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27 m of lake ice on an Antarctic lake reveals past hydrologic variability

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Abstract

Lake Vida, located in Victoria Valley, is one of the largest lakes in the McMurdo Dry Valleys. Unlike other lakes in the region, the surface ice extends at least 27 m, which has created an extreme and unique habitat by isolating a liquid-brine with salinity of $195\,g\,L^{-1}$. Below 21 m, the ice is marked by well-sorted sand layers up to 20 cm thick, within a matrix of salty ice. From ice chemistry, isotopic abundances of ^{18}O and ^{2}H, ground penetrating radar profiles, and mineralogy, we conclude that the entire 27 m of ice formed from surface runoff, and the sediment layers represent the accumulation of fluvial and aeolian deposits. Radiocarbon and optically stimulated luminescence dating limit the maximum age of the lower ice to 6300 ^{14}C yr BP. As the ice cover ablated downwards during periods of low surface inflow, progressive accumulation of sediment layers insulated and preserved the ice and brine beneath; analogous to the processes that preserve shallow ground ice. The repetition of these sediment layers reveals climatic variability in Victoria Valley during the mid- to late Holocene. Lake Vida is an excellent Mars analog 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). 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 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 limits the maximum age of the surface ice to ~ 5 yr (Dugan et al., 2013). However, the Lake Vida ice cover is at least partially grounded (Doran et al., 2003), so the ice does not turn over in the same way. Water that is flowing to the lake is trapped on the surface of the ice where it freezes and is later ablated or buried by subsequent ice buildup. In this way, the thick ice on Lake Vida may record past hydrological changes, similar to a glacier; but, unlike a glacier, intermittent accumulation may lead to large discontinuities in the ice cover during prolonged cold/dry periods.

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

The aim of this study is to reconstruct the history of the ice cover on Lake Vida. We examine the isotopic and ion geochemistry, texture, mineral 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. In order to determine the provenance of sediment onto the lake, we compare mineralogy and granulometry of the sediment layers with those of sand samples collected along the floor and stream beds of Victoria Valley. Both radiocarbon dating and optically stimulated luminescence (OSL) dating 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 (340 m a.s.l.) lakes in the McMurdo Dry Valleys (Fig. 1). The lake is endorheic (closed basin), and receives inflow via streams originating from Victoria Upper, Victoria Lower, and Clark Glaciers. Lake Vida occupies a unique climatological niche where summer temperatures can rise slightly above 0 °C to generate stream flow, yet unusually cold winters (compared to the other major valleys in the region) maintain a thick ice cover on the lake. From 1995 to 2000, the mean annual air temperature at Lake Vida (–27.4 °C) was 7 to 10 °C lower than valley bottom temperatures in Taylor and Wright Valleys, but mean summer temperatures were similar (Doran et al., 2002).

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.

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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 liquid water isotope analyzer for isotopic abundances of ^{2}H and ^{18}O. 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-Wenworth scale (Wentworth, 1922). Selected samples were treated with 10 % HCl and photographed under a CamScan CS 44 scanning electron microscope (SEM). Sediment layers were also subsampled to evaluate diatom assemblages and absolute abundance (valves g^{-1} dry weight) via light microscopy. Preparation methods followed standard techniques (Scherer, 1994). A known mass of freeze dried sediment was reacted with 10 % HCl and 10 % H_2O_2 to remove carbonates and organics. Abundance per gram was extrapolated from diatom counts on coverslips in a beaker of known area.

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

Catchment samples were obtained from twelve areas surrounding Lake Vida (Fig. 1) to examine sediment provenance. Four samples were fluvial bedload deposits from river channels (streams were not flowing at the time), seven were from aeolian sources outside of stream channels, and one was collected on the ice surface. Grain size was evaluated on a Malvern particle size analyzer. Mineralogy of sediments in all of the catchment and nine lake samples was identified via binocular optical microscopy. Approximately 300–800 grains were counted per sample.

To confirm the continuity of horizons and sediment layers noted in the ice cores, 55 km of GPR transects were recorded over the surface of Lake Vida in 2010 (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 meters above sea level (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.

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 $< 1\,g\,L^{-1}$ to $34\,g\,L^{-1}$. Although brine inundated the borehole up to 10.5 m depth, and all of the cores below 16 m depth were in contact with the brine while being extracted, it appears from the ratio of major anions/cations (Fig. 2), that the ice was not substantially contaminated by brine in the 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 as 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 SL26.28 revealed abundant diatom frustules of the genera *Pinnulaira and Luticola*.

Mineralogy of sediment grains revealed a higher percentage of quartz and plagioclase feldspar in the ice core sediment layers than in Victoria Valley samples (Fig. 6a) The grain size distribution of the ice core sediments was more aligned with stream bed sediments than aeolian deposits (Fig. 6b). In comparison, the sole sediment sample collected on the surface of the lake was coarser than ice core sediments.

SEM images from the lower sediment layers show no differentiation in microtextures between layers. Grains surfaces, especially those of quartz, ranged from well-rounded to heavily abraded and showed evidence of glacial, aeolian, and fluvial transport (Fig. 7). Deeply imbedded features, such as conchoidal fractures and multiple striations may suggest glacial transport (Fig. 7a, b and e); v-shaped percussion cracks are indicative of fluvial transport (Fig. 7c and d), and rounded grains with surface craters due to collisions may result from aeolian transport (Fig. 7f–h) (Mahaney, 2011).

The upper sediment layers noted in two ice cores retrieved in 2010 correspond to those cored in 1996 (Doran et al., 2003) and 2005 (Taylor, 2009) (Fig. 4). The radiocarbon dates vary significantly, especially in the upper ice where there is no correlation between age and depth. The dates from SL26.28 do show increasing age with depth (4909 ± 46 to 6300 ± 49 ^{14}C yr BP) and provide a maximum age of the lower ice. A portion of SL21.62 and SL25.59 showed no evidence of exposure to sunlight during collection, and returned OSL dates of 320 ± 40 and 1200 ± 100 yr BP, respectively; dates which are younger than the radiocarbon ages from the respective horizons. It is noted that ^{13}C values are unreliable and not reported due to high fractionation during AMS measurement.

The δ^{18}O and δ^{2}H values of both the ice and sediment pore water fell below the global meteoric water line, but on a slope consistent with the sublimation of ice (Sokra-

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tov 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. 8). Most of the ice and sediment pore water samples have $\delta^{18}O$ 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. 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.

GPR profiles distinctly map the edge of the lake basin until an impenetrable basal reflector around 21 m (Fig. 9). Along the basin edges there are features resembling ancient terraces, especially at 8 m (Fig. 9). From 8 to 12 m, synchronous, wavy reflectors are spaced approximately 1 m apart. In all transects, ice and sediment layers appear to be continuous across the lake.

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 yr, and has a hydrologic history similar to Lake Bonney (Fig. 10), which has been documented to have risen ∼ 16 m from 1903 to 2010 (Chinn, 1993, Doran, unpublished data). If a linear extrapolation is applied to the Vida record based on the correlation of volumetric change to Lake Bonney, the surface of Lake Vida would have risen ∼ 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.

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 (Fig. 8). Combined, this is a signature of freshwater freezing, which increases the dissolved salt load and depletes heavy isotopes in the water that remains unfrozen (Horita, 2009).

The isotopic value for the fraction of water that remains unfrozen during the freezing of ice downward (δ_f) can be approximated by Eq. (1), where α is the fractionation factor, f is the fraction of water that remains unfrozen, δ_o is the original isotopic value of the water.

$$(\delta_f - \delta_o) \cong 10^3(\alpha - 1)(\ln f) \tag{1}$$

We employ Eq. (1) with $\delta^{18}O = -25.0$ ‰ (isotopic value at 10.5 m) and $\alpha = 1.0029$ (Horita, 2009) to test if the high salinity and low isotopic values at 12.75 m could have resulted from the downward freezing of 3 to 4 m of ice above. With these parameters and $f = 0.96$, $\delta^{18}O = 34.3$ ‰. This fits the observed $\delta^{18}O$ at 12.75 m of −34.6 ‰ and along with the high salinity and lack of sediment particles, indicates a high likelihood of downward freezing in this section of the ice core. A 3 to 4 m layer of water on the surface of the lake could easily result from the combination of a large surface flood and the melt generated at the water/ice contact. This record may be affirmation that anomalous warming events, such as the flood year of 2001/02 (Barrett et al., 2008; Doran et al., 2008), are not unprecedented in the Dry Valleys.

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

Three lines of evidence support that the entirety of the ice core was formed from surface inflow rather than brine freezing.

1. Sediment layers found in the lower portion of the core (21 to 27 m) must have been deposited on the surface of the ice. These sediment layers are underlain by ice which is too thick to be segregation ice, as ice lenses formed within basal sediments beneath glaciers are typically thin at 0.01–0.1 m and are spaced more than 0.5 m apart (Christoffersen and Tulaczyk, 2003).

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2. The ice in SL26.28 contains abundant diatom frustules, which are photosynthetic organisms. All genera are freshwater diatoms associated with Dry Valley streams (McKnight et al., 1999; Warnock and Doran, 2013).

3. If the ice was formed from brine freezing, the lower ice would be expected to have similar ionic ratios to the underlying brine, yet these are clearly distinct (Fig. 2). This final argument may not hold if the brine underwent significant mineral precipitation following freezing that altered the relative proportions of ions in the brine.

Based on these observations, our interpretation is that the entire 27 m of ice cored on Lake Vida has formed from the bottom up.

The sediment layers in the lower ice are also unusual, given we hypothesize the ice formed from surface inflow. In January 2002, anomalously high stream discharge flooded the surface of Lake Vida with turbid water; yet, only a thin band of sediment < 1 cm is evident in the upper 3 m of the ice core. Therefore, it is improbable that sediment layers 20 cm in thickness formed from a single surface flooding event. Additionally, it has been noted in Taylor Valley (where lake ice is formed at the bottom of the ice cover, not the surface), that during the austral summer the low albedo of surface sediment can cause it to warm and move downward in the ice cover (Hendy, 2010). This movement tends to aggregate sediment into layers and pockets approximately 2 m below the surface (Priscu et al., 1998). From this, it is hypothesized that the thick sediment layers in Lake Vida formed from repeated deposition rather than individual events, and were amassed during periods having a negative water balance as the ice cover ablated. This is further supported by the isotopic enrichment of water contained within the sediment layers that reveals an evaporative signature in comparison to the ice layers (Fig. 8) (Horita, 2009).

From the overall mineralogy of stream and aeolian samples it was not possible to ascertain the provenance of the sediments in the ice core; however, the majority of sediment contained within the ice had a mean grain size smaller than sediment collected from the ice surface and most of the aeolian deposits collected on the valley

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floor (Fig. 6). This may be the result of smaller particles preferentially melting through the ice cover (Hendy, 2010) or the prevalence of riverine input. Field observations support that during high flow years, turbid river water flows over the surface of the lake. A saturated lake surface provides a mechanism for sediment to both infiltrates cracks in the ice, as well as become frozen beneath a new layer of water. Aeolian transport of sediment is common in Victoria Valley (Speirs et al., 2008); however, aeolian deposition onto a dry, flat ice cover has a high probability of further redistribution by wind and may not be readily entrained into the ice column. This conclusion is supported by field observations that windblown sediment largely does not get trapped on the frozen ice surface of Lake Vida, but saltates across the lake.

During periods of dry climatic conditions in Victoria Valley, the sediment layers may have been visible near the surface of the lake, and would have been analogous to the ice-cemented soils found at higher elevations in the Victoria Valley. At 450 m 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 three years of study (Hagedorn et al., 2007). At the edge of Lake Vida, soil temperatures at 10 cm depth, rise only slightly above 0 °C during the short summer (MCM LTER). On Lake Vida, we propose that similar thermal conditions preserved the ice found below and between the thick sediment layers. As sediment layers on the ice thickened to almost 20 cm, the amassed sediment provided insulation for the ice beneath, and allowed the ice to remain below freezing during the summer.

Only in SL26.28, does the mean grain size significantly decrease toward the base of the core (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.

In all GPR transects, the radar signal is attenuated below 21 m. Doran et al. (2003) interpreted this horizontal reflector as the top of a large brine body. From the drilling detailed in this study, we now interpret this impenetrable basal reflector to be the SL21.62 sediment layer, as the thickness and salt content likely inhibited radar penetration (Frolov, 2003). Above SL21.62, the continuity of the horizontal reflectors across the lake validates the extrapolation of ice core records to the entire lake body. The noticeable undulations in the reflectors between 8 to 12 m are interpreted as density contrasts in ice layers (Arcone and Kreutz, 2009), which may have been formed from 3 to 4 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. 9). 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. 9a) further suggests the lake level has mostly risen since this time, which based on our previous calculation encompasses the past 100 yr.

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 or strong salinity gradients, or the inclusion of old organic material reworked into modern stream water (Doran et al., 1999; Hendy and Hall, 2006). In Victoria Valley, water travels more than 1 km from glacial sources to Lake Vida, which should allow waters to equilibrate with modern atmosphere before reaching the lake. Moat waters too should contain mostly modern carbon, although Doran et al. (2014) found moat waters in Taylor Valley lakes with apparent DIC ages as high as 3790 ^{14}C yr BP. Only

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lakes with large open water moats like Lake Fryxell seem to have modern DIC. Hall and Henderson (2001) found that at Lake Vida, lacustrine carbonates along the shorelines with uranium/thorium ages of 9600 yr BP had a ^{14}C reservoir age of +3600 yr BP. They concluded that old CO_2 was likely input into the system with meltwater from the Ross Ice Sheet.

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

OSL dates at SL21.62 and SL25.59 represent the date at which the minerals in the sediment layers were last exposed to solar radiation, or more specifically, an interval when the ice cover was thin enough to allow sunlight to penetrate to the dated sediment layer, followed by a period of ice growth or further sediment burial which extinguished the light source to the sediment layer. Therefore, the OSL dates indicate a lake level drawdown and rebound at 1200 and 320 yr BP. A lowering at 1200 yr 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 yr is further hindcast, it suggests that a 21.62 m lake level rise is not improbable over 320 yr.

The discrepancy between OSL and radiocarbon dating techniques has been documented in the Dry Valleys (Berger et al., 2010, 2013). Contaminated OSL samples

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yield artificially young dates, whereas contaminated radiocarbon samples tend to yield artificially old dates. Together, the two dating techniques constrain evaporation events, and suggest that the current Lake Vida system is a few millennia in age.

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

6 Conclusions

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 at least four major drawdowns that led to the accumulation of thick sediment layers in the lower ice cover. These drawdowns may have occurred as early as 6300 ^{14}C yr BP, but OSL ages and a presumed reservoir effect in radiocarbon ages suggest these events were likely constrained to the last 2–3 millennia.

2. The accumulation of discrete sediment pockets into $\sim$ 20 cm thick layers insulates the ice beneath. This is a significant process for the preservation of ancient ice beneath sediment in cold and dry climates, and may have consequence for the preservation of ancient ice and brine on other icy planets.

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

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

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

Lake Vida represents a unique lacustrine system, which during wet periods, records a hydrologic history in the growing ice cover. As climate is projected to 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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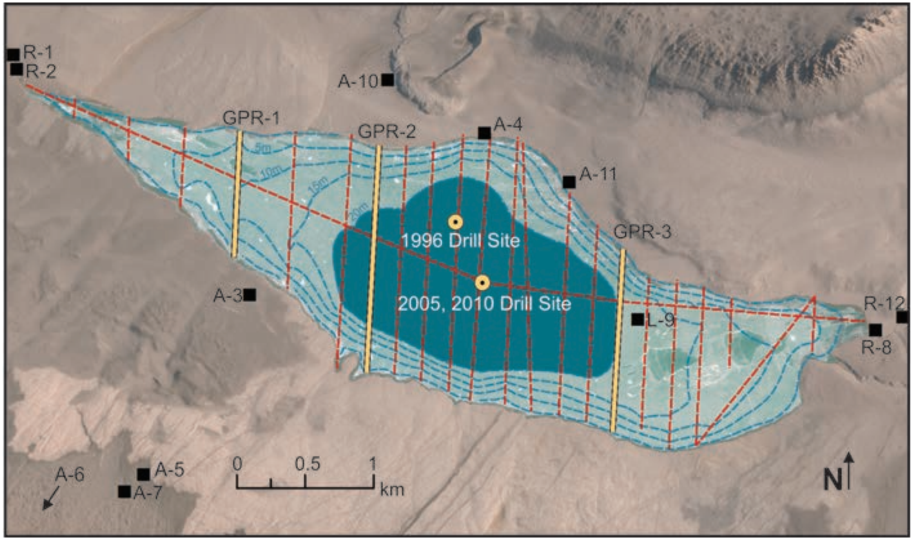

Figure 1. Location of drill sites and GPR transects (red dashed line) on Lake Vida in central Victoria Valley ©DigitalGlobe, Inc. (2011). Highlighted GPR transects 1–3 are presented in the manuscript. Bathymetric lines 0–20 m were digitized and interpolated from GPR profiles. The dark blue area below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

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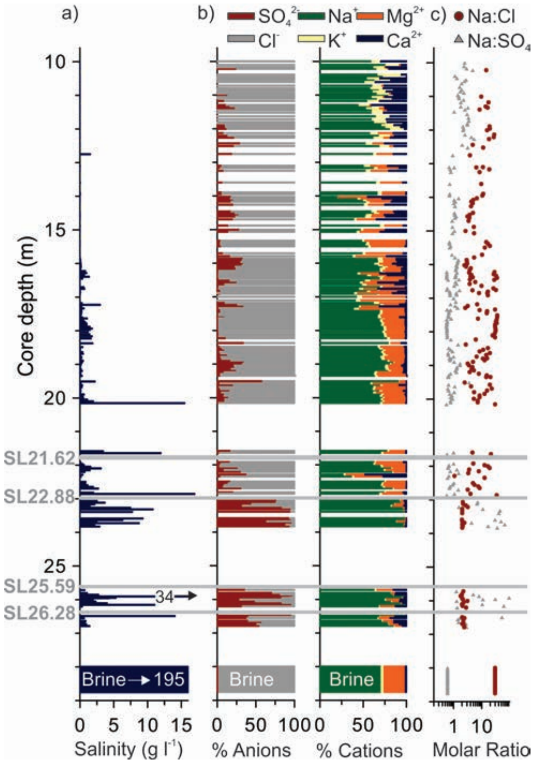

Figure 2. (a) Salinity of ice samples below 10 m subsampled from the Lake Vida ice cores. Salinities ranged from < 1 to 34 g L^{-1}. Brine salinity is 195 g L^{-1}. SL21.62, SL22.88, SL25.59, and SL26.28 represent sediment layers > 10 cm thick, present in the 27 m ice core. **(b)** Total percentage of major anions and cations present below 10 m in the Lake Vida ice cores. Ion percentages in the brine are noted at the base of the figure. **(c)** Ratio of $Na^+ : Cl^-$ and $Na^+ : SO_2^-$ (mol : mol).

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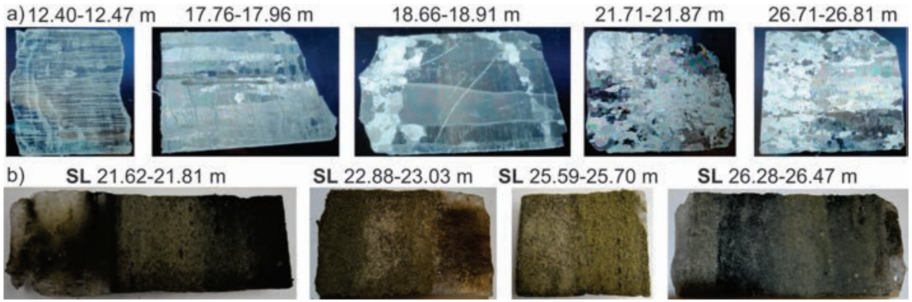

Figure 3. (a) Images of thick sections of Lake Vida ice between two sheets of polarized film. At 12.40–12.47 m, the entire section is a single ice crystal. With depth, the average grain size of ice crystals decreases to < 1 cm. **(b)** Photographs of sediment sections SL21.62, SL22.88, SL25.59, and SL26.28.

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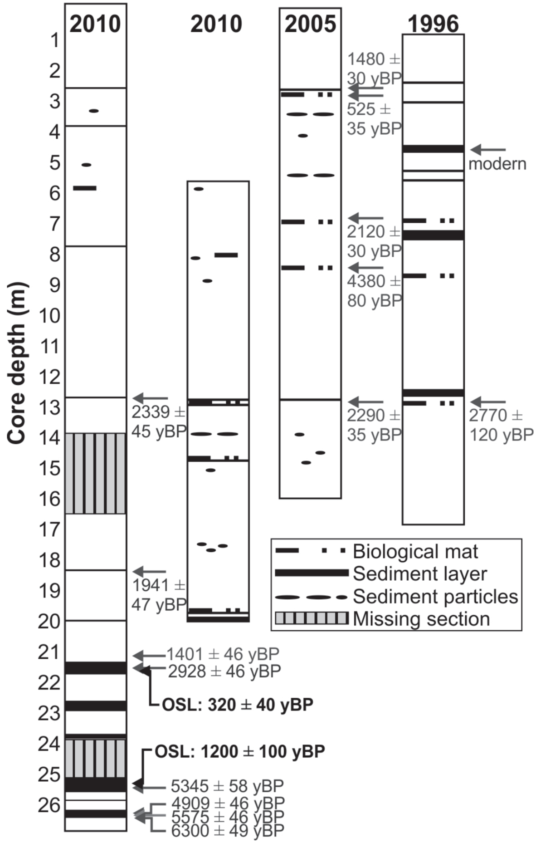

Figure 4. Lake Vida ice core logs from 1996, 2005 and 2010. Two ice cores were drilled in 2010. For core locations in 1996 and 2005, see Doran et al. (2003) and Murray et al. (2012). Surface heights are adjusted to the 2010 lake elevation.

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27 m of lake ice on an Antarctic Lake

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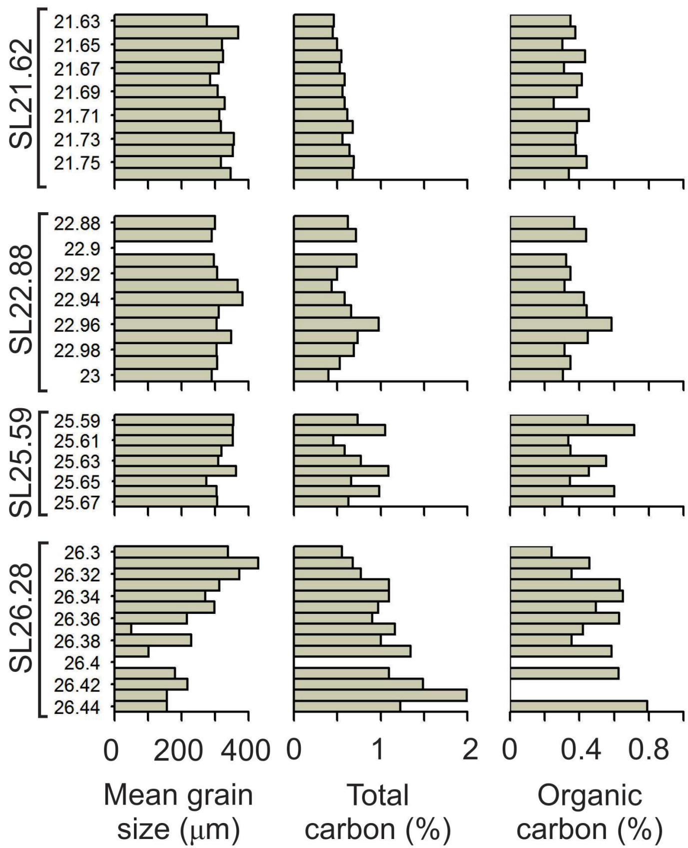

Figure 5. Mean grain size (µm), percentage of total carbon (%), and percentage of organic carbon (%) in sediment sections removed from the 27 m ice core from Lake Vida.

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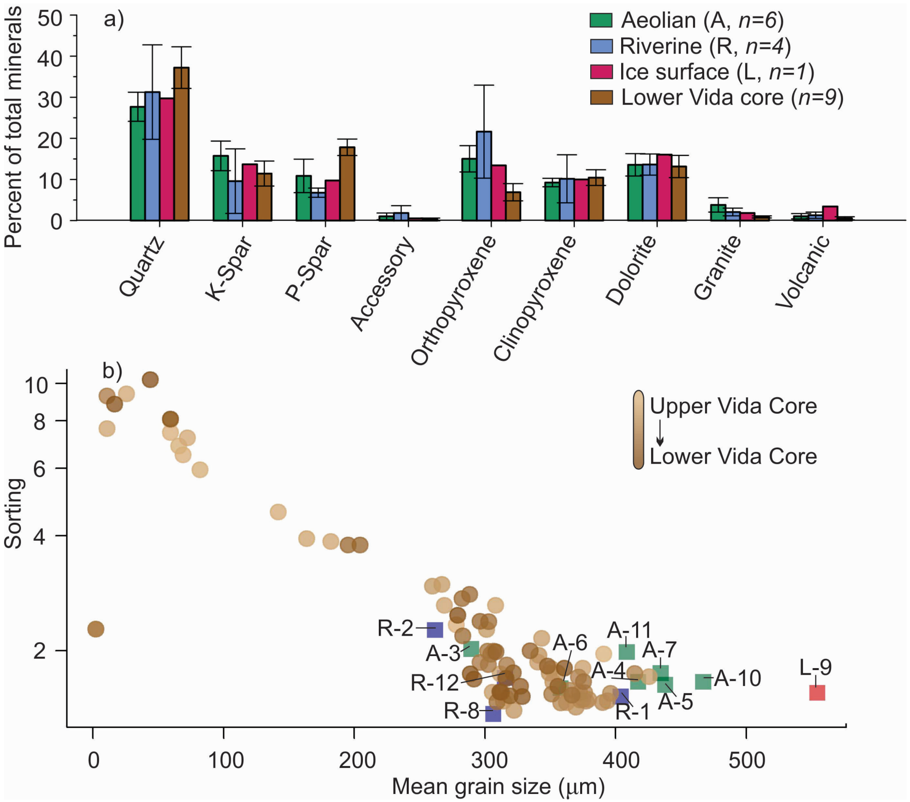

Figure 6. (a) Mineralogy of sediment samples from Victoria Valley and the Lake Vida ice core. Fluvial samples were collected within the stream channels and likely represent bedload. Aeolian samples were outside the channels, and the ice surface was sampled near the middle of the lake (Fig. 1). The ice core samples were obtained from sediment layers within the ice core. **(b)** Mean grain size and sorting of the samples represented in **(a)**. The darker shading of circles represents a lower depth in the Vida ice core.

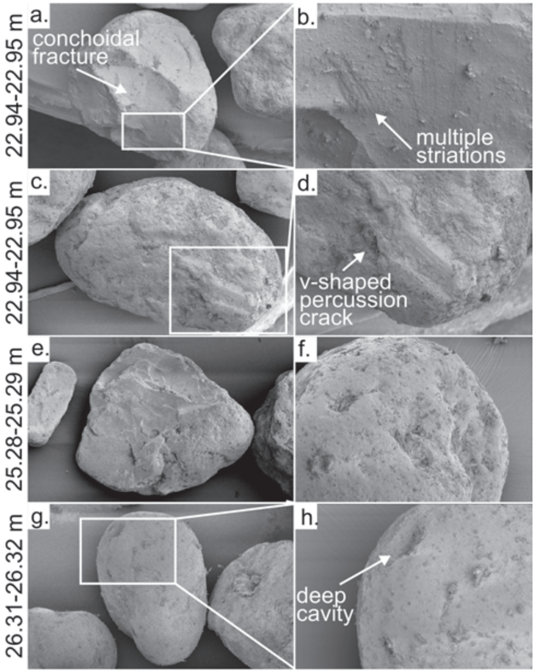

Figure 7. SEM images of select grains from sediment layers in Lake Vida. **(a, b)** Grain shows a large conchoidal fracture and multiple striations. **(c, d)** Sub-rounded grain with abraded surface. Micro-features include a v-shaped percussion crack. **(e)** Grain with high relief features indicative of glacial transport. **(f)** Rounded grain with multiple collision features. **(g, h)** Rounded grain with several deep cavities.

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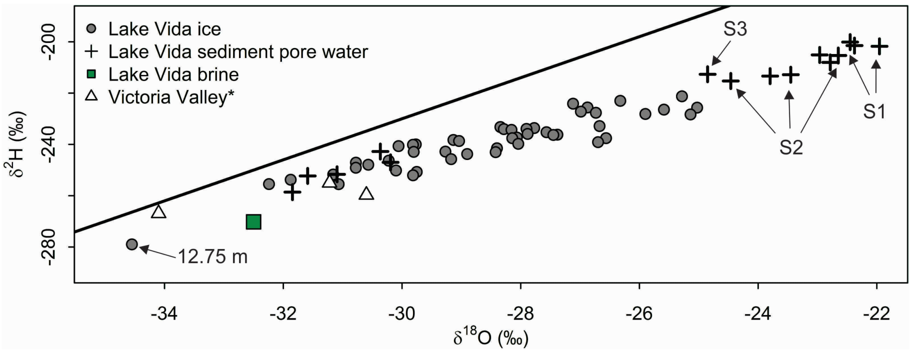

Figure 8. Stable isotope concentrations for deuterium (δ^2H) and δ^{18}O in Lake Vida ice, sediment pore water, and brine. Published values for Victoria Valley surface water and snow are denoted by triangles (Hagedorn et al., 2010). The global meteoric water line (GMWL) is plotted for reference.

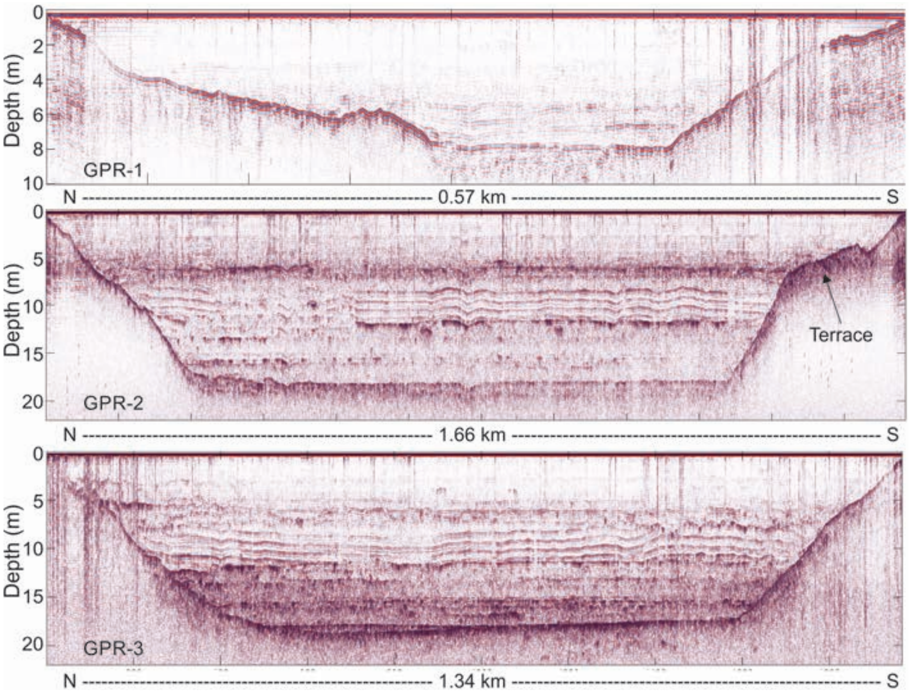

Figure 9. GPR transects recorded north to south across the surface of Lake Vida (Fig. 1).

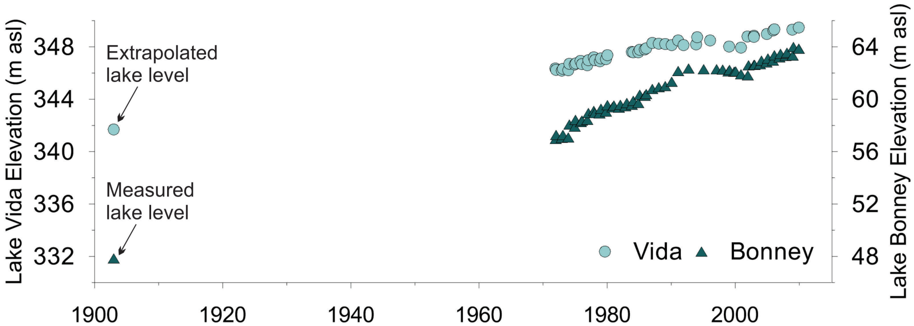

Figure 10. Lake Vida and Lake Bonney surface elevations. The 1903 Lake Bonney elevation is inferred from a measurement at Lake Bonney narrows by Robert Falcon Scott (Chinn, 1993). The 1903 Lake Vida elevation is hindcasted from the correlation between the two lakes for 1971–2010.

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