Repeated Self-Potential Profiling of Izu-Oshima Volcano, Japan

Tsuneo ISHIDO, Tsumeo KIKUCHI, Nobuo MATSUSHIMA, Yusaku YANO, Shinsuke NAKAO, Mituhiko SUJIRRA, Toshiyuki TOSHA, Shinichii TAKAKURA, and Yasuo OGAWA

Geological Survey of Japan, 1-1-3 Higashi, Tsukuba, Ibaraki 305, Japan

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Between March 1989 and March 1994, annual self-potential (SP) surveys were carried out on Izu-Oshima, a small volcanic island. A terrain-related SP distribution of about -1 mV per meter of elevation was observed outside the caldera in all five surveys. Inside the caldera, SP increases from about -350 mV to near 0 mV (relative to the coastline) as the summit crater is approached, although negative anomalies of small spatial extent are manifest. Self-potential inside the caldera decreased by about 100 mV between the March 1989 and the March 1990 surveys, which appears to be correlated with a significant decline in the degassing rate from the summit crater. After 1990, the SP distribution is quite steady along the entire survey line which extends from the west coast through the southern part of the caldera, and ends east of Ura-sabaku. Recently a postprocessor has been developed to calculate space/time distributions of electrokinetic potentials resulting from histories of underground conditions (pressure, temperature, salt concentration, flowrate etc.) computed by multi-phase multi-component unsteady geothermal reservoir simulations (Ishido and Pritchett, 1996). We applied this postprocessor to a simple two-dimensional model of hydrothermal activity in a volcanic island. The low potentials in areas of high elevation are reproduced in the model, and are caused by downflow of meteoric waters. The high potential centered at the summit crater is found to be produced by upflows of volcanic gas and vapor which diminish meteoric water downflow near the volcanic conduit.

1. Introduction

Self-potential (SP) surveys of a number of geothermal areas in Japan were carried out in the 1980's by the Geological Survey of Japan. Each of the surveys covered an area of 50–100 km² with survey lines about 100 km in total length and data sampling intervals around 100 m. SP anomalies of various types were recorded, and obvious anomalies of positive polarity were observed in several different areas: the Kutcharo caldera (Ishido, 1989) and the Nigorikawa caldera (Ishido, 1981) in Hokkaido, the Sumikawa and Okuaizu geothermal areas in the northern part of Honshu (Ishido et al., 1989), and the Unzen-seibu (NEDO, 1988) and Kirishima (Ishido et al., 1990) geothermal areas in Kyushu. Most of these cases provided additional support for correlations reported in the 1970's between positive anomalies and high temperature upflow zones (Zohdy et al., 1973; Zablocki, 1976; Anderson and Johnson, 1976). Electrokinetic potential (streaming potential) generated by hydrothermal circulation is the most probable cause of these positive anomalies (Ishido, 1981; Ishido and Pritchett, 1996).

Terrain-related self-potentials were observed in all of the fields listed above; SP usually decreases between 1 and 100 mV as elevation increases 10 m. This observation can also be attributed to electrokinetic mechanisms. Fluid flow induced by spatial variations in the elevation of the water table is thought to produce SP which decreases as the ground surface elevation increases (Ishido, 1989).

We carried out six annual SP surveys on Izu-Oshima island (Fig. 1), starting in March 1989 and continuing until March 1994. The main purpose of the repeat SP surveys is to detect
changes in SP anomalies caused by hydrothermal activity taking place near the volcanic conduit of Mt. Mihara. Data from repeat SP surveying, along with data from various geophysical and geochemical measurements, is expected to be very useful for modeling the magma-hydrothermal system.

Izu-Oshima is a basaltic volcanic island located about 110 km south of Tokyo. It is made up of several Quaternary volcanoes (e.g., Sakaguchi et al., 1987). Significant volcanic activity has taken place during the last decade. After the summit eruption (lava fountaining) of Mt. Mihara in November 1986, the summit crater filled with lava. After the formation of a new pit crater by a rapid “drain-back” of the pooled magma in November 1987, degassing activity increased. The active degassing phase started in January 1988 (Kazahaya et al., 1993). During the active degassing period, volcanic gas was continuously released at high rates from the summit crater without any eruptions. The degassing rate decreased in March 1990, and weak degassing activity continues at present.

2. Repeated SP Measurements

The survey lines on the island are shown in Fig. 1. SP measurements were made with silver-silver chloride nonpolarizing electrodes and a high-impedance voltmeter. For each survey line, the maximum wire length (from a fixed base electrode) and the data sampling intervals were 500 and...
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100 meters respectively. Telluric activity was monitored by recording potentials across stationary dipoles on the island; no significant telluric variation was observed during each survey period. Closure offsets were relatively small: for example, 2600 m and 3400 m loop traverses were closed with an error of 14.5 mV and 16.1 mV respectively. A common ground reference was chosen near the southwestern coast of the island; recorded SP was featureless and essentially constant for both southwest and northwest survey lines along the shoreline.

SP measurements were repeated five times between 1989 and 1994 along line A–A’ (Fig. 1) (no survey was performed along line A–A’ in 1991). As seen in Fig. 2, SP decreases from -50 to -350 mV along the western part of the survey line (between 0 and 3 km), corresponding to the ground surface elevation increase from 100 to 400 meters. This terrain-related SP is steady, and a similar SP-elevation correlation is also seen along the eastern part of the survey line (between

Fig. 2. Self-potential profiles along survey line A–A’. Also shown is elevation along the survey line.
7 and 8.5 km). The magnitude of the observed terrain-related SP is about 1 mV/meter, which can be produced by streaming potentials associated with downward percolation of meteoric water from higher elevations (Ishido, 1989).

Inside the caldera (the central part of line A-A' between 3 and 6.5 km), SP does not decrease further but instead increases (despite the continuing increase in surface elevation) and small-scale SP anomalies (less than 1 km) are manifest. The maximum value of each short-wavelength anomaly seems to be dependent upon the distance between the observation point and the center of summit crater, becoming greater as the summit crater is approached. Although these short-wavelength features were quite stable, SP within the caldera decreased about 100 mV between March 1989 and March 1990, which seems to correspond to the substantial decline of degassing rate from the summit crater (Kazahaya et al., 1993). These observations indicate that the high

![Diagram of self-potential profiles along survey line B-B'. Also shown is elevation along the survey line.](image-url)
SP in the caldera is probably associated with the hydrothermal activity beneath the summit crater (the electrokinetic mechanism will be described in the next section).

SP profiles along line B-B' (Fig. 3) exhibit similar features. Terrain-related SP was again observed in the northwestern part of 1990 survey between 0 and 3 km and in the eastern part of 1994 survey between 10.5 and 11.5 km (the eastern part of 1990 survey line between 11 and 13 km is the same as in Fig. 2). Although large short-wavelength negative anomalies were present in the caldera, the highest observed potential within the caldera was located near the summit crater (at 8.3 km of the 1994 survey).

3. Numerical Simulation of Electrokinetic Potentials

Recently a postprocessor has been developed to calculate space/time distributions of electrokinetic potentials resulting from histories of underground conditions (pressure, temperature, salt concentration, flowrate etc.) computed by multi-phase multi-component unsteady geothermal reservoir simulations (Ishido and Pritchett, 1996). (We use the term “geothermal reservoir simulation” hereafter to represent numerical simulations of heat and mass transfer in underground porous rocks.) We applied this “EKP” postprocessor to a two-dimensional mathematical model of a volcanic island to understand the mechanisms responsible for the high SP values observed on Izu-Oshima island.

The general relations between the electric current density $I$ and fluid volume flux $J$ (on the one hand), and the electric potential gradient $\nabla \phi$ and pore pressure gradient $\nabla \xi$ forces (on the other) are

\[
I = -L_{ee} \nabla \phi - L_{ev} \nabla \xi, \quad (1)
\]

\[
J = -L_{ve} \nabla \phi - L_{vv} \nabla \xi \quad (2)
\]

where the $L_{ab}$ are phenomenological coefficients (e.g., Ishido and Mizutani, 1981). The first term on the right-hand side in Eq. (1) represents Ohm’s law and the second term in Eq. (2) represents Darcy’s law. The cross-coupling terms (involving the $L_{ev}$ and $L_{ve}$ coefficients) represent the electrokinetic effect; $L_{ev} = L_{ve}$ according to Onsager’s reciprocal relations. Equation (1) describes the total current density, composed of a conduction current density $I_{cond}$ caused by electric conduction, and a drag (convection) current density $I_{drag}$ caused by charges moved by fluid flow; hence,

\[
I = I_{cond} + I_{drag} \quad (3)
\]

where

\[
I_{cond} = -L_{ee} \nabla \phi,
\]

\[
I_{drag} = -L_{ev} \nabla \xi.
\]

In the absence of external current sources, $\nabla \cdot I = 0$, so from Eq. (3):

\[
\nabla \cdot I_{cond} = \nabla \cdot (-L_{ee} \nabla \phi) = -\nabla \cdot I_{drag} \quad (4)
\]

Equation (4) (Poisson equation for $\phi$) represents sources of conduction current which are a prerequisite for the appearance of electric potential at the surface.

The EKP postprocessor simulates electric potentials caused by subsurface fluid flow using a two-step procedure. First, it calculates the distributions of drag current $I_{drag}$ and $L_{ee}$ from the reservoir-simulation results using the same spatial grid used for the reservoir simulation calculation (hereafter called the RSV-grid). The electrokinetic coupling coefficient $L_{ev}$ is computed by the postprocessor using formulations based on experimental work reported by Ishido and Mizutani (1981). Next, the postprocessor calculates the electric potential distribution by solving Eq. (4) within a finite-difference grid which is usually much greater in spatial extent than the RSV-grid.
We will describe two computations. The first mathematical model simulates steady groundwater flow due to topographic relief on an island. The second simulates transient hydrothermal circulation due to magma intrusion, starting from the final state of the first model.

3.1 SP before magma intrusion

The same two-dimensional RSV-grid was used for both models (Fig. 4). It consists of 30 grid blocks in the horizontal direction; the block size is the smallest for the central 10 blocks (each 100 m) and ranges from 250 m to 1000 m for outer blocks on both sides. The grid extends vertically from -3100 to 600 m RSL; each of the lower 12 layers is 250 m in thickness and each of the uppermost 7 layers (from -100 to 600 m RSL) is 100 m in thickness. The upper surface of the grid is uneven to represent topographic relief. The rock is heterogeneous; the entire region is divided into seven sub-regions with different rock properties. All exterior boundaries except the bottom surface are open. Pressure and temperature are maintained along the side boundaries; any "sea water" which flows inward into the grid through the side surface contains a dilute tracer to permit its identification. ("Dilute tracer" means its presence does not influence the physical properties of the fluid.) Constant downward flux of "fresh water" (similarly tagged with a dilute tracer) was imposed along the top surface above sea level.

Starting from an appropriate initial condition, groundwater flows from higher to lower elevations were simulated using the "STAR" geothermal reservoir simulator (Pritchett, 1995). The system reached steady state after about 50 years; in Fig. 5, the distributions of mass flux, pressure, gas saturation and mass fraction of "seawater tracer" are shown for 100 years. The water table is located 100–300 meters below the ground surface and an unsaturated zone is created in the central high elevation areas. The water table depth depends mainly on the recharge rate of meteoric water and the permeability of the shallow part of the island; in the present case, the recharge rate is assumed to be equivalent to annual precipitation of 3 meters, and a permeability of 100 millidarcy is assigned to the region above -1.6 km RSL. As seen in Fig. 5(d), a structure with a fresh-water layer overlying a sea-water layer is produced.

For the self-potential calculations, the fresh water and seawater are assumed to contain
NaCl; the NaCl concentration is $1.7 \times 10^{-3}$ mol/l and 0.6 mol/l in pure fresh water and pure seawater respectively. The postprocessor calculates $\text{Lee}$, $\text{Lev}$ and $\text{Idrag}$ from these distributions of composition and other results from the STAR simulation (such as temperature, pressure and fluid mass flux) within the RSV-grid. Within that portion of the SP-grid overlapped by the RSV-grid, the distribution of electrical conductivity is obtained directly from RSV-grid values. Elsewhere within the SP-grid, the electrical conductivity distribution is user-specified and time-invariant; 0.1 S/m is assumed below $-600$ m RSL and 5 S/m is prescribed above that. Then, the distribution of electric potential is calculated within the SP-grid; the results are shown in Fig. 6(b).

A large negative SP anomaly appears in the high elevation region where meteoric water flows downward. Since the zeta-potential, to which $\text{Lev}$ is proportional, is negative (Ishido and Mizutani, 1981), the descending meteoric water removes positive charge from the neighborhood of the ground surface and thus produces a conduction current sink. The magnitude of the negative
anomaly is sensitive to the electrical conductivity $Lee$ of the shallow zone; it is more than 300 mV at elevations above 300 m RSL in the present calculation (in which $Lee \approx 10^{-3} \text{ S/m}$). If $Lee$ is adjusted to $10^{-2} \text{ S/m}$ by increasing the rock matrix conductivity, the magnitude is reduced to around 100 mV.

SP is slightly higher in the central summit region than the surrounding area; this is due to the relatively deep location of the conduction current sink corresponding to the deep water table in the summit region. In the unsaturated zone, the drag current density $Idrag$ is assumed to be zero unless the liquid water saturation (volume fraction of pore space occupied by liquid
phase) exceeds 0.5, so meteoric water downflow does not induce \( I_{drag} \) in the shallow part of unsaturated zone where the liquid water saturation is less than 0.5.

3.2 SP after magma intrusion

The second model simulates the evolution of hydrothermal convection caused by heat transfer from magma, starting from the steady state of the first model (Fig. 5). A 200-meter wide magma body extending from \(-3100\) to \(-1100\) m RSL was introduced to the central zone at time = 0; the temperature of the magma body was held at 780°C for the first two years. To represent magma degassing, a constant upward flux of water (780°C, tagged with “magmatic dilute tracer”) released from the top of magma was assumed. This “magma degassing” was also terminated at \( t = 2 \) years. Then, the cooling of the magma body and associated geothermal phenomena were simulated from 2 to 10 years after the magma intrusion. In Fig. 7, the distributions of temperature, fluid mass flux, gas (vapor) saturation and mass fraction of seawater tracer (computed using the

![Fig. 7. Five years after magma intrusion: distributions of (a) fluid mass flux, (b) temperature (solid contour interval is 100°C), (c) gas (vapor) saturation (contour interval is 0.05) and (d) mass fraction of “seawater dilute tracer” (contour interval is 1/18 of source fluid concentration) in central part (~3500 m in the horizontal direction) of the RSV-grid. Thick solid line in each diagram shows a zone of magma intrusion.](image-url)
STAR simulator) are shown for \( t = 5 \) years. The magmatic temperature is still high at this stage and strong upward flow takes place along and above the magma body. As seen in Fig. 7(c), a vapor-dominated zone about 1 km in height develops under the summit region.

The SP distribution calculated by the EKP postprocessor for \( t = 5 \) years is shown in Fig. 6(c). A significant change is seen in the central region (compare Fig. 6(c) to Fig. 6(b)); a SP increase of more than 300 mV is produced near the summit. This change is mainly brought about by reduction of the sink of conduction current associated with meteoric water downflow which was present under the summit region prior to the magma intrusion (Fig. 6(b)). Also, the upward flow of liquid-phase water induces \( \mathbf{I}_{\text{drag}} \) and produces a source of conduction current at the bottom of the vapor dominated zone, but this effect is relatively minor due to the deep location of the source.

The calculated result shown in Fig. 6(c) is based on a simple two dimensional model, but it still reproduces the main features of the observed SP on Izu-Oshima island shown in Figs. 2 and 3. (Causes of the short wavelength anomalies inside the caldera will be discussed in the next section.) As seen in Fig. 2, the SP distribution (including the relatively high potential inside the caldera) was almost steady from March 1990 to March 1994. This feature is well reproduced in the present model; the calculated SP distributions for times from 6 to 10 years are almost the same as to that for \( t = 5 \) years shown in Fig. 6(c). The potential at the summit shows a slight decrease as the upward flux of high temperature fluid (fumarolic activity at the summit) decreases, but remains near zero millivolts. It takes a long time to cool the heated zone beneath the summit region. As long as meteoric water downflow is prevented by these buoyant effects, the high potential in the summit region will persist.

4. Discussion

The electrical conductivity \( \epsilon \) calculated for the present model is as follows. Regions saturated with fresh water have about 0.001 S/m, which is roughly given by multiplying the porosity (0.05 is assumed for the shallow rocks) by the pore fluid conductivity (calculated as about 0.02 S/m from NaCl concentration and temperature). Deep regions saturated with seawater have high conductivities (0.03 to 0.3 S/m), since seawater conductivity is about 10 S/m for temperatures between 50 and 100°C. As shown in Fig. 7(d), seawater is driven to shallow levels due to thermal buoyancy in the central region. Although \( \epsilon \) is very small within the vapor zone, the surrounding zone has high \( \epsilon \)—more than 0.1 S/m.

Resistivity surveys (Ogawa et al., 1989; Mitsuhata et al., 1994) reported a high resistivity (low \( \epsilon \)) at shallow levels and low resistivities at deeper levels near Mt. Mihara; the present model provides a possible explanation of these observations. Yukutake et al. (1994) reported changes in resistivity in the eastern part of caldera between 1982 and 1987; the resistivity above sea level was about 2 k \( \Omega \cdot \text{m} \) in 1976 and 1982, but it was observed as about 200 \( \Omega \cdot \text{m} \) in March 1987 and in the three subsequent surveys between 1988 and 1991. This observed resistivity change can be explained by the present model as resulting from replacement of fresh water with high salinity water in the peripheral region of the vapor zone beneath the summit crater.

As seen in Figs. 2 and 3, persistent local short-wavelength SP anomalies are observed inside the caldera. These features cannot be explained by the present simple model. If we assume that these features result from the presence of local negative anomalies (not by positive anomalies), the most probable interpretation is as follows. A relatively thick unsaturated layer is thought to overlie the water table inside the caldera. However, local liquid-phase meteoric water downflows will in reality take place through the unsaturated zone owing to the heterogeneous small-scale permeability structures (fractures) present in volcanic formations. These local liquid downflows, in turn, will produce sinks of conduction current localized near the ground surface, resulting in negative SP anomalies of limited spatial extent. The anomalies will persist so long as the fractures...
5. Concluding Remarks

Repeat self-potential surveys of active volcanoes have been carried out in many areas during the last two decades; Kilauea (Zablocki, 1976), Etna (Massenet and Pham, 1985), Usu (Nishida and Tomiya, 1987; Nishida et al., 1996), Akita Yake-yama (Ishido et al., 1989), Hokkaido Komaga-take (Matsushima et al., 1990), Soufriere (Zlotnicki et al., 1994), Unzen (Hashimoto and Tanaka, 1995), Tarumae (Miyamura et al., 1995), and Miyake-jima (Nishida et al., 1996). Obvious positive-polarity SP anomalies were often observed around volcanic vents or fissures where fumarole activity was taking place. In most cases, the streaming potential associated with high temperature upflow was thought to be the primary cause of these positive anomalies. In cases like Akita Yake-yama, Miyake-jima, Usu and Tarumae, the SP anomaly distributions have similar features to those observed at the Izu-Oshima volcano: SP first decreases several hundred millivolts as one climbs the slopes of the volcano, then rapidly recovers to the level measured on the flank of volcano as the summit crater is approached.

Upflow of volcanic gas and vapor (which diminishes meteoric-water downflow near the volcanic conduit) is believed to be the primary cause of the relatively high potential inside the Izu-Oshima caldera (compared to that outside). Although the upflow itself produces a source of conduction current, the effect of suppressing the sink of conduction current associated with meteoric-water downflow is dominant. We believe this mechanism is also applicable to observations at the other volcanoes.

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REFERENCES

Anderson, L. A. and G. R. Johnson, Application of the self-potential method to geothermal exploration in Long Valley, California, J. Geophys. Res., 81, 1527–1532, 1976.

Hashimoto, T. and Y. Tanaka, A large self-potential anomaly on Unzen Volcano, Shimabara peninsula, Kyushu island, Japan, Geophys. Res. Lett., 22, 191–194, 1995.

Ishido, T., Streaming potential associated with hydrothermal convection in the crust: a possible mechanism of self-potential anomalies in geothermal areas, J. Geothermal Res. Soc. Japan, 3, 87–100, 1981 (in Japanese with English abstract).

Ishido, T., Self-potential generation by subsurface water flow through electrokinetic coupling, in Detection of Subsurface Flow Phenomena, Lecture Notes in Earth Sciences, vol. 27, edited by G.-P. Merkler et al., pp. 121–131, Springer-Verlag, 1989.

Ishido, T. and H. Mizutani, Experimental and theoretical basis of electrokinetic phenomena in rock-water systems and its applications to geophysics, J. Geophys. Res., 86, 1763–1775, 1981.

Ishido, T. and J. W. Pritchett, Numerical simulation of electrokinetic potentials associated with subsurface fluid flow, in Proc. 81st Workshop on Geothermal Reservoir Engineering, pp. 143–149, Stanford University, 1996.

Ishido, T., T. Kikuchi, and M. Sugihara, Mapping thermally driven upflows by the self-potential method, in Hydrological Regimes and Their Subsurface Thermal Effects, Geophys. Monogr. 47, IUGG vol. 2, pp. 151–158, AGU, 1989.

Ishido, T., T. Kikuchi, Y. Yano, M. Sugihara, and S. Nakao, Hydrogeology inferred from the self-potential distribution, Kirishima geothermal field, Japan, Geothermal Resources Council Transactions, 14, 916–926, 1990.

Kazahaya, K., M. Takahashi, and A. Ueda, Discharge model of fumarolic gases during post-eruptive degassing of Izu-Oshima volcano, Geochim. J., 27, 261–270, 1993.

Massenet, F. and V. N. Pham, Mapping and surveillance of active fissure zones on a volcano by the self-potential method, Etna, Sicily, J. Volcanol. Geotherm. Res., 24, 315–338, 1985.
Matsushima, N., M. Michiwaki, N. Okazaki, R. Ichikawa, A. Takagi, Y. Nishida, and H. Y. Mori, Self-potential studies in volcanic areas (2)—Usu, Hokkaido Komaga-take and Me-akan, J. Fac. Sci., Hokkaido Univ., Ser. VII(Geophysics), 8, 465-477, 1990.

Mitsuhata, Y., T. Uchida, T. Yokota, and A. Tanaka, Investigation of Izu Oshima volcano by audio magnetotelluric (AMT) and transient electromagnetic (TEM) surveying, in Proc. 90th SEGJ Conf., pp. 536-539, 1994 (in Japanese).

Miyamura, J., Y. Tajima, Y. Yamaguchi, and K. Tamura, Self-potential measurement in Tarumae volcano, Quarterly J. Seismology, Japan Meteorological Agency, 58, 79-90, 1995 (in Japanese with English Abstract).

NEDO (New Energy and Industrial Technology Development Organization), Unzen-seibu Area, Report of Survey to Promote Geothermal Development, No. 15, 1988 (in Japanese).

Nishida, Y. and H. Tomiya, Self-potential studies in volcanic areas (1)—Usu volcano, J. Fac. Sci., Hokkaido Univ., Ser. VII(Geophysics), 8, 173-190, 1987.

Nishida, Y., N. Matsushima, A. Goto, Y. Nakayama, A. Oyamada, M. Utsugi, and H. Oshima, Self-potential studies in volcanic areas (3)—Miyake-jima, Easan and Usu, J. Fac. Sci., Hokkaido Univ., Ser. VII(Geophysics), 10, 63-77, 1996.

Ogawa, Y., S. Takakura, T. Uchida, T. Tosha, T. Nakatsuka, T. Soya, J. Nakai, Y. Ninomiya, and R. Morijiri, Magnetotelluric profiling of Izu Oshima volcano (1), in Proc. 81st SEGJ Conf., pp. 331-334, 1989 (in Japanese).

Pritchett, J. W., STAR: a geothermal reservoir simulation system, in Proc. World Geothermal Congress, Florence, pp. 2959-2963, 1995.

Sakaguchi, K., K. Okumura, T. Soya, and K. Ono (ed.), The 1986 Eruption of Izu-Oshima Volcano 1:25,000, Miscellaneous map series 26, Geological Survey of Japan, 1987.

Yukutake, T., T. Yoshino, H. Utada, Y. Sasai, T. Shimomura, and S. Koyama, Variations in the electrical resistivity of Izu-Oshima volcano observed by a direct current method, Bull. Earthq. Res. Inst., 69, 107-120, 1994 (in Japanese with English abstract).

Zablocki, C. J., Mapping thermal anomalies on an active volcano by the self-potential method, Kilauea, Hawaii, in Proc. 2nd U.N. Symposium on the Development and Use of Geothermal Resources, vol. 2, pp. 1299-1309, 1976.

Zlotnicki, J., M. Peuillard, and G. Hammouya, Water circulations on la Soufriere volcano inferred by self-potential surveys (Guadeloupe, Lesser Antilles). Renew of volcanic activity?, J. Geomag. Geoelectr., 46, 797-813, 1994.

Zohdy, A. A. R., L. A. Anderson, and L. J. P. Muffler, Resistivity, self-potential, and induced-polarization surveys of a vapor-dominated geothermal system, Geophysics, 38, 1130-1144, 1973.