

# Magnetic petrophysics and magnetic petrology: aids to geological interpretation of magnetic surveys

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The present capability to acquire, process and display very large sets of high-quality magnetic survey data has far outstripped the capacity to extract all the geological information from the data. Interpretation of magnetics has lagged behind, largely because of a lack of information on magnetic properties within the survey areas and because the geological factors and processes that determine magnetic signatures are often poorly understood. In the absence of magnetic property measurements, simplistic rules-of-thumb concerning magnetic properties of different rock types can be deceptive. Furthermore, empirical correlations between mapped geology and magnetics in one area cannot necessarily be extrapolated to other areas of poor exposure if

changes in depositional environment, metamorphic grade or structural setting are ignored. Magnetic modelling is afflicted by the inherent non-uniqueness of source geometry. Information on magnetic properties in the area of interest is important for constraining interpretation in order to resolve this ambiguity. Magnetic petrology integrates rock magnetism and conventional petrology in order to define the processes that create, alter and destroy magnetic minerals in rocks. By relating magnetic mineralogy, bulk magnetic properties, petrology and geochemistry to observed magnetic anomalies an understanding of the geological factors that control magnetic signatures is obtained, which can be used to improve *geological* interpretation of magnetic surveys.

## Introduction

Although the magnetic method has been widely used in mineral exploration for decades, recent improvements in magnetic data acquisition, processing and presentation, and reduced airborne acquisition costs have increased the utility and importance of magnetic surveys, particularly high-resolution aeromagnetic surveys. Increasingly, high-quality surveys of large areas are available at reasonable cost. This has led to increasing emphasis on magnetics in area selection and regional mapping, as well as prospect-scale mapping and drill targeting.

We have now reached a stage where the ability to acquire, process and present magnetic survey data far outstrips the capacity to interpret the surveys. There is often far more geological information in these very large data sets than can presently be extracted in the time available for interpretation. Better understanding of the relationships between magnetic signatures and geology can facilitate the interpretation process and produce more reliable geological interpretations.

Magnetic anomalies arise from sources at all depths within the crust and, therefore, magnetic surveys provide 3D geological information. Unlike other airborne geophysical mapping tools, which detect properties of superficial layers (radiometrics, remote sensing) or have limited depth penetration (EM), magnetics have the ability, in principle, to see to great depths. This aspect of magnetic survey data is not yet fully exploited, because of the lack of adequate 3D modelling tools and automated interpretation techniques for large data sets. With the increased affordability of computing power, the opportunity exists to develop software aimed at extracting true three-dimensional pictures of the Earth from magnetics. By constraining such models with other geodata, fully integrated interpretations of multiple data sets will be achievable.

A crucial limitation of 3D interpretations of magnetic surveys arises from the fundamental non-uniqueness of potential field source distributions. This ambiguity in source geometry can only be addressed by constraining models. The most important control on the reliability of magnetic models is information on magnetic properties. Understanding of the factors that determine magnetisation intensities and directions for the geological units within the survey area is essential for resolving geological ambiguity in order to produce a reliable interpretation of subsurface geology.

## Induced, remanent and viscous magnetisation

Magnetised matter contains a distribution of microscopic

magnetic moments. Unpaired electron spins are the most important sources of magnetic moment. Magnetisation  $\mathbf{J}$  is defined as the magnetic dipole moment per unit volume of the material. *Induced magnetisation*  $\mathbf{J}_I$  is the component of magnetisation produced in response to an applied field. The induced magnetisation varies sympathetically with changes in the applied field and vanishes when the field is removed. *Remanent magnetisation* or *remanence*  $\mathbf{J}_R$  is the 'permanent' magnetisation that remains when the applied field is removed, and is essentially unaffected by weak fields.

The total magnetisation is the vector sum of the induced and remanent magnetisations:

$$\mathbf{J} = \mathbf{J}_I + \mathbf{J}_R$$

For sufficiently weak fields, such as the geomagnetic field, the induced magnetisation is approximately proportional to the applied field. The constant of proportionality is known as the *susceptibility*,  $k$ . Thus if the applied field is  $\mathbf{F}$ , the induced and total magnetisations are given by:

$$\mathbf{J}_I = k\mathbf{F}; \quad \mathbf{J} = k\mathbf{F} + \mathbf{J}_R$$

For most rocks, the induced magnetisation is essentially parallel to the applied field, irrespective of the field direction. In this case the susceptibility is a scalar quantity, i.e. it is characterised simply by its magnitude and is isotropic. The case of anisotropic susceptibility is discussed below. The *Koenigsberger ratio* ( $Q$ ) is a convenient parameter for expressing the relative importance of remanent and induced magnetisations. It is given by:

$$Q = J_R / J_I = J_R / kF.$$

Thus  $Q > 1$  indicates that remanence dominates induced magnetisation, whereas  $Q < 1$  implies that induced magnetisation is dominant.

The distinction between induced and remanent magnetisations is not completely clear-cut because, strictly speaking, remanent magnetisation is metastable, not permanent. A wide range of time constants characterises the equilibration of magnetisation in many magnetic materials. Thus, magnetisation responds over time to changes in the applied field. The time-varying part of the remanence is called *viscous remanent magnetisation*. Figure 1 illustrates the distinction between induced magnetisation, viscous remanence and stable remanence. Figure 1a shows the changes in magnetisation of an initially demagnetised sample in response to an applied field, which is switched on and off as shown in Figure 1b. Small applied fields, comparable in strength to the geomagnetic field, produce small, reversible changes in magnetisation when they are applied briefly, i.e. the magnetisation vanishes on removal of the field. This induced magnetisation is approximately

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proportional to the strength of the applied field. Thus, the susceptibility, which is defined as the induced magnetisation divided by the applied field, is approximately independent of the field. If a larger field is applied and then removed an irreversible change in magnetisation occurs—an *isothermal remanent magnetisation*  $J_{IRM}$  has been imparted to the sample. This magnetisation can be regarded as permanent on the experimental time scale. Figure 1c illustrates this behaviour in terms of the hysteresis loop, plotting magnetisation  $J$  versus applied field  $H$ . The initial portion of the  $J$ - $H$  curve is approximately linear and is traversed reversibly when a small field is applied and removed. When the larger field is applied

and instantaneously removed, the sample follows the trajectory  $abc$ . At point  $b$ , the total magnetisation is the sum of the induced magnetisation, which by definition is the component that vanishes when the field is removed, and the isothermal remanance.

If the field is again switched on, the magnetisation follows the path  $cd$  and returns initially to its former value ( $J_{IRM} + J_{IND}$ ). If the applied field remains on, however, the total magnetisation increases gradually with time, from  $d$  to  $e$ . After an initial period, which depends on the grain size distribution, the increase in magnetisation is usually found to be approximately proportional to the logarithm of time, over several

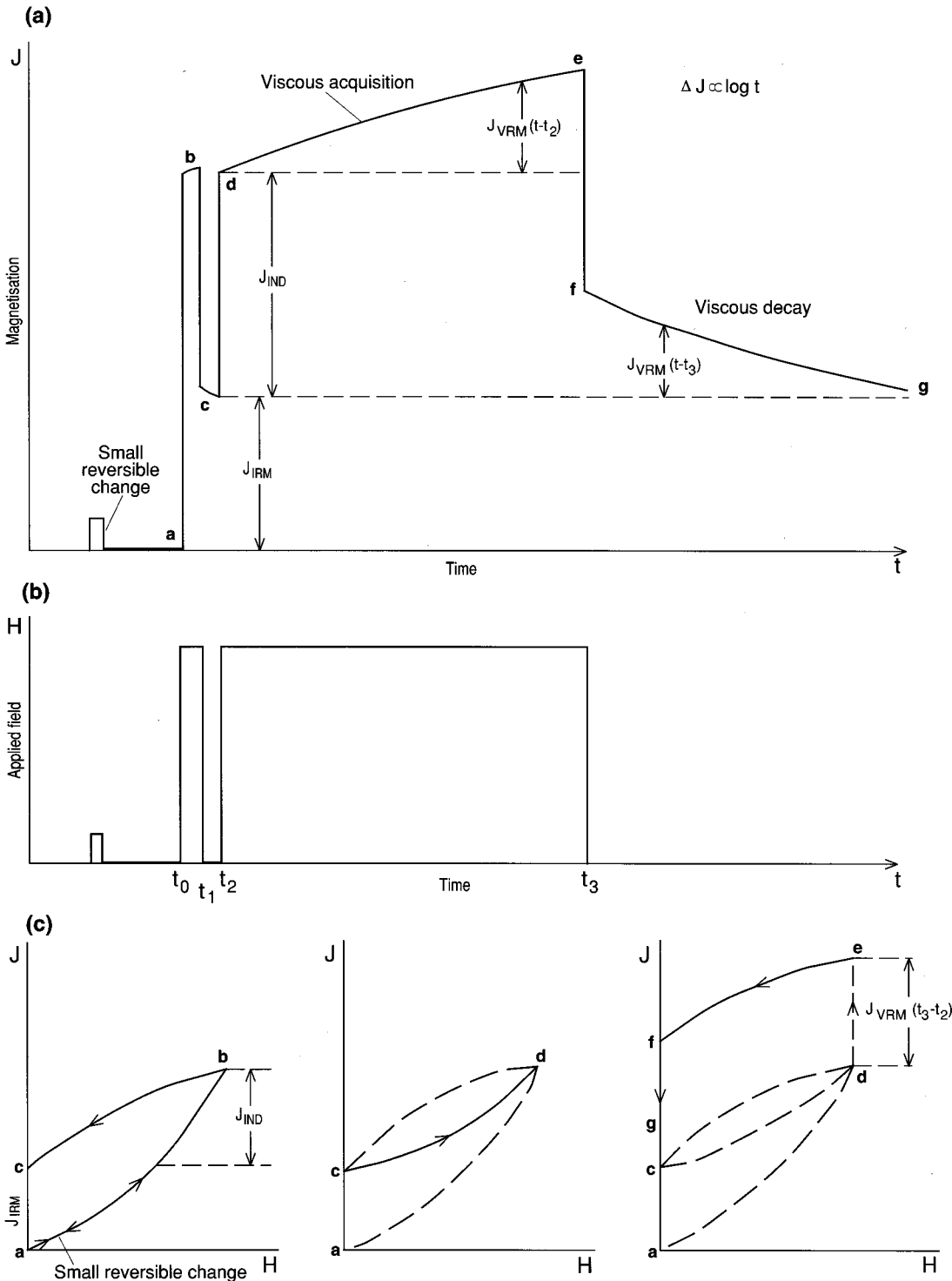


Figure 1. Distinction between induced, viscous and remanent magnetisations (see text).

decades of  $t$ . When the field is removed, this excess magnetisation, the viscous remanent magnetisation (VRM), remains and augments the isothermal remanence, but decays at a rate comparable to the acquisition rate. After sufficient time in zero field, the VRM has decayed completely and only the stable IRM remains. Thus, viscous remanence is a temporary magnetisation that is intermediate in character between induced magnetisation and more stable forms of remanence. The separation of magnetisation into induced, viscous and stable components is not exact, because the distinction depends on the time scale under consideration.

In general, the sample may not be initially in a demagnetised state. If the sample carries a stable remanence component initially, the magnetisation plots in Figures 1a and 1c are simply shifted upwards.

### Types of remanent magnetisation

The natural remanent magnetisation (NRM) of a rock reflects its history and the ambient field. Some rock samples carry isothermal remanent magnetisation as a result of exposure to magnetic fields. In this case, the NRM is not representative of the remanence carried by the bulk of the rock unit. Lightning-affected samples often carry intense NRM, characterised by anomalously high  $Q$  and directions that are scattered for samples separated by a few metres or more. NRM measurements from surface exposures should, therefore, be regarded with caution. Drilling, logging with pencil magnets, mining activities and exposure to fields during or after collection can contaminate sample NRMs with IRM components of no geological significance. Standard palaeomagnetic cleaning techniques are strongly recommended to detect and preferentially remove these sources of palaeomagnetic noise.

During initial cooling of an igneous rock, magnetic mineral grains make the transition from paramagnetism to ferromagnetism as they cool through their Curie points. A spontaneous magnetisation appears that is initially in equilibrium with the applied field, but becomes 'frozen in' or blocked at a somewhat lower temperature, called the blocking temperature, when the relaxation time of the grain's magnetic moment increases prodigiously. Below the blocking temperature, the magnetisation is a stable remanence, which is known as *thermoremanent magnetisation* (TRM). Chemical changes to these ferromagnetic minerals at elevated temperature, but below the Curie temperature, produce *thermochemical remanent magnetisation* (TCRM).

Detrital magnetic mineral grains align their magnetic moments with the ambient field as they settle through water. As a result, sediments acquire detrital remanent magnetisation (DRM), which is subparallel to, but slightly shallower than, the ambient field. Subsequent rotation of grains within pore spaces produces a *post-depositional remanent magnetisation* (pDRM) which more closely reflects the ambient field. Crystallisation of new magnetic minerals or physico-chemical changes to existing magnetic minerals during diagenesis produces *chemical remanent magnetisation* (CRM) in many sedimentary rocks. Neglecting anisotropy, TRM and CRM are acquired parallel to the ambient field at the time of acquisition. For each of these types of remanence the remanent intensity is approximately proportional to the magnitude of the field, for typical geomagnetic field strengths.

The NRM of a rock may consist of components, carried by different subpopulations of magnetic minerals, acquired at different times. For example low-grade or medium-grade metamorphism may overprint a primary remanence with a partial TRM. Alteration or weathering may superimpose a CRM on an earlier remanence.

## Magnetic minerals in rocks

### *Diamagnetism, paramagnetism and ferromagnetism*

Diamagnetism arises from the response of orbiting electrons in atoms to a magnetic field. The orbital motions adjust to oppose the applied field, i.e. diamagnetism is characterised by a negative susceptibility. Diamagnetism is present in all materials, but is very weak and is overwhelmed by paramagnetism when permanent magnetic dipoles, usually due to unpaired electron spins, are present. Diamagnetic minerals (e.g. calcite and quartz) have very weak, negative susceptibilities (about  $-10^{-6}$  G/Oe or  $-10^{-5}$  SI) and can usually be regarded as nonmagnetic for geophysical purposes. Minerals that contain iron, manganese or a number of other transition metal and rare-earth ions are paramagnetic, provided that the electron spins of adjacent ions are weakly correlated. Paramagnetic minerals (e.g. olivines, pyroxenes and pure ilmenite) have weak positive susceptibilities (maximum  $\sim 10^{-3}$  G/Oe  $\approx 10^{-2}$  SI; generally less than  $10^{-4}$  G/Oe  $\approx 10^{-3}$  SI) and do not carry remanence. They were, therefore, normally unimportant for magnetic interpretation, given that most rocks only contain a few per cent of these minerals. In recent times, however, the enhanced sensitivity of magnetometers and improved resolution of magnetic surveys permit detection of susceptibility contrasts between rocks of differing paramagnetic mineral contents.

Antiferromagnetism occurs when magnetic moments are parallel within a sublattice, but antiparallel on different sublattices, such that the net moment is zero, in the absence of an applied field. The susceptibility of antiferromagnetic minerals is low, similar to that of paramagnetic substances, and they do not acquire remanence. Magnetically ordered phases that possess spontaneous magnetisation can be ferromagnetic *sensu stricto* (e.g. iron) or ferrimagnetic (e.g. magnetite). Ferromagnetism occurs when exchange interactions tend to align electron spins. Ferrimagnetism occurs when there is antiferromagnetic coupling, but the moments of the two magnetic sublattices are unequal. For simplicity, all these strongly magnetic minerals will be referred to hereafter as ferromagnetic. Ferromagnetic minerals lose their spontaneous magnetisation at a characteristic temperature, the Curie point ( $T_C$ ), which is a function only of composition and which can, therefore, be used to detect particular ferromagnetic phases. Below its Curie temperature, a ferromagnetic mineral has high susceptibility and can carry remanence. Above  $T_C$  the mineral becomes paramagnetic, with low susceptibility and no remanence. Paramagnetic minerals have greater susceptibility at low temperatures, reflecting a  $\sim 1/T$  dependence, and may become magnetically ordered, i.e. ferromagnetic or antiferromagnetic, below a transition temperature (the Curie point or the Néel point, respectively) which is composition dependent and can be used to determine the presence of particular paramagnetic minerals. Table 1 gives susceptibilities of common diamagnetic and paramagnetic minerals. The properties of ferromagnetic minerals are discussed below. Readers are referred to O'Reilly (1984) for detailed discussion of the atomic basis of magnetism and the properties of natural minerals.

### *Intrinsic magnetic properties of spinel group minerals*

Banerjee (1991) has recently reviewed the magnetic properties of Fe-Ti oxide minerals. Bleil & Petersen (1982) compiled extensive data on magnetic properties of natural minerals, including spinels. Figure 2a shows the compositions of the Fe-Ti oxide minerals that play a major role in rock magnetism. Many cations other than titanium can substitute into magnetite, with important effects on magnetic properties. Spontaneous magnetisation at room temperature ( $J_S$ ) and Curie temperature

( $T_C$ ), which are intrinsic properties dependent only on composition, are given in Table 2 for a number of minerals, including some end-member spinel phases. Titanomagnetites with less than ~80 mole% ulvospinel are ferromagnetic at room temperature. Although the spontaneous magnetisation and Curie temperature of titanomagnetites decrease steadily with increasing titanium content (Fig. 2b, c), the susceptibility and specific remanent intensity of TRM are not strongly dependent on composition for the ferromagnetic phases. Thus, moderate titanium content in magnetite does not generally

**Table 1. Common diamagnetic and paramagnetic minerals (data from Bleil & Petersen 1982).**

Mineral	Type of magnetism	Susceptibility (G/Oe)	Susceptibility (SI)
Quartz*	Diamagnetic	$-1.2 \times 10^{-6}$	$-1.5 \times 10^{-5}$
Orthoclase*	Diamagnetic	$-1.1 \times 10^{-6}$	$-1.4 \times 10^{-5}$
Forsterite*	Diamagnetic	$-1.0 \times 10^{-6}$	$-1.3 \times 10^{-5}$
Calcite*	Diamagnetic	$-1.0 \times 10^{-6}$	$-1.3 \times 10^{-5}$
Gypsum*	Diamagnetic	$-2.3 \times 10^{-6}$	$-2.9 \times 10^{-5}$
Anhydrite*	Diamagnetic	$-4.7 \times 10^{-6}$	$-5.9 \times 10^{-5}$
Halite*	Diamagnetic	$-0.8 \times 10^{-6}$	$-1.0 \times 10^{-5}$
Galena*	Diamagnetic	$-2.6 \times 10^{-6}$	$-3.3 \times 10^{-5}$
Sphalerite*	Diamagnetic	$-1.0 \times 10^{-6}$	$-1.3 \times 10^{-5}$
Apatite*	Diamagnetic	$-0.85 \times 10^{-6}$	$-1.1 \times 10^{-5}$
Fayalite*	Paramagnetic	$390 \times 10^{-6}$	$490 \times 10^{-5}$
Ferrosilite*	Paramagnetic	$260 \times 10^{-6}$	$330 \times 10^{-5}$
Hedenbergite*	Paramagnetic	$220 \times 10^{-6}$	$270 \times 10^{-5}$
Olivine	Paramagnetic	$10...430 \times 10^{-6}$	$12...540 \times 10^{-5}$
Orthopyroxene	Paramagnetic	$10...260 \times 10^{-6}$	$12...330 \times 10^{-5}$
Clinopyroxene	Paramagnetic	$50...220 \times 10^{-6}$	$60...280 \times 10^{-5}$
Actinolite	Paramagnetic	$40 \times 10^{-6}$	$50 \times 10^{-5}$
Hornblende	Paramagnetic	$60...110 \times 10^{-6}$	$75...130 \times 10^{-5}$
Sodic amphiboles	Paramagnetic	$270 \times 10^{-6}$	$340 \times 10^{-5}$
Pyrope	Paramagnetic	$40 \times 10^{-6}$	$50 \times 10^{-5}$
Almandine	Paramagnetic	$210...530 \times 10^{-6}$	$260...660 \times 10^{-5}$
Spessartine	Paramagnetic	$540 \times 10^{-6}$	$680 \times 10^{-5}$
Andradite	Paramagnetic	$180...350 \times 10^{-6}$	$230...440 \times 10^{-5}$
Biotite	Paramagnetic	$70...260 \times 10^{-6}$	$90...330 \times 10^{-5}$
Phlogopite	Paramagnetic	$15...25 \times 10^{-6}$	$20...30 \times 10^{-5}$
Muscovite	Paramagnetic	$3...60 \times 10^{-6}$	$4...75 \times 10^{-5}$
Cordierite	Paramagnetic	$15...90 \times 10^{-6}$	$20...110 \times 10^{-5}$
Epidote	Paramagnetic	$80 \times 10^{-6}$	$100 \times 10^{-5}$
Sphene	Paramagnetic	$20 \times 10^{-6}$	$30 \times 10^{-5}$
Psilomelane	Paramagnetic	$270 \times 10^{-6}$	$340 \times 10^{-5}$
Ilmenite	Paramagnetic	$150 \times 10^{-6}$	$190 \times 10^{-5}$
Hausmannite	Paramagnetic	$60 \times 10^{-6}$	$75 \times 10^{-5}$
Chromite	Paramagnetic	$225...580 \times 10^{-6}$	$280...730 \times 10^{-5}$
Spinel	Paramagnetic	$2 \times 10^{-6}$	$3 \times 10^{-5}$
Siderite	Paramagnetic	$210...810 \times 10^{-6}$	$260...1020 \times 10^{-5}$
Magnesite	Paramagnetic	$5 \times 10^{-6}$	$6 \times 10^{-5}$
Dolomite	Paramagnetic	$1 \times 10^{-6}$	$1.2 \times 10^{-5}$
Rhodochrosite	Paramagnetic	$380 \times 10^{-6}$	$480 \times 10^{-5}$
Pyrite	Paramagnetic	$3.4 \times 10^{-6}$	$4.3 \times 10^{-5}$
Marcasite	Paramagnetic	$5...20 \times 10^{-6}$	$6...25 \times 10^{-5}$
Sphalerite	Paramagnetic	$0...160 \times 10^{-6}$	$0...200 \times 10^{-5}$
Chalcopyrite	Paramagnetic	$25...35 \times 10^{-6}$	$30...40 \times 10^{-5}$
Bornite	Paramagnetic	$45...70 \times 10^{-6}$	$55...90 \times 10^{-5}$
Arsenopyrite	Paramagnetic	$3...50 \times 10^{-6}$	$4...60 \times 10^{-5}$

\* pure phases

produce weaker magnetic properties, contrary to popular opinion. In fact, grain size is a more important factor influencing magnetisation of titanomagnetite-bearing rocks (Fig. 2d, e).

Note that the Curie point of maghaemite cannot be observed directly, because maghaemite inverts to haematite below its Curie temperature, at temperatures as low as ~300°C, depending on impurities. The Curie temperature of magnetite-bearing spinel minerals varies systematically with magnetite content. To a first approximation, the Curie point of a particular spinel composition can be estimated by linear interpolation between the  $T_C$  of the constituent end members. Diamagnetic minerals can be assigned a nominal  $T_C$  of absolute zero (-273°C) for this purpose. For the titanomagnetite series, a more accurate expression relating Curie temperature to mole fraction of ulvospinel (x) is:

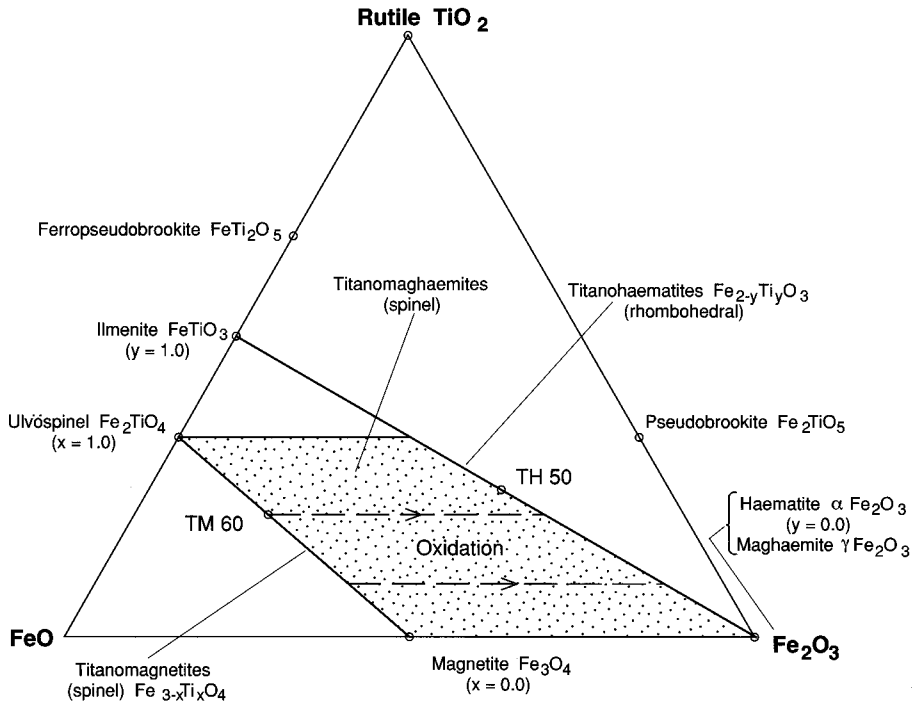
$$\text{Fe}_{3-x}\text{Ti}_x\text{O}_4: T_C (\text{°C}) = 578 - 580x - 150x^2.$$

**Table 2. Intrinsic magnetic properties of ferromagnetic minerals.**

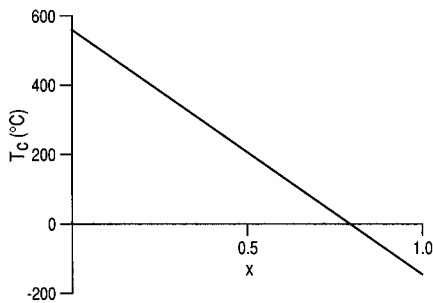
Mineral	Chemical formula	Spontaneous magnetisation (G)	Curie temperature (°C)
Iron	Fe	1710	770
Awaruite	Ni <sub>3</sub> Fe	950	620
Magnetite	Fe <sub>3</sub> O <sub>4</sub>	480	578
Ulvospinel	Fe <sub>2</sub> TiO <sub>4</sub>	0	-153
Titanomagnetite	Fe <sub>3-x</sub> Ti <sub>x</sub> O <sub>4</sub>	~480-600x (x<0.8); 0 (x≥0.8)	~578-580x -150x <sup>2</sup>
Maghaemite	γFe <sub>2</sub> O <sub>3</sub>	440	>>300 (750?)
Kenomagnetite (cation-deficient magnetite)	Fe <sub>3-y</sub> O <sub>4</sub> (0 < y < 1.3)	~460	~600...630
Titanomaghaemite	Fe <sub>(3-R)</sub> Ti <sub>R</sub> O <sub>4</sub> (0.89 < R < 1)	10...70 (not monotonic)	150...450 (common range)
Magnesian ferrite	MgFe <sub>2</sub> O <sub>4</sub>	≤220 (function of cation dist)	≤420 (function of cation dist)
Chromite	FeCr <sub>2</sub> O <sub>4</sub>	0	-185
Ferrichromite/Cr-magnetite	Fe <sub>3-x</sub> Cr <sub>x</sub> O <sub>4</sub> 0 ≤ x ≤ 1.2 e.g. Fe <sub>2</sub> CrO <sub>4</sub>	ferromag for	-185...578
Hercynite	FeAl <sub>2</sub> O <sub>4</sub>	0	-265
Magnesian ulvospinel	Mg <sub>2</sub> TiO <sub>4</sub>	0	diamagnetic
Picrochromite	MgCr <sub>2</sub> O <sub>4</sub>	0	-258
Spinel	MgAl <sub>2</sub> O <sub>4</sub>	0	diamagnetic
Jacobsite	MnFe <sub>2</sub> O <sub>4</sub>	397	300
Jacobsite/Mn-magnetite	Fe <sub>3-x</sub> Mn <sub>x</sub> O <sub>4</sub>	ferromag for 0 ≤ x ≤ 2.5	
Trevorite	NiFe <sub>2</sub> O <sub>4</sub>	330	595
Coulsonite	FeV <sub>2</sub> O <sub>4</sub>	0	-164
Haematite	αFe <sub>2</sub> O <sub>3</sub>	2	680
Ilmenite	FeTiO <sub>3</sub>	0	-205
Titanohaematite	Fe <sub>2-x</sub> Ti <sub>x</sub> O <sub>3</sub>	0 ≤ x < 0.5: antiferro 0.5 ≤ x ≤ 0.8: ferro 0.8 < x ≤ 1: para	~680-885x
Monoclinic pyrrhotite	Fe <sub>7</sub> S <sub>8</sub>	90	320
Hexagonal pyrrhotite	Fe <sub>9/10</sub> ...Fe <sub>11</sub> S <sub>12</sub>	0	270...210
Smythite	Fe <sub>9</sub> S <sub>11</sub>	~50	>300 (~400?)
Greigite	Fe <sub>3</sub> S <sub>4</sub>	~30	~350

1 G = 1 kAm<sup>-1</sup> (SI)

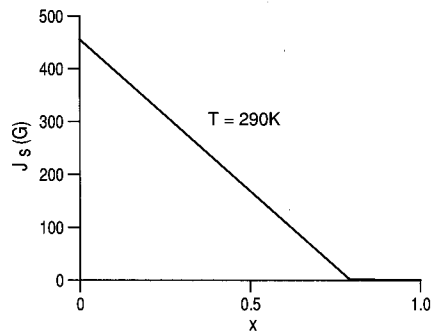
(a) Iron-titanium oxide minerals



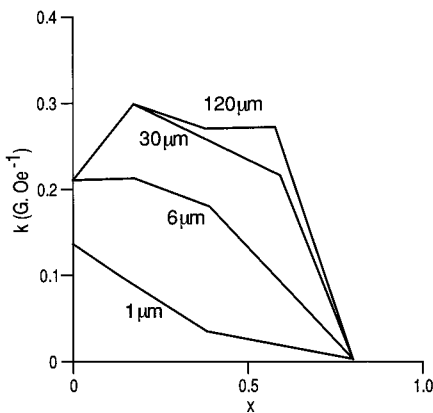
(b) Curie temperature of TMs



(c) Spontaneous magnetisation of TMs



(d) Susceptibility of TMs



(e) TRM of TMs (H = 1 Oe)

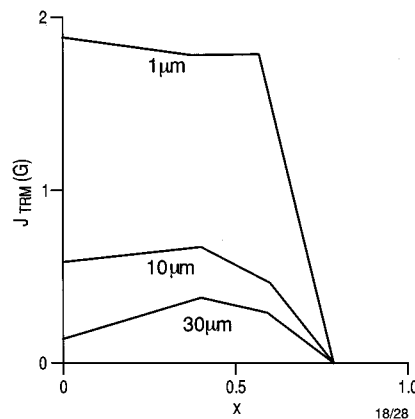


Figure 2. (a) Composition of the Fe-Ti oxide minerals that play a major role in rock magnetism. Composition dependence of: (b) Curie temperature, (c) spontaneous magnetisation, (d) susceptibility and (e) specific remanent intensity of TRM for titanomagnetites. Note the grain-size dependence of susceptibility and TRM intensity.

The effect on  $T_C$  of substitution of Cr, Al and V into magnetite is broadly similar to that of Ti substitution. Substitution of Ni increases the Curie point slightly, as does cation deficiency. Thus cation-deficient magnetites ('kenomagnetite'), representing compositions intermediate between stoichiometric magnetite and maghaemite, can have  $T_C$  above 600°C. Magnesioferrite is the only important spinel end member, other than magnetite, that is ferromagnetic at room temperature. The magnetic properties of magnesioferrite depend strongly on the cation distribution, which reflects thermal history. The Curie point of magnesioferrite is given by:

$$\text{MgFe}_2\text{O}_4: \quad T_C(\text{°C}) = 417 - 490f,$$

where  $f$  is the fraction of  $\text{Mg}^{2+}$  ions on tetrahedral sites. Because of the elevated  $T_C$  and high spontaneous magnetisation, magnesioferrite-rich spinels and Mg-magnetites are strongly magnetic and can be important contributors to the magnetic properties of rocks (e.g. some kimberlites) in which they occur.

A note of caution needs to be sounded concerning the magnetic properties of spinels. Although a number of three-component solid-solution series have been studied, in most cases the composition is restricted to the edges of the spinel prisms. Large portions of the interiors have not been systematically investigated. Although there is some theoretical basis for predicting the properties of complex spinel compositions, unusual behaviour has been found in special cases and it is conceivable that some of the exotic spinels found in some mafic and ultramafic rocks may have unexpected behaviour.

#### ***Intrinsic magnetic properties of titanohaematites and picroilmenites***

The rhombohedral titanohaematites are second in importance only to the titanomagnetites. The magnetic phase diagram of the titanohaematite series is complex and the magnetic properties of the compositions that are ferromagnetic at room temperature are influenced by thermal history.  $\text{FeTiO}_3\text{-Fe}_2\text{O}_3$  solid solutions have been thoroughly studied. The Curie temperature decreases approximately linearly with increasing ilmenite content. Titanohaematites containing between 50 mole% and 80 mole% ilmenite are strongly magnetic and are efficient carriers of remanence. An interesting property of titanohaematites that contain between 50 mole% and 73 mole% ilmenite is that they can acquire self-reversed thermoremanent magnetisation. Although full self-reversal of remanence is a rare phenomenon in rocks, the majority of well-documented cases are attributable to titanohaematites within this composition range. Compositions closer to ilmenite are paramagnetic at room temperature and haematite-rich compositions are only weakly magnetic.

Pure haematite has a diagnostic magnetic transition (the Morin transition) at  $-20^\circ\text{C}$ . Below this temperature, atomic magnetic moments are aligned with the crystallographic  $c$ -axis and haematite is antiferromagnetic. Above the transition the moments lie in the basal plane, but are slightly canted out of antiparallelism, giving rise to weak ferromagnetism and an increased susceptibility. The Morin transition can also be exploited for diagnosis of well-crystallised haematite by the effect of low temperature demagnetisation on remanence. As well as the intrinsic weak ferromagnetism due to spin-canting, haematite has an additional ferromagnetism associated with crystal defects. This defect moment is dependent on grain size, crystallinity and stress history, but is not affected by the Morin transition.

The magnetic properties of picroilmenites (solid solutions between ilmenite, haematite and geikielite,  $\text{MgTiO}_3$ ) have not been systematically studied outside the former Soviet Union. Most of the data have been gathered from kimberlitic ilmenite megacrysts, but the Russian literature on this subject is often difficult to interpret, even in translation, and some inconsistencies between different studies are evident.

Results reported by Frantsesson (1970), Garanin et al. (1978) and Yakubovskaya & Ilyupin (1982) indicate that kimberlitic picroilmenites from particular pipes or clusters tend to have an inverse relationship between haematite and geikielite contents, with ilmenite approximately constant. The Curie temperature increases monotonically with haematite content. The picroilmenites in each case had somewhat different compositions. These studies found that the investigated picroilmenites became ferromagnetic at room temperature when the haematite content exceeded  $\sim 10$  mole% (ilmenite  $\sim 68$  mole%),  $\sim 15$  mole% (ilmenite  $\sim 50$  mole%) and  $\sim 20$  mole% (ilmenite  $\sim 60$  mole%) respectively. In all cases picroilmenites with less than 20 mole% haematite were found to have low  $T_C$ , below  $200^\circ\text{C}$ . The maximum Curie temperatures reported for the most haematite-rich picroilmenites were  $230^\circ\text{C}$ ,  $240^\circ\text{C}$  and  $220^\circ\text{C}$  for these three studies respectively. Approximate Curie temperature for picroilmenites with  $\sim 15$  mole% haematite was found to be, respectively,  $95^\circ\text{C}$ ,  $0^\circ\text{C}$  and  $-110^\circ\text{C}$ . It seems established, therefore, that kimberlitic picroilmenites with 10 mole% haematite are paramagnetic and those with  $>20$  mole% haematite are ferromagnetic at room temperature, but have relatively low Curie temperatures. The properties of intermediate compositions require further study.

#### ***Magnetic properties of iron sulphide minerals***

The crystallography and mineralogy of the pyrrhotite group have been reviewed by Ward (1970), Ribbe (1974), Power & Fine (1976), Vaughn & Craig (1978) and Vaughn & Lennie (1991). The general formula of pyrrhotites is  $\text{Fe}_{1-x}\text{S}$ , with  $0 \leq x \leq 0.13$ . Crystal structures in the pyrrhotite group are all superstructures of the NiAs structure. At elevated temperatures, above  $300^\circ\text{C}$ , Fe vacancies in the structure are disordered and the  $\text{Fe}_{1-x}\text{S}$  solid solution adopts the hexagonal NiAs structure, which is denoted 1C. At lower temperatures the random distribution of vacancies gives way to various vacancy ordering schemes (depending on composition and thermal history) accompanied by formation of superstructures, usually with a lowering of crystal symmetry.

The only common pyrrhotite type that is ferrimagnetic at ambient temperatures is monoclinic pyrrhotite with 4C superstructure and approximate composition  $\text{Fe}_7\text{S}_8$ . The ferrimagnetism of monoclinic pyrrhotite arises from the ordering of Fe vacancies onto alternate, antiferromagnetically coupled, layers of Fe atoms. The spontaneous magnetisation of monoclinic pyrrhotite is  $\sim 90$  G (90 kA/m) and its Curie temperature is  $\sim 320^\circ\text{C}$ . The spontaneous magnetisation is confined to the basal plane, except at cryogenic temperatures, by strong magnetocrystalline anisotropy. Within the basal plane there is also substantial intrinsic anisotropy, which controls magnetic domain structure and structure-dependent properties, such as susceptibility, remanence intensity and stability of remanence. Other common types, such as 5C pyrrhotite with composition  $\sim \text{Fe}_9\text{S}_{10}$  and 6C pyrrhotite ( $\sim \text{Fe}_{11}\text{S}_{12}$ ), are antiferromagnetic at room temperature. The temperature dependence of the crystal structure of pyrrhotite is very sensitive to composition. The variation of magnetic properties with temperature is, therefore, diagnostic of composition. Thermomagnetic analysis of pyrrhotites has been discussed at length by Schwarz & Vaughn (1972).

Monoclinic pyrrhotite with composition  $\sim \text{Fe}_7\text{S}_8$  is a relatively common ferromagnetic mineral, particularly in areas with sulphide mineralisation. The magnetic susceptibility of monoclinic pyrrhotite is less than that of magnetite, but is nevertheless substantial and strongly dependent on grain size. The susceptibility of monoclinic pyrrhotite decreases monotonically with decreasing grain size from  $\sim 0.1$  G/Oe (1.3 SI) for very coarse ( $\geq 1$  mm) grains,  $\sim 0.025$  G/Oe (0.31 SI) for  $\sim 80$   $\mu\text{m}$  grains, 0.01 G/Oe (0.13 SI) for  $\sim 10$   $\mu\text{m}$  grains, to  $\sim 0.004$  G/Oe (0.05 SI) for single domain ( $\leq 2$   $\mu\text{m}$ ) grains

(Clark 1983b). The grain-size dependence of magnetic hysteresis properties of natural pyrrhotites has been studied by Clark (1983b, 1984) and Dekkers (1988, 1989).

Rocks and ores containing monoclinic pyrrhotite often carry relatively intense remanent magnetisation, characterised by Koenigsberger ratios greater than unity (Kropacek 1971; Kropacek & Krs 1971; Schwarz 1974; Clark 1983b; Thomson et al. 1991; Clark & Tonkin 1994). Clark (1983b) has shown that strong specific remanent intensities, high coercivity and high Koenigsberger ratios are predictable consequences of the intrinsic properties of monoclinic pyrrhotite, in particular the moderately high spontaneous magnetisation and the strong magnetocrystalline anisotropy. Theoretical, experimental and petrophysical studies have demonstrated that remanence carried by monoclinic pyrrhotite, particularly by fine grains, is magnetically hard and can be stable for geologically long periods at low temperatures (Clark 1983b). Remanence carried by monoclinic pyrrhotite is easily thermally reset, however, because of the low Curie temperature and the instability of the monoclinic phase with respect to antiferromagnetic pyrrhotite phases + pyrite above  $\sim 250^\circ\text{C}$ . It is quite common, therefore, for pyrrhotite-bearing rocks to have a magnetisation dominated by an ancient remanence, which dates from the last significant thermal event or events experienced by the rocks and which may be highly oblique to the present field. Temperatures as low as  $\sim 100^\circ\text{C}$  can superimpose an overprint remanence on an older magnetisation carried by monoclinic pyrrhotite (Clark 1983b). Interpretation of magnetic anomalies associated with such rocks may be seriously in error if the effects of remanence are ignored.

Because of the high intrinsic anisotropy of monoclinic pyrrhotite, any preferred orientation of pyrrhotite grains produces substantial susceptibility anisotropy of the rock or ore within which the pyrrhotite occurs. The effect of anisotropy on the magnitude and direction of induced and remanent magnetisations may also be significant for magnetic interpretation, particularly when the pyrrhotite is coarse-grained (Clark 1983b; Thomson et al. 1991).

Monoclinic pyrrhotite exhibits significant field-dependence of susceptibility (Worm 1991). Therefore, susceptibilities of pyrrhotite-bearing rocks should be measured in fields that are comparable in intensity to the geomagnetic field. In addition, the susceptibility of massive sulphide ores with pyrrhotite, but not disseminated pyrrhotite-bearing rocks, is dependent on measurement frequency, owing to eddy currents induced in the specimens (Worm et al. 1993). Thus, susceptibility of sulphide ores should be measured at the lowest practicable frequency. Frequencies of  $\sim 1$  kHz or less are acceptable.

Greigite is iron thiospinel,  $\text{Fe}_3\text{S}_4$ , the sulphur analogue of magnetite. The spontaneous magnetisation and inferred Curie temperature of greigite are similar to those of monoclinic pyrrhotite. In recent times it has been realised that greigite is a common magnetic sulphide mineral in young sediments and in sedimentary rocks from a variety of environments (Snowball 1991; Krs et al. 1990, 1992; Roberts & Turner 1993; Reynolds et al. 1994). The magnetic properties of greigite have been characterised by Snowball (1991) and Hoffman (1992).

Smythite, which has approximate composition  $(\text{Fe}, \text{Ni})_9\text{S}_{11}$ , is a relatively rare ferromagnetic mineral associated with some magmatic ores and is also present in some sedimentary rocks. The magnetic properties of smythite probably resemble those of monoclinic pyrrhotite. Krs et al. (1993) have studied magnetic properties of smythite from organic-rich Lower Miocene claystones of Bohemia.

### Domain structure

The bulk magnetic properties of rocks reflect the modal proportions, composition and microstructure of the magnetic mineral grains, which are usually present in only minor

amounts. Microstructure includes, inter alia, grain size and shape, degree of crystallinity and textural relationships, and strongly influences the magnetic domain state of the grains. The most important control on domain structure is the effective grain size, which is equivalent to the actual grain size in a homogeneous grain, but is related to lamella size in grains with exsolution lamellae and the size of the ferromagnetic zone in a zoned grain.

Sufficiently small grains are uniformly magnetised, i.e. they have single domain (SD) structure. Ultrafine SD grains ( $< 0.03 \mu\text{m}$  for equant magnetite) are sufficiently perturbed by thermal fluctuations that the orientation of the spontaneous magnetisation flips rapidly between two or more easy directions. Such grains cannot retain a stable remanence and their magnetisation tends to relax rapidly towards any applied field, leading to a very high susceptibility. This behaviour is called superparamagnetism, and the grains are termed superparamagnetic (SPM). Rocks and soils that contain substantial quantities of SPM grains exhibit substantial frequency-dependence of susceptibility, often greater than 10 per cent and up to 24 per cent decrease in susceptibility per decade of frequency (Thompson & Oldfield 1986). Lateritic soils that contain SPM grains of cation-deficient magnetite or maghaemite are responsible for anomalous decays in coincident-loop TEM surveys (Buselli 1982).

The relaxation time for superparamagnetism increases exponentially with grain volume. Thus, slightly larger grains have very long relaxation times, even on a geological time scale, and can retain remanent magnetisation indefinitely. These stable SD grains, typically in the submicron size range, are important carriers of remanent magnetisation in many rocks. Acicular grain shape, or elongated lamellar shape, favours SD behaviour and extends the maximum size for SD behaviour to the micron range.

With increasing size it becomes energetically favourable for the grain to subdivide into a number of magnetic domains with differently oriented magnetisations. These multidomain (MD) grains have susceptibilities, controlled by self-demagnetisation, which in the case of strongly magnetic minerals, such as magnetite, are comparable to the susceptibilities of SD grains with similar composition. The remanence of MD grains is more easily demagnetised ('softer') than that carried by SD grains and is of lower specific intensity. The case of magnetite will be considered in some detail, but other magnetic minerals exhibit qualitatively similar behaviour. Magnetite grains larger than  $\sim 20 \mu\text{m}$  exhibit true MD behaviour. The coercivity, which is a measure of the ease of demagnetisation, and the remanent intensity decrease steadily with increasing grain size until they level out for grain sizes greater than  $\sim 100 \mu\text{m}$ .

Grains smaller than  $\sim 20 \mu\text{m}$  have properties intermediate between those of SD and true MD grains and are called pseudosingle domain (PSD). Small PSD grains, a few microns in size, are relatively hard and carry relatively intense remanence. For this reason, small PSD grains are the most important remanence carriers in many rocks, in spite of the fact that they constitute a minor proportion of the magnetic mineral assemblage and even though relatively large MD grains may dominate the susceptibility of the rocks.

The behaviour of some important magnetic properties with increasing grain size (above the SPM threshold size) can be summarised as: remanence, coercivity and Koenigsberger ratio (remanent/induced magnetisation) decrease; susceptibility increases slightly. Typical values of susceptibility, remanent intensity and Koenigsberger ratio of various domain states for a number of magnetic minerals are summarised in Fig. 3 (from Clark 1983a). The theoretical maximum size for superparamagnetic and single domain behaviour for equant grains of a number of magnetic minerals are summarised in Table 3.

It has become apparent in recent years that magnetic grains frequently occupy metastable domain states of anomalously low domain multiplicity. The above values for the critical SD size assume equilibrium, i.e. that the grain is in the absolute

energy minimum state. In fact, grains an order of magnitude larger than the theoretical size can remain in a metastable SD state, because formation of a domain wall requires an energy barrier to be overcome. Thus the effective upper limit for SD

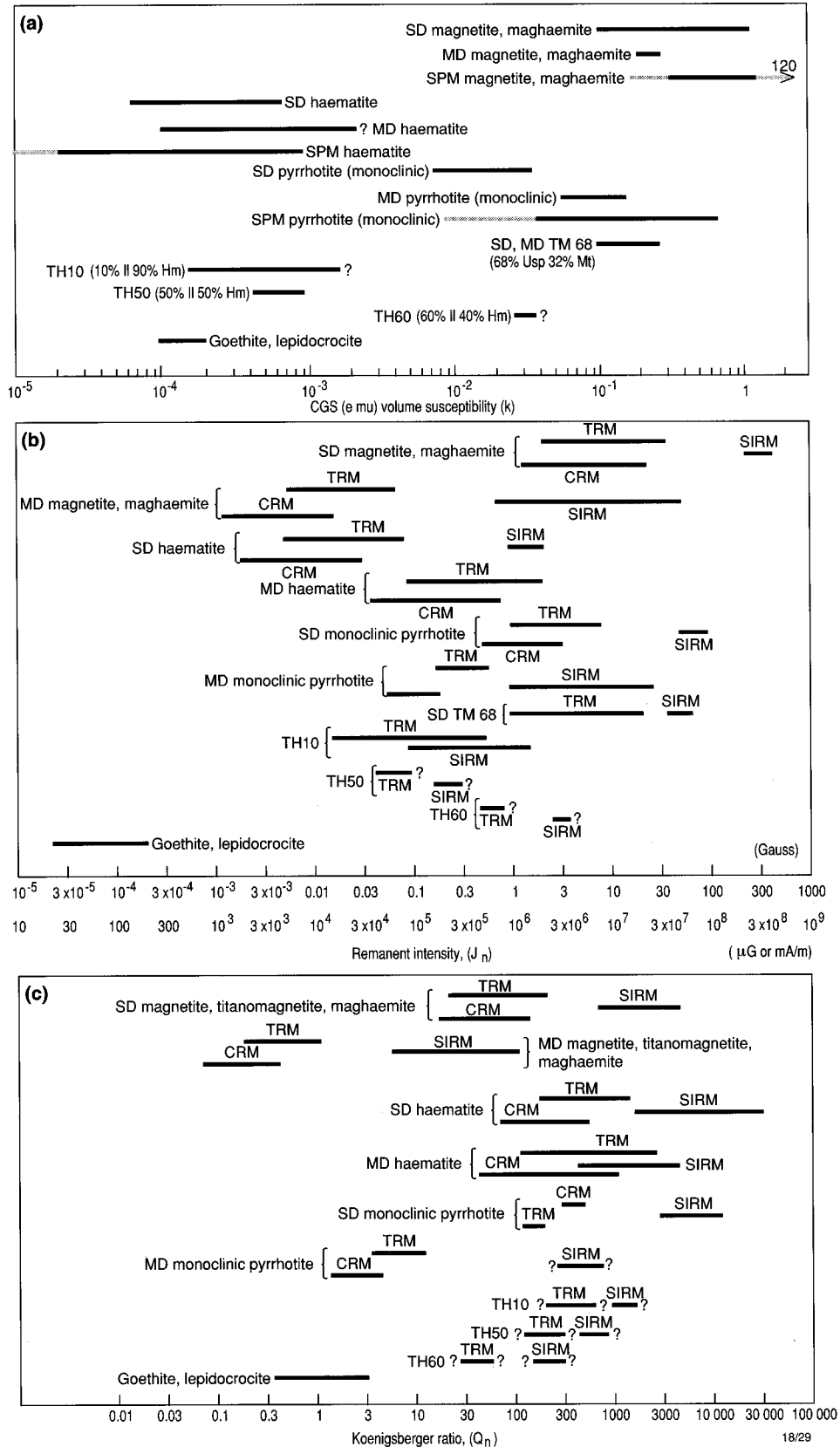


Figure 3. Typical values of (a) susceptibility, (b) remanent intensity and (c) Koenigsberger ratio of various domain states for common magnetic minerals. SIRM = saturation IRM. (from Clark 1983).



**Table 3. Superparamagnetic threshold and critical single domain sizes.**

Mineral	SPM threshold size ( $\mu\text{m}$ )	Critical SD size ( $\mu\text{m}$ )
Iron	<0.008	0.023
Magnetite	0.03	0.06
Maghaemaite	0.02	0.06
Titanomagnetite (60% usp)	0.08	0.4
Haematite	0.03	15
Monoclinic pyrrhotite	0.018	1.6
Greigite	-0.04	-0.8

**Table 4. Domain state size ranges and coercivities for magnetite.**

Domain state	Size ( $\mu\text{m}$ )	Coercivity (Oe)
SPM	<0.05	0
(Acicular) SD	0.05–1	>600
PSD	~1–20	100–600
MD	>20	100

1 Oe = 0.1 mT (SI)

behaviour has been extended to  $\sim 1 \mu\text{m}$  for magnetite and by a similar factor for other minerals. The threshold sizes are also larger for elongated grains than for equant grains. The domain structure transition sizes for titanomagnetites and other spinel phases with lower spontaneous magnetisations are larger than for magnetite. Thus the single domain/two domain transition, the upper limit of the PSD size range etc. occur at larger grain sizes for spinels with decreasing magnetite contents.

Taking metastable behaviour into account, the approximate size ranges and coercivities for SPM, (stable) SD, PSD and MD behaviour for magnetite are listed in Table 4.

### Palaeomagnetic cleaning techniques

Collinson (1983) and Butler (1992) described demagnetisation techniques routinely used for resolving palaeomagnetic components and for characterising magnetic mineralogy. Alternating field (AF) demagnetisation is similar to degaussing of permanent magnetisation of ships or tape recorder heads. The sample is subjected to an alternating magnetic field that is gradually reduced to zero, thereby randomising the moments of all grains with coercivity less than the initial AF amplitude. This procedure is repeated at successively higher fields, demagnetising successively harder fractions of the magnetic mineral assemblage and isolating the most stable component of remanence. AF demagnetisation is particularly effective at removing isothermal remanent magnetisation (IRM) acquired by exposure to strong magnetic fields, such as those produced by mining equipment or lightning strikes etc. AF demagnetisation is sometimes effective in selectively removing VRM.

Thermal demagnetisation involves heating the sample and recooling to room temperature, in zero magnetic field. This randomises the magnetic moments of all grains with unblocking temperatures less than the heating temperature. The procedure is repeated at successively higher temperatures, thereby demagnetising successively higher unblocking-temperature fractions. Thermal demagnetisation is particularly effective at unravelling the thermal history of the rock, for instance by resolving primary TRM from a later metamorphic overprint. Thermal demagnetisation is especially useful for removing

VRM components acquired at low temperatures, but is not so effective against IRM noise.

Low-temperature demagnetisation involves cooling the sample to very low temperatures (e.g. the boiling point of liquid nitrogen) and rewarming to room temperature, in zero field. This technique exploits low-temperature magnetic transitions of particular magnetic minerals, to isolate remanence components carried by different minerals or by different grain-size fractions of magnetite, for example. The effect of low-temperature demagnetisation can be characterised by the parameter  $R_K$ , which is the ratio of the remanence remaining at room temperature, after cooling to  $-196^\circ\text{C}$  in zero field, to the original remanence.

Because different remanence components are carried by subpopulations of grains with different characteristics, they can usually be distinguished by their different responses to various demagnetisation techniques. Palaeomagnetic cleaning techniques have three main applications, which can be termed palaeomagnetic, petrophysical and rock magnetic, respectively:

- Resolution of remanence components acquired at different times, allowing estimation of palaeofield directions and palaeopole positions at the time of formation, at the time of metamorphism etc.,
- Removal of palaeomagnetic noise components, which are not representative of the bulk in-situ properties, allowing characteristic NRM of the rock to be determined,
- Identification of magnetic minerals by their demagnetisation characteristics. Information on compositions, domain states and grain sizes of the magnetic minerals can be obtained.

When two or more components of remanence are present in a rock unit, the NRM direction and intensity of individual samples can be very variable, depending on the relative proportion of each component in the sample. Estimating the effective remanence of a large volume of rock from NRM measurements alone is difficult in these circumstances. Statistical techniques for estimating average vectors and errors in these circumstances are in their infancy. However, if the NRM of each sample is resolved into discrete components, a well-grouped set of directions and moderate range of intensities can be determined for each generation of remanence. Standard statistical techniques can then be used to estimate an average remanence vector, and its uncertainty, for each component. The estimated remanence vector for the rock unit can then be reconstructed from the components.

The difference in AF demagnetisation behaviour of acicular SD, small PSD and large MD magnetite is shown in Figure 4. The remanence being demagnetised is an artificially imparted saturation IRM (SIRM), produced by placing the samples in a very strong field. The plot shows remanent intensity, normalised to the initial value before AF treatment, versus AF field. The corresponding coercivity spectra can be obtained from the demagnetisation curves by differentiation, as the coercivity spectrum is the magnitude of the slope of the demagnetisation curve. Thus, the coercivity spectrum of the large MD grains peaks at low fields (less than 5 mT), the spectrum for the small PSD grains peaks at  $\sim 15$  mT and that of the SD grains peaks above 40 mT. The Lowrie-Fuller test exploits the differences between AF demagnetisation behaviour of small and large grains. For SD and PSD grains, weak field remanence, such as TRM, is more resistant to demagnetisation than strong field remanence, particularly SIRM. For true MD grains the relative stability of weak and strong field remanence is reversed.

Demagnetisation behaviour can be used to detect the presence of other magnetic minerals. For example, haematite is very hard (coercivity  $\gg 1000$  Oe) and does not thermally demagnetise until close to  $670^\circ\text{C}$ . Goethite has even greater coercivity than haematite, but has a Curie temperature of  $\sim 120^\circ\text{C}$ , and is, therefore, very easily thermally demagnetised.

### Thermomagnetic analysis of magnetic mineralogy

The variation of magnetic properties of magnetic minerals with temperature depends on composition and, in some cases, on domain state and microstructure. This variation can, therefore, be used for analysis of magnetic minerals and for optimising thermal demagnetisation steps (Schmidt 1993). The variation of susceptibility with temperature is particularly useful, because of the rapid change in susceptibility close to  $T_C$ , which enables well-defined Curie points to be determined, and because of the sensitivity of susceptibility to domain state and microstructure.

The characteristic susceptibility ( $k$ ) versus temperature ( $T$ ) behaviour of different magnetic minerals is shown in Figure 5. The  $k$ - $T$  curve for paramagnetic minerals is hyperbolic, reflecting the  $1/T$  dependence of paramagnetic susceptibility. The  $k$ - $T$  curve of magnetite with MD structure, including PSD grains as well as true MD grains, is very diagnostic. There is a prominent peak at  $-155^\circ\text{C}$ , which corresponds to the isotropic point of magnetite. Below this temperature, the easy magnetisation directions are along the  $\langle 100 \rangle$  cubic axes, whereas above it the easy directions lie along  $\langle 111 \rangle$  body diagonals of the cubic unit cell. At the isotropic point, the magnetisation rotates freely to align with an applied field, giving rise to an increase in susceptibility. The susceptibility of MD grains is limited by self-demagnetisation, and the observed susceptibility remains almost constant until just below the Curie temperature, when it plummets to paramagnetic values. The isotropic point is very sensitive to composition, and substitution of cations other than iron or departures from stoichiometry tend to lower the isotropic point. Titanomagnetites containing more than  $\sim 10$  mole% ulvospinel have isotropic points below liquid-nitrogen temperature. Thus the presence of a well-defined peak around  $-155^\circ\text{C}$  is diagnostic of the presence of nearly pure PSD and/or MD magnetite.

The titanomagnetite for which the  $k$ - $T$  curve is shown has a Curie point of  $\sim 200^\circ\text{C}$  and contains  $\sim 60$  mole% ulvospinel. The  $k$ - $T$  curve is irreversible on cooling from high temperature (not shown), owing to exsolution of more magnetite-rich and magnetite-poor titanomagnetites than the original composition. Thus two Curie points, one above  $500^\circ\text{C}$  and the other shifted somewhat lower than the original  $T_C$ , would be seen in the cooling curve.

The  $k$ - $T$  curve for SD magnetite does not exhibit a peak at the isotropic point, because the properties of SD grains are controlled largely by shape anisotropy, rather than by magnetocrystalline anisotropy. The susceptibility is almost constant at low temperatures, but increases as the Curie temperature is approached. This increase in  $k$  reflects unblocking of fine grains below  $T_C$ .

The susceptibility of grains increases sharply at the unblocking temperature, because the relaxation time suddenly decreases, allowing the magnetic moments of the grains to equilibrate rapidly with the applied field. Above the unblocking temperature, the superparamagnetic susceptibility of grains of specified volume is proportional to  $J_s^2/T$  and is much higher than the room temperature susceptibility, until the Curie temperature is approached (at which point  $J_s \rightarrow 0$ , so  $k \rightarrow 0$ ). Thus, the presence of significant unblocking of remanence well below the Curie temperature may indicate that this portion of the remanence is carried by very fine (submicron) single domain grains.

Even smaller grains unblock at much lower temperatures and exhibit superparamagnetism at room temperature. The  $k$ - $T$  curve shown for SPM grains is idealised for an assemblage of identical grains. In rocks, there is always a distribution of grain sizes, and the superposition of unblocking peaks over a wide range of temperature leads to a steady increase in susceptibility from below room temperature up to the maximum unblocking temperature of the ultrafine SPM + SD assemblage.

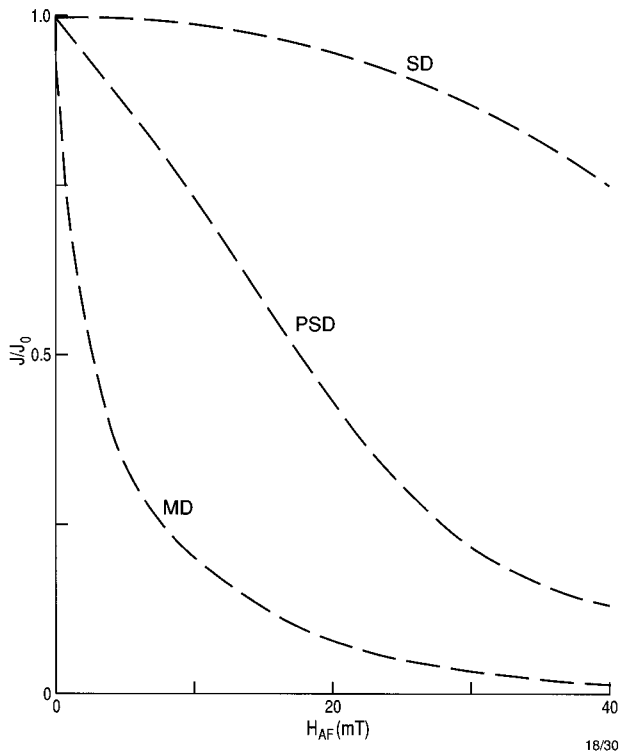


Figure 4. Characteristic AF demagnetisation behaviour of acicular SD, small PSD and large MD magnetite.  $1\text{mt} = 10\text{ Oe}$ .

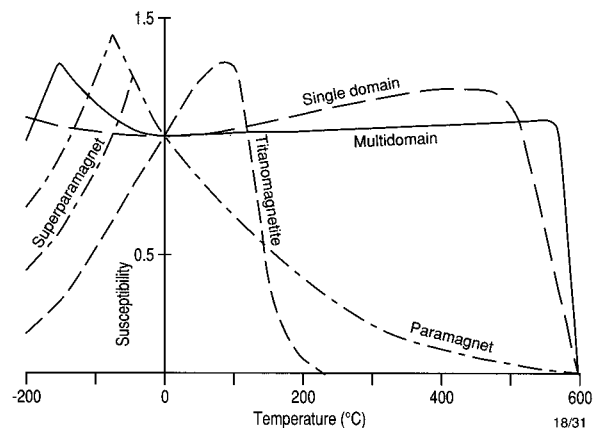


Figure 5. The characteristic susceptibility ( $k$ ) versus temperature ( $T$ ) behaviour of different magnetic minerals (after Thompson & Oldfield 1986).

This behaviour is commonly seen in soil samples, particularly lateritic soils.

Large multidomain grains also exhibit substantial unblocking below the Curie temperature, owing to rearrangements of domain structure with changes in temperature. The component unblocked below the Curie temperature has much lower coercivity if it is carried by MD grains than by SD grains. The unblocking behaviour of partial TRMs is quite diagnostic of SD versus MD carriers. For SD grains the maximum unblocking temperature for a PTRM acquired at a given temperature below the Curie point is approximately the maximum temperature of acquisition. For MD grains, on the other hand, unblocking occurs from low temperatures right up to the Curie point.

As the Curie temperature of a magnetic mineral is approached, there is a rapid decrease in spontaneous magnetisation and an even more rapid decrease in magnetocrystalline anisotropy. As a consequence, the remanence of even the most

stable grains unblocks, but without an increase in  $k$ . In fact, the susceptibility plummets until it attains paramagnetic values at  $T_C$ .

Thermal demagnetisation of remanence provides another analytical technique. MD grains exhibit a spectrum of unblocking temperatures right up to  $T_C$ , whereas the unblocking-temperature spectrum of SD grains cuts off below the Curie point. However, given that the grain-size range of a particular mineral extends at least to the upper end of the SD range, the maximum unblocking temperature lies just below  $T_C$ . Thus, Curie points can also be estimated from thermal demagnetisation data. Prominent inflexions in the demagnetisation curve, corresponding to a sharp peak in the blocking-temperature spectrum, indicate the approximate Curie temperature of a particular phase. Comparison of the blocking-temperature spectra with  $k$ - $T$  curves enables phases originally present in samples to be distinguished from phases created during heating. Low-temperature demagnetisation also allows magnetic transitions characteristic of magnetite and haematite to be detected. Judicious application of a variety of rock magnetic techniques, including thermomagnetic analysis, allows the relative contributions to susceptibility and remanence of different compositions and grain-size ranges to be estimated. Lowrie (1990) has presented an elegant technique for identifying magnetic minerals, even in trace amounts, which exploits differences in Curie temperature and coercivity for different minerals.

### Magnetic susceptibility of minerals and rocks

Figure 3a plots susceptibility for dilute dispersions of common ferromagnetic minerals, in various domain states, normalised to 100 per cent by volume. That is, if the specific volume susceptibility indicated is  $\chi$  and the volume fraction of the mineral is  $f$ , the contribution of that mineral to the susceptibility of the rock is  $\sim \chi f$ , provided  $f \ll 1$ . For strongly magnetic minerals, such as magnetite, the susceptibility versus volume fraction relationship departs substantially from linearity for  $f > 0.1$ . This is because the effective susceptibility of isolated strongly magnetic grains is mainly controlled by self-demagnetisation and depends only weakly on the intrinsic susceptibility of the grain material. Thus the microstructure and size of the grains does not greatly affect the observed susceptibility of sparsely dispersed magnetite particles. At high concentrations, however, susceptibility is influenced by interactions between grains and increases faster than the extrapolated linear relationship applying for low concentrations. On the other hand, for weakly magnetic minerals, such as haematite, the contribution to the susceptibility is essentially proportional to the volume content, for all  $f$ .

A simple model of grain interactions adequately accounts for the concentration-dependence of susceptibility for magnetite-bearing rocks. Each magnetite grain can be regarded as residing in a cavity within a magnetised medium (see Fig. 6). The environment of each grain is a magnetic 'plum pudding' of randomly dispersed magnetic grains, of average magnetisation  $J$ , within a non-magnetic matrix. Replacing the actual environment with a continuous medium that has the same volume-averaged magnetisation, viz.  $fJ$ , the nett magnetic field within each grain is:

$$H' = H - NJ + NfJ,$$

where  $H'$  is the internal field of the grain,  $H$  is the applied field, and  $N$  is the self-demagnetising factor of the grain. The second term on the RHS is the self-demagnetising field and the third term is the Lorentz field term, which corrects for the grain interactions.

Thus, the induced magnetisation of each grain obeys:

$$J = \chi H' = \chi[H - N(1 - f)J],$$

$$\therefore J = \chi H / [1 + (1 - f)N\chi].$$

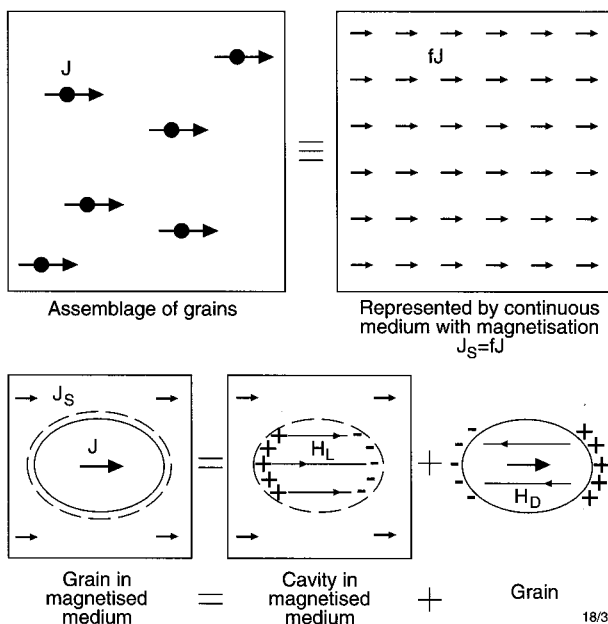


Figure 6. Simple model of effect of grain interactions on susceptibility of ferromagnetic mineral-rich rocks. The self-demagnetising field of each grain is opposed by the Lorentz field of the cavity occupied by the grain.

The magnetisation of the rock is  $fJ$ , which implies that the observed susceptibility of the rock is:

$$k = fJ/H = f\chi/[1 + (1 - f)N\chi].$$

For low concentrations of magnetite, this reduces to:

$$k = f\chi/[1 + N\chi]. \quad (f < 1)$$

For MD magnetite, the intrinsic susceptibility depends on microstructure, but is always high ( $\chi \geq 1$  G/Oe = 13 SI), and  $N$  is of the order of 3–4 Oe/G (0.25–0.3 SI). Therefore  $N\chi \gg 1$  and the susceptibility of the rock is almost independent of  $\chi$  at low concentrations of magnetite:

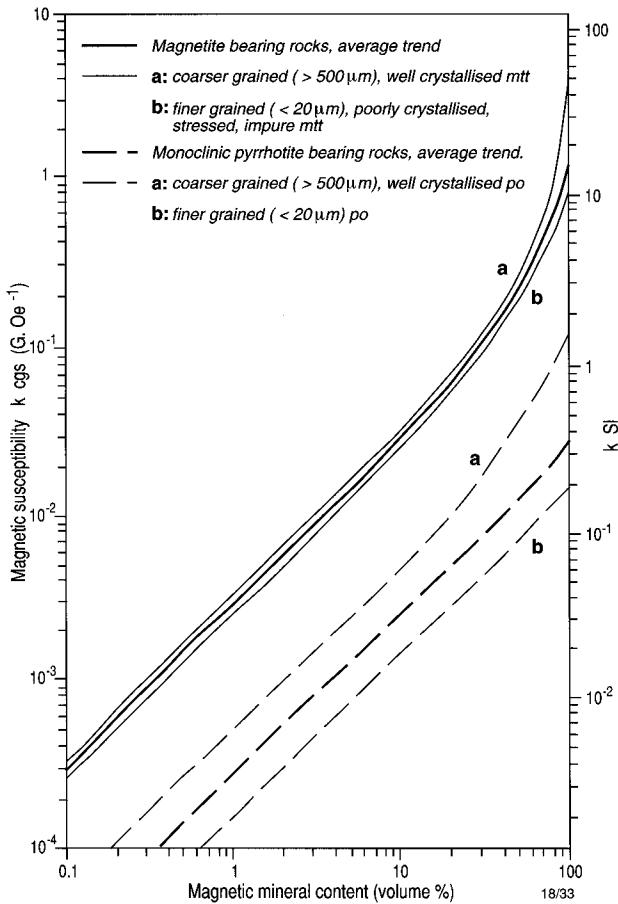
$$k \approx f/N. \quad (f \ll 1)$$

Thus, the susceptibility of a rock containing low to moderate amounts of MD magnetite is approximately proportional to magnetite content, irrespective of the magnetite grain size or microstructure. Conversely, the susceptibility can be used to infer the magnetite content for moderately magnetic magnetite-bearing rocks. According to Puranen (1989), a reasonable average empirical relationship between susceptibility and magnetite content for magnetite contents up to a few per cent is:

$$k \approx 0.276f \text{ G/Oe} = 2760v \text{ } \mu\text{G/Oe} = 3.47f \text{ SI} = 3470v \times 10^{-5} \text{ SI},$$

where  $v = 100f$  is the volume per cent magnetite content.

Significant departures from the linear  $k$ - $f$  relationship occur for magnetite content greater than about 10 per cent. In the limit of 100 per cent magnetite, the susceptibility of the ore is simply the intrinsic susceptibility of the magnetite crystals, which depends on average grain size, crystallinity, dislocation density etc. The susceptibility of pure magnetite ore approaches  $\chi$ , which ranges from  $\sim 1$  G/Oe (13 SI) for fine-grained, poorly crystalline, inhomogeneous or stressed grains to  $\sim 10$  G/Oe (130 SI) for very coarse, well-crystallised magnetite. Parry (1980) estimated the average effective demagnetising factor,  $N$ , for MD magnetite grains to be 3 Oe/G (0.24 SI). Puranen's relationship is consistent with the theoretical relationship if a typical value of intrinsic susceptibility of natural magnetite grains in rocks is  $\sim 1.6$  G/Oe (20 SI), which agrees well with Stacey & Banejee's (1974) estimates of intrinsic susceptibility



**Figure 7** Susceptibility versus magnetic mineral content for magnetite and monoclinic pyrrhotite.

of large multidomain magnetite grains. Figure 7 shows curves of susceptibility versus magnetic mineral content for magnetite and monoclinic pyrrhotite, based on the theory presented above and confirmatory experimental data.

The paramagnetic susceptibility of rocks containing no ferromagnetic minerals can usually be estimated simply and reasonably accurately from their total iron content (Puran 1989). In this case,

$$k_{SI} \approx 2.86 \times 10^{-8} \times \text{wt\% Fe} \times \text{density (kg/m}^3) = 2.22 \times 10^{-8} \times \text{wt\% FeO}^T \times \text{density (kg/m}^3)$$

$$k \text{ (G/Oe)} \approx 2.27 \times 10^{-6} \times \text{wt\% Fe} \times \text{density (g/cm}^3) = 1.77 \times 10^{-6} \times \text{wt\% FeO}^T \times \text{density (g/cm}^3).$$

As examples, a paramagnetic granite with 2 wt% total iron as FeO and a density of 2.65 g/cm<sup>3</sup> = 2650 kg/m<sup>3</sup> has a susceptibility of 9 μG/Oe = 12 × 10<sup>-5</sup> SI, whereas a paramagnetic gabbro with 12 wt% total iron as FeO and a density of 3.0 g/cm<sup>3</sup> = 3000 kg/m<sup>3</sup> has a susceptibility of 64 μG/Oe = 80 × 10<sup>-5</sup> SI (Clark & Emerson 1991). The maximum anomaly that could be associated with a contact between rock units with these susceptibilities is ~20 nT. Susceptibility contrast between paramagnetic rock units with differing total iron contents is certainly easily detectable, if the background variation due to ferromagnetic rocks in the area is sufficiently subdued.

Figure 8 plots observed susceptibility ranges and common susceptibility ranges for various rock types. The implications of these data will be discussed below in the section on magnetic petrology.

**Remanent magnetisation of minerals and rocks**

The susceptibility of most rocks primarily reflects their magnetite content. Remanence intensity, while also correlated

with modal magnetite, is sensitive to other factors, particularly grain size and microstructure of magnetic minerals and geological history. The natural remanent magnetisation of rocks is often multicomponent in character, i.e. it represents the sum of remanence components, each carried by a different subpopulation of magnetic grains, acquired at different times and, therefore, generally with differing directions. Variation in the relative proportions of remanence components throughout a rock unit produces scatter in the remanence direction, as well as variation in intensity. Specific remanent intensity is highest for acicular submicron magnetite grains in the single domain size range. Pseudosingle domain (titano)magnetite grains, up to ~20 μm diameter, also carry relatively intense remanence and are the dominant remanence carriers in many rocks. Larger magnetite grains have relatively weaker remanence, corresponding to Q values less than unity.

Remanence carried by haematite and monoclinic pyrrhotite is characterised by high Q values, but haematite is only weakly magnetic and, therefore, haematite-rich rocks are rarely responsible for substantial anomalies, unless the haematite is associated with magnetite and/or maghaemite. In cases where minor amounts of magnetite or maghaemite are intimately intergrown with haematite a relatively intense remanence, characterised by exceptionally high Q, may be present.

Pyrrhotite-bearing rocks often carry a relatively intense remanence, which may be ancient and quite oblique to the present field. Greigite also seems to be associated with high Q values (Reynolds et al. 1990b). The remanence carried by magnetically soft multidomain magnetite, which is the dominant magnetic phase in many rocks, is dominated by viscous magnetisation. This remanence is subparallel to the present field and, therefore, augments the induced magnetisation, enhancing the effective susceptibility. Thus, most anomalies can be interpreted in terms of magnetisation by induction, even when typical Koenigsberger ratios are comparable to unity. However, for a given source geometry, the anomaly amplitudes may be larger than measured susceptibilities indicate, owing to the viscous remanent magnetisation. Neglect of remanence may, therefore, mislead quantitative interpretation, even though the anomaly form is consistent with magnetisation parallel to the present field.

Koenigsberger ratios for viscous remanence carried by multidomain magnetite have an upper limit of about unity, but are typically much lower, averaging ~0.2. Rocks containing predominantly somewhat harder multidomain magnetite grains may carry a stable ancient remanence, characterised by a larger Q value. The Koenigsberger ratio of thermoremanence carried by an igneous rock containing predominantly such multidomain grains is typically ~0.5. In this case, the remanence direction records the geomagnetic field direction at the time of initial cooling. This direction can be of either normal or reversed polarity and may be highly oblique to the present field, depending on the age of the rock. Figure 3c summarises Koenigsberger ratios for various types of magnetisation carried by the major magnetic minerals, as a function of domain state. Figure 9 shows ranges of Koenigsberger ratio found for a variety of rock types.

Estimation of the bulk remanent magnetisation of a rock unit is not straightforward. The scatter of direction must be taken into account, as well as the distribution of intensity. Remanence makes a greater contribution to the anomaly associated with a unit showing consistent remanence direction and moderate Koenigsberger ratios than a unit with highly scattered remanence directions on a mesoscopic scale, even though the samples may all have high Q values. Measurements of raw NRMs can also be quite misleading. Surface samples are often affected by lightning, which imparts unrepresentatively high remanent intensities and Q values. Drill-core samples may carry spurious remanence imparted by drilling. Estimation

of representative remanence vectors requires palaeomagnetic cleaning of samples to remove spurious components and to identify the components that correspond to bulk in-situ

properties. The remanent magnetisation identified in this way should then be analysed statistically as *vectors*, rather than as directions and scalar intensities, separately.

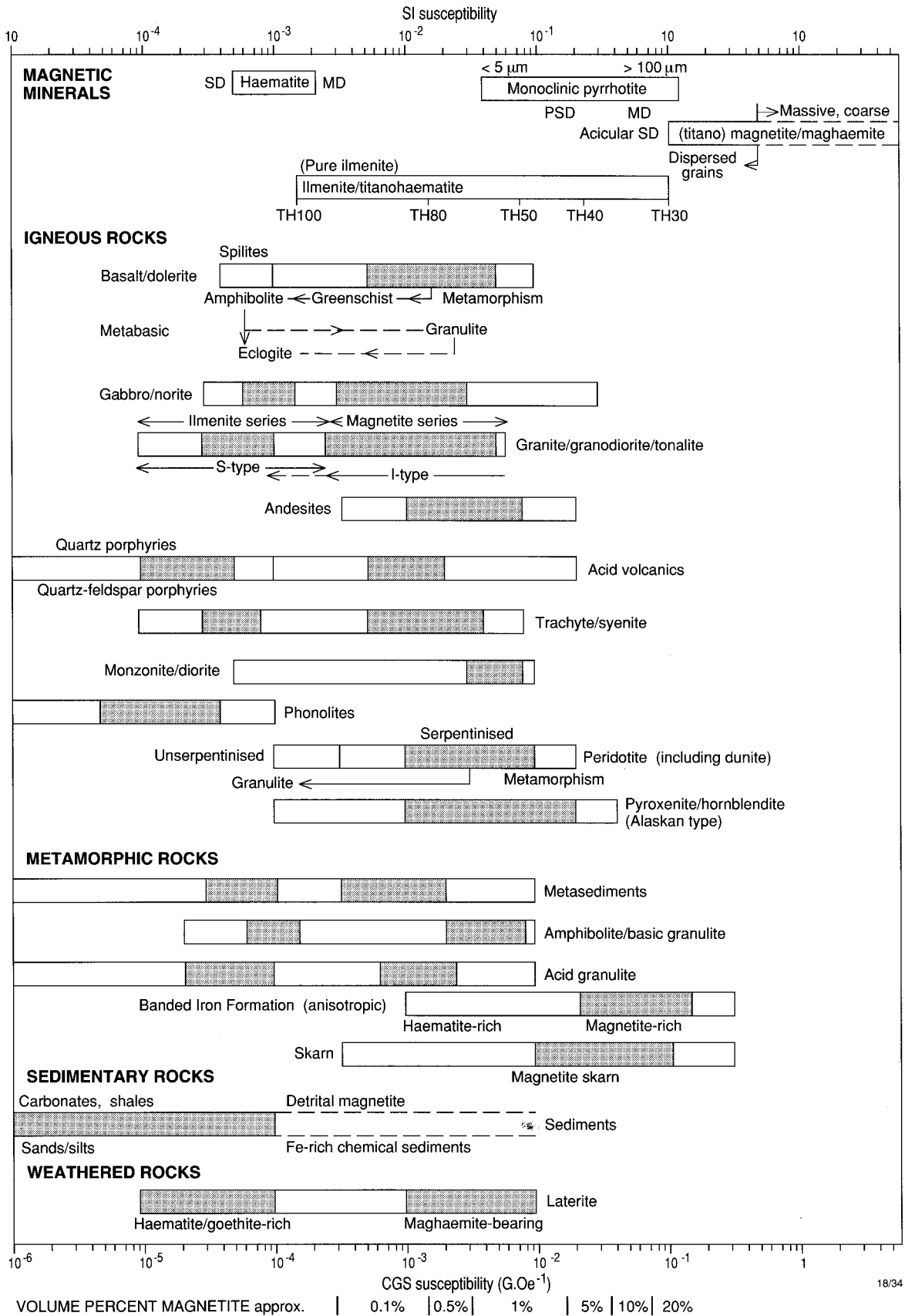


Figure 8. Observed and common susceptibility ranges for various rock types.

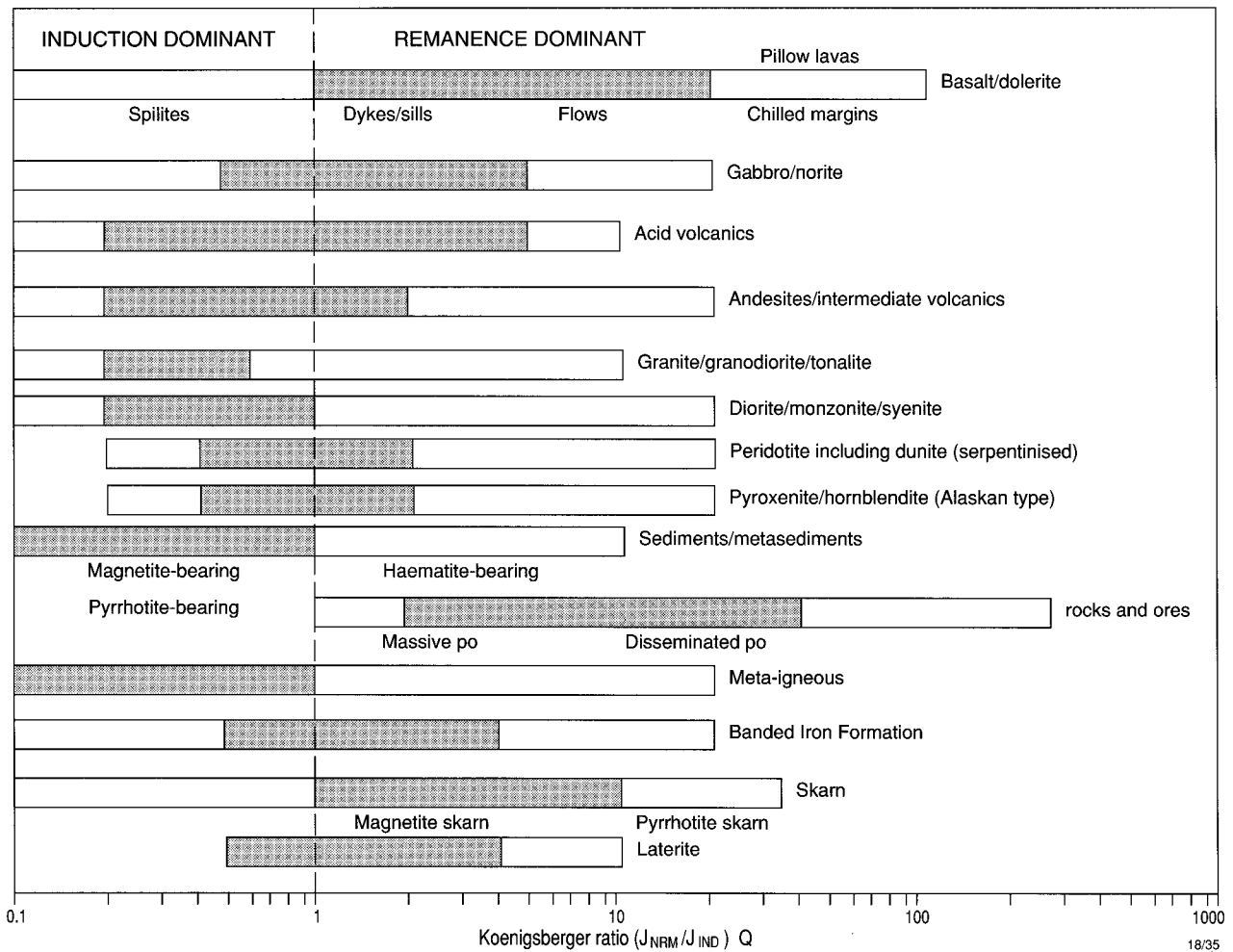


Figure 9. Observed and common ranges for Koenigsberger ratios of various rock types.

**Magnetic anisotropy of minerals and rocks**

Anisotropic susceptibility cannot be described by a scalar quantity, but takes the form of a symmetric second-order tensor. For any crystal and, similarly, for any homogeneous but anisotropic rock, three mutually orthogonal axes can be defined, along which the susceptibility has maximum, intermediate and minimum values. These principal axes (the *major*, *intermediate*, and *minor axes*, respectively) define an ellipsoid, the *susceptibility ellipsoid* (Nye 1957), which describes the behaviour of the susceptibility and induced magnetisation uniquely and also characterises the *magnetic fabric* of the rock. The *magnetic foliation* is the plane containing the major and intermediate susceptibility axes. It reflects planar structures in the rock. The *magnetic lineation* coincides with the major susceptibility axis and reflects linear structures. Figure 10 illustrates the concept of the susceptibility ellipsoid.

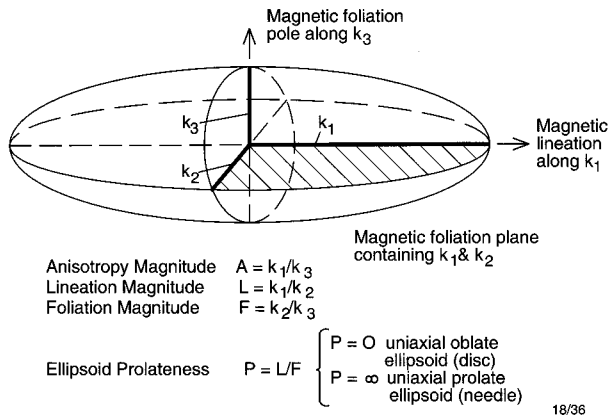
The susceptibility of paramagnetic and diamagnetic minerals can differ along different crystallographic directions, depending on crystal symmetry (Nye 1957). Cubic minerals have isotropic susceptibility, whereas that of minerals with lower crystallographic symmetry is anisotropic. The anisotropy can be characterised by the parameter A, which is the ratio of the major susceptibility to the minor susceptibility. Common paramagnetic minerals for which anisotropy has been measured have A ranging from ~1.1 to ~1.8. In highly deformed rocks, for which paramagnetic phases, such as mica, dominate the susceptibility, the high degree of preferred orientation of mica can produce substantial anisotropy of the rock. The orientation of major susceptibility axes reflects, and can be used to infer,

the planar and linear petrofabric elements in such rocks.

Ferromagnetic minerals of low crystal symmetry, such as haematite and monoclinic pyrrhotite show very strong susceptibility anisotropy. For both these minerals, susceptibility in the basal plane is orders of magnitude greater than along the c-axis. Therefore, even a slight preferred crystallographic orientation of haematite or monoclinic pyrrhotite can produce substantial anisotropy for the rock.

The susceptibility anisotropy of (titano)magnetite-bearing rocks arises from self-demagnetisation. In the case of dispersed grains, the effective susceptibility is greatest along the long axis of the grain and least along the short axis, because the grain demagnetising factor is least along the long axis and greatest along the short axis. Preferred dimensional orientation of magnetite grains, therefore, leads to anisotropic susceptibility of the rock, with maximum susceptibility parallel to the preferred alignment of long axes. When magnetite grains are inhomogeneously distributed, self-demagnetisation of the magnetite aggregates can produce strong anisotropy. The classic case is banded iron formations with magnetite-rich bands alternating with weakly magnetic bands. The susceptibility normal to the banding is greatly reduced by self-demagnetisation, whereas the susceptibility parallel to the banding is unaffected by self-demagnetisation.

Although almost all rocks exhibit slightly anisotropic magnetic susceptibility, which can be interpreted in terms of petrofabric, the degree of anisotropy is generally insufficient to significantly influence the form of magnetic anomalies. Exceptions include banded iron formations and some rocks



**Figure 10. Relationship between the susceptibility ellipsoid and magnetic fabric elements.**

and ores that contain pyrrhotite with a strong preferred orientation. In rocks with substantial magnetic anisotropy, susceptibility varies significantly with the field direction and, in general, magnetisation is not parallel to the applied field, because it tends to be deflected away from an axis of relatively low susceptibility towards an axis of higher susceptibility.

Acquisition of remanent magnetisation is also affected by anisotropy. The palaeomagnetic direction recorded by anisotropic rocks is deflected from the palaeofield direction. Thus palaeomagnetic results obtained from strongly anisotropic rocks are unreliable, unless they are corrected for anisotropy. Strong anisotropy can complicate magnetic interpretation in areas where remanence is important. If anisotropy effects are neglected, the inferred timing of remanence acquisition relative to folding can be misinterpreted (Clark & Schmidt 1994).

A recent book (Tarling & Hrouda 1993) is devoted the subject of magnetic anisotropy in rocks and readers are referred to this work for further information and a guide to the extensive literature on applications of magnetic anisotropy.

#### **Properties of strongly magnetic rocks and ores**

Magnetite-rich and monoclinic pyrrhotite-rich rocks and ores present particular difficulties for magnetic interpretation. Interpretation of structure and source geometry for these rocks is unreliable unless complicating factors are taken into account. These factors include:

- self-demagnetisation,
- anisotropy of susceptibility,
- Koenigsberger ratio,
- palaeofield direction during remanence acquisition,
- anisotropy of remanence acquisition,
- age of remanence with respect to deformation,
- perturbation of the local geomagnetic field direction by intense anomalies.

It should be noted that existing commercial modelling packages do not address these problems and are unsuitable for interpreting structures in these environments.

In Australia, types of mineralisation where these considerations are important include BIF-derived supergene-enriched iron ores, Tennant Creek-type discrete ironstone Au–Cu orebodies, BIF-hosted gold in Archaean greenstone belts, magnetite-rich Archaean VMS deposits, pyrrhotite-rich metasediments and sulphide ores in many areas, including the Pine Creek Inlier, the Cobar District and the Paterson Province, magnetite-rich ironstone-associated Cu–Au mineralisation, and BIF-associated Pb–Zn–Cu–Ag mineralisation of the east Mount Isa Inlier. The function of replaceable iron-rich lithology as chemical traps and the reducing power of magnetite are now recognised as important controls on mineralisation in the east Mount Isa Inlier, the Cullen Mineral Field of the Pine Creek

Inlier, and other areas.

Rock magnetic research has concentrated on the properties of dilute dispersions of magnetic grains in rocks. The theoretical and experimental basis for understanding the magnetic properties of magnetite and pyrrhotite-rich rocks is, therefore, inadequate. In such rocks, the properties are strongly influenced by grain interactions, greatly modifying the magnetic behaviour.

The magnetic petrophysics and petrology of iron formations and ironstones is also a fertile area for research. Little is known about the magnetic properties of different facies of these rocks and the effects on them of alteration. The magnetic expression of transition from magnetite-facies to haematite-facies or sulphide-facies iron formations in mineralised environments is poorly known.

Clark & Schmidt (1994) have reviewed the magnetic properties of BIFs. Magnetite grains in BIFs are in the large multidomain size range. In most rocks, for which magnetite is an accessory mineral occurring as widely dispersed grains, large multidomain grains are associated with Koenigsberger ratios (*Q* values) less than unity, owing to inhomogeneous magnetisation and strong self-demagnetisation of such grains. When magnetite occurs in very high concentrations, however, interactions between grains reduce the effective self-demagnetising factors of the grains and enhance the specific susceptibility and, to an even greater extent, the specific remanence. Thus, the Koenigsberger ratio exhibited by BIFs can be substantial. *Q* values vary widely for BIFs, but predominantly fall in the range 1–2, with higher values occurring occasionally. Because remanence contributes substantially to total magnetisation, it should not be ignored in quantitative interpretation of magnetic signatures of BIFs.

The effective susceptibility of BIFs parallel to bedding exceeds the susceptibility normal to bedding, typically by a factor of 2–4. Bedding-parallel susceptibilities of magnetite-rich BIFs are typically 0.5–2.0 SI (0.05–0.16 G/Oe). Similarly, remanence is acquired more readily along layering than across it, leading to deflection of the remanence direction away from the palaeofield direction, towards the bedding plane. The anisotropy of remanence acquisition generally exceeds the susceptibility anisotropy and can produce large deflections of palaeomagnetic directions acquired in moderate to high (but not polar) palaeolatitudes. For this reason, remanence direction in BIFs usually lies fairly close to the bedding plane.

The strong anisotropy of BIFs means that the *Q* value of a BIF unit is sensitive to the orientation of the bedding with respect to the geomagnetic field. Remanence direction and the relative age of remanence and folding are best determined by palaeomagnetic study of fresh BIF samples (from mines or drill core), backed up by detailed modelling of magnetic anomalies over BIFs of known structure.

Magnetic properties of outcropping BIFs are usually greatly modified by weathering, which substantially decreases the bulk susceptibility, the degree of anisotropy and the remanence intensity. Deeper and more intense weathering of BIFs is assisted by faulting and may be associated with reduced magnetic response over intensely faulted zones. The magnetisation of haematite-rich supergene-enriched iron ores is much lower than that of their BIF precursors.

Much magnetic modelling ignores the effects of anisotropy and remanence and proceeds on the assumption of magnetisation parallel to the geomagnetic field. Dips of magnetic units are then interpreted from the shape of associated magnetic anomalies. This approach is inappropriate for interpretation of anomalies associated with BIFs, because anisotropy and remanence effects are demonstrably important and the direction of the resultant magnetisation may depart substantially from geomagnetic field direction. As an illustration, Figure 11 shows theoretical magnetic profiles over idealised fold structures, indicating the differences in anomalies produced by isotropic

induced magnetisation, anisotropic induced magnetisation (for  $A = 2.5$ ), and anisotropic induced magnetisation plus pre-folding remanence (with intensity 1.5 times that of the isotropic induced magnetisation). The anisotropy and Koenigsberger ratio used are typical values for BIFs. Clearly, an interpretation of the magnetic signatures that ignores anisotropy and remanence will lead to serious errors, particularly in interpreted dips.

## Magnetic petrology

Magnetic petrology integrates rock magnetism and conventional petrology to characterise the composition, abundance, microstructure and paragenesis of magnetic minerals, and to define the processes that create, alter and destroy magnetic minerals in rocks. By relating magnetic mineralogy, bulk magnetic properties and petrology to observed magnetic anomalies, an understanding of the geological factors that control magnetic signatures can be obtained, which can be used to improve geological interpretation of magnetic surveys.

Magnetic properties of rocks reflect, above all else, the partitioning of iron in the rock between strongly magnetic oxides and/or sulphides and weakly magnetic phases (silicates, carbonates etc.). This partitioning depends on chemical composition, oxidation ratio of the iron and petrogenetic conditions. Thus, a host of geological factors influence magnetic properties, and simplistic correlations between magnetic properties and lithotype are generally unreliable. Rules-of-thumb, such as 'susceptibilities go according to basic extrusive > basic intrusive > acid igneous > sedimentary' have so many exceptions that they are almost useless. It is also dangerous to extrapolate empirical correlations between mapped geology and magnetics in one area to another area, ignoring changes in depositional environment, metamorphic grade or structural setting.

A magnetic petrological classification that can provide a sound basis for geological interpretation of magnetics in a given province must rely on an understanding of the factors that control partitioning of iron. The most direct approach to this problem is to characterise the ferrous and ferric iron content of whole rocks and of all the minerals in the rocks, as a function of the geological variable being studied. This requires wet chemical determination of ferrous and ferric iron, petrographic determination of modal mineralogy and microprobe determinations of the mineral chemistry of each phase, to produce a complete inventory of iron in the sample. As yet, such studies are in their infancy, owing to the rarity of wet chemical analyses to determine whole rock ferrous and ferric iron and to difficulties in determining ferric/ferrous ratios in individual minerals, particularly hydrous silicates.

The data of Figures 8 & 9 are based on magnetic property measurements at the CSIRO Division of Exploration and Mining over the last 18 years and published studies and compilations. The systematic collection of petrophysical data by the geological surveys of Scandinavian countries, in particular, has greatly expanded the size and scope of the information available. It is evident from Figure 8 that each rock type exhibits a wide range of susceptibility and that the values are not generally diagnostic of lithology. Classical rock names are in fact much too broad to be useful for classification of magnetic properties. This is because the susceptibility of most rocks reflects the abundance of accessory minerals, particularly magnetite (*sensu lato*), which are ignored in petrological classification.

At a more refined level, however, there is significant geological information in basic magnetic properties, especially if the statistical characteristics of large collections are considered and if the measurements are supplemented by rock magnetic experiments to characterise the composition and microstructure of the magnetic minerals. The magnetic minerals in a meta-igneous rock, for example, are sensitive to its

geological history, including the bulk composition and petrogenetic affinities of the magma, the degree of differentiation, conditions of emplacement, degree and type of hydrothermal alteration and conditions of metamorphism (temperature, pressure, fugacities of oxygen, water, sulphur, CO<sub>2</sub> etc.). Differences in magnetic properties can, therefore, reflect subtle variations in some or all of these influences. More detailed classification schemes, based on the most important of these factors, may, therefore, allow more meaningful interpretation of magnetic surveys in terms of geology. In some cases, observed differences in magnetic anomaly patterns within single mapped units have indicated hitherto unsuspected heterogeneity, which has then been confirmed by remapping.

## *Bimodal susceptibility distributions reflect ferromagnetic and paramagnetic populations*

The variation in magnetic properties for a given lithology is generally greater between geological provinces than within them, although large variations are also possible over smaller areas, even down to the outcrop scale. A notable feature of Figure 8 is that the susceptibilities of a number of rock types have distinctly bimodal distributions. More generally, bimodal susceptibility distribution represents distinct subpopulations within each rock type, for which ferromagnetic minerals are absent and present, respectively. Iron in the weakly magnetic subpopulation is incorporated into paramagnetic silicate minerals, predominantly as Fe<sup>2+</sup>, whereas similar rocks that are moderately to strongly magnetic contain significant Fe<sup>3+</sup>, which is incorporated into magnetite. Very highly oxidised rocks, however, tend to contain haematite rather than magnetite and are, therefore, also weakly magnetic. Puranen (1989) presents results from very large petrophysical sampling programs in Finland. His data show that all broad field names, such as 'granite', 'gabbro', 'mica schist', 'amphibolite' etc., exhibit distinctly bimodal susceptibility distributions.

Within each magnetic subpopulation, susceptibility tends to increase with basicity. The greater abundance of paramagnetic mafic minerals in rocks with lower SiO<sub>2</sub> increases the paramagnetic contribution to the susceptibility. This produces a small, but systematic difference, in the susceptibility of paramagnetic acid and basic rocks. The increasing sensitivity of modern magnetometers and the trend to more detailed magnetic surveys suggests that magnetic mapping may become useful even in very weakly magnetic terrains, where the rocks would hitherto have been classified as 'non-magnetic' on the basis of their apparent flat and featureless magnetic patterns in low-resolution surveys. More commonly, however, the observed magnetic signatures reflect variation in the abundance of ferromagnetic minerals.

When varietal mineralogy is incorporated into the rock classification, the bimodal susceptibility distribution tends to resolve into a paramagnetic subpopulation associated with particular varietal minerals and a ferromagnetic subpopulation associated with a distinct mineralogy. Bimodality often also reflects the fact that important geological factors, such as geochemical affinity, alteration and metamorphic grade, are not considered in the simple classification schemes used for most petrophysical summaries. A truly meaningful magnetic petrological classification scheme must include chemical and/or mineralogical data for protoliths, plus information on metamorphic grade and/or alteration.

## *Magnetic properties of igneous rocks*

In the case of granitoids, bimodal susceptibility reflects the existence of two distinct categories, the magnetite-series and ilmenite-series granitoids of Ishihara (1977). Magnetite-series granitoids are relatively oxidised and correspond broadly to the I-type granitoids of Chappell & White (1974), whereas ilmenite-series granitoids are more reduced and are usually S-type. The new classifications, which have important petro-



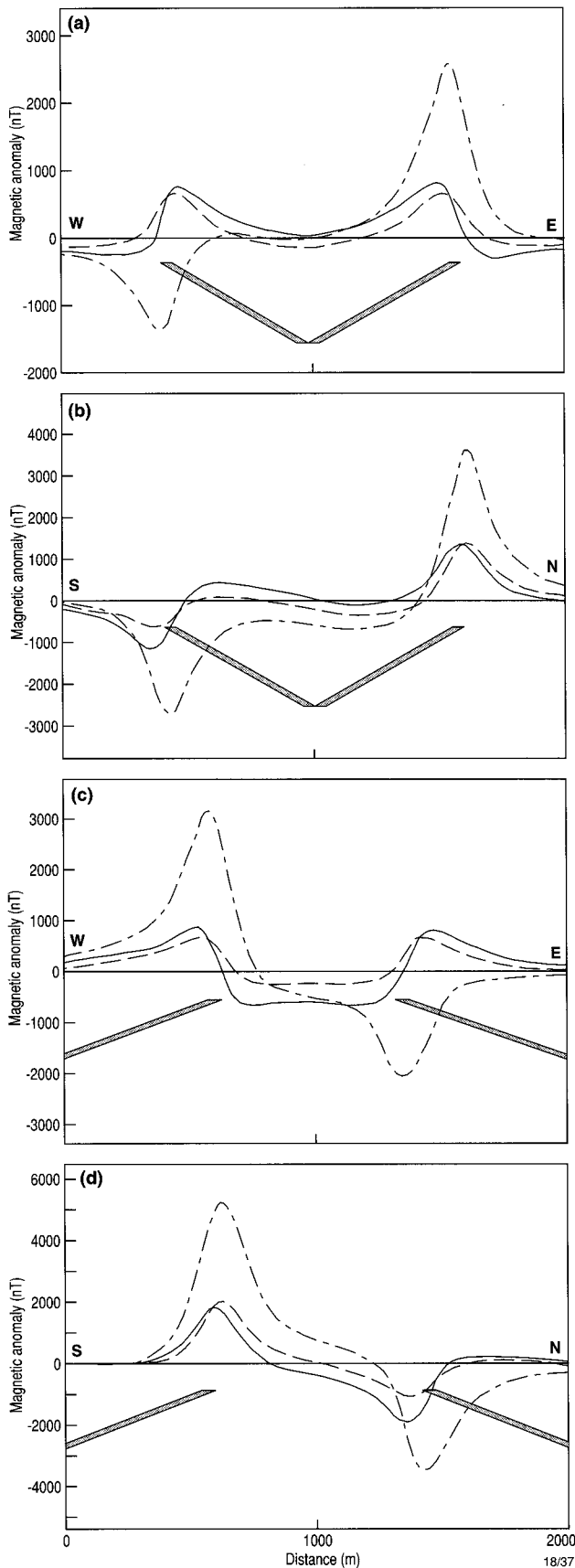


Figure 11. Magnetic anomalies of folded BIF units, showing the effects of anisotropy and pre-folding remanence on the total field anomalies. (a) E-W profiles over a N-S striking syncline, (b) N-S profiles over an E-W striking syncline, (c) E-W profiles over a N-S striking eroded anticline, (d) N-S profiles over an E-W striking eroded anticline. The regional geomagnetic field intensity is 54 000 nT, the field declination is  $0^\circ$  and the field inclination is  $-56^\circ$ .

genetic and metallogenic implications, have led to the concept of mapping granitoid terrains using a hand-held susceptibility meter or a magnetometer, and provide a good example of the utility of categories based on magnetics.

Although broad rock names such as 'granodiorite' show only a weak correlation with magnetic properties, more detailed classifications that include varietal minerals are much more predictive. For example, hornblende-biotite granodiorites are predominantly ferromagnetic, with moderate susceptibility, whereas muscovite-biotite granodiorites are usually paramagnetic. Accessory minerals that occur commonly in ferromagnetic felsic granitoids include sphene, epidote, allanite, pyrite and haemoilmenite or Mn-rich ilmenite. The biotite is  $\text{Fe}^{3+}$  and Mg-rich and is generally brown, black or olive green.

Accessory minerals often associated with paramagnetic granitoids include cordierite, garnet or aluminosilicate (which are characteristic of S-type granites), reduced (low  $\text{Fe}^{3+}$ ) ilmenite, pyrrhotite, and biotite that is  $\text{Fe}^{2+}$  and Al-rich, often with a 'foxy red' colour. Hornblende-bearing granitoid that is too reduced to contain magnetite is less common overall and tends to occur within specific provinces or 'basement terranes', for which the source material at depth is relatively reduced (Chappell et al. 1988). Susceptibility of felsic granitoid also correlates with feldspar colour (Blevin 1994). For granites with white plagioclase, relatively oxidised ferromagnetic granite generally has salmon pink K-feldspar, whereas reduced paramagnetic granite has white K-feldspar. There is a systematic increase in susceptibility with increasing pinkness of the K-feldspar. On the other hand, brick-red K-feldspar usually indicates hydrothermal alteration and is associated with low susceptibility. This simple relationship is disturbed if green plagioclase is present. The green colour reflects replacement of plagioclase by chlorite, sericite, albite, epidote and calcite. This alteration also tends to destroy magnetite. Granites with green plagioclase and pink K-feldspars exhibit variable susceptibility, reflecting destruction of some or all of the magnetite originally present in the granite.

There is a clear association between granite metallogeny and oxidation state, which translates into a correlation between magnetic susceptibility of granitoids and their associated mineralisation (Ishihara 1981; Blevin & Chappell 1992). In order of reducing susceptibility, Cu and Au are associated with magnetite-series, intermediate I-type granitoid suites; Mo, with more fractionated and oxidised magnetite-series I-type suites; and Sn, with paramagnetic, reduced, fractionated I- or S-type suites.

The overall tendency for magnetite content of ferromagnetic rocks to increase with basicity is somewhat obscured when igneous rocks from different provinces are compared. For consanguineous rocks, in particular, there is a general correlation between susceptibility and basicity. Andesites generally have similar, or slightly lower, susceptibility than related basalts. Rhyolites have a distinctly bimodal susceptibility distribution. Ferromagnetic rhyolites tend to be somewhat less magnetic than the more basic members of the series, and many rhyolites are paramagnetic (e.g. fayalite rhyolites, strongly peraluminous and strongly peralkaline rhyolites). Trachyandesites and trachytes generally have moderate to high susceptibility, comparable to or somewhat less than that of related alkali basalts, but corresponding phonolites are usually weakly magnetic. Within the ferromagnetic subpopulation of each

Solid curves represent pure induced magnetisation with isotropic susceptibility ( $k = 0.754 \text{ SI} = 0.06 \text{ G/Oe}$ ), dashed curves represent induced magnetisation with anisotropy (susceptibility parallel to bedding  $0.942 \text{ SI} = 0.075 \text{ G/Oe}$ ,  $A = 2.5$ ), and dash-dot curves incorporate anisotropy and pre-folding remanent magnetisation (declination normal to strike, original inclination  $-10^\circ$ , intensity  $48.6 \text{ A/m} = 0.0486 \text{ G}$ ). Because of the anisotropy, the Koenigsberger ratio depends on the field direction and bedding attitude, but it is about 1.5. After Clark & Schmidt (1994).

lithology, magnetic properties can also be related to geochemistry. For tholeiitic rocks in both oceanic and continental settings, iron and titanium-rich variants that are interpreted as related to hotspots have been found to have substantially higher susceptibility, reflecting greater modal titanomagnetite, than similar rocks with lower Fe and Ti contents (e.g. Anderson et al. 1975; de Boer & Snider 1979).

Granitic rocks and metamorphic rocks with secondary magnetite usually contain relatively coarse-grained multidomain magnetite, accounting for the generally low  $Q$  values of these rocks. On the other hand, young, rapidly chilled basaltic rocks (e.g. pillow lavas) exhibit very high Koenigsberger ratios, owing to the fine grain size of the titanomagnetites. In basaltic rocks the  $Q$  value of the primary thermoremanence is essentially a function of cooling rate, being highest for subaqueous chilled margins and small pillows and decreasing with distance from the margin. However, even thick doleritic sills and dykes are characterised by relatively high  $Q$  values, typically 1–10, provided the primary remanence has not been substantially modified by thermal or chemical overprints.

Plutonic rocks generally have low Koenigsberger ratios, owing to their coarse grain size, with the notable exception of some gabbroic to dioritic intrusives, for which the remanence is dominated by SD and PSD grains of magnetite (*sensu lato*). These efficient remanence carriers may be either fine (titanomagnetite inclusions in silicate grains or large, originally homogeneous, titaniferous magnetite grains that have undergone deuteric oxidation, which produces abundant oxyexsolution lamellae of ilmenite within relatively low-Ti magnetite. This subdivision of the magnetite grains by non-magnetic lamellae produces an effective grain size in the PSD range.

Ultramafic rocks (pyroxenites, hornblendites, serpentinised dunites etc.) in zoned Alaskan-type complexes are generally highly magnetic. The associated mafic and intermediate rocks (gabbro, diorite, monzonite) in these intrusives are also moderately to highly magnetic. Primary magnetite is the main magnetic mineral in these intrusive complexes. On the other hand, the primary spinel phase in unaltered komatiitic lavas (including spinifex textured peridotite, olivine orthocumulate and adcumulate (dunite) zones) and alpine-type peridotites is paramagnetic chromite and they are, therefore, weakly magnetic. However, komatiites are usually serpentinised, as are alpine-type peridotites. Serpentinisation usually creates substantial quantities of magnetite, accounting for the high susceptibility of serpentinised ultramafic rocks. This magnetite is generally multidomain, well-crystallised, almost pure  $Fe_3O_4$ , which is magnetically soft and carries relatively weak remanence ( $Q < 1$ ). The NRM of serpentinities is often dominated by viscous remanence, subparallel to the present field. Large dunite bodies sometimes have a weakly magnetic, unserpentinised core within an envelope of highly magnetic serpentinite.

### ***Magnetic properties of sedimentary rocks***

'Clean' carbonates and clastic sediments have very low susceptibility. Some immature sandstones are magnetic because they contain significant quantities of detrital magnetite. Sediments deposited in the presence of metal-bearing solutions, associated with volcanic activity for example, may contain appreciable magnetite or, possibly, pyrrhotite and, therefore, be magnetic. Such sediments may be transitional to syngenetic massive mineralisation or to banded iron formation. Magnetite-rich banded iron formations are not only strongly magnetic, but characterised by high anisotropy of susceptibility. The susceptibility values for banded iron formation shown in Figure 8 are bulk susceptibilities, i.e. the average susceptibility along any three orthogonal directions.

Traditionally, aeromagnetic surveys over sedimentary basins were used primarily to determine depth to basement. In recent

times it has become evident that modern high-resolution aeromagnetic surveys over sedimentary basins also provide considerable information about sedimentary lithology and structure. Little is known about the intrasedimentary sources responsible for the observed anomaly patterns. There is an urgent need for research into the magnetic petrophysics and petrology of intrasedimentary sources, directed at improving understanding of the geological significance of magnetic anomaly patterns. Gay (1992) has given examples of intrasedimentary anomalies associated with distributions of detrital magnetite. Reynolds et al. (1990a,b; 1994) have demonstrated that diagenetic monoclinic pyrrhotite and greigite are possible sources of low-amplitude, high-frequency anomalies, which in some cases may be associated with hydrocarbon seepage, over some sedimentary basins. There is, as yet, no definitive evidence of diagenetic magnetite produced by hydrocarbon seepage in quantities sufficient to produce observable anomalies (Reynolds et al. 1990b), but Machel & Burton (1991) have described chemical and microbial processes that could, in principle, produce authigenic magnetite and pyrrhotite, at the expense mainly of haematite, in the presence of hydrocarbons.

### ***Effects of hydrothermal alteration***

Studemeister (1983) pointed out that the redox state of iron in rocks is a useful indicator of hydrothermal alteration. Large volumes of fluid or high concentrations of exotic reactants, such as hydrogen or oxygen, are required to shift  $Fe^{3+}/Fe^{2+}$ . When reactions associated with large water/rock ratios occur, the change in redox state of the rocks produces large changes in magnetic properties, owing to the creation or destruction of ferromagnetic minerals.

Zeolite-grade hydrothermal alteration of mafic to silicic igneous rocks tends to decrease the susceptibility of these rocks according to the degree of development of zeolites (Emerson & Welsh 1987). Zeolite-rich rocks are generally paramagnetic, even when derived from strongly magnetic protoliths. Regional hydrothermal alteration of volcanic piles produces progressive demagnetisation from zeolite grade to greenschist grade. For example, primary titanomagnetite and ilmeno-magnetite in basalt progressively develops ferrirutile granules at  $\sim 150^\circ C$ , with sphene replacing ilmenite lamellae by  $\sim 250^\circ C$ , and polycrystalline titanohaematite replacing titanomagnetite above  $300^\circ C$  (Ade-Hall et al. 1971). However, contact metamorphism of hydrothermally altered, demagnetised igneous rocks, for instance by dyke injection, may produce secondary magnetite (Hall & Fisher 1987).

The available evidence suggests that both major types of alteration associated with epithermal systems, acid-sulphate and adularia-sericite alteration, tend to demagnetise volcanic rocks through replacement of magnetite by paramagnetic phases. Similarly, phyllic alteration and intense propylitic alteration associated with porphyry intrusions tend to destroy magnetite in the intrusion and in surrounding rocks. On the other hand, the potassic alteration zone associated with oxidised, magnetic felsic intrusions is often magnetite-rich. This is commonly observed for Au-rich porphyry copper systems (Sillitoe 1979). Early potassic (biotite-rich) alteration around the gold-mineralised Mount Leyshon complex (Queensland), which comprises intrusive breccias, porphyry plugs and dykes, produced abundant magnetite in metasedimentary and doleritic host rocks and is largely responsible for the Mount Leyshon magnetic anomaly (Sexton et al. 1995).

Serpentinisation of olivine-rich ultramafic rocks tends to produce abundant magnetite. At low grades, initial serpentinisation of olivine produces Fe-lizardite plus brucite. With further serpentinisation, the maximum iron content of lizardite is exceeded and magnetite is produced, along with lizardite, chrysotile and brucite. At higher grades, antigorite and magnetite are produced. For serpentinities, there is generally

an inverse relationship between density, which decreases with progressive serpentinisation, and susceptibility. Saad (1969) found that weakly serpentinised peridotites (~10%) are weakly magnetic ( $k \sim 100 \mu\text{G}/\text{Oe} = 126 \times 10^{-5} \text{ SI}$ ), partially (~75%) serpentinised peridotites are moderately ferromagnetic ( $k \sim 500 \mu\text{G}/\text{Oe} = 6300 \times 10^{-6} \text{ SI}$ ) and fully serpentinised peridotites are substantially more magnetic ( $k \sim 5000 \mu\text{G}/\text{Oe} = 63\,000 \times 10^{-6} \text{ SI}$ ).

Carbonate alteration of serpentinised ultramafics initially redistributes magnetite, without destroying it, and has little effect on susceptibility. Intense talc-carbonate alteration, however, consumes the magnetite, with iron entering magnesite as a siderite component, and demagnetises the rock.

### Effects of metamorphism

Metamorphism has profound effects on magnetic properties. It follows that magnetic interpretation should take changes in metamorphic grade across the study area into account. Effects of metamorphism on magnetic properties have been discussed by McIntyre (1980), Grant (1985), Shive et al. (1988), Wasilewski & Warner (1988), Urquhart (1989), Frost (1991a,b), Olesen et al. (1991), Skilbrei et al. (1991) and Clark et al. (1992).

Fresh basalt and dolerite have moderate to high susceptibility. Burial metamorphism of igneous rocks to zeolite or prehnite-pumpellyite grade does not demagnetise them, in the absence of circulating hydrothermal fluids, but regional metamorphism to greenschist and lower to mid-amphibolite grades tends to demagnetise basic igneous rocks. In gabbro, the fine magnetic grains within silicates may be protected by their silicate hosts from metamorphic breakdown, so that gabbro may be somewhat less sensitive to low and medium-grade metamorphism than its extrusive and hypabyssal equivalents.

Where magnetite is an abundant cumulus phase, it appears to be much less prone to metamorphic breakdown than if it is present as a minor intercumulus mineral. Thus highly differentiated ferrogabbro and ferrodiorite, which contain abundant primary magnetite, remain strongly magnetic throughout medium to high-grade metamorphism. This probably reflects the fact that secondary metamorphic minerals become rapidly saturated in  $\text{Fe}^{3+}$  when magnetite starts to break down, ensuring that the rest of the magnetite remains stable during metamorphism. Felsic plutons seem to be more resistant to metamorphic destruction of magnetite in the greenschist and amphibolite facies than mafic rocks and felsic volcanic rocks.

Amphibolite-grade metamorphism overall produces heterogeneous magnetic properties with bimodal susceptibility, although dominated by weakly magnetic rocks. At this grade, magnetic mafic rocks are more common than magnetic silicic rocks. Chlorite and/or biotite-bearing amphibolite tends to be weakly magnetic, whereas hornblende-rich amphibolite may have much higher susceptibility.

Granulite-facies metamorphism of mafic protoliths frequently produces secondary magnetite and large increases in susceptibility. High-pressure granulite and eclogite are generally paramagnetic. Magnetite breakdown in these high pressure rocks generally occurs at 10–20 kb. The reactions occur at generally lower pressure for undersaturated basaltic rocks than for quartz tholeiite, for lower Mg/Fe ratios and for more reduced rocks.

These changes in properties reflect redistribution of ferric iron in metabasites during metamorphism. Ferric iron originally present in magnetite goes largely into haematite, epidote and chlorite during greenschist-grade metamorphism, then into biotite and amphibole in the amphibolite facies, into metamorphic magnetite in the granulite facies and, finally, into paramagnetic silicates such as garnet and clinopyroxene in the eclogite facies. The magnetite-in isograd reflects P-T-t conditions and appears to occur within lower granulite facies,

above the orthopyroxene-in isograd for prograde metamorphism, followed by isobaric cooling (Olesen et al. 1991). For retrograde metamorphism or for isothermal decompression during rapid uplift the magnetite-in isograd lies within the upper amphibolite facies, above the biotite isograd (Skilbrei et al. 1991).

Decompression of high-pressure granulite during rapid uplift can produce fine-grained magnetite by breakdown of garnet and clinopyroxene. This magnetite may carry relatively strong and stable remanence. In general, however, metamorphic magnetite is relatively coarse-grained, well crystallised and fairly pure, i.e. it occurs as magnetically soft MD grains, which are associated with low Q values and unstable, predominantly viscous, remanence.

Prograde metamorphism of serpentinised ultramafics causes increasing substitution of Mg and Al into the magnetite, eventually shifting the composition into the paramagnetic field. Thus, metamorphism progressively demagnetises serpentinite, which becomes paramagnetic at granulite grade (Shive et al. 1988). Subsequent retrograde serpentinisation, if it occurs, can produce a magnetic rock again. Clark et al. (1992) noted a progressive decrease in magnetic anomaly amplitude associated with a major serpentinised ultramafic unit as metamorphic grade increases from prehnite-pumpellyite facies to upper amphibolite facies.

The iron content of sediments (generally higher for pelites than for psammites) and the  $\text{Fe}^{3+}/\text{Fe}^{2+}$  ratio, which reflects the redox conditions during deposition and diagenesis, have a major bearing on the capacity of sedimentary rocks to develop secondary magnetite during metamorphism (McIntyre 1980). Thus, magnetic patterns over metasedimentary rocks tend to reflect sedimentary facies variations, as well as metamorphic conditions. These patterns can be very useful for mapping, although the relationship between the magnetic marker units and conventional lithological units may be quite tenuous (McIntyre 1980).

Magnetite formation is favoured by high total iron and, therefore, tends to be associated with metapelites more frequently than metapsammites. Substantial chemical input by exhalative metal-bearing solutions, including iron, increases the potential for magnetite formation during subsequent metamorphism. Magnetite formation is favoured by intermediate redox state/oxidation ratio. Low oxidation states are associated with ilmenite in metasediments and very high oxidation states, with haematite. Very iron oxide-rich assemblages, for example BIFs, are self-buffered during metamorphism, preserving fine layering with magnetite- and haematite-rich bands.

Organic-rich sediments produce reduced graphitic metasediments, which are magnetite-free, but commonly contain pyrrhotite above greenschist grade. Red (haematite-bearing) sediments produce intermediate redox metasediments, which may be magnetite-bearing. Haematite goes to magnetite in the biotite or lower garnet zones for these rocks. The maximum ferromagnetic proportion in metasediments generally occurs at granulite grade.

Pyrrhotite is the main magnetic mineral in many metasedimentary rocks, particularly in mineralised areas. Clark & Tonkin (1994) discuss magnetic anomalies associated with monoclinic pyrrhotite in metasediments of the Cobar area, Australia. Remanent magnetisation and susceptibility anisotropy are often important in monoclinic pyrrhotite-bearing metasediments (Clark 1983b).

### Conclusions

The aim of magnetic interpretation is to elicit geological information from magnetic survey data. There is a continuing need for magnetic petrophysical studies to constrain interpretation of magnetic surveys and for fundamental research in

magnetic petrology to improve understanding of the geological factors that create, alter and destroy magnetic minerals in rocks. A large-scale, systematic effort is required to create physical property databases, including comprehensive magnetic property data, classified according to magnetic petrological principles and integrated with geological and geochemical databases. New developments in 3D modelling, including inversion, and visualisation will assist efficient interpretation of large magnetic data sets. However, meaningful geological interpretation using these tools should allow incorporation of petrophysical information and magnetic petrological concepts. This paper notes some of the important ideas upon which such a research effort should be based.

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