A case study of resistivity and self-potential signatures of hydrothermal instabilities, Inferno Crater Lake, Waimangu, New Zealand

A. Legaz, J. Vandemeulebrouck, A. Revil, A. Kemna, A.W. Hurst, R. Reeves, R. Papasin

A case study of resistivity and self-potential signatures of hydrothermal instabilities, Inferno Crater Lake, Waimangu, New Zealand

A. Legaz,[1] J. Vandemeulebrouck,[1] A. Revil,[1,2] A. Kemna,[3] A. W. Hurst,[4] R. Reeves,[4] and R. Papasin[4]

Received 23 February 2009; revised 31 March 2009; accepted 7 April 2009; published 18 June 2009.

[1] Inferno Crater Lake, Waimangu, one of the largest hot springs in New Zealand, displays vigorous cyclic behavior in lake level and temperature. It provides a natural small-scale laboratory for investigating the geo-electrical signature of fluid flows. We measured self-potential and electrical resistivity to see whether the huge variations of fluid volume, approximately 60,000 m^3 during a mean cycle period of 40 days, could be detected. Electrical resistivity measurements revealed spectacular changes over time, with the medium becoming more conductive as the lake receded. This result is consistent with analog models, where the vapor phase is replaced by liquid at recession. The self-potential survey did not detect temporal changes related to fluid movements. This can be explained by the pH of the pore water ($\sim$2.3), which is close to the point of zero charge of silica. **Citation:** Legaz, A., J. Vandemeulebrouck, A. Revil, A. Kemna, A. W. Hurst, R. Reeves, and R. Papasin (2009), A case study of resistivity and self-potential signatures of hydrothermal instabilities, Inferno Crater Lake, Waimangu, New Zealand, *Geophys. Res. Lett.*, *36*, L12306, doi:10.1029/2009GL037573.

1. Introduction

[2] The characterization of hydrothermal fluids has fundamental implications for geothermal explorations and volcanology. In the geothermal case, operators look for strategies in order to get the most benefit from reservoirs and minimize their production costs. Identifying the distribution of phases in the subsurface is essential to characterize the reservoir prior to exploration, and during exploration and exploitation of the field [*Goff and Janik*, 1999]. In volcanology, resumption of volcanic activity is often accompanied by heating episodes [*Trunk and Bernard*, 2008] and the ascent of multiphase fluid [*Hurst et al.*, 1991]. A reliable detection of these phase changes can thus provide evidence for renewal of activity or a warming process. A proper identification of different phases or phase changes requires a good understanding of physical phenomena, as well as detection and monitoring of the area of interest. For this purpose, a preliminary step consists of testing techniques on a laboratory site well adapted to the issue. In order to study the instabilities of a hydrothermal system, we focused on Inferno Crater Lake, a small crater lake (about 80 meters in diameter) located within the Waimangu hydrothermal system (Figure 1), which displays cyclic behavior in lake level and temperature. This site offers three main advantages i) Its cyclic behavior has been established thanks to permanent monitoring of the temperature and water level using pressure and temperature gauges immersed in the lake ii) Its cycles are clearly visible to the naked eye, enabling the adaptation of the measuring device with respect to the phase of interest iii) The relatively short period behavior allows us to assess the reproducibility of measurements over a number of cycles in a reasonable time.

[3] Inferno is a very suitable natural laboratory to test the effectiveness of self-potential and electrical resistivity as methods to monitor deep hydrothermal source processes. This approach is supported by the large lake level variations (up to 5 m in a week), that involve volume fluctuations up to about 60,000 m^3. Previous resistivity measurements in geothermal areas, including Waimangu, have been performed at a regional scale (650 km^2), using mainly Schlumberger arrays with electrode spacings of 500 and 1000 m [*Bibby et al.*, 1994], in order to determine the electrical properties of the geothermal fields and to define their spatial extent. Electrical resistivity methods are commonly used for imaging hydrothermal systems [*Finizola et al.*, 2006; *Saba et al.*, 2007], or as a monitoring technique, most often in ground water pollution studies [*Leroux and Dahlin*, 2006]. To the best of our knowledge, this study is the first attempt that aims to detect and characterize the electric signals associated with physical processes within a cyclic hydrothermal system.

2. Geological Settings

[4] The Waimangu Thermal Valley is located in the Taupo Volcanic Zone in the North Island of New Zealand. It was created during the eruption of Mt. Tarawera on June 10, 1886, with the opening of a south–west radial line of craters (Figure 1). Thermal activity in the valley includes the Inferno Lake, which occupies a sub-circular crater excavated from the side of Mt Hazard, a small rhyolitic dome, and shows a cyclic behavior of lake level and temperature [*Scott*, 1994; *Vandemeulebrouck et al.*, 2008]. This unusual cyclic behavior has an approximate six-week period, during which the water level oscillates between overflowing and about 9 m below the overflow level. Thus, the length of its period is between that of a geyser, about one hour at Old Faithful Geyser, Yellowstone, USA [*Hurwitz et al.*, 2008], and the period of larger crater lakes, such as six months to a year at Mt. Ruapehu, New-Zealand [*Hurst et al.*, 1991]. During overflows, the lake water has a

[1]LGIT, Université de Savoie, CNRS, Le Bourget-du-Lac, France.
[2]Department of Geophysics, Colorado School of Mines, Golden, Colorado, USA.
[3]Department of Geodynamics and Geophysics, University of Bonn, Bonn, Germany.
[4]GNS Science, Lower Hutt, New Zealand.

0094-8276/09/2009GL037573

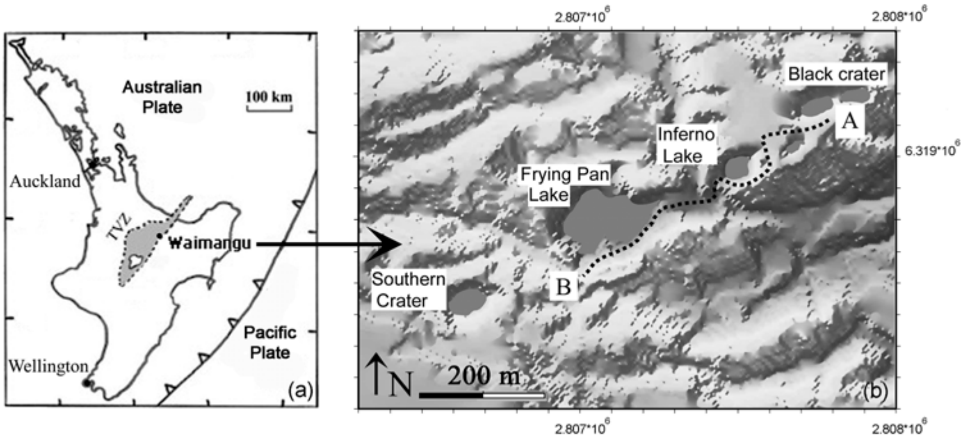

Figure 1. Location of the studied area. (a) Map showing the Waimangu geothermal system and the Taupo Volcanic Zone (TVZ) in the North Island of New-Zealand. (b) Map showing location of resistivity outline along geothermal features, performed from (A) to (B); New Zealand Map Grid Projection coordinates in meters.

maximum temperature of about 75° and a pH around 2.3 [*Glover et al.*, 1994], while at minimum lake level it has a temperature of approximately 35° and a pH of 2.5.

3. A Multi-temporal Sequence of Electrical Resistivity

[5] Electrical resistivity surveys were performed along the profile shown in Figure 1, with a resistivity-meter ABEM SAS 4000, and using the Wenner-α array. We used a set of 64 brass electrodes, with a spacing of 20 m along a 1260 m long profile, following the footpath from Raupo Pond (Point A) to Frying Pan Lake (Point B). A pilot study was conducted in March, 2006, during which two profiles were performed, when Inferno level was high, and during recession, respectively. A second study was performed in March, 2007, during which we repeated the measurements during the whole cycle with nine surveys. As the mid-point of the profile was located on the Inferno viewing platform during the 2006 and 2007 studies, it was quite easy to repeat the survey accurately. Electrode location differences were less than one meter. Due to the topography and dense vegetation, it was not possible to conduct any straight profile across the line of the thermal features. The resulting electrode layout was thus quasi 3-D, with a horizontal distortion larger than the vertical one. For the longer electrode spacings, that produced a mismatch between the theoretical geometric factor and the one corresponding to the true set-up. We thus re-calculated more realistic geometric factors by taking into account the real locations of the electrodes. Each electrical resistivity tomography was computed independently with RES2DINV code [*Loke and Barker*, 1996], using the finite-element method for the forward analysis. The topography was included in the inversion. For the present study, we choose to show two key stages of the cycle: we represent the decimal logarithm of resistivity ratio, which is the measured resistivity when lake level was about 1–1.5 m below the overflow (point "N" in Figure 2b) over the reference resistivity at low lake level (see, point "Ref" at recession, in Figure 2b).

[6] The results show a notable change in electrical resistivity between the two stages, with a 40-fold increase of conductivity at recession, confined to the subsurface around Inferno, at a mean depth of 100 m (Figure 3). Such a behavior is seen in both the 2006 and 2007 resistivity surveys. The main changes are concentrated in an area whose lateral extent is about 150 m, and which is centered on Inferno Crater Lake and becomes deeper towards a south–westerly direction (point B). Three causes can produce resistivity changes during the cycle. The temperature increase during the heating phase that accompanies the lake level rise would produce a resistivity decrease, but we observe an opposite change. A second cause would be the change in the fluid composition, but the 20% variation in sulphate concentration as measured by *Glover et al.* [1994] cannot explain the 40-fold increase of resistivity during the cycle. The last cause would be the fluid phase change that occurs during the cycle as suggested by *Vandemeulebrouck et al.* [2005], who explained the Inferno cycling as a consequence of thermal and gravitational instabilities in a heat pipe system below the lake. In this model, thermal boundary conditions induce a layered structure in the porous heated medium below the lake, with a dense liquid saturated layer lying over a two-phase layer. During the cycle, the relative size of the two layers changes; when the two-phase layer thickness increases, some liquid is transformed into steam producing a lake level rise. When the two-phase layer thickness decreases, steam condenses at depth, and the lake level decreases. The resistivity changes that we observe between the two stages are consistent with the analog modeling; during recession, steam is replaced by liquid in the upper part of the two-phase layer, and saturation increases in the other part (Figure 4), both leading to downward migration of the conductive layer. We present in Figure 5 the evolution of the average resistivity below Inferno during the cycle. The resistivity of the host rock is in part controlled by the resistivity of the saturant fluid and its saturation, according to Archie's law. We measured the conductivity of the lake water, which is 1.7 Ω m referred to a temperature of 35°. Archie's law, for saturation equal to 1,

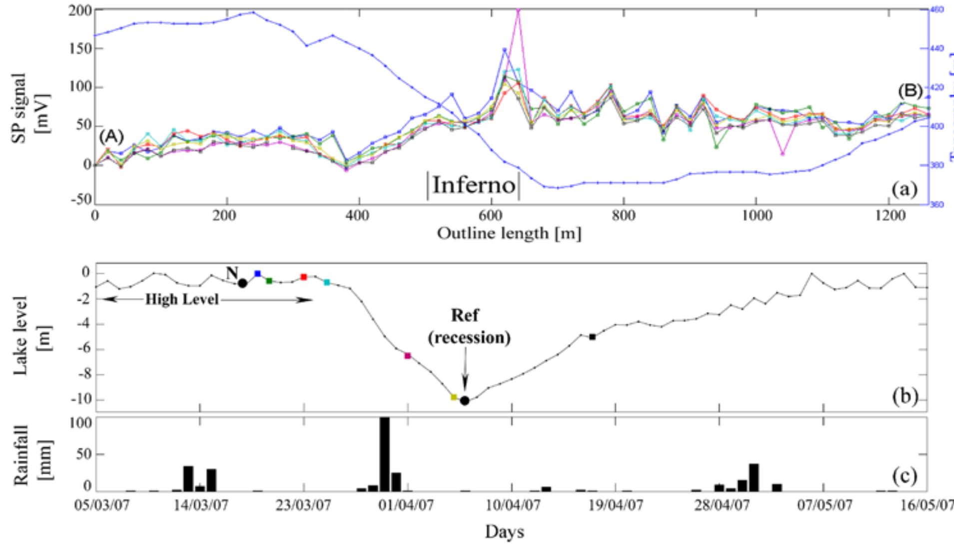

Figure 2. Results for self-potential survey. (a) Self-potential measurements along profile A–B in Figure 1, taken at different stages of Inferno cycle (see the corresponding colors above in Figure 2b). (b) Time at which the surveys have been performed over a whole cycle. "Ref" stands for the state chosen as reference to compute the resistivity ratio in Figure 3, and "N" stands for the high lake level state. (c) Rainfall data.

and a typical porosity for volcanic tuffs of 30% leads to a rock resistivity around 30 Ω m, very close to the values found at recession in Figure 5. We thus obtain resistivity values during recession which correspond well to those of the Inferno fluid in a liquid phase, in agreement with the analog model. However, it is difficult to follow precisely the evolution of the resistivity during the recession stage due to the lack of measurements at intermediate lake levels.

4. Self-Potential Monitoring

[7] In March, 2007, we conducted a series of seven self-potential survey at different stages of the cycle (Figure 2b),

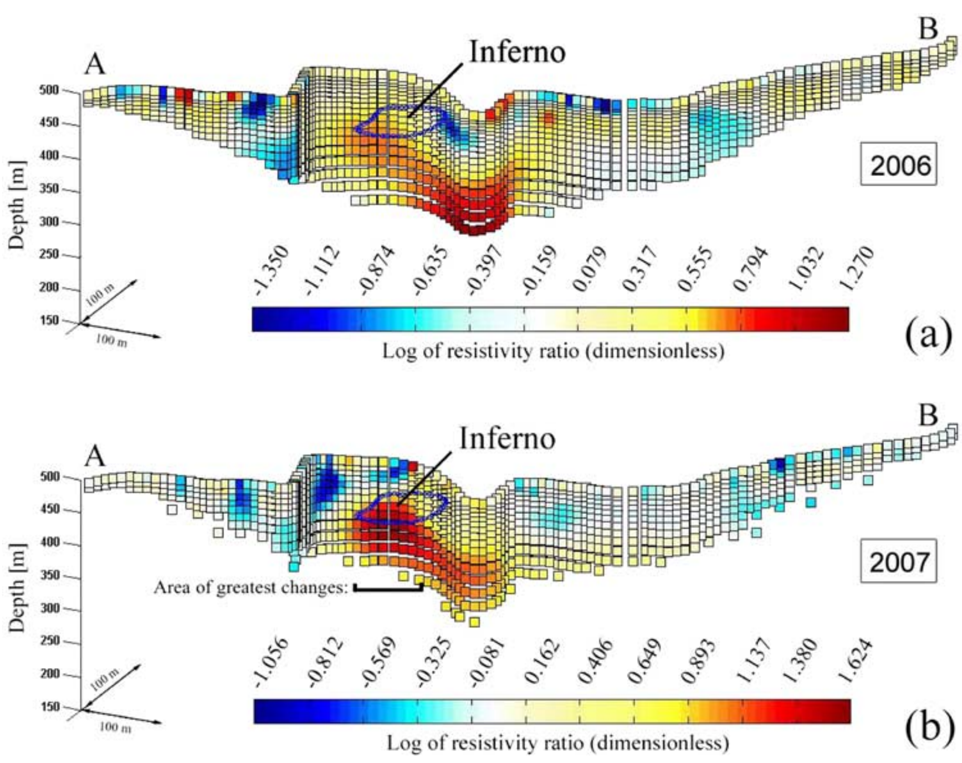

Figure 3. The distribution of the logarithm of resistivity ratio between overflow stage and recession stage (chosen as reference), performed in 2006 and 2007 along profile A–B shown in Figure 1; the blue circle represents the Inferno Lake location.

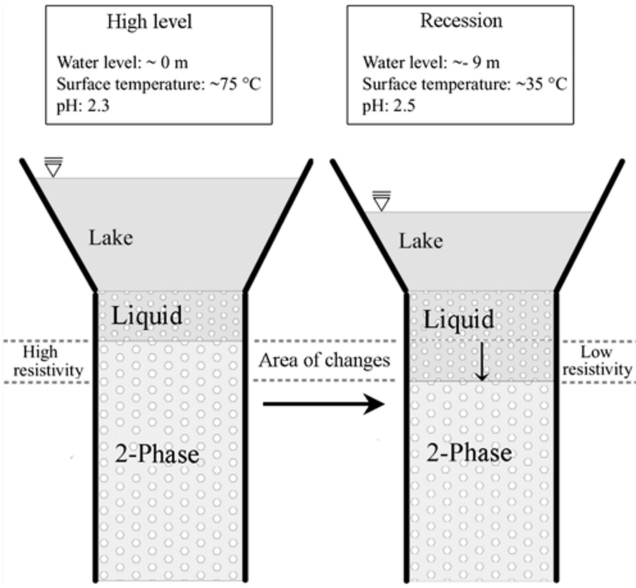

Figure 4. Sketch of phase changes occurring at Inferno, with a downward migration of the liquid head at recession. From *Vandemeulebrouck et al.* [2005], modified.

along the same resistivity profile as described above, with 20 m spacing. We used Petiau Pb/$PbCl_2$ electrodes and a high impedance (100 MΩ) voltmeter Metrix MX20. The reference electrode was set up outside the thermal area (Point "A" in Figure 1). We used a saturated bentonite mud in order to improve the contact between the electrodes and the ground. The results are presented in Figure 2a. There is some relationship between altitude and the self-potential data: the signal along the survey path between 380 and 520 m shows a negative correlation with altitude of -1.8 mVm^{-1}, most likely due to a downflow created by infiltrating waters occurring along the slope.

[8] The most notable result of the surveys is that self-potential signal did not change with lake level, with a mean standard deviation over the entire signal of about 8 mV. In one of the surveys, after 100 mm of rainfall were recorded in the area (Figure 2c), we observed a 200 mV amplitude change at only one point, which can likely be attributed to a change in sub-surface resistivity or to an artifact. Taking into account the constant shape of the SP signal vs. time, the self-potential method does not seem to reflect changes in the subsurface near Inferno.

[9] Two mechanisms are likely to produce self-potential anomalies in this hydrothermal area: firstly, thermoelectric currents, which can be driven by heat flow. In our case, the permanent heat flow is very large in this area; however we do not expect large temporal variations of temperatures at depth during the cycle, indicating the changes due to the thermoelectric effect would be small. Assuming a mean thermoelectric coupling coefficient of 0.2 mV/°C [*Corwin and Hoover*, 1979], a temperature change of 35°C during the cycle would give rise to a 7 mV SP difference, which is of the same order of magnitude as the standard deviation. It will be then very difficult to distinguish the thermoelectric effect variations from noise. The second mechanism expected to generate SP anomalies is the streaming current which corresponds to the drag of the diffuse layer at the pore/liquid interface by the flow of the pore water. In this process, the strength of self-potential signal is controlled by the streaming potential coupling coefficient C (mV m^{-1}), which is proportional to the ζ potential, an electrochemical property of the mineral surface [*Revil et al.*, 1999], which in turn, is a function of water pH. In March 2006, we measured a pH in the lake of 2.23, consistent with *Glover et al.* [1994] values. Moreover, *Grange* [1937] found Inferno Crater to have a high silica concentration of 741 mg kg^{-1}. For a pH range between 2 and 3, the ζ potential is nearly equal to zero about the isoelectric point for silica [*Leroy et al.*, 2008]. Consequently C = 0 suggesting that self-potential signal of electrokinetic origin should not be expected in response to lake level changes. For a similar geysering mechanism, but at a pH of 9, *Legaz et al.* [2009] observed that a cyclic upflow within a pipe gave rise to a clear and cyclic self-potential signature. Inferno Crater is therefore an unusual case where strong fluid flow in the subsurface does not produce any self-potential anomalies. We are not aware of such a documented case in the literature.

5. Conclusion

[10] We studied geo-electrical signals associated with lake level variations at Inferno Crater Lake. The resistivity survey reveals a clear signature of phase changes, reinforced by the reproducibility of results over two years. Lake level variations that can be observed at the surface are likely to be the consequence of saturation changes in the porous medium below, which occur in the vicinity of Inferno, at a mean depth of 100 m. No obvious changes could be seen in self-potential surveys, probably because of the low pH of the pore water. In contrast, electrical resistivity tomography highlighted deep changes of resistivity with time, with a more conductive fluid at recession. This finding is very consistent with a contraction of the 2-phase zone, as suggested by analog models. This result is very promising for the recognition of fluid circulation and phase changes

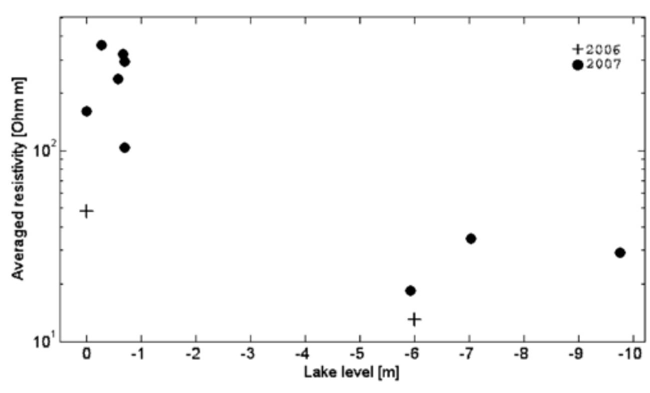

Figure 5. Averaged resistivity versus lake level in 2006 and 2007 computed on an area indicated on Figure 3 where the greatest changes are observed in 2007. The greatest changes in 2006 are located just outside this area (see Figure 3), which explains the lower resistivity values for 2006.

within hydrothermal fields or geysers, as well as in geothermal systems, where similar studies could help to recognize the fraction of recharging water which is transformed into vapor.

References

Bibby, H., S. Bennie, V. Stagpoole, and T. Caldwell (1994), Resistivity structure of the Waimangu, Waiotapu, Waikite and Reporoa geothermal areas, New Zealand, *Geothermics*, *23*, 445–471.

Corwin, R. F., and D. B. Hoover (1979), The self-potential method in geothermal exploration, *Geophysics*, *44*, 226–245.

Finizola, A., A. Revil, E. Rizzo, S. Piscitelli, T. Ricci, J. Morin, B. Angeletti, L. Mocochain, and F. Sortino (2006), Hydrogeological insights at Stromboli volcano (Italy) from geoelectrical, temperature, and CO_2 soil degassing investigations, *Geophys. Res. Lett.*, *33*, L17304, doi:10.1029/2006GL026842.

Glover, R. B., et al. (1994), The relationship of chemical parameters to the cyclic behaviour of Inferno Crater lake, Waimangu, New-Zealand, *Geothermics*, *23*, 583–597.

Goff, F., and C. Janik (1999), Geothermal systems, in *Encyclopedia of Volcanoes*, edited by H. Sigurdsson, pp. 817–834, Academic, San Diego, Calif.

Grange, L. I. (1937), The geology of the Rotorua-Taupo sub-division, *N. Z. Geol. Surv. Bull.*, *37*, 91–94.

Hurst, A., et al. (1991), The heat source of Ruapehu Crater Lake: Deductions from the energy and mass balances, *J. Volcanol. Geotherm. Res.*, *46*, 1–20.

Hurwitz, S., A. Kumar, R. Taylor, and H. Heasler (2008), Climate-induced variations of geyser periodicity in Yellowstone National Park, USA, *Geology*, *36*, 451–454.

Legaz, A., et al. (2009), Self-potential and passive seismic monitoring of hydrothermal activity: A case study at Iodine Pool, Waimangu geothermal valley, New Zealand, *J. Volcanol. Geotherm. Res.*, *179*, 11–18, doi:10.1016/j.jvolgeores.2008.09.015.

Leroux, V., and T. Dahlin (2006), Time-lapse resistivity investigations for imaging saltwater transport in glaciofluvial deposits, *Environ. Geol.*, *49*, 347–358, doi:10.2007/s00254-005-0070-7.

Leroy, P., et al. (2008), Complex conductivity of water-saturated packs of glass beads, *J. Colloid Interface. Sci.*, *321*, 103–117.

Loke, M., and R. Barker (1996), Rapid least-squares inversion of apparent resistivity pseudosections using a quasi-Newton method, *Geophys. Prospect.*, *44*, 131–152.

Revil, A., P. A. Pezard, and P. W. J. Glover (1999), Streaming potential in porous media:1. Theory of the zeta potential, *J. Geophys. Res.*, *104*, 20,021–20,031.

Saba, M., et al. (2007), Development of geothermal field following the 2000 eruption of Usu volcano as revealed by ground temperature, resistivity and self-potential variations, *Ann. Geophys.*, *50*, 79–92.

Scott, B. (1994), Cyclic activity in the crater lakes of Waimangu hydrothermal system, New-Zealand, *Geothermics*, *23*, 555–572.

Trunk, L., and A. Bernard (2008), Investigating crater lake warming using ASTER thermal imagery: Case studies at Ruapehu, Poás, Kawah Ijen, and Copahue volcanoes, *J. Volcanol. Geotherm. Res.*, *178*, 259–270, doi:10.1016/j.jvolgeores.2008.06.020.

Vandemeulebrouck, J., D. Stemmelen, T. Hurst, and J. Grangeon (2005), Analogue modeling of instabilities in crater lake hydrothermal systems, *J. Geophys. Res.*, *110*, B02212, doi:10.1029/2003JB002794.

Vandemeulebrouck, J., A. W. Hurst, and B. J. Scott (2008), The effects of hydrothermal eruptions and a tectonic earthquake on a cycling crater lake (Inferno Crater Lake, Waimangu, New Zealand), *J. Volcanol. Geotherm. Res.*, *178*, doi:10.1016/j.jvolgeores.2008.06.021.

A. W. Hurst, R. Papasin, and R. Reeves, GNS Science, 1 Fairway Drive, Avalon, P.O. Box 30368, Lower Hutt, 6315 New Zealand.

A. Kemna, Department of Geodynamics and Geophysics, University of Bonn, Nussallee 8, D-53115 Bonn, Germany.

A. Legaz, A. Revil, and J. Vandemeulebrouck, LGIT, Université de Savoie, CNRS, F-73376 Le Bourget-du-Lac CEDEX, France. (aurle@univ-savoie.fr)