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2. *Aspects of Rock Magnetism — General*

Thermal Enhancement of Magnetic Susceptibility

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Abstract. The interpretation of geomagnetic anomalies with deep-seated sources sometimes requires postulating magnetic susceptibilities larger than those measured for common rock types at the earth's surface. A possible explanation is that rocks buried at depths approaching the Curie point isotherm exhibit enhanced susceptibility due to the Hopkinson effect. In measurements on a sample of single-domain magnetite (0.04 μm particles), the susceptibility increased by a factor 2 between 20 and 500 °C and by a factor 3 at 550 °C. The Hopkinson peak was less pronounced in multidomain magnetites: the relative increase in susceptibility at 550 °C was by a factor 2 in 0.1 μm particles and a factor 1.5 in 0.25 μm particles. Single-domain hematite (0.1—1 μm) gave a spectacular Hopkinson peak, with relative susceptibility enhancement by a factor 5 at 530 °C and a factor 20 at 640 °C. However, rocks containing fine-grained maghemite and magnetite showed an enhancement of 50—70 % at most. The reasons for this variability in the height of the Hopkinson peak are not understood, but the width and shape of the peak are clearly related to the blocking temperature spectrum. Distributed blocking temperatures are associated with a broad peak, while discrete blocking temperatures are accompanied by a sharp susceptibility peak within 50—100 °C of the Curie point. A corollary is that remanent magnetization decreases roughly in inverse proportion to increase in susceptibility, so that the Koenigsberger Q_n ratio decreases sharply at high temperature. For this reason, deep-seated anomalies can almost certainly be interpreted in terms of induced magnetization only. Finally, somewhat shallower bodies (temperatures of 200—400 °C) may exhibit thermally enhanced magnetization for two reasons: first, titanomagnetites have widely varying Curie points depending on titanium content, and second, observed anomalies are the result of a geomagnetic field applied over 10^6 years and viscous magnetization is also known to be enhanced at high temperature.

Key words: Magnetic Anomalies — Susceptibility — Induced Magnetization — Remanent Magnetization — Hopkinson Effect — Curie Point Isotherm.

Introduction

Relatively intense regional magnetic anomalies whose sources are apparently 10—30 km deep in the crust are widespread in the Precambrian Shields of the world. The effective magnetizations of the sources as deduced by Bhattacharya and Morley (1965), Hall (1968), McGrath and Hall (1969) and Krutikhovskaya *et al.* (1973), among others, correspond to susceptibilities of $10^{-3} - 10^{-2}$ emu/cm^3. Since the susceptibilities of the surface rocks in shield areas are commonly $10^{-4} - 10^{-3}$ emu/cm^3, one is forced to conclude that either remanent magnetization is responsible for most regional anomalies or else that susceptibility is somehow enhanced at depth in the crust. The latter possibility is explored in this paper.

Susceptibility and Domain Structure

The magnetization induced by a given field H depends on whether the domain structure must be rotated, as in single-domain (SD) grains, or whether domain wall displacements are also possible, as in multidomain (MD) grains. The theoretical initial susceptibility χ_0 of unaxial, randomly oriented SD grains is (Stoner and Wohlfarth, 1948; Dunlop, 1969)

$$\chi_0 = 0.349 \frac{M_S}{H_R} \tag{1}$$

where the remanent coercive force H_R (the reverse field required to reduce to zero the remanence following saturation in the forward direction) essentially measures the resistance of anisotropy forces to rotation of the spontaneous magnetization M_S. The intrinsic susceptibility χ_i of MD grains is likewise inversely proportional to coercive force (in this case, the resistance to wall motion) but the self-demagnetizing field reduces the observed susceptibility to (Neel, 1955; Nagata, 1961; Stacey, 1963)

$$\chi_0 = \frac{\chi_i}{1+N\chi_i} \approx \frac{1}{N} \tag{2}$$

where N is the demagnetizing factor, determined by the shape of the grain. Both Nagata (1961) and Stacey and Banerjee (1974) suggest an average value $N=3.8$–3.9 for titanomagnetite grains in igneous rocks, very little different from the value $4\pi/3$ appropriate to equidimensional grains.

Fig. 1 compares measured values of χ_0 for single-domain and two-domain grains of magnetite and other strongly magnetic minerals with Eqs. (1) and (2). It is interesting that while some of the SD samples have extreme M_S/H_R values seldom, if ever, encountered in nature, the resulting range of χ_0 is only 0.1–0.6 emu/cm^3. χ_0 for SD magnetite is thus comparable to χ_0 for MD magnetite. The great contrast between the Koenigsberger ratios of SD and MD magnetite reflects the more intense TRM (thermoremanent magnetization) of SD grains, not a susceptibility contrast.

Although two-domain grains display some SD-like features (Dunlop, 1974), Fig. 1 shows that their susceptibilities are greatly reduced by self-demagnetization. Grains with higher domain multiplicities have χ_0 essentially independent of H_R, as predicted by Eq. (2).

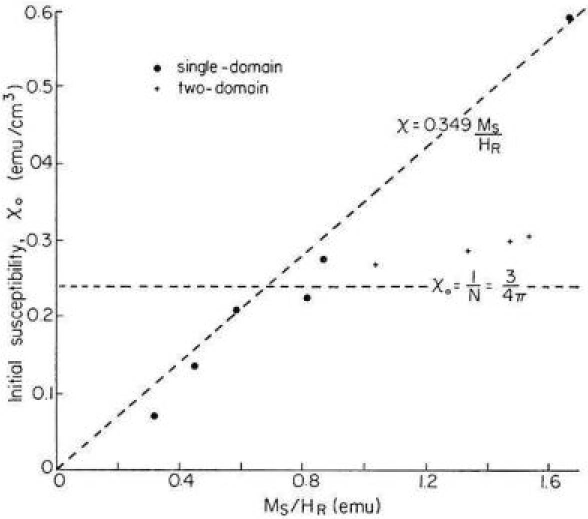

Fig. 1. Initial susceptibilities of some natural and synthetic samples of magnetite and maghemite, as a function of the ratio M_S/H_R. The theoretically expected dependences for SD and MD materials (Eqs. 1 and 2) are shown as dashed lines. SD materials agree well with theory, while two-domain materials are intermediate between SD and MD behaviour. (After Dunlop, 1974)

Above a critical blocking temperature T_B, both SD and MD grains become superparamagnetic (SPM): that is, barriers to domain rotation or wall motion are overcome by thermal energy and very large susceptibilities can result. For SD grains (Bean and Livingston, 1959),

$$\chi_0 = \frac{M_S}{H} L\left(\frac{vM_SH}{kT}\right) \tag{3}$$

where v is grain volume, k is Boltzmann's constant and T is temperature. The Langevin function $L(\alpha)$ equals α for $\alpha \lesssim 0.2$ and equals 1 for large α.

Since the blocking temperature for MD grains is invariably close to the Curie point, M_S is rather small throughout the blocking region and SPM enhancement of χ_0 is not as significant as in the SD case.

Susceptibility of Crustal Rocks

Table 1 lists measured values of χ_0 for a variety of crustal igneous and metamorphic rocks ranging in age from Archean to Tertiary. The rocks were originally measured because of their paleomagnetic interest and by no means comprise a representative sampling of the continental crust.

Table 1. Susceptibilities of some crustal rocks. Both the observed susceptibility and the calculated value corresponding to 100% magnetite are tabulated. The numbers (I through VI) accompanying the Steens and Iceland samples refer to the oxidation index (Wilson and Watkins, 1967) of these basalts

Rock type	Volume % Fe_3O_4, p	χ_0 as observed (10^{-4}emu/cm^3)	χ_0 for $p = 100\%$ (emu/cm^3)
Tertiary basalts			
1) Cobb seamount[a]	0.41	8.55	0.210
2) Cape Dyer basalts[b]	<0.01	0.18	0.295
	0.02	0.44	0.184
	0.03	0.86	0.324
	0.11	1.96	0.179
3) Steens basalts I[c]	0.18	6.02	0.344
II	0.18	5.27	0.291
III	0.17	6.39	0.377
IV	0.23	4.85	0.211
V	0.18	3.88	0.221
4) Iceland basalts I[d]	0.22	10.0	0.455
II	0.27	11.8	0.436
III	0.26	5.97	0.234
IV	0.29	5.98	0.204
V	0.23	4.69	0.201
VI	0.23	4.95	0.214
Archean metabasalts			
1) Kirkland Lake	<0.01	0.14	0.185
greenstones[e]	0.01	0.26	0.202
	0.08	2.86	0.344
Diabases			
1) Matachewan dike[e]	0.01	0.23	0.216
2) Logan sills[f]	0.03	0.68	0.237
3) Glamorgan dike[g]	0.02	0.56	0.276
4) Coronation sills[h]	0.27	6.83	0.258
	0.44	9.45	0.214
5) Tasmanian dolerite[i]	0.11	2.55	0.244
6) Lambertville diabase[j]	0.12	3.44	0.287
Gabbros, anorthosites			
1) Modipe gabbro[k]	0.02	0.41	0.255
2) Sudbury norite[l]	0.02	0.39	0.204
3) Michikamau anorthosite[m]	0.25	9.78	0.387
4) Michael gabbro[n]	0.44	14.7	0.332
5) Glamorgan gabbro[g, o]	0.01	0.18	0.206
	0.06	2.13	0.335
Diorites, granites			
1) Kirkland Lake diorite[e]	0.02	0.30	0.189
2) Dudmon diorite[g, o]	0.79	33.8	0.428
3) Bark Lake granodiorite[g, o]	0.24	10.3	0.428
4) Spavinaw granite[p]	0.33	8.50	0.259

Nevertheless several significant trends appear in the data. First of all, since the volume percent of magnetic material was known from saturation magnetization measurements, it was possible to find the value of χ_0 corresponding to 100 % magnetite for each rock. Despite the variety of rock types, these all lay between 0.18 and 0.46 emu/cm^3, a range entirely consistent with the theory of the previous section and the data of Fig. 1.

A second point is that all the rock types examined, even those generally considered to be very magnetic (eg. Tertiary continental and submarine basalts), contained less than 1 % by volume magnetite. As a result, the observed susceptibilities (with one or two exceptions) were 10^{-3} emu/cm^3 or less. It is interesting that the Spavinaw granite and the Bark Lake granodiorite (both of Precambrian age) each contain about 0.3 % magnetite and are as strongly magnetic as most of the more mafic rocks. Although these two rocks are probably atypical, they do demonstrate that felsic rocks cannot be dismissed as possible sources of regional anomalies.

Finally, Table 1 shows that despite great variability in observed susceptibilities, χ_0 tends to be lower for older, more metamorphosed rocks. For example, the Kirkland Lake greenstones contain about one-tenth the magnetic material found in Tertiary volcanic equivalents. Precambrian diabases like the Matachewan and Glamorgan dikes and the Logan sills are 5—10 times less magnetic than the Phanerozoic Tasmanian dolerite and Lambertville diabase. Bearing this trend in mind, we have even more difficulty in reconciling the effective magnetizations of deep magnetic anomaly sources in Precambrian terrain with the susceptibilities typical of surface rocks in these areas.

Remanent and Induced Magnetization

There are two ways of resolving the paradox that typical crustal rocks seem to have susceptibilities that are insufficient to account for regional magnetic anomalies. One way is to appeal instead to the TRM of rocks contaning SD or small MD grains of titanomagnetite. These rocks can readily supply the requisite magnetization, the "enhancement factor" being the Koenigsberger ratio Q_t between TRM and induced magnetization produced by the earth's field. Stacey (1967) has shown that Q_t is much

←

[a] Merrill and Burns, 1972; [b] Deutsch, Kristjansson and May, 1971; [c] Watkins, 1969; [d] Watkins and Haggerty, 1968; [e] Pesonen, 1973; [f] Robertson and Fahrig, 1971; [g] Dunlop, Hanes and Buchan, 1973; [h] Fahrig, Irving and Jackson, 1971; [i] Stott and Stacey, 1960; [j] Hargraves and Young, 1969; [k] Evans and McElhinny, 1966; [l] Hood, 1961; [m] Murthy, Evans and Gough, 1971; [n] Fahrig and Larochelle, 1972; [o] Buchan and Dunlop, 1973; [p] Spall and Noltimier, 1973.

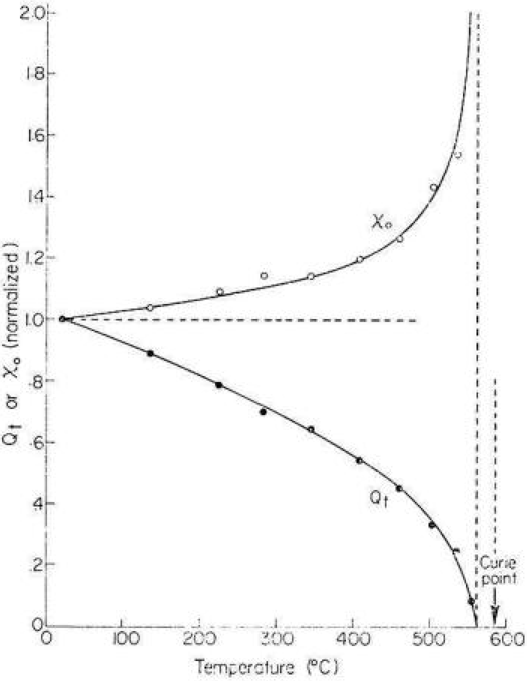

Fig. 2. The normalized temperature dependence of initial susceptibility χ_0 and Koenigsberger ratio Q_t for a typical submicron magnetite. Between the Curie point and the blocking temperature (the dashed line about 30 °C below the Curie point), Q_t is zero but χ_0 is considerably enhanced

greater than 1 for magnetite grains smaller than a few microns but less than 1 for larger grains.

Appealing to TRM as the source of anomalies has of course the unfortunate consequence that the magnetization direction is unrelated to that of the present field, complicating the interpretation process. The alternative is to suppose that induced magnetization is enhanced deep in the crust. Susceptibility changes significantly with hydrostatic pressure (Nagata, 1961, Girdler, 1963) but the increase of susceptibility with temperature (the Hopkinson (1889) effect) is a more important effect if the rocks are buried close to the depth of the Curie point isotherm. Fig. 2 illustrates the effect. The enhancement is generally significant only within 100 °C or so of the Curie temperature.

There is at present no evidence so compelling that we can choose one explanation over the other. However it is worth noting that since regional anomalies are very widespread in Precambrian Shields, it is only natural to favour as sources rock types that are similarly widespread. Obvious candidates would be the deep equivalents of the granitic rocks which are ubiquitous at the surface. Some authors (e.g. Krutikhovskaya *et al.*, 1973) favour very large basic intrusives as sources. In either case, the rocks are coarse-grained and would be expected to exhibit generally low Q_t values. Furthermore, as Fig. 2 illustrates, Q_t decreases steadily as the temperature increases and is zero throughout the region between the blocking temperature and the Curie point. This is of course the same region in which susceptibility is maximum because of superparamagnetism. Thus even SD grains behave like low-Q_t materials at high temperature.

On the other hand, it is not clear that susceptibility enhancement by the Hopkinson effect is sufficient or occurs over a sufficiently wide range of temperature to significantly enhance the magnetization of deeply buried rocks. We shall examine these questions and the physical factors controlling the Hopkinson effect in the next section.

The Hopkinson Effect in Rocks and Minerals

In SD grains and small MD grains with pseudo-SD magnetic moments (Dunlop, 1973a), the susceptibility increases with temperature for two reasons. Below the blocking temperature T_B of a particular grain, Eq. (1) predicts that χ_0 is proportional to M_S/H_R. Above T_B, χ_0 jumps to the value given by (3), eventually falling to zero with M_S at the Curie point. The grains in a rock have a spectrum of sizes and hence of T_B, with the result that both effects contribute to χ_0 of the rock in the blocking region.

The height of the Hopkinson peak, which determines the maximum enhancement of susceptibility, is controlled both by H_R at room temperature, T_0, and by the minimum blocking temperature. The larger $H_R(T_0)$ and the higher $(T_B)_{\min}$, the larger the enhancement factor $H_R(T_0)/H_R(T_B)$. On the other hand, if $(T_B)_{\min}$ is too high, M_S is small and so is the SPM susceptibility. The width of the Hopkinson peak, which determines over what range of temperatures (and by extension, over what range of crustal depths) there is significant enhancement of χ_0, is controlled almost entirely by the width of the blocking temperature spectrum.

Large MD grains invariably have a narrow T_B spectrum just below the Curie point, and according to Eq. (2), there is no susceptibility increase below $(T_B)_{\min}$. We can expect therefore a rather narrow and not particularly high Hopkinson peak in MD materials.

Figs. 3, 4 and 5 illustrate these principles. Fig. 3 compares χ_0 and M_S/H_R data and the blocking temperature spectrum (determined from stepwise thermal demagnetization of a 1-oersted TRM) for two submicron magnetites described by Dunlop (1973a, b). The small MD grains (mean size about 0.1 μm) have a narrow range of T_B just below the Curie temperature of 583 °C, and except in this range, χ_0 and M_S/H_R have identical

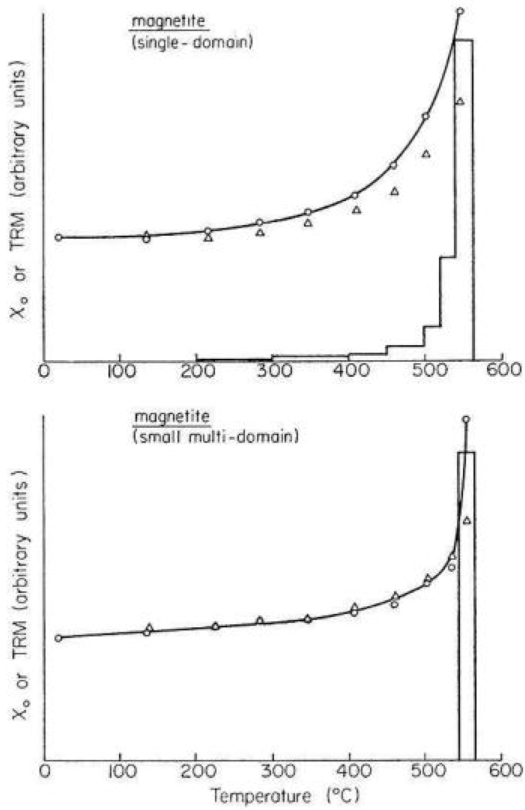

Fig. 3. A comparison of the temperature dependences of χ_0 (circles) and M_S/H_R (triangles) and the distribution of blocking temperatures (the derivative spectrum of TRM) for two fine-grained magnetites. The samples are numbers 2 and 4 of Dunlop, 1973a, b. χ_0 and M_S/H_R have similar temperature variations, as expected theoretically, and deviations between the two functions are closely correlated with the blocking-temperature spectrum

temperature dependences. The Hopkinson peak is narrow but relatively high (enhancement factor about 3). The SD grains (mean size about 0.04 μm) have a broad range of T_B and a correspondingly broader Hopkinson peak. The enhancement factor is again about 3. Throughout the blocking range, χ_0 is distinctly higher than M_S/H_R, the difference being clearly correlated with the spectrum of blocking temperatures.

Fig. 4 makes a similar comparison of data except that here (and in Fig. 5) there is no continuous record of the T_B spectrum. Instead the form of the spectrum is indicated by a series of partial TRM's. Because it contains relatively large needle-like grains (0.2—0.3 μm long), the SD magnetite has a T_B spectrum reminiscent of MD grains. The Hopkinson peak is narrow

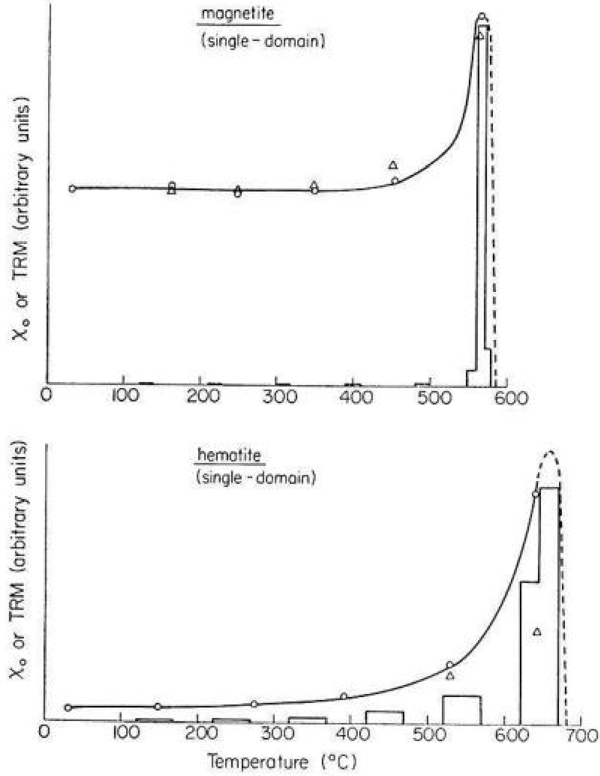

Fig. 4. Hopkinson effect data for SD magnetite and hematite (samples 3 and 6 of Dunlop and West, 1969). Symbols are as in Fig. 3. Only a partial indication of the blocking temperature distribution is given, a series of partial TRMs being plotted. Note the spectacular Hopkinson enhancement factor of the hematite.

and there is an enhancement in χ_0 of less than a factor 2, apparently confined to the blocking region. The SD hematite (grain size 0.1—1 μm) has a broader T_B spectrum and Hopkinson peak, with spectacular enhancement of χ_0 — by a factor 5 at 530 °C and a factor 20 at 640 °C. The enhancement reflects the fact that the room-temperature value of χ_0 is very low (because of the large coercive force of hematite at T_0) compared to the SPM susceptibility. Unfortunately hematite is not of interest in the context of crustal anomalies, being at least two orders of magnitude less magnetic than magnetite, but it does provide a spectacular illustration of the principles involved.

Fig. 5 shows susceptibility data for two rocks with fairly broad T_B spectra and correspondingly broad Hopkinson peaks. Rhadhakrishnamurty and Likhite (1970) show a number of similar examples, which they classify

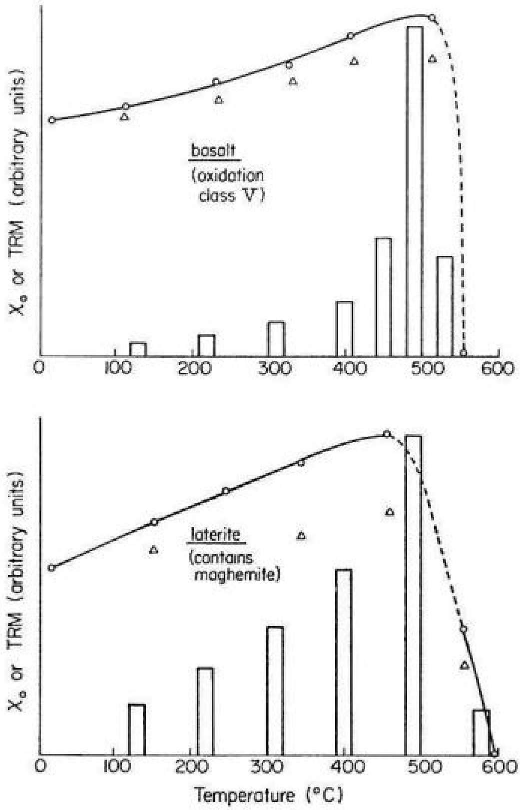

Fig. 5. Hopkinson effect data for two rocks (samples 1 and 2 of Dunlop and West, 1969). Symbols, etc. as in Fig. 4. The broad Hopkinson peaks reflect the broad blocking temperature distributions

as being of SD-type. (However, since we have seen in Figs. 3 and 4 that SD magnetites can have quite narrow T_B spectra and Hopkinson peaks, this classification is a little misleading). The susceptibilities of the rocks in Fig. 5 are only enhanced by 50—70%, but both Radhakrishnamurty and Likhite (1970) and Nagata (1961, pp. 98, 143, 163) show examples of basalts with enhancement factors of 1.5—2.5.

The Source of Deep-Seated Anomalies

The Hopkinson effect data presently available is scanty and refers specifically to basalts and fine-grained magnetites. It clearly needs to be supplemented by data on MD magnetites and coarse-grained rocks. Nevertheless a modest enhancement, by a factor 2 or so, of the susceptibility of deeply buried crustal rocks seems to be entirely possible. Larger enhancements appear unlikely. Since the range of significant susceptibility increase is typically no more than 100 °C and often more like 50 °C, the depth extent of crustal bodies whose effective magnetization is enhanced by this mechanism can scarcely be more than a few km.

The actual depth in the crust at which enhancement occurs is less certain. The surface geothermal gradient in shield areas is about 30 °C/km but a lower figure is appropriate for the whole crust, perhaps as low as 10 °C/km. The Curie point (580 °C) isotherm for magnetite could then be as deep as 50 km, although thermal evolution models (e.g. MacDonald, 1963) and phase equilibrium considerations (Turner and Verhoogen, 1951) favour a depth in the range 25 to 40 km.

Of course, somewhat shallower bodies (temperatures of 200—400 °C) may exhibit enhanced magnetization if they contain titanomagnetites with a high titanium content and correspondingly lower Curie temperature. This is particularly likely to be true of deep intrusions in which the titanomagnetites have retained their primitive compositions.

A final point to note is that observed anomalies reflect the magnetization produced by the geomagnetic field acting over the past 10^6 years. Viscous (time-dependent) magnetization is known to be enhanced, sometimes spectacularly, at high temperature (e.g. Shashkanov and Metallova, 1970) and could well result in an effective susceptibility appreciably higher than that measured in short-term experiments.

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