Numerical simulation of piezomagnetic changes associated with hydrothermal pressurization and its application to volcanomagnetic variations observed at Merapi Volcano.

Ayako OKUBO* and Wataru KANDA**

* Geological Survey of Japan, AIST, ** Sakurajima Volcano Research Center, DPRI, Kyoto Univ.

We have developed a computer program to calculate geomagnetic field changes due to the piezomagnetic effect, which is caused by hydrothermal pressurization. The program uses the distribution of pore fluid pressure that is the result of a numerical simulation for hydrothermal fluid circulation after magma intrusion. Our method was applied to actual volcanomagnetic field variations observed at the Merapi Volcano in Indonesia. In this report, we aim to explain these volcanomagnetic variations by applying our method because the variations are relatively slow, implying pressurization/depressurization processes within the volcano. Special attention is devoted to the period from August 1990, when a gas plume emission was observed, to January 1992, when an eruption fed a large pyroclastic flow occurred. During this period, a step like positive magnetic change, starting from May 1991, was observed at both the IJO and CEM stations. Zlotnicki and Bof (1998) ascribed this anomalous change to over pressure within the volcano, as inferred from the increase in seismic velocity (Ratdomopurbo & Poupinet, 1995).

A numerical model of the Merapi Volcano utilizing HYDROTHERM had already been proposed by Harmoko *et al.* (2007) in order to know the temperature and pressure distributions during a period of continuous volcanic activity. In this study, we refer to Harmoko *et al.*(2007)'s model and use similar model parameters for the hydrothermal simulation (Fig. 1). Our numerical model uses 2-D axi-symmetric radial coordinates to simulate a 3-D domain. For simplicity, we assumed a 2-D topography representing the Merapi Volcano by the function in consideration of the slope. A deep magma chamber is located at a depth of 8.7 km beneath the summit, corresponding to the geodetic source (Beauducel & Cornet, 1999), and a magmatic conduit with a radius of 10 m extends from the chamber to the depth of 1 km beneath the summit. In addition, a shallow magma chamber is located at depths between 1.5 km and 2.5 km as proposed by Ratdomopurbo & Poupinet (2000). The HYDROTHERM simulation was run with 40 by 40 grids in the vertical (z) and the horizontal (x) directions. The bottom boundary was impermeable rock with a constant basal heat flux of 120 mW/m^2. In the simulations, we assumed the temperature dependents permeability and heat capacity we used in the previous section, while other rock properties (porosity, thermal conductivity, density and Poisson's ratio) were constant throughout the simulations following Harmoko *et al.*(2007)'s model (Table 1). Permeability was set to 2×10^{-15} m^2 below 360◦C; this value was used in Harmoko *et al.*'s model, although they assumed temperature-independent permeability. In the temperature range of 360-500◦C, permeability was assumed to decrease log-linearly and a negligibly low value of 2×10^{-21} m^2 was used above 500◦C. In addition, the heat capacity (2700$J/(kgK)$) for temperatures between 900◦C and 750◦C is double the lower temperature value. The direction of the initial magnetization was assumed to be the same as that of the geomagnetic field. The inclination

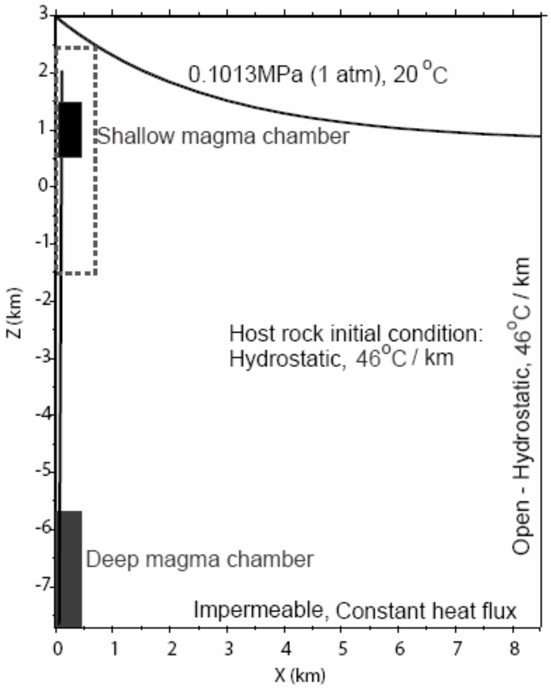

Fig. 1 Configuration of the numerical model of the Merapi Volcano modified from Harmoko et al.(2007), showing boundary and initial conditions for hydrothermal fluid circulation. The model consists of axi-symmetric 2-D coordinates. The area shown by a dashed line corresponds to the area shown in Fig. 3.

and the declination were set to -30 and 0 degrees, respectively. The stress sensitivity was assumed to be 10^{-3}MPa^{-1} following Zlotnicki & Bof (1998). Since Merapi Volcano is an andesitic stratovolcano, the initial magnetization was assumed to be uniformly magnetized with intensities of 1.0, 3.0 or 5.0 A/m.

Simulations begin with the instantaneous intrusion of magma at a temperature of 900$\circ C$ and with a lithostatic pressure (t = 0) into the deep chamber. After the hydrothermal activity has sufficiently declined to a quasi-stationary state ($t = t_0$; t_0=10,000 years), the following four cases are examined the source of pressurization.

Case 1: The deep magma chamber is pressurized by a new intrusion.

Case 2: The deep magma chamber and the conduit are pressurized by a new intrusion.

Case 3: The conduit and the shallow magma chamber are pressurized by an intrusion from the deep chamber.

Case 4: The entire magmatic system including the conduit from the deep chamber to the shallow magma chamber is pressurized by a new intrusion.

At t=t_0, the new intrusion occurs at 900$\circ C$ and with lithostatic pressure. Pore-pressure distribution is calculated at each time step ($t = t_0+\Delta t$). The magmatic system is then pressurized at Δt=0.75 years (*Cases* 1−4). The magma in the chambers or the conduit is reset to 900$\circ C$ and 1.5 times lithostatic pressure. This corresponds to observational evidence that the overpressure within the volcanic edifice was inferred in May 1991 from the speedup of seismic velocity (Ratdomopurbo & Poupinet, 1995). They suggest that the variation in velocity is probably related to a change in pressure within the magma chamber and the conduit, and that the pressure is at least higher than the lithostatic pressure. They also observed a decrease in seismic velocity after September 1991, which we simulate by a decrease in pressure in the magmatic system to 0.8 times lithostatic pressure at 900$\circ C$ at Δt=1.1 years.

Table 1. Rock properties used for hydrothermal model of Merapi volcano, Indonesia. Those values are the same used in Harmoko *et al.*(2007)'s model except for permeability.

Property	both host-rock and magma
Permeability [m^2]	temperature dependent (details in text)
Porosity [%]	10
Heat capacity [$J/(kgK)$]	temperature dependent (details in text)
Thermal conductivity [$W/(mK)$]	2.6
Rock density [kg/m^3]	2100
Poisson's ratio ν (μ=λ)	0.25

Figures 2(a-d) show temporal changes in the expected piezomagnetic fields calculated for three different initial magnetizations of 1.0, 3.0 and 5.0 A/m. In comparison with the simulation results of four cases, Figs. 2(c) and (d) show a similar pattern to the observed steplike magnetic signal (Fig. 6 of Zlotnicki & Bof, 1998). A common feature of two cases (*Cases* 3 and 4) is that the shallow chamber is pressurized. In the other two cases, when pressurization does not occur in the shallow chamber (*Case* 1 and *Case* 2) they show a pattern opposite to the observed signal (Figs. 2a and b). This means that if the observed volcanomagnetic signal is of piezomagnetic origin as proposed by Zlotnicki & Bof (1998), the shallow magma chamber should be the pressurized source. Seismic activity during this period was high. A number of volcano-tectonic A-type (VTA) and B-type (VTB) earthquakes occurred at depths of 2.5-5.5 km and at depths above the shallow chamber, respectively, and their seismic energy reached a maximum in September 1991 (Δt=1.1), respectively (Ratdomopurbo & Poupinet, 2000).The changes in seismic velocity were detected from analyses of the VTB earthquakes. Thus, our results are consistent with seismic activity and velocity changes. The results of *Case* 3 and *Case* 4 assuming a uniformly magnetized media of 3 A/m better explain the observed volcanomagnetic field amplitudes.

Figure 3 shows the distributions of the pore-fluid pressure change (ΔP) from a quasistationary state ($t = t_0$) and the temperature near the intruded magma (dashed square region in Fig. 1) at Δt=0.8 years for *Case* 4 (Fig. 3a) and *Case* 2 (Fig. 3b), respectively. The simulation shown in Fig. 3 demonstrates that both pressure changes and temperatures within the conduit or the shallow magma chamber are significantly greater compared to the surroundings. In contrast, an intrusion into the shallow magma chamber yields significant differences in pressure distribution at the shallower depth as shown in Fig. 3(a). The pressure changes in and around the shallow magma chamber, which strongly influence the piezomagnetic field change on the ground as discussed in *Cases* 1−4, are greater by ΔP=10-60 MPa compared to a system without the intrusion (Fig. 3b). Although the temperature of the magma still exceeds 800 ◦C at Δt=0.8 years after the intrusion, the temperature gradient around the chamber is very large and an order of 10 MPa of overpressure is obtained around the contour of 200 degrees, at which enough magnetization remains. Ratdomopurbo & Poupinet(1995) suggested that a pressure change of 50 MPa would be required to explain the observed change in an S-wave velocity of 1.2%. It is consistent with our simulation result within a range of that order.

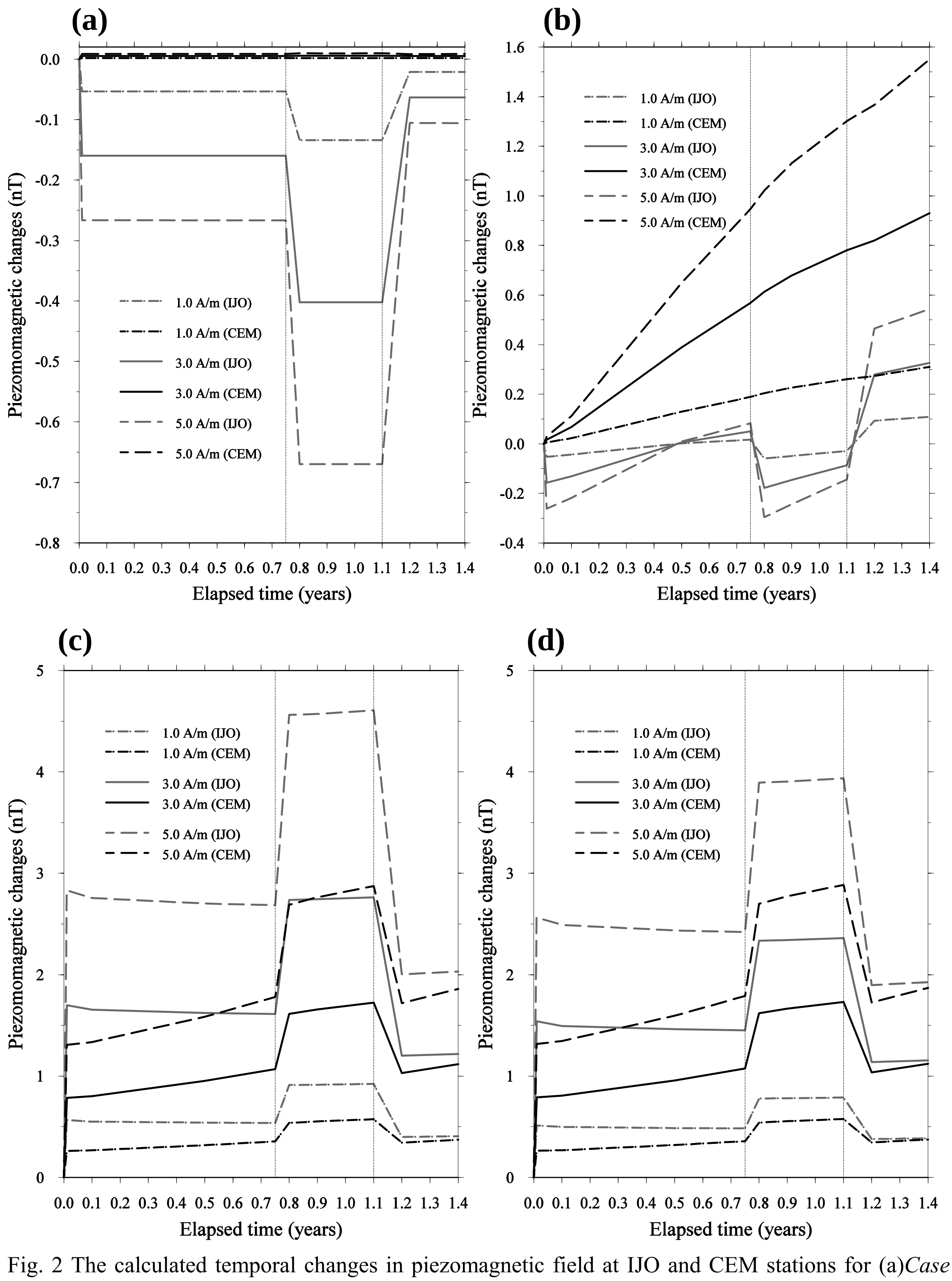

Fig. 2 The calculated temporal changes in piezomagnetic field at IJO and CEM stations for (a)*Case 1*, (b)*Case 2*, (c)*Case 3* and (d)*Case 4*, corresponding to the period 1-3 in Fig. 6 of Zlotnicki & Bof (1998).

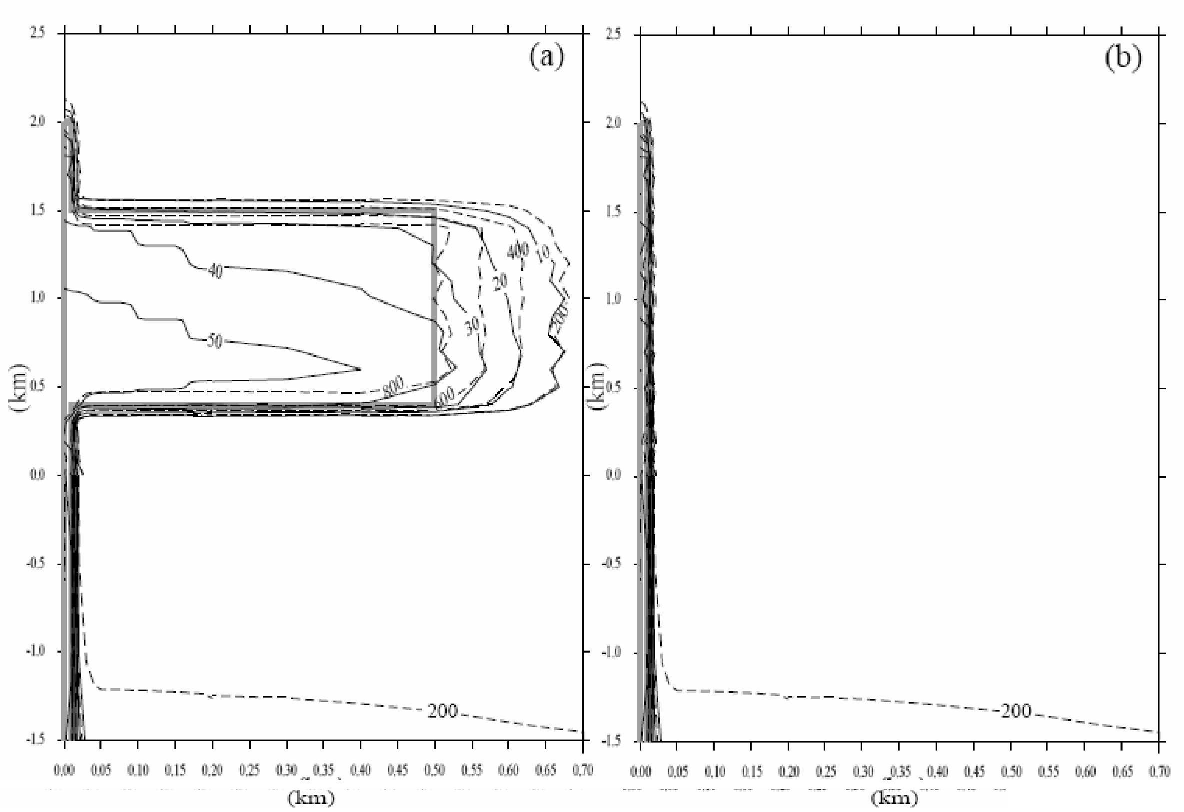

Fig. 3 Distributions of pore-fluid pressure change of Δ P=P(t=0.8)-P(t=0) (solid contours) and temperature (dotted contours) for (a) *Case 4* and (b) *Case 2*. The section along the north-south direction at Y=0 km is drawn for both cases. The contour intervals are 10 MPa in pressure changes and 200°C in temperature. The gray thick line indicates the outline of the shallow magma chamber and the conduit.

REFERENCES

Beauducel, F. and Cornet, F. H., 1999, Collection and three-dimensional modeling of GPS and tilt data at Merapi volcano, Java, *J. Geophys. Res.*, **104**, 725-736.

Harmoko, U., Fujimitsu, Y. & Ehara, S., 2007. Shallow ground temperature anomaly and thermal structure of Merapi volcano, central Java, Indonesia, *J. Geotherm. Res. Soc. Japan*, **29**, 25-37.

Ratdomopurbo, A. & Poupinet, G., 1995. Monitoring a temporal change of seismic velocity in a volcano: application to the 1992 eruption of Mt. Merapi (Indonesia), *Geophs. Res. Lett.*, **22**, 775-778.

Ratdomopurbo, A. & Poupinet, G., 2000. An overview of the seismicity of Merapi Volcano (Java, Indonesia), 1983-1994, *J. Volcanol. Geotherm. Res.*, **100**, 193-214.

Zlotnicki, J., & Bof, M., 1998. Volcanomagnetic signals associated with the quasi-continuous activity of the andesitic Merapi volcano (Indonesia): 1990-1995, *Phys. Earth Planet. Int.*, **105**, 119-130.