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Key Points:
- Geophysical survey finds low resistivities beneath a lake in Antarctic Dry Valleys
- Liquid brine abundant beneath Antarctic lake
- Aquifer provides microbial refugium in cold desert environment

Supporting Information:
- Text S1 and Figure S1

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Subsurface imaging reveals a confined aquifer beneath an ice-sealed Antarctic lake

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Abstract Liquid water oases are rare under extreme cold desert conditions found in the Antarctic McMurdo Dry Valleys. Here we report geophysical results that indicate that Lake Vida, one of the largest lakes in the region, is nearly frozen and underlain by widespread cryoconcentrated brine. A ground penetrating radar survey profiled 20 m into lake ice and facilitated bathymetric mapping of the upper lake basin. An airborne transient electromagnetic survey revealed a low-resistivity zone 30–100 m beneath the lake surface. Based on previous knowledge of brine chemistry and local geology, we interpret this zone to be a confined aquifer situated in sediments with a porosity of 23–42%. Discovery of this aquifer suggests that subsurface liquid water may be more pervasive in regions of continuous permafrost than previously thought and may represent an extensive habitat for microbial populations.

1. Introduction

The McMurdo Dry Valleys (MDV, 77–78°S, 160–164°E) of East Antarctica are one of the coldest unglaciated environments on Earth and a possible planetary analogue to the cold and icy conditions that once existed on Mars [*Head and Marchant*, 2014]. In this habitat, terrestrial hydrologic inputs are limited to summertime snow and glacial melt that feed surface streams and lakes. Permafrost is continuous, and ground ice is abundant close to the surface at lower elevations [*Bockheim et al*., 2007].

In the MDV, groundwater is an unidentified component of the hydrologic cycle, yet it is critical to the biosphere, as it can act as a microbial refugium by providing a stable habitat for microbial populations that could not survive at the surface. For groundwater to remain liquid below 0°C, the freezing point must be depressed by increased salinity. This can occur through cryoconcentration, whereby solutes are excluded during the freezing process and concentrated in the remaining liquid. Even in continuous permafrost, saline groundwater may pervade through tortuous pathways in sediments with low hydraulic conductivities [*Seyfried and Murdock*, 1997; *Dickinson and Rosen*, 2003].

In the MDV, Lake Bonney, Lake Vanda, and Lake Vida are known to contain hypersaline water at depth [*Chinn*, 1993]. From geochemical data at Lake Bonney, it has been interpreted that the high salinity results largely from cryoconcentration events [*Lyons et al*., 2005]. Resistivity measurements from the Dry Valleys Drilling Project (DVDP) in the 1970s indicated that Lake Vanda and Lake Bonney were not underlain by frozen ground, and authors suggested that a hydrological connection between the lakes and a deep groundwater source was possible [*McGinnis and Jensen*, 1971]. Recent studies on brine geochemistry have found little evidence of a deep groundwater source to most of the lakes [*Lyons et al*., 1998b; *Witherow et al*., 2010], and consequently, research has shifted toward seasonal shallow subsurface water [*Levy et al*., 2011; *Dickson et al*., 2013; *Gooseff et al*., 2013].

Lake Vida is unique among all lakes investigated to date, because of the presence of a pressurized brine system within and beneath at least 27 m of lake ice. The brine was sampled via surface drilling in 1996, 2005, and 2010 and was found to harbor a microbial population [*Murray et al*., 2012]. However, as the depth of lake ice and volume of brine beyond 27 m was unknown, the extent of the biosphere remained undefined.

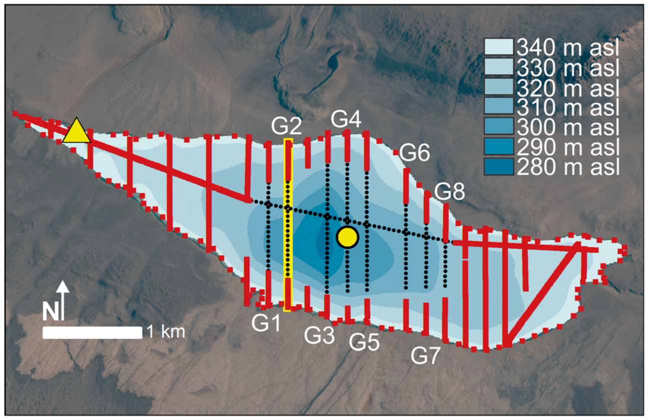

Figure 1. Lake Vida bathymetry acquired from interpolation of GPR transects. This is an estimate of potential lake depth (maximum ~60 m). The thick red squares near the edges of the lake represent known lake basin contact depths from GPR data, whereas the black dots in the center represent the eight transects where data were estimated by spline interpolation. The yellow dot corresponds to the 2010 Lake Vida drill site, and the yellow triangle locates DVDP-6. The highlighted yellow transect is shown in Figure 2.

Here we use ground penetrating radar (GPR) and an airborne transient electromagnetic (AEM) survey to map the ice stratigraphy and ice-bed contact of Lake Vida and, for the first time, constrain the volume and spatial extent of brine in and beneath the lake.

2. Study Site

Lake Vida, at 6.8 km^2, is one of the largest closed basin lakes in the MDV and is located in the middle of Victoria Valley at 350 m above sea level (asl) (77°23′S, 161°56′E). Unlike other lakes in the MDV, the ice cover is extremely thick, at least 27 m and punctuated by thick (~20 cm) sediment layers. During both the 2005 and 2010 coring campaigns, brine entered the drill hole below 16 m and rose to 10.5 m below the surface [*Murray et al*., 2012]. The brine was −13.4°C with a salinity of 195 g L^{-1} [*Murray et al*., 2012]. Lake Vida is the only lake in the world known to contain pressurized brine confined beneath ice and sediment layers.

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

3. Methods

GPR profiles were recorded across Lake Vida with a Geophysical Survey Systems, Inc. SIR-3000 acquisition unit and handheld GPS (error ± 5 m). Fifty kilometer of transects were recorded at 400 MHz in 2010, and several transects were repeated at 900 MHz in 2011 (Figure 1). Bathymetry was mapped from the ice/basin contact depths resolved from the GPR transects (Figure 2). In general, this was possible down to 20 m depth before the signal attenuated, and no lake bed reflection was returned. For transects G1 to G8, the ice/bed contact below 20 m was approximated by a fitting a cubic spline function using R [*R Core Team*, 2012] (see supporting information). Fitted points were merged with the GPR data set and spatially interpolated into a bathymetric map using natural neighbor interpolation in Environmental Systems Research Institute ArcGIS 10.1.

The AEM survey was performed using a SkyTEM504 system flown underneath a Bell 212 helicopter [*Sørensen and Auken*, 2004]. Electromagnetic signals were present only in soundings above brine, as next to no signal was produced from the very resistive permafrost or crystalline bedrock. In total, 727 soundings, covering 16.7 km, were left after processing, split between five flights lines (ST1 to ST5; Figures 3 and 4). The soundings were inverted using a multilayer inversion approach discretizing the models in 28 logarithmically distributed bins from 0 to 600 m below the surface [*Auken and Christiansen*, 2004; *Viezzoli et al*., 2008], although in approximately half of the locations, the depth of investigation (DOI) determined during data inversion was limited to 150 m [*Christiansen and Auken*, 2012]. To estimate the resistivity of the area between flight lines, known resistivity values were interpolated using a natural neighbor method over a ~40 × 40 m *x-y* grid bounded by the flight lines at a 1 m depth interval.

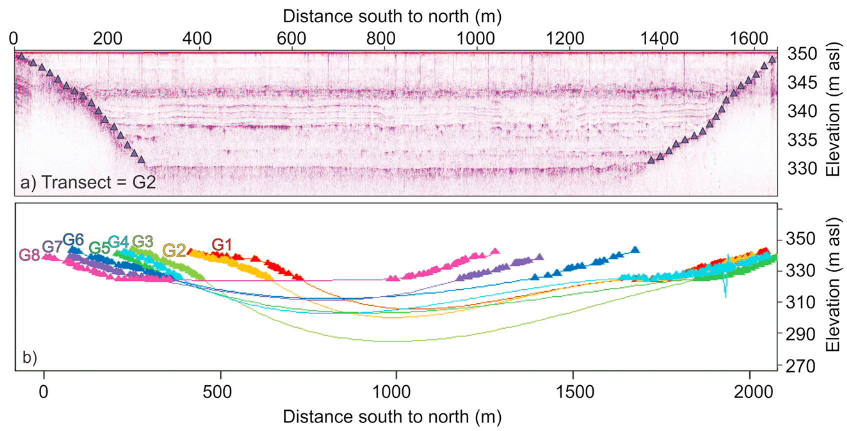

Figure 2. (a) GPR profile across the middle of Lake Vida (location of G2 shown in Figure 1). The ice/basin contact, marked by triangles, was digitized for all transects. At 330 m asl, the basin contact is lost when the GPR cannot penetrate a strong horizontal reflector. (b) Cubic spline interpolation across the eight transects where the GPR did not return the ice/basin contact below 330 m asl.

4. Results

The GPR data show the lakebed where ice is less than ~20 m thick (Figure 2a), providing bathymetry for 69% of the area (Figure 1). For the eight north-south GPR transects, which crossed the center of the lake, spline fitting returned maximum depths of 40, 41, 60, 42, 42, 33, 34, and 21 m from west to east (Figure 2b).

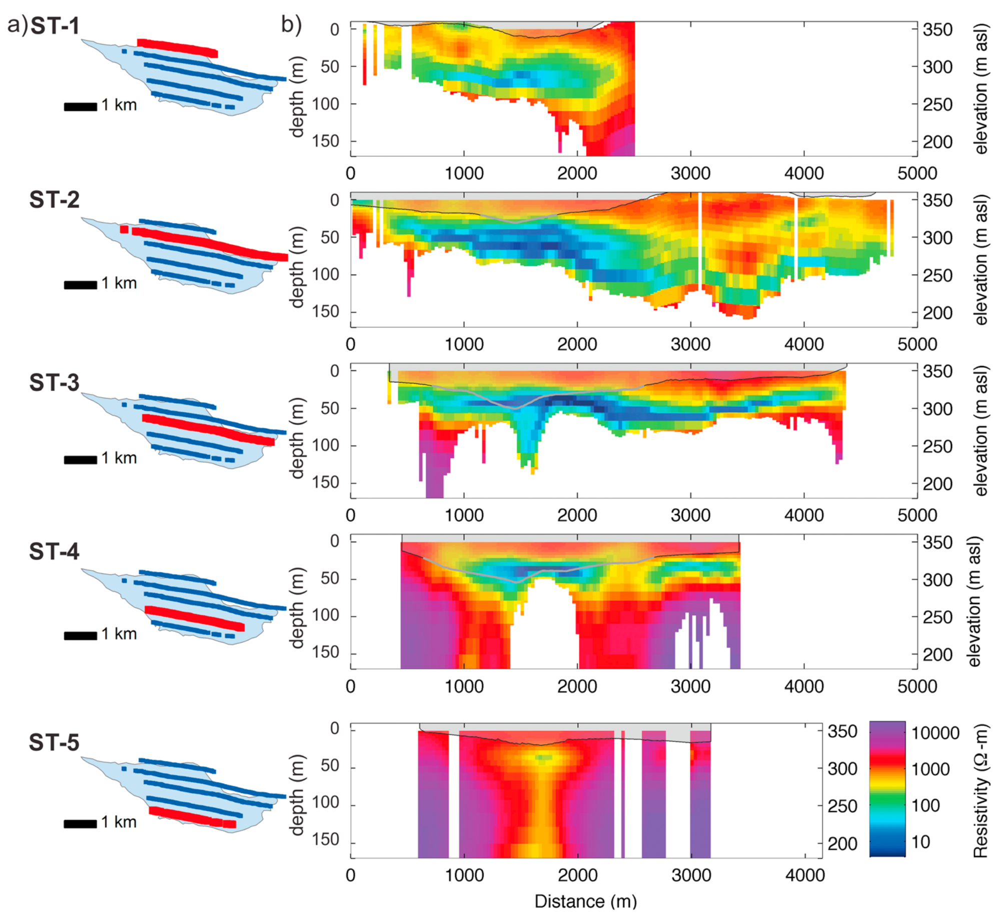

Figure 3. (a) Location of AEM flight lines on Lake Vida. (b) Resistivity cross sections returned from the AEM flight lines over Lake Vida from west to east, referenced to a common easting. The black and grey lines superimposed on the profiles are known and estimated bathymetric depths, respectively. The area above this line represents the lake ice.

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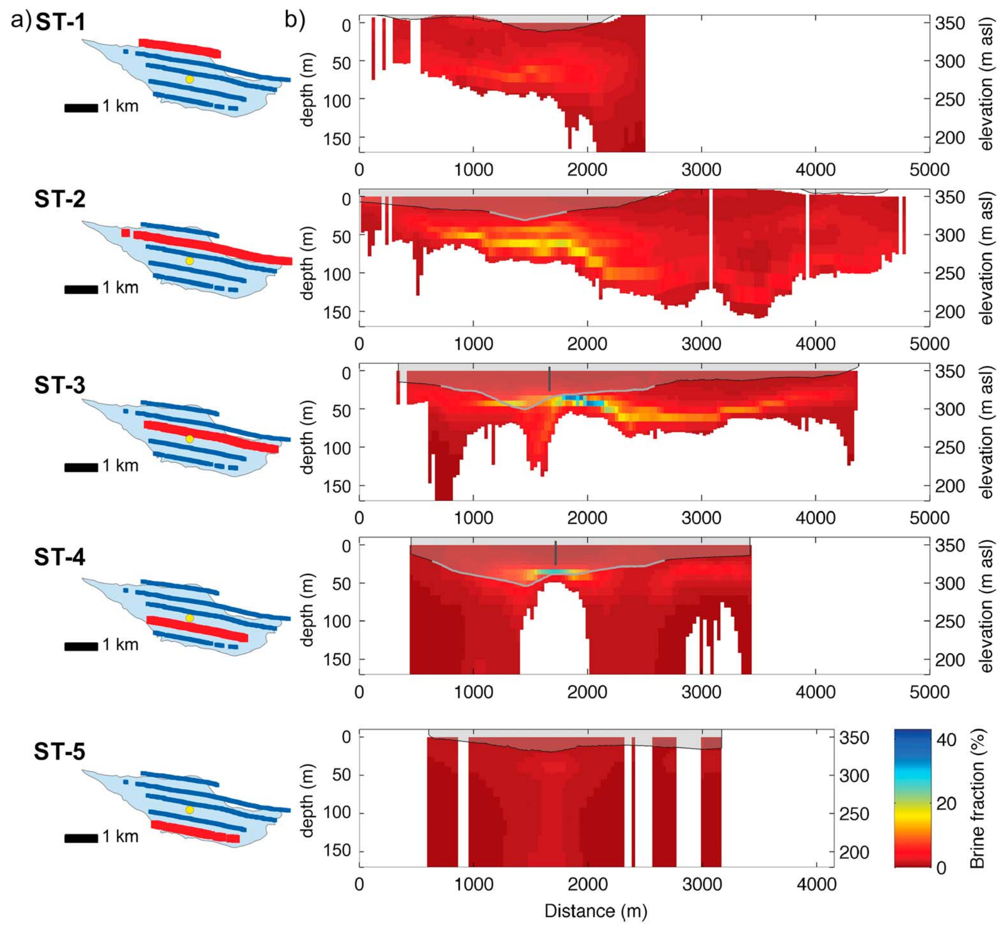

Figure 4. (a) Location of AEM flight lines on Lake Vida. The 2010 Lake Vida drill site is shown as a yellow dot between ST3 and ST4. (b) Brine fraction estimated from Archie's law with a cementation factor of $m = 2$. The black and grey lines superimposed on the profiles are known and estimated bathymetric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

Resistivity values obtained from the AEM survey range from 1.3 to 20,000 Ω m. The minimum in the first 20 m beneath the surface is 84 Ω m. The largest zone of low resistivity lies 30 to 100 m beneath the center of Lake Vida (Figure 3), where values approach 1.3 Ω m at 33–47 m depth. The low resistivities begin to dissipate near the north and south edges of the lake. Flight line ST5 did not register any values below 250 Ω m. It must be noted that weakening resistivity at the upper or lower corners of a strong brine signal may be exaggerated as the inversion is based on local 1-D models [*Goldman et al*., 1994; *Auken et al*., 2008]. All AEM flight lines were completely within the footprint of Lake Vida, with the exception of the edges of ST1 and the eastern half of ST2. Based on our estimated bathymetry, the low-resistivity zone resides immediately below the modeled lake bottom in ST1, ST2, and ST3. Only in ST4 is low overall resistivity found above the lake bottom (Figure 3).

5. Discussion

Natural freshwaters have variable resistivities, 5–100 Ω m, depending largely on solute content (Table 1). Typical seawater at its freezing point has resistivity slightly higher than 0.3 Ω m. When water freezes, the resistivity of the remaining liquid gradually decreases, while the resulting ice has high resistivity. Insulating materials, such as unweathered rock and ice, have resistivity values $>$1000 Ω m and upward of 100,000 Ω m (Table 1). The only geologic deposits with resistivities $<$10 Ω m are extremely weathered rocks, clays, or economical mining targets, such as sulfide deposits [*Palacky*, 1988].

Table 1. Resistivity of Common Substrate and Surficial Waters in the McMurdo Dry Valleys

Substrate	Resistivity (Ω m)
Lake Vanda (Wright Valley) surface water[a]	111–180
Lake Vanda ice 0.1 m[a]	19,500
Lake Vanda ice 0.9 m[a]	18,690
Wright Valley unsorted soils[a]	90–413
Beacon sandstone[b]	103,800
Granite Gneiss (21°C) [b]	100,360
Diorite (21°C/−25°C) [b]	102,200/452,000
Diorite (21°C/−25°C) [b]	33,100/27,700
Seawater (20°C)	0.20
Seawater (−1.9°C)	0.36
Lake Vida brine[c]	0.15

[a]Field measurements with Wheatstone bridge [*McGinnis and Jensen*, 1971].
[b]Laboratory measurements [*McGinnis and Jensen*, 1971].
[c]Inverse of brine conductivity measured in 2010.

Beneath Lake Vida, the presence of brine in the subsurface is revealed by low overall resistivity, as there is no other conductive media. We assume the brine beneath Lake Vida has a resistivity (ρ_b) equal to 0.15 Ω m, determined from the conductivity (6.5 S m^{-1}) measured at the in situ temperature of −13°C [*Murray et al.*, 2012]. The resistivity minimum of 1.3 Ω m indicates that only a fraction of the subsurface volume consists of the highly concentrated 0.15 Ω m brine observed in the drill holes. We do not consider it plausible that 1.3 Ω m signifies open water, as 1.3 Ω m water would have an in situ salinity half that of seawater and could not underlie the observed brine without inducing an unstable density stratification.

The volume fraction of brine in each bin can be estimated using a simplified Archie's law [*Archie*, 1950]:

$$\varphi = (\rho_o/\rho_b)^{(-1/m)} \tag{1}$$

where φ is liquid brine fraction, ρ_o is the bulk resistivity determined from AEM, ρ_b is the resistivity of the brine, and m is the cementation factor. All pore space is occupied either by unfrozen brine or by ice (>10,000 Ω m). Cementation factors for permafrost are unknown, but published values for other substrates typically range from 1.4 (highly fractured rock) to 2.8 (rock with spherical pores or increased tortuosity). To account for uncertainties in the cementation factor, we performed calculations with $m = 1.5$, 2, and 2.5. Maximum brine fraction ranges between 23 and 42% (Figure 4). The total brine volume beneath Lake Vida is 5.3×10^6 m^3, 1.58×10^7 m^3, and 3.23×10^7 m^3, when $m = 1.5$, 2, and 2.5, respectively. From our modeled bathymetry and resistivity values alone, it is difficult to distinguish if the brine exists in a network of bedrock, ice, or permafrost, as these are all highly insulating materials. Here we examine each of these possibilities in detail.

5.1. Brine in Bedrock

The calculated maximum brine fraction of 23–42% beneath Lake Vida indicates porosity much higher than found in massive biotite gneiss documented in the near surface of the DVDP-6 core. Therefore, we infer that the valley continues to have steep relief beneath Lake Vida and the low-resistivity anomaly is situated above local bedrock. We cannot rule out that brine may be present in the cracks of the bedrock. At 1% porosity, the overall resistivity would equal 1500 Ω m ($m = 2$), which is well within the range returned by the AEM sensor. In DVDP-6, thermal fluctuations were attributed to groundwater intrusion into the borehole at 230–250 m depth (100–120 m asl) in an area of massive biotite-granite [*Cartwright and Harris*, 1981]. In most areas covered by the AEM survey, the depth of investigation was less than 150 m, and brine at 230–250 m cannot be confirmed.

5.2. Brine in Lake Ice

The GPR performed on Lake Vida was unable to penetrate beneath a reflector in the lake at 20 m depth (Figure 2a). Initially assumed to be the top of a liquid brine body [*Doran et al.*, 2003], we believe that this reflector represents a thick sediment layer present in the lower ice. Below 20 m, our extrapolated bathymetry reaches a maximum depth of 60 m (transect G3; Figures 1 and 2). All other extrapolated GPR transects do not attain depths greater than 42 m. Based on these constraints, Lake Vida is likely 40–45 m deep. If lake ice does not extend below 45 m, the majority of brine is below the lake bottom, and therefore, the brine sampled in the Lake Vida drill holes represents only a small fraction of the total brine volume in the basin. However, considering that the lowest resistivity was recorded in ST3 at 33–47 m depth, we cannot resolve whether the highest concentration of brine exists in an ice matrix at the very bottom of the lake or within lake sediments.

5.3. Brine in Unconsolidated Sediments

Our calculations indicate that the brine network is focused between 30 and 100 m depth (250 to 320 m asl) and largely within the footprint of the lake surface. Our preferred scenario is that below 350 m asl (2010 lake level), the deep valley bottom is filled with porous unconsolidated sediments. The calculated brine fractions are comparable to measured soil porosities of 30–36% [*Levy et al*., 2011] in the MDV.

5.4. Confined Aquifer Beneath Lake Vida

We propose that the brine within the lower ice and sediments is isolated from the atmosphere based on the following:

1. The brine should not be pressurized if an atmospheric connection exists.
2. There is 16 m of freshwater ice on Lake Vida that is not penetrated by the pressurized brine below.
3. AEM data do not show any low resistivity ($<100\ \Omega$ m) in the upper 15 m. At shallow depths, the resolution of the instrument is sufficiently high that we would expect to see a signal from concentrated brine seepage from the near surface into the lake.
4. DVDP borehole data show that temperatures are <20°C in the upper 100 m of the subsurface [*Decker and Bucher*, 1982]. At these temperatures, warm brines from the active layer would freeze.
5. There is no evidence that saline water in the active layer are able to penetrate below the ice table.
6. Tritium levels in the brine are 0.5 ± 0.9 tritium unit (P. T. Doran, unpublished data).

Based on these observations, we proposed that the brine beneath Lake Vida is analogous to a confined aquifer.

This aquifer, which we define as the entire low-resistivity zone beneath Lake Vida, contains between 15×10^6 m^3 and 32×10^6 m^3 of brine. In comparison to Taylor Valley lakes, this is roughly 2–5 times the volume of liquid beneath the chemocline in brackish Lake Fryxell (6.7×10^6 m^3) and 1–2 times that beneath the chemocline in hypersaline east Lake Bonney (1.9×10^7 m^3), exclusive of sediments. The source of the Lake Vida brine must stem from either external input or evaporation/freeze concentration of lake water. Since Victoria Valley has not been inundated by the ocean since the Miocene [*Chinn*, 1993], and there is no evidence of extensive groundwater connectivity beyond the lake, the majority of the brine presumably formed from the concentration of lake water that permeated the sediments. Research in Taylor Valley supports repeated evaporation events in both Lake Fryxell and Lake Bonney [*Lyons et al*., 1998a; *Poreda et al*., 2004; *Whittaker et al*., 2008], and repeated lake level fluctuations have been documented in the Vida basin [*Hall et al*., 2010].

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

Two further observations support the idea that the Vida aquifer has cooled and concentrated over time, similar to ancient aquifers on Mars. First, the unstructured shape of the aquifer suggests heterogeneous freezing of the surrounding permafrost inward [*Gaidos and Marion*, 2003]. Second, in aquifers confined by basement rock and frozen sediments, the volumetric expansion of ice during freezing can increase the hydrostatic pressure [*Gaidos*, 2001; *Gaidos and Marion*, 2003]. The pressurized brine entering the drill holes from 2005 and 2010 corroborates an increase in hydrostatic pressure, since at the time of ice formation, the potentiometric level of the brine could not have been higher than the surface of the lake, which was at least 27 m below the current surface. Furthermore, the inferred absence of a free brine body even in the deepest part of the lake basin negates the previous interpretation that the potentiometric level results from a partial floatation of the ice cover [*Doran et al*., 2003].

6. Conclusions

Data collected using an AEM survey allowed us to define the limits of a confined aquifer beneath Lake Vida and estimate the porosity at 23–42% and the brine volume at 15 to 32 million m^3. The lack of a strong hydrologic connection beyond the edges of the lake to a larger groundwater system in Victoria Valley is an important insight into the regional hydrogeology, and the legacy of past shifts in climate on the present-day ecosystem [*Moorhead et al*., 1999]. Our results:

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

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

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References

Archie, G. (1950), Introduction to petrophysics of reservoir rocks, *Am. Assoc. Pet. Geol. Bull.*, *34*(5), 943–961.

Auken, E., and A. V. Christiansen (2004), Layered and laterally constrained 2-D inversion of resistivity data, *Geophysics*, *69*, 752–761, doi:10.1190/1.1759461.

Auken, E., A. V. Christiansen, L. H. Jacobsen, and K. I. Sørensen (2008), A resolution study of buried valleys using laterally constrained inversion of TEM data, *J. Appl. Geophys.*, *65*, 10–20, doi:10.1016/j.jappgeo.2008.03.003.

Bockheim, J. G., I. B. Campbell, and M. McLeod (2007), Permafrost distribution and active-layer depths in the McMurdo Dry Valleys, Antarctica, *Permafrost Periglac.*, *18*, 217–227, doi:10.1002/ppp.588.

Cartwright, K., and H. J. H. Harris (1981), Hydrogeology of the Dry Valley region, Antarctica, in *Dry Valley Drilling Project. Antarctic Research Series 33*, vol. 33, edited by L. D. McGinnis, pp. 193–214, AGU, Washington D. C.

Chinn, T. J. (1993), Physical Hydrology of the Dry Valley Lakes, in *Physical and Biogeochemical Process in Antarctic Lakes, Antarctic Res. Ser.*, vol. 59, edited by W. Green and E. I. Friedmann, pp. 1–52, AGU, Washington, D. C.

Christiansen, A. V., and E. Auken (2012), A global measure for depth of investigation, *Geophysics*, *77*, 171–177.

Decker, E. R., and G. J. Bucher (1982), Geothermal Studies in the Ross Island-Dry Valley Region, in *Antarctic Geoscience*, vol. 1, edited by C. Craddock, pp. 887–894, Univ. of Wisconsin Press, Madison, Wis.

Dickinson, W. W., and M. R. Rosen (2003), Antarctic permafrost: An analogue for water and diagenetic minerals on Mars, *Geology*, *31*(3), doi:10.1130/0091-7613(2003)031<0199:APAAFW>.

Dickson, J. L., J. W. Head, J. S. Levy, and D. R. Marchant (2013), Don Juan Pond, Antarctica: Near-surface CaCl2-brine feeding Earth's most saline lake and implications for Mars, *Sci. Rep.*, *3*, 1166, doi:10.1038/srep01166.

Doran, P. T., C. P. McKay, G. D. Clow, G. L. Dana, A. G. Fountain, T. Nylen, and W. B. Lyons (2002), Valley floor climate observations from the McMurdo Dry Valleys, Antarctica, 1986–2000, *J. Geophys. Res.*, *107*(D24), 4772, doi:10.1029/2001JD002045.

Doran, P. T., C. H. Fritsen, C. P. McKay, J. C. Priscu, and E. E. Adams (2003), Formation and character of an ancient 19 m ice cover and underlying trapped brine in an "ice-sealed" east Antarctic lake, *Proc. Natl. Acad. Sci. U.S.A.*, *100*(1), 26–31, doi:10.1073/pnas.222680999.

Gaidos, E., and G. M. Marion (2003), Geological and geochemical legacy of a cold early Mars, *J. Geophys. Res.*, *108*(E6), 5055, doi:10.1029/2002JE002000.

Gaidos, E. J. (2001), Cryovolcanism and the recent flow of liquid water on Mars, *Icarus*, *153*, 218–223.

Goldman, M., L. Tabarovsky, and M. Rabinovich (1994), On the influence of 3-D structures in the interpretation of transient electromagnetic sounding data, *Geophysics*, *59*(6), 889–901, doi:10.1190/1.1443648.

Gooseff, M. N., J. E. Barrett, and J. S. Levy (2013), Shallow groundwater systems in a polar desert, McMurdo Dry Valleys, Antarctica, *Hydrogeol. J.*, *21*, 171–183, doi:10.1007/s10040-012-0926-3.

Hall, B. L., G. H. Denton, A. G. Fountain, C. H. Hendy, and G. M. Henderson (2010), Antarctic lakes suggest millennial reorganizations of Southern Hemisphere atmospheric and oceanic circulation, *Proc. Natl. Acad. Sci. U.S.A.*, *107*, 21,355–21,359, doi:10.1073/pnas.1007250107.

Head, J. W., and D. R. Marchant (2014), The climate history of early Mars: Insights from the Antarctic McMurdo Dry Valleys hydrologic system, *Antarct. Sci.*, *26*, 774–800, doi:10.1017/S0954102014000686.

Kurasawa, H., Y. Yoshida, and M. G. Mudrey Jr. (1974), Geologic log of the Lake Vida core—DVDP 6, *Dry Val. Drill.Proj. Bull.*, *3*, 92.

Levy, J. S., A. G. Fountain, M. N. Gooseff, K. A. Welch, and W. B. Lyons (2011), Water tracks and permafrost in Taylor Valley, Antarctica: Extensive and shallow groundwater connectivity in a cold desert ecosystem, *Geol. Soc. Am. Bull.*, *123*, 2295–2311, doi:10.1130/B30436.1.

Lyons, W. B., S. W. Tyler, R. A. Wharton Jr., D. M. McKnight, and B. H. Vaughn (1998a), A late Holocene desiccation of Lake Hoare and Lake Fryxell, McMurdo Dry Valleys, Antarctica, *Antarct. Sci.*, *10*(3), 247–256.

Lyons, W. B., K. A. Welch, and P. Sharma (1998b), Chlorine-36 in the waters of the McMurdo Dry Valley Lakes, Southern Victoria Land, Antarctica: Revisited, *Geochim. Cosmochim. Acta*, *62*(2), 185–191.

Lyons, W. B., K. a. Welch, G. Snyder, J. Olesik, E. Y. Graham, G. M. Marion, and R. J. Poreda (2005), Halogen geochemistry of the McMurdo Dry Valleys lakes, Antarctica: Clues to the origin of solutes and lake evolution, *Geochim. Cosmochim. Acta*, *69*, 305–323, doi:10.1016/j.gca.2004.06.040.

McGinnis, L. D., and T. E. Jensen (1971), Permafrost-hydrogeologic regimen in two ice-free valleys, Antarctica, from electrical depth sounding, *Quat. Res.*, *1*, 389–409.

Moorhead, D. L., P. T. Doran, A. G. Fountain, W. B. Lyons, D. M. Mcknight, J. C. Priscu, R. A. Virginia, and D. H. Wall (1999), Ecological legacies: Impacts on ecosystems of the McMurdo Dry Valleys, *BioScience*, *49*, 1009–1019.

Murray, A. E., et al. (2012), Microbial life at −13 °C in the brine of an ice-sealed Antarctic lake, *Proc. Natl. Acad. Sci. U.S.A.*, *109*, 20,626–20,631, doi:10.1073/pnas.1208607109.

Palacky, G. (1988), Resistivity characteristics of geologic targets, in *Electromagnetic Methods in Applied Geophysics*, edited by M. Nabighian, pp. 513, Society of Exploration Geophysicists, Tulsa, Okla.

Poreda, R. J., A. G. Hunt, W. B. Lyons, and K. A. Welch (2004), The helium isotopic chemistry of Lake Bonney, Taylor Valley, Antarctica: Timing of late Holocene climate change in Antarctica, *Aquat. Geochem.*, *10*, 353–371, doi:10.1007/s10498-004-2265-z.

R Core Team (2012), R: A language and environment for statistical computing, Vienna, Austria: R Foundation for Statistical Computing; 2012. [Available at http://cran.r-project.org.]

Seyfried, M. S., and M. D. Murdock (1997), Use of air permeability to estimate infiltrability of frozen soil, *J. Hydrol.*, *202*(1), 95–107.

Sørensen, K. I., and E. Auken (2004), SkyTEM—a new high-resolution helicopter transient electromagnetic system, *Explor. Geophys.*, *35*, 194–202.

Viezzoli, A., A. V. Christiansen, E. Auken, and K. I. Sørensen (2008), Quasi-3D modeling of airborne TEM data by spatially constrained inversion, *Geophysics*, *73*, F105–F113, doi:10.1190/1.2895521.

Whittaker, T. E., B. L. Hall, C. H. Hendy, and S. A. Spaulding (2008), Holocene depositional environments and surface-level changes at Lake Fryxell, Antarctica, *Holocene*, *18*, 775–786, doi:10.1177/0959683608091797.

Witherow, R. A., W. B. Lyons, and G. M. Henderson (2010), Lithium isotopic composition of the McMurdo Dry Valleys aquatic systems, *Chem. Geol.*, *275*, 139–147.