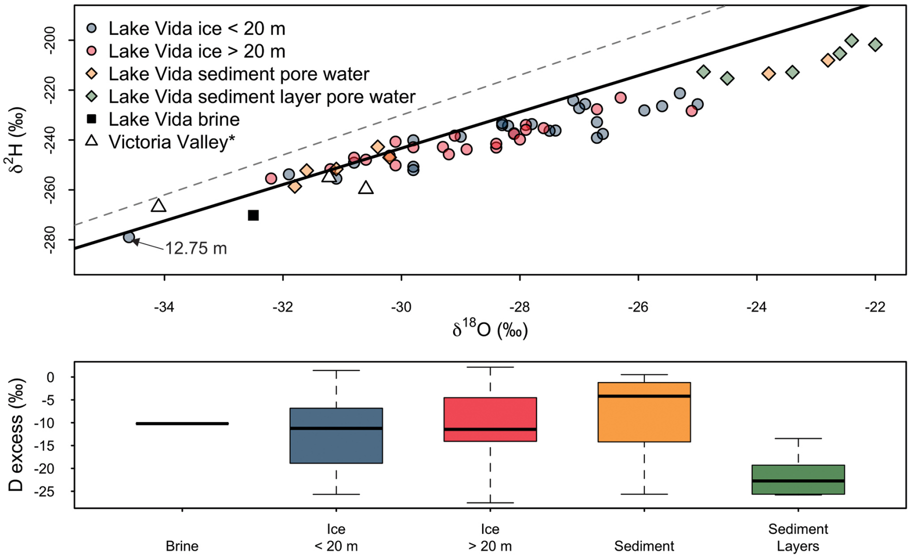

Figure 6. (a) Stable isotope composition of hydrogen (δ^2H) and oxygen (δ^{18}O) in Lake Vida ice, pore water in thin sediment pockets and thick sediment layers, and brine. Published values for Victoria Valley surface water and snow are denoted by black triangles (Hagedorn et al., 2010). The local meteoric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

Stable isotope (δ^{18}O and δ^2H) values of both the ice and sediment pore water fell on or below the local meteoric water line but on a slope consistent with the sublimation of ice (Sokratov and Golubev, 2009; Hagedorn et al., 2010; Lacelle et al., 2011, 2013) and isotopic values reported around the dry valleys (Gooseff et al., 2006; Harris et al., 2007) (Fig. 6). Most of the ice and sediment pore water samples have δ^{18}O between −25 and −32 ‰. There are, however, two notable exceptions. The first is the sample at 12.75 m that is significantly depleted in ^{18}O versus all other ice samples. The second is the pore water of the four thick sediment layers below 21 m, as well as some thinner sediment layers above, that are relatively enriched in ^{18}O in comparison to the ice.

GPR profiles distinctly map the edge of the lake basin until an impenetrable basal reflector around 21 m (Fig. 7). Along the basin edges there are features resembling ancient terraces, especially at 8 m (Fig. 7). From 8 to 12 m, synchronous, wavy reflectors are spaced approximately 1 m apart. In all profiles, ice and sediment layers appear to be continuous across the lake.

5 Discussion

The low ion concentration, absence of large sediment layers, and clear GPR returns in the top 8 m of ice suggest that the upper ice has formed recently under a positive water balance. The level of Lake Vida has risen 3.5 m in the last 40 years (Doran, 2015) and has a hydrologic history similar to Lake Bonney (Fig. 8), which has been documented to have risen ∼ 16 m from 1903 to 2010 (Chinn, 1993; Doran, 2015). If a linear extrapolation is applied to the Vida record based on the correlation of volumetric change to Lake Bonney, the surface of Lake Vida would have risen ∼ 7.7 m from 1903 to 2010 when our cores were collected. Therefore, the 8 m contour in Fig. 1 may be an approximate representation of the lake shore in 1903. This is an indication of the rapidity with which the level of Lake Vida can change over time.

5.1 Upper ice

The ice between 9 and 13 m contains almost no sediment. In addition, the ice sample at 12.75 m has a salinity of 3.5 g L^{-1} and is heavily depleted in ^{18}O and ^{2}H versus all other ice samples (Fig. 6). Both an increase in salinity and the relative depletion of heavy isotopes are a signature of freshwater freezing. During freezing at ice–water interface, equilibrium isotope fractionation preferentially retains heavy isotopes in the ice, which leads to a depletion in heavy isotopes in the unfrozen water (Gallagher et al., 1989; Horita, 2009). Under equilibrium conditions, the isotopic composition for the fraction of water that remains unfrozen during the freezing of ice downward (δ_f) can be approximated by Rayleigh fractionation (Eq. 1), where α is the fractionation factor between ice and water (measured at 0 °C), f is the fraction of water that remains unfrozen, and δ_o is the original isotopic value of the water (Miller and Aiken, 1996; Fritz et al., 2011).

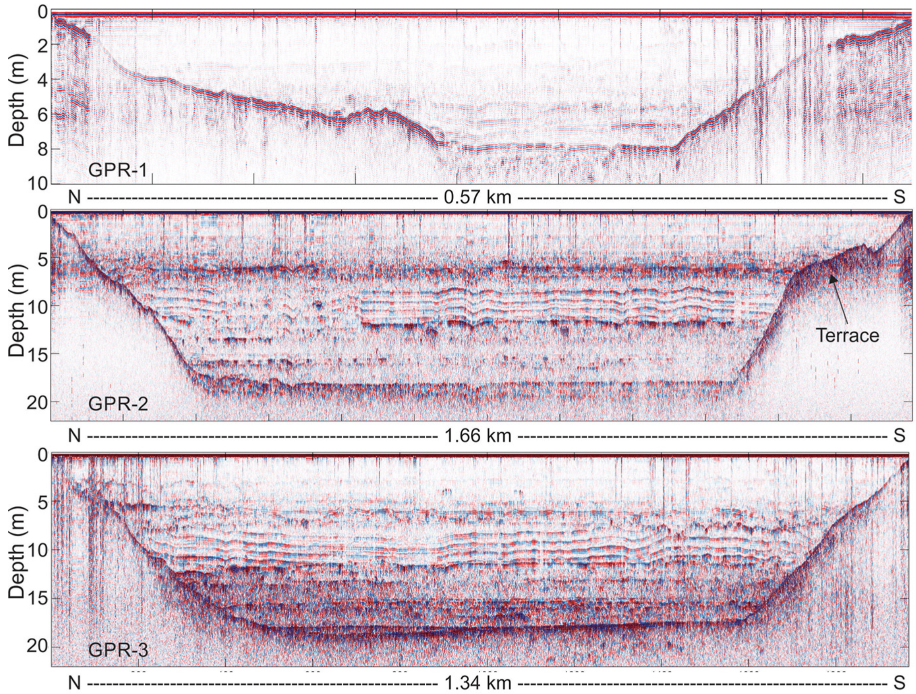

Figure 7. GPR profiles recorded north to south across Lake Vida (Fig. 1).

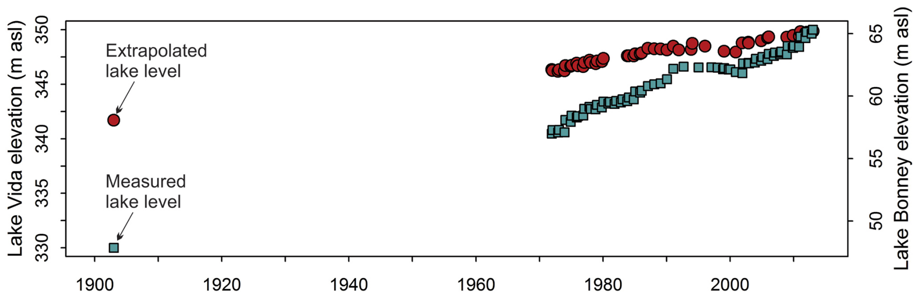

Figure 8. Lake Vida and Lake Bonney surface elevations. The 1903 Lake Bonney elevation is inferred from a measurement at Lake Bonney narrows by Robert Falcon Scott (Chinn, 1993). The 1903 Lake Vida elevation is extrapolated from the correlation between the two lakes for 1971–2010.

$$(\delta_\mathrm{f} - \delta_\mathrm{o}) \cong 10^3(\alpha - 1)(\ln f) \quad (1)$$

As a simple model for the depletion of stable isotopes during freezing, we employ Eq. (1) with $\delta_\mathrm{o} = -25$ and -30 ‰ (within the range of δ^{18}O of Lake Vida ice) and $\alpha = 1.0029$ (average value from Horita, 2009) to test if the high-salinity and low-isotopic values at 12.75 m could have resulted from the downward freezing of 3 to 4 m of ice above. To obtain the observed δ^{18}O at 12.75 m of -34.6 ‰, parameter f must equal 0.04 and 0.20 for an original source δ_o of -25.0 and -30, respectively. This illustrates that the relatively depleted ^{18}O at 12.75 m could easily be generated through freezing processes. This, along with the high salinity and lack of sediment particles, indicates a high likelihood of downward freezing in this section of the ice core. A 3 to 4 m layer of water on the surface of the lake could easily result from the combination of a large surface flood and the melt generated at the water/ice contact. Particles settling out from this event may have generated the sediment seen between 12 and 13 m (Fig. 4). This record may be affirmation that anomalous warming events, such as the flood year of 2001/2002 (Barrett et al., 2008; Doran et al., 2008), are not unprecedented in the dry valleys.

In the GPR profiles, the continuity of the horizontal reflectors across the lake validates the extrapolation of ice core records across the entire lake body. The noticeable undulations in the reflectors between 8 to 12 m are interpreted as density contrasts in ice layers (Arcone and Kreutz, 2009), which may have been formed from 3 to 4 m of liquid water freezing downwards, as discussed previously in regards to isotopic composition. Initially horizontal, these bands were later forced into their present configuration by pressure due to freezing below. Also evident in the radar profiles is a preserved paleo-terrace at 8 m depth along the south end of the lake (Fig. 7). This may be evidence that Lake Vida maintained an elevation of 8 m below present for a prolonged period. The lack of downcutting along the lake margins below 8 m (Fig. 7) further suggests the lake level has mostly risen since this time, which, based on our previous calculation, encompasses the past 100 years.

5.2 Lower ice

In all GPR profiles the radar signal is attenuated below 21 m. Doran et al. (2003) interpreted this horizontal reflector as the top of a large brine body. From the drilling detailed in this study, we now interpret this impenetrable basal reflector to be the SL21.62 sediment layer, as the thickness and salt content likely inhibited radar penetration (Frolov, 2003).

Below 16 m in the ice core, salinity increases and the ice appears to have recrystallized. Recrystallization is induced by temperature changes, stress or strain on the ice, and/or the presence of debris (Samyn et al., 2008). There are three processes by which the lower ice and sediment layers may have formed: (1) the repeated freezing of surface water and deposition of sediment layers (increased salinity is generated from the concentration of salts through evaporation/sublimation), (2) the formation of segregated ice in lake sediments from the freezing of brine from beneath, or (3) the lower ice is remnant glacial ice.

The following were a priori hypotheses:

1. Victoria Valley has not been occupied by a valley glacier since the Miocene (Fountain et al., 1998). Combined with the presence of a glacial lake, there should be no remnant glacial ice near the surface of Victoria Valley.

2. If the ice were segregated ice, we would expect:

   a. a gradient in isotopic composition throughout the sediment and ice layers, as discussed in French and Harry (1990) and observed in closed system freezing of massive ground ice (Fritz et al., 2011);

   b. diatoms present in the sediment layers, which were originally lake sediments;

   c. the lower ice to have similar ionic ratios to the underlying brine.

However, we must note that segregated ice has not been researched in a lacustrine context.

3. If the sediment layers were formed from surface deposition, we would expect the layers to be roughly horizontal.

The following results support that the lower lake ice was formed from surface deposition and is not segregated ice formed from brine seeping upwards through lake sediments.

1. Only the deepest sediment layer, SL26.28, contains abundant diatom frustules. This layer also begins to fine downward in grain size and increases in total carbon content (Fig. 5). This is the only layer that resembles lake sediments and yet is underlain by ice with similar chemical composition to all ice below 21 m.

2. The GPR profiles reveal a lake-wide horizontal layer at 21 m (Fig. 7).

3. The lower ice is chemically distinct from the brine (Fig. 2).

4. The sediment layers are relatively enriched in the stable isotopes of oxygen and hydrogen versus the surrounding ice (Fig. 6).

Therefore, we hypothesize the entire 27 m of ice and sediment was formed from surface processes, and the brine that enters the drill holes is sourced from below 27 m. Here, we explore processes that might have led to the formation of interspersed sediment layers in 27 m of lake ice.

If the lower ice formed from surface inflow rather than brine, the sediment layers in the lower ice are unusual. Field observations at Lake Vida during the 1990s, when lake levels were static, and during the 2000s, when lake levels were rising, reveal that the surface of the lake is largely flat and free of surface sediment. Aeolian transport of sediment is common in Victoria Valley (Speirs et al., 2008); however, aeolian deposition onto a dry, flat ice cover has a high probability of further redistribution by wind and may not be readily entrained into the ice column. This conclusion is supported by observations that windblown sediment largely does not get trapped on the frozen ice surface of Lake Vida but saltates across the lake. Therefore, we propose that sediment is mainly delivered onto the surface of the lake through fluvial transport as melt streams flood the lake surface. A saturated lake surface provides a mechanism for sediment to both infiltrate cracks in the ice and freeze beneath a new layer of water.

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

The hypothesis that sediment layers were formed from long-term evaporation and sublimation is further supported by the isotopic enrichment and low deuterium excess of water contained within the sediment layers (Fig. 6). When water evaporates from a water body to the atmosphere, the remaining water becomes enriched in the heavy isotopes of oxygen and hydrogen (Horita, 2009). The same has been shown for the sublimation of snow and ice (Neumann et al., 2008; Sokratov and Golubev, 2009; Lacelle et al., 2011). Likewise, negative deuterium excess values suggest that evaporative fractionation has more strongly modified oxygen isotopes over hydrogen isotopes.

During periods of significant lake drawdown, the sediment layers may have been visible near the surface of the lake and would have been analogous to the ice-cemented permafrost found at higher elevations in the Victoria Valley. At 450 m a.s.l. in Victoria Valley, the ice-rich permafrost in a 1.6 m soil profile had similar sediment characteristics to the lower sediment layers in Lake Vida, with a median grain size range of 357–510 µm and water content $< 13\,\%$ (Hagedorn et al., 2007). The dry permafrost/ice-cemented contact was found at 22 cm below the surface, which was also the approximate maximum depth of the 0 °C isotherm during the 3 years of study (Hagedorn et al., 2007). At the edge of Lake Vida, soil temperatures at 10 cm depth rise only slightly above 0 °C during the short summer (Fountain, 2015). On Lake Vida, we propose that similar thermal conditions preserved the ice found below and between the thick sediment layers. As sediment layers on the ice thickened to almost 20 cm, the amassed sediment provided insulation for the ice beneath and allowed the ice to remain below freezing temperatures during the summer. This processes allowed for the repeated deposition and preservation of sediment and ice layers.

Only in SL26.28 does the mean grain size significantly decrease toward the base of the core (Fig. 5). The occurrence of silts is not common in the dry valleys but is found in the sediment beneath ice-covered lakes (Wagner et al., 2006, 2011). The increase in TC concurrent with the decrease in grain size, as well as the presence of freshwater diatom frustules, points to the possible occurrence of more open water conditions during this time.

5.3 Constraining the age of the lower ice

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

Of the eight radiocarbon samples taken from the upper 13 m of ice in Lake Vida, only one sample returned a concentration indicative of modern carbon (Fig. 4). All indications point to erroneously old carbon dates as we assume that at least the upper 7.7 m of ice formed during the last century. However, the radiocarbon dates impart a maximum age constraint on the ice, as it is almost impossible that samples were contaminated by young carbon, and indicate the top 27 m of ice formed after 6300 ^{14}C yr BP. This aligns with the geomorphic reconstructions that indicate that Victoria Valley was filled with a deep (> 200 m) glacial lake prior to 8600 yr ^{14}C BP (Hall et al., 2002, 2010), and the Ross Ice Shelf retreated from the mouth of the dry valleys between 6500 and 8340 ^{14}C yr BP (Hall and Denton, 2000). Furthermore, it may be that the ice cover originated much later, as radiocarbon dates of the dissolved organic carbon fractions in the Lake Vida brine date between 2955 and 4150 ^{14}C yr BP (Murray et al., 2012).

OSL dates at SL21.62 and SL25.59 represent the date at which the minerals in the sediment layers were last exposed to solar radiation or, more specifically, an interval when the ice cover was thin enough to allow sunlight to penetrate to the dated sediment layer, followed by a period of ice growth or further sediment burial that extinguished the light source to the sediment layer. To constrain the amount of time between sediment deposition and burial, we draw on SL26.28, where the relative difference in radiocarbon ages along the length of the layer suggests formation spanned 1400 years. Using this inference and the assumption that light could only penetrate a few centimeters into a layer, we assume burial lag of < 300 years. Therefore, the OSL dates indicate a lake level drawdown and rebound at 1200 (+300) and 320 (+300) yr BP. A lowering at 1200 years matches previous paleolimnological studies of lake levels in Taylor and Wright Valleys (Wilson, 1964; Lyons et al., 1998). Also, if our previous interpolation of a 7.7 m lake level rise in 103 years is further hindcast, it suggests that a 21.62 m lake level rise is not improbable over 320 years.

All errors associated with the quality of dating methods are subsumed when radiocarbon and OSL techniques are considered together. The discrepancy between the two dating techniques has been documented before in the dry valleys (Berger et al., 2010, 2013). Contaminated OSL samples yield artificially young dates and therefore results are considered a minimum age, whereas contaminated radiocarbon samples tend to yield artificially old dates and therefore are considered maximum ages. When viewed together, the two dating techniques constrain evaporation events between 6300 and 320 yr BP and suggest that the current Lake Vida system is a few millennia in age.

From the ages available in this study, there is no discernible correlation between lake highstands/lowstands and the temperature proxy record from neighboring Taylor Dome (Mayewski et al., 1996; Stager and Mayewski, 1997; Steig et al., 2000). This lack of synchronicity between runoff/lake level and temperature has been noted in other lake level reconstructions (Whittaker et al., 2008) and throughout the instrumental climate record (Levy et al., 2013).

6 Conclusions

Our analyses point toward the formation of sediment layers in the Lake Vida ice cover from the accumulation of sediment entrained in the ice cover accumulating during periods of ice ablation and lake level drawdown. The capacity of Lake Vida to integrate watershed processes presents a fundamental framework for understanding hydrological and climatological shifts over time.

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

1. A hydrologically variable climate is not unique to recent times. Lake Vida has experienced major drawdowns that led to the accumulation of four thick sediment layers in the lower ice cover. These drawdowns may have occurred as early as 6300 ^{14}C yr BP, but OSL ages and a presumed reservoir effect in radiocarbon ages suggest these events were likely constrained to the last 1–3 millennia.

2. 27 m of ice was produced from glacial streams flooding the ice surface. In the ice collected to date, there is no freezing from the bottom of the ice cover downwards as suggested by Doran et al. (2003).

3. The brine that entered the drill holes at 16 m and rose to 10.5 m in both 2005 and 2010 appears to be hydrologically connected and sourced from below 27 m.

Lake Vida represents a unique lacustrine system that has recorded a hydrologic history in the growing ice cover. As climate is projected to change (Thompson et al., 2011), Lake Vida may provide an ideal environment for tracking the influence of climate on hydrology in the dry valleys.

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