Notes on Piezo-remanent Magnetization of Igneous Rocks

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Abstract

The acquisition of piezo-remanent magnetization (PRM) of igneous rocks under uniaxial compression is experimentally studied. It has been found that the rate of acquisition of PRM is several times as large as that of assemblages of magnetic minerals. This is in remarkable contrast to another observed fact that the rate of decrease of magnetic susceptibility owing to an increase of uniaxial compression in rocks is about one order smaller than that in magnetic minerals.

The rate of acquisition of PRM is larger in a smaller magnetic field than in a larger magnetic field, and it is larger for rock samples which are magnetically harder, having a larger value of magnetic coercive force and a slower rate of decrease by AC-demagnetization.

It is qualitatively shown that all these experimental results can be consistently explained based on a simple model of configuration of magnetic domains of magnetic minerals under the effect of inverse magnetostriction.

1. Introduction

Until the present time, the study of piezo-remanent magnetization has largely been carried out on magnetites or titanomagnetites (e.g. *Domen* 1957; *Nagata and Kinoshita* 1965), although experimental examination of the effect of compression on magnetic susceptibility has been made using natural rocks (e.g. *Kapitsa*, 1955). The main purpose in the present study lies in examining the acquisition of piezo-remanent magnetization (PRM) owing to uniaxial compression in natural rocks.

The authors have in mind the possible changes of magnetization of the earth's upper crust caused by crustal deformations which often result in the occurrence of earthquakes. As the crushing strength of natural rocks ranges from several hundred bars to 1 kb (*Handin* 1966) and the breaking stress of the upper crust is estimated to be somewhere between 50 and 200 kg/cm² (e.g. *Tsuboi* 1933, 1956), the possible changes in magnetization of natural rocks in this stress range is of great interest. Hence the PRM acquisition behaviors of igneous rocks in this stress range in weak magnetic fields are of particular concern in the present work.

2. Experimental procedures

In this preliminary experiment series igneous rock samples in the shape of a circular cylinder have been used. Their dimensions are 10 mm in diameter and 15 mm in length. Two extreme types of basalt have been examined particularly in detail: one (A) having a large coercive force ($H_c=245$ Oe) and the other (B) a small coercive force ($H_c=71$ Oe).

The magnetic hysteresis curves of these two samples are illustrated in Fig. 1. Sample A is an olivine basalt sampled near Boulder, Colorado and Sample B is a fine grained basalt from Somerest County, New York.

Before every experimental procedure, the sample which is to be examined is AC-demagnetized up to the maximum peak intensity of 500 Oe and then the residual magnetization is measured to confirm that the remanent magnetization is sufficiently small and will thereby not appreciably effect the following experimental procedure which begins with the approximate virgin state of the sample in so far as the magnetization is concerned.

The rock sample is given a uniaxial compression by means of a hydraulic press with a hand pump. This system affords a maximum pressure of about 300 kg/cm². The anvils of the press are made of non-magnetic stainless steel whose magnetic susceptibility is similar to that of a para-magnetic metal or of Cu-Be alloy in later experiments. A magnetic field is given on samples in a direction either parallel or perpendicular to the axis of compression by means of a small Helmholtz coil or a solenoid.

The intensity of magnetization of the sample is measured by means of a spinner magnetometer. The geomagnetic field in the space of a rotating system of the AC-demagnetizer, that between the anvils of the press and that of a rotating system of the spinner magnetimeter are approximately nullified respectively by Helmholtz coil systems, the remaining field being less than 1000 γ . Although every possible step was taken to assure the greatest possible precision, the results of measurements are still subject to errors of about one percent.

It has been shown by Nagata and Kinoshita (1965) that PRM is essentially an irreversible phenomenon which is extremely sensitive to the ordering of the application and release of uniaxial compression (P) and of a magnetic field (H). In the present series of experiments two different sequences of procedure have been specifically selected. They are (a) the application and release of a uniaxial compression (P) in the presence of a magnetic field (H) where the remanent magnetization after removal of the magnetic field is denoted by

Fig. 1 Magnetic hysteresis curves of Sample A (olivine basalt) and sample B (fine-grained basalt).

$J_r(H_+P_+P_0H_0)$; (b) the application and release of a magnetic field (H) in the presence of a uniaxial compression (P) where the remanent magnetization after removal of the compression is denoted by $J_r(P_+H_+H_0P_0)$. When the direction of H is parallel or perpendicular to the axis of compression, the acquired J_r are denoted by $J_r^{\parallel}$ or $J_r^{\perp}$ respectively. In the present terminology the ordinary IRM without any effect of external mechanical stress is denoted by $J_r(H_+H_0)$.

3. Results of experiments

(i) Typical examples of the experimental results obtained in a magnetic field of $H=36$ Oe are shown in Fig. 2, where the direction of magnetic field is parallel to the axis of compression. The remarkable characteristics of PRM shown in the figure may be summarized as follows:

(a) $J_r^{\parallel}(H_+P_+P_0H_0)$ always increases with P . It may be remarkable that in basalts the rate of increase in $J_r^{\parallel}(H_+P_+P_0H_0)$ with an increase of P is very large in comparison with that of magnetic minerals. (See Table I)

(b) In the case of Sample A, in which $H_c=245$ Oe, $J_r^{\parallel}(H_+P_+P_0H_0)$ increases sharply with P , but $J_r^{\parallel}(P_+H_+H_0P_0)$ remains almost invariant in the range of P between 0–200 kg/cm².

(c) In the case of Sample B, in which $H_c=71$ Oe, the rate of increase of $J_r^{\parallel}(H_+P_+P_0H_0)$ with P is much less than that of Sample A. However, $J_r^{\parallel}(P_+H_+H_0P_0)$ in this case

Fig. 2 Acquisition of piezoremanent magnetization, $J_r(H_+P_+P_0H_0)$, and reduction of IRM by uniaxial compression, $J_r(P_+H_+H_0P_0)$, of Samples A and B.

Table I The intensity of PRM, $J_r (H_+P_+P_0H_0)$, and its rate of acquisition $R (H, P)$, dependent on the magnetic field (H) and the uniaxial compression (P)

(I) Basaltic rocks (Sample A*)

$P(\text{kg/cm}^2)$	$H=18\text{ }Oe$		$H=36\text{ }Oe$		$H=45\text{ }Oe$		$H=54\text{ }Oe$	
	$J_r (H_+P_+P_0H_0)$	$R (H, P)$	$J_r (H_+P_+P_0H_0)$	$R (H, P)$	$J_r (H_+P_+P_0H_0)$	$R (H, P)$	$J_r (H_+P_+P_0H_0)$	$R (H, P)$
0	3.9×10^{-4}	0	7.3×10^{-4}	0	12.0×10^{-4}	0	17.1×10^{-4}	0
69	6.6 "	0.66	9.1 "	0.24	13.8 "	0.15	19.5 "	0.14
130	9.1 "	1.71	11.8 "	0.59	15.6 "	0.30	20.6 "	0.20

(II) Coagulated titanomagnetite grains (Mesh 230-)

$P (\text{kg/cm}^2)$	$H=10\text{ }Oe$		$H=30\text{ }Oe$		$H=50\text{ }Oe$	
	$J_r (H_+P_+P_0H_0)$	$R (H, P)$	$J_r (H_+P_+P_0H_0)$	$R (H, P)$	$J_r (H_+P_+P_0H_0)$	$R (H, P)$
0	2.1×10^{-2}	0	7.4×10^{-2}	0	13.4×10^{-2}	0
250	3.7 "	0.76	10.9 "	0.47	18.6 "	0.39
500	6.3 "	2.00	17.4 "	1.35	26.4 "	0.97
1000	8.0 "	2.80	22.9 "	2.09	—	—
2000	11.2 "	4.33	26.9 "	2.64	41.0 "	2.06
4000	12.5 "	4.95	29.4 "	2.98	45.1 "	2.36

definitely decreases with P .

(d) The intensity of saturation magnetization (J_s) of Sample A is about a half of that of Sample B, whereas H_c of Sample A is about 3.5 times as large as that of Sample B. As shown in Fig. 2, $J_r(H+H_0)$ for Sample A is much smaller than that of Sample B, the former being about 1/40 of the latter. It may be concluded therefore that a magnetically harder rock can acquire a larger rate of PRM. Another sample studied (Sample C) whose intensity of $J_r(H+H_0)$ and magnetic hardness are between those of Samples A and B shows that its rate of increase in $J_r(H+P_0H_0)$ with P is between those of (A) and (B) and that its rate of decrease in $J_r(P+H_0P_0)$ is also between those of (A) and (B), as illustrated in Fig. 3.

(e) As shown in Fig. 4, the rate of acquisition of PRM increases with a decrease in the applied magnetic field intensity. This is one of the remarkable general characteristics of PRM. This behavior of acquisition of PRM holds also for a coagulated assemblage of magnetic minerals. Table I shows the intensity of $J_r(H+P_0H_0)$ in units of emu/cc and the acquisition rate of PRM, $R(H, P)$ which is defined as

Fig. 3 Acquisition of piezoremanent magnetization, $J_r(H+P_0H_0)$, and reduction of IRM by uniaxial compression, $J_r(P+H_0P_0)$, of Sample C (basalt).

Fig. 4 Acquisition rate of PRM as dependent on an applied magnetic field.

Fig. 5 AC-demagnetization of Samples A, B and C.

$$R(H, P) = \frac{J_r(H+P+P_0H_0) - J_r(H+H_0)}{J_r(H+H_0)}, \quad (1)$$

in various intensities of the magnetic field for a basaltic rock sample A* (the same rock as Sample A) and a coagulated assemblage of titaniferous magnetites.

(ii) To clarify the dependence of the acquisition ratio of PRM on the magnetic hardness of rock samples, the AC-demagnetization curves of the saturated IRM of those samples are shown in Fig. 5. Obviously, Sample A is the hardest, Sample B the softest and Sample C is in between in the range of 0–150 Oe of the AC-demagnetization field ($\tilde{H}$). In other words the spectrum of microscopic coercive force (hc) is broad in Sample A, whereas in Sample B it is sharp and its maximum peak takes place at a little less than 20 Oe. The AC-demagnetization coercive force of the saturated IRM, which is defined as the intensity of H_c for $J_r(H_c)/J_r(0)=1/2$, is 98, 22 and 58 Oe respectively in Samples A, B and C. As shown in Fig. 6, Sample A can hardly acquire IRM in such a comparatively weak field as several tens of oersteds, $J_r(H+H_0)/J_r(\infty)$ for $H=40$ Oe being only 1.5×10^{-3} , for example. On the other hand, Sample B can easily acquire IRM, $J_r(H+H_0)/J_r(\infty)$ for $H=40$ Oe amounting to about 0.1. In this respect also, Sample C is between Samples A and B.

It is concluded from the above-mentioned results that rocks having the larger magnetic coercive force and slower rate of decay of IRM by AC-demagnetization can acquire PRM at a faster rate, the rate of acquisition being faster in a weaker magnetic field.

(iv) The behavior of acquisition of PRM in igneous rocks in a weak magnetic field near the intensity of the geomagnetic field will be particularly interesting for the uniaxial compression which is parallel to the direction of the magnetic field. As shown in Fig. 7, the intensity of $J_r''(H+P+P_0H_0)$ under a constant strength of compression is approximately proportional to the field intensity in the range of a magnetic field up to about 2 Oe. The lowest curve in Fig. 7, represents the intensity of IRM of the same sample, namely, $J_r(H+0H_0)=J_r(H+H_0)$, whose dependence on H can be expressed by Rayleigh's formula,

$$J_r(H+H_0)=bH^2, \quad (2)$$

provided that H is small. As $J_r''(H+P+P_0H_0)$ in the range of P examined in the present work (i.e. $P=37\text{--}118$ kg/cm²) is approximately proportional to H , $J_r(H+P+P_0H_0)$ may be

Fig. 6 Relative intensity of IRM acquired in various intensities of applied magnetic field.

Fig. 7 Dependence of the intensity of $J_r(H+P_0H_0)$ on the applied magnetic field intensity, H , for different values of P . Sample B.

Table II Comparison of the rate of increase in $J_r''(H+P_0H_0)$ with H defined as $C(P)J_r(H+P_0H_0)$ with H defined as $C(P)J_r''(H+P_0H_0)/H$ with the initial magnetic susceptibility (χ_0), the coefficient in the Reyleigh's loop for IRM, $b \equiv J_r''(H+H_0)/H^2$, and the rate of increase in IRM with H ($\Delta J_r(H+H_0)/\Delta H$) for the range of remanent magnetization intensity of PRM.

Sample	$C(P)$	χ_0	b	$\Delta J_r/\Delta H$
A continental basalt	$0.20 \times 10^{-4} \text{ emu/cc/Oe}$	$1.9 \times 10^{-3} \text{ emu/cc}$	$0.6 \times 10^{-6} \text{ emu/cc/Oe}^2$	$1.3 \times 10^{-5} \text{ emu/cc/Oe}$
B continental basalt	13.0×10^{-4}	5.3×10^{-3}	140×10^{-6}	71×10^{-5}
D oceanic basalt	0.35×10^{-4}	0.3×10^{-3}	1.7×10^{-6}	2.1×10^{-5}
E Diorite	2.1×10^{-4}	5.9×10^{-3}	6.0×10^{-6}	8.7×10^{-5}

(Remarks: $P = 118 \text{ kg/cm}^2$ for $C(P)$)

empirically expressed as

$$J_r(H+P_0H_0) \approx C(P)H. \quad (3)$$

Then the rate of acquisition of PRM, $R(H, P)$, in the ranges of P and H concerned are expressed as

$$R(H, P) = \frac{C(P)}{bH} - 1. \quad (4)$$

Eq. (4) suggests that the rate of acquisition of PRM increases with a decrease in the magnetic field intensity as observed in Fig. 4. It will be obvious in Fig. 7 (also in Fig. 9 which is discussed later) that the effect of uniaxial compression and its release on an acquisition

Fig. 8 Dependence of the intensity of $J_r(H+P_0H_0)$ on the applied magnetic field, H , at $P=118 \text{ kg/cm}^2$.

Fig. 9 Dependence of the intensity of $J_r(H+P_0H_0)$ on the applied magnetic field intensity. Sample B.

of remanent magnetization becomes relatively much more conspicuous in a smaller magnetic field.

It was experimentally difficult to measure precisely how the linear relationship between $J_r'(H+P_0H_0)$ and H approaches the parabolic relation between $J_r(H+H_0)$ and H when P approaches zero. However, the plots of $J_r(H+P_0H_0)$ versus H in Fig. 7 show a tendency that these values corresponding to smaller values of H are gradually deviated from a linear relation to a parabolic one in accordance with a decrease in P .

(v) The dependence of $J_r'(H+P_0H_0)$ on H in case of $P=118 \text{ kg/cm}^2$ is shown in Fig. 8 for four different kinds of typical igneous rocks, where Sample D is an oceanic basalt collected from an Atlantic Ocean bed and Sample E a diorite collected in Oregon. For Samples B and E, the dependence of $J_r^\perp(H+P_0H_0)$ on H also is shown in the same figure.

(a) It may be concluded in Fig. 8 that $J_r'(H+P_0H_0)$ and $J_r^\perp(H+P_0H_0)$ are in general, approximately proportional to H . The coefficients for the proportionality, $C(P)$, for these samples are given in Table II together with their initial magnetic susceptibility, χ_0 , and the coefficient (b) in the Rayleigh's relation. The magnitude of $C(P=\text{const})$ may depend on various factors, not only on χ_0 but also on the coercive force (H_c) and other parameters, as suggested in Figures 1, 5 and 6.

$C(P)$ is proportional neither to χ_0 nor to b in Table II, though the largest and smallest values of $C(P)$ correspond respectively to the largest and smallest values of b .

As shown in Fig. 9, however, the intensity of IRM for the intensity range of $J_r(H+P_0H_0)$ concerned in the present work no longer exhibits the Rayleigh's parabolic relation with H , but the rate of its increase is roughly proportional to the rate of increase in H . Then, one may take the rate of increase in $J_r(H+H_0)$ with an increase in H in this

Fig. 10 Mechanical demagnetization curves for IRM. Dependence of $J_r(H+H_0, P+P_0)$ upon P . Sample B.

linear range as a measure of capability of additional acquisition of remanent magnetization caused by the procedures of compression and its release in the presence of a magnetic field.

The rate, $J_r(H+H_0)/\Delta H$, thus defined is given for each sample in Table II. Roughly speaking, $C(P)$ is proportional to $J_r(H+H_0)/\Delta H$. This result may suggest that the process to produce $J_r(H+P+P_0H_0)$, namely an application and a release of compression on a rock sample in the presence of a magnetic field, is approximately equivalent to enlarging the magnetic field.

(b) Fig. 8 shows that $J_r^\perp(H+P+P_0H_0)$ is not much different from $J_r^\parallel(H+P+P_0H_0)$, though the former is usually a little smaller than the latter. Although the demagnetizing field is ignored in the present work, the field is always smaller than 1% of the external field. Hence, the discrepancy between $J_r^\parallel$ and $J_r^\perp$ in Fig. 8 is appreciably larger than the possible effect of the demagnetizing field. It may be safely argued here, however, that $J_r^\perp(H+P+P_0H_0)$ has nearly the same magnitude as $J_r^\parallel(H+P+P_0H_0)$.

(c) As shown in Fig. 9, the rate of increase of $J_r(H+P+P_0H_0)$ with H , i.e. $\Delta J_r(H+P+P_0H_0)/\Delta H$, decreases gradually with an increase in H , although the rate is approximately constant, being represented by $C(P)$, in the range of weak magnetic field. This tendency of $J_r(H+P+P_0H_0) \sim H$ curve seems to be consistent with the idea that the process to produce $J_r(H+P+P_0H_0)$ is approximately equivalent to enlarging the magnetic field, because the $J_r(H+H_0) \sim H$ curve also has the same tendency insofar as the magnitude of $J_r(H+H_0)$ is similar to that of $J_r(H+P+P_0H_0)$.

(iv) Mechanical demagnetization

Fig. 10 illustrates an experimental demonstration that IRM of natural rocks can be demagnetized by a uniaxial compression in a non-magnetic space. IRM acquired in a magnetic field H is denoted by $J_r(H+H_0)$, whereas the residual remanent magnetization after

applying a compression (P) and effecting its release on a sample having $J_r(H_+H_0)$ in a non-magnetic space will be denoted by $J_r(H_+H_0, P_+P_0)$. In Fig. 10, $J_r(H_+H_0, P_+P_0)$ decreases with an increase in P regardless of whether $P \parallel H$ or $P \perp H$. This experimental result may suggest that the magnetic orientation of domains is rearranged by the uniaxial compression so as to be equally distributed in azimuth within a plane perpendicular to the axis of compression. The rate of decay of $J_r(H_+H_0, P_+P_0)$ with P depends on the initial value of $J_r(H_+H_0)$ for the same sample; the rate being smaller for $J_r(H_+H_0)$ acquired by a larger field H .

4. Some discussions on the experimental results.

The effect of a uniaxial compression upon the configuration of magnetic domains in magnetites has recently been experimentally demonstrated (e.g. *Soffel*, 1967). According to the result, the total volume of domains with direction of spontaneous magnetization perpendicular to the axis of compression increases with an increase of compression P , just as predicted from the theory of magnetostriction of magnetite. This change of domain configuration should result in a reduction of magnetization along the direction of the compression axis and consequently a reduction of magnetic susceptibility along that direction.

On the other hand, a release of the compression from a sample in the presence of a magnetic field along the compression axis should result in a preference of orientation of spontaneous magnetization of domains towards the direction of the magnetic field. Under the effect of the uniaxial compression a space in the magnetic minerals which was initially occupied by those domains whose spontaneous magnetization was pointed in a direction opposite to the field is refilled by those domains whose spontaneous magnetization is perpendicular to the field; the release of compression in the field results in a refilling of the space by the domains which are oriented towards the field; thus the application and release of a uniaxial compression in the presence of a magnetic field parallel to the compression axis results in a 180° rotation of spontaneous magnetization of some domains.

When the direction of uniaxial compression is perpendicular to that of the applied magnetic field, the mechanism of acquisition of $I_{r\perp}(H_+P_+P_0H_0)$ may be somewhat different from the above-mentioned one. As experimentally demonstrated in Fig. 10, a uniaxial compression of a rock sample having a remanent magnetization in a non-magnetic space results in a rearrangement of configurations of magnetic domains so as to approach the non-magnetized state of homogeneous distribution of the domain orientations. These experimental results may suggest that the force of inverse magnetostriction impressed by a uniaxial compression overcomes, to a certain extent, the force which blocks the domain walls so that the domain configuration approaches the state of minimum energy in the presence of a demagnetization field which corresponds to a reduced intensity of remanent magnetization in the case of a non-magnetic space. If a uniaxial compression is applied on a non-magnetized sample in the presence of a magnetic field which is perpendicular to the compression axis, the intensity of magnetization along the field should be larger than in the case of non-compression, because the domains whose spontaneous magnetization is originally near the compression axis ought to additionally contribute to the magnetizations. Con-

sequently, the remanent magnetization after the release of the compression and then of the field becomes larger than the ordinary IRM.

The magnetostatic energy of magnetite in a magnetic field H is of the order of $J_s H \simeq 5 \times 10^2 H \text{ cgs}$, whereas its magnetoelastic energy under stress P is of the order of $\frac{3}{2} \lambda_s P \simeq 60 P$ where P is expressed in units of kg/cm^2 . Hence, the energy owing to the mechanical stress of $P = 1 \text{ kg/cm}^2$ may be roughly equivalent to that caused by a magnetic field of $H = 0.12 \text{ Oe}$ with respect to the energetics of magnetization processes of dominant magnetic minerals in rocks.

It may then be obvious that, if there exists any effect of mechanical stress upon magnetization of rocks in a certain way, the effect caused by a constant value of stress will become relatively larger in a weaker magnetic field.

Actually, experimentally measured values of $J_r(H + P + P_0 H_0)$ of basalts are markedly larger than their ordinary $J_r(H + H_0)$ in a magnetic field less than 1 Oe, as illustrated in figures 7 and 9. Since the mechanical stress within the earth's crust is estimated to be $10 - 10^2 \text{ kg/cm}^2$ and the geomagnetic field intensity is between 0.3 and 0.7 Oe, one can expect that possible effects of mechanical stress on the acquisition of remanent magnetization of the earth's crust are markedly large.

It may be worthwhile to note that the effect of uniaxial compression on the magnetic susceptibility of rocks is appreciably reduced from the case of the effect on rock-forming magnetic minerals themselves. As well known, the dependence of magnetic susceptibility χ on a uniaxial compression P along the applied magnetic field is expressed as

$$\chi(P) = \frac{\chi_0}{1 + \beta P}$$

where χ_0 denotes the magnetic susceptibility at $P = 0$. The order of magnitude of β is $1 \times 10^{-4} \text{ kg/cm}^2$ for rocks, whereas it is $1 \times 10^{-3} \text{ kg/cm}^2$ for magnetite. The reduction of β in the case of rocks is interpreted as due to the effect of demagnetizing field of magnetic minerals scattered in rocks (Nagata 1966a).

On the contrary, the acquisition of piezo-remanent magnetization is due to a reduction of the microscopic coercive force of some domains caused by a uniaxial compression, resulting in a preference of magnetic orientation towards the field direction as compared with the opposite direction, where the magnetic field works merely as a choice-maker for the domain's orientation either to or opposite to the field direction. The effect of mechanical stress upon the change of microscopic coercive force is not essentially different between naked magnetic minerals and those contained in rocks. The main criterion for acquiring the additional remanent magnetization may be whether or not the effective magnetic field is beyond the reduced microscopic coercive force. As the effective magnetic field is smaller in the case of magnetic minerals in rocks than in the case of naked minerals insofar as the bulk demagnetizing factor of a sample only is taken into consideration, the relative effect of mechanical stress upon the acquisition rate of PRM becomes larger in rocks than in magnetic minerals themselves.

5. Geophysical discussions of experimental results

It has been frequently reported that the geomagnetic field near the epicenter of a shallow earthquake has changed appreciably before and after the occurrence of the quake. Among a number of papers dealing with this problem *Kato-Utashiro's* report (1948) on the local geomagnetic change accompanying the great Nankaido earthquake in 1946 seems to give the most reliable data. According to their result, the westward geomagnetic declination at Katsuura abruptly decreased by about $4'$ on the day (December 21, 1946) when the great earthquake took place off shore of the observing site. The earthquake was accompanied by an abrupt upheaval of the Katsuura seashore by about 60 cm. The shore had continued to subside gradually during more than 30 years before the quake. It seems likely that the westward declination had been slowly increasing, at least during three months, before the quake, the total amount of the local change being about $+1'$ in angle. This result indicates that the sense of the abrupt geomagnetic change accompanied by a great earthquake with an abrupt upheaval of the land is opposite to that of the pre-quake slow change accompanied by a slow subsidence of the land, and also that the magnitude of the abrupt change is much beyond the total amount of the pre-quake slow change. It has been interpreted that the pre-quake stage accompanied by the subsidence of the land represents an accumulation of mechanical stress within the earth's crust whereas the abrupt upheaval of the same land corresponds to a sudden release of the stress.

If we assume a simple model of a rock piece being gradually compressed by an increasing compression and then being suddenly broken at a certain value of compression in the presence of a magnetic field nearly parallel to the compression axis, the result of the present experiment may predict that the intensity of magnetization of the rock is gradually decreasing before the breaking down and it abruptly increases by acquiring PRM when the compression is suddenly released by the breaking. It seems very likely that the changes in a magnetic field caused by the above-mentioned changes in the rock's magnetization correspond well to the local geomagnetic changes observed at Katsuura.

Based on this simple model, one may further be able to estimate possible changes of magnetization of the earth's crust before and accompanied by the breaking-down. Assume that the breaking stress of the earth's crust is $P=118 \text{ kg/cm}^2$ and the total geomagnetic field intensity is $F=0.5 \text{ Oe}$, and that the average value of β for rocks is $1.0 \times 10^{-4} \text{ kg/cm}^2$. Then, the total amount of pre-breaking change in magnetization may be given by $\frac{\beta P}{1 + \beta P} \chi_0 F$, whereas the co-breaking abrupt change in magnetization is given by $C(P)F$. Table III gives the numerical values of these two kinds of magnetization for four rock samples, A, B, D and E. As shown in the table, the co-breaking abrupt change is much larger than the pre-breaking change in three rock samples, and the former is nearly the same as the latter in one case. Although there is a wide scattering of the ratio of the co-breaking change to the pre-breaking one in these examples, one may say that the co-breaking change is generally larger than the pre-breaking change. It seems likely that this result can reasonably interpret the pre-quake and co-quake variations in the geomagnetic field observed at Katsuura.

Table III Pre-breaking change in magnetization, $\frac{\beta p}{1+\beta p} \chi_0 F$, and co-breaking abrupt change in magnetization, $C(p) F$, of igneous rocks in the presence of a magnetic field $F=0.5$ Oe (The breaking stress P is assumed to be 118 kg/cm²)

Sample	Pre-breaking change	Co-breaking change
A	1.1×10^{-5} gauss	1.0×10^{-5} gauss
B	3.1×10^{-5}	65.0×10^{-5}
D	0.18×10^{-5}	1.8×10^{-5}
E	3.6×10^{-5}	10.5×10^{-5}

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