

Stability of Remanent Magnetization of Igneous Rocks

E. Larson*, Mituko Ozima, Minoru Ozima, T. Nagata and D. Strangway

(Received 1968 November 15)†

Summary

Most tests of the magnetic stability of a rock provide only a crude evaluation. A better estimation could be obtained by measuring the spectrum of coercivities possessed by a rock; a simple means of obtaining the coercivity spectrum is presented.

It is also shown that a relation between the distribution of grain size and the spectrum of coercivities exists, not only for prepared samples, but for rocks as well. An evaluation of the sizes of the opaque grains effectively controlling the magnetic properties must not only take account of initial crystallization processes but must include consideration of post-consolidation, high- and low-temperature oxidation modifications. The development of effective grain size is discussed.

Results of this study have provided additional insight into the relation of the shape of a J_s - T curve and the opaque mineralogy.

Introduction

The subject of the stability of remanent magnetization is of primary importance to palaeomagnetism. Presently, however, confusion exists as to exactly where the stable components of remanence reside and the best procedure for testing stability. Laboratory studies performed on prepared samples show a definite dependence of stability on the grain size of magnetic materials. Yet, studies on natural rocks generally suggest surprisingly little relation between these two parameters. Most igneous rocks contain iron-titanium grains varying in size from 1 mm to several microns. In spite of this common size distribution, the magnetic stabilities of individual rocks vary widely. The apparent lack of grains small enough to act as single domains, has in fact, required various workers to postulate the presence of small regions within the opaque grains which behave as single domains.

In addition, there has always been some confusion about the best way of determining the relative stability of rocks in order to get the most useful data. To clarify some of the confusion, 45 igneous rock samples were subjected to a series of stability tests; thermomagnetic curves were obtained for each, and their microscopic properties were carefully examined by reflected light. The results of these studies are presented in three parts.

Section 1 contains a discussion and comparison of the various criteria for magnetic stability. Test samples include prepared samples of known grain-size distribution. In Section 2, stability data for the 45 rock samples are compared with the results of microscopic observation; conclusions drawn in Section 1 are tested and discussed. In Section 3 of the paper, the thermomagnetic curves of the samples are analysed in relation to magnetic stability.

*Present address: Department of Geological Sciences, University of Colorado, Boulder, Colorado.

†Received in original form 1967 February 20.

1. Coercivity spectra

In order to obtain meaningful palaeomagnetic information, it is imperative to examine as completely as possible the stability of remanent magnetization of rocks. a.c. demagnetization, storage tests, and low-temperature demagnetization have been used successfully for this purpose in the past. In addition, coercive force has commonly been utilized as a physical parameter to characterize the stability of magnetization. But, how well does a single number represent the stability? Rocks generally contain several ferrimagnetic minerals and each mineral grain may differ from another in size, shape, mineralogy, and magnetic behaviour. Therefore, the coercive force must vary from grain to grain and from subregion to subregion within a single rock. Hence, the stability of the remanent magnetization of rocks is most usefully represented by a spectrum of coercive forces rather than by a bulk coercive force, as discussed by Graham (1953).

Coercivity spectrum

A coercivity spectrum may be obtained by successive partial thermal demagnetization of total TRM, as suggested by Everitt (1962). The heating of a rock, however, often results in chemical change of the samples, and the data thus obtained may not represent the primary magnetic properties of the rock. Instead, successive a.c. demagnetization of total ARM can be used to obtain a coercivity spectrum, as explained below. The latter method seems superior to the former since it is relatively simple and does not lead to chemical alteration of the sample.

Many rocks contain a large number of ferrimagnetic grains each consisting of at least one domain. Every domain or subregion probably can be characterized by a particular coercivity h_c . If a steady magnetic field larger than the h_c of a particular subregion is applied, the magnetization of the subregion will be aligned in the direction of that field. If a rock is subjected to the simultaneous action of a steady d.c. field (h) and a decreasing a.c. field (\tilde{H}) then all the subregions whose coercivity is smaller than $(\tilde{H} + h)$ will be aligned in the direction of the applied field h . On the reversed part of the cycle, the resultant maximum field will be $(\tilde{H} - h)$ and some of the subregions which were aligned by the positive maximum field will not be realigned. If the a.c. field is reduced to zero and subsequently the d.c. field is removed, the subregions which were aligned in the direction of h will provide an anhysteretic remanent magnetization, that is, ARM.

Only those domains or subregions whose coercivity lies between \tilde{H}_1 and \tilde{H}_2 will be given partial ARM during the reduction of the a.c. field from \tilde{H}_1 to \tilde{H}_2 . It has been experimentally demonstrated (Patton & Fitch 1962) that the total ARM is the summation of all the partial ARM's obtained over each interval of decreasing a.c. field. Accordingly, during subsequent a.c. demagnetization of the acquired ARM the domains which are demagnetized by increasing the field from \tilde{H}_2 to \tilde{H}_1 will be those which were fixed during a decrease of the field from \tilde{H}_1 to \tilde{H}_2 . Consequently, a coercivity spectrum can be obtained by differentiating the a.c. demagnetization curve of the total ARM with respect to \tilde{H} .

Fig. 1 represents a coercivity spectrum of basalt obtained by the procedure described above. The total ARM was produced by reducing the a.c. field from a peak of $\tilde{H} = 625$ Oe (the highest field strength attainable at the University of Tokyo) to zero in the presence of a steady field of 5 Oe. Although a field of 625 Oe will not in general be sufficient to produce a maximum ARM, it is clear from the foregoing discussion that the procedure should yield a spectrum of coercivities up to the peak field strength used. For the basalt sample shown, the distribution of the coercivities is roughly Gaussian.

The spectrum shifts to the left with increased temperature (Fig. 2). From the analysis of the temperature change of the coercivity curve, a theory has previously

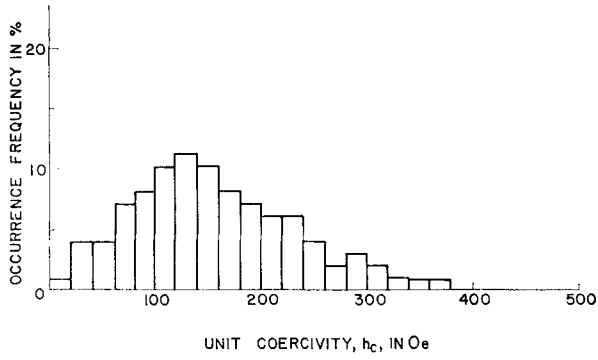


FIG. 1. Coercivity spectrum of basalt at room temperature.

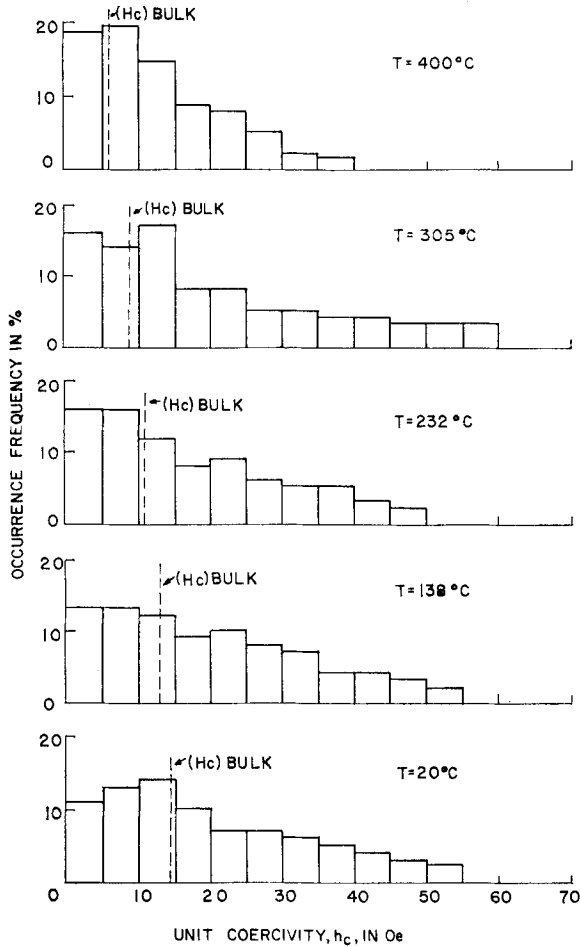


FIG. 2. Coercivity spectrum of Kamaishi magnetite at various temperatures.

Downloaded from https://academic.oup.com/gji/article/17/3/263/2182047 by guest on 16 August 2022

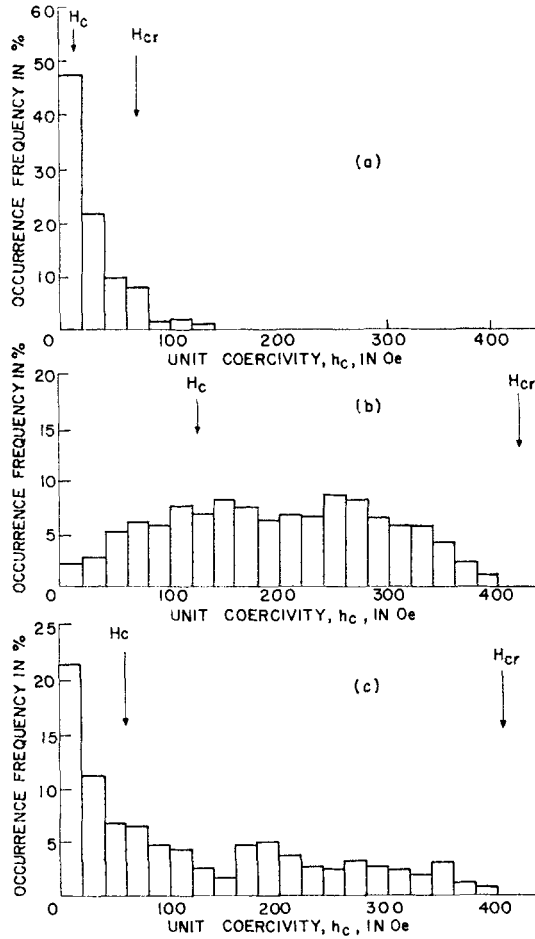


FIG. 3. (a) Coercivity spectrum of coarse grained magnetite. (b) Coercivity spectrum of fine grained magnetite. (c) Coercivity spectrum of a mixture.

been offered to explain the acquisition of remanent magnetizations—IRM and inverse TRM—which are primarily caused by domain-wall shifts (Ozima, Yama-ai & Nagata 1963). In this theory, it is assumed that a distribution of h_c is Gaussian in nature and gives a bulk coercivity $H_c(T)$ which is a function of temperature.

Since the bulk coercive force $H_c(T)$ decreases with increasing temperature, the h_c distribution also shifts towards a lower coercive-force value and maintains its Gaussian nature. Fig. 2 shows the coercive force spectrum obtained at various temperatures for a magnetite-grain assemblage (mean grain size $\sim 100\mu$). This result supports the assumption which underlies the theory of the remanent-magnetization acquisition of multi-domain grains.

Coercivity spectrum and stability of remanent magnetization

It has been pointed out by two of the authors (Ozima & Ozima 1965) that magnetic characteristics at low-temperature seem to indicate the dominant single-domain origin of TRM. The grain-size effect of susceptibility also supports the idea that TRM resides in single-domain regions (Verhoogen 1959). It is experimentally established that the finer the ferrimagnetic grains in rocks, the more stable is the

TRM. Apparently, the finer the grain size, the more chance for single-domain grains or subregions to exist. Moreover, even if the ferrimagnetic grains are multi-domain, the coercive force increases with decreasing grain size by some mechanism which is not yet clearly understood (Kittel 1956; Stacey 1962). This, then, explains why the finer grains of multi-domain size can also acquire a stable and strong IRM or inverse TRM, both of which are shown to be acquired by domain-wall movements (Ozima & Ozima 1965). Hence, in both types of remanent magnetization, whether acquired by rotation of a single domain or by domain-wall movements, it may be expected that the grain-size distribution will control the stability of the remanence. A coercivity spectrum which characterizes the distribution of stability of remanent magnetization in a rock should then reflect the grain-size distribution of ferrimagnetic minerals.

If this is so, microscopic observation of polished sections should yield a direct measure of the stability of remanent magnetization. Optical observation would then provide a rapid but effective means of evaluating the stability of palaeomagnetic samples. This hypothesis will be examined in Section 2; in this section we will study some prepared samples of known grain size and show how the grain-size distribution correlates with the coercivity spectrum.

For this study, three grain-size assemblages, coarse, fine, and a mixture of the two, were prepared. The coarse and fine samples have an average grain size of 1 mm and 0.5μ , respectively. About equal parts of coarse and fine grains were mixed to constitute the mixed sample. Each volume of sample was diluted with 20 volumes of very fine silica powder (grain size $\sim 1\mu$) in order to simulate the relations of ferrimagnetic minerals in rocks. Fig. 3 shows the coercivity spectra of the three samples. As seen in the figure, these spectra reflect the grain-size distribution of each case. There is a single peak in the curve of the coarse and fine samples, corresponding to the single-grain-size mode of each of these samples.

The spectrum of the mixed sample represents a combination of the curves of the coarse and fine samples. The coercivity value corresponding to the peak in the spectrum of the coarse sample is between 0 and 10 Oe, which is about $\frac{1}{10}$ of the value of the peak of the spectrum of the fine sample. This fits well with the fact that the bulk H_c of the coarse sample is 13 Oe, whereas that of the fine sample is 126 Oe.

On the figures is also indicated the value of H_{cr} . H_{cr} is the remanent coercive force, which is defined as the d.c. field necessary to reduce to zero a remanence produced by a saturating d.c. field.

At this point, it may be instructive to show how a bulk-coercivity measurement commonly gives a deceptive measure of the stability of remanent magnetization of rocks. It is frequently observed that rocks which have coercive forces significantly

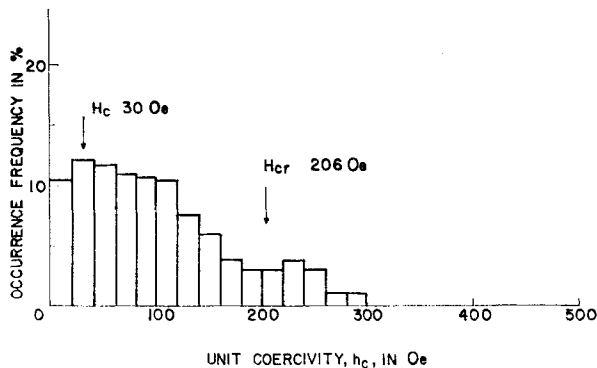


FIG. 4. Coercivity spectrum of Warabidaira dolerite.

higher than the Earth's magnetic field still possess very unstable magnetization. As an example, consider the Warabidaira Dolerite which has been extensively studied by Akimoto & Kushiro (1960). The bulk coercive force of the dolerite is about 30 Oe, which is much larger than the Earth's magnetic field. In spite of this relatively high coercive force, the remanent magnetization is extremely unstable and the direction of the magnetization quickly follows the direction of the Earth's magnetic field. The h_c spectrum of this sample is shown in Fig. 4. As seen in the figure, more than 10 per cent of the domains have a coercive force lower than 20 Oe, whereas about 40 per cent of the distribution have coercive forces above 100 Oe. It is probable that the former component is responsible for the extremely unstable component of the remanent magnetization, whereas the tail of the spectrum contributes to the observed high bulk coercivity.

Conclusion

A bulk coercivity may not be an adequate measure of the remanent magnetization stability of rocks in which numerous ferrimagnetic grains with different compositions and grain size exist; the stability is better represented by a coercivity spectrum. Experiments, using prepared samples, show that the coercivity spectrum is a direct reflection of the grain-size distribution. Therefore, it is suggested that a reliable stability criterion of the remanent magnetization of rocks could be obtained by a microscopic observation of the grain-size distribution of their ferrimagnetic grains.

2. Coercivity spectra and microscopic studies of natural samples

In Section 2 of this paper, magnetic stability relations of the suite of 45 igneous rocks are presented. For each sample a comparison was made between grain-size distribution as observed with the reflecting microscope and measured magnetic stability properties using a.c. demagnetization, low-temperature cleaning, determination of the coercive and remanent coercive forces, and the coercivity spectrum.

The last part of Section 2 is a discussion of effective grain size in rocks and the various factors which control it.

Description of samples

In order to provide a meaningful sampling, 45 igneous rocks of diverse petrologic character were selected from Japan, California, Oregon and New Mexico. They include granitic and basic intrusives, rhyolite, welded tuff, dacitic tuffaceous sediments, and andesitic and basaltic lavas. Except for some granitic samples of Cretaceous Age, all rocks are Cenozoic.

Microscopic studies

Polished sections of these rocks were studied under oil with an Olympus reflecting microscope at magnifications of 1200 to 2400—grains as small as 0.5μ could be observed. The polished sections were made from fragments of the same samples which had been subjected to the stability tests.

Detailed investigation of the opaque mineral assemblages was made, including estimation of their initial grain-size distribution and effective grain-size distribution. The last estimate takes into account processes of high-temperature oxidation which may affect the titanomagnetite grains after initial crystallization and which lead to the subdivision of large grains into a number of smaller, magnetically independent ones.

For rocks containing an abundance of ferrimagnetic crystals, the first 100 to 200 grains that passed the field of view were described. However, for samples containing few opaque grains—mostly granitic rocks—an estimate of the number of

grains per unit area and the ratio of large to small grains was obtained by counting all grains encountered during a series of traverses across the sample. Undoubtedly so few grains were described that only a rough estimate of the distribution of the opaque minerals in each sample was obtained. This is particularly true for extrusive rocks which show marked variability in initial grain-size distribution and subsequent high-temperature oxidation.

Titanomagnetite ($\text{Fe}_2\text{TiO}_4 \cdot \text{Fe}_3\text{O}_4$ solid solution), was the predominant opaque in most samples. In many samples titanomaghemite was identifiable in small amounts and in a few samples the original titanomagnetite was greatly altered to titanomaghemite. Ilmenite, which was commonly present, occurred as single crystals (either free or intergrown with titanomagnetite) and as thin to thick lamellae in the (111) planes of magnetite. In those samples affected by high-temperature oxidation, the ilmenite commonly had been altered to metalmenite and pseudobrookite. Hematite was generally present in minor amounts as a filling or cracks in silicates and opaque grains, as a partial to complete replacement of titanomagnetite, and as a replacement around the margins of highly oxidized olivine grains. In no case was olivine oxidized sufficiently to cause the ferrous silicate to break down to magnetite as described by Wilson & Haggerty (1966). Small amounts of opaque sulphide minerals were also present in some samples.

As verified by a study of the thermomagnetic (J_s - T) curves, titanomagnetite or titanomaghemite is the principal carrier of remanent magnetization of all samples. Therefore, detailed observation was limited to these opaques.

Stability tests

Table 1 provides for each sample a brief summary of the opaque mineralogy, effective grain size, Curie point, and magnetic stability determined in several ways.

The bulk-coercive-force values (H_c) fall in two groups, one above and one below 100 Oe. A histogram of these is shown in Fig. 5(a). Similarly, there is a double set of remanent-coercive-force values (H_{cr}), one over and one less than 300 Oe, as shown in Fig. 5(b). Samples occurring in either the low or high groups of H_c measurements usually show a similar placement in the H_{cr} measurements. a.c. demagnetization data as provided in Table 1 also show good correlation with the coercive-force results. In fact, of all these tests, a.c. demagnetization is the most meaningful. However, it commonly fails to define the low-coercive-force part of the spectrum.

It is clear from Table 1 that all of the various tests provide some measure of relative stability. Those specimens containing abundant small grains, either as a result of initial crystallization or subsequent oxidation generally proved to be more stable, whatever the test applied. However, as pointed out in Section 1, rocks contain a range of stability components. Therefore, a single coercivity measurement cannot adequately characterize this range.

For this reason, the coercivity spectrum was also determined for each specimen in the manner described in Section 1. The ARM was produced by a steady field of

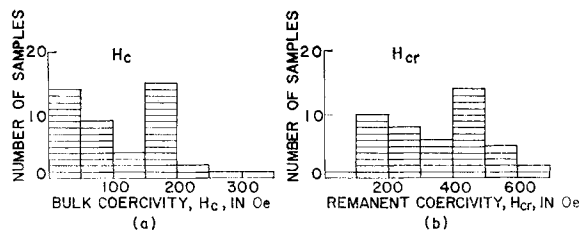










FIG. 5. (a) Histogram of bulk coercivities. (b) Histogram of remanent coercivities.

Table 1

Synopsis of rock-magnetic and microscopic data

•	Measurement on separated opaques.		Essentially only large simple grains; 20 μ.
†	Sample heated to 600 °C.		Large and small grains.
	Mostly small grains; 15 μ.		Combination of small grains and oxidized larger grains.
	Initially large grains showing effects of oxidation, i.e. hematite or ilmenite lamellae and planes.	L, M, H	Low, moderate, high state of oxidation.

E. Larson *et al.*

Sample	Rock type	Opaque minerals	NRM (e.m.u. cm ⁻³)	H _c (Oe)	H _{cr} (Oe)	a.c. demagnetization	Curie temperature	Effective grain size	Oxidation
1	Rhyolite welded tuff	Mt, Ht	2 × 10 ⁻³	94	480	—	585		L-M
2	Diabase (sill)	Mt, Il, Mh, Ht	4 × 10 ⁻³	71	240	Intermediate	515		L-M
3	Basalt	Mt, Mh, Il	4 × 10 ⁻⁴	78	180	Unstable	505 596		L
4	Basalt	Mt, Ht, Il	5.2 × 10 ⁻⁴	265	600	—	560		M-H
5	Basalt	Mt, Il, Mh, Ht	4.6 × 10 ⁻⁴	187	520	Stable	560		L-M
6	Basalt	Mt, Il, Ht, Mh	2.6 × 10 ⁻²	161	460	Stable	570		M-H
7	Basalt	Mt, Il, Ht	2.6 × 10 ⁻³	161	520	Stable	555		H


























8	Basalt	Mt, Il, Ht	1.4×10^{-3}	161	450	Stable	555		H
9	Basalt	Mt, Il, Mh, Ht	2×10^{-3}	184	402	Intermediate	565		M
10	Basalt	Mt, Mh, Il, Ht	4.6×10^{-4}	81	372	---	210 580		L-LM
11	Basalt	Mt, Il, Mh, Ht	9.6×10^{-4}	123	442	Intermediate	510 > 580		L-LM
12	Basalt	Mt, Il, Ht	2.8×10^{-4}	346	570	Stable	555		H
13	Basalt	Mt, Mh, Il	1.2×10^{-3}	45	108	Very unstable	225 515 590		VL
14	Basalt	Mt, Mh, Il	8.2×10^{-4}	55	148	Very unstable	265 510		VL
15	Basalt	Mt, Mh, Il	5.8×10^{-4}	71	260	Very unstable	300 510 > 580		VL
16	Basalt	Mt, Il, Ht	2×10^{-3}	223	560	Stable	580		H
17	Basalt	Mt, Il, Ht	1.5×10^{-3}	174	480	Stable	570		H
18	Basalt	Mt, Ht	7.0×10^{-3}	242	650	Stable	580		H
19	Basalt	Mt, Il, Ht	3.5×10^{-3}	161	410	Stable	565		L-M
20	Basalt	Mt, Il, Ht	2.4×10^{-3}	168	464	Stable intermediate	575		L-M

Table 1 (continued)

Sample	Rock type	Opaque minerals	NRM (e.m.u. cm ⁻³)	H _c (Oe)	H _{cr} (Oe)	a.c. demagnetization	Curie temperature	Effective grain size	Oxidation
21	Basalt	Mt, Ht, Mh (?)	4.4 × 10 ⁻³	187	416	Stable	305 585		M
22	Basalt	Mt, Il, Mh, Ht	4.6 × 10 ⁻²	161	398	Intermediate	565 > 600		L-M
23	Basalt	Mt, Il, Ht	2.4 × 10 ⁻³	152	424	Stable	555		M-H
24	Basalt	Mt, Il, Mh, Ht	2.15 × 10 ⁻³	123	340	Unstable	350 550		L
25	Basalt	Mt, Il, Ht	—	184	386	—	515 580		H
26	Granitic	Mt, Ht	—	23*	196	—	570		L
27	Granitic	Mt, Ht	—	16*	94	—	570		L
28	Granitic	Mt, Ht	—	19*	180	—	570		L
29	Granitic	Mt, Ht, Il	—	26*	160	—	570		L
30	Granitic	Mt, Ht	—	16*	107	—	570		L
31	Granitic	Mt, Ht	3 × 10 ⁻⁴	78*	520	High intermediate	570		L

32	Granitic	Mt, Ht	10^{-6}	11*	106	Unstable	570		L
33	Granitic	Mt, Ht	—	197*	380	—	570		L
34	Basalt	Mt, Il, Ht	3×10^{-2}	96	420	—	500		L
35	Tuffaceous sediment	Mt	1×10^{-4}	61	250	—	525 590		
36	Tuffaceous sediment	Mt	2×10^{-4}	32	170	—	310 525		
37	Dolerite intrusive	Mh, Il	1×10^{-4}	39 103†	254 360†	Unstable	150 565		L-M
38	Dolerite intrusive	Mh, Il	1×10^{-4}	48 181†	232 418†	Unstable	535 > 600		L-M
39	Dolerite intrusive	Mh, Il	1×10^{-4}	31 174†	206 400†	Unstable	555 600		L-M
40	Andesite	Mt, Il, Mh, Ht	9×10^{-4}	142	434	—	400 570		M
41	Andesite	Mt, Il, Mh, Ht	6×10^{-4}	35	188	—	540		L-LM
42	Andesite	Mt, Il, Mh, Ht	6×10^{-4}	29	174	—	560 > 600		L-LM
43	Andesite	Mt, Il, Mh, Ht	6×10^{-4}	45	208	—	350 > 540		L-LM

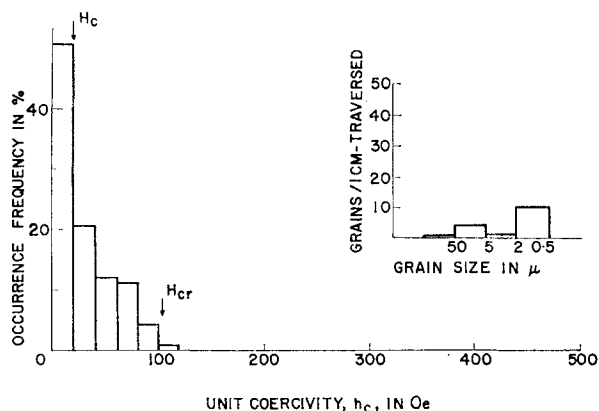


FIG. 6. Coercivity spectra and grain size distribution. Sample 32—coarse-grained granite.

10 Oe acting upon the sample, as an a.c. field was lowered from a peak intensity of 625 Oe. Comparison of the curve showed that each type of grain-size distribution was associated with a characteristic coercivity spectrum. These 'typical' spectra curves are shown in Figs 6–13. Figs 6–8 show variation in acidic (granitic composition) rocks and are arranged in order of decreasing grain size. Fig. 9 shows a curve obtained from a relatively coarse-grained basic intrusive; Figs 10–13 show results obtained from basalts in order of decreasing effective grain size.

Observed grain-size distribution

With each coercivity curve (Figs 6–13) are plotted estimated grain size distribution curves for that sample. The solid line outlines the distribution of grains resulting from initial crystallization. Changes in this distribution curve can result from high-temperature oxidation whereby non-magnetic ilmenite (in some cases, hematite) lamellae are formed which effectively subdivide the larger crystals into a multitude of smaller ones. It is difficult to estimate the number and sizes of the small grains formed in this manner. For that reason, the dashed effective grain size distribution curve shows only those changes that occurred in the larger grain size fractions. It is evident, however, that a homogeneous grain 50μ on an edge could readily produce

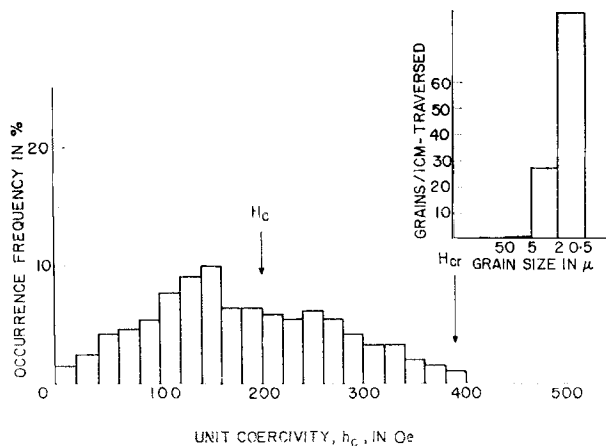


FIG. 7. Coercivity spectra and grain size distribution. Sample 33—fine-grained granite.

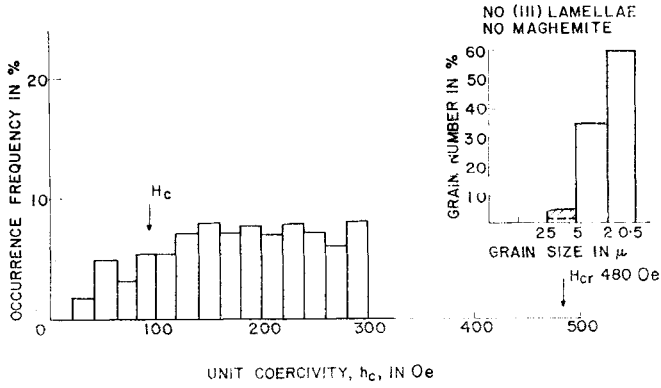


FIG. 8. Coercivity spectra and grain size distribution. Sample 1—welded tuff.

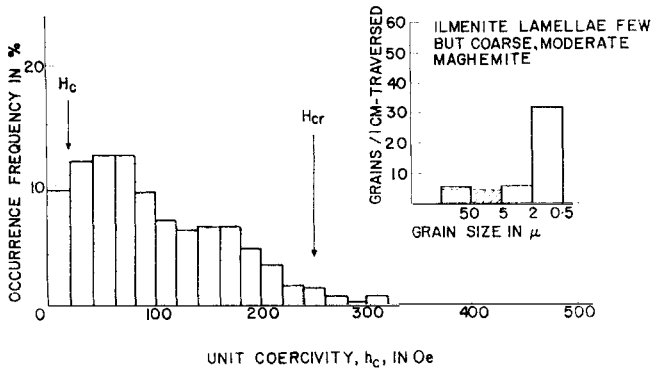


FIG. 9. Coercivity spectra and grain size distribution. Sample 37—dolerite intrusive, coarse-grained.

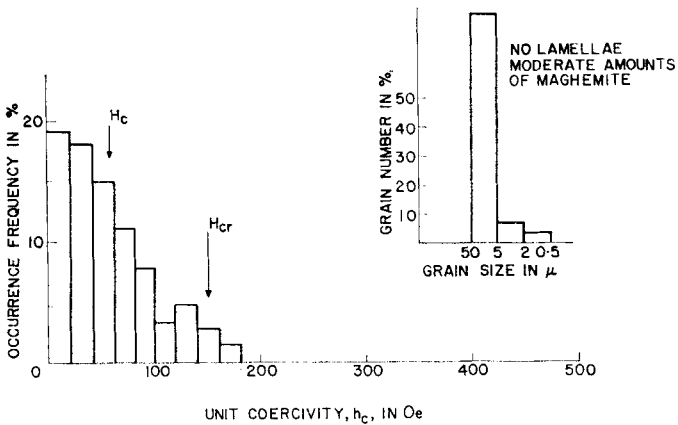


FIG. 10. Coercivity spectra and grain size distribution. Sample 14—basalt, large effective grain size.

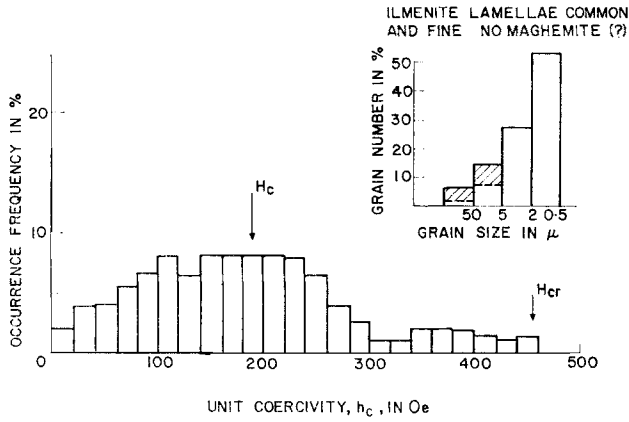


FIG. 11. Coercivity spectra and grain size distribution. Sample 20—basalt, mixture of effective grain sizes.

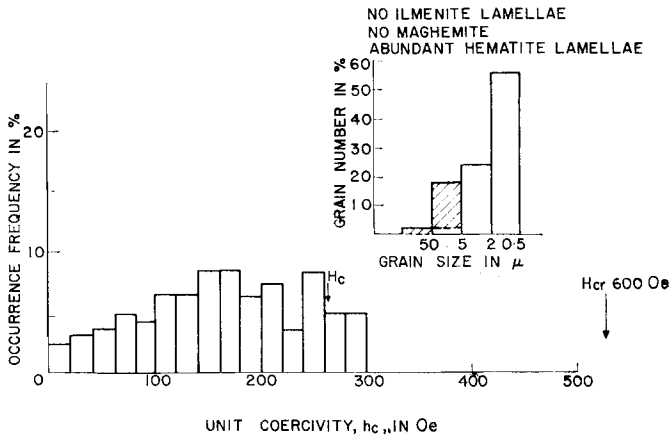


FIG. 12. Coercivity spectra and grain size distribution. Sample 4—basalt, predominantly small effective grain size.

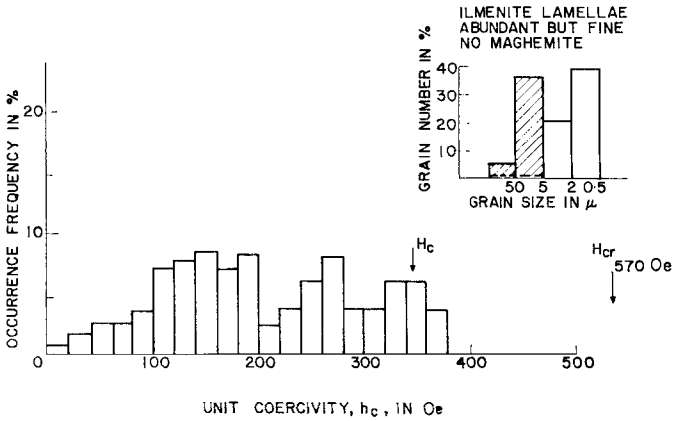


FIG. 13. Coercivity spectra and grain size distribution. Sample 12—basalt, small effective grain size.

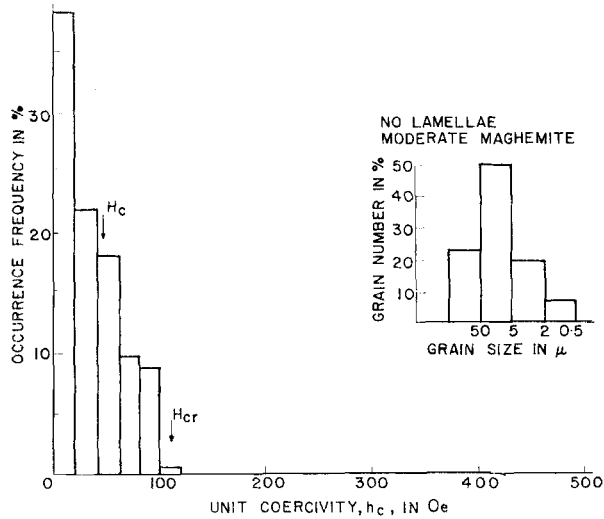


FIG. 14. Coercivity spectra and grain size distribution. Coercivity spectrum and grain size distribution before heating. Sample 13—basalt.

500 grains 5μ or less in size. Hence, even small changes in the large grain size fraction could produce large increases in the small size fractions.

In the initial and effective grain size curves the ordinate is given in number percentage of grains of each size range or as number of grains of each size range per centimetre traversed.

Correlation of grain size and coercivity spectrum

As is shown in Figs 6 and 10, where large simple grains are predominant the spectrum is composed mostly of low coercive force components and the sample loses most of its ARM at very low a.c. field strengths—i.e. less than 100 Oe. Where most of the opaque occurs as small crystals, 5μ or less (e.g. Figs 8 and 13), few low coercive force components are present. Rather, the magnetic materials are composed primarily of an intermediate coercivity fraction and, to a lesser extent, a high coercivity portion. For samples showing mixtures of the two extremes of grain size intermediate spectrum curves are obtained.

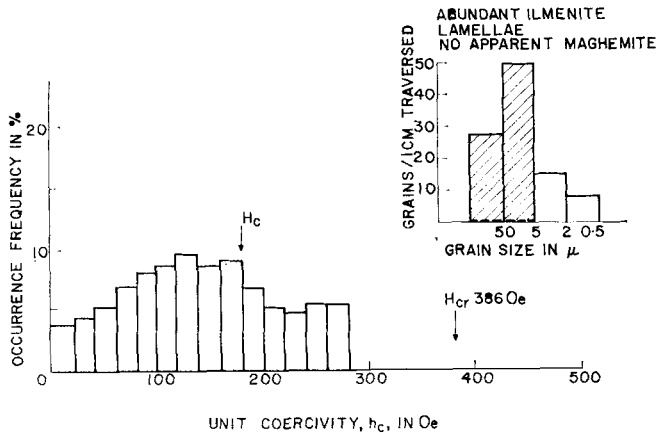


FIG. 15. Coercivity spectra and grain size distribution. Coercivity spectrum and grain size distribution after heating for 14 h at 800°C . Sample 13.

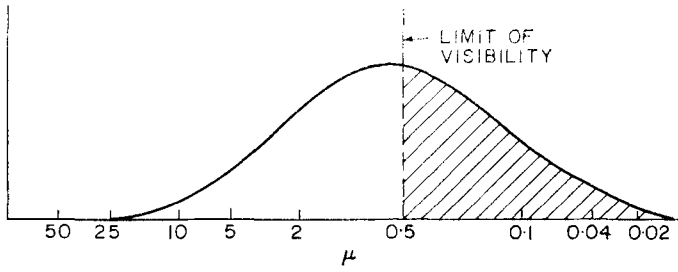


FIG. 16. Coercivity spectra and grain size distribution. Grain size distribution curve. Sample 1—welded tuff.

An interesting case is presented by sample 13 which originally mainly contained large homogeneous opaque grains and displayed a low stability spectrum (Fig. 14). After heating for 14 h at 800°C in air, however, all grains were oxidized to some degree—about 20–25 per cent were completely altered to pseudobrookite and hematite, and about 50–60 per cent developed abundant ilmenite lamellae. Subdivision of the large grains by the lamellae effectively altered the coercivity spectrum to that characteristic of a stable sample (Fig. 15).

As early as 1953, Graham recognized that the presence of ilmenite lamellae would provide a means of reducing grain size and increasing coercivity in magnetite; our experimental evidence supports this idea. Powell (1963) pointed out more recently that in the dikes he studied the higher coercivities were associated with a higher density of ilmenite lamellae in the magnetite grains.

As noted in Section 1 and verified by our results on natural and prepared samples, the coercive force, and therefore the magnetic stability of a magnetic material, increases as the grain size diminishes. For samples containing abundant small grains, either from initial crystallization or from subdivision of large grains, mechanically or by the development of oxidation lamellae, the microscopic coercivity spectrum shifts toward the higher coercive force values. The general increase in stability, with decreasing grain size, must in part be related to increased restrictions being placed on domain-wall movements and in part to the presence of more abundant single-domain grains. Morrish & Yu (1955) calculated that equant single-domain grains of magnetite can be no larger than 0.14μ and probably near $0.02\text{--}0.03\mu$. This is below the limit of visual resolution of our microscope (0.5μ), and such grains can hardly be observed. However, the abundance of grains near the limit of visibility in all stable samples makes it highly probable that there are many grains smaller than the visual limit. Grain-size distribution curves are shown in Figs 16 and 17 for two samples, 1 and 16, respectively. Sample 1 is a welded tuff in which the grain

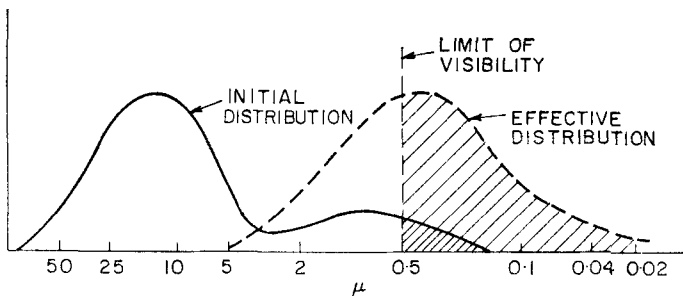


FIG. 17. Coercivity spectra and grain size distribution. Grain size distribution curve showing effect of lamellae development. Sample 16—olivine basalt.

size is generally small as a result of an explosive mode of origin. As seen, an extension of a normal curve (shaded part of graph) leads to the conclusion that many grains less than 0.5μ should exist.

Sample 16 is a basalt in which the original distribution, determined by initial crystallization (solid line) is double peaked; relatively few grains less than 0.5μ would have been present initially. As a result of later high-temperature lamellae growth, however, the large grains ($> 5\mu$) have been virtually eliminated and the visible numbers of small grains increased to such proportions that many grains less than 0.5μ are probably present. Therefore, it seems probable that grains small enough to be single-domains do exist in rocks and could be responsible for the highly stable components of remanent magnetization.

Development of initial grain size

In any igneous rock the prime factors controlling the initial grain size of titanomagnetite are the rate of cooling and, to a lesser degree, the amount and oxidation state of the iron oxides available.

In a large deep-seated magma, cooling rates are slow and few centres of crystallization normally develop: most crystals grow to large size. Deep-seated rocks, therefore, contain an abundance of large grains, commonly 0.1 to 1 mm in diameter but there is also present a scattering of small grains at least as small as 0.5μ —the limit of visibility. Near the margins of plutons where cooling may be more rapid small crystals may become relatively abundant.

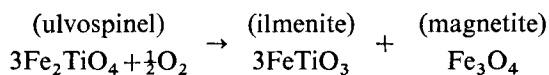
Volcanic rocks commonly display two generations of opaque grains. Crystals formed either prior to eruption (slow cooling) or shortly after eruption (more-rapid high-temperature diffusion) tend to be larger and better formed than those formed during the final stages of solidification. Typically, abundant, exceedingly-fine (less than 1μ) opaque crystallites (commonly of skeletal habit) will be found associated with the last silicate phases to solidify.

The other factor that influences initial crystal size is the ratio of $\text{Fe}^{3+}/\text{Fe}^{2+}$. For magnetite to precipitate, there must be an excess of iron in the ferric state. The oxidizing environment and therefore the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio is generally controlled by the fugacity of oxygen which in turn is predominantly dependent on the amount of water present in the magma (Buddington & Lindsley 1964). In weakly oxidizing environments little Fe^{3+} is present and most iron (Fe^{2+}) finds its way into silicate structures. The resultant size of opaque crystals will depend to some degree on how readily the magnetite precipitates from the melt and at what stage in the cooling process it does so.

Post-crystallization grain-size changes due to high-temperature oxidation

Most magnetite precipitated from natural melts contains titanium. (For a more complete development of these ideas the reader is referred to Buddington & Lindsley 1964). Generally, the higher the temperature and the lower the $f(\text{O}_2)$ (oxygen fugacity) the greater the content of titanium in the structure. Basaltic magnetite, for example, which crystallizes at about 1050–1200°C, initially contains 60–80 mole per cent ulvospinel, whereas, magnetite in a granite (which commonly crystallizes at 5–600°C) contains as little as 5–15 per cent.

Subsequent changes in temperature and $f(\text{O}_2)$ after crystallization of the opaque will tend to be accompanied by a readjustment in the Ti/Fe ratio. In a basalt, as the temperature decreases and if small percentages of volatiles, mostly water, are present, the titanomagnetite begins to oxidize according to the following approximate equation:



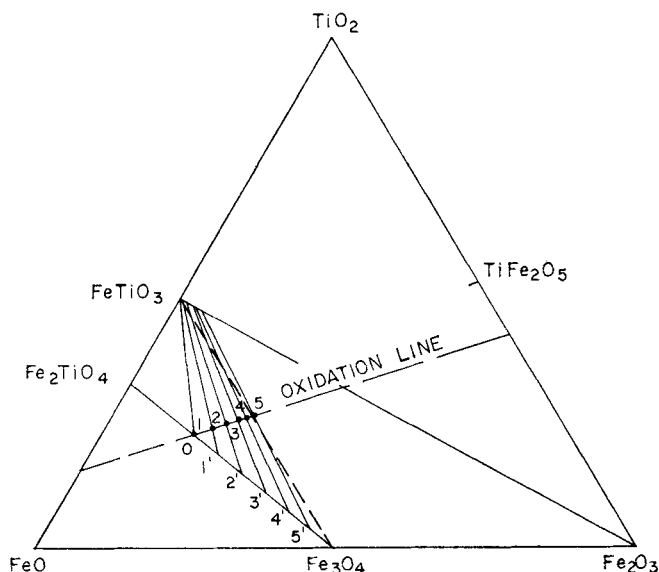


FIG. 18. Coercivity spectra and grain size distribution. Progressive high-temperature oxidation of a titanomagnetite grain to ilmenite and magnetite.

This reaction is generally restricted to temperatures above 600°C. The resultant ilmenite is structurally incompatible with the titanomagnetite host and at elevated temperatures diffuses into the (111) planes of the titanomagnetite where it forms lamellae. As temperature continues to decrease and if oxidizing conditions prevail, most of the ulvospinel component will have been altered to ilmenite and the original titanomagnetite will have been converted to nearly pure magnetite. Further high-temperature oxidation will cause alteration of the ilmenite and magnetite phases to hematite, pseudobrookite, or possible rutile. A fine display of photomicrographs representing the progressive stages of high-temperature oxidation of titanomagnetite are presented by Wilson & Haggerty (1966).

The progressive oxidation of a basaltic titanomagnetite is shown in Fig. 18. It is evident in the figure that as progressive oxidation occurs, the bulk composition of the opaque grain moves toward the Fe_3O_4 - FeTiO_3 (dashed) join, finally reaching this line (possibly a little past it) when maximum oxidation of the ulvospinel component has occurred.

If a sufficient increase in $f(\text{O}_2)$ occurs, it may be possible, at high temperatures, for the original titanomagnetite to oxidize directly to titanohematite (and perhaps pseudobrookite). Hematite, like ilmenite, has a hexagonal structure; it also tends

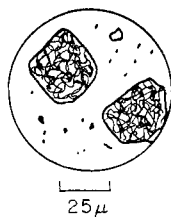


FIG. 19. Coercivity spectra and grain size distribution. Development of irregular septa and lamellae of hematite in titanomagnetite as a result of high-temperature oxidation.

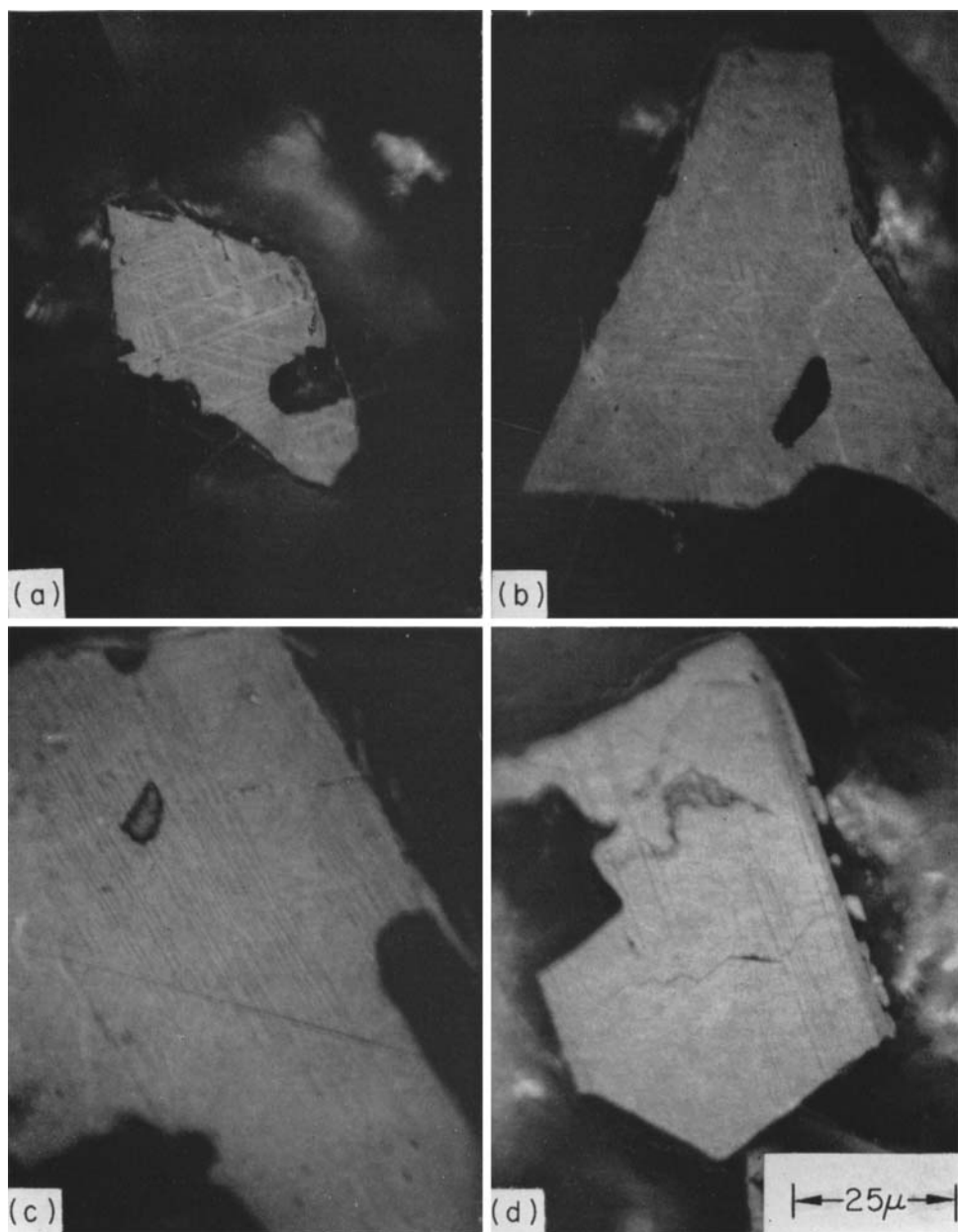


FIG. 20. Photomicrographs showing typical lamellae development.

to diffuse out of the spinel structure, forming regular or irregular and wavy lamellae and septa, or, in some cases, small blebs. In some instances, the hematite alteration is ubiquitous but on such a fine scale that it can be seen only under close inspection at high magnification (Fig. 19). This type of high-temperature alteration has affected sample 4 (Fig. 12) and has produced a very stable magnetization. It is, of course, possible that oxidation does not occur. Where extremely quick chilling occurs or where, because of the presence of sulphur, hydrogen, or carbon (from assimilation of vegetation during flow) reducing conditions prevail, the original titanomagnetite may be little affected.

Polished-section studies reveal that all titanomagnetite grains in a rock are not oxidized to the same degree. The particular stage of oxidation reached by any one grain is the result of the local micro-oxidational environment. In specimens showing low oxidation a few grains may be completely oxidized, whereas in highly oxidized samples some grains may be unoxidized.

The end result of the post-crystallization high-temperature oxidation is the development of essentially non-magnetic lamellae within the titanomagnetite grains. Since magnetostatic and exchange interactions are probably minimal across them, they effectively compartmentalize the larger grains into a great number of smaller, magnetically independent regions. As seen in the photomicrographs, Fig. 20, magnetic compartments at least as small as 0.5μ can be formed by this process. The resultant grains can be equant, tabular, or elongate, the particular case determined by the degree of development of the appropriate sets of (111) lamellae.

Lamellae formation is the result of diffusion, a temperature-dependent process, which is particularly active at temperatures above 600°C . (Buddington & Lindsley 1964). Therefore, the formation of lamellae and the subdivision of the large crystals should occur at temperatures above the Curie point and, therefore, the magnetization of oxidized samples should represent highly stable TRM acquired during the initial cooling of the lava.

As previously noted, the Ti/Fe ratio in the initial titanomagnetite depends upon the temperature at which the rocks solidified. In acidic rocks which generally contain only small amounts of TiO_2 initially, the growth of ilmenite lamellae by oxidation is of small importance.

Low-temperature oxidation

As pointed out by Verhoogen (1962) the stable iron-titanium oxide minerals in an oxidizing environment are hematite and rutile. At high temperatures and in an adequately oxidizing environment, titanomagnetite alters to these phases in a matter of hours. However, in natural conditions, these reactions do not generally go to completion. Commonly, the oxidation of ulvospinel to ilmenite and some magnetite to hematite is as far as the reaction proceeds. And in some cases, very little oxidation of any sort has occurred at high temperatures.

During the low-temperature portion of the cooling process and later, by circulation of ground water and during weathering, the unoxidized portions of the iron-titanium opaques are vulnerable to continued oxidation.

Low-temperature oxidation sometimes involves a direct alteration to hematite. Perhaps more commonly, however, it involves a single-phase alteration of titanomagnetite to titanomaghemite (Katsura & Kushiro 1961; Buddington & Lindsley 1964). Pure maghemite ($\gamma\text{Fe}_2\text{O}_3$) is a metastable iron oxide which possesses the hematite composition but an inverse spinel structure. As oxidation occurs, the bulk composition of a titanomagnetite phase moves off the ulvospinel-magnetite join towards the ilmenite-hematite join along an oxidation line (see Fig. 21). High-grade titanomaghemite would be found near and beyond the $\text{FeTiO}_3\text{-Fe}_2\text{O}_3$ join.

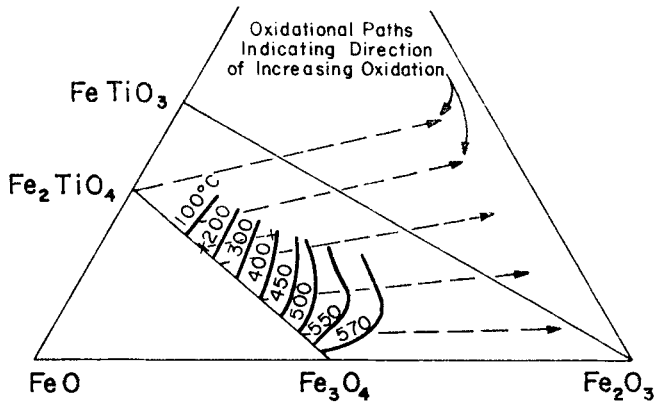


FIG. 21. Variation of Curie temperature with Ti content and low-temperature oxidation (after Akimoto *et al.* (1957)); progressive oxidation path of a sample from position \times to position $+$.

As shown in Table 1, 20 of the 45 samples have been affected by low-temperature titanomaghemite formation. Under the microscope, titanomaghemite—distinguished by its blue-grey colour—is seen to occur on prominences along the crystal margins or along fractures. Often the brown-grey of the titanomagnetite grades imperceptibly into the blue-grey colour. Kushiro (personal communication) notes that only titanomaghemite close to the FeTiO_3 – Fe_2O_3 join possesses the characteristic blue-grey hue. The presence of diffuse light-grey transition zones and mottlings within the brownish titanomagnetite grains probably indicates partially oxidized portions (low-grade titanomaghemite).

Titanomaghemite is a chemically metastable phase which tends to change to hematite plus rutile. It may be possible by this transformation to form essentially non-magnetic lamellae and walls, thereby subdividing large grains into small magnetically stable grains. Similar results could be obtained by initial formation of hematite by low-temperature oxidation. Stable components of magnetization could then be formed after (perhaps long after) the bulk TRM was acquired and may bear no relation to the TRM direction. As an example, Akimoto & Kushiro (1960) heated several samples of highly unstable dolerite (samples 37, 38 and 39) which contained large amounts of titanomaghemite. After heating them to 600°C , they cooled them in the Earth's field. The resultant TRM was extremely stable as is shown in Table 1. Fig. 22 shows the coercivity spectrum before heating for

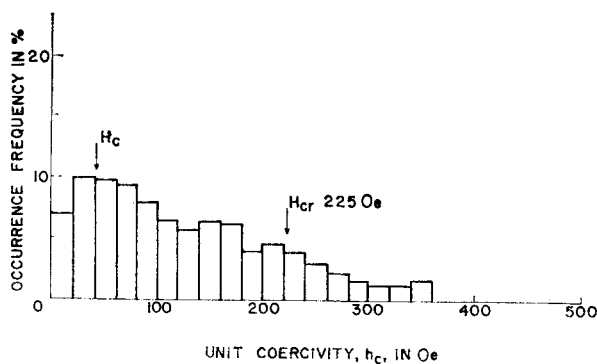


FIG. 22. Coercivity spectrum before heating. Sample 38—dolerite (Sidara II).

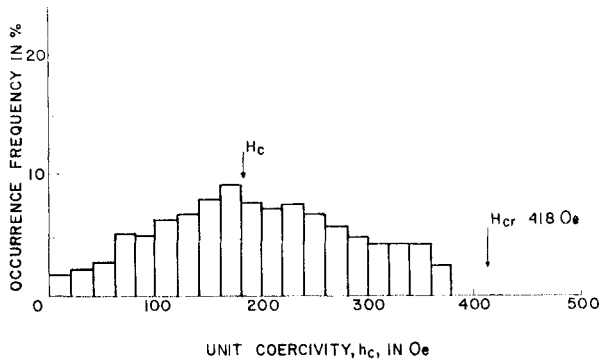


FIG. 23. Coercivity spectrum after heating Sample 38 to 600 °C.

sample 38 and Fig. 23 the spectrum after heating. Microscopic observation showed that some of the titanomagnetite had broken down to hematite thereby producing numerous tiny lamellae walls and effectively increasing the number of stable small grains.

Conclusions

(1) Our results clearly indicate a close relation between the magnetic stability of a rock and the effective grain size of its constituent titanomagnetite as viewed under the microscope. Rocks containing mainly large grains in excess of 10μ are magnetically soft, whereas rocks containing grains mostly less than 10μ behave in a hard manner. Rocks containing portions of both size fractions contain both soft and hard components. Lavas and welded tuffs generally contain larger abundances of small grains than do plutonic rocks and, on the whole, contain larger components of stable magnetization.

(2) The effective grain size of the magnetic opaques in a rock is dependent on (1) the initial conditions during crystallization (primarily slowness of cooling) and (2) on post-consolidation oxidational changes. High-temperature oxidation of titanomagnetite commonly results in the development of ilmenite lamellae which effectively subdivide the initial magnetic crystals into a great many smaller, magnetically more stable grains.

(3) Of all stability tests, the coercivity spectrum appears to provide the most useful representation of the range of stability possessed by any sample. It is particularly suited to the study of rocks, which possess a wide variation of effective grain size and titanomagnetite composition.

(4) The consistency of the relation between stability of magnetization and effective grain size visible under the microscope seems to indicate that the carriers of the stable components of remanent magnetization reside in actual small grains rather than in hypothetical submicroscopic strain regions which act as single-domain grains (Verhoogen 1959). It has been suggested that the slow growth and thorough annealing of the opaques in granitic rocks accounts for their instability. Yet, as experiments have shown granitic rocks possessing stable components contained greater numbers of small grains than those showing little stability. Furthermore, in a sequence of basalt flows which presumably were erupted under similar conditions, those flows possessing unstable magnetization contain very few small grains. In the case of sample 13 it was possible to produce stable magnetization by a heating process which effectively reduced the initial grain size.

3. J_s - T curves and rock stability

Determination of the variation of saturation magnetization with temperature change has long been standard procedure in rock magnetic and palaeomagnetic studies. However, because of the complexity of the opaque mineralogy of most rocks, the relation between a J_s - T (saturation magnetization versus temperature) curve and mineralogy is not always clear. Studies by Kawai *et al.* (1954, 1955) and Ade-Hall *et al.* (1965) have provided partial solutions.

As part of the present study on magnetic stability we have determined a J_s - T curve for each of the 45 samples. When these curves are correlated with the microscopic studies, it is generally possible to interpret the J_s - T curves in terms of mineralogy.

Single Curie point samples

Without exception those rocks possessing a single Curie point have Curie temperatures above 515°C, and in most cases, above 550°C (similar to the results of Ade-Hall *et al.* 1965). A typical J_s - T curve is shown in Fig. 24; its squarish shape indicates a single magnetic phase with only a small range in composition. As revealed by microscopic observation (see Table 1), rocks with a single, high Curie point were either acidic in composition or highly oxidized.

The Curie point of titanomagnetite varies with the degree of low-temperature oxidation and the amount of contained titanium, as shown in Fig. 21. As explained in Section 2, low-temperature oxidation refers to the single-phase alteration of titanomagnetite to titanomaghemite. As oxidation proceeds, the Curie temperature progressively increases (Akimoto *et al.* 1957).

The Curie point of titanomagnetite also varies with the content of Ti; it is low (100–200°C) for titanium-rich grains and close to 580°C for nearly pure magnetite. As discussed in Section 2, the amount of titanium contained in an opaque is partly a function of the consolidation temperature of the magma and of subsequent high-temperature oxidation. Acidic rocks crystallize at temperatures between 500–600°C, and little Ti is contained in the spinel; high Curie points are the result. If the rocks crystallize at higher temperatures (for example, basalt), large amounts of titanium are generally incorporated in the magnetite structure. During the early, high-temperature phases of cooling, if oxidizing conditions prevail, most of the ulvospinel component will be oxidized to ilmenite. By this process, a progressive purification of the magnetite occurs and the Curie point increases commensurately. Therefore, if an opaque relatively rich in Ti undergoes either high or low-temperature oxidation, its Curie point will increase.

Since the development of ilmenite lamellae promotes increased magnetic stability, rocks possessing high Curie points as a result of high-temperature oxidation and concomitant exsolution are generally stable. This is particularly noticeable for the basaltic rocks in Table 1. The stability of acidic rocks, however, which possess a high Curie point as a result of initial crystallization or basic rocks which contain few lamellae and have a high Curie point as a result of low-temperature oxidation, will be determined by the initial grain size. Rhyolitic welded tuff, rapidly cooled rhyolite flows, or chilled intrusive contacts may display high stability, whereas slowly cooled rocks containing large crystals would be highly unstable.

Multiple Curie point samples

Several samples showed multiple Curie points; one between 200°C and 350°C, and one or more above 500°C (Fig. 25). Results similar to these were also obtained by Ade-Hall *et al.* (1965). Rocks with multiple Curie points are generally basic, i.e. basalt, dolerite or andesite, in which the titanomagnetite has undergone little

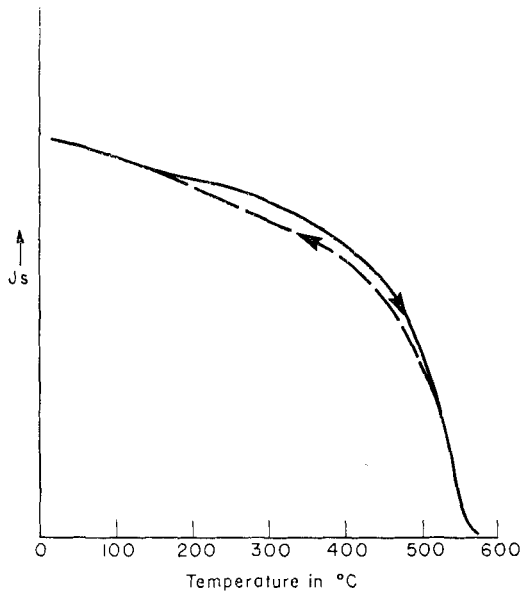


FIG. 24. J_s - T curve for basaltic Sample 7.

high-temperature oxidation and, therefore, possess few ilmenite exsolution lamellae or little hematite (see Table 1).

The low Curie points reflect relatively high Ti-Fe ratios which are probably close to those resulting from primary crystallization of a basic rock. As shown by Buddington & Lindsley (1964), if TiO_2 was plentiful in the magma, the mole per cent of ulvospinel in basaltic titanomagnetite should be 55 to 70—an amount that would result in a Curie temperature of 100–250°C. Samples with Curie points between 250° and 350°C, most probably have undergone partial low-temperature oxidation in which the original titanomagnetite has been altered to a low-grade titanomaghemite.

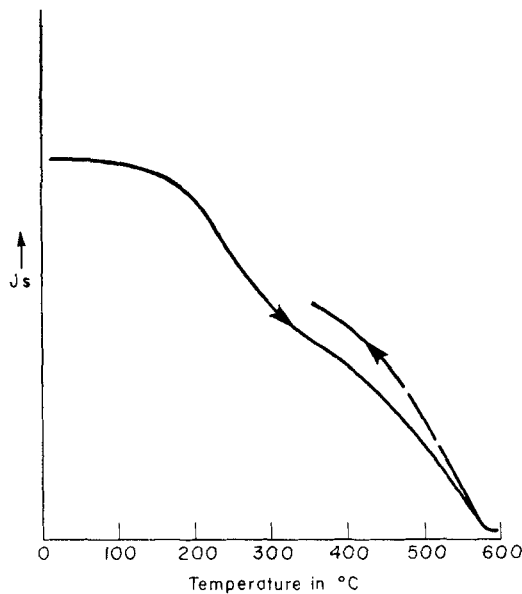


FIG. 25. J_s - T curve showing two distinct Curie temperatures—basaltic Sample 21.

In Fig. 21 is plotted the compositional point of a titanomagnetite with 65 mole per cent ulvospinel—point \times —this composition corresponds to that typical of a basalt. As low-temperature oxidation progresses, the composition moves, along an oxidation line, off the $\text{Fe}_2\text{TiO}_4\text{--Fe}_3\text{O}_4$ join, toward the $\text{FeTiO}_3\text{--Fe}_2\text{O}_3$ join and the Curie point increases. Partial oxidation to point $+$ would account for the observed Curie point of about 350°C .

Microscopic observation reveals that low-temperature oxidation has affected all low Curie point samples to some extent—all samples contain some (up to 25 per cent) high-grade maghemite. Moreover, the lower Curie points are never sharp; rather, the transition is a gentle, broad deflection, indicative of a range in compositions and probably due to variation in the degree of oxidation.

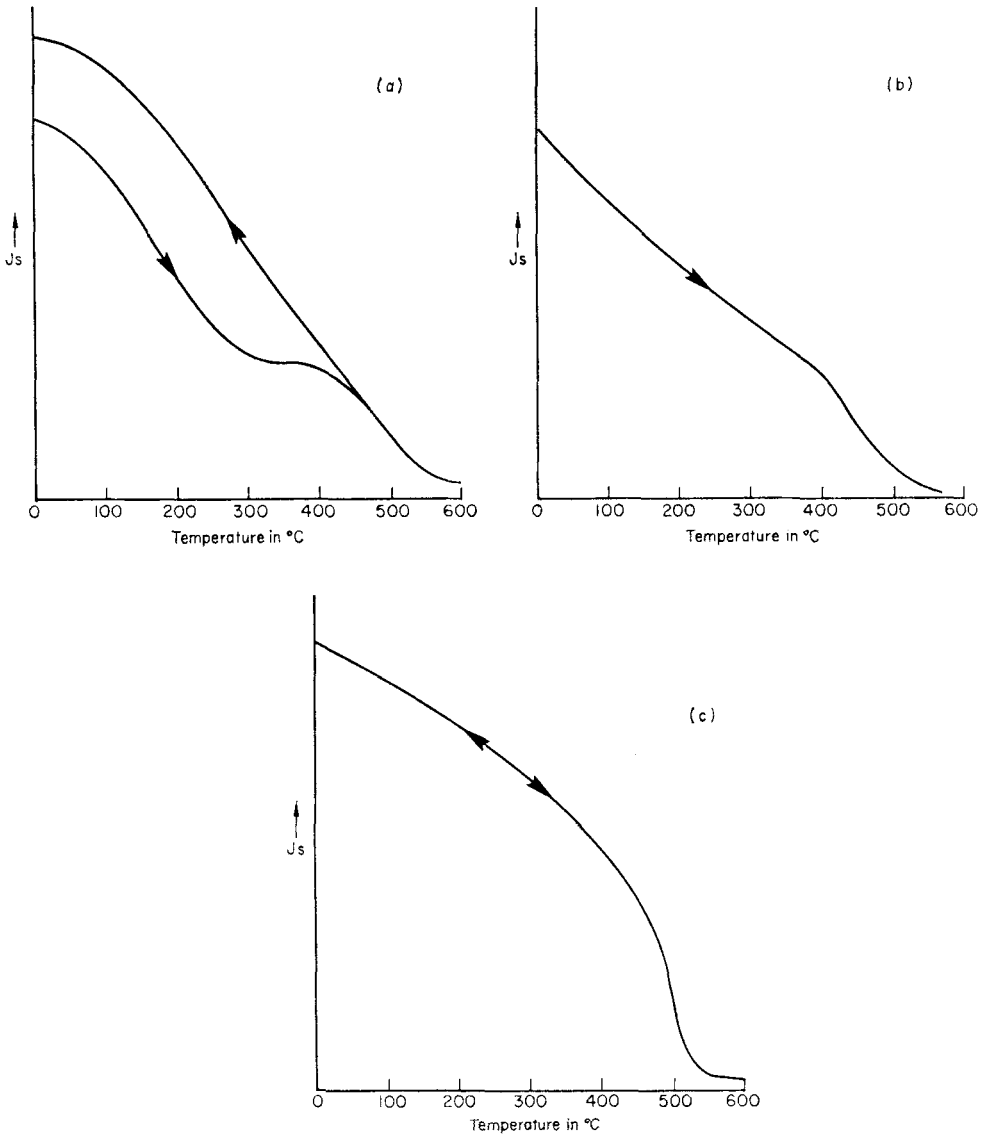


FIG. 26. Increase in high Curie-point phase by progressive low-temperature oxidation. (a) Sample 24, basalt. (b) Sample 15, basalt. (c) Sample 38, dolerite intrusive.

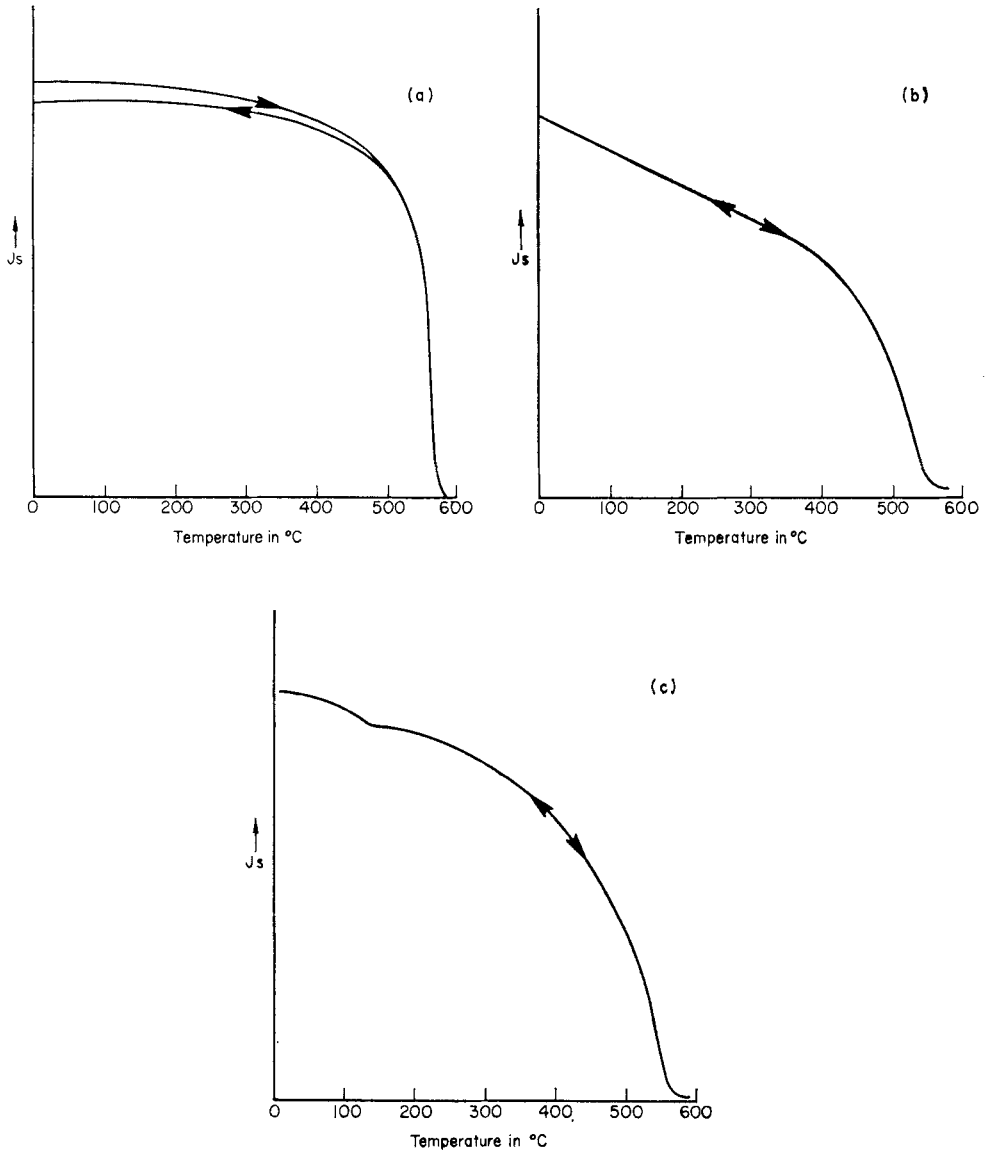


FIG. 27. Reversible J_s - T curves of different rock types. (a) Sample 30, granite. (b) Sample 23, basalt. (c) Sample 37, dolerite intrusive.

Where little low-temperature oxidation has occurred, the low Curie point deflection is more pronounced (Fig. 26(a)); with increased oxidation the deflection becomes more obscure (Fig. 26(b)). Where essentially complete oxidation to titanomaghemite has occurred (Fig. 26(c)), the curve assumes the blocky shape characteristic of a rock with magnetic opaque minerals of a single composition.

The higher Curie points of the multiple Curie point rocks seem associated with components of nearly pure magnetite or high-grade titanomaghemite, resulting from the production of ilmenite lamellae by high-temperature oxidation or the production of titanomaghemite by low-temperature oxidation, respectively. In some samples, microscopic studies indicate that both high Curie point phases are present; in many, only high-grade titanomaghemite is visible.

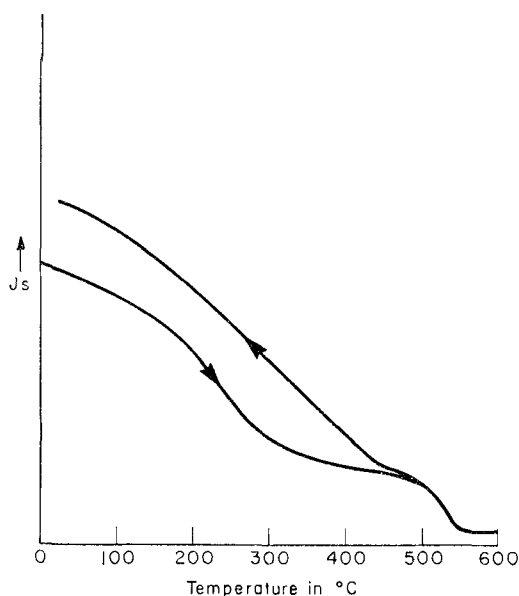


FIG. 28. Irreversible upshift of thermomagnetic curve. Sample 24, basalt.

Several J_s - T curves possess a small tail extending to 600°C and beyond (see Fig. 26). The component responsible for this tail does not seem to be hematite since little of this phase is microscopically detectable. Moreover, even in samples containing abundant hematite, no such high-temperature tail exists. In every case, its appearance is coincident with the presence of high-grade maghemite.

Reversible samples

The J_s - T curves of many specimens show little difference between their heating and cooling curves. In most cases this behaviour is associated with blocky curves and single Curie point specimens with Curie points in excess of 500°C; that is, the rocks contain (1) nearly pure magnetite—as a result of initial crystallization (acidic rocks) or high-temperature oxidation and exsolution (basic rocks)—or (2) nearly pure titanomagemite (see Fig. 27).

The fact that titanomagemite remains stable at elevated temperatures is noteworthy since the transition temperature is usually quoted as about 275°C (Nagata 1961). The stability of the γ phase is now believed to be dependent largely upon the amount of titanium in the structure (Kushiro, personal communication). If little Ti is contained, the transition temperature is near 275°C; if large amounts are present, the mineral can be stable up to 800°C.

The fact that the samples containing titanium-deficient titanomagnetite showed little or no downshift of the J_s - T curves on cooling indicates clearly that no maghemite was present in those samples.

Irreversible changes

During the course of heating a specimen above 600°C, the magnetic minerals commonly undergo chemical changes—notably, decomposition of metastable phases, oxidation, and reduction—so that the cooling curve does not retrace the heating curve. In some cases, variation in the thermomagnetic curve cannot be attributed to any of the above causes. Changes in the internal crystal structure such

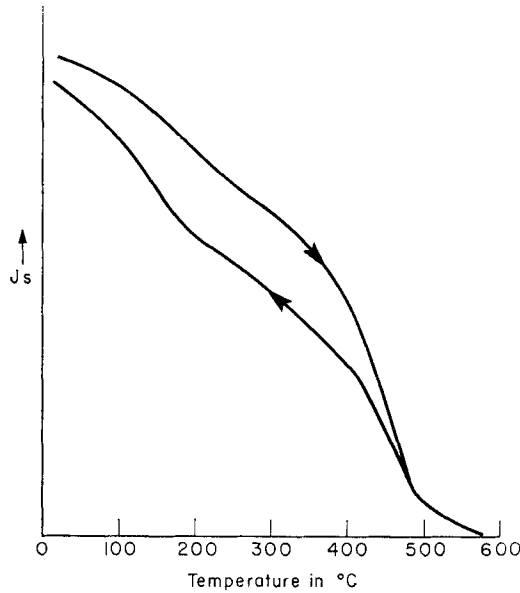


FIG. 29. Irreversible downshift of thermomagnetic curve. Sample 14, basalt.

as separation of unlike phases, ordering of the lattices, or variation of electronic configuration, are possible causes.

Decomposition to a stable phase occurs whenever the temperature exceeds the transition temperature of the metastable mineral. Oxidation and reduction, however, are processes dependent upon the nature of the atmosphere surrounding the sample during heating. In air, essentially all samples are subject to oxidation. However, in an evacuated tube the $P(\text{O}_2)$ is so limited that the oxidation potential is probably determined primarily by the nature and amount of the volatile components within the rock sample. In some cases, the environment can even become slightly reducing. Because the internal state of the opaques and the nature of the microenvironment around the sample are not well known during the experiment, definite causes cannot always be assigned to irreversible changes in the J_s - T curve.

Large-scale irreversible changes—representing a marked variation in saturation magnetization—are restricted to samples containing a low Curie point constituent; that is, to samples containing titanium-rich titanomagnetite or low-grade titaniferous magnetite.

Upshifts of the low Curie point portion of the J_s - T curves (as shown in Fig. 28) are probably the result of low-temperature oxidation, but the possibility of ilmenite lamellae formation or internal ordering of the opaques cannot be dismissed. In most cases, the cooling curve is displaced until it is tangent to that of the high Curie point titanomagnetite toe on the heating curve. This argues in favour of titanomagnetite formation.

In some cases the cooling curve is downshifted with respect to the heating curve (see Fig. 29). Any one of several reactions could account for this behaviour: decomposition of a metastable phase, reduction of a partially oxidized phase, and internal homogenization or ordering by diffusion processes. It is generally difficult to single out any one of these causes. In most cases, the most plausible explanation is decomposition of some component. In acidic rocks, which contain little titanium, temperatures reached during heating to 600°C should certainly exceed the transition

temperature of a maghemite component, and it should alter to the antiferromagnetic stable phase—hematite (see Ozima & Larson 1967). The slight lowering of the cooling curves of some granite samples may be due to this cause, or some small grains of magnetite may be directly oxidized to hematite.

In the case of basic rocks, however, the titanomaghemite may or may not decompose at temperatures achieved during the experiment. As noted earlier, high-grade, titanium-rich maghemite can be stable up to 800°C. But the stability may also be dependent on other, less recognizable factors and break-down of even richly titaniferous members may occur at temperatures below 610°C (reviewed in Nagata (1961)).

Conclusions

(1) Blocky curves with one high Curie point (over 500°C) are characteristic of samples containing nearly pure magnetite or high-grade titanomaghemite only. The titanium-deficient magnetite can result either from original crystallization in acidic rocks or from high-temperature auto-oxidation and ilmenite production in basic rocks. Basalts and other basic rocks which possess high Curie points because of the presence of lamellae are magnetically stable. In contrast, the stability of acidic rocks is determined only by initial crystallization. A fine-grained rhyolite or welded tuff will probably be highly magnetically stable whereas a coarse-grained granitic rock probably will not.

Rocks mainly containing high-grade titanomaghemite will normally also display a blocky curve and a high Curie point. Oxidation of titanomagnetite to titanomaghemite is a low-temperature process which causes progressive elevation of the Curie temperature but does not generally disturb the single-phase homogeneity of the spinel structure. Therefore, rocks containing large homogeneous grains of titanomaghemite will probably be magnetically unstable.

(2) Two or more Curie points are generally associated with basic rocks which have been little affected by high-temperature oxidation and which are, therefore, generally magnetically unstable. The low Curie point corresponds to titanium-rich titanomagnetite or low-grade titanomaghemite whereas the higher Curie points correspond either to high-grade titanomaghemite or magnetite (resulting from high-temperature development of ilmenite) in some grains—or both.

(3) Samples containing only titanium-poor magnetite (single high Curie point samples) showed little or no difference between their heating and cooling curves, indicating complete chemical stability of their opaques over the range of experimental conditions.

Samples containing high-grade titanium-rich titanomaghemite also proved to be chemically stable and had reversible J_s - T curves. This supports the idea that titanium-rich titanomaghemite can possess a transition temperature above 600°C.

It is evident that basic rocks with chemically stable, nearly pure magnetite as a result of high-temperature (111) lamellae growth will probably be magnetically stable as well, whereas, acid rocks with chemically stable magnetite and basic rocks with chemically stable titanium-rich titanomaghemite will possess homogeneous opaques, and their magnetic stability will depend upon initial grain size.

(4) Large-scale irreversible changes, either up or down, were associated with basic rocks which showed the effects of high-temperature oxidation only slightly, if at all. Rocks of this nature should generally behave unstably.

Acknowledgments

This work was done as part of the U.S.–Japan Scientific Co-operation administered by the Japan Society for the Promotion of Science and by the National Science Foundation under Grant No. GF-206.

E. Larson and D. Strangway:
*Department of Geology and Geophysics,
 Massachusetts Institute of Technology,
 Cambridge,
 Massachusetts.*

Mituko Ozima, Minoru Ozima and
 T. Nagata:
*Geophysical Institute,
 University of Tokyo,
 Japan.*

1968 November.

References

- Ade-Hall, J. M., Wilson, R. L. & Smith, P. J., 1964. The petrology, Curie points and natural magnetizations of basic lavas, *Geophys. J. R. astr. Soc.*, **9**, 323–336.
- Akimoto, S., Katsura, T. & Yoshida, M., 1957. Magnetic properties of TiFe_2O_4 – Fe_3O_4 system and their change with oxidation, *J. Geomagn. Geoelect.*, **9**, 165.
- Akimoto, S. & Kushiro, I., 1960. Natural occurrence of titanomaghemite and its relevance to the unstable magnetization of rocks, *J. Geomagn. Geoelect.*, **11**, 94–110.
- Buddington, A. F. & Lindsley, D. H., 1964. Iron–titanium oxide minerals and synthetic equivalents, *J. Petrol.*, **5**, 310–357.
- Everitt, C. W. F., 1962. Thermoremanent magnetization—II. Experiments on multidomain grains, *Phil. Mag.*, **7**, 583–597.
- Graham, J. W., 1953. Changes in ferromagnetic minerals and their bearing on magnetic properties of rocks, *J. geophys. Res.*, **58**, 243–260.
- Katsura, T. & Kushiro, I., 1961. Titanomaghemite in igneous rocks, *Am. Miner.*, **16**, 134–145.
- Kawai, N., 1955. Magnetism of rocks and solid phase transformation in ferromagnetic minerals—III, *Proc. Japan Acad.*, **31**, 346.
- Kawai, N., Kume, S. & Sasajima, S., 1954. Magnetism of rocks and solid phase transformation in ferromagnetic minerals—I and II, *Proc. Japan Acad.*, **30**, 588, 864.
- Kittel, C., 1956. *Solid State Physics*, p. 511, Academic Press, New York.
- Morrish, A. H. & Yu, S. P., 1955. Dependence of the coercive force on the density of some oxide powders., *J. appl. Phys.*, **26**, 1049.
- Nagata, T., 1961. *Rock Magnetism*, Maruzen, Tokyo.
- Ozima, M. & Ozima, M., 1965. Origin of thermoremanent magnetization, *J. geophys. Res.*, **70**, 1363–1369.
- Ozima, M., Yama-ai, M. & Nagata, T., 1963. Role of crystalline anisotropy energy on the acquisition of stable remanent magnetization: inverse type of thermoremanent magnetization, *Geofis. pura Appl.*, **55**, 77–90.
- Ozima, M. & Larson, E. E., 1967. Study on irreversible change of magnetic properties of some ferromagnetic minerals, *J. Geomagn. Geoelect.*, **19**, 117–127.
- Patton, B. J. & Fitch, J. L., 1962. Anhysteretic magnetization in small steady field, *J. geophys. Res.*, **67**, 307–311.
- Powell, D. W., 1963. Significance of differences in magnetization along certain dolerite dikes, *Nature, Lond.*, **199**, 674–676.

- Stacey, F. D., 1962. Generalized theory of thermoremanence covering the transition from single domain to multidomain magnetic grains, *Phil. Mag.*, **7**, 1887–1900.
- Verhoogen, J., 1959. Origin of thermoremanent magnetism, *J. geophys. Res.*, **64**, 2441–2449.
- Verhoogen, J., 1962. Oxidation of iron–titanium oxides in igneous rocks, *J. Geol.*, **70**, 168–181.
- Wilson, R. L. & Haggerty, S. E., 1966. Reversals of the Earth's magnetic field, *Endeavour*, **25**, 104–109.