Resistivity structure of Unzen Volcano derived from time domain electromagnetic (TDEM) survey

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Time-domain electromagnetic (TDEM) surveys were carried out around Unzen Volcano, Shimabara Peninsula, South-west Japan in 2001 and 2002 in the eastern part of the peninsula. The surveys were a completion of the previously interpreted TDEM data collected from surveys in 1995 in the western part of the peninsula. Interpretation of the recorded transient magnetic fields resulted in a general feature of the resistivity structure which are composed of three main layers: a resistant surface layer, a conductive layer (below 10 m) from about sea level to 2.5 km below the surface, and a third resistive layer. The conductive layer is considered to be a complex of water-saturated and altered layers which spreads widely beneath Shimabara Peninsula. The spatial conductance distribution shows a W–E trending pattern of high conductance from Tachibana Bay to the summit of Unzen (Fugen–Dake). This evidence indicates that volcanic gas was supplied to shallow layer in the western part of the peninsula, which is in accordance with the path of magma migration derived from seismic and deformation studies. Another high conductance zone was also found near Mayu–Yama in the eastern part of Shimabara Peninsula, indicating magma degassing process due to the possibility of the presence of a magma system beneath it.

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1. Introduction

Unzen is a composite volcano composed of a group of volcanic cones, situated in the center of Shimabara Peninsula, Kyushu, SW Japan (32.75 N, 130.30 E). It has developed within the E–W trending regional tectonic graben system called Unzen Graben, of which basement has subsided more than 1000 m below sea level due to N–S extensional stress and is bounded by Chijiwa Fault in the north and regional tectonic graben system called Unzen Graben, of which the 198 yr of dormancy, Fugen Dake, Myoken Dake, and Mayu–Yama volcanoes, in order from the oldest to the youngest (Hoshizumi et al., 1999). After the 198 yr of dormancy, Fugen–Dake (Mt. Fugen), which is the main peak of Unzen Volcano, began small phreatic eruption in November 1990, followed by successive different types of eruptions, and finally ended in 1995 with a total volume of erupted lava of 2.1×10⁸ km³ including Heisei–Shinzan, a huge lava dome in the eastern flank of Fugen–Dake (Nakada et al., 1999).

Several models on magma supply system from deeper structure of the region have been proposed which can be represented by two main models: First, a W–E ascending magma supply system concentrated at about 15 km beneath Tachibana bay and reached the shallow part of Unzen beneath Fugen–Dake based on volcano-tectonic seismicity (Umakoshi et al., 2001) supported by the results of leveling surveys (Ishihara, 1993) and GPS surveys (Nishi et al., 1999) by which the main pressure source was located to the west of the summit. The other one is a magma supply system proposed by Ohmi and Lees (1995), inferred from the study of three-dimensional P- and S-wave velocity structure below the volcano. It was proposed that a NW–SE trending elongated system of dykes starting at 30 km northwest of Unzen, striking to the southeast and dipping to about 15 km below the Shimabara bay area.

Although the chemical and physical properties of dacite magma of Unzen were similar to those of Mount Pinatubo and Mount St. Helens, which erupted explosively in 1991 and 1980, respectively, the recent eruption of Unzen was effusive (Nakada and Motomura, 1999; Nakada, 2000), and the difference was caused by the difference of the degassing efficiency of the ascending magma. The volcanic gases may have escaped
Extensive magneto-telluric (MT) observations had been conducted around Unzen volcano since the first eruption on 17 November 1990 (Kagiyama et al., 1999). It was revealed that the conductivity structure beneath the volcano comprises a very resistive layer of several hundreds to several thousands $\Omega \cdot m$ from the surface to about 500 m above sea level, a conductive second layer (about 10 $\Omega \cdot m$), followed by a third resistive layer which begins at about 1 ± 0.5 km below seal level (about 100 $\Omega \cdot m$), and another conductive layer (10–30 $\Omega \cdot m$) starting from depths of 10 ± 3 km below sea level. The conductive layer beneath the surface layer was interpreted as water-saturated layer, which had played important role in generating precursory phenomena and in controlling the temporal change of the eruption types. Kagiyama et al. (1999) explained that at a rate of about 20 m/day a magma column reached the bottom of the saturated layer causing an intense volcanic tremor and phreatic eruptions in January–February 1991, followed by a phreatomagmatic explosion at the time the magma approached the upper boundary of the layer on April 9, 1991. They also suggested that volcanic gas was continuously released when the magma began to rise at about 10 km depth in the western part of the Shimabara Peninsula. The gas reached and heated the saturated layer instigating the first occurrence of volcanic tremor in July 1990. However, there is still no evidence regarding the supply of volcanic gas into the water saturated layer in the western part of the peninsula.

The purpose of this study is to infer information on the general feature of resistivity structure in the shallow part of Shimabara peninsula from the interpretation of TDEM data set. We try to assess signatures of magmatic volatiles supply from depth to the shallow...
surrounding in the peninsula based on the revealed resistivity structure. The data set were obtained from TDEM measurements conducted previously in 1995 in the western part of Shimabara peninsula (Kanda, 1997), and from the measurements performed in 2001 and 2002 in the eastern part of the peninsula.

2. TDEM survey and data interpretation

2.1. TDEM method

Electrical conductivity $\sigma$, expressed in S/m (or its reciprocal $\rho$, the resistivity, expressed in $\Omega\text{m}$) significantly depends on temperature, rock porosity, fluid saturation, and pore fluid conductivity (e.g. Hyndman and Shearer, 1989; Llera et al., 1990; Hermance, 1995). Commonly, a volcanic region is characterized by the presence of a high conductivity contrast due to the combination of the mostly cold, dry and resistive host rock and the high temperature, conductive melted magma reservoir and the overlying hydrothermal system associated with it (Wright et al., 1985; Newman et al., 1985). The contrast leads to the high effectiveness of the application of electromagnetic methods in such region. The time domain electromagnetic (TDEM) method that uses an artificial source of electromagnetic excitation (Kaufman and Keller, 1983; Keller et al., 1984; Ward and Hohmann, 1988; Strack, 1992; Kanda et al., 1996) is one of the most effective electromagnetic exploration techniques, that is highly applicable for exploring 10 km depths in a region where artificial electromagnetic noises are high and electrical field measurement is difficult to carry out because of steep topography and stiff lava on the surface.

As to compile the overall TDEM surveys around Unzen Volcano which were conducted in the western part in 1995 (Kanda, 1997) and in the eastern part in 2001 and 2002, we will first describe the method briefly. At Unzen, a grounded horizontal electric dipole (HED) of about 1–2 km length was utilized as the transmitter where a bipolar square wave current with 50% duty cycle was injected to the ground. One cycle of current was completed in 20 s, consisted of 5 s of turn-on interval and 5 s of turn-off interval in an alternating polarity (Fig. 2). The transmitter was fed by a three phased power supply (200 V AC) and the maximum of output current was 20 A (peak-to-peak) at 800 V. The receivers were located at about 2–8 km from the transmitter, where the three components of the induced magnetic fields were measured as a function time. The receivers were fluxgate magnetic sensors (Tierra Technica U43 System) powered by 12 V car battery, with resolution of 1 pT. Earlier fluxgate magnetometers version (U36 System) were used in the 1995 surveys. The transmitter and receivers were synchronized by a stable clock system that has accuracy of $2\times10^{-8}$ s.

Although this method is less affected by the near-surface inhomogeneities (Pellerin and Hohmann, 1990; Poddar 1999), the transmitter–receiver offsets control the trade-off between the signal-to-noise ratio ($S/N$) and the length of decaying signal. Small offset will cause the transient signal to decay rapidly, containing only information on the shallower part of the ground conductivity distribution and it is strongly influenced by the characteristic of the measurement system. On the other hand, large offset will cause the $S/N$ to decrease. The effects of measurement system can be removed by performing the signal deconvolution procedure in the latest stage of data processing, and the $S/N$ can be improved by collecting sufficiently large data cycles to stack. To obtain the filter for the deconvolution, calibration at the vicinity of the transmitter (about 50–100 m) should be carried out upon assuming that the induced magnetic fields can be ignored and that the measured waveforms are regarded purely as the convolution of the whole measurement system with the magnetic field generated by the transmitted current.

2.2. Survey locations and data processing

The distribution of TDEM receiver stations at Unzen is shown in Fig. 3. Solid circles denote the measurement sites in the western part of Shimabara Peninsula in January and December 1995, where some of them were close to the most active summit during the eruptions of Unzen (KZA, AZM, NTG, and KNM) and some were near to the Unzen Spa (SKM and SUS). A grounded-wire source of 1.8 km was installed in the northwest of the volcano, indicated by the line CD. The current intensity were about 17 A and 15 A with the ground resistance of 40 $\Omega$ and 45 $\Omega$ in the first and second experiments respectively. About 1000–7000 cycles of transient wave forms of 8 Hz sampling rate were collected in 2–4 days from the first survey, while about 2500–5000 cycles were obtained in the second. About 180–300 cycles of transient signals measured with 128 Hz sampling rate were also collected to resolve the shallower part of the conductivity structure. Only the early time part of the 128 Hz is required for the analysis so the measurements were carried out for only few hours.

The surveys in the eastern part were conducted in February–March 2001 and 2002. The receivers were distributed mainly to cover the eastern part of Shimabara Peninsula, eastward from the...
Fugen–Dake, forming an almost semi-circular shape of receiver distribution, covering also Mayu–Yama (Mt. Mayu). The 2001 measurement sites are denoted by 13 open circles named as UZ01–UZ13, while those in the 2002 survey are denoted by 12 rectangles named as N101–N112. The closest station to the summit of Fugen–Dake (Heisei Shinzan) is N112 in south of the summit (2.1 km from the summit at elevation 756 m above sea level) and the farthest is N107 in the east (5.5 km at elevation 81 m). AB in Fig. 3 is the position of the 1.4 km long transmitter’s grounded edges. Approximately 7 A of current at ground resistance of 85 Ω was injected during the surveys.

Fig. 4. (a) Example of the stacked data from the 1995 TDEM survey at AZM, merged from the 128 Hz and 8 Hz data and (b) The resampled AZM data and the inverted data. (c) The stacked data from the 2002 survey at N106, obtained by merging the 128 Hz and the 32 Hz data. (d) The resampled and inverted data of N106.

Fig. 5. Comparison between the observed data and the inversion result for several sites from the western part of Unzen (upper left) and the eastern part (upper right). Their rms misfits for each iteration are shown in the lower diagram.
and transient magnetic data were collected at the total 24 sites. The sampling rates of the receivers used were 128 Hz and 32 Hz and they were operated for about 3–4 h and 3–4 days respectively. There were total about 15,000 cycles of 32 Hz data and 1900 cycles of 128 Hz data collected in the eastern part.

The data from each site were filtered using the pre-stack digital recursive filter (Strack et al., 1989) that removes the periodic noises without attenuating the amplitude of the transient signals and then selectively stacked by rejecting the sporadic noises by using the standard deviation of the recorded signals (Strack et al., 1989; Kanda et al., 1996). For the 1995 data, after applying the deconvolution process to remove the integrated system response effect, the early time part of the 128 Hz data (up to 0.1875 s after the current switching) were merged with the 8 Hz data. Similar way was applied to the 2001 and 2002 data, where the early time part 128 Hz data (up to 0.0234 s) were merged with the 32 Hz data. The last curve of transient response for each station was obtained by resampling the merged data in a log time scale. The examples of the merged and resampled data for AZM (in the western part of the peninsula) and N106 (in the eastern part) are shown in Fig. 4a–d. In most of the sites, the amplitude of the 128 Hz data and those of the higher frequency data have small differences and the data were connected smoothly.

### 2.3. 1-D inversion

The forward formulation for an earth model of horizontal layers that relates the dipole moment of a grounded horizontal wire with the generated vertical component of the magnetic field intensity at the surface can be found in Kaufman and Keller (1983) and Ward and Hohmann (1988). As mentioned, the TDEM interpretation in this paper is intended mainly to seek for the general features or large spatial scale of the conductivity structure, assuming that the conductivity structure can roughly be approximated using individual 1-D inversion at each site. The inverse problem of transforming the induced magnetic field as a function of transient time into conductivity beneath the receiver as a function of depth is nonlinear. Hence, we used an iterative procedure that solves a linearized approximation of the full problem at each iteration using a weighted least-square scheme with a smoothing constraint (Constable et al., 1987; Smith and Booker, 1988; Farquharson and Oldenburg, 1993). The 2001 and 2002 data were interpreted by adopting the Occam-type algorithm (Constable et al., 1987) which uses maximum smoothness constraints, yielding a smooth model of conductivities that fitted the observed data within a certain level of misfit. The implementation of the algorithm is straightforward, requires only the nonlinear forward model and its Jacobian. We discretized the subsurface electrical conductivity profile from the depth from 0 m down to 10 km into 40 layers with the logarithmically increasing thicknesses and it is terminated by a semi-infinite layer below the 10 km depth. Thus we have $N=41$ model parameters to estimate.

The 1995 data previously were interpreted by Kanda (1997), based on Bayesian statistic (Uchida, 1993; Mitsuhashi, 1994; Ogawa and Uchida, 1996) with an assumption that the spatial derivative of the prior distribution for the model parameter is Gaussian. A Bayesian likelihood is introduced to evaluate the validity of the model objectively, in which the optimum model is determined so as to maximize the Bayesian likelihood by selecting a hyper-parameter using the ABC information criterion (Akaike, 1980), that combines the data misfit and the prior information. In this inversion scheme, the target region of 0.1 km to 15 km depths was equally divided into 30 layers.

In both inversion schemes, by assuming that the data errors are unbiased, independent and Gaussian, the $rms$ (root mean square) misfit at each iteration, given by

$$
rm_s_{\text{misfit}} = \sqrt{\frac{1}{M} \| W_d - WG(m) \|_2^2}
$$

is expected to be equal to 1. $W$ is the diagonal weighting matrix whose elements are the reciprocal of the uncertainty of each datum, $d=(d_1, d_2, ..., d_M)$ is a set of the observed data whose total number is $M$, and $G(m)$ is a set of the calculated data which is a function of a model vector $m=(m_1, m_2, ..., m_N)$. Two examples of the comparison of the observed and inverted data are shown in Figs. 4b, d, and 5. Generally the calculated data fitted well to the observed ones except for the latter time. This discrepancy is probably caused by the 3-D nature of the site.
Fig. 7. (a) Cross section of resistivity structure at Unzen as shown in Fig. 3, Line L1 along Fugen–Dake, (b) L2 for the western part from ID2 to SUS, (c) L3 along Mayu–Yama, and (d) L4 for the southeastern part.
which cannot be represented by a layer. The rms misfits for each iteration at several stations are shown in Fig. 5, for an area in the western and eastern part of Shimabara Peninsula, respectively, the value generally tends to converge to the desired misfit tolerance.

A comparison between resistivity values from two borehole sites namely USDP-1 and USDP-2 with the inverted results from sites UZ01 and UZ09, which were located very near from the boreholes is shown in Fig. 6. Lithologies of the USDP-1 are mostly pyroclastic flow and mud flow deposit, interbedded by lava blocks and pumice flow exist between the strata, whereas the lithologies of USDP-2 are mostly mud flow deposit and rock avalanches, with phreatomagmatic deposits from about 600 to 800 m depth (Hoshizumi et al., 2002). The comparison shows an acceptable agreement in values and pattern, with decreasing tendency of resistivity with depth.

3. Results

Almost similar to the MT results, the most prominent results from both TDEM measurements in the western and eastern part of Unzen is the existence of the conductive layer beneath a resistive surface layer in most of the sites. The resistivity structure can roughly be estimated by a 3 layer model: i.e., a resistive or moderately resistive surface layer (several tens to several hundreds 1Ωm) followed by a conductive layer (<10 Ωm) and third resistive layer. The conductive layer is located at about depths of 300 m to 3 km below the surface. They are equivalent to depths of sea level to 2.5 km below sea level depending on the altitude of each site. At several sites to the west of Fugen–Dake (KNM, KT2, KT4, HKB, JY, ID1, and ID2) and in the eastern part of Unzen (UZ01, UZ03, UZ06, UZ13, N103, N104, N106, N107, N108, and N109) minimum resistivity values attain 1 Ω m. The conductive fourth layer detected at depths of ≈10 km by MT method was not found by the present study probably because it is deeper than the depth of penetration of TDEM method. To see the local features of the resistivity structure in more detail, several 1-D sections denoted by L1, L2, L3, and L4 are drawn as shown in Fig. 7. The location and orientation of each section can be seen in Fig. 3. Along the L1 section from the west to around the summit (Fugen–Dake), the depth of the conductive layer from the sea level tends to decrease, whereas the thickness of the layer increases. However, minimum value of resistivity (1 Ω m) is distributed around KNM to KT2, indicating that conductivity enhancement occurs at points aside from the summit. Along L2 cross section in the western part, the conductive structure was located at about 1 to 2 km depth. The resistivity along the northern side of this section is lower than the southern side and the minimum value of resistivity is found at ID2 (1 Ωm). Beneath Mayu–Yama the low resistivity layer extents to the deeper part as can be seen at two sites, UZ02 and UZ03 in L3 section. The resistivity values show decreasing pattern towards Mayu–Yama both from the UZ01 in the north and UZ08 in the south. Increasing thickness of the low resistivity layer is found from N104 near Mayu–Yama to UZ108, which is located around Akamatsudani Fault. The L4 section in the southeastern part of the peninsula also exhibits almost the same feature as that of the western part. A conductive layer was found at about 1 km to 2 km in this section. It is also shown that the resistivity value of this layer tends to decrease at points near from Mayu–Yama.

4. Discussion

There are several possible explanations on the existence of the extensive conductor in the shallow part of a volcanic region. The causes may be attributed to the presence of shallow magma chamber (Newman et al., 1983), water-saturated layer (e.g. Kagiyama et al., 1999), hydrothermal zones (Wright et al., 1985), hydrothermally altered minerals (e.g. Jones and Dumas, 1993; Ogawa et al., 1998; Lenat et al., 2000), or combination of the above features. The lava emergence took place in the vicinity of the Fugen–Dake. The emergence is associated with the presence of the smallest magma chamber located just beneath the Fugen–Dake at about 1.5 km depth as inferred from geodetic study (Ishihara, 1993). A high temperature, ductile and low-Q body exists only within 4 km west from the summit as suggested from the seismicity study (Umakoshi et al., 2001). These features lead to an unlikely imagery that magma would occupy a large horizontal extent in shallow depth in the peninsula. Aside from the summit area, there are three manifestations of geothermal system in Shimabara Peninsula: i.e. Obama hot spring, Unzen hot spring, and Shimabara hot spring. The first two are located in the southwest direction from the summit and the latter is located to the east from summit. These geothermal expressions indicate the presence of fluids as media to transfer heat to the surrounding or to the surface. NEDO (1988) also suggested the presence of high temperature hydrothermal fluids inferred from drilling data at several sites within about 5 km west from the summit. Based on the above features and the previous MT results, we interpret the conductive shallow layer derived from TDEM data as a combination of a water-saturated layer and the thermally altered rock resulted from long term interaction between the hydrothermal fluid and the surrounding rock. The decrease of resistivities to their minimum values at several sites might be attributed to supply of hot ionic gases or fluids to the water-rich layer. TDEM method has higher sensitivity to detect a conductive layer (Kauffman and Keller, 1983) but it has lower reliability to determine resistive zones above a conductor in volcanic environment as shown in Lenat et al. (2000). Furthermore, electromagnetic methods are well perceived to be more sensitive to conductance (conductivity–thickness product) rather than to conductivity variations with depth (e.g. Jones, 1987; Kusi et al., 1998; Korja et al., 2001). Hence, we use the conductance distribution derived from the inversion results to allow for locating the regional anomalous areas, where the cumulative conductance changes substantially caused by higher supply of magmatic gas or volatiles into the water-saturated layer and/or the presence of alteration products. The conductance value is obtained by vertical integration of the conductivity with depth:

\[ \int_a^b \sigma dz = \sum_{i=1}^k a_i z_i \]

where \( a_i \) and \( z_i \) are conductivity and thickness of the \( i \)th layer respectively, and \( k \) is the number of layer in the interval (a,b). In this study the value \( a \) is assigned as the height of the surface, whereas \( b \) 5000 m depth, which is assumed to cover the total thickness of the water-saturated layer as well as the altered zone in the area.

The contour of spatial conductance is shown in Fig. 8a. Two zones of high conductance are observed within the map: an elongated zone trending W–E starting from around Tachibana Bay toward Fugen–Dake and a zone in the northeastern part of the map. Fugen–Dake and its surrounding are indicated by a dashed rectangle in the center. Conductance in this zone ranges from about 300 to 450 S. The section of resistivity along Fugen–Dake (Fig. 7a) shows that a low resistivity layer is shallower and also thicker beneath the summit. This shallow conductivity region below Fugen–Dake is probably related to the volcanic activity including the circulating hydrothermal system in a limited area (within 1 km from the summit) as inferred from the VLF-MT (Kagiyama et al., 1999) and self-potential (SP) survey. Hashimoto (1997) identified that a large SP anomalies of 1000 mV in the summit vicinity (Fig. 8b) is attributed to the streaming potential caused by subsurface hydrothermal upflows. In this conductance map, the decreasing tendency of conductance in the radial direction from the summit, which can be regarded as signature of the escape of volcanic gas to the edifice through conduit walls, is not clearly seen. Maximum value of conductance is not found around Fugen–Dake which is located on the eastern tip of the W–E high conductance zone. This is
probably due to higher contribution of ionic substances in some locations west from the summit area.

The elongated high conductance zone trending W–E from around Tachibana Bay to the summit can be seen in Fig. 8a, with values from 450 to 700 S. This elongated zone is in accordance with the W–E fault systems near from Take Fault, around ID1, ID2, KT2, KT3, KNM, and several other sites near by. This high conductance region corresponds to high seismic region around the shore of Tachibana Bay to Take Fault as indicated by a dashed line in Fig. 8a. The dashed line indicates the distribution of earthquake hypocenters from 1989–1991 (Umakoshi et al., 1994). There are no TDEM data to the north from this elongated zone for comparing the conductance value there. However, results of MT survey by Kagiyama et al., (1992) along the Chijiwa Fault revealed that thick accumulation of resistive layer exists, which might confirm that the elongated zone is bounded between two resistive zones within the graben. The seismic activity in Unzen has long been recorded before the eruption. Two large events of Shimabara Earthquake (M 6.9 and M 6.5) took place during the sequence of earthquakes in 1922. Since then earthquake occurrences took place in all the following decades, including the earthquake swarm during 1968–1974 whose hypocenters were distributed beneath the Tachibana Bay and in the Unzen Graben. Another earthquake swarm occurred in 1984–1985, of which magnitudes varied from M 4.3 to M 5.7 and widely distributed in the western shore of the peninsula and an earthquake at a depth of 6.6 km with magnitude of 4.2 occurred in 1985. An earthquake swarm with maximum magnitude of 3.5 also took place in the central part of Tachibana Bay in 1987. On November 1989, one year before the eruption, an earthquake swarm with a maximum magnitude of 3.3 took place beneath the SW part of Tachibana Bay and the depth of hypocenters ranged between 12 and 15 km (Umakoshi et al., 2001; Japan Meteorological Agency, 2006). Most of earthquakes occurred around Unzen Graben were identified as volcano-tectonic earthquakes (Nakada et al., 1999) and all of the above frequent earthquake occurrences were not accompanied by eruptions, which indicated that the occurrences were related to the intrusive events of the magma system in the western part of Shimabara Peninsula. These earthquake occurrences before the eruption probably were seated in many fracture zones in the western part of the peninsula, and these fracture zones became the pathways of volcanic gases to escape and dissolve into the water-saturated layer. The elongated high conductance zone which is related to a high ionic concentration might be attributed to the continuous supply of magmatic volatiles that initiated long before the eruption. Whereas, the conductance zone around Fugen–Dake whose value lower than that of the elongated zone was mainly due to the latest eruption activity. This also implies that during the magma ascent process the escape route of magma degassing was established mainly in region to the west of the summit and thus contributed to the non-explosiveness of the recent eruption type.

Within the investigated area, a very high conductance zone occupies the northeastern part of the investigated area, near from Mayu–Yama.
with conductance value higher than 1000 S. In relation to the 1990–1995 eruptive activity, the presence of magmatic chamber in shallow part of Shimabara Peninsula is related to the migration of magma from beneath Tachibana Bay to the summit of Fugen–Dake as previously mentioned. Therefore, high conductance in the eastern part of the investigated area can be attributed to other causes that might not be directly correlated to the ascending magma in the western part.

Beside salts, dissolved CO₂ gas in crust fluid can have significant impact in raising the conductivity (Nesbit, 1993). The study of magmatic CO₂ concentration, flux, and dissolved inorganic carbon from soil and groundwater around Unzen Volcano indicated higher contribution of magmatic gas to the eastern region of Unzen Volcano, especially around Chijiwa Fault and Akamatsudani Faults, which implies that the volcanic gas may migrate from magmatic source along the fracture zones (Ohsawa et al., 2002; Takahashi et al., 2004). The high concentration of CO₂ fits with the presence of high conductance in the area as can be seen from Fig. 8a, b, and c. Notsu et al. (2001) noted that the helium isotopic ratio (³He/⁴He), which is the tracer to identify source materials of volcanic products, shows an increasing pattern from west to east inside the Unzen Graben with Obama hot spring at the west side, Unzen hot spring around the center, and Shimabara hot spring at the east side. They analyzed the isotopic ratio from the bubbling gas of the hot springs and suggested that since the ratio should be highest in the shortest distance from the magma, the results might indicate either the presence of magma below Shimabara Bay or the migration of magmatic gas through the faults system from the volcanic edifice to the Shimabara hot spring. The spatial correlation between the high concentration of CO₂, high helium isotopic ratio, and high conductance distribution suggest that magmatic gas is supplied to the water-saturated layer reflecting the possibility of the presence of deep magmatic source beneath Mayu–Yama, supporting the magmatic model of Ohmi and Lees (1995) in the eastern part of Shimabara Peninsula. Geological studies also suggested the presence of an old magma in the eastern part of the peninsula that created Mayu–Yama about 4000 years ago. This magma has not been cooled down yet and caused the low seismicity in the eastern part of the peninsula (Nakada, 1996).

5. Conclusions

TDEM surveys in Unzen region have revealed the large scale resistivity structure of the Shimabara Peninsula. In the shallow part of the peninsula, a conductive layer is found and distributed widely around Shimabara Peninsula at depth ranging for about 100 m to 2.5 km below sea level indicating the existence of a water-saturated and altered layers in the area. High conductivity due to the supply of volcanic gases into the water saturated layer, represented as high conductance zones is found between inside the Unzen Graben. A high conductance zone trending W–E direction in the shallow part is also found at around the western part of the peninsula towards the summit. This zone is correlated strongly with the pattern of the frequent swarm earthquakes that started relatively long before the 1991–1995 eruption indicating the repeating emission of volcanic gases into the shallow water layer through the fault system in the western part of Unzen, which affects the degassing efficiency during the magma migration of the latest eruption. Another zone, the highest conductance zone, is located around the northern part of Mayu–Yama. This zone is correlated with the high spatial distribution of volcanic gases as inferred from geochemical studies that traced the escape of volcanic gas from a deeper magmatic source, and might indicate the presence of magma beneath Mayu–Yama.

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