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Key Points:

- Lake Vostok is not expected to discharge liquid water in climatic timescales
- Discharge would lead from east of the lake's southern tip to Ross Ice Shelf
- Lake Vostok would be significantly deeper and larger without ice sheet

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Subglacial Lake Vostok not expected to discharge water

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Abstract The question whether Antarctica's largest lake, subglacial Lake Vostok, exchanges water is of interdisciplinary relevance but has been undecided so far. We present the potential pathway, outlet location, and threshold height of subglacial water discharge from this lake based on a quantitative evaluation of the fluid potential. If water left Lake Vostok, it would flow toward Ross Ice Shelf. Discharge would occur first to the east of the southern tip of the lake. At this location the bedrock threshold is 91 ± 23 m higher than the hydrostatic equipotential level of Lake Vostok. It is concluded that Lake Vostok is not likely to reach this level within climatic timescales and that no discharge of liquid water is to be expected. We show that in absence of the ice sheet the Lake Vostok depression would harbor a lake significantly deeper and larger than the present aquifer.

1. Introduction

Among the subglacial lakes discovered in Antarctica, more than a hundred have been shown to change their volume as a result of their active participation in subglacial water transport [*Siegert et al*., 2005; *Smith et al*., 2009; *Pattyn*, 2011]. Whether the largest subglacial lake on Earth, Lake Vostok in central East Antarctica [*Kapitsa et al*., 1996] (Figures 1a and 1b), is also involved in subglacial exchange of liquid water is an important question for several reasons:

First, given its large size (according to the shoreline by *Popov and Chernoglazov* [2011]: 16,250 km^2 including 370 km^2 islands) and volume (according to the bathymetry model by *Popov et al*. [2011]: 6,250 km^3), an active Lake Vostok would be capable of an enormous impact on the subglacial hydrological network, in particular in the case of a rapid, possibly catastrophic outburst. A subglacial flooding in the East Antarctic interior, in turn, may affect large-scale ice flow dynamics [*Bell et al*., 2007] and thus the continental ice-mass balance. Second, subglacial inflow or discharge of liquid water would affect substantially the isotope and chemical budget of Lake Vostok's aquifer. If that occurred, it would have to be accounted for in the inference of the physical, chemical, and limnological conditions and the assessment of viable life in the lake from the analysis of the accreted lake ice retrieved by drilling at Vostok station. For example, the variation in isotope composition observed in the accreted ice has recently been explained based on the assumption that exchange and renewal of water is restricted to melting and accretion at the ice-water interface [*Ekaykin et al*., 2010]. Third, evidence for or against regular water exchange would provide constraints for the explanation of the formation of Lake Vostok [*Duxbury et al*., 2001]. Finally, the proposed use of the ice surface at Lake Vostok as a calibration area for satellite laser altimetry data [*Ewert et al*., 2012; *Shuman et al*., 2006] assumes the stability of the ice surface height and would thus be impaired by water volume fluctuations.

Geodetic monitoring of the ice surface height suggests that Lake Vostok has not experienced significant water volume changes over the last years (i.e., lake level changes do not exceed a few centimeters over the last decade; *Richter et al*. [2008]; *Ewert et al*. [2012]). However, the relatively short period covered by the observations together with the uncertainties inherent to the measurements limit the detectability of small water volume fluctuations (i.e., fractions of mm/yr). Water inflow into the lake, if present, would likely be slow and steady. In view of the lake's huge area, an afflux of an implausibly large water volume would be needed to produce a measurable surface height change over a short time. In contrast, a sudden, substantial discharge of lake water might seem more feasible.

Making use of the theory of a captured ice sheet and the fluid potential, *Erlingsson* [2006] proposed an approach to assess the potential of Lake Vostok for subglacial water discharge. That work concluded that

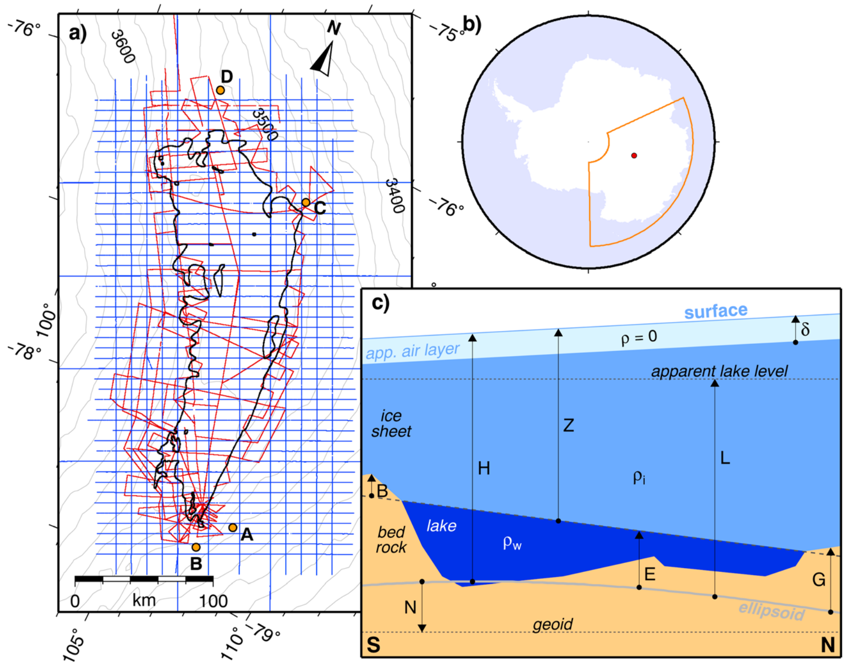

Figure 1. Geographical setting of subglacial Lake Vostok. (a) Map of the Lake Vostok region. Black line: lake shore according to *Popov and Chernoglazov* [2011]; red lines: ice thickness profiles derived from ground-based RES [*Popov et al*., 2012]; blue lines: ice thickness profiles derived from airborne RES [*Studinger et al*., 2003]; orange dots: potential discharge outlets; and contours: ellipsoidal ice surface elevation (with respect to WGS84) from the BEDMAP2 DEM [*Fretwell et al*., 2013]. (b) Overview over Antarctica. Red dot: location of subglacial Lake Vostok and orange outline: area depicted in Figure 2b. (c) Schematic south-north cross section of the Lake Vostok system showing the geometric relationship between the quantities referred to in the text.

Lake Vostok might be close to discharge and that this could be a common feature during interglacials. However, the scarcity of input data at hand at that time limited the significance of these results. Since then, satellite missions (ICESat, GOCE), dedicated fieldwork [*Popov et al*., 2012; *Richter et al*., 2008], and modeling efforts [*Salamatin et al*., 2009; *Ewert et al*., 2012; *Schwabe et al*., 2014] have substantially improved this situation.

2. Methods and Data

Water flow at the surface of the solid Earth is directed along the gradient from high to low fluid potential. In the case of subglacial water flow, the fluid potential Φ is composed of the bedrock elevation z with respect to an arbitrary datum and the pressure head of the overlying ice sheet of thickness Z:

$$\Phi = gz + \frac{\rho_i}{\rho_w} gZ \tag{1}$$

with g being the gravity acceleration and ρ_i and ρ_w the densities of the ice and the lake water, respectively (Figure 1c). Subglacial Lake Vostok represents a local minimum of fluid potential. As reference surface for the bedrock elevation we adopt the hydrostatic equipotential level at the lake water table, with a height

$$E = L - \frac{\rho_i}{\rho_w - \rho_i}(H - L) \tag{2}$$

that shall be zero all over the lake surface (Figure 1c). Here H is the ellipsoidal height of the ice surface and L is the ellipsoidal height of the "apparent lake level" [*Ewert et al*., 2012] ("floating level" according to *Erlingsson* [2006]). This apparent lake level represents the equilibrium level about which the floating ice sheet is hydrostatically balanced and coincides with the fictive lake water level, if the ice above the lake was melted. According to *Ewert et al*. [2012, equations (9) and (10)] (dedicated to the evaluation of the fulfillment of the hydrostatic equilibrium), the apparent lake level can be considered as composed of the geoid height N, a constant offset b, and the thickness of an apparent air layer δ [*Horwath et al*., 2006]:

$$L = N + b - \frac{\rho_i}{\rho_w}\delta. \tag{3}$$

The apparent air layer (of thickness δ) accounts for the deviation of the vertically integrated ice density from the density of pure ice ρ_i. If the vertical density profile is uniform over the lake surface (as we assume), then δ is constant. According to *Horwath et al*. [2006] and *Ewert et al*. [2012], we adopt the densities of 1016 kg m^{-3} and 917 kg m^{-3} for the lake water and pure ice, respectively.

Water can leave a subglacial lake only if and where the bedrock elevation is below the hydrostatic equipotential level E. The bedrock elevation, G, is determined relative to the ellipsoid by subtraction of the observed ice thickness from the observed ellipsoidal surface elevation: $G = H - Z$. If the ellipsoid was chosen as datum for z in equation (1), G and z would be identical. Outside the lake, the comparison of the bedrock elevation G against E results in the relative hydrostatic bedrock height B. Expressing the bedrock elevation in terms of observable quantities (H, Z), and subsequently, converting H and L to orthometric heights by subtracting N, the hydrostatic bedrock height, B, becomes

$$B = G - E = H - Z - L + \frac{\rho_i}{\rho_w - \rho_i}(H - L) = \frac{\rho_w}{\rho_w - \rho_i}(H - N - b) - Z. \tag{4}$$

In a first step, the offset b is determined within the portion of the ice sheet floating in hydrostatic equilibrium above the lake σ_L:

$$b = \frac{1}{\sigma_L}\int_{\sigma_L} H - N + Z\left(\frac{\rho_i}{\rho_w} - 1\right) d\sigma_L. \tag{5}$$

Close to the grounding line bounding the lake and its islands the ice surface deviates from the hydrostatic equilibrium position [*Ewert et al*., 2012]. Further deviations are caused by local inconsistencies between the grounding line location as detected along radio echo sounding (RES) profiles [*Popov and Chernoglazov*, 2011] and the ice thickness and surface elevation data with their limited spatial resolution. Therefore, in order to ensure that only hydrostatically equilibrated parts of the ice surface are considered in the offset determination, the integral in equation (5) is taken over the central test area defined in *Ewert et al*. [2012] (Figure 2a). This yields an offset b of 3127.66 ± 0.39 m, very close to the 3129.69 m obtained by *Ewert et al*. [2012] and with some variability depending on the location (i.e., distance from grounding line). In a second step, equation (4) is applied to derive the relative hydrostatic bedrock heights B throughout the area under investigation.

Three ingredients are needed to determine the spatial variation in fluid potential at the ice-bedrock interface (equation (1)) and to assess the water discharge of subglacial Lake Vostok (equation (4)): a model of the ice surface elevation H (digital elevation model, DEM), a geoid model, and an ice thickness model. For a large part of East Antarctica a map of the fluid potential (Figure 2b) is derived from the BEDMAP-2 [*Fretwell et al*., 2013] ice thickness and surface height data sets and the underlying geoid model EIGEN-GL04C [*Foerste et al*., 2008]. Based on this map the pathways of potential subglacial water flow from Lake Vostok down to the Antarctic coast are traced following the downward gradient.

Within a map of the hydrostatic bedrock height B, the lowest saddle point around the periphery of the lake marks the location and threshold height for water drainage. For a detailed determination of the hydrostatic bedrock height around Lake Vostok, new regional models are brought together with higher resolution and quality than the continental-scale data sets. We have generated a new DEM by combining a model based on crossover-adjusted ICESat satellite laser altimetry data [*Ewert et al*., 2012] with a DEM inferred from ERS-1 satellite radar altimetry [*Roemer et al*., 2007]. The applied geoid model [*Schwabe et al*., 2014] combines local airborne gravity [*Studinger et al*., 2003] and GOCE satellite data [*Bruinsma et al*., 2013]. Our ice thickness model is derived from ground-based [*Popov et al*., 2012] and airborne [*Studinger et al*., 2003] RES data (Figure 1a). It results from a common crossover adjustment of both data sets and is tied to the exactly known ice thickness at the Vostok drilling site [*Lipenkov et al*., 2012].

The accuracy of the obtained hydrostatic bedrock heights is estimated by propagation of the uncertainties of the input quantities. Systematic effects affecting the model domain uniformly (e.g. biases in the ice

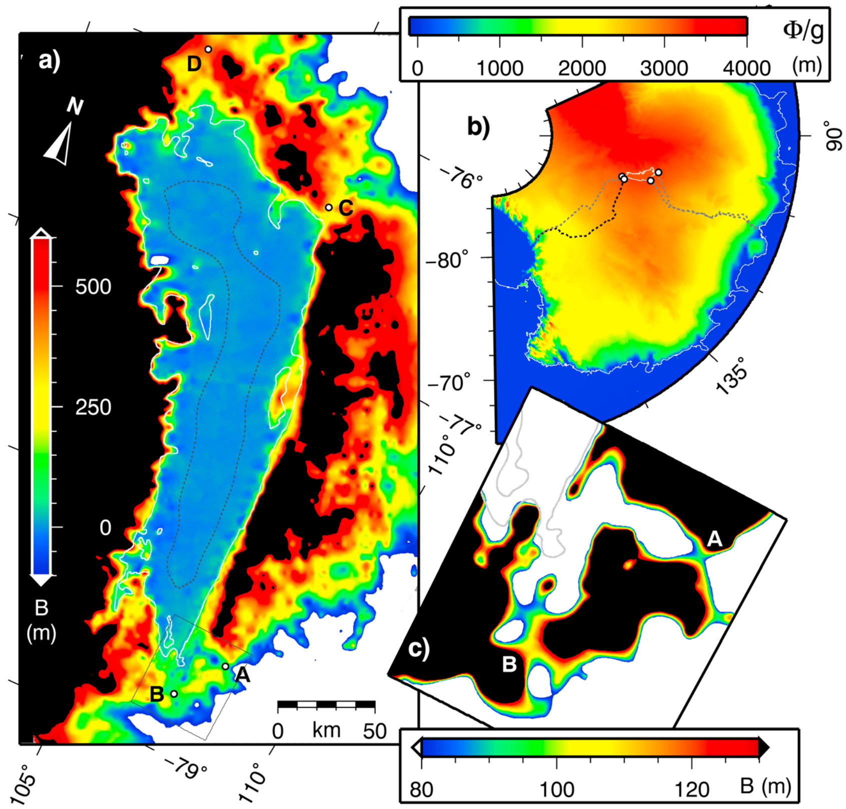

Figure 2. Maps of fluid potential and threshold height for subglacial Lake Vostok. (a) Hydrostatic bedrock height *B* relative to the hydrostatic equipotential level of Lake Vostok. Areas with hydrostatic bedrock heights exceeding 700 m are shown in black. Circles: potential discharge outlets; white line: lake shore; dashed grey line: central test area for the determination of the apparent lake level; and grey box: area shown in inset c. (b) Fluid potential over a sector of East Antarctica based on BEDMAP2 data sets [*Fretwell et al*., 2013]. Potential discharge pathways are shown for outlet A (dashed black line) and outlets B–D (dashed grey lines). (c) As Figure 2a, zoom into the area of outlets A and B (note different color scales in Figures 2a and 2c). Areas with hydrostatic bedrock heights exceeding 140 m are shown in black. Grey line: the lake shore.

thickness, surface elevation and geoid height as well as in the densities of water and ice) are largely accommodated by the adjustment of offset *b*. The uncertainty budget is dominated by the contribution of the ice thickness data, the spatially most variable and most unevenly sampled of the introduced quantities. The crossover differences of the adjusted ice thickness profiles yield a standard deviation of 10 m for a single ice thickness value outside the lake boundary (3.5 m within the lake area). The uncertainty of the gridded ice thickness values increases with increasing distance from the RES profiles and is therefore not uniform throughout the model domain. Based on the output of the applied kriging algorithm, we adopt a mean uncertainty of 20 m for a single ice thickness grid value. The uncertainty of the regional geoid model amounts to 5 cm [*Schwabe et al*., 2014]. For the ICESat-only ice surface DEM an uncertainty of 2.1 m is reported [*Ewert et al*., 2012]. The larger part of this value results from the interpolation between the relatively sparse ICESat ground tracks. The applied DEM is much less prone to interpolation errors since the meshes between the ICESat tracks are filled with ERS-1 radar altimetry data. The bias between laser-based and radar-based surface heights has been considered in the way that the ERS-1 model was tied to the ICESat model. We estimate an uncertainty of 1 m for a single orthometric height value, that is, including the contribution of the geoid height. The standard deviation of the *b* value obtained by equation (5) of 39 cm is considered as additional source of uncertainty. This results in a total uncertainty of 23 m for the relative hydrostatic bedrock heights.

3. Results and Discussion

The map of the relative hydrostatic bedrock heights around Lake Vostok (Figure 2a) depicts large (>600 m) bedrock elevations to the west of the lake. The eastern shore of the lake is flanked by an elongated, high

Table 1. Potential Outlet Locations for Subglacial Water Discharge From Lake Vostok[a]

	ϕ	λ	H (m)	Z (m)	N (m)	d (km)	B (m)
A	78°30.3′	108°06.0′	3430	3210	−19.2	3.2	91
B	78°43.8′	107°19.6′	3441	3306	−19.2	2.5	109
C	76°24.0′	105°39.0′	3495	3593	−15.5	0.8	340
D	75°57.8′	102°17.3′	3519	3694	−11.9	5.3	446

[a] ϕ: latitude S; λ: longitude E; H: ellipsoidal ice surface elevation (WGS84); Z: ice thickness; N: geoid height; d: distance to the nearest ice thickness profile; and B: hydrostatic bedrock height relative to the hydrostatic equipotential level of Lake Vostok (threshold height).

ridge. Beyond this ridge the bedrock drops steeply below Lake Vostok's hydrostatic equipotential level. As expected, B values around zero are obtained for the area of the lake. The mean value amounts to -0.04 ± 4.04 m within the central test area, and to $+2.8 \pm 15.2$ m over 90% of the lake surface. The recovered standard deviations prove that our accuracy estimate is conservative. On this map, four saddle points A–D are identified around the lake's periphery which indicate potential outlets for water discharge. The quantitative results for these locations are summarized in Table 1. The violation of the hydrostatic equilibrium along the grounding line may affect the determination of the hydrostatic equipotential level E. This would lead to a uniform offset of all the hydrostatic bedrock heights B, thus changing slightly the threshold heights, but not the outlet locations. Our restriction to the central test area in the determination of E should minimize this effect.

Subglacial water discharge from Lake Vostok is closest at location A, to the east of the southern tip of the lake. There, the threshold is 91 m above the lake's hydrostatic equipotential level. The second potential outflow, B, is situated to the south of the lake's southern tip, about 40 km southwest from A and 18 m higher than A. The thresholds at locations C (east of the northern part of the lake) and D (north of the northern shore) are much higher (340 and 446 m, respectively). Moreover, the obtained fluid potential and hydrostatic bedrock heights indicate, that Lake Vostok is not a "captured lake" [*Erlingsson*, 2006] and, thus, not prone to a jökulhlaup. Even if the lake level rose above the threshold (91 m), only the amount of water exceeding the threshold height would drain through the outlet. If water left the lake through one of the southern outlets A or B, it would reach the ocean beneath Ross Ice Shelf (Figure 2b). Despite the proximity of both outlets in location and threshold height, the discharge would take different pathways from Lake Vostok down to Ross Ice Shelf depending on the outlet through which the water would flow. The discharge pathways from the northern outlets C and D are practically identical, leading to the coast of Wilkes Land close to 110° east.

These results suggest that the water level of the subglacial lake must rise by about 90 m until water is discharged at location A. This would imply an increase in water volume of 1589 km^3, about 25% of the present water body. This water input must be provided by additional basal melting, either within the lake or in the surrounding, hydrostatically higher areas. For comparison, the detected discharge of a small subglacial water cavity east of Lake Vostok [*Smith et al.*, 2009] would raise Vostok's lake level by only 1.2 mm if all the water reached Lake Vostok. On the other hand, thermodynamic modeling suggests the base of the grounded ice around Lake Vostok to be below the pressure melting point [*Salamatin et al.*, 2009]. This is consistent with the widespread basal accretion along the western lake shore revealed by RES [*Tikku et al.*, 2004]. If this is true, it would limit the source of water input affecting Vostok's lake level to the immediate lake area. Among all the quantities involved, the ice thickness seems to be the most variable in time. While bedrock and geoid configuration may change over geological time scales ($\geq 10^6$ years), the climatically driven ice thickness changes with glacial-interglacial cycles ($\approx 10^5$ years) [*Petit et al.*, 1999]. It is assumed that the ice thickness in the Lake Vostok region was close to the present one during interglacials and about 100–200 m less during glacial maxima [*Salamatin et al.*, 2009]. A decrease in ice thickness reduces the pressure at the ice base, thus increasing the pressure melting point and reducing basal melting in favor of accretion. This, in principle, would extract liquid water from the lake and make discharge less likely. In fact, conditions and processes relevant for mass exchange at the ice-water interface above Lake Vostok are thought to be stable for at least 1 Ma [*Salamatin et al.*, 2009]. In agreement with *Thoma et al.* [2008], an approximate estimation has shown that basal melt/accretion rates in Lake Vostok are insensitive to ice thickness changes of up to ±2000 m, far beyond realistic values in the area under investigation since the formation of the stable Antarctic ice sheet about 13 Ma ago [*Leichenkov and Popkov*, 2012]. We conclude that under present conditions the water level of Lake Vostok is not expected to reach the discharge threshold height within climatic time scales.

As shown in Table 1, the geoid height varies by about 7 m among the four discharge outlets; between outlets A and B the difference keeps below 10 cm. This shows that geoid height variations reach the same order of magnitude (several meters) as the hydrostatic threshold height differences. They must therefore be accounted for in general in the assessment of potential subglacial lake discharge, even though for our particular case their omission would not change the results significantly.

In contrast, the consideration of the pressure head due to the overlying ice sheet is indeed crucial for the identification of subglacial discharge scenarios for Lake Vostok given the considerable increase in ice thickness of about 400 m from the southern to the northern lake shores. If the pressure head term was neglected, water outflow from the lake would be governed solely by the orthometric bedrock height and water discharge would first occur at the northern tip of the lake at 75°50.5′S, 102°21′E, close to location D. In fact, this case corresponds to a Lake Vostok without ice cover and thus would represent the hydrological setting before the onset of glaciation in the region (34 Ma ago) [*Leichenkov and Popkov*, 2012] if differential crustal deformations since that time due to glacial-isostatic adjustment and tectonics are neglected. Considering this scenario, the maximum depth of Lake Vostok, at present 1166 m according to the bathymetry model by *Popov et al*. [2011], would increase by 297 m to then 1463 m, and the lake's water volume would amount to 14000 km^3, more than twice the present volume. This would place Lake Vostok at the third rank among the freshwater lakes on Earth both in terms of maximum depth and water volume (after Lakes Baikal and Tanganyika). Finally, our finding of a threshold height of several tens of meters may support the suggestion of *Zotikov and Duxbury* [2000] that the aquifer of Lake Vostok could predate the onset of glaciation and may have frozen only partially from above.

4. Conclusions

Precise regional models of surface height, ice thickness, and geoid height allow the determination of the location and threshold height of water discharge from subglacial Lake Vostok by a quantitative evaluation of the fluid potential. The agreement found between the estimates by *Erlingsson* [2006] and our results in terms of the location and the threshold height (150 ± 100 m versus 91 ± 23 m) of the outlet and the apparent lake level (3151 m versus 3128 m) is remarkable considering the very limited data available in the earlier study. The primary benefit of the new data lies in a significantly reduced uncertainty of the derived threshold heights. Whereas *Erlingsson* [2006] concluded that Lake Vostok might be close to a jökulhlaup, our results demonstrate that the largest subglacial lake on Earth should not be expected to discharge liquid water and can most likely not be regarded as an active lake.

A future increase in the accuracy of the presented results would require additional ice thickness data by dedicated high-resolution RES profiling. Our work confines the extent of the profiling to a relatively small area around outlets A and B and along the predicted discharge pathways. Increasingly reliable conclusions about the presence or absence of subglacial water inflow into Lake Vostok will be provided by the continuation of geodetic in situ observations of height changes above Lake Vostok [*Richter et al*., 2008].

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