1. *The Natural Remanent Magnetism of Volcanic Rocks and Its Relation to Geomagnetic Phenomena.*

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CONTENTS.

General Introduction.

Since discovery of the magnetic properties of the lodestone, those of minerals and rocks have engaged the attention of a number of investigators, sometimes in connexion with the origin of the earth's permanent magnetic field, sometimes in connexion with the physical interpretation of the magnetic field, and at others in connexion with the physical interpretation of the magnetism of various substances. However, even after the general aspect of a geomagnetic field, especially its diurnal variation and its disturbances, such as magnetic storms, were clarified from the theory of electro-magnetic induction, which was developed chiefly by B. Stewart,[1] A. Schuster,[2] and S. Chapman,[3] the problem of the possible relation between the geomagnetic field and the magnetization of the earth's crust had been rather neglected, until recent years when the need of examining the magnetic properties of rocks in order to interpret correctly the results of magnetic prospecting of geological structure became pressing. As is well known, the interpretation of local geomagnetic surveys is based primarily on a knowledge of the magnetic properties of geological structure. Since discovery of the marked geomagnetic anomaly in Kursk,[4] both the technique of measuring the local geomagnetic anomaly and the knowledge of the magnetic behaviour of rocks were advanced by numerous investigators, especially, the magnetic susceptibility of various minerals, which was fully examined by H. Reich.[5] While it was re-discovered along with the progress in our knowledge of the local magnetic anomaly and the magnetism of rocks and ores that the local geomagnetic anomaly is not uniquely due to the magnetization of subterranean rocks induced by the present geomagnetic field, the necessity was felt of taking into consideration the existence of remanent magnetization of rocks, resulting in, for example, H. Haalck's[6] calling attention to the case of the big anomaly in Kursk.

1) B. STEWART, *Encyclopaedia Britanica*, 9th edition (1882), 36.
2) A. SCHUSTER, *Phil. Trans. London. A*, **208** (1908), 163.
3) S. CHAPMAN, *Phil. Trans. London. A*, **214** (1914), 295; **218** (1919), 1; and others.
4) A. D. ARCHAGELSKI, *Transact. of the Special Comm. for the Study of Kursk Magne. Anomaly*, Vol. 7 (1926).
P. LASAREFF, *Gerl. Beitr. Geophys.*, **15** (1926), 71.
G. A. GANBURZEFF u. POLIKAEPOFF, *Gerl. Beitr. Geophys.*, **19** (1928), 210, 219.
H. HAALCK, *Gerl. Beitr. Geophys.*, **22** (1929), 219, 241.
5) H. REICH, *Handb. Geophy.*, Bd. VI, pp. 50 Berlin (1919), *ZS. Deutsch. Geol. Gesel.*, **83** (1931), 502.
6) H. HAALCK, *Gerl. Beitr. Geophys.*, **22** (1929), 241, 385.

On the other hand, that igneous rock has more or less remanent magnetism was believed, at any rate, by a circle of geophysicists. Moreover, G. Folgheraiter,[7] P. Mercanton,[8] J. G. Königsberger,[9] M. Matuyama,[10] and others,[11] have studied the direction of remanent magnetization of old pottery, igneous rocks, and lava fields, for the purpose of presuming the direction of the geomagnetic field in the past under the assumption that these rocks retain the permanent magnetization which they had acquired during cooling in a geomagnetic field, exactly as in the case of a piece of hot steel when it is cooled by being plunged into cold water under the inducing influence of the magnetic field.

Königsberger[12] was the first to point out that the intensity of remanent magnetization of igneous rocks is fairly large compared with their induced magnetization in the earth's magnetic field, and that the cause of the remanent magnetism is probably owing to a particular irreversible magnetization of ferro-magnetic minerals during cooling from a sufficiently high temperature in the earth's magnetic field. Although Königsberger's conclusion was the result of his comprehensive experimental studies of a number of igneous rocks, his interpretation of the experimental resuls is based also on a few assumptions derived from the classical phenomenological theory of ferro-magnetism. Although the recent extensive studies of R. Chevallier[3] and E. Thellier[4] have advanced our knowledge of the magnetic properties of rocks, their conclusions do not always represent correctly the magnetic behaviour of rocks in the earth's magnetic field, for the reason that their experiments were made in a magnetic field fairly more intense than that of the earth.

In short, the magnetic properties of igneous rocks, especially, their fairly intense remanent magnetization, is not yet fully understood. For example, according to S. Chapman and J. Bartels,[5] in their recent-

7) G. Folgheraiter, *Real. Acc. Lincei.*, **4** (1895), 78, 203, **5** (1897), 64.

8) P. Mercanton, *C.R. Acad. Sci.*, **143** (1906), 138; **182** (1926), 859, 1231.

9) J. G. Königsberger, *Geil. Beitr. Geophys.*, **35** (1932), 51, 204; *Beitr. Angew. Geophys.*, **4** (1934), 385.

10) M. Matuyama, *Proc. Imp. Acad. Japan*, **5** (1929), 203; *Proc. 4th Pacific Sci. Cong.*, (Java 1929), 567; and others.

11) Ch. Maurain, F. Pockels, P. David, S. Nakamura, R. Chevallier, E. Thellier and others. See the literatures in the foot-notes in Chapters II and III.

12) J. G. Königsberger. *loc. cit.* The results of his studies are summarized in his paper "Natural Remanent Magnetism of Eruptive Rocks." (Terr. Mag. **43** (1938), 119, 299.)

13) R. Chevallier, *Ann. Physique*, **4** (1925), 5.

14) E. Thellier, *Ann. Phys. Globe*, **16** (1938), 157.

15) S. Chapman and J. Bartels, "Gemagnetism" I. pp. 154, Oxford (1940).

ly published well known book, "Geomagnetism", "it is certain that igneous rocks acquire a remanent magnetization during their cooling when the temperature passes through the Curie-point ; the magnetic elementary particle, free to move in the molten lava, align themselves in the direction of the magnetizing field, and keep that direction when rock solidifies....." That this explanation of the causation of the remanent magnetization of rocks does not agree with facts established from the results of actual experiments with rocks, will be clear from the present study.

On the other hand, we have at present an interesting phenomenon of geophysics that seems to be closely related to the magnetic properties of the earth's crust, especially to the mode of its change with temperature. It is the local anomalous variation in the geomagnetic field that follows severe volcanic activities, a number of actual examples of marked anomalous variation having been observed without the least doubt, for example, by a few investigators in Japan,[16] which is what accounts for our keen interest in the mode of change in the magnetism of rocks with temperature, i. e. changes in the susceptibility and in the development of remanent magnetization, seeing that these phenomena can be presumed to occur in the earth's crust accompanying changes in temperature due to volcanic activity, with the result that they cause disturbances of considerable magnitude in the geomagnetic field in the neighbourhood of volcanoes.

If the formulae, in which the magnetic behaviour is influenced by temperature, external magnetic field and pressure, could be physically established, we should be able to form some idea of the physical conditions and their changes in the earth's crust by measuring the anomalous distribution of the geomagnetic field on the earth's surface and its changes. In order to do this, it is necessary to examine experimentally the magnetic properties of various igneous rocks, the petrological and chemical constitutions of which have been accurately determined. The physical aspect of the causes that gave rise to the intense remanent magnetism, especially, will be interesting both from the geophysical standpoint mentioned above and from the standpoint of the physics of the magnetism of rock, say an aggregate of various complex solid solutions and crystals.

16) S. T. NAKAMURA, *Proc. Imp. Acad. Japan*, **11** (1935), 102.
H. NAGAOKA and T. IKEBE, *Proc. Imp. Acad. Japan*, **13** (1937), 62, 251.
R. TAKAHASI and K. HIRANO, *Bull. Earthq. Res. Inst.*, **19** (1941), 82.
T. MINAKAMI, Zisin, **10** (1938), 430.
Y. KATO, *Proc. Imp. Acad. Japan*, **16** (1940), 440.
T. NAGATA, Zisin, **10** (1938), 221. *Bull. Earthq. Res. Inst.*, **19** (1941), 335.

In this report, therefore, the characters of remanent magnetism, the most interesting problem in rock magnetism, will be the chief topic. The general magnetic behaviour of volcanic rocks, namely, susceptibility and its change with temperature, and the hysteresis phenomenon, are summarized in Chapter I, with a brief description of the petrological and chemical characters of the rocks that were examined. Although the method of determining the remanent magnetism of rocks, together with a discussion of the observed results, and the mode of development of the remanent magnetism with decrease in temperature in a magnetic field, will be found in Chapters II and III respectively, in the latter the possible physical mechanism of this phenomenon is also discussed. In the final Chapter, the relation of the remanent magnetism of volcanic rocks and its change with temperature to a number of geophysical phenomena are discussed in connexion with actual examples obtained by the writer and others.

CHAPTER I. GENERAL MAGNETIC BEHAVIOUR OF VOLCANIC ROCKS.

§ 1. Brief description of the rock samples examined in the present investigation.

In order to examine their magnetic properties, many kinds of volcanic rocks ejected from various volcanoes in the Kantô District were collected. Of these, 47 rock specimens from volcanoes Huzi, Taga, Usami, Amagi, Asama, Hakone, and Mihara have already been studied petrologically by H. Tsuya,[1] H. Kuno,[2] and I. Iwasaki,[3] the chemical composition of these samples having been determined also by them. Here, since about half of the collected sample, each about 10 *cm* in mean diameter, was petrologically and chemically studied, and the remaining half magnetically examined, it follows that the chemical composition of that part of the rock that was magnetically examined is exactly the same as that obtained from the remaining half.

The classification of these rocks in the following list, where their localities and chemical compositions are also given, is the same as that of the foregoing petrologists.

1) H. TSUYA, *Bull. Earthq. Res. Inst.*, **15** (1937), 215.
2) H. KUNO, "Kazan", **3** (1936), 53.
3) I. IWASAKI, *Journ. Chem. Soc. Japan*, **56** (1935), 1511.

(i) Ejecta from Asama Volcano.[4]

No. 5. Bomb ejected in 1929,
Olivine-bearing hypersthene-augite andesite,

No. 6. Temmei lava ejected in 1783,
Olivine-bearing hypersthene-augite andesite,

No. 8. One of the Kurohuyama lava,
Olivine-bearing two-pyroxene-andesite,

No. 11. Lava flow in Mugen-no-tani,
Hornblende-two-pyroxene-dacite, (obsidian),

Table I–1–I. Chemical composition of ejecta from Volcano Asama.

	No. 8	No. 14	No. 5	No. 6	No. 15	No. 11
SiO_2	58·39	59·67	60·24	62·37	69·93	72·01
Al_2O_3	17·89	16·83	16·43	15·90	14·12	14·37
Fe_2O_3	2·45	1·44	1·54	1 14	2·45	0·73
FeO	5·33	5·51	5·26	4·79	0·45	1·81
MgO	3·30	3·92	3·92	3·43	0·58	0·57
CaO	7·23	7·21	7·07	6·13	2·42	2·56
Na_2O	3·04	2·81	3·00	3·20	4·04	4·35
K_2O	0·71	1·00	1·16	1·49	2·25	2·39
H_2O+	0·36	0·28	0·17	0·30	2·00	0·26
—	0·12	0·13	0·06	0·12	0·97	0·08
TiO_2	0·71	0·70	0·70	0·66	0·38	0·41
P_2O_5	0·15	0·14	0·13	0·13	tr.	0·07
MnO	0·14	0·12	0·11	0·11	0·09	0·08
Total	99·82	99·76	99·79	99·77	99·68	99·69
Q	15·20	15·92	15·98	17·84	31·77	30·39
Or	4·45	6·12	6·68	8·90	13·36	13·91
Ab	25·69	23·59	25·17	27·26	24·08	36·70
An	32·82	30·22	28·09	24·48	11·96	12·79
Di	1·83	3 87	5·02	4·12	—	—
C	—	—	—	—	0·71	—
Hy	14·19	15·89	14·66	13·63	1·24	3·65
Mt	3·49	2·08	2·31	1·62	0·23	1,16
Hm	—	—	—	—	2·24	—
Ap	0·31	0·31	0·31	0·31	tr.	—
Il	1·37	1·37	1·37	1·21	0·76	0·61

4) TSUYA, *Bull. Earthq. Res. Inst.*, **11** (1933), 578.

No. 14. Yunotaira lava, an older lava of the Maekakeyama centre,
Hypersthene-augite-andesite,

No. 15. Ko-Asama lava forming a small domed hill,
Hornblende-two-pyroxene-dacite.

(ii) Ejecta from Huzi Volcano.[5)]

No. 17. One of the lava flows exposed on the western wall of the Hôei explosion-crater,
Olivine-basalt (almost aphyric),

No. 18. One of the lava flows at Makuiwa on the southeastern slope,
Hypersthene-augite-olivine-basalt,

Table I-1-II. Chemical composition of ejecta from Volcano Huzi.

	No. 17	No. 19	No. 20	No. 23	No. 21	No. 18	No. 22	No. 25	No. 24
SiO_2	49·60	49·60	50·28	50·64	50·66	51·05	51·09	51·30	63·84
Al_2O_3	16·96	16·14	18·30	18·58	18·25	18·35	17·62	18·74	15·82
Fe_2O_3	5·40	3·67	4·50	3·04	4·78	2·76	2·64	1·83	0·95
FeO	6·65	9·90	6·89	7·29	5·72	7·72	8·42	8·34	5·02
MgO	5·92	4·79	3·80	5·58	4·94	4·63	5·09	4·80	1·67
CaO	10·03	8·80	9·76	10·00	9·98	9·90	9·68	9·76	4·88
Na_2O	2·48	2·90	2·87	2·64	2·78	2·81	2·80	2·55	3·88
K_2O	0·58	0·93	0·94	0·61	0·77	0·81	0·76	0·71	2·12
H_2O+	0·50	0·55	0·20	0·20	0·38	0·40	0·28	0·22	0·45
H_2O-	0·12	0·14	0·08	0·06	0·13	0·11	0·06	0·06	
TiO_2	1·40	1·97	1·78	1·15	1·38	1·41	1·38	1·43	0·87
P_2O_5	0·20	0·31	0·34	0·16	0·25	0·24	0·26	0·29	0·22
MnO	0·21	0·23	0·20	0·17	0·17	0·18	0·21	0·28	0·17
Total	100·05	99·93	99·94	100·12	100·19	100·37	100·29	100·31	99·89
Q	3·78	1·26	3·96	1·98	4·14	2·34	2·04	2·82	18·14
Or	3·34	7·57	5·57	3·34	4·45	5·01	4·45	4·45	12·80
Ab	20·97	24·64	24·14	22·55	23·59	23·59	23·59	21·50	33·03
An	33·38	28·09	34·49	36·99	35·05	35·05	33·38	37·55	19·19
Wo	6·50	5·69	5·11	4·88	5·34	5·23	5·46	3·83	1·28
En	14·76	11·95	9·43	13·85	12·34	11·54	12·65	11·95	4·11
Fs	5·80	12·27	6·47	9·24	4·62	9·89	11·35	12·01	7·25
Mt	7·87	5·33	6·48	4·40	6·95	3·94	3·94	2·55	1·39
Il	2·73	3·79	3·34	2·12	2·58	2·73	2·58	2·73	1·67
Ap	0·33	0·65	0·65	0·33	0·65	0·65	0·65	0·65	0·65

5) H. TSUYA, *Bull. Earthq. Res. Inst.*, **13** (1935), 645; **15** (1937), 302; **16** (1938), 638. Kazan, **2** (1935), 149.

No. 19. The lava flow at Tawarano-taki,
Olivine-basalt (almost aphyric),
No. 20. One of the lava flows at Makuiwa,
Olivine-basalt,
No. 21. One of the lava flows at Makuiwa,
Augite-bearing olivine-basalt,
No. 22. Scoriaceous lava-block ejected from Hôei crater in 1707,
Augite-bearing olivine-basalt,
No. 23. One of the lava flows at Makuiwa,
Two-pyroxene-bearing olivine-basalt,
No. 24. Karasu-isi lava at Nagamine on the southeastern slope,
Aphanitic andesite,
No. 25. Aokigahara-lava,
Two-pyroxene-olivine-basalt.

(iii) Ejecta fr m Amagi volcano.[6]

No. 27. Zizôdô lava, a lava flow that poured out of Maruno-yama,

Table I-1-III. Chemical composition of ejecta from Volcano Amagi.

	No. 27	No. 28	No. 34	No. 29	No. 30	No. 32	No. 31
SiO_2	49·51	51·48	54·81	59·11	60·09	62·60	71·16
Al_2O_3	18·19	18·25	18·09	16 31	16·64	15·98	14·72
Fe_2O_3	2·89	1·92	1·54	3·94	3·03	2·66	1·12
FeO	7·66	6·77	7·00	2·86	3·30	3·40	2·15
MgO	7·07	5·93	4·50	3·49	3.27	2·84	1·30
CaO	9·83	10·32	9·03	6·02	6·59	5·84	3·50
Na_2O	2·49	3·21	2·92	2·97	3·60	3·00	3·61
K_2O	0·48	0·34	0·57	1·47	1·40	1·66	2·06
H_2O+	0·45	0·43	0·50	1·48	0·68	0·45	0·27
H_2O-	0·27	0·25	0·50	1·00	0·70	0·24	0·10
TiO_2	0 64	1 03	0·79	0·76	0·80	0·65	0·36
P_2O_5	0·17	0·12	0·13	0·17	0·19	0·17	0·08
MnO	0·28	0·27	0·14	0·13	0·11	0·11	0 08
Total	99·93	100·32	100·18	99·71	100·40	99·60	100·51
Q			7·33	18·38	15·44	21·62	32·19
Or	2·78	2·23	3·34	8·90	8·35	10·02	12·24
Ab	20·97	27·26	24·64	25·17	30·41	25·17	30·41
An	36·99	34·21	34·49	26·70	25·03	25·31	16·41
Wo	4·53	6·74	3·95	0·93	2·91	1·16	C=0·51
En	16·06	14·30	11·24	8·73	8·13	7·02	3·21
Fs	10·29	9·30	10·42	0·92	2·51	3·17	2·51
Fo	1·06	0·39					
Fa	0·71	0·25					
Mt	4·17	2·55	2·32	5·79	4·40	3·94	1·52
Il	1·21	1·97	1·52	1·52	1·52	1·21	0·76
Ap	0·33	0·33	0·33	0·33	0·33	0·33	0·33

6) H. TSUYA, *Bull. Earthq. Res. Inst.*, 15 (1937), 252.

parasitic cone on the northern flank of Amagi,
Olivine-basalt.

No. 28. Hatikubo-yama lava, a lava flow that issued from Hatikubo yama, a parasitic cone on the north-western flank of Amagi,
Olivine-basalt,

No. 29. Yawatano lava, the lava on the northeastern flank of Amagi,
Qua.'z-olivine-bearing two-pyroxene-andesite,

No. 30. Inatori lava, a lava flow forming a terrace at Inatori,
Olivine-bearing two-proxene-andesite,

No. 31. Iwano-yama lava, the lava forming the parasitic lava dome, (Iwano-yama),
Olivine-bearing two-pyroxene-hornblende-dacite,

No. 32. Yanagase lava, a lava flow on the northern flank of Amagi,
Two-pyroxene-andesite.

No. 33. Yahadu-yama lava, the lava forming the twin dome, Yahadu-yama,
Olivine-bearing two-pyroxene-hornblende-dacite,

No. 34. Akasawa lava, a lava flow that poured out of a parasitic volcano, either Togasa or Iio-yama,
Olivine-two-pyroxene-bearing andesite.

(iv) Ejecta from Mihara Volcano.[7]

No. 35. Old lava of Oosima exposed on the sea-cliff, immediately west of Okata,
Hypersthene-bearing olivine-basalt,

No. 36. Somma lava of Mihara Volcano, southwestern part of ring-wall of the somma,
Hypersthene-bearing olivine-basalt,

No. 37. An'ei lava at the northern foot of the central cone,
Hypersthene-bearing basalt,

No. 38. Meizi-Taisyo lava ejected during 1911–1914,
Basalt,

No. 39. Old lava of Oosima exposed on the sea-cliff immediately west of Okata,
Aphyric basalt.

7) H. TSUYA, *Bull. Earthq. Res. Inst.*, **15** (1937), 296.
I. IWASAKI, *Journ. Chem. Soc. Japan*, **56** (1935), 1511.

Table I-1-IV. Chemical composition of ejecta from Volcano Mihara.

	No. 35	No. 39	No. 36	No. 37	No. 38
SiO_2	48·02	51·25	51·23	52·45	52·53
Al_2O_3	20·48	14·73	15·24	13·48	15·25
Fe_2O_3	2·13	3·82	4·18	4·60	2·69
FeO	7·05	10·22	8·99	10·02	10·57
MgO	5·79	5·47	4·85	4·78	4·54
CaO	14·14	11·73	11·14	10·23	10·76
Na_2O	1·28	1·85	2·19	1·99	1·89
K_2O	0·30	0·26	0·54	0·52	0·43
H_2O	n.d.	n.d.	n.d.	n.d.	n.d.
TiO_2	0·78	0·81	0·83	1·39	0·74
P_2O_5	0·01	0·13	0·03	0·33	0·41
MnO	0·26	0·28	0·33	0·31	0·24
Total	100·363	100·708	99·68	100·319	100·157
Q	0·24	5·41	5·35	9·61	7·87
Or	1·67	1·67	3·34	3·34	2·78
Ab	11·01	15·73	18·35	16·78	15·73
An	49·23	30·88	30.04	26·15	31·99
Wo	8·71	11·03	10·57	9·41	7·78
En	14·46	13·65	12·05	11·95	11·34
Fs	10·42	14·77	12·01	12·80	16·36
Mt	3·01	5·56	6·02	6·71	3·91
Il	1·52	1·52	1·52	2·58	1·37
Ap	tr.	0·32	tr.	0·65	0·99

(v) Ejecta from Usami Volcano.[8]

No. 49. The lava flow on the southern and southwestern part of this volcano,
Olivine-two-pyroxene-andesite,

No. 50. The lava flow forming the lower part of the western slope, (one of the earlier lavas of this volcano),
Aphyric two-pyroxene-andesite,

No. 51. The lava flow in a small area immediately north of Isigami-yama,
Two-pyroxene-andesite,

No. 52. The lava flow underneath the lava flow of No. 49,
Hypersthene-bearing olivine-augite-andesite,

8) H. TSUYA, *Bull. Earthq. Res. Inst.* 15 (1937), 264.

No. 53. The lava 5 *km* northeast of Hiyekawa,
Augite-bearing olivine-hypersthene-andesite,

No. 54. The lava flow on the west slope of this volcano,
Olivine-bearing two-pyroxene-andesite,

No. 55. The lava forming the hill that lies between Itô and Kawana,
Olivine-bearing two-pyroxene-andesite,

No. 56. Massive lava situated between Usami and Kameisi-tôge,
Augite-hypersthene-andesite.

Table I–1–V. Chemical composition of ejecta from Volcano Usami.

	No. 55	No. 52	No. 54	No. 56	No. 49	No. 53	No. 50	No. 51
SiO_2	50 57	50·59	51·67	53·08	55·96	57·61	59·00	62·70
Al_2O_3	19·38	19·73	19·88	20·01	16·92	15·88	15·26	16·09
Fe_2O_3	3·81	3·18	3·25	3·48	2·76	2·79	4·60	2·61
FeO	7·25	6·73	6·97	5·70	6·03	5·28	5·90	4·61
MgO	3·89	4·15	3·30	2·80	4·39	5·92	2·14	1·49
CaO	11·43	10·31	9·46	9 80	8·06	7·58	6·36	5·25
Na_2O	1·86	1·94	2·18	2·29	2·07	2·59	2·83	3·26
K_2O	0·16	0·24	0·40	0·31	0·57	0·71	0·58	0·86
H_2O+	0·45	1·13	1·26	0·47	1·07	0·36	0·90	1·25
H_2O-	0·30	0·88	1·00	0·75	1·18	0·28	0·69	0·85
TiO_2	0·76	0·85	0·92	0·73	0·73	0·71	1·04	0·82
P_2O_5	0·10	0·10	0·10	0·06	0·08	0·07	0·11	0·17
MnO	0·13	0·15	0·16	0·17	0·17	0·15	0·18	0·12
Total	100·09	99·98	100·55	99·65	99·99	99·93	99·59	100·08
Q	7·57	7·75	9·31	11·65	15·68	13·33	21·98	24·80
Or	1·11	1·67	2·23	1·67	3·34	4·45	3·34	5·01
Ab	15·73	16·25	18·35	19·40	17·30	22·02	24·12	27·79
An	43 95	44·50	43·39	43·39	35·33	29·48	27·26	25·31
Wo	4·99	2·44	1·16	2·21	1·63	3·37	1·39	C=0·51
En	9·64	10·33	8·23	6·92	10·94	14·76	5·32	3·71
Fs	9·10	8·58	8·83	6·60	7·91	6·36	6·67	5·28
Mt	5·56	4·63	4·63	5·09	3·94	3·94	6·71	3·70
Il	1·52	1·67	1·82	1·37	1·37	1·73	1·97	1·52
Ap	0·33	0·33	0·33	0·33	0·33	0·33	0·33	0·33

(vi) Ejecta from Hakone Volcano.[9]

No. 78. The upper part of an old somma lava (near Hakone-tôge),

9) H. KUNO, "Kazan" (Volcano), 3 (1936), 53.

quartz- and hornblende- bearing augite-pigeonite-hypersthene-labradorite-andesite,

No. 79. Hutago-yama lava in the central cone,
Olivine-bearing hypersthene-augite-labradorite-andesite,

No. 80. The upper part of an old somma lava (northwestern foot of Kurakake-yama),
Augite-hypersthene-labradorite-andesite,

No. 81. The upper part of an old somma lava, (Iwato-yama north of Atami),
Augite-hypersthene-bytwonite-andesite,

No. 82. The upper part of a new somma lava, (Bunko-yama east of Hakone-mati),
Hypersthene-augite-andesite.

Table I–1–VI. Chemical composition of ejecta from Volcano Hakone.

	No. 80	No. 79	No. 78	No. 81	No. 82
SiO_2	55·83	57·07	57·22	59·77	67·37
Al_2O_3	17·75	17.53	17·46	17·00	15·28
Fe_2O_3	2·76	2·59	2·45	2·49	1·13
FeO	6·01	5·44	6·15	5·59	3·86
MgO	3·90	3·87	3·77	2·24	1·20
CaO	8·62	8·77	7·93	7·05	4.46
Na_2O	2·98	2·80	2·87	3·29	4·67
K_2O	0·55	0·52	0·75	0·62	0·92
$H_2O(+)$	0·19	0·27	0·30	0·41	0·14
$H_2O(-)$	0·17	0·12	0·23	0·51	0·09
TiO_2	0·61	0·77	0·73	0 63	0·72
P_2O_5	0·13	0·08	0·11	0·10	0·18
MnO	0·17	0·14	0·16	0·16	0·15
Total	99·67	99·97	100·13	99·86	100·17
Q	10·68	13·68	12·96	17·94	24·06
Or	3·34	2·78	4·45	3·34	5·56
Ab	25.15	23·85	24·10	27·77	39·30
An	33·64	33·92	32·80	30·02	18·07
Wo	3·48	3·60	2·44	1·74	1·39
En	9·80	9·70	9·40	5·60	3·00
Fs	7·79	6·73	8·32	7·39	5·28
Mt	4·18	3·71	3·71	3·71	1·62
Il	1·22	1·52	1·37	1·22	1·37
Ap	0·34	0·34	0·34	0·34	0·34

(vii) Ejecta from Taga Volcano.[10]

No. 91. The lava flow on the southwest slope of this volcano,
Augite-olivine-basalt,

No. 92. The lava flow on the southwest slope of the above volcano,
Augite-bearing olivine-basalt,

No. 93. The lava flow on the south slope of do,
Two-pyroxene-olivine-basalt,

No. 94. One of the earlier lavas exposed on a sea-cliff near Uomi-saki,
Two-pyroxene-andesite,

No. 95. A lava flow sandwitched between lavas of porphyric augite-olivine-basalt, 1 *km* east of Nirayama.
Aphanitic basalt,

No. 96. A massive lava, about 15 *m* thick, in the vicinity of Ono, Kita-Kanomura,
Olivine-two-pyroxene-bearing andesite.

Table I-1-VII. Chemical composition of ejecta from Volcano Taga

	No. 91	No. 92	No. 93	No. 94	No. 95	No. 96
SiO_2	50·17	50·38	50·71	50·94	52·13	52·85
Al_2O_3	19·65	20·36	18·84	19·32	15·68	18·37
Fe_2O_3	2·54	2·86	2·42	4·14	4·36	3·48
FeO	7·92	7·26	7·13	5·52	9·81	7·83
MgO	4·54	4·45	6·08	4·20	4·08	2·96
CaO	11·29	11·28	10·05	10·62	8·87	10·12
Na_2O	1·66	1·95	2·03	2·15	2·44	2·07
K_2O	0·29	0·33	0·24	0·32	0·32	3·36
H_2O+	0·41	0·36	0·69	0·73	0·67	0·39
H_2O-	0·20	0·25	0·38	1·04	0·50	0·16
TiO_2	0·87	0·84	0·67	0·78	0·74	0·93
P_2O_5	0·10	0·10	0·13	0 07	tr.	tr.
MnO	0·17	0·15	0·15	0·18	0·32	0·17
Total	99·81	100·57	99·52	100·01	99·92	99·69
Q	5·77	4·93	4·44	7·87	8·35	11·11
Or	1·67	2·23	1·67	1·67	1·67	2·23
Ab	14·16	16·25	17·30	18·35	20·45	17·30
An	45·34	45·89	41·44	42·28	31·15	39 78
Wo	4·06	3·83	3·14	4·30	5·34	4·30
En	11·35	11·04	15·16	10·44	10·14	7·33
Fs	11·21	9·77	10·29	5·80	13·98	10·16
Mt	3·70	4·17	3·47	6·02	6·25	5·09
Il	1·67	1·67	1·21	1·52	1·37	1·82
Ap	0·33	0·33	0·33	0·33	tr.	tr.

10) H. TSUYA, *Bull. Earthq. Res. Inst.*, 15 (1937), 270.

Although the various magnetic properties of about 150 rock samples were examined in the present study, those of 47 rock specimens, the petrological and chemical characters of which are known at present, will be chiefly dealt with in the present report.

On the other hand, some rock specimens from Volcanoes Huzi, Mihara, Miyake-sima, Asama, Amagi, and Taga were examined for the particular purpose of ascertaining the possible relation between the magnetic property of rock and the local geomagnetic anomaly, and to determine the direction of residual permanent magnetization of massive ejecta. The localities and petrological characters of these rock samples which will be dealt with in the paragraphs to follow, are

Nos. 1, 2, 3, Meizi-Taisyô lava ejected from the central cone of Volcano Mihara during the period from 1911 to 1914, Basalt,[11]

Nos. 1′, 2′, 3′, 4′, 10′, 11′, and 12′. An'ei lava ejected from the central one of volcano Mihara in 1778, Olivine-basalt,[11]

Nos. 58, 59. Akabakkyo lava flow ejected from Volcano Miyake-sima in 1941, Olivine-hypersthene-pyroxene-basaltic andesite,[12]

Nos. 60, 60′, 60″. Yoridai-sawa lava ejected from Volcano Miyake-sima in 1941, Olivine-hypersthene-pyroxene-basaltic andesite,[12]

Nos. 63, 64, 65. Bombs from the crater of Hyôtan-yama, (a new parasitic cone resulting from the volcanic activity of Miyake-sima in 1941), Pyroxene-olivine-basalt,[12]

Nos. 67, 68, 69, 70, 71. Large bombs ejected from the central cone of volcano Mihara in 1941, (petrologically almost the same as Meizi-Taisyo lava),

Nos. 101–110. Aokigahara lava flow ejected from Nagaoyama (one of the parasitic cones of volcano Huzi), Two-pyroxene-olivine-basalt,

Nos. 131, 132, 133. Yahatano-lava flow from volcano Amagi, the petrological characters are nearly the same as that of sample No. 29,

Nos. 134, 135. Lava ejected from Omuro-yama,

11) S. TSUBOI, *Journ. Col. Sci. Tokyo Imp. Univ.*, **43** (1920), 6.

12) H. TSUYA, *Bull. Earthq. Res. Inst.*, (1941), 492. *Zisin*, **12** (1940), 474.

Basalt,

Nos. 141, 142, 143. Lava ejected from volcano Taga, which was exposed at Aziro.

§ 2. Magnetic susceptibility of igneous rocks in a weak magnetic field.

(i) For the purpose of describing the magnetic behaviour of various kinds of igneous rocks in the earth's magnetic field, we first measured the extent of magnetic susceptibility of rocks in such a field. The apparatus used in the present experiment was a modified Stschodro type[1] ballistic instrument, a general view of which is given in Fig. 2-1. The apparatus[2] in outline is as follows:

The field coil of the apparatus is a circular cylindrical solenoid, 100 *cm* long and 7.0 *cm* in diameter, and the secondary coil also a cylindrical solenoid set coaxially with the field coil, 7.5 *cm* long, 3.23 *cm* and 4.45 *cm* in inner and outer diameters, the number of total turns of the coil being 31000.

In order to eliminate the effect of the earth's magnetic field, the two coaxial soleniods were so set that their axes were perpendicular to the direction of the earth's magnetic force in the laboratory, actually the intensity of component of the earth's magnetic force in the direction of the axes being less than 0.005 *Oersteds*.

The sensitivity of the apparatus depends on both the number of total turns of the secondary coil and the sensitivity of the galvanometer. Actually, a ballistic type galvanometer, about 10^{-11} in current sensitivity and 12 *sec.* in its free oscillation period, was used in its state of critical damping.

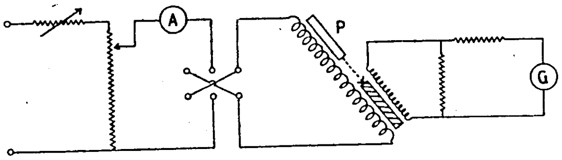
Some of the rock samples to be measured had a circular cylindrical shape, 15 *cm* long and 2.0 *cm* in diameter. These were cut off from the rock mass, while others were pulverized into small particles ranging from 0·3 *mm* to 0·5 *mm*. Since, in the latter case, the mean diameter of the ferro-magnetic minerals, magnetite, hematite, ilmenite, etc. contained in these rocks were less than those of the particles, judging from microscopic examinations of these samples, the crystals composing the minerals were probably unimpaired. Moreover, since the specific intensity of magnetization of a cylindrical specimen was practically

1) N. STSCHODRO, *Gerl. Beitr. Geophys.*, **17** (1927), 148.

2) The apparatus used in the preliminary study (T. NAGATA, *Bull. Earthq. Res. Inst.*, **18** (1940), 102) was reconstructed, a number of parts of apparatus being improved.

the same as that of the pulverized particles of the same sample, in many cases we used the latter form of sample. In this case, a cylindrical tube of bakelite was filled with these particles, the length of the tube being 16 *cm*, which was large enough compared with that of the secondary solenoid. Here, the demagnetizing factor at the centre of the tube was 0.250, whereas if the cylindrical tube were replaced by a rotational ellipsoid of suitable size, the factor was 0.313. Then, since the magnetic susceptibility of rocks examined in the present experiment was always less than 3×10^{-3} e.m.u., the intensity of the effective magnetic field may be taken to be almost the same as that of the applied external magnetic field, provided we neglect an error of the order of 0.1 per cent.

In measuring the intensity of magnetization of rocks induced by the applied magnetic field due to the field coil, the specimen to be measured slides down from position P (in Fig. 2-1) into the secondary coil within the short time of less than 0.5 *sec*.



Fig, I-2 1. Electric circuit of the apparatus for measuring the magnetic susceptibility of volcanic rocks.

Then, the magnetic flux ϕ_s in the space inside the secondary coil, in which a sufficiently long specimen of κ of susceptibility is inserted, is given by

$$\phi_s = AH + 4\pi\kappa HS \tag{2-1}$$

where A, S, and H denote respectively the area of effective cross section of the secondary coil, that of the inserted specimen, and the intensity of the magnetic field.

On the other hand, if the specimen is situated at P in Fig. 2-1, that is, a point on the axis line of the secondary coil as well as of the field coil, and is outside the secondary coil, the magnetic flux passing through a cross section of that part of the secondary coil which is Δr thick and r distant from the centre of the specimen, is given approximated by

$$\phi_r \simeq A\left(H + \frac{2M}{r^3}\right), \tag{2-2}$$

where M denotes the total magnetic moment of the specimen in H, i.e. $M=\kappa HSL$, where L is the length of specimen.

Then, the change in flux passing through the cross-section of that part of the secondary coil owing to displacement of the specimen from P into the secondary coil, is given by

$$\phi_s-\phi_r=4\pi\kappa HS-\frac{2M}{r^3}A=\kappa HS\left(4\pi-\frac{2LA}{r^3}\right). \tag{2-3}$$

The number of turns of secondary coil per cm and the electric resistance in the secondary cercuit being then expressed by n and R respectively, the electric charge Δq induced in the part of secondary coil Δr thick accompanying the change in flux just mentioned is

$$\Delta q=\frac{n\kappa HS}{R}\Delta r\left(4\pi-\frac{2LA}{r^3}\right) \quad \text{(in e.m.u.)} \tag{2-4}$$

Integrating the right hand-side of eq. (2–4) from a to b with respect to r, where a and b are distances of the two ends of the secondary coil from P, the electric charge q induced in the whole of the secondary coil is given by

$$q=\frac{n\kappa HS}{R}\int_a^b\left(4\pi-\frac{2LA}{r^3}\right)dr=\frac{N\kappa HS}{R}\left(4\pi-V\frac{a+b}{a^2b^2}\right), \tag{2-5}$$

where N and V denote respectively the number of total turns in the secondary coil and the volume of the space given by $L\times A$.

Since in the present instrument $a=40\ cm$, $b=47{\cdot}5\ cm$ and $V=110\ cm^3$,

$$V\frac{a+b}{a^2b^2}\Big/4\pi\simeq2\times10^{-4},$$

showing that the second term in eq. (2–5) is negligible compared with the first term.

Thus, the observed deflection of the galvanometer θ is given by

$$\theta=CSJ=CS\kappa H\,, \tag{2-6}$$

in which the intensity of magnetization $J=\kappa H$ can be obtained from θ, where C is a constant substituted for the magnitudes of N and R and for the characterics of the galvanometer. The actual value of C was determined experimentally with the aid of the deflection of the galvanometer θ_0 accompanying the sudden production of magnetic field H_0 due to the field coil, or to its sudden disappearance, in which case

$$\theta_0=CAH_0\,. \tag{2-7}$$

In the actual measurement of J, we adopted not only the method of measuring by placing the sample from P to the secondary coil, but also the reverse operation of suddenly pulling up the specimen, the galvanometer in this case deflecting in the direction opposite to that of the first operation. The arithmetical mean of the readings of these two deflections was adopted as the reading of one measurement, three or five measurements forming a set of observations. Finally, it will be noted here that the lowest limit of the observable value of intensity of magnetization of a specimen in the present apparatus is 1×10^{-5} *e.m.u.*, that is, an error of 10^{-5} in magnitude seems inevitable.

(ii) Results.

The observed values of the specific susceptibilities of various rocks in a magnetic field of from 0 to 19 *Oersteds* are given in Table 2–I and in Fig. 2–2.[3] As will be clear from these results, the specific susceptibility of igneous rocks, generally, increases with increase in magnetic field in the range from zero to 19 *Oersteds*. For a few specimens, susceptibility in a magnetic field smaller than one *Oersted* was especially examined, the result being given in Table 2–II and in Fig. 2–3. From this result, it will be seen that the range of initial susceptibility, where it is constant notwithstanding any change in the magnetic field and is reversible with respect to the magnetic field, is relatively small—say less than 0·5 *Oersteds*.

For convenience, however, we adopted the value of specific susceptibility in the special magnetic field of 1·26 *Oersteds* as representing the magnetic susceptibility in a weak magnetic field, it being denoted by χ_0.

The magnitudes of χ_0 of various rock specimens are given in Table 2–III, where the amounts of Fe_2O_3, FeO, and TiO_2, and the normative amounts of magnetite and ilmenite as calculated from the chemical composition of these rock specimens are also given, both being in weight percentage. Now, plotting the value of χ_0 of each rock against the corresponding value of normative magnetite, we get the result shown in Fig. 2–4, from which it will be seen that, generally speaking, the more normative magnetite, the larger the specific susceptibility, although this law does not exactly hold.

If, then, we assume that χ_0 is approximately proportional to the normative amount of magnetite q_m, it is possible to establish the relation

3) Before measuring the magnetization curve, the rock specimen was always demagnetized by means of alternating magnetic field.

Table I–2–I, a.

Susceptibility of ejecta from Volcano Asama

H \ No.	No. 5	No. 6	No. 8	No. 11	No. 14	No. 15
1·89 *Oe.*	$0{\cdot}84\times10^{-3}$	$0{\cdot}58\times10^{-3}$	$1{\cdot}18\times10^{-3}$	$0{\cdot}169\times10^{-3}$	$0{\cdot}57\times10^{-3}$	$0{\cdot}064\times10^{-3}$
3·77	0·88 〃	0·59 〃	1·21 〃	0·182 〃	0·57 〃	0·068 〃
6·28	0·92 〃	0·59 〃	1·24 〃	0·176 〃	0·57 〃	0·070 〃
8·79	0·95 〃	0·60 〃	1·25 〃	0·176 〃	0·58 〃	0·070 〃
11·30	0·95 〃	0·61 〃	1·26 〃	0·175 〃	0·59 〃	0·069 〃
13·82	0·97 〃	0·63 〃	1·27 〃	0·183 〃	0·58 〃	0·068 〃
16·33	1·00 〃	0·64 〃	1·28 〃	0·183 〃	0·59 〃	0·070 〃
18·84	1·03 〃	0·63 〃	1·29 〃	0·184 〃	0·59 〃	0·071 〃

Table I–2–I, b.

Susceptibility of ejecta from Volcano Huzi.

H \ No.	17	18	19	20	21	22	23	24	25
1·26 *Oe.*	$2{\cdot}25\times10^{-3}$	$1{\cdot}31\times10^{-3}$	$1{\cdot}80\times10^{-3}$	$1{\cdot}28\times10^{-3}$	$1{\cdot}55\times10^{-3}$	$0{\cdot}70\times10^{-3}$	$1{\cdot}29\times10^{-3}$	$0{\cdot}402\times10^{-3}$	$0{\cdot}90\times10^{-3}$
2·51	2·26	1·46	1·97	1·29	1·63	0·719	1·30	0·394	1·04
3·77	3·29	1·54	2·05	1·30	1·66	0·728	1·30	0·397	1 14
5·02	2·31	1·63	2·17	1·31	1·69	0·776	1·31	0 412	1·23
6·28	2·33	1·68	2·23	1·31	1·72	0·79	1·33	0 413	1·30
7·54	2·36	1·72	2·31	1·32	1·73	0·81	—	—	1·33
8·79	2·36	1·75	2·34	1·33	1·74	0·833	1·37	0·411	1 37
11·30	2·38	1·82	2·40	1·34	1·77	0·855	1·39	0·418	1 43
13·82	2·41	1·87	2·46	1·34	1·78	0·859	1·41	0·422	1·49
16·33	2·43	1·91	2·52	1·35	1·80	0·875	1·44	0 427	1·54
18·84	2·44	1·95	2·56	1·36	1·82	0·[illegible]9	1·45	0 431	1·58

Table I-2-I, c.

Susceptibility of ejecta from Volcano Amagi.

H \ No.	27	28	29	30	31	32	33	34
1·26 *Oe.*	1·28×10⁻³	0·775×10⁻³	0·78×10⁻³	0·94×10⁻³	0·56×10⁻³	0·65×10⁻³	0·82×10⁻³	0·62×10⁻³
2·51	1·30	0·837	0·81	0·97	0·575	0·66	0·84	0·68
3·77	1·33	0·896	0·80	0·98	0·575	0·66	0·84	0·70
5·02	1·38	0·922	0·82	0·98	0·59	0·67	0·85	0·73
6·28	1·42	0 969	0·82	0·99	0·595	0 67	0·85	0·75
7·54	1.43	0·980	0·82	0·99	0·60	0·67	0·85	0·76
8·79	1·45	1·005	0·83	0·99	0·61	0·67	0·85	0·77
11·30	1·50	1·035	0·84	0·99	0·61	0·67	0·86	0·79
13·82	1·53	1·058	0·85	0·99	0·61	0·68	0 86	0·81
16·33	1·56	1·082	0·86	1·00	0·615	0·68	0·86	0·83
18·84	1·59	1·106	0·86	1·01	0·615	0·69	0·86	0·84

Table I-2-I, d.

Susceptibility of ejecta from Volcano Mihara.

H \ No.	35	36	37	38	39
1·26 *Oe.*	$0{\cdot}40\times10^{-3}$	$1{\cdot}86\times10^{-3}$	$1{\cdot}46\times10^{-3}$	$1{\cdot}01\times10^{-3}$	0.34×10^{-3}
2·51	0·4	1·92	1·60	1·11	0·38
3·77	0·4	1·97	1·72	1·14	0·39
5·02	0·39	2·02	1·81	1·19	0·42
6·28	0·39	2·03	1·88	1·21	0·43
7·54	0·40	2·05	1·92	1·24	0·45
8·79	0·40	2·08	1·97	1·26	0.47
11·30	0·40	2·12	2·05	1·30	0·49
13·82	0·40	2·15	2·11	1·34	0·51
16·33	0 40	2·17	2 16	1·37	0·53
18·84	0·41	2·19	2·21	1·39	0·56

Table I–2–I, e.

Susceptibility of ejecta from Volcano Usami.

H \ No.	49	50	51	52	53	54	55	56
1·26 *Oe.*	0·835× 10^{-3}	0·95× 10^{-3}	0·37× 10^{-3}	1·09× 10^{-3}	0·566× 10^{-3}	0·847× 10^{-3}	1·14× 10^{-3}	0·677× 10^{-3}
2·51	0·97	1·00	0·43	1·09	0·583	0 877	1·17	0·688
3·77	1·00	0·99	0·46	1·09	0·580	0·877	1·18	0·727
5·02	1 03	1·00	0·47	1·10	0·594	0·899	1·18	0·729
6·28	1·06	1·02	0·49	1·11	0·597	0·902	1·19	0·726
7·54	1·08	1·02	0·51	1·12	0·607	0·914	1·20	0·720
8·79	1·10	1·03	0·52	1·13	0·612	0·920	1·21	0·718
11·30	1·12	1·04	0·55	1·15	0·621	0·932	1·24	0·723
13·82	1·14	1·05	0·57	1·16	0·629	0·944	1·26	0·730
16·33	1·16	1·06	0·59	1·18	0·637	0·955	1·28	0·743
18·84	1·18	1·08	0·61	1·18	0·641	0·968	1·29	0·749

Table I–2–I, f.

Susceptibility of ejecta from Volcano Hakone.

H \ No.	78	79	80	81	82
1·26 *Oe.*	0·76×10^{-3}	1·65×10^{-3}	1·56×10^{-3}	0·698×10^{-3}	0·423×10^{-3}
2·51	0·84	1·7	1·58	0·791	0·452
3·77	0·87	1·83	1·61	0·817	0·449
5·02	0·89	1·87	1·64	0·839	0·465
6·28	0·91	1 89	1·67	0·845	0 473
7·54	0·926	1·91	1·68	0·841	0·490
8·79	0·930	1·93	1·70	0·853	0 500
11·30	0·940	1·95	1·7	0·860	0·516
13·82	0·94	1·97	1·71	0·874	0·527
16·33	0·95	1·99	1·74	0·888	0·538
18·84	0·958	2·01	1·76	0·903	0·549

Table I–2–I, g.

Susceptibility of ejecta from Volcano Taga.

H \ No	91	92	93	94	95	96
1·26 Oe.	$0·88\times10^{-3}$	$0·955\times10^{-3}$	$0·54\times10^{-3}$	$0·78\times10^{-3}$	$1·37\times10^{-3}$	$0·29\times10^{-3}$
2·51	0·92	0·963	0·62	0·80	1·46	0·75
3·77	0·96	0·970	0·64	0·80	1·49	0·81
5·02	1·01	0·975	0·65	0·83	1·54	0·88
6·28	1·05	0·995	0·66	0·84	1·58	0·92
7·54	1·08	0·995	0·67	0·83	1·64	0·95
8·79	1·10	1·00	0·68	0·83	1·65	0·98
11·30	1·18	1·01	0·70	0·84	1·69	1·02
13·82	1·25	1·032	0·71	0·83	1·75	1·07
16·33	1·28	1·04	0·73	0·84	1·77	1·09
18·84	1·30	1·04	0·74	0·85	1·79	1·10

$$\chi_0 = (2{\cdot}43 \pm 0{\cdot}75^*) \times 10^{-2}\, q_{Mt}, \tag{2–8}$$

as shown in Fig. 2–4, the frequency curve of χ_0/q_{Mt} being shown in Fig. 2–5.

Thus, we may presume that the largest part of the magnetic behaviour of igneous rock is due to the presence of ferro-mangetic minerals, especially to that of the magnetite contained in the rock. The actual ferro-magnetic minerals in the rock, however, are believed to consist not only of pure magnetite, but also of titano-magnetite, or a more complex solid solution of magnetite and ilmenite, hematite, and sometimes pyrrhotite, etc., so that the magnitude of normative magnetite calculated from the chemical composition of a rock cannot be expected to give the true actual amount of ferro-magnetic minerals contained in the rock. Moreover, the magnetic behaviour of igneous rocks will be influenced by internal stress and the demagnetizing factor, namely, the shape of the mineral particles.

It is, therefore, natural that the law expressed by eq. (2–8) does not exactly hold in practice. It is possible, however, to take the normative amount of magnetite as a measure of the order of magnetic susceptibility of igneous rocks, at any rate, as a first approximation.

By imagining such a condition in the magnetic behaviour of igneous rock as that in which a large number of microcrystals of ferro-magnetic minerals are uniformly distributed in almost non-magnetic

* Mean deviation.

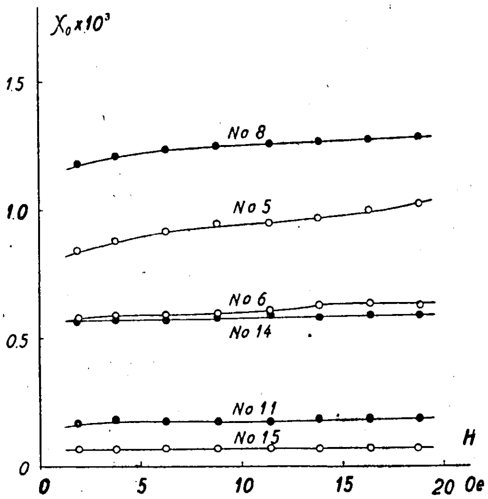


Fig. I-2-2, a. Susceptibility of ejecta from Volcano Asama.

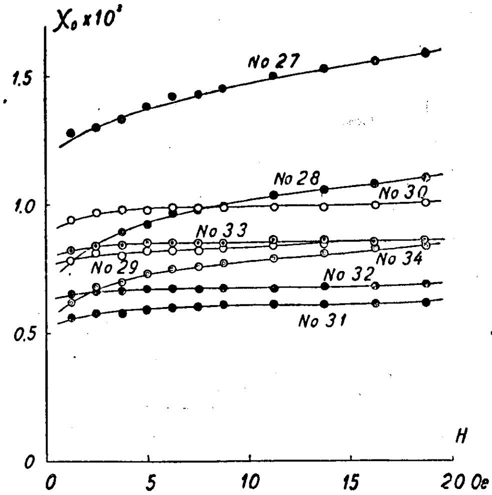


Fig. I-2-2, c. Susceptibility of ejecta from Volcano Amagi.

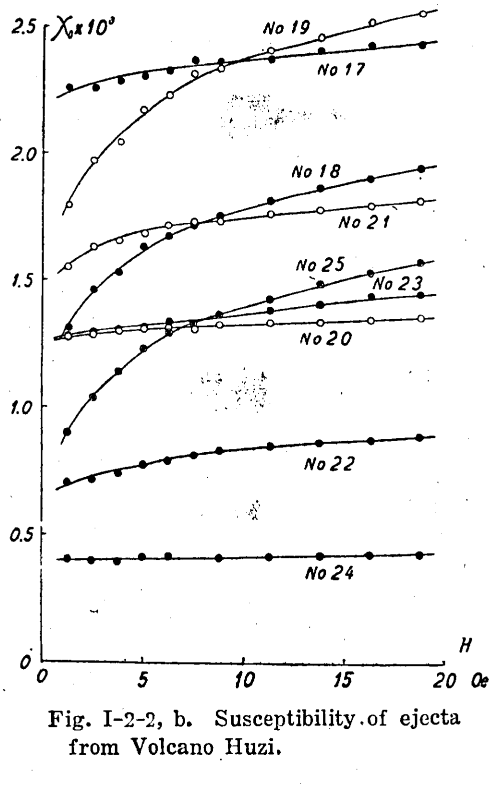


Fig. I-2-2, b. Susceptibility of ejecta from Volcano Huzi.

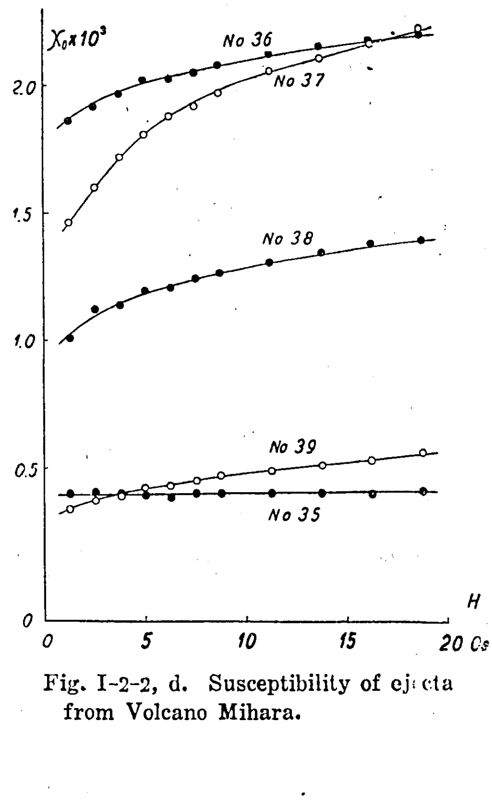


Fig. I-2-2, d. Susceptibility of ejecta from Volcano Mihara.

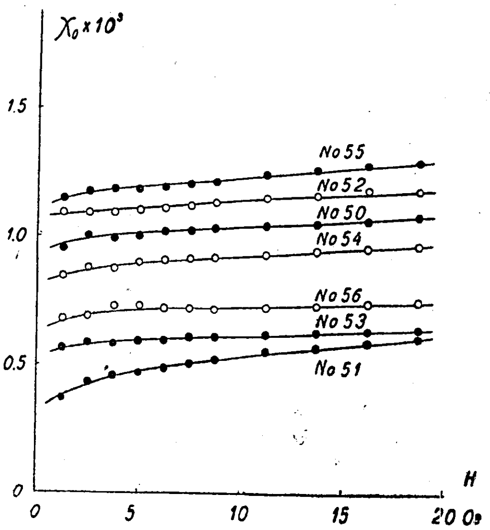


Fig. I-2-2, e. Susceptibility of ejecta from Volcano Usami.

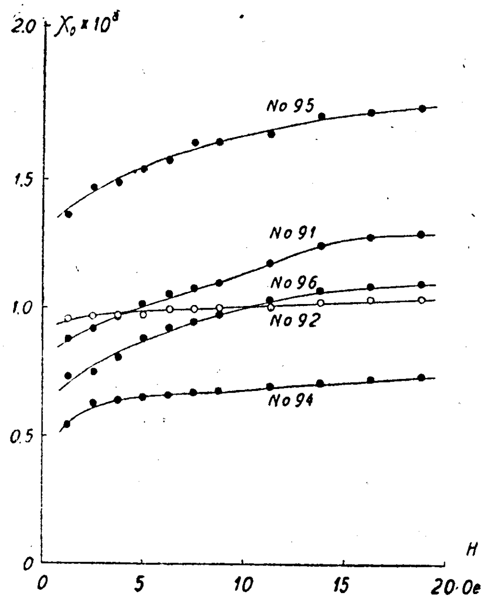


Fig. I-2-2, g. Susceptibility of ejecta from Volcano Taga.

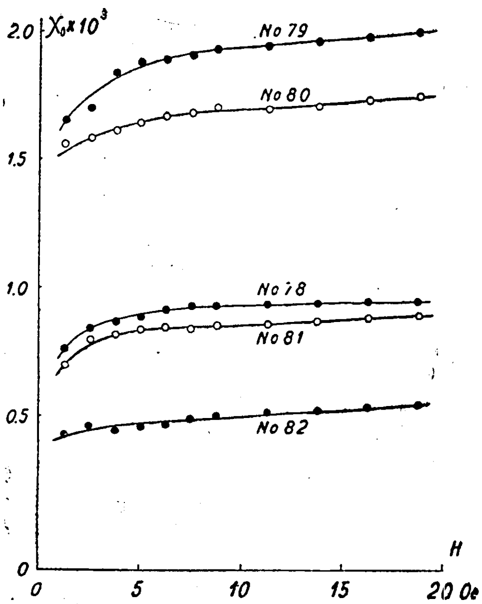


Fig. I-2-2, f. Susceptibility of ejecta from Volcano Hakone.

space, we can calculate theoretically the apparent magnetic susceptibility of rock, given the magnitude of susceptibility and the demanetizing factor of the ferro-magnetic minerals. With the aid of the analogy of the electrostatic polarization of the medium, the effective magnetic field $\vec{H}_e$ in a hollow cavity in the magnetic medium, the susceptibility of which is κ, is given by

$$\vec{H}_e = \vec{H} + \nu \vec{I}\,, \quad \vec{I} = \kappa \vec{H}\,, \qquad (2\text{–}9)$$

where ν denotes the demagnetizing factor depending on the geometrical shape of the cavity.

Table I–2–II.

Susceptibility of volcanic rocks in weak magnetic field.

H \ No	No 20	No 21	No 23	No 27	No 28	No 30
0·13 *Oe.*	1·20	1·19	0·94	1·01	—	0·81
0·25	1·20	1·28	0·94	0·99	0·73	0·81
0·38	1·22	1·38	1·05	1 07	—	0·88
0·50	1·20	1·45	1·08	1·13	0·74	0·88
0·63	1·22	1·50	1·09	1·20	—	0·90
0·75	1·23	1·50	1·12	1·23	0·74	0·91
1·00	1·23	1·54	1·17	1·26	0·76	0·92
1·26	1·25	1·55	1·20	1·28	0·78	0·94
1·57	1·28	1·61	1 21	—	—	—
1·88	1·29	1·62	1·23	1·29	0·80	0·95
2·51	1 31	1·63	1·27	1·30	0·84	0·97
3·14	1·35	1·64	1·29	1·32	0·88	0·97
3·76	1·35	1·66	1 30	1·33	0·89	0·98

If H_i, I_i, κ_m, and λ denote respectively the effective magnetic field, intensity of magnetization, susceptibility, and the demagnetizing factor, each being the respective mean value of the microcrystals of the ferro-magnetic mineral that is present in the hollow cavity, as shown in Fig. 2–6, we get the relation

$$\vec{H}_i=\vec{H}_e-\lambda\vec{I}_i\,, \qquad \vec{I}_i=\kappa_m\vec{H}_i\,. \tag{2–10}$$

The effective magnetic field in the ferro-magnetic mineral is then given by

$$\vec{H}_i=\frac{1+\nu\kappa}{1+\lambda\kappa_m}\vec{H}\,. \tag{2–11}$$

And, since

$$\vec{I}=p\vec{I}_i \tag{2–12}$$

where p denotes the volume content of ferro-magnetic minerals in the rock, we get from eqs. (2–9) and (2–10)

$$\kappa=\frac{p\kappa_m}{1+(\lambda-\nu p)\kappa_m}\,. \tag{2–13}$$

The special case of this formula wherein $\nu=\lambda$ exactly agrees with

Ollendorf's formula[4] for the permeability of dust core, well known in the field of electric engineering.[5]

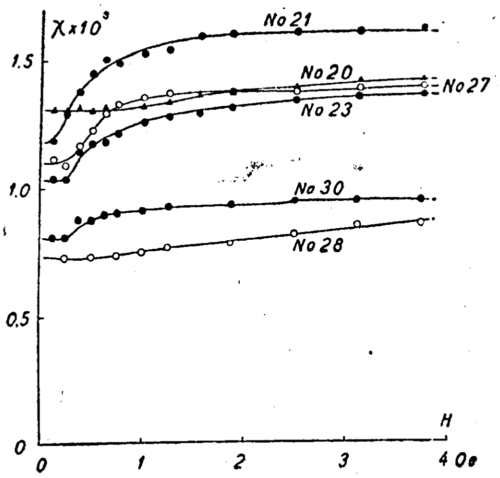


Fig. I-2-3. Magnetic susceptibility of volcanic rocks in weak magnetic field.

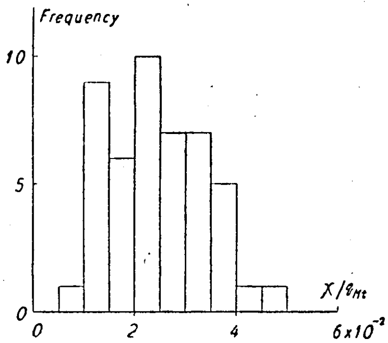


Fig. I-2-5. Frequency of distibilition of χ_0/q_{Mt}

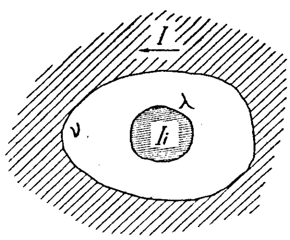


Fig. 1-2-6.

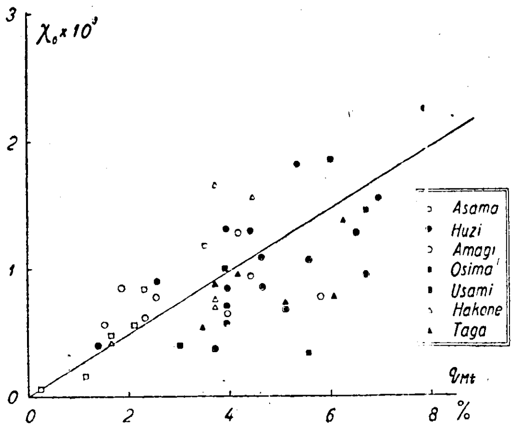


Fig. I 2-4. The relation between specific susceptibility of rocks and the amount of normative magnetite contained in them.

The apparent specific susceptibility of rock, that of a ferro-magnetic mineral and the weight content of the ferro-magnetic mineral in rock being denoted by χ, χ_m, and q respectively, we get, in place of eq. (2-13), the relation

$$\chi=\frac{q\chi_m}{1+(\lambda\rho_m-\nu\rho)\chi_m}, \qquad (2\text{-}14)$$

where ρ and ρ_m denote respectively the density of rock and that of

4) F. OLLENDORF, *Arch. Elek. Tech.*, **25** (1931), 436

5) The Ollendorf's formula was already applied for interpreting magnetization of rock by J. G. KÖNIGSBERGER, (*Beitr. Angew. Geophys.*, **4** (1932), 385.)

the ferro-magnetic mineral. In many actual rocks, however, q is fairly small, amounting at most to only 3∼4 per cent, whence we can put, approximately,

$$\chi \doteqdot \frac{q\chi_m}{1+\lambda\rho_m\chi_m}, \tag{2–14'}$$

Now, we assume an ideal igneous rock, where the ferro-magnetic mineral contained in it is only pure magnetite, the amount of which agrees with the normative amount of magnetite calculated from its chemical composition, and the mean magnitude of the demagnetizing factor of the mineral is always constant regardless of the kind of rock. Then, eq. (2–14') shows that there is a linear relation between the specific susceptibility of rocks and their normative magnetite. That this relation approximately holds was already shown in eq. (2–8).

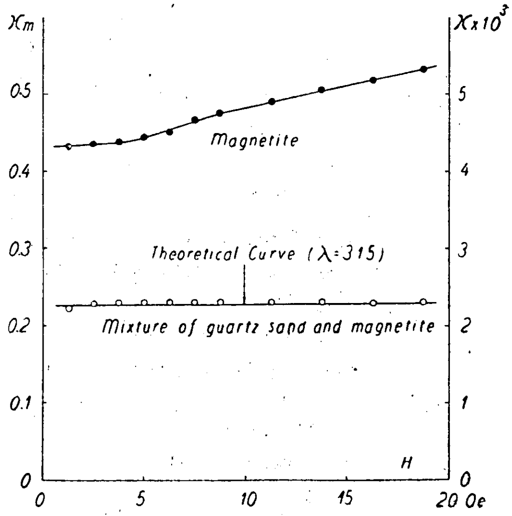


Fig. I–2–7. Susceptibility of magnetite and that of quartz-sand containg a few percent of the magnetite.

In order to prove directly the validity of eq. (2–14), polycrystalic magnetite of 0·43 specific susceptibility was pulverized into small grains, 0·10 *mm* mean diameter, mixed with almost pure silica of like size, and the susceptibility of magnetite and the apparent susceptibility of the mixture were measured in various magnetic fields, an example of the measurement being given in Table 2–IV and in Fig. 2–7, where p in eq. (2–13) was 0·0124 and 0·0250, while ρ_m was 4·8. From this result, we find that eq. (2–14), or at any rate, eq. (2–14'), holds in the present case, so long as we choose a suitable value for the demagnetizing factor of the mean value of the demagnetizing factor of a large number of grains, λ, although the magnitude of ν was not obtained from the present experiment. Actually, the magnitude of λ suitable for the present case was 3·15. Since, on the other hand, the magnitude of λ determined by Königsberger[6] was

6) J. G. Königsberger, *Beitr. Angew. Geophys.*, 4 (1932), 385.

$3{\cdot}0 \sim 4{\cdot}9$, and since, further, in the special case that the mean effect of demagnetization due to the shape of the magnetite particle is assumed to be the same as that of a sphere, λ ought to be $\frac{4\pi}{3} \doteq 4{\cdot}2$, it may be presumed that the λ of the ferro-magnetic minerals in the actual rock is about $3 \sim 4$.

Taking, then, $\lambda = 4{\cdot}0$, $\rho_m = 5{\cdot}2$ in eq. (2–14′), we get from eq. (2–8)

Table I–2–III.

No.	Locality	χ_0	Mt	Fe_2O_3	FeO	TiO_2	χ_0/q_{Mt}
5 □	Asama, Bomb	$0{\cdot}84 \times 10^{-3}$	2·31 %	1·54 %	5·36 %	0·70 %	$3{\cdot}64 \times 10^{-2}$
6	Asama, Temmei lava	0·49	1·62	1·14	4·79	0·66	3·58
8	Asama, Kurohuyama lava	1·18	3·49	2 49	5·38	0·71	3·38
11	Asama, Mugeno tani	0·17	1·16	0·73	1·81	0·41	1·47
14	Asama, Yunotaira lava	0 57	2·08	1·44	5·51	0·70	3·01
15	Asama, Ko-Asama lava	0·064	0·23	2·45	0·45	0·38	2·78
17 •	Huzi, Hôei crater	2·25	7·87	5·40	6·65	1·40	2·86
18	Huzi, Makuiwa 1	1·31	3·94	2·76	7·72	1·41	3·33
19	Huzi, Tawaranotaki	1·82	5·33	3·67	9·90	1·97	3·41
20	Huzi, Makuiwa 2	1·28	6·48	4·50	6·89	1·78	1·98
21	Huzi, Makuiwa 3	1·55	6·95	4·78	5·72	1·38	2·23
22	Huzi, Hôei bomb	0·70	3·94	2·64	8·42	1·38	1·78
23	Huzi, Makuiwa 4	1·29	4·40	3·04	7·29	1·15	2·93
24	Huzi, Karasuiwa	0·40	1·39	0·95	5·02	0·87	2·88
25	Huzi, Aokigahara	0·90	2·55	1·83	8·34	1·43	3·53
27 ○	Amagi, Zizôdô	1·28	4·17	2·89	7·66	0·64	3·07
28	Amagi, Hatikubo-yama	0·78	2·55	1·92	6·77	1·03	3·06
29	Amagi, Yahatano	0·78	5·79	3·94	2·86	0·76	1·35
30	Amagi, Inatori	0·94	4·40	3·03	3·30	0·80	2·04
31	Amagi, Iwano-yama	0·56	1·52	1·12	2·15	0·36	3·68
32	Amagi, Yanagase	0·65	3·94	2·66	3·40	0·65	1·65
33	Amagi, Yahazu-yama	0·85	1·85	1·25	2·32	0·40	4·57
34	Amagi, Akazawa	0·62	2·32	1·54	7·00	0·79	2·66
35 ■	Oosima, Okata	0·40	3·01	2·13	7·05	0·78	1·33
36	Oosima, Somma of Mihara	1·86	6·02	4·18	8·99	0·83	3·09
37	Oosima, Aniei lava	1·46	6·71	4·60	10·02	1·39	2·18
38	Oosima, Meizi-Taisyô lava	1·01	3·91	2·69	10·57	0·74	2·58
39	Oosima, Okata	0·34	5·56	3·82	10·22	0·81	0·61

(to be continued.)

Table I-2-III. (*Continued.*)

No.	Locality	χ_0	Mt	Fe_2O_3	FeO	TiO_2	χ_0/q_{Mt}
49 ◎	Usami	0.84×10^{-3}	3·94 %	2·76 %	6·03 %	0·73 %	2.12×10^{-2}
50	Usami	0·95	6·71	4·60	5·90	1·04	1·42
51	Usami	0·37	3·70	2·61	4·61	0·82	1·00
52	Usami	1·09	4·63	3·18	6·73	0·85	2·36
53	Usami	0·57	3·94	2·79	5·28	0·71	1·44
54	Usami	0·85	4·63	3·25	6·97	0·92	1·83
55	Usami	1·07	5·56	3·18	7·25	0·76	1·92
56	Usami	0·68	5·09	3·48	5·70	0·73	1·34
78 △	Hakone, Hakone-tôge	0·76	3·71	2·45	6·15	0·73	2·05
79	Hakone, Hutago-yama	1·65	3·71	2·59	5·44	0·77	4·45
80	Hakone, Kurakake-yama	1·56	4·18	2·76	6·01	0·61	3·73
81	Hakone, Atami, Iwatoyama	0·70	3·71	2·49	5·59	0·63	2·32
82	Hakone, new-somma	0·42	1·62	1·13	3·86	0·72	2·61
91 ▲	Taga	0·88	3·70	2·54	7·92	0·87	2·38
92	Taga	0·96	4·17	2·86	7·26	0·84	2·30
93	Taga	0·54	3·47	2·42	7·13	0·67	1·56
94	Taga	0·78	6·02	4·14	5·52	0·98	1·30
95	Taga	1·37	6·25	4·36	9·81	0·74	2·19
96	Taga	0·73	5·09	3·48	7·83	0·93	1·43

Table I-2-IV.

H	κ_m	κ (obs.) $p=0.0124$			κ (calc.) $p=0.0124$ $\lambda=3.15$	κ (obs.) $p=0.0250$ $0.3<\delta<0.6$ mm
		$0.1<\delta<0.2$ mm	$0.3<\delta<0.6$ mm	$0.7<\delta<1.0$ mm		
1·26 *Oe.*	0·431	2.12×10^{-3}	2.23×10^{-3}	2.10×10^{-3}	2.27×10^{-3}	0.448×10^{-3}
2·52	0·435	2·30 〃	2·28 〃	2·14 〃	2·27 〃	0·460 〃
3·77	0·437	2·29 〃	2·29 〃	2·10 〃	2·27 〃	0·462 〃
5·02	0·444	2·30 〃	2·29 〃	2·18 〃	2·28 〃	0·463 〃
6·28	0·455	2·30 〃	2·30 〃	2·17 〃	2·28 〃	0·462 〃
7·54	0·469	2·30 〃	2·30 〃	2·19 〃	2·28 〃	0·460 〃
8·59	0·475	2·29 〃	2·30 〃	2·21 〃	2·28 〃	0·462 〃
11·30	0·490	2·30 〃	2·29 〃	2·20 〃	2·29 〃	—
13·82	0·505	2·30 〃	2·30 〃	2·20 〃	2·30 〃	—
16·33	0·518	2·30 〃	2·31 〃	2·21 〃	2·31 〃	—
18·84	0·533	2·31 〃	2·31 〃	2·22 〃	2·32 〃	—

(δ: the diameter of magnetite grains)

$$\chi_m=0.049\,, \quad \text{or} \quad \kappa_m=\rho_m\chi_m=0.26\,.$$

This value of κ_m seems to agree with the mean initial susceptibility of polycrystalline magnetite,[7] which ranges from 0·1 to 3 according to the amount of impurity contained in it.

We may thus conclude that the magnetic susceptibility of volcanic rock is interpreted, in the first approximation, as the apparent susceptibility of such a substance as consists largely of non-magnetic material plus small particles of ferro-magnetic minerals scattered almost uniformly in that non-magnetic material, where the ferro-magnetic minerals are presumed to be nearly magnetite or its solid solution with FeO, Fe_2O_3, $FeTiO_3$, etc.[8]

Supplementary note to § 2

In order to determine the approximate magnetic susceptibility of rock in a comparatively short time, we constructed[9] an apparatus in which the magnetic susceptibility of rock specimen in an alternating magnetic field was measured. The electric circuit of this apparatus is shown in Fig. 2–8. The frequency of the alternating magnetic

Table I–2–V. Magnetic susceptibility of volcanic rocks in weak alternating magnetic field.

H	$\chi \times 10^3$			
	No. 36	No. 37	No. 38	No. 39
0·06 *Oe.*	1·65	1·10	0·78	0·26
0·13	1·65	1·14	0·83	0·26
0·19	1·70	1·21	0·86	0·26
0·25	1·71	1·22	0·88	0·26
0·31	1·73	1·27	0·91	0·27
0·38	1·75	1·25	0·89	0·28
0·44	1·76	1·35	0·90	0·28
0·50	1·83	1·36	0·91	0·29
0·57	1·84	1·35	0·93	0·29
0·63	1·85	1·37	0·92	0·30
0·69	—	—	—	0·31
0·75	1·85	1·38	0·92	0·33
0·88	1·85	1·39	0·94	—
1·01	1·89	1·42	0·92	—
1·26	—	1·43	0·98	—

7) H. REICH, *Handb. Geophys.*, Bd. VI, pp. 62.

8) See § 4, this Chapter, § 3 and § 4 of Chapter III.

9) T. NAGATA, *Bull. Earthq. Res. Inst.*, **18** (1940), 102.
The similar apparatus was made by Y. Kato, a number of igneous rocks being examined with the aid of it.
(Y. KATO, *Journ, Fac. Sci. Tohoku Imp. Univ.*, **27** (1938), 93.)

field was kept at 50 cycles/sec., because the magnifying ratio of the amplifier is nearly constant in the range of frequency around this

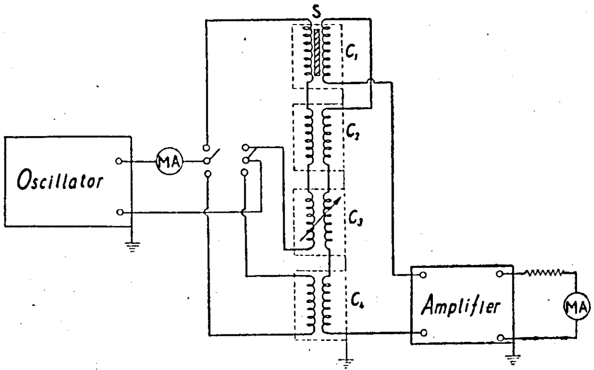


Fig. I-2-8. Apparatus measuring magnetic susceptibility of rocks in alternating magnetic field.

value. A few examples of the observed results are shown in Fig. 2-9.

Although this apparatus seemed to be suitable for determining the approximate magnetic susceptibility of rock in a weak magnetic field,

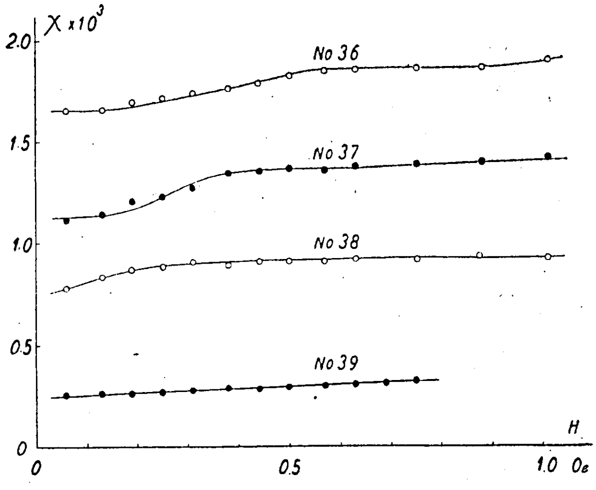


Fig. I-2-9. Susceptibility of volcanic rocks in a weak alternating magnetic field.

there was difficulty in measuring the true susceptibility, probably owing to dielectric loss in various parts of the whole measuring apparatus, especially in C_1, C_2, C_3 and C_4, and to the core loss of S shown in Fig. 2-8.

§ 3. The hysteresis phenomenon in the magnetization curve of volcanic rocks.

In the preceding paragraph, we concluded that the magnetic susceptibility of rock seems to be largely due to the micro-crystals of ferro-magnetic minerals in the rock. Here, the magnetization curve of volcanic rocks will be examined.

(i) First, we have the intensity of magnetization of rock specimen J during the processes of increase of magnetic field H from zero to 18·8 *Oe*, decreasing from +18·8 to −18·8 *Oe*, and increasing from −18·8 to +18·8 *Oe*. with the aid of the same apparatus as that used in § 2. The observed result of the experiment just mentioned shows that all the rock specimens examined showed the hysteresis phonomenon in J-H relation. Here, only the typical examples of observed results will be given Figs. 3–1.[1] From these results, we find that the hysteresis loop corresdonding to the range of change in magnetic field of from −18·8 to +18·8 *Oe* shows the particular form of magnetic behaviour of each rock specimen, that is, some rocks are magnetically "hard", while others are "soft". This fact may show that the internal stress in the ferro-magnetic minerals is not uniquely given. Since, however, the hysteresis curves shown in Fig. 3–1 are not complete, they being only hysteresis loops in the range of magnetic field of from about −20 to +20 *Oe*., where the magnetization of rock is far below its saturation, quantitative examination of the magnetization curve cannot be attempted.

Here, we shall adopt a measure of "hardness" of the rock's magnetism merely for the sake of convenience, that is, the coersive force H_c and the ratio of remanent magnetization J_r to initial susceptibility, where J_r and H_c are defined in the above-mentioned hysteresis loop analogously in the case of a complete hysteresis curve. The actual values of these various rocks are given in Table 3–1, where $J_r/H\chi$ (H=1·26 *Oe*.) and H_c are given.

(ii) Next, the magnetization curves of some rock specimen were obtained in a magnetic field of from zero to 220 *Oe*. In this measurement, the ballistic method, exactly the same in principle as that mentioned in the foregoing paragraph, was used. It consists of two cylindrical solenoids, the field coil, 60 *cm* long and 5·0 *cm* in its

1) A number of other examples were given in the writer's previous paper. (T. NAGATA, *Bull. Earthq. Res. Inst.*, **18** (1940), 112.)

Table I–3–I.

No.	$Jr/J_{1\cdot3}$	Hc	No.	$Jr/J_{1\cdot3}$	Hc
5	0·98	1·13 *Oe.*	39	2·76	2·00 *Oe.*
6	0·87	1·01	49	1·80	1·76
8	0·58	0·75	50	0·79	1·00
11	0·68	0·88	51	3·44	2·51
14	0·36	0·50	52	0·94	1·16
17	0·58	0·78	53	1·23	1·44
18	2·50	2·05	54	1·41	1·55
19	2·23	1·97	55	1·03	1·02
20	0·61	0·68	56	0·94	1·13
21	1·11	1·13	58	2·15	1·91
22	1·85	1·88	73	0·81	1·01
23	1·17	1·26	74	2·66	2·51
24	1·06	1·13	75	1·94	1·91
25	4·08	3·03	76	1·61	1·63
26	0·76	1·03	77	2·61	2·07
27	1·94	2·01	78	1·12	1·14
28	2·13	1·93	79	0·91	0·88
29	0·86	1·04	80	0·40	0·53
30	0·41	0·48	81	0·78	0·79
31	0·43	0·50	82	1·47	1·48
32	0·81	0·98	91	1·82	1·63
33	0·49	0·65	92	1·57	1·88
34	1·99	1·88	93	2·33	1·95
35	0·74	1·21	94	0·47	0·59
36	1·60	1·76	95	2·29	2·24
37	2·85	2·42	96	3·13	2·51
38	2·29	2·14			

inner diameter and 360/*cm* in turn of solenoid, and the secondery coil, 6·0 *cm* long, 2·0 *cm* in its inner diameter and 10000 total turns. In the present apparatus, however, the water cooling equipment was set between the field and the secondary coils in order that the heat produced by the electric current through the field coil shall be absorbed. The specimens examined were circular cylinders 10 *cm* long and 1·5 *cm* in diameter.

The observed hysteresis curves in the range of the applied magnetic field of from -220 *Oe.* to $+220$ *Oe.* are given in Fig. 3–2. As will be seen from these results, although the magnetization and demagnetization curves of the examined rock specimens show a marked hysteresis phenomenon, magnetization of the rock is not yet

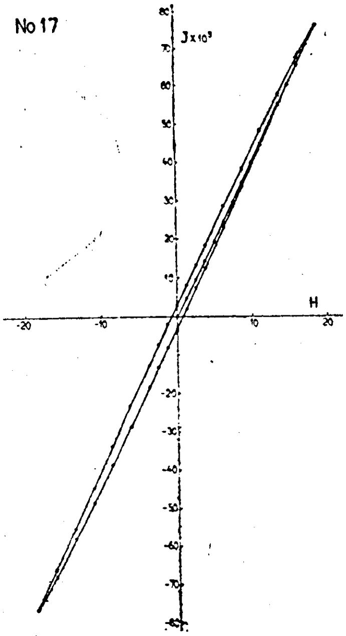


Fig. I-3-1. (a) Specimen No 17.

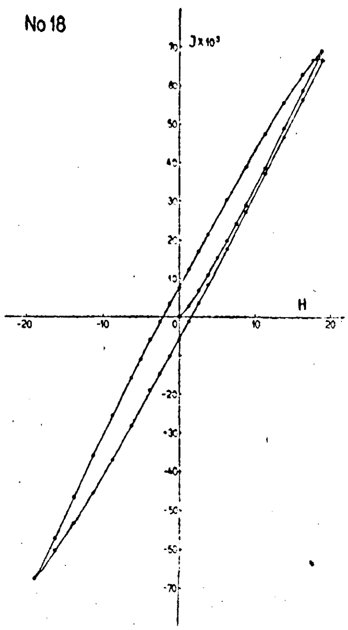


Fig. I-3-1. (b) Specimen No 18.

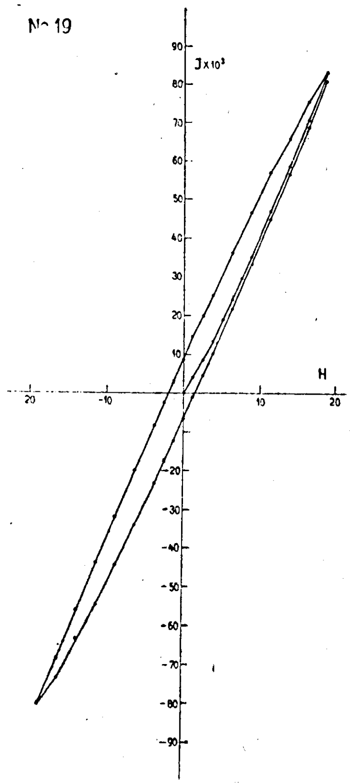


Fig. I-3-1. (c) Specimen No 19.

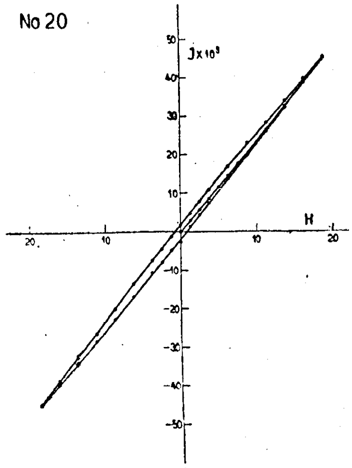


Fig. I 3-1. (d) Specimen No 20.

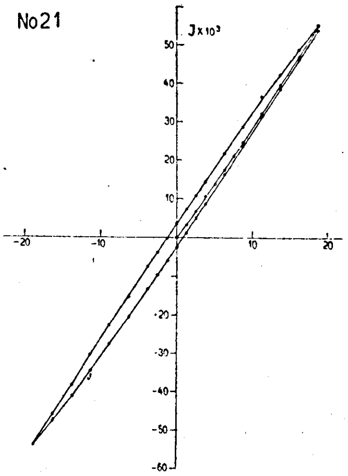


Fig. I-3-1. (e) Specimen No 21.

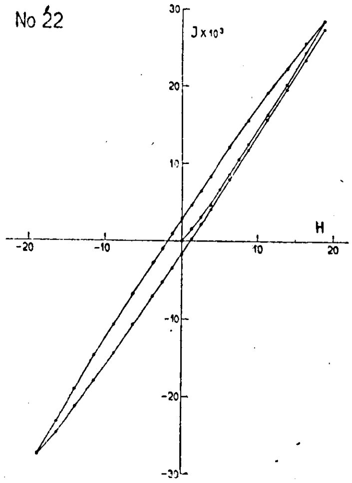


Fig. I-3-1. (f) Specimen No 22.

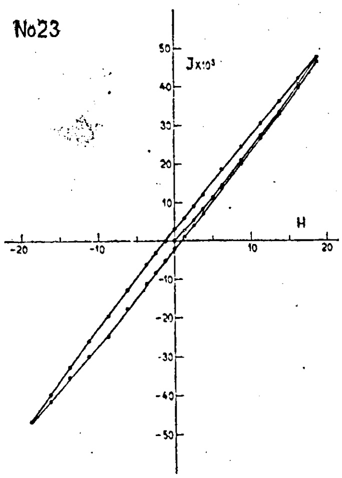


Fig. I-3-1. (g) Specimen No 23.

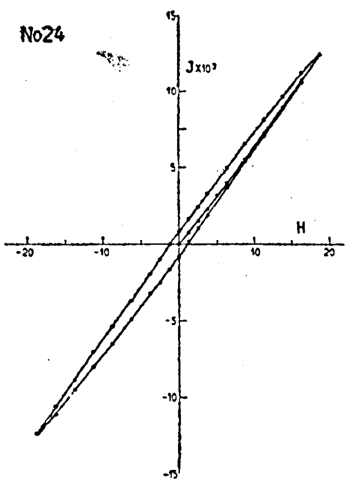


Fig. I-3-1. (h) Specimen No 24.

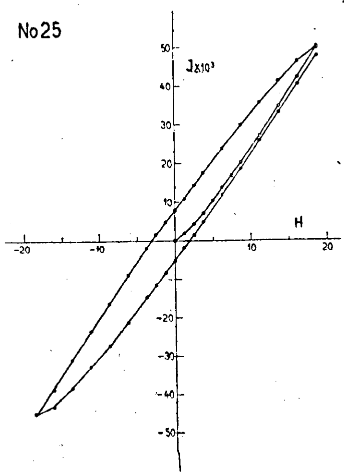


Fig. I-3-1. (i) Specimen No 25.

saturated even at the largest value of applied magnetic field, say 220 *Oe.*[2]

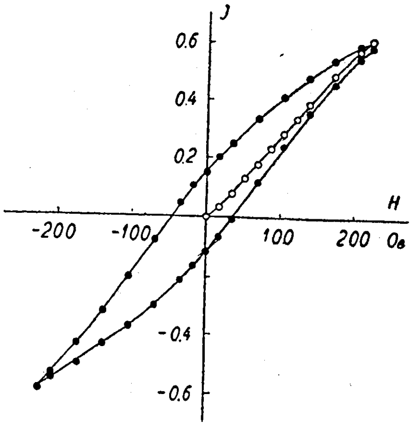


Fig. I-3-2. (a) Specimen No 19.

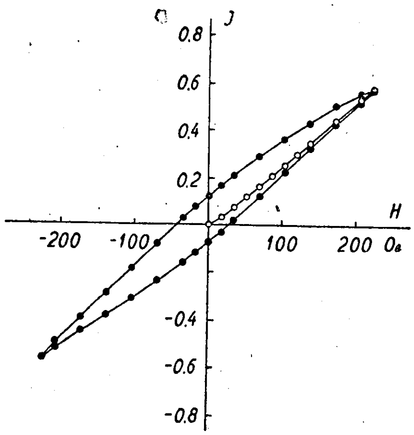


Fig. I-3-2. (b) Specimen No 21.

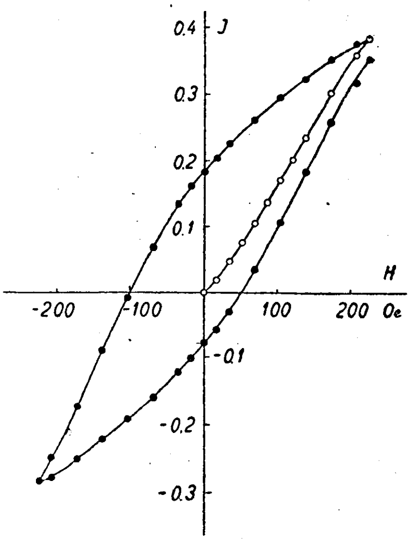


Fig. I-3-2. (c) Specimen No 22.

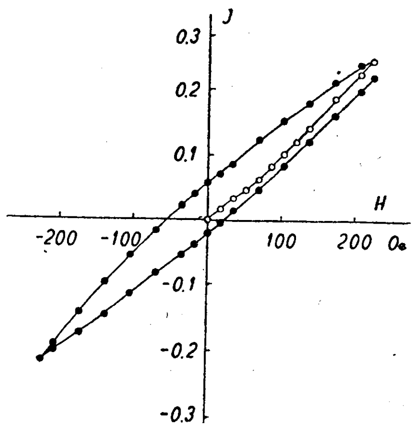


Fig. I-3-2 (d) Specimen No 28.

2) If we assume that the ferro-magnetic mineral in the rock is magnetite alone, we may expect the intensity of applied magnetic field, where the magnetization of rock is saturated, to be 3000 *Oe*, for since

$$H_e \simeq H - \lambda I_i \,, \tag{2-10'}$$

the intensity of saturated magnetization of magnetite=590 Gauss, the value of H in which the magnetization of magnetite is almost saturated=600 *Oe.* and

$$\lambda \doteqdot 4 \,,$$

so that the amount of H in which the magnetization of the rock is saturated must exceed 2800 *Oe.*

Comparing now the hysteresis curves given in Fig. 3–2 with the corresponding curves given in Fig. 3–1, we find that the amount of hysteresis loss in the $-220 \sim +220$ *Oe.* magnetization curve is not always proportional to that in the $-19 \sim +19$ *Oe.* magnetization curve, whence it may be said that the magnetic hardness given by Figs. 3–1, 3–2, and the quantities $J_r/H\chi$, H_c given in Table 3–1 is not necessarily trustworthy. It has a definite physical meaning, showing only the general approximate tendency of the hysteresis phenomenon of rock in a relatively weak magnetic field. Further, hardly any successful result was obtained in the search for a definite relation between the chemical composition and the measures of magnetic hardness defined in the present paragraph, for which reason it will be presumed that the magnetization and hysteresis curves in the range of magnetic field up to the saturation of the rock's magnetization alone can give indicators for the magnetic behaviour of ferro-magnetic minerals in rocks,[3] the results obtained in this paragraph merely supporting the conclusion obtained in the foregoing paragraph that the magnetic behaviour of volcanic rocks is largely due to the ferro-magnetic minerals contained in them.

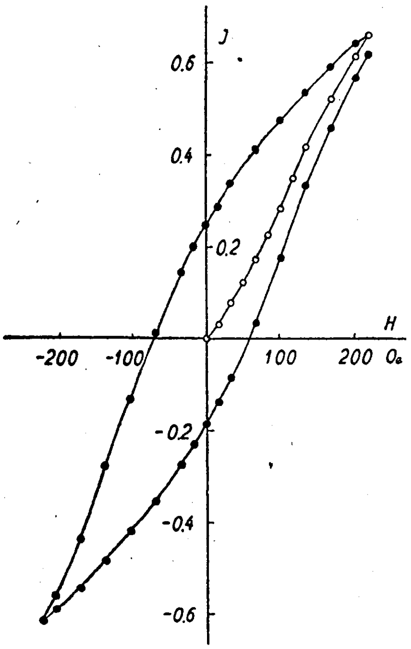


Fig. I-3-2. (e) Specimen No 75.

§ 4. Change in magnetic susceptibility with temperature.

(i) The change in magnetic susceptibility of various rocks with temperature in a weak magnetic field were measured by means of a

3) Hence, it will be most required to determine the complete hysteresis curve of volcanic rocks. Until the present, we have nothing of complete hysteresis curves of volcanic rocks, although the magnetization curve in the magnetic field less than 500 *Oe.* have been fuquently reported. (For example, TZU-CHANG-WANG, *ZS. Geophys.*, 16 (1940) 160.)

However, an apparatus for measuring the magnetization of rock in the field of from zero to 5000 *Oe* is now being constructed in the writer's laboratory. The writer hopes that some fruitful results are obtained in future by means of this apparatus

Adding note. (Dec. 20, 1942). The magnetization curve of various volcanic rocks in $0 \sim 5000$ *Oe.* has recently been obtained. According to the experimental result, the rocks magnetization is almost saturated at the magnetic field of $3000 \sim 4000$ *Oe.* The details of this study will shortly be published.

ballistic method. Since the magnetic susceptibility is usually very small (less than 3×10^{-3} in specific value), several technical improvements were required in order to determine the magnetic susceptibility in a weak magnetic field with an error of less than a few per cent.

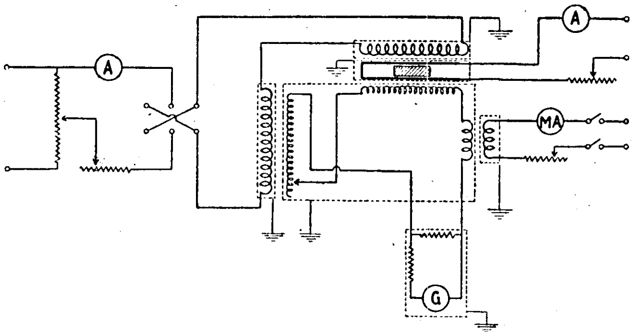


Fig. I-4-1. Electric circuit in the apparatus for measuring the magnetic susceptibility of rock at any temperature.

Details of the measuring apparatus having already been given by the writer in another paper,[1] only an outline of it will follow here. The general electric circuit in the ballistic method is shown in Fig. 4-1, and a schematic view of the main part of the apparatus shown in Fig. 4-2. The primary coils of mutual inductance M_1 and M_2 are Helmholtz coils of 35 *cm* radius, the intensity of the magnetic field at the centre of the coil being 3·85 *Oe.* corresponding to unit electric current through the coil. The secondary coils in M_1 and M_2 are cylindrical solenoids set coaxially with the Helmholtz coil, 9·0 *cm* long and 6·7 *cm* in inner diameter, the total number of turns of wire in coil M_1 being 43000, while that in M_2 may be varied from 40000 to 46000 with every two turns, so that the mutual inductance of M_2 can be made equal to that of M_1 with a deviation of about only 2/43000 $\simeq 5\times10^{-5}$. A vacuum electric furnace, which was set coaxially with M_1, was

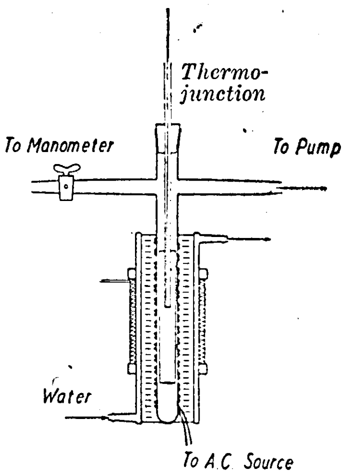


Fig. I-4-2. The schematic view of main part of apparatus for measuring the magnetic susceptibility of rock at any temperature.

1) T. NAGATA, *Bull. Earthq. Res. Inst.*, 19 (1941), 579.

made of fuzed silica and platinum wire, which last was wound non-inductively on a silica tube. The rock specimen to be examined was pulverized into small particles of 0·1 *mm*, mean diameter, and poured into a tube of fuzed silica which was set as shown in Fig. 4–2, where the temperature at the centre of the specimen was determined by means of Pt-Pt.Rh thermo-junction. In short, every part of the apparatus was sufficiently non-magnetic, bearing a small error of the order of 10^{-5} *e.m.u.*.

The electric current I through the primary coils of M_1 and M_2 being turned to $-I$, the electric charge Q due to electro-magnetic induction between the two coils of M_1, containg a rock specimen as a core, and between those of M_2 could be measured by means of a ballistic galvanometer, where Q is given by

$$Q=2kI(M_1\mu-M_2\mu_0)$$
$$\mu=1+4\pi CS\kappa/A\,, \qquad \mu_0=1\,, \tag{4-1}$$

where k, S, κ, A and C denote respectively a constant, depending on the resistance of the secondary circuit, a cross section of the inserted specimen, its magnetic susceptibility, effective cross section of the secondary coil of M_1, and a constant depending on the dimension of the specimen with respect to that of the secondary coil, that is, $CS\kappa/A$ gives the effective susceptibility of the entire space inside the secondary coil of M_1.

Since $M_1 \simeq M_2$, we finally get

$$\kappa=\frac{\theta}{KI}-\varepsilon=\frac{\theta}{K'H}-\varepsilon\,, \tag{4-2}$$

where θ, K, K', H, and ε denote the deflection of the ballistic galvanometer, the constants depending on the dimensions of the instrument as well as of the specimen, the intensity of the magnetic field corresponding to I through the primary coils, and a constant due to the inequality when M_2 is compared with M_1.

In practice, the values of constants K and ε were determined experimentally for each specimen at room temperature, where the κ of these specimens at room temperature was accurately determined by another instrument.

The largest source of observational error was the fluctuation in the value of θ, which was probably due to a slight leakage from the alternating current in the electric furnace or from some other source to the secondary circuit. This fluctuation depends on the temperature in the furnace, the higher the temperature the greater the fluctuation.

The mean of the observed values of $\dfrac{\theta}{K'H}-\varepsilon$ at various temperatures, where $\mu=\mu_0$, are

$$\text{temperature}=\ 0\sim350°C,\qquad \kappa=(0{\cdot}6\pm1{\cdot}6)\times10^{-5},$$
$$400\sim700°C,\qquad \kappa=(0{\cdot}0\pm2{\cdot}7)\times10^{-5},$$

where $H=2{\cdot}2$ *Oersteds*, while in determining κ from the values of Q and H, K' was assumed to be the average value of its magnitude determined for a number of rock specimens. Thus, the observational error in determining κ in a magnetic field of about 2 *Oersteds* at various temperatures of from $0°C$ to $700°C$ would always be less than 3×10^{-5} *e.m.u.*.

Since, on the other hand, the mean deviation of temperature in the specimen from its mean value during heating or cooling at the rate of 100 degrees per hour are such as shown in the following table,

Mean value of temp.	110·°6C	240·°9	425·°6	585·°3
Mean deviation	±0·°6	±3·°1	±5·°2	±6·°5

we may say that the temperature of a specimen was uniform with an error of less than 10° in a temperature range of from $0°C$ to $650°C$.

(ii) Several typical specimens of basaltic and andesitic rocks were examined with the aid of the present apparatus. The observed magnetic susceptibilities of these rocks are various temperatures from $20°C$ to $660°C$ are given in Table 4–I, and in Figs. 4–3∼4–14, where the scale of the ordinate gives the specific susceptibility, (I) and (II) denote the heating and cooling processes respectively, while (III) and (IV) in Fig. 4–5 show the second heating and cooling processes respectively.

As will be seen from these figures, the susceptibility of all rock specimens examined becomes zero when the temperature exceeds 580°C or 600°C, provided we neglect small quantities less than the observational error——say 3×10^{-5} *e.m.u.* The fact that the magnetism of a rock disappeared when its temperature exceeded a certain critical value is the most positive proof for concluding that almost the whole of the magnetism of rock is of ferro-magnetic character, whence the critical temperature at 580°C∼600°C in the case of rock seems to be the magnetic transition point (Curie-temperature) in usual ferromagnetic meterial.[2]

2) Actually, the intensity of magnetization at temperature higher than the critical point cannot be estimated, from the reason that its magnitude would be much smaller than the observational error.

Table I–4–I a.

Temp.	No. 17		No. 18		No. 19		No. 20 (1)		No. 20 (2)		No. 21	
	H	C	H	C	H	C	H_1	C_1	H_2	C_2	H	C
20°∼25°C	2·25× 10^{-3}	2·25× 10^{-3}	1·42× 10^{-3}	1·41× 10^{-3}	1·86× 10^{-3}	1·77× 10^{-3}	1·29× 10^{-3}	1·27× 10^{-3}	1·28× 10^{-3}	1·27× 10^{-3}	1·62× 10^{-3}	1·63× 10^{-3}
40	2·28	2·25	1·44	1·45	1·97	1·83	1·29		1·28		1·62	1·64
60	2·28	2·26	1·48	1·48	2·17	1·93	1·29		1·30	1·26	1·61	1·65
80	2·31	2·28	1·57	1·50	2·32	2·05	1·31		1·33		1·66	1·66
100	2·31	2·30	1·63	1·59	2·53	2·13	1·32		1·31		1·70	1·67
120	2·34	2·32	1·79	1·76	2·69	2·25	1·32		1·32	1·27	1·71	1·69
140	2·38	2·35	1·87	1·83	2·86	2·40	1·36		1·31		1·74	1·70
160	2·44	2·36	1·94	1·88	2·92	2·54	1·36		1·30		1·75	1·70
180	2·47	2·39	2·03	1·93	2·95	2·53	1·37		1·31	1·29	1·77	1·72
200	2·51	2·41	2·16	1·97	2·69	2·41	1·37		1·31		1·78	1·73
220	2·52	2·42	2·16	1·89	2·38	2·11	1·42	1·32	1·33	1·29	1·82	1·75
240	2·57	2·46	1·98	1·74	2·00	1·79	1·41		1·31		1·84	1·76
260	2·60	2·44	1·79	1·46	1·63	1·51	1·43		1·31		1·87	1·77
280	2·62	2·47	1·43	1·15	1·38	1·20	1·41		1·30		1·87	1·78
300	2·64	2·47	0·98	0·79	1·27	1·01	1·44	1·34	1·30	1·29	1·91	1·78
320	2·63	2·42	0·57	0·58	1·19	0·99	1·47	1·35	1·30	1·30	1·93	1·78
340	2·48	2·32	0·54	0·49	1·14	0·93	1·44	1·35	1·29	1·30	1·95	1·79
360	2·32	2·25	0·51	0·46	1·11	0·86	1·47	1·35	1·29	1·31	1·98	1·80
380	2·22	2·10	0·50	0·42	1·05	0·83	1·47	1·34	1·27	1·28	2·02	1·82
400	2·14	1·93	0·50	0·41	0·96	0·78	1·49	1·30	1·28	1·29	2·03	1·82
420	1·96	1·57	0·46	0·39	0·89	0·70	1·47	1·29	1·26	1·27	2·04	1·77
440	1·40	1·18	0·40	0·39	0·79	0·63	1·37	1·20	1·19	1·23	2·03	1·62
460	0·90	0·80	0·38	0·36	0·70	0·60	1·20	1·03	1·06	1·06	1·89	1·51
480	0·57	0·53	0·35	0·33	0·57	0·48	0·99	0·85	0·65	0·79	1·61	1·25
500	0·36	0·31	0·30	0·28	0·48	0·41	0·60	0·60	0·41	0·57	1·21	0·92
520	0·15	0·17	4·22	0·18	0·38	0·30	0·41	0·39	0·30	0·38	0·78	0·58
540	0·11	0·13	0·13	0·15	0·19	0·16	0·23	0·22	0·16	0·20	0·37	0·34
560	0·05	0·04	0·05	0·06	0·10	0·08	0·11	0·11	0·07	0·08	0·21	0·19
580	0·00	−0·01	−0·01	0·01	0·01	0·00	0·00	0·01	0·01	0·01	0·05	0·07
600	0·00	0·00	0·00	0·00	0·00	0·00	0·01	0·02	−0·01	0·00	0·00	0·02
620	0·01	−0·01	0·00	−0·01	0·00	0·00	0·00	0·00	0·00	0·00	0·00	0·00
640	0·00	0·00	0·00	−0·01	0·00	0·00	−0·02	−0·01	−0·01	−0·01	0·00	0·01
660	0·00	0·00	−0·01	0·00	−0·01		−0·01	−0·01	−0·01	−0·01	0·00	0·00

It will be noticed, however, that the $\chi(t)\sim t$ curves of the rock specimens are not always so simple as those of the usual ferromagnetic material, the susceptibility, sometimes, changing stepwise with temperature as shown in Fig. 4–7∼4–14, for example. In some of the specimens, the thermal change in susceptibility has a fairly simple

Table I-4-I b.

Temp.	No. 22		No. 23		No. 24		No. 25		No. 60′ (H=2·09 Oe.)	
	H	C	H	C	H	C	H	C	H	C
20°∼25°C	0·72×10⁻³	0·70×10⁻³	1·30×10⁻³	1·28×10⁻³	0·40×10⁻³	6·40×10⁻³	0·94×10⁻³	0·95×10⁻³	1·12×10⁻³	1·13×10⁻³
40	0·75	0·74	1·30	1·29	0·43	0·43	1·15	1·10	1·15	1·17
60	0·80	0·78	1·28	1·29	0·46	0·47	1·42	1·39	1·17	1·20
80	0·87	0·81	1·32	1·30	0·49	0·49	1·66	1·63	1·22	1·25
100	0·91	0·91	1·32	1·31	0·52	0·53	1·93	1·77	1·25	1·29
120	0·96	0·96	1·35	1·31	0·55	0·57	2·06	1·91	1·29	1·34
140	0·01	1·01	1·37	1·32	0·60	0·62	2·14	1·96	1·34	1·39
160	1·09	1·03	1·37	1·32	0·63	0·65	2·11	1·89	1·39	1·43
180	1·10	1·05	1·38	1·32	0·67	0·66	1·72	1·66	1·45	1·49
200	1·11	1·03	1·40	1·34	0·69	0·67	1·29	1·26	1·51	1·55
220	1·02	1·02	1·41	1·36	0·56	0·56	0·97	1·08	1·53	1·60
240	0·93	0·93	1·42	1·37	0·36	0·37	0·81	0·82	1·56	1·63
260	0·80	0·77	1·44	1·39	0·28	0·25	0·68	0·62	1·57	1·63
280	0·68	0·64	1·46	1·39	0·21	0·18	0·61	0·47	1·53	1·59
300	0·55	0·52	1·49	1·38	0·15	0·13	0·53	0·44	1·47	1·52
320	0·49	0·45	1·51	1·39	0·13	0·13	0·50	0·43	1·39	1·45
340	0·41	0·38	1·50	1·37	0·12	0·12	0·49	0·42	1·30	1·34
360	0·40	0·35	1·52	1·38	0·12	0·13	0·48	0·42	1·15	1·19
380	0·37	0·33	1·52	1·38	0·13	0·12	0·48	0·40	1·03	1·06
400	0·35	0·30	1·54	1·34	0·11	0·12	0·48	0·40	0·94	0·92
420	0·33	0·28	1·49	1·28	0·12	0·12	0·47	0·39	0·83	0·79
440	0·30	0·25	1·34	1·13	0·12	0·12	0·46	0·39	0·68	0·64
460	0·23	0·21	1·05	0·96	0·12	0·12	0·43	0·38	0·53	0·52
480	0·21	0·18	0·83	0·71	0·11	0·12	0·39	0·36	0·41	0·39
500	0·18	0·15	0·[illegible]4	0·42	0·11	0·10	0·34	0·34	0·31	0·31
520	0·14	0·11	0·34	0·25	0·09	0·10	0·31	0·28	0·21	0·20
540	0·06	0·07	0·21	0·15	0·06	0·06	0·21	0·18	0·12	0·12
560	0·02	0·03	0·09	0·07	0·03	0·03	0·10	0·09	0·05	0·06
580	0·01	−0·01	0·03	0·03	0·01	0·00	0·01	0·01	0·02	0·04
600	0·00	0·01	0·00	−0·01	0·00	0·00	0·00	0·00	−0·01	0
620	0·00	0·00	0·00	0·00	0·00	−0·01	0·00	−0·01	0	0·01
640	−0·01	0·00	0·00	−0·01	−0·01	0·01	−0·01	0·00	0·01	−0·01
660	0·01	0·00	−0·01	−0·01	0·01		0·00		0	

character, like those given in Fig. 4–3∼4–8, in which $\chi(t)$ slowly increases with temperature from $0°C$ to about $400°C$, whence it decreases rather abruptly. Further, comparing the $\chi(t)\sim t$ relations of these specimens in the heating processes, they are nearly similar, excepting a slight difference of that the susceptibility in cooling is usually a little less than that in the heating, that is to say, the change in susceptibility of these rocks is nearly reversible with respect to temperature. This type of change in susceptibility, as just mentioned, we shall call "the standard type" of rock magnetism, so that in the $\chi(t) \sim t$ relation of the standard type rock specimen, the critical temperature at which the ferro-magnetism of a rock specimen is believed to disappear almost agrees with the Curie-temperature of pure magnetite, namely, 585°C. This critical temperature will be called here the apparent Curie-point, θ_a. However, since the temperature at which the rate of change in susceptibility with respect to temperature is

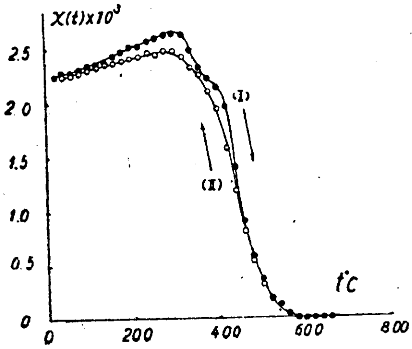


Fig. I-4-3. Specimen No. 17.

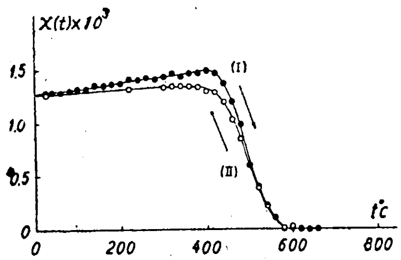


Fig. I 4-4. Specimen No. 20.

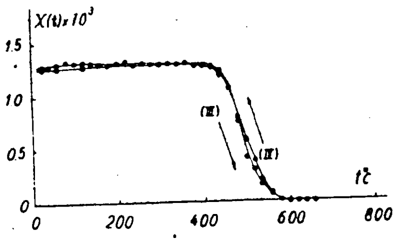


Fig. I 4-5. Specimen No. 20,

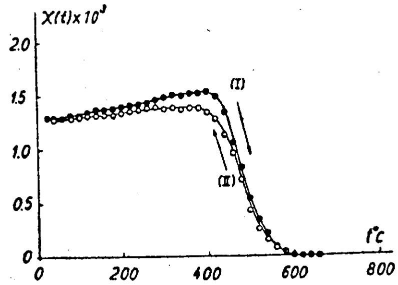


Fig. I-4-6. Specimen No. 21.

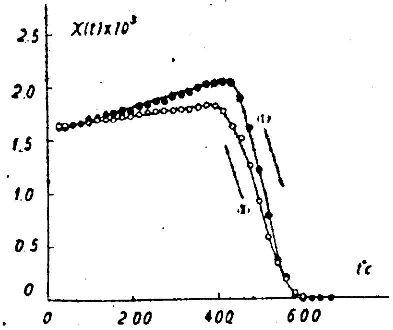


Fig. I-4-7. Specimen No. 23.

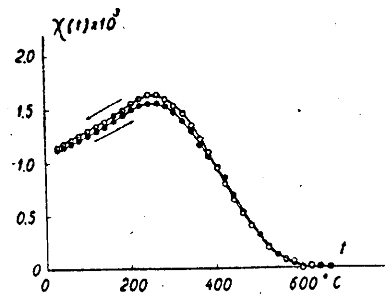


Fig. I-4 8. Specimen No. 60′,

maximum does not agree with the critical point, the former usually being slightly lower than the latter, by taking into consideration that the ferro-magnetic minerals in the rocks are composed not only of pure-magnetite, but also of solid solutions of Fe_2O_3, FeO, TiO_2 and other elements in various phases, it may be presumed that the magnetic transition (Ferro$\rightleftarrows$Para) of the rock will not be unique, but, instead, a superposition of many transitions at various different temperatures that are distributed, macroscopically speaking, almost continuously in a finite range of temperature.

In other words, by letting x and $f(x)$ respectively denote a measure

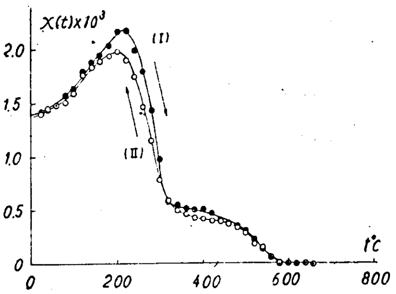


Fig. I-4-9. Specimen No. 18.

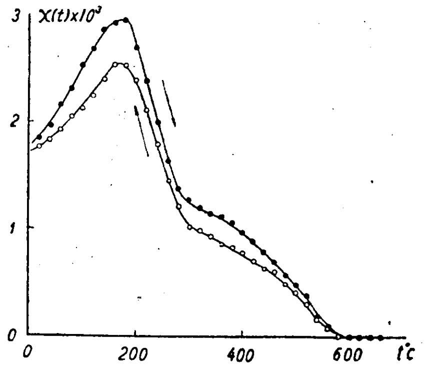


Fig. I-4-10. Specimen No. 19.

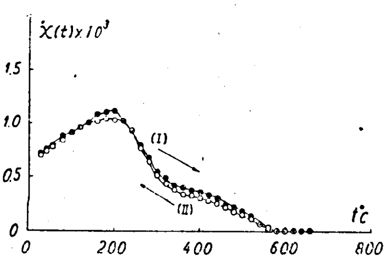


Fig. I 4 11. Specimen No. 22.

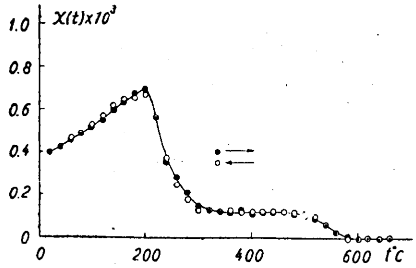


Fig. I-4-12. Specimen No. 24.

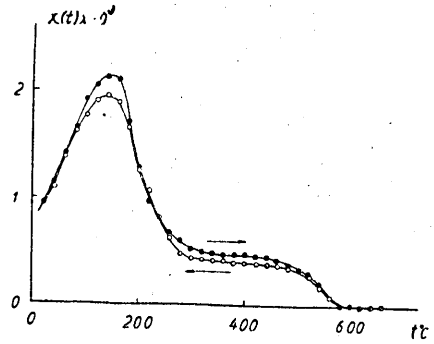


Fig. I-4-13. Specimen No. 25.

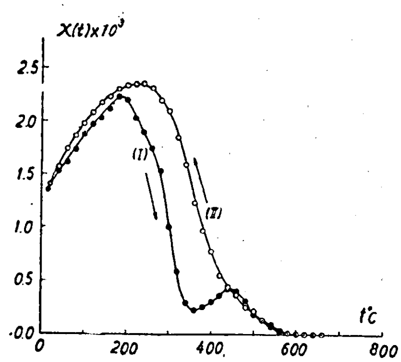


Fig. I-4-14. Specimen No. 59.

of the chemical composition of the ferro-magnetic minerals and the distribution function of the mineral that has x, the apparent mode of change in susceptibility with temerature is given by

$$\chi(t)=\int_{x_1}^{x_2} f(x)\chi(x,t)\,dx\,, \tag{4-3}$$

where $\chi(x,t)$ denotes the susceptibility of an elemental micro-crystal of a ferro-magnetic mineral, being a function of x as well as of t, while x_1, x_2 are the limiting values of possible distribution of x. Here, the Curie-point corresponding to $\chi(x,t)$ is assumed to be uniquely determined as a function of x.[3]

We then take here the temperature at which

$$-\frac{\partial}{\partial t}\chi(t)=\text{maximum}, \tag{4-4}$$

as the mean Curie-point of a rock specimen, denoting it by θ_m, though the value of θ_m thus defined must, in general cases, differ slightly from the true mean value of the Curie-points given by $\frac{1}{S}\int_{x_1}^{x_2}\theta(x)f(x)\,dx$.[4]

3) Needless to say, the total amount of ferro-magnetic mineral S in unit volume of rock is given by

$$S=\int_{x_1}^{x_2} f(x)\,dx\,.$$

4) Since $\theta=\theta(x)$, inversely $x=x(\theta)$.
Then, putting

$$f(x)dx=g(\theta)d\theta\,,\quad \chi(x,t)=\lambda(\theta,t)\,,$$

we get

$$\chi(t)=\int_{\theta_1}^{\theta_2} g(\theta)\lambda(\theta,t)d\theta\,,\qquad \text{where}\quad \theta_i=\theta(x_i),\ i=1,2\,. \tag{4-3'}$$

Assuming that the distribution function $g(\theta)$ has such a parabolic form with respect to temperature as

$$g(\theta)=k^2-\frac{(\theta-\theta_0)^2}{a^2}\,,\qquad \text{where}\quad ka=\theta_2-\theta_0=\theta_0-\theta_1\,,$$

and further taking the special form of $\lambda(\theta,t)$ as that

$$\lambda(\theta,t)=\begin{cases}\lambda_0\ (\text{constant}), & 0\leqq t\leqq\theta\,,\\ 0 & \theta\leqq t\,,\end{cases}$$

we get

$$\chi(t)=\begin{cases}\frac{4}{3}\lambda_0k^3a\,, & 0\leqq t<\theta_1\,,\\ \lambda_0k^2\left[\frac{2}{3}ka-(t-\theta_0)\left\{1-\frac{(t-\theta)^2}{3k^2a^2}\right\}\right], & \theta_1\leqq t<\theta_2\,,\\ 0\,, & \theta_2\leqq t\,.\end{cases}$$

(*to be continued.*)

The apparent Curie-point θ_a may then be interpreted as the upper limit of $\theta(x)$, which point will corresponds to the Curie-point temperature of that part in all the ferro-magnetic minerals that has the composition of pure magnetite. The actual θ_m and θ_a of various rock specimens are given in Table 4–II. As will be seen from Table 4–II and from Fig.

Table I-4-II.

No of Specimen	θ_{ms}	θ_{me}	θ_{as}	θ_{ae}
17	460°C	(350°C)	580°C	(390°C)
18	520	285	580	320
19	530	235	580	300
20	500	—	580	—
21	515	—	590	—
22	520	260	580	310
23	495	—	580	—
24	530	230	585	300
25	530	195	585	290
59 heating	500	290	580	360
59 cooling	360	—	580	—
60′	420	—	600	—

4–3∼4–8, the θ_m of standard type rocks is always 40–80 degrees lower than θ_a, which fact is interpreted as showing that the chemical composition of every micro-crystal of the ferro-magnetic minerals does not differ much from that of pure magnetite, although there may be slight differences the one from the other.

In contrast to this, the $\chi(t)\sim t$ relations in some rock specimens are not simple, their susceptibilities changing stepwise with temperature, as shown in Fig. 4–9∼4–14. As to these curves of the

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(*continued.*)
Then, from these relations we obtain

$$-\frac{\partial}{\partial t}\chi(t)=\begin{cases}\chi_0 k_2\left\{1-\dfrac{(t-\theta_0)^2}{k^2a^2}\right\}, & \theta_1\leqq t<\theta_2\\ 0, & t<\theta_1 \text{ and } t\geqq\theta_2,\end{cases}$$

with the result that $-\frac{\partial}{\partial t}\chi(t)=$maximum at $t=\theta_0$.

On the other hand, the mean Curie-point in this special case is given by

$$\theta=\frac{1}{S}\int_{x_1}^{x_2}f(x)\,\theta(x)\,dx=\frac{1}{S}\int_{\theta_1}^{\theta_2}\left\{k^2-\frac{(\theta-\theta_0)^2}{a^2}\right\}\theta\,d\theta=\theta_0.$$

Thus, in this special case, θ agrees exactly with the temperature at which $-\frac{\partial}{\partial t}\chi(t)$ is maximum. Generally speaking, however, they are not always identical.

$\chi(t) \sim t$ relations, they may be regarded as the result of superposition of two fundamental modes of $\chi(t)$, one of which is the standard type mode, while the other, called here the extraordinary mode, is shown by a curve, the mode of which is similar to the standard type curve, having only one mean Curie-point, although its θ_a does not agree with the Curie-point of pure-magnetite which, as a rule, differs considerably from the last named. The mean and apparent Curie-points of these two fundamental curves, i.e. the standard and extraordinary ones, being denoted by θ_{ms}, θ_{as}, and θ_{me}, θ_{ae} respectively, the observed values of these four quantities are given also in Table 4–II, where it will be seen that, in most cases, θ_{me} and θ_{ae} are about $200° \sim 280°C$ and $290 \sim 320°C$ respectively, while θ_{ms} and θ_{as} almost agree with θ_m and θ_a respectively in the standard type rock. Expressing this conclusion by formula, we get

$$\chi(t) = \chi_s(t) + \chi_e(t)$$

where

$$\chi_s(t \geqq \theta_{as}) = 0\,, \qquad \chi_e(t \geqq \theta_{as}) = 0\,.$$

As the most reliable explanation of the stepwise change in susceptibility with temperature, we may presume that the ferro-magnetic minerals in the rock consist of two mineral groups of different chemical composition. One of them, which corresponds to the standard, will have nearly the same composition as magnetite, while that corresponding to the extraordinary mode will differ greatly from that of magnetite.

Generally speaking, however, it is possible that a $\chi(t) \sim t$ curve is composed of more than two fundamental curves, the mean Curie-point of which are not alike. Thus, generally,

$$\chi(t) = \sum_j \chi_j(t)\,. \tag{4–6}$$

Further, a more complex mode of $\chi(t)$ may be possible provided we take an arbitrary functional form, such as $f(x)$ in eq. (4–3) or $g(\theta)$ in eq. (4–3′). So far as the present experimental results go, however, the $\chi(t) \sim$t curve shows either the standard mode or one to be seperated into two fundamental modes.

In order to see the relation between mode of $\chi(t)$ and the chemical composition of the rock, if any, the amount of normative magnetite and ilmenite calculated from the chemical composition of the particular rock and the relative content of ilmenite, ilmenite/(magnetite + ilmenite),

Table I-4-III.

No.	$\chi_s(0)$	$\chi_e(0)$	$\frac{\chi_e}{\chi_s+\chi_e}$	Mt	Il	$\frac{Il}{Mt+Il}$
17	2.25×10^{-3}	0×10^{-3}	0	7.87%	2.73%	0.26
18	0.72 "	0.70	0.49	3.94	2.73	0.41
19	1.30 "	0.56	0.30	5.33	3.79	0.42
20	1.29 "	0	0	6.48	3.34	0.34
21	1.62 "	0	0	6.95	2.58	0.27
22	0.40 "	0.32	0.46	3.94	2.58	0.40
23	1.30 "	0	0	4.40	2.12	0.33
24	0.14 "	0.26	0.65	1.39	1.67	0.55
25	0.42 "	0.52	0.55	2.55	2.73	0.52

are given in Table 4-III. Comparing each of these values with the respective corresponding $\chi(t)$ curve, it will be seen that the extraordinary mode in $\chi(t)$ begins to appear when the relative content of ilmenite just exceeds 0.34——a fact probably showing that the extraordinary mode in $\chi(t)$ is closely related to the relative proportion of ilmenite or TiO_2 in the ferro-magnetic minerals contained in the rock.[5] It should be noted, however, that this effect of ilmenite on the magnetic behaviour of rocks may not be due to the proportion of pure ilmenite in the rock, but to the presence of solid solutions of ilmenite and magnetite, seeing that the appearance of the extraordinary mode depends on the ratio $Il/(Il+Mt)$, and not on the proportion of ilmenite itself, which is nearly the same in all specimens as shown in Table 4-III.[6]

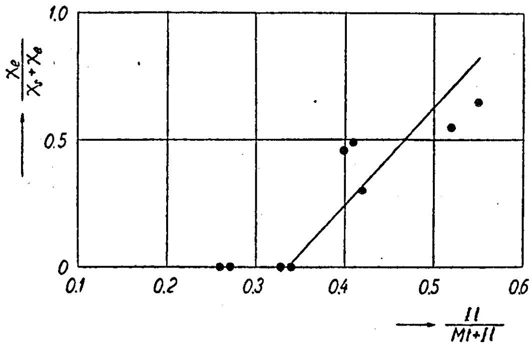

Fig. I-4-15. Relation between $\frac{\chi_e}{\chi_s+\chi_e}$ and $\frac{Il}{Mt+Il}$, (ejecta from Volcano Huzi).

In all these examples just mentioned, the susceptibility of rocks is almost reversible with respect to temperature, regardless of whether

5) See, for comparison, §3 of Chapter III.

6) Since the equilibrium condition between magnetite and ilmenite, or between Fe_2O_3, FeO and TiO_2 has not yet been cleared, more detailed analysis of this phenomenon seems to be impossible at present. (However, see Fig. III-4-9 in Chapter III).

it is of the standard mode or not. On the other hand, we have an example in which the change in susceptibility is irrevers ble with respect to temperature. It is shown in Fig. 4–14, that the susceptibility changes stepwise with temperature in the initial heating process, θ_{me}, θ_{ms}, θ_{ae} and θ_{as} being 270°*C*, 500°*C*, 360°*C*, and 580°*C* respectively, while in cooling it changes rather monotonously, θ_m being 360°C, though θ_a is 580°C in this case also. In such rock, the chemical composition of ferro-magnetic minerals would be unstable in the case of thermal agitation, whence it is presumed that, owing to heat treatment, ilmenite separates out from the magnetite-ilmenite solid solution, the chemical composition of the remaining ferro-magnetic minerals becoming rich in magnetite.

Lastly, it may be noted here that the susceptibility of rock in a weak magnetic field, say, about 2 *Oersteds*, increases with temperature in a certain initial temperature range from 0°*C*, the rate of increase being marked especially in the case of the $\chi(t)$ containing the extraordinary mode, in which case that rate is from 0·2 to 0·5 percent/degree. Since this phenomenon of increase of magnitude in magnetic susceptibility with temperature would not be seen in the $\chi(t)\sim t$ relations in a magnetic field of about 150 *Oersteds*,[7] it is believed that the phenomenon[8] is peculiar to the rock in a weak field.

Summarizing the results obtained in the present paragraph, we may say that the rock composing the earth's crust is generally a kind of ferro-magnetic material, although its apparent magnetization is not so intense as in the usual ferro-magnetic material, and further that the ferro-magnetic behaviour is largely due to the presence of a large number of micro-crystals, the chemical composition of which is not always uniquely known, although a part of these micro-crystals seems to have the same composition as magnetite.

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7) R. Chevallier et J. Piére, *Ann. Physique*, **18** (1932), 353.

8) A few possible causes of this phenomenon may be considered, one of which is that the susceptibility itself increases with temperature just as is in the Hopkinson effect of iron in a weak magnetic field, while another cause may be that the demagnetizing factor of micro-crystals of a ferro-magnetic mineral decreases with increases in temperature, where the mean susceptibility of rock according to the result discussed in § 2 is assumed to be given by $\chi(t)=\dfrac{p\chi_m(t)}{1+p\lambda(t)\chi_m(t)}$.

CHAPTER II. NATURAL REMANENT MAGNETISM OF VOLCANIC ROCKS.

§ 1. Historical remarks and the definition of natural remanent magnetism of igneous rocks.

That igneous rocks, especially the effusive ones, probably have permanent magnetization was known in Italy since the middle of the XIX Century; it was certainly known to G. Folgheraiter[1] in 1897, whence a number of investigators attempted to measure the intensity and direction of the residual permanent magnetization of igneous rocks collected from various volcanic regions scattered over the earth's surface. Of these investigators mention will be made of G. Folgheraiter,[2] Ch. Maurain,[3] P. Mercanton,[4] F. Pockels,[5] P. David,[6] J. G. Königsberger,[7] R. Chevallier,[8] B. Brunlies,[9] and E. Thellier.[10]

But since a specimen cut off from a large mass of rocks is usually of an arbitrary complex shape, it would be difficult to determine accurately the intensity and direction of magnetization of such a rock specimen. To overcome this difficulty, S. Nakamura and S. Kikuchi[11] first attempted to apply the method of spherical harmonic analysis in determining the direction and mean intensity of magnetization of any form of rock specimen, exactly as in the case of analysis of the earth's magnetic field; that is, they assumed that the direction and intensity of the magnetic moment of a centred dipole obtained from an analysis of the observed values correspond respectively to the direction and total magnetic moment of the residual permanent magnetization of the specimen. This method of analysis was adopted by M. Matuyama[12]

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1) G. FOLGHERAITER, *Real. Accad. Lincci.*, Vol. 5, 1 *sem.*, *serie.*, 5 (1897), 64.

2) G. FOLGHERAITER, *Journ. Phys.*, 8 (1899), 660; Intensita orizzontale del Magnetisms Terrestre, Rome (1899).

3) CH. MAURAIN, *Journ. Phys.*, 10 (1901), 123.

4) P. MERCANTON, *Comp. Rend.*, 143 (1906), 138; 182 (1926), 859, 1231.

5) F. POCKELS, *Phys. ZS.*, 2 (1901), 306.

6) P. DAVID, *Comp. Rend.*, 138 (1904), 41.

7) J. G. KÖNIGSBERGER, *Gerl. Beitr. Geophys.*, 35 (1932) 51, 204; *Beitr. Angw. Geophys.*, 4 (1934), 385; *ZS. Geophys.*, 8 (1932), 322; *Phys. ZS.* 33 (1932), 468.

8) R. CHEVALLIER, *Ann. Physique*, 4 (1925), 5.

9) B. BRUNLIES, *Journ. Physique*, 5 (1906), 705.

10) E. THELLIER, *Comp. Rend.*, 197 (1933), 232, 1399. *Ann. Phys. Globe*, 10 (1932), 112; 16 (1938), 157.

11) S. NAKAMURA and S. KIKUCHI, *Proc. Tokyo Math. Phys. Soc.*, 6 (1912), 268.

12) M. MATUYAMA, *Proc. Imp. Acad. Japan*, 5 (1929), 203; *Proc. 4th Pacific Sci. Cong.*, (*Java* 1929), 567; *Nippon Gakuzyutu Kyōkai Hōkoku*, 3 (1927), 252.

in his comprehensive study on the direction of residual permanent magnetization of basalt in both Japan and Manchoukuo.

As this measuring method proposed by Nakamura and Kikuchi seemed the most reliable one at present, virtually the same method was adopted in our investigations. In the present study, however, the order of magnitude of possible errors was first estimated, after which the uniformity of magnetization of pieces of rock specimen was examined with the aid of spherical harmonic analysis. In this way, it should be possible to obtain a reliable value of the direction and intensity of residual permanent magnetization of a piece of rock specimen.

Generally speaking, the residual permanent magnetization that a rock mass has in its natural state in the field can be called its natural remanent magnetization. Then, there is the contingency that a rock mass very near the earth's surface has rather intense residual magnetization as the result of sudden occurrence of a strong magnetic field caused by a thunderbolt falling near that rock mass. F. Pockels,[13] who made a special experimental study of this problem, found that pieces of basalt lying around the lightning-conductor were permanently magnetized after lightning passed through this conductor. According to him, the direction of magnetic force around the lightning-rod produced by the lightning through it, is circular with respect to the axis line coinciding with the rod, the intensity of the magnetic force estimated from the magnitude of residual magnetization of pieces of basalt not being so very intense, that is, the mean current intensity corresponding to a lightning passing through the rod amounts to 10000–20000 Amperes.

J. G. Königsberger[14] and A. F. Hallimond[15] also pointed out that the residual magnetization of igneous rocks lying very close to the earth's surface was sometimes due to the falling of thunderbolt, while a few examples of the same phenomenon have also been described by the writer.[16]

In these cases, needless to say, the direction of residual magnetization is not uniform, varying from point to point in a very small limited region—a few meters in linear dimensions—of the earth's surface, its intensity, moreover, changing also from point to point.

On the other hand, however, there are many cases in which a widely distributed rock mass, such as a lava flow, is permanently

13) F. Pockel, *Ann. Chem. Phys.*, **63** (1897), 63; *Phys. ZS.*, **2** (1901), 334.

14) J. G. Königsberger, *Gerl. Beitr. Geophys.*, **35** (1932), 204.

15) A. F. Hallimond, *Proc. Roy. Soc. London*, **141** (1933), 302.

16) T. Nagata, *Zisin*, **13** (1941), 243.

magnetized in almost uniform direction. For example, any rock specimen collected from any part of the An'ei lava flow in the Mihara Volcano,[17] which covers an area of about 5 km^2, has residual permanent magnetization of nearly the same direction, say, 42°∼50° in dip and 2°W∼14°E in declination from the present geomagnetic meridian. In this latter case, the magnetic field around the rock mass would not change much during the period from the time of its ejection to the present, except the various kinds of periodic and secular variations in geomagnetic field and magnetic storms. Briefly speaking, the external magnetic field in connexion with residual magnetization seems to be limited to the earth's magnetic field.

In the present report, only the residual permanent magnetization of rock in the latter case will be called natural remanent magnetization.[18] Here, the extraordinary residual magnetization which is due to lightning can easily be distinguished from the newly defined natural remanent magnetization by examining its heterogeneous distribution in direction and intensity.

§ 2. Measurement of natural remanent magnetization.

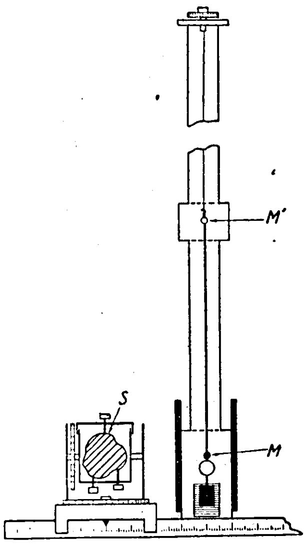


Fig. II-2-1. Schematic view of astatic magnetometer.

The instrument for measuring the natural remanent magnetization of rocks used in the present study was an astatic magnetometer,[1] in which the vertical distance between two magnets was 700 mm. The sensitivity of the astatic magnetometer could be varied from 0·3 to 3 mm/γ in scale value by using suspension wires of various diameters. In actual measurements, the sensitivity of the magnetometer was calibrated by means of Helmholtz coil before and after each measurement. A general schematic view of the instrument is shown in Fig. 2–1, where S is the test sample of rock. From the figure, it will be seen that the sample is held on a brass universal

17) see § 4, this Chapter.

18) As is given in Chapter III of this report, th natural remanent magnetization of rocks is reproducible in the laboratory experiment, in which the rock specimen is slowly cooled from a sufficiently high temperature in a magnetic field of 0·45 *Oersteds*.

1) The apparatus used in the preliminary study, (T. NAGATA, *Bull. Earthq. Res. Inst.*, 18 (1942), 281), was reproduced and improved.

stage, on which the sample could be rotated around both its horizontal and vertical axes. Since the axes of magnets M and M' nearly coincide with the geomagnetic meridian, and the position of sample S is either magnetic east or west of M, the deflection of M is, approximatelly speaking, proportional to the S–M direction component of the magnetic force at point M, which is due to the magnetization of S. Hence, the magnetic intensity at M of the component in the direction through the centre and any point on the surface of the sample could be measured by rotating the sample around the vertical axis from 0 to π, and around the horizontal axis from 0 to 2π, where the distance between S and M has always been kept constant (r). In other words, the operation just mentioned is nothing else but measuring the radial magnetic force at various points (r, θ, ϕ) around the sample, where the spherical coordinates (r, θ, ϕ) are fixed to the sample, the $\theta=0$ axis, especially, being taken as the horizontal axis of the sample.

Since magnets M and M' were sufficiently small, say, about 2 *cm* long, compared with r (16 *cm*~20 *cm*), magnets M and M' can be assumed to be almost a magnetic dipole. While, since the distance between the compensating magnet M' and sample S was much larger than that between M and S, the effect of the magnetic field due to magnetization of S upon M' can be neglected, as the first approximation, compared with that upon M.

Then, since the magnetic potential due to the magnetization of a rock sample is taken as being

$$W=R\sum_{n=1}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r}\right)^{n+1}P_n^m(\cos\theta)\left\{A_n^m\cos m\phi+B_n^m\sin m\phi\right\}, \qquad (r\geqq R) \tag{2-1}$$

where R is the radius of the sphere that envelopes the rock specimen, the radial component of magnetic force at a point (r, θ, ϕ) is given by

$$F=-\frac{\partial W}{\partial r}=\sum_{n=1}^{\infty}\sum_{m=0}^{n}(n+1)\left(\frac{R}{r}\right)^{n+2}P_n^m(\cos\theta)\left\{A_n\cos m\phi+B_n^m\sin m\phi\right\}. \tag{2-2}$$

On the other hand, the deflection of the magnet system δ observed at various points (θ, ϕ), where r is constant, can also be expressed by a series of surface spherical harmonics. Putting

$$\delta=KF\,, \qquad F=\sum_{n=1}^{\infty}\sum_{m=0}^{n}P_n^m(\cos\theta)\left\{\alpha_n^m\cos m\phi+\beta_n^m\sin m\phi\right\},$$

we get

$$A_n^m=\left(\frac{r}{R}\right)^{n+2}\frac{\alpha_n^m}{n+1}, \qquad B_n^m=\left(\frac{r}{R}\right)^{n+2}\frac{\beta_n^m}{n+1}. \tag{2-3}$$

Thus, we can determine the magnetic potential due to the magnetization of a rock sample from $\delta(\theta, \phi)$.

Before dealing with the results of actual observation, we shall examine here the extent of possible errors in our measurements.

(i) Effect of the magnetic field of the sample upon compensating magnet M'.

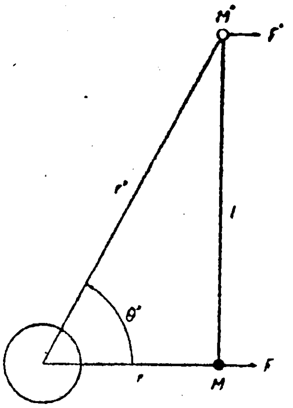


Fig. II-2-2.

The magnet system MM' is subjected to the rotational moment C, which is due to the couples acting on M and M' owing to the magnetic field expressed by eq. (2-1), it being given by

$$C=MF-M'F', \tag{2-4}$$

where F and F' denote the components of the magnetic field in the direction of SM through M and M' respectively. Taking for simplicity, the $\theta=0$ axis and the $\phi=0$ plane as the SM and SMM' plane respectively, we get

$$F=-\left(\frac{\partial W}{\partial r}\right)_{\substack{r=r\\ \theta=0\\ \phi=0}}=\sum_{n=1}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r}\right)^{n+2}(n+1)P_n^m(1)A_n^m,$$

$$F'=-\left(\cos\theta'\frac{\partial W}{\partial r}-\frac{\sin\theta'}{r}\frac{\partial W}{\partial\theta}\right)_{\substack{r=r'\\ \theta=\theta'\\ \phi=0}}$$

$$=\sum_{n=1}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r'}\right)^{n+2}\left\{(n+1)\cos\theta'\cdot P_n^m+\sin\theta'\left(\frac{dP_n^m}{d\theta}\right)_{\theta=\theta'}\right\}A_n^m$$

$$=\sum_{n=1}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r'}\right)^{n+2}\sqrt{(n+1)^2-m^2}\,P_{n+1}^m(\cos\theta')A_n^m. \tag{2-5}$$

Since $M\simeq M'$, we get

$$C=MF(1-\varepsilon)$$

$$\varepsilon=\frac{\sum_{n=1}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r'}\right)^{n+2}\sqrt{(n+1)^2-m^2}\,P_{n+1}^m(\cos\theta')A_n^m}{\sum_{n=1}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r}\right)^{n+2}(n+1)P_n^m(1)A_n^m}. \tag{2-6}$$

Because, in the actual case, the higher degree terms ($n \geqq 2$) of A_n^m is alway negligible compared with the first degree terms, as shown in the latter paragraph,[2] we can put $A_n^m = 0$ for $n \geqq 2$, whence we get

$$\varepsilon \doteqdot \left(\frac{r}{r'}\right)^3 \left\{ P_2^0(\cos\theta') + \frac{\sqrt{3}}{2} P_2^1(\cos\theta') \frac{A_1^1}{A_1^0} \right\}. \tag{2-7}$$

Since r'/r is $3{\cdot}7 \sim 4{\cdot}5$, ε has a magnitude less than 5×10^{-4}, provided A_1^1/A_1^0 is not much larger than unity. In the special case that $A_1^1/A_1^0 \gg 1$, however, C is not zero when $F=0$, that is, the position on the spheri cal coordinate where $C=0$ slightly differs from that corresponding to the case $F=0$. Here from eqs. (2-6) and (2-7),

$$C \doteqdot M\left[2\left(\frac{R}{r}\right)^3 A_1^0 - \left(\frac{R}{r'}\right)^3 \left\{2A_1^0 P_2^0(\cos\theta') + \sqrt{3} A_1^1 P_2^1(\cos\theta')\right\}\right]. \tag{2-8}$$

And if $A_1^0 = A$, and $A_1^1 = 0$,

$$CM \doteqdot F = 2M\left(\frac{R}{r}\right)^3 A\,,$$

whereas if $A_1^1 = A$ and $A_1^0 = 0$,

$$C = -MF' = -\sqrt{3} MA\left(\frac{R}{r'}\right)^3 P_2^1(\cos\theta')\,.$$

Then, the angular distance of the line $C=0$ from that of $F=0$ being taken as τ, we get

$$\sin\tau \doteqdot \frac{-\sqrt{3} MA\left(\frac{R}{r'}\right)^3 P_2^1(\cos\theta')}{2MA\left(\frac{R}{r}\right)^3} = -\frac{\sqrt{3}\, r^3 P_2^1(\cos\theta')}{2r'^3}\,. \tag{2-9}$$

Since, in the actual case, $\left(\frac{r'}{r}\right)^3 = 50 \sim 90$, $P_2^1(\cos\theta') = 0{\cdot}38 \sim 0{\cdot}46$, we obtain

$$\tau \doteqdot 0{\cdot}01 \doteqdot 30'\,. \tag{2-10}$$

This augular distance τ is negliglble in the present studies.

(ii) Dimensions of magnet.

The length, 2·0 *cm*, of a suspended magnet is not very small compared with its distance, 16∼20 *cm*, from the centre of the sample. If,

2) Moreover, the contribution to ε from higher degree terms is, as already given in eq. (2-6), much smaller than that from the first degree terms.

however, we assume that the magnet is a schematic magnet (mathematical magnet), we get, according to calculation by Ad. Schmidt[3] and G. Angenheister,[4] the ratio of the moment of such magnet due to the magnetic field of another schematic magnet of the same dimensions as that of the magnetic dipole due to the same field is $(1+\varepsilon):1$, where ε is of the order of $(l/r)^2$, l denoting the half-length of the schematic magnet. Since in the present case $(l/r)^2 \doteqdot 0{\cdot}0025 \sim 0{\cdot}004$, the correction term ε is negliglble.

(iii) Effect of magnetization of rock specimen induced by the earth's magnetic field.

In the present method of measurement, the rock sample has not only a permanent magnetization, called natural remanent magnetism, but also a magnetization induced by the earth's magnetic field. If, however, we take into consideration that induced magnetization, such as that just mentioned, is always nearly uniform in the direction of the earth's magnetic force, most of it does not affect the suspended magnet, the axis of which coincides with the magnetic meridian plane at a point magnetic east or west of the rock sample, which will now be proved. Take centre O of the spherical coordinate (r, θ, ϕ) at the centre of sample, where the $\theta=0$ axis agrees with the direction of geomagnetic force through O. Then, gravitational potential V, due to the mass of the sample, is given by

$$V=-R\sum_{n=0}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r}\right)^{n+1}P_n^m(\cos\theta)\left\{c_n^m\cos m\phi+d_n^m\sin m\phi\right\}. \qquad (2\text{–}11)$$

$$(r \geqq R)$$

By assuming that the rock specimen is uniformly magnetized by the earth's magnetic field, we get, with the aid of Poisson's law, the magnetic potential due to that magnetization, namaly,

$$\begin{aligned} W &= \frac{\kappa H}{k^2\rho}\left(\cos\theta\frac{\partial}{\partial r}-\frac{\sin\theta}{r}\frac{\partial}{\partial\theta}\right)V \\ &= \frac{\kappa H}{k^2\rho}\sum_{n=0}^{\infty}\sum_{m=0}^{n}\left(\frac{R}{r}\right)^{n+2}\sqrt{(n+1)^2-m^2}P_{n+1}^m(\cos\theta) \\ &\quad \times\left\{c_n^m\cos m\phi+d_n^m\sin m\phi\right\}, \end{aligned} \qquad (2\text{–}12)$$

3) Ad. Schmidt, *Terr. Mag.*, **17** (1912), 181; **18** (1913), 65; **29** (1924), 109.
4) G. Angenheister, *Handb. Exp. Phys. XXV. 1 Teil,* (1928), 562.

where κ, ρ, k^2, and H are respectively the magnetic susceptibility, mean density of rock specimen, gravitational constant, and the total geomagnetic force, consequently, the magnetic moment of the suspended magnet M, owing to W, is given by

$$-M\left(\frac{\partial W}{\partial r}\right)_{\theta=\frac{\pi}{2}}=\frac{\kappa HM}{Rk^2\rho}\sum_{n=0}^{\infty}\sum_{m=0}^{n}(n+2)\sqrt{(n+1)^2-m^2}\left(\frac{R}{r}\right)^{n+3}P_{n+1}^m(0)$$
$$\times(c_n^m\cos m\phi+d_n^m\sin m\phi)\,. \qquad (2\text{–}13)$$

In the case of an actual rock sample, there holds always the relation,

$$c_0^0 \gg c_n^m,\ d_n^m$$

where $R^2c_0^0$ is nothing but the total mass of rock specimen. While on the other hand, since

$$P_1^0(0)=P_3^0(0)=\cdots\cdots=P_{2n+1}^0(0)=0\,,$$

and

$$P_{n'}^{m'}(0)=0 \text{ when } n'+m'=\text{odd number},$$

$$-M\left(\frac{\partial W}{\partial r}\right)_{\theta=\frac{\pi}{2}}=\frac{\kappa HM}{Rk^2\rho}\left[3\left(\frac{R}{r}\right)^4c_1^0\right.$$
$$\left.-6{\cdot}93\left(\frac{R}{r}\right)^5\left\{c_2^1\cos\phi+d_2^1\sin\phi\right\}+\cdots\cdots\right]. \qquad (2\text{–}14)$$

Fortunately, since the intensity of natural remanent magnetization always exceeds that of induced magnetization $\kappa\times H(H=0{\cdot}45\,Oe.)$, the former being from 2 to 100 times the latter (that is, the magnitude of $\sqrt{(A_1^0)^2+(A_1^1)^2+(B_1^1)^2}$ in eq. (2) is from twice to one hundred times $\frac{\kappa H}{k^2\rho}c_0^0$), it may safely be said that the effect of magnetization of a rock sample induced by the earth's magnetic field given by eq. (2–14) is quite negligible compared with its natural remanent magnetization.

(iv) Error in determining the distance between the magnet and the centre of the sample.

As will be seen from eq. (2–2), the intensity of magnetic field due to the rock's magnetization depends directly upon the distance between the magnet and the centre of sample, r. Taking the first degree terms alone in the right-hand side of eq. (2–2), which outstands most in actual cases, we can estimate the relative observational error as the result of error in determining the distance r, as

5) $\frac{r}{R}$ was 4~6 in the actual case.

$$\frac{\Delta F}{F} = -3\frac{\Delta r}{r}, \qquad (2\text{–}15)$$

where r was defined for the practical case as the distance between the centre of magnet M and the centre of the rock holder. In an actual experiment, determination of natural remanent magnetization of a rock sample consists of two sets of measurements, the one in which the sample is mounted at a point P magnetic east of the magnet, and the other mounted at a point P' nearly symmetric with P with respect to the magnet, that is, at almost the same distance but magnetic west of the magnet.

Then the distance between P and P' being taken as $2r$, the mean value of the corresponding readings of δ in these two sets of measurement was taken as the true value of δ. It will be worth-while to note here that the above mentioned procedure which eliminates the largest part of error in determining r, should also eliminate other possible errors resulting from the assymmetry of the magnet with respect to the sample, consisting of the non-uniformity in magnetization of the magnet, deviation of the magnet from the geomagnetic meridian in its initial state, etc., whence the chances of error in F owing to the uncertainty in r is estimated to be less than 0·1 percent, seeing that

$$\Delta r/r \leqq 0{\cdot}0002\,.$$

This error is quite negligible compared with that due to other causes.

(v) Disturbance due to fluctuation in the external magnetic field.

Owing to disturbance in the magnetic field (its space gradient as well as its force at a point) due to external sources, such as tram-cars, etc., the equilibrium position of a magnet freely suspended moves fluctuatively. In order to eliminate as much as possible the short period fluctuation, the suspended magnet system was over-damped by means of an oil-damper. With this damper, much of the fluctuation was eliminated, the amplitude of residual fluctuation being smaller than 0·1 γ in the scale of magnetic intensity.

Finally, we get the residual magnetic intensity $F(\theta, \phi)$ from the observed values of $\delta(\theta, \phi)$ by means of the relation $F = \delta/K$ with an error of few per cent, where δ was observed every 20 or 30 degrees in angle of both of θ and ϕ.

Several typical examples of observed data are shown in Figs. 2–3∼2–6. As will be seen from these figures, the natural remanent magnetization of a piece of volcanic rock, generally speaking, seems to be fairly uniform, the dipole field being markedly prominent.

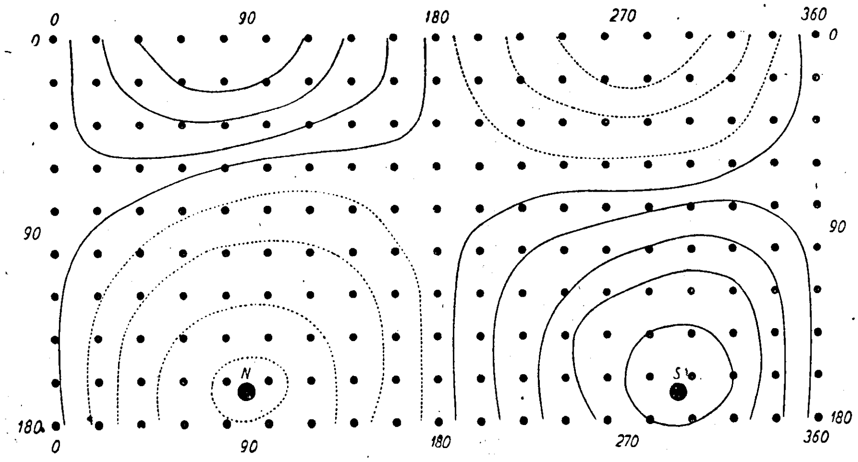


Fig. II-2-3. Specimen No. 2′.

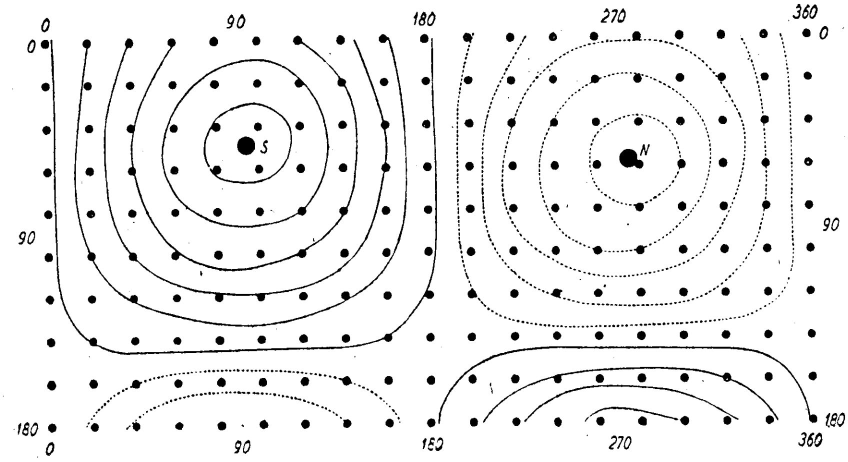


Fig. II-2-4. Specimen No. 11′.

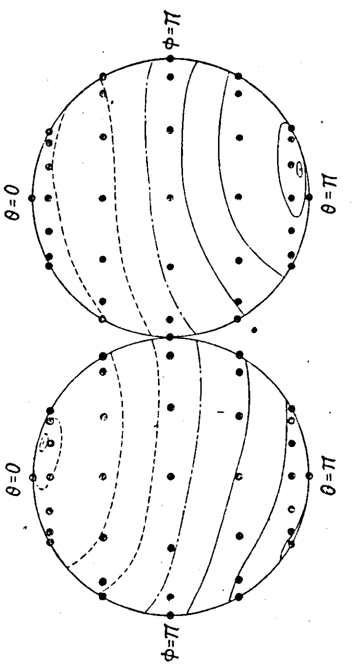


Fig. II-2-5. $F(\theta, \phi)$ of Specimen No. 23.

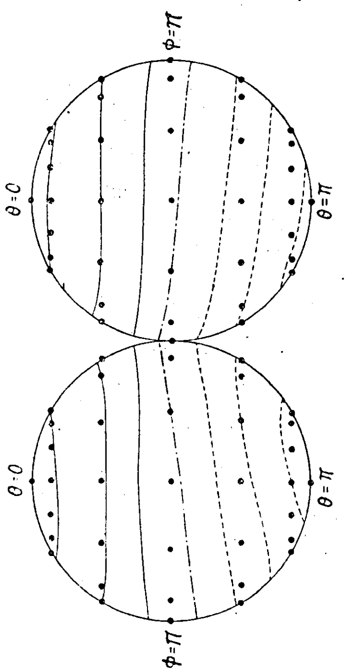


Fig. II-2-6. $F(\theta, \phi)$ of Specimen No. 73.

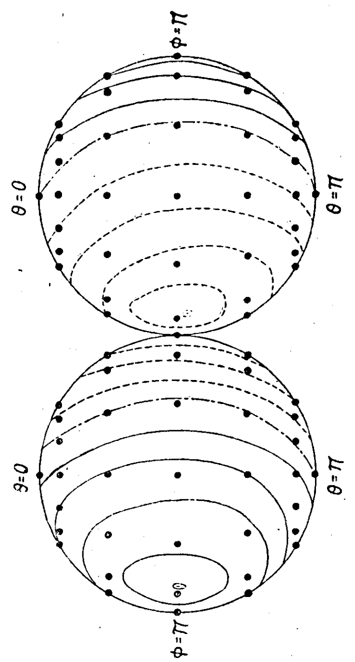


Fig. II-2-7. $F(\theta, \phi)$ of Specimen No. 69.

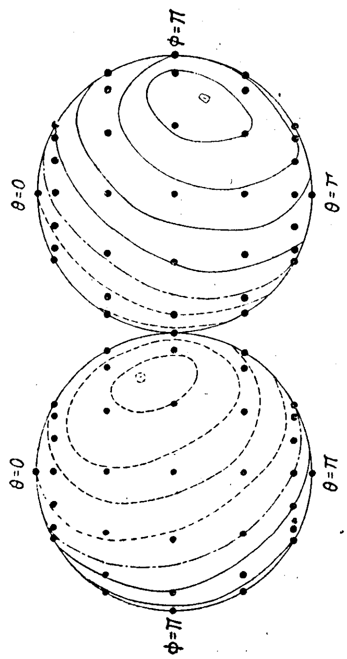


Fig. II-2-8. $F(\theta, \phi)$ of Specimen No. 66.

§ 3. Uniformity of natural remanent magnetization of rocks.

The observed values of $F(\theta, \phi)$ of rock specimens were subjected to spherical harmonic analysis, $F(\theta, \phi)$ being assumed to be given by

$$F(\theta, \phi)=\sum_{n=0}^{4}\sum_{m=0}^{n}P_n^m(\cos\theta)\,(\alpha_n^m\cos m\phi+\beta_n^m\sin m\phi)\,. \tag{3-1}$$

If, further, the magnetic potential at $r=r$ (from which eq. (3-1) ought to be derived) is expressed by

$$(W)_{r=r}=r\sum_{n=0}^{4}\sum_{m=0}^{n}P_n^m(\cos\theta)\,(g_n^m\cos m\phi+h_n^m\sin m\phi)\,, \tag{3-2}$$

then between α_n^m, β_n^m and g_n^m, h_n^m we have the relations

$$g_n^m=\frac{\alpha_n^m}{n+1}\,,\qquad h_n^m=\frac{\beta_n^m}{n+1}\,.$$

These spherical harmonic coefficients g_n^m, h_n^m [1] of the magnetic potential of a number of typical rock specimens are shown in Table 3-I∼3-IV, from which it will be seen that the higher degree terms in g_n^m, h_n^m are so small compared with the first degree terms, g_1^0, g_1^1 and h_1^1,

$[g_n^m]$ Table II-3-I. Specimen No. 23.

n \ m	0	1	2	3
0	− 0·08			
1	−18·65	+4·04		
2	+ 0·04	−0·68	−0·07	
3	+ 0·64	−0·03	−0·13	+0·13

Unit=γ

$[h_n^m]$

n \ m	1	2	3
1	−3·56		
2	+0·18	−0·01	
3	+0·07	+0·03	+0·07

Unit=γ

1) The normalized spherical harmonic function is always used throughout the calculation in this Chapter. AD. SCHMIDT. (*Tafeln der Normierten Kugelfunktionen* (1935).)

$[g_n^m]$ Table II-3-II. Specimen No. 73.

n \ m	0	1	2	3
0	+ 3·26			
1	+52·58	−4·99		
2	− 0·10	+2·44	−0·53	
3	+ 0·04	−0·38	+0·24	−0·23

Unit=γ

$[h_n^m]$

n \ m	1	2	3
1	−0·21		
2	+0·13	+0·18	
3	+0·20	+0·23	−0·02

Unit=γ

giving the dipole magnetic field, that, in the first approximation, they may be neglected.

It must be remembered, however, that the higher degree terms in the magnetic potential W at points near the rock sample ought to reach to considerable values compared with the first degree terms at these points. These higher degree terms may chiefly be due to the lack of uniformity in both intensity and direction of natural remanent

$[g_n^m]$ Table II-3-III. Specimen No. 69.

n \ m	0	1	2	3
0	+156·9			
1	−2·80	−109·88		
2	−0·09	−2·84	−4·97	
3	−0·44	+1·94	+0·29	−0·22

Unit=γ

$[h_n^m]$

n \ m	1	2	3
1	+65·16		
2	+ 0·70	−4·47	
3	− 0·67	+1·61	−0·04

Unit=γ

$[g_n^m]$ Table II–3–IV. Specimen No. 66.

$n \backslash m$	0	1	2	3
0	+ 4·90			
1	−38·49	−79·91		
2	− 0·03	−0·60	+1·21	
3	− 1·85	+4·18	+1·20	+1·57

Unit=γ

$[h_n^m]$

$n \backslash m$	1	2	3
1	+90·32		
2	+ 1·09	+0·54	
3	− 1·18	−2·55	+3·63

Unit=γ

magnetization of rock as well as to the particular form of the examined rock piece. The pieces of rock specimens examined in the present study were of a variety of complex arbitrary forms, although they did not deviate markedly from the sphere. So long as it is required that the physical condition of the rock sample shall, as near as possible, be that of its natural state in the field, the arbitrary form of the sample to be examined seems inevitable, whence it behooves us to examine here the effect of shape of rock specimen on the magnetic potential W of its natural remanent magnetization.

Let the shape of the rock sample be arbitrary, although differing but little from the sphere. The equation of surface of the sample will be

$$l(\theta, \phi) = \sum_{n=0}^{\infty} \sum_{m=0}^{n} P_n^m (\cos\theta) \left\{ p_n^m \cos m\phi + q_n^m \sin m\phi \right\}, \tag{3–3}$$

where l denotes the distance of the surface from the apparent centre of the sample. Let, further, the radius of the mean sphere, which has the same volume as the sample, the centre of the former agreeing with that of the latter, be a, and the difference l–a be d. Then

$$d(\theta, \phi) = l(\theta, \phi) - a = \sum_{n=1}^{\infty} \sum_{m=0}^{n} P_n^m (\cos\theta) \left\{ p_n^m \cos m\phi + q_n^m \sin m\phi \right\}, \tag{3–4}$$

$$a = p_0^0 .$$

Provided d/a is small, the effect of mass distribution of the sample responsible for its gravitational potential in the external space of the sample is almost equal to that of surface mass $\sigma(\theta,\phi)$ distributed on the spherical surface of $r=a$, where

$$\sigma(\theta,\phi)=\frac{a\rho}{3}+\rho d(\theta,\phi)\ , \tag{3-5}$$

and where ρ is the mean density of the sample.

On the other hand, the gravitational potential V at $r>a$ due to the mass of the sample is expressed by

$$V=-a\sum_{n=0}^{\infty}\sum_{m=0}^{n}\left(\frac{a}{r}\right)^{n+1}P_n^m(\cos\theta)\left\{u_n^m\cos m\phi+v_n^m\sin m\phi\right\}. \tag{3-6}$$

Then, from eqs. (3-4), (3-5), (3-6) we get, with the aid of the usual boundary condition of the single charge layer,

$$u_0^0=\frac{4\pi k^2\rho a}{3}$$

$$u_n^m=\frac{4\pi k^2\rho n}{2n+1}p_n^m\ ,\qquad v_n^m=\frac{4\pi k^2\rho n}{2n+1}q_n^m\ . \tag{3-7}$$

If the magnetization of the sample is uniform[2], the potential W due to its magnetization is derived from V by means of the well known Poisson's relation

$$W=\frac{J}{k^2\rho}\frac{\partial V}{\partial\nu}\ , \tag{3-8}$$

where J and ν denote respectively the intensity and direction of magnetization.

Taking direction ν as being parallel to the $\theta=0$ axis, that is,

$$\frac{\partial}{\partial\nu}=\frac{\partial}{\partial z}=\cos\theta\frac{\partial}{\partial r}-\frac{\sin\theta}{r}\frac{\partial}{\partial\theta}\ ,$$

we get

$$W=\sum_{n=0}^{\infty}\sum_{m=0}^{n}\left(\frac{a}{r}\right)^{n+2}\left\{\sin\theta\frac{dP_n^m}{d\theta}+(n+1)P_n^m\right\}\left\{u_n^m\cos m\phi+v_n^m\sin m\phi\right\}$$

$$=4\pi J\sum_{n=0}^{\infty}\sum_{m=0}^{n}\left(\frac{a}{r}\right)^{n+2}\frac{n\sqrt{(n+1)^2-m^2}}{2n+1}P_{n+1}^m\left\{p_n^m\cos m\phi+q_n^m\sin m\phi\right\}. \tag{3-9}$$

2) Since the specific intensity of natural remanent magnetization is always less than 0·03, the effect of demagnetizing factor due to the shape of rock sample will be negligibly small, whence in seems that the Poisson's relation holds approximately.

Substituting n in eq. (3–9) by n–1, we obtain

$$W=4\pi J\sum_{n=1}^{\infty}\sum_{m=0}^{n-1}\left(\frac{a}{r}\right)^{n+1}\frac{n-1}{2n-1}\sqrt{n^2-m^2}P_n^m$$

$$\times\left\{p_{n-1}^m\cos m\phi+q_{n-1}^m\sin m\phi\right\}. \qquad (3\text{–}10)$$

Comparing term by term the right-hand side of eq. (3–2) with that of eq. (3–10), we get

$$\left.\begin{matrix}g_n^m\\h_n^m\end{matrix}\right\}=\frac{4\pi J}{r}\left(\frac{a}{r}\right)^{n+1}\frac{n-1}{2n-1}\sqrt{n^2-m^2}\left\{\begin{matrix}p_{n-1}^m\\q_{n-1}^m\end{matrix}\right.$$

$$(n\geqq 1, n-1\geqq m\geqq 0)\,, \qquad (9\text{–}11)$$

and

$$g_n^n=h_n^n=0\,. \qquad (9\text{–}11')$$

Whereupon, provided the shape of the sample, $l(\theta,\phi)$, is given, and its magnetization is uniform in the direction parallel with the $\theta=0$ axis, the magnetic potential is uniquely given by the relations (3–3), (3–10), and (3–11), where g_n^n and h_n^n must always be zero.

Conversely, if the magnetic potential W at $r=r$ is given, where g_n^n and h_n^n in its spherical harmonic expansion are zero, then p_n^m and q_n^m are given by

$$\left.\begin{matrix}p_n^m\\q_n^m\end{matrix}\right\}=\frac{r}{4\pi J}\cdot\frac{2n+1}{(n+1)\sqrt{(n+1)^2-m^2}}\left(\frac{r}{a}\right)^{n+2}\left\{\begin{matrix}g_{n+1}^m\\h_{n+1}^m\end{matrix}\right. \qquad (n\geqq 1)\,,$$

$$\frac{a}{3}=\frac{r}{4\pi J}\left(\frac{r}{a}\right)^2 g_1^0 \qquad (3\text{–}12)$$

with the result that $d(\theta,\phi)$ is given by putting these terms p_n^m and q_n^m in eq. (3–4)

A few samples were examined by the present method. For example, since the direction of magnetization of samples No. 23 and No. 73 nearly agrees with the $\theta=0$ axis, as shown in Tables 3–I and 3–II, where $g_1^0\gg g_1^1$, h_1^1 although the magnitudes of g_n^n, h_n^n are not exactly zero, the present method can be applied to them.

The coefficients of spherical harmonic expansion of $d(\theta,\phi)$, i. e. p_n^m and q_n^m, which were calculated with the aid of eq. (3–12), are given in Tables 3–V and 3–VI, where a was about 3·5 *cm* in both cases, the $d(\theta,\phi)$ of these two samples obtained with the synthesis of p_n^m and q_n^m by means of eq. (3–4) is graphically shown in Figs. 3–1 and 3–2.

$[p_n^m]$ Table III-3-V. Specimen No. 23.

n \ m	0	1	2
0	+37·1		
1	− 0·6	+11·8	
2	−53·2	+ 2·5	+14·4

$[q_n^m]$

n \ m	1	2
1	−3·1	
2	−6·0	−8·0

$[p_n^m]$ Table III-3-VI. Specimen No. 73.

n \ m	0	1	2
0	+66·99		
1	− 0·76	+21·56	
2	+ 1·43	−13·65	+11·01

$[q_n^m]$

n \ m	1	2
1	+1·02	
2	+6·73	+10·39

From these results, the figures of cross section of the calculated shape of the rock sample along the planes $\theta=\frac{\pi}{2}$ and $\phi=0$ (or $\phi=\pi$) are shown in Figs. 3-3 and 3-4,[3] where, for comparison, their actual shape which was directly measured, are also shown.

As clearly seen from these results, the calculated values of $l(\theta, \phi)$ are in good agreement with the directly observed values, which agreement leads to the conclusion that the magnetization of rock specimen is almost uniform throughout all examples, where the higher degree

3) The magnitude of J, which is necessary for calculating p_n^m and q_n^m by means of eq. (3 12), can be obtained by the equation

$$J=\frac{3}{4\pi}\left(\frac{r}{a}\right)^3 g_1^0 .$$

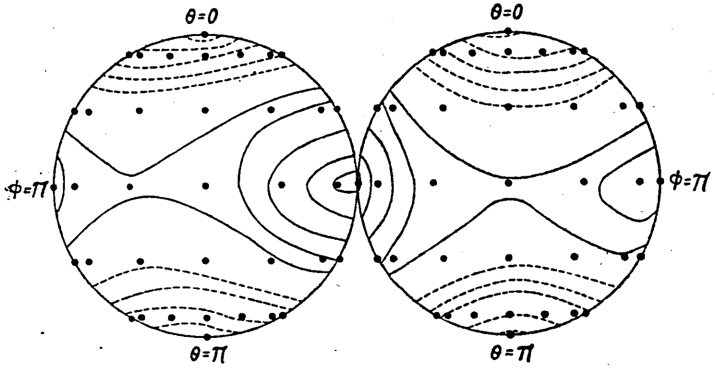


Fig. II-3-1. $d(\theta, \phi)$ of Specimen No. 23.

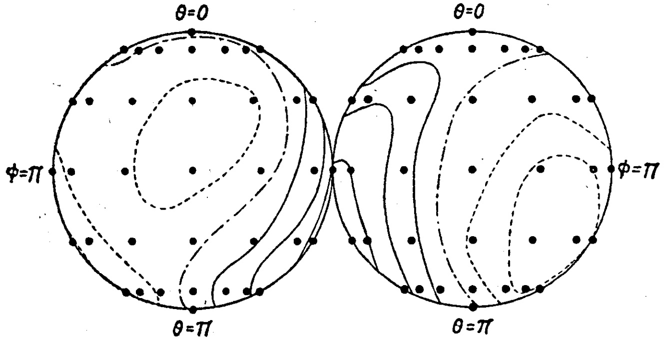


Fig. II-3-2. $d(\theta, \phi)$ of Specimen No. 73.

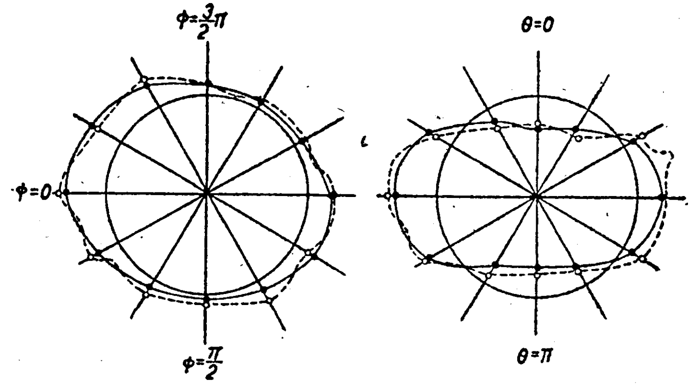


Fig. II-3-3. Left, $l\left(\frac{\pi}{2}, \phi\right)$; Right, $l(\theta, 0)$. Specimen No. 23.

Full line; calculated.
Dotted line; measured.

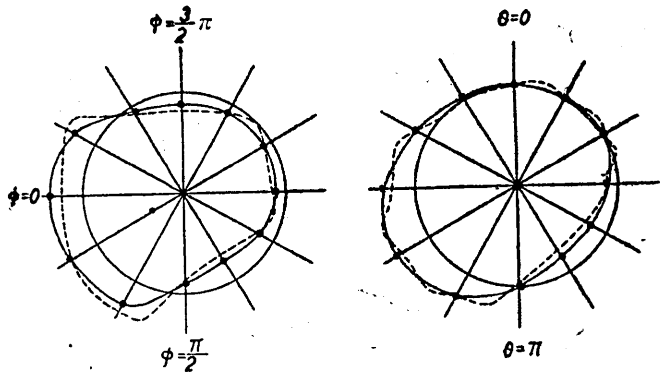


Fig. II-3-4. Left, $l\left(\frac{\pi}{2}, \phi\right)$; Right, $l(\theta, 0)$. Specimen No. 73.

Full line; calculated.
Dotted line; measured.

terms in the observed magnetic potential W are largely due to arbitrariness in the shape of the rock specimen. That is to say, the conclusion from this result is that the centred dipole which can be calculated from the first degree terms, g_1^0, g_1^1, and h_1^1, represents almost exactly the direction and intensity of natural remanent magnetization of a rock specimen, regardless of the higher degree terms in W.

It is worth while noting here that the first degree terms of the spherical harmonic series is invariant with respect to its (parallel) translation of the coordinate.[4] That is, if

$$x' = x - x_0\,, \qquad y' = y - y_0\,, \qquad z' = z - z_0\,,$$

and

$$z = r\cos\theta\,, \qquad x = r\sin\theta\cos\phi\,, \qquad y = r\sin\theta\sin\phi\,,$$

$$z' = r'\cos\theta'\,, \qquad x' = r'\sin\theta'\cos\phi'\,, \qquad y' = r'\sin\theta'\sin\phi'\,,$$

and

$$W(x, y, z) = \frac{a}{r^2}\left\{g_1^0\cos\theta + (g_1^1\cos\phi + h_1^1\sin\phi)\sin\theta\right\} + O\left(\frac{1}{r^3}\right), \qquad (3\text{–}13)$$

the spherical harmonic expansion of W with respect to the (x', y', z') coordinates becomes

$$W(x', y', z') = \frac{a}{r'^2}\left\{g_1^0\cos\theta' + (g_1^1\cos\phi' + h_1^1\sin\phi')\sin\theta'\right\} + O\left(\frac{1}{r'^3}\right).$$

Hence, the first degree terms in W are always constant, regardless of any small deviation in the true centre of the rock sample from that of the holder, which coincides with the centre of coordinate.

For this reason, in many cases of practical analysis of the observed values of F, we calculated the first degree terms alone, from which we derived the intensity and direction of magnetization.

§ 4. The direction of the natural remanent magnetization of volcanic rocks.

For determining the direction of natural remanent magnetization of rocks in their natural condition in the field the direction of a piece of rock must be determined in the field with respect to its geographic coordinates. In practice, the geomagnetic meridian through the piece of rock and the horizontal plane that it occupied in the field was measured and noted. This was measured with magnetic compass and

4) AD. SCHMIDT, *Gerl. Beitr. Geophys.*, 41 (1934), 346.

level, the observational error being less than one degree. Since, however, the rock samples to be examined were cut off from a large lava sheet, some difficulty was experienced in determining the direction of the geomagnetic meridian at that spot from which the sample was taken, for the reason that the magnetic declination directly on the surface of a lava block usually deviated a few degrees from the normal value obtaining in its neighbourhood owing to the presence of the magnetized lava block.

In order to estimate the magnitude of this deviation, the distribution of the declination along a normal from the surface of the lava sheet was observed on the spot. Several typical examples of these observations at the An'ei lava flow in Mihara Volcano and the Aokigahara lava flow in Huzi Volcano are shown in Figs. 4–1 and 4–2,

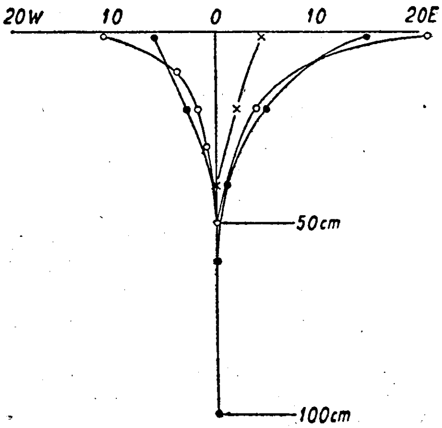


Fig. II-4-1. Anomalous deviation of geomagnetic meridian near the surface of An'ei lava.

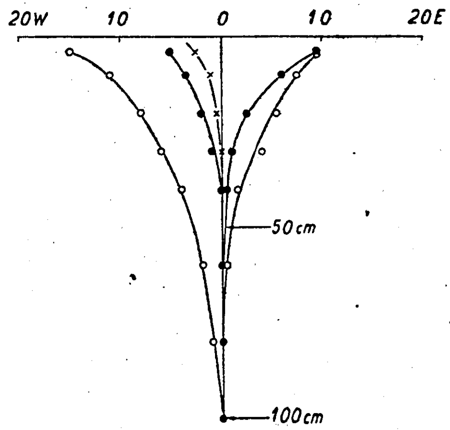


Fig. II-4-2. Anomalous deviation of geomagnetic meridian near the surface of Aokigahara lava.

where the ordinate respresents the distance from the surface of the lava sheet. As will be clear from these examples, sometimes the declination at a point a few *cm* from the surface of a large lava block differed as much as 20 degrees from the normal value. This anomalous deviation, however, vanishes rapidly with increasing distance from the surface of the lava sheet, the local deviation in declination dropping to less than one degree at a point 40∼80 *cm* distant from the surface. Hence, in our case, we adopted the magnetic meridian at a point 100 *cm* distant from the surface of the lava as the present magnetic meridian at the place where the rock sample was taken. In the case of the rock from Mihara, however, the direction of the magnetic compass at a point 50 *cm* distant from the rock surface was adopted as that of the normal geomagnetic meridian, seeing that the distance seemed sufficient, as will be seen from Fig. 4–1.

The direction of the natural remanent magnetization which was determined by the method discussed in previous paragraphs, was compared with the present magnetic meridian and the horizontal plane. The observed results will now be dealt with as follows:

(i) Direction of natural remanent magnetization of new ejecta.[1]

(a) Volcano Mihara, Meizi-Taisyō lava (ejected 1911∼1914).

That the volcanic bombs ejected in the severe eruptions of Volcano Mihara during 1911∼1914 were permanently magnetized in the direction of the geomagnetic field has already been pointed out by S. Nakamura and S. Kikuchi.[2] In our case, three specimens from the sheet of lava in the bottom of the crater which flowed out also in the volcanic activity during 1911∼1914[3], will be examined. This rock is basalt (Miharite).[4]

The angle of deviation $\Delta\delta$ in the horizontal plane of the natural remanent magnetization from the geomagnetic meridian and the dip angle θ of the former with respect to the horizontal plane are given in Table 4–I, while $\Delta\delta$ and θ are graphically shown in Figs. 4–3, and 4–4. These figures give the projection on horizontal and N-S geomagnetic vertical planes of the natural remanent magnetization, which is expressed in the form of a vector.

(b) Ejecta from Volcano Miyake-sima in July, 1940.

Volcano Miyake-sima erupted violently in July 1940, a full report of this volcanic activity and the results of geophysical observations were published in detail by the members of the expedition party from

Table II–4–I. Natural remanent magnetization of Meizi-Taisyō lava of Volcano Mihara.

Specimen	χ_0	$\Delta\delta$	θ	J_n	Q_n
No. 1	$0{\cdot}34\times10^{-3}$	0°	48°	$1{\cdot}62\times10^{-2}$	106
No. 2	0·31 〃	3°W	52°	1·89 〃	135
No. 3	—	2°E	47	1·62 〃	—

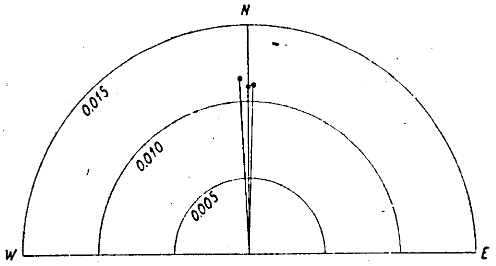
1) The natural remanent magnetization of the rock samples examined in the preliminary works, (T. NAGATA, *Bull. Earthq. Res. Inst.*, **18** (1940), 281; **19** (1941), 304.) was again measured by means of the new apparatus.

2) S. NAKAMURA and S. KIKUCHI, *loc. cit.*

3) F. OMORI, *Sinsai-Yobō Tyōsakai Hōkoku*, No. 81 (1915).

4) S. TSUBOI, *Journ. Col. Sci. Tōkyo Imp. Univ.*, **43** (1920) [6].
J. IWASAKI, *Journ. Chem. Soc. Japan*, **56** (1935), 1511.

Fig. II-4-3. Projection of the vectors of natural remanent magnetization on a harizontal plane.
Mihara, Meizi-Taisyō lava.



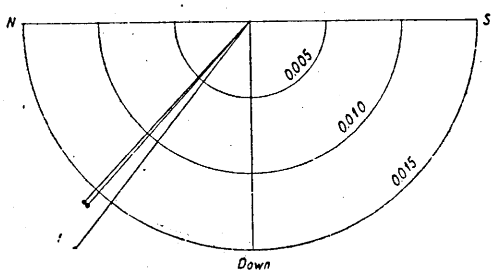
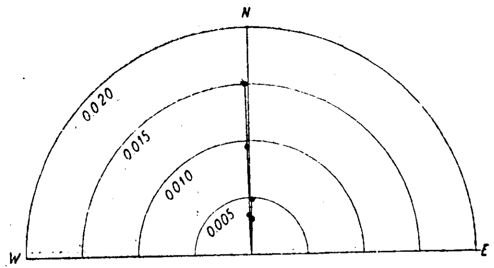


Fig. II-4-4. Projection of the vectors of natural remanent magnetization on a vertical plane containing their mean vector.
Mihara, Meizi-Taisyō lava.

Fig. II-4-5. Miyake-sima, new ejecta. (1940)



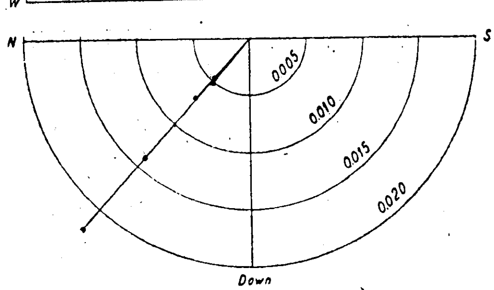
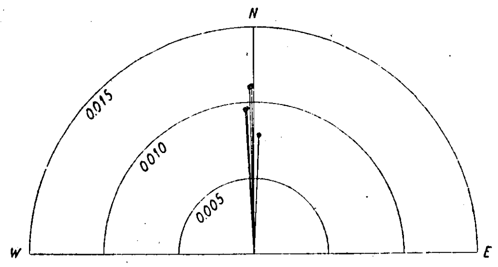


Fig. II-4-6. Miyake-sima, new ejecta. (1940)

Fig. II-4-7. Volcano Mihara, new ejecta. (1940)



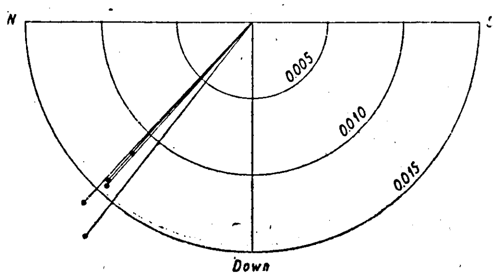
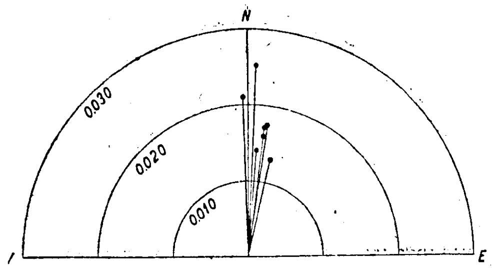


Fig. II-4-8. Volcano Mihara, new ejecta. (1940)

Fig. II-4-9. Volcano Mihara, An'ei lava.



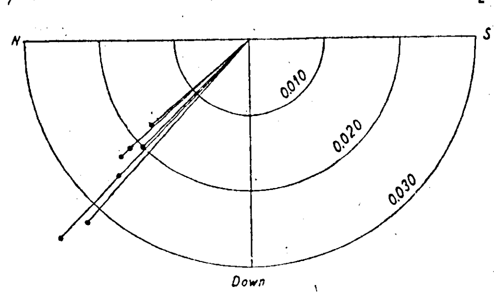
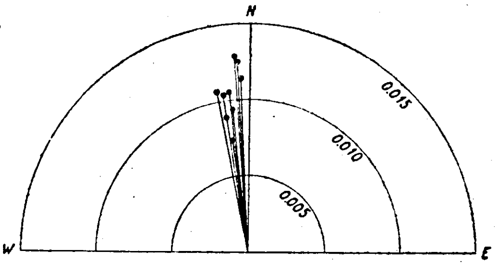


Fig. II-4-10. Volcano Mihara, An'ei lava.

Fig. II-4-11. Volcano Huzi, Aokigahara lava.



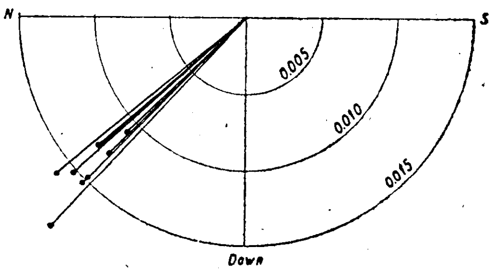
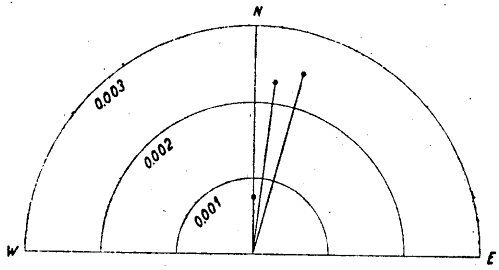


Fig. II-4-12. Volcano Huzi, Aokigahara lava.

Fig. II-4-15. Omuro-yama lava.



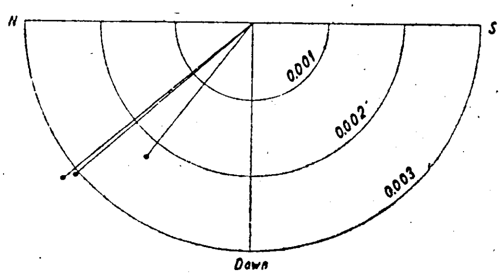
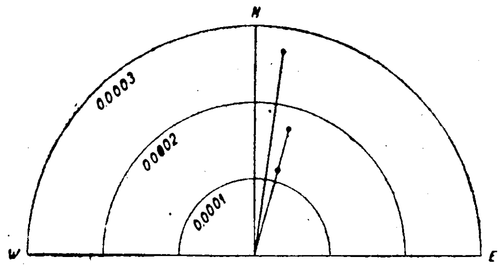


Fig. II-4 16. Omuro-yama lava.

Fig. II-4-13. Volcano Amagi, Yahatano lava.



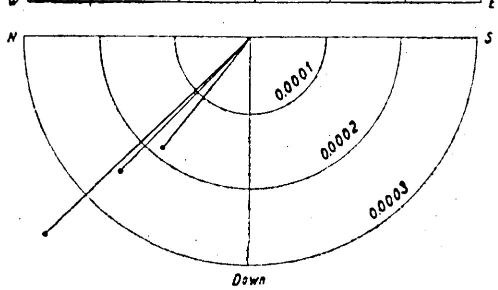
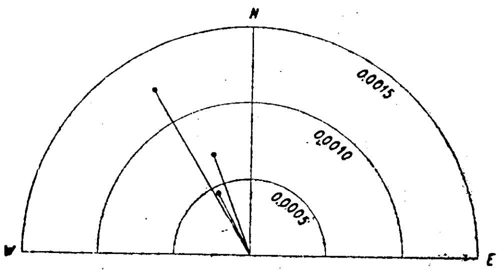


Fig. II-4-14. Volcano Amagi, Yahrtano lava.

Fig. II-4-17. Aziro lava.



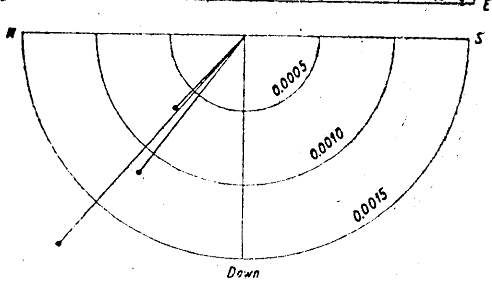


Fig. II-4-18. Aziro lava.

the Earthquake Research Institute.[5] The natural remanent magnetization of five specimens, two from the lava flow and three from large volcanic-bombs, were examined, where the lava flow and the large bombs appeared to be in the same position and direction that they occupied when the lava cooled and solidified, and when the bombs rained down on the earth's surface.[6] The values of $\Delta\delta$ and θ of these samples are given in Table 4–II, while they are graphically shown in Figs. 4–5, and 4–6.

Table II–4–II. Natural remanent magnetization of new ejecta from Miyake-sima. (1940)

Specimen	χ_0	Q_n	$\Delta\delta$	θ	J_n	ρ
No. 58	$0{\cdot}76\times10^{-3}$	42	1°W	48°	$1{\cdot}40\times10^{-2}$	2·29
No. 59	1·31 〃	39	1°W	48°	2·24 〃	2·57
No. 63	0·63 〃	17	2°E	48°	0·47 〃	1·50
No. 64	0·80 〃	20	1°E	47°	0·71 〃	1·38
No. 65	0·74 〃	16	1·5°W	50°	0·51 〃	1·38

(c) Ejecta from Volcano Mihara, in August, 1940.

Thirty-eight days after the beginning of the severe eruption of Miyake-sima, Volcano Mihara in Oosima Island also began to show activity.[7] In this activity, however, lava did not flow, only a number of large bombs being ejected. Five speceimens from these ejected bombs that were believed to have been in the same position and direction as when they fell on the earth's surface, were examined.[8] Their $\Delta\delta$ and θ are given in Table 4–III and in Figs. 4–7 and 4–8.

Table II–4–III. Natural remanent magnetization of new ejecta from Volcano Mihara. (1940)

Specimen	χ_0	$\Delta\delta$	θ	J_n	Q_n
No. 67	$0{\cdot}34\times10^{-3}$	1·5°W	52·0°	$1{\cdot}79\times10^{-2}$	115
No. 68	0·80 〃	3·0°W	48·0°	1·44 〃	40
No. 69	0·97 〃	2·0°E	47·5°	1·17 〃	27
No. 70	0·37 〃	1·0°W	47·0°	1·63 〃	99
No. 71	0·68 〃	3·5°E	47·5°	1·41 〃	46

5) H. TSUYA and other, *Bull. Earthq. Res. Inst.*, **19** (1941), 260.

6) According to H. TSUYA, these rocks are *olivine-hypersthene-pyroxene-basaltic andesite* (the lava flow) and *pyroxene-olivine-basalt*, (the bombs).

7) T. NAGATA, *Bull. Earthq. Res. Inst.*, **19** (1941), 402.

8) According to H. TSUYA, these rocks are, roughly speaking, *olivine-basalt* nearly the same as Meizi-Taisyō lava ejected 1911~1914.

As will be seen from Figs. 4–3∼4–8, the horizontal direction of the natural remanent magnetization of newly ejected rocks almost agrees in every case with the present geomagnetic meridian, namely, $\Delta\delta \simeq 0$, while the dip angle of the former also nearly agrees with the inclination of the geomagnetic force.

The mean values of $\Delta\delta$ and θ in the three cases of Mihara 1911∼1914, Miyake-sima 1940, and Mihara 1940 are respectively

$$\Delta\delta = 0.^\circ 3W \pm 1.^\circ 1, \quad \theta = 49.^\circ 0 \pm 2.^\circ 1, \quad \text{(Mihara, 1911∼1914)},$$

$$\Delta\delta = 0.^\circ 2W \pm 0.^\circ 6, \quad \theta = 48.^\circ 2 \pm 0.^\circ 5, \quad \text{(Miyake-sima, 1940)},$$

$$\Delta\delta = 0.^\circ 0W \pm 1.^\circ 2, \quad \theta = 48.^\circ 4 \pm 1.^\circ 4, \quad \text{(Mihara, 1940)},$$

where the error represents the mean error in each case.

It may be said from this result that the direction of natural remanent magnetization agrees almost exactly with the present geomagnetic meridian plane, provided the observational error of about one degree in angle is disregarded.

Although it is not possible at present to go into the details of the magnitude of dip angle θ, since the precise inclination angles of the geomagnetic force at the spots where the rock samples were taken are unknown, according to the magnetic survey in Oosima Island made by R. Takahasi and the writer,[9] the geomagnetic dip at Oosima in 1936 was from 46° 40′ to 51° 25′, the mean dip observed at 12 points being $48.^\circ 8 \pm 0.^\circ 5$.

Comparing this mean geomagnetic dip with that of the natural remanent magnetization of ejecta from volcano Mihara in 1911∼1914, and 1941, we can say that, within the limits of observational error, the latter almost agrees with the former.

Similarly, for the geomagnetic dip in Miyake-sima, the results of dip survey made by T. Minakami,[10] in 1941, will be used. According to him, the mean geomagnetic dip observed at 43 points in Miyake-sima was $48.^\circ 0 \pm 0.^\circ 4$, the individual values at each of the 42 points varying from 44° 42′ to 54° 38′, except one point situated at the summit of this volcano where the dip amounted to 58° 19′. In this case of Miyake-sima, the θ of the natural remanent magnetization also agrees with the geomagnetic dip.

Finally, we can conclude that, generally speaking, the direction of natural remanent magnetism of newly ejected rock agrees almost

9) R. TAKAHASI and T. NAGATA, *Bull. Earthq. Res. Inst.*, **15** (1937), 441.
10) T. MINAKAMI, *Bull. Earthq. Res. Inst.*, **19** (1941), 356.

exactly with the direction of the present geomagnetic force in dip as well as in declination, in the field of which the rock cooled from a high temperature and solidified—seeming to point to an important fact in connexion with the natural character of the so-called natural remanent magnetization of igneous rocks.

(ii) Direction of natural remanent magnetism of effusive rocks ejected during historic times.

a) An'ei lava ejected from Mihara Volcano in 1778.

The northern and north-eastern parts of the caldera of Volcano Mihara are at present covered with a large lava flow. According to old records relating to Oosima Island,[11] this lava which was ejected from the north-western part of the central cone in March 1778, flowed down through the northern and north-eastern parts of the caldera to the eastern coast of this island, whence it has been usually called the An'ei lava flow of Volcano Mihara. It is also recorded in the archives that this lava flow was red hot until about November 1778 since its ejection. Four rock samples were taken from this lava flow, and the characters of their natural remanent magnetism examined.

On the other hand, since another lava flow, ejected also in September 1778 from the crater, flowed down over the southern slope of this volcano, the southern part of the central cone at present is covered with this lava flow, which also goes by the name of An'ei lava flow. Three specimens from this lava flow were also examined. These rocks are, according to petrographical studies by S. Tsuboi[12] and H. Tsuya[13], hypersthene-bearing basalt.

The $\Delta\delta$ and θ of these seven specimens are given in Table 4–IV

Table II–4–IV. Natural remanent magnetization of An'ei lava, Volcano Mihara.

Specimen	χ_0	$\Delta\delta$	θ	J_n	Q_n
No. 1′	—	5°E	45°	$2{\cdot}00\times10^{-3}$	—
No. 2′	$0{\cdot}52\times10^{-3}$	8°E	42°	2·31 ″	99
No. 3′	0·48 ″	9°E	46°	2·50 ″	116
No. 4′	0·56 ″	2°W	48°	3·23 ″	118
No. 10′	2·21 ″	3°E	46°	3·63 ″	37
No. 11′	—	8°E	42°	2·14 ″	—
No. 12′	0·76 ″	14°E	41°	1·72 ″	50

11) F. Omori, *Sinsai-Yobō Tyōsakai Hōkoku*, No. 81 (1915), pp. 19.

12) S. Tsuboi, *loc. cit.*

13) H. Tsuya, *Bull. Earthq. Res. Inst.*, 15 (1937), 296.

and shown in Figs. 4–9 and 4–10, while the mean $\Delta\delta$ and θ are respectively

$$\Delta\delta = 6\cdot{}^{\circ}4E \pm 1\cdot{}^{\circ}9\,, \qquad \theta = 44\cdot{}^{\circ}3 \pm 1\cdot{}^{\circ}6\,.$$

Comparing $\Delta\delta$ and θ with those of the ejecta at the present time, we find that the declination of natural remanent magnetism of the old lava deviates considerably from that of the present geomagnetic force, while the dip of the former does not differ so much from the present value as does the declination. Since we have already seen that the effusive rock recently ejected was permanently magnetized in the direction of its external magnetic field in which that rock cooled and solidified, we can believe that the direction of natural remanent magnetization of the An'ei lava indicates the direction of geomagnetic force at the time this lawa cooled—say 1778 –, provided there has been no movement of the lava flow since the time it solidified. Then, the $\Delta\delta$ of the An'ei lava must show the amount of secular change in geomagnetic declination during the period from 1778 to the present day, while θ represents the geomagnetic dip there in 1778.

(b) Aokigahara lava ejected from volcano Huzi in 864.

The Aokigahara lava flow, which is widely distributed on the northern slope of volcano Huzi, was ejected in 864 from Nagaoyama,[14] one of the parasitic cones of this volcano. According to Tsuya,[15] this lava is generally two-pyroxene olivine-basalt. Ten samples were obtained from various points in the north-western part of this lava flow, and their natural remanent magnetism was examined.

Table II–4–V. Natural remanent magnetization of Aokigahara lava, Volcano Huzi.

Specimen	$\Delta\delta$	θ	J_n
No. 102	4°W	42°	$1\cdot53\times10^{-3}$
No. 103	7 〃	41·5°	1·26 〃
No. 104	9 〃	44°	1·08 〃
No. 105	10 〃	45·5°	1·48 〃
No. 106	4 〃	39·5°	1·62 〃
No. 107	10 〃	45°	1·27 〃
No. 108	12 〃	45·5°	1·53 〃
No. 109	4·5〃	47°	1·88 〃
No. 110	7·5〃	41°	1·39 〃

14) F. OMORI, *Nippon-Hunka-Si (I)*, *Sinsai-Yobō Tyōsakai Hōkoku*, No. 86 (1918), 90.

15) H. TSUYA, *Bull. Earthq. Res. Inst.*, 15 (1937), 302; **16** (1938), 638.

The values of $\Delta\delta$ and θ of these ten rock specimens are given in Table 4–V, and graphically shown in Figs. 4–11, 4–12. The mean values of $\Delta\delta$ and θ are respectively

$$\Delta\delta=7.^{\circ}9W\pm0.^{\circ}9, \qquad \theta=43.^{\circ}3\pm0.^{\circ}6.$$

As will be clear from this result, although the mean value of θ of Aokigahara lava ejected in 864 does not differ much from that of the An'ei lava ejected in 1778, the mean value of $\Delta\delta$ of the former differs considerably from that of the latter.

(iii) Direction of natural remanent magnetization of volcanic rocks ejected in remote geologic times.

The direction of natural remanent magnetization of some basalt ejected in remote geologic times in Japan has already been measured by M. Matuyama,[16] who from experiments found that the directions of natural remanent magnetism of some basalt nearly agree with that of the present geomagnetic force (within the limit of deviation of about 30 degrees), while those of other rocks are almost opposite to those of the latter. The writer and S. Murauti also measured the $\Delta\delta$ and θ of a few volcanic rocks in Izu Peninsula, which rocks were ejected in the beginning of the Quarternary Age.

The rock samples examined were three from Yahatano lava ejected from volcano Amagi in Pleistocene ages, three from the lava ejeted from Omuro Yama, a parasitic volcano of Amagi, ejected also during the Pleistocene, and three from lava exposed at Aziro, probably ejected from volcano Taga during latest Tertiary Age.[17]

The observed values of $\Delta\delta$ and θ of these nine samples are given in Table 4–VI, while they are graphically shown in Figs. 4–13 ~4–18, their mean values of $\Delta\delta$ and θ being

Yahatano lava,	$\Delta\delta=12.^{\circ}7E\pm1.^{\circ}5$,	$\theta=47.^{\circ}3\pm1.^{\circ}6$,
Omuroyama lava,	$\Delta\delta=8^{\circ}E\pm4^{\circ}$,	$\theta=45^{\circ}\pm3^{\circ}$,
Aziro lava,	$\Delta\delta=26.^{\circ}7W\pm2.^{\circ}0$,	$\theta=49.^{\circ}7\pm1.^{\circ}5$.

It is clear from these results that the declination of natural remanent magnetism of those old ejecta fairly differ from that of the present geomagnetic field, while the dip of the former nearly agrees with that of the latter, though the values of $\Delta\delta$ and θ of Omuroyama lava seem to contain a fairly large error.

16) M. Matuyama, *loc. cit.*

17) H. Tsuya, *Bull Earthq. Res. Inst*, 15 (1937), 243.

Table II–4–VI. Natural remanent magnetization of some Quarternary ejecta.

Specimen	$\Delta\delta$	θ	J_n
(Yahatano)			
No. 131	15°E	52°	$1{\cdot}86\times10^{-4}$
No. 132	15°E	46°	2·46 〃
No. 133	8°E	44°	3·76 〃
(Omuro-Yama)			
No. 134	0°	52°	$1{\cdot}24\times10^{-3}$
No. 135	7°E	41°	3·04 〃
No. 136	15°E	40°	3·21 〃
(Aziro)			
No. 141	21°W	53°	$1{\cdot}17\times10^{-3}$
No. 142	31°W	49°	1·90 〃
No. 143	28°W	47°	0·67 〃

Thus, it is established in all the examples mentioned above that the difference between the directions of natural remanent magnetization of any two rock samples taken from the same massive ejecta is less than 10 degrees, the mean values of $\Delta\delta$ and θ of a number of these rock samples having been determined with an error of a few degrees.

These mean values of $\Delta\delta$ and θ, then, must have a certain definite physical meaning, such as discussed with those of new ejecta and An'ei lava. The observed values of $\Delta\delta$ and θ of rocks ejected during both historic and geologic times will be dealt with again in Chapter IV in connexion with secular variation in the geomagnetic field.

§ 5. Intensity of natural remanent magnetism of volcanic rocks.

The most interesting character of natural-remanent magnetism of volcanic rock is that its intensity is fairly large compared with its induced magnetism in the earth's magnetic field. That is to say, letting J_n denote the specific intensity of natural remanent magnetization, we found that the J_n of a volcanic rock is usually larger than its induced magnetization χF, where F is the total intensity of the earth's magnetic field. In Table 5–I, the magnitude of J_n of various rock specimens are given, where the χ_0 of the corresponding rock specimens are also given for comparison. In order to show more clearly the ratio of J_n to the intensity of induced magnetization of the corresponding rock in the earth's magnetic field, we introduce here the quantity

$$Q_n = \frac{J_n}{\chi_0 \times 0{\cdot}45}\,^{1)},$$

where the numeral 0·45 represents the mean value of total intensity of the geomagnetic field in central Japan.

The amount of Q_n of various rocks are also given in Table 5–I. The values of J_n ad Q_n of some rock samples of ejecta from Volcanoes Mihara, Miyake, and Huzi, the $\Delta\delta$ and θ of which were determined in the preceding paragraph, are already given in Tables 4–I∼4–VI.

As will be seen from this result, Q_n amounts to 2∼10 in general cases, while it amounts to more than 26 in basaltic effusive rocks, although there are few cases in which Q_n is slightly less than one.

This fact may show that more than half the magnetization of a piece of volcanic rock placed in the earth's magnetic field is residual permanent magnetization, the magnetization induced in the rock by the earth's magnetic field being only a half or less.

Here, we must take into consideration that, according to the definition of natural remanent magnetism of rock mentioned in §1 of this Chapter, the magnetic force that has affected the rock throughout the whole period from the time of its formation to the present is only the earth's magnetic force. While, the total intensity of the earth's magnetic force in the neighbourhood of Japan is about 0·45 *Oersteds*, its change due to a severe magnetic storm usually does not exceed a few thousauds γ, whence we can say that the external magnetic force in connexion with the development of natural remanent magnetism of rock was about 0·5 *Oersteds* or less.

Then, even if the coersive force of rock is infinite, the intensity of residual magnetization after it has once been magnetized by the earth's magnetic field cannot exceed the amount $H\chi$, where H is the total intensity of the earth's magnetic force. We must remember, however, that the susceptibility of rock, generally, increases with increase in temperature over a fairly wide range from 0°C. Hence, we assume the most extreme case in which the net amount of induced magnetization $H\chi_{max}$ at temperature T_{max}, where χ is maximum, remains as it is, as the residual permanent magnetism after the rock is cooled to atmospheric temperature, where the coersive force also is infinite. The result of our experiment determining $\chi(t)$ shows

1) This quantity had been already introduced by J. G. KÖNIGSBERGER. (J. G. KÖNIGSBERGER, *Terr. Mag.*, 42 (1938), 119, and others).

Table II–5–I. Intensity of natural remanent magnetization of volcanic rocks.

Specimen No.	J_n	χ_0	Q_n
17	4.20×10^{-3}	2.55×10^{-3}	3·7
18	3·69 〃	1·31 〃	6·3
19	4·51 〃	1·82 〃	5·5
21	5·51 〃	1·55 〃	7·9
23	2·08 〃	1·29 〃	3·6
25′	2·20 〃	0·90 〃	5·4
27	0·58 〃	1·28 〃	1·0
29	0·86 〃	0·78 〃	2·5
30	0·56 〃	0·94 〃	1·3
31	0·39 〃	0·56 〃	1·6
50	0·55 〃	0·95 〃	1·3
51	0·75 〃	0·37 〃	4·5
52	0·72 〃	1·09 〃	1·5
53	0·21 〃	0·57 〃	0·8
54	2·61 〃	0·85 〃	6·9
55	1·28 〃	1·07 〃	2·7
73	4·89 〃	0·50 〃	21·7
91	2·38 〃	0·88 〃	6·0
92	6·23 〃	0·96 〃	14·5
93	0·48 〃	0·54 〃	2·0
94	1·73 〃	0·78 〃	4·9
95	1·51 〃	1·37 〃	2·5
96	1·65 〃	0·73 〃	5·1

that χ_{max} is, at most, twice or less compared with $\chi(t=\text{atmospheric temperature})$. Therefore, not even this hypothesis of development of natural remanent magnetism can explain a permanent magnetization larger than twice $H\chi$. Next, we further assume that in the above-mentioned hypothetical process, the intensity of magnetization $H\chi_{max}$ increases with decrease in temperature, following the Curie-Weiss law of ferro-magnetism. Since, however, according to our experiment t_{max} is less than $420°C$, the ratio of intensity of saturation magnetization of rock at $0°C$ to that at t_{max}, i.e. $I(0)/I(t_{max})$ is, at the most, 2 or less, where the Curie-temperature is $585°C$. The residual permanent magnetization then in this hypothetical case does not exceed four or five times $H\chi$.

Notwithstanding these circumstances, the $Q_n=J_n/H\chi$ of rocks in nature is frequently larger than 10, sometimes exceeding 100. Hence, it is inevitable that we assume that there is a process for developing the permanent magnetization that may be peculiar to the rock, which process regarding the ferro-magnetism of metals and their alloys is yet unknown.

Lastly, we shall deal briefly with the amount Q_n of various rocks. Comparing the values of Q_n given in Tables 4–I$\sim$4–IV and 5–I with the petrographic descriptions of corresponding rocks given in § 1, Chapter I, we find that, generally speaking, the Q_n of basalt is fairly larger than that of andesite, and that the Q_n of such basaltic rocks that seem to cool fairly rapidly is larger than that of the other basaltic rocks. In other words, the amount of ferro-magnetic minerals (magnetite) in a rock scattered as micro-crystals in the ground-mass is greater (consequently the phenocrysts of magnetite are less) the larger the Q_n of that rock.

Summarizing the facts obtained in this paragraph, we can say, at any rate, that the most marked peculiarity of natural remanent magnetism of volcanic rocks is that its intensity is anomalously large.

CHAPTER III. THE MECHANISM OF DEVELOPMENT OF NATURAL-REMANENT MAGNETISM IN IGNEOUS ROCKS.

§ 1. Thermo-remanent magnetism of igneous rocks.

It has been discovered that the natural-remanent magnetization of igneous rocks is generally the residual magnetization after the rock was cooled from a certain high temperature in the earth's magnetic field.

Since G. Folgheraiter[1] experimented in first reproducing the natural-remanent magnetization of igneous rocks by cooling a rock specimen from a certain high temperature to room temperature in the earth's magnetic field in the laboratory, a number of investigators[2] have repeated the experiment.

J. G. Königsberger,[3] however, was the first to point out that usually the intensity of natural-remanent magnetization far exceeds the magnitude of the residual magnetization that remains in the rock after a magnetic field of 0·45 *Oersteds* (say, the total intensity of geomagnetic field in middle latitude regions) has been applied and then removed at room temperature,[4] the former amounting to 1000∼10000 times the latters. From experiments with a number of various igneous rocks, Königsberger concluded that the relatively intense magnetization as natural-remanent magnetism can be reproduced in the laboratory only by cooling the rock specimen from a certein high temperature to room temperature, so long as the intensity of the external magnetic field is limited to less than 0·45 *Oersteds*, which magnetization he called thermo-remanent magnetism, presuming that there would be a certain effect of thermal agitation upon development of this relatively intense magnetization in rocks.

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1) G. FOLGHERAITER, *Reale Accademia dei Lincei*, **4** [I] (1895), 203.
2) G. E. ALLAN, *Phil. Mag.*, **7** (1904), 45; **17** (1909), 572.
G. E. ALLAN and J. BROWN, *Proc. Roy. Soc. Edinburgh*, **23** (1912), 69.
E. THELLIER, *Comp. Rend.*, **197** (1933), 1390.
F. LOWINSON-LESSING, *Comp. Rend. Aca. Sci. U. S. S. R.*, (1930), 239.
S. NAKAMURA and S. KIKUCHI, *Proc. Tokyo Math. Phys. Soc.*, **6** (1912), 268.
E. THELLIER, *Thèse* (L'université de Paris), 1938.
3) J. G. KÖNIGSBERGER, *ZS. Geophys.*, **6** (1930), 190. *Terr. Mag.*, **35** (1930), 145.
Gerl. Beitr. Geophys., **35** (1932), 51, 204; **38** (1933), 47.
Beitr. Angew. Geophys., **5** (1935), 193.
Phys. ZS., **33** (1932), 468. *Terr. Mag.*, **43** (1938), 119, 299.
4) As mentioned in Chapter I of this report, since the range of initial susceptibility of volcanic rock is usually smaller than 0·45 *Oe.*, there remains a residual magnetization after the rock is once magnetized by a small magnetic field of only 0·45 *Oe.*

As to the mechanism of development of thermo-remanent magnetization, he assumed that the magnitude of susceptibility of the ferromagnetic minerals in the rock, the magnitude of their demagnetizing factor, and their coersive force depend on their temperature, and that, at a certain high temperature near the Curie-point of ferro-magnetic minerals, the intensity of magnetization induced by the external magnetic field reaches a large value, owing to the increase in susceptibility and to decrease in the demagnetizing and coersive forces, whereas by cooling, the coersive force increases, the relatively intense magnetization that is obtained at high temperature remaining after the rock has been cooled down to room temperature and the external magnetic field has been removed.[5]

This assumption was founded on the results of experimental measurements of the remanent magnetization of rock specimens that were produced by cooling from various temperatures in a magnetic field of 0·45 *Oersteds.*

According to our experimental study, however, the magnitude of apparent susceptibility of igneous rock at any temperature lower than its apparent Curie-point never exceeds twice that at room temperature, notwithstanding that the intensity of natural remanent magnetization amounts to more than a few times, sometimes more than one hundred times, that of its induced magnetization in a magnetic field of 0·45 *Oe.* at room temperature, i. e. the product of the intensity of the magnetic field and susceptibility at room temperature.

Moreover, our experimental study on the general magnetic behaviour of igneous rock shows that most of its ferro-magnetism results from a few groups of a large number of ferro-magnetic microcrystals, the chemical composition of which is not uniquely determined. For this reason, we shall examine the mode of development of thermoremanent magnetism in igneous rocks from our viewpoint on the magnetic behaviour of igneous rocks found in our experiments, and then investigate the physical mechanism of its development as far as the present state of our knowledge permits.[6]

(i) Apparatus.

The apparatus for measuring the thermo-remanent magnetization produced by cooling from various temperatures in a magnetic field of various intensities is, in outline, as follows.

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5) J. G. KÖNIGSBERGER, *Terr. Mag.*, **43** (1938), 119, 299.

6) The preliminary works of this study were given in the following papers.
T. NAGATA, *Bull. Earthq. Res. Inst.*, **19** (1941), 49; *ibid.*, **20** (1942), 192; *Zisin*, **12** (1940), 285 545; **13** (1941), 55.

A non-inductive, non-magnetic electric furnace, 22 *cm* long and 2·5 *cm* inner diameter, is set coaxially with a Helmholtz-Gaugain Coil of 25 *cm* radius, the resistance wire in the furnace of which is pure nichrom wire of magnetic susceptibility less than 10^{-4} *e.m.u.*, the axes of the furnace and of the coil, obviously, being exactly perpendicular to the geomagnetic meridian.

A few of the rock specimens examined were circular cylinders 10 *cm* long and 2·0 *cm* in diameter, while the others were pulverized into small particles, 0·5∼1·0 *mm* in their mean diameter. In the latter case, these particles were poured into a silica tube 11 *cm* long and 1·8 *cm* inner diameter, the tube being shielded by fuzing after it was evacuated.

The rock specimen of cylindrical shape or the tube containing the rock particles were placed in the centre of the furnace coaxially as shown Fig. 1–1. In the space occupied by the rock piece or by the tube, the magnetic field supplied by the Helmholtz-Gaugain coil was nearly uniform, with an error of less than 0·1 per cent. The temperature of the specimen was measured at its centre in the case of pulverized rock, and at the surface nearest its centre in the case of a cylindrical specimen, both by *Pt∼Pt·Rh* thermo-junction, the temperature throughout the sample being uniform with an error of less than 5 degrees, provided the heating or cooling velocity did not exceed 100 degrees/hour.

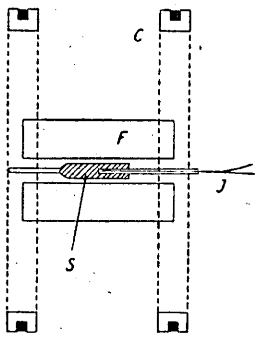

Fig. III-1-1. The apparatus for developing thermo-remanent magnetism.
C. Helmhortz coil;
F. furnace;
S. specimen;
J. thermo-junction.

The intensity of the thermo-remanent magnetism produced in this apparatus was measured with the aid of another instrument by a ballistic method, exactly the same in principle as that described in § 2, Chapter I. For measuring the intensity of magnetization of the pulverized rock in the silica tube, the method of relative measurement was adopted exactly as in the case of measuring susceptibility at high temperature (§ 4 in Chapter I), seeing that a slight error is unavoidable in the absolute determination of the constant that depends on the shape of the tube.

We now define the residual permanent magnetization at 0°C which remains after the rock specimen is cooled from temperature t to 0°C in a uniform magnetic field of H, after which the magnetic field is removed at 0°C, as the thermo-remanent magnetism of temperature t

and of magnetic field H, indicated here by $J_{t,H}$ for unit mass of sample.[7]

In measuring the $J_{t,H}$ of various values of t, successively, the $J_{t,H}$ obtained in the first experiment was purposely nullified by heating the specimen to a temperature higher than t and cooling it in non-magnetic space before proceeding with the next experiment. In this way, the intensity of the thermo-remanent magnetism of various values of t in a constant magnetic field H was measured.

(ii) The relation between intensity of thermo-remanent magnetization and the initial temperature.

The direction of the thermo-remanent magnetism always agreed with the direction of the external magnetic field applied, that is

$$\vec{J}_{t,H} \,/\!/\, \vec{H}. \tag{1–1}$$

Several typical examples of the observed data are given in Table 1–I

Table III–1–I. Thermo-remanent magnetization $\dfrac{J_{tH}}{H}$ of various volcanic rocks.

No. of specimen / Temperature	No. 20	No. 21	No. 24	No. 28	No. 60	No. 60′	No. 75	No. 76
20°C	0·00× 10^{-3}	0·00× 10^{-3}	0·00× 10^{-3}	0·00× 10^{-3}	0·00× 10^{-3}	0·00× 10^{-3}	0·00× 10^{-3}	0·00× 10^{-3}
100	0·10 ″	0·36 ″	0·12 ″	0·13 ″	—	0·11 ″	0·06 ″	0·08 ″
150	0·14 ″	—	0·25 ″	0·28 ″	—	—	0·32 ″	0·17 ″
200	0·23 ″	0·39 ″	1·87 ″	1·40 ″	0·65 ″	0·34 ″	0·70 ″	0·50 ″
250	0·26 ″	—	4·88 ″	2·90 ″	—	—	1·20 ″	3·00 ″
300	0·51 ″	0·83 ″	5·86 ″	3·70 ″	1·02 ″	0·97 ″	7·24 ″	10·41 ″
350	0·63 ″	0·84 ″	5·96 ″	4·00 ″	—	1·51 ″	12·04 ″	13·74 ″
400	0·96 ″	1·48 ″	6·19 ″	4·16 ″	2·65 ″	3·26 ″	13·70 ″	14·06 ″
450	1·29 ″	2·61 ″	6·27 ″	4·33 ″	8·64 ″	7·07 ″	14·06 ″	14·43 ″
500	3·38 ″	5·25 ″	6·44 ″	4·51 ″	18·65 ″	14·12 ″	14·44 ″	14·70 ″
550	7·56 ″	9·80 ″	6·97 ″	5·91 ″	35·8 ″	28·5 ″	14·82 ″	14·92 ″
600	10·37 ″	12·56 ″	7·80 ″	7·29 ″	46·1 ″	38·0 ″	15·18 ″	14·98 ″
650	10·49 ″	12·40 ″	7·85 ″	7·38 ″	48·0 ″	38·1 ″	15·22 ″	15·04 ″
700	10·53 ″	12·48 ″	7·80 ″	7·40 ″	47·7 ″	38·0 ″	15·20 ″	15·02 ″

as also in Figs. 1–2~1–5.

7) Since the intensitiy of the remanent magnetism that is produced by cooling from t to ordinary room temperature (less than 20°C) and that to 0°C are practically the same, the remanent magnetism produced in the former case will frequently be taken as $J_{t,H}$.

As will be seen from these results, $J_{t,H}$ in a constant magnetic field H increases with increase of temperature t, where some of the $J_{t,H}\sim t$ curves increase monotonously, while the others do so stepwise, similarly as in the $\chi(t)\sim t$ relations of the rock specimens. In the standard type mode of the $J_{t,H}\sim t$ relation, which means the mode $J_{t,H}\sim t$ that increases monotonously with temperature, the magnitude of $J_{t,H}$ reaches its highest value at t_c, $J_{t,H}$ of higher temperature than t_c being nearly always constant and equal to $J_{t_c,H}$,[8] whence we may say that the thermo-remanent magnetization $J_{t,H}$ develops during cooling through range of temperature from the critical temperature to 0°C, in a magnetic field H, while, in ranges of temperature higher than t_c, it seems that a magnetic field has scarcely any effect on the development of thermo-remanent magnetism, for which reason, we shall call $J_{t_c,H}$ the saturated thermo-remanent magnetism in H.

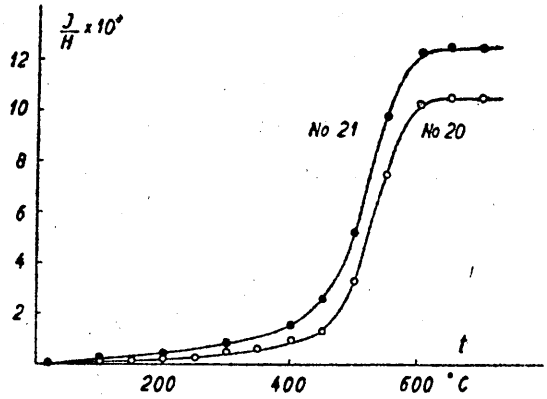

Fig. III-1-2. Thermo-remanent magnetization of specimen Nos. 20 and 21.

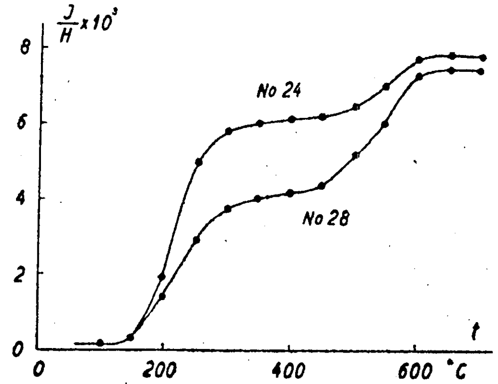

Fig. III-1-3. Thermo-remanent magnetization of specimen Nos. 24 and 28.

On the other hand, the differential cofficient of the curve for $J_{t,H}$ of the standard mode with respect to temperature is largest at t_o, $\frac{\partial^2}{\partial t^2} J_{t,H}$ being positive for temperature range lower than t_o, while it is negative at temperatures higher than t_o, so that temperature t_o may be taken as the mean transition point for development of thermo-remanent magnetism.

8) t_c will be called here the *critical temperature* of the thermo-remanent magnetism.

As to the curve of the $J_{t,H} \sim t$ relations, in which $J_{t,H}$ changes stepwise with temperature, they may be regarded exactly in the same way as in the case of the $\chi(t) \sim t$ relation, that is, the superposition of two funda-

Table III-1-II

No	t_o	t_c
17	{520°C (380)	580°C
18	{530 250	580
19	{555 240	600
20	530	600
21	545	600
22	{535 240	600
23	530	580
24	{530 240	580
25	{520 185	600
28	{530 250	580
60	530	600
60′	530	600
75	300	600
76	320	600

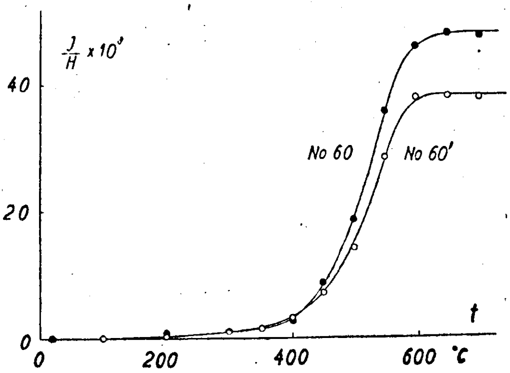

Fig. III-1-4. Thermo-remanent magnetization of specimens, Nos. 60 and 60′.

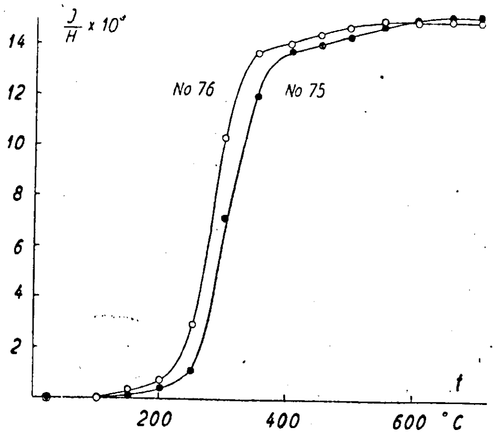

Fig. III-1-5. Thermo-remanent magnetization of specimens, Nos. 75 and 76.

mental modes, one of which is the standard type mode, while the other, called here also the extraordinary mode, is a curve such that its general mode is almost similar to the standard type mode, having only one mean transition point, but its critical temperature is considerably lower than that of the standard type mode. Consequently, one critical temperature and one transition temperature correspond to one fundamental curve of $J_{t,H}$, those of the standard type mode being denoted by t_{cs} and t_{os}, and those of the extraordinary mode by t_{ce} and t_{oe}.

The actual values of t_c and t_o are given in Table 1-II (see also Table III-3-I), where it will be noted that the critical temperature of the standard type mode of $J_{t,H}$ is always about 580°$\sim$600°C, which exactly agrees with the apparent Curie-point temperature of igneous rocks, and which probably shows that the development of thermo-remanent magnetism in igneous rocks is interpreted, at any rate, as irreversible magnetization during cooling in a magnetic field through a range of temperature at which the rock assumes a ferro-magnetic character. These characteristic properties of thermo-remanent magnetism, however, will be discussed in detail in the next paragraph.

(iii) The dependence of thermo-remanent magnetization on the magnetic field applied.

Table III–1–III. Specimen No. 60.

t.	$J_{tH_1}(H=0{\cdot}458)$	$J_{tH_2}(H_2=0{\cdot}282)$	J_{tH_1}/J_{tH_2}
20°C	$0{\cdot}00\times10^{-3}$	$0{\cdot}00\times10^{-3}$	—
100	0·15 〃	0·10 〃	1·5
200	0·30 〃	0·18 〃	1·7
300	0·45 〃	0·28 〃	1·6
400	1·22 〃	0·75 〃	1·65
450	3·97 〃	—	—
500	8·58 〃	5·40 〃	1·59
550	16·45 〃	—	—
600	21·20 〃	13·15 〃	1·61
700	22·10 〃	13·45 〃	1·64
750	21·95 〃	13·35 〃	1·64

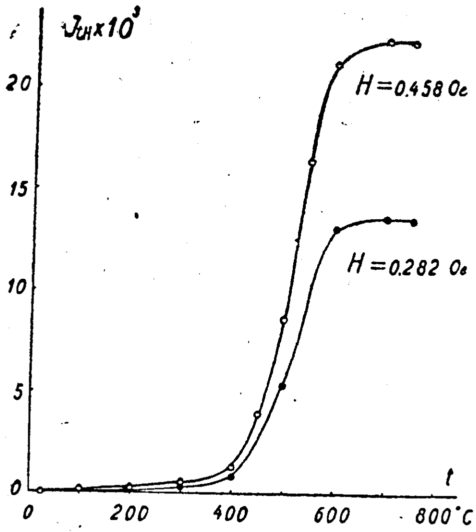

Fig. III–1–6. Intensity of thermo-remanent magnetization in different magnetic fields. Specimen No. 60.

From the foregoing results, the relation between the intensity of thermo-remanent magnetism and temperature was established, where the magnetic field applied during cooling was always kept constant.

Now, Table 1–III and Fig. 1–6 give the relation between $J_{t,H}$ and t in two different values of H. From the result given in Table 1–III and Fig. 1–7, we get the relation

$$\frac{J_{t,H_1}}{J_{t,H_2}} \fallingdotseq \frac{H_1}{H_2}, \qquad (1\text{–}2)$$

provided both H_1 and H_2 are small, i. e., about one *Oersted* or less.

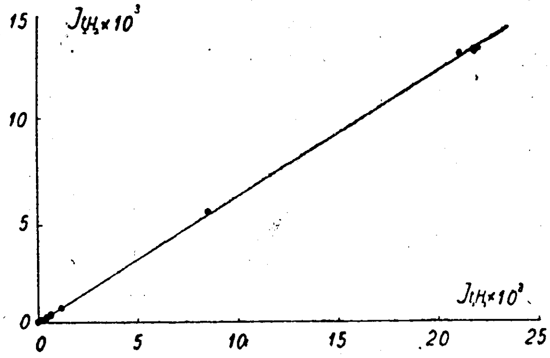

Fig. III-1-7. Relation between J_{t,H_1} and J_{t,H_2}.

Since, further, the thermo-remanent magnetism was always zero when $H=0$, we may write

Table III-1-IV (a)
Specimen No. 46.

H	$J_{tc,H}$	$J_{tc}=J_{tc,H}/H$
0 *Oe.*	0.00×10^{-2}	~
0·27	1·21 〃	4.5×10^{-3}
0·63	2·78 〃	4·4
0·98	4·07 〃	4·2
1·34	5·62 〃	4·2
1·70	5·68 〃	4·1

Table III-1-IV (b)
Specimen No. 47.

H	$J_{tc,H}$	J_{tc}
0 *Oe.*	0.00×10^{-2}	~
0·14	0·96 〃	6·9
0·28	2·10 〃	7·5
0·46	3·53 〃	7·6
0·64	5·07 〃	7·9
0·82	5·90 〃	7·2
1·00	7·20 〃	7·2
1·36	9·22 〃	6·8
1·72	11·33 〃	6·6

Table III-1-V (a)
Specimen No. 59′.

H	$J_{tc,H}$	$J_{tc,H}/H$
1·51 *Oe.*	2.67×10^{-2}	1.84×10^{-2}
3·02	4·44 〃	1·47 〃
6·04	8·04 〃	1·33 〃
10·6	11·9 〃	1·13 〃
16·6	16·1 〃	0·97 〃
27·2	21·9 〃	0·81 〃
42·3	27·5 〃	0·65 〃

Table III-1-V (b)
Specimen No. 60.

H	$J_{tc,H}$	$J_{tc,H}/H$
0·28 *Oe.*	1.36×10^{-2}	4.91×10^{-2}
0·46	2·22 〃	4·62 〃
3·02	9·33 〃	3·09 〃
6·04	13·35 〃	2·21 〃
9·06	16·2 〃	1·78 〃
16·3	20·7 〃	1·26 〃
27·2	25·2 〃	0·93 〃
42·3	30·1 〃	0·71 〃

$$\vec{J}_{t,H}=\vec{H}\cdot J_t, \tag{1-3}$$

where J_t is a function of temperature alone, independent of $\vec{H}$, provided H is small. In order to see more generally the dependence of thermo-remanent magnetization on the magnetic field applied, the magnitude of the saturated thermo-remanent magnetism in various values of H are given in Tables I–IV and I–V, and plotted in Figs.

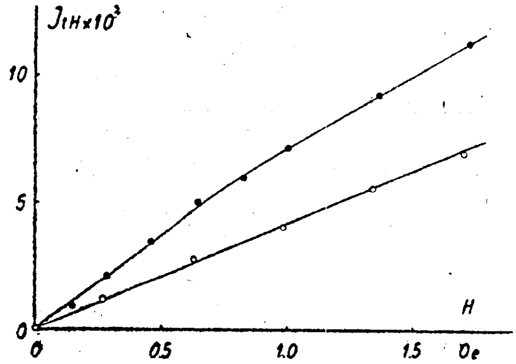

Fig. III-1-8. Intensity of saturated thermo-remanent magnetization corresponding to various values of H. Specimens. Nos. 46 and 47.

1–8, 1–9, and 1–10, where H varies from 0 to 1·7 *Oe.* and 0 to 42 *Oe.* in the former and latter two respectively. As shown in Fig. 1–8, eq. (1–3) seems to hold approximately for the values of H from 0 to about one *Oersted*. Hence, so long as we deal only with the develop-

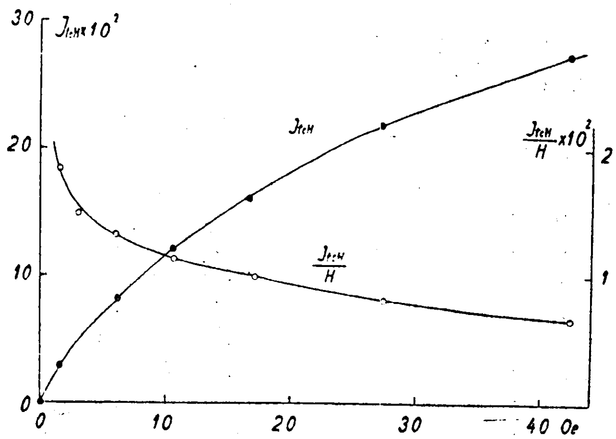

Fig. III-1-9. Intensity of saturated thermo-remanent magnetization. (Specimen No. 59′.)

ment of thermo-remanent magnetization in the earth's magnetic field, eq. (1–3) holds as a sufficient approximation.

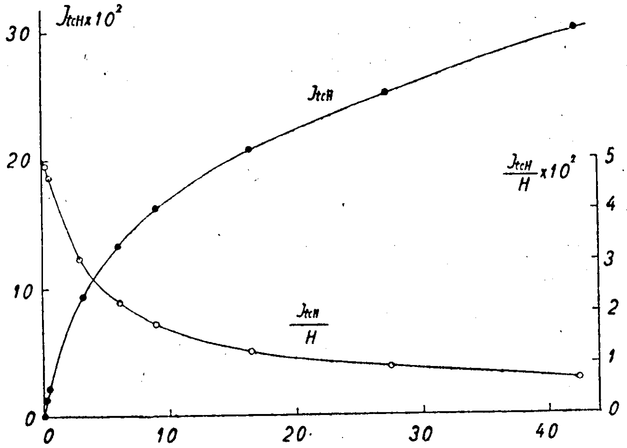

Fig. III-1-10. Intensity of saturated thermo-remanent magnetization. (Specimen No. 60.)

However, as will be seen from Figs. 1–9 and 1–10, if H is large, $J_{tc,\,H}$ is not proportional to H, although the former increases with the latter. Thus, generally speaking, there will be the relation,

$$J_{tc,\,H}=F(H)\cdot J_{tc}$$

instead of eq. (1–3).

In order to show more clearly the functional form of $F(H)$, the magnitudes of $J_{tc,\,H}/H$ corresponding to various values of H are given in Table I–V, and Figs. 1–9 and 1–10, where it will be seen that $J_{tc,\,H}/H$ approaches a finite value when $H\to 0$, whereas it tends to zero when $H\to\infty$.

On the other hand, since thermo-remanent magnetization $J_{tc,\,H}$ can never exceeds the intensity of saturation magnetization, it follows that

$$\lim_{H\to\infty} J_{tc,\,H}\leqq J_s,$$

where J_s denotes the intensity of saturation magnetization.

Summarizing then the experimental results, which are expressed in Figs. 1–8, 1–9, and 1–10, it may be said that

$$\lim_{H\to\infty} J_{tc,\,H}=J_0\leqq J_s, \tag{1–4}$$

$$\lim_{H\to 0}\frac{J_{tc,\,H}}{H}=J_{tc}. \tag{1–5}$$

The general formula expressing the relation between $J_{t,H}$ and H throughout the whole range from zero to infini y may not be simple, since it must not only fit all values given in Table V, but must also satisfy the conditions of eqs. (1–5).[9] In a certain range of H, say 10 *Oe.*, however, the following formula is probably to be preferred, namely

$$J_{t,H}=\frac{H}{1+kH}J_{tc}. \tag{1–6}$$

At any rate, it is concluded that the relative effect of development of thermo-remanent magnetization, which can be measured by $J_{tc,H}/H$, is greater the smaller the external magnetic field, and that the effect of the applied magnetic field on the development of thermo-remanence separates from the temperature effect, at any rate in the case where H is small.

The intensity of saturated thermo-remanent magnetization in a magnetic field of unit intensity, J_{tc}, then, which is reduced from $J_{tc,H}$ in a weak magnetic field by means of eq. (1–3), will be taken as a parameter in order to show the character of the thermo-remanent magnetism of materials, seeing that it is a constant peculiar to each rock specimen, independent of t as well as of H. Examples of actual values of J_{tc} are also shown in Table 1–II.

(iv) Intensity of saturated thermo-remanent magnetization in the earth's magnetic field.

From the foregoing fact it will be seen that the effusive rocks that cooled in the earth's magnetic field from a sufficiently high temperature ought to have thermo-remanent magnetization, the intensity of which is approximately given by $H\cdot J_{tc}$, where H is the total intensity of the earth's magnetic field, while its direction agrees with that of the geomagnetic field.

For example, several actual values of $H\cdot J_{tc}$ of volcanic rocks ejected in recent years, where H is assumed to be 0·45 *Oersteds*, are given in Table 1–VI,[10] where, for comparison, the intensity of natural-remanent magnetization of the same rocks is also given.

As will be seen from this result, the saturated thermo-remanent magnetism in a magnetic field of 0·45 *Oersteds* well agrees with the

9) It should be noted here that determination of the pure effect of thermo-remanent magnetization is almost impossible in a strong magnetic field, because ordinary remanent magnetism, which does not depend on heat treatment, appears relatively intense in this case.

10) A part of the magnetic behaviours of these samples given in this Table have been already reported. (T. NAGATA, *Bull. Earthq. Res. Inst.*, **19** (1941), 304). Those samples, however, were re-examined in the present study.

natural-remanent magnetism, so long as we are dealing with recent volcanic ejecta, whence it is concluded with confidence that the so-called natural remanent magnetism of igneous rocks is nothing but thermo-remanent magnetism in the earth's magnetic field.

It should be remembered, however, that there will sometimes be a certain difference in the intensities of natural-remanence and thermo-remanence in a field of 0·45 *Oersteds*, probably owing to such causes as the following; (1) the intensity of the earth's magnetic field in which the rock mass cooled from high temperature was not 0·45 *Oersteds*; (2) the rock mass moved during cooling from the critical temperature of thermo-remanent magnetism to ordinary atmospheric temperature; (3) the conditions of ferro-magnetic crystals in the rock in the case of re-heating and re-cooling in the laboratory differ from those in the case of primary cooling in the field, probably due to their instability.

Table III-1-VI

No of Specimen	$H \cdot J_{tc}$ ($H=0{\cdot}45$)	J_n
58	$1{\cdot}34\times10^{-2}$	$1{\cdot}40\times10^{-2}$
59	1·72 〃	2·24 〃
63	0·53 〃	0·47 〃
65	0·54 〃	0·51 〃
68	1·50 〃	1·44 〃
70	1·57 〃	1·63 〃
71	1·28 〃	1·41 〃
73	$5{\cdot}60\times10^{-3}$	$4{\cdot}89\times10^{-3}$

The few cases in which J_n differs from $J_{tc}\times0{\cdot}45$ in recent volcanic ejecta seem to belong to the last two cases, namely (2) and (3). This problem is discussed in greater detail in § 1, Chapter IV.

Here, we shall introduce a characteristic quantity of thermo-remanent magnetism, namely, the coefficient of thermo-remanent magnetism Q_t,[11] which is given by

$$Q_t \equiv J_{tc}/\chi_0\,, \tag{1-7}$$

where χ_0 denotes the specific magnetic susceptibility in a weak magnetic field, say, about one *Oersted*. Needless to say, the coefficient of thermo-remanent magnetism Q_t corresponds exactly to the coefficient of natural-remanent magnetism, since

$$\left.\begin{aligned} &\frac{J_{tc}}{\chi}=\frac{J_{tc,\,H}}{H\chi}\,, \\ \text{and}\qquad & \\ &J_{tc,\,0\cdot45}=J_n\,. \end{aligned}\right\} \tag{1-8}$$

11) The notation Q_t has been already used by Königsberger. (J. G. Königsberger, loc. cit.).

Table III-1-VII

No. of Specimen	χ_0	J_{tr}	Q_t
20	$1\cdot28\times10^{-3}$	$10\cdot5\times10^{-3}$	8·2
21	1·55 〃	12·5 〃	8·1
24	0·40 〃	7·8 〃	19·5
28	0·78 〃	7·4 〃	9·5
60	1·11 〃	47·6 〃	43·0
60′	1·10 〃	38·0 〃	34·5
75	1·19 〃	15·2 〃	12·8
76	1·27 〃	15·0 〃	11·8

The actual Q_t of several typical rock specimens are given in Table 1-VII.[12]

One of the most important facts in the phenomenon of thermo-remanent magnetism is, then, that in almost every case Q_t is larger than unity, frequently amounting to 3~20, and sometimes to more than 100.

To give an example, the Q_t of rock specimen No. 21 (Huzi, Makuiwa lava) is 8·1, where J_{tr} and χ are $12\cdot5\times10^{-3}$ and $1\cdot55\times10^{-3}$ respectively, whereas, from Table 4–I, Chapter I, the largest value of its specific susceptibility $\chi(t)$ throughout the whole range of temperature from 0°C to the apparent Curie-point is only $2\cdot04\times10^{-3}$, which occurs at about 420°C, whence the ratio of J_{tr} to this largest value of $\chi(t)$ amounts to 6·4; that is to say, thermo-remanent magnetization in a weak magnetic field is 6·1 times the intensity of the largest value of induced magnetization that the rock specimen shows during cooling from its apparent Curie-point in the same magnetic field.

Another example is rock specimen No. 60, where J_{tr} and the largest value of $\chi(t)$ are $4\cdot76\times10^{-2}$ and $2\cdot15\times10^{-3}$ (at 280°C) respectively, the ratio of the former to the latter being 24·5.

In all the remaining specimens, the general features are much the same as these two examples. It is interesting to note that the intensity of remanent magnetization of rock after cooling down in a magnetic field greatly exceeds that of the induced magnetization which the rock shows at any temperature in the process of cooling, a fact that seems to suggest that the causation of thermo-remanent magnetism is due, probably, to certain particular physical and chemical conditions in the ferro-magnetic minerals in the rocks, and not to mere increase in remanent magnetization (in the ordinary meaning) owing to increase in coersive force.

12) See also Table III-3-I in §3, this Chapter, and the following papers; T. NAGATA, *Bull. Earthq. Res. Inst.*, **19** (1941), 49, 304, 402; **20** (1942), 192.

§ 2. The general mode of development of thermo-remanent magnetism. (Partial magnetization during cooling.)

In the preceding paragraph, thermo-remanent magnetism $J_{t,\,H}$ was defined as the residual permanent magnetization that remains after the material cools from temperature t to $0°C$ in a magnetic field of H.

Now, in the cooling of the specimen, let magnetic field H be applied only in the temperature range of from t_i to t_{i-1}, while in other temperature ranges, higher than t_i and lower than t_{i-1}, let the external magnetic field be kept always at zero. The rock specimen that was cooled under the conditions just mentioned also showed residual permanent magnetism, so that we shall call this permanet magnetism the partial magnetization during cooling, and refer to it as $\Delta J_{t_{i},\,H}^{t_{i-1}}$.

The experiment for bringing about partial magnetism during cooling was made with the same apparatus as that for producing thermo-remanent magnetism, its intensity being measured also by the ballistic method. In this case, also, tests were made before each experiment in order to see that the intensity of permanent magnetization of the specimen in its initial state was zero.

Thus, the $\Delta J_{t_{i},\,H}^{t_{i-1}}$ for various successive values of t_i was measured, the temperature range $\Delta t = t_i - t_{i-1}$ being always kept constant.

As examples, the observed values of $\Delta J_{t_{i},\,H}^{t_{i-1}}$ for $\Delta t = 50$ degrees are shown in Tables 2–I∼2–V, and in Figs. 2–1∼2–5, where, as already mentioned, the values of $\Delta J_{t_{i},\,H}^{t_{i-1}}$ are arranged in the order of temperature $\overline{t}_i = \frac{t_i + t_{i-1}}{2}$, where,

$$\left.\begin{aligned} H&=0, &\quad &\text{for} &\quad & t \geqq t_i, \\ H&=H, && \text{for} && t_i > t \geqq t_{i-1}, \\ H&=0, && \text{for} && t_{i-1} > t \geqq 0. \end{aligned}\right\} \quad (2\text{–}1)$$

As will be seen from these figures, in some rock specimens, the curves showing the values of $\Delta J_{t_{i},\,H}^{t_{i-1}}$ against the various values of $\overline{t}_i$ in a constant magnetic field H have only one maximum, while in others they have two maxima. Further, comparing the relation between $\Delta J_{t_{i},\,H}^{t_{i-1}}$ and t_i with that between $J_{t,\,H}$ and t of the same specimen, we find that the $\Delta J_{t_i, H}^{t_{i-1}} \sim t_i$ curves which have one maximum correspond to the standard type in the $J_{t, H} \sim t$ relation, while the former, which has two maxima, corresponds to the complex type mode in the latter, all of which can be interpreted as the superposition of standard and extraordinary modes of thermo-remanent magnetism. Consequently, as will be seen

Table III–2–I. Specimen No. 21. ($H = 0{\cdot}92$ *Oe.*)

Range of Temp. Δt	$\dfrac{\Delta J}{H}$	$\dfrac{\partial J}{H \cdot \partial t} \times \Delta t$	$\Sigma \dfrac{\Delta J}{H}$	$\dfrac{J}{H}$	Temp.
100~50°C	$0{\cdot}20 \times 10^{-3}$	——	$0{\cdot}20 \times 10^{-3}$	$0{\cdot}36 \times 10^{-3}$	100°C
150~100	0·20 〃	$0{\cdot}02 \times 10^{-3}$	0·40 〃		150
200~150	0·23 〃	0·02 〃	0·63 〃	0·39 〃	200
250~200	0·20 〃	0·22 〃	0·83 〃		250
300~250	0·13 〃	0·[illegible]2 〃	0·96 〃	0·83 〃	300
350~300	0·23 〃	0·01 〃	1·19 〃	0·84 〃	350
400~350	0·15 〃	0·64 〃	1·34 〃	1·48 〃	400
450~400	0·45 〃	·13 〃	1·79 〃	2·61 〃	450
500~450	2·01 〃	2·61 〃	3·80 〃	5·25 〃	500
550~500	3·99 〃	4·55 〃	7·79 〃	9·80 〃	550
600~550	4·22 〃	3.76 〃	12·01 〃	12·56 〃	600
650~600	0·29 〃	−0·16 〃	12·30 〃	12·40 〃	650
700~650	0·13 〃	0·08 〃	12·43 〃	12·48 〃	700

Table III–2–II. Specimen No. 60. ($H = 0{\cdot}46$ *Oe.*)

Range of Temp. Δt	$\dfrac{\Delta J}{H}$	$\dfrac{\partial J}{H \cdot \partial t} \times \Delta t$	$\Sigma \dfrac{\Delta J}{H}$	$\dfrac{J}{H}$	Temp.
100~50°C		$0{\cdot}22 \times 10^{-3}$		$0{\cdot}00 \times 10^{-3}$	20°C
150~100	$0{\cdot}15 \times 10^{-3}$	0·22 〃	$0{\cdot}15 \times 10^{-3}$		150
200~150	0·46 〃	0·22 〃	0·61 〃	0·65 〃	200
250~200	0·46 〃	0·22 〃	1·06 〃		250
300~250	0·56 〃	0·43 〃	1·63 〃	0·98 〃	300
350~300	0·69 〃	0·69 〃	2·32 〃		350
400~350	0·93 〃	0·98 〃	3·39 〃	2·65 〃	400
450~400	0·95 〃	5·97 〃	4·25 〃	8·61 〃	450
500~450	4·77 〃	10·00 〃	9·03 〃	18·62 〃	500
550~500	24·74 〃	17·08 〃	33·77 〃	35·70 〃	550
600~550	14·89 〃	10·31 〃	48·65 〃	46·00 〃	600
650~600	0·54 〃	0·43 〃	49·20 〃		650
700~650	0·30 〃	0·22 〃	49·50 〃	47·96 〃	700
750~700	0·17 〃	−0·33 〃	49·67 〃	47·63 〃	750

from Tables 2–I∼2–V and Figs. 2–6∼2–10, the sum of $\Delta J^{t_{i-1}}_{t_i, H}$ of the whole range from 0 to t, that is, the sum of $\Delta J^{t_{i-1}}_{t_i, H}$ from $t_{i-1}=0$ to $t_i=t$, almost always agrees with the amount of $J_{t, H}$, the thermo-remanent magnetization developed by cooling from t to 0 in the same magnetic field H.

Table III–2–III. Specimen No. 60′. ($H=0{\cdot}46$ *Oe.*)

Range of Temp. Δt	$\frac{\Delta J}{H}$	$\frac{\partial J}{H\cdot\partial t}\times\Delta t$	$\Sigma\frac{\Delta J}{H}$	$\frac{J}{H}$	Temp.
100∼50°C	$0{\cdot}26\times10^{-3}$	$0{\cdot}11\times10^{-3}$	$0{\cdot}26\times10^{-3}$	$0{\cdot}11\times10^{-3}$	100°C
150∼100	0·52 〃	0·11 〃	0·78 〃	----	150
200∼150	0·59 〃	0·1 〃	1·37 〃	0·34 〃	200
250∼200	0·59 〃	0·28 〃	1·96 〃	——	250
300∼250	0·96 〃	0·35 〃	2·92 〃	0·97 〃	300
350∼300	1·05 〃	0·54 〃	3·97 〃	1·51 〃	350
400∼350	1·44 〃	1·75 〃	5·41 〃	3·26 〃	400
450∼400	3·27 〃	3·81 〃	8·68 〃	7·07 〃	450
500∼450	9·50 〃	7·05 〃	18·18 〃	14·12 〃	500
550∼500	14·42 〃	14·4 〃	32·60 〃	28·5 〃	550
600∼550	5·34 〃	9·50 〃	37·94 〃	38·0 〃	600
650∼600	0·00 〃	0·10 〃	37·94 〃	38·1 〃	650
700∼650	0·09 〃	−0·10 〃	38·03 〃	38·0 〃	700

Table III–2–IV. Specimen No. 24. ($H=0{\cdot}92$ *Oe.*)

Range of Temp. Δt	$\frac{\Delta J}{H}$	$\frac{\partial J}{H\cdot\partial t}\times\Delta t$	$\Sigma\frac{\Delta J}{H}$	$\frac{J}{H}$	Temp.
100∼50°C	$0{\cdot}10\times10^{-3}$	$0{\cdot}12\times10^{-3}$	$0{\cdot}10\times10^{-3}$	$0{\cdot}12\times10^{-3}$	100°C
150∼100	0·21 〃	0·13 〃	0·31 〃	0·25 〃	150
200∼150	0·21 〃	1·62 〃	0·52 〃	1·87 〃	200
250∼200	3·54 〃	3·01 〃	4·06 〃	4·88 〃	250
300∼250	1·64 〃	0·98 〃	5·70 〃	5·86 〃	300
350∼300	0·16 〃	0·10 〃	5·86 〃	5·96 〃	350
400∼350	0·21 〃	0·23 〃	6·07 〃	6·19 〃	400
450∼400	0·13 〃	0·08 〃	6·20 〃	6·27 〃	450
500∼450	0·16 〃	0·17 〃	6·36 〃	6·44 〃	500
550∼500	0·98 〃	0·53 〃	7·34 〃	6·97 〃	550
600∼550	0·44 〃	0·83 〃	7·78 〃	7·80 〃	600
650∼600	0·04 〃	0·05 〃	7·82 〃	7·85 〃	650
700∼650	0·05 〃	−0·05 〃	7·87 〃	7.80 〃	700

Table III-2-V. Specimen No. 28. ($H=0{\cdot}92$ *Oe.*)

Range of Temp. Δt	$\frac{\Delta J}{H}$	$\frac{\partial J}{H\cdot\partial t}\times\Delta t$	$\Sigma\frac{\Delta J}{H}$	$\frac{J}{H}$	Temp.
100~50°C	$0{\cdot}11\times10^{-3}$	—	$0{\cdot}11\times10^{-3}$	$0{\cdot}13\times10^{-3}$	100°C
150~100	0·20 〃	$0{\cdot}15\times10^{-3}$	0·31 〃	0·28 〃	150
200~150	0·63 〃	1·12 〃	0·94 〃	1·40 〃	200
250~200	1·32 〃	1·30 〃	2·26 〃	2·90 〃	250
300~250	1·16 〃	0·80 〃	3·42 〃	3·70 〃	300
350~300	0·20 〃	0·30 〃	3·62 〃	4·00 〃	350
400~350	0·20 〃	0·16 〃	3·82 〃	4·16 〃	400
450~400	0·28 〃	0·17 〃	4·10 〃	4·33 〃	450
500~450	0·46 〃	0·18 〃	4·56 〃	4·51 〃	500
550~500	1·44 〃	1·40 〃	6·00 〃	5·91 〃	550
600~550	1·69 〃	1·38 〃	7·69 〃	7·29 〃	600
650~600	0·03 〃	0·09 〃	7·72 〃	7·38 〃	650
700~650	0·00 〃	0·02 〃	7·72 〃	7·40 〃	700

This relation holds for any value of t, regardless of whether the mode of thermo-remanent magnetism is of the standard or of the

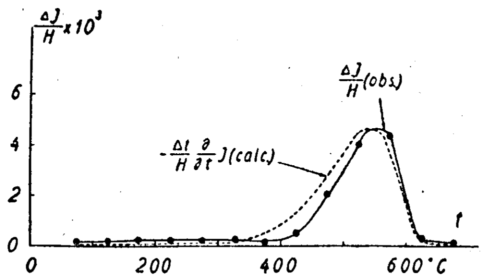

Fig. III-2-1. Specimen No. 21.

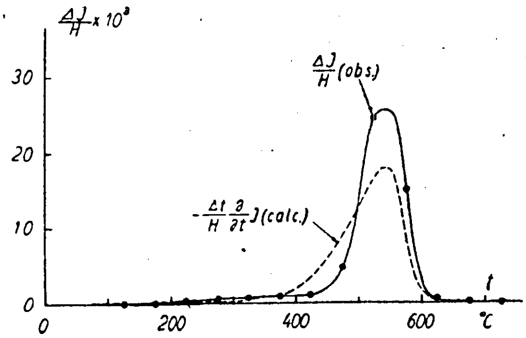

Fig. III-2-2. Specimen No. 60.

complex mode, although there is a discrepancy of a few per cent between $J_{t,H}$ and the sum of $\Delta J_{t_i,H}^{t_{i-1}}$, while the general tendency of the $J_{t,H}\sim t$ relation and that of $\sum_i \Delta J_{t_i,H}^{t_{i-1}}\sim t_i$ seems always to be parallel.

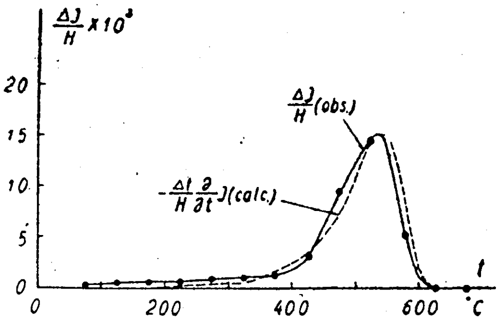

Fig. III-2-3. Specimen No. 60′.

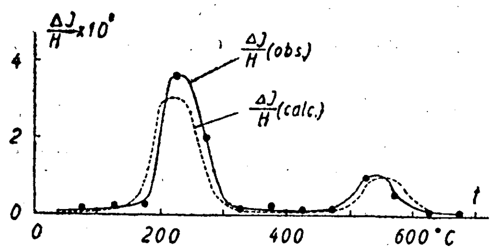

Fig. III-2-4. Specimen No. 24.

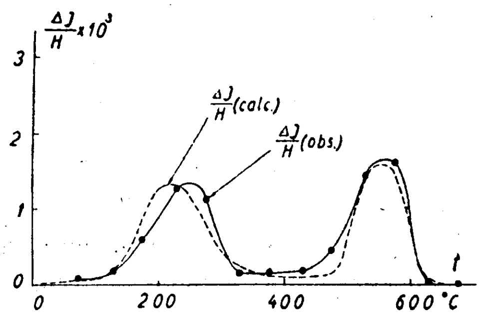

Fig. III-2-5. Specimen No. 28.

Neglecting the small discrepancy, as it may probably be the result of experimental error, we can establish the following relation, as a first approximation,

$$\sum_{t_{i-1}=0}^{t_i=t} \Delta J_{t_i,H}^{t_{i-1}} = J_{t,H} \tag{2-2}$$

Conversely, we compare $\Delta J_{t_i,H}^{t_{i-1}}$ with the gradient of the $J_{t,H}\sim t$ curve for the same range of temperature, $\Delta t = t_i - t_{i-1}$. The difference $J_{t_i, H} - J_{t_{i-1}, H}$ and the value of $\Delta J_{t_i, H}^{t_{i-1}}$ of the same specimen are given in Tables 2–I~2–V and in Figs. 2–1~2–5, Δt being 50 degrees in

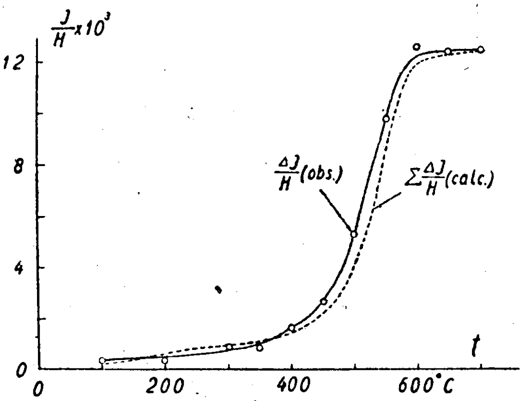

Fig. III-2-6. Specimen No. 21.

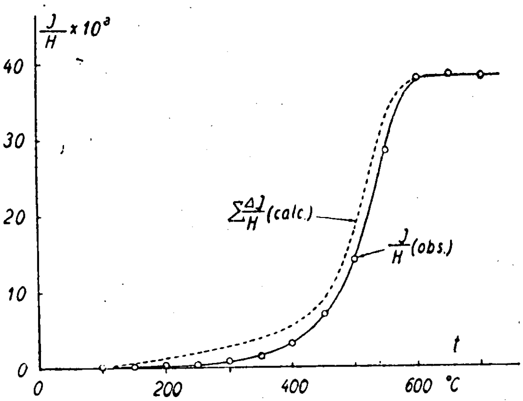

Fig. III-2-7. Specimen No. 60′.

actual experiments. Although there is also a small discrepancy between these two quantities for the same temperature range and in the same magnetic field, the general tendency is identical. Hence, we can write approximately

$$J_{t_i, H} - J_{t_{i-1}, H} = \Delta J_{t_i, H}^{t_{i-1}} \qquad (2\text{–}3)$$

In the foregoing results, the magnitude of the partial temperature range was always taken as 50 degrees. As the next step, the magnitude of the partial magnetism during cooling which corresponds to a temperature range of 25 degrees will be dealt with. The observed values are given in Table 2–VI and in Figs. 2–11, 2–12.

In this experiment, the temperature range of 50 degrees for the preceding experiment was divided into two equal ranges of $\Delta'_1 t = \Delta'_2 t = 25$ degrees. The partial magnetizations during cooling corresponding

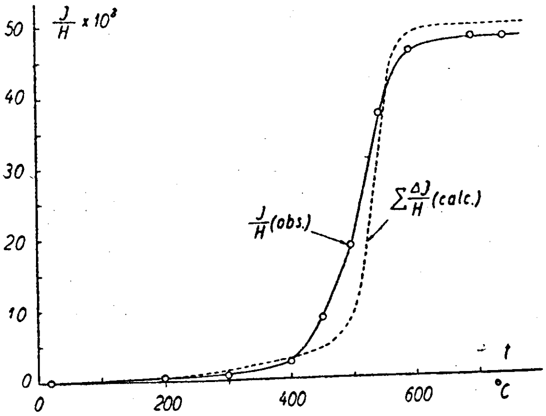

Fig. III-2-8. Specimen No. 60.

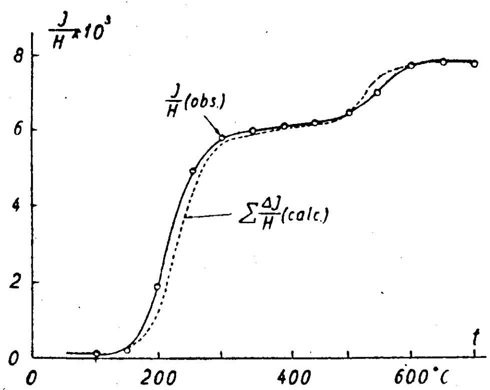

Fig. III-2-9. Specimen No. 24.

to $\Delta'_1 t$ and $\Delta'_2 t$ are denoted respectively by $\Delta' J^{t_{i-1}}_{t'_i, H}$ and $\Delta' J^{t_i}_{t'_i, H}$, where $t'_i = \frac{t_i + t_{i-1}}{2}$. Comparing, then, these values with $\Delta J_{t_i, H}$ of the same rock specimen, we get approximately for any value of t_i

$$\Delta' J^{t_{i-1}}_{t'_i, H} + \Delta' J^{t'_i}_{t_i, H} = \Delta J^{t_{i-1}}_{t_i, H}. \qquad (2\text{–}4)$$

The validity of this relation is graphically shown in Fig. 2–11. Needless to say, the law of composition of thermo-remanent magnetism,

$$\sum_{t_{i-1}=0}^{t_i'=t}\left(\Delta' J_{t_i,\,H}^{t_{i-1}'} + \Delta' J_{t_i',\,H}^{t_i}\right) = J_{t,\,H}, \tag{2-2'}$$

and the law of its decomposition

$$\Delta' J_{t_i,\,H}^{t_{i-1}'} = J_{t_i',\,H} - J_{t_{i-1},\,H}, \qquad \Delta' J_{t_i,\,H}^{t_i'} = J_{t_i,\,H} - J_{t_i',\,H}, \tag{2-3'}$$

also hold approximately in this case, as shown in Fig. 2-12. Thus, the laws of the composition and decomposition of thermo-remanent magnetism are established as a first approximation in our quest.[1] Since it was practically difficult to measure quantitatively the partial

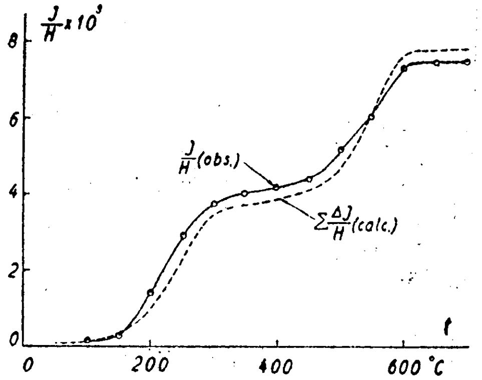

Fig. III-2-10. Specimen No. 28.

magnetization during cooling corresponding to the narrower temperature range by means of the present apparatus, whether the composition law expressed by eqs. (2-2) and (2-4) and the decomposition law expressed by eq. (2-3) hold or not in partial magnetization during cooling corresponding to temperature range narrower than 25 degrees, has not been ascertained. However, it will be concluded from the foregoing experimental results that the relation

$$\Delta J_{t_i,\,H}^{t_{i-1}} \doteqdot -\left(\frac{\partial}{\partial t} J_{t,\,H}\right)_{t=\bar{t}_i} \times \Delta t \tag{2-5}$$

holds for any temperature range $\Delta t = t_i - t_{i-1}$ at any mean temperature $\bar{t}_i = \frac{1}{2}(t_i + t_{i-1})$, provided Δt does not greatly exceed 50 degrees, or is perhaps not very much less than 25 degrees.

Now, assuming that eq. (2-5) holds in the extreme case that

1) Another example of the proof of this law is given in the writer's previous paper. T. NAGATA, *Bull. Earthq. Res. Inst.* **19** (1941), 49.

Table III-2-VI. Specimen No. 60′. ($H=0{\cdot}92\ Oe.$)

Range of Temp.	$\Delta'J$	$\Delta'J_1+\Delta'J_2$	ΔJ	$\Sigma\Delta'J$	J	Temp.
75~50°C	$0{\cdot}00\times10^{-3}$	$0{\cdot}05\times10^{-3}$	$0{\cdot}12\times10^{-3}$	$0{\cdot}00\times10^{-3}$	$0{\cdot}00\times10^{-3}$	20°C
100~75	0·05 〃			0·05 〃	0·06 〃	100
125~100	0·15 〃	0·28 〃	0·24 〃	0·20 〃		
150~125	0·13 〃			0·33 〃		150
175~150	0·13 〃	0·24 〃	0·27 〃	0·46 〃		
200~175	0·11 〃			0·57 〃	0·16 〃	200
225~200	0·13 〃	0·26 〃	0·27 〃	0·70 〃		
250~225	0·13 〃			0·83 〃		250
275~250	0·13 〃	0·24 〃	0·44 〃	0·96 〃		
300~275	0·11 〃			1·07 〃	0·47 〃	300
325~300	0·11 〃	0·24 〃	0·48 〃	1·18 〃		
350~325	0·13 〃			1·31 〃	0·75 〃	350
375~350	0·19 〃	0·32 〃	0·66 〃	1·50 〃		
400~375	0·13 〃			1·63 〃	1·50 〃	400
425~400	0·21 〃	1·16 〃	1·50 〃	1·84 〃		
450~425	0·95 〃			2·79 〃	3·24 〃	450
475~450	1·81 〃	4·37 〃	4·35 〃	4·60 〃		
500~475	2·56 〃			7·16 〃	6·52 〃	500
525~500	3·45 〃	7·14 〃	6·61 〃	10·61 〃		
550~525	3·69 〃			14·30 〃	13·10 〃	550
575~550	2·82 〃	2·95 〃	2·45 〃	17·12 〃		
600~575	0·13 〃			17·25 〃	17·46 〃	600
625~600	0·03 〃	0·03 〃	0·00 〃	17·28 〃		
650~625	0·00 〃			17·28 〃	17·50 〃	650
675~650	0·00 〃	0·00 〃	0·04 〃	17·28 〃		
700~675	0·00 〃			17·28 〃	17·46 〃	700

$\Delta t\to0$, we get

$$\lim_{\Delta t\to0}\frac{\Delta J^{t_{i-1}}_{t_i,\,H}}{\Delta t}=-\left(\frac{\partial}{\partial t}J_{t,\,H}\right)_{t=t_i}\equiv P(t_i,H)\,, \tag{2-6}$$

where P is a function of temperature, magnetic field, and the petrological characteristic of the rock. Eq. (2–6) is believed to be a good approximation for the mode of development of thermo-remanent magnetism in volcanic rock, so long as we deal with its phenomenological character alone, although whether or not it exactly holds in the extreme case in which $\Delta t\to0$ is not clear.

On the other hand, the intensity of $\Delta J^{t_{i-1}}_{t_i,\,H}$ corresponding to a given temperature range is almost proportional to that of the applied magnetic field H, provided H is small. The relation of thermo-remanent magnetization to the applied magnetic field, as established in the preceding paragraph, is given by

$$\vec{J}_{t,H} = F(\vec{H}) \cdot J_t$$

for the general case, and

$$\vec{J}_{t,H} = \vec{H} \cdot J_t,$$

when H is small.

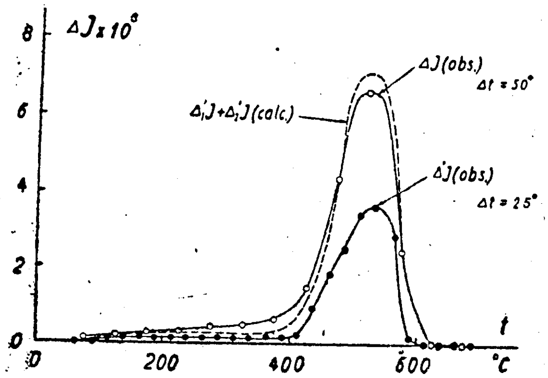

Fig. III-2-11. Specimen No. 60′.

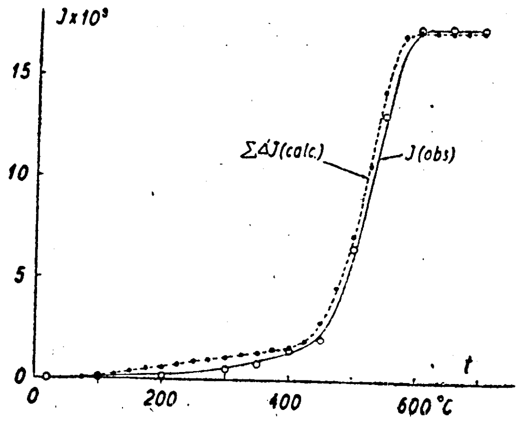

Fig III-2-12. Specimen No. 60′.

Since thermo-remanent magnetism is given by the sum of the partial magnetism during cooling for each temperature range, obtained by dividing the whole range into minor ranges through which the rock must cool in order to bring about the thermo-remanent magnetism, the effect of the applied magnetic field on partial magnetism during cooling will be regarded as being quite similar to that on the thermo-remanent magnetism.

Consequently, we get for the case where H is small,

$$\Delta \vec{J}^{t_{i-1}}_{t_i,\,H} = \vec{H} \cdot \Delta J^{t_{i-1}}_{t_i}, \tag{2-7}$$

$$\sum_{t_{i-1}=0}^{t_i=t} \Delta J^{t_{i-1}}_{t_i} = J_t, \tag{2-8}$$

$$\lim_{\Delta t \to 0} \frac{\Delta \vec{J}^{t_{i-1}}_{t_i,\,H}}{\Delta t} = -\vec{H} \cdot \left(\frac{\partial}{\partial t} J_t\right)_{t=t_i} = \vec{H} \cdot P(t_i)\ , \tag{2-9}$$

where $P(t)$ is a function of temperature as well as of the petrological character of the rock specimen, and called here the characteristic function of thermo-remanent magnetism.

Summarizing, now, the facts proved in the foregoing experiments, we establish the relation

$$\vec{J}_{t,\,H} = \vec{H} \cdot J_t = H \cdot \int_0^t P(t)\,dt\,. \tag{2-10}$$

More generally, if the residual permanent magnetization at 0°C, developed during cooling from t to t' in a magnetic field of H, (that is, in the process of cooling, magnetic field H is applied during cooling from t to t' alone), is dentoed by $J^{t'}_{t\ H}$, we get

$$\vec{J}^{t'}_{t,\,H} = \vec{H} \cdot \int_{t'}^{t} P(t)\,dt \qquad (t' \leqq t) \tag{2-11}$$

Here, the general tendency of function $P(t)$ is given by the curves shown in Figs. 2–1∼2–4, that is, $P(t)$ takes maximum value at one or two points with respect to temperature, which last, needless to say, corresponds to transition temperature t_0 of thermo-remanent magnetization, which was defined as the temperature at which $\frac{\partial}{\partial t} J_{t,\,H}$ is maximum. That is to say,

$$P(t_0) = \text{maximum}\,. \tag{2-12}$$

Next, for the critical temperatuer t_c, we obtained

$$P(t \geqq t_c) = 0\,. \tag{2-13}$$

Whence, the saturated thermo-remanent magnetization in H, $J_{t_c,\,H}$, which was defined in the preceding paragraph, is given by

$$\vec{J}_{t_c, H} = \vec{H} \cdot \int_0^{t_c} P(t)\,dt\,. \tag{2-14}$$

Thus, from our studies of the general phenomenological mode of development of thermo-remanent magnetization in volcanic rocks, we conclude that this residual permanent magnetization is the result of superposition of many irreversible magnetizations caused in the various temperature ranges during cooling in a magnetic field H, and that it is not merely the residual part of an intense magnetization at a particular temperature in the process of cooling.

(ii) In order to describe quantitatively the mode of development of thermo-remanet magnetism the particular form of the characteristic function would be the most fundamental property. Actually, the functional form of $P(t)$ of various rock specimens were determined with the aid of the approximate relation

$$P\left(\frac{t_i + t_{i-1}}{2}\right) \doteq \frac{1}{H} \cdot \frac{\Delta J^{t-1}_{t_i, H}}{\Delta t},$$

where Δt was frequently taken as 50 degrees. The actual $P(t)$ at various temperatures thus obtained are given in Tables 2–VII∼2–IX and in Figs. 2–13 and 2–15, from which it will be seen that, generally speaking, the mode of $P(t)$ is classified into two types, one the standard mode and the other the complex mode (the superposition of

Table III–2–VII. $P(t)$ of ejecta from Volcano Huzi.

Specimen / Range in Temp.	No. 17	No. 18	No. 19	No. 20	No. 21	No. 22	No. 23	No. 24	No. 25
°C 100∼ 50	0·28 $\times 10^{-5}$	0·24 $\times 10^{-5}$	0·26 $\times 10^{-5}$	0·20 $\times 10^{-5}$	0·36 $\times 10^{-5}$	0·08 $\times 10^{-5}$	0·26 $\times 10^{-5}$	0·20 $\times 10^{-5}$	0·40 $\times 10^{-5}$
150∼100	0·18 ″	0 32 ″	0·36 ″	0·08 ″	0·40 ″	2·38 ″	0 30 ″	0·42 ″	3·08 ″
200∼150	0·28 ″	0·32 ″	0·28 ″	0·18 ″	0·46 ″	6·66 ″	0·42 ″	0·42 ″	6·74 ″
250∼200	0·42 ″	3·98 ″	2·66 ″	0·12 ″	0·40 ″	11·68 ″	0·42 ″	7·08 ″	3·18 ″
300∼250	0·44 ″	4·74 ″	1·02 ″	0·50 ″	0·26 ″	9·74 ″	0·88 ″	3·28 ″	2·34 ″
350∼300	1·24 ″	0·88 ″	0·98 ″	0·24 ″	0·46 ″	5·68 ″	0·10 ″	0·32 ″	0·98 ″
400∼350	2·34 ″	0·96 ″	0·62 ″	0·66 ″	0·30 ″	5·36 ″	0·52 ″	0·42 ″	0·84 ″
450∼400	1·90 ″	0·88 ″	0·80 ″	0·66 ″	0·90 ″	6·72 ″	0·16 ″	0·26 ″	1·02 ″
500∼450	4·98 ″	1·60 ″	2·08 ′	4·18 ″	4·02 ″	8·52 ″	2·12 ″	0 32 ″	3·50 ″
550∼500	10 04 ″	5·08 ″	4·64 ″	8·36 ″	[illegible]·98 ″	15·10 ″	3·60 ″	1·96 ″	4·16 ″
600∼550	3·74 ″	4·50 ″	5·28 ″	5·62 ″	8·44 ″	10·36 ″	1·40 ″	0·88 ″	3 08 ″
650∼600	0·34 ″	0·16 ″	0·28 ″	0·24 ″	0·58 ″	1·28 ″	0 06 ″	0·08 ″	0 18 ″
700∼650	0·18 ″	0·22 ″	0·06 ″	0·08 ″	0·26 ″	0·28 ″	0·16 ″	0·10 ″	0·00 ″
750∼700	—	—	—	—	—	0·42 ″	—	—	—

Table III-2-VIII. $P(t)$ of various volcanic rocks.

Specimen / Range in Temp.	No. 8	No. 27	No. 28	No. 35	No. 36	No. 38	No. 93	No. 94	No. 25′
°C 100~ 50	0·17 $\times10^{-5}$	0·02 $\times10^{-5}$	0·22 $\times10^{-5}$	0·08 $\times10^{-5}$	0·21 $\times10^{-5}$	0·23 $\times10^{-5}$	0·80 $\times10^{-5}$	0·52 $\times10^{-5}$	0·58 $\times10^{-5}$
150~100	0·16 ″	0·04 ″	0·40 ″	0·18 ″	0·31 ″	1·60 ″	0·24 ″	0·58 ″	0·34 ″
200~150	0·23 ″	0·11 ″	1·26 ″	0·18 ″	0·31 ″	3·06 ″	0·19 ″	0·52 ″	0·28 ″
250~200	0·26 ″	0·09 ″	2·64 ″	0·18 ″	0·37 ″	5·81 ″	0·24 ″	0·24 ″	0·44 ″
300~250	0·39 ″	0·13 ″	2·32 ″	0·28 ″	0·25 ″	6·88 ″	0·34 ″	0·15 ″	0·52 ″
350~300	0·40 ″	0·13 ″	0·40 ″	0·21 ″	0·37 ″	5·87 ″	0·35 ″	0·09 ″	0·42 ″
400~350	0·33 ″	0·26 ″	0·40 ″	0·34 ″	0·37 ″	3·23 ″	0·39 ″	0·00 ″	0·46 ″
450~400	0·53 ″	0·26 ″	0·56 ″	0·29 ″	0·55 ″	3·29 ″	0·52 ″	0·00 ″	0·58 ″
500~450	0·54 ″	1 56 ″	0·92 ″	0·57 ″	0·92 ″	6·70 ″	0·80 ″	0·45 ″	2·24 ″
550~500	2·14 ″	2·93 ″	2·88 ″	2·69 ″	2.21 ″	10·08 ″	2·24 ″	2·30 ″	2·70 ″
600~550	1·05 ″	1·98 ″	2·38 ″	2·00 ″	2·42 ″	5·88 ″	2·91 ″	4·35 ″	1·16 ″
650~600	0·13 ″	0·04 ″	0·06 ″	0·21 ″	0·00 ″	0 08 ″	0·15 ″	0·32 ″	0·24 ″
700~650	0·00 ″	0·05 ″	0·05 ″	0·04 ″	0·01 ″	0·04 ″	0·00 ″	0·00 ″	0·00 ″

Table III-2-IX. $P(t)$ of various volcanic rocks.

Specimen / Range in Temp.	No. 16	No. 45	No. 58	No. 59	No. 60	No. 60′	No. 73	No. 75
°C 100~ 50	0·07 $\times10^{-5}$	0·00 $\times10^{-5}$	0·30 $\times10^{-5}$	0·64 $\times10^{-5}$	0·20 $\times10^{-5}$	0·51 $\times10^{-5}$	0·00 $\times10^{-5}$	0·14 $\times10^{-5}$
150~100	0·15 ″	0·48 ″	0·08 ″	0·60 ″	0·60 ″	1·04 ″	0·00 ″	0·56 ″
200~150	0 54 ″	2·72 ″	0·30 ″	0·74 ″	0·92 ″	1·18 ″	11·47 ″	0·82 ″
250~200	1·76 ″	5·30 ″	3·08 ″	3·60 ″	0·92 ″	1·18 ″	8·58 ″	1·08 ″
300~250	0·15 ″	13·86 ″	12·46 ″	12·76 ″	1·12 ″	1·92 ″	1·09 ″	13·10 ″
350~300	0·20 ″	14·02 ″	9·98 ″	30·90 ″	1·38 ″	2·10 ″	0·83 ″	10·42 ″
400~350	0·24 ″	5·46 ″	6·90 ″	10·58 ″	1·86 ″	2·88 ″	0·87 ″	3·60 ″
450~400	0·46 ″	3·20 ″	4·52 ″	5·08 ″	1·90 ″	6·54 ″	0·57 ″	0·78 ″
500~450	0·83 ″	5·74 ″	8·68 ″	4·26 ″	9·54 ″	19·00 ″	0·39 ″	0·82 ″
550~500	1·00 ″	7·53 ″	7·46 ″	4·64 ″	49·48 ″	28·84 ″	0·96 ″	0 82 ″
600~550	0·30 ″	2·96 ″	4·16 ″	1·52 ″	29·78 ″	10·68 ″	0·13 ″	0·78 ″
650~600	0·06 ″	0·40 ″	0·86 ″	0·52 ″	0·60 ″	0·00 ″	0·00 ″	0·08 ″
700~650	0·00 ″	0·06 ″	0·44 ″	0·66 ″	0·60 ″	0·19 ″	0 00 ″	0·00 ″
750~700	—	—	0·30 ″	0·34 ″	0·34 ″	—	—	—

the standard and extraordinary modes). To the former correspond one transition temperature t_0 and one critical temperature t_c, while to the latter correspond two transition temperatures and two critical temperatures, being denoted by t_{os}, t_{oe}, and t_{cs}, t_{ce} for the component of the standard mode and that of the extraordinary mode respectively.

As will be clear from the relation between $P(t)$ and $J_{t,H}$, all these characteristic temperatures ought exactly to identify the same properties named in the mode of the $J_{t,H}\sim t$ relation.

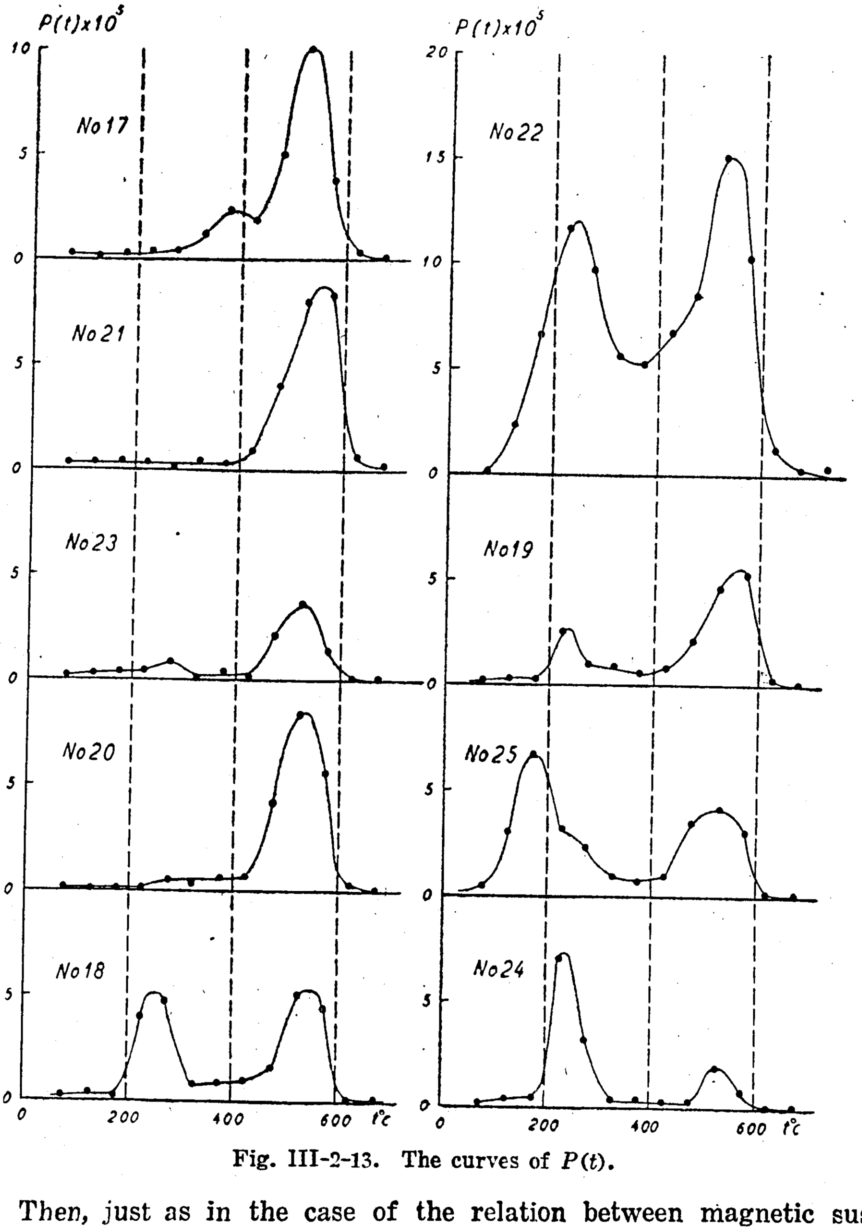

Fig. III-2-13. The curves of $P(t)$.

Then, just as in the case of the relation between magnetic susceptibility and temperature, $P(t)$ can generally be written

$$P(t)=P_s(t)+P_e(t)\,, \tag{2–15}$$

where

$$P_s(t_{os}) = \text{maximum}, \qquad P_e(t_{oe}) = \text{maximum},$$

$$P_s(t \geqq t_{cs}) = 0, \qquad P_e(t \geqq t_{ce}) = 0,$$

and each of the fundamental curves ($P_s(t)$ or $P_e(t)$) has a rather simple form, that is nearly the same as the error function with its centre at the transition temperatures t_{oe} and t_{os}. As already discussed

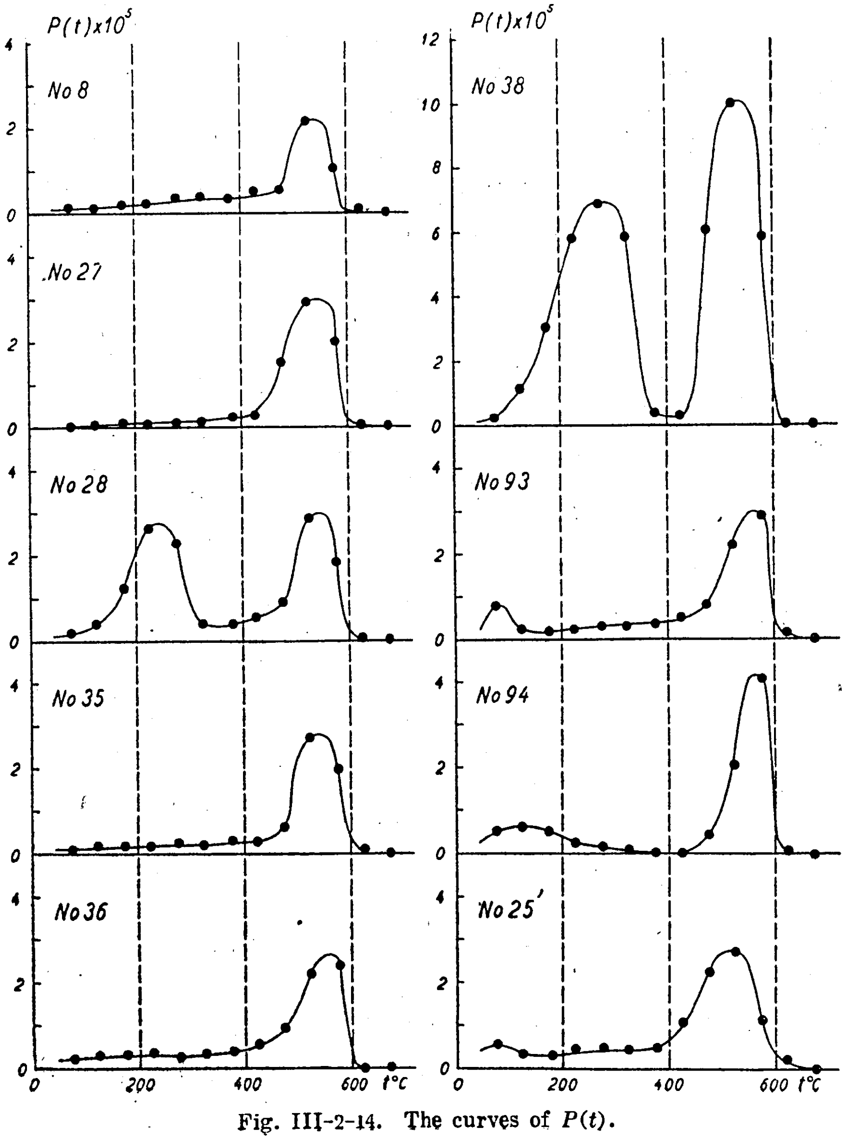

Fig. III-2-14. The curves of $P(t)$.

in the explanation of the $\chi(t) \sim t$ relation, it is presumed that the ferro-magnetic minerals in the rock consist of a few mineral groups

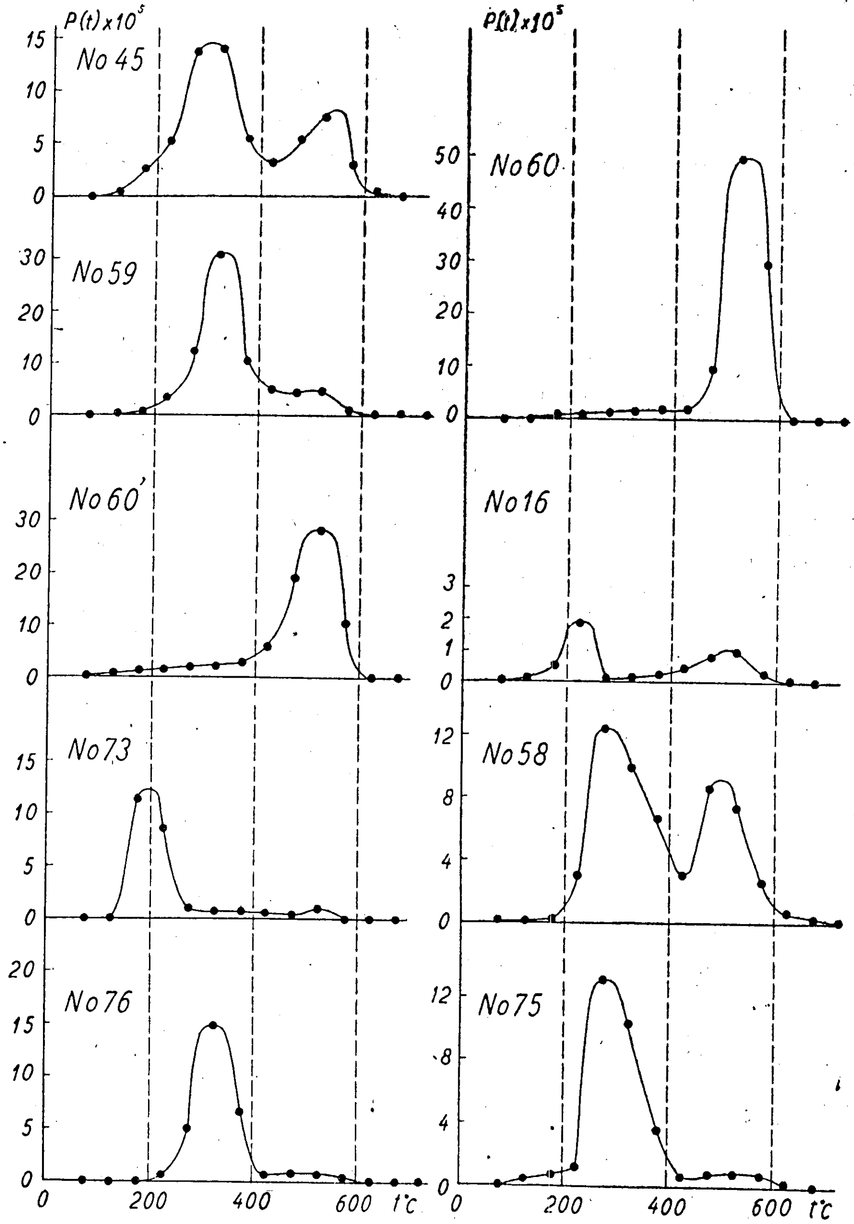

Fig. III-2-15. The Curves of $P(t)$

of different chemical composition, whence it is further presumed that the two fundamental modes of $P(t)$ correspond to these two mineral groups, so that in this case also, it may be possible that a $P(t)$ curve is composed of more than three fundamental curves, the transition temperatures of which are not alike, whence generally

$$P(t)=\sum_j P_j(t)$$

and

$$\vec{J}_{t,\,II}^{t'}=\vec{H}\cdot\int_{t'}^{t}\sum_j P_j(t)\,dt=\vec{H}\cdot\sum_j\int_{t'}^{t}P_j(t)\,dt. \qquad (2\text{–}16)$$

In many actual cases, however, the $P(t)$ function is approximately given by eq. (2–15), where $P_e(t)$ sometimes vanishes, while there are a few cases in which $P_e(t)$ alone appears, $P_s(t)$ vanishing, and in which $P(t)$ consists of more than three fundamental modes. Typical examples of these cases are given in Table 2–IX and in Fig. 2–15. That eqs. (2–10) and (2–11) hold in this case also was proved in the writer's previous paper.

However, so far as the present experimental data of the writer go, the occurrence of the latter two cases is limited to volcanic bombs of small size, i.e. $0{\cdot}5\sim1{\cdot}0\,m$ in mean diameter, which cooled very rapidly, with the result that crystalization of the minerals in the rock seems scarcely complete or perfect.

(iii) Effect of time required for cooling on thermo-remanent magnetism.

We must examine next the dependence of thermo-remanent magnetization on the time required for cooling. For this purpose, the intensity $\Delta J_{t_i,\,II}^{t_{i-1}}$ of a rock from Mihara Volcano, corresponding to a cooling range of from 400°C to 350°C in 0·458 *Oe.*, was measured, the cooling velocity being varied. In actual treatment, the time consumed in cooling through a range of 50 degrees was varied from about one minute to 80 minutes. The result of experiment is shown in Fig. 2–16 and in Table 2–X, from which we find that the intensity of partial magnetization during cooling for various periods of it seems to be nearly constant, provided the cooling velocity does not greatly exceed 40 degrees per hour, the deviation forming only a small percentage, with a tendency for $\Delta J_{t_i,\,II}^{t_{i-1}}$ to become smaller the shorter the time consumed in cooling. It must not be overlooked here that the experimental error may not be small, should the time consumed in cooling be shorter than ten minutes, since in this case the temperature distribution in the specimen will not be uniform. It will be concluded that, as a first approximation, the development of thermo-remanent magnetization is independent of the cooling velocity, whence we get

$$\frac{\partial}{\partial \tau}\left(\Delta J_{t_i, H}^{t_{i-1}}\right)=0,$$

with the result that

$$\frac{\partial}{\partial \tau}P(t)=0, \qquad (2\text{–}17)$$

where τ denotes the time required for cooling.

Needless to say, eq. (2–17) may hold where the cooling velocity is rather large compared with the small velocity required for maintaining (at any temperature during cooling) thermo-dynamical equilibrium between the many phases due to the various compositions shown by the rocks.

We may thus conclude that the general phenomenological formula must be established independent of the time of cooling, provided the external magnetic field applied is small and the cooling velocity is neither too small nor too large.

Table III-2-X The relation between cooling velocity and the thermo-remanent magnetization.

Cooling velocity	τ	ΔJ
degree/sec.	m s	
0·73	1 08	$3{\cdot}14\times10^{-3}$
0·089	9 20	3·47 〃
0·046	18 08	3·93 〃
0·024	35 12	3·78 〃
0·022	38 29	3·74 〃
0·011	77 33	3·95 〃

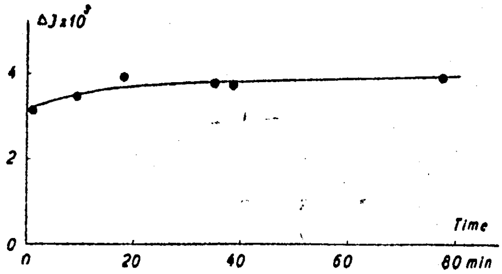

Fig. III-2-16. Relation between ΔJ_{tH} and the cooling time (τ) through Δt.

(iv) The law of vanishing of thermo-remanent magnetization with increase in temperature.

As the next step, we shall examine the general mode of change in thermo-remanent magnetization with increase in temperature, the experiment for studying which was made by two different methods, that described here being one which, however, is rather an indirect way for the purpose of the present study, while the other is dealt with in § 4 of this Chapter.

The method of experiment was as follows. A rock specimen that was heated to temperature higher than t_c in non-magnetic space was cooled to $0°C$ in a magnetic field of H, the specimen acquiring the residual permanent magnetization of saturated thermo-remanent magnetization in H, $\vec{J}_{t_c, H}$, as already proved. Next, the specimen with permanent magnetization of J was heated from 0°C to t in a magnetic field of H', and then cooled again to 0°C from t in non-magnetic space. The residual permanent magnetization, which is here denoted by $(\vec{J}_{t_c, H})_{t, \vec{H}'}$, usually differs from $\vec{J}_{t_c, H}$. The experiments were made with the same apparatus as that used for causing thermo-remanent magnetism.

Before dealing with the experimental results, it must be remembered that a rock specimen cooled from any temperature to *0°C* in non-magnetic space shows no residual magnetization. We may, therefore, presume that cooling from t to *0°C* in non-magnetic space scarcely participates in the change in magnetization, so far as thermo-remanent magnetization is concerned, the change being chiefly due to heating from *0°C* to t in H'.

In an actual experiment, the temperature was varied from 100°C to 700°C, an example of the measurement being given in Table 2–XI and in Fig. 2–17, where H' was taken as zero. For comparison, in Table 2–XI and in Fig. 2–17 are also given the values of the residual magnetization of the same rock specimen, which was obtained after successive heat treatments; after measuring $(\vec{J}_{t_c, H})_{t_1, 0}$, the specimen was again heated to t_2 in $H'=0$ and then cooled in non-magnetic space also, where the residual magnetization is denoted by $((J_{t_c, H})_{t_1, 0})_{t_2, 0}$. This process was repeated n-times. The residual magnetization after the n-th heat treatment is denoted generally by $((\cdots(J_{t_c, H})_{t_1, 0}\cdots)_{t_{n-1}, 0})_{t_n, 0}$, where $t_n > t_{n-1} > \cdots > t_1$. Then, as will be seen from the result in Table 2–XI, we get

$$(\cdots((\vec{J}_{t_c, H})_{t_1, 0})_{t_2, 0}\cdots)_{t_n, 0} = (\vec{J}_{t_c, H})_{t_n, 0}. \qquad (2\text{–}18)$$

Assuming then that the law given by eq. (2–18) holds in the general case, we examined by means of successive heat treatment the $(\vec{J}_{t_c, H})_{t_n, \vec{H}'}$ of other rock specimens for various values of t in the three cases $\vec{H}'=\vec{H}$, $\vec{H}'=-\vec{H}$, and $H'=0$, the result being given in Table 2–XII and in Fig. 2–18, from which we get

$$(\vec{J}_{t_c, H})_{t_n, H} = (\vec{J}_{t_c, H})_{t_n, -H} = (\vec{J}_{t_c, H})_{t_n, 0}, \qquad (2\text{–}19)$$

whence, extending the conclusion given by eq. (2–19), we assume that for any value of H'

$$(\vec{J}_{t_c, H})_{t_n, \vec{H}'} = (\vec{J}_{t_c, H})_{t_n, 0}, \qquad (2\text{–}20)$$

Table III-2-XI. $(J_{t_e,H})_{t_n,0}$. Specimen No. 60′ ($H=+0{\cdot}46$ *Oe.*)

Temp.	$(J_{t_c,H})_{t_n,0}\times10^3$							$(\cdots(J_{t_c,H})_{t_1,0}\cdots)_{t_n,0}\times10^3$
20°C	17·40	17·40	17·36	17·33	17·38	17·45	17·38	17·55
200	17·21	↓						17·41
300		17·10	↓					17·20
400			15·92	↓				16·22
500				12·38	↓			11·17
550					4·63	↓		4·23
600						0·11	↓	0·19
700							0·04	0·02

Table III-2-XII. $(\cdots(J_{t_cH})_{t_1H'}\cdots)_{t_nH'}$ in the three cases of $H'=0$, $H'=+0{\cdot}46$ *Oe.* and $H'=-0{\cdot}46$ *Oe.* ($H=+0{\cdot}46$ *Oe.*) Specimen No. 60′.

Temp.	$H'=0$	$H'=+0{\cdot}46$ *Oe.*	$H'=-0{\cdot}46$ *Oe.*
20°C	$17{\cdot}55\times10^{-3}$	$17{\cdot}65\times10^{-3}$	$17{\cdot}71\times10^{-3}$
100	17·50 ″	——	——
200	17·41 ″	17·48 ″	17·68 ″
300	17·20 ″	17·12 ″	17·40 ″
350	16·86 ″		——
400	16·22 ″	16·10 ″	16·02 ″
450	14·25 ″	14·46 ″	14·58 ″
500	11·17 ″	11·53 ″	12·10 ″
550	4·23 ″	5·24 ″	4·30 ″
600	0·19 ″	0·13 ″	0·16 ″
650	0·08 ″	——	——
700	0·02 ″	0·05 ″	0·05 ″

holds for any value of H, provided H and H' are small. That is to say, $(\vec{J}_{t_c,H})_{t_n,\vec{H'}}$ is independent of the magnetic field $\vec{H'}$ applied while heating the specimen, whence we shall omit the suffix H' in the following description.

Comparing, next, the value of $(J_{t_c,H})_{t_n}$ with that of $\Delta J^{t_{i-1}}_{t_i,H}$ of the same rock specimen, we get, as shown in Table 2-XIII and in Figs. 2-19, 20,

$$\Delta\vec{J}^{t_{i-1}}_{t_i,H}=(\vec{J}_{t_c,H})_{t_{i-1}}-(\vec{J}_{t_c,H})_{t_i}\,, \tag{2-21}$$

with the result that

$$(\vec{J}_{t_c,H})_t-(\vec{J}_{t_c,H})_{t'}=-\vec{H}\cdot\int_{t'}^{t}P(t)\,dt\,, \tag{2-22}$$

or in another form, putting

$$\lim_{\Delta t \to 0} \frac{(\vec{J}_{t_c,H})_t - (\vec{J}_{t_c,H})_{t'}}{\Delta t} = \frac{\partial}{\partial t}(\vec{J}_{t_c,H}) ,$$

we get

$$\frac{\partial}{\partial t}(\vec{J}_{t_c,H})_t = -\vec{H} \cdot P(t) . \tag{2-23}$$

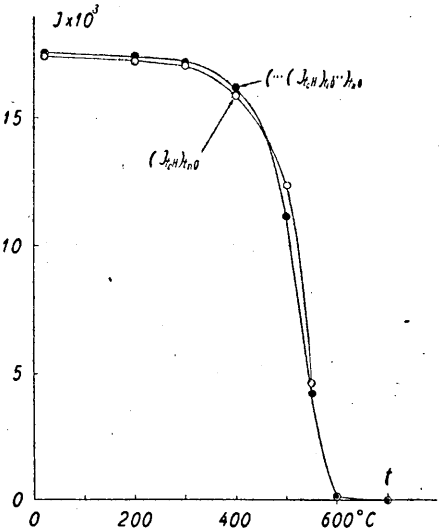

Fig. III-2-17. The mode of vanishing of thermo-remanent magnetization. Relation between $(J_{t_cH})_{t_n0}$ and $(\cdots(J_{t_cH})_{t_10}\cdots)_{t_n0}$. Specimen No. 60′.

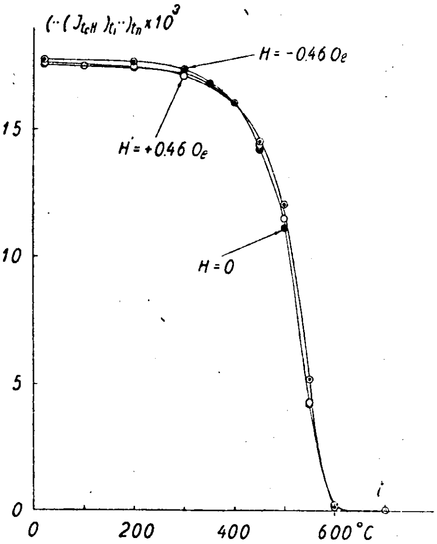

Fig. III-2-18. $(\cdots(J_{t_cH})_{t_1H'}\cdots)_{t_nH'}$ in the three cases of $H'=0$, $+0.46$ and -0.46 *Oe*. Specimen No. 60′.

We may find, then, that the law given by eq. (2–18) is the result naturally derived from the conclusion given by eq. (2–23), since

$$(\vec{J}_{t_c,H})_{t_1} = \vec{J}_{t_c,H} - \vec{H} \cdot \int_0^{t_1} P(t)\,dt$$

$$((\vec{J}_{t_c,H})_{t_1})_{t_2} = \vec{J}_{t_c,H} - \vec{H} \cdot \int_0^{t_1} P(t)\,dt$$

$$-0 \times \int_0^{t_1} P(t)\,dt - \vec{H} \cdot \int_{t_1}^{t_2} P(t)\,dt = \vec{J}_{t_c,H} - \vec{H} \cdot \int_0^{t_2} P(t)\,dt$$

$$((\cdots(\vec{J}_{t_c,H})_{t_1}\cdots)_{t_{n-1}})_{t_n} = \vec{J}_{t_c,H} - \int_0^{t_n} P(t)\,dt = (\vec{J}_{t_c,H})_{t_n} .$$

Thus, the general law of disappearance of thermo-remanet magnetization with increase in temperature, is established, the conclusion being

Table III–2–XIII. Comparison of the mode of vanishing of thermo-remanent magnetism accompanying heating with that of its development accompanying cooling.
(Specimen No 60′, H=0·46 Oe.)

Temp.	$\frac{1}{H}(\cdots(J_{tcH})_{t_10}\cdots)_{tn0}$	$\frac{1}{H}(J_{tcH}-J_{tH})$	$\sum_{t=700}^{t-1=t}\frac{\Delta J}{H}$	$\frac{1}{H}(\cdots(J_{tH})_{t_10}\cdots)_{t_{n-1}0} - \frac{1}{H}(\cdots(J_{tcH})_{t_10}\cdots)_{t_n0}$	$\frac{1}{H}(J_{t_nH}-J_{t_{n-1}H})$	$\frac{\Delta J}{H}$	Range in Temp.
700°C	0 04×10⁻³	0·00×10⁻³	0 00×10⁻³	0·13×10⁻³	−0·1×10⁻³	0 09×10⁻³	700∼650°C
650	0·17 〃	−0·1 〃	0·09 〃	0·24 〃	0·1 〃	0·00 〃	650∼600
600	0·41 〃	0·0 〃	0·09 〃	8·79 〃	9·5 〃	5 34 〃	600∼550
550	9·20 〃	9·5 〃	5·41 〃	15·05 〃	14·4 〃	14·42 〃	550∼500
500	24·25 〃	23·9 〃	19·83 〃	6·75 〃	7·0 〃	9 50 〃	500∼450
450	31·0 〃	30·9 〃	29·33 〃	4·2 〃	3·8 〃	3·27 〃	450∼400
400	35·2 〃	35·7 〃	32 60 〃	1·4 〃	1·8 〃	1·44 〃	400∼350
350	36·6 〃	36·5 〃	34·04 〃	0·8 〃	0·5 〃	1·05 〃	350∼300
300	37·4 〃	37·0 〃	35·09 〃	0·3 〃	0 4 〃	0·96 〃	300∼250
250	——	——	36 05 〃	0 1 〃	0·2 〃	0·59 〃	250∼200
200	37·8 〃	37·7 〃	36·64 〃	0·1 〃	0·1 〃	0 59 〃	200∼150
150	——	——	37·23 〃	0 1 〃	0·1 〃	0·52 〃	150∼100
100	38·0 〃	37·9 〃	37·75 〃				

that the thermo-remanent magnetization which was caused by cooling the specimen from t to t' in H disappears with increase in tempera-

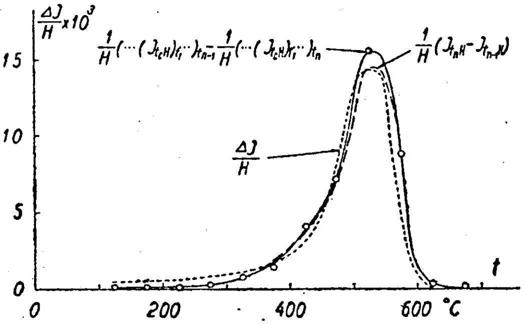

Fig. III-2-19. Comparison of the amount of $\frac{1}{H}\left[(\cdots(J_{t_c}H)_{t_1}\cdots)_{t_{n-1}}-(\cdots(J_{t_c}H)_{t_1}\cdots)_{t_n}\right]$ with that of $\frac{\Delta J}{H}$. Specimen No. 60′.

ture from t' to t, regardless of the intensity or direction of the magnetic field applied during heating. This conclusion of the observed facts may suggest that, phenomenologically speaking, the physical state that is capable of being permanently magnetized as $P(t_i)\,\Delta t$ corresponds to the elemental cooling range Δt at t_i, the external magnetic field $\vec{H}$ affecting the physical state in such a way as to result in permanent magnetization $\vec{H}\cdot P(t_i)\Delta t$ (at 0°C), whereas the said physical state vanishes when it corresponds to heating of Δt at t_i, resulting in the disappearance of magnetization $\vec{H}\cdot P(t_i)\,\Delta t$, independent of the magnetic field H' which is applied during heating; that is to say, the development and disappearance of physical state $P(t_i)\,\Delta t$ is reversible with respect to change in temperature.

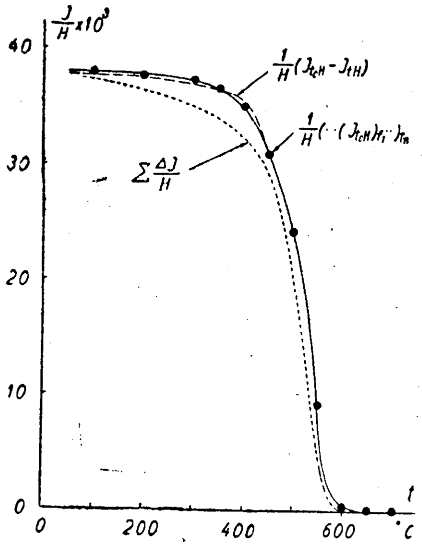

Fig. III-2-20. Comparision of $\frac{1}{H}(\cdots(J_{t_c}H)_{t_1}\cdots)_{t_n}$ with $\sum\frac{\Delta J}{H}$. Specimen No. 60′.

§ 3. Relation between the mode of change in magnetic susceptibility with temperature and that of the development of thermo-remanent magnetism.

In § 4, Chapter I, the mode of change in magnetic susceptibility of vorious rocks with temperature in a rather weak magnetic field was obtained by means of the ballistic method. Since magnetic susceptibility at a temperature t is uniquely defined as the ratio of induced magnetization to the intensity of magnetic field at constant temperature, provided the magnetic field is weak, the intensity of induced magnetization at t in H ought to be reversible with respect to the applied magnetic field, while in igneous rocks, thermo-remanent magnetization appears during the course of cooling and disappears during heating, its development and disappearence corresponding respectively to cooling or heating through a certain range of temperature. If the magnetic field H is applied to a rock sample kept at t, at which temperature the specimen is cooled from a temperature higher than the Curie-point in non-magnetic space, or heated from 0°C after it has been demagnetized by heat treatment in non-magnetic space, the specific magnetization of the specimen ought to be induced magnetization, $\vec{H} \cdot \chi(t)$, alone, and it must become zero again when the magnetic field is removed under the same conditions. It is now clear that the intensity of magnetization obtained by the ballistic method, in which the magnetic field is instantaneously reversed at t during heating or cooling with finite velocity, almost agrees with the induced magnetization.

On the other hand, thermo-remanent magnetism or magnetization during cooling, as defined in the present study, is purely the permanent (rigid) component of magnetization, since it was obtained by measuring residual magnetization in non-magnetic space by means of the ballistic method.

Thus, the conceptions of induced magnetization and thermo-remanent magnetization are rigidly separated, the one excluding the other. Now in order to compare these two quantities, the various characteristic properties of thermo-remanent magnetism as well as of the thermal change in magnetic susceptibility of various rock specimens are summarized in Table 3–I.

In this Table, the notations denote the characteristic properties as follows.

t_0 =the transition temperature of thermo-remanent magnetism.

J_{t_c} =saturated thermo-remanent magnetism, which is given by

$\int_0^{t_c} P(t)\,dt\,.$

$P(t_0)$ = the magnitude of $P(t)$ at transition temperature, showing the maximum value of each fundamental mode $P_j(t)$ of thermo-remanent magnetism.

2ϑ = the temperature range between the half-value points of $P(t_0)$—a measure for expressing the sharpness of the $P_j(t)$ curve, which is expressed by

$$2\vartheta = t_2 - t_1, \qquad P(t_1) = P(t_2) = \frac{1}{2}P(t_0)$$

$$t_2 > t_0 > t_1\,.$$

Table III–3–1. Characteristic quantities of thermo-remanent magnetism.

No. of Specimen	χ_0	χ_m	t_0	θ_m	θ_a	J_{t_c}	$P(t_0)$	2ϑ	Q_t
17	$2{\cdot}25\times10^{-3}$	$2{\cdot}64\times10^{-3}$	520°C (380)	460°C (350)	580°C (390)	$12{\cdot}12\times10^{-3}$	$10{\cdot}1\times10^{-5}$ (2·3) ″	80° 80	5·4
18	1·31 ″	2·16 ″	535 250	520 285	580 320	10·94 ″	5·4 ″ 4·8 ″	75 65	8·4
19	1·82 ″	2·95 ″	550 240	530 235	580 300	9·66 ″	5·5 ″ 2·7 ″	80 85	5·3
20	1·28 ″	1·49 ″	520	500	580	10·5 ″	8·4 ″	110	8·2
21	1·55 ″	2·04 ″	540	515	590	12·95 ″	8·9 ″	100	8·4
22	0·70 ″	1·11 ″	530 250	520 260	580 310	41·7 ″	1·52 ″ 11·7 ″	140 125	59·5
23	1·29 ″	1·54 ″	525	495	580	5·18 ″	4·0 ″	100	4·0
24	0·40 ″	0·69 ″	530 240	530 230	585 300	7·87 ″	2·1 ″ 7·2 ″	60 60	19·7
25	0·90 ″	2·14 ″	520 190	530 195	585 290	14·75 ″	4·2 ″ 0·8 ″	125 90	16·4
8	$1{\cdot}18\times10^{-3}$	—	535	—	—	$3{\cdot}29\times10^{-3}$	$2{\cdot}2\times10^{-5}$	85	2·8
27	1·28 ″	—	540	—	—	3·74 ″	3·0 ″	105	2·8
28	0·78 ″	—	530 245	—	—	7·40 ″	3·0 ″ 2·8 ″	85 110	9·5
35	0·40 ″	—	540	—	—	3·80 ″	2·8 ″	110	9·5
36	1·86 ″	—	550	—	—	4·20 ″	2·7 ″	95	2·3
38	1·01 ″	—	525 275	—	—	27·6 ″	10·0 ″ 6·9 ″	105 140	27·4
54	0·85 ″	—	540	—	—	9·05 ″	5·4 ″	90	11·6
93	0·54 ″	—	555 (100)	—	—	4·61 ″	3·0 ″ (0·7) ″	85	8·5
94	0·78 ″	—	560 120	—	—	4·79 ″	4·2 ″ (0·6) ″	75	6·2

(*to be continued.*)

Table III-3-I. (*continued*).

No. of Specimen	χ_0	χ_m	t_0	θ_m	θ_a	J_{tc}	$P(t_0)$	2θ	Q_t
16	$0{\cdot}61\times10^{-3}$	—	520°C 220	—	—	$2{\cdot}89\times10^{-3}$	$1{\cdot}1\times10^{-5}$ 1·8 ″	110° 70	4·7
25′	1·14 ″	—	520	—	—	5·07 ″	2·7 ″	120	4·4
45	0·25 ″	—	540 290	—	—	29·8 ″	8·1 ″ 14·6 ″	110 120	119·0
58	0·76 ″	—	495 295	—	—	29·8 ″	9·1 ″ 12·6 ″	110 105	39·2
59	1·31 ″	$2{\cdot}22\times10^{-3}$	520 320	500 290	580 360	38·3 ″	4·8 ″ 30·8 ″	105 80	29·3
60	1·11 ″	2·20 ″	525	460	590	47·6 ″	49·4 ″	85	42·9
60′	1·10 ″	1·57 ″	520	420	600	38·0 ″	28·8 ″	95	34·5
63	0·63 ″	—	495 395 300	—	—	11·7 ″	5·0 ″ 5·0 ″ 5·1 ″	70 80 80	18·5
65	0·74 ″	—	495 395 300	—	—	11·9 ″	3·9 ″ 5·3 ″ 4·8 ″	80 75 80	16·1
68	0·80 ″	—	420	—	—	33·5 ″	13·45 ″	220	42·0
70	0·37 ″	—	430	—	—	34·8 ″	21·1 ″	160	94·0
71	0·68 ″	—	430	—	—	28·4 ″	10·8 ″	210	41·7
73	0·50 ″	—	520 200	—	—	12·47 ″	0·9 12·4	70	24·9
75	1·19 ″	1·76 ″	295	530 270	580 350	16·5 ″	13·1 ″	85	13·8
76	1·27 ″	—	320	—	—	15·2 ″	14·8 ″	85	12·0

θ_m =mean Curie-temperature.

θ_a =apparent Curie-temperature.

χ_0 =specific susceptibility in a weak magnetic field at ordinary atmospheric temperature.

χ_m =maximum value of specific susceptibility in range of temperature of from 0°C to θ_a.

Q_t =coefficient of thermo-remanent magnetism, given by

$$Q_t \equiv J_{t_c}/\chi_0 .$$

Further, in order to compare more directly the mode of thermal change in susceptibility with that of thermo-remanent magnetism, the curves of $\chi(t)$ in the heating process and the $J_{t_c}^{t} \equiv \int_{t}^{t_c} P(t)\,dt$ curves of the same specimen are shown together in Fig. 3–1 and 3–2. while θ_m, θ_a, and t_0 are graphically shown in Figs. 3–3 and 3–4 in the form of spectra with respect to temperature.[1]

1) A part of these results was already reported in the writer's previous paper. (T. NAGATA, *Bull. Earthq. Res. Inst.*, **20** (1942), 192.)

Comparing the mode of $\int_t^{t_c} P(t)\,dt$ with that of $\chi(t)$, or t_{oj} with θ_{mj} and θ_a, we find that the standard mode of $P(t)$ corresponds to that of $\chi(t)$, and the complex mode in the former to that of the latter, each fundamental curve $P_j(t)$ in thermo-remanent magnetism corresponding to each $\chi_j(t)$ in susceptibility, where, as already mentioned,

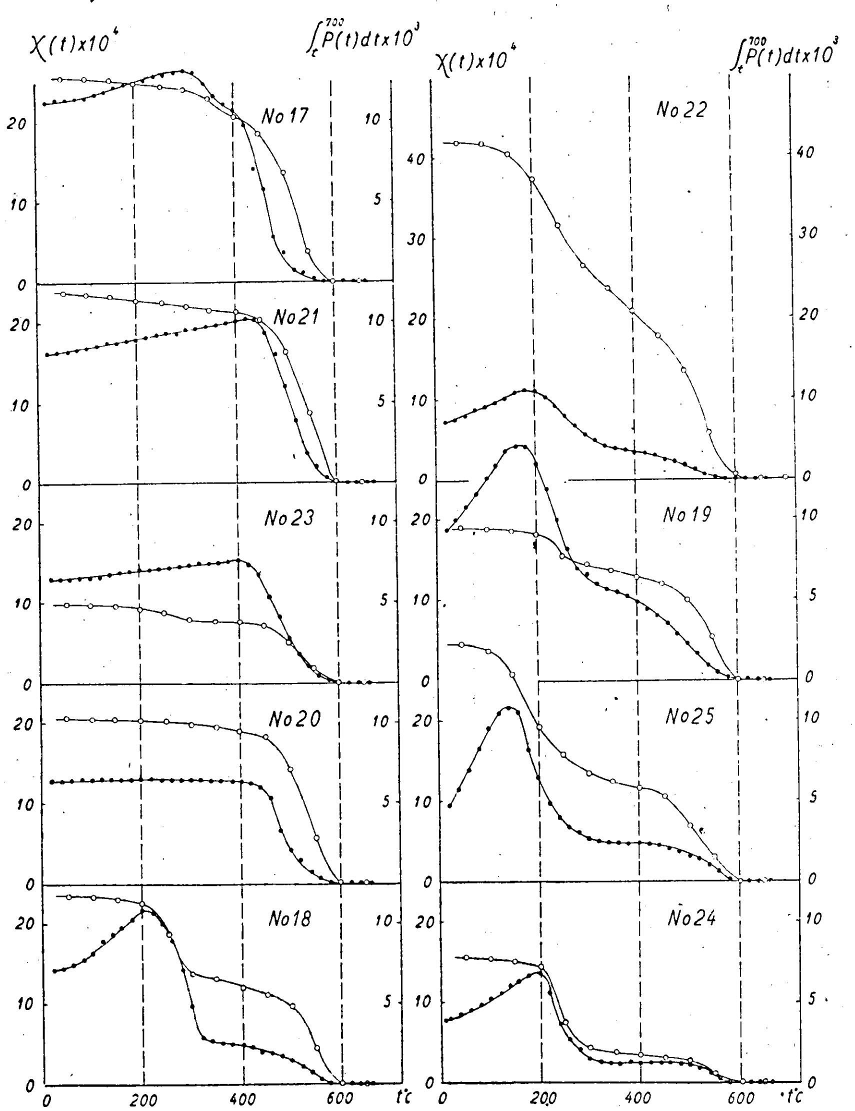

Fig. III-3-1. Comparison of susceptibility with thermo-remanent magnetism.
Full circles; specific susecptibility $\chi(t)$.
Hollow circles; thermo-remanent magnetism, $\int_t^{700} P(t)\,dt$.

$$P(t)=\sum_j P_j(t)\ , \qquad \chi(t)=\sum_j \chi_j(t)\ .$$

In other words, a transition temperature t_{oj} in thermo-remanent magnetism always corresponds to a mean Curie-point θ_{mj}, and to an apparent Curie-point θ_{aj}, although the former does not exactly agree with θ_{mj} nor with θ_{aj}. This correspondence will be seen more clearly in Fig. 3–3. Since the error in determining the temperature of the specimen in both cases of measuring $P(t)$ and $\chi(t)$ amounted to about 5 degrees, no rigorous values of t_{oj}, θ_{mj}, and θ_{aj} have yet been obtained. Generally speaking, however, the transition temperature is 40–80 degrees lower than the apparent Curie-temperature.

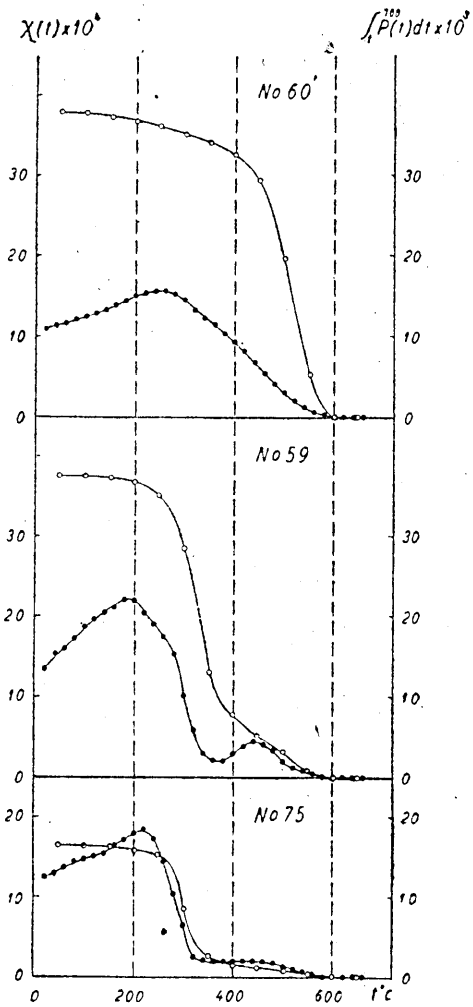

Fig. III-3-2. Comparison of susceptibility with thermo-remanent magnetism.
Full circles; specific susceptibility $\chi(t)$.
Hollow circles; thermo-remanent magnetism, $\int_t^{700} P(t)dt$.

Suppose now that a rock specimen is cooled to t from a temperature higher than the apparent Curie-point in non-magnetic space, and that a magnetic field H is applied at t. The specific intensity of magnetization of the specimen at this instant ought then to be $\vec{H}\cdot\chi(t)$. As the next step, let the specimen be cooled further in magnetic field H through a range of temperature δt, when the total magnetization J of the specimen at $t-\delta t$ is given by

$$\vec{J}(t-\delta t)\simeq\vec{H}\cdot\chi(t-\delta t)+\vec{H}\cdot p(t,\ t-\delta t)\,\delta t \tag{3–1}$$

where $p(t,\ t-\delta t)$ denotes the characteristic function of thermo-remanent magnetism of t measured at $t-\delta t$ which, generally speaking, could differ from $P(t)$—the characteristic function of thermo-remanent magnetism measured at 0°C.

Here, $P(t)$ should be

$$P(t) \equiv p(t, o), \quad (3\text{–}2)$$

since permanent magnetization $\vec{H} \cdot p(t, t-\delta t)\delta t$ becomes $\vec{H} \cdot P(t)\delta t$ when, as metioned in the preceding paragraph, the specimen is cooled from $t-\delta t$ to 0°C in non-magnetic space.

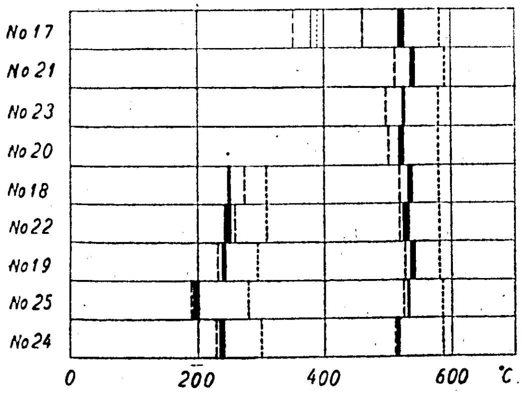

Fig. III-3-3. Charcteristic temperatures of ejecta from Volcano Huzi.
Full line; transition temperature t_0.
Dotted line; apparent Curie-point θ_a.
Chain line; mean Curie-temperature θ_m.

Then, $p(t, t)$[2] may, at any rate, be a function of $\chi(t)$, with the result that $P(t)$ is also a function of the latter, because any permanent magnetization developed during cooling from t to $t-\delta t$ in $\vec{H}$ must be the residue of the whole or a part of the induced magnetization $\vec{H} \cdot \chi(t)$ at t (which is the total magnetization at t), so long as δt is infinitesimally small.

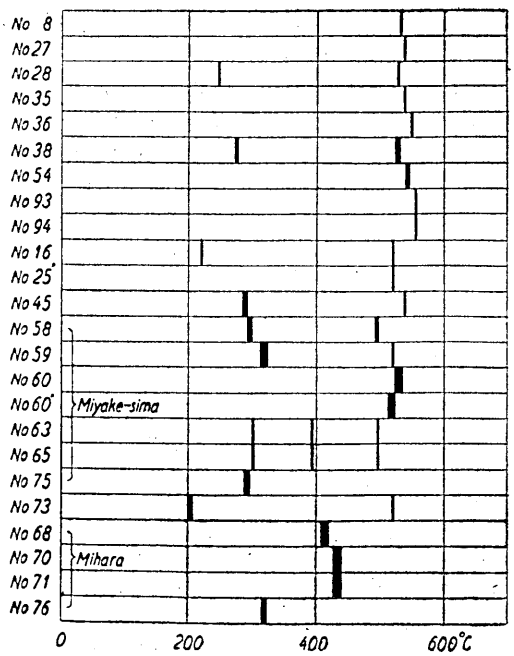

Fig. III-3-4. Transition temperature of various volcanic rocks.

From this point of view, we compare the values of $\chi(t)$ with those of $P(t) = p(t, o)$ corresponding to the same temperature, after which we may find that always $P(t) < \chi(t)$, although $\int_t^{t_c} P(t)\,dt$ far exceeds $\chi(t)$ corresponding to the low values of t, and that $P(t)$

2) $p(t, t)$ is defined by

$$p(t, t) = \lim_{\delta t \to 0} p(t, t-\delta t).$$

is nearly parallel to $-\frac{\partial}{\partial t}\chi(t)$ in wide ranges of temperature, except that in which $-\frac{\partial}{\partial t}\chi(t)$ is negative.

It may thus be concluded that thermo-remanent magnetism increases immediately after it has been transformed from the paramagnetic state into the ferro-magnetic state, or speaking more practically, during cooling from the apparent Curie-temperature through a temperature-range of about 100 degrees. We shall, however, deal again with the relation between thermo-remanent magnetism and susceptibility after we have shown the functional relation of $P(t) = p(t, o)$ to $p(t, t)$ (see § 4 of this Chapter).

(ii) Dealing especially with the modes of $P(t)$ of Huzi ejecta, we may say that they belong either to the standard mode or to the complex mode that is constituted of two fundamental modes, the standard and the extraordinary.

In the extraordinary modes of thermo-remanent megnetism and the susceptibility of these rocks, i. e., $P_e(t)$ and $\chi_e(t)$ respectively, the transition point t_{oe}, mean Curie-point θ_{me}, and apparent Curie-point θ_{ae} are about 200–250°C, 200～280°C, and 290–320°C respectively. As mentioned in § 4, Chapter I, the standard mode of susceptibility $\chi_s(t)$ is presumed to be chiefly due to the group of ferro-magnetic minerals having nearly the same chemical composition as pure magnetite, whence, from the law of correspondence between $\chi(t)$ and $P(t)$, the standard mode of thermo-remanent magnetization $P_s(t)$ may also be due to that group of minerals, while the extraordinary mode $P_e(t)$ may be due to the group of ferro-magnetic minerals composed not only of Fe_2O_3, and FeO, but also of TiO_2, just as in the case of $\chi_e(t)$.

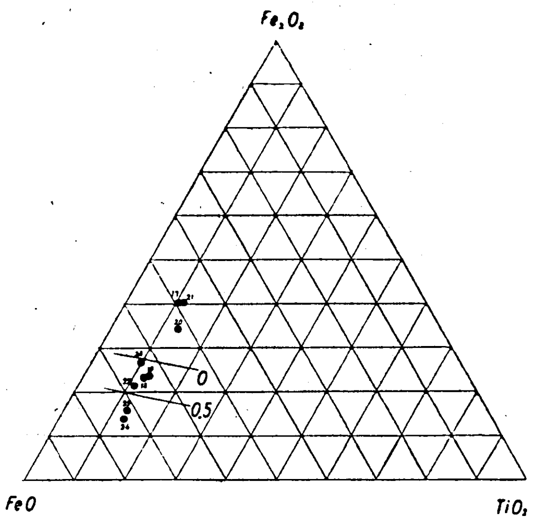

Fig. III-3-5. Relation between $Pe(t_{oe})/\{P_s(t_{os}) + P_e(t_{oe})\}$ and Fe_2O_3: FeO: TiO_2. Ejecta from Huzi Volcano.

In order to see directly the relation between the mode of $P(t)$ and chemical composition, if any, the values of $P_e(t_{oe})/\{P_s(t_{os})+P_e(t_{oe})\}$ were plotted against the ratios of $FeO:Fe_2O_3:TiO_2$, as shown in Fig. 3–5, where these amounts of FeO, etc, are those contained in the total rock, and not those composing the ferro-magnetic minerals alone. It will be seen from this figure that the extraordinary mode of thermo-remanent magnetization grows as the relative amount of Fe_2O_3 decreases and that of FeO increases, that of TiO_2 being kept nearly constant. On the other hand, taking into consideration that the minerals concerned with the magnetic behaviour of rock are chiefly magnetite $(FeO \cdot Fe_2O_3)$, ilmenite $(FeO \cdot TiO_2)$, and their solid solutions, the values of $P(t_{oe})/\{P(t_{os})+P(t_{oe})\}$ were again plotted against the relative ratio $Il/\{Il+Mt\}$ as shown in Fig. 3–6, where Il and Mt denote respectively the normative contents of ilmenite and magnetite in the rocks as calculated from their chemical compositions.

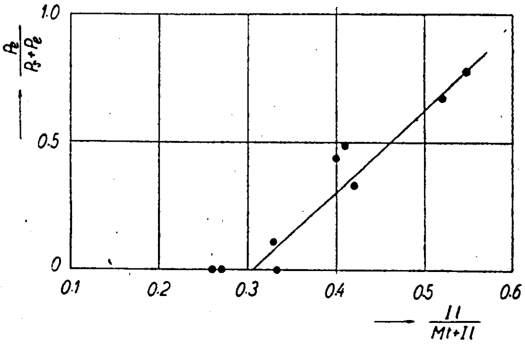

Fig. III–3–6.
Relation between $\frac{P_e}{P_s+P_e}$ and $\frac{Il}{Mt+Il}$.
(Huzi ejecta)

Exactly as in the case of the relation between $\chi(t)$ and the chemical composition, the extraordinary mode $P_e(t)$ begins to appear when the relative content of ilmenite exceeds about 0·34, whence it will be said that the transition temperature of thermo-remanent magnetism is related to the chemical composition of rock in exactly the same way that the Curie-point is related to the latter.

Table III–3–II (a)
Huzi ejecta.

No.	$P_e/(P_s+P_e)$	Mt	Il	$Il/(Mt+Il)$
17	0	7·87%	2·73%	0·26
18	0·49	3·94	2·73	0·41
19	0·34	5·33	3·79	0·42
20	0	6·48	3·34	0·34
21	0	6·95	2·58	0·27
22	0·44	3·94	2·58	0·40
23	0·11	4·40	2·12	0·33
24	0·79	1·39	1·67	0·55
25	0·68	2·55	2·73	0·52

Table III–3–II (b)

No.	$P_e/(P_s+P_e)$	Mt	Il	$Il/(Mt+Il)$
8	0	3·49%	1·37%	0·28
27	0	4·17	1·21	0·[illegible]2
28	0·48	2·55	1·97	0·44
35	0	3·01	1·52	0·34
36	0	6·02	1·52	0·20
38	0·42	3·91	1·37	0·26
93	(0·22)	3·47	1·21	0·26
94	(0·14)	6·02	1·52	0·20

Therefore, summarizing the two foregoing results, namely, the relation between θ and t_0 and that between $P(t_0)$ and the chemical composition, we may conclude that the temperature at which the bulk of the thermo-remanent magnetization develops is largely subject to the Curie-point temperature of rock, although the conditions that determine the magnitude of $P(t_0)$ have not yet been ascertained.

§ 4. Discussion on the physical mechanism of development of thermo-remanent magnetism.

As mentioned in Chapter I of this report, the magnetic behaviour of volcanic rocks is largely due to that of the ferro-magnetic minerals contained in them, whence it is presumed that thermo-remanent magnetism is also a magnetic property of ferro-magnetic minerals—a presumption based on the result given in § 3 of this Chapter. A direct proof of the ferro-magnetic character of volcanic rock was obtained in the following manner.

A rock specimen (No 60′) of cylindrical shape, 0·8 *cm* in diameter and 5 *cm* long, was subjected to a strong uniform magnetic field (about 4000 *Oe.*), where the axis of the specimen coincided with the direction of the applied magnetic field, the rock specimen after removal of the magnetic field having intense residual magnetization M. The specimen was next heated slowly up to exactly 500°C in a non-magnetic, non-inductive electric-furnace in non-mangetic space, and then slowly cooled down to room temperature under like conditions. The intensity of residual magnetization M' following the foregoing treatment was less than M.

On the other hand, according to the result given in § 3, Chapter I, the ferro-magnetic mineral in the rock is believed to consist of a large number of micro-crystals having different Curie-points; that is, letting $I(\theta, t)$ denote the intensity of residual magnetization at t of a ferro-magnetic mineral having its Curie-point at θ, we get the total intensity of residual magnetization at 0°C from

$$M(0)=\int_{\theta_1}^{\theta_2} I(\theta,\ t)\,g(\theta)\,d\theta\,, \tag{4–1}$$

where $g(\theta)$ denotes the frequency of distribution in volume of ferro-magnetic minerals with respect to their Curie-point temperature θ, so that in the case of a rock specimen of $M(0)$ at 0°C in the initial state heated to 500°C in non-magnetic space, the residual permanent magnetization of such part of ferro-magnetic minerals as have Curie-points lower than 500°C vanishes, with the result that $M'(0)$ after this heat treatment becomes

$$M'(0)=\int_{\theta_1}^{\theta_2} I(\theta,\ 0)\,g(\theta)\,d\theta-\int_{\theta_1}^{500^\circ\mathrm{C}} I(\theta,\ 0)\,g(\theta)\,d\theta$$

$$=\int_{500^\circ\mathrm{C}}^{\theta_2} I(\theta,\ 0)\,g(\theta)\,d\theta<M(0)\,. \qquad (4\text{–}2)$$

Now, the rock specimen of $M'(0)$ at 0°C was again heated to $t'<500°\mathrm{C}$ (actually $t'=420°\mathrm{C}$) in non-magnetic space, the intensity of magnetization $M'(t)$ at any temperature during heating being measured with an astatic magnetometer, while from $t'=420°\mathrm{C}$, the specimen was cooled in non-magnetic space, the magnetization during cooling also having been measured. The results of this experiment are given in Fig. 4–1, from which we find that $M'(t)$ decreases monotonously with increase in temperature t, and that $M'(t)$ is almost reversible with respect to change in temperature, though there is a slight discrepancy in $M'(t)$ during heating and cooling. That is to say, $M'(t) =\int_{500^\circ\mathrm{C}}^{\theta_2} I(\theta,\ t)\,g(\theta)\,d\theta$ is almost reversible with change in temperature so long as $t<500°\mathrm{C}$, and

$$\frac{\partial}{\partial t}M'(t)=\frac{\partial}{\partial t}\int_{500^\circ\mathrm{C}}^{\theta_2} I(\theta,\ t)\,g(\theta)\,d\theta\,. \qquad (4\text{–}3)$$

This fact seems to show that volcanic rock follows the Curie-Weiss law of ferro-magnetism. If we assume in eq (4–2) that, corresponding to any Curie-point θ, $g(\theta)$ denotes the product of the volume of elementary domains in ferro-magnetic minerals having θ and the direction-cosine of the applied magnetic field with respect to the direction of their easiest magnetization, we can take $I(\theta, t)$ as the intensity of saturation magnetization of t

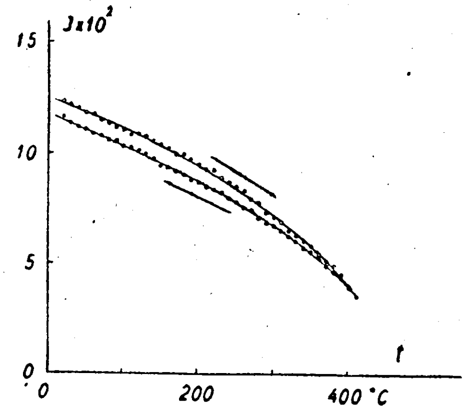

Fig. III-4-1. The mode of change in remanent magnetization with tempereture. Specimen No. 60′.

of the ferro-magnetic mineral of Curie-point θ. We may, then, say that, in heating and cooling in non-magnetic space, the saturation magnetization of each ferro-magnetic mineral $I(\theta, t)$ changes reversibly with temperature, following the Curie-Weiss law, a conclusion leading to the result that the elementary domains in the ferro-magnetic minerals in rocks are not affected by temperature, only the intensity of saturation magnetization changing with it. It was then found that the mode of change of residual magnetization of rock with temperature well agrees with that derived from the theory of ferromagnetism based on the magnetic behaviour of ordinary ferro-magnetic substances.

(ii) On the other hand, from the result given in § 3, this Chapter, the thermo-remanent magnetization of a ferro-magnetic mineral is believed to be brought about at a temperature just below its Curie-point. The following conception may possibly explain the development of thermo-remanent magnetism during cooling in a magnetic field:

In cooling in a magnetic field, a part of the elementary domains in every micro-crystal that is contained in a rock takes, irreversibly, the direction of easiest magnetization nearest that of the applied magnetic field at a temperature[1] below the Curie-point peculiar to it, where the number and dimensions of these domains is a function of the applied magnetic field H, the internal stress S, and the chemical composition of the micro-crystal, and consequently the corresponding Curie-temperature θ. Hence, if $\Psi(H, S, \theta)$ and $I(\theta, t)$ denote the probable volume of the elementary domains satisfying the condition just-mentioned and the intensity of saturation magnetization of the domain respectively, the magnetization of such a micro crystal of unit volume is given by

$$I(\theta, t)\Psi(H, S, \theta)\,,$$

where $I(\theta, t)$ decreases, as is well known, with increase in temperature, being zero when $t=\theta$.

If, then, in the process of cooling of the rock sample from a sufficiently high temperature, a magnetic field H is applied during its cooling from t_2 to t_1, the remanent magnetization of a whole rock specimen at temperature t is given by

1) In ordear to comlete this action of domains in taking the direction of the applied magnetic field, the cooling of sample through a finite range of temperature and finite time will be necessary. For simplicity, however, we assume here that the process takes place at a temperature independently of time.

$$J(t)=\int_{t_1}^{t_2} I(\theta, t)\Psi(H, S, \theta)g(\theta)d\theta\ , \tag{4–4}$$

the special case of eq (4–4) that $t=0$ becoming

$$J(0)=\int_{t_1}^{t_2} I(\theta, 0)\Psi(H, S, \theta)g(\theta)d\theta\ . \tag{4–4}$$

On the other hand, according to the concluding remark in the preceding paragraph in connexion with eq. (3–1), the total magnetization of a rock specimen cooled from t_2 to t_1 in a magnetic field H, as measured at t_1, is given by

$$\vec{J}=\vec{H}\chi(t_1)+\vec{H}\int_{t_1}^{t_2} p(\theta, t_1)d\theta\ , \tag{4–5}$$

where $t_2 \geqq t_1$, and $p(\theta,\ t)=0$ when $t>\theta$. In eq. (4–5), however, the first term on the right-hand side gives the induced magnetization, which is reversible with respect to any change in the applied magnetic field, the irreversible residual magnetization corresponding only to the second term.

Consequently, the remanent magnetization $J(t)$ at temperature t of a rock specimen which is cooled from t_2 to t_1 in H is given by

$$\vec{J}(t)=\vec{H}\int_{t_1}^{t_2} p(\theta, t)d\theta\ . \tag{4–6}$$

Comparing then, eq. (4–4) with (4–6), we get

$$Hp(\theta, t)=I(\theta, t)\Psi(H, S, \theta)g(\theta)\ , \tag{4–7}$$

or putting

$$\Psi(H, S, \theta)=F(H)\,\Phi(S, \theta)\ ,$$

we get

$$p(\theta, t)=I(\theta, t)\,\Phi(S, \theta)g(\theta)\ , \tag{4–8}$$

provided H is small, whereas from eq. (4–3) we have found that

$$\frac{\partial}{\partial t}I(\theta, t)<0\ . \tag{4–9}$$

Then, so long as we assume that $\Psi(H,\ S,\ \theta)$ is constant, independently of temperature t, we find from eqs. (4–7) and (4–8) that $p(\theta,\ t)$ ought to satisfy the conditian

$$\frac{\partial}{\partial t}p(\theta, t) < 0\,, \qquad (t < \theta)\,,$$

with the result that also

$$\frac{\partial}{\partial t}\left[\frac{J(t)}{H}\right] = \frac{\partial}{\partial t}\int_{t_1}^{t_2} p(\theta, t)\,d\theta < 0, \qquad t < t_1 < t_2\,. \tag{4–10}$$

(iii) In order to test whether or not the above-mentioned hypothesis is correct, several experiments were made, for which purpose an astatic magnetometer was used, a schematic view of this apparatus being shown in Fig. 4–2.

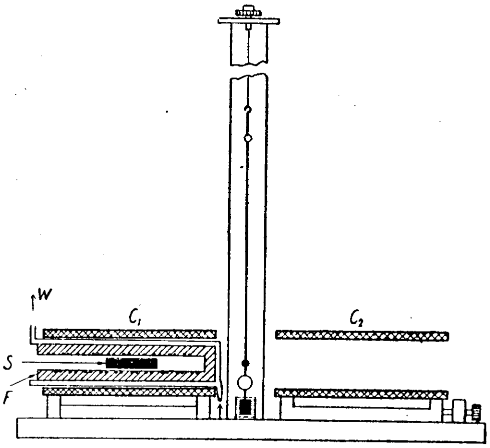

Fig. III-4-2. Schematic view of astatic magnetometer.
S, specimen; F, furnace; W, water cooling; C_1, C_2, field coils.

First, after a rock specimen of cylindrical shape (No. 60″) was heated to 600°C in non-magnetic space, it was then cooled from 600°C to 500°C in a magnetic field of 1·82 Oe, while, from thence to room temperature, it was cooled (at the rate of 2 degree/minute) in non-magnetic space, the intensity of magnetization $J(t)$ at any temperature during the final process being measured with an astatic magnetometer. The observed results are given in Table 4–I and in Fig. 4–3. Here, the observed value of $J(t)$ ought to be given by

$$J(t) = 1{\cdot}82 \times \int_{500^\circ C}^{600^\circ C} p(\theta, t)\,d\theta\,, \qquad t \leq 500^\circ C\,.$$

Then, as will be seen from Fig. 4–3, the observed value of $J(t)$ seems to satisfy the condition given by eq. (4–10) throughout the entire range of temperature from 500°C to 0°C, and the intensity of residual magnetization at room temperature is about 2·2 times 490°C. These facts seem to show that the assumed mechanism of development of thermo-remanent magnetism given by eq. (4–4) holds approximately in the actual case, at any rate, when the magnetic force H is small.

Table III–4–I. (Specimen No 60′).

Temp.	J	Temp.	J
480	$6{\cdot}58\times10^{-3}$	250	$11{\cdot}93\times10^{-3}$
470	6·93 〃	240	12·10 〃
460	7·31 〃	230	12·29 〃
450	7·68 〃	220	12·35 〃
440	8·11 〃	210	12·55 〃
430	8·52 〃	200	12·61 〃
420	8·80 〃	190	12·73 〃
410	9·07 〃	180	12·85 〃
400	9·29 〃	170	12·96 〃
390	9·49 〃	160	13·08 〃
380	9·72 〃	150	13·16 〃
370	9·96 〃	140	13·23 〃
360	10·16 〃	130	13·31 〃
350	10·43 〃	120	13·39 〃
340	10·56 〃	110	13·45 〃
330	10·66 〃	100	13·57 〃
320	10·88 〃	90	13·62 〃
310	11·05 〃	80	13·74 〃
300	11·21 〃	70	13·80 〃
290	11·32 〃	60	13·91 〃
280	11·49 〃	50	13·98 〃
270	11·61 〃	40	14·08 〃
260	11·75 〃	30	14·15 〃

Secondly, an experiment was made to test whether or not $p(\theta, t)$ is independent of time τ after removal of the magnetic field H. A rock sample (No 60″), which was cooled from 600°C to 550°C in non-magnetic space, and from 550°C to 500°C in a magnetic field of 1·82 Oe, was kept at 500°C (±0·5°) in non-magnetic space, the change in magnetization J(500°C) being measured every five minutes: The observed results are shown in Table 4–II and in Fig. 4–4. It will be seen from this result, that $p(\theta, t)$ is subject to time-effect, decreasing with time, at any rate, at high

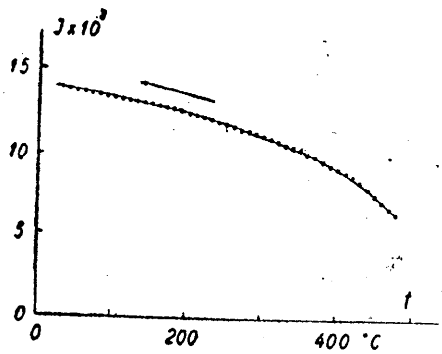

Fig. III–4–3. The mode of increase of thermo-remanent magnetization with decrease in temperature in the space $H=0$. Specimen No 60′.

Table III-4-II. Time effect in development of thermo-remanent magnetism. Specimen No 60′.

Time	J/ρ	Time	J/ρ
2·5 minutes	$2{\cdot}91\times10^{-3}$	67·5 minutes	$2{\cdot}52\times10^{-3}$
7·5 〃	2·90 〃	72·5 〃	2·49 〃
12·5 〃	2·87 〃	77·5 〃	2·47 〃
17·5 〃	2·86 〃	82·5 〃	2·41 〃
22·5 〃	2·83 〃	87·5 〃	2·39 〃
27·5 〃	2·77 〃	92·5 〃	2·38 〃
32·5 〃	2·72 〃	97·5 〃	2·37 〃
37·5 〃	2·71 〃	102·5 〃	2·37 〃
42·5 〃	2·69 〃	107·5 〃	2·38 〃
47·5 〃	2·68 〃	112·5 〃	2·37 〃
52·5 〃	2·64 〃	117·5 〃	2·35 〃
57·5 〃	2·61 〃	122·5 〃	2·35 〃
62·5 〃	2·57 〃		

temperature, though it seems to approach to almost a stable final state within a relatively short time, say about 80 minutes. It must be noted here that the existence of the above-mentioned time-effect does not run counter to the fact that $\frac{\partial}{\partial\tau}P(t)\simeq0$, as obtained in § 2, this Chapter, seeing that the former fact shows that thermo-remanent magnetization at a high temperature after removal of the external magnetic field is unstable, decreasing with time, while the latter shows that the development of $p(\theta,t)\Delta\theta$ during cooling through the range of temperature $\Delta\theta$ at θ under the effect of the external magnetic field does not depend on the rapidity of cooling.

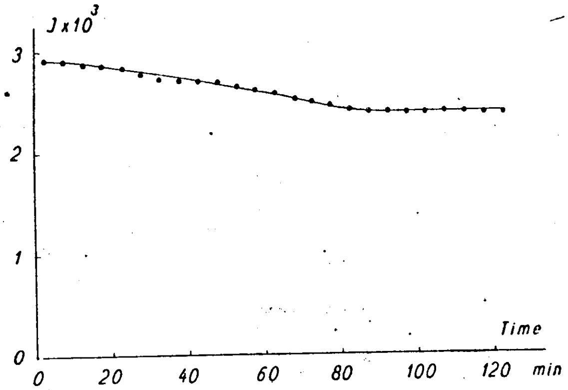

Fig. III-4-4. Time effect in development of thermo-remanent magneeism. Specimen No 60′.

Now, we assume that the time effect of $p(\theta,t)$ is due to that of $\Psi(H,S,\theta)$ in eq. (4–7), and not to $I(\theta,t)$, nor to $g(\theta)$, that is, $\Psi=\Psi(H,S,\theta,\tau)$, where the effect of τ on H is probably the function of temperature. In other words, if the temperature is higher, the time effect will be larger, since the phenomenon of time effect is probably the result of thermal agitation. Further, it is presumed that the time-effect is affected by the magnetic field, since if, for example, a magnetic field H parallel to $J(t)$ does exist, a decrease in Ψ must do some work against H.*

The mode of development of thermo-remanent magnetism must then be very complex, depending on the character of the ferro-magnetic minerals, temperature, magnetic field, and time.

However, in the actual experiments in thermo-remanent magnetism dealt with in §1 and §2 of this Chapter, the rapidity of heating and cooling was always kept at a nearly constant rate, about 2 degrees/minute, whence all the experimental results of measuring $\vec{J}_t$ or $\Delta\vec{J}_t'$ (i.e. $J(0)$ or $\Delta J(0)$), is subject to nearly the same amount of time effect. The conclusion arrived at in §1∼§3 of this Chapter then holds, at any rate, as a first approximation in the case of a weak magnetic field, notwithstanding the time effect. Further, the fact that the intensity of the natural remanent magnetization J_n of a freshly ejected rock, which was developed during cooling in a geomagnetic field, probably at an extremely slow rate of cooling, almost agrees with that of the thermo-remanent magnetization developed during a fairly rapid rate of cooling, may mean that the time effect integrated with respect to temperature from t_c to ordinary room temperature depends to a slight extent on the time required for cooling.

Strictly speaking, however, the time effect ought to cause a slight discrepancy between the intensities of thermo-remanent magnetization brought about as the result of differences with regard to the time required for cooling, etc., whence it will be only natural that the intensity of thermo-remanent magnetism and partial magnetism during cooling should invariably show an error of a few percent.

(iv) In the next step, the sample that had been kept at 500°C in $H=0$ for 120 minutes in the foregoing experiment was cooled in non-magnetic space, the intensity of magnetization $J(t)$ at various temperatures during cooling also being measured, which results are given in Table 4–III and in Fig. 4–5. (Here, this operation of cooling will be called process (1).) In this case, $J(t)$ also increases markedly with decrease in temperature.

Now, the rock sample that was cooled in the above-mentioned experiment was heated (2) from 30°C to 450°C, (3) then cooled to

* Therefore, the above mentioned time-effect will be extremely small in the case of cooling in a magnetic field.

Table III–4–III. The change in thermo-remanent magnetism with temperature in non-magnetic space. Specimen No 60′.

Temp.	cooling	heating	cooling	heating	cooling	heating
30°C	$5{\cdot}64\times10^{-3}$	$5{\cdot}61\times10^{-3}$				
40	5·62 ″	5·64 ″				
50	5·60 ″	5·62 ″				
60	5·58 ″	5·59 ″				
70	5·55 ″	5·55 ″				
80	5·52 ″	5·52 ″				
90	5·49 ″	5·46 ″				
100	5·46 ″	5·44 ″				
110	5·44 ″	5·37 ″				
120	5·41 ″	5·33 ″				
130	5·37 ″	5·28 ″				
140	5·34 ″	5·23 ″				
150	5·30 ″	5·20 ″				
160	5·25 ″	5·13 ″				
170	5·21 ″	5·09 ″				
180	5·16 ″	5·01 ″				
190	5·11 ″	4·99 ″				
200	5·06 ″	4·92 ″				
210	5·02 ″	4·92 ″				
220	4·99 ″	4·85 ″				
230	4·98 ″	4·81 ″				
240	4·93 ″	4·75 ″				
250	4·90 ″	4·70 ″	$4{\cdot}60\times10^{-3}$	$4{\cdot}58\times10^{-3}$		
260	4·87 ″	4·64 ″	4·55 ″	4·52 ″		
270	4·83 ″	4·62 ″	4·52 ″	4·48 ″		
280	4·79 ″	4·56 ″	4·48 ″	4·43 ″		
290	4·74 ″	4·48 ″	4·41 ″	4·38 ″		
300	4·69 ″	4·41 ″	4·33 ″	4·31 ″		
310	4·66 ″	4·37 ″	4·26 ″	4·25 ″		
320	4·62 ″	4·31 ″	4·23 ″	4·19 ″		
330	4·56 ″	4·25 ″	4·18 ″	4·12 ″		
340	4·50 ″	4·18 ″	4·13 ″	4·06 ″		
350	4·46 ″	4·12 ″	4·08 ″	4·03 ″		
360	4·42 ″	4·07 ″	4·02 ″	3·96 ″		
370	4·38 ″	4·00 ″	3·95 ″	3·89 ″		
380	4·30 ″	3·92 ″	3·89 ″	3·81 ″		
390	4·22 ″	3·87 ″	3·81 ″	3·72 ″		
400	4·15 ″	3·79 ″	3·71 ″	3·63 ″		
410	4·06 ″	3·67 ″	3·63 ″	3·54 ″		

(to be continued.)

Table III-4-III. (*continued*)

Temp.	cooling	heating	cooling	heating	cooling	heating
420°C	$3{\cdot}97\times10^{-3}$	$3{\cdot}55\times10^{-3}$	$3{\cdot}52\times10^{-3}$	$3{\cdot}43\times10^{-3}$		
430	3·86 〃	3·38 〃	3·38 〃	3·30 〃		
440	3·76 〃	3·26 〃	3·27 〃	3·19 〃		
450	3·67 〃	3·12 〃	3·15 〃	3·06 〃		
460	3·53 〃			2·91 〃		
470	3·41 〃			2·77 〃		
480	3·19 〃			2·57 〃	$-0{\cdot}03\times10^{-3}$	$-0{\cdot}03\times10^{-3}$
490	2·94 〃			2·36 〃	−0·02 〃	−0·03 〃
500	2·58 〃			2·07 〃	−0·01 〃	−0·03 〃
510				1·72 〃	0·01 〃	−0·01 〃
520				1·24 〃	0·02 〃	−0·02 〃
530				0·82 〃	0·01 〃	0·00 〃
540				0·46 〃	0·00 〃	0·01 〃
550				0·25 〃	0·03 〃	−0·01 〃
560				0·16 〃	0·03 〃	0·01 〃
570				0·10 〃	0·05 〃	−0·02 〃
580				0·08 〃	0·08 〃	−0·02 〃
590						−0·02 〃
600						−0·03 〃
610						−0·02 〃
620						−0·01 〃

250°*C*, (4) from which temperature it was again heated to 580°*C*, (5) then cooled to 480°*C*, (6) and heated again to 630°*C*, all the heating and cooling being done in non-magnetic space. The intensity of magnetization during these operations is shown in Table 4–3 and Fig. 4–5, from which it will be clear that the intensity of magnetization $J(t)$ in operation (2) decreases with temperature, the $J(t) \sim t$ relation in this case almost agreeing with that in operation (1), although $J(t)$ in the former is always slightly less than that in the latter. During the range of temperature

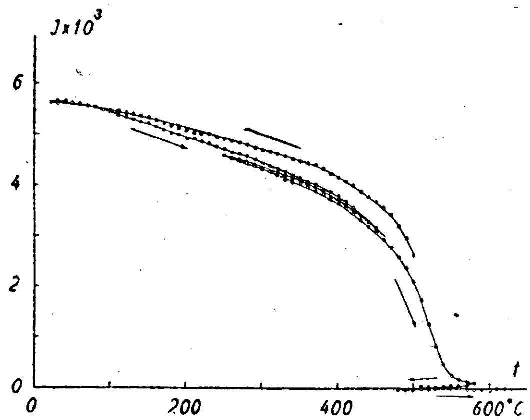

Fig. III-4-5. Change in intensity of thermo-remanent magnetism with temperature. Specimen No 60′.

from 250°C to 450°C, the values of $J(t)$ in heating operation (2), cooling operation (3), and in re-heating operation (4) almost agree; in other words, $J(t)$ is reversible with respect to change in temperature. After $J(t)$ becomes zero (within the limits of observational error) at 580°C in operation (4), it is kept at zero notwithstanding that it was cooled from 580°C to 480°C in (5), that is, during this range of temperature $J(t)$ is irreversible with respect to t.

As will be clear from the given conditions in this measurement, $J(t)$ in this case is given by

$$\frac{J(t)}{H}=\int_{500°C}^{550°C} p(\theta, t)\,d\theta\,. \tag{4–11}$$

Then, putting relation (4–8) into eq. (4–11), we get

$$\frac{J(t)}{H}=\int_{500°C}^{550°C} I(\theta, t)\,\Phi(s, \theta)\,g(\theta)\,d\theta\,. \tag{4–12}$$ [2]

Although the present experiment shows that $J(t)$ changes reversibly with temperature, provided $t \leqq 500°C$, we can conclude that only the intensity of magnetization of the elementary domains $I(\theta, t)$ changes reversibly with temperature (following the Curie-Weiss law), $\Phi(s, \theta)$ and $g(\theta)$ being kept almost constant.

If, however, t exceeds 500°C, $\Phi(s, \theta')g(\theta')$ in eq. (4–12) or $\Psi(H, S, \theta')g(\theta')$ in eq. (4–4), where $500°C \leqq \theta' \leqq t$ ought to disappear, with the result that $\int_{500°C}^{t} I(\theta', t)\,\Phi(s, \theta')\,g(\theta')\,d\theta'=0$. According to its definition, only during cooling through temperature θ, as the effect of the magnetic field H, $\Phi(s, \theta)$ can be caused in the form of $H\cdot\Phi(s, \theta)$, or generally $\Psi(H, S, \theta)$. Hence, in cooling in non-magnetic space, no thermo-remanent magnetization should develop, with the result that $J(t)$ in heating during a range of temperature exceeding 500°C must be given by

$$\frac{J(t)}{H}=\int_{t}^{550°C} I(\theta, t)\,\Phi(s, \theta)\,g(\theta)\,d\theta\,,$$

whereas, in cooling in non-magnetic space from $t>500°C$, $J(t')$ at temperature t' is given by

2) Here, we assume that $\Phi(s, \theta)$ in eq. (4–12) includes the time-effect discussed in the preceding example.

$$\frac{J(t')}{H}=\int_{t}^{550°C} I(\theta, t')\,\Phi(s, \theta)\,g(\theta)\,d\theta$$

Then, after $\Phi(s, \theta)\cdot g(\theta)$ entirely disappears as the result of heating the specimen to more than $550°C$, $J(t)$ must be zero regardless of whether it is heating or cooling, so long as no magnetic field is supplied. The facts observed from operations (4), (5), and (6) in Fig. 4–5 well agree with the above-mentioned theory, as also the experimental facts given by eqs. (2–18)$\sim$(2–23) in §2 of this Chapter. Summing up, then, we may conclude that

$$P(\theta)=I(\theta, 0)\,\Phi(s, \theta)\,g(\theta)\,, \tag{4–13}$$

$$\vec{J}_{t,H}^{t'}=\vec{H}\int_{t'}^{t} I(\theta, 0)\,\Phi(s, \theta)\,g(\theta)\,d\theta\,, \tag{4–14}$$

$$(\vec{J}_{t_c,H})_{t_n}=\vec{H}\int_{t_n}^{t_c} I(\theta, 0)\,\Phi(s, \theta)\,g(\theta)\,d\theta\,, \tag{4–15}$$

provided H is small and the time-effect is neglected.

(v) So long as the intensity of the applied magnetic field is small, say less than 2 *Oersteds,* the time-effect in the development of thermo-remanent magnetism seems negligible, at any rate as a first approximation, seeing that the results of all the various experiments harmonize well with one another, notwithstanding that the time effect was scarcely taken into consideration.

On the other hand, the mode of development of thermo-remanent magnetization in a fairly large magnetic field seems very complex. For example, a rock sample (No. 60″) was cooled from $550°C$ to $500°C$ in a magnetic field $H=15$ *Oe*, and then cooled in non-magnetic space, $J(t)$ in (1) this cooling operation and those in (2) re-heating (from 30°C to 460°C) and (3) re-cooling (from 460°C to 30°C) in non-magnetic space, being shown in Fig. 4–6. It will be seen from these

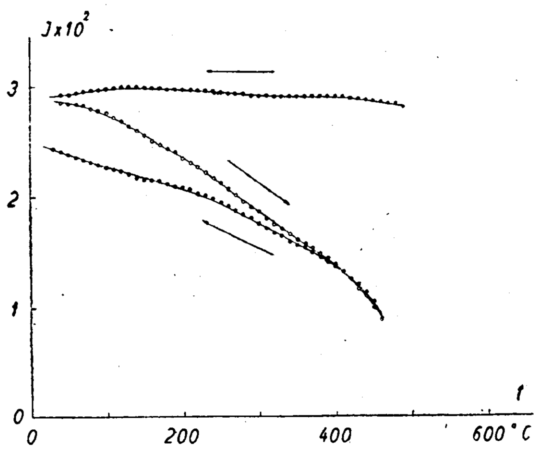

Fig. III-4·6. Specimen No 60′. $H=15$ *Oe*.

results that, although $J(t)$ in re-heating and re-cooling seems to follow approximately the Curie-Weiss law, in the initial cooling process it changes but little notwithstanding the decrease in temperature, whence it may be presumed that the time-effect in the initial cooling operation (1) is fairly marked, the increase in $p(\theta, t)$ due to that in $I(\theta, t)$ with decrease in temperature being almost nullified by its fairly large decrease with time. In re-heating and re-cooling, however, the time effect at high temperatures is not so marked as in operation (1), since the larger part of the unstable magnetization disappears during the initial cooling.

In Fig. 4–7 are shown, for comparison, the values of $J(t)$ in the operation of cooling from 490°C to 30°C in non-magnetic space, which magnetization was caused in $H=30{\cdot}5$ *Oe.* during the cooling from 600°C to 500°C at a slow rate, 1 degree/minute, say one-half the standard rate. Since the time-effect in this case ought apparently to diminish compared with the case shown by Fig. 4–6, the discrepancy between the curves in Figs. 4–6 and 4–7 is only natural.

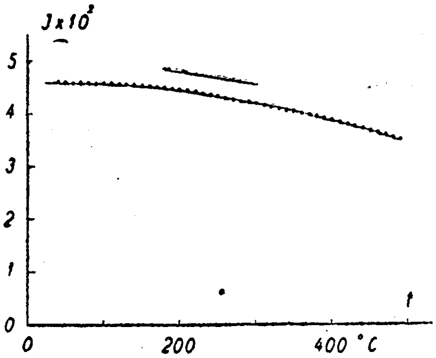

Fig. III-4-7. Specimen No 66'.
$H=30{\cdot}5$ *Oe.*

(vi) We may thus finally assume that eq. (4–4) approximately holds in the case of development of thermo-remanent magnetization in a weak magnetic field. The physical conditions for determining the function $\Psi(H, S, \theta)$, however, has not yet been cleared. We must mention here an important fact, pointed out by S. Kaya and Z. Harada,[3] that a single crystal of magnetite retains no thermo-remanent magnetism, notwithstanding its cooling from its Curie-point in a magnetic field, a fact clearly showing that the phenomenon of thermo-remanent magnetism is not a character of magnetite itself. On the other hand, according to A. Brun,[4] the so-called actual magnetite contained in volcanic rocks is not pure Fe_3O_4, but consists of Fe_3O_4, Fe_2O_3, FeO, and TiO_2 in various proportions, the ratio of FeO to Fe_2O_3 sometimes amounting to 4·7, whence the natural presumption that the thermo-remanent magnetism of rock is reduced to the character

3) The writer wishes to express his sincere thanks to Prof. S. Kaya and Prof. Z. Harada for acquainting him with their experimental results.

4) A. BRUN, *Arch. Sci. Ph. et Nat.* **6** (1924), 244.

of a solid solution or of an eutectic mixture of magnetite and other elements, TiO_2, Fe_2O_3, for example—a possibility also pointed out by S. Kaya.[5] If so, experimental examination of the magnetic properties of these ferro-magnetic minerals then becomes very necessary.[6]

Unfortunately, however, no systematic examination into the magnetic behaviours of the solid solution of Fe_2O_3 in Fe_3O_4 nor of that of TiO_2 in Fe_3O_4 have yet succeeded. We have so far at present but little experimental data of the thermo-remanent magnetism of impure magnetite; that is, a polycrystalline magnetite (containing a few per cent each of Fe_2O_3 and TiO_2) examined in our laboratory[7] showed a little thermo-remanent magnetization in a weak magnetic field (Curie-point about 610°C and $\frac{J_{t.}}{\chi} \simeq 1$), whereas, according to J. G. Königsberger,[8] two samples of polycrystalline magnetite (89% Fe_3O_4, 1·7% FeO and 5·3% SiO_2 in compostion) had also thermo-remanent magnetization; in this case the Curie-point was 590°C and $J_{t.}/\chi_0=6{\cdot}5$. These experimental facts would suggest that only impure polycrystalline magnetite is capable of retaining thermo-remanent magnetization. On the other hand, the general tendency of the equilibrium diagram between Fe_3O_4 and Fe_2O_3 was obtained by J. W. Grig, E. Posnjak, H. E. Merwin, and R. B. Sosman,[9] while that between Fe_3O_4 and $FeTiO_2$ was roughly determined by M. Kamiyama,[10] whose equilibrium diagrams are given in Figs. 4–8 and 4–9. From these results, we see that Fe_3O_4 and $FeTiO_2$ are soluble in each other to a limited content, likewise Fe_3O_4 and Fe_2O_3. (If, however, we consider more generally the equilibrium between Fe_2O_3, FeO, and TiO_2, the equilibrium will be found to be much more complex.) Besides, Brun's analysis[11] of magnetite contained in a number of effusive basaltic rocks showed that the magnetite consisted not only of Fe_3O_4 and $FeTiO_2$, but also of FeO in excess, as shown in Table 4–IV. Although the equilibrium condition between Fe_3O_4 and FeO is not known for certain at present,

5) S. Kaya, *Buturigaku-Kōensyū*, II (1942), pp. 100.

6) Although the experiment of separating the ferro-magnetic minerals from a few volcanic rocks was tried in the writer's laboratory also, we have not yet succeeded in seperating them perfectly, the ferro-magnetic minerals obtained always containing a few tens of per cent of SiO_2, Al_2O_3 and other elements.

7) T. Nagata, *Zisin*, 12 (1940), 301.

8) J. G. Königsberger, *Phys. ZS.*, 33 (1932), 468.

9) J. W. Grig, E. Posnjak, H. E. Merwin, and R. B. Sosman, *Amer. Journ. Sci.* 30 (1935), 239.

10) M. Kamiyama, *Journ. Geol. Soc. Japan* 36 (1929), 12.

11) A. Brun, *loc. cit.*

Table III-4-IV. Chemical compositions of ferro-magnetic minerals contained in volcanic rocks and their Curie-points. (after A. Brun and R. Chevallier).

Locality of sample.	Chemical composition				The relative amount of ferric-and ferrous oxides.		Curie-point
	TiO_2	Fe_2O_3	FeO	TiO_2+FeO	Fe_2O_3	FeO	θ
1 Krakatoa	13·85%	45·2%	41 %	54·8%	52·4%	47·6%	355°C
2 Gountour	12·3 ″	37·9″	49·7″	61·9″	43·4″	56·6″	305″
3 Kilauea	22 ″	31·8″	45·2″	67·2″	41·3″	58·8″	250″
4 Teyde	13 ″	37·5″	49·5″	62·5″	43·1″	56·4″	225″
5 Guimar	24·2 ″	21·3″	54·5″	78·7″	28·1″	72 ″	75″
6 Chaharra	16·9 ″	20·5″	62·5″	79·4″	24·7″	75·3″	40″
7 Chinyero	37 ″	11 ″	52 ″	89 ″	17·5″	82·5″	−30″

from the result of magnetic examination of these magnetites by R. Chevallier and J. Piére,[12] the relative amount of FeO in excess with reference to Fe_3O_4 was found to be larger, and the apparent Curie-point lower, or more precisely speaking, letting $C(Fe_2O_3)$,

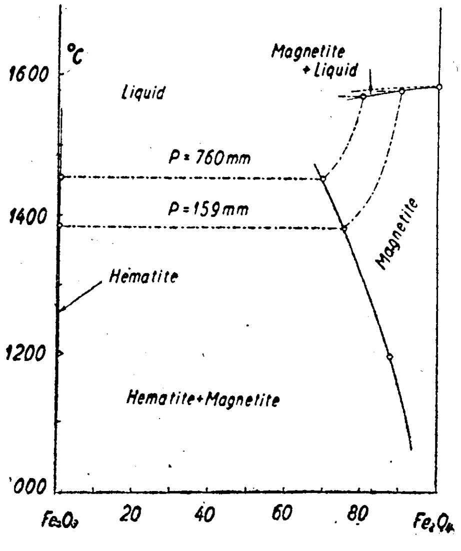

Fig. IV-4-3. Equilibrium diagram between Fe_3O_4 and Fe_2O_3. (After. J. W. Grig, E. Posnjak, H. E. Merwin, and R. B. Sosman).

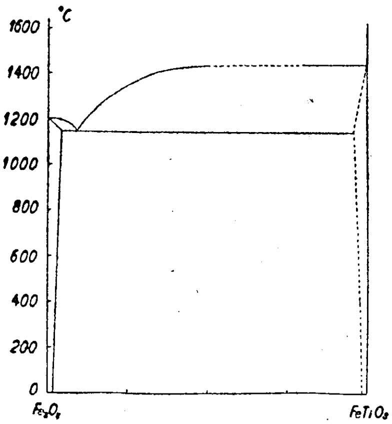

Fig. III-4-9. Equilibrium diagram between Fe_3O_4 and $FeTiO_3$. (After M. Kamiyama.)

$C(FeO)$ and θ denote respectively the weight percentages of Fe_2O_3 and FeO in the impure magnetite and its Curie-point temperature, we get the relation

12) R. CHEVALLIER et J. PIÉRE, *Ann. Physique*, 18 (1932), 383.

$$\theta(in^\circ K)\simeq s\frac{\theta_m}{0{\cdot}69}, \qquad s=\frac{C(Fe_2O_3)}{C(Fe_2O_3)+C(FeO)},$$

where θ_m denotes the Curie-point of pure magnetite (i.e. $\theta_m=580°C$ $=853°K$), the numeral 0·69 giving the critical value of s at which Fe_2O_3 and FeO compose the pure magnetite, Fe_3O_4. This experimental result seems to show that Fe_3O_4 and FeO also compose their solid solution, the Curie-point of which decreases with increse in impurity of Fe_3O_4 in exactly the same manner as in the case of the ferro-magnetic alloys of metals.

H. L. Alling,[13] J. G. Königsberger,[14] and R. Chevallier[15] also pointed out the existence of minerals composed of Fe_2O_3, FeO, and TiO_2 for example, ilmenite in the general form $m(FeO\cdot TiO_2)nFe_2O_3$, and titano-magnetite, $(Fe,Ti)_2O_3\cdot FeO$. Actually, the fact that the mean Curie-point or the transition temperature of the extraordinary mode of $\chi(t)$ and $P(t)$ in the Huzi ejecta, which are rich in TiO_2, is alway kept stable at temperature 200∼250°C probably suggests the existence of a fairly stable compound of Fe_2O_3, FeO, and TiO_2, such as those mentioned above. Here, the intergrowth of magnetite and ilmenite known as titano-magnetite (or titanoferrous magnetite) is due to exsolution.[14] The conclusion from the foregoing experimental facts may be that the ferro-magnetic minerals in volcanic rocks consist of pure magnetite, besides the intergrowth, the solid solution, and the eutectic mixtures of magnetite, ilmenite, ferric and ferrous oxides.

Taking, then, into consideration that the development of thermo-remanent magnetism is restricted to impure magnetite, it may be presumed that the development of thermo-remanent magnetization is probably due to the effect of exsolution of the impurities from magnetite or titanoferrous magnetite,[17] although the occurence of super-lattice phenomenon at temperatures below the Curie-point is also not

13) H. L. Alling, "*Interpretive Petrology of the Igneous Rocks*" pp. 140, New York (1936).

14) J. G. Königsberger, *loc. cit.*

15) R. Chevallier et J. Piére, *loc. cit.*

16) H. L. Alling, *loc. cit.*

17) Königsberger seems to be of the opinion that thermo-remanent magnetism is due to "crystallization remanence" (his own expression, Terr. Mag. **43**, 122, 1938.): that is to say, owing to the exsolution of impurites (chiefly Fe_2O_3), the remaining magnetite re-crystallizes, its lattice-constant changing at a high temperature where the coercive force is small and the susceptibility large, with the final result that the intense magnetization induced by the external magnetic at high temperature remains as remanent magnetization. However, the physical meaning of crystallization-remanence is still not yet clear.

disregarded. For instance, due to the exsolution of the impurities from magnetite, the lattice constant of the crystal will change, resulting in change also in its internal stress, which phenomenon is probably capable of bringing about $\Psi(H, S, \theta)$ in eq. 4–4 under the influence of an external magnetic field.[18]

Although the foregoing hypothesis is not unique, nor has its reliability been directly proved, yet the fact that the natural remanent magnetization of basic effusive rocks were subjected to relatively rapid cooling, and that the ferro-magnetic minerals in them consequently presented micro-crystals of indefinite shape in the ground mass (for example, bombs and the upper parts of lava flows) is always particularly intense, seems to support the possibility of this hypothesis being correct.

The foregoing discussion of the physical mechanism of the development of $\Psi(H, S, \theta)$, however, is not based on any experimental fact that directly proves its reliabitity. A trustworthy answer to this problem, therefore, is possible only after we have accumulated a sufficiently large number of experimental data of the magnetic behaviour of various kinds of solid solutions and of eutectic mixtures of Fe_2O_3, FeO, and TiO_2.

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18) Any conclusive theory of the physical mechanism of development of thermo-remanent magnetism ought to explain the fact that the development of magnetization under the influence of a magnetic field scarcely depends on the rate of cooling, provided the velocity is sufficiently small compared with that which is necessary to maintain the thermo-dynamical equilibrium between the various phases of all the elements in the rocks. Since the present hypothesis does not concern this important conclusion that the development of thermo-remanent magnetism does not depend on the rate of cooling, it is not conclusive.

Adding note. (*May*, 1943).

The hysteresis curve of various volcanic rocks (in magnetic field of from -4000 to $+4000$ *Oe.*) was recently determined. From this result, it was found that there is a close relation between the magnetic hardness and the ability of causing the thermo-remanent magnetization; that is, the rock sample having a large coersive force H_c always retains the intense thermo-remanence, the relation between H_c and Q_t being approximately proportional.

CHAPTER IV. RELATION OF NATURAL REMANENT MAGNETISM TO GEOMAGNETIC PHENOMENA.

§ 1. The Natural remanent magnetization of rocks as an indicator of secular variation in the geomagnetic field.

(i) As already mentioned in § 1, Chapter II, a number of investigators[1] have tried to trace the secular variation in the geomagnetic field from the direction of natural remanent magnetism of rocks ejected in ancient time or from that of baked earth in remote times. The important assumption used by these investigators that the natural remanent magnetization of rock has the direction of geomagnetic force affecting the rock during its cooling was confirmed by the result of our present study on a few recent volcanic ejecta, whence, we are justified in the belief that the curve showing secular variation in declination and dip of geomagnetism in various regions (chiefly in Europe) as presumed from the direction of natural-remanent magnetism of rocks and baked earths shows, at least approximately, its true value.

Now, since all volcanoes from which the rock samples examined by us were ejected are situated in the neighbourhood of the Kantō District, we can also trace the secular variation in geomagnetism in this district from the values of $\Delta\delta$ and θ of those rock samples. For this purpose the $\Delta\delta$ of the An'ei lava ejected in 1778 and of the Aokigahara lava ejected in 864 will be again shown, namely,

$$\Delta\delta = 6{\cdot}^{\circ}4E \pm 1{\cdot}^{\circ}9\,, \qquad (1778)\,,$$

$$\Delta\delta = 7{\cdot}^{\circ}9W \pm 0{\cdot}^{\circ}9\,, \qquad (864)\,.$$

As to declination, we have old records relating to the magnetic determination in various districts in Japan. These values of declination in olden times are summarized in Table 1–I, where the measurements by Sinzan Tani (Kôti, 1694) and by Tadataka Inô (Tokyo, 1802) were investigated historically and brought to notice by N. Sinozaki[2] and R. Otani[3] respectively.

The declination, in 1933, at the various places shown in Table 1–I were calculated by means of the empirical formula for the dis-

1) See the foot notes in § 1. Chapter II.
2) N. SINOZAKI, *Kagaku*, 8 (1938), 258.
3) R OTANI, "INŌ TADATAKA" pp. 516, (1917), *Tokyō*.

Table IV–1–I.

Place	Date	Observer		δ	$\Delta\delta$
Tokyo	1883	Y. Wada	a)	4° 16′·8W	1° 18′ E
〃	1882	I. Arai	b)	4° 24′ 〃	1° 11′ 〃
〃	1860	〃	c)	3° 11′ 〃	2° 24′ 〃
Yokohama	1854	M. C. Perry	d)	2° 44′ 〃	2° 45′ 〃
Ogiura	1827	British Navy	e)	1° 03′ E	4° 25′ 〃
Edo (Tokyo)	1802	T. Ino	f)	0° 19′ 〃	5° 55′ 〃
Kōti	1694	S. Tani	g)	5° 40′ 〃	10° 55′ 〃
Hirato	1613	J. Saris	h)	2° 50′ 〃	7° 35′ 〃

a), b), c) C. G. Knott. *Journ. Col. Sci. Tokyo Imp. Univ.* **3**.
d) Introduced by N. Sinozaki (*Kagaku* **8** (1938), 258).
e) *Bull. Hydro. Dep. Jap. Navy.* **8** (1936).
f) Historically researched by R. Otani. ("Inō Tadataka" pp 516).
g) Historically researched by N. Sinozaki (loc. cit.).
h) Introduced by N. Sinozaki (ibid.).

tribution of δ in the neighbourhood of Japan, which was supplied by the Hydrographic Department, Imperial Japanese Navy.[4] The differences in δ in the respective years given in Table 1–I from that in 1933 is denoted by $\Delta\delta$, which value is also given in Table 1–I.

On the other hand, as will be clear from Table 1–II, the secular variation in declination during the past 50 years at any point in south-western Japan (including Ogiura in the Ogasawara Islands, Hirato, Kôti, and Tokyo) seems to be almost parallel to each other, so far as there has been no marked change in the upper part of the earth's crust near the observing point caused by volcanic eruptions or earthquakes. More precisely speaking, however, there will be a difference of less than 20′ between the amounts of secular variation in declination during 50 years at two different points in the abovementioned region. Then, by extrapolating the foregoing result, we may presume that the $\Delta\delta$ given in Table I–1 will show the general tendency of secular variation in declination in the south-western districts of Japan, with an error of about one degree. Now, comparing these values of $\Delta\delta$ with the amount of $\Delta\delta$ in 1778 as determined from the direction of natural remanent magnetization of the An'ei lava, we find that they are in harmony; that is to say, plotting $\Delta\delta$ against the corresponding time in Fig. 1–1, we can express the relation between $\Delta\delta$ and time with a smooth curve, so that we can conclude that Fig. 1–1 shows the approximate tendency of secular

4) *Bull. Hydro. Dep. Jap. Navy*, **8** (1936).

Table IV-1-II. Average values of geomagnetic declination in a few regions in Japan.*

Region / Year	Northern part of Honsyū δ	Northern part of Honsyū diff.	Southern part of Honsyū δ	Southern part of Honsyū diff.	Middle part of Honsyū δ	Middle part of Honsyū diff.	S-W part of Honsyū and Sikoku δ	S-W part of Honsyū and Sikoku diff.	Kyūsyū δ	Kyūsyū diff.	Ogasawara Islands δ	Ogasawara Islands diff.
1895~1897	5°07·′6						4°37·′5					
1912~1913		+44·′4						+27·′2				
	5°52·′0		5°09·′8		5°23·′1		5°04·′7		4 29·′8		2°16·′4	
1922~1923		+22·′6		+24·′4		+20·′6		+20·′5		+17·′7		+20·′1
	6°14·′6		5°34·′2		5°43·′7		5°25·′2		4°47·′5		2°36·′5	
1932~1933		+12·′4		+ 9·′9		+ 8·′3		+ 9·′8		+ 8·′0		+ 5·′1
	6°27·′0		5°45·′1		5°52·′0		5°35·′0		4°55·′5		2°41·′6	

* (From *Bull. Hydro. Dep. Jap. Navy.* **8** (1936), 129).

variation in δ in the neighbourhood of Japan during the last 300 years.

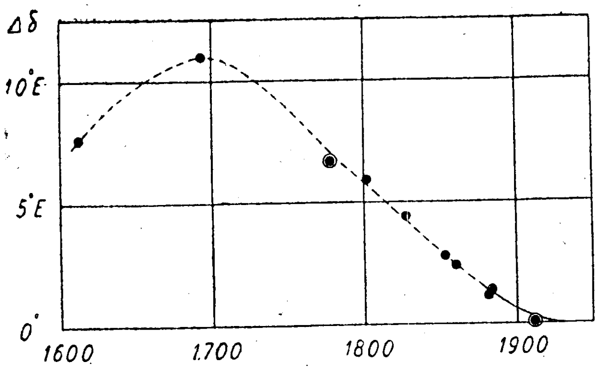

Fig. 1-1. The approximate tendency of secular variation in geomagnetic declination in the neighbourhood of Japan.

Consequently, that the $\Delta\delta$ determined from the natural remanent magnetization of old ejecta gives the amount of secular variation in declination during the time from its ejection to the present day is conclusively established.

Since, unfortunately, the data of $\Delta\delta$ during the time from 900 to 1600 is lacking, it is not possible to trace back the secular variation in δ continuously to 900 A.D. However, it may be said from the $\Delta\delta$ of Aokigahara lava that the geomagnetic meridian in Japan about 1100 years ago was about 8° west of that at the present time. This is the oldest data of geomagnetic declination for any known year at any known place so far obtained for various places in the world.

As to the geomagnetic dip in Japan prior to 1850, no historical record of measurement has been found. Judging only from our data of θ of the ejecta from Volcano Mihara (1778, 1911, and 1940), we may say that, generally speaking, the dip angle increased about 4 degrees during a period of nearly 150 years from 1778 to 1940, although this amount of decrease does not greatly exceed the limits of observational error, as will be seen from the following table.

Year	1778	1911	1940
θ	44·°3±1·°6	49·°0±2·°1	48·°4±1·°4

In contrast to what has just been said, the θ' of Aokigahara lava is fairly small compared with the geomagnetic dip observed by T, Minakami,[5] in 1941, at a point near where our samples were taken, that is,

$$\theta \text{ of lava, } 865 = 43^\circ\!\cdot 3 \pm 0^\circ\!\cdot 6,$$

$$\text{Mean value of dip observed in } 1941 = 51^\circ\!\cdot 1.$$

We may then conclude that the geomagnetic dip in Japan 1100 years ago was about 8° less than that at the present time. This result is also the oldest record of geomagnetic dip.

Lastly, we shall deal briefly with the direction of natural remanent magnetization of several old rocks that were ejected probably during Pleistocene Age from volcanoes Amagi and Taga. The deviation of the geomagnetic meridian $\varDelta\delta$ and the geomagnetic dip θ obtained in Chapter II are again given in Table 1–III.

Table IV–1–III.

Place	$\varDelta\delta$	θ
Aziro	26·°7W±2·°0	49·°7±1·°5
Omuro	8°E±4°	45°±3°
Yahatano	12·°7E ±1·°5	47·°3±1·°6

Although the $\varDelta\delta$ and θ in Izu peninsula during these remote times seem to differ a few tens of degrees from these at the present time, roughly speaking, the former approximately agree with the latter with a deviation of less than 30 degrees.

Here, we must mention the interesting conclusion given by Matuyama that the direction of natural remanent magnetization of some samples of basalt, which were probably ejected during the Tertiary Age, is nearly opposite to the direction of the present geomagnetic force. In contrast to this, the result of our study on the volcanic rocks ejected during the period from latest Tertiary to the present shows that the $\varDelta\delta$ and the deviation of θ from the present geomagnetic dip in central Japan are less than 30° and 10° respectively, which fact may show that the secular variation in geomagnetism during the Quarternary Age seems to be limited to a few tens of degrees in declination as in dip. It is interesting, however, that the geomagnetic meridian at the time the lava at Aziro was ejected (probably in later Tertiary) seems to deviate 27° west from the present one, although the dip angle at that time seems to be nearly the same as that at present.

(ii) Since the intensity of natural remanent magnetization J_n caused in a weak magnetic field is almost proportional to that of the

5) The writer wishes to express his hearty thanks to Dr. T. MINAKAMI for acquainting him with the result of geomagnetic survey.

applied magnetic field H, in which the rock was cooled from a temperature higher than its t_c to atmospheric temperature, the ratio of J_n to its intensity of saturated thermo-remanent magnetization J_{t_c} ought to agree with H, that is

$$\frac{J_n}{J_{t_c}}=H\,.$$

In §1, Chapter III, we already found that the specific intensity of natural remanent magnetization of "newly ejected" lava always approximately agrees with that of saturated thermo-remanent magnetization in a magnetic field of 0·45 Oersteds; that is,

$$J_n \doteqdot J_{t_c}\times 0{\cdot}45\,.$$

As to the volcanic rocks ejected in ancient times, there is, however, the possibility that its J_n fairly differs from its $J_{t_c}\times 0{\cdot}45$, since the total intensity of geomagnetic field at the time that rock was ejected might differ from the present value, and since the natural remanent magnetization once completely developed in remote time may slowly demagnetize during the extremely long interval from the time the rock solidified to the present.

Here, in order to examine the above-mentioned problem, the ratio of J_n to J_{t_c} of various volcanic rocks are given in Table 1–IV, where the ratio of J_n to $J_{t_c}\times 0{\cdot}45$ is also shown. As will be clear from these results, the J_n/J_{t_c} of newly ejected rocks (including volcanic rocks ejected in historical times), take the values of from 0·40 to 0·51, $J_n/(J_{t_c}\times 0{\cdot}45)$ being from 0·9 to 1·15, while the J_n/J_{t_c} of the volcanic rocks ejected in Pleistocene times (ejecta from Volcanoes Amagi, Usami, and Taga) assume values of 0·30∼0·35, that is, the $J_n/(J_{t_c}\times 0{\cdot}45)$ are from 0·4 to 0·8. The mean values of J_n/J_{t_c} and $J_n/(J_{t_c}\times 0{\cdot}45)$ of these two groups of rock ejected in the different epochs are respectively

$$J_n/J_{t_c}=0{\cdot}46\pm 0{\cdot}02\,,\qquad J_n/(J_{t_c}\times 0{\cdot}45)=1{\cdot}05\pm 0{\cdot}10\,,$$

$$J_n/J_{t_c}=0{\cdot}32\pm 0{\cdot}03\,,\qquad J_n/(J_{t_c}\times 0{\cdot}45)=0{\cdot}68\pm 0{\cdot}05\,.$$

So long, then, as we assume that the natural remanent magnetization of these rocks that were ejected and had cooled down to atmospheric temperature in remote geological times have scarcely been demagnetized during the long time interval until the present age, we can presume that the total intensity of geomagnetic field in Izu Peninsula was about 0·32 *Oe*, at that geological epoch.

Table IV-1-IV.

Specimen	J_n	J_{tc}	$J_{tc} \times 0{\cdot}45$	J/J_{tc}	$J_n/(J_{tc} \times 0{\cdot}45)$
(Old Ejecta)					
No. 17	$4{\cdot}20 \times 10^{-3}$	$12{\cdot}12 \times 10^{-3}$	$5{\cdot}45 \times 10^{-3}$	0·345	0·77
18	3·69 〃	10·94 〃	4·93 〃	0·336	0·75
21	5·51 〃	12·95 〃	5·83 〃	0·425	0·94
23	2·08 〃	5·18 〃	2·32 〃	0·402	0·90
27	0·58 〃	3·74 〃	1·68 〃	0·155	0·35
52	0·72 〃	2·76 〃	1·24 〃	0·260	0·58
54	2·61 〃	9·05 〃	4·07 〃	0·288	0·64
93	0·48 〃	4·61 〃	2·07 〃	0·104	0·23
94	1·73 〃	4·79 〃	2·16 〃	0·362	0·80
95	1·51 〃	5·09 〃	2·29 〃	0·297	0·66
96	1·65 〃	4·61 〃	2·07 〃	0·358	0·80
(Recent Ejecta)					
19	$4{\cdot}51 \times 10^{-3}$	$9{\cdot}66 \times 10^{-3}$	$4{\cdot}35 \times 10^{-3}$	0·456	1·04
25′	2·20 〃	5·07 〃	2·28 〃	0·434	0·96
58	$1{\cdot}40 \times 10^{-2}$	$2{\cdot}98 \times 10^{-2}$	$1{\cdot}34 \times 10^{-2}$	0·470	1·05
59	2·24 〃	3·83 〃	1·72 〃	0·585	1·30*
63	0·47 〃	1·17 〃	0·53 〃	0·402	0·89
65	0·51 〃	1·19 〃	0·54 〃	0·429	0·93
68	1·44 〃	3·35 〃	1·50 〃	0·430	0·96
70	1·63 〃	3·48 〃	1·57 〃	0·468	1·04
71	1·41 〃	2·84 〃	1·28 〃	0·496	1·10
73	$4{\cdot}89 \times 10^{-3}$	$12{\cdot}47 \times 10^{-3}$	$5{\cdot}60 \times 10^{-3}$	0·391	0·88

* The sample No. 59 was taken off from the east end of a large mass of Akabakkyō lava flow which was flowed down from the crater and accumulated in sea, whence it was quenched in the sea water. (See H. Tsuya and others, *Bull. Earthq. Res. Inst.*, 19 (1941), 260). The marked inequality between J_n and $J_{tc} \times 0{\cdot}45$ may be due to the above-mentined special condition in its cooling.

Since it will be rather illogical to assume that the remanent magnetization of rocks have not been demagnetized during such a long period as that far exceeding 10,000 years, it would be more in order to presume that the phenomenon that the $J_n/(J_{tc} \times 0{\cdot}45) < 1$ in the cases of these old ejecta is largely due to the effect of demagnetization during the long period.

A trustworthy answer to this problem,[6] however, can be made only after we have accumulated a large number of similar data covering the various rocks ejected in various epochs for the largest possible number of places on the earth's surface.

6) This problem has been studied only by Königsberger. (*Terr. Mag.*, 43 (1938), 307).

§ 2. Local magnetic anomaly and natural-remanent magnetism.

A large number of geophysicists have observed the local anomalous distribution of geomagnetic field in various regions over the earth's surface. The unanimous conclusion of all of these investigators is that the local anomaly in the geomagnetic field is closely related to the geological structure near the earth's surface.[1] Especially, it is a well known fact that the geomagnetic field in the neighbourhood of basaltic rocks is markedly disturbed.[2] These local anomalies are believed to be due to the heterogeneous distribution of magnetization of the earth's crust near its surface, and scarcely to electro-magnetic induction due to earth currents or to the earth-air current, since, in practice, magnetic field intensity induced by such kinds of electric current as just mentioned is negligible[3] compared with that of the observed geomagnetic anomaly. There is, however, a question regarding the cause of the magnetization of the earth's crust, that is, whether it is reversible magnetization induced by the normal geomagnetic field in the rocks composing the earth's crust or by permanent magnetization of the rock itself such, for example, as natural remanent magnetism. If we adopt the former assumption, the intensity of magnetization of rock is approximately given by $\kappa\vec{H}$, where H denotes the present normal geomagnetic force at the very point of observation, whereas if we take the latter also into consideration, it becomes $\kappa\vec{H}+\vec{J}_n$, where the direction of $\vec{J}_n$ does not always agree with that of $\vec{H}$.

Obviously, it is not possible to judge from the distribution of the anomalous geomagnetic field alone which assumption best fits in with the natural phenomenon, judgement being possible only by comparing the magnetic properties of the rock experimentally ascertained with the value of the magnetization it requires from a magnetic survey. On the other hand, from our studies on the magnetic properties of volcanic rocks, almost all the rocks examined have rather intense natural remanent magnetization, showing that the presence of this permanent magnetization in volcanic rocks ought to disturb in a fairly marked manner the geomagnetic field in space near a large mass of volcanic rock. From this point of view, a few examples of local geomagnetic anomaly in volcanic regions will now be discussed.

1), 2) See the literatures given the foot notes (15)～(25) of this paragraph.
3) L. A. BAUER, *Terr. Mag.*, **25** (1920), 145; **28** (1923), 1.
G. ANGENHEISTER, *Phys. ZS.*, **26** (1925), 305.
W. J. PETERS, *Terr. Mag.*, **28** (1923), 83.

(i) Anomalous distribution of geomagnetic declination around a volcanic crater.

The distribution of geomagnetic declination at 19 points around the crater of Volcano Mihara, in Oosima Island, was measured with the aid of a magnetic compass,[4] the result being shown in Fig. 2–1

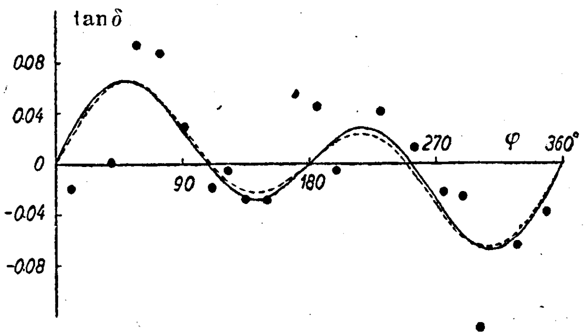

Fig. 2–1. Anomalous distribution of geomagnetic declination around the crater of Volcano Mihara.
Full line; approximate curve for observed values.
Dotted line; theoretically calculated.

and in Table 2–I, where φ denotes the azimuthal angle of the observing point measured clockwise from the geomagnetic south, as view from the centre of the crater. As will be seen from this figure, the distribution around the crater is nearly antisymmetric with respect to the geomagnetic meridian, whence the relation between φ and $\tan\delta$ is approximately given by

$$\tan\delta = 0{\cdot}0283\sin\varphi + 0{\cdot}0468\sin 2\varphi\,. \qquad (2\text{–}1)$$

Table IV–2–I.

Azimuth φ	Anomaly in Declination δ	φ	δ
12	−1° 23′	186°	+2° 35′
40	0 00	200	−0 20
58	+5 20	232	+2 20
75	+5 00	255	+0 40
92	+1 40	277	−1 20
112	−1 10	290	−1 30
123	−0 20	310	−7 20
135	−1 30	328	−3 40
152	−1 40	348	−2 15
170	+3 10		

4) T. NAGATA, *Bull. Earthq. Res. Inst.*, **16** (1938), 288. *Zisin*, **10** (1938), 87; 214.

This fact may suggest that the observed geomagnetic anomaly is largely due to the particular topography around the crater, it being nearly symmetric with respect to the centre of the crater.[5]

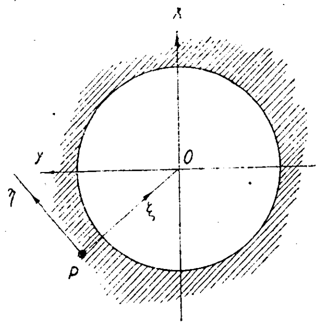

Fig. 2-2.

Here, as an interpretation of this observed result, we shall assume an anomalous distribution of geomagnetic declination which is due to uniform magnetization of the central cone of the volcano, the topography of which is quite symmetric with respect to its center. In Fig. 2-2. the circle represents the boundery of the crater, the coordinate systems (x, y, z) and (ξ, η, ζ) being taken as shown in the figure, where x and y point respectively to geomagnetic north and west. Then, H_0, Z_0, and V denote respectively the horizontal and vertical intensities of the mean geomagnetic field and the gravitational potential at P due to the mass of the central cone, the ξ- and η-components of the geomagnetic field at P, H_ξ and H_η, being given by

$$H_\xi = H_0 \cos\varphi + \frac{J}{k^2\rho}\left(\alpha\frac{\partial^2 V}{\partial\xi^2} + \beta\frac{\partial^2 V}{\partial\xi\partial\eta} + \gamma\frac{\partial^2 V}{\partial\xi\partial\zeta}\right),$$

$$H_\eta = H_0 \sin\varphi + \frac{J}{k^2\rho}\left(\alpha\frac{\partial^2 V}{\partial\xi\partial\eta} + \beta\frac{\partial^2 V}{\partial\eta^2} + \gamma\frac{\partial^2 V}{\partial\eta\partial\zeta}\right), \tag{2-2}$$

where J, ρ, k^2 and (α, β, γ) denote respectively the intensity of magnetization, the mean density of the central cone, the gravitational constant, and the direction-cosines of magnetization with respect to ξ, η, and ζ. Now, if we assume that the direction of magnetization agrees with that of the geomagnetic meridian and that the topography around the crater is symmetric with respect to its centre, eq. (2-2) becomes

5) The low degree terms in the harmonic series of $\tan\delta$ with respect to φ is not perfectly antisymmetric with regard to the geomagnetic meridian, as shown in the following formula.

$$\tan\delta = 0{\cdot}0302\sin\varphi - 0{\cdot}0195\cos\varphi + 0{\cdot}0455\sin 2\varphi$$
$$- 0{\cdot}0062\cos 2\varphi - 0{\cdot}0053\sin 3\varphi - 0{\cdot}0122\cos 3\varphi + 0{\cdot}0004.$$

Since, however, if we take into consideration the inequality of phase of each degree term, the mathematical form becomes complex, the amplitude of a certain degree terms which are necessary for physical interpretation being the same order as the observational error, here we deal only with the first approximating expression of $\delta \sim \varphi$ relation, as given by eq. (2-1).

$$H_\xi = H_0 \cos\varphi + \frac{J}{k^2\rho}\left\{\cos I_0 \cos\varphi \frac{\partial^2 V}{\partial \xi^2} + \sin I_0 \frac{\partial^2 V}{\partial\xi\partial\zeta}\right\}, \tag{2-3}$$

$$H_\eta = H_0 \sin\varphi ,$$

because in this case

$$\frac{\partial^2 V}{\partial\xi\partial\eta} = \frac{\partial^2 V}{\partial\eta^2} = \frac{\partial^2 V}{\partial\eta\partial\zeta} = 0$$

where I_0 is the geomagnetic dip, i.e. $\tan I_0 = Z_0/H_0$. The geomagnetic north and west components of the total geomagnetic force, H_x and H_y, are then given by

$$\begin{aligned} H_x &= H_\xi \cos\varphi + H_\eta \sin\varphi \\ &= H_0 + \frac{J\cos I_0}{k^2\rho}\left\{\frac{\partial^2 V}{\partial\xi^2}\cos^2\varphi + \frac{\partial^2 V}{\partial\xi\partial\zeta}\tan I_0 \cos\varphi\right\}, \\ H_y &= -H_\xi \sin\varphi + H_\eta \cos\varphi \\ &= -\frac{J\cos I_0}{k^2\rho}\left\{\frac{\partial^2 V}{\partial\xi\partial\zeta}\tan I_0 \sin\varphi + \frac{1}{2}\frac{\partial^2 V}{\partial\xi^2}\sin 2\varphi\right\}. \end{aligned} \tag{2-4}$$

Here, if δ denotes the anomaly in declination at a station, we get

$$\tan\delta = \frac{H_y}{H_x} = \frac{J}{F_0 k^2\rho}\left\{\frac{\partial^2 V}{\partial\xi\partial\zeta}\tan I_0 \sin\varphi + \frac{1}{2}\frac{\partial^2 V}{\partial\xi^2}\sin 2\varphi\right\}, \tag{2-5}$$

provided $\frac{H_0 k^2\rho}{J} \gg \frac{\partial^2 V}{\partial\xi^2}, \frac{\partial^2 V}{\partial\xi\partial\zeta}$, and F_0 denotes the total intensity of the geomagnetic field, i.e. $H_0 \sec I_0 = F_0$.

This relation agrees with the empirical formula given by eq. (2–1). Calculating the actual values of $\frac{\partial^2 V}{\partial\xi^2}$ and $\frac{\partial^2 V}{\partial\xi\partial\zeta}$ due to the mean topography around the crater by numerical integration, we get from eq. (2–5).

$$\tan\delta = \frac{J}{F_0}(2{\cdot}41 \tan I_0 \sin\varphi + 3{\cdot}26 \sin 2\varphi), \tag{2-6}$$

Then, comparing eq. (2–1) with eq. (2–6), where I_0 and F_0 are taken as 45° and 0·45 *Oe.* respectively, we get for the most probable value of J

$$J = 0{\cdot}0058\, e.m.u., \tag{2-7}$$

with the result that eq. (2-6) becomes

$$\tan\delta = 0{\cdot}031\sin\varphi + 0{\cdot}043\sin 2\varphi\,.$$

Thus, the anomalous distribution of geomagnetic declination around the crater of Volcano Mihara can be interpreted as magnetic anomaly, largely due to the uniform magnetization of the central cone in the direction of geomagnetic force.

By assuming, next, that the magnetization of the rock composing the central cone is solely induced magnetization in the present geomagnetic field, the magnetic susceptibility of rock ought to be $\chi = J/H_0\rho = 5.2\times10^{-3}$ e.m.u., where ρ is assumed to be 2·5.

While the actual value of χ of the central cone of Volcano Mihara, which was experimentally obtained, amounts only to from $0{\cdot}4\times10^{-3}$ to $1{\cdot}4\times10^{-3}$, these values being too small compared with that expected from the geomagnetic anomaly. On the other hand, the direction of natural remanent magnetization of the Meizi-Taisyō lava and the An'ei lava almost agrees with that of the present geomagnetic force, while their specifiic intensity amounts to from $1{\cdot}6\times10^{-2}$ to $3{\cdot}6\times10^{-2}$, these values being too large compared with the value of J/ρ, i.e. $2{\cdot}3\times10^{-3}$. The central cone of Volcano Mihara, however, is formed not only of continuous ropy lava, but also of bombs, lapilli, ashes, and other pyroclastic materials, so that by assuming that a lava sheet 6 or 7 m thick with natural remanent magnetization $J_n = 0{\cdot}025$ and specific susceptibility $\chi = 6{\cdot}5\times10^{-4}$ covers the body of the central cone, composed of pyroclastic ejecta, the magnetic susceptibility of which is almost the same as that of lava, but without any remanent magnetization in the sense of average character of the whole body, the interpretation of the magnetic survey is in good agreement with the result of experimental studies on the magnetic properties of Mihara rocks; that is, the value of $\tan\delta$, which was numerically calculated by means of eq. (2-5) under the above-mentioned assumption, exactly agrees with eq. (2-7)

A very similar phenomenon was observed around the crater of Volcano Asama.[6] The distribution of $\tan\delta$ around the crater of Asama, observed by T. Fukutomi, is shown in Fig. 2-3 and Table 2-II, from which we get the empirical formula of the relation between φ and $\tan\delta$ as

6) T. FUKUTOMI, *Zisin*, 2 (1930), 641.
T. NAGATA, *Bull. Earthq. Res. Inst.*, 16 (1938), 288.

Table IV–2–II.

Azimuth φ	Anomaly in declination δ	φ	δ
0°	+0° 03′	198	+0° 40′
24	+0 35	210	+0 40
52	+1 35	226	+0 35
72	+2 25	246	+0 20
88	+0 50	262	+0 10
103	−0 25	275	−0 35
118	−0 55	288	−1 50
128	−0 55	300	−2 05
145	+0 08	322	−1 20
179	+0 08	343	−0 25

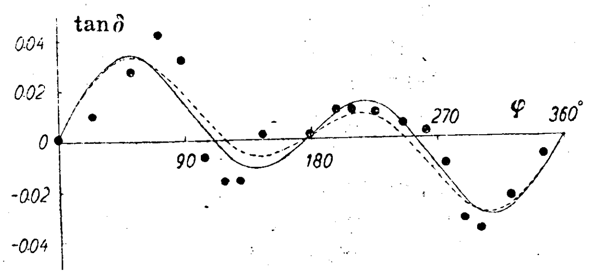

Fig. 2–3. Anomalous distribution of geomagnetic declination around the crater of Volcano Asama.

$$\tan\delta = 0{\cdot}0133\sin\varphi + 0{\cdot}0221\sin 2\varphi\,,^{7)} \tag{2–8}$$

while the result of numerical calculation by means of eq. (2–5) gives

$$\tan\delta = \frac{J_0}{H_0}(2{\cdot}54\tan I_0\sin\varphi + 3{\cdot}05\sin 2\varphi)\,. \tag{2–9}$$

Then, assuming also that $I_0 = 45°$, $H_0 = 0{\cdot}45\,Oe.$, we get from eq. (2–9)

$$J = 0{\cdot}0025\,,$$

$$\tan\delta = 0{\cdot}015\sin\varphi + 0{\cdot}017\sin 2\varphi\,. \tag{2–10}$$

On the other hand, the specific susceptibility of the ejecta from Asama is $0{\cdot}5\times10^{-3}\sim1{\cdot}1\times10^{-3}$, while the specific intensity of natural remanent

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7) Expressing $\tan\delta$ as the form of general Fourier series, we get

$$\tan\delta = 0{\cdot}0141\sin\varphi - 0{\cdot}0019\cos\varphi + 0{\cdot}0196\sin 2\varphi$$
$$- 0{\cdot}0032\cos 2\varphi - 0{\cdot}0012\sin 3\varphi - 0{\cdot}0019\cos 3\varphi + 0{\cdot}0006\,.$$

So far as we concern the low degree terms, the cos-terms are negligibly small compared with the sine-terms.

magnetization is $1{\cdot}3\times10^{-3}\sim2{\cdot}5\times10^{-3}$, as given in Table 2–III. Comparing these observed values of χ and J_n with the specific values of susceptibility and intensity of magnetization expected from eq. (2–10), namely, $\frac{J}{\rho H_0}=2{\cdot}2\times10^{-3}$ and $\frac{J}{\rho}=1{\cdot}0\times10^{-3}$, we see that χ is fairly smaller than $\frac{J}{\rho H_0}$, while J_n is a little larger than $\frac{J}{\rho}$. Hence in this case also, we may presume that the central cone of Volcano Asama is composed of 40% lava with a magnetization of $\rho(J_n+H_0\chi)$ and 60% pyroclastic ejecta with only induced magnetization $\rho H_0\chi$, where we assume that J_n and χ take the mean of their observed values, i.e. $J_n=1{\cdot}7\times10^{-3}$ and $\chi=0{\cdot}7\times10^{-3}$.

Table IV–2–III.

No.	χ_0	J_n
5	$0{\cdot}84\times10^{-3}$	—
6	0·88 〃	—
8	1·18 〃	$1{\cdot}5\times10^{-3}$
11	0·17 〃	—
14	0·57 〃	—
16	0·61 〃	$1{\cdot}3\times10^{-3}$
73	0·50 〃	2·5 〃

(ii) Magnetic anomaly near volcanoes.

The geomagnetic field around a few Japanese volcanoes were measured by a number of investigators, all of whose results showed marked anomalous distribution of geomagnetic force near the volcanoes.

(1) The magnetic survey of Oosima Island was made by R. Takahasi and the writer[8] in 1936, from the result of which survey, it was found that the anomalous magnetic field in Oosima is chiefly due to the mass of Mihara volcano and its subterranean mass, which are magnetized in direction nearly the same as that of the present geomagnetic force.[9]

On the other hand, as the result of a torsion balance survey on this volcanic island made by the writer,[10] in 1938, where eight out of the eighteen stations almost agree with those eight out of the twelve stations where the magnetic measurement was made, so that provided we assume that the observed geomagnetic anomaly is due to uniform magnetization of the excess mass that is the cause of the observed anomaly in the second derivatives of the gravity field, between these two quantities observed at the same station, we get the relation

$$\varDelta H_y=\frac{J}{k^2\rho}\left(\alpha\frac{\partial^2V}{\partial x\partial y}+\beta\frac{\partial^2V}{\partial y^2}+\gamma\frac{\partial^2V}{\partial y\partial z}\right), \qquad (2\text{–}11)$$

the notations being the same as in eq. (2–2).

8) R. TAKAHASI and T. NAGATA, *Bull. Earthq. Res. Inst.*, **15** (1934), 441.
9) R. TAKAHASI and T. NAGATA, *Zisin*, **9** (1937), 434.
10) T. NAGATA, *Bull. Earthq. Res. Inst.*, **17** (1939), 93.

Table IV-2-IV

Observed Station	$\frac{\partial^2 V}{\partial x \partial y} \times 10^9$	$\frac{\partial^2 V}{\partial y \partial z} \times 10^9$	ΔH_y in *e.m.u.*	$\left(\cos I_0 \frac{\partial^2 V}{\partial x \partial y} + \sin I_0 \frac{\partial^2 V}{\partial y \partial z}\right) \times 10^9$	J/ρ
Motomura.	−35·6	+49·6	+0·0009	+ 5·9	0·010
Nomasi.	−16·6	+37·4	+0 0054	+13·1	0·027
Senzu.	+80·0	−77·4	+0·0031	+ 9·1	0·023
Dōbutuen.	+19·1	− 3·6	−0·0027	+11·9	(−0·015)
Sabaku. I	−14·3	+90·9	+0·0218	+48·0	0·030
〃 III	+31·0	−68·9	−0·0130	−22·1	0·039
〃 IV	−76·8	−18·2	−0·0303	−79·7	0·026
〃 V	−30·0	+130·5	+0·0275	+63·4	0·029

By taking (x, y, z) as the magnetic north, east, and the downward rectangular coordinate system,[11] eq. (2–11) becomes

$$\Delta H_y = \frac{J}{k^2 \rho}\left(\cos I_0 \frac{\partial^2 V}{\partial x \partial z} + \sin I_0 \frac{\partial^2 V}{\partial y \partial z}\right), \tag{2–12}$$

because in this case $\beta = 0$. Here, the magnitude of $\frac{\partial^2 V}{\partial x \partial y}$ and $\frac{\partial^2 V}{\partial y \partial z}$ were known from the results of the torsion balance measurement, while that of ΔH_y found by the magnetic survey.

Then, putting these observed quantities in eq. (2–11), it is possible to calculate the specific intensity of magnetization J/ρ for each observing station. The actual values of $\frac{\partial^2 V}{\partial x \partial y}$, $\frac{\partial^2 V}{\partial y \partial z}$, and ΔH_y are given in Table 2–IV, while $k^2 \Delta H_y$ are plotted against $\left(\cos I_0 \frac{\partial^2 V}{\partial x \partial y} + \sin I_0 \frac{\partial^2 V}{\partial y \partial z}\right)$ as shown in Fig. 2–4, where the angle of dip I_0 is assumed to be 48°46′, the mean of the dip observed at 12 stations.

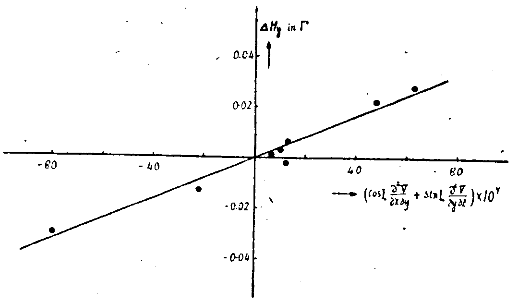

Fig. 2–4. The relation between the east component of amomalous geomagnetic field and the Eötvös quantity. (Oosima Island).

Now, allowing for a possible slight error in assuming the mean value of

11) The direction of x-axis deviates 5° 28′ west from the geographic north, since the mean value of declinati n observed at 12 stations takes that value.

the observed declination to be the azimuth of normal declination in the region considered, we can write

$$k^2 \Delta H_y = C + \frac{J}{\rho}\left(\cos I_0 \frac{\partial^2 V}{\partial x \partial y} + \sin I_0 \frac{\partial^2 V}{\partial y \partial z}\right).$$

Applying the method of least square with the aid of eq. (2–12′), to the observed values given in Table 2–IV, we get

$$\frac{J}{\rho} = 0{\cdot}027_2 \pm 0{\cdot}001_4 . \qquad (2\text{–}12')$$

The magnitude of the specific intensity of magnetization thus established is fairly large, being almost the same as that of the natural remanent magnetization of An'ei lava which, obviously, far exceeds the induced magnetization $H\chi$ of any rock specimen observed in our experiment. From this result, we may presume that the subterranean rock under the earth's surface of volcanic island Oosima ought to have fairly intense remanent magnetization.

(2) The magnetic survey of the neighbourhood of Volcano Asama was carried out by T. Minakami,[12] from which result, he found that the largest part of the magnetic anomaly in this region seems to be due to the magnetized mass of Asama Volcano. Replacing the effect of the mass of volcano by a rotational ellipsoid of suitable dimension, uniformly magnetized in the direction of the present geomagnetic force he found that the intensity of magnetization is about 0·0022 *e.m.u.* Since this intensity of magnetization nearly agrees with that obtained from the magnetic survey around the crater of Asama Volcano, the interpretation given in the latter case holds, as also in this case.

(3) In 1940, Takahasi and Hirano[13] made a magnetic survey of Volcano Miyakesima, using a Z-variometer, Minakami[14] doing so with a dip-circle. From the magnetic survey, the mean intensity of magnetization of Miyakesima is estimated to be about 0·015∼0·020 *e.m.u.*, which value greatly exceeds the induced magnetization of Miyakesima rock, while it is less than that of their natural remanent magnetization (Table 2–V), so that, in this case also, the natural conclusions are that Miyakesima is composed of lava flow of $\rho(J_n + \chi H)$ and pyroclastic ejecta of only $\rho\chi H$, that is, it is a composite stratified volcano.

12) T. MINAKAMI, *Bull. Earthq. Res. Inst.*, **16** (1938), 100; **18** (1940), 178.
13) R. TAKAHASI and K. HIRANO, *ibid.*, **19** (1941), 82, 373.
14) T. MINAKAMI, *ibid.*, **19** (1941), 356.

We have, at present, a large number of other data of the geomagnetic anomaly near volcanoes or volcanic regions, for examples, A Tanakadate's[15] work on Mt. Huzi, H. Hatakeyama's[16] on Mt. Nasu and Mt. Daiton, W. Yoda's[17] on Mt. Mayu, M. Hasegawa's[18] on Mt. Aso, S. T. Nakamura's[19] on Mt. Komagadake, T. Minakami's[20] on Mt. Sakurazima and Mt. Kusatu-sirane, T. Yumura's[21] on Towada Lake, the writer's[22] on Niisima (a volcanic island), J. P. Rothe's[23] on several volcanoes in France, H. Reich's and A. Nippoldt's[24] on a number of volcanoes in Germany, the members' of D. T. M.[25] on the Guatemala volcanic region, and others. All their results show that there is a marked anomalous distribution of geomagnetic field in volcanic regions. We can presume that this anomalous field is due to both induced magnetization and natural remanent magnetization of the rocks composing volcanoes. On the other hand, from our experimental studies, the amount of magnetic susceptibility of effusive rock is always smaller than 0·01 *e.m.u.* per cm^3, whence it may be said that if the intensity of magnetization of volcanic rocks as estimated from the geomagnetic anomaly exceeds 0·005 *e.m.u.*, a part of the magnetization, at any rate, is due to the presence of natural remanent magnetization in rocks.

Table IV-2-V.

No. of Specimen	χ_0	J_n
58	$0{\cdot}76\times10^{-3}$	$1{\cdot}40\times10^{-2}$
59	1·31 〃	2·24 〃
60	1·12 〃	(2·15 〃)
63	0·63 〃	0·47 〃
64	0·80 〃	0·71 〃
65	0·74 〃	0·51 〃
75	1·19 〃	(0·68 〃)

15) A. TANAKADATE, *Journ. Col. Sci. Tōkyo*, 2 (1889). *Proc. Tōkyo Math. Phys. Soc.*, 2 (1904), 149, 405.
16) H. HATAKEYAMA, *Bull. Cent. Meteoro. Obs. Japan, Vol. No.* 2. (1937).
17) W. YODA, *Tikyū-Buturi*, 3 (1939), 26.
18) H. HASEGAWA, Read before the meeting (June 1942) of geophysical section of Gakuzyutu-Kenkyū Kaigi.
19) S. T. NAKAMURA, *Proc. Imp. Acad. Japan*, 11 (1935), 102.
20) T. MINAKAMI, *Bull. Earthq. Res. Inst.*, 16 (1938), 117; H. TSUYA and T. MINAKAMI, *ibid*, 18 (1940), 335.
21) T. YUMURA, *Kakioka Tiziki-Kansokusyo Yōhō*, (*Mem. Kakioka Mag. Obs.*) *Vol.* 2, *No.* 4.
22) T. NAGATA, *Bull. Earthq. Res. Inst.*, 15 (1937), 497.
23) J. P. ROTHE, *Ann. Inst. Phys. Globe*, 15 (1937), 1.
24) H. REICH, *ZS. Geophys.*, 4 (1928), and others.
A. NIPPOLDT, *Naturwiss.*, 3 (1915), 349.
25) A. G. MCNISH, *Rep. Intern. Assoc. Volcanology*, (*Congr. Sept.* 1939, *Washington*).
J. A. FLEMING, *Carnegie Inst. Washington, Year book*, 39 (1939~1940), 92. (*Ann. Rep. of Director of D.T M.* 1940).

It must be noted here, however, that the magnetic properties of igneous rocks sometimes fairly differ owing to slight differences in their crystallization, even if there are but slight differences in their chemical composition. For example, it was sometimes found that a piece cut off from a large mass of lava flow fairly differed in natural remanent magnetism as well as in magnetic susceptibility from another piece obtained from the same mass a few tens of meters distant from the position of the first piece, notwithstanding that the petrographic characters of these two pieces were virtually the same. Hence in dealing with the details of anomalous distribution of the geomagnetic field in a volcanic region it may turn out to be very complex as the natural result of such heterogeneous distribution of magnetization. The general tendency of geomagnetic anomaly, however, can be said to be the result of nearly uniform magnetization, permanent and induced, of rocks composing the volcanoes as mentioned in connexion with the preceding few examples.

§ 3. Local anomalous change in geomagnetism and development of thermo-remanent magnetism.

Since it is an established fact, as already mentioned in the foregoing paragraph, that the geological structure near the earth's surface, which consists of various kinds of rock, generally accompanies local geomagnetic anomaly of the neighbourhood of that structure, it should also be possible that the geomagnetic field changes locally, accompanying changes in physical and chemical conditions of the earth's crust, for example, volcanic activity and earthquakes. The question whether or not any marked geomagnetic change is caused by occurrence of earthquakes or volcanic activities has been raised by a number of investigators, for example, A. Tanakadate and H. Nagaoka,[1] S. T. Nakamura and Y. Kato,[2] S. Chapman,[3] J. P. Rothé,[4] H. Nagaoka,[5] T. Minakami,[6] R. Takahasi,[7] and the writer.[8] In answer to this

1) A. TANAKADATE and H. NAGAOKA, *Journ. Col. Sci. Tōkyo Imp. Univ.*, **1** (1893), 149.
2) S. T. NAKAMURA and Y. KATO, *Proc Imp. Acad Japan*, **10** (1934), 256; Y. KATO, *Sci. Rep. Tōhoku Imp. Univ.* **27** (1938), 1.
3) S. CHAPMAN, *Terr. Mag.*, **35** (1930), 81.
4) J. P. ROTHE, *Ann. Inst. Phys. Globe*, **15** (1937), 1.
5) H. NAGAOKA and T. IKEBE, *Proc. Imp. Acad. Japan*, **13** (1937), 62, 251.
6) T. MINAKAMI, *Zisin*, **10** (1938), 430.
7) R. TAKAHASI and K. HIRANO, *Bull. Earthq. Res. Inst.*, **19** (1941), 82.
8) T. NAGATA, *Bull. Earthq. Res. Inst*, **15** (1937), 497; **19** (1941), 335, 402. and others.

question, a number are of the opinion that volcanic activity, sometimes at any rate, results in considerable local geomagnetic changes, for example, Nakamura[9] in the case of Komagadake Volcano, Nagaoka[10] and the writer[11] in the case of Asama Volcano, as also Takahasi,[12] Minakami,[13] Kato,[14] and the writer[15] respectively in the case of Miyake-sima Volcano. The fact that all geomagnetic observations made after the great eruption of Miyake-sima, in 1940, which were carried out independently by the four investigators just mentioned, show the existence of local geomagnetic changes almost alike in the order of magnitude, is probably the most positive answer to that question. At the same time, we have not yet any sufficiently reliable results of studies made on the relation between geomagnetic change and the occurrence of earthquakes, although a few investigators have attempted to find a possible correlation between these two phenomena, the result being sometimes positive, and at other times negative.

For this reason, only a few typical examples of local anomalous changes in geomagnetism, presumed to accompany volcanic activity, will be briefly dealt with here, with a discussion on its possible relation to changes in the magnetic behaviour of the rock composing the volcano.

(i) Anomalous geomagnetic changes observed on Volcano Miyake-sima after its severe eruption.

Soon after the severe volcanic activity of Miyake-sima began in July 1940, a geomagnetic survey of this island was made; Takahasi and Hirano measured the distribution of vertical intensity of the geomagnetic field, using a vertical field balance, Minakami a dip-circle, Y. Kato an earth-inductor, and obtained the differences in geomagnetic field for different times, while the writer, with the aid of a magnetograph, made continuous observations of changes in inclination in geomagnetic field at a point near the volcano.

The magnetograph was a simplified type dip-variometer,[16] designed and constructed by the writer, containing, as the principal part of the instrument, only a bar magnet of M.K. steel supported on agate plates by a knife edge, with which the centre of gravity of the magnet

9) S. T. NAKAMURA, *Proc. Imp. Acad. Japan*, **11** (1935), 102.
10) H. NAGAOKA, *loc. cit.*
11) T. NAGATA, *Bull. Earthq. Res. Inst.*, **19** (1941), 350.
12) R. TAKAHASI, *loc. cit.*
13) T. MINAKAMI, *Bull. Earthq. Res. Inst.*, **19** (1941).
14) Y. KATO, *Proc. Imp. Acad. Japan*, **16** (1940), 440.
15) T. NAGATA, *loc. cit.*
16) T. NAGATA, *Bull. Earthq. Res. Inst.*, **15** (1937,, 185.

system is brought in line. The deflection of the magnet was magnified solely with the aid of the usual optical lever, 120 *cm* long, the actual scale-value being 1·43 *min./mm.*

As already discussed in detail in the writer's previous paper,[17] the general approximate equation of equilibrium of the instrument is given by

$$\Delta I=\left(1+\frac{mg\,p\cos\varphi}{MH\cos\theta}\right)\frac{\Delta n}{2L}-\tan\theta\frac{\Delta H}{H}-(\alpha-\beta)\tan\theta\,\Delta t\,, \qquad (3\text{–}1)$$

where θ = deviation of direction of magnet from that of geomagnetic force,
M = magnetic moment of magnet in standard temperature,
m = total mass of magnet system,
H = initial value of total intensity in geomagnetism,
p = distance between centre of gravity of the magnet system and the line of the knife edge,
φ = angle between the vertical and the p-line,
L = length of optical lever,
α = temperature coefficient of magnetic moment of bar magnet,
β = linear expansion coefficient of the magnet system.

From this equation, we find that since both temperature coefficient and the spurious factor are proportional to $\tan\theta$, these two terms could be made negligible provided θ be taken to be almost zero in the initial state.

Actually, in the case of observations at Miyake-sima, these two terms were estimated to be

Temperature coefficient,

$$-\left(\frac{\partial I}{\partial t}\right)_0=(\alpha-\beta)\tan\theta_0\lesssim 1{\cdot}4\times10^{-5}\,Rad./degree\,,$$

Spurious factor,

$$-\left(\frac{\partial I}{\partial H}\right)_0=\tan\theta_0/H_0\lesssim 5\times10^{-7}\,Rad./\gamma\,.$$

This dip-variometer was set up in a wooden house at the north-eastern coast of Miyake-sima, and systematic observation was begun from July 17, five days after the beginning of the violent eruption of this volcanic island.

17) T. NAGATA, *Bull Earthq. Res. Inst.*, 19 (1941), 336.

During the period of this observation from July 17 to August 3, Minakami made five absolute measurements of geomagnetic dip by means of the dip-circle at a point 20 *m* south of the observing station.[18]

Table IV-3-I. Geomagnetic Dip at Miyake-sima during its active period.

Date (in G.M.T.)	Geomagnetic Dip				Mean Value of ΔI of Kakioka and Suzuki	Corrected Value of Dip at Hōdai
	Hōdai Miyake	Suzaki	Kakioka	Toyohara		
	min	min	min	min	min	min
July 16	47°	47° 02·7	49° 31·1	60° 41·3	+0·9	(48°09 3)
17	58 5	02·3	31·1	41·1	+0·7	47°57·8
18	54·9	02·3	31·2	41·0	+0·8	54·1
19	51·5	01·5	30·2	40·6	+0·2	51·3
20	47·6	00·9	29·5	40·3	−0·8	48·4
21	47·8	01·4	30·2	40·3	−0·2	48·0
22	48·0	02·2	31·3	41·2	+0·8	47·2
23	45·6	02·1	30·6	40·9	+0·4	45·2
24	43·6	01·8	30·2	—	0	43·6
25	44·0	01·7	30·1	41·0	−0 1	44·1
26	(43·9)	01·6	30·1	40·3	−0·2	(44·2)
27	47·2	01·2	29·8	39·9	−0·5	47·7
28	(46·0)	00·7	29·1	39·9	−1·1	(47·1)
29	48·8	00·4	29·0	39·7	−1·3	50·1
30	47 6	00·9	29·4	39·8	−0·9	48·5
31	48·4	01·5	30·5	40·7	0	48·4
Aug. 1	46·7	01·8	30·7	40·5	+0·3	46·3
2	44·3	01·8	30·2	40·0	0	44·3

The daily mean values of the hourly observed geomagnetic dip are given in Table 3-I and Fig. 3-1, where the time is referred to G.M.T. In Table 3-I and in Fig. 3-1, are also given the daily mean dip at the Magnetic Station, the Mitui Geophysical Institute at Suzaki (about 80 *km* north from Miyake), at the Kakioka Magnetic Observatory (about 250 *km* N-E-N), and the Toyohara Magnetic Observatory (about 1500 *km* N-E-N). As will be seen from this result, the day-to-day variation in dip at Miyake-sima was distinctly large compared with those at the other three stations, while those of the three stations were similar in their magnitudes as well as in their forms.

Assuming, then, that the mean of the daily variation at Suzaki and Kakioka gives the general variation for Central Japan and vicinity,

18) T. MINAKAMI, *Bull. Earthq. Res. Inst.*, 19 (1941), 363.

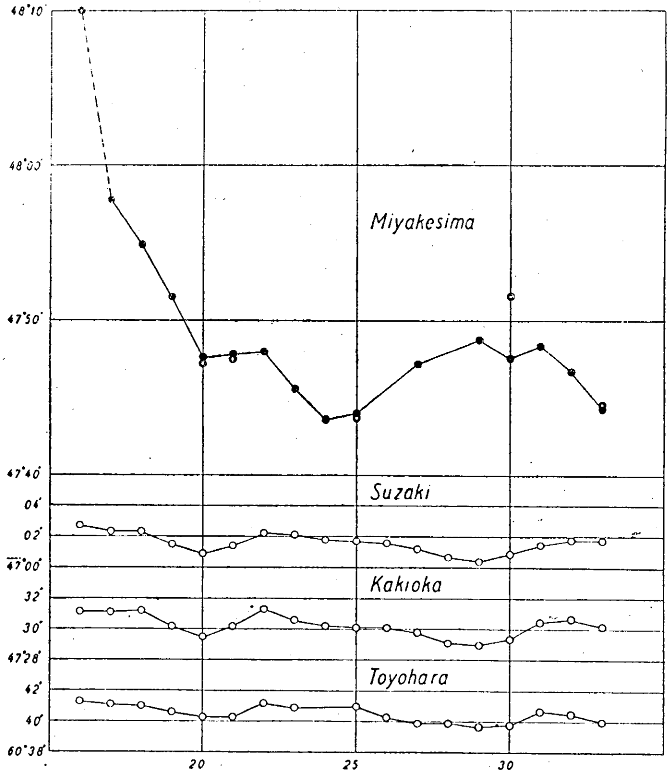

Fig. 3-1. The daily mean values of geomagnetic dip at Miyake-sima after the severe volcanic eruption.

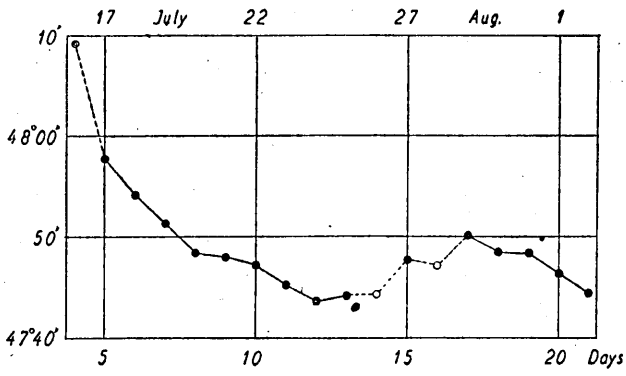

Fig. 3-2. Local anomalous change in geomagnetic dip at Miyake-sima after the severe volcanic eruption.

it is possible to estimate the amount of local anomalous change at Miyake-sima by subtracting the general variation from the observed value at our station, the actual values of local anomalous change being shown in Fig. 3–2, where it is assumed to have been zero on Aug. 2 (G.M.T.).

The error in the final corrected values is estimated to be less than 0.′5. Now, from this result, it will be seen that the geomagnetic dip at our station decreased anomalously 14.′2 during the 7 days from July 17th to 24th, whereas it increased from 25th to 29th, the increase being 6.′5, after which it again decreased, the decrease during the period from July 29th to Aug. 2nd amounting to 5.′8.

If, further, we take into consideration the value on July 16, which is not so certain compared with those on other days, although its observational error may be less than 3′, the decrease in dip during the 8 days from July 16 to July 24 amounted to as much as nearly 25′.

That the geomagnetic field changed remarkably during a period of less than ten days is worthy of notice. As already mentioned, the results obtained by other investigators during nearly the same period shows also a geomagnetic change of nearly the same magnitude as ours. For example, Minakami's intermittent measurements near our station naturally agrees well with ours, his measurement at other points in this island also showing decrease in dip of 15′∼35′ during the same period. Kato's measurements[19] covering a period of 10 years from July 1930 to July 1940, also show change in dip of like order. On the other hand, according to Takahasi and Hirano,[20] the difference in vertical intensity of the geomagnetic field at various points in this island during the period from July 20 to 24 and those from July 26 to 29 were distributed rather regularly, the largest change being a decrease of about 500 γ.

We can thus conclude, at any rate, that there was marked local anomalous change in geomagnetism in Miyake-sima immediately following the severe volcanic activity, the change amounting to about 20 minutes in dip or to a few hundred gamma in vertical intensity.

Since this anomalous change in geomagnetic field is presumed to be closely related to the volcanic activity, its cause probably lies in the change in magnetization of material underneath this volcanic island. (It will be worth while to note here that the change in earth current

19) Y. KATO, *loc. cit.*
20) R. TAKAHASI and K. HIRANO, *loc. cit.*

observed by T. Hagiwara[21] during the same period was too slight to produce the observed geomagnetic change.)

We must here take into consideration that the temperature distribution underneath the volcano ought to change accompanying severe volcanic activity. Since the temperature of the earth's surface nearly agrees with atmospheric temperature, and since that of magma in the interior of the volcano ought to be higher than 1000°C, as directly proved by the fact that the temperature of the Akabakkyō lava that flowed over the earth's surface exceeded 900°C,[22] there must exist a layer having a Curie-point temperature or a transition point of thermo-remanent magnetism of igneous rock (about 600°C) between the earth's surface and magma. If so, then the rock between the earth's surface and the isothermal surface of Curie-point temperature is in a ferro-magnetic state, with thermo-remanent magnetization as well as magnetization induced by the earth's magnetic field, while in rock below that isothermal surface, the ferro-magnetic properties vanish, being almost non-magnetic. Should volcanic activity be increasing, the temperature inside the volcano would increase owing to the ascent of magma and gases of high temperature, the isothermal surface of Curie-point temperature reaching nearer the earth's surface. In contrast to this, the interior of the volcano would gradually cool, owing to loss of heat energy through the crater and the surface of the volcano, when the supply of heat of subterranean origin will decrease or will have already stopped, resulting in the isothermal surface of Curie-point receding from the earth's surface.

From the experimental results obtained in the foregoing Chapter, it can be said that a decrease of temperature Δt of unit mass of subterranean rock at temperature t_i is accompanied with increase of magnetization

$$\Delta\vec{J} \simeq -\vec{H}\left(\frac{\partial\chi}{\partial t}\right)\Delta t + \vec{H}P(t_i)\Delta t,$$

provided Δt is not large, where, needless to say, $\frac{\partial\chi}{\partial t} = P(t) \fallingdotseq 0$, if $t \geqq \theta_s = t_c$. This process of creation of magnetization is, as already proved, reversible with respect to change in temperature, that is, increase of temperature Δt results in derease of magnetization ΔJ, whence it can be presumed that total magnetization underneath the volcano,

21) T. HAGIWARA, *Bull. Earthq. Res. Inst.* **19** (1941), 367.
T. HAGIWARA, T. MINAKAMI and T. NAGATA, *ibid.*, **19** (1941), 393.

22) T. NAGATA, *Bnll. Earthq. Res. Inst.*, **19** (1941), 295.

generally speaking, decreases with increase of volcanic activity, while the former increases with decrease of the latter.

For the purpose of estimating the change in magnetization due to fluctuation of the isothermal surface, we assume, for simplicity, the special case of change with temperature, in which thermo-remanent magnetism and susceptibility take the following approximate forms

$$\left.\begin{array}{llll} \chi(t)=0\,, & P(t)=0\,, & & t\geqq t_c=\theta_a\,, \\ \chi(t)=\chi_0\dfrac{t_c-t}{t_c-t_e}\,, & P(t)=P=\text{constant}, & & t_c\geqq t\geqq t_e\,, \\ \chi(t)=\chi_0=\text{constant}, & P(t)=0\,, & J^t_{t_c}=J_{t_c}=\text{const}, & t\leqq t_e\,, \end{array}\right\} \quad (3\text{–}2)$$

the general mode of change in $\chi(t)$ and $J^t_{t_c}$ with temperature being shown in Fig. 3–3, so that if the depth of the isothermal surface of t_c changes from D_1 to D_1+S_1 and that of t_e from D_2 to D_2+S_2, the change in magnetization is given by

$$\rho\Delta M=\frac{S_1+S_2}{2}\rho H(\chi_0+J_{t_c})\,, \quad (3\text{–}3)$$

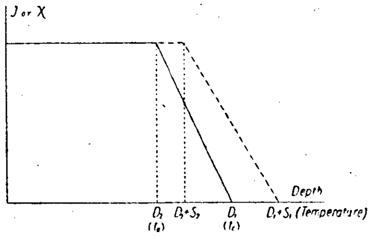

Fig. 3–3.

provided the temperature varies linearly with depth in the layer between D_1 and D_2 in the initial state and in that between D_1+S_1 and D_2+S_2 in the final state.

Since the actually observed J_{t_c} and χ_0 of the ejecta from Miyake-sima are

$$J_{t_c}\lesssim 5\times10^{-2}\,, \qquad \chi_0\lesssim 1{\cdot}5\times10^{-3}\,,$$

assuming $\rho=2{\cdot}7$, we get

$$\rho\Delta J=\frac{\rho\Delta M}{S}\lesssim 6\times10^{-2}, \qquad (3\text{–}4)$$

where

$$S=\frac{1}{2}(S_1+S_2)\,.$$

On the other hand, by assuming that the earth's surface is plane and the subterranean layer, in which magnetization changes, is sufficiently thin and parallel to the earth's surface, we get a relation between the distribution of anomalous change in dip on the earth's surface and that of change in magnetization of the layer,[23] the distribution of anomalous change in dip ΔI and that of magnetization ΔM being expressed respectively by

$$\left.\begin{aligned}\Delta I=\sum_m\sum_n\Big(&A_{mn}\cos\frac{2\pi}{L}mx\cos\frac{2\pi}{L}ny+B_{mn}\sin\frac{2\pi}{L}mx\cos\frac{2\pi}{L}ny\\&+C_{mn}\cos\frac{2\pi}{L}mx\sin\frac{2\pi}{L}ny+D_{mn}\sin\frac{2\pi}{L}mx\sin\frac{2\pi}{L}ny\Big),\\\Delta M=\sum_m\sum_n\Big(&a_{mn}\cos\frac{2\pi}{L}mx\cos\frac{2\pi}{L}ny+b_{mn}\sin\frac{2\pi}{L}mx\cos\frac{2\pi}{L}ny\\&+c_{mn}\cos\frac{2\pi}{L}mx\sin\frac{2\pi}{L}ny+d_{mn}\sin\frac{2\pi}{L}mx\sin\frac{2\pi}{L}ny\Big).\end{aligned}\right\}\qquad(3\text{–}5)$$

Where the positive direction of x is taken as agreeing with the geomagnetic north, we get the relation[23]

$$\left.\begin{aligned}\begin{matrix}a_{mn}\\c_{mn}\end{matrix}\Big\}&=\frac{\exp\left(\sqrt{m^2+n^2}\frac{2\pi}{L}d\right)}{\pi\sqrt{m^2+n^2}\,(\lambda_{mn}^2+\mu_{mn}^2)}\left(\lambda_{mn}\begin{Bmatrix}A_{mn}\\C_{mn}\end{Bmatrix}-\mu_{mn}\begin{Bmatrix}B_{mn}\\D_{mn}\end{Bmatrix}\right),\\\begin{matrix}b_{mn}\\d_{mn}\end{matrix}\Big\}&=\frac{\exp\left(\sqrt{m^2+n^2}\frac{2\pi}{L}d\right)}{\pi\sqrt{m^2+n^2}\,(\lambda_{mn}^2+\mu_{mn}^2)}\left(\mu_{mn}\begin{Bmatrix}A_{mn}\\C_{mn}\end{Bmatrix}+\lambda_{mn}\begin{Bmatrix}B_{mn}\\D_{mn}\end{Bmatrix}\right),\end{aligned}\right\}\qquad(3\text{–}6)$$

where d is the depth of the layer from the earth's surface, and

$$\left.\begin{aligned}\lambda_{mn}&=\frac{\sin 2I_0}{H_0}\left(1+\frac{m^2}{m^2+n^2}\right),\\\mu_{mn}&=\frac{\sin 2I_0}{H_0}\frac{m}{\sqrt{m^2+n^2}}(\cot I_0-\tan I_0)\,,\end{aligned}\right\}\qquad(3\text{–}7)$$

where H_0 and I_0 denote respectively the total intensity and dip of the average geomagnetic field, so that given the distribution of the anomalous change on the earth's surface, it is possible to obtain approximately the distribution of change in magnetization in the assumed subteranean layer with the aid of these relations, so long as its depth has been reasonably assumed.

In the absence, unfortunately, of sufficiently trustworthy data of the space distribution of geomagnetic changes in Miyake-sima we shall

23) T. NAGATA, *Bull. Earthq. Res. Inst.*, **16** (1938), 550; *Proc. Imp. Acad. Japan*, **14** (1938), 176

merely estimate the order of magnitude of a_{mn}, etc., that correspond to the given values of A_{mn} etc., in the probable special cases.

Putting $I=45°$ in equation (3–7), we get from equation (3–6)

$$a_{mn}=\frac{H_0\sqrt{m^2+n^2}\exp\left(\sqrt{m^2+n^2}\frac{2\pi}{L}d\right)}{\pi(2m^2+n^2)}A_{mn}, \quad \text{etc.}$$

Then, H_0 and A_{mn} being taken as 0·45 *Oe.* and $20'=0{\cdot}006$ respectively, the magnitude of a_{mn} for various values of m, n and d/L are given in Table 3–II.

Table IV–3–II.

d/L	$m=1$			$m=2$			$m=3$		
	$n=1$	$n=2$	$n=3$	$n=1$	$n=2$	$n=3$	$n=1$	$n=2$	$n=3$
0·1	9·8 $\times10^{-4}$	13·0 $\times10^{-4}$	17·0 $\times10^{-4}$	8·5 $\times10^{-4}$	11·8 $\times10^{-4}$	17·4 $\times10^{-4}$	9·7 $\times10^{-4}$	13·5 $\times10^{-4}$	19·3 $\times10^{-4}$
0·05	6·3 〃	6·4 〃	6·5 〃	4·2 〃	4·9 〃	5·6 〃	3·7 〃	4·4 〃	5·1 〃
0·01	4·4 〃	3·7 〃	3·0 〃	2·4 〃	2·4 〃	2·3 〃	1·7 〃	1·8 〃	1·8 〃

Here magnetization $a_{mn}\cos\frac{2\pi}{L}mx\cos\frac{2\pi}{L}ny$ may be interpreted as the result of uniform magnetization of the subterranean layer, $S_{mn}\cos\frac{2\pi}{L}mx\cos\frac{2\pi}{L}ny$ in thickness, with specific intensity ΔJ, where

$$\frac{a_{mn}}{\rho\Delta J}=\frac{2\pi}{L}S_{mn}. \tag{3–8}$$

Putting $\rho\Delta J=0{\cdot}06$ in this equation, we can obtain the S_{mn}/d, corresponding to various values given in Table 3–II, with the result that

$$\frac{S_{mn}}{d}=0{\cdot}02\sim0{\cdot}12\,.$$

It may, therefore, be said that if the depth of the isothermal surface that corresponds to a temperature between t_c and t_e changes to the form $S_{mn}\cos\frac{2\pi}{L}mx\cos\frac{2\pi}{L}ny$, where S_{mn} is from 2 to 12 percent of the initial value of depth d, it is possible for the geomagnetic dip on the earth's surface to show a maximum change of about 20′, so far as concerns cases where the values of m, n and d/L are restricted to those given in Table 3–II. That is to say, if $d=1\,km$, a change of 20′ in dip requires a change in the depth of the isothermal surface of t_c (about 600°C) in at least a few tens of meters, and even if $d=500\ m$, S must be $10\sim50\ m$.

If, however, we assume that the change in temperature distribution in the earth crust is caused largely by thermal conduction through the rock composing the volcano, it is scarcely possible for an isothermal surface of about 600°C to rise or fall a few tens of meters during such a short period as almost a week, seeing that the amount of change ought to be of the order of a few *cm* during a week, provided the thermal conductivity of rock has its ordinary value, say, about 4×10^{-3}.[24] Then, so long as we assume that the observed geomagnetic change is mainly due to development or disappearance of thermo-remanent magnetism in the subterranean rock, it will be presumed that the change in distribution of temperature in this volcano was sufficiently large, owing to rapid heating due to successive intrusion of magma and gases of high temperature or to rapid cooling due to gas emission and other phenomena, the apparent diffusivity amounting to at least 0·05. There is, however, the possibility that the rock inside the volcano under high pressure and stress would be intensly magnetized under the effect of the geomagnetic field and temperature, although we have at present only a few experimental data covering the effects of pressure and stress on the mode of development of thermo-remanent magnetism.

We must then accept the conclusion that the geomagnetic change observed in Miyake-sima after its great eruption was too large to be explained as that due to development or disappearance of thermo-remanent magnetism through rise of temperature of isothermal surface from where it is t_c, or its fall, which phenomena would be caused by heating or cooling of rock through thermal conduction alone. It will be noted, here, that Takahasi[25] reached the same conclusion from analysis of his own data of geomagnetic change in Miyake-sima. Since, however, within the limits of phenomena so far known with certainty from experiments and actual field observations, there is no other possible reason for the anomalous geomagnetic change except the development and disappearance of thermo-remanent magnetism and the change with temperature of magnetic susceptibility in the subterranean rock, and since, here, the latter is negligible compared with the former, the writer is of opinion that the change in thermo-remanent magnetization in the subterranean rock is the result of rather rapid heating by

24) H. REICH, *Handb. d. Geophysik, Bd.* 4, Lieferung 1, pp. 79.
25) R. TAKAHASI and K. HIRANO, *loc. cit.*

intrusion of magma and gas or rapid cooling through emission of ejecta and gas, and that it is the most plausible cause of the observed geomagnetic change, although this presumption is not without its drawbacks.

(ii) Interpretation of remarkable geomagnetic change observed near the cooling lava flow.

According to Minakami,[26] the geomagnetic dip measured at a spot 2 *m* south of the southern margin of the Yoridai-no-sawa lava flow, which flowed out in the great eruption of Miyake-sima in July 1940, changed markedly during the ten months following the ejection of lava, the observed results being shown in Fig. 3–4. The change in dip during the period from July 22, 1940 to May 16, 1941, reached the remarkable value of 1°30′, as will be seen from Fig. 3–4.[27]

Since in July 1940, immediately after the ejection of the lava flow, the temperature of the bulk of the lava flow was believed to be sufficiently high, say about 1000°C although the surface of the lava flow had already cooled to about 100°C five or six days after its ejection, it may be presumed that in July 1940, the larger part of the lava flow was almost non-magnetic.

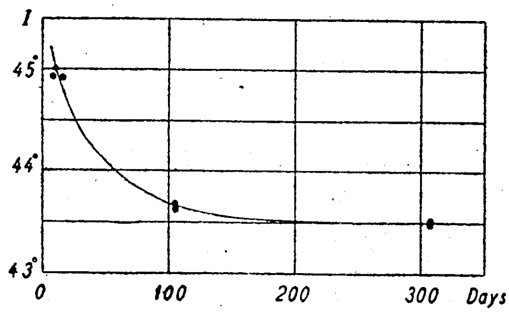

Fig. 3-4. The change in geomagnetic dip at a spot near the Yoridai-no-sawa lava flow during a year after the flowing out of that lava. (After T. Minakami)

Since, according to Minakami, the lava flow had cooled considerably in May 1941, it would be in the ferro-magnetic state, with the thermo-remanent magnetization that had developed during cooling in the earth's magnetic field, whence it will be presumed that the observed geomagnetic change was due mainly to development of magnetization in the lava flow that accompanied its cooling, because no other phenomena could possibly cause the geomagnetic change. Actually, assuming that the geomagnetic change was due to uniform magnetization of the lava flow in the direction of geomagnetic force, Minakami estimated the change in magnetization during the above mentioned 10 months, as being about 0·012 *e.m.u.* per cm^3.

26) T. MINAKAMI, *Bull. Earthq. Res. Inst.* 19 (1941), 612.

27) Quite similar phenomenon was observed at a place near the northern margin of Akabakkyo lava flow ejected with the same volcanic activity. (*Minakami, ibid.*)

On the other hand, the characteristic magnetic properties of two rock pieces near the surface of Yoridai-no-sawa lava was,

$$J_{t_c}=\begin{cases}4{\cdot}77\times10^{-2}\\3{\cdot}80 \quad ''\end{cases}, \qquad \chi_0=\begin{cases}1{\cdot}11\times10^{-3}\\1{\cdot}10 \quad ''\end{cases},$$

$$t_0=\begin{cases}525^\circ\mathrm{C}\\530\end{cases}, \qquad t_c\doteqdot\theta_a=\begin{cases}580^\circ\mathrm{C}\\580\end{cases},$$

while the density of the rock piece ρ was 2·25.

As frequently pointed out, the change in magnetization of a unit volume of rock in passing from from t_2 to t_1 is given by

$$\rho\Delta\vec{J}=\rho\vec{H}\left\{\chi(t_1)-\chi(t_2)\right\}+\rho\vec{H}\int_{t_1}^{t_2}P(t)\,dt\,,$$

so that even by assuming that $t_2>\theta_a$ and $t_1=0^\circ\mathrm{C}$, the change in induced magnetization in a magnetic field of $H=0{\cdot}45$ *Oe.*, the first term on the right-hand side of the above equation, amounts to only $1{\cdot}1\times10^{-3}$, this value being only about one-tenth that of the change in magnetization expected from the geomagnetic measurements. While, the intensity of thermo-remanent magnetism in the same conditions of t_2 and t_1 amounts to $4{\cdot}3\times10^{-2}$, this value of thermo-remanent magnetization much exceeds the expected value of, say, $1{\cdot}2\times10^{-2}$.

However, we must not forget to take into consideration (1) that the part near the surface of the lava flow had already cooled and magnetized at the time of the first measurement in July 1940, (2) that the interior near the centre of the lava flow could not yet have sufficiently cooled in May 1941, as is to be expected from the very slow rate at which a lava flow cools, owing to the smallness of its thermal conductivity, and lastly (3) that the mean bulk density of lava flow is less than the density of an individual rock piece. It is then rather natural that the observed geomagnetic change corresponds to a part of the saturated thermo-remanent magnetization of the whole lava flow in the earth's magnetic field. It should, however, be noted that the intensity of thermo-remanent magnetization of the interior of a lava flow does not necessarily agree with that of a rock piece near the surface,[28] which was used in our experiment, since the former cooled comparatively slowly, while the latter was cooled rapidly or rather quenched. Since intense thermo-remanent magnetization is generally the property of rapidly cooled basaltic rock, in which the crystallization of ferro-magnetic minerals is insufficient, while the

28) See, R. CHEVALLIER, *Ann. Physique*, 4 (1925), 5.

slowly cooled basalt and andesite, in which they are fairly well crystallized, have comparatively weak thermo-remanence, we may expect the intensity of thermo-remanent magnetization of the interior of the lava flow to be less than that of its exterior.

In order to discuss more quantitatively the physical mechanism of geomagnetic change due to cooling of the lava flow, more systematic data of both geomagnetic measurement and the magnetic properties of a large number of rock specimens are needed.

Here we can conclude that the anomalous change in geomagnetism observed about the Yoridai-no-sawa lava flow was, no doubt, chiefly due to the development of thermo-remanent magnetization that accompanied the cooling of the lava in the earth's magnetic field.

(iii) Anomalous geomagnetic changes observed at Volcano Asama.

During the summer of 1937 and 1938, continuous recording of the geomagnetic dip was made near the Asama Volcano Observatory by means of a magnetograph.[29]

The instrument for measurement was also that dip variometer constructed by the writer,[30] with a magnifying device based on the bifilar suspension system. Expressing the general equation of equilibrium of this magnetograph by

$$\Delta I = S\Delta n + K\Delta H + T\Delta t ,$$

where ΔI, Δn, ΔH and Δt denote respectively the deviations in geomagnetic dip, in scale reading, in total intensity of geomagnetic force, and in temperature from their standard values.

We get for the actual magnitude of S, K and T

Scale value,	$S = 0{\cdot}232\ min./mm$,
Spurious factor,	$K = 0.001\ min./\gamma$,
Temperature coefficient,	$T = 0{\cdot}052\ min./degree$.

This dip variometer was set in a wooden cottage in front of Asama Volcano Observatory, about 4 km east of the crater of the volcano. The daily mean values referred to the Greenwich mean time of dip at the station during the period from July to October, in 1937, and during the same period in 1938, are given in Table 3–III and in Figs. 3–5 and 3–6. In the same table and figures, the daily mean dip at the Kakioka Magnetic Observatory (Lat. $= 36°14'$ N, Long. $= 140°11'$ E)

29) T. NAGATA, *Zisin*, 10 (1938), 221.
30) T NAGATA, *Bull. Earthq. Res. Inst.*, 15 (1937), 185.

Table IV-3-III (a). Day-to-day variation in geomagnetic dip in 1937.

July 1937

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′	′	′	′	′
1	49° 37·9	60° 42·4	49° 32·2	0° 5·7	11° 10·2
2	37·4	42·0	31·9	5·5	10·1
3	36·4	42·2	31·6	4·8	10·6
4	35·7	41·5	30·4	5·3	11·1
5	35·0	40·5	30·0	5·0	10·5
6	36·3	41·8	31·3	5·0	10·5
7	37·0	42·2	31·7	5·3	10·5
8	—	41·4	30·7	—	10·7
9	—	41·2	30·3	—	10·9
10	37·1	42·7	32·0	5·1	10·7
11	35·9	41·8	30·6	5·3	11·2
12	36·1	41·6	30·7	5·4	10·9
13	36·1	41·5	30·6	5·5	10·9
14	37·9	43·3	32·5	5·4	10·8
15	37·3	42·7	31·9	5·4	10·8
16	36·5	42·1	31·3	5·2	10·8
17	36·0	41·3	30·5	5·5	10·8
18	35·9	—	29·8	6·1	—
19	35·7	40·7	29·5	6·2	11·2
20	36·3	41·3	30·4	5·9	10·9
21	36·4	41·5	30·9	5·5	10·6
22	39·7	44·3	34·2	5·5	10·1
23	38·4	43·3	33·0	5·4	10·3
24	39·2	44·3	33·9	5·3	10·4
25	38·3	43·1	32·9	5·4	10·2
26	37·3	43·0	32·1	5·2	10·9
27	37·6	—	31·4	6·2	—
28	37·1	41·7	30·8	6·3	10·9
29	36·2	41·8	31·0	5·2	10·8
30	—	—	31·2	—	—
31	—	41·0	30·7	—	10·3

Aug. 1937

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′		′	′	
1	49° 34·7	—	49° 29·9	0° 4·8	—
2	34·2	—	29·2	5·0	—

(to be continued.)

Aug. 1937 Table III-3-III (a). (*continued*).

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′	′	′	′	′
3	49° 38·1	60° 44·0	49° 32·6	0° 5·5	11° 11·4
4	39·3	44·0	34·1	5·2	9·9
5	36·0	43·1	33·1	2·9	10·0
6	35·6	—	33·7	1·9	—
7	34·3	—	32·8	1·5	—
8	34·2	42·8	31·7	2·5	11·1
9	35·1	42·6	31·0	4·1	11·6
10	34·6	42·4	30·7	3·9	11·7
11	34·0	41·7	29·8	4·2	11·9
12	32·9	41·2	28·9	4·0	12·3
13	33·5	41·0	29·0	4·5	12·0
14	33·4	—	29·1	4·3	—
15	33·5	41·7	30·0	3·5	11·7
16	32·8	41·4	29·8	3·0	11·6
17	32·7	41·6	29·8	2·9	11·8
18	32·2	40·8	28·9	3·3	11·9
19	33·0	41·0	29·1	3·9	11·9
20	32·7	40·8	28·9	3·8	11·9
21	33·1	41·3	29·8	3·3	11·5
22	38·5	46·6	35·6	2·9	11·0
23	37·1	44·6	33·6	3·5	11·0
24	35·9	43·7	32·3	3·6	11·4
25	34·6	42·7	31·1	3·5	11·6
26	33·0	41·5	29·7	3·3	11·8
27	33·8	41·9	30·5	3·3	11·4
28	35·1	42·8	31·7	3·4	11·1
29	34·9	—	31·4	3·5	—
30	34·6	41·8	31·1	3·5	10·7
31	33·3	41·4	29·8	3·5	11·6

Sept. 1937

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′	′	′	′	′
1	49° 33·0	60° 41·4	49° 29·4	0° 3·6	11° 12·0
2	33·5	41·8	29·8	3·7	12·0
3	33·6	41·7	30·2	3·4	11·5
4	33·3	42·0	30·2	3·1	11·8
5	34·6	42·3	30·5	4·1	11·8

(*to be continued.*)

Sept. 1937 Table III–3–III (a). (*continued*).

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
6	49° 34·9	—	49° 30·2	0° 4·7	—
7	35·9	60° 42·1	31·0	4·9	11° 11·1
8	34·7	—	30·3	4·4	—
9	33·6	41·2	30·3	3·3	10·9
10	35·8	41·2	30·3	5·5	10·9
11	—	44·5	33·8	—	10·7
12	—	42·9	32·0	—	10·9
13	—	42·5	31·8	—	10·7
14	—	42·1	31·6	—	10·5
15	—	42·4	31·6	—	10·8
16	—	42·2	31·3	—	10·9
17	34·7	42·4	30·8	3·9	11·6
18	33·6	42·1	30·3	3·3	11·8
19	34·3	—	31·0	3·3	—
20	33·7	41·9	30·6	3·1	11·3
21	—	41·6	31·2	—	10·4
22	34·5	41·2	31·4	3·1	9·8
23	34·9	42·0	32·4	2·5	9·6
24	—	43·2	33·9	—	9·3
25	—	41·8	32·1	—	9·7
26	—	41·7	31·9	—	9·8
27	35·5	41·8	32·2	3·3	9·6
28	35·9	42·0	32·7	3·2	9·3
29	—	41·8	31·9	—	9·9
30	35·9	41·9	31·9	4·0	10·0

Oct. 1937

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
1	49° 39·2	60° 45·0	49° 35·9	0° 3·3	11° 9·1
2	38·0	43·8	34·1	3·9	9·7
3	37·6	43·7	33·6	4·0	10·1
4	42·2	47·8	38·3	3·9	9·5
5	39·1	44·9	35·2	3·9	9·7
6	—	43·4	33·8	—	9·6
7	—	43·2	33·6	—	9·6
8	40·6	44·7	36·4	4·2	8·3
9	40·7	44·3	36·5	4·2	7·8
10	39·9	45·2	36·3	3·6	8·9

(*to be continued.*)

Table IV-3-III (b). Day-to-day variation in geomagnetic dip in 1938.

July 1938

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
1	—	60° 41′·6	49° 30′·8	—	11° 10′·8
2	49° 34′·5	41·3	30·5	0° 4′·0	10·8
3	32·7	—	29·6	3·1	—
4	33·7	41·7	30·7	3·0	11·0
5	—	41·1	29·9	—	11·2
6	—	41·7	—	—	—
7	—	40·9	—	—	—
8	31·0	41·2	30·3	0·7	10·9
9	30·2	—	29·6	0·6	—
10	—	41·0	30·0	—	11·0
11	—	40·8	30·1	—	10·7
12	—	40·4	29·3	—	11·1
13	—	40·1	29·1	—	11·0
14	—	41·3	31·1	—	11·2
15	28·0	42·8	32·3	−4·3	10·5
16	29·8	43·3	33·8	−4·0	9·5
17	30·7	42·1	32·7	−2·0	9·4
18	31·7	41·7	32·9	−1·2	8·8
19	33·2	41·6	32·9	0·3	8·7
20	33·5	40·9	32·9	0·6	8·0
21	34·5	40·8	32·5	2·0	8·3
22	—	40·9	32·3	—	8·6
23	—	40·8	32·4	—	8·4
24	—	40·5	32·2	—	8·3
25	—	40·6	31·8	—	8·8
26	—	40·7	32·2	—	8·5
27	36·4	40·2	31·7	4·7	8·5
28	36·4	40·1	31·5	4·9	8·6
29	37·0	40·1	31·9	5·1	8·2
30	41·1	43·3	36·0	5·1	7·3
31	40·7	42·4	34·8	5·9	7·6

Aug. 1938

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
1	49° 39′·8	60° 41′·6	49° 33′·7	0° 6′·1	11° 7′·9
2	39·2	42·8	34·1	5·1	8·7

(to be continued.)

Aug. 1938 Table IV-3-III (b). (*continued*).

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′	′	′	′	′
3	49° 36·7	60° 41·6	49° 32·7	0° 4·0	11° 7·9
4	42·1	43·5	34·7	7·4	8·8
5	—	42·7	34·0	—	8·7
6	—	42·1	33·3	—	8·8
7	41·1	—	32·7	8·4	—
8	39·3	41·2	31·5	7·8	9·7
9	38·6	41·1	30·8	7·8	10·3
10	38·4	40·9	30·1	8·3	10·8
11	40·7	43·8	33·5	7·2	10·3
12	40·8	42·7	32·7	8·1	10·0
13	39·6	—	31·6	8·0	—
14	38·8	41·7	30·8	8·0	10·9
15	39·2	41·7	31·2	8·0	10·5
16	38·9	41·4	30·8	8·1	10·6
17	39·2	41·5	30·6	8·6	10·9
18	39·3	41·1	30·8	8·5	10·3
19	39·2	41·2	30·8	8·4	10·4
20	39·5	40·6	30·6	8·9	10·0
21	39·8	40·7	30·8	9·0	10·1
22	39·3	40·6	31·1	8·2	9·5
23	40·0	41·7	32·2	7·8	9·5
24	40·3	41·5	32·2	8·1	9·3
25	41·0	—	32·5	8·5	—
26	40·8	41·4	32·0	8·8	9·4
27	40·2	40·8	31·2	9·0	9·4
28	39·1	—	30·4	8·7	—
29	39·4	40·3	30·9	8·5	9·4
30	40·1	41·2	31·8	8·3	9·4
31	40·4	41·4	32·0	8·4	9·4

Sept. 1938

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′	′	′	′	′
1	49° 40·3	60° 41·7	49° 31·5	0° 8·8	11° 10·2
2	39·1	40·7	30·9	8·2	9·8
3	40·4	41·5	32·4	8·0	9·1
4	37·9	40·9	32·0	5·9	8·9
5	38·6	41·2	32·2	6·4	9·0

(*to be continued.*)

Sept. 1938 Table IV-3-III (b). (*continued*).

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′		′	′	
6	49° 38·5	—	49° 31·3	0° 7·2	—
7	37·6	—	30·7	6·9	—
8	—	—	31·7	—	—
9	40·6	—	30·8	9·8	—
10	40·4	—	31·4	9·0	—
11	—	60° 41′·3	32·7	—	11° 8′·6
12	41·4	41·4	33·4	8·0	8·0
13	41·8	41·2	32·8	9·0	8·4
14	43·1	43·3	34·9	8·2	8·4
15	44·7	47·8	37·6	7·1	10·2
16	43·3	43·8	35·0	8·3	8·8
17	41·6	43·2	34·3	7·3	8·9
18	41·8	42·7	33·3	8·5	9·4
19	39·2	42·2	32·3	6·9	9·9
20	38·2	41·6	31·6	6·6	10·0
21	37·9	41·6	31·1	6·8	10·5
22	38·5	—	31·5	7·0	—
23	39·2	41·6	30·3	8·9	11·3
24	28·1	—	29·3	8·8	—
25	35·7	41·5	27·7	8·0	13·8
26	37·7	41·3	30·5	7·2	10·8
27	37·5	43·5	31·1	6·4	12·4
28	40·0	45·8	34·0	6·0	11·8
29	39·5	44·0	31·8	7·7	12·2
30	38·4	43·0	30·5	7·9	12·5

Oct. 1938

Date	Asama θ_A	Toyohara θ_T	Kakioka θ_K	$\theta_A-\theta_K$	$\theta_T-\theta_K$
	′	′	′	′	′
1	49° 41·0	60° 45·0	49° 33·0	0° 8·0	11° 12·0
2	39·5	43·7	31·8	7·7	11·9
3	40·0	42·6	31·3	8·7	11·3
4	40·5	42·9	32·0	8·5	10·9
5	38·8	42·2	31·3	7·5	10·8
6	38·4	41·8	30·4	8·0	11·4
7	42·2	45·8	34·2	8·0	11·6
8	—	45·5	35·2	—	10·3
9	—	43·3	33·0	—	10·3
10	29·1	43·1	31·9	7·2	11·2

and those at the Toyohara Magnetic Observatory (Lat.=46°58′ N, Long.=142°45′ E) are also shown.[31]

Comparing the values of the dip at these three stations, we notice that in the summer of 1937, the calm period of Volcano Asama, the day-to-day variation in dip at both Asama and Toyohara generally ran

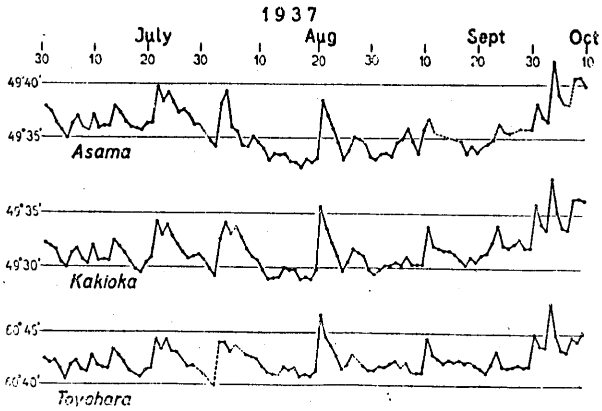

Fig. 3-5. The day-to-day variation in geomagnetic dip at Asama, Kakioka, and Toyohara during the summer of 1937.

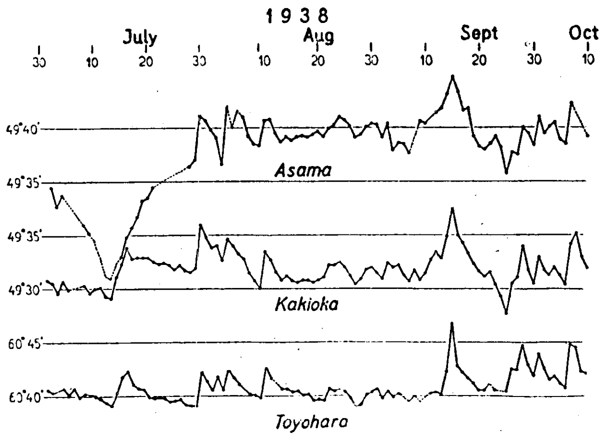

Fig. 3-6. The day-to-day variation in geomagnetic dip at Asama, Kakioka, and Toyohara during the summer of 1938.

parallel with that at Kakioka, while during the same season of 1938, the active period of this volcano, a fairly marked change of 13 minutes in dip occurred, preceding the onset of the group of frequent violent

31) *Kisyô-Yôran*, 1937 and 1938.

explosions of Mt. Asama, although during the same period the day-to-day variation in dip at Toyohara was also parallel to that at Kakioka. In order to bring out these results more clearly, the differences between the values of the dip at Asama and Kakioka ($\theta_A - \theta_K$) and between those at Toyohara and Kakioka ($\theta_T - \theta_K$) are plotted in Fig. 3–7, from

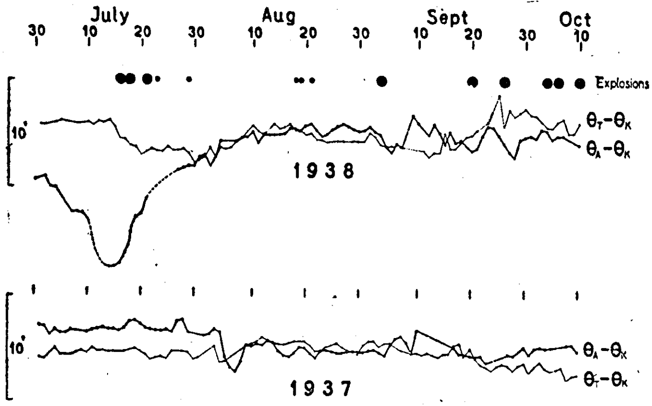

Fig. 3–7. The day-to-day change in the differences between geomagnetic dip at Asama and that at Kakioka, and between dip at Toyohara and that at Kakioka during summer of 1937 and 1938.

which fluctuations in magnitude are seen, amounting to several minutes in the difference between the daily mean magnetic dip at widely separated stations. These fluctuation may be due mainly to inequalities in the amplitudes of magnetic variations depending on the latitude of the stations, and in less degree to observational errors. In these circumstances, therefore, it may not be possible to discuss the correlation between the volcanic activities of Mt. Asama and the day-to-day variation in dip less than 5 minutes in its amplitude. However, the value of $\theta_A - \theta_K$ changed in magnitudes exceeding 10 minutes immediately preceding the successive violent explosions of Volcano Asama in the summer of 1938, that is to say, from July 3, $\theta_A - \theta_K$ became smaller and smaller until on July 14 it reached minimum, 8 minutes less than the initial value, after which it again increased in the middle of August. On the other hand, the $\theta_T - \theta_K$ values for the same period showed no marked anomalous change, the difference between the maximum and minimum values in this period amounting to only 3·5 minutes.

In contrast to this, in the summer of 1937, through which period Volcano Asama was comparatively quiet, there was no change larger than 5 minutes in either the $\theta_A - \theta_K$ or the $\theta_T - \theta_K$ values, whence it may be reasonable to conclude that the anomalous variation in dip observed at Asama, in July 1938, was a local magnetic disturbance connected with its volcanic activity. Volcano Asama had been quiet for about nine months since the last minor eruption of June 29, 1937, up to the violent eruption on March 25, 1938. In April and May 1938, the volcano was very active, erupting violently several times, while in July it was comparatively calm, erupting only once on the 15th. Its activity during the period that followed, namely, from July to October, is shown in Fig. 3–7.[32] Comparing the change in volcanic activity with that of $\theta_A-\theta_K$, we find that the latter gradually began to decrease from about fifteen days preceding the first violent explosion of July 16, its minimum having been reached on July 14, and then began to increase. It may be concluded from this result that the observed geomagnetic change was connected with a group of frequent explosions, that is, the active state of the volcano continued for a few months. As to the origin of this observed geomagnetic change, the development and disappearance of thermo-remanent magnetization in the rock of the volcano could also be the cause.

Just as in the case of Volcano Miyake-sima, this explanation is not without its drawbacks, since the characteristic magnetic properties of a few rock samples ejected from Volcano Asama are the same as those given in Table 3–IV.

Table IV–3–IV.

No. of Specimen	χ_0	J_{tc}	t_0	$P(t_0)$
8	$1{\cdot}18\times10^{-3}$	$2{\cdot}29\times10^{-3}$	535°C	$2{\cdot}2\times10^{-5}$
16	0·61 〃	2·89 〃	{520 {220	{ 1·1 〃 { 1·8 〃
73	0·50 〃	12·47 〃	{520 {200	{ 0·9 〃 {12·4 〃

(IV) The result of the geomagnetic variation on Volcano Mihara.

Although for two weeks following the minor eruption of Volcano Mihara, in Oosima Island, on Aug. 19, 1940, geomagnetic observations with the aid of the dip-variometer were made at the north-western summit of the somma, no marked geomagnetic change could be observed, although there was a fluctuation of about 5 minutes in the difference between the dip values at Mihara and Suzaki.[33]

From September 1941, geomagnetic observations by means of H

32) *Kisyō-Yōran*, 1937 and 1938.

33) T. NAGATA, *Bull. Earthq. Res. Inst.*, 19 (1941), 402.

and D variometers have been continued at the geomagnetic station where the dip was observed. During the period from Sept. 1941 to June 1942, Volcano Mihara had been quiet, and no marked local geomagnetic change had been observed.

It will be seen from the foregoing examples, that marked local geomagnetic changes sometimes occurred in the neighourhood of that locality where it is certain that the temperature of the rock changes, our experiments showing that change in temperature of rock is usually accompanied with change in thermo-remanent magnetization as well as with that in magnetic susceptibility, provided the temperature is lower than the apparent Curie-point.

In the case of anomalous geomagnetic change near the cooling lava flow, the development of thermo-remanent magnetism in the lava flow with its cooling seems to be the cause of the observed geomagnetic change, although in the case of changes that occur simultaneously with severe volcanic activity, the magnitude of the observed change seems greatly to exceed those due to probable changes in thermo-remanent magnetization, estimated on the basis of a few assumptions. In order to explain the observed geomagnetic change as the result of change in thermo-remanent magnetization in the subterranean rock, we must assume either that the change in subterranean temperature is much larger than that owing to thermal conduction through the rock, or that the intensity of magnetization of subterranean rock is greatly in excess of that experimentally measured under the ordinary atmospheric pressure that obtains in our laboratory. It seems that this is a phenomenon that needs to be carefully studied in the future by as many investigators as possible.

Summary and Conclusion.

In Chapter I, the general magnetic properties of a variety of volcanic rocks, the petrological and chemical constitutions of which are already known, were examined. The magnetic susceptibility of volcanic rocks is approximately proportional to the normative amount of magnetite contained in them, while it disappears at a certain temperature below 600°C. Further, the magnetization curves of these rock samples clearly show hysteresis phenomenon. From all these experimental facts, it can be concluded that volcanic rock is, in its magnetic behaviour, a kind of ferro-magnetic substance. Taking into consideration that the temperature at which the susceptibility of volcanic rocks almost perfectly disappears is about 600°C, and that the susceptibility is nearly proportional to the norm of magnetite, we can presume that the ferro-magnetic character of volcanic rocks is largely due to magnetite or its solid solution with other elements that are scattered as phenocrysts or micro-crystals in the groundmass among various minerals that are almost non-magnetic. Strictly speaking, however, susceptibility χ is not exactly proportional to the norm of magnetite, and the Curie-point is not always uniquely given at 580°C$\sim$600°C, some rock specimens having a second Curie-point at temperatures of 200°C$\sim$300°C. These facts will show that the constitution of ferro-magnetic minerals in volcanic rocks is very complex. A comparison of the chemical composition of some rocks with their mode of thermal change in susceptibility leads to the conclusion that not only magnetite, but TiO_2, or ilmenite, also affects the magnetic property of volcanic rocks to a marked degree, probably through the effect of solid solution between magnetite and ilmenite. Moreover, the fact that the hysteresis tendency in the magnetization curves of various rock samples differ the one from the other may suggest that the amount of internal stress and its distribution in ferro-magnetic minerals varies with the chemical and petrological constitutions of rocks and with the physical conditions during their crystalization from magma of high temperature.

Nevertheless, we can substitute the magnetic properties of volcanic rocks, at least in a first approximation, as a simple model of a dust core specimen of rare content, constituted of magnetite and its solid solutions with other elements, whence it must be pointed out that, although the demagnetizing factor of rock pieces is generally small, owing to its apparent low susceptibility (in the order of $10^{-4}\sim10^{-3}$ *e.m.u.*), that of ferro-magnetic minerals themselves is fairly high since the order of their susceptibility is 0·1$\sim$1 (actually the demagnetizing factor of ferro-magnetic minerals in the rocks was estimated as 3~4).

In the first half of Chapter II, we discussed the method of determining the intensity and direction of remanent magnetization, which the volcanic rocks generally retain in their natural condition in the field. Since the apparatus for measuring the remanent magnetization of a rock sample is of arbitrary form, an astatic magnetometer was adopted. In this apparatus, letting the spherical coordinate (r, θ, ϕ) be fixed to the sample examined, where the origin of the coordinate is situated at the apparent centre of the sample, the radial force at $r=$constant, due to the remanent magnetization of the rock, was measured at points every 20 or 30 degrees with respect to θ and ϕ. The observed values being then subjected to spherical harmonic analysis, the intensity and direction corresponding to the first degree terms were adopted as those of the rock specimen, since we found that the higher degree terms were largely due to the shape of the examined rock, this fact showing that a piece of rock is almost uniformly magnetized. Estimating here the various possible errors, we found that the direction and intensity of remanent magnetization of a rock sample of arbitrary shape could be determined respectively with an error of less than one degree and only few percent.

Toward the end of Chapter II will be found the observed values of intensity and direction of natural remanent magnetization of various rocks ejected from Japanese volcanoes. Here, the natural remanent magnetization of rocks is defined as the residual permanent magnetization of a rock, whose direction is almost constant at every part of a fairly large mass of rock, excluding, of course, the residual magnetization due to lightning and artificial magnetization.

As to the direction of natural remanent magnetization of volcanic rocks, it was concluded that its direction in newly ejected lavas always agrees with that of the geomagnetic field at the place where the rocks were ejected, and that both declination and dip of natural remanent magnetization of rocks ejected in historical and remote geological times deviate a few degrees from those of the present geomagnetic field. As to its intensity, an interesting fact discovered was that the intensity of natural remanent magnetization of volcanic rocks is generally larger than that of their magnetization induced by the present geomagnetic force, say the product of their susceptibility and the total intensity of the geomagnetic field. The fact that the former magnetization frequently exceeds 20 times the latter seems to suggest that some peculiar process of developing the remanent magnetization exists in volcanic rocks.

In Chapter III, it was shown that the natural remanet magnetization of volcanic rocks can be reproduced in the laboratory by cooling the rock from a sufficiently high temperature in a geomagnetic field. Here, the residual permanent magnetization, which remains after the test rock sample is cooled in a weak magnetic field, is called thermo-remanent magnetization. The direction, then, of the thermo-remanent magnetization of volcanic rocks agrees with that of the magnetic field applied to them during the cooling, while its intensity is almost proportional to that of the magnetic field H, provided H is small and the range of temperature through which the rock sample is cooled is kept constant. On the other hand, so long as H is kept constant, the temperature t, from which a rock is cooled to 0°C in H, is higher, the larger the thermo-remanent magnetization. If, however, t should exceed a certain critical temperature t_c, the thermo-remanent magnetization in H will remain almost constant. Actually the t_c of the rock sample is less than 600°C. The intensity of thermo-remanent magnetization of any newly ejected rock, which develops during cooling from a temperature higher than t_c to ordinary atmospheric temperature in a magnetic field same as the geomagnetic field, nearly agreed with that of its natural remanent magnetization. The development of thermo-remanent magnetization seems most marked during cooling in a magnetic field through a temperature range of 100° degrees or so around a certain temperature t_0 peculiar to that rock specimen, where t_0 is called the transition temperature of thermo-remanent magnetism. Comparing, then, the mode of change in magnetic susceptibility due to temperature with that of the development of thermo-remanent magnetism, we found that every transition temperature of a rock corresponded with its apparent Curie-point, the former being always a few tens of degrees less than the latter. In other words, if a rock sample has two apparent Curie-points, the susceptibility changing stepwise with temperature, there are two transition temperatures in the mode of development of thermo-remanent magnetism, it also developing stepwise with decrease of temperature. It is therefore presumed that the thermo-remanent magnetization of a rock is due to the irreversible magnetization caused at temperatures just below its Curie-point. However, the fact that the intensity of thermo-remanent magnetization caused during cooling in H greatly exceeds that of induced magnetization $H\chi(t)$ at any temperature during the cooling process (the former amounting sometimes to a few tens times the latter), seems to need some explanation.

Let $J_{t_i,H}^{t_{i-1}}$ denote the thermo-remanent magnetization caused during the experiment in which a rock specimen is cooled from a sufficiently high temperature (above 600°C) to t_i in $H=0$ space, from t_i to t_{i-1} in $H=H$, and from t_{i-1} to 0°C again in $H=0$.

Then, from experiment we have the relation

$$J_{t,H}^{t'}=\sum_{t_{i-1}=t'}^{t_i=t} J_{t_i,H}^{t_{i-1}},$$

which always approximately holds in any example, so long as in the experimental results, $\Delta t=t_i-t_{i-1}$ exceeds 25 degrees. Further it was established that $J_{t_i,H}^{t_{i-1}}$ is, in first approximation, independent of the cooling velocity during the process of cooling from t_i to t_{i-1}.

From these results, we assumed an ideal case of development of thermo-remanent magnetism where $\lim\limits_{t_i\to t_{i-1}} \dfrac{J_{t_i,H}^{t_{i-1}}}{t_i-t_{i-1}}$ exists and is expressed by $P(H,t)$. It was finally established that, provided H is small, an ideal phenomenological aspect of development of thermo-remanent magnetism is expressed by

$$\vec{J}_{t,H}^{t'}=\vec{H}\cdot\int_{t'}^{t} P(\theta)\,d\theta\,,$$

where $P(t)$ is a function of temperature, the from of which is peculiar to each rock sample, taking the maximum value at t_0. This result leads to the conclusion that when the temperature of rock is cooled to $t-\Delta t$, where $p(\theta,0)=P(\theta)$, a part of the induced magnetization $H\chi(t)$ at any temperature t remains as residual permanent magnetization $Hp(t,t-\Delta t)\Delta t$; that is to say, $H\chi(t)$ does not correspond to the induced magnetization of a unique material, but probably does so to the various states of material or to those of a group of materials whose nature changes with temperature. Hence, it is possible for the intensity of thermo-remanent magnetization to exceed by far the amount of $H\chi(t)$, provided $P(t)/\chi(t)$ is not very small. Next, the results of other experiments show that the thermo-remanent magnetization $J_{t_i,H}^{t_{i-1}}$, caused during cooling from t_i to t_{i-1} in H, always disappears during heating through the same temperature range, regardless of the intensity and direction of the magnetic field that is applied during the heating process.

The last part of Chapter III dealt with the modes of development and disappearance of thermo-remanent magnetization, directly observed at any temperature during its development in cooling and disappearance in heating.

From these results, we found that the intensity of thermo-remanent magnetization at temperature t, $J_{t_2, H}^{t_1}(t)$, caused during cooling from t_2 to t_1 in H, is approximately given by

$$\vec{J}_{t_2, H}^{t_1}(t) = \vec{H} \cdot \int_{t_1}^{t_2} p(\theta, t)\, d\theta\,, \qquad (t < t_1 < t_2)\,,$$

$$p(\theta, t) = I(\theta, t)\, \Phi(s, \theta)\, g(\theta)\,,$$

$$\frac{\partial}{\partial t} I(\theta, t) < 0\,,$$

where $I(\theta, t)$, $\Phi(s, \theta)$ and $g(\theta)$ denote respectively the intensity of saturation magnetization at t of the ferro-magnetic mineral that has its Curie-point at θ, the probability of irreversible magnetization of the elemental domains in the ferro-magnetic minerals having Curie-point θ and internal stress s (at temperature just below the Curie-point), and the density of distribution of ferro-magnetic minerals with respect to the Curie-point temperatures. Although it seems that a little time-effect always accompanies the causation of thermo-remanent magnetization, so long as H is small, the above-mentioned hypothesis harmonizes well with all the observed facts. The physical conditions for determining the amounts of functions $\Phi(s, \theta)$ and $g(\theta)$ is not clear at present. Since, however, a single crystal of magnetite has no thermo-remanent magnetization, it will be clear that $\Phi(s, \theta)$ is due to the particular character of solid solution of magnetite with other elements. Several possible mechanisms for causing $\Phi(s, \theta)$ were discussed, for example, the exsolution of other elements (FeO, Fe_2O_3, TiO_2, etc.) from their solid solution with magnetite, and the phenomenon analogous to superlattice in metalic alloys.

In Chapter IV, a few examples of the application of natural remanent magnetization of volcanic rocks to geophysical phenomena were dealt with. Of these, the first is the fact that the direction of natural remanent magnetization agrees with that of the magnetic field applied during cooling of the rock is related to the direction of the geomagnetic field in the past by a few investigators, at the spot whence the rock samples were obtained. Tests were made to determine the past geomagnetic direction from that of remanent permanent magnetization of volcanic rocks and baked earths especially with fruitful results in the case of baked earth. In the present study, however, the assumption that effusive rock has a natural remanent magnetization having the same direction as that of the geomagnetic force that affected the rock during its cooling, was ensured from magnetic examinations of a number of rocks recently ejected from a few active Japanese volcanoes.

From this point of view, the direction of geomagnetic force in remote times in central Japan was presumed, where the data of declination in recent historical times obtained from the magnetization of rocks seem to harmonize with the data given in ancient records of geomagnetic declination in Japan. Further, from the direction of natural remanent magnetization of rocks ejected in historical and geological times, it was concluded that the secular variation in geomagnetism during the Quarternary Age (from its beginning to the present) seems to be restricted to a few tens of degrees in declination as well as in dip,—a result agreeing well with the conclusion obtained by the writer, K. Akasi, and T. Rikitake from the direction of natural remanent magnetization of sedimentary deposit (sand-stone), probably of middle Pleistocene. The fact that the intensity of natural remanent magnetization of volcanic rocks ejected probably in the Pleistocene Age is fairly smaller than that of their thermo-remanent magnetization developed in a magnetic field of 0·45 *Oe.* shows that the discrepancy is due to the effect of demagnetization during a very long period, or else that the intensity of the geomagnetic field during the Pleistocene Age was considerably less than that at present, although the latter contingency seems scarcely possible.

The second is the relation between local anomaly in the geomagnetic field and the natural remanent magnetization of rocks, which were discussed with a few actual examples. The anomalous distributions of the geomagnetic declination around the craters of Volcanoes Mihara and Asama, and around the central cone of Volcano Mihara were interpreted as that, although the magnetization induced in the rock by the present geomagnetic force contributes to a part of the observed geomagnetic anomalies, they are largely due to the natural remanent magnetization of the rocks that compose the respective volcanoes, where the direction of remanent magnetization nearly agrees with that of the present geomagnetic force.

Comparing the amount of the anomalous geomagnetic field in the neighbourhood of Volcanoes Asama, Miyakesima, and others, as observed by various investigators, with the magnetic characters of the rocks composing the respective volcanoes, we found that the outstanding cause responsible for this marked geomagnetic anomaly in volcanic regions is, in general cases, the natural remanent magnetization of effusive rocks.

Last, we discussed the possible relation between the anomalous local change in geomagnetic field accompanying volcanic activities and the development or disappearance of thermo-remanent magnetization of rock with changes in its temperature.

In the case of an anomalous geomagnetic change observed at a place near a cooling lava flow in Miyakesima, it was found that the development of thermo-remanent magnetism in the lava flow with its cooling sufficiently explained the observed anomalous change, where the increase in induced magnetization due to increase in susceptibility with decrease in temperature contributed but slightly to the anomalous change. The marked change in geomagnetic field following the severe activity of Volcano Miyakesima seemed to be too high in intensity to be interpreted as a result of development or disappearance of thermo-remanent magnetization in the rock underneath this volcano, accompanying the change in temperature due to the severe activity, where the extent of the latter phenomenon was estimated under a few reliable assumptions based on the observed magnetic properties of rocks. The relation between the anomalous change in geomagnetic field accompanying the activity of Volcano Asama and the character of the thermo-remanent magnetism of the Asama rocks came about in the same way. In order to interpret the observed geomagnetic change as a result of change in thermo-remanent magnetization of subterranean rocks, we must assume either that the change in temperature in subterranean rocks far exceeds that caused by thermal conduction through the rocks, or that the intensity of magnetization of subterranean rocks in nature is considerably higher than that experimentally determined in the laboratory under ordinary atmospheric pressure. It must be noted, however, that, so long as we concern ourselves with the phenomena known with certainty from experiments and actual field observations, there is no other possible phenomenon to account for the observed geomagnetic change, except the development or disappearance of thermo-remanent magnetization, seeing that change in magnetic susceptibility of rock with temperature is, generally, negligible compared with that in thermo-remanent magnetism.

In the problem of secular variation in a geomagnetic field (including general secular variation of large dimensions and the above-mentioned local change), we invariably get into difficulties when we try to explain its amplitude of intensity. A way to circumvent this difficulty must be found in future.

At any rate, in this study, the general phenomenological aspect of natural remanent magnetization of volcanic rocks has been cleared, and the geophysical phenomena within the limit of influence of the natural-remanent magnetization of rocks were dealt with, where the way to correlate both phenomena was always experimentally examined.

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Throughout the course of this investigation, valuable suggestions, criticisms and encouragement were given the writer from a number of his senior colleagues. The writer takes this opportunity of thanking them all for their kindness. His hearty thanks are, especially, due to Professor H. Tsuya, who kindly allowed the writer to use various volcanic rocks collected by him, besides giving many valuable suggestions and criticisms from the petrological point of view, and to Professor C. Tsuboi for his valuable criticisms from the geophysical point of view and much encouragement throughout the course of this study.

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1. 火山岩の自然殘留磁氣及びその地球磁氣學的諸問題への應用に就いて

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地殼を構成する岩石の磁氣的諸性質は，斷片的には古くから多くの人々によつて測定されて來た．特に火山岩が野外に於ける自然狀態に於いて保持する殘留磁氣は Melloni, Folgheraiter, Mercanton, Königsberger, 中村淸二博士，松山基範博士等によつて屢々測定され，この自然殘留磁氣は火山岩が噴出後，徐々に冷却する際に地球磁場の影響によつて帶磁し，その帶磁の一部が殘留するのである事がほゞ明らかになつた．從つて火山岩の自然殘留磁氣に關する殆んどすべての研究は，その帶磁方向の決定から，火山岩噴出當時の地球磁場の方向を求めんとする事に向けられて來た．

然し，火山岩の自然殘留磁氣の一つの著しい特徵は，その強さが同一岩石の地球磁場内に於ける誘導磁氣の強さに比して一般に著しく大きいといふ事實である．從つて岩石の自然殘留磁氣は何等かの特殊の事情の下に生成されると想像される．この點に關しては，僅かに Königsberger の研究があるのみであつて，火山岩の自然殘留磁氣生成の物理學的機構は未だ明瞭でない．

他方，地球磁氣學の發達に伴ひ，地球磁場の局部的異常と地殼構造との關聯が次第に明らかになり，將來他の諸種の地球物理學的方法の併用と相俟つて，地表に於ける地球磁力分布及びその變化の測定結果から地殼構造をある程度迄明瞭に推定し得る可能性が增して來た．この場合に地殼を構成する諸岩石の磁氣的性質の知識が問題の根底を爲す可きは論を俟たないが，Reich, Königsberger 等の一部の地球物理學者を除いては，從來磁力探礦に於ける習慣の延長として，主として岩石の帶磁率のみが問題とされて來た．この場合比較的強大な火山岩の自然殘留磁氣を充分考慮する必要が當然考へられる．特に火山地域に於ける，火山活動に伴ふ地球磁場の異常變化の解釋に當つては，自然殘留磁氣の發生及び消滅を主要原因の一つとして擧げる可きであらう．

上述の如き要請によつて，この研究に於いては岩石學的，化學的性質の明らかな多くの火山岩に就いて，先づ一般的な磁氣的諸性質を吟味し，然る後主として自然殘留磁氣の諸性質を確める可き諸種の實驗を行つた．本論文各章の内容の概略は次の如くである．

第 I 章

この研究に用ひた岩石試料は，淺間，富士，天城，三原（伊豆大島），宇佐美，箱根，多賀，三宅島諸火山の噴出物であり，その產地，岩石學的分類，化學成分は第 1 節に示す通りである．之等の岩石試料に就いて，先づ弱磁場に於ける帶磁率を測定し，火山岩の帶磁率はその岩石に含まれる磁鐵鑛のノルム値に大略比例するが，ある岩石に於いては平均値よりかなり大きな偏倚を示す事を知つた（§2）．又 $\pm 19\ Oe.$ 及び $\pm 220\ Oe.$ の範圍の磁場變化に對する磁化曲線を求めた結果，すべての火山岩は磁氣履歷現象を示すが，磁場の強さ 220 *Oe.* では未だ飽和磁氣に達するには遙かに及ばず，殘留磁氣の強さ，抗磁力等を正確に求める事は出來ない．但し，之等の磁氣履歷曲線によつても各種岩石の磁氣的硬さの大勢を見る事は出來る（§3）．次に弱磁場に於ける帶磁率の溫度による變化を各種の岩石に就いて測定した．その結果，すべての火山岩は 580°C 乃至 600°C で磁性を失ひ（但し實驗誤差の範圍内で），且つ大部分の岩石に於いては溫度の增減

に對して帶磁率の變化がほゞ可逆的である．然しある岩石に於いては室溫より溫度の上昇と共に帶磁率は若干の增加を示し 400°C～450°C より比較的急激に減少する（之を標準型變化とする）のに對してある種の岩石に於いては 200°C～300°C と 500°C～600°C との二回の急激な減少を爲す等の段階的變化を示す場合がある．(第 I-4-3～I 4-14 圖).

以上の諸事實から，火山岩の磁性は，非强磁性鑛物の中にはさまれた數パーセント量の强磁性鑛物に主として起因する事，その强磁性鑛物は磁鐵鑛（磁氣變態點約 580°C）を主とするかなり複雜な固熔體，共晶である事等が推定される．例へば 200°C～300°C に於ける磁氣變態は，TiO_2 の含有率の多い岩石に限つて行はれる．(第 I-4-15 圖) 更に帶磁率の溫度による變化の形から，標準型變化の場合と雖も，お互ひに僅かながら磁氣變態點を異にする多くの强磁性鑛物粒のそれぞれに於ける變化の平均を表はすものと考へられる．(§4)

第 II 章

火山岩の自然殘留磁氣を次の如く定義する．“十分廣い範圍（$(100\ m)^2$ 以上の廣さ）に擴る熔岩流に於いて，どの部分から採集した岩石片もほゞ同一方向の殘留磁氣を保持する場合，この殘留磁氣を自然殘留磁氣と呼ぶ” この定義は，落雷，放電等による强磁場の發生に起因する岩石の殘留磁氣を除外する爲である．(§1)

採集した岩石片の殘留磁氣の强さ及び方向は無定位式磁力計に依つて測定する．岩石片を中心とする球面上の各點に於いて中心線方向の磁力を測定し，球函數分析によつて，その第一次項を採りその方向及び强さを岩石片の自然殘留磁氣の方向及び强さとして採用する．測定方式に於ける各誤差の大きさを吟味した結果，方向に於いて 1°，强さに於いて 1% 以下の測定誤差となり．(§2) 又測定結果の球函數分析に於ける第二次項以上は岩石片の形に主として起因し，岩石片の帶磁そのものはかなり一様である事が分つた．(§3)．測定結果として先づ最近の火山活動による噴出岩の帶磁方向は常にその場所の地球磁場の方向と一致する事，及び過去歷史時代，地質時代に於ける噴出岩は偏角，伏角共に現在の地球磁力の方向から數度乃至數十度偏倚してゐる事實を三原，三宅，富士，天城等の火山岩に就いて明らかにした．(§4)．更に火山岩の自然殘留磁氣のその岩石の地球磁場に於ける誘導磁氣に對する比 Q_n が，小數の場合を除いては，普通には 5～20 に及び，ある場合には 100 以上に達する事實を明らかにした．(§5)

第 III 章

次に火山岩の自然殘留磁氣の生成を實驗室に於いて再現せしめ，所與外部條件と自然殘留磁氣との關係を明らかにした．

一様な弱磁場內に於いて 600°C 以上の高溫より徐冷した火山岩は强い殘留磁氣を持ち，その磁化方向は所與磁場に等しい．之を熱殘留磁氣と呼ぶ事にする．外部磁場の强さを地球磁力のそれ（中部日本に於いて約 0·45 *Oe.*）にすれば，最近の噴出岩の熱殘留磁氣の强さは，その岩石の自然殘留磁氣にほゞ等しい．600°C 以下の溫度から弱磁場冷却を行つても，やはり熱殘留磁氣を持つ．所與磁場を一定に保てば，磁場冷却を行ふ最初の溫度が高い程熱殘留磁氣は大になり，600°C に於いて飽和値に達する．(第 III-1-2～III-1-5 圖)．但しその增加の様子は必しも單調ではなく，ある場合には段階的に增加する．溫度の條件を一定して，所與磁場を變化せしめれば弱磁場（2 *Oe.* 以下）に於いては磁場の强さと熱殘留磁氣の强さは比例するが，更に磁場が强くなるに從つて熱殘留磁氣の相對的强さは減少する．(§1)．熱殘留磁氣の强さは，0·2～5 度/分の範圍に於いては磁場冷却の速度に無關係である．

實驗條件を制限して，溫度 t_i から t_j の間のみ磁場 H 內で冷却し，後の溫度區間は無磁場冷却をする事によつて得た熱殘留磁氣を $\vec{J}^{t_j}_{t_i,\,H}$ とすれば，一般に

$$\vec{J}^{t_1}_{t_n,\,H}=\vec{J}^{t_1}_{t_2,\,H}+\vec{J}^{t_2}_{t_3,\,H}+\cdots\cdots+\vec{J}^{t_{i-1}}_{t_i,\,H}+\cdots\cdots+\vec{J}^{t_{n-1}}_{t_n,\,H}$$

が成立する．但し $\Delta t=t_i-t_{i-1}=25°$ 以上の溫度區間での實驗結果である．上述の關係が Δt の更に小さい範圍迄成立するとする理想的な場合を考へると

$$\vec{J}^{t_1}_{t_n,\,H}=\int_{t_1}^{t_n}P(t,\,\vec{H})\,dt$$

で表はされ，更に H が小さければ

$$\vec{J}^{t_1}_{t_n,\,H}=\vec{H}\cdot\int_{t_1}^{t_n}P(t)\,dt$$

となる．一度磁場內で生成された熱殘留磁氣は，磁場 H' 內で加熱する事によつて消滅するが，その際に於ける溫度と帶磁の强さとの關係は，その熱殘留磁氣が生成された場合の溫度對帶磁の强さの關係をそのまゝ逆にたどり，H' の如何に關係しない．(§2)

$P(t)$ は岩石によつて定る溫度の函數であるが，普通には 500〜550° の範圍に極大を有する簡單な形を持ち，600°C 以上，及び室溫附近では $P(t)=0$ である．然し，例へば 300°C 附近にも磁氣變態點を持つ岩石に於いては，之に對應してやはり 200〜300°C に $P(t)$ の極大が表はれる帶磁率の溫度による變化と熱殘留磁氣の生成率 $P(t)$ とを比べた結果（第 III-3-1〜III-3-2 圖）前者の急激な變化には必ず後者の極大が對應する．$P(t)$ の極大を示す溫度を熱殘留磁氣の遷移溫度と名付ければ，その遷移溫度は常に平均磁氣變態溫度にほゞ等しく，又見かけの磁氣變態點より約 50° 低い．（第 III-3-3 圖）．斯くして熱殘留磁氣の發生は，火山岩が高溫より冷却して强磁性狀態に爲つて後數十度以內の冷却の進行中に行はれる事が明らかになつた．(§3).

次に岩石片を 600°C 以上の高溫から無磁場冷却をして溫度 t_2 に至り，t_2〜t_1 の溫度區間のみ弱い磁場 H 內で冷却する．溫度 t_1 に於いて磁場を除去すれば，岩石試料はやはり殘留磁氣 (J) を持つが，更に t_1 より室溫迄無磁場空間で冷却する際，この殘留磁氣 (J) の强さは溫度の低下と共に增大する．（第 III-4-2 圖）．然も，溫度が t_1 より小なる限り無磁場空間に於いては，J は溫度 t の變化に對して殆んど可逆的に變化するが，（第 III-4-4 圖），一度 t_1 を越える高溫にすれば，最早 J〜t 關係は t に對して非可逆的である．以上の諸事實を綜合すると，火山岩の熱殘留磁氣の生成過程は次の樣であると考へられる．今，岩石內の磁氣變態溫度 θ を持つ一つの强磁性鑛物粒に注目する．弱磁場冷却の途中に於いて溫度が θ を越へその鑛物結晶が强磁性狀態に入つた直後，所與磁場の作用の下に，結晶內の若干の磁域が磁場の方向に逆轉する．而して更に溫度が降下すればこの磁域の向きは固定され，磁場を取去つても舊の狀態に戻らない．從つて，溫度降下によつて例へば內部歪の變化が生じ，磁場の働いてゐる間は，磁氣エネルギーが，歪の變化による仕事のエネルギーに打勝つて磁域を磁場の方向に固定するが，溫度が僅か降下して後磁場を取除いた時には，磁域の向きが舊に戻る爲には歪の力に對して仕事を爲さればならない故に，そのまゝ固定されてゐる，といふ物理的機構が考へられる．磁氣的に言へば溫度降下に因る抗磁力の增大である．而してこの變化自身は溫度變化に對して可逆的であると考へる．從つて，磁域の飽和磁氣の强さを I，磁域逆轉の確率を Ψ とすれば，磁氣變態點 θ の强磁性鑛物單位體積の

熱殘留磁氣は $I(\theta,t)\Psi(H,S,\theta)$ で與へられる．玆に S は內部歪を表はす量であつて，Ψ は H,S 及び θ によつて定ると考へられ，又熱殘留磁氣の重疊の法則が成立つ事實から，θ より僅か下のあまり廣くない溫度區間の降下中に熱殘留磁氣生成の殆んどすべてが終つてしまふと考へられる故に，t によらず θ のみで與へられるとした．Ψ は θ に對して可逆的であるから熱殘留磁氣 $I\Psi$ を加熱して θ に達すれば，0 に戻る筈である．又 $I(\theta,t)$ は $t<\theta$ の範圍では Curie-Weiss の法則に從つて可逆的變化する筈である．扨て一種類の岩石中の强磁性鑛物粒のその磁氣變態點 θ に關する分布函數を $g(\theta)$ とすれば，溫度 θ_2 から θ_1 迄磁場 H 內で冷却して得た熱殘留磁氣 J は

$$J(t)=\int_{\theta_1}^{\theta_2}I(\theta,t)\Psi(H,S,\theta)g(\theta)d\theta$$

で與へられ，H が小さい範圍では

$$\vec{J}(t)=\vec{H}\cdot\int_{\theta_1}^{\theta_2}(I\theta,t)\Psi(H,S,\theta)g(\theta)d\theta$$

となる．從つて無磁場空間に於いて $t<\theta_1$ では $J(t)$ は t に對して可逆的に變化し，$\theta_1<t<\theta_2$ では溫度上昇により熱殘留磁氣を消失する．今 $I(\theta,t)\Psi(S,\theta)=p(\theta,t)$ と置けば

$$\vec{J}(t)=\vec{H}\cdot\int_{\theta_1}^{\theta_2}p(\theta,t)d\theta$$

であり從つて，第 III 章 §1 及 §2 の P は

$$P(\theta)=p(\theta,0)=I(\theta,0)\Psi(S,\theta)$$

と解釋される．詳しい吟味によれば熱殘留磁氣の生成にも時效があり，高溫では時間と共に熱殘留磁氣が減少する．この時效は磁場 H 從つて J が小さい範圍ではあまり著しくないが，H が大きくなると極めて顯著になる．Ψ の意味に就いて，種々吟味をしたが，上述の如き現象論的な解釋以上に明確する事には未だ成功してゐない．(§4)

第 IV 章

火山岩の自然殘留磁氣と地球磁氣學上の二，三の問題との密接な關係を具體的實例に就いて論じた．§1 では從來の歐米に於ける研究と同樣に火山岩の自然殘留磁氣の方向から日本附近に於ける地球磁場の永年變化を推定する．三原火山安永熔岩 (1778) から推定した偏角の値は，磁氣子午線測定に關する古記錄の結果と良く調和する（第 IV-1-1 圖）富士，青木ケ原熔岩 (864)，伊豆の二，三の熔岩流の偏角及伏角の値から，少くとも第四紀の始めから現在迄の間では，日本附近に於ける地球磁力の方向の變動は數十度以內に止つてゐる事を結論した．卽ち地球磁力の方向の逆轉はこの期間には見られない樣である．

自然殘留磁氣の强さと，同一岩石を實驗室內で 600°C 以上から 0·45 Oe の磁場で冷却して得た熱殘留磁氣の强さとを比べると，有史時代以後の噴出物では前者對後者の比が 0·9～1·1 であるが，地質時代（第三紀末及び第四紀初め）の岩石に於いては 0·2～0·9 で明らかに小さい．之は地質時代の地球磁場の强さが小さかつたと考へるより，永い年月の間の消磁作用の結果と考へる方が妥當の樣である．

§2 に於いては火山附近に於ける局部的地磁氣異常の原因の解釋を行ふ. 三原，淺間兩火山の火口周邊，三原山，淺間山，三宅島等の磁氣測量の結果と，それぞれの火山を構成する岩石の磁氣的性質を對比した結果，何れの場合にもその帶磁率は小に過ぎ，地球の主磁場による火山の誘導磁化のみでは局部的異常を說明するに不充分である. 以上の如き成層火山に於いては，ほゞ齊一方向の自然殘留磁氣を持つ熔岩層と，假令自然殘留磁氣を持つてゐても多數の平均に於いてはその影響を打消し合つてゐると考へられる火山礫，火山灰等の火山碎層とから成つてゐると考へるのが至當であるといふ結論を得た.

§3 に於いては，火山活動に伴つて觀測された地球磁場の局部的異常變化の解釋を行つた. 三宅島の大噴火に伴つた著しい局部的地球磁場變動及び淺間火山の活動期に現はれる地球磁場變動の原因は，火山內部の溫度變化に因る火山體又は內部の帶磁の變化と考へるのが最も自然である然し實測された地球磁場變動量と，火山岩の磁氣的諸性質とを對比して見ると，若し岩石の溫度變化が單なる熱傳導のみに依るとすると，帶磁率の溫度變化のみは勿論，熱殘留磁氣の生成又は消滅を以てしても，到底實測された如き地磁氣變動は起し得ない事が分つた. 從つて現在知られてゐる範圍の火山岩の磁氣的性質から出發する限りに於いては，火山活動に伴つて，火山內部に著しい溫度變化が生じ（例へば赤熱熔岩，高溫の火山ガス等の貫入，流動等），熱殘留磁氣の生成又は消滅が行はれると解釋するのが，最も有力であると思ふ. 他方，例へば水上理學士が觀測された如き，地表に於ける熔岩流の冷却に伴ふその附近の地球磁場變化は，火山岩の冷却に伴ふ自然殘留磁氣の生成がその主因をなすといふ事は，殆んど疑ふ余地のない事實となつた.