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Experiments on the Stress Sensitivity of Natural Remanent Magnetization

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Ten rock samples were progressively demagnetized by uniaxial compressions up to 300 bars, and the results were compared with the alternating field demagnetization of the same rocks. Relative change of the NRM intensity during the stress demagnetization ranges from 0.4% to 47.7% at 100 bars. The stress sensitivity correlates with the stability to AF demagnetization, and the directional changes during the two types of demagnetization were similar to each other. These observations indicate that the stress level which affect a remanence relates with the magnetic coercivity of the remanence. For a quantitative comparison, equivalent magnetic fields of the uni-axial stress were calculated as a ratio of the stress sensitivity to the magnetic stability. The equivalent field ranges from 2.3 oe/100 bars to 37.5 oe/100 bars, and the porous rocks tended to have larger equivalent fields. The porosity effect can be attributed to the stress intensification on the magnetic minerals embedded in rocks. Based on the present observations, it is concluded that the magnetic stability and the porosity of rocks have dominant effect on the stress sensitivity of remanent magnetization.

1. Introduction

Local changes of the earth's magnetic field associated with earthquakes have been reported by several authors (KATO and UTASHIRO, 1949; RIKITAKE, 1968; FUJITA, 1965; TAZIMA, 1968; ISPIR and UYAR, 1971; SMITH and JOHNSTON, 1976; for reviews, NAGATA, 1969; RIKITAKE, 1976). Most of the observed changes were attributed to the earthquake-related stress effects. The phenomena is considered to be useful to understand and to predict earthquakes. Although the fundamental mechanism of the stress effects on the magnetic properties of rocks has been investigated by extensive laboratory experiments (for example, KAPITSA, 1955; NAGATA and CARLETON, 1968; 1969a; NAGATA and KINOSHITA, 1965; OHNAKA and KINOSHITA, 1968b; KEAN *et al.*, 1976; POZZI, 1977; REVOL *et al.*, 1977, 1978; MARTIN, 1980; HAO *et al.*, 1982), more investigations are required to apply the experimental results to *in situ* field observations.

One of the main purpose of the further laboratory study is to find some criteria for good magnetic stress sensors, which enable us to choose appropriate site for the magnetic field observations. Previously, OHNAKA and KINOSHITA (1968a) de-

monstrated that the rate of the stress change of the initial susceptibility increases with the increase of Ti-content in the titanomagnetite solid solutions. Besides this compositional effect, KEAN *et al.* (1976) pointed out that the susceptibility of large multi-domain grains are much more sensitive to stress than small single-domain grains. The grain size effect has also been observed in the stress change of remanences (OHNAKA, 1969). In addition to this effect, OHNAKA and KINOSHITA (1968b) found that magnetically soft remanences such as IRM (Isothermal Remanent magnetization) are more susceptible to stress than the hard remanences such as TRM (Thermal Remanent Magnetization).

In spite of these previous studies, recent experiments are rather concentrated in hard remanences (MARTIN *et al.*, 1978; REVOL *et al.*, 1977). Main reason of this neglection of the soft remanences in the study of the stress effect is that the soft remanences change irreversibly with stress (NAGATA, 1969), and, therefore, these remanences are regarded as relatively unimportant in the earthquake regions, where the rocks have been exposed to cyclic stress changes. However, this assumption may not be true since VRM (Viscous Remanent Magnetization) is probably the most important secondary remamence in nature. The VRM can be acquired repeatedly if the period of the stress cycle is larger than the characteristic acquisition time of the VRM. Moreover, the directional coherency of the VRM can be better than the original remanences. Based on the above consideration, I think that the role of the soft remanences in the observed magnetic field variation should not be dismissed.

In order to find a good stress sensor, it is practically useful to relate the stress sensitivity to a standard measure of the magnetic hardness. Since the alternating field demagnetization has been a common procedure to test the magnetic stability of remanences, the stress demagnetization and the AF demagnetization of NRM were compared in the present paper, where irreversible changes of the NRM were mainly observed from the point of views presented in the preceding part.

2. Samples and Experimental procedure

The rocks studied in the present work were all collected in Japan. Description of the rock samples and their physical and magnetic properties are summarized in Tables 1 and 2. The rock types are basalt, andesite, welded tuff and scoria. Their sampling localities are also shown in Table 1. Most of the rocks were previously sampled for paleomagnetic studies.

The bulk density and porosity of the present samples were measured because the physical properties of the rocks affect the internal stress when the rocks are stressed. As evident from Table 1, rocks with various porosities were selected. Measurements of the bulk density and porosity were made by a standard immersion technique. Weights of a rock sample in dried and water-saturated conditions, and the bulk volume of the sample give these properties. Although this process can measure only the open pore porosity, the true porosity might be close to the observed value for the present porous rocks. The observed porosity varies from about 6% to 47% among the present samples, where scoria, welded tuff and vesicular andesite have the higher porosities. As will be shown later, this wide range

Table 1. Bulk density, porosity, Curie temperature, magnetic susceptibility, and description of the present samples.

Sample No.	Bulk density (g/cm^3)	Porosity (%)	T_c (°C)	Susceptibility (10^{-3} G/oe)	Description
A	1.36	47.7	492	2.65	scoria, Hakone
B	2.05	10.7	551	3.50	dacite, Shuzenji
C	2.32	7.4	401	0.16	welded tuff, Shibutami
D	2.73	6.1	536	3.18	andesite, Usami
E	2.38	11.3	505	3.33	scoria, Myoko
F	2.72	6.1	172	0.46	basalt, Kinosaki
G	2.06	23.1	545	3.08	tuff, Shirakawa
H	2.78	5.8	536	2.25	basalt, Izu
I	2.15	16.7	536	3.35	andesite, Asama
J	1.95	23.3	503	2.14	andesite, Asama

Table 2. Magnetic hysteresis properties.

Sample No.	J_s (emu/g)	J_r/J_s	H_c (oe)	H_{rc}/H_c
A	0.86	0.032	26.5	4.9
B	1.95	0.018	19.5	9.0
C	0.05	0.051	42	7.3
D	1.73	0.120	87.5	2.4
E	1.59	0.042	36.5	4.8
F	0.25	0.262	82.5	1.8
G	2.23	0.051	53	3.9
H	1.70	0.181	187.5	1.9
I	2.17	0.118	120.5	2.7
J	1.06	0.069	57	14.4

of the porosity value enabled us to investigate the effect of porosity on the stress change of the remanences. During the selection of the samples, sedimentary rocks were excluded because of their low concentration of magnetic minerals, although high porosity are expected.

The Curie temperature was observed with a standard Curie balance in a high vacuum ($<10^{-5}$ Torr.), where the rate of the temperature variation was 500°C/hr. Except one sample(Sample F), the observed Curie temperature is higher than 400°C, and most of them are around 500°C. The result indicates that magnetic minerals in these rocks are Ti-poor titanomagnetite, whereas the magnetic mineral in Sample F is Ti-rich titanomagnetite. J_s-T curves of all the samples are reversible type, suggesting that the samples were not severely altered at low temperature.

The magnetic hysteresis properties listed in Table 2 were measured by using a PAR vibration magnetometer. Large variation of the properties were observed among the present samples. The high J_r/J_s ratio ($=0.262$) as well as the low Curie temperature ($=173$°C) suggest that the magnetic minerals in Sample F are fine

grain titanomagnetites with a x-value around 0.6. In other samples, the observed low J_r/J_s ratio and the high H_{rc}/H_c ratio both indicate that magnetic grains in these samples are large enough to have multi-domain structures.

For the demagnetization experiments, standard core samples with 2.54 cm diameter and about 2.4 cm height were used. Two core samples were cut from each rock and the two ends of each sample were polished as parallel as possible (less than 2/100 mm difference on the surface). The intensity and the direction of the remanences in the two samples from each rock were checked to coincide within a limit (intensity difference $< 10\%$ and angle difference < 5 degrees). The direction of the core axis was selected so that the magnetization in the axial direction is comparable with that in the lateral direction.

In the demagnetization experiments, one specimen from each rock was subjected to AF demagnetization up to 2,000 oe, whereas the other specimen was demagnetized by a uniaxial compression up to 300 bars. The remanences were measured by a SSM-1A Schonstedt spinner magnetometer. The AF demagnetizations were conducted with a three-axis tumbler system, where fine steps were used at the low magnetic field range of less than 50 oe (5 oe interval). For the stress demagnetization a non-magnetic press made of Beryllium–Copper and stainless steel was used. The compressive stress was applied along the axial direction of the samples. A Helmholtz coil system surrounding the press cancels out the ambient field during the compression. After each stress level of compression, the stress was released and the magnetization of the sample was measured by a spinner magnetometer. This process was repeated for the successively higher stresses. This procedure measures the irreversible change of the magnetization during the compression. Hence, it is straightforward to compare the result with the AF demagnetization.

3. Experimental Results

The variation of the NRMs during the AF and the stress demagnetizations are compared in Fig. 1, where the relative changes of the magnetizations along the axial direction and the lateral direction are separately shown. As evident from the figure, the samples are ordered as the magnetic stability increases. Hence, Sample A has the lowest coercivity spectrum and Sample J has the highest coercivity among the present samples.

The remanence in Sample A is magnetically very soft and the MDF (Median Destructive Field) is less than 40 oe. But the directional change during the AF demagnetization is small. The intensity change due to stress in this sample is as large as 20% at 40 bars, and the directional change is also small. Since this sample is scoria and is very fragile, maximum stress attained before fracture was 70 bars. In samples B and C, the remanences contain soft secondary components and the directional change during the AF demagnetization is appreciable. The large directional change is also observed in the stress demagnetization. In sample B, the magnetization in the axial direction remains constant, whereas the magnetization in the lateral direction decreases linearly with the increase of stress. In sample C, the axial magnetization increases and the other component decreases. These features

can be observed in the initial part of their respective AF demagnetization curves.

Samples D, E and F show more stable characteristics against the AF and the stress demagnetizations. Directional change during the demagnetizations are small. The MDFs of the three samples are around 100 oe, and slightly increases with this order. However, Sample E shows the largest change during the stress demagnetization among these three samples. It is worth noting that the porosity of this sample is larger than the other two samples. Directional change of Sample G is large, where the magnetization in the axial direction decreases and that in the lateral direction increases. This type of the stress change is commonly observed in the reversible change of the hard remanences and the susceptibility, and can be explained by a single-domain rotation model. But it is to be noted that the presently observed change is irreversible and the AF demagnetization of this sample shows the same tendency.

The last three samples (H, I and J) are more stable than the previous samples. Their stability to both the demagnetizations increases with this order. Although the variation in Sample H is small, larger decreasing rate in the axial direction than in the lateral direction is observable in the two demagnetization curves. The stress variation in Samples I and J is very small and their magnetic stability is also large.

As overviewed above, the stress variation of the NRMs relates with the magnetic stability of the NRMs. The difference of the variation in the axial direction and the lateral direction also relates between the two demagnetizations. These results indicate qualitatively that the magnetically soft component is selectively demagnetized at low stresses. However, the observed non-linearity in the variations prohibits a quantitative comparison. Since the apparent non-linearity arises because of the non-parallelness between the soft component and the hard component, the difference vector during the demagnetization can be simpler. In Fig. 2 the relative magnitude of the difference vector of the magnetization are plotted as a function of the applied stress (Fig. 2a) and of the alternating field (Fig. 2b). At these ranges of the stress and the magnetic field, the variation of the difference vector can be reasonably approximated by a linear line. Therefore, the gradient of the fitted lines can be defined as a stress sensitivity and a magnetic field sensitivity of NRM. The least squares fit was made to obtain the sensitivities. In all the samples but Sample A, all the data points in the stress variation and the points up to 100 oe in the AF demagnetization were used for the fitting. Because of the large change of the magnetization in Sample A, only the points up to 50 oe was used in this sample. The calculated sensitivity is listed in Table 3 and the two sensitivities in each sample are compared in Fig. 3. As mentioned earlier, correlation between the stress and the AF sensitivities is apparent. However, large scatter of the data points suggests that other factors may also control the stress sensitivity. In order to clarify this point, "equivalent magnetic field of stress" was calculated as a ratio of the stress sensitivity to the magnetic field sensitivity. The equivalent field means a magnetic field which can cause the same effect on the magnetization as a specified amount of stress. Hence, if the stress sensitivity is solely determined by the coercivity of the remanence, the equivalent field should be constant. However, as shown in Table 3, this value is highly variable and ranges from 2.3 oe/100 bars to 37.5 oe/100 bars. The range covers more than one order of magnitude. This result indicates that some

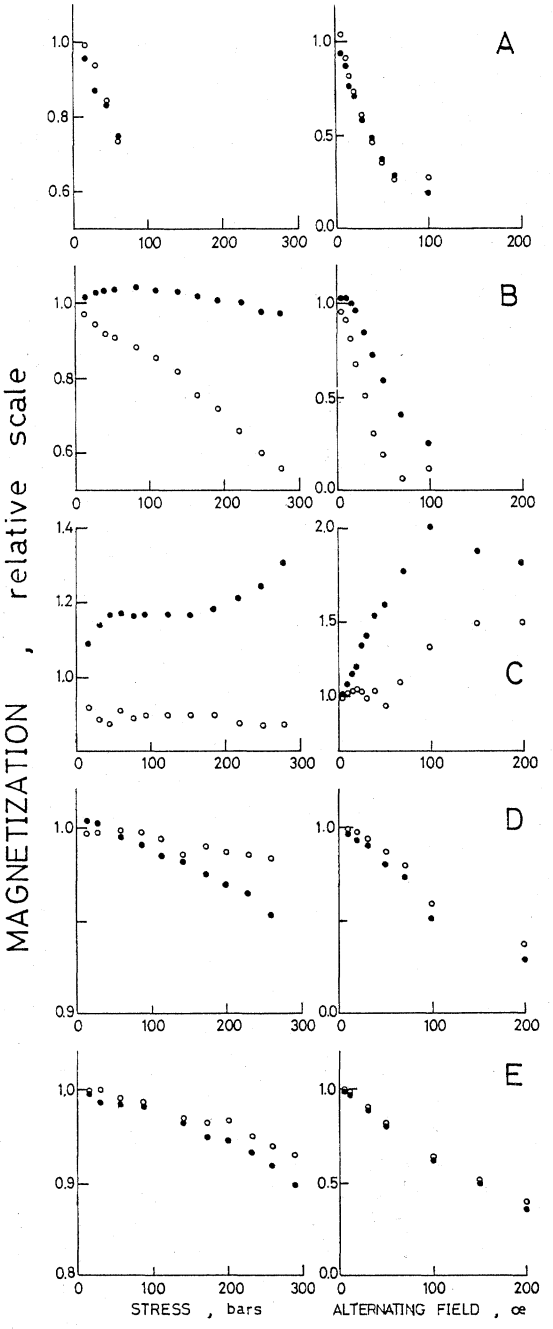

A
B
C
D
E
MAGNETIZATION , relative scale
STRESS , bars
ALTERNATING FIELD , oe

Fig. 1

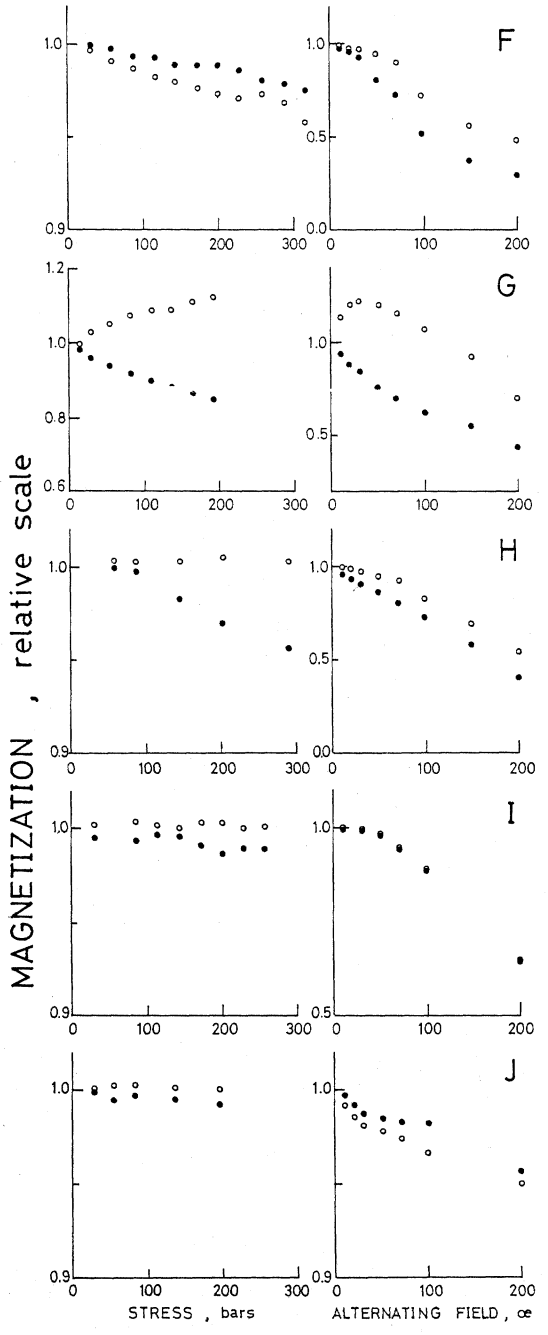
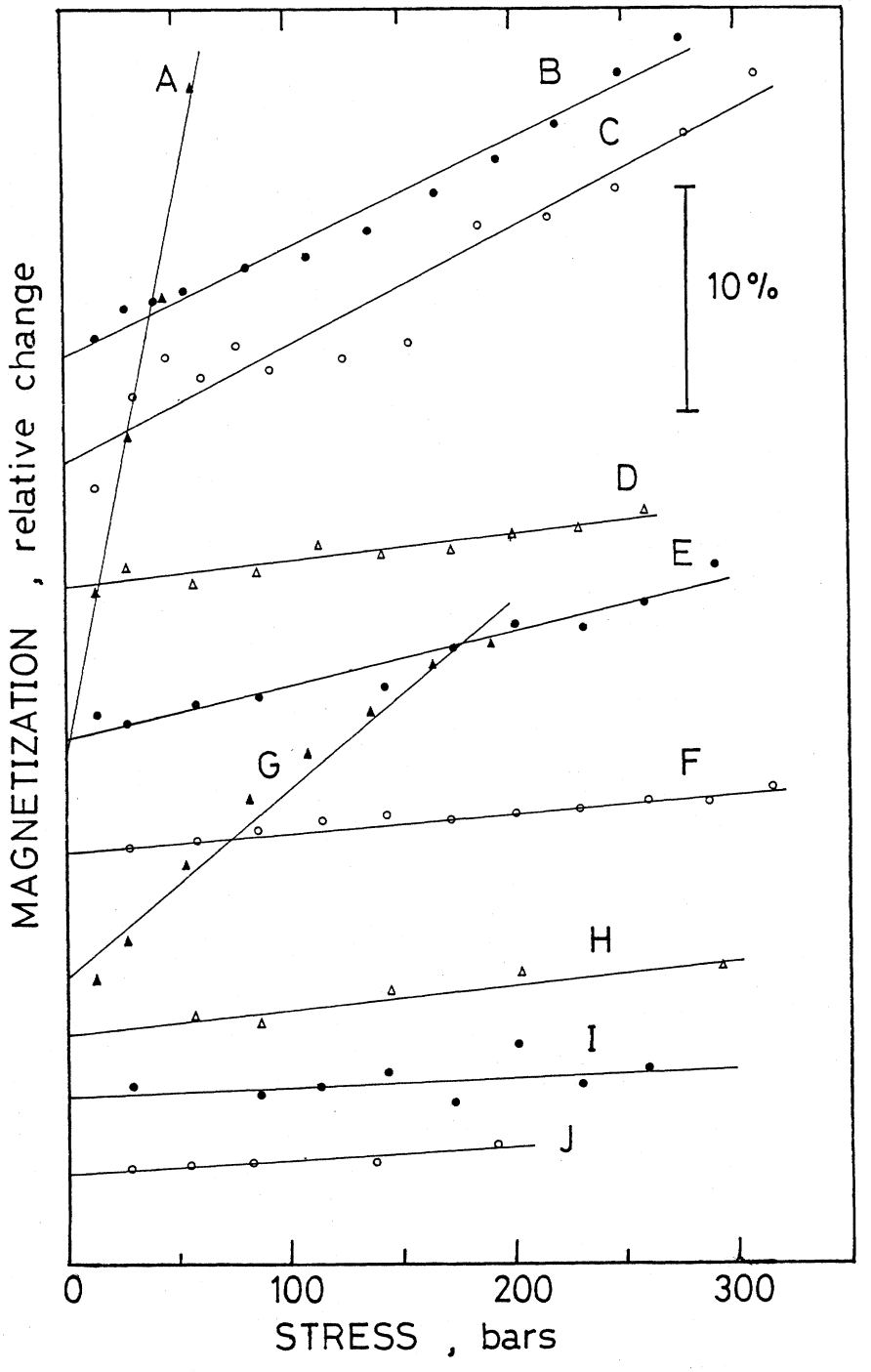
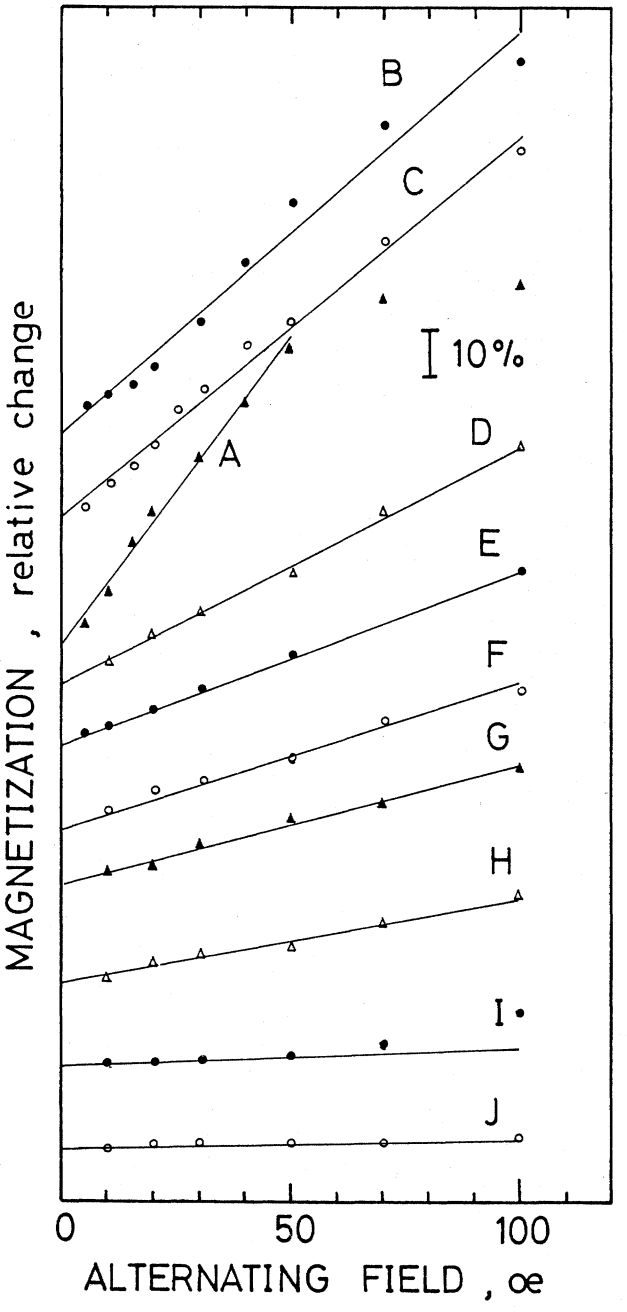


Fig. 1. The relative variation of the magnetizations along the axial (●) and the lateral (○) directions are plotted against the compressive stress (left) and the alternating field (right).


A
B
C
D
E
F
G
H
I
J
10 %
MAGNETIZATION , relative change
0
100
200
300
STRESS , bars

(a)

Fig. 2



(b)

Fig. 2. The change of the relative magnitude of the difference vectors are shown as a function of the compressive stress (a) and of the alternating field (b). The linear lines were fitted by the least squares method.

other factors control the stress sensitivity. As a candidate of the factor, the stress concentration on the magnetic minerals in rocks can be considered. Since the stress concentration depends on the difference in the elastic properties between the bulk rock and the magnetic minerals, the observed sensitivity may depend on the porosity of rocks if the above assumption is valid. We plotted the equivalent fields observed in each rock against the porosity of the rock in Fig. 4. The strong correlation in Fig. 4 suggests the validity of the above assumption.

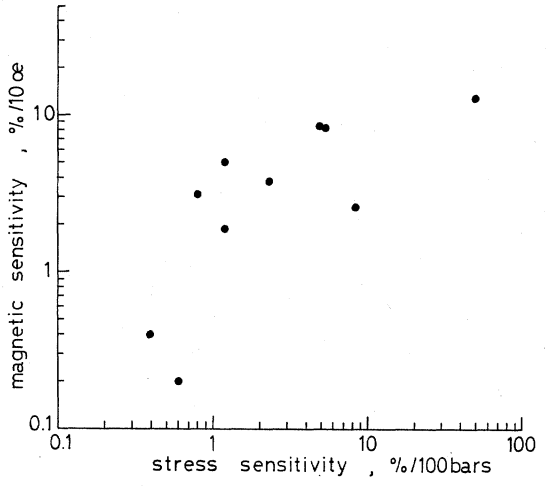


Fig. 3. The stress sensitivity and the magnetic sensitivity are compared for the present ten samples. The values are listed in Table 3.

Table 3. Stress sensitivity, magnetic sensitivity and equivalent magnetic field.

Sample No.	Stress sensitivity (%/100 bars)	Magnetic sensitivity (%/10 oe)	Equivalent field (oe/100 bars)
A	49.5	13.2	37.5
B	4.8	8.7	5.5
C	5.2	8.3	6.3
D	1.2	5.1	2.3
E	2.3	3.8	6.2
F	0.8	3.2	2.4
G	8.4	2.6	31.7
H	1.2	1.9	6.0
I	0.4	0.4	11.1
J	0.6	0.2	37.5

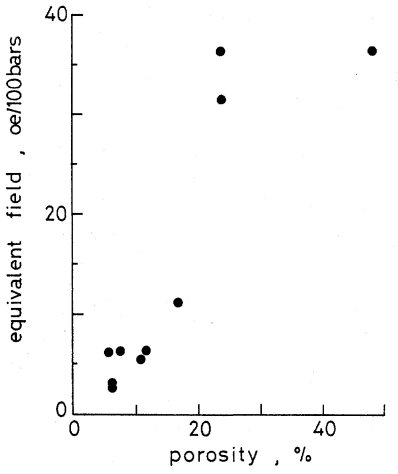


Fig. 4. The equivalent magnetic fields are plotted as a function of the porosity.

4. Equivalent Magnetic Field of Stress

The irreversible magnetic effects of stress was first examined by BROWN (1949). He considered the domain wall movement as a cause of the magnetic response to tensile stresses. His theory was extended to a general stress system by BRUGEL and RIMET (1966). Independently NAGATA and CARLETON (1969b) used the same concept in order to explain their experiments (NAGATA and CARLETON, 1968, 1969a). Recently REVOL (1979) explained the BRUGEL and RIMET (1966)'s theory in detail and made numerical calculations on the magnetic responses in various cases. In the theory originated by BROWN (1949), the effect of stress is replaced by a fictious magnetic field equivalent to the stress, which enables us to calculate the stress effect with respect to the magnetic properties of rocks. Since the theory is the only available one for truly multi-domain grains, and the concept of the equivalent field is compatible with the present experiment, it is appropriate to apply the theory to the interpretation of the present results.

The fictious field equivalent to a uniaxial stress can be expressed as

$$h = W \cos(p) \sin(p) \tag{1}$$

with

$$W = 3g_{111}\, s/J_s \cos(d/2)$$

where p is the angle of the stress direction to the wall surface, d is the angle between the two magnetization vectors separated by the wall, g_{111} is the magneto-striction coefficient in (lll) directions, s is the applied stress and J_s is the saturation magnetization. As evident from Eq. (1), the equivalent field varies with the stress direction and ranges from $-W/2$ to $W/2$. In magnetite grains, the magnetization direction is in (lll) because of the negative first magneto-crystalline anisotropy

coefficient. Therefore, three types of the domain walls with d equals to 180°, 70° 30' and 109° 30' exist. For the 180° walls, the equivalent field given by Eq. (1) is zero irrespective of the stress direction. Hence, the 180° walls are not affected by stress. This result is reasonable considering the uni-axial nature of the stress and the uni-directional nature of the magnetic field. Therefore, the stress affect the other two types of the walls. These two walls are commonly not differentiated and considered as 90° walls. For magnetite particles g_{111} and J_s are given by 78×10^{-6} and 480 emu/cm^3, respectively (SYONO, 1961). Hence, the maximum value of the equivalent field to a stress of 100 bars for the 90°walls ($d = 90°$) becomes about 17 oe. If we assume an isotropic distribution of these 90 degree walls within a rock, the equivalent field for the whole rock reduces to about 9 oe.

As can be easily understood from the discussion above, the equivalent field derived from the experimental results have the same meaning as the equivalent field defined by Eq. (1). However, the values shown in Table 3 can not be directly compared with the value estimated above. In the experimental case, the stress sensitivity was compared with the result of the AF demagnetization, where all the domain walls were affected by the alternating field. On the other hand, the stress demagnetization only affect the 90 degree walls in the sample. Hence, the theoretically estimated value (9 oe) should be reduced depend on the fraction of the 90 degree walls in each sample. The fraction of the wall area of the 90 degree walls is generally not known. BRUGEL and RIMET (1966) estimated that 40% of the total wall area was the 90 degree walls and that half of the walls were favourably situated. In this case, the equivalent field for a total rock becomes about 2 oe. Although the fraction of the 90 degree walls may be different in each rock, 2 to 3 oe of the equivalent field can be a good estimate.

Comparing to the above estimated value, the equivalent field for the most of the present samples is much greater. The large equivalent field is observed in porous rocks (Fig. 4). Hence, the descrepancy may be explained if we consider the stress concentration on magnetic minerals in rocks.

5. Stress Concentration Effect

If a homogeneous material is stressed by a constant surface force, the stress and the strain within the material are constant. However, the rocks are generally considered as an assemblage of minerals with different elastic properties. In these composite materials, the stress and the strain varies with position even if the external force is constant. Therefore, the stress exerted on the magnetic minerals in a rock can be different from the stress exerted on the whole rock.

Mathematical calculation of the elastic properties of composite material has been received much attention in many fields of geophysics. The most elegant and the most useful formulation was given by ESHELBY (1957). For simplicity, we consider magnetic minerals in a rock as inclusions embedded in a homogeneous matrix. If a composite material is uniaxially compressed by a constant external force, the stress within the inclusions is different from the external stress. The internal stress depends on the shape and the orientation of the inclusions and, therefore, is different in each grain. However, the upper and the lower bound of the stress can be estimated. If we

define a stress intensification factor, f, as a ratio of the internal stress to the external stress, the range of the factor can be expressed by the elastic constants of the matrix and the inclusion. In the piezomagnetic effect, only the shear component of stress is important. Then the range of f is from 1 to G_1/G, where G_1 and G denote the rigidity of the inclusion and the matrix, respectively. The factor f in each grain has a different value within the limits, and the factor averaged in a rock is generally not determined. However, HILL (1952) demonstrated that a simple mean of the upper and the lower bounds can be a good estimate. Therefore, we put the factor f as

$$f = (1 + G_1/G)/2. \tag{2}$$

As evident from Eq. (2), the internal stress is intensified when $G_1 > G$, and reduced when $G_1 < G$.

The elastic constants of a single crystalline magnetite have been measured by HEARMON (1956). Since the magnetite particles are assumed to be isotropically distributed in rock samples, the averaged isotropic elastic constants are more meaningful for the present discussion. SIMMONS and WANG (1972) calculated and tabulated the averaged values for minerals. The averaged rigidity for magnetite is $G = 0.91$ Mb. The elastic constants of rocks at atmospheric pressure is highly variable because of the pores and cracks in the rocks. However, the effect of the pores and the cracks can be reduced by applying a hydrostatic pressure, and the elastic constants under high hydrostatic pressures (~ 2 kb) are rather systematic and mainly determined by the mineral assemblage (for example, see BIRCH, 1960, 1961; SIMMONS, 1964; and KANAMORI and MIZUTANI, 1965). The intrinsic elastic properties varies with the density of the rocks and the minerals (ANDERSON *et al.*, 1968). In the present samples, the grain density (porefree intrinsic density) obtained during the porosity measurement are similar and around 2.7 g/cm^3 except two basalt samples (Samples F and H). The intrinsic rigidity of the rocks estimated from the rigidity-density systematics is $G = 0.3 \sim 0.4$ Mb. The value is smaller than that in magnetite by about a factor of two. Hence, the stress concentration on magnetite is expected even under high pressure.

At lower pressures, the pores and the cracks in rocks reduce the bulk rigidity of the rocks. Hence, further concentration of stress on the magnetite is possible. In general, the bulk elastic constants decrease with the increase of the amount of the pores and the cracks in the rocks. This qualitatively explains the apparent dependence of the equivalent field on the porosity of the present rocks. For more quantitative discussion, the effect of porosity on the bulk elastic constants should be calculated. This problem is also important in many respects of geophysical studies, and investigated by many authors. The effect of pores on the elastic constant depends on the shape and the orientation of the pores. Hence, a unique solution can not be obtained without complete knowledge of the pore configuration, which is generally impossible for natural rocks. Therefore, some assumption on the configuration is required. WU (1966) calculated the effective elastic constants for isotropical distribution of ellipsoidal pores. His result is only applicable for low concentration of pores. For higher concentration of pores, O'CONNEL and BUDIANSKII (1974) invented a self-consistent method, which enables us to calculate

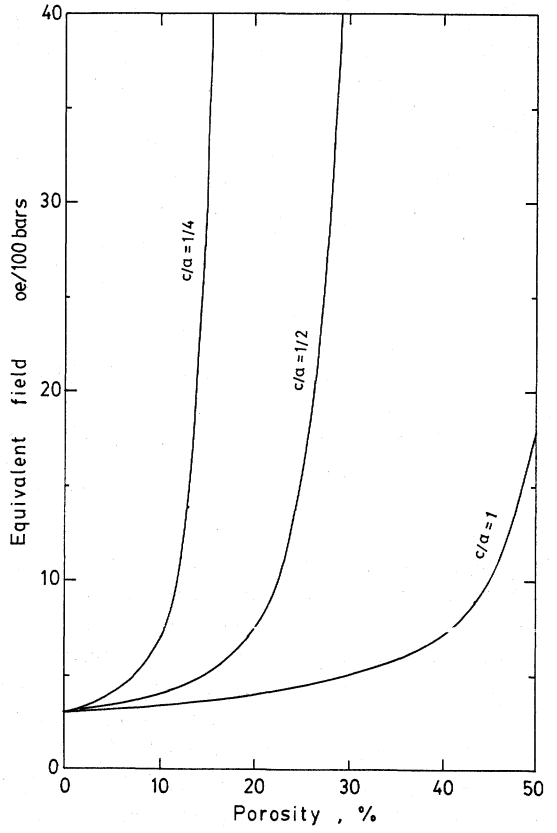


Fig. 5. Results of the model calculation about the variation of the equivalent field as a function of the porosity. c/a denotes the aspect ratio of the pores.

the effective elastic constants up to about 50% of porosity. By using their formulation, it is possible to calculate the variation of the rigidity G as a function of the porosity. Then, Eq. (2) can be used to calculate the stress intensification factor for a given porosity. Further, if we assume the value of the intrinsic equivalent field of magnetite as 2 oe, the porosity dependence of the apparent equivalent field can be calculated. Figure 5 shows the result of the calculation, where an aspect ratio of the pores is used as a parameter. The results indicates that higher concentration of stress is expected with the decrease of the aspect ratio (i. e. more flat pores). Comparing Figs. 4 and 5, we can conclude that the stress concentration effect can explain the observed correlation between the equivalent field and the porosity.

6. Discussion

Since STACEY (1964), possible geomagnetic changes associated with earthquakes have been calculated based on the fault model of the earthquakes by many authors (SHAMSI and STACEY, 1969; TALWANI and KOVACH, 1972;

HILDENBRAND and BHATTACHARYYA, 1974; NAGATA, 1976; SASAI, 1980). In all of the studies, the reversible changes of the remanent magnetization and the induced magnetization were assumed. The reversible magnetization change under a uniaxial compression is represented as

$$\begin{aligned} J^{//} &= J_0^{//}(1 - bs) \\ J^{\perp} &= J_0^{\perp}(1 + (b/2)s) \end{aligned} \tag{3}$$

where superscripts // and $\perp$ denote the magnetization component parallel and perpendicular to the applied stress, respectively, and 0 indicates the unstressed state. Equation (3) can be easily generalized to a general stress system (STACEY *et al.*, 1965). In most of the papers above, b is assumed as 10^{-4} bar^{-1} and J_0 as 10^{-3} emu/cm^3. The estimated changes of the magnetic field are at most 10 nT at the occurrence of the earthquakes. The result is somewhat discouraging for the use of the seismomagnetic effect as an earthquake precursor. Because the rate of the field change before an earthquake is usually assumed to be much less than the change at the earthquake.

In the present paper we observed the irreversible change of the remanent magnetization. The irreversible change can be expressed as

$$J_i = J_i(1 - c|s|) \tag{4}$$

where s is the maximum stress difference in each point, and i (= 1, 2, and 3) denotes each component in a rectangular coordinate system. The observed sensitivity c ranges from 0.4×10^{-4} bar^{-1} to 49.5×10^{-4} bar^{-1} with a mean value of 7.4×10^{-4} bar^{-1}. The absolute value of the magnetization change, $c|J|$, ranges from 1.4×10^{-6} to 1.5×10^{-4} emu/cm^3/bar with a mean of 3.6×10^{-5} emu/cm^3/bar, where nine samples out of ten gives the magnetization change greater than 10^{-5} emu/cm^3/bar. Because of the small number of the samples ($N = 10$), the mean value can not have a great meaning. However, it can be said that many samples show the irreversible magnetization change much larger than the commonly used reversible change. Therefore, larger geomagnetic field variation can be expected due to the irreversible change of the remanent magnetization, if we choose an appropriate site for the observation. In order to find a good observational site, magnetic properties of the crustal rocks at the site can be a useful clue, where the AF demagnetization of the NRM of the rocks at low magnetic field (< 10 oe) can be used as a measure of the stress sensitivity of the rocks.

Because of the different manner of the variation in the magnetization, the change of the geomagnetic field due to the reversible and the irreversible piezomagnetic effect is also different. If we fix the coordinate system of X, Y, and Z as North, East and Down, respectively, and assume that a shear stress of s_{xy} within a sphere is reduced to zero, the change of the magnetization in each case is calculated. In both cases we assume that the direction of the original magnetization is parallel to the present field. For example, we can assume that Inclination $I = 45°$ and Declination $D = 0°$. Then the direction of the magnetization due to the reversible effect is given by $I = 0°$ and $D = -90°$, whereas the direction in the irreversible case is given with

$I = -45°$ and $D = 180°$. The difference in the magnetization causes the difference in the observed geomagnetic field change. In more realistic case, we must consider the stress change accompanied with a fault motion. In this case, the stress along the surface of the fault reduces, whereas the stress increases at the chip of the fault. Therefore, the change of the magnetization in these two parts is in opposite senses if the change is reversible. Because the geomagnetic field change can be obtained by an integration of the magnetization change, the field change accompanied with the fault motion generally decreases compared with the simple sphere model above. On the other hand, the magnetization change of the irreversible effect is in the same direction at all the part around the fault. Hence, larger field change is expected in the irreversible case even if the sensitivity is the same.

At present, we do not know which effect is more dominant under *in situ* condition. However, since many observations have not been explained by the reversible model, the calculation of the field change based on the presently observed irreversible model may give more insight to the geomagnetic change associated with earthquakes.

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