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A study of annual variations in the geomagnetic total intensity with special attention to detecting volcanomagnetic signals

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This paper investigates the cause of annual variations in the geomagnetic total intensity that are often seen especially in the volcanic areas. As a hypothesis of the cause, a model was proposed, in which such a change is produced by changes in the inhomogeneous magnetization of near-surface rocks due to temporal changes of the atmospheric temperature. This hypothesis was tested by field and laboratory experiments. First, amplitude and phase difference of annual variations in the total intensity and ground temperature data were determined by time series analyses. Considering thermal diffusion from the surface into the ground, the phase difference between the total intensity and temperature was converted to a characteristic depth, and then the amplitude of annual temperature variation at the depth was estimated. Finally, the observed total intensity variations were compared with the expected change on the basis of the temperature dependence of rock's magnetization obtained by a laboratory experiment and the local magnetic anomaly obtained by a magnetic survey at each magnetometer site. A good agreement between the observed and expected changes was obtained, which strongly suggests that the hypothesis is correct. It was also shown that a correction of annual variations by using temperature data will enable us more accurate detection of volcanomagnetic signals.

1. Introduction

Continuous measurement of the geomagnetic total intensity by an array of proton magnetometers is regarded as a powerful tool for monitoring volcanic activity. During the last a few decades, a number of successful results have been obtained from various kinds of volcanic environments. From Izu-Oshima volcano (Fig. 1), central Japan, for example, significant changes in the total intensity were observed associated with its 1986 activity; i.e., changes which correspond to thermal demagnetization around the conduit prior to the eruption (Yukutake *et al*, 1990a), volcanomagnetic effects that accompanied the explosive events during the eruption (Sasai *et al*., 1990), and gradual variations associated with the thermal diffusion in the ground after the eruption (Hamano *et al*., 1990). From Aso volcano (Fig. 1) in Kyushu, SW Japan, Tanaka (1993) argued that implication for the eruption mechanism is reflected in the observed total intensity changes during a most active period of 1989–1990. Tanaka (1995) also detected remarkable changes in the total intensity just before a dome formation of Unzen volcano (Fig. 1) in Kyushu, SW Japan.

Zlotnicki *et al*. (1993) observed total intensity changes of different temporal and spatial scales during the 1986–1990 activity of Piton de la Fournaise volcano in Réunion island. Magnetometer array observations are used to mitigate a possible disaster by destructive eruptions on volcanoes such as Etna volcano in Italy (Del Negro *et al*., 1997), la Soufriere volcano, French Guadaloupe island (Pozzi *et al*., 1979), and Montagne Pelée volcano, Martinique island (Zlotnicki *et al*., 1986) in Lesser Antilles.

Most of the geomagnetic changes reported in these works prior to or associated with surface eruptions were well explained by a thermal demagnetization model (Yukutake *et al*., 1990a; Hamano *et al*., 1990; Tanaka, 1993; Zlotnicki *et al*., 1993). On the other hand, geomagnetic changes associated with the earthquake swarm activity and crustal deformation in Long Valley Caldera where no surface eruption occurred were successfully interpreted in terms of piezomagnetic effects due to inflating pressure source (Mueller and Johnston, 1998). There is an implication that some of the events observed at Piton de la Fournaise volcano might have been caused by electrokinetic effects (Zlotnicki and Le Mouël, 1990). Thus, volcanomagnetic studies give us various kinds of information to understand the thermal condition, stress or pressure distribution, and flow of pore fluids that are directly related to the volcanic activity in the ground.

For accurate detection, it is necessary to separate volcanomagnetic signals from natural geomagnetic fluctuations of external origin. The spatial distribution of external field fluctuations originated from the ionospheric and magnetospheric current systems (Campbell, 1997) is uniform compared to the scale of a volcano, and therefore simple differences of the total intensity with respect to the simultaneous value at a remote reference are used to reduce these effects. Even if the effects of external field are properly eliminated, however, we sometimes see periodic or non-periodic geomagnetic changes irrespective of the state of the volcanic activity. An annual variation is one of the most typical examples of such a periodic non-volcanic change.

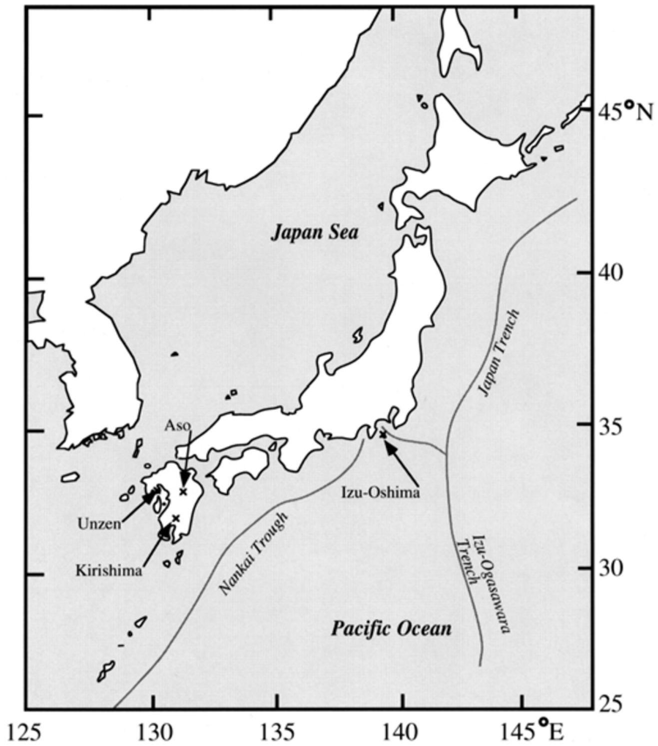

Fig. 1. Location of the study areas, Kirishima and Izu-Oshima volcanoes. Other volcanoes in Japan, Aso and Unzen, are also shown, in which successful volcanomagnetic studies were done.

Observed volcanomagnetic signals larger than 10 nT (Yukutake *et al.*, 1990a) and up to 100 nT (Tanaka, 1993) are reported in the above examples. In these cases, we can ignore the non-volcanic component in the site differences of the geomagnetic total intensity. However, when we expect only a small volcanomagnetic signal either due to a weak volcanic activity or due to the large distance between the source zone and the observation sites, the presence of non-volcanic changes not only makes signal detection more difficult but also may lead us to a misinterpretation of data. Our experience in Kirishima volcano (Fig. 1) in Kyushu, SW Japan, is one of such examples.

Kirishima volcano is a group of active andesitic volcanoes located in the southern part of Kyushu. In late 1991, there was a small seismo-volcanic crisis accompanied by an earthquake swarm activity and a steam explosion in Shinmoe-dake crater, which is one of the most active craters in the volcano group during the past a few thousand years (Imura, 1994). During this event, we started continuous observation of the geomagnetic total intensity by using proton magnetometers around the crater to monitor the activity. Our results show a gradual geomagnetic change from late 1991 to early 1992 that was possibly related to the thermal demagnetization just below the crater, though the activity itself soon calmed down (Kagiyama *et al.*, 1992). The observations were continued after installing an additional site, and revealed repeated occurrence of similar changes. Masutani and Kagiyama (1996) inferred that these changes are due to the thermal demagnetization, because small-scale swarm activities also took place repeatedly. However, this inference was turned out incorrect, as the geomagnetic data changed in a similar manner following 1996 in spite of having no swarm activity.

Annual variations can be seen in the total intensity data from other places especially in volcanic areas (e.g., Hamano *et al.*, 1990; Fukushima *et al.*, 1990). Although some works attempted to study the cause of such variations (Ozima *et al.*, 1996), its physical mechanism is not well understood yet. This paper aims to solve this problem of annual variations by means of field and laboratory experiments and to find a practical method for data correction, which will enable us more accurate detection of volcanomagnetic signals. For this purpose, we analyzed field data and examined the magnetic properties of rock samples from Kirishima and Izu-Oshima volcanoes (Fig. 1), as will be shown in the following sections.

It is well known that the differences in magnetic field orientation and induction sometimes cause apparent variations in the total intensity difference (Davis *et al.*, 1981; Tanaka *et al.*, 1977). Before this study, we examined if this effect can explain the observed annual variations by using nearby 3-component geomagnetic data. However, our results indicated that the amplitude of expected change is negligibly small for annual variations.

2. Data

Figure 2 shows the location of the proton magnetometer sites in the Kirishima volcano area. Two sites, SMS and SMW, were installed in November, 1991, while SMN a year later (Masutani and Kagiyama, 1996). The geomagnetic total intensity is measured every 20 minutes at each site and transferred to our institute by ARGOS satellite telemetry (Neki *et al.*, 1998). Correction of external geomagnetic disturbances was made by taking a simple difference of simultaneous data with reference to KNY about 60 km to the south (Fig. 2). EBI is a site for ground temperature measurement since 1989, and the temperature data will be referred in the data analyses later.

Figure 3 shows daily means of total intensity variations for 5 years from December 1992 to December 1997 observed at SMN, SMW, and SMS, relative to KNY. Arrows in the top diagram indicate the occurrence time of swarm earthquakes below Shinmoe-dake crater. A significant change is seen at the beginning of the SMN record that may correspond to the swarm activity. However, changes in the following years show similar feature even when there was no swarm activity. This strongly suggests non-volcanic origin of this variation with periodicity of one year.

Compared to andesitic rocks of Kirishima volcano, basaltic rocks of Izu-Oshima volcano are known to have intense magnetization of 10 A/m or more (Ohno, 1988), which is nearly one order of magnitude higher than that of Kirishima rocks. Observation by a proton magnetometer in Izu-Oshima volcano was initiated in 1965 (Rikitake, 1966), because large volcanomagnetic effects were expected to be caused by heating or cooling of underground rocks with such intense magnetization. The observation network was reinforced after its 1986 eruption by installing a large number of magnetometers as shown in Fig. 4 (Hamano *et al.*, 1990). In this study, the total intensity record from MI2 on the caldera floor south of the Mihara-yama crater is analyzed. This site was chosen because the total intensity record is almost continuous and shows a clear annual variation (Fig. 5). Here, simple site

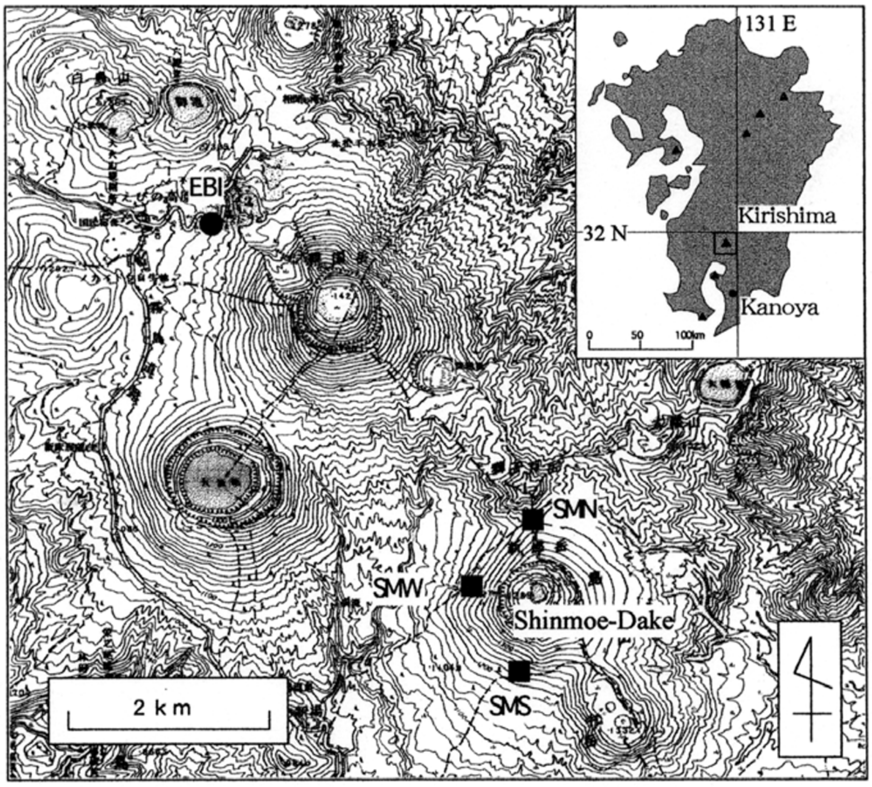

Fig. 2. Location of proton magnetometer sites in Kirishima volcano, SMN, SMW, and SMS. Ground temperature is also measured at EBI. Effects of the external geomagnetic disturbances are corrected by making a simple difference with reference to KNY.

difference of the total intensity was applied with reference to NOM on the western coast of the island to correct the effect of external geomagnetic disturbances (Yukutake *et al.*, 1990b).

Each record section shown in Figs. 3 and 5 was divided into subsections with 1 year length, detrended, and stacked for five years to extract an annual change (Figs. 6(a)–6(d)). The amplitude at MI2 is nearly 10 times larger than those at magnetic sites in Kirishima. Among three sites in Kirishima, annual change is the most prominent at SMN. It should also be pointed out that changes at SMN and MI2 are almost antiphase to those at SMW and SMS. Since annual variations at SMW and SMS show similar features both in amplitude and in phase, we did not use data at SMS.

3. A Model of Thermally-induced Near-surface Magnetization Change: A Hypothesis

A qualitative interpretation in the previous section indicated that the annual variation found in the geomagnetic total intensity data is supposed to be of non-volcanic environmental origin. We decided to examine the effect of temperature as it has clear physical causality with magnetization change. It is well known that the magnetization of ferromagnetic minerals in rocks changes with temperature. However, usual rock magnetic studies pay most attention to the dependence in a wide temperature range. Most of volcanomagnetic studies also ignore small temperature change on the surface, because underground rocks at depths may actually be heated above the Curie temperature by volcanic activities such as magma intrusion.

Seasonal change in the atmospheric temperature penetrates into the ground by thermal diffusion. Figure 7 shows the result of ground temperature measurements at 0.5 m and 1 m depths at EBI. Differences in the amplitude and phase are supposed to be reflecting the thermal diffusion process. As shown in this result, the annual temperature change near the surface has an amplitude of 30 degrees or so, which may cause a local magnetic field change of observable intensity due to the temperature dependence of the magnetization of near-surface rocks. Detectability depends on the intensity of the local magnetic anomaly due to near-surface magnetic inhomogeneity and its temperature dependence.

Suppose that the sensor of a proton magnetometer is placed above the earth's surface near a local magnetic inhomogeneity with temperature variation penetrating from the surface (Fig. 8). The total intensity measured by using this sensor can be derived as,

$$|\mathbf{F}(\mathbf{r}, t)| = |\mathbf{F}_0(\mathbf{r}) + \mathbf{F}_1(\mathbf{r}, t)|, \tag{1}$$

where $\mathbf{F}_0$ is the regional magnetic field and $\mathbf{F}_1$ is the local magnetic anomaly due to the heterogeneity. If the inhomogeneous magnetization distribution is $\mathbf{J}_1(\mathbf{r}_1, t)$, $\mathbf{F}_1$ can be expressed as,

$$\mathbf{F}_1(\mathbf{r}, t) = -\nabla \int_V \mu \mathbf{J}_1(\mathbf{r}_1, t) \cdot \frac{(\mathbf{r} - \mathbf{r}_1)}{4\pi |\mathbf{r} - \mathbf{r}_1|^3} dV_1. \tag{2}$$

If the geothermal profile has a time variation, $\Delta T(z, t)$, $\mathbf{J}_1(\mathbf{r}_1, t)$ will also have a time variation that may be approximated in a small temperature range of present interest as,

$$\mathbf{J}_1(\mathbf{r}_1, t) = \mathbf{J}_1^{(0)}(\mathbf{r}_1)[1 + \delta J_1 \Delta T(z_1, t)] \tag{3}$$

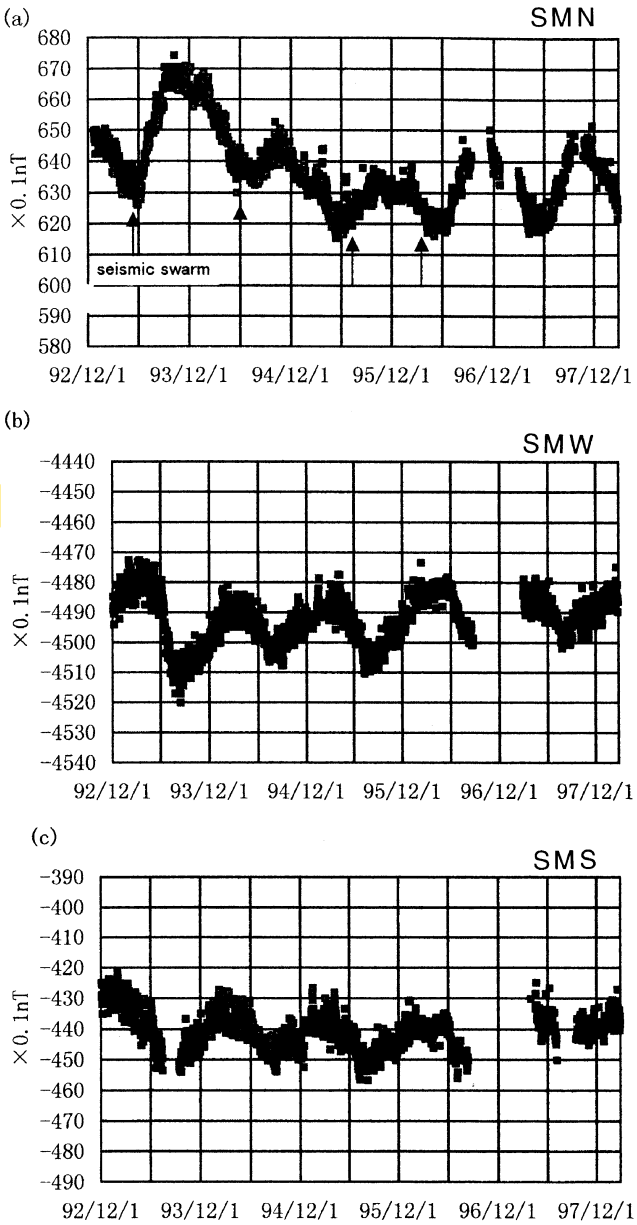

Fig. 3. Daily means of the total intensity at SMN (a), SMW (a) and SMS (c) from December 1992 to December 1997, with reference to KNY.

where $\mathbf{J}_1^{(0)}$ is the magnetization at a reference temperature, and δJ_1 is its temperature coefficient. Note that we assumed that only the magnetization intensity changes with temperature. The total intensity anomaly at this site can be defined as,

$$\begin{aligned}\Delta F(\mathbf{r}, t) &= \mathbf{n}_0(\mathbf{r}) \cdot \mathbf{F}_1(\mathbf{r}, t) \\ &= \Delta F_0(\mathbf{r}) + \Delta F_T(\mathbf{r}, t),\end{aligned} \tag{4}$$

where $\mathbf{n}_0(\mathbf{r}) = \frac{\mathbf{F}_0(\mathbf{r})}{|\mathbf{F}_0(\mathbf{r})|}$ is a unit vector parallel to the geomagnetic field at $\mathbf{r}$, and ΔF_0 and ΔF_T are time independent and dependent parts of the anomaly. Substituting (2) and (3) into (4), we have,

$$\Delta F(\mathbf{r}, t) = -\mathbf{n}_0(\mathbf{r}) \cdot \nabla \int_V \mu \mathbf{J}_1^{(0)}(\mathbf{r}_1)[1 + \delta J_1 \Delta T(z_1, t)]$$

$$\cdot \frac{(\mathbf{r} - \mathbf{r}_1)}{4\pi |\mathbf{r} - \mathbf{r}_1|^3} dV_1. \tag{5}$$

If we assume a two-dimensional magnetization distribution on a plane at depth d, (5) can be rewritten as,

$$\begin{aligned}\Delta F(\mathbf{r}, t) &= -\mathbf{n}_0(\mathbf{r}) \cdot \nabla \int_S \mu \mathbf{M}_1^{(0)}(x_1, y_1)[1 + \delta J_1 \Delta T(d, t)] \\ &\quad \cdot \frac{(\mathbf{r} - \mathbf{r}_1)}{4\pi |\mathbf{r} - \mathbf{r}_1|^3} dS_1 \\ &= \Delta F_0(\mathbf{r})[1 + \delta J_1 \Delta T(d, t)],\end{aligned} \tag{6}$$

where $\mathbf{M}_1^{(0)}(x_1, y_1)$ is the surface magnetization distribution. Thus, we have an expression for the time dependent part of the total intensity anomaly in (4) as,

$$\Delta F_T(\mathbf{r}, t) = \Delta F_0(\mathbf{r}) \cdot \delta J_1 \Delta T(d, t). \tag{7}$$

By using this expression, it is possible to quantitatively estimate the effect of temperature variation if we have the total intensity anomaly, temperature variation data, and the temperature coefficient of the magnetization.

This paper proposes a hypothesis described by this model for the cause of annual changes in the total intensity and tries to test it by field and laboratory experiments as follows:
(1) Estimating amplitude ratio and phase difference between annual variations in the total intensity and the ground temperature.
(2) Measuring the temperature dependence of magnetization of rock samples taken from each magnetometer site.
(3) Making a magnetic survey at each site to extract the local total intensity anomaly due to near-surface magnetic heterogeneity.
(4) Comparing the values observed and expected from the temperature coefficient and survey result.

4. Observed Annual Changes

In this section we quantitize annual changes in the total intensity and temperature, and determine the relation between them by assuming a simple linear system. Strictly speaking, temperature has to be measured at each magnetometer site to make correlation analysis. However, temperature observed at EBI was used to analyze all data, as it is the only ground temperature data that we have. Although there could be a difference in annual temperature variations among these magnetometer sites, especially between the sites in Kirishima and MI2 in Izu-Oshima, we simply assumed the same temperature. This assumption may cause an error in the amplitude ratio by up to a few tens %, which will not be so serious for the present purpose.

Assuming a simple 1-D thermal diffusion (Carslaw and Jaeger, 1959), temperature variation $\Delta T(z, t)$ in (5) can be expressed as,

$$\Delta T(z, t) = \Delta T_0 \exp[-nz] \cos(\omega t - nz), \tag{8}$$

where ΔT_0 is the amplitude of temperature variation at the surface, ω is the angular frequency, and

$$n = \sqrt{\frac{\omega}{2\kappa}}, \tag{9}$$

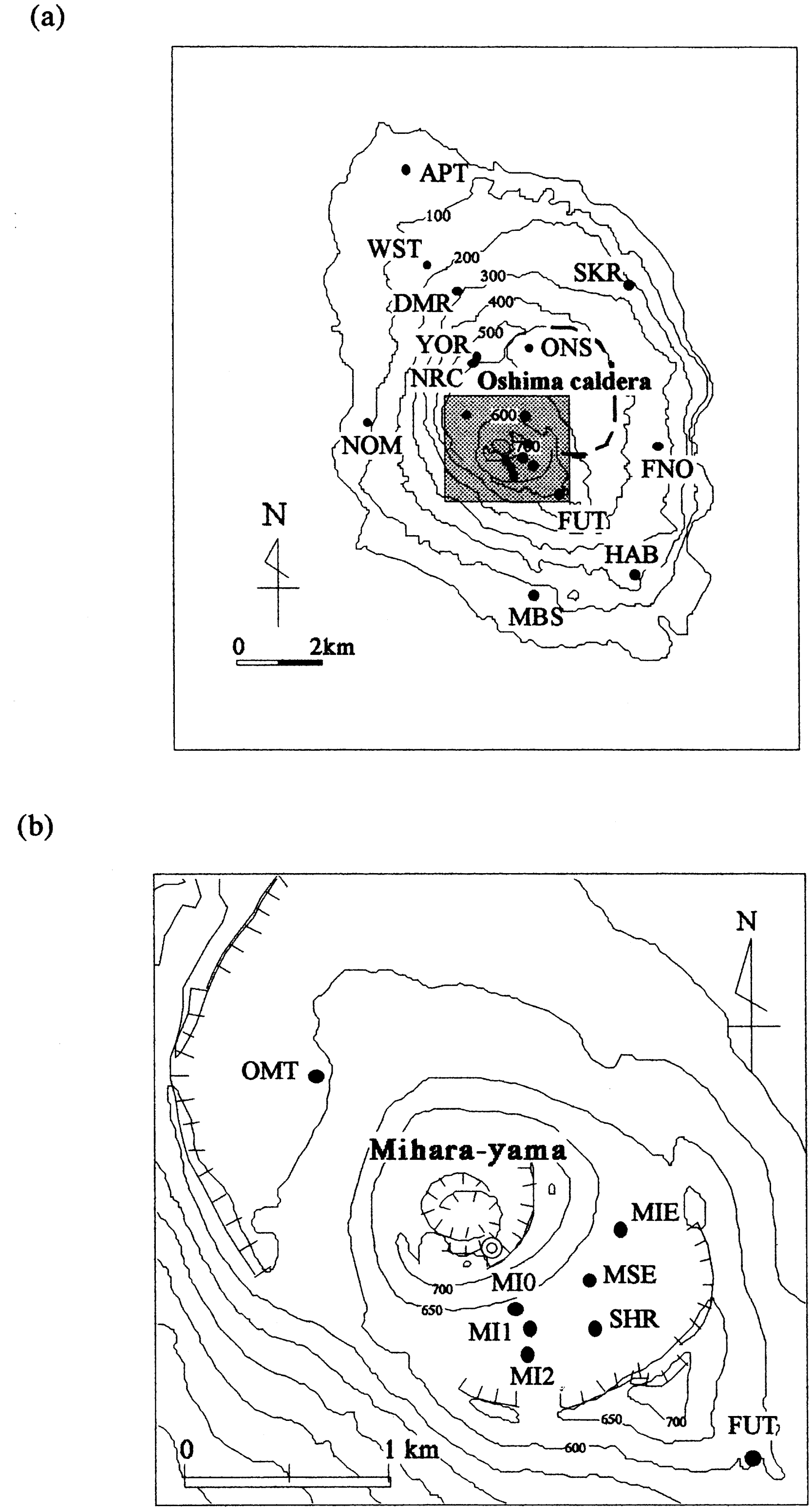

Fig. 4. Location of proton magnetometer sites in Izu-Oshima volcano (after Hamano *et al.*, 1990). Total intensity data from MI2 with reference to NOM are used in this study.

where κ is the thermal diffusivity. By using the two temperature records at EBI (Fig. 7), we determined the thermal diffusivity as $\kappa = 1.34 \times 10^{-7} \text{m}^2/\text{s}$. Although the thermal diffusivity could also be different among these sites, we simply adopted this value to each site. Because difference in given value of the thermal diffusivity does not affect the estimation of annual geomagnetic variation as shown below.

The temperature effect due to a model shown in Fig. 8 is also simplified as follows. As shown in (6), the spatial distribution of the magnetization $\mathbf{J}_1(\mathbf{r}_1, t)$ was replaced by a lateral distribution on a plane at a representative depth, d. We further assumed that the temporal magnetization change

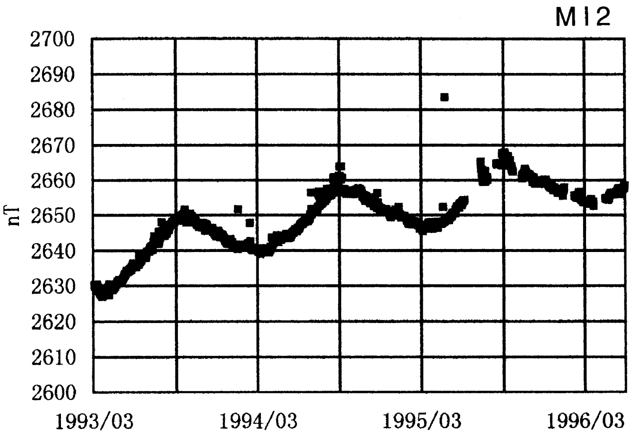

Fig. 5. Daily means of the total intensity at MI2 from December 1992 to December 1997, with reference to NOM.

simply follows the temperature change $\Delta T(d, t)$. Thus simplified, Fourier analyses of the total intensity (Figs. 6(a)–(c)) and ground temperature (Fig. 7) data give us the amplitude ratio of annual changes and the phase difference between the total intensity and temperature. From the phase difference, we can estimate a characteristic depth, d, of the source of annual change and the amplitude of temperature change, $\Delta T(d)$ ($= \Delta T_0 \exp[-nd]$) at this depth by using (8) and (9). The result is summarized in Table 1.

The same value of κ was used for the calculations, which may result in error of parameters to be determined, d and $\Delta T(d)$. However, this effect is supposed to be not so severe, as κ appears in a square root in (9). If the thermal diffusivity at MI2 is κ' and $\kappa' = a\kappa$, the characteristic depth d' will be estimated as $d' = \sqrt{a}d$. However, it is obvious from (8) and (9) that error in κ only affects the estimate of the characteristic depth but not the estimate of the amplitude of temperature change at this depth (i.e., $\Delta T(d') = \Delta T(d)$).

Thus, we substituted both temperature data and the thermal diffusivity obtained at EBI in Kirishima volcano for those at MI2. Overall error due to this assumption is, however, only the difference in the amplitudes of annual temperature change between the two areas, which is not serious for the present purpose.

5. Laboratory Experiment

Here we examine the temperature dependence of the magnetization of rocks sampled from each magnetometer site. The magnetization in ferromagnetic minerals in volcanic rocks consists of the remanent and induced magnetization, which in general cannot be separated in the presence of the ambient magnetic field. However, under a limited condition of a room temperature and a weak magnetic field, we assume we can treat them as independent. In this case, the magnetization **J** can be expressed by a summation,

$$\mathbf{J} = \mathbf{J}_R + \mathbf{J}_I, \tag{10}$$

where $\mathbf{J}_R$ and $\mathbf{J}_I$ are the remanent and induced magnetization, respectively. Using this assumption, we independently measured the temperature dependence of each component.

In Kirishima, we have taken two and three rock samples around the sensors of SMN and SMW, respectively. At MI2 in Izu-Oshima, we took two samples, one from a volcanic bomb of 1986 ejecta and the other from an older lava flow. From each rock sample, three or four cylindrical specimens (length: 2 cm, diameter: 2.4 cm) were taken for magnetic measurements (Neki, 1999).

Figure 9 illustrates an experimental setup to measure the temperature dependence of the induced magnetization. With this system, the magnetic susceptibility was measured by changing the temperature of a rock sample from 0 to 40°C. Temperature was controlled by water circulating around the sample holder. At each measurement, water circulation was done for 15 minutes after water temperature in the tank became constant, in order to make the temperature inside the sample uniform. As shown in Fig. 10, for example, the susceptibility showed a good linearity to temperature but the dependence was not so strong. The temperature dependence was represented by a mean gradient between 24 and 40°C. Results are summarized in Table 2 together with the result of remanent magnetization measurements described below.

According to the single domain theory of the thermoremanent magnetization (TRM) by Néel (1949), the TRM intensity J_R in a ferromagnetic mineral at a room temperature, T, below a blocking temperature can be given by,

$$J_R = vJ_s \tanh\left[\frac{vJ_sF}{kT}\right]_{\text{blocking}}, \tag{11}$$

where v, J_s, F, and k denote the volume of the particle, the intensity of the spontaneous magnetization, the ambient magnetic field intensity, and Boltzmann constant, respectively. Note that the parameters in the argument of the hyperbolic tangent should be measured at the blocking temperature (Merrill *et al.*, 1996). Therefore, the TRM intensity expressed by (8) will have two kinds of temperature dependence, one due to the temperature change of the spontaneous magnetization J_s and the other due to the difference in the blocking temperature among ferromagnetic rock-forming minerals. However, the former effect turned out to be negligibly small by a measurement of the saturation magnetization, J_S (Table 3), to which J_s is supposed to be proportional (Merrill *et al.*, 1996). Therefore we neglect this effect hereafter.

Effects of the blocking temperature distribution were estimated by thermal demagnetization experiments at a room temperature (24°C) and a high temperature (40°C). In the actual situation, temperature of rocks changes with a typical time scale of one month. However, it may not be realistic to undertake an experiment that takes a month for one measurement. In this study, we measured the remanent magnetization after heating each rock sample at 150°C for 10 minutes. During the heating stage, the intensity exponentially decays with time following the equation,

$$J_R(T, t) = J_R^0 \exp\left[-\frac{t}{\tau}\right], \tag{12}$$

where J_R^0 is the initial value of the remanent magnetization. τ is a relaxation time constant given by,

$$\tau = \frac{1}{C} \exp\left[\frac{vh_cJ_s}{2kT}\right], \tag{13}$$

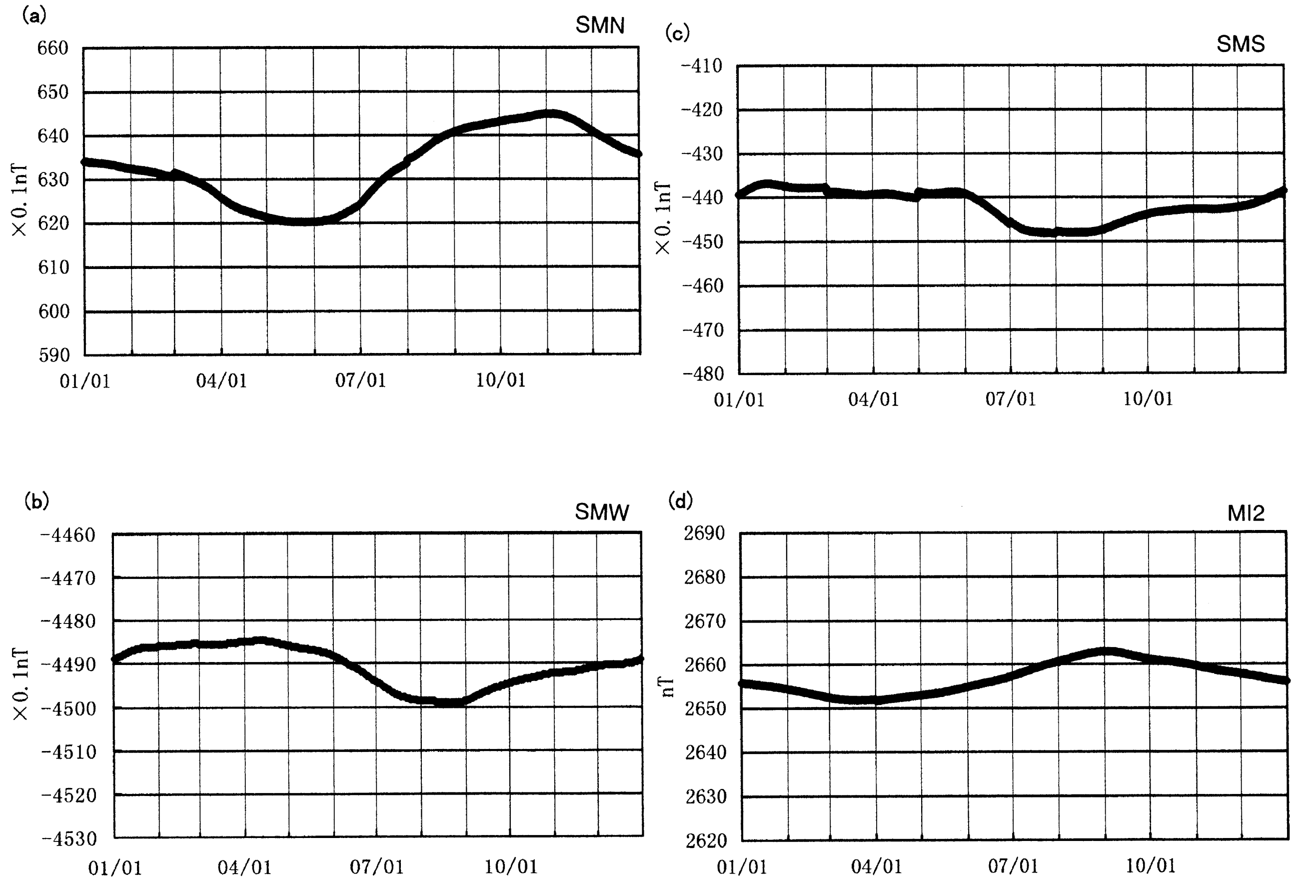

Fig. 6. Stacked annual variation of the total intensity at SMS (a), SMW (b), SMS (c) and MI2 (d).

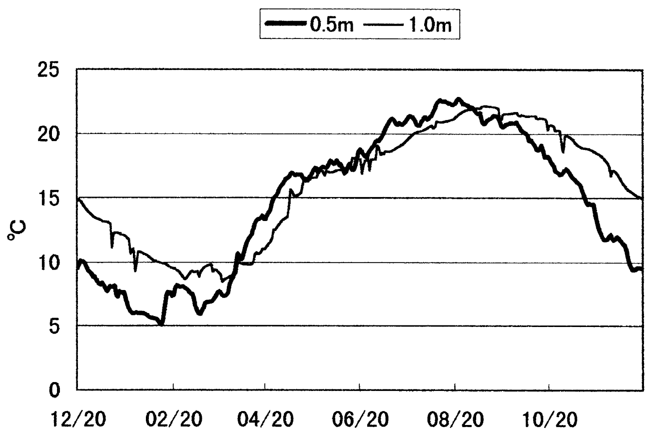

Fig. 7. Ground temperature variations at 0.5 and 1 m depths measured at EBI from December 1997 to December 1998.

where C is the frequency factor (10^8 sec^{-1}), and h_c denotes the microscopic coercive force of a single domain crystal (Néel, 1949). From (10) we estimated a heating time at 150°C for 10 minutes equivalent to a heating at 40°C for a month. For each sample, we repeated several heating/cooling measurements and confirmed that results show a good agreement within 10% difference. This indicates that the irreversible component is not dominant.

Table 1. Result of Fourier analyses of the total intensity data at three sites and ground temperature at EBI for 1 yr period. Negative amplitude indicates that the polarity is reversed.

	SMN	SMW	MI2
Amplitude of F, nT	1.1	−0.6	4.9
Phase (F-T), deg.	70	9	27
d, m	2.4	1.2	1.5
$\Delta T(d)$, deg.	4.5	12.8	9.4

As a result, we found that the temperature dependence of the remanent magnetization has the most dominant effect in overall variations of the magnetization of rock samples as shown in Table 2. In the table, $Amp(d)$ is relative amplitude of annual variation of the magnetization calculated from the gradient in this table and the amplitude of annual temperature variation, $\Delta T(d)$, in Table 1, assuming a linear relation in the small temperature range.

6. Magnetic Structure Around the Sensor and Its Seasonal Variation

In the hypothesis that we are testing in this paper, measured annual variation in the total intensity is supposed to be caused by a temporal change of heterogeneous magnetization due to a temperature variation in the ground just below the

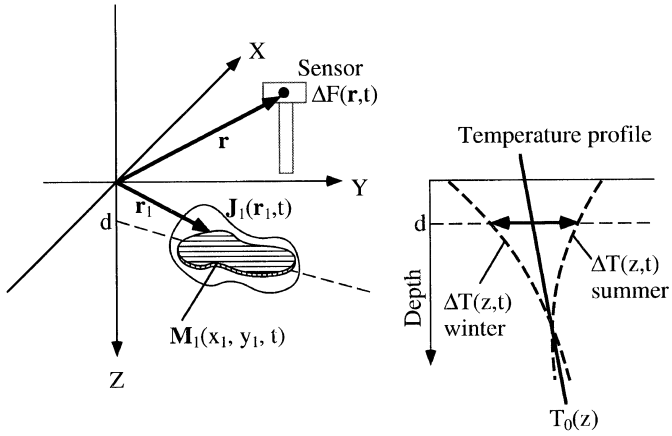

Fig. 8. A schematic model describing the present hypothesis. The magnetization of near surface rocks $_b f J_1$ creates magnetic anomaly of short wavelength. The volume magnetization distribution can be approximated by a surface distribution $_b f M_1$ shown by the dark area at depth d. Temporal changes in the magnetization due to temperature variation that penetrates from the surface will cause a seasonal variation in the local magnetic anomaly and thus the observed total intensity.

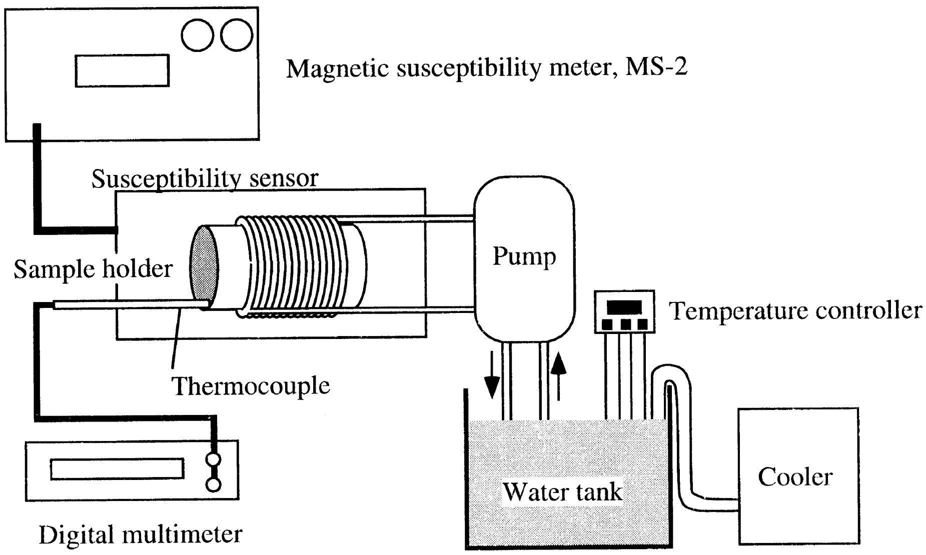

Fig. 9. An instrumental setup of the laboratory experiment to examine the temperature dependence of the induced magnetization, in which the susceptibility of a rock sample is measured at different temperatures. Temperature of the sample is controlled by water circulating around the sample holder.

sensor of a proton magnetometer. The inhomogeneous magnetization at such a shallow depth creates a local magnetic anomaly of short wave lengths (typically a few meters). If the source magnetization changes are exclusively due to the temperature effect, the local magnetic anomaly is expected to change with the same spatial scale. In order to identify the source magnetic anomaly, we carried out a magnetic survey at each site. Each measurement was taken at every 1 m interval in a square area of 10m×10m by using a portable proton magnetometer. The height of measurement was the same as the sensor height of each continuous observation. Resulting maps of total intensity distributions are shown in Figs. 11–13. Values shown in these figures are the difference between each measurement and simultaneous value of continuous measurement at the center of each survey area. The local short-wavelength anomaly was extracted by fitting a spatially linear polynomial,

$$\Delta F(x, y) = \Delta F_0 + f_x x + f_y y, \tag{14}$$

to each map. Residuals are shown in Figs. 14–16.

Assuming that the magnetic anomaly shown in these figures is caused by an equivalent source (Dampney, 1969) at a characteristic depth d given in Table 1, amplitude of its annual change at each site can be estimated by the anomaly value at the sensor position and the relative amplitude of annual variation of the magnetization (i.e., $Amp(d)$ in Table 2). For Izu-Oshima, we took the value of MI2(1) because we need an anomaly source responsible for the most dominant effect. Thus estimated amplitudes of the total intensity variations were compared to the observations given in Table 1 (Fig. 17). A good agreement between the estimated and observed values indicates that the present hypothesis is correct.

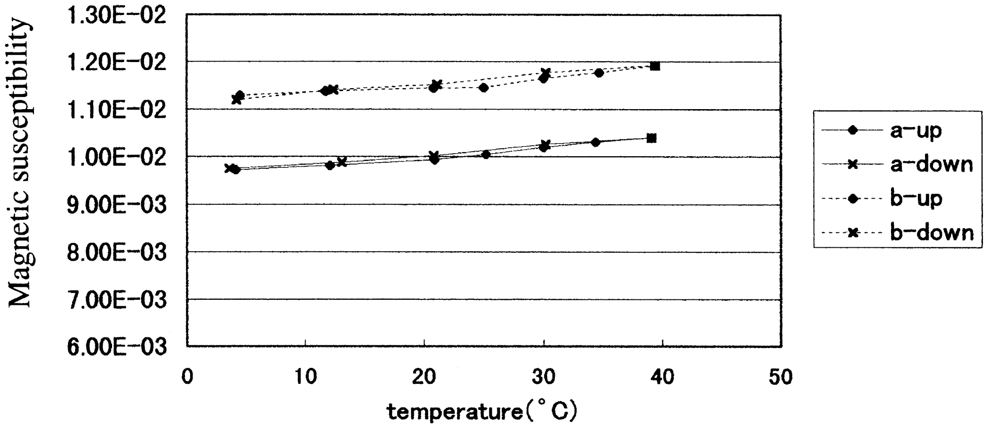

Fig. 10. An example of susceptibility-temperature relation. Two curves indicate the temperature dependence of the susceptibility of two rock samples from SMN.

Table 2. Temperature dependence of the magnetization. Values of the induced magnetization were calculated from measured susceptibility and the total intensity at each site. MI2(1) and (2) are referred to the sample from 1986 ejecta and older lava flow, respectively.

	J_R, A/m		J_I, A/m		$\frac{1}{J}\frac{dJ}{dt}$, $\frac{1}{°C}$	$Amp(d)$
	24°C	40°C	24°C	40°C		
SMN	1.099	0.819	1.157	1.224	−0.094	0.385
SMW	8.896	8.586	1.061	1.131	−0.024	0.035
MI2(1)	15.04	12.87	0.803	0.844	−0.134	0.084
MI2(2)	22.52	22.35	0.373	0.386	−0.007	0.003

Table 3. Temperature dependence of the saturation magnetization measured by a vibration magnetometer with the ambient magnetic field of 0.5 T.

	J_S, $\times 10^3$A/m		$\frac{1}{J_S}\frac{dJ_S}{dt}$, $\frac{1}{°C}$
	20°C	40°C	
SMN	2.812	2.822	0.010
SMW	4.456	4.462	0.006
MI2(1)	2.743	2.750	0.007
MI2(2)	2.883	2.902	0.019

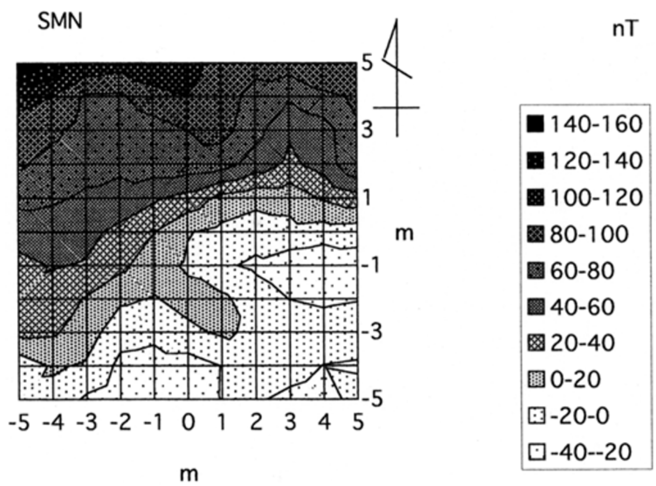

Fig. 11. A total intensity map around the sensor of a proton magnetometer at SMN. Measurements were done at 1 m interval in 10 m×10 m area by using a portable proton magnetometer. Contour indicates difference between each measured value and the simultaneous value at SMN (at the center of the survey area).

7. Discussion

By means of field and laboratory experiments, we obtained a good agreement between annual variations of the geomagnetic total intensity expected by a hypothetical model and those observed. The following two points are especially noteworthy that the model explains:
(1) The large difference in amplitudes at Kirishima and Izu-Oshima.
(2) The anti-phase feature at SMW.
The first point can be ascribed to the intense magnetization of basaltic rocks in Izu-Oshima volcano. The second point turned out to be due to the relative position of the magnetometer sensor with respect to the position of the major source of the local magnetic anomaly. If the near surface magnetic structure is approximated by a single dipole source aligned to the direction of the present geomagnetic field, it creates a pair of local total intensity anomalies, one positive to the south and the other negative to the north in middle latitudes (Fig. 18). When the dipole moment changes periodically with time, total intensity values measured in the areas of positive and negative anomalies will change with reversed

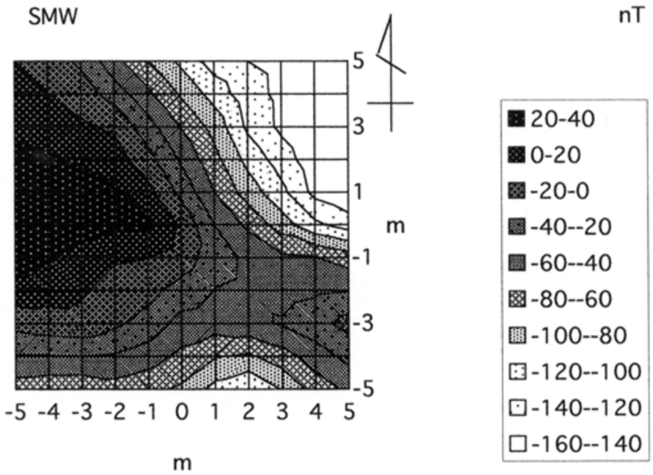

Fig. 12. A total intensity map at SMW.

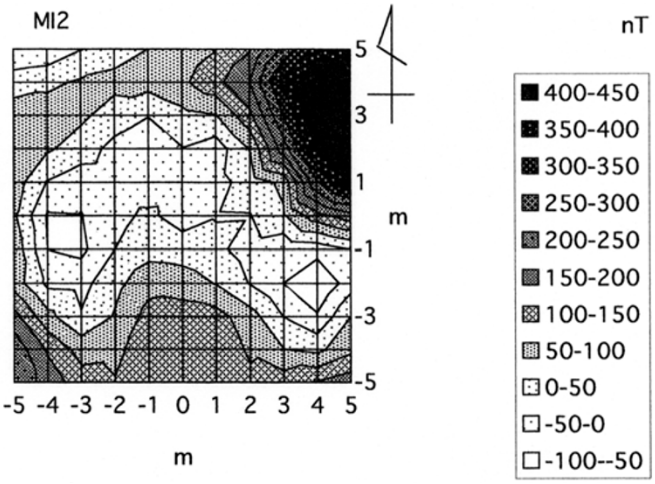

Fig. 13. A total intensity map at MI2.

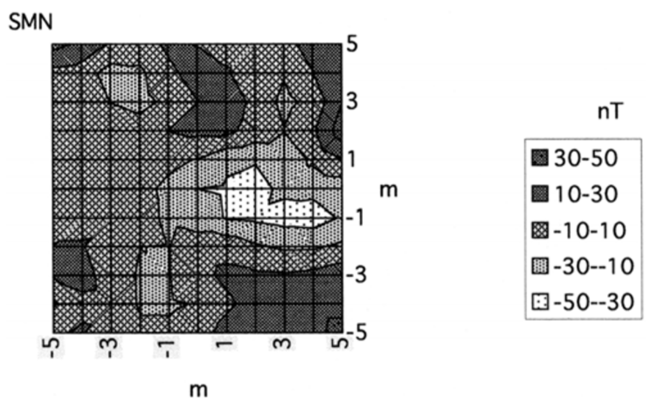

Fig. 14. Local magnetic anomaly around SMN.

polarity. This simply explains the observed feature.

Ozima *et al*. (1996) qualitatively considered a similar temperature effect to explain the feature of annual variations of the total intensity observed in Aso volcano. However, they took into account the effect of the regional magnetic anomaly to explain the different polarity of annual variations. As a

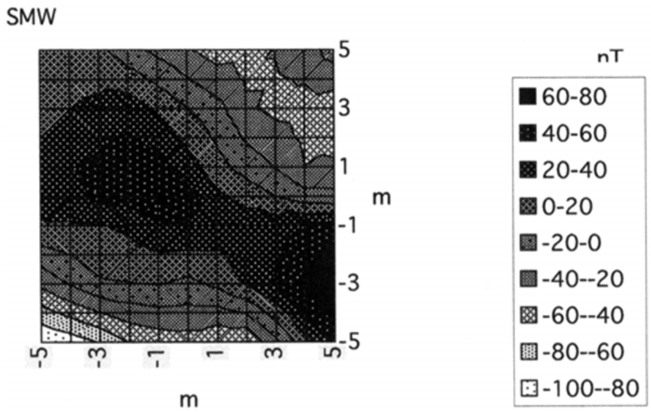

Fig. 15. Local magnetic anomaly around SMW.

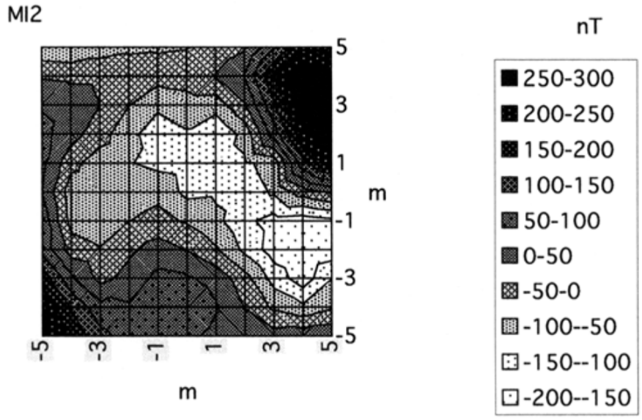

Fig. 16. Local magnetic anomaly around MI2.

result, they did not explain some of the important features such as phase differences and polarity reversals, although their basic idea is the same as that of this study.

In the proposed model, annual variation in the total intensity is a local phenomenon simply due to the near-surface heterogeneous magnetic anomaly. Its amplitude and polarity simply reflect the amplitude and pattern of the total intensity anomaly of a local scale. Its phase difference is simply reflects the time delay due to the diffusion of atmospheric temperature variation to the depth of the source magnetization inhomogeneity. Therefore, in order to avoid such effects, it is recommended that the sensor of a proton magnetometer be installed where the magnetic anomaly is as weak as possible, or in other words, where the field gradients are as small as possible (Campbell, 1997).

Magnetic observatories such as those used as a reference in this study were selected in this way. In volcanic environments, however, it is often hard to find such a place with sufficiently weak magnetic anomaly because of the strongly magnetized volcanic rocks. Therefore contamination of some amount of annual variation would be inevitable as we have seen in this study. Our results showed that these variations are ascribed to seasonal changes of the atmospheric temperature, and therefore indicate that data correction of temperature effects will be effective. The proposed model implies that the effect, $\Delta F_T(t)$, in temporal changes of the total intensity can

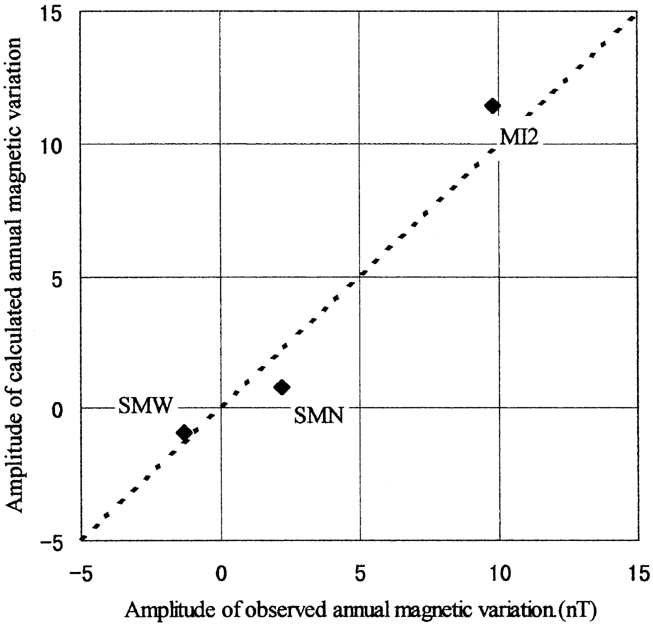

Fig. 17. Comparison between amplitudes of total intensity annual variation estimated from the model and those observed at SMN, SMW and MI2. Negative amplitude means that the polarity of this variation is reversed.

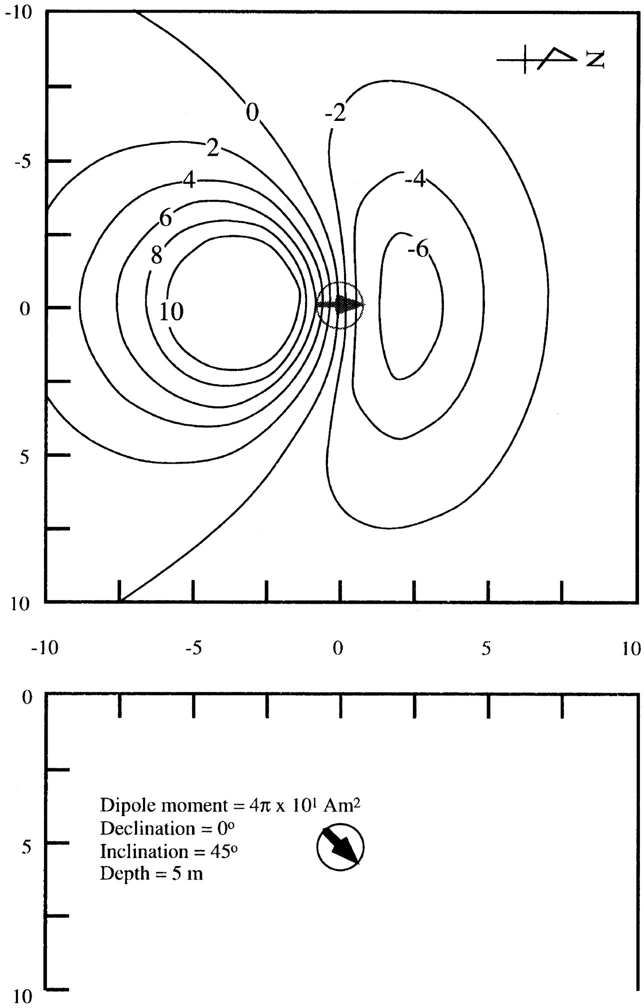

Fig. 18. A map of local magnetic anomaly due to a magnetic dipole placed 5 m below the center with contour interval of 2 nT (above). The magnetic dipole is assumed to be parallel to the ambient geomagnetic field with intensity, declination and inclination as given in the figure (below).

be expressed by a convolution form as,

$$\Delta F_T(t) = \int_{-\infty}^{\infty} h(\tau)\Delta T(t-\tau)d\tau, \tag{15}$$

where $h(\tau)$ is the impulse response and ΔT is time variation either in the atmospheric or in the ground temperature. If $h(\tau)$ is determined beforehand by a least squares method, the temperature effect can be removed from original data by using (15). Note that a rapid temperature change will not have large contribution to geomagnetic change, because of the large time constant of the remanent magnetization (13) at this temperature range. This situation will be much changed in the presence of a new extrusion of high temperature where the time constant becomes much smaller.

In a study of volcanomagnetic effects, correction of the temperature effects will provide more reliable result especially when the expected signal amplitude is comparable to or smaller than the amplitude of annual variations. As an application of the present result, an example will be shown below in which we try to estimate the volume of a demagnetized region from the observed total intensity data in Kirishima. For this purpose, we retried to extract a volcanomagnetic signal by removing annual variations with a simple one-coefficient filter for (15) as,

$$\Delta F_T(t) = h_0 \Delta T(t-t_0), \tag{16}$$

where h_0 is the estimated amplitude ratio and t_0 is time lag calculated from the phase differences in Table 1.

Figure 19 shows that even such a simple filtering is effective enough and that the changes seen in 1992–1993 are the only significant volcanomagnetic signal during the observation period. Using the modeling method proposed by Hamano *et al.* (1990), the source position for the magnetic change was determined by a grid search (Neki, 1999). A contour map of χ^2-misfit between observed and calculated total intensity changes (Fig. 20) suggests that the source exists to the northwest of Shinmoe-dake crater at about 500 m below the surface. The change in the magnetic moment is estimated as 5×10^6Am^2. If we take the magnetization of 1 A/m as a typical value for rocks in Kirishima volcano, the observed change corresponds to the volume of a totally demagnetized region of 5×10^6m^3. Details will be reported elsewhere (Kagiyama *et al.*, 2000, in preparation).

Above example has shown that a simple linear filter correction of the temperature effect dramatically improves the detectability of volcanomagnetic signals. However, there still remains periodic annual variation especially in SMN data (Fig. 18a), which is supposed to be ascribed to non-linearity of the temperature effect. Simply speaking, it is due to the temperature dependence of the relaxation time constant given by (13). This dependence is supposed to result in asymmetric seasonal variation in the total intensity during summer and winter times, though that of temperature is almost symmetric. It is impossible to model this effect by a simple linear filter but a non-linear filter with a temperature

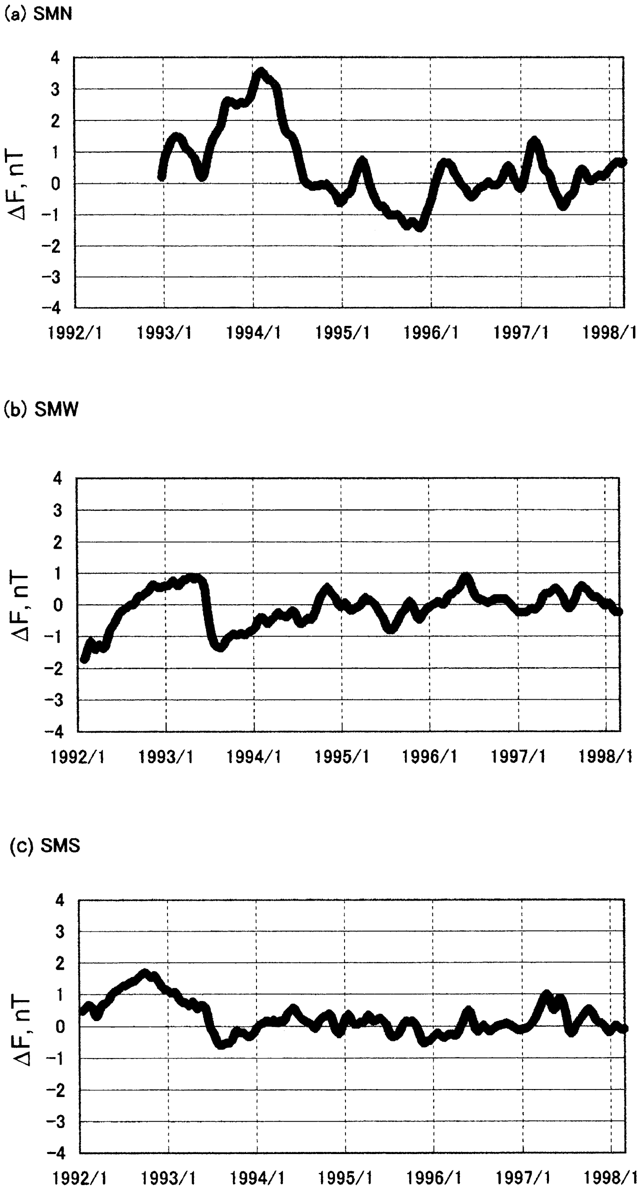

Fig. 19. Geomagnetic total intensity changes at SMN (a), SMW (b) and SMS (c) around Shinmoe-dake crater in Kirishima volcano after removal of temperature effects.

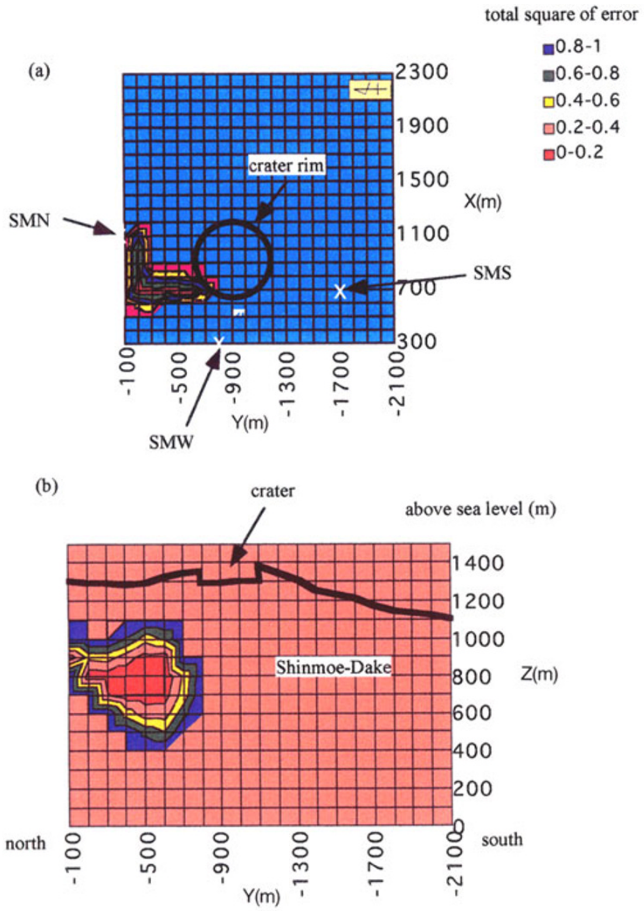

Fig. 20. Plan view at Z = 800 m (a) and N-S cross section at Y = 700 m (b) of the contour map of χ^2 misfit between total intensity changes observed at three sites, SMN, SMW, and SMS and those estimated by a dipole model.

dependent coefficient $h({}_tau)$ will be required.

8. Conclusion

This paper proposed a model for the cause of annual variations in the geomagnetic total intensity with special attention to detecting volcanomagnetic signals. It was assumed in the model that the annual geomagnetic variation is caused by seasonal changes in the near-surface heterogeneous magnetization due to a diffusion of atmospheric temperature change into the ground. The hypothetical model was tested by analyzing field data from Kirishima and Izu-Oshima volcanoes and by laboratory experiments on the rock samples from these two volcanoes. Results have shown that the features of annual variations can be quantitatively explained by the proposed model. By applying this model, a method was recommended to correct the effect from original total intensity data. Finally, the volcanomagnetic signal was detected more accurately from original data in Kirishima volcano associated with an activity in 1992–1993, and the signal was shown to correspond to the totally demagnetized volume of $5\times10^6\text{m}^3$.

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