Terrestrial smokers: Thermal springs due to hydrothermal convection of groundwater connected to surface water

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[1] Thermal springs are ubiquitous features whose underground kinematic structure is mostly unknown but are typically thought to originate from deep sources. We documented a type of thermal springs at the banks of a volcanic lake that are discharge zones of hydrothermal convection cells circulating groundwater within the near shore environment. The convection captures lake water through the lakebed, mixes it with deeper groundwater at velocities of 100s of m d−1, then returns the water to the lake via the spring. The convection cell is flushed in a few hours and turns over the lake’s volume in a few days. Most volcanic lakes and other relatively cool surface water bodies in areas of elevated geothermal heat fluxes meet the conditions for the occurrence of local hydrothermal circulation of groundwater. The type of spring we studied, the terrestrial version of black smokers, is likely present but perhaps unrecognized at many areas. Citation: Bayani Cardenas, M., A. M. F. Lagmay, B. J. Andrews, R. S. Rodolfo, H. B. Cabria, P. B. Zamora, and M. R. Lapus (2012), Terrestrial smokers: Thermal springs due to hydrothermal convection of groundwater connected to surface water, Geophys. Res. Lett., 39, L02403, doi:10.1029/2011GL050475.

1. Introduction and Background

[2] We know little about the underground kinematic structure of thermal springs discharging at or near surface water bodies such as lakes and rivers. Volcanic lakes are an ideal setting for studying the interactions of thermal springs with surface water since volcanic lakes occur in ~12% of Holocene-age volcanoes [Pasternack and Varekamp, 1997] and typically host springs. Moreover, the springs in this setting are worth studying since volcanic lakes provide a window into the inner workings of volcanoes. Several sources and pathways, including springs and fumaroles, affect the mass and energy balance of volcanic lakes and may ultimately determine the activity of volcanoes [Brantley et al., 1987; Brown et al., 1989; Pasternack and Varekamp, 1997; Pearson et al., 2008; Varekamp et al., 2000].

[3] Numerous studies have pointed to the important role of subsurface inflows and outflows from volcanic lakes [Brantley et al., 1987; Brown et al., 1989; MacNeil et al., 2007; Rowe et al., 1995, 1992; Sanford et al., 1995] which remains the largest source of uncertainty when closing mass and energy budgets [Rowe et al., 1995]. Bulk chemical and heat balance analyses of some volcanic lakes only require percolation of lake water into the volcano [Brantley et al., 1987], while others suggest the inflow of heated brines into the lake is balanced by extremely large percolation rates out of the lake [Rowe et al., 1992]. These studies suggest very pronounced convection or circulation of groundwater without providing direct evidence.

[4] Despite its recognized importance, there have been no direct observations of groundwater convection and associated heat and solute transport at the perimeter of volcanic lakes, or near any other surface water bodies such as rivers in geothermal areas. Convection black smokers near mid-ocean ridges have been studied using numerical models which showed that the discharge is recirculated seawater [Coulomb et al., 2008]. Hydrogeologic investigations in volcanic areas have also largely relied on computer models and suggest deep, density driven groundwater flow towards a volcano’s edifice [MacNeil et al., 2007; Sanford et al., 1995] but regional flow largely away from the crater towards the flank of the volcano. Little is known regarding how regional flow interacts with local hydrothermal systems and how springs fit into this complex picture.

[5] Here, we report the discovery of a new type of thermal spring along the shore of a volcanic crater lake. We also present a detailed analysis of the coupled fluid flow and heat transport associated with hydrothermal groundwater convection producing the spring. Analysis of conditions suitable for spring development shows that the type of thermal spring we studied may be quite common.

2. Methods

[6] The field component of our study was conducted at the Main Crater Lake (MCL) of Taal Volcano, which is a 311-m high complex volcano that has erupted 33 times since its first recorded activity in 1572. Snapshot and time-lapse thermal images with resolution of 640 × 480 pixels were collected with a handheld thermal infrared camera. Corrections for atmospheric absorbance and other factors were conducted following standard procedures. Velocity fields were calculated from time-lapse thermal images using particle image velocimetry.

[7] In-situ groundwater pressure, total dissolved solids (TDS), and temperature were measured with a probe that simultaneously measured all three parameters. The probe measures electrical conductivity and converts it to TDS with a linear calibration curve. PVC piezometers with a 3.175 cm (1.25 in) diameter and with a screened point (screen length=30.48 cm) were manually driven into the sediment. After sufficient time for thermal equilibration, typically at least 30 minutes, the probe was lowered to the level of the
The coupled flow and temperature field were modeled by simultaneously solving the groundwater flow and heat advection-conduction-dispersion equation using a finite-element model. Groundwater age was similarly modeled with a finite-element model. The numerical models were used for calculation of fluid and heat fluxes across the lake bed and for a broader sensitivity study. A complete discussion of our methods can be found in the auxiliary material. 

3. Results and Discussion

Ground-based thermal imaging of the MCL with a handheld radiometer identified hot areas along the lake shore associated with warm lake water (Figure 1a). Thermal imaging at the shore revealed abundant diffuse non-fumarolic thermal springs (Figures 1b and 1c and Figure S1 of the auxiliary material); we only show a few here although these springs were found throughout the area we accessed. Analysis of time-lapse thermal infrared images of the springs indicated velocities of up to 3 cm s$^{-1}$ within the plumes of warm water discharging from the bank (Figures 1b and 1c and Movies S1 and S2 of the auxiliary material), with most of the water moving at 1–2 cm s$^{-1}$ (or on the order of 1000 m d$^{-1}$). The thermal images (Figures 1b and 1c) show that discharge is very localized as we did not observe any upwelling of warm water in the lake surface further away from the banks. In order to sustain the fluxes we observed, extremely large groundwater flow rates need to be present and/or groundwater is captured from a large area and somehow funneled into the discharge point of the springs.

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Figure 1. Visual and thermal images of the Main Crater Lake (MCL) and springs along its shore. (a) Visual and thermal images of the MCL. (b and c) Visual and thermal images of two springs in the area indicated by the white arrow in Figure 1a. Superposed on the thermal images are vectors representing the time-averaged velocity determined by particle image velocimetry.

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1Auxiliary materials are available in the HTML. doi:10.1029/2011GL050475.
Measurements of groundwater pressure, TDS, and temperature from piezometers driven to increasing depths and at increasing distances along a transect perpendicular to the shore revealed part of a convection cell (Figures 2b and 2c). Even though the water table in the lakeshore follows the topography and dips towards the lake, indicating regional groundwater flow to the lake (Figure 2 and Figure S2 of the auxiliary material), calculated groundwater flow is dominantly upwards towards the water table due to buoyancy but with a strong horizontal component away from the lake in the deeper part of the transect (Figure 2b). Some groundwater also flows downwards and deeper into the aquifer (lower middle portion of Figure 2b). The velocity of upwelling groundwater is on the order of hundreds of m d⁻¹ which is consistent with the rates required by the fluxes observed at the thermal spring.

The isotherms within the 2D vertical transect across the bank (Figure 2a) show a pattern typical of free convection [Nield and Bejan, 2006]. Temperatures within the transect reach 74.7°C in the deepest sampling point (Figure 2a) and decreases to a minimum of 47.1°C at the water table (Figure S2 of the auxiliary material). Lake TDS is 16,330 mg L⁻¹ whereas relatively deep groundwater furthest away from the lake reaches 37,700 mg L⁻¹ (Figure 2c). The TDS distribution suggests percolation by lake water followed by its buoyant rise after mixing with groundwater. With distance inland, both the TDS and temperature decrease to values as low as 1,300 mg L⁻¹ and 47.1°C. High
Effects of geothermal heat flux on the local hydrothermal circulation. Ambient groundwater flux from the bottom and right side is held at $9.84 \times 10^{-6}$ m s$^{-1}$ and permeability is $9.2 \times 10^{-11}$ m$^2$. The geothermal heat fluxes are (a) 2, (b) 10, (c) 20, (d) 30, and (e) 40 W m$^{-2}$.

Figure 3. Effects of geothermal heat flux on the local hydrothermal circulation. Ambient groundwater flux from the bottom and right side is held at $9.84 \times 10^{-6}$ m s$^{-1}$ and permeability is $9.2 \times 10^{-11}$ m$^2$. The geothermal heat fluxes are (a) 2, (b) 10, (c) 20, (d) 30, and (e) 40 W m$^{-2}$.

TDS and high temperatures along the water table at $x = 2$ m (top of transect in Figures 2a and 2c) reflect buoyant upflow of saline water.

[12] The thermohaline Rayleigh Number ($Ra$) – the sum of the solute and thermal Rayleigh Numbers – indicates whether convection is expected to take place. A conservative value for the critical $Ra$ above which convection occurs is $4n^2$ (39.8); this theoretical threshold is for an infinitely wide insulated and bounded domain. The critical $Ra$ normally decreases in the presence of permeability anisotropy and heterogeneity and complex boundary conditions [Nield and Bejan, 2006]. The calculated $Ra$ values for the three vertical profiles in Figure 2 range from 56 to 199 (Table S1 of the auxiliary material) and far exceeds the critical $Ra$. The effects of solute concentration are negligible in all cases and convection is primarily due to temperature differences.

[13] A fully coupled groundwater flow and heat transport numerical model which considers all of the available observations verifies the occurrence of convection (Figures 2e and 2f). The model shows that colder lake water percolates along a broad section of the lake bed, circulates through the shore sediment while mixing with any deeper fluids, and then returns as hotter water along the shore. The hydrothermally circulated water is released as focused spring discharge along the shallowest portion of the bank (Figure 2d) at seepage velocities consistent with those calculated from velocimetry of the spring (Figure S3a of the auxiliary material). The model also indicates a series of convection cells further away from the shore. Assuming that age is 0 when water is recharged into the aquifer, the modeled groundwater age distribution shows that water in the convection cell along the lake shore is mostly less than a half a day old (Figure S3b of the auxiliary material). The mean flux-weighted age of water returning back to the lake is approximately an hour. This rapid circulation of water could turn over the MCL volume of $45 \times 10^6$ m$^3$ [Zlotnicki et al., 2009] in as little as $\sim$11 days if this convective exchange occurs throughout the lake’s perimeter. The total heat flux from the seepage zones, integrated over the lake perimeter, is 82 MW which is larger than the estimated 50 MW which the lake itself dissipates [Zlotnicki et al., 2009]. While convection throughout the MCL’s perimeter may not be the case, the thermal image of the MCL does show that throughout its perimeter the shore is warmer than the lake (Figure 1a).

[14] Additional simulations with increasing basal heat fluxes (from 2–40 W m$^{-2}$) shows that convection occurs even at the lowest heat flux considered (Figure 3). The total heat output of the MCL is 39 W m$^{-2}$ [Zlotnicki et al., 2009]. Since the volcanic input in volcanic lakes ranges from $\sim$5–86% of total output [Pasternack and Varekamp, 1997], the volcanic heat input at the MCL is therefore expected to be smaller than 39 W m$^{-2}$ which is at the low end of calculated values for other volcanic lakes which range from 72–2500 W m$^{-2}$ [Pasternack and Varekamp, 1997]. Nevertheless, the MCL’s average temperature of 29.5°C [Zlotnicki et al., 2009] is similar to the average of 28.4°C for well-studied volcanic lakes [Pasternack and Varekamp, 1997]. Therefore, the presence of bank convection at MCL despite its relative low heat input suggest the likelihood of this phenomena occurring in many volcanic lakes with banks comprised of unconsolidated permeable materials.

[15] Simulations where bank slopes were varied from 10° to 90° show that slope is not a significant causal factor for the hydrothermal circulation (Figures 4b–4f); this range of slopes encompasses what might also be encountered in rivers. A gentler slope enlarges the capture zone of the convection cell under the surface water body, but the water coming from a larger area is still funneled into a spring whose discharge zone remains small and is largely insensitive to shore slope. The larger capture zones for the gentler
sloping banks result in more pronounced flow focusing at the spring (Figure 4a). However, simulations considering an infinitely long domain (with periodic vertical boundaries) with finite depth did not produce convection even at heat fluxes as high as 200 W m$^{-2}$. This condition represents the lake floor far from the shore. Therefore, the convection and springs at the shore result from the nexus of factors at that location: a no flow boundary at the top (the water table) adjacent to sub-vertical to vertical boundary open to a relatively colder water body (a lake or river), and elevated temperatures at depth and/or elevated geothermal heat flux.

[16] These requirements are met by most surface water bodies in geothermal areas. For example, heat flux in the greater Yellowstone area is 2 W m$^{-2}$ [Fournier, 1989] whereas most rivers and lakes in the area have low temperatures; less intense thermal anomalies are found throughout the western United States [Ingebritsen et al., 1989, 2001]. The simulations in Figure 3 include cases where the geothermal heat flux is a relatively low 2 W m$^{-2}$ and the convection and thermal springs are present in all these cases. Therefore, the average conditions needed for local convection and the type of thermal spring we identified are met in Yellowstone, assuming that the rivers or lakes have beds and banks comprised of materials of sufficient permeability.

[17] The thermal springs we observed are basically terrestrial counterparts of submarine black smokers. While natural free convection in porous media has been theoretically and experimentally investigated for centuries, there have been few direct observations of it in natural field settings [Simmons, 2005] except for geophysical evidence for salinity-driven fingers [Van Dam et al., 2009]. Moreover, our study provides the first direct evidence for and analysis of groundwater convection around a volcanic lake previously inferred only from fluid, chemical, and energy balance calculations for volcanic lakes [Brantley et al., 1987; Brown et al., 1989; Pasternack and Varekamp, 1997; Rowe et al., 1995]. The type of thermal springs we studied may dominantly source their water locally and within a stone’s throw of the spring.

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Figure 4. (a) Effects of bank slope on the local hydrothermal circulation and Darcy flux along the bank. Ambient groundwater flux is held at 1 x 10$^{-5}$ m s$^{-1}$, geothermal heat flux is held at 30 W m$^{-2}$, permeability is 9.2 x 10$^{-11}$ m$^{2}$. The slopes are (b) 90° (purple line in Figure 4a), (c) 60° (blue line in Figure 4a), (d) 45° (green line in Figure 4a), (e) 30° (orange line in Figure 4a), and (f) 10° (red line in Figure 4a).

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