MAGNETIC PROPERTIES OF YAMATO-73-04 AND YAMATO-73-07 METEORITES

Takesi NAGATA,

National Institute of Polar Research, Kaga 1-chome, Itabashi-ku, Tokyo 173

Naoji SUGIURA

Geophysical Institute, University of Tokyo, Yayoi 2-chome, Bunkyo-ku, Tokyo 113

and

F. C. SCHWERER

U. S. Steel Corporation Research Laboratory, Monroeville, Pennsylvania, U.S.A.

Abstract: Two Yamato meteorites collected in 1973 in Antarcica, Yamato-73-04 and Yamato-73-07, may be identified to an olivine-bronzite chondrite and a hypersthene achondrite respectively from their petrographic and chemical compositions.

Magnetic analyses indicate that native irons in Yamato-73-04 chondrite comprise 2.6 wt% of 7 wt% Ni kamacite and 4.3 wt% of 13 wt% Ni kamacite, and that Yamato-73-07 achondrite contains 0.24 wt% of Fe^0 with small amounts of Ni and Co and less than 0.02 wt% of 60 wt% Ni taenite or $Fe_{3-x}Cr_xO_4$.

The natural remanent magnetization, NRM, of Yamato-73-07 achondrite can be distinctly separated into the original NRM of the undisturbed interior of 6.3×10^{-6} emu/gm in intensity and the secondary NRM of the surface skin fusion crust of about 0.5 mm in thickness. The secondary NRM is attributable to the thermoremanent magnetization acquired in a magnetic field of 0.44 Oe—probably the geomagnetic field.

1. Introduction

The glaciology research party of the 14th Japanese Antarctic Research Expedition (1972–1974) collected 12 pieces of meteorite in a bare ice area of the "Meteorite Ice Field" near the Yamato Mountains and its vicinity in Antarctica in December 1973 (SHIRAISHI *et al.*, 1976). Within a very limited area of about 5 km × 10 km centered at 71°50′S and 36°20′E in the "Meteorite Ice Field", which is located on the southeastern side of the Yamato Mountains, 9 pieces of meteorite were found and collected in 1969 (YOSHIDA *et al.*, 1969; KUSUNOKI, 1975). In December 1973, 8 pieces of meteorite were found in the same area as in 1969, and four more meteorites were found and collected from bare ice surfaces on the western and southwestern sides of the Yamato Mountains. Among 12 meteorite pieces collected in 1973, four are of relatively large masses, weighing 650 gm (Yamato-73-01), 900 gm (Yamato-73-04), 480 gm (Yamato-73-07) and 500 gm (Yamato-73-12) respectively, whereas the other eight pieces are much smaller fragments, weighing less than 40 gm. As the preliminary test, the large four meteorite pieces were subjected to analysis of γ-ray emissions of cosmogenic radioactive nuclides. All of the four samples showed the characteristic Al^{26} γ-ray emission of a considerable intensity. This result may be considered as evidence for the extraterrestrial origin of these samples. All 12 samples collected in 1973 are covered with a thin fusion crust of brown colour, which is another characteristic of meteorite. Chemical and petrographic characteristics of two samples, Yamato-73-04 and Yamato-73-07, have been studied to a certain extent to identify them to an olivine-bronzite chondrite and an achondrite respectively.

In the present short note, magnetic properties and characteristics of the natural remanent magnetization of these two meteorite samples will be systematically described with some discussions. The aim of studies on the magnetic properties of meteorites may generally comprise several different aspects.

(a) First, the magnetic analysis is useful in quantitatively detecting metallic components such as kamacite, taenite, plessite, schribersite, etc. in lunar rocks and meteorites (*e.g.* NAGATA *et al.*, 1972, 1974, 1975). It is because the magnetic analysis can indicate the total amounts of compositionally different phases of the ferromagnetic components separately and regardless of their grain size.

(b) Abundance of superparamagnetically fine particles of metals evaluated either by the magnetic viscosity or by the ferromagnetic resonance technique is a good measure-scale for the impact metamorphism of lunar rocks and meteorites (*e.g.* NAGATA and CARLETON, 1970; HOUSLEY *et al.*, 1975; GOSE *et al.*, 1976).

(c) The paramagnetic susceptibility can well represent the total abundance of FeO in silicate minerals in meteorites.

(d) The natural remanent magnetization of meteorites may indicate the paleomagnetic field of the parent meteorite planets if its magnetic characteristics are sufficiently well established. It must be noted, however, that only very few meteorites have been magnetically examined to date, the total number of magnetically examined samples of meteorites being much less than that of lunar samples. It is hoped therefore that this short note can add at least two new data to the "meteoritic magnetism".

2. Chemical and Petrographic Compositions of Samples

Results of chemical and petrographic analyses of Yamato-73-04 and Yamato-73-07 meteorites will be separately described in detail by the respective investigators. However, several major points of chemical and petrographic properties of these two meteorites are briefly summarized here to be compared with their magnetic properties.

(i) *Yamato-73-04* meteorite is covered with a thin fusion crust of brownish gray colour and the outer-surface has a number of "finger-prints". A number of chondrule of various shapes are observed in a thin section of this sample. The chondrules consist mostly of olivine and bronzite (orthopyroxene) with small amounts of clinopyroxene and plagioclase. The matrix comprises olivine as the most abundant mineral and bronzite as the second dominant mineral. Clinopyroxene and plagioclase also are present, but their abundance is much smaller than that of olivine and bronzite. Metal grains of FeNi are present; these FeNi grains are not appreciably weathered. From these petrographic data, this meteorite can be provisionally identified to an *olivine-bronzite chondrite.*

(ii) *Yamato-73-07* meteorite also is covered with a thin fusion crust of dark gray colour. No chondrule is observable under a microscope. The most dominant minerals in this meteorite are clinopyroxene and orthopyroxene is the next in abundance, only a small amount of olivine being observable. Plagioclase of comparatively large grain size also is present in a little larger amount than olivine. The

Table 1. Chemical composition of Yamato-73-04 and Yamato-73-07 meteorites.

	Yamato-73-04	Yamato-73-07
(Silicate phase)		
SiO_2	39.1 wt%	55.2 wt%
MgO	23.4	26.4
FeO	12.3	12.5
Al_2O_3	2.0	0.6
CaO	1.5	1.2
Na_2O	0.8	trace
K_2O	0.1	trace
Cr_2O_3	0.5	1.3
MnO	0.3	0.5
TiO_2	trace	0.1
P_2O_5	0.2	trace
(Metal phase)		
Fe	12.5	0.5
Ni	1.6	trace
Co	0.1	trace
(Sulfide phase)		
Fe	3.5	0.9
S	2.0	0.5
Total	100.0	99.6

(*After M. SHIMA*)

content of FeNi metal grains is very small, but they are still observable under a microscope. From these petrographic data, this meteorite can be provisionally indentified to *a Ca-poor achondrite.*

The bulk chemical compositions of the two meteorites are summarized in Table 1. The bulk chemical composition of Yamato-73-04 meteorite is in between those of *H* type and *L* type meteorite groups (UREY and CRAIG, 1953). A plot of weight percent of iron in metal and FeS against that of oxidized iron (FeO) in this sample also is in between the plot group of olivine-bronzite chondrites and that of olivine-hypersthene chondrites in the MASON's diagram (1962). From the chemical data, this meteorite may be identified to either olivine-bronzite chondrite or olivine hypersthene chondrites.

The chemical composition of Yamato-73-07 meteorite seems to be typical as a Ca-poor achondrite and is almost the same as that of Yamato (b) meteorite which is a hypersthene achondrite (SHIMA and SHIMA, 1975; TAKEDA *et al.*, 1975). If the classification system to divide the Ca-poor achondrites into four categories, namely, enstatite-, hypersthene-, olivine- and olivine-pigeonite achondrites (MASON, 1962) is adopted, the chemical composition of this meteorite is closest to that of the hypersthene achondrites (YAVNEL and DYAKONOVA, 1958). Attention may have to be paid to a comparatively rich abundance of Cr_2O_3 (1.3 wt%) in this sample. Since the chemical composition of this achondrite is almost exactly the same as Yamato (b) achondrite which contains chromites (TAKEDA *et al.*, 1975), Yamato-73-07 achondrite also may contain chromites and/or their solid solutions with Fe_3O_4 (NAGATA *et al.*, 1975).

In connection with the magnetic properties of the meteorites, the metallic phase of Fe-Ni-Co system plays the most significant role for their ferromagnetic characteristics. If the metal phase in Yamato-73-04 meteorite is assumed to consist of only a single phase, it can be presumed from the chemical data to be 11 wt% Ni kamacite. The composition of metal phase in Yamato-73-07 meteorite cannot be predicted from the chemical data alone, because its abundance is so small that Ni abundance is not available.

The iron sulfide can be ferromagnetic, if its composition is within the range of pyrrhotite (FeS_{1+x} with $0.1<x<1/7$). However, the chemical compositions of iron sulfides in both meteorites are well represented by FeS (troilite). As troilite is antiferromagnetic, it may be practically impossible to quantitatively detect it by the magnetic method when the ferromagnetic kamacites coexist with it.

Finally, iron oxides (FeO) contribute simply to the paramagnetism provided they are constituents of silicate minerals such as olivine, orthopyroxene, clinopyroxene, etc. If, however, some parts of them were members of spinels such as for example, Fe_3O_4, they should contribute to the ferromagnetism (*e.g.* NAGATA, 1961) with a characteristic Curie temperature.

These key points will be magnetically examined in the succeeding section.

3. Basic Magnetic Properties

3.1. Magnetization curve at room temperature

The magnetization curves of the two Yamato meteorites measured at room temperature by a vibration magnetometer are illustrated in Figs. 1 and 2. From

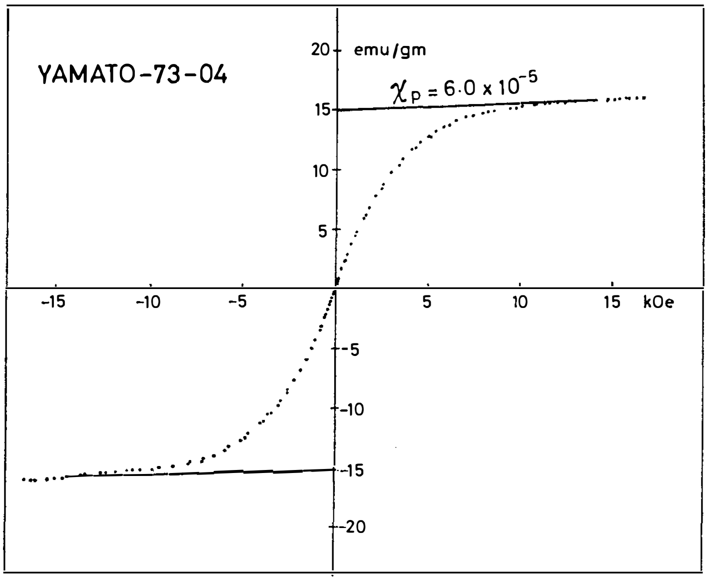

Fig. 1. Magnetic hysteresis curves of Yamato-73-04 meteorite.

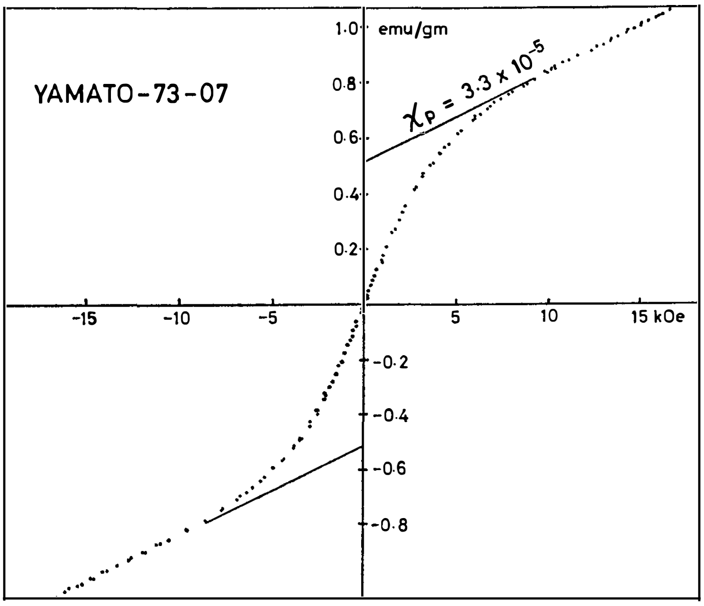

Fig. 2. Magnetic hysteresis curves of Yamato-73-07 meteorite.

Table 2. Magnetic properties.

Parameter	Yamato-73-04	Yamato-73-07	Unit
Initial susceptibility (χ_0)	4.1×10^{-3}	0.088×10^{-3}	emu/gm/Oe
Paramagnetic susceptibility (χ_p)	6.0×10^{-5}	3.3×10^{-5}	emu/gm/Oe
Saturation magnetization (I_s)	15.0	0.527	emu/gm
Saturation remanence (I_R)	0.045	0.0027	emu/gm
Coercive force (H_c)	8	13	Oersteds
Remanence coercive force (H_{Rc})	375	—	Oersteds
Curie point (Θ)	—	792 568	°C
Phase transition ($\Theta_{\alpha\gamma}$)	740 659	None	°C
Phase transition ($\Theta_{\gamma\alpha}$)	624 403	None	°C
Ni in kamacite	7.0 13.0		wt%

these data, the saturation magnetization (I_s), the saturation remanent magnetization (I_R), and the paramagnetic susceptibility (χ_p), all at room temperature, are determined. By use of a special experimental technique to enlarge the abscissa scale for the magnetic field intensity near the zero field intensity, the coercive force (H_c) and the remanence coercive force (H_{Rc}) also were determined by the same measuring system. The initial magnetic susceptibility (χ_0) was separately measured by a susceptibility bridge with 1 kHz in frequency and 9 Oe in peak field intensity. The observed results are summarized in Table 2. As expected from the chemical and petrographic data of these samples, ferromagnetic component in Yamato-73-07 meteorite is much less than that in Yamato-73-04 meteorite. However, the observed saturation magnetizations (I_s) of the two meteorites are considerably smaller (about a half) than their theoretically expected values based on the Fe^0 content given in Table 1.

3.2. *Thermomagnetic curve*

The thermomagnetic curves of the two samples are illustrated in Figs. 3 and 4 respectively. It is clearly seen in Fig. 3 that the ferromagnetic component in Yamato-73-04 chondrite consists of two phases, *i.e.* (a) a phase having magnetic transition temperatures at 740°C and 624°C in the heating and cooling curves respectively and (b) a phase of magnetic transition temperatures at 659°C and 403°C. The magnetic transition temperatures in the heating curve can be inter-

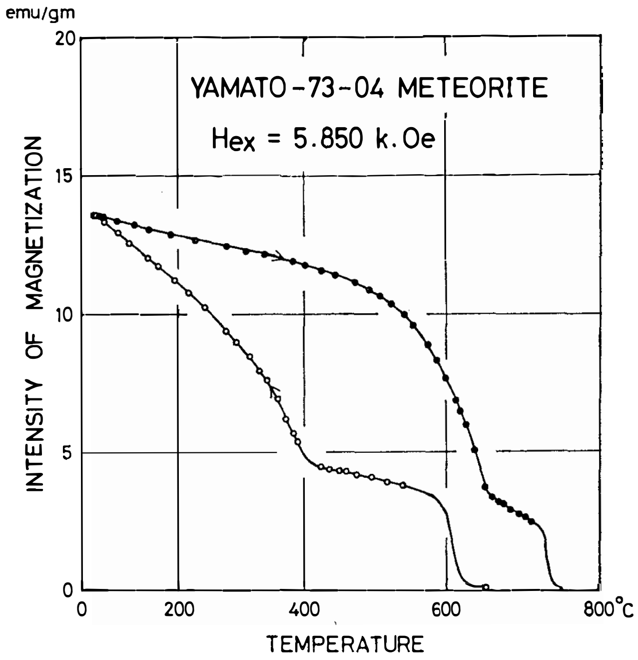

Fig. 3. Thermomagnetic curves of Yamato-73-04 meteorite (Remarks: Full lines in higher temperature range represent results of continuous measurements of magnetization with varying temperature).

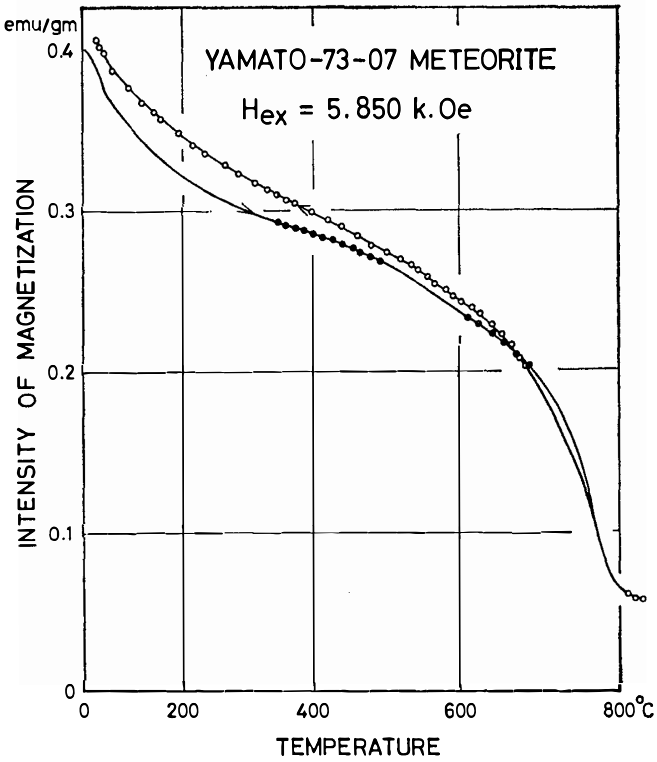

Fig. 4. Thermomagnetic curves of Yamato-73-07 meteorite (Remarks: Continuous full lines represent results of continuous measurements of magnetization with varying temperature).

preted to represent the $\alpha \rightarrow \gamma$ transition temperature of kamacite ($\Theta_{\alpha\rightarrow\gamma}$), while those in the cooling curve represent the $\gamma \rightarrow \alpha$ transition temperature ($\Theta_{\gamma\rightarrow\alpha}$). Then, Curie points of the two phases, which are higher than the $\alpha \rightarrow \gamma$ transition temperatures, cannot be observed in the thermomagnetic curve. Thus, the observed thermomagnetic curves in Fig. 3 can be decomposed into a thermally irreversible thermomagnetic hysteresis loop of 7% Ni kamacite ($\Theta_{\alpha\rightarrow\gamma}=740°C$, $\Theta_{\gamma\rightarrow\alpha}=624°C$) and that of 13% Ni kamacite ($\Theta_{\alpha\rightarrow\gamma}=659°C$, $\Theta_{\gamma\rightarrow\alpha}=403°C$). At room temperature, the magnetization intensities of 7% Ni- and 13% Ni-kamacite phases amount to 5.4 emu/gm and 8.9 emu/gm respectively. Since the spontaneous magnetization intensities at room temperature of 7% Ni- and 13% Ni-kamacites are about 212 emu/gm and 208 emu/gm respectively, the weight percentages of the two phases in this chondrite are given by m (7% Ni kamacite)=2.55 wt% and m (13% Ni kamacite) =4.28 wt%. Then, the average value of the Ni content in the kamacite phases is estimated as about 11 wt%. This magnetically estimated value of the Ni content in the metal is in approximate agreement with the chemically estimated value described in the preceding section. As already mentioned, however, the total content of kamacite phases magnetically evaluated is about a half that derived from the chemical data.

The thermomagnetic curve of Yamato-73-07 achondrite is almost thermally reversible, having a major Curie point at 792°C and another Curie point at 568°C for a minor ferromagnetic component. The observed value of magnetization at various temperatures in Fig. 4 is a sum of ferromagnetic magnetization, $I(T)$, and paramagnetic magnetization, $\chi_p(T)H$, where $\chi_p(T)$ is expressed by $\chi(T)=C/T$ with T in unit of degree Kelvin. The numerical value of C derived from Fig. 4 amounts to $C=9.4\times10^{-3}$ (emu/gm/Oe) degree K, which gives $\chi_p(T=300°K)=3.1\times10^{-5}$ emu/gm/Oe. The $\chi_p(T=300°K)$ value thus estimated is in approximate agreement with the paramagnetic susceptibility determined from the magnetization curve at room temperature (Fig. 2 and Table 2). The high value of Curie point 792°C of the main ferromagnetic phase suggests a metallic Fe alloy with small amounts of Ni and Co. The ferromagnetic magnetization at room temperature, $I(T=300°K)$ in Fig. 4 amounts only to 0.17 emu/gm. Since, $I(T=300°K)$ value at $H=5.85$ kOe is about 94% of the saturation magnetization (I_s), as shown in Fig. 2, the discrepancy between $I(T=300°K)$ value in Fig. 4 and I_s value in Fig. 2 is considerably large. The discrepancy could be attributed only to inhomogenuity of the metallic grain content in this achondrite, because the sample mass used for the magnetization curve measurement and the thermomagnetic curve one was only 104 mgm and 35 mgm respectively. Simliar inhomogenuity of metallic component abundance has been suggested in Yamato (b) achondrite also (NAGATA *et al.*, 1975). The second ferromagnetic component having Curie point at 568°C is about 0.02 emu/gm in intensity at room temperature.

A possible identification of this second ferromagnetic component will be a substituted magnetite, in which some parts of Fe^{3+} in Fe_3O_4 are substituted by

Cr^{3+}, because the abundance of Cr^{3+} is considerably rich in this sample. Another possibility of the identification will be a taenite of about 60 wt% Ni content or a kamacite of about 30 wt% Ni content. In any case, the content of the second ferromagnetic component is less than 0.02 wt%.

4. Characteristics of Natural Remanent Magnetization

Changes in the intensity and the direction of natural remanent magnetization (NRM) of the two Yamato meteorites in the course of the AF-demagnetization are shown in Figs. 5 and 6, together with the AF-demagnetization curve of the isothermal remanent magnetization (IRM) of almost the same initial intensity. The initial value of NRM intensity and the remanent magnetization intensity after the AF-demagnetization process up to 50 Oe rms and 100 Oe rms are listed in Table 3, together with a static magnetic field intensity (h) to make the sample acquire IRM of the same intensity as the initial value of NRM.

Table 3. Characteristics of natural remanent magnetization.

Parameters	Yamato-73-04	Yamato-73-07	Unit
NRM intensity (I_n)	10.6×10^{-5}	1.15×10^{-5}	emu/gm
I_n ($\tilde{H}=50$)	1.65×10^{-5}	0.63×10^{-5}	emu/gm
I_n ($\tilde{H}=100$)	1.36×10^{-5}	0.49×10^{-5}	emu/gm
h	2.9	10.5	Oersteds

(Remarks)

I_n ($\tilde{H}=50$): NRM intensity after AF-demagnetization up to $\tilde{H}=50$ Oe rms.

I_n ($\tilde{H}=100$): NRM intensity after AF-demagnetization up to $\tilde{H}=100$ Oe rms.

h: Static magnetic field to make sample acquire IRM of the same intensity as the initial value of NRM (I_n).

Although the initial intensity of NRM of Yamato-73-04 chondrite is much larger than that of Yamato-73-07 achondrite, the stability of NRM of the former against the AF-demagnetization is much worse than that of the latter. As the stability of the direction of NRM is particularly low in Yamato-73-04, no paleomagnetic significance could be put on NRM of this sample. On the contrary, the stability of NRM of Yamato-73-07 achondrite is so good in both intensity and direction for the remaining component after the initial AF-demagnetization up to 10 Oe rms, that NRM of this achondrite sample could be used for the paleomagnetic study for the parent meteorite planet.

However, a significant problem in the paleomagnetic study should be concerned with NRM of the fusion crust, because it is certain that the fusion crust was produced when this achondrite piece was reheated on entry into the terrestrial atmosphere with a high speed so that the secondary thermoremanent magnetization

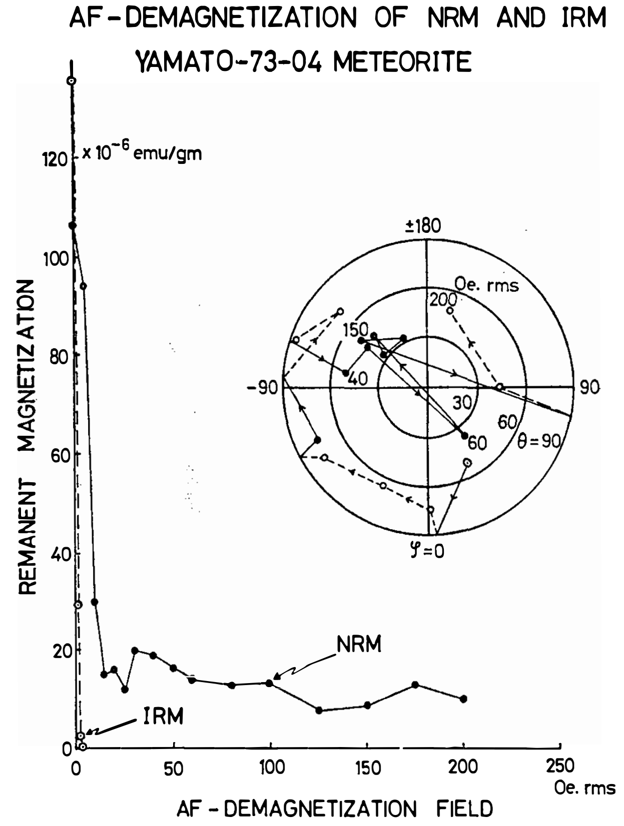

Fig. 5. *Changes in intensity and direction of NRM in the course of AF-demagnetization (Yamato-73-04 meteorite).*

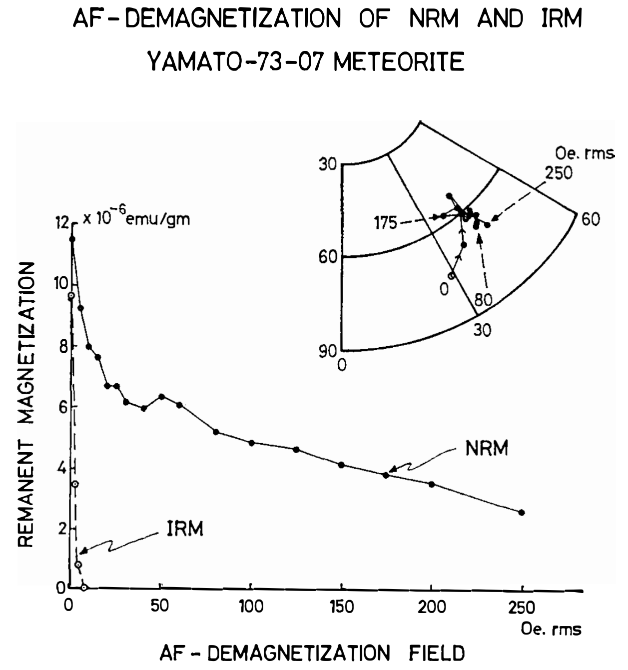

Fig. 6. *Changes in intensity and direction of NRM in the course of AF-demagnetization (Yamato-73-07 meteorite).*

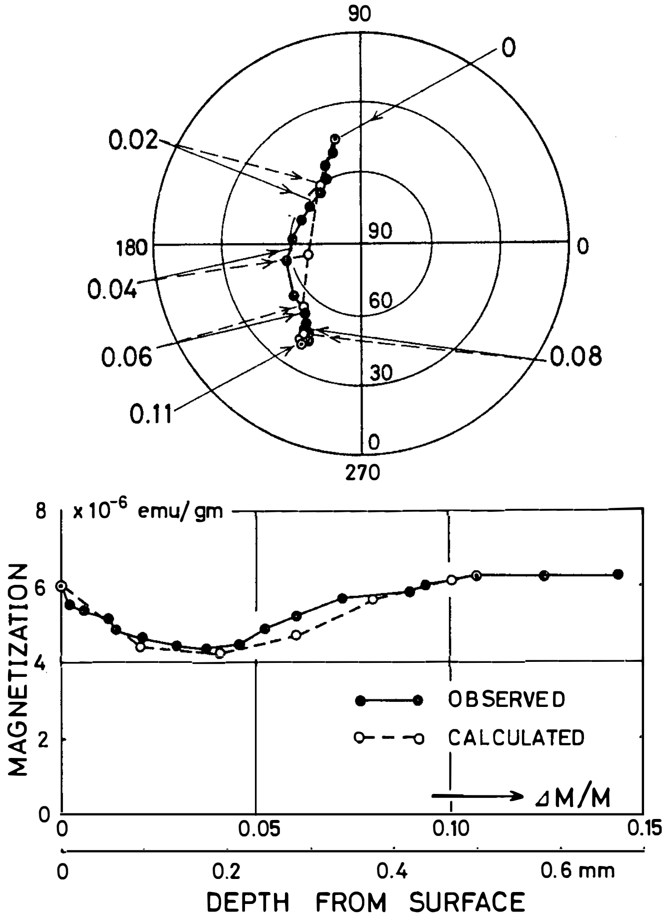

Fig. 7. Changes in the direction (top) and the intensity (bottom) of NRM with successive removal of the fusion crust skin of Yamato-73-07 meteorite.
Full circles: experimentally observed values.
Hollow circles: theoretically calculated values.

(TRM) of the fusion crust ought to be acquired in the presence of geomagnetic field if the orientation of this achondrite piece was roughly kept constant with respect to the geomagnetic field during the cooling process. Weaving (1962) specifically examined the secondary TRM of the surface skin of the Brewster meteorite, which is definitely distinguished from the original TRM of the central parts of the meteorite. In the present work also, a special study was made on a distinction of the secondary NRM of the fusion crust from the original NRM of the central parts of the meteorite piece.

A small piece (2.06 gm in weight) of Yamato-73-07 achondrite including a fusion crust of surface skin was subjected to NRM measurements by successively taking off thin layers from the surface. In Fig. 7, changes in the intensity and the direction of residual NRM after successively taking off thin layers from the surface are shown as a function of $\Delta M/M$, where M and ΔM represent respectively the initial mass of the examined sample and the total weight of taken-off portion. $\Delta M/M=0.15$ corresponds to about 0.7 mm in thickness of the taken-off layer. It is clearly seen in the figure that the original NRM of the interior of this sample is represented by 6.3×10^{-6} emu/gm in intensity and declination$=240°$ and inclination$=40°$ in direction after taking off the surface skin fusion crust by about $\Delta M/M=0.11$ (*i.e.* equivalent to 0.5 mm in thickness). Then, the initial NRM vector $(\vec{I_n})$ can be expressed by a vector sum of the original NRM vector $(\vec{I_o})$ of the interior and the secondary NRM vector $(\vec{I_f})$ of the fusion crust part, namely

$$\vec{I_n}=\vec{I_o}+\vec{I_f}\,. \tag{1}$$

The secondary NRM $(\vec{I_f})$ thus defined is estimated to be 7.6×10^{-5} emu/gm in average intensity and declination$\simeq80°$ and inclination$\simeq0°$ in direction. If the distribution of the secondary NRM in the fusion crust assumes that the direction is practically uniform but the intensity decreases linearly from the surface to practically zero value at $\Delta M/M=0.11$, the theoretically estimated intensity and direction of the total NRM as a function of $\Delta M/M$ fit well with the observed ones, as illustrated in Fig. 7. It will be concluded therefore that (a) the original NRM of the undisturbed interior of this metorite is 6.3×10^{-6} emu/gm in intensity and (b) the meterorite piece is covered with a fusion crust of about 0.5 mm in thickness, which has the secondary NRM whose direction is uniformly different by about 130° from that of the original NRM and whose intensity decreases from about 1.5×10^{-4} emu/gm at the surface to zero at the bottom (0.5 mm in depth from the surface) of the fusion crust.

Then, the most probable acquisition mechanism of the secondary NRM of the fusion crust may be attributable to production of the thermoremanent magnetization (TRM) in the geomagnetic field in the cooling process after this meteorite surface was reheated to result in formation of the fusion crust on entry into the earth's atmosphere. The acquisition characteristic of partial TRM of this sample is shown in Fig. 8. Although the second minor blocking temperature range is observed in addition to the main blocking temperature range for the nearly pure metallic iron phase, the observed secondary NRM of the fusion crust may be attributable mostly to TRM due to the main blocking temperature range, because it is believed that the surface temperature at least was raised over 800°C on entry into the earth's atmosphere. Then, the ambient magnetic field affecting this meteorite piece during its cooling process can be evaluated from the surface value of the secondary NRM, *i.e.* 1.5×10^{-4} emu/gm, to be about 0.44 Oersteds. The magnetic field intensity thus estimated may be reasonable as an approximate value of the geomagnetic field at high latitudes.

The thermal demagnetization experiment also was carried out on the original

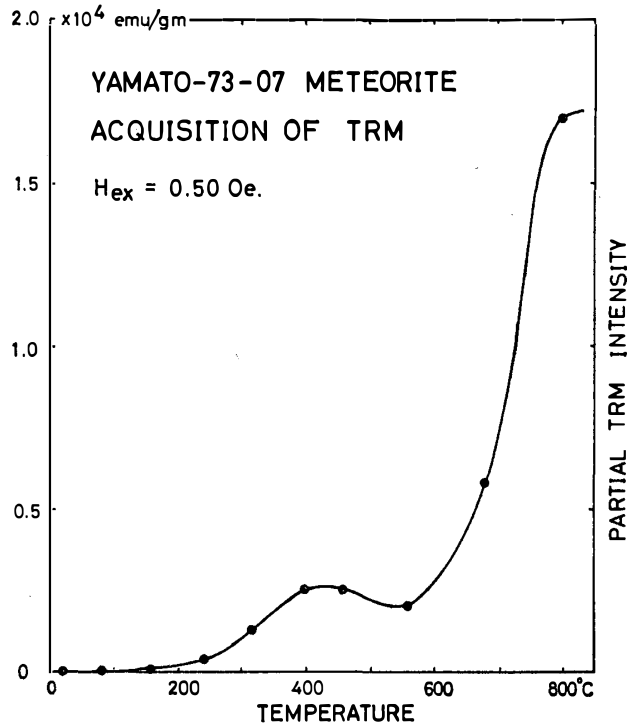

Fig. 8. Partial TRM acquisition curve of Yamato-73-07 meteorite.

TRM of an undisturbed interior part by means of Königisberger-Thellier method. However, the thermal demagnetization is not of a monotonous decreasing characteristic with an increase in temperature but an increase of remanence takes place in a temperature range of 400–480°C, suggesting the presence of magnetic interaction between at least two different ferromagnetic phases (*e.g.* PEARCE *et al.*, 1967). The TRM acquisition mechanism should be complicated in the case of meteorites, because the cooling rate of meteorites is so small, being $1 \sim 10$ degree/million years, that the Neel's relaxation time can become significantly long to be able to block the spontaneous magnetization of single-domain iron particles of 300 Å in mean diameter (namely, the maximum grain size for a single-domain iron particle) at temperatures only below about 700°K. In other words, it may not be expected that TRM acquired in the main blocking temperature range at about 800°C, shown in Fig. 8, is present as the original NRM of meteorites. It seems likely further in the case of Yamato-73-07 meteorite that TRM acquired in the second minor blocking temperature range at about 400°C is magnetically interacted with the neighbouring ferromagnetic metal grains. In such circumstances, an exact evaluation of the paleomagnetic field of the parent meteorite planet seems to be scarcely possible. By a comparison of the thermal demagnetization curve of the original NRM with the partial TRM acquisition curve, however, it could be roughly concluded that the paleomagnetic field should not be larger than 0.13 Oersteds.

5. Viscous Magnetization

The viscous magnetization of the two meteoritic samples was examined by measuring the time-decay of the isothermal remanent magnetization (IRM) in non-magnetic space. IRM of Yamato-73-04 chondrite is associated with a considerable amount of viscous magnetization (about one third). An example of the Richter-type viscous decay of the remanent magnetization is shown in Fig. 9. In this case, the intensity of remanent magnetization at time (t) after the acquisition of remanence is represented by

$$I(t)=I_0+\frac{\Delta I_v}{\ln(\tau_2/\tau_1)}\int_{\tau_1}^{\tau_2}\frac{e^{-t/\tau}}{\tau}d\tau, \qquad (2)$$

where I_0 and ΔI_v represent respectively the stable component of IRM and the total viscous remanent magnetization (VRM), and τ, τ_1, and τ_2 denote respectively the relaxation time and its lower and upper limits of an assumed uniform distribution of $\ln\tau$ (BECKER and DÖRING, 1939; NAGATA and CARLETON, 1970). Results of analysis of the observed data shown in Fig. 9 (where H_{ex}=4.7 Oe) by eq. (2) are represented by $\Delta I_v/I_0=0.46$, $\tau_1=240$ sec and $\tau_2=1\cdot2\times10^4$ sec. The observed magnetic viscosity can be attributed to the presence of very fine grains of single-domain native iron, as frequently confirmed in lunar surface materials (*e.g.* NAGATA and CARLETON, 1970). The Neel's relaxation time (τ) for single-domain ferromagnetic particles of v in volume is expressed by

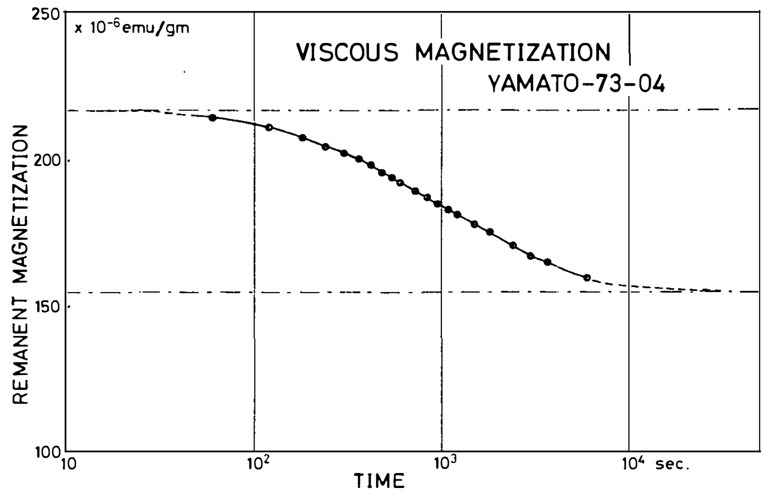

Fig. 9. Richter-type viscous magnetization characteristic curve of Yamato-73-04 meteorite.

$$\frac{1}{\tau}=f_0\exp\left(-\frac{vJ_sH_{Rc}}{2kT}\right), \tag{3}$$

where J_s denotes the spontaneous magnetization of the ferromagnetic particles and $f_0\simeq 10^9\,\mathrm{sec}^{-1}$. Then, the assumed approximation of a uniform distribution of $\ln\tau$ from τ_1 to τ_2 corresponds to a uniform distribution of v from v_1 to v_2. The estimated values of τ_1 and τ_2 correspond to $v_1=3.40\times 10^{-18}\,\mathrm{cm}^3$ and $v_2=3.88\times 10^{-18}\,\mathrm{cm}^3$ respectively. It may be concluded therefore that the native iron grains in this sample consist of a group of multi-domain grains larger than 300 Å in mean diameter and the other group of single-domain grains of 187–197 Å in mean diameter.

The viscous magnetization component in IRM of Yamato-73-07 achondrite is much smaller than that of Yamato-73-04 chondrite and is approximately represented simply by

$$I(t)=I_0+\Delta I_v \exp(-t/\tau)\,. \tag{4}$$

For the same external magnetic field (H_{ex}) as in the case of Fig. 9, $\Delta I_v/I_0=0.12$ and $\tau=170$ sec (corresponding to $v=3.3\times 10^{-18}\,\mathrm{cm}^3$) for Yamato-73-07 achondrite. Thus, this achondrite sample also contains a smaller amount of single-domain fine grains of native iron. The abundance of the single-domain grains relative to the multi-domain grains cannot be derived directly from a comparison of ΔI_v to I_o, because the acquisition mechanism of VRM is different from that of IRM. It has been discussed for the case of lunar materials (NAGATA *et al.*, 1972), however, that an extremely small value of H_c in comparison with H_{Rc} is mainly due to the co-existence of superparamagnetically fine grains of ferromagnetics. When a portion β of the total ferromagnetic component is in a form of the superparamagnetically fine particles and the remaining portion, $1-\beta$, is in a form of the multi-domain ferromagnetics, an apparent value of the coercive force (H_c) should be approximately given by

$$I_R\left(1-\frac{H_c}{H_{Rc}/2}\right)=\beta\frac{vJ_s}{3kT}I_sH_c, \tag{5}$$

where the true value of coercive force is reasonably assumed to be given by $H_{Rc}/2$. Putting observed values of I_R, H_c, H_{Rc} and I_s, given in Table 2, and the average value of v obtained from analysis of the viscous magnetization into eq. (5), β is evaluated to be 0.7 wt% for Yamato-73-04. Since H_{Rc} of Yamato-73-07 achondrite could not be measured because of its small value of magnetization, β cannot be estimated for this sample.

6. Concluding General Remarks

It is an established fact that the total abundance of iron in metals, iron oxides and iron sulfides is approximately constant throughout various types of chondrite (*e.g.* UREY and CRAIG 1953; MASON, 1962a). The metallic iron is the richest in enstatite chondrites and its content successively decreases in olivine-bronzite chondrites, in olivine-hypersthene chondrites and in olivine-pigeonite chondrites in the order, being the poorest in carbonaceous chondrites. Hence, the saturation magnetization (I_s) which is approximately proportional to the metallic iron content in chondritic meteorites, can be one of reasonable measures in classifying chondrites. Since nickel and cobalt in these chondrites compose mostly metallic alloys together with iron, *i.e.* kamacite and/or taenite, Curie point (Θ) and the $\alpha \rightleftarrows \gamma$ transition temperatures ($\Theta_{\alpha\to\gamma}$ and $\Theta_{\gamma\to\alpha}$) of kamacite phase also become reasonably good criteria for the classification. It is because the Ni content also is approximately constant through all types of chondrites, so that enstatite chondrites contain Ni-poor kamacites and olivine-bronzite- and olivine-hypersthene-chondrites contain 5–10 wt% Ni kamacites, whereas the main ferromagnetic component in most carbonaceous chondrites comprises magnetites (Fe_3O_4). Yamato-73-04 meteorite examined in the present study has the characteristic magnetic properties of the ordinary chondrite (either olivine-bronzite chondrite or olivine-hypersthene chondrite); namely, its main ferromagnetic constituents are kamacites of 6 wt% Ni and 13 wt% Ni, and the I_s-value amounts to 15 emu/gm at room temperature. The same magnetic criteria could be applied in classifying Yamato (a), (c) and (d) meteorites also (NAGATA *et al.*, 1975); Yamato (a) meteorite (enstatite chondrite) is magnetically characterized by a large value of I_s amounting to 48 emu/gm and the $\alpha \to \gamma$ transition representing 3 wt% Ni kamacite which are typical characters of enstatite chondrite; Yamato (c) meteorite (carbonaceous chondrite) by Curie point 540°C of magnetite and a very small amount of ferromagnetic metal which correspond to the typical magnetic properties of carbonaceous chondrites; Yamato (d) meteorite (olivine-bronzite chondrite) by $I_s=32$ emu/gm and the $\alpha \to \gamma$ transition representing 6 wt% Ni kamacite, which are typical magnetic parameters for olivine-bronzite chondrites.

On the other hand, the magnetic properties of Ca-poor achondrites can be characterized first by an considerably small content of ferromagnetic metals compared with chondrites, though a small amount of metallic iron is always present in achondrites. Yamato-73-07 meteorite has this typical magnetic characteristics of achondrites, and Yamato (b) meteorite (Ca-poor achondrite) also has the same characteristic. Both Yamato-73-07 and Yamato (b) meteorites are magnetically characterized by kamacite of poor Ni and Co and additional ferromagnetic component of about 570°C in Curie point. The second ferromagnetic phase in the two achondrites has not yet been exactly identified. It seems likely from the observed Curie point, however, that the main part of this second ferromagnetic phase is Fe_3O_4, but there is also a possibility that some parts of Fe^{3+} of the magnetite are substituted by Cr^{3+}. Further analysis of this second ferromagnetic phase in more detail will be important in relation to the magnetic properties of achondrites.

Another significant problem in connection with the magnetic characteristics of meteorites is the secondary remanent magnetization of their fusion crust. In the present study, the secondary NRM of the fusion crust is separated out from the original NRM of the undisturbed interior. The natural remanent magnetization of the fusion crust and that of the interior of the Brewster meteorite (olivine-hypersthene chondrite) were examined in detail by WEAVING (1962). In his result, the NRM intensity of the fusion crust is much larger than that of the central part, but the TRM interpretation of the intense fusion crust NRM requires an ambient magnetic field of about 10 Oe during its cooling process. Results of the present study have shown in the case of Yamato-73-07 achondrite that the direction of the secondary NRM of the fusion crust is definitely different from that of the original NRM of the interior part and the intensity of the former is much larger than that of the latter, and that the TRM interpretation of the fusion crust NRM requires a magnetic field of about 0.44 Oe. Thus, the comparatively intense NRM of the fusion crust of Yamato-73-07 achondrite is naturally attributable to an effect of the geomagnetic field during the cooling process of the skin fusion crust. As all collected Yamato meteorites are entirely or partially covered with fusion crusts, the problem on the secondary NRM of fusion crusts will be more systematically examined with more samples in the future, in order to definitely clarify the aquisition mechanism of the secondary NRM. It can be argued at the present stage, however, that the secondary NRM of the fusion crust must be eliminated in the paleomagnetic study of meteorites for discussing the paleomagnetic field of the parent meteorite planet. As for the paleomagnetic field corresponding to the original NRM of the interior, WEAVING estimated it as about 0.1 Oe for the Brewstar chondrite, while the present study suggests it to be less than 0.13 Oe for the Yamato-73-07 achondrite.

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