

ACTIVE system for monitoring volcanic activity: A case study of the Izu-Oshima Volcano, Central Japan

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Abstract

A system is proposed for the monitoring of changes in the underground structure of an active volcano over time by applying a transient electromagnetic method. The monitoring system is named ACTIVE, which stands for Array of Controlled Transient-electromagnetics for Imaging Volcano Edifice. The system consists of a transmitter dipole used to generate a controlled transient electromagnetic (EM) field and an array of receivers used to measure the vertical component of the transient magnetic field at various distances, with automatic operation of both units. In order to verify the performance of the proposed system, numerical and field experiments were carried out by application of the system to the Izu-Oshima volcano, where a remarkable change in the apparent DC resistivity over time had been detected in association with the eruption in 1986.

Assuming that the next eruption will follow a scenario similar to that of 1986, an array of five receivers was constructed around the summit crater of the central cone, Mihara-yama (Mt. Mihara) and a transmitter dipole with a 700 m long grounded cable was installed approximately 1 km southeast of the summit crater in 2002. A long-term field experiment was then carried out. With the transmitter and receivers both synchronized by a global positioning system (GPS) clock, this field experiment has shown that over approximately 4 years, the daily value of the accuracy of the system was as high as 1% or better, while the accuracy of the monthly mean of the response functions was approximately 0.1%. Many problems with the instruments, the device software, and/or the electronic circuits have also been solved during this field experiment.

For data interpretation, a three-dimensional (3-D) forward modeling code was built to calculate theoretical responses. By applying this forward code, a sensitivity analysis was performed for a realistic 3-D conductivity distribution model, which showed that the receiver array of the ACTIVE system in Izu-Oshima has a relatively high sensitivity to conductivity changes below the summit crater. A simple and rapid imaging method based on the Born approximation was developed for monitoring purposes, and a numerical test showed that the change in conductivity over time, supposed to have occurred before the 1986 eruption, could be imaged by the present system, which features improved spatial resolution.

During the 4-year-long field experiment, changes in the response function over time of up to 7% were observed, from which two source regions, R1 and R2, were imaged at a shallower part of the caldera between the transmitter and the crater, and at approximately 500 m below the crater, respectively. A preliminary interpretation suggests that the present receiver array does

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not have resolution for R1, although the time-lapse change due to R1 is statistically significant, while the time-lapse change due to R2 is shown to be statistically insignificant, even though the receiver array has sufficient resolution for recording the time-lapse change.

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1. Introduction

A number of field examples have shown that volcanological knowledge is greatly enhanced by the application of electromagnetic (EM) methods, which can provide a heterogeneous distribution of the electrical conductivity (or resistivity as its reciprocal) in the ground and its change over time. These electrical features are useful for characterizing a volcano and its activity, because of the large variability in the conductivity of volcanic rocks. For example, most volcanic rocks have extremely low conductivity in the dry state (typically lower than 0.001 S/m); however, the conductivity can be enhanced by several orders of magnitude by the presence of water and/or high temperature, etc. Moreover, laboratory experiments have shown that the conductivity of molten basalt is as high as 1 S/m or higher (e.g. Tyburczy and Waff, 1983), which implies that EM methods are promising tools for monitoring the presence of molten magma and its movement.

The bulk conductivity of rocks saturated with pore water is known to be controlled by the conductivity of pore water, which is usually higher than that of volcanic rocks by several orders of magnitude (e.g., Quist and Marshal, 1968), a relationship known as Archie's law. Geothermal or non-thermal aquifers are often found in volcanic regions, and their interaction with a volcanic heat source is considered to be one of the most important factors controlling volcanic activity (e.g., Kagiya et al., 1999). Therefore, EM methods, such as the natural source magnetotelluric (MT) method and the direct current (DC) resistivity method, have often been applied in recent years for the study of volcanic structures (e.g. Hoffmann-Rothe et al., 2001; Matsushima et al., 2001; Manzella et al., 2004; Nicollin et al., 2006). Most past studies have only attempted to investigate the electrical structure of volcanoes; however, a limited number of studies (e.g. Zlotnicki et al., 2003) have been concerned with time variations in the conductivity despite the volcanological importance. It is considered that technological difficulties of long-term operations with high accuracy have mainly prevented most workers from attempting to monitor the time evolution of volcanic structures in terms of the electrical conductivity.

A number of geophysical and volcanological studies have been carried out on Izu-Oshima island, a basaltic volcano along the Izu-Bonin volcanic chain about 100 km southwest of Tokyo, Japan (Fig. 1(a) and (b)). Among the various kinds of measurements, the DC resistivity measurement at the central cone, Mihara-yama (or Mt. Mihara), provides a rare example of successful detection of changes in the subsurface structure over time that are associated with the volcanic activity. This measurement, first attempted in 1975 by Yukutake et al. (1978), was also in operation during the 1986–1987 activity (Yukutake et al., 1990) and continues today. A remarkable time-lapse change in the apparent resistivity was observed for approximately six months until the onset of the 1986 summit eruption. Utada (2003) attempted to interpret this observation in terms of time-lapse changes in the underground conductivity structure by three-dimensional (3-D) numerical modeling. As a possible interpretation, Utada (2003) suggested that the observed time variation can be explained by the ascent of the head of a very conductive (2 S/m) body below the summit crater. The time-lapse change also implied that a significant amount of magma with a volume of approximately $5 \times 10^6 \text{ km}^3$ should have been present just below the crater a few months before the summit eruption, and this volume was shown to be in good agreement with that of the drain-back magma estimated by repeated gravimetry (Watanabe et al., 1998).

Although this result from Izu-Oshima has shown the potential of an EM method to monitor the pre-eruption and eruption processes, especially at shallow depths down to 1 km or so, it is quite difficult to further improve the monitoring capability of the DC resistivity measurement system, especially its spatial resolution. This is because improvement of the resolution in both the lateral and vertical directions requires an increase in the number of receiver dipoles (a pair of grounded electrodes connected to a recording instrument by a cable approximately 200 m long in the present case) distributed over a wide area including the area of interest, while it is easy to imagine the difficulty of maintaining such a system for years. Thus, an alternative system has been anticipated.

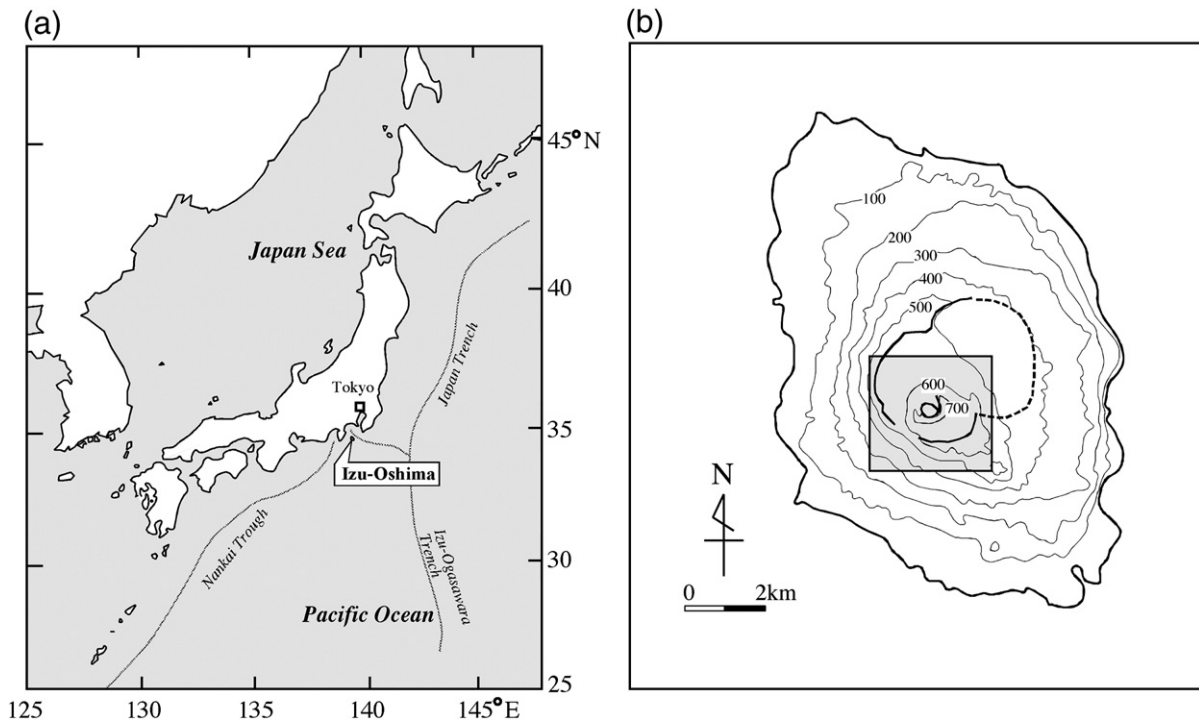


Fig. 1. Location (a) and topographical map (b) of the Izu-Oshima volcano. The shaded area in (b) indicates the central part of the island, a detailed map of which is shown in Fig. 5. Contours give elevation in meters. The rim of the caldera and crater is shown by a thick line, but it is shown by a dashed line where the rim is not visible on the surface topography.

Considering the requirements for accuracy and resolution, the application of controlled source EM induction methods (Kaufman and Keller, 1983) is expected to be effective, as it provides high accuracy due to the controlled source and the sufficient resolution in the vertical direction, without increasing the number of receivers owing to the skin effect (the frequency dependence of EM response functions can resolve the vertical layering of conductivity). There are examples of volcanological applications of such a method in the literature (Lenat et al., 2000; Savin et al., 2001; Bibby et al., 2002; Commer et al., 2005; Srigutomo et al., *in press*). However, achieving long-term continuous operation in a volcanic environment remains a challenging problem to be overcome. Here we propose a transient EM system to be applied for the monitoring of active volcanoes, which is a variation of the LOTEM (Long Offset Transient Electromagnetics) method that was originally designed for deep crustal soundings (Strack et al., 1990; Hordt et al., 2000). This paper aims to provide a theoretical basis for monitoring time-lapse changes in 3-D conductivity, and discusses the efficiency of the proposed system for monitoring the activity of the Izu-Oshima volcano by showing the results of numerical and field experiments.

2. System configuration

The proposed system is named ACTIVE (Array of Controlled Transient-electromagnetics for Imaging a Volcanic Edifice). The ACTIVE monitoring system consists of a transmitter that generates a controlled transient electromagnetic field and a number of receivers, each of which detects and records the vertical component of the transient magnetic field (Fig. 2). All units of the system are programmed for automatic operation at certain time intervals.

The transmitter (Fig. 3(a)) is composed of an electronics unit that generates a controlled current signal, a clock synchronized by a global positioning system (GPS) signal, eight 12 V batteries powered by a wind generator or solar panels, two grounded electrodes, and an insulated single metallic cable that connects the electrodes to the transmitter and carries the transmitted current. The receiver (Fig. 3(b)) consists of a sensor which is a 1 m long solenoidal induction coil buried vertically in the soil, an amplifier-recorder unit with a GPS synchronized clock, one 12 V battery, and a solar panel. Only the vertical component of the transient magnetic field is measured at each receiver site, although any of three components

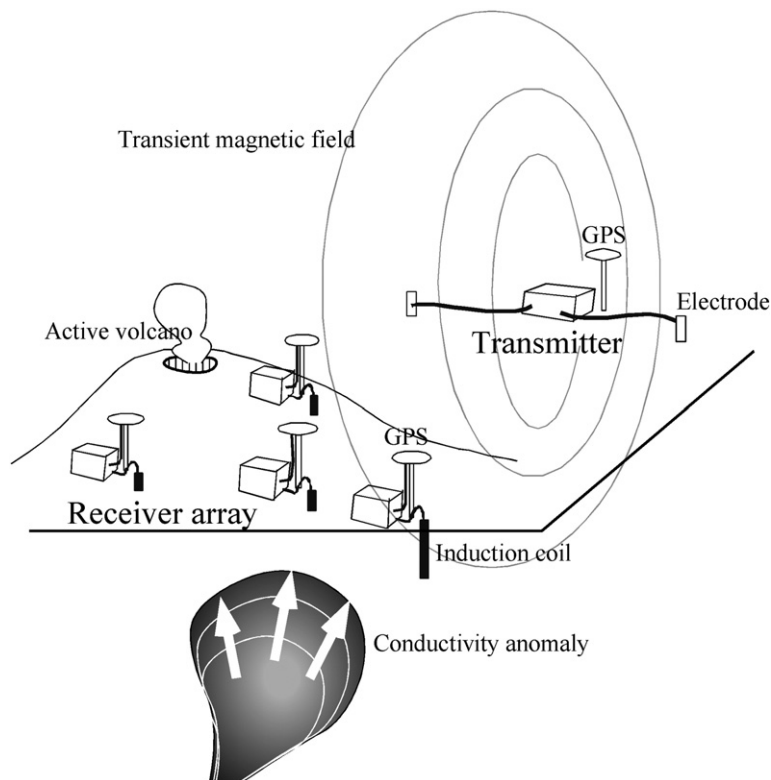


Fig. 2. Schematic showing the concept of the ACTIVE system. A transmitter with a grounded electric dipole generates a transient magnetic field which causes electromagnetic induction in the conductive Earth. If there is a conductivity anomaly that changes with time related to the volcanic activity, the time-lapse changes in the induction responses can be detected by receivers distributed around the active crater.

(magnetic and two electric field components) can be used as the response function to infer conductivity in the ground structure. There are two major reasons for this simple receiver design: (1) a simpler system reduces the frequency of problems, and (2) using fewer components reduces power consumption. We may increase the number of magnetic components for each receiver site, as this can improve spatial resolution and imaging accuracy. However, electric field measurements are not included, because our experience of DC resistivity monitoring for nearly 30 years shows the extreme difficulty of maintaining a measurement system with long grounded wires in a volcanic environment.

Accurate and precise measurement of the transmitter current, $j(t)$, is crucial for the overall performance of the present system. It is measured as a voltage difference over a standard in-series $2\ \Omega$ resistor, which is inserted into the current circuit. The signal is received as the electromotive force or output voltage from the induction coil, which is proportional to a time derivative of the magnetic field component in an ideal case. However, for an actual receiver system, the frequency characteristics are more complex than a simple time derivative. There-

fore in the present system, a precise calibration was made before installation to obtain the frequency dependence of each receiver system, so that the recorded signal time series could be deconvolved to the time variation of the magnetic field. Thus, received signals are treated as a time series of the magnetic field in the following derivation.

Both transmitted and received signals are simultaneously sampled by a 16-bit A/D (analog/digital) converter at 1000 Hz, and the timing for sampling is controlled by the GPS synchronized clock. Therefore, neither highly accurate control nor stability is required for either the waveform or the intensity of the transmitter current. If we let the transmitter current waveform be $J(\omega)$, the waveform of the vertical component of the magnetic field $H_z(\mathbf{r}_i; \omega)$ at a receiver site $\mathbf{r}_i$ can be written in the frequency domain as,

$$H_z(\mathbf{r}_i; \omega) = R(\mathbf{r}_i; \omega) \cdot C_i(\omega) \cdot J(\omega) + N(\mathbf{r}_i; \omega) \quad (1)$$

where $R(\mathbf{r}_i; \omega)$ and $C_i(\omega)$ represent the response function of the Earth at the angular frequency ω that contains the information for the underground electrical conductivity,

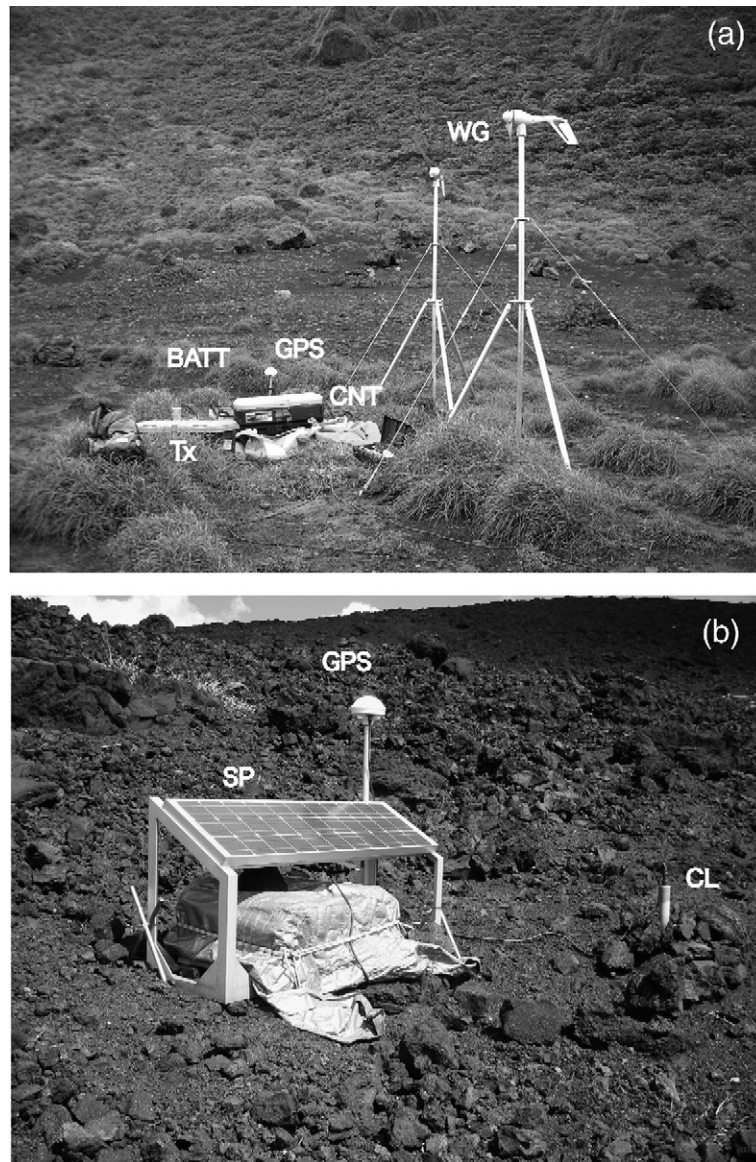


Fig. 3. (a) A view of the transmitter site (TRM). Wind power generators (WG), main transmitter unit (Tx) and controller (CNT), a GPS antenna (GPS) and eight 12 V batteries (BATT) are placed on the ground. Each of the two wind power generators has a capacity of 400 W. (b) A view of one of the receiver sites (C1). An induction coil (CL) is buried vertically in the soil, and is connected to the receiver unit placed inside the frame of a 24 W solar panel (SP). The GPS signal is received by an antenna (GPS) for time synchronization.

and the system response that includes the coil and filter characteristics, respectively, and $N(\mathbf{r}_i; \omega)$ is noise. The system response is accurately determined by performing a calibration of each receiver unit. Because of the clock synchronization by the GPS signal, the error in time between any pair of simultaneous time series is negligibly small compared to the sampling interval (1 ms).

$H_z(\mathbf{r}_i; \omega)$ and $J(\omega)$ in Eq. (1) can be obtained by a Fourier transform of a stacked time series of field measurement records. As shown in Fig. 4, the transmitter

current $j(t)$ is an alternating direct current with a 50% duty cycle with the period $T_0=1$ s and amplitude of approximately 1 A. The received signal is generally contaminated with much more noise, and therefore requires noise reduction for the accurate estimation of $H_z(\mathbf{r}_i; \omega)$. Therefore, in data processing, a weighted stacking method (Nagao et al., 2004) is applied for noise reduction and for the accurate estimation of the response function $R(\mathbf{r}_i; \omega)$ at each receiver site. Dividing the whole time series of the received signal into segments

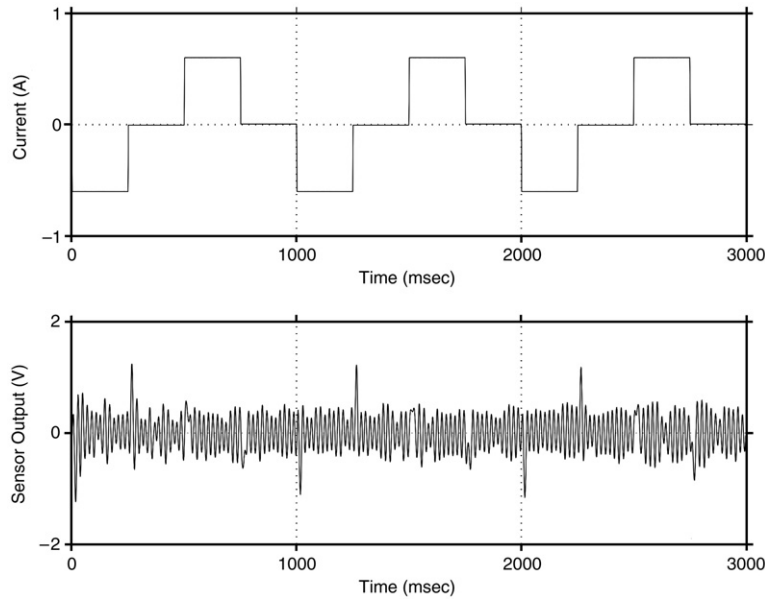


Fig. 4. Waveforms of the transmitted current (above) and received signal (below).

$h_z^{(k)}(\mathbf{r}_i; t)$, ($k=1, \dots, K$) with a transmitter period T_0 ($T_0=1000$ ms in the present case so that $K=3600$), the stacked signal can be derived by

$$\bar{h}_z(\mathbf{r}_i; t) = \frac{1}{W_i} \sum_{k=1}^K w_k(\mathbf{r}_i) h_z^{(k)}(\mathbf{r}_i; t), \quad t = 0, 1, \dots, T_0 - 1 \quad (2)$$

where $W_i = \sum_{k=1}^K w_k(\mathbf{r}_i)$, w_k is the stacking weight for each data segment at the receiver site $\mathbf{r}_i$, and t is the time in ms. Here the weight is chosen as

$$w_k(\mathbf{r}_i) = \frac{1}{s_k(\mathbf{r}_i)^2} \quad (3)$$

where $s_k(\mathbf{r}_i)^2$ is the noise variance of each data segment of the receiver site. The most prominent noise in the raw time series (Fig. 4) is monochromatic noise (at 50 Hz and its higher harmonics in this case), which can effectively be reduced by taking the difference between the former and latter half periods of each data segment (Kanda et al., 1996). Because of the symmetry of the signal, the variance of noise remaining in each segment can be estimated by

$$s_k(\mathbf{r}_i)^2 = \sum_v |H_z^{(k)}(\mathbf{r}_i; f_v)|^2, f_v = 2, 4, \dots, f_c \text{ Hz} \quad (4)$$

where $H_z^{(k)}(\mathbf{r}_i; f_v)$ is a complex Fourier transform of the observed signal at a frequency f_v , and f_c is the cut-off

frequency of the receiver amplifier filter (120 Hz in the present case). The background noise level can be estimated by Eq. (4), because the signal is composed of line spectra at a fundamental frequency of 1 Hz and its odd higher harmonics. By applying the same weight Eq. (3) to a simultaneous record of the transmitted current, the stacked current waveform is obtained by

$$\bar{j}_i(t) = \frac{1}{W_i} \sum_{k=1}^K w_k(\mathbf{r}_i) j^{(k)}(t), \quad t = 0, 1, \dots, T_0 - 1 \quad (5)$$

where $j^{(k)}(t)$ is a segment of the entire time series of the transmitted current. The response function in Eq. (1) can be obtained by converting stacked signals, $\bar{h}_z(\mathbf{r}_i; t)$ and $\bar{j}_i(t)$, to the frequency domain. Note that the stacked current waveform should be independently estimated for each site, because there is a fluctuation in current amplitude during the period of signal transmission while the noise variance may vary among receiver sites.

3. Field experiment

In the proposed monitoring system, the EM field is transmitted through a grounded dipole (a cable grounded at a pair of electrodes), so a good electric contact between the electrodes and the ground is important. However, this is usually not easily achieved in a volcanic environment where the ground surface is covered by non-conducting

materials such as lava, scoria, pumice, and so forth; therefore, installing a well-grounded transmitter dipole is one of the major difficulties in applying controlled source EM methods. In the present case of the Izu-Oshima volcano, a transmitter dipole has to be installed within a distance of approximately 1 km from the center of the crater, taking the depth and location of the target into account. A high-frequency MT survey (Utada and Shimomura, 1990) revealed that the typical conductivity of the thick scoria layer, which mostly covers the surface of the southern part of the study area, is 1×10^{-4} S/m. The northern part is covered mostly by recent lava flows and is not appropriate for electrode installation. Finally, the southeastern part of the caldera floor near the rim was selected as the transmitter location (Fig. 5), where the MT result indicated the presence of a less resistive (ca. 5×10^{-4} S/m) surface layer. Copper wire netting of 1×15 m² in size was buried approximately 1 m deep in the soil (fine and partly altered scoria) as an electrode. Two electrodes, separated by approximately 700 m, were connected to the transmitter unit installed between them by a cable. The total resistance of the transmitter dipole was approximately 500 Ω , including the contact resis-

tance at the electrodes and the cable resistance. The cable was buried 20–30 cm deep in the soil to protect it from damage. Five receiver units were installed around the summit crater of the central cone of Mihara-yama (or Mt. Mihara) as shown in Fig. 5. The measurement is programmed to function for 1 h from 03:00 to 04:00 AM local time once every 2 days. This time window was chosen so that contamination by man-made noise would be minimal.

The field monitoring experiment was started on July 25, 2002. At the beginning of this field experiment, both the transmitter and receivers were only equipped with a recorder which registers digital data into a compact flash memory card with a capacity of 512 MB. Later in 2005, both the transmitter and receivers were connected to a wireless internet system (IEEE802.11b). There are three wireless HUB stations equipped with non-directional antenna that form a wireless LAN (local area network) system in the Izu-Oshima caldera, which has been in operation since November 2005. This LAN system was designed so that any observations anywhere in the caldera, such as seismic, geomagnetic and GPS observations as well as the ACTIVE monitoring, can be linked

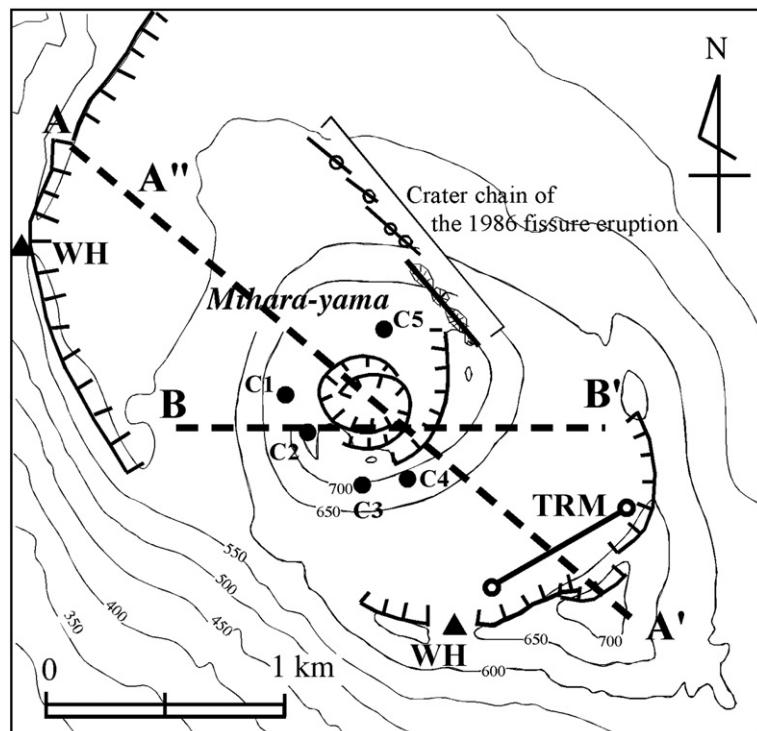


Fig. 5. Locations of the transmitter (TRM) and five receiver (C1–C5) sites in the Izu-Oshima caldera. All receiver sites are located on the crater floor of the central cone, Mihara-yama. Two wireless HUB (WH) stations used for the data transmission are indicated by black triangles. The 1986 eruption started from the southern wall of the pit crater, followed by a fissure eruption that formed a chain of small craters aligned in a SE–NW direction as indicated in the figure. Cross-sections of the electrical parameters along the dashed lines A–A' and B–B' are presented in later sections.

to the internet. All transmitter and receiver sites are equipped with a high gain directional antenna to connect to either of the two nearest HUB stations (Fig. 5), which are connected to the internet by another one-to-one wireless link or by a digital line. Thus, the wireless LAN system enables not only the collection of data in close to real time, but also remote control of the instruments (changing parameters etc.).

By processing each dataset as shown in the previous section, time variations of the response functions at five receiver sites have been obtained as presented in Fig. 6. The time-lapse changes in amplitude (left) and phase

(right) of the complex response function $R(\mathbf{r}_i; \omega)$ at each receiver site $\mathbf{r}_i$ ($i=1, \dots, 5$) are shown. The standard error in stacking the time series of the received signals was transferred to the error of the response function, because the error in estimating current is much smaller. The error is estimated as approximately 1% or less at frequencies lower than 50 Hz. The error at higher frequencies becomes 2–3% at sites C1, C2, and C5, but sometimes exceeds 5% at sites C3 and C4. The large estimation error at high frequencies for sites C3 and C4 is due to enigmatic noise synchronous to the transmitted signal (Takahashi, 2006).

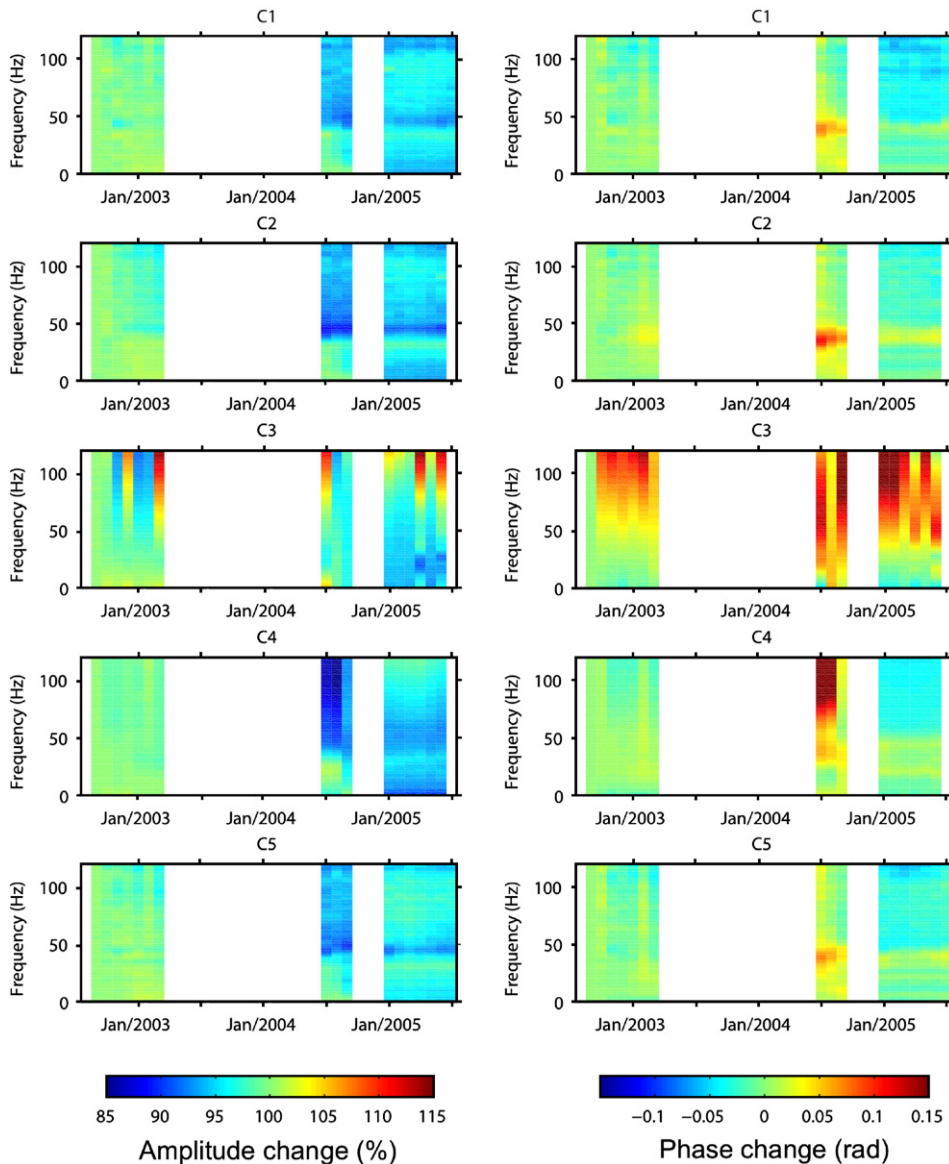


Fig. 6. Time-lapse changes of the amplitude ratio in % (left) and the phase in radians (right) for the observed responses relative to the values of the initial measurement on July 25, 2002.

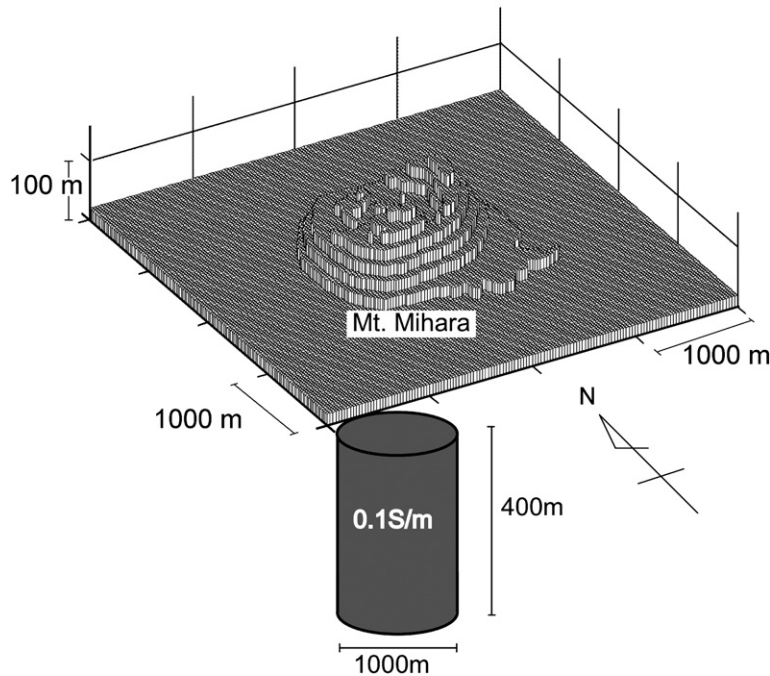


Fig. 7. Topography of the 3-D conductivity model of Izu-Oshima and the conductive body below the summit crater of Mt. Mihara.

Since the beginning of the monitoring experiment, all receiver units have been operating without any serious problems. However, there have been many problems with the transmitter due to lightning strikes, mechanical trouble with the wind generator, and so forth. Nearly 500 days, from July 2003 to November 2004, was spent reinforcing the electrical circuits for protection against lightning damage. During this period different types of wind generators were also tested, but finally the use of the wind generator was abandoned and it was replaced with solar panels. Since this modification, the entire measurement system has been in operation without any serious trouble. Data were also recorded for 3 months from May to July of 2004, but the quality of this interval was lower than that during other periods, because the circuit repair was still in progress.

4. Numerical experiments

Because the surface topography and subsurface structure are highly heterogeneous, a 3-D approach is indispensable for forward modeling and inversion in this study. This means the Maxwell equations,

$$\begin{aligned} \nabla \times \mathbf{E}(\mathbf{r}; \omega) &= -i\omega\mu\mathbf{H}(\mathbf{r}; \omega) \\ \nabla \times \mathbf{H}(\mathbf{r}; \omega) &= \sigma_{3D}(\mathbf{r})\mathbf{E}(\mathbf{r}; \omega) + \mathbf{J}_S(\mathbf{r}; \omega) \end{aligned} \quad (6)$$

have to be solved for a 3-D heterogeneous conductivity distribution $\sigma_{3D}(\mathbf{r})$. The Maxwell equation given in Eq. (6) is a frequency domain expression in which the current density due to the transmitter $\mathbf{J}_S(\mathbf{r}; \omega)$ is given and the displacement current is neglected. The magnetic permeability μ is assumed to be constant and equal to that of a vacuum. $\mathbf{E}(\mathbf{r}; \omega)$ and $\mathbf{H}(\mathbf{r}; \omega)$ are the electric and magnetic fields as a function of the position, $\mathbf{r}$, and the angular frequency, ω , respectively. If a transmitter current of unit amplitude is given in each calculation, the vertical component of the calculated magnetic field can be directly compared with the observed response. From the many available mathematical schemes, the integral equation (IE) method (e.g., Wannamaker, 1991) is employed. In the following derivation, the Cartesian

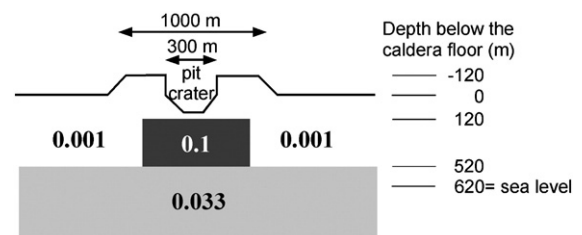


Fig. 8. A cross-section of the 3-D conductivity model.

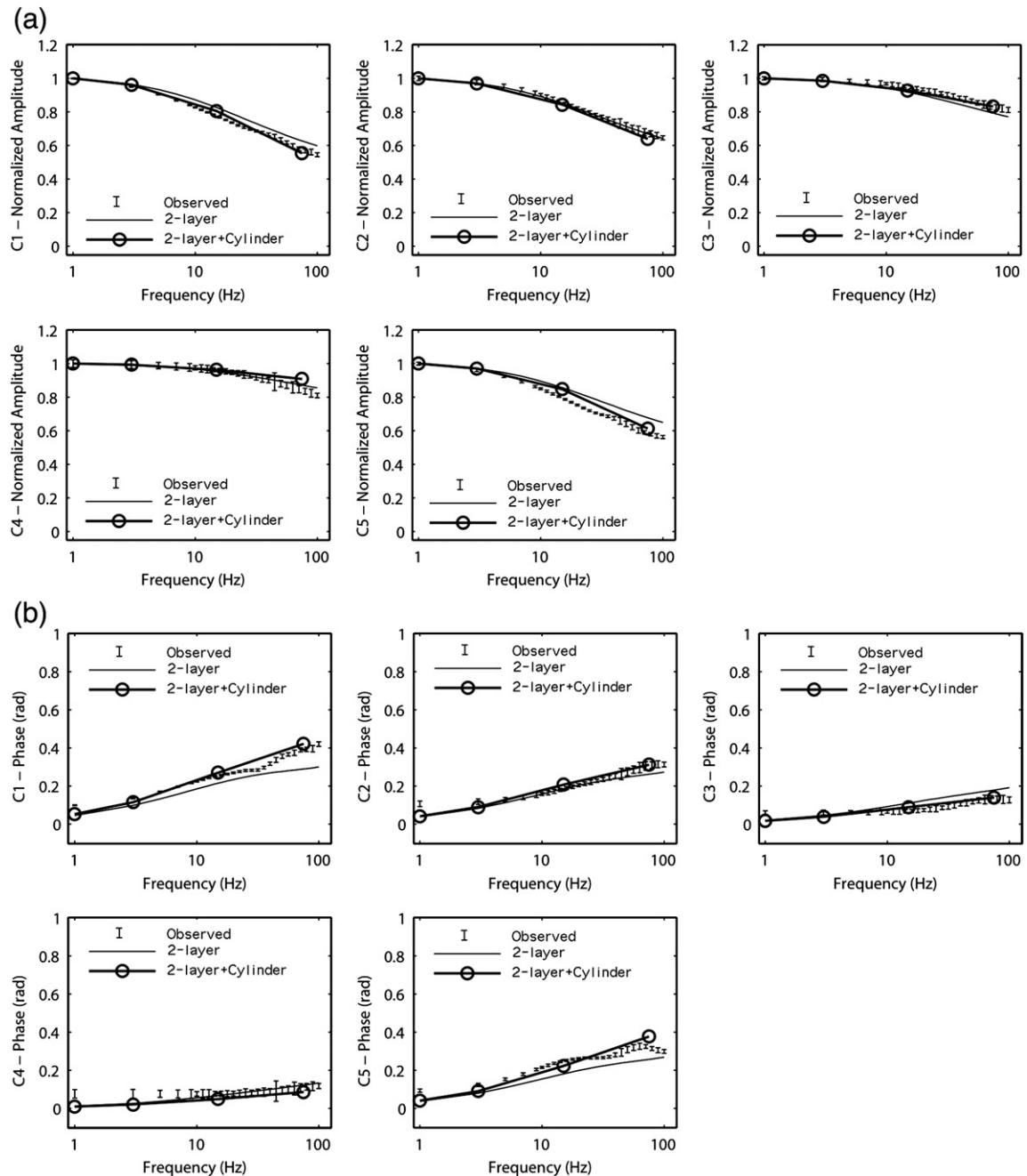


Fig. 9. (a) Frequency dependence of the observed amplitude responses (dots with error bars) and those calculated for the background (thin lines), and for the 3-D conductivity model with a cylindrical conductor (solid lines with open circles) shown in Fig. 8. The amplitude is normalized by the value at 1 Hz. (b) Frequency dependence of the observed and calculated phase responses, relative to the value at 1 Hz.

coordinate system is employed, in which the x - and y -directions are taken to be northward and eastward, and the z -direction is taken to be vertically downward. The center of the transmitter dipole is placed at the origin.

Any 3-D heterogeneous Earth structure can be mathematically separated into the following two components: a 1-D background conductivity distribution, $\sigma_{1D}(z)$, varying only in the vertical direction, and lateral perturbation from this distribution. If we let $\sigma_{3D}(\mathbf{r})$ be

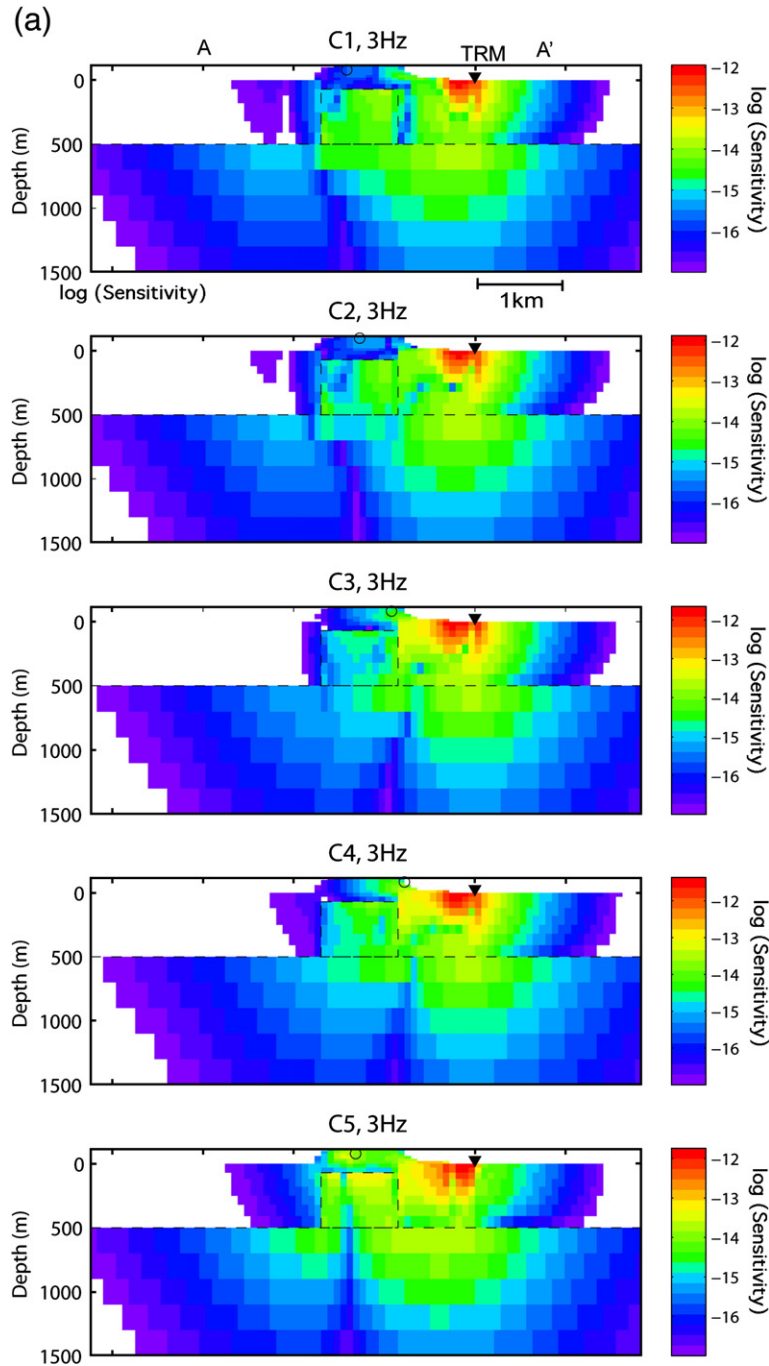


Fig. 10. Cross-sections (along A–A') of the sensitivity defined by Eq. (21) plotted in a logarithmic scale for frequencies of 3 Hz (a), 15 Hz (b), and 75 Hz (c), for receiver sites C1, C2, C3, C4 and C5, respectively. The triangle (TRM) indicates the location of the transmitter, while the open circle indicates the location of each receiver.

the 3-D heterogeneous electrical conductivity of arbitrary distribution, then the background 1-D structure, $\sigma_{1D}(z)$, can be estimated by horizontally averaging the conductivity in each appropriate depth slice. The later-

al perturbation of conductivity can be uniquely defined as

$$\delta\sigma(\mathbf{r}) = \sigma_{3D}(\mathbf{r}) - \sigma_{1D}(z) \quad (7)$$

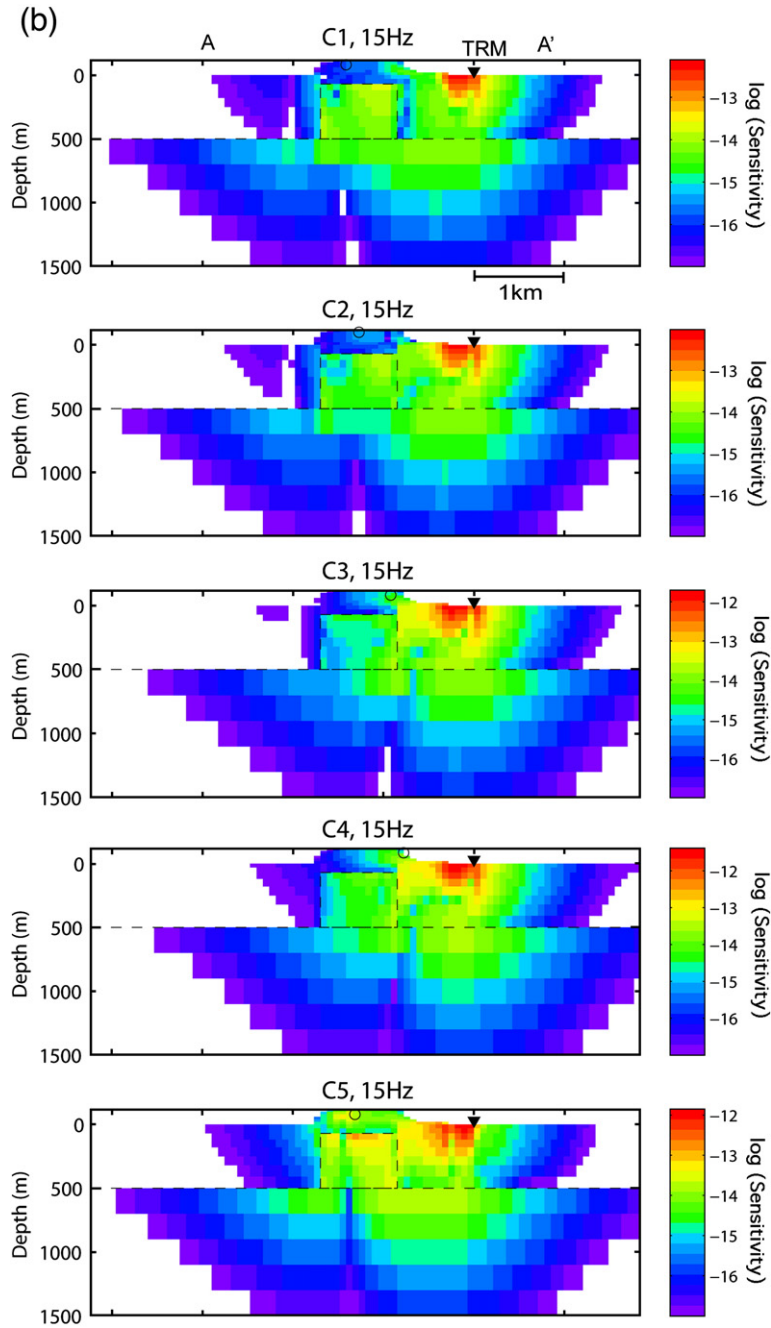


Fig. 10 (continued).

Then the frequency domain Maxwell equations in the 1-D background structure can be written as

$$\begin{aligned} \nabla \times \mathbf{E}_B(\mathbf{r}; \omega) &= -i\omega\mu\mathbf{H}_B(\mathbf{r}; \omega) \\ \nabla \times \mathbf{H}_B(\mathbf{r}; \omega) &= \sigma_{1D}(z)\mathbf{E}_B(\mathbf{r}; \omega) + \mathbf{J}_S(\mathbf{r}; \omega) \end{aligned} \quad (8)$$

where $\mathbf{E}_B(\mathbf{r}; \omega)$ and $\mathbf{H}_B(\mathbf{r}; \omega)$ are the electric and magnetic fields for the background structure. Solu-

tions for Eq. (8) can be represented by the integral form as

$$\begin{aligned} \mathbf{E}_B(\mathbf{r}; \omega) &= \int G^{EE}(\mathbf{r}, \mathbf{r}'; \omega) \mathbf{J}_S(\mathbf{r}'; \omega) d\mathbf{r}' \\ \mathbf{H}_B(\mathbf{r}; \omega) &= \int G^{EM}(\mathbf{r}, \mathbf{r}'; \omega) \mathbf{J}_S(\mathbf{r}'; \omega) d\mathbf{r}' \end{aligned} \quad (9)$$

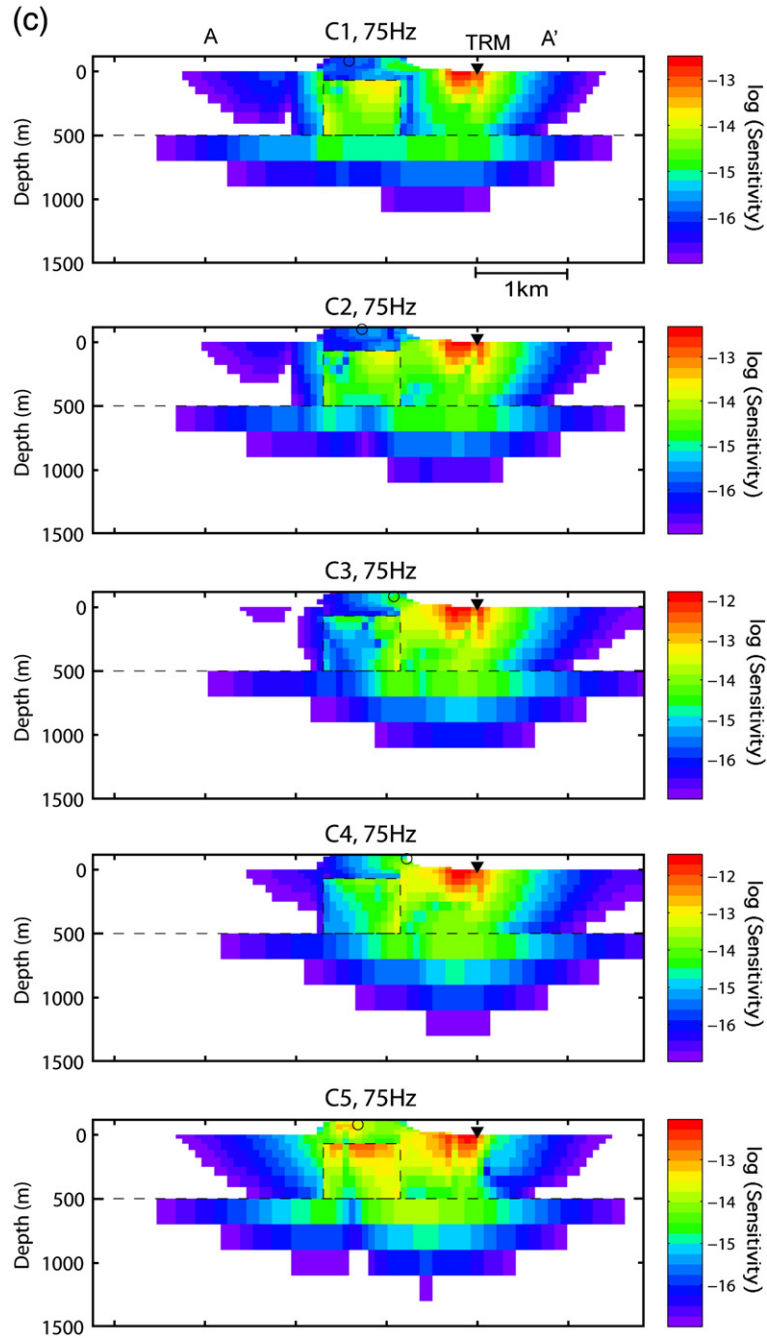


Fig. 10 (continued).

where $G^{EE}(\mathbf{r}, \mathbf{r}'; \omega)$ and $G^{EM}(\mathbf{r}, \mathbf{r}'; \omega)$ are Green's tensor (Weidelt, 1975) for the electric and magnetic fields excited by the electric dipole, respectively. Integration is carried out over the entire space where the source current density $\mathbf{J}_S(\mathbf{r}'; \omega)$ is distributed.

If the electric and magnetic fields in the 3-D heterogeneous structure, $\sigma_{3D}(\mathbf{r})$, are given by

$$\mathbf{E}(\mathbf{r}; \omega) = \mathbf{E}_B(\mathbf{r}; \omega) + \delta\mathbf{E}(\mathbf{r}; \omega)$$

$$\mathbf{H}(\mathbf{r}; \omega) = \mathbf{H}_B(\mathbf{r}; \omega) + \delta\mathbf{H}(\mathbf{r}; \omega) \quad (10)$$

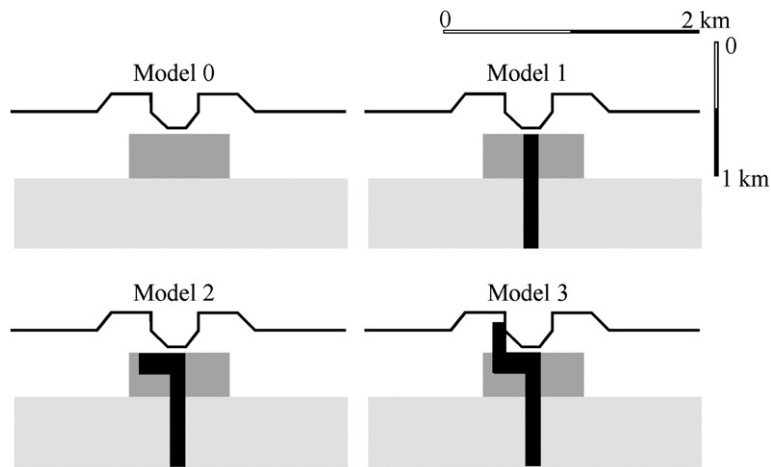


Fig. 11. Models for the conductivity change over time before the 1986 summit eruption (after Utada, 2003). Highly conductive regions of 2 S/m shown in black were estimated to account for the observed time changes in the DC apparent resistivity, as shown in models 1–3.

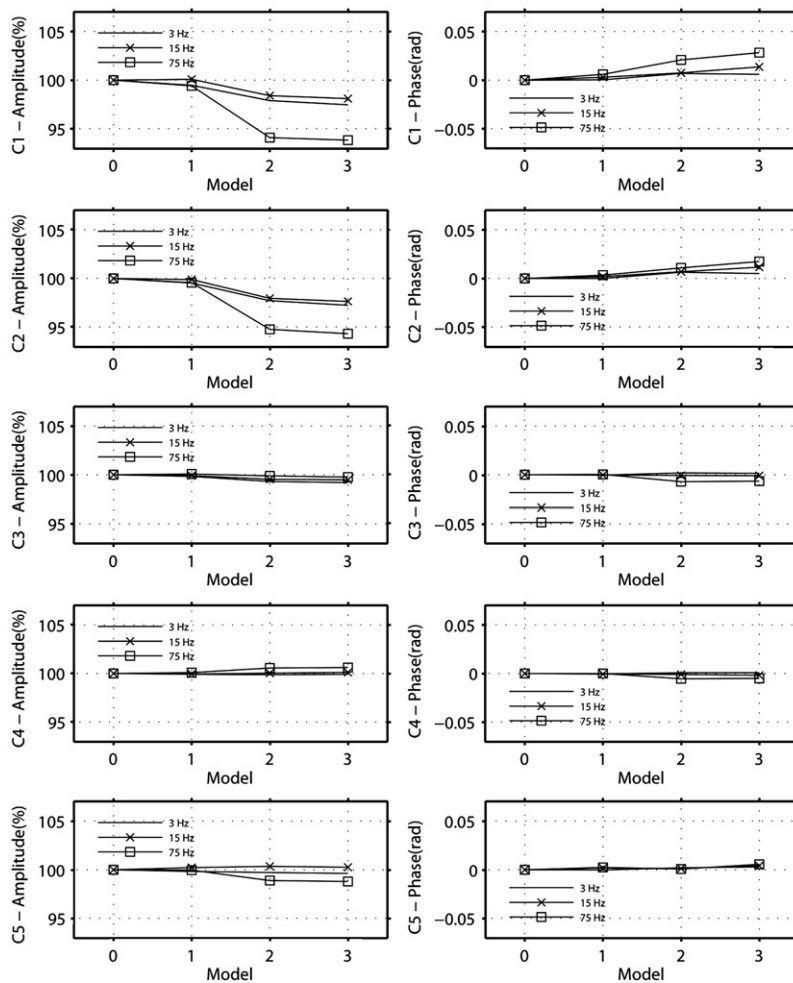


Fig. 12. Amplitude and phase responses predicted by models 0–3 at receiver sites C1–C5 for frequencies of 3, 15 and 75 Hz, respectively. The amplitude ratio and phase difference are shown relative to those of model 0.

where $\delta \mathbf{E}(\mathbf{r}; \omega)$ and $\delta \mathbf{B}(\mathbf{r}; \omega)$, respectively, are the perturbations of the electric and magnetic fields due to the heterogeneity, the Maxwell equations in a 3-D heterogeneous medium, Eq. (6), can be rewritten as

$$\begin{aligned}\nabla \times \{\mathbf{E}_B(\mathbf{r}; \omega) + \delta \mathbf{E}(\mathbf{r}; \omega)\} &= -i\omega\mu\{\mathbf{H}_B(\mathbf{r}; \omega) + \delta \mathbf{H}(\mathbf{r}; \omega)\} \\ \nabla \times \{\mathbf{H}_B(\mathbf{r}; \omega) + \delta \mathbf{H}(\mathbf{r}; \omega)\} &= \sigma_{3D}(\mathbf{r})\{\mathbf{E}_B(\mathbf{r}; \omega) + \delta \mathbf{E}(\mathbf{r}; \omega)\} + \mathbf{J}_S(\mathbf{r}; \omega)\end{aligned}\quad (11)$$

Subtracting Eq. (8) from Eq. (11) yields the following equations for the perturbation field:

$$\begin{aligned}\nabla \times \delta \mathbf{E}(\mathbf{r}; \omega) &= -i\omega\mu\delta \mathbf{H}(\mathbf{r}; \omega) \\ \nabla \times \delta \mathbf{H}(\mathbf{r}; \omega) &= \sigma_{3D}(\mathbf{r})\delta \mathbf{E}(\mathbf{r}; \omega) \\ &\quad + \{\sigma_{3D}(\mathbf{r}) - \sigma_{1D}(z)\}\mathbf{E}_B(\mathbf{r}; \omega) \\ &= \sigma_{1D}\delta \mathbf{E}(\mathbf{r}; \omega) + \delta \mathbf{J}(\mathbf{r}; \omega)\end{aligned}\quad (12)$$

Here the perturbation current density is introduced and is defined as

$$\begin{aligned}\delta \mathbf{J}(\mathbf{r}; \omega) &= \{\sigma_{3D}(\mathbf{r}) - \sigma_{1D}(z)\}\{\mathbf{E}_B(\mathbf{r}; \omega) + \delta \mathbf{E}(\mathbf{r}; \omega)\} \\ &= \delta \sigma(\mathbf{r})\mathbf{E}(\mathbf{r}; \omega)\end{aligned}\quad (13)$$

Using Green's tensor, the integral form of the electric perturbation field can be represented as

$$\begin{aligned}\delta \mathbf{E}(\mathbf{r}; \omega) &= \int G^{EE}(\mathbf{r}, \mathbf{r}'; \omega) \delta \mathbf{J}(\mathbf{r}'; \omega) d\mathbf{r}' \\ &= \int G^{EE}(\mathbf{r}, \mathbf{r}'; \omega) \delta \sigma(\mathbf{r}') \mathbf{E}(\mathbf{r}'; \omega) d\mathbf{r}' \\ &= \mathbf{E}_0(\mathbf{r}'; \omega) + A^{EE} \delta \mathbf{E}(\mathbf{r}'; \omega)\end{aligned}\quad (14)$$

where A^{EE} is an operator defined by

$$A^{EE}(\mathbf{r}; \omega) = \int G^{EE}(\mathbf{r}, \mathbf{r}'; \omega) \delta \sigma(\mathbf{r}') \mathbf{E}(\mathbf{r}'; \omega) d\mathbf{r}' \quad (15)$$

and

$$\mathbf{E}_0(\mathbf{r}; \omega) = A^{EE} \mathbf{E}_B(\mathbf{r}; \omega) \quad (16)$$

Similarly, the integral form of the magnetic perturbation field can be expressed as

$$\begin{aligned}\delta \mathbf{H}(\mathbf{r}; \omega) &= \int G^{EM}(\mathbf{r}, \mathbf{r}'; \omega) \delta \mathbf{J}(\mathbf{r}'; \omega) d\mathbf{r}' \\ &= A^{EM} \mathbf{E}(\mathbf{r}; \omega)\end{aligned}\quad (17)$$

where A^{EM} is an operator defined by

$$A^{EM} \mathbf{E}(\mathbf{r}; \omega) = \int G^{EM}(\mathbf{r}, \mathbf{r}'; \omega) \delta \sigma(\mathbf{r}') \mathbf{E}(\mathbf{r}'; \omega) d\mathbf{r}' \quad (18)$$

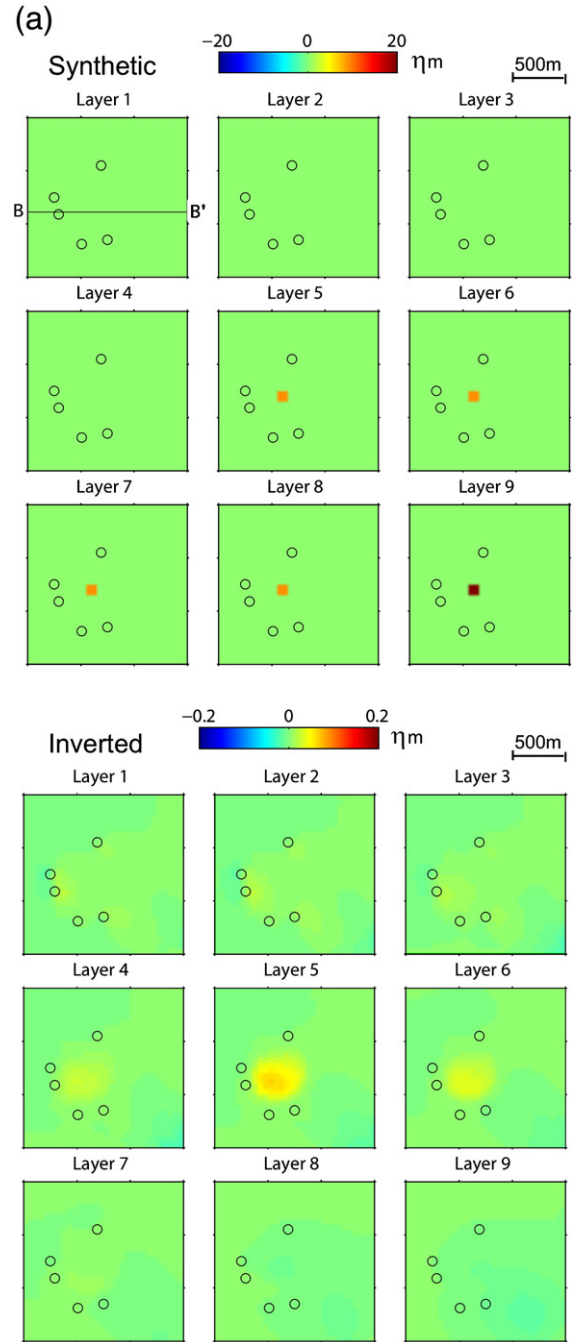


Fig. 13. Results of a synthetic test of the simple imaging method for a conductivity change from model 0 to model 1. (a) Plan views of synthetic conductivity distributions (above) and those inverted by the simple imaging method (below) at each depth slice. (b) A cross-section (along B–B') of synthetic conductivity distributions (above) and those inverted by the imaging method (below). The conductivity increase is shown in red. The locations of the receivers are indicated by open circles, and the transmitter is located at the bottom right corner of each plan view.

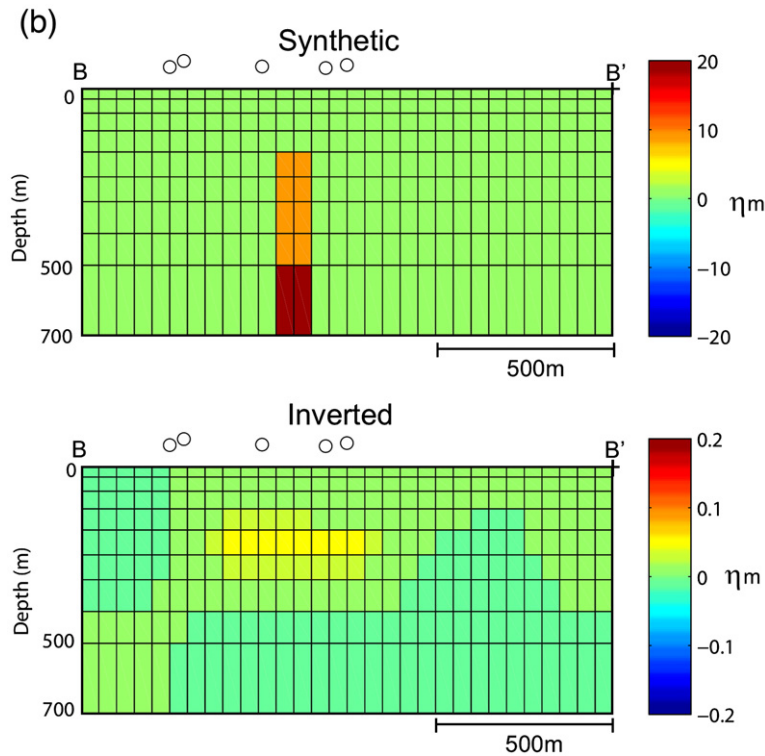


Fig. 13 (continued).

Among several solvers, the Modified Iterative Dissipative Method (Singer, 1995; Pankratov et al., 1995; Avdeev et al., 1997; Munekane, 2001) is employed in this study to solve the integral equations (Eqs. (14) and (17)). The ACTIVE system only concerns the vertical component of the magnetic field signal given by Eq. (17).

By separating the 3-D conductivity distribution into a 1-D background structure, σ_{1D} , and its lateral perturbation, $\delta\sigma$, the general 3-D Maxwell equations can be separated into a 1-D induction equation and a 3-D integral equation for the perturbation field. There are two major reasons for taking this approach, one mathematical and the other physical. From a mathematical viewpoint, it is possible to independently determine a model of 1-D structure, σ_{1D} , so that the variance of $\delta\sigma$ is smaller than that of σ_{3D} , which reduces the computational burden of solving the 3-D induction equation. On the other hand, the crustal conductivity is, in general, strongly affected by the presence of water that is supposed to have an almost horizontal distribution. In a volcanic environment, a geothermal anomaly at an active crater will cause lateral perturbation of the conductivity in an otherwise nearly horizontally stratified structure. Thus the separation of Eq. (7) is considered to be reasonable in the present situation.

The underground conductivity structure of the study area has been investigated by various exploration methods, such as DC resistivity (Yukutake et al., 1983), MT (Utada and Shimomura, 1990), CSAMT (controlled source audio-frequency magnetotellurics) (Ogawa and Takakura, 1990), and transient EM (Takahashi, 2006). Considering the previous results, as well as his own transient EM results, Takahashi (2006) obtained a model of the 3-D conductivity distribution beneath this area by a forward modeling study. This 3-D model consisted of the topography, the background 1-D structure, and a cylindrical conductive body beneath the summit crater. Fig. 7 shows a bird's-eye view of the topography and the location of the conductive body, while Fig. 8 shows an E–W cross-section of the model. The background structure is composed of two layers with conductivity of 0.001 S/m at the shallower part and 0.033 S/m at the deeper part. The boundary between these layers is at approximately 100 m above sea level, and the deeper conductive layer is interpreted as a water-bearing layer (Takahashi, 2006). The anomalous conductive body is a cylindrical columnar structure with a diameter of 1000 m, and its top surface is at approximately 120 m below the mean level of the caldera floor. The conductivity of this body is 0.1 S/m.

The Earth model was divided into 90 blocks in the x - and y -directions and 20 blocks in the z -direction. The topography of Mihara-yama and its summit crater were modeled by six-layered blocks divided in the z -direction. Each block was $50 \times 50 \text{ m}^2$ in the horizontal direction, and variable in the z -direction, from 20 m for those near the surface to 200 m for those deep in the Earth. Forward calculations were made for this 3-D model to predict the theoretical amplitude and phase responses at monitoring sites C1–C5, which were compared with the observed responses from the first measurement made on July 25, 2002 (Fig. 9(a) and (b)). It is clear that the model with a conductive anomaly below the crater provides a better fit to the observation than the 2-layered background structure. The RMS (root mean square) misfit between the observed and calculated responses at 3, 15, and 75 Hz was estimated to be 1.83 for the background structure, while it was reduced to 1.13 by adding a conductive anomaly.

5. Sensitivity map and an approximate imaging scheme

By using the 3-D conductivity model of Izu-Oshima thus obtained, the sensitivity of each of the five receivers was evaluated as follows: In the above formulation, the 3-D conductivity structure $\sigma_{3D}(\mathbf{r})$ is expanded into discrete form by

$$\sigma_{3D}(\mathbf{r}) = \sum_{m=1}^M \sigma_m \psi_m(\mathbf{r}) \quad (19)$$

where σ_m and $\psi_m(\mathbf{r})$ are the conductivity (a real constant) and basis function, respectively, and M is the number of blocks of discretization. In this study, the target region is divided into rectangular blocks, v_m ($m=1, \dots, M$), and the basis function is defined as

$$\psi_m(\mathbf{r}) = \begin{cases} 1 & \mathbf{r} \in v_m \\ 0 & \mathbf{r} \notin v_m \end{cases} \quad (20)$$

The sensitivity, $S_m(\mathbf{r}_i; \omega)$, can then be defined as a variation of the response function (the vertical component of the magnetic field generated by a source current of unit intensity in the present case) at a site, $\mathbf{r}_i$, due to the small variation in the 3-D conductivity as

$$S_m(\mathbf{r}_i; \omega) = \frac{1}{V_m} \left| \frac{\partial H_z(\mathbf{r}_i; \omega)}{\partial \log \sigma_m} \right| = \frac{1}{V_m} \left| \sigma_m \frac{\partial H_z(\mathbf{r}_i; \omega)}{\partial \sigma_m} \right| \quad (21)$$

where

$$V_m = \int \psi_m(\mathbf{r}') d\mathbf{r}' \quad (22)$$

To evaluate the sensitivity given by Eq. (21), a reciprocity approach (McGillivray and Oldenburg, 1990; McGillivray et al., 1994), which is widely used in 2-D and 3-D induction problems (e.g., Rodi and Mackie, 2001), was employed. Using the reciprocity of

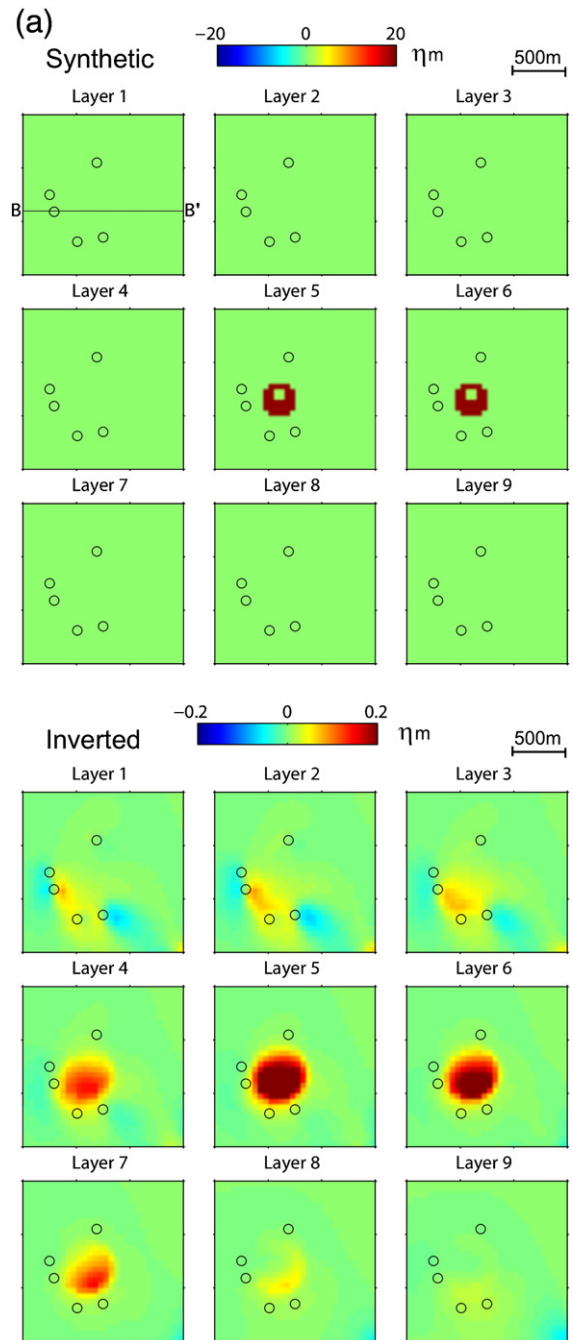


Fig. 14. Plan views (a) and cross-sections (b) of the synthetic test results for the imaging method of a conductivity change from model 1 to model 2.

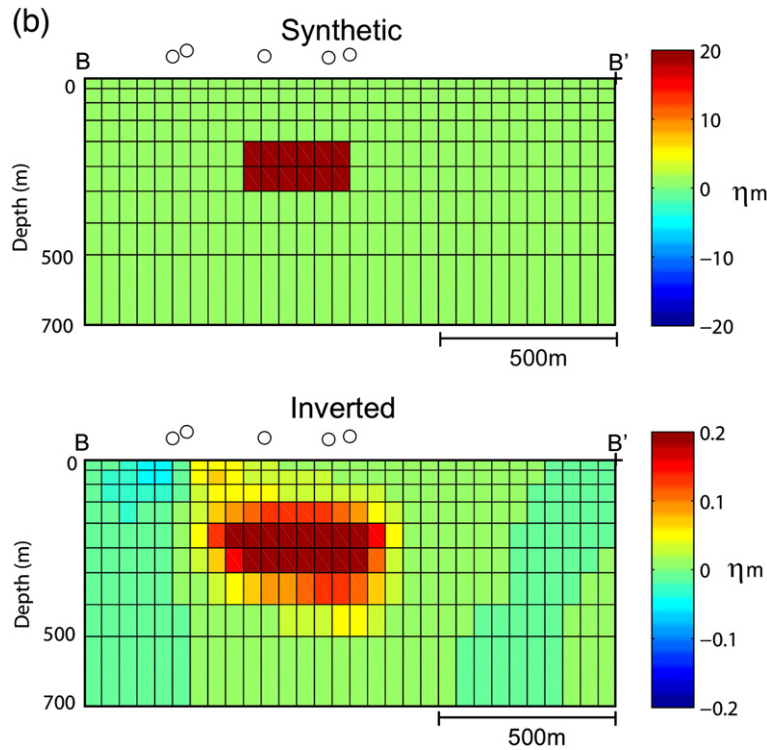


Fig. 14 (continued).

EM fields, the partial derivative in Eq. (21) is represented as

$$\frac{\partial H_z(\mathbf{r}_i; \omega)}{\partial \sigma_m} = \int \mathbf{E}^{(i)}(\mathbf{r}'; \omega) \cdot \mathbf{E}(\mathbf{r}'; \omega) \psi_m(\mathbf{r}') d\mathbf{r}' \quad (23)$$

where $\mathbf{E}(\mathbf{r}'; \omega)$ denotes the electric field excited by the actual transmitter, while $E^{(i)}(\mathbf{r}'; \omega)$ is the electric field excited by a pseudo-source located at the receiver site, $\mathbf{r}_i$. The pseudo-source is a virtual source with the same component as the receiver, i.e., a vertical magnetic dipole with a unit moment. For Eq. (23), $\mathbf{E}(\mathbf{r}'; \omega)$ can be calculated using Eqs. (10)–(16) given above. $\mathbf{E}^{(i)}(\mathbf{r}'; \omega)$ can be obtained by the following equation with a pseudo-source,

$$\begin{aligned} \mathbf{E}^{(i)}(\mathbf{r}'; \omega) &= \mathbf{E}_{1D}^{(i)}(\mathbf{r}'; \omega) \\ &+ \int G^{EE}(\mathbf{r}', \mathbf{r}''; \omega) \delta\sigma(\mathbf{r}'') \mathbf{E}^{(i)}(\mathbf{r}''; \omega) d\mathbf{r}'' \end{aligned} \quad (24)$$

$E_{1D}^{(i)}(\mathbf{r}'; \omega)$ denotes the electric field at $\mathbf{r}'$ excited by a pseudo-source at $\mathbf{r}_i$ in the 1-D background structure, and can be represented by elements of the Green tensor as

$$\mathbf{E}_{1D}^{(i)}(\mathbf{r}'; \omega) = \begin{bmatrix} G_{zx}^{ME}(\mathbf{r}_i, \mathbf{r}'; \omega) \\ G_{zy}^{ME}(\mathbf{r}_i, \mathbf{r}'; \omega) \\ G_{zz}^{ME}(\mathbf{r}_i, \mathbf{r}'; \omega) \end{bmatrix} \quad (25)$$

where $G_{zx}^{ME}(\mathbf{r}_i, \mathbf{r}'; \omega)$, for instance, denotes the X component of the electric field at $\mathbf{r}'$ excited by a unit vertical magnetic dipole source at $\mathbf{r}_i$. The reciprocal relation (e.g., Weaver, 1994; Newman and Commer, 2005) shows that the right-hand side of Eq. (25) can be rewritten as

$$\begin{bmatrix} G_{zx}^{ME}(\mathbf{r}_i, \mathbf{r}'; \omega) \\ G_{zy}^{ME}(\mathbf{r}_i, \mathbf{r}'; \omega) \\ G_{zz}^{ME}(\mathbf{r}_i, \mathbf{r}'; \omega) \end{bmatrix} = \begin{bmatrix} G_{xz}^{EM}(\mathbf{r}', \mathbf{r}_i; \omega) \\ G_{yz}^{EM}(\mathbf{r}', \mathbf{r}_i; \omega) \\ G_{zz}^{EM}(\mathbf{r}', \mathbf{r}_i; \omega) \end{bmatrix} \quad (26)$$

Green's tensor G^{EM} in the right-hand side of Eq. (26) is identical to the tensor appearing in Eqs. (9), (17) and (18).

Fig. 10(a), (b), and (c) shows distributions of the sensitivity of the receivers at sites C1–C5 in a cross-section along the line A–A' (Fig. 5) for frequencies of 3, 15, and 75 Hz, which represent low, middle and high frequencies, respectively. High and low sensitivity is indicated by red and blue, respectively. As clearly seen in these figures, each receiver has relatively higher sensitivity for more conductive structures, and the sensitivity decreases with increasing depth. Aside from the high sensitivity commonly observed for the shallower part between the transmitter and each receiver location, a relatively high sensitivity is also noticeable within the conductive body beneath the summit crater. Also site-to-site variation in the sensitivity distribution is noticeable.

Qualitatively speaking, therefore, the present receiver array is rather sensitive to a possible increase in conductivity within this body that is supposed to have happened before the 1986 summit eruption (Yukutake et al., 1990; Utada, 2003).

Next, we consider the time dependence for a particular σ_m in Eq. (19), with which the magnetic perturbation field in Eq. (17) becomes

$$\delta \mathbf{H}(\mathbf{r}_i; \omega; \tau) = [\sigma_m(\tau) - \sigma_{1D}(z_m)] \times \int G^{EM}(\mathbf{r}_i, \mathbf{r}'; \omega) \psi_m(\mathbf{r}') \mathbf{E}(\mathbf{r}'; \omega; \tau) d\mathbf{r}' \quad (27)$$

where z_m is the Z coordinate of the center of the corresponding block. Here we propose an approximate method to image the time variations in the conductivity using this relation. Subtracting the perturbation field at a time τ , from that at a later time τ' , we derive

$$\begin{aligned} \delta \mathbf{H}(\mathbf{r}_i; \omega; \tau') - \delta \mathbf{H}(\mathbf{r}_i; \omega; \tau) &= \int G^{EM}(\mathbf{r}_i, \mathbf{r}'; \omega) \psi_m(\mathbf{r}') \\ &\times \{ [\sigma_m(\tau') - \sigma_{1D}(z_m)] \mathbf{E}(\mathbf{r}'; \omega; \tau') \\ &- [\sigma_m(\tau) - \sigma_{1D}(z_m)] \mathbf{E}(\mathbf{r}'; \omega; \tau) \} d\mathbf{r}' \end{aligned} \quad (28)$$

A full 3-D inversion (e.g., Newman and Commer, 2005) is necessary to obtain the change over time, $\sigma_m(\tau') - \sigma_m(\tau)$, using this equation. However, Eq. (28) can be simplified as

$$\begin{aligned} \delta \mathbf{H}(\mathbf{r}_i; \omega; \tau') - \delta \mathbf{H}(\mathbf{r}_i; \omega; \tau) &= [\sigma_m(\tau') - \sigma_m(\tau)] \\ &\times \int G^{EM}(\mathbf{r}_i, \mathbf{r}'; \omega) \psi_m(\mathbf{r}') \mathbf{E}(\mathbf{r}'; \omega; \tau) d\mathbf{r}' \end{aligned} \quad (29)$$

if the approximation $\mathbf{E}(\mathbf{r}'; \omega; \tau') = \mathbf{E}(\mathbf{r}'; \omega; \tau)$ is allowed for a small variation in the conductivity (the Born approximation). If we perform a forward calculation of the integral in the right-hand side of Eq. (29) at a certain time τ , then Eq. (29) becomes a linear equation such as

$$\delta \mathbf{H}(\mathbf{r}_i; \omega; \tau') - \delta \mathbf{H}(\mathbf{r}_i; \omega; \tau) = A_{i,m} [\sigma_m(\tau') - \sigma_m(\tau)] \quad (30)$$

where

$$A_{i,m} = \int G^{EM}(\mathbf{r}_i, \mathbf{r}'; \omega) \psi_m(\mathbf{r}') \mathbf{E}(\mathbf{r}'; \omega; \tau) d\mathbf{r}' \quad (31)$$

In the actual imaging calculation, signals at discrete frequencies (3, 15, and 75 Hz in the present case) are chosen. To solve the linear matrix inversion given by

Eq. (30), we applied a data space approach (e.g., Parker, 1994; Siripunvaraporn and Egbert, 2000; Siripunvaraporn et al., 2005), because the number of data, N (the number of receivers times the number of frequencies), is much smaller than the number of parameters, M . It is thus possible to monitor time variations in the 3-D

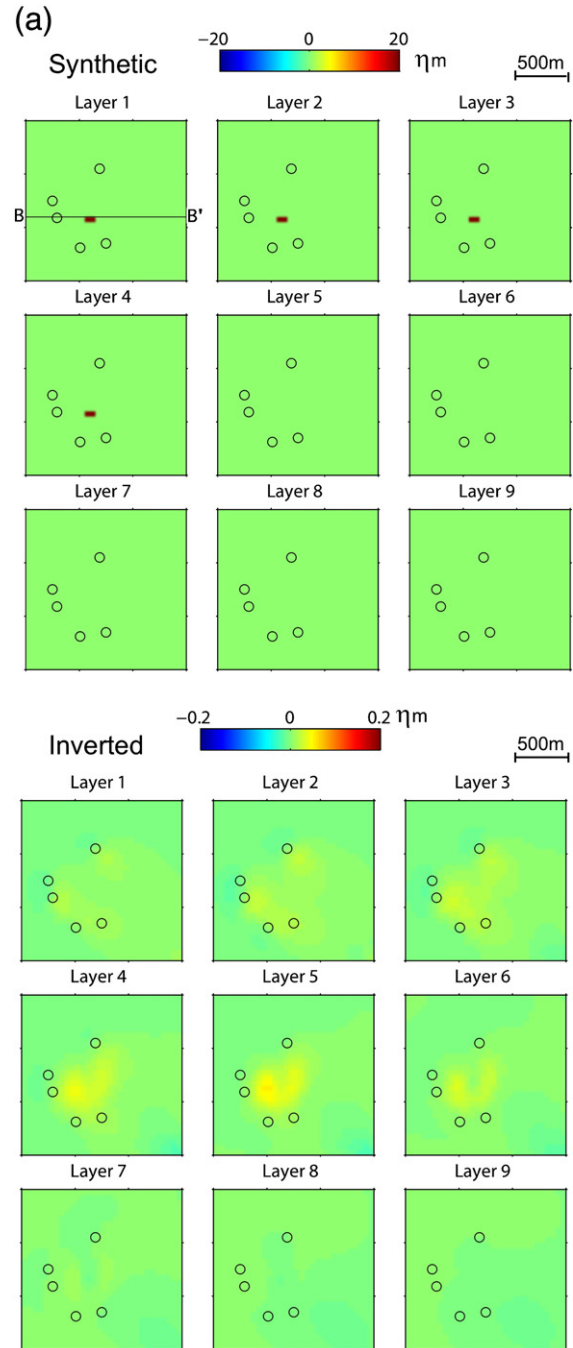


Fig. 15. Plan views (a) and cross-sections (b) of the synthetic test results for the imaging method of a conductivity change from model 2 to model 3.

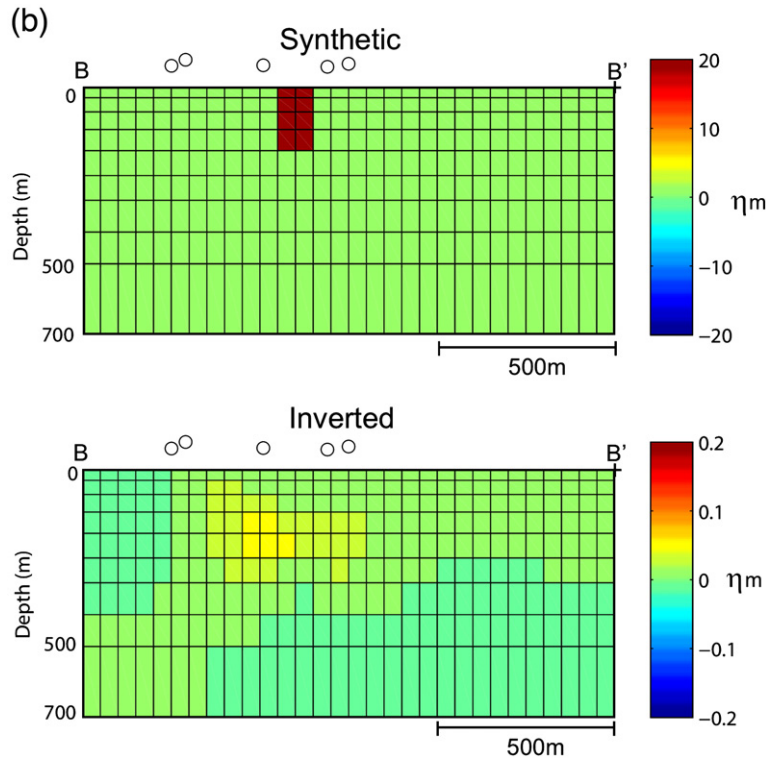


Fig. 15 (continued).

conductivity distribution by successively performing an approximate imaging given by Eq. (30).

6. Numerical results and discussion

Utada (2003) presented 3-D conductivity models (Fig. 11) that account for the time variations in the apparent DC resistivity of Mihara-yama observed before the 1986 summit eruption. Here we carry out full 3-D calculations to estimate ACTIVE responses for these models. As the position of each conductive body present in these models was not precisely determined in Utada (2003), for the present study, we assumed that sites C1 and C2 are located behind the body appearing in model 2 relative to the transmitter, where the sensitivity maps suggest that high sensitivity to conductivity below the crater is to be expected (Fig. 10(a), (b), and (c)). On the other hand, sites C3, C4 and C5 are either beside or in front of the body, where the sensitivity is expected to be much lower.

Fig. 12 shows the relative amplitude and phase responses expected at receiver sites C1–C5 at three frequencies, 3, 15 and 75 Hz, in the respective models. As shown in these diagrams, significant changes, both in amplitude and phase, of the response functions can be expected at C1 and C2 corresponding to model 2, especially at high frequencies. The changes in amplitude at sites C3 and C4

will be smaller than those at C1 and C2. The changes in phase at C3 and C4 are of similar magnitude to those at other sites, but will be of opposite sign. Taking the actual observation error into account, this forward modeling result indicates the feasibility of the present ACTIVE system for detecting underground conductivity changes in the pre-eruptive stage, if the eruption process follows a scenario similar to that observed in 1986.

The feasibility was further tested by performing imaging that converts these changes in model responses to changes in anomalous conductivity distribution. Amplitude and phase responses at 3 frequencies (3, 15 and 75 Hz) at 5 receiver sites were used as input data. Therefore the dimension, N , of the column vector in the left-hand side of Eq. (30) became 30. An error of 1% was assumed to be independent of frequency. The underground conductivity structure below the level of the caldera floor was divided into $90 \times 90 \times 9$ blocks, assuming that the conductivity above this level was invariable. This assumption is reasonable because the shallower structure is composed of highly resistive materials. Hence the dimension, M , of the column vector on the right-hand side of Eq. (30) was 72,900. Each imaging calculation was performed in approximately 10 s by a Linux workstation, except the calculations of Green's tensor and the initial electromagnetic fields,

which were much more time consuming, but only had to be performed once before the imaging calculations.

Fig. 13(a) shows a plan view of the given (above) and imaged (below) conductivity distributions at each of the 9 layers, while Fig. 13(b) shows cross-sections of the given (above) and inverted (below) views, both corresponding to the change from model 0 to model 1. Note that a different color scale is used for the given and inverted values of the parameter

$$\eta_m(\tau'; \tau) = \frac{\sigma_m(\tau') - \sigma_m(\tau)}{\sigma_m(\tau)} \quad (32)$$

which represents the relative time-lapse change of the discretized 3-D conductivity. Although the contrast is

weaker and the shape is flatter, a region of positive η_m (conductivity increase) was imaged at the location almost exactly where a change in conductivity was given. This means that the location of the image is accurate, but its value is not, probably because of the heavy non-linearity of the basic equation (the Born approximation is not a good approximation).

Fig. 14(a) and (b) are the same plots as Fig. 13(a) and (b), but for the change from model 1 to model 2, which provides a more impressive image than the previous case. A large conductive anomaly was imaged at around the given location with much stronger contrast than in the previous case. This can be recognized as a significant change and is sufficient to indicate that something unusual is occurring below the crater. This

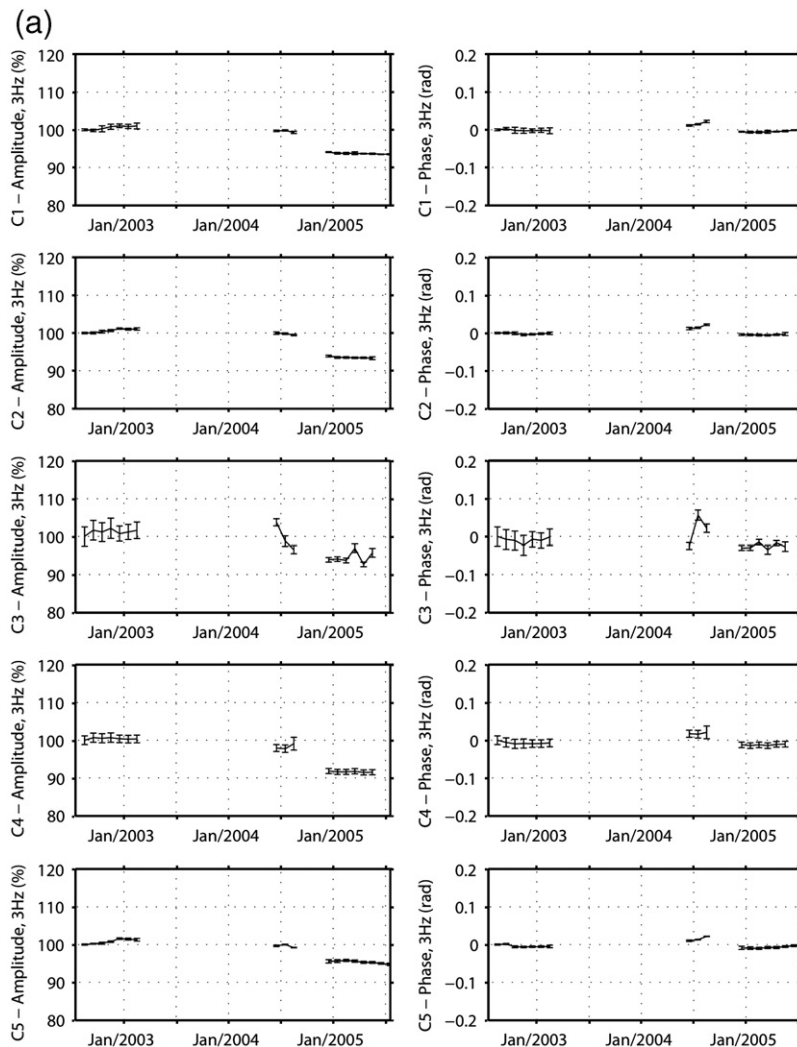


Fig. 16. Time-lapse changes in the amplitude and phase of the monthly mean for the responses at 3 Hz (a), 15 Hz (b), and 75 Hz (c) observed at C1, C2, C3, C4 and C5. The amplitude ratio and phase difference are shown relative to those of the initial measurement.

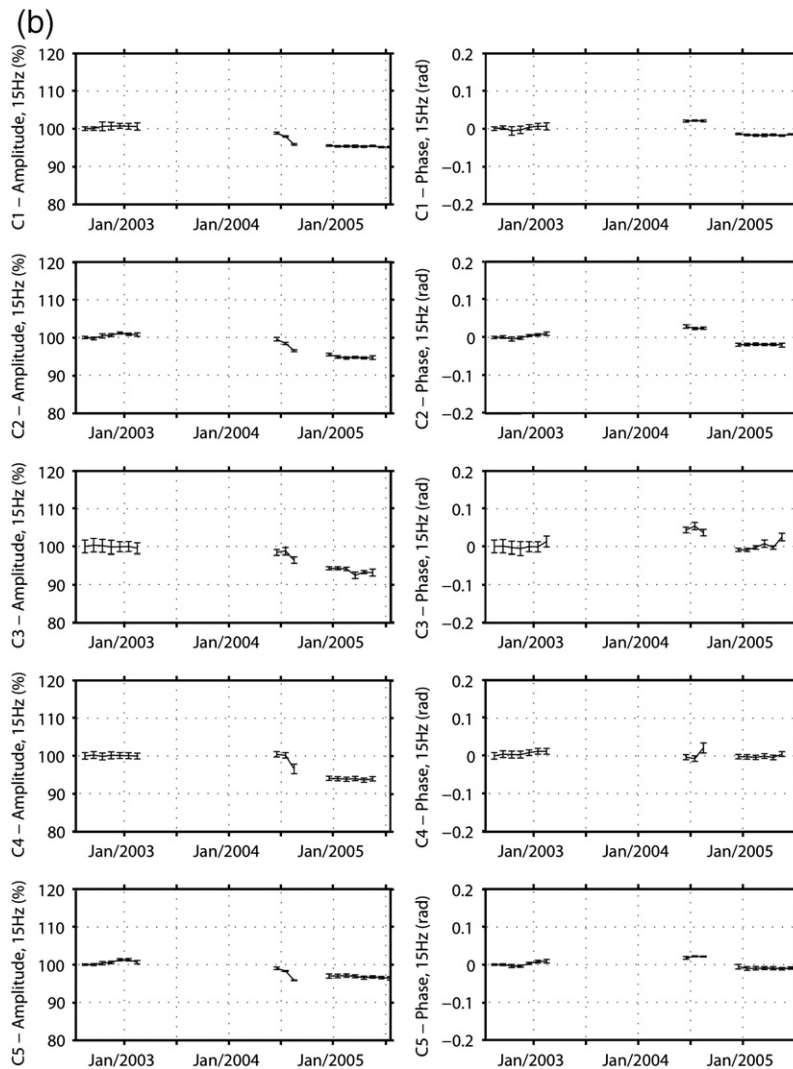


Fig. 16 (continued).

result is not surprising, because the conductive body that corresponds to this change is rather flat and occupies a large volume, so that it is effective in causing an induction anomaly as shown in Fig. 12. However, the imaged contrast is much weaker (similar to 1/100 of what was calculated) and gradates away toward the outside of the modeled conductive body. The presence of this feature is also ascribed to significant non-linearity. A full 3-D inversion is required to solve this problem to obtain a clear image and an accurate value for the conductivity change.

A rather blurred image was again obtained for the change from model 2 to model 3, as shown in Fig. 15(a) and (b). The location error is larger than that for the two previous cases, both horizontally and vertically. This is probably because the horizontal distance between the

receiver sites and the conductive body is larger than the depth of the body, which makes it difficult to image at the correct location. The installation of more receivers, especially closer to the target, would solve this problem, although such installation is not a simple task to complete at the actual location.

It is now possible for us to apply this imaging scheme to real data from Izu-Oshima as shown in Fig. 6. Because of the long data gap and the rather smooth time variations during the period in which data exists, we calculated the monthly mean for the responses at frequencies of 3, 15, and 75 Hz (Fig. 16). The standard error for these estimates was as small as 0.1% or so for low frequency responses at C1, C2, and C5. Large changes between the responses observed before and after the large data gap were recognized at all 5 sites.

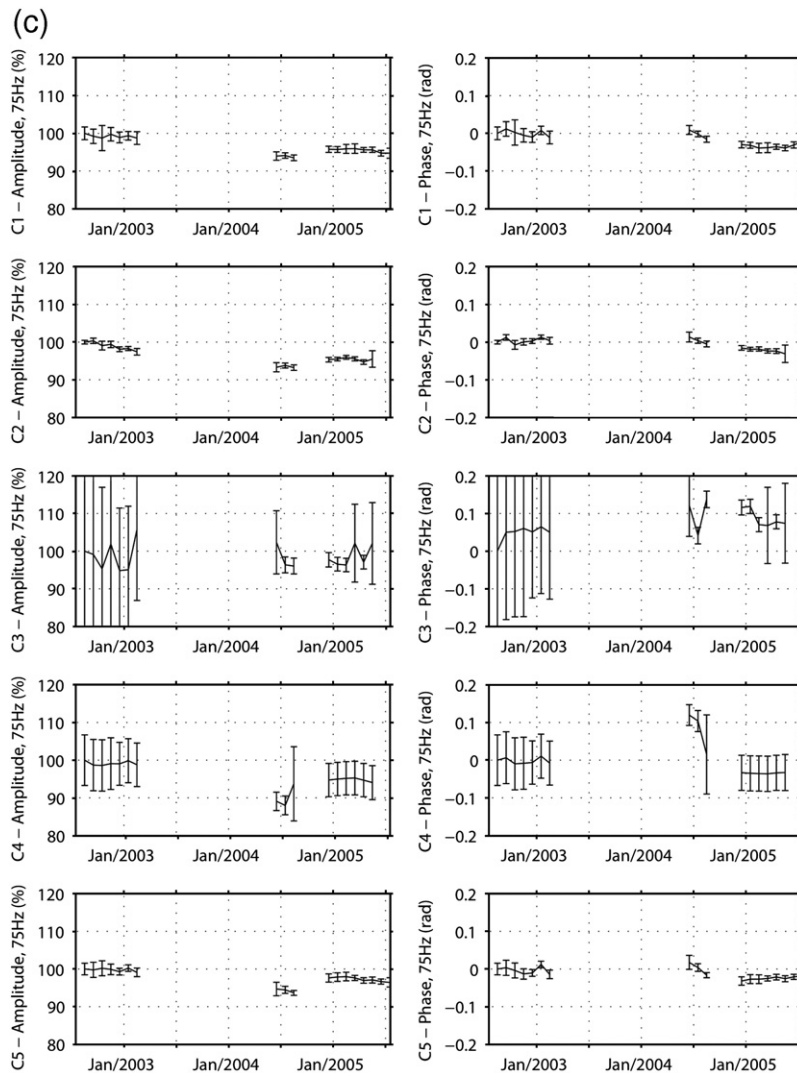


Fig. 16 (continued).

The numerical experiment performed using the 1986 eruption model shown above indicates that a conductivity increase below the summit crater will give locally variable effects between its northern (sites C1, C2, and C5) and southern (sites C3 and C4) sites. However, the observed response at each site commonly shows a decrease in amplitude with a negative phase change (with the exception of C3) where the data are rather noisy. Therefore, a qualitative analysis of these observed time-lapse changes suggests that their major cause is not conductivity changes below the crater.

To quantitatively examine the consequence of this qualitative analysis, we attempted to convert these time-lapse changes into anomalous conductivity images, by letting τ and τ' in Eq. (32) be the dates of the first (July

2002) and last (April 2005) measurements in Fig. 16. The dimensions of the input data and parameters are the same as those used for the numerical experiment. The resulting plan views and cross-section of the parameter η_m are presented in Fig. 17, and indicate that the observed changes can be mainly ascribed to an increase in the near-surface conductivity in the vicinity of the transmitter, and partly to a slight increase in the conductivity below the crater. Hereafter, the source regions for the former and latter conductivity changes are called R1 and R2, respectively.

The location of R1 almost coincides with the region of high sensitivity common to all receivers (Fig. 10(a), (b) and (c)). A spatially uniform change could be obtained if there was a systematic error in the values of the recorded

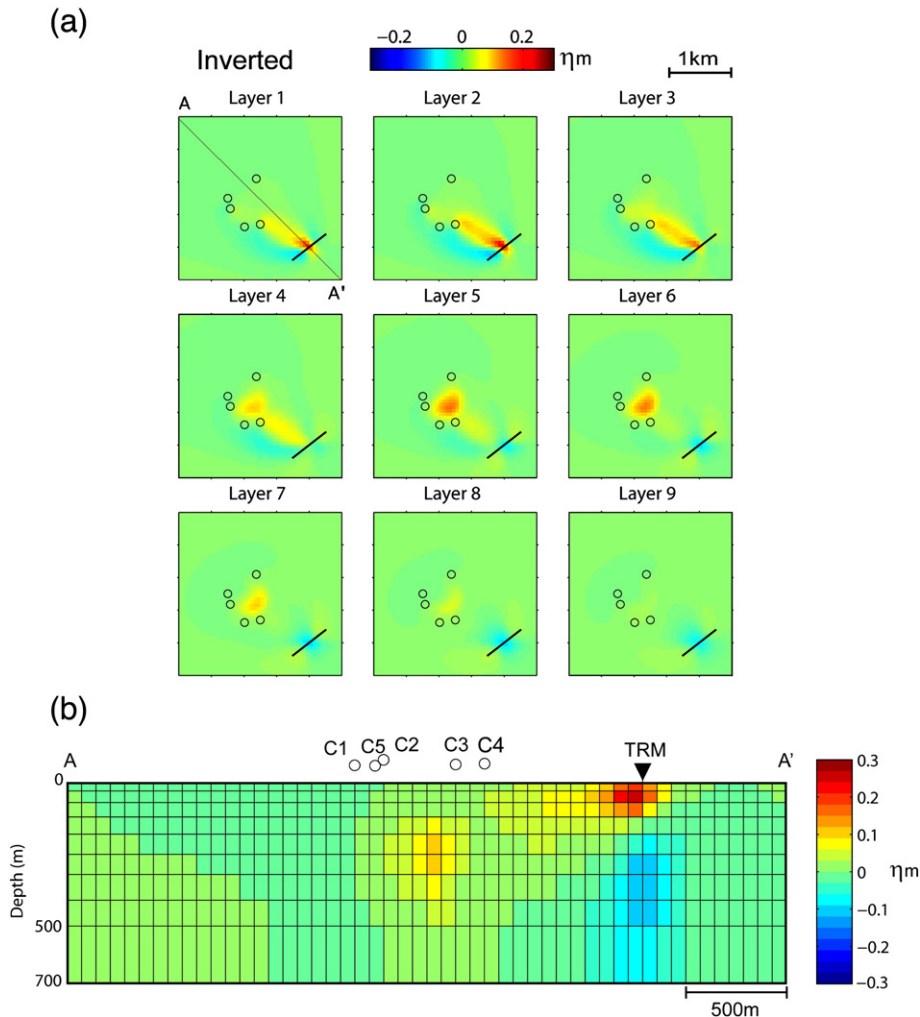


Fig. 17. Plan views (a) and a cross-section along the line A''–A' (b) of the conductivity distribution converted from the time-lapse change in the observed responses between the first and last measurements (Fig. 16).

transmitter current between measurements performed before and after the data gap. However, this possibility can be rejected, because there is small but significant lateral variation of the amplitude change, and because the measurement accuracy of the transmitter current is better than 1% in the ACTIVE system. Therefore, we may conclude that R1 reflects a real conductivity change in the Earth. Considering the spatial and frequency dependence and the imaging scheme applied here, the source region is not necessarily at the imaged location, but could be somewhere near the surface and outside the receiver array where there is little sensitivity. In other words, R1 may not be related to the volcanic process of the presently active summit crater. Additional receivers must be installed in a wider area around the present receiver array to address this problem.

We also examined whether or not the effect of R1 is statistically significant, by comparing the RMS misfits between the observed and calculated responses for models with and without R1. First, we performed a 3-D forward calculation for the conductivity model shown in Fig. 17, and estimated the RMS misfit to be 0.68. Next, another forward calculation was made by letting the parameter η_m equal zero for blocks corresponding to R1, from which an unacceptable RMS misfit of 1.91 was estimated. However, contrary to the case of R1, it is possible to conclude that R2 is related to the volcanic process below the central cone, if R2 is real and significant. The statistical significance of R2 was tested again by comparing the RMS misfits. When the parameter η_m was set to zero for blocks corresponding to R2, the RMS misfit was estimated to be 0.91. This result suggests that a

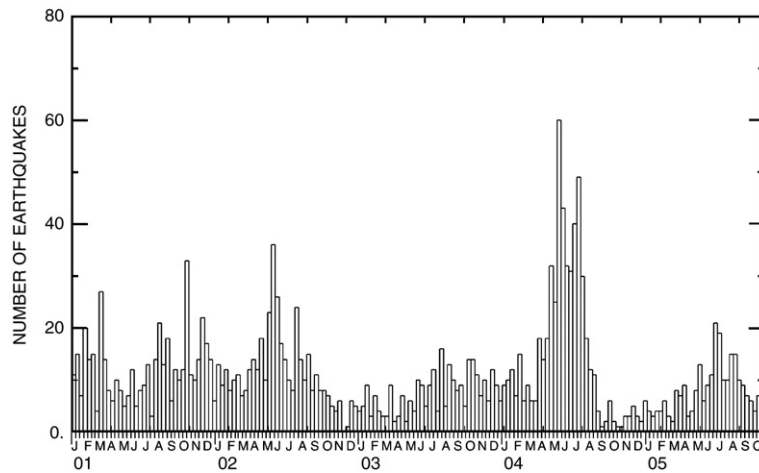


Fig. 18. Time variation of seismicity (the monthly number of earthquakes that occurred in the Izu-Oshima caldera) between January 2001 and December 2005.

model without R2 is also acceptable, and that the effect of R2 is not statistically significant.

The DC resistivity measurements on Mihara-yama were carried out with only small data gaps during this period. The apparent resistivity showed a decrease of approximately 5% from the beginning of 2003 to the middle of 2004, and then a recovery of approximately 2–3% (H. Watanabe, 2006, personal communication). Seismic observations also showed that there was a sudden enhancement of seismicity in the caldera region during April–August 2004 (Fig. 18). Although further detailed discussion in relation to the ACTIVE data is not appropriate at this time, because of the large data gap during 2003 and 2004, three kinds of independent evidence may commonly indicate a slight elevation of volcanic activity during this period that enhanced the seismicity in the caldera and caused heating (conductivity increase) at some depth below the crater. Such a phenomenon is not unusual for this volcano and may be related to a change in the regional stress field as reported by Yukutake et al. (1983).

7. Conclusion

This paper presented a transient EM system, named ACTIVE, and a simple and rapid method of imaging conductivity changes using the ACTIVE data in order to monitor the activity of the Izu-Oshima volcano in Japan. Numerical experiments, using a model of conductivity changes resulting from the 1986 eruption, have shown that the changes in conductivity below the crater over time can be monitored with improved spatial resolution by the proposed ACTIVE system. The system is very accurate at determining the location of such changes

unless the changes are very shallow. A field experiment performed over a period of approximately 4 years showed the capability and efficiency of the long-term operation of this system for monitoring, nearly in real time, in combination with the wireless LAN system developed in the Izu-Oshima caldera. Significant time-lapse changes in ACTIVE responses were observed from July of 2002 to April of 2005, and the difference between the responses observed in 2002 and in 2005 was converted into conductivity changes that indicated two anomalous regions: one at the shallower (down to approximately 100 m) part of the caldera floor between the transmitter and the array of receivers, and the other at approximately 500 m below the central cone. The latter anomaly suggests an increase in conductivity due to heating below the summit crater that, although it was not statistically significant, indicates temporarily elevated volcanic activity. This interpretation is qualitatively consistent with other observations, such as the decrease in the apparent DC resistivity and the enhanced seismicity during this period. Thus, we may conclude that the basic development of the ACTIVE monitoring system is complete. However, the imaging method presented here is still in a premature stage, and further technological improvement is necessary if the method is to be used for more detailed volcanological discussion.

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